

**Imperial College
London**

Kinematics of Normal Faults

A thesis submitted to
Imperial College London

For the degree of
Doctor of Philosophy

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Declaration

I certify that the research presented within this thesis is my own, except where the work of others is cited. The contents of this thesis have not previously been submitted for a degree at Imperial College London or any other university.

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Acknowledgements

I would first like to thank my supervisors Chris Jackson, Becky Bell, and Atle Rotevatn for their support over the past few years. Thank you for sharing your knowledge with me.

This PhD would not have been possible without the constant support and inspiration from the Basins Research Group. In particular, I would like to thank Sian for being an amazing friend, colleague, and flatmate, and Mike and Tim for teaching me the wonders of Petrel.

I would like to thank everyone that I've GTAed for throughout my time at Imperial: Philippa, Alex, Valentin, Becky, Al, John, Mark, Alan, and Emma. Teaching has been an incredibly formative experience, and I am lucky to come out of a PhD as a more well-rounded geologist. Thanks for teaching me how to teach and letting me see real rocks a few weeks a year.

Many thanks to the Fun Huns (& extended huns): Ruth, Sinead, Jenny, Tom, Sian, Evie, Graham, Harriet, Alex, Cecily and Catherine, aka my British family, aka my wine drinking pals. You are all brilliant people and scientists.

I would like to thank my support network from Nebraska, Dr B and Claire. Dr B, you inspired me to pursue geology and went above and beyond with your support and I would not be here without you.

I am indebted to my parents, Joe and Pam, for their constant support from across the world. Thank you for the Sunday Facetimes and being my home base as I follow my dreams.

Last but not least, I would like to acknowledge that I would not be here without the massive amount of privilege that has allowed me to pursue my passions; It is not a luxury that I take lightly. There is a long way to go in making science more equitable, and I will spend the rest of my career working for progress.

Abstract

Continental extension is accommodated by normal faults, and fault growth is achieved by increases in length (L) and displacement (D). When normal fault length and displacement were initially studied in the 1970's and 80's, it was assumed that faults maintain a constant D/L ratio from initiation to cessation, growing via sympathetic increases in displacement and length (the propagating fault model). Since the advent of 3D geophysical seismic data, it has been suggested that faults reach their maximum length before accruing significant displacement (the constant-length model). Several uncertainties remain, such as (i) how normal fault displacement and length relate to each other in different settings and whether a universal scaling law can be used, (ii) how different geological factors affect fault geometry and growth. To address these issues, I first assemble and interrogate a global database of normal faults that includes displacement and length, as well as tectonic setting, host rock lithology, fault maturity, and D/L through time where available. I next present a case study that uses growth faults imaged in 3D seismic data from offshore NW Australia to track fault lengthening, throw, and changes in slip rate through time. Observations on fault growth from the global database and seismic-based interpretations were then tested using physical analogue models, designed to investigate fault displacement/throw and length through time in higher resolution than seismic data allows. Results demonstrate that 'one-size-fits-all' D/L scaling laws are not accurate because the fault geometry and growth is affected by fault size, host rock lithology, fault reactivation, and fault maturity in particular. Universal scaling laws are inherently problematic because the relationship between length and displacement/throw is ever-changing throughout a fault's life. Individual normal faults follow the following growth pattern: a lengthening stage (10-30% of a fault's active life), displacement stage (30-75% of a fault's active life), and possible tip retreat stage (final 25% of a faults' active life). On a fault array scale, faults follow a cyclical growth pattern, where faults alternate between the lengthening and displacement stages.

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Chapter 1

Introduction

1.1. Rationale

As continental extension occurs, strain is accommodated by normal faults, which grow by increasing in length (L) and displacement (D). Even though the relationship between normal fault length and displacement and how it changes through time has been a topic of discussion amongst the structural geology community for the past 30+ years, there are only 5 published natural examples that show incremental normal fault growth through time (Meyer et al., 2002; Walsh & Watterson, 2002; Tvedt et al., 2013; Jackson et al., 2017; Pan et al., 2021), and how normal faults grow is still widely unknown.

Understanding how faults grow, the relationship between fault D and L, and how it changes through time is important for many aspects of structural geology. Faults are conduits for fluids, so understanding their geometries and growth histories has implications in areas that require knowledge of fluid flow, such as assessing geothermal potential and finding secure locations for nuclear waste and CO₂ storage (Nicol et al., 2010; Torabi & Berg, 2011; Tanner & Brandes, 2020). Having specific D/L scaling equations is important when only one value is available, for example, in an isolated field exposure or when only 2D seismic data are available. There are also implications for earthquake geohazards; knowledge of fault growth is important in estimating the location, magnitude, and recurrence interval of potentially damaging earthquakes in areas of continental extension (Walsh et al., 2003; Soliva et al., 2008; Nicol et al., 2010). Knowledge of fault growth is also important for understanding the tectonostratigraphic development of sedimentary basins, as rifts control drainage patterns and facies distributions (Gawthorpe & Leeder, 2000; Ge et al., 2017; Henstra et al., 2017; Jackson et al., 2017; Rotevatn et al., 2019).

Historically, the relationship between normal fault displacement and length has been characterised by one-size-fits-all scaling laws, which were taken from power-law averages of D/L data from global databases. Previously compiled D/L databases (e.g., Walsh & Watterson, 1988; Cowie & Scholz, 1992a; Schlische et al., 1996; Bailey et al., 2005; Torabi & Berg, 2011) typically have a large amount of scatter, where a fault of a given length can vary in displacement by 4+ orders of magnitude. One source of scatter is likely due to the nature of D/L ratios; they are a constantly changing value throughout the life of a fault, and a static law does not capture this. Another source of scatter is that different types of faults (normal, reverse, and strike-slip) have been grouped together in global datasets, along with faults of all tectonic settings, ages, and host rock lithologies. One-size-fits-all scaling laws are likely not a

precise way to estimate fault length or displacement, thus it is necessary to investigate how factors such as host rock lithology, tectonic setting, fault maturity, and size affect faulting.

The aim of this research is to answer a broad structural geology question: how do normal faults grow and what geological factors affect fault growth? In this chapter, I will introduce my research aims and rationale. In the Chapter 2, I will offer an introduction to normal fault growth and take a brief dive into the literature to describe our current understanding. The subsequent three chapters contain three stand-alone scientific papers. Some of the work has already been peer-reviewed and published, and some of it is in preparation to be submitted in the near-future. The final chapter will review my work as a whole to address withstanding gaps in knowledge and look to future work in the field.

1.2. Research aims

The overall aim of this research is to understand how normal faults grow. This thesis aims to answer the following questions:

1. What is the relationship between normal fault length and displacement?
2. How do different external geological factors affect the relationship between displacement and length?
3. How do normal faults grow through time?

In order to answer these questions, I utilised a multi-faceted approach combining i) the creation and analysis of a large global database of normal fault length and displacement, ii) seismic reflection data from offshore NW Australia, and iii) physical analogue models.

1.3. Specific aims and objectives

1.3.1. What is the relationship between normal fault length and displacement? (Chapter 3)

- Review existing global fault databases
- Identify potential sources of measurement error that could affect D/L relationships
- Compile a new global database showing normal fault displacement and length

1.3.2. How do different external geological factors affect the relationship between displacement and length? (Chapter 3)

- Make database including this information
- Complete a statistical analysis of how different lithologies, etc. affect D/L

- Compare findings to different case studies

1.3.3. How do normal faults grow through time? (Chapters 3, 4 & 5)

- Assess the validity of two competing normal fault growth models: the propagating fault growth model and the constant-length growth model
- Compile known examples of normal fault displacement and length through time, including natural examples as well as numerical and physical models
- Undertake a seismic reflection case study from offshore NW Australia to assess how syn-sedimentary normal faults grow through time and determine which fault growth model is appropriate
- Quantify how normal fault geometries change as they become inactive and identify methods of how to identify this in seismic
- Use physical analogue models run inside a medical grade CT scanner to view fault growth through time
- Describe early-stage fault growth

1.4. Data and methods

The overarching aim of my research is to determine how fault length and displacement relate to each other, what factors can affect it, and how faults grow through time. In order to understand this and how different geological factors affect D/L ratios, I compiled a global database of displacement and length data from 4046 normal faults in outcrop, seismic, and numerical and physical models from 69 different sources. I also collected additional information about the faults such as host rock lithology, tectonic setting, and fault maturity so that I could determine how those factors affected D/L ratios. Global datasets have been assembled before (Walsh & Watterson, 1988; Cowie & Scholz, 1992a; Schlische et al., 1996; Bailey et al., 2005; Torabi & Berg, 2011), however, this additional information was not collected previously and there were some issues with the data that was included in these compilations, which I will highlight in Chapter 3. I also compiled D and L through time data where it was available in order to see how faults grow through time.

As revealed in my compilation of normal fault displacement and length through time, there are very few known examples of natural normal faults with length and displacement are tracked through time, so I used a 3D seismic dataset from the Exmouth Plateau from the NW

Shelf of Australia as a case study. I chose this study area and dataset because it is an open access, high-quality and well-constrained 3D seismic dataset that images growth faults.

The Glencoe dataset is post-stack time-migrated and spans 3900 km² and contains four time-constrained wells. The Glencoe 3D seismic reflection survey was proposed by Hess Exploration Australia and conducted by CGG Veritas (CGGV) in 2007 with the aim of ascertaining the potential of subsurface oil and gas deposits in the study area. The data is available open access through the Australian government (<https://www.ga.gov.au/nopims>) which has made it a popular area in recent years to study fault growth (Pan et al., 2021), as well as other geological phenomena, such as igneous sills (Magee et al., 2017; Mark et al., 2019; Köpping et al., 2021; Norcliffe et al., 2021), fluid flow and polygonal faulting (Dirstein et al., 2013; Velayatham et al., 2019), and mass transport complexes (Hengesh et al., 2013; Nugraha et al., 2018).

The Exmouth Plateau is located in the North Carnarvon Basin, offshore NW Shelf of Australia. The North Carnarvon Basin was formed due to Late Carboniferous-Permian rifting of Pangea (Stagg & Colwell, 1994; Longley et al., 2002; Gibbons et al., 2012). The Exmouth Plateau formed as a result of rifting between Australia and Greater India from Late Triassic (Rhaetian)-Early Cretaceous (Stagg & Colwell, 1994; Longley et al., 2002; Gibbons et al., 2012; Bilal et al., 2018), during which time, an array of N-S trending faults formed, which offset thick pre-rift fluvial-deltaic to marginal marine succession (the Triassic Mungaroo Formation) (Longley et al., 2002; Stagg et al., 2004). Syn-rift extension of these faults lasted from the Rhetian-Early Cretaceous; I focus on these syn-sedimentary normal faults in Chapter 4.

Even with high-quality 3D seismic data, resolution is not high enough to see the early stages of fault growth and at short time scales. Sandbox modelling can be used to understand concepts that are difficult to resolve in outcrop and in seismic. I ran a series of analogue models inside of a medical-grade CT scanner at the University of Bern in Switzerland. These models used the same apparatus used in Zwaan et al, (2016, 2019): a sandbox with a fixed base plate and computer-controlled motorised sidewalls. There is compressed foam at the bottom of the model, which allows extension to be spread across the model. Each model was extended a total of 4 cm, and the models were scanned every 2 mm of extension.

Deformation at the top of the model was tracked using time-lapse photography

1.5. Thesis structure

In the following chapter, I will present a literature review of normal fault growth. The subsequent three chapters contain three stand-alone scientific papers. Some of the work has already been peer-reviewed and published, and some of it is in preparation to be submitted in the near-future. The final chapter will review my work as a whole, address withstanding gaps in knowledge, and look to future work in the field. There will be some repetition of the literature review in the scientific chapters as they have been written as stand-alone papers.

The research chapters in this thesis (Chapters 3-5) use ‘we’/’our’ pronouns in order to acknowledge the work of the co-authors who are included in my published and/or soon to be published work. However, I confirm that I conceptualised the projects, carried out the data collection and analysis, wrote the manuscripts, and made the associated figures. Professor Chris A.-L. Jackson, Dr Rebecca Bell, and Professor Atle Rotevatn supervised my PhD research and edited my manuscripts. Additional acknowledgements are included at the beginning of each scientific chapter.

The following chapters of the thesis are organised as follows:

Chapter 2: Literature Review

In this chapter I will offer an introduction to normal fault growth and take a brief dive into the literature to describe where the field is now.

Chapter 3: Displacement/length scaling relationships for normal faults; a review, critique, and revised compilation

The first study takes a broad look at normal fault geometry and kinematics and how they are influenced by geological factors such as tectonic history, fault maturity, host rock lithology, and fault size. I first review current knowledge on the relationship between normal fault length and displacement and critique previous works. I then present a newly collected large global database of fault length and displacement data, along with the previously mentioned geological factors and use it to suggest new and improved fault scaling laws. This will soon be submitted to *Earth Science Reviews*.

Chapter 4: Normal fault kinematics and the role of lateral tip retreat: an example from offshore NW Australia

The second study provides a detailed example of normal fault growth through time and investigates how faults become inactive through a 3D seismic-based case study of the offshore NW Shelf of Australia. This paper has been published in *Tectonics*.

Chapter 5: Modelling Chapter

The final study addresses the resolution limitations of outcrop and 3D seismic studies of normal fault growth by running analogue models run in a CT scanner to understand early stages of normal fault growth in a way that cannot be resolved otherwise. I track fault length and throw through time and take a detailed look into relay ramp development through time. This will soon be submitted to *Solid Earth*.

Chapter 6: Discussion

In the discussion section, I interrogate the findings from my three scientific studies and critically evaluate normal fault growth, from initiation to death, and how geological factors affect it. I then detail how I will continue to develop this research after the completion of my PhD and discuss future areas of work for the field of fault kinematics as a whole.

Chapter 2

Literature Review

2.1. Continental extension and rift basins

Rift basins form as a result of extension of the continental lithosphere (Allen & Allen, 2005). They are formed by the 3D interaction of normal faults arrays and drainage catchments, and are affected by water level and changing climate (Gawthorpe & Leeder, 2000). Rift basins form in both continental and marine settings, with sedimentary facies evolving from continental rift, to lacustrine, to marine (Gawthorpe & Leeder, 2000).

The rifting process can be broken into three broad phases: the pre-rift, syn-rift, and post-rift. The pre-rift is a period of tectonic quiescence and a precursor to rifting. Before rifting starts, there are potential pre-existing weaknesses that can affect later faulting, such as pre-existing structures in the crystalline basement (Morley, 2002; Bird et al., 2015; Phillips et al., 2016), faults from previous rifts that could be reactivated (Walsh et al., 2002; Vétel et al., 2005, Baudon & Cartwright, 2008, Giba et al., 2012, Whipp et al., 2014), and even structures at the grain to micro-grain scale (Phillips et al., 2016). The syn-rift begins at the start of lithosphere extension, when rift basins form (Gawthorpe & Leeder, 2000). The post-rift marks the end of lithosphere extension, during which subsidence occurs across the basin (Allen & Allen, 2005).

2.2. Rift initiation

Gawthorpe & Leeder (2000) broke the rifting into three phases: a rift initiation phase, a fault interaction and linkage phase, and a through-going fault zone phase (Figure 2.1) Faults first nucleate as many small, isolated fault segments with small displacements (Cowie, 1998; Gawthorpe & Leeder, 2000). Initially, faults grow via tip-propagation. As syn-rift packages develop, wedge-shaped syn-sedimentary packages form along the hanging walls of active faults. The presence of pre-existing structures may cause fault rift initiation to happen quickly as new faults exploit existing weaknesses (Walsh et al., 2002; Vétel et al., 2005).

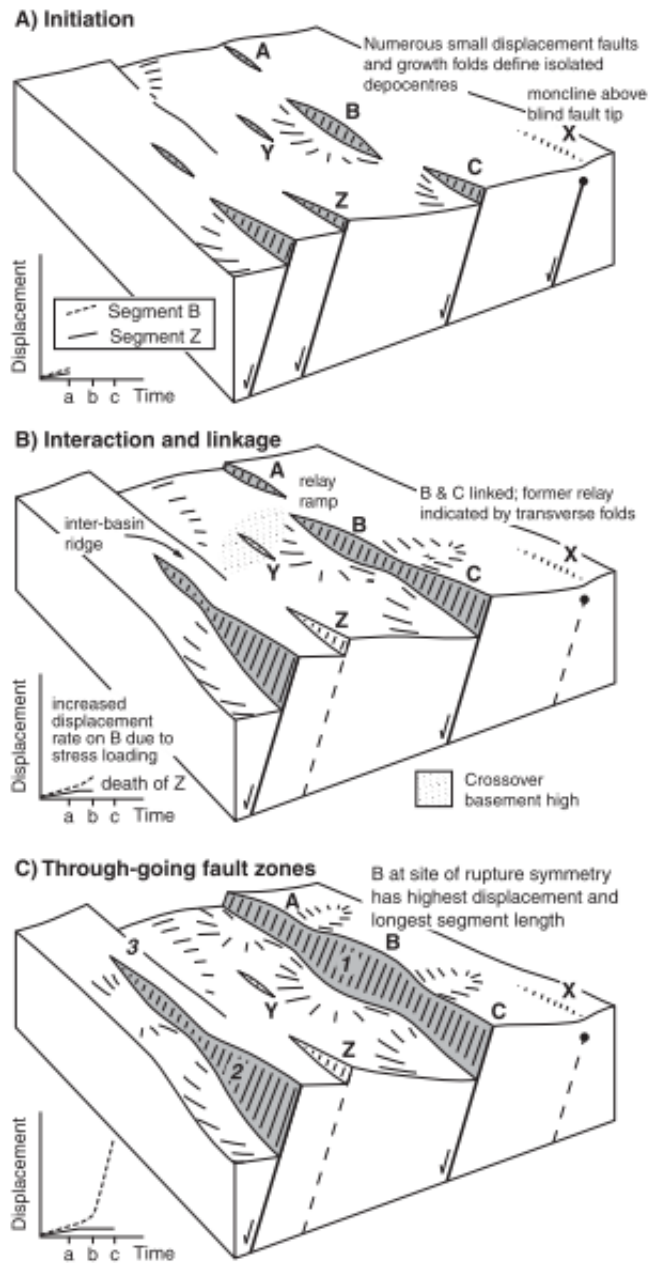


Figure 2. 1. Schematic 3D evolution of normal fault arrays. A) Fault initiation characterised by many low-displacement fault segments. B) Fault interaction and linkage, showing faults interacting and linking, while other fault segments in stress shadows become inactive. C) Through-going fault zones, showing localised deformation on select faults. Figure from Gawthorpe & Leeder, (2000).

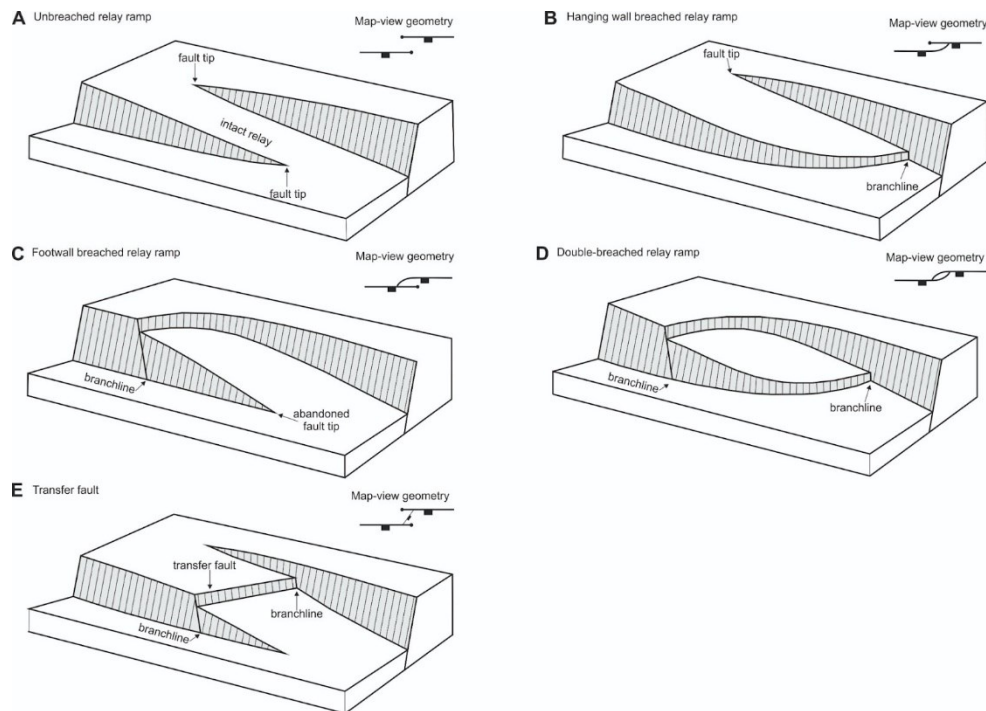


Figure 2.2. Schematic showing different types of relay ramp geometries, including A) Unbreached relay ramp, B) Hanging wall breached relay ramp, C) Footwall breached relay ramp, D) Double breached relay ramp, and E) Relay breaching via a transfer fault. Figure modified from Whipp et al. (2016).

2.3. Fault propagation and interaction

Segment linkage is an important mechanism for fault growth (Cartwright et al., 1995; Wilkins & Gross, 2002; Kim & Sanderson, 2005). As faults propagate laterally, they will inevitably interact with neighbouring faults (Figure 2.1b) and form through-going fault zones (Figure 2.1c).

Relay ramps are locations of fault interactions where strain is transferred from one structure to another (Fossen & Rotevatn, 2016), and this allows the system to act as a whole, maintaining a laterally constant amount of strain across the rift (Walsh & Watterson, 1991; Fossen & Rotevatn, 2016). When faults initially interact and a relay ramp forms, they are soft-linked, i.e., they do not have an obvious physical connection (Figure 2.2a; Trudgill & Cartwright, 1994; Walsh et al., 1999; Fossen & Rotevatn, 2016; Peacock et al., 2016; Whipp et al., 2016; Nicol et al., 2020). Soft-linked faults can cause interacting fault tips to become temporarily pinned, and while they are not physically connected with neighbouring faults, transfer zones are created between them (Figure 2.2a; Trudgill & Cartwright, 1994; Walsh et al., 1999; Fossen & Rotevatn, 2016; Peacock et al., 2016; Nicol et al., 2020). Faults may then become hard linked when the relay ramp between them breaches, physically connecting the fault segments (Figure 2.2b-e; Trudgill & Cartwright, 1994; Walsh et al., 1999; Fossen &

Rotevatn, 2016; Peacock et al, 2016; Nicol et al., 2020). Relay ramps can form in several different ways, which results in a variety of different relay ramp geometries and different displacement-length profiles. Ramps can be breached via the hanging wall (Figure 2.2b), the footwall (Figure 2.2c), double breached (i.e., when both the hanging wall and the footwall; Figure 2.2d), or via a separate transfer fault (i.e., when a fault forms between the two fault tips; Figure 2.2e) (Walsh & Watterson, 1991; Trudgill & Cartwright, 1994; Childs et al., 1995; Fossen & Hesthammer, 1997; Fossen & Rotevatn, 2016; Whipp et al., 2016; Nicol et al., 2020).

There are mechanical drivers behind fault propagation and linkage. Normal faults have idealized bell-shaped D/L profiles, with displacement at their highest in the center and decreasing to zero at its tips (Cowie & Scholz, 1992b). According to the Dugdale elastic-plastic fracture model (Dugdale, 1960), stress at the tips of the fault approaches the shear strength of the surrounding rock and there is a zone of elastic stress around the fault tips (referred to as the breakdown zone) (Cowie & Scholz, 1992b). The fault will propagate if stress at the fault tips exceeds the shear strength of surrounding rock (Cowie & Scholz, 1992b; Gupta & Schulz, 2000; Cowie & Shipton, 2001).

In the case of interacting fault tips, there must be enough strain energy to compensate for the drop in shear stress produced by the adjacent fault. If there is enough strain energy, fault tips will be able to break new surface and eventually link (Gupta & Schulz, 2000). Once a fault segment has linked with an adjacent segment, the structure will no longer laterally propagate until the displacement gradient has increased enough to drive further propagation (Cowie & Roberts, 2001). In order to maintain a constant regional strain, some individual faults become more active and accommodate more strain, while others surrounding them will become inactive/less active (Cowie & Roberts, 2001).

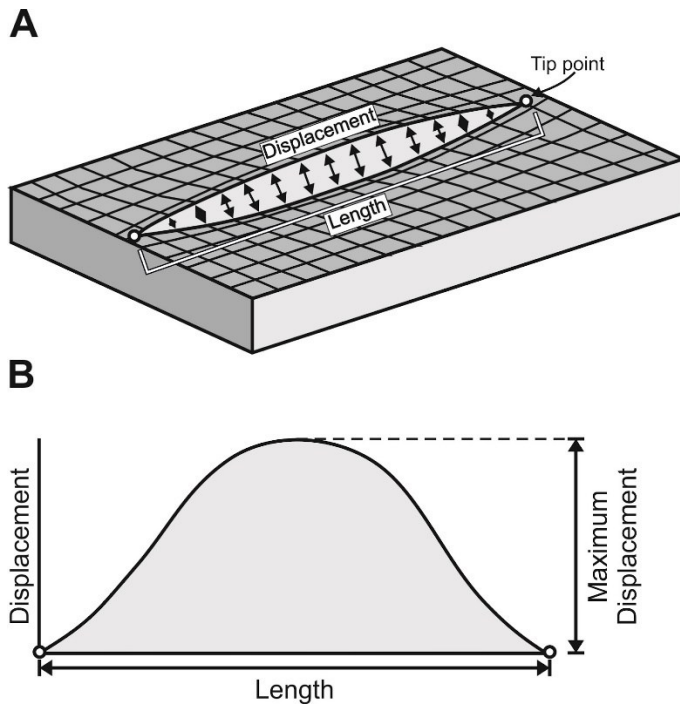


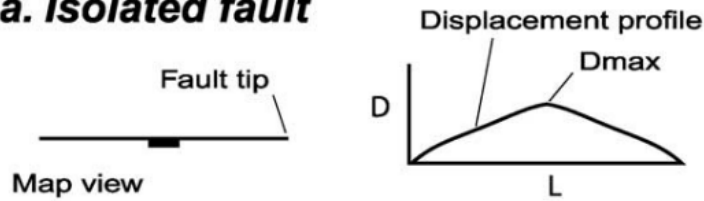
Figure 2. 3. Figure showing the idealised geometry of an isolated normal fault. A) Schematic showing the displacement and length of a normal fault. B) Strike-parallel displacement-length plot showing the idealised bell-shaped curve that is common in normal faults. Modified from Nicol *et al.* (2020).

2.4. Normal fault geometries

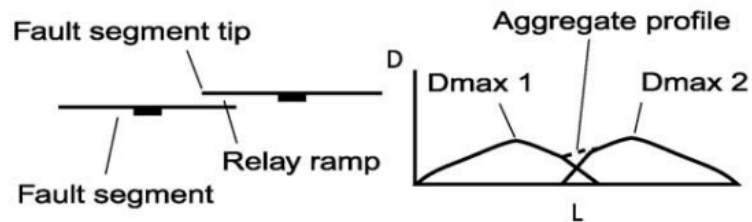
The displacement-length profile of a normal fault has an idealised bell-shaped displacement profile, where displacement is at its highest in the centre of the fault and decreases to zero at the tips of the fault (Figure 2.3 and 2.4a; e.g., Schlische *et al.*, 1996; Childs *et al.*, 2003; Kim & Sanderson, 2005). However, fault geometries are often more complex (Figure 2.4b).

Displacement-length geometries are often linked to fault growth histories. For example, if a fault has two peaks in a displacement-length plot, it likely originally formed as two fault segments that linked via relay ramp linkage (Figure 2.4b-c; Soliva & Benedicto, 2004; Kim & Sanderson, 2005).

a. Isolated fault



b. Fault segment array



c. Relay ramp

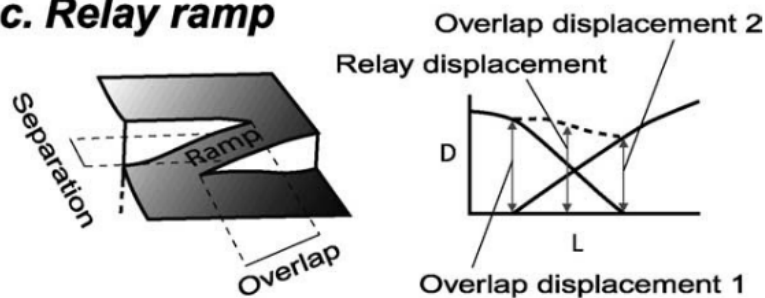


Figure 2. 4. Geometries of idealised normal faults. A) Map-view and displacement-length profile of an isolated normal fault. Note that maximum displacement is in the center of the displacement profile. B) Map-view and displacement-length profile of a fault array with two segments. Note the two maximum displacements in the center of each segment. The aggregate profile of the overlapping zone between the two segments is the sum of the displacements. C) Relay ramp schematic and its associated displacement-length profile. The displacement of the relay and the displacement of the overlapping fault tips is summed together on the displacement-length profile. From Soliva & Benedicto, 2004.

Growth faults are key structures in understanding the kinematics of normal faults. Growth faults, or syn-sedimentary faults, are normal faults that breach the surface and accumulate sedimentation while active (Figure 2.5; e.g., Walsh et al., 2002; Childs et al., 2003, 2017; Taylor et al., 2008; Jackson et al., 2017; Reilly et al., 2017). If fault slip rate is less than or equal to sedimentation, thicker packages of strata will be deposited in the hanging wall compared to the footwall, which is an indicator of when a fault was or was not active (Figure 2.5; Childs et al., 2003). This results in a ‘scoop-shaped’ hanging-wall depocenters which thicken towards the point of highest displacement on the fault (e.g., Schlische, 1995; Young et al., 2001; Jackson et al., 2002; Morley, 2002; Withjack et al., 2002; Gawthorpe et al., 2003; Jackson et al., 2017). Growth faults, especially in 3D seismic data, have allowed us to

track the relationship between fault length and displacement through time (Walsh et al., 2002; Nicol et al., 2005; Giba et al., 2012; Tvedt et al., 2013; Pan et al., 2021).

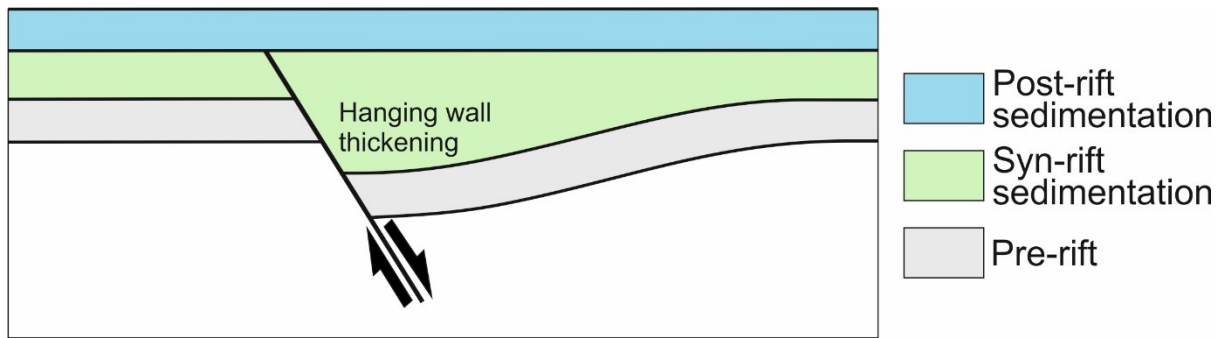


Figure 2. 5. Schematic showing the pre-rift, syn-rift, and post-rift sedimentation of a normal growth fault. Note the hanging wall thickening associated with syn-rift sedimentation. Modified from Reilly et al. (2017).

2.5. Fault displacement and length scaling

The relationship between fault length and displacement in normal faults is typically described by the generalised equation:

$$D_{max}=cL^n$$

Where D is maximum fault displacement, L is maximum fault trace length, and c is a constant, hypothesised to be related to rock material properties such as shear strength and elasticity of the rock. Published values of c range between 0.0001 and 1 (Torabi & Berg, 2011), but usually fall between 0.001 and 1 (Schultz et al., 2008). The value of n falls between 0.5 and 2 ($n=0.5$, Fossen & Hesthammer, 1997; $n=1$, Cowie & Scholz, 1992a; Dawers et al., 1993; Scholz et al., 1993; Clark & Cox, 1996; Schlische et al., 1996; Kim & Sanderson 2005; Xu et al., 2006; $n=1.5$, Marrett & Allmendinger, 1991; Gillespie et al., 1992; $n=2$, Watterson, 1986; Walsh & Watterson, 1988). A number of global databases that investigate D/L relationships exist (Figure 2.6; Walsh & Watterson, 1988; Cowie & Scholz, 1992a; Schlische et al., 1996; Bailey et al., 2005; Torabi & Berg, 2011). These databases are not exclusively for normal faults, as strike-slip and reverse faults are also included. I critically review each of these databases in Chapter 3.

Despite $D-L$ scaling relationships being static measurements, they were initially used as a basis for understanding fault growth. $D-L$ data plotted in log-log space appears to be positively correlated across all orders of magnitude (Figure 2.6), which initially led to the assumption that faults maintain a constant D/L ratio throughout their lives and accumulate displacement and length synchronously (e.g., Watterson, 1986; Gillespie et al., 1992; Cowie

& Scholz, 1992a; Peacock & Sanderson, 1996). When this method of fault growth was first assumed in the 1980s-1990s, faults could only be viewed in 2D and map view, where constraining fault growth through time was difficult (Nicol et al., 2020). This style of growth was later shown to be incompatible with the displacement-length ratios of individual earthquakes ($D=5 \times 10^{-5}L$; Wells & Coppersmith, 1994; Walsh et al., 2002), and once 3D data imaging syn-sedimentary growth faults became widely available from the 1990's onwards, additional fault growth models were introduced.

There are a number of geological factors that can affect the relationship between displacement and length in normal faults, including fault reactivation (Walsh et al., 2002; Vétel et al., 2005; Baudon & Cartwright, 2008; Giba et al., 2012, Whipp et al., 2014; Worthington & Walsh, 2017), the amount of time that the fault has been active (Mouslopoulou et al., 2009; Nicol et al., 2010), host rock lithology (Walsh & Watterson, 1988, 1989; Cowie & Scholz, 1992b; Wibberley et al., 1999; Gudmundsson, 2004), mechanical stratigraphy (Muraoka & Kamata, 1983; Nicol et al., 1996; Gross et al., 1997; Schulz & Fossen, 2002; Soliva et al., 2006; Roche et al., 2013), fault size (Cowie & Scholz, 1992a; Torabi & Berg, 2011; Kolyukhin & Torabi, 2012), along-fault changes in dip (Delokgos et al., 2020), sills reactivating normal faults (Norcliffe et al. 2021), segmented fault geometries and linked fault geometries (Peacock & Sanderson, 1991; Gillespie et al., 1992; Cartwright et al., 1995; Dawers et al., 1995; Acocella et al., 2000; Xu et al., 2006; Faure Walker et al., 2009), decompaction (Taylor et al., 2008), and different types of faults grouped together (Kim & Sanderson, 2005; Torabi & Berg, 2011). I will go into greater detail on how these geological phenomena affect D/L in Chapter 3.

Many common fault measurement errors can skew D/L data, such as not measuring the true maximum length or displacement (i.e., 'cutting effect') (Kim & Sanderson, 2005). Some common errors in D/L measurement are fault throw being measured but included as displacement in a dataset, inaccurate velocity models used in 3D seismic studies, individual earthquakes being included as tectonic faults (Wells & Coppersmith, 1994; Iezzi et al., 2018), deformation bands being included as normal faults (Wibberly et al., 2000b; Schulz et al., 2008; Fossen & Rotevatn, 2012), and fault lengths being under-estimated due to seismic resolution (Yielding et al., 1996; Pickering et al., 1997; Rotevatn & Fossen, 2011). More details on errors in D/L measurement and their effects on D/L scaling can be found in Chapter 3.

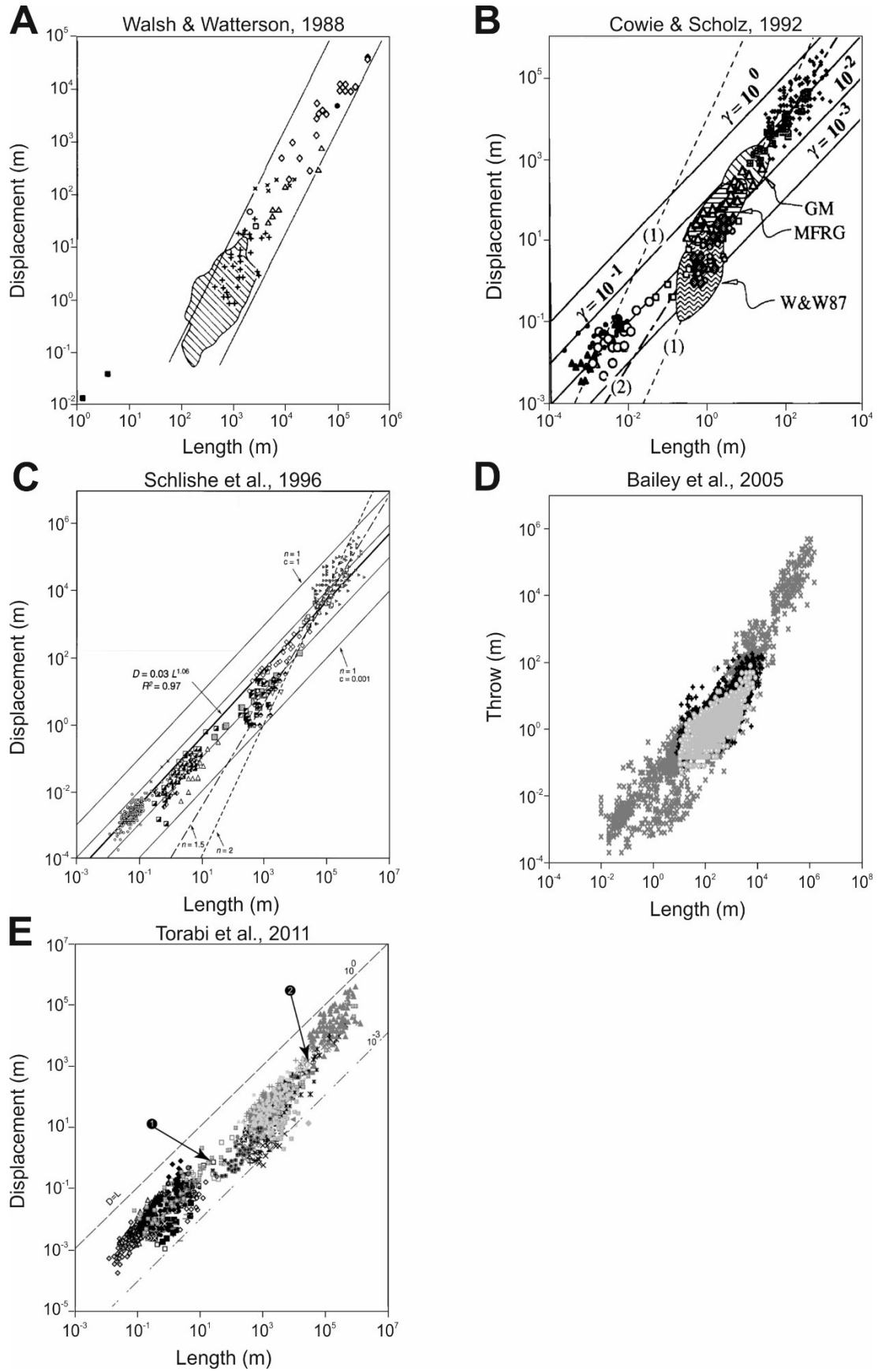


Figure 2. 6. Previously published displacement-length/throw-length fault databases. A) Figure from Walsh et al. (1988). B) Figure from Cowie & Scholz. (1992). C) Figure from Schlische et al. (1996). D) Bailey et al. (2005). E) Torabi et al. (2011).

There are issues associated with scaling normal faults with individual earthquake events (either ancient or active). Faults grow via earthquakes, however D/L ratios in normal faults and singular earthquake events vary greatly. The average D/L relationship for a singular earthquake event is $D=5 \times 10^{-5}L$ (Wells & Coppersmith, 1994), which plots significantly beneath D/L in a normal fault database (Walsh et al., 2002). As faults continue to be active and experience more earthquakes, they tend to have an increasingly higher D/L value (Nicol et al. 2010). Faults also experience aseismic slip, which can account for scaling relationships between normal faults and earthquakes (e.g., Gudmundsson, 2004; Cowie et al., 2013; Fagereng et al., 2014)

2.6. Normal fault growth models

There are two competing models that describe normal fault growth in which faults accrue length and displacement differently. I here describe these end-member models: the *propagating fault model* and *constant-length model*. I also describe a *hybrid model* of the two, as well as other recent developments in normal fault growth.

The *propagating fault model* describes a model in which fault growth occurs via a synchronized increase in fault length and displacement (Figure 2.7a; e.g., Walsh and Watterson, 1988; Dawers et al., 1993; Cartwright et al., 1995; Cowie et al., 2000; Walsh et al., 2003; Childs et al., 2017b; Rotevatn et al., 2019; Nicol et al., 2020). Faults accrue displacement whilst lengthening occurs via tip propagation and segment linkage, and relatively constant D/L ratio is maintained throughout (e.g., Walsh and Watterson, 1988; Dawers et al., 1993; Cartwright et al., 1995; Walsh et al., 2003; Childs et al., 2017b; Rotevatn et al., 2019; Nicol et al., 2020). This method of fault growth has also been referred to as the *conventional fault model* (Walsh et al., 2002), *isolated fault model* (Walsh et al., 2003; Jackson & Rotevatn, 2013; Ghalayini et al., 2017; Morley, 2017), *non-coherent propagating model* (Childs et al., 2017b), and *isolated propagating fault model* (Nicol et al., 2020). I will refer to it as the propagating fault model throughout this thesis.

The *constant-length model* states that normal faults reach their near-maximum length rapidly, and then spend the rest of their lives accumulating displacement without further extensive tip propagation or segment linkage (Figure 2.7b; e.g., Walsh et al., 2002, 2003; Nicol et al., 2005, 2017, 2020; Jackson & Rotevatn, 2013; Henstra et al., 2015; Fossen & Rotevatn, 2016; Hemelsdaël & Ford, 2016; Tvedt et al., 2016; Childs et al., 2017b; Rotevatn et al., 2019).

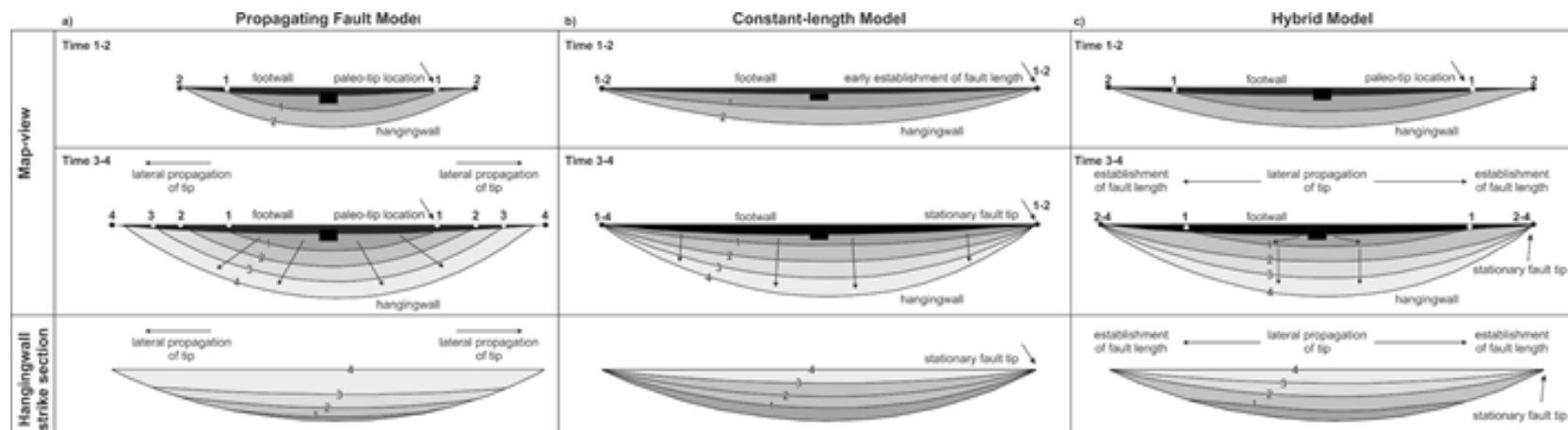


Figure 2. 7. Schematic diagrams illustrating normal fault growth in map-view and along a hanging wall strike section according to the A) propagating fault model, B) constant-length model, and C) hybrid model. Modified from Jackson et al. (2017).

After lengthening, mechanical interactions with adjacent faults retard fault tips, making them unable to laterally propagate (e.g., Gupta & Scholz, 2000; Walsh et al., 2002). The notion that normal faults could form rapidly was first suggested by Cowie (1998). Motivated by the observation that earthquakes do not follow the propagating fault model, Walsh et al. (2002) first presented an example of faults growing in this manner in a 3D seismic study looking at reactivated growth faults in the Timor Sea. The phrase '*constant-length model*' was first used in Nicol et al. (2005). Faults growing via the constant-length model has been confirmed in seismic studies (Giba et al., 2012; Tvedt et al., 2016; Jackson et al., 2017; Pan et al., 2021), physical analogue studies (Schlagenhauf et al., 2008); and a numerical modelling studies (Finch & Gawthorpe, 2017).

Historically, faults have also been described as non-coherent and coherent; non-coherent fault segments form in physical and mechanical isolation of each other, while coherent faults can have several segments within the same structure (Figure 2.8; Walsh & Watterson, 1991; Walsh & Watterson, 2003; Nicol et al., 2017; Childs et al., 2017b). Within these classifications, faults can either lengthen by propagation (Figure 2.8a and 2.8b) or by the constant-length model (Figure 2.8b and 2.8c).

The propagating fault model and the constant-length model are likely end-member models and are not mutually exclusive. Rotevatn et al. (2019) suggests a hybrid model of the propagating fault model and the constant-length model where (i) during the first 20-30% of the faults active lives, they follow the propagating fault model: faults grow via tip propagation and segment linkage to reach their maximum length while 10-60% of maximum displacement is accrued, and (ii) during the remaining 70-80% of the faults life, they follow the constant-length model: faults accumulate the remaining 40-90% of their displacement and fault tips do not propagate any further (Figure 2.7c). A hybrid fault growth model has been shown in 3D seismic studies (Tvedt et al., 2016; Jackson et al., 2017; Rotevatn et al., 2019) and in outcrop studies (Bramham et al., 2021).

It has been suggested that fault arrays grow in cyclical stages where faults alternate between quick lengthening stages and prolonged displacement stages (Pan et al., 2021). During the lengthening stage, faults grow via the constant-length model, lengthening quickly by linking with an adjacent fault, followed by a period of displacement without additional tip propagation. As rifts continue to develop, smaller faults in stress shadows (i.e., faults that are not optimally positioned to accommodate strain) become inactive as strain is partitioned and

localised onto larger faults (Cowie et al., 2000; Gawthorpe & Leeder, 2000; Meyer et al, 2002; Pan et al., 2021). This pattern likely continues until extension stops in the area.

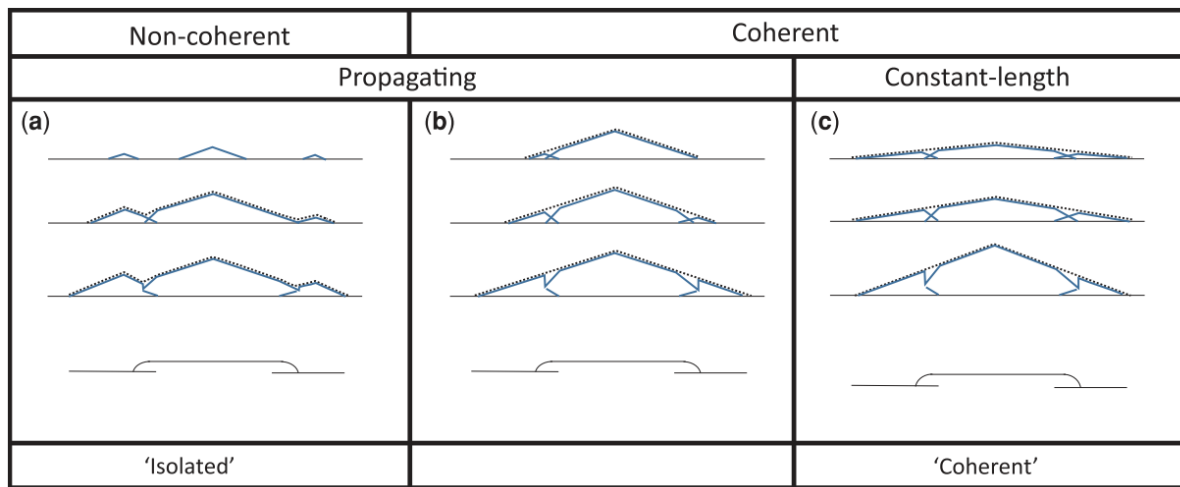


Figure 2. 8. Displacement-length profiles illustrating three possible normal fault growth histories that would lead to the same final fault geometry. The three cases are categorised by if the fault segments are coherent or non-coherent and whether fault tips propagate continually throughout the fault's life or maintain a constant length. Non-coherent propagating faults are sometimes referred to as the 'isolated fault growth model' and coherent constant-length faults are sometimes rereferred to as the 'coherent fault growth model' (Childs et al., 2017b). Propagating coherent faults are rarely referenced in the literature. Figure from Childs et al. (2017b).

2.7. Gaps in knowledge

There is still some uncertainty in our understanding of normal fault growth. First, even though there is mounting evidence that faults do not grow strictly via the propagating fault model, it still appears to be preferred amongst the wider geological community and is still featured in most modern textbooks (e.g., Davis et al., 2011; Fossen, 2016). Second, the details of the early stages of fault growth (the first 20-30% of a faults life) are still widely unknown; the early lengthening stage of fault growth typically cannot be seen in seismic data, as it often happens on a shorter timescale than a resolvable growth package (Nicol et al., 2020). Last, there very few natural examples that show fault D-L through time, so more examples are needed.

2.8. Methodologies used to understand fault kinematics

There are several techniques that can be used to understand fault growth in 3D seismic studies. Here I will detail how isochron analysis, displacement/throw backstripping, and Expansion Index (EI) analysis have been used to track fault length and displacement through time. More information about 3D seismic fault analysis techniques can be found in Jackson et al. (2017).

Isochrone/isochore analysis involves creating stratal thickness maps derived from 3D seismic data; isochrone maps are created using seismic data in milliseconds (ms) or seconds (s) TWT, and isochore maps are created using seismic data in depth (Jackson et al., 2017; Nicol et al., 2020). Surfaces are generated for stratigraphic intervals of growth faults, and isochrone maps show the amount of thickness of the hanging-wall depocenters, and thus the growth history of syn-sedimentary normal faults (Jackson et al., 2017). The propagating fault model would predict the hanging-wall depocenter to get progressively longer at each interval, while the constant-length model would see an early establishment of the depocenter across the entire length of the fault (Jackson et al., 2017).

Displacement/throw backstripping involves subtracting the displacement of a shallower horizon from the displacement of a deeper horizon (Chapman & Meneilly, 1991; Peterson et al., 1992; Jackson et al., 2017). This is an important tool for fault kinematics because it can show the geometry of a fault through time. There are two methods of displacement backstripping that have been used historically, which are extensively outlined in Jackson et al. (2017). When using the *original method*, the throw/displacement from a shallower horizon is directly subtracted from the deeper horizon (Chapman & Meneilly, 1991; Peterson et al., 1992); this method does not assume the fault growth model of the fault and can help show fault tip propagation and how the fault grew through time (Jackson et al., 2017). According to the *modified method*, the throw/displacement along a horizon is subtracted from the entire fault surface (Rowan et al., 1998; Dutton & Trudgill, 2009). The modified method assumes that faults grew via the propagating fault method (Jackson et al., 2017). Since tracking fault growth through time is an important result of displacement/throw backstripping, the original displacement model is the preferred method of displacement backstripping (Tvedt et al., 2016; Jackson et al., 2017).

EI analysis measures variations in stratal thickness in growth packages across the fault (Jackson et al., 2017). It records across-fault thickness changes through time by dividing the thickness of the hanging wall stratal unit by the equivalent thickness of the footwall stratal unit (Thorsen, 1963; Cartwright et al., 1998; Bouroullec et al., 2004; Jackson & Rotevatn, 2013; Jackson et al., 2017). An $EI > 1$ suggests that there was hanging wall thickening and syn-depositional fault activity, which indicates that accommodation space was created by active faulting (Jackson et al., 2017). If $EI = 1$, the fault was not active during the deposition of that strata (Jackson et al., 2017). This is an effective and simple method for determining if a

fault was active without having to solve for decompaction (Taylor et al., 2008; Nicol et al., 2020).

These are all effective strategies to visualise and quantify fault growth, however there are some drawbacks. Firstly, these are all very time-intensive; it can take weeks to pick high-quality horizons apt for making high-resolution isochrons and conducting throw/displacement backstripping can take weeks as well. Secondly, these methods are only useful for specific datasets and areas. The data must image syn-sedimentary growth packages that are not sediment-starved and ideally age-constrained by wells.

There is some error associated with fault analysis using 3D seismic data, even with high-quality, high-resolution data. First, burial-related compaction of sedimentary rocks can result in throw values being under-estimated by up to 20%, especially when the studied interval is deeply buried or has a high shale content (Taylor et al., 2008). Second, fault tips that have a displacement lower than the resolution of the seismic data will not be imaged, which could result in fault lengths being under-estimated by a few hundred meters, depending on fault size (see Pickering et al., 1997).

Given the difficulties of studying normal fault growth from nucleation to cessation in nature, physical and numerical analogue modelling have also been used to understand how normal fault geometries change through time. Sandbox physical analogue models (Filbrandt et al., 2007; Schlagenhauf et al., 2008), plaster-based physical analogue models (Mansfield & Cartwright, 2001; Rotevatn et al., 2019), clay-based physical analogue models (Ackerman et al., 2001; Whipp et al., 2016), and numerical models (Finch & Gawthorpe, 2017) have been used to track fault displacement and length through time, and typically have shown faults growing via a hybrid fault growth model. More details on these physical analogue models can be found in chapter 4.

Chapter 3

Displacement/length scaling relationships for normal faults; a review, critique, and revised compilation

We are planning to submit this chapter to **Earth Science Reviews** in the near future:

Lathrop, B. A., Jackson, C. A. L., Bell, R. E., & Rotevatn, A. (in prep). Displacement/length scaling relationships for normal faults; a review, critique, and revised compilation.

Acknowledgements

We thank Imperial College for providing Bailey Lathrop with the President's Scholarship to fund their PhD research. Whilst compiling this data, we received an outpouring of support from Twitter and the Geotectonics mailing list, and we are grateful to each person that reached out to share data with us. We thank Alex Lipp and Sinead Lyster for their advise during the data analysis portion of this research. We thank Sophie Pan, Ahmed Alghuraybi, and the rest of the Imperial College Basins Research Group (BRG) for their feedback and help throughout this research.

Data availability statement

The databases compiled for this chapter are available open-access via Figshare. All sources used in these databases are cited within. The global normal fault database can be accessed at <https://doi.org/10.6084/m9.figshare.17087273>. Young's Modulus values for different lithologies were also compiled, which can be accessed at <https://doi.org/10.6084/m9.figshare.17087342.v1>.

Abstract

The relationship between normal fault displacement (D) and length (L) varies due to fault size, maturity, basin tectonic history, and host rock lithology. Understanding how fault D and L relate to each other is useful, as D/L scaling laws are often used in conventional and green energy applications, including basin modelling, CO₂ and nuclear waste storage, and groundwater and geothermal studies. Here we provide a review of D/L relationships for normal faults, discuss factors that could influence these relationships, including both geological factors and errors in measurement, and provide a critique of previously published D/L databases. We then present our newly assembled database of 4035 normal faults from 64 sources that include information on: (i) fault length and displacement, (ii) host rock lithology, (iii) host basin tectonic history, and (iv) maturity, as well as fault D and L through time when these data are available. We find an overall fault scaling equation where $D=0.3L^{0.92}$, which is similar to previously published scaling equations and varies in response to the aforementioned factors. Small faults (<1 m length) tend to be over-displaced compared to larger faults, active faults tend to be over-displaced compared to inactive faults, and faults with stiffer host rock lithologies, like volcanic and carbonate rocks, tend to be under-displaced with respect to faults within softer, more compliant host rocks, like clastic sedimentary rocks. Our D-L through time data show that faults follow the *hybrid fault growth* models, where they have a lengthening stage, where faults will appear under-displaced, followed by a displacement accrual stage. To the best of our knowledge, this is the first attempt at such a comprehensive, integrated study of the style of and these factors influencing the growth of normal faults. These revised relationships can now be utilized for predicting fault length or displacement when only one variable is available, and for the general understanding of normal fault geometries. This database will be open-access and available for analysis and manipulation by the broader structural geology community.

3.1. Introduction

The relationship between normal fault length (L) and displacement (D) has been widely researched over several decades (Walsh & Watterson 1988; Cowie & Scholz 1992a; Dawers et al., 1993; Clark and Cox, 1996; Schultz & Fossen, 2002; Kim & Sanderson, 2005; Schultz et al., 2008; Torabi & Berg, 2011; Xu et al., 2006). The empirical relationship between D and L is often described by:

$$D_{max} = cL^n$$

The value n may range from 0.5 to 2.0 ($n=0.5$, Fossen & Hesthammer, 1997; $n=1$, Cowie & Scholz, 1992a; Dawers et al., 1993; Scholz et al., 1993; Clark & Cox, 1996; Schlische et al., 1996; Kim & Sanderson 2005; Xu et al., 2006; $n=1.5$, Marrett & Allmendinger, 1991; Gillespie et al., 1992; $n=2$, Watterson, 1986; Walsh & Watterson, 1988). $n=1$ indicates a linear scaling law, which implies that faults of different sizes act similarly and $n \neq 1$ indicates a scale-dependent geometry, i.e., large faults scale exponentially larger or smaller than small faults (Kim & Sanderson, 2005; Schultz et al., 2008). Cowie & Scholz (1992b) argued that $n=1$ is controlled by Dugdale's fracture model, which indicates that deformation at the tip of a tensile crack in an elastic-plastic material is inelastic deformation.

The value c (sometimes written as P or γ) is an expression of fault displacement at unit length and is hypothesized to be related to rock material properties such as shear strength and elasticity of the rock, as well as the driving stress; for example, as rock shear strength increases, for example from a mudstone to a granite, c increases (Walsh & Watterson, 1988; Cowie & Scholz, 1992b; Gillespie et al., 1992; Ackermann et al., 2001; Kim & Sanderson, 2005; Schultz et al., 2008; Torabi & Berg, 2011). Reported values of c range from 0.0001-1 (Schulz et al., 2008; Torabi & Berg, 2011); the value of c typically falls between 0.001 and 0.1 (Schultz et al., 2008; Torabi & Berg, 2011). High values of c (i.e., $c=1$) have come from strike-slip faults (MacMillan, 1975; Torabi & Berg, 2011).

This scaling relationship has typically been used to: i) assess the way in which normal faults form, with applications to geohazard analysis (Cowie & Scholz, 1992b), and ii) to allow better prediction of fault dimensions, with applications to energy resource exploration and extraction, nuclear waste, and CO₂ storage, which all rely on robust structural models which are commonly constructed from incomplete datasets (Torabi & Berg, 2011; Kolyukhin & Torabi, 2012). We may need to estimate L when only D (or vice versa) can be constrained from an isolated field exposure or when only 2D seismic data are available. For example, in

basin modelling, fault connectivity impacts fluid flow from source to reservoir, thus knowing how fault length might impact that, and how displacement may influence fault seal, is key.

When plotted in log-log space, the relationship between displacement and length appears strongly positively correlated across several orders of magnitude (see D/L plots in Walsh & Watterson, 1998; Cowie & Scholz, 1992a; Schlische et al., 1996; Kim & Sanderson, 2005; Torabi & Berg, 2011). However, the relationship between normal fault length and displacement is highly variable, and a one-size-fits all equation to describe D/L scaling is likely imprecise. Understanding how factors such as tectonic history, fault maturity, host rock lithology, and fault size effect D/L scaling, and using these observations to create bespoke D/L equations, will improve our ability estimate fault D or L.

A linear relationship (i.e., $n=1$) between D and L was used to justify a model of normal fault growth where faults accumulated displacement and length synchronously; this was originally referred to as the *isolated fault model* but is now commonly referred to as the *propagating fault model* (e.g., Walsh & Watterson, 1988; Morley et al., 1990; Dawers et al., 1993; Cartwright et al., 1995; Walsh et al., 2003; Childs et al., 2017b; Rotevatn et al., 2019). However, numerous studies have since challenged the notion that fault growth follows a linear trajectory in D-L scaling space and have instead postulated that faults grow in accordance with the *constant-length model*. The constant-length model suggests that faults reach their near-final length rapidly and then accrue displacement without significant further tip propagation (e.g., Walsh et al., 2002, 2003; Nicol et al., 2005, 2017; Jackson & Rotevatn, 2013; Henstra et al., 2015; Fossen & Rotevatn, 2016; Hemelsdaël & Ford, 2016; Tvedt et al., 2016; Childs et al., 2017b; Rotevatn et al., 2019; Pan et al., 2021). Faults have also been shown to grow in accordance with the *hybrid fault model*; this combines the propagating and constant-length models, suggesting that faults grow in two distinct phases: (i) an initial phase (20-30% of the faults life), when maximum fault length is reached and 10-60% of displacement is accrued, (ii) a second stage (the remaining 70-80% of the faults life) when 40-90% of displacement is accrued (Rotevatn et al. 2019). Some faults will also experience a stage of lateral tip-line retreat in the last ~25% of their lives, where slip is concentrated along their central portions (Meyer et al., 2002; Morley 2002; Nicol et al., 2020; Lathrop et al., 2021).

Global databases of D/L data result in a range of scaling relationships with different values for both c and n . There are several possible reasons for this. First, existing fault global

databases have included faults with errors in measurement that could influence D/L ratios; scaling relationships based on these imprecise data will not be as reliable as we wish. Second, there has been little research into how D/L scaling relationships change for faults of different size, forming in differing tectonic settings (i.e., if a fault forms due to the reactivation of an older structure, or whether it is newly formed in previously undeformed or only weakly deformed host rock), in different host rock lithologies, and fault maturity (i.e., if the fault is in a tectonically active area or not). It has been noted that these factors can cause high variability in global datasets (e.g., Cowie & Scholz, 1992a; Nicol et al., 2010; Rotevatn et al., 2019), but this variability has not yet been quantified (see section 2). Finally, if faults really do grow via a constant-length or hybrid fault growth model, D/L ratios will vary greatly throughout the life of a fault, and thus D/L ratios from faults of different stages in their development are less meaningful, and a compilation of D and L through time will more accurately show how faults grow than a singular measurement taken: (i) at the end of a fault's life, once it has become inactive, or (ii) as a snapshot at a specific, possibly unknown time in the fault's development.

It is clear there are numerous factors that may cause variability in the important, widely used relationship between fault displacement and length. In this paper we look closely at normal fault D and L, isolating various factors that could affect it and put forward improved scaling laws for different scenarios. We first summarise the contributions and discuss inconsistencies in previous fault databases and quality check the included data. We next provide a new open-source normal fault database that includes factors such as fault maturity, tectonic history, and host rock lithology, which previous work suggests may be important to consider when establishing and ultimately applying D/L relationships. We also compile data on normal fault D and L through time (i.e., from structures flanked by growth strata that permit displacement and length backstripping; Meyer et al., 2002; Tvedt et al., 2016; Jackson et al., 2017; Lathrop et al., 2021; Pan et al., 2021, physical analogue studies; Schlagenhauf et al., 2008; and numerical modelling studies; Finch & Gawthorpe, 2017) to show how faults grow and how D/L ratios change through time. Finally, we interrogate our new database and discuss how the factors size, host rock lithology, tectonic history, and fault maturity affect fault growth and D/L scaling. This is the first attempt at a comprehensive, integrated study of these factors on normal fault growth. Our new database of normal fault properties demonstrates that one-size-fits-all scaling relationships are overly simplistic and that D/L scaling relationships should not be used indiscriminately.

3.2. How might geological factors and measurement errors influence scaling laws?

There are a range of geologic phenomena that can cause normal faults to be over or under-displaced, and that are known to influence D/L databases. Several common errors in measurement can also influence D/L ratios and equations. We briefly outline these and illustrate how the related data would theoretically plot in D/L scaling space (Figure 3.1), before addressing measurement errors in published datasets.

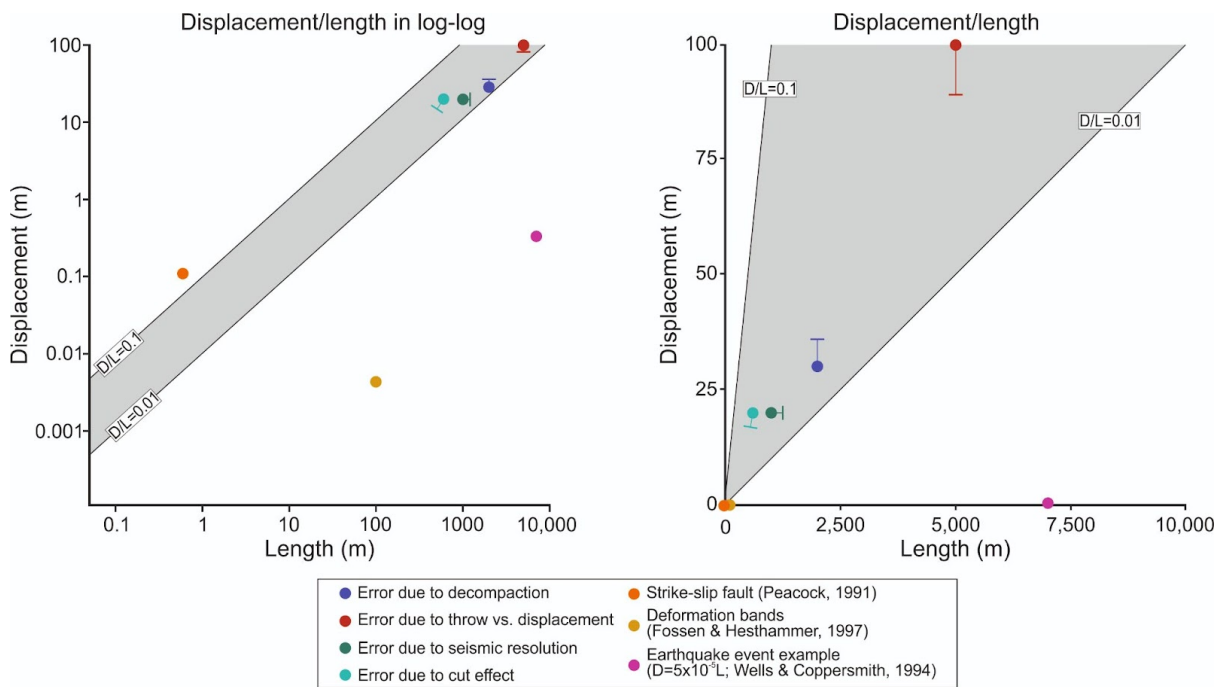


Figure 3. 1. Schematic showing how errors in measurement and data from structures other than normal faults that can affect D/L scaling in and out of log-log space.

3.2.1. Geological factors

Different geological phenomena could affect the relationship between fault length and displacement. Tectonic setting is said to affect the relationship between D and L (Cowie & Scholz, 1992a). Specifically, reactivated faults can establish their maximum length more quickly than non-reactivated faults, which makes reactivated normal faults have an especially low D/L ratio, at least in the early stages of fault growth (Walsh et al., 2002, Vétel et al., 2005, Baudon & Cartwright, 2008, Giba et al., 2012, Whipp et al., 2014).

The amount of time that a fault has been active can affect D/L ; Mouslopoulou et al. (2009) noted that fault displacement rates vary through time, especially in ‘young’ faults (<20 Kyr), which can result in ~an order of magnitude of D/L scatter. Nicol et al. (2010) demonstrated that active faults are under-displaced in the early stages of growth and then D/L increases with age.

Host rock lithology can change the D/L ratio of a fault, and host rock lithology is tied to shear modulus and Young's Modulus. Walsh & Watterson (1988, 1989), Cowie & Scholz (1992b) and Wibberley (1999) compared fault D/L and host rock shear modulus, and they showed that stiffer lithologies (i.e., high shear modulus) are under-displaced compared to softer lithologies (i.e., low shear modulus). Gudmundsson et al. (2004) suggested that faults with a low Young's Modulus and D/L are inversely related; faults with softer and/or more deformed host rocks have a lower Young's Modulus and higher D/L ratios, and faults with stiffer host rocks have a higher Young's Modulus and lower D/L ratios. Several studies have also shown that mechanical stratigraphy can affect D/L (Muraoka & Kamata, 1983; Nicol et al., 1996; Gross et al., 1997; Schulz & Fossen, 2002; Soliva et al., 2006; Roche et al., 2013). Faults being stratigraphically confined can prevent faults from vertical propagation, and thus can cause them to be under-displaced (Schulz & Fossen, 2002), but the effects of mechanical stratigraphy are dependent on factors such as changes in rock properties between the lithological contacts (Roche et al., 2013, 2014) and the thickness of the layers (Soliva et al., 2006).

Fault size could affect D/L ratios, however, there has been some disagreement in the literature in what way. Schlische et al. (1996) did not find a relationship between D/L, however Cowie & Scholz (1992a) found that large faults (>1 km) are over-displaced compared to smaller faults, and Torabi & Berg (2011) noted that small faults (<1 m) and large faults (>1 km) have a higher D/L ratio than medium faults (between 1 m and 1 km).

When faults have along-fault changes in dip (i.e., fault dip changes in cross-section) strain is partitioned onto bends, which causes faults to appear either over or under-displaced.

According to estimations by Delokgos et al. (2020), fault bends are predicted to typically cause throw to be under-estimated by approximately 10%, and up to 50% in extreme cases.

Sills reactivated under normal faults can affect D/L ratios. Sills often form underneath pre-existing normal faults and force increased fault displacement post-rift, which does not reflect the faults tectonic history (Norcliffe et al., 2021).

Segmented fault geometries and the interaction and linkage of segments can cause faults to have two smaller displacement maxima instead of one large maximum displacement value. This can cause the faults to appear under-displaced (Peacock & Sanderson, 1991; Gillespie et al., 1992; Cartwright et al., 1995; Dawers et al., 1995; Acocella et al., 2000; Xu et al., 2006; Faure Walker et al., 2009).

Fault drag can reduce the amount of displacement measured on a normal fault, especially on larger faults (Walsh & Watterson, 1987; Gross et al., 1997; Kim & Sanderson, 2005; Childs et al., 2017a; Delogkos et al., 2017). Delogkos et al. (2017) noted that fault drag accounted for up to ~24% of total throw in faults between 35-550 m in the Ptolemais Basin, Greece.

3.2.2. Measurement errors

The relationship between displacement and length could be affected by where on a fault surface these values are measured, as it is possible that the true maximum length and displacement have not been recorded (Kim & Sanderson, 2005). Maximum displacement is typically located at a fault center, however an arbitrary section of the fault exposed in outcrop may not cut through the center, which is referred to as 'cutting effect' (Kim & Sanderson, 2005). The 3D nature of faults measured in outcrop and 2D seismic could lead to under-estimation of maximum displacement or length (Kim & Sanderson, 2005; Torabi et al., 2019).

If fault offset is measured as throw instead of displacement and is included in a D/L database without knowledge of fault dip, the D/L ratio would not be accurate (Figure 3.1). This does not alter the position of a plot point greatly (Figure 3.1), but it could affect scaling equations.

D/L ratios can be skewed if different types of faults are plotted together. For example, strike-slip faults tend to be over-displaced compared to normal faults, with D/L ratios being as high as 1:1, whereas normal faults haven't been known to have a ratio higher than 1:2 (Kim & Sanderson, 2005; Torabi & Berg, 2011), so this could skew normal fault scaling laws towards being more over-displaced (Figure 3.1). The higher D/L ratios in this case are possibly due to fault length being measured parallel to slip direction; whereas normal fault length is measured perpendicular to dip (Kim & Sanderson, 2005; Torabi & Berg, 2011).

Displacement and length relationships measured from individual earthquakes scale differently than measurements taken from faults, i.e., the average slip to rupture length scaling relationship for individual earthquake events is $D=5 \times 10^{-5} L$ (Wells & Coppersmith, 1994; Iezzi et al., 2018; Figure 3.1), thus singular earthquakes should not be added to fault databases.

Deformation bands are mechanically different than tectonic faults; deformation bands experience strain hardening after formation due to grain interlocking, and strain then tends to localize elsewhere and form new bands instead of increasing displacement on existing bands (Fossen & Rotevatn, 2012). This causes deformation bands to be under-displaced compared

to tectonic faults, usually having a value of $n=0.05$, and thus they would skew D/L scaling relationships in a global database (Wibberly et al., 2000b; Schulz et al., 2008; Fossen & Rotevatn, 2012; Figure 3.1).

There is also error associated with D/L measurements in 3D seismic; length could be underestimated by a few hundred meters to a few kilometers, depending on fault size, due to displacement at fault tips being under seismic resolution (Yielding et al., 1996; Pickering et al., 1997; Rotevatn & Fossen, 2011). If fault displacement is measured without an accurate velocity model in seismic data, the D/L ratio may not be accurate; if the velocity is too high, displacement will be under-estimated for values measured in time-migrated data. Additionally, compaction could decrease throw values (up to 20% compaction according to Taylor et al., 2008; Figure 3.1), which can be an issue for deeply buried faults measured in seismic or outcrop.

The measurement errors described above can both significantly change calculated D/L equations and visually skew plotted data. D/L equations are undoubtedly important to making D and L estimations, but D/L plots are also important as they are often used to qualitatively check if a D/L relationship is strong or weak. These errors can skew D/L plots to different extents, depending on how the data is presented (Figure 3.1). The effects of errors such as measuring throw instead of displacement, cutting effect, decompaction, and fault tip resolution in seismic, result in changes that are visible in a graph not in log-log space (Figure 1a), but make little difference in a log-log D/L graph (Figure 3.1b). Data in log-log space tend to ‘hide’ fluctuations due to small measurement errors, as data will move very little and will still lie within the range of values in the global database. These errors in measurement could change D/L scaling equations, but unless the error is more than an order of magnitude than the correct value, it likely will not be seen in log-log space. However, when structures besides normal faults, such as strike-slip faults and deformation bands, or measurements from singular earthquake events are included in a database, they fall significantly outside the typical range of D/L values (Figure 3.1). While more error is visible in a plot not in log-log space, there is a bias towards larger faults if the plot spans several orders of magnitude, as only the largest faults are visible when faults of all sizes are included on one plot (see Figure 3.1).

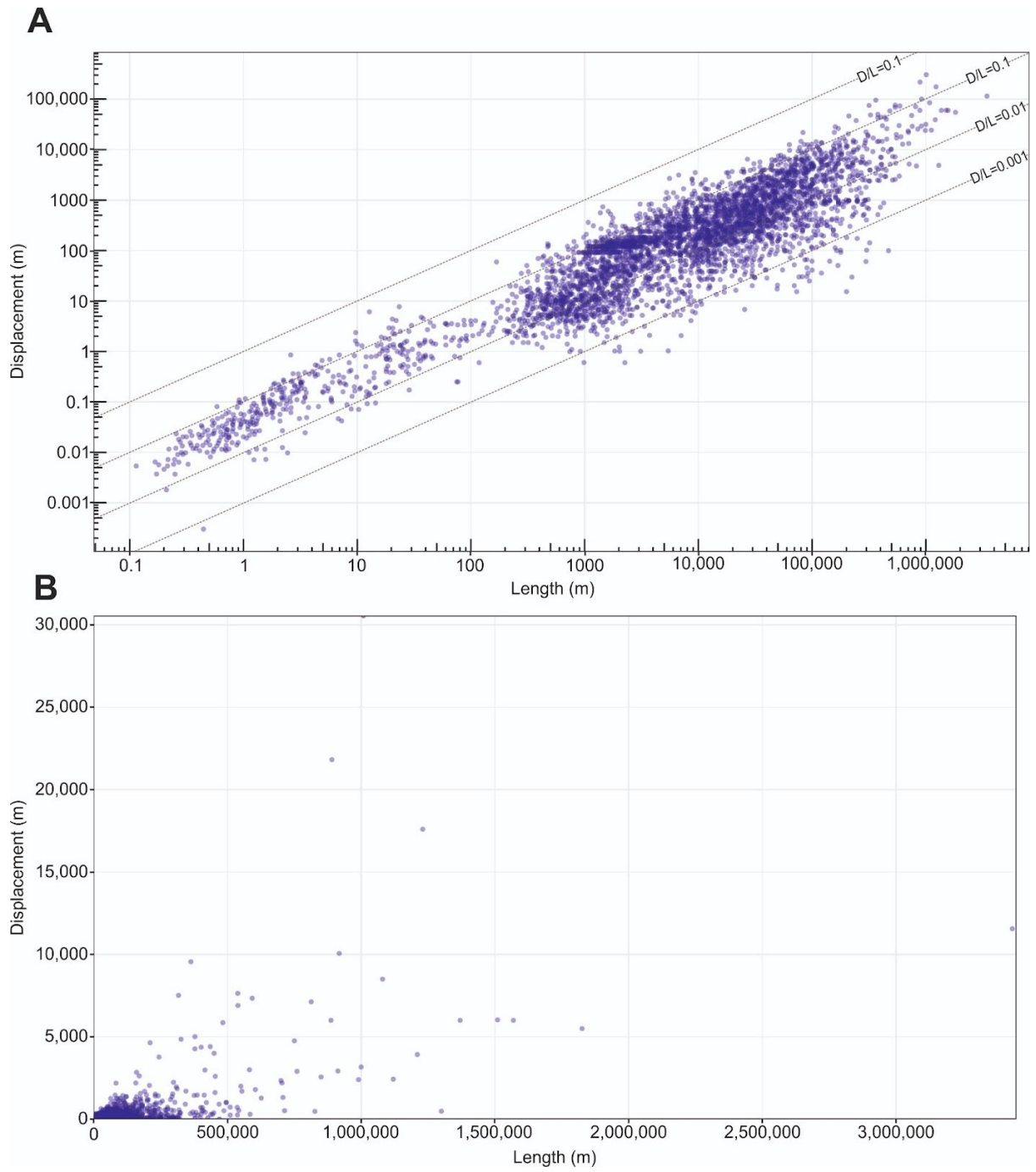


Figure 3. 2. Plots showing fault length vs displacement for all data in our database. A) Data in log-log space. B) Data not in log-log space.

Database	# of faults	Sources Used	n	C
Walsh & Watterson, 1988	308	Beck, 1929; Reeves, 1929; Teas, 1929; Babenroth & Strahler, 1945; Brunstrom, 1963; Tschopp, 1967; Janoschek & Gotzinger, 1969; Mayuga, 1970; Bond et al., 1971; Huntoon, 1974; Cave, 1977; Shepherd & Burns, 1978; Shoemaker et al., 1978; Frost & Smart, 1979; Drozdowski, 1980; Frost & Halliday, 1980; Nelson, 1980; Van den Bark & Thomas, 1980; Verdier et al., 1980; Muroaka & Kamata, 1983; Aitkenhead et al., 1985	2	not given
Cowie & Scholz, 1992	210	MacMillan, 1975; Elliott, 1976; Muraoka & Kamata, 1983; Walsh & Watterson, 1987; Krantz, 1988; Opheim & Gudmusson, 1989; Peacock, 1991; Peacock & Sanderson, 1991; Villemin et al., 1995	1	0.006-0.17 (range)
Schlische et al., 1996	547	McMillan, 1975; Elliott, 1976; Muroaka & Kamata, 1983; Walsh & Watterson, 1987; Krantz, 1988; Opheim & Gudmusson, 1989; Peacock, 1991; Peacock & Sanderson, 1991; McGrath, 1992; Dawers et al., 1993; Villemin et al., 1995	1.06	0.03 (average)
Bailey et al., 2005	9618	Beck, 1929; Reeves, 1929; Teas, 1929; Babenroth & Strahler, 1945; Fox, 1959; Brunstrom, 1963; Woodland & Evans, 1964; Tschopp, 1967; Janoschek & Gotzinger, 1969; Wood et al., 1969; Freund, 1970; Mayuga, 1970; Bond et al., 1971; MFRG, 1973; Huntoon, 1974; MacMillan, 1975; Elliott, 1976; Cave, 1977; Ruzhich, 1977; Shepherd & Burns, 1978; Frost & Smart, 1979; Drozdowski et al., 1980; Nelson, 1980; Verdier et al., 1980; Van den Bark & Thomas, 1981; Muroaka & Kamata, 1983; Aitkenhead et al., 1985; Villemin & Sunwoo, 1987; Opheim & Gundmundsson, 1989; Walsh & Watterson, 1988; Gillespie, 1991; Marret & Allmendinger, 1991; Peacock, 1991; Gillespie et al., 1992; Gillespie et al., 1993; Dawers et al., 1993; Davison, 1994; Dawers & Anders, 1995; Cartwright et al., 1996; Jackson et al., 1996; Nicol et al., 1996; Schlische et al., 1996; Rowan, 1997; Fossen & Hesthammer, 1998	1.19	not given
Torabi et al., 2011	not given	McMillan, 1975; Elliot, 1976; Muroaka & Kamata, 1983; Watterson, 1986; Villemin & Sunwoo, 1987; Walsh & Watterson, 1987; Krantz, 1988; Opheim & Gudmundsson, 1989; Peacock, 1991; Peacock & Sanderson, 1991; Gillespie et al., 1992; McGrath, 1992; Scholz & Cowie, 1992; Dawers et al., 1993; Gillespie, 1993; Cartwright et al., 1995; Villemin et al., 1995; Nicol et al., 1996; Schlische et al., 1996; Yielding et al., 1996; Gross et al., 1997; Wibberly et al., 1999; Kim, 2000; Walsh et al., 2002; Wilkins & Gross, 2002; Davis et al., 2005; Schultz et al., 2008	1	0.0001-1

Table 3. 1. Table showing previously published global databases with the number of faults used and their sources, along with their values of *n* and *C* if given.

3.3. Issues with previous D/L databases

Several highly cited displacement/length and throw/length databases have been compiled in the past 35 years (Walsh & Watterson, 1988; Cowie & Scholz, 1992a; Schlische et al., 1996; Bailey et al., 2005; Torabi & Berg, 2011). Some of the data included in these contributions do not measure true D or L, despite these data being reused in newer publication compilations. As a result, D/L scaling ratios could be affected. Additionally, the way that the data was presented tightly in log-log space made the data non-replicable. We here review these complications and suggest which data points could inaccurately skew D/L ratios and should not be included in future databases.

Walsh & Watterson (1988) was the first contribution that compared the relationship between length (referred to as fault width) and displacement using a global compilation of faults. 308 normal faults from the British Coalfields were compared to a global dataset of 58 faults from 22 sources (Table 3.1). In that paper, the relationship between fault length and displacement was described as $D=L^2/P$, where P (equivalent to c) is a variable related to rock properties, such as host rock shear modulus, (e.g., Cowie & Scholz, 1992b; Bailey et al., 2005; Kim & Sanderson, 2005; Nicol et al., 2020). They use the assumption that $n=2$ because all of their data was bounded by a slope of 2, however, their data had an overall regression line of $n=1.58$ (Walsh & Watterson, 1988). An average best fit equation was not given, so the average value of c is not known. Of the 22 sources included in their dataset, nine had included data where D or L was not explicitly given, which could have skewed their final D/L trends. For example, fault L and/or D were not included in some of the original sources (Teas, 1929; Babenroth & Strahler, 1945; Brunstrom, 1963; Mayuga, 1970; Huntoon, 1974; Van den Bark & Thomas, 1980; Aitkenhead et al., 1985). We note that Teas (1929) lists the measurement of the closure around the fault, which was likely included as displacement, and Huntoon (1974) does not explicitly state fault length and displacement, therefore Walsh & Watterson (1988) may have established fault length and displacement from a rough schematic map of the study area (see Figures 2, 3, 5, and 6 in Huntoon, 1974). In some papers, D and/or L were given as a range of values and not precise measurements (i.e., displacement ranges from 100-500 m; Shepherd & Burns, 1978; Frost & Halliday, 1980), and Walsh & Watterson (1988) may have picked a mid-point or maximum value of the range, which could possibly change the derived scaling relationship, making the data appear over or under-displaced depending on what value was chosen. The data from Babenroth & Strahler (1945) and

Huntoon (1974) was also originally given as throw, and was included in this dataset as displacement, which could make the data look slightly over-displaced; throw data could be converted to displacement if the fault dip is known or assumed, however this is not discussed in the methodology. It is also entirely possible that correct fault length and displacement values were given to Walsh & Watterson (1988) via personal correspondence with the authors, however that was not included in the methodology or indicated by an in-text citation.

Cowie & Scholz (1992a) subsequently compared D/L relationships of ~210 faults from nine different sources, one of which overlaps with the sources used in Walsh & Watterson (1998; Table 3.1). Their data suggest a linear D/L scaling relationship ($n=1$) (Cowie & Scholz, 1991a), which would suggest an equation of $D=cL$; average values of c are not given. They note that large faults (defined as faults longer than 1 km) have a higher D/L ratio, possibly due to faults that cut through the brittle upper crust (usually faults with $L > 10$ km) having a higher displacement (Cowie & Scholz, 1992a). This could also be due to large faults being two-dimensional and only pinned at their ends, whereas small faults are three-dimensional and are pinned along their entire parameter, or possibly because large faults rupture stronger rock on average (Cowie & Scholz, 1992b).

Of the nine sources used in Cowie & Scholz (1992a), four had potential errors in measurement or included data other than normal faults, which could have affected their D/L ratios. For example, both thrust faults (Elliot, 1976) and strike-slip faults (MacMillan, 1975; Peacock, 1991) were included in Cowie & Scholz (1992a). Additionally, neither D nor L data was presented in the data from Krantz (1988) (which contributed ~12 of ~210 data points) so we cannot be sure where these values are coming from. Again, it is possible that the correct fault length and displacement values were obtained via personal correspondence.

Schlische et al. (1996) compared 201 normal faults from the Dan River Basin (USA) to a global database of 346 faults from 11 sources, nine of which overlap with the earlier Walsh & Watterson (1988) and Cowie & Scholz (1992a) databases (Table 3.1). One of the key aims of this paper was to compare the D/L relationship of small ($L < 1.25$ m) and larger faults. They found that D/L did not vary based on fault size. Of the faults in their global compilation, 174 were strike-slip faults from two different sources, and 172 were normal faults from 11 different sources (Table 3.1). They note a broadly linear relationship between D and L ($n=1$), with c values between 0.001 and 1; some of the variability in c could be due to the inclusion of strike-slip faults in the dataset, which typically have a higher D/L ratio than normal faults

(Kim & Sanderson, 2005). The best fit curve through the compiled data is arithmetically defined by $D=0.03L^{1.06}$, with the authors noting that there is no significant change in the D/L scaling relation across many orders of magnitude.

Bailey et al. (2005) compared throw-length (rather than displacement-length) relationships of their 7862 normal faults from the East Pennine Coalfield, UK to a global dataset of 1756 faults from 46 different sources, 22 of which overlap with Walsh & Watterson (1988), Cowie & Scholz (1992a), or Schlische et al. (1996) (Table 3.1). The aim of the study was to map strain distribution across the East Pennines, UK (Bailey et al., 2005). Of the 46 sources used, 29 had potential errors in measurement, included data that was not from normal faults, or was from a source that was not publicly available, which could have affected D/L scaling. For example, length and/or displacement/throw were not listed in the original sources of several datasets (Beck, 1929; Teas, 1929; Babenroth & Strahler, 1945; Brunstrom, 1963; Woodland & Evans, 1964; Wood et al., 1969; Mayuga, 1970; Huntoon, 1974; Van den Bark & Thomas, 1980; Aitkenhead et al., 1985; Gillespie et al., 1993). For example, Beck (1929) only had length and displacement shown in a rough schematic cross-section, Krantz (1988) only measured slip vector direction, Gillespie et al. (1993) measured fault spacing, and Gross et al. (1997) measured maximum dip separation, yet all these values were included as throw. Thrust faults were included in the database (Fox, 1959; Elliott, 1976; Rowan, 1997), as well as strike-slip faults (Freund, 1970; MacMillan, 1975; Peacock, 1991). Data from unpublished (and still publicly inaccessible) theses were also included (MacMillan, 1975; Gillespie et al., 1991), as were data from individual earthquakes (Jackson et al., 1996). Some faults had either displacement or length listed as a range of values instead of a precise measurement (see Figure 3.1) (Shepherd & Burns, 1978; Frost & Halliday, 1980). There were also some duplicate data, where the same faults were studied in two separate papers and both were included; note that this does not visually affect the data plot but can influence scaling relationship calculations (Dawers et al., 1993; Dawers & Anders, 1995). Deformation bands were also included as faults (Fossen & Hesthammer, 1997), with these structures having displacements up to two orders of magnitude smaller than tectonic faults of the same length. Several sources measured fault displacement in their original sources (Muroaka & Kamata, 1983; Opheim & Gudmundsson, 1989; Walsh & Watterson, 1988; Marret & Allmendinger, 1991; Dawers et al., 1993; Nicol et al., 1996; Schlische et al., 1996), but were included in Bailey et al. (2005) as throw. Despite these issues, the data compiled by Bailey et al. (2005)

has been used in several subsequent papers (Nicol et al., 2010; Nicol et al., 2017; Reilly et al., 2017; Rotevatn et al., 2019; Bramham et al., 2021).

The most recent compilation by Torabi & Berg (2011) compared the relationship between fault length and displacement of faults in siliciclastic rocks from 27 sources, 16 of which had overlap with Walsh & Watterson (1988), Cowie & Scholz (1992a), Schlische et al. (1996), or Bailey et al. (2005) (Table 3.1). The total number of faults they included is unclear, as the data is very tightly spaced in the presented scatterplot and the raw data not available for analysis. However, in the text they state these data are normal faults from 22 sources, reverse faults from four sources, and strike-slip faults from three sources (some sources had more than one type of fault; Table 3.1). This paper considered the potential causes of scatter in the data, such as the underestimation of the frequency of small faults (truncation effect), and the under-estimation of the frequency of long faults due to sample line limitations (censoring effect) (Torabi & Berg, 2011). They found that small faults ($L < 1$ m) and large faults ($L > 1$ km) have a similar D/L ratio, and that medium-sized faults (1-1000 m) tend to be comparatively under-displaced (Torabi & Berg, 2011). They suggest this difference arises because medium-sized faults are still growing by segment linkage, and that their D/L ratio will eventually match that of larger faults as they mature (Torabi & Berg, 2011). They also found that strike-slip faults are over-displaced compared to normal and reverse faults, and that cataclastic deformation bands are under-displaced compared to faults (Torabi & Berg 2011). Length and/or displacement was also not listed in the original sources of several datapoints (Krantz, 1988; Gillespie et al., 1993. Vertical offset (i.e., throw) was measured in Villemin & Sunwoo (1987), which would vary slightly from displacement.

3.4. Methodology

Our D/L database includes 4046 normal faults from 69 sources (Table 3.1), ranging in length from 10 mm to 245 km, in age from the Carboniferous to presently active faults, and in duration of activity from faults that were active for >100 Myr to those that have been active for <1 Myr, from natural faults as well as physical and numerical models (Table 3.2).

Maximum length and maximum displacement are noted in our database, along with fault host rock lithology, fault maturity, and tectonic history. We focused on these parameters because they are known to affect fault growth (e.g., Cowie & Scholz, 1992a; Torabi & Berg, 2011), and they provide a relatively easy and replicable way of characterizing and comparing faults. All the data are provided in raw format and are publicly available, such that the wider geologic community can easily access and analyze it and, in the future, build upon it.

Source	Number of Faults	Data Type	Dominant Lithology	Reactivated	Size Range (length in m)	Active/inactive
Alghuraybi et al., 2021	18	3D seismic	Fine-grained clastic	No	4714-42,673 m	Inactive
Balsamo et al., 2016	23	Outcrop	Mixed carbonate & clastic		412-9290 m	Inactive
Blaekkan, 2016	43	Physical analogue			0.1-2 m	N/a
Bramham et al., 2021	768	Satellite imagery & topo data	Volcanic	Yes	16-2009 m	Active
Cartwright et al., 1995	91	Outcrop	Sedimentary & evaporites	No	280 m	Active
Crider & Pollard, 1998	2	Outcrop	Volcanic		1800-2200 m	Active
Dawers et al., 1993	15	Outcrop	Clastic w/ volcanic		20-2200 m	Active
Delokgos et al., 2017	16	Outcrop	Fine-grained carbonate		79-772 m	Active
Densmore et al., 2004	9	Outcrop	Clastic w/ volcanic		32,421-344,800 m	Active
Duffy et al., 2017	3	3D seismic	Mixed carbonate & clastic	Yes	4250-6738 m	Inactive
Ellis & Barnes, 2015	4	Outcrop	Carbonate		15,700-55,000 m	Varies
Ellis & Barnes, 2015	6	Outcrop	Clastic w/ volcanic		23,200-182,600 m	Active
Ellis & Barnes, 2015	1	Outcrop	Mixed carbonate & clastic		58,200 m	Active
Ellis & Barnes, 2015	3	Outcrop	Metamorphic, igneous & sedimentary		55,400-99,000 m	Varies
Finch & Gawthorpe, 2017	10	Numerical model			55,400-99,000 m	N/a
Gauthier & Lake, 1993	380	3D seismic	Clastic	Yes	71-1904 m	
Ghalayini et al., 2017	82	3D seismic	Fine-grained carbonate	No	5580-63,530 m	Inactive
Gillepsie et al., 1992	54	2D seismic	Fine-grained clastic		0.05-2.6 m	Inactive
Gross et al., 1997	121	Outcrop	Fine-grained clastic			Inactive
Gudmundsson, 2004	24	Outcrop	Volcanic	No	629-8982 m	Active
Hollinsworth et al., 2019	1	Outcrop & seismic	Metamorphic	Yes	15,440 m	Active
Hus et al., 2005	25	Physical analogue			0.14-0.4 m	N/a
Jackson & Rotevatn, 2013	4	3D seismic	Sedimentary & evaporites	Yes	5000-12,300 m	Active
Jackson et al., 2017	1	3D seismic	Sedimentary & evaporites		1950 m	Inactive
Karp et al., 2012	2	2D seismic	Metamorphic & clastic	Yes	23,873-27,140 m	Active
Khalil & McClay, 2017	3	3D seismic	Sedimentary & evaporites	Yes	12,000-23,000 m	Inactive
Kicono, 2005	1	2D Seismic	Metamorphic		6199 m	Active
Lamarche et al., 2005	3	2D Seismic	Clastic		225-436 m	Active
Lathrop et al., 2021	7	3D seismic	Mixed carbonate & clastic		8800-42,100 m	Inactive

Source	Number of Faults	Data Type	Dominant Lithology	Reactivated	Size Range (length in m)	Active/inactive
Marrett & Allmendinger, 1991	133	GPS & 2D seismic	Sedimentary & evaporites		1957-34,464 m	Inactive
McClymont et al., 2009	2	2D & 3D Seismic	Volcanic & clastic		30-47 m	Active
McGlue et al., 2006	2	2D Seismic	Metamorphic	Yes	17,014-51,000 m	Active
McLeod et al., 2000	32	3D seismic	Clastic		368-111,570 m	Inactive
Meyer et al., 2002	84	3D seismic	Mixed carbonate & clastic		690-8592 m	Varies
Morley et al., 2007	4	2D seismic	Clastic	Yes	6075-37,529 m	Inactive
Morley 2017	78	3D seismic	Mixed carbonate & clastic	Yes	3028-82,704 m	Inactive
Morley 2017	14	3D seismic	Fine-grained clastic	Yes	20,900-123,400 m	Inactive
Morley 2017	33	Outcrop	Sedimentary & evaporites	Yes	231-100,838 m	Inactive
Morley 2017	7	3D seismic	Clastic	Yes	30,845-91,792 m	Inactive
Muraoka & Kamata, 1983	14	Outcrop	Volcanic		0.6-2.5 m	Active
Nicol et al., 1996	112	3D seismic	Mixed carbonate & clastic	No	174-8926 m	Inactive
Nicol et al., 2005	1	2D seismic	Clastic	Yes	70,000 m	Active
Nicol et al., 2010	29	Aerial photography	Clastic w/ volcanic		2271-70,742 m	Active
Nicol et al., 2020	122	Outcrop & DEM	Carbonate		32-46,993 m	Active
Opheim & Gudmundsson, 1989	8	Outcrop	Volcanic	No	351-3383 m	Active
Pan et al., 2021	147	3D seismic	Mixed carbonate & clastic		307-18,182 m	Inactive
Peacock & Sanderson, 1991	6	Outcrop	Carbonate	No	7.2-226 m	Inactive
Poulimenos, 2000	45	Outcrop	Mixed carbonate & clastic		1383-17,633 m	Active
Reeve et al., 2015	3	3D seismic	Clastic		1874-17,299 m	Inactive
Reilly et al., 2017	75	2D & 3D seismic	Clastic		1207-91,281 m	Inactive
Rippon, 1985	36	Outcrop	Clastic		200-4600 m	Inactive
Robert & Michetti, 2004	6	Outcrop	Carbonate		17,720-29,800 m	Active
Roche et al., 2017	5	Outcrop	Carbonate		0.26-3.7 m	Inactive
Schlishe et al., 1996	116	Outcrop	Fine-grained clastic		0.02-1.0 m	Inactive
Sieburg et al., 2020	432	Lidar	Volcanic		25-9707 m	Active
Solvía et al., 2005	36	Outcrop	Carbonate		0.3-50 m	Inactive
Solvía et al., 2006	50	Outcrop	Mixed carbonate & clastic		19-51 m	
Solvía et al., 2008	2	Outcrop	Carbonate		0.2-0.5 m	Inactive?

Source	Number of Faults	Data Type	Dominant Lithology	Reactivated	Size Range (length in m)	Active/inactive
Solvia et al., 2008	2	Outcrop	Mixed carbonate & clastic		0.8-1 m	Inactive
Solvia & Schultz, 2008	25	Outcrop			0.4-5 m	Active
Solvia & Schultz, 2008	34	2D seismic	Volcanic	No	66-53,700 m	Active
Shunshan et al., 2011	17	3D seismic	Sedimentary & evaporites	Yes	477-3298 m	Inactive
Torabi et al., 2019	21	3D seismic	Sedimentary & evaporites		238-23,255 m	Inactive
Tschopp 1967	1	2D seismic	Sedimentary & evaporites		45,000 m	Inactive
Tvedt et al., 2016	3	3D seismic	Sedimentary & evaporites	Yes	12,208-16,000 m	Inactive
Vétel et al., 2016	28	Satellite imagery & topo data	Volcanic	Yes	8772-29,467 m	Active
Villemin et al., 1995	26	Outcrop	Clastic		350-27,586 m	Inactive
Walsh & Watterson, 1988	32	Outcrop	Clastic		451-4985 m	Inactive
Walsh et al., 2002	22	Outcrop	Mixed carbonate & clastic	Yes	1200-18900 m	Active
Watterson, 1986	7				4799-53,645 m	
Wedmore et al., 2020	6	Outcrop	Metamorphic	Yes	13,000-85,000 m	Active
Whipp et al., 2014	176	3D seismic	Mixed carbonate & clastic		892-54,227 m	Inactive
Wibberly et al., 1999	28	Outcrop	Unlithified sand		0.17-10 m	Inactive
Wilkins & Gross, 2002	41	Outcrop	Clastic		0.9-490 m	Inactive
Willemse, 1997	7	Outcrop	Clastic w/ volcanic		47-151 m	Active
Williams et al., 2021	1	Outcrop	Metamorphic	Yes	130,000 m	Active
Worthington & Walsh, 2017	11	3D seismic	Fine-grained clastic	Yes	554-10,000 m	Inactive
Yielding et al., 1996	114	3D seismic	Fine-grained clastic		253-24,574 m	Inactive
Young et al., 2001	2	2D seismic	Clastic	Yes	11,00-12,000 m	Inactive
Zygouri et al., 2008	93	2D seismic & outcrop	Mixed carbonate & clastic		1764-15,147 m	Active

Table 3. 2. Table showing all of the sources used in our global database, along with the number of faults from each source, type of data, host rock lithology, if the fault was reactivated or not, the range of fault sizes, and if the fault was active or inactive. Not all information was available for every source.

When displacement and length were not explicitly stated in the original sources, we used data acquisition software (Quintessa Graph Grabber; <https://www.quintessa.org/software/downloads-and-demos/graph-grabber-2.0.2>) to pick the displacement and length from graphs. This yields a certain level of error, especially when taking values from a graph in log-log space, because: (i) several overlapping data points may only yield one datapoint; and (ii) there is some minor imprecision on where the extracted data lie on the X (length) and Y (displacement) axis, which in a log-log plot could be moderately significant (see Figure 3.1).

In order to be included, faults had to be normal (i.e., extensional) faults dominated by dip-slip kinematics; reverse and strike-slip faults were not included. Fault length is defined as ‘the longest horizontal or sub-horizontal dimension along the fault plane, perpendicular to slip direction (Watterson, 1988; Kim & Sanderson, 2005). Fault displacement describes the movement between two fault blocks, calculated by measuring an offset marker bed separated by a fault (Walsh & Watterson, 1988; Xu et al., 2006). Displacement should be measured at its maximum point on the fault. If throw was listed in the original source, it was converted to displacement using the listed fault dip, or an average 55 degrees when fault dip was not explicitly stated. All data are from geologic faults, not data derived from individual earthquakes. Faults have been sorted and analyzed by size. We use length as a measure of fault size, defining three classes: *small* (>1 m), *medium* (1 m-1 km), and *large* (>1 km) (see also Torabi & Berg, 2011).

Since host rock lithology might influence scaling laws, we sorted D/L data into the following groupings: clastic (fine-grained sand and coarser), fine-grained clastic (siltstone and finer), carbonate (specifically a carbonate ‘coarser’ than lime-mud), fine-grained carbonate (e.g., lime mud), mixed carbonate-clastic, evaporite-bearing sedimentary rocks, volcanic, volcanic with clastic, and unlithified sand. Information on host rock lithology could not be found for every fault, and it is only included in the database when explicitly listed by the author or found in another source documenting the same basin. Faults often offset a variety of host rock lithologies, especially for large faults, but they were categorized by the dominant lithology (i.e., over c. 50%). ‘Carbonate’ host rocks are those with >50% carbonate material that is coarser than lime-mud. Faults with host rocks classified as ‘clastic sedimentary’ have host rocks whose lithologies are >50% clastic sedimentary rock, with sand-sized or coarser grains. Faults with host rocks classified as sedimentary with evaporites have host rocks whose lithologies are sedimentary rocks in areas with evaporites; not every fault is necessarily

physically linked to an evaporite detachment. Faults with host rocks classified as ‘fine-grained clastic’ have host rocks whose lithologies are over 50% clastic sedimentary rock with silt-sized or smaller grain sizes. Faults in rocks classified as ‘fine-grained carbonate’ have host rocks whose lithologies are over 50% carbonate rock with fine-grained lithologies, such as lime-muds. Faults in rocks classified as ‘mixed carbonate and clastic’ have host rocks whose lithologies are roughly 50/50 clastic and carbonate. Faults with host rocks classified as ‘unlithified’ were formed in unlithified sediment at the time of active faulting. Faults with host rocks classified as ‘volcanic’ have volcanic host rocks. Faults in rocks classified as ‘sedimentary with volcanics’ have both sedimentary and volcanic host rocks. Faults in metamorphic host rocks were included in the overall dataset, however, there were not enough of them to be statistically significant, so they are not separated in their own sub-group. To compare the relationship between D/L to lithology and Young’s Modulus, we compiled a list of known Young Modulus for different lithologies from published sources to find a range of possible values and average value for each lithology (data can be downloaded at: https://figshare.com/articles/dataset/Young_s_Modulus/17087342).

Faults were also classified based on tectonic history to see how end-member tectonic histories might affect their length and displacement. More specifically, we categorized them as *reactivated* and *no pre-existing structures*; the former are from areas where faults clearly reactivated structures that, after a period of quiescence, became active again. These faults may have formed in response to the reactivation of structures that previously experienced extensional, compressive, strike-slip deformation, or a combination of these, before being reactivated as normal faults. Faults categorized as having no pre-existing structures are from areas thought to have not experienced significant earlier deformation. Information on tectonic history is not always available, so not every fault is categorized this way.

Faults were classified as *active* and *inactive*; this allowed us to assess whether active faults show different length and displacement relationships compared to inactive (dead) faults. Faults categorised as active are from study areas where faults are currently active in tectonically deforming regions, although every fault might not necessarily be active. Faults categorised as inactive are from areas that are not tectonically active, i.e., inactive rifts now buried and imaged in seismic reflection data or in outcrop. This information is not available for every fault in the database, so not every fault is included in this categorization. Care must be taken with these data because it is possible for an active fault to be more to have been

active for a long period of time and thus be over-displaced compared to an inactive fault, if the latter became inactive prematurely due to the removal of the driving stress.

Care must be taken when evaluating how these factors affect D/L scaling relationships as it might be difficult or impossible to isolate the role of each. For example, if a large fault is newly active, has a volcanic host rock, and formed due to reactivation of a pre-existing structure, it may be difficult or impossible to determine which factors has the most influence on its D/L ratio.

For each subcategory, we present the data in four ways: (i) in log-log space – even though data can visually ‘hide’ in log-log space (Rotevatn et al., 2019), they allowed us to view all data in one plot where all orders of magnitude can be seen together (ii) in non-log-log space, with data shown all together in one graph spanning all orders of magnitude – this allowed us to show overall D/L average trendlines, even though smaller faults cannot be visualised, (iii) non-log-log space, broken up by order-of-magnitude so that all of the data can be seen more clearly, (iv) and in a probability density plot. Probability density plots calculated the probability density of D/L values in the each of the different aforementioned categories. We used a kernel density estimation (KDE), which is a non-parametric method of estimating the probability density of a function of a random value, in this case D/L. The height of each plot (y-axis) corresponds to the probability density of the data at a given value of D/L (x-axis). The peaks of the density plot are at the D/L values with the highest probability. A log-log linear model (linear regression) was conducted to calculate a scaling law relationship of the entire dataset, as well as each sub-category (i.e., fault size, tectonic history, fault maturity, host rock lithology). Power law relationships were used because that is the standard in the literature when relating fault displacement and length, and because it tended to fit the data best. When describing faults throughout the paper, we refer to faults as over-displaced if $D/L > 0.1$ and under-displaced if $D/L < 0.01$, as most faults lie within those parameters, so it is a logical cut-off when speaking generally.

3.5. Results

In our database, faults are 0.01-344,800 m long (Figure 3.2) and have a power-law trendline of $D_{max} = 0.03L^{0.92 \pm 0.01}$ (i.e., $n = 0.92$ and $c = 0.03$; Table 3.3). Our value of n is thus broadly consistent with the estimate of Cowie & Scholz (1992) and others ($n = 1$) for normal faults. However, there is a large amount of scatter in the data, with displacements for a given fault length ranging across 1.5-4 orders of magnitude (Figure 3.2). In this section we investigate how normal fault length and displacement relationships are affected by fault size, maturity,

tectonic history, and host rock lithology. We also look at examples of how D and L (and their related scaling relationship) change through time and plot it against the D/L global database.

Category	Sub-category	Number	Number of Sources	Power-law Equation	R-squared
All faults	—	4035	65	$D_{max}=0.03L^{0.92\pm0.01}$	0.85
Size	Small (0-10 m)	395		$D_{max}=0.04L^{0.97\pm0.02}$	0.81
Size	Medium (10-10,000 m)	3246	48	$D_{max}=0.03L^{0.94\pm0.01}$	0.63
Size	Large (10,000+ m)	394	35	$D_{max}=0.007L^{1.3\pm0.01}$	0.24
Maturity	Active	1959	27	$D_{max}=0.02L^{0.92\pm0.01}$	0.74
Maturity	Inactive	2059	38	$D_{max}=0.05L^{0.93\pm0.01}$	0.92
Tectonic Setting	Reactivated	1620	8	$D_{max}=0.03L^{0.96\pm0.01}$	0.74
Tectonic Setting	Not Reactivated	265	15	$D_{max}=0.04L^{0.87\pm0.05}$	0.56
Lithology	Clastic Sedimentary	644	13	$D_{max}=0.11L^{0.84\pm0.02}$	0.71
Lithology	Clastic w/ Evaporites	344	12	$D_{max}=0.02L^{1.04\pm0.02}$	0.85
Lithology	Carbonates	181	7	$D_{max}=0.04L^{0.79\pm0.03}$	0.97
Lithology	FG Carbonates	220	2	$D_{max}=0.06L^{0.93\pm0.01}$	0.20
Lithology	FG Clastic	324	6	$D_{max}=0.03L^{0.94\pm0.01}$	0.95
Lithology	Mixed Carbonate & Clastic	124	14	$D_{max}=0.04L^{0.92\pm0.02}$	0.78
Lithology	Unlithified Sand	28	1	$D_{max}=0.01L^{0.72\pm0.02}$	0.31
Lithology	Sedimentary w/ Volcanics	130	7	$D_{max}=0.007L^{1.0\pm0.05}$	0.77
Lithology	Volcanics	1317	8	$D_{max}=0.04L^{0.82\pm0.02}$	0.63

Table 3. 3. Table listing of all the results for each of the studied categories, including number of faults, number of sources, power-law equation that can be used to estimate fault length or displacement, and the R-squared for that equation.

3.5.1. Size

A total of 395 small faults were included from 11 different sources, 3246 medium faults were included from 48 sources, and 394 large faults were included from 35 sources (seen in Figure 3.2). The dataset includes small faults from areas such as the High Atlas, Morocco and the Dan River Rift, USA, medium faults from areas such as the Pyrenees and Utah, USA, and large faults were included from areas such as the Levant Basin, Lebanon, and the North Sea. Our data show that small faults have a higher D/L ratio, with a power-law trendline indicating $D_{max}=0.04L^{0.97\pm0.02}$, $n=0.97\pm0.02$ and $c=0.04$; Table 3.3). Medium and large faults have similar power-law trendlines of $D_{max}=0.03L^{0.94\pm0.01}$ and $D_{max}=0.007L^{1.3\pm0.01}$, respectively, $n=0.94\pm0.01$ and $c=0.03$ for medium faults and $n=1.3\pm0.11$ and $c=0.007$ for large faults (Table 3.3). The values of n of small are within the same confidence interval (Table 3.3).

There is a significant amount of spread in the relationship between D and L, especially for larger faults (Figure 3.2). Large faults show the greatest amount of scatter, with displacements varying across 3-4 orders of magnitude (Figure 3.2). For example, faults that are 10,000 m (± 200 m) long have displacements ranging from 4 m to 999 m and a standard

deviation of 303 m. In contrast, medium faults only vary by 1-2 orders of magnitude (Figure 3.2). For example, faults that are 50 (± 1 m) long have displacements ranging between 0.3 m and 7 m and a standard deviation of 1.6 m. Small faults have the least amount of scatter, with displacements that vary by only 1-1.5 orders of magnitude (Figure 3.2). For example, faults that are 0.1 ± 0.05 m long have displacements between 0.002 m and 0.01 m and a standard deviation of 0.003 m.

Medium and large faults plot similarly in a probability density plots (Figure 3.3); there is a $\sim 27\%$ probability and $\sim 20\%$ probability, respectively, of a D/L value of ~ 0.02 , i.e., medium to large faults in the dataset are most likely to have a displacement that is $\sim 2\%$ of fault length. More small faults in the dataset were over-displaced compared to medium and large faults: small faults have a $\sim 16\%$ probability of a D/L value of ~ 0.035 , i.e., small faults in this dataset are most likely to have a displacement that is $\sim 3.5\%$ of fault length (Figure 3.3). The shape of the distribution of small faults is relatively longer tailed, which means that there are more small faults with a higher D/L value than medium or large faults.

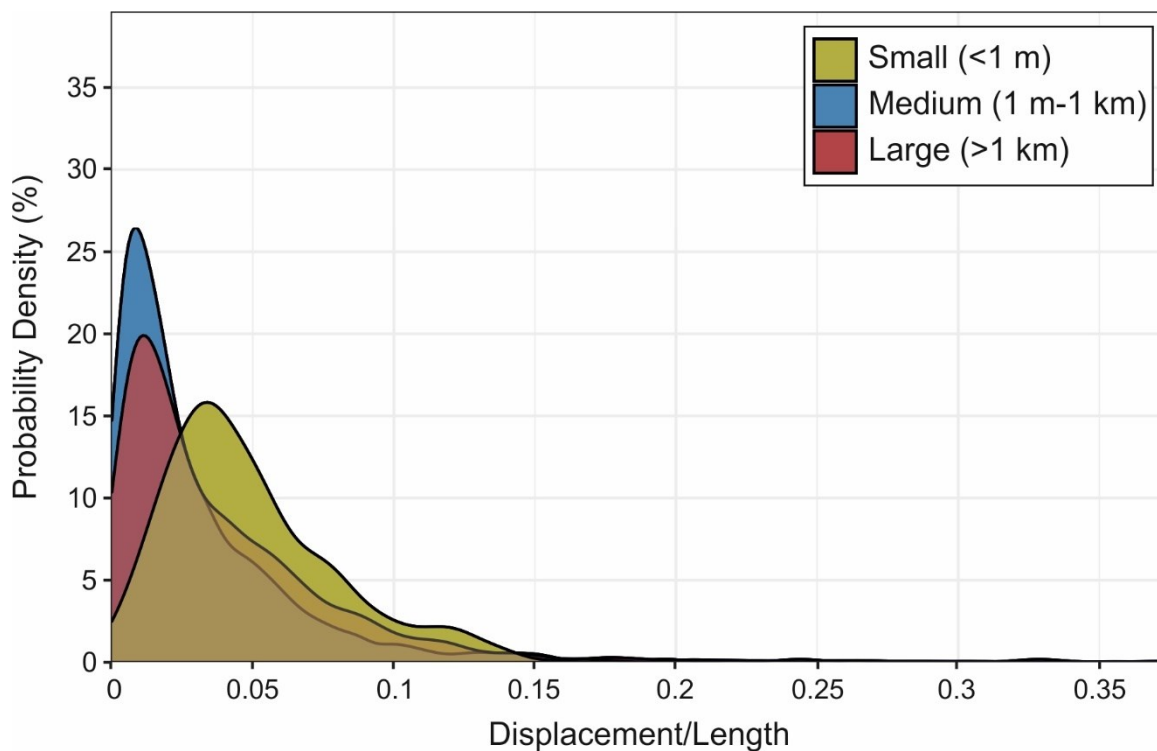


Figure 3. 3. Density estimates of the D/L value of small, medium, and large faults in our dataset. Peaks in the density plot are at the D/L values with the highest probability.

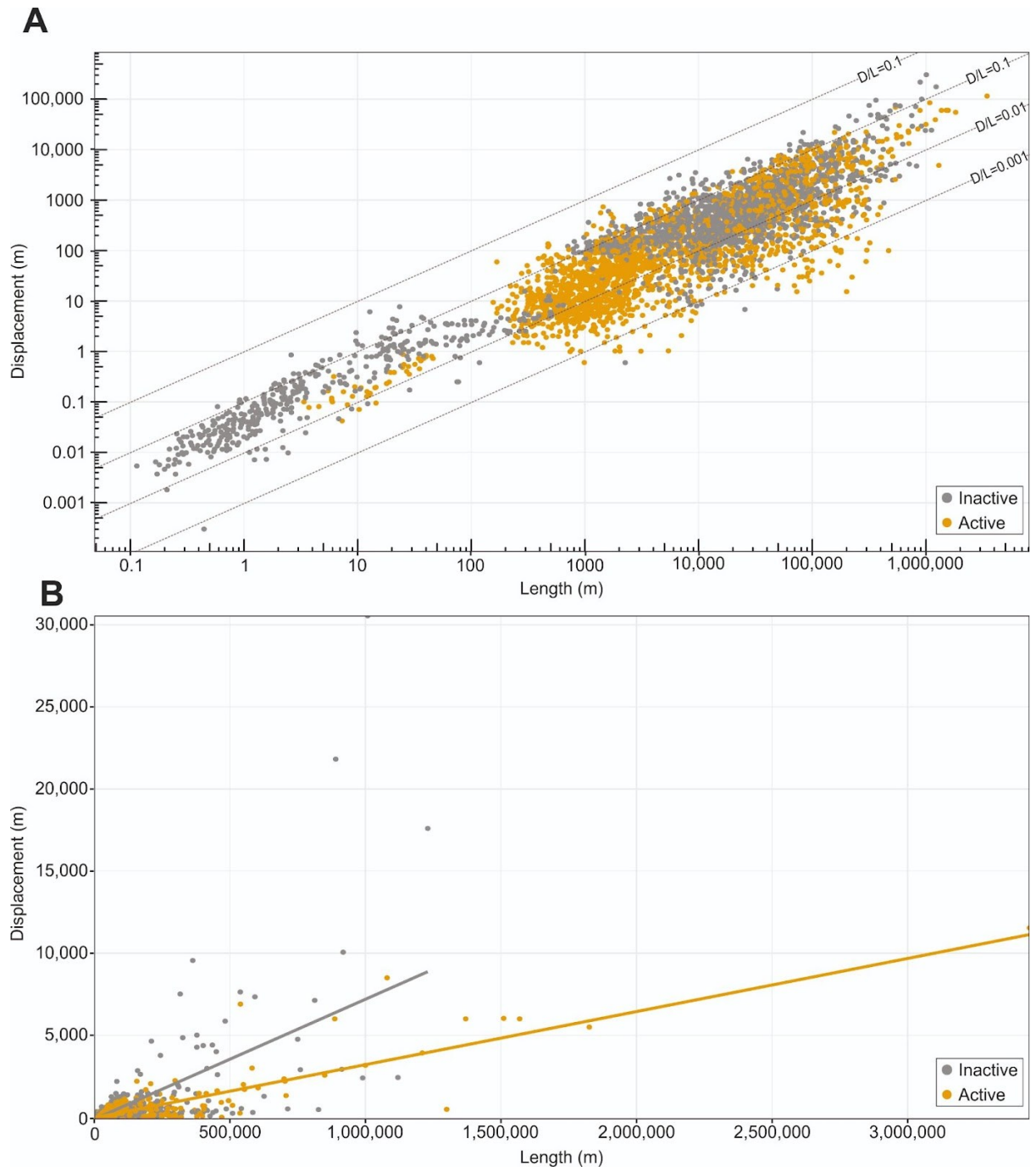


Figure 3. 4. Plots showing fault length vs displacement for active and inactive normal faults in our database. A) Data in log-log space. B) Data not in log-log.

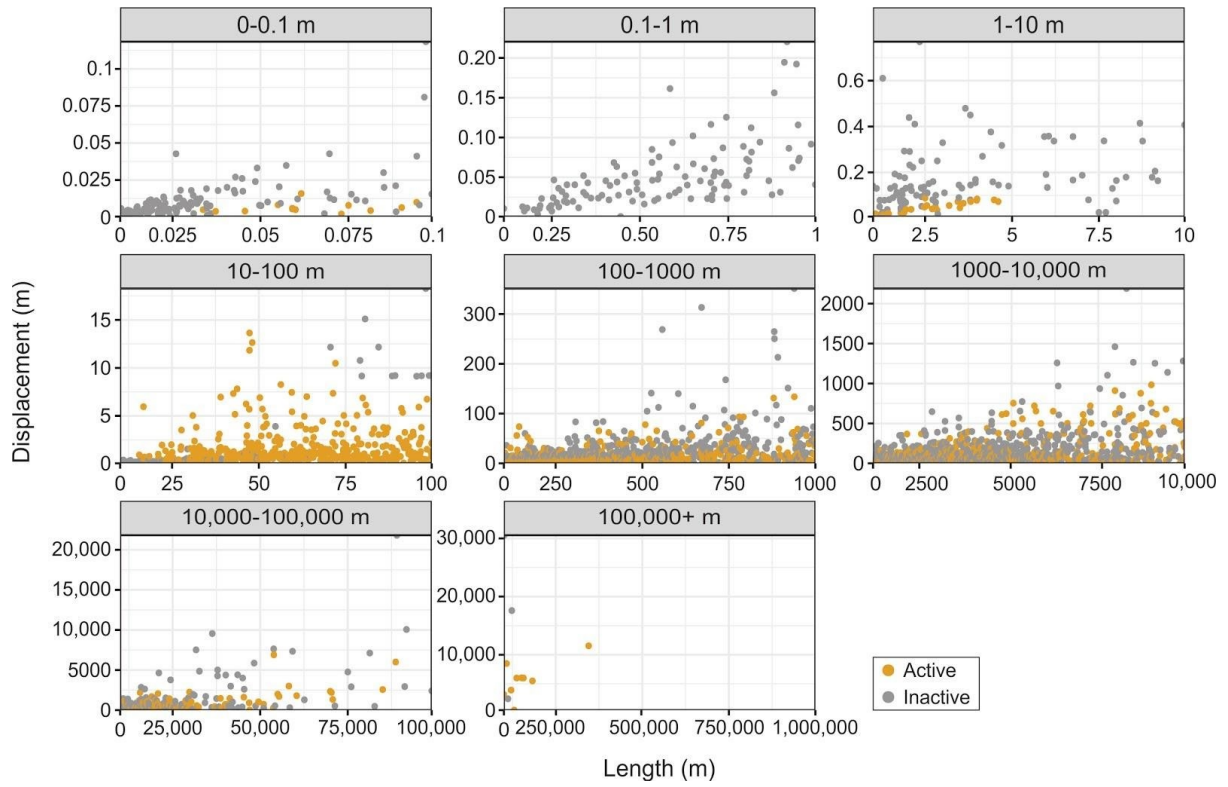


Figure 3. 5. Plots showing active and inactive normal fault D/L data, separated by order of magnitude.

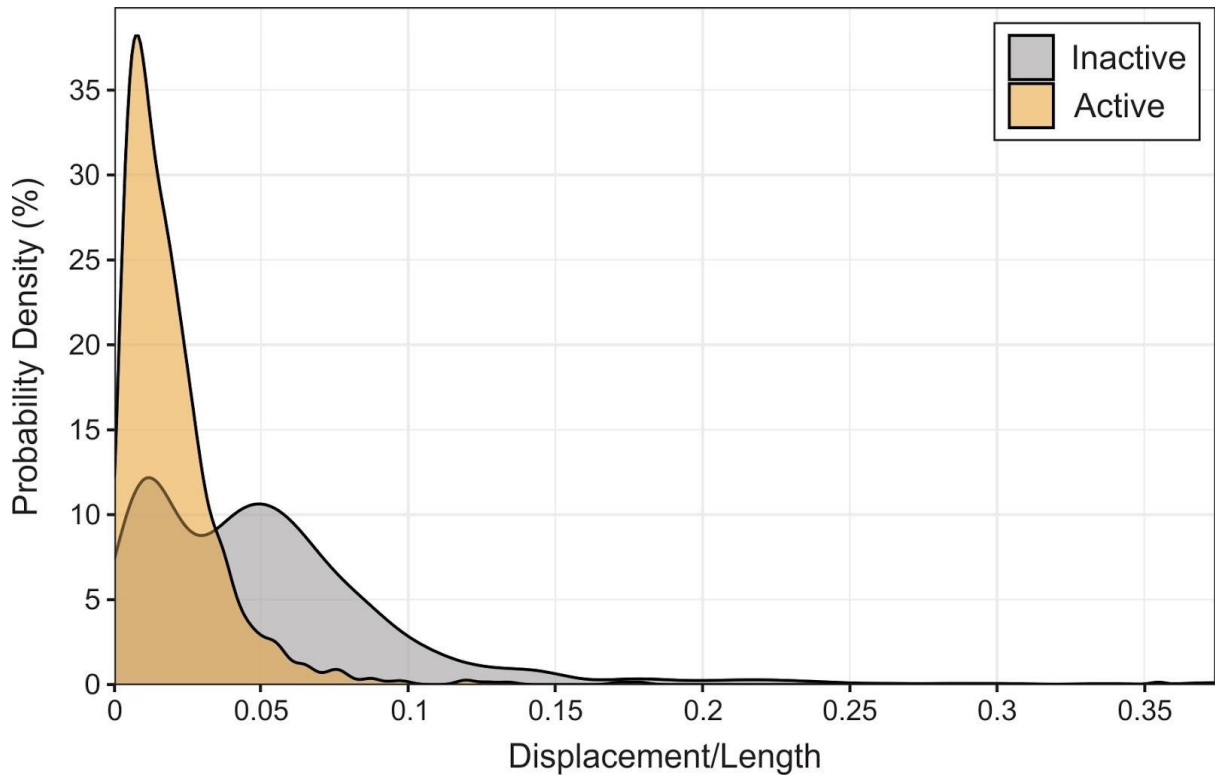


Figure 3. 6. Density estimates of the D/L value of active and inactive faults in our dataset. Peaks in the density plot are at the D/L values with the highest probability.

3.5.2. Maturity

1959 active faults from 27 sources were included, ranging in size from 0.3 m to 345 km in length, with data from areas such as Crete, the Apennines (Italy), and the Turkana Rift (Kenya) (Figures 3.4-5). A total of 2059 inactive faults were included from 38 sources, ranging in size from 0.01 m to 123.4 km in length, with data from areas such as the Exmouth Plateau (Offshore NW Australia), Horda Platform (Offshore Norway), and the Levant Basin (Lebanon) (Figures 3.4 and 3.5).

The active faults have a power-law trendline of $D_{max}=0.02^{0.92\pm 0.01}$, which requires $n=0.92\pm 0.02$ and $c=0.02$, and inactive faults have a trendline of $D_{max}=0.05^{0.93\pm 0.01}$, which requires $n=0.93\pm 0.01$ and $c=0.05$ (Table 3.3). The confidence values of n of inactive and active faults overlap (Table 3.3). Inactive faults have a higher displacement/length ratio than active faults (Figure 3.4b).

According to the probability density plot (Figure 3.6), there is a ~37% probability of active faults having a D/L value of ~0.025, i.e., active faults are most likely to have a displacement that is 2.5% of length. Inactive faults have two probability peaks; there is ~12% probability of a D/L value of 0.025 and ~11% probability of a D/L value of 0.05, i.e., inactive faults are most like to have a displacement that is ~2.5% or ~5% of length. The density plot of inactive faults has a longer tail, which means that higher D/L values are more probable in inactive faults than active faults.

3.5.3. Tectonic history

1620 reactivated faults from eight sources were included, ranging in size from 17 m to 123 km in length, with data from areas such as the Porcupine Basin (Offshore Ireland) and the North Malay Basin (Thailand) (Figures 3.7 and 3.8). 265 faults with no pre-existing structures were taken from 15 sources from areas such as Canyonlands (Utah, USA) and the East Pacific Rise (Figures 3.7 and 3.8). Faults range in size from 0.2 m to 54 km. The reactivated faults have a power-law trendline of $D_{max}=0.03^{0.96\pm 0.01}$, $n=0.96\pm 0.1$ and $c=0.03$, and the non-reactivated faults have a power-law trendline of $D_{max}=0.04^{0.87\pm 0.05}$, $n=0.87\pm 0.05$ and $c=0.04$ (Table 3.3). Reactivated faults have a higher D/L ratio on average than faults with no pre-existing structures (Figure 3.7b). This is unusual, given several authors have suggested that reactivated faults tend to be under-displaced (Walsh et al., 2002; Vétel et al., 2005). We discuss the possible reasons for this in sub-section 5.3.

In probability density plots (Figure 3.9), reactivated faults and faults with no pre-existing structures plot similarly; for reactivated faults and faults with no pre-existing structures, there

is a ~27% and ~24% probability, respectively, of a D/L value of ~0.025, i.e., both reactivated faults and faults with no pre-existing structures in this dataset are most likely to have a displacement that is ~2.5% of length. The distribution of reactivated faults has a slightly longer tail, which means that there is a slightly higher probability of reactivated faults having a higher D/L value.

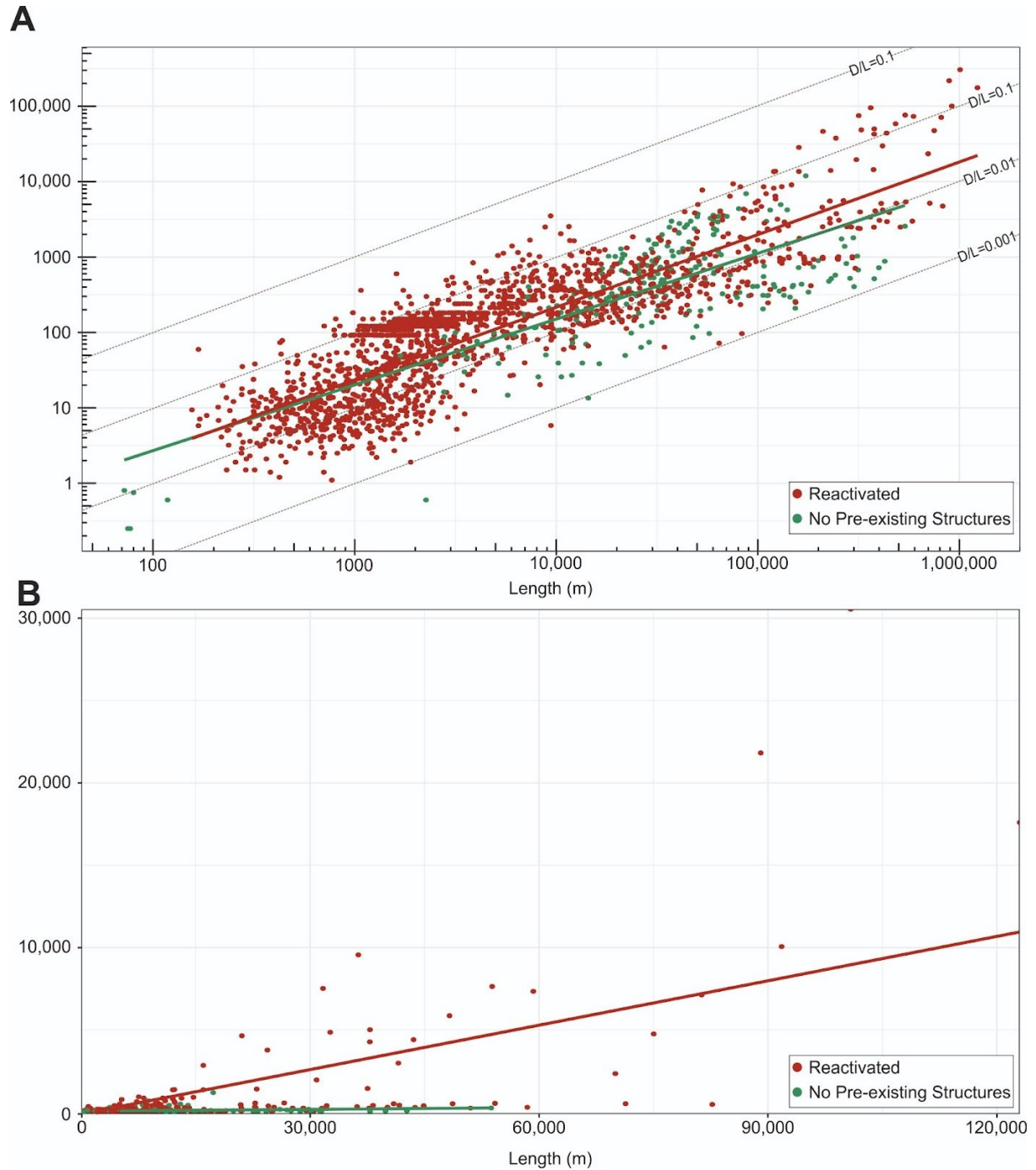


Figure 3. 7. Plots showing fault length vs displacement for reactivated and non-reactivated normal faults in our database. A) Data in log-log space. B) Data not in log-log space.

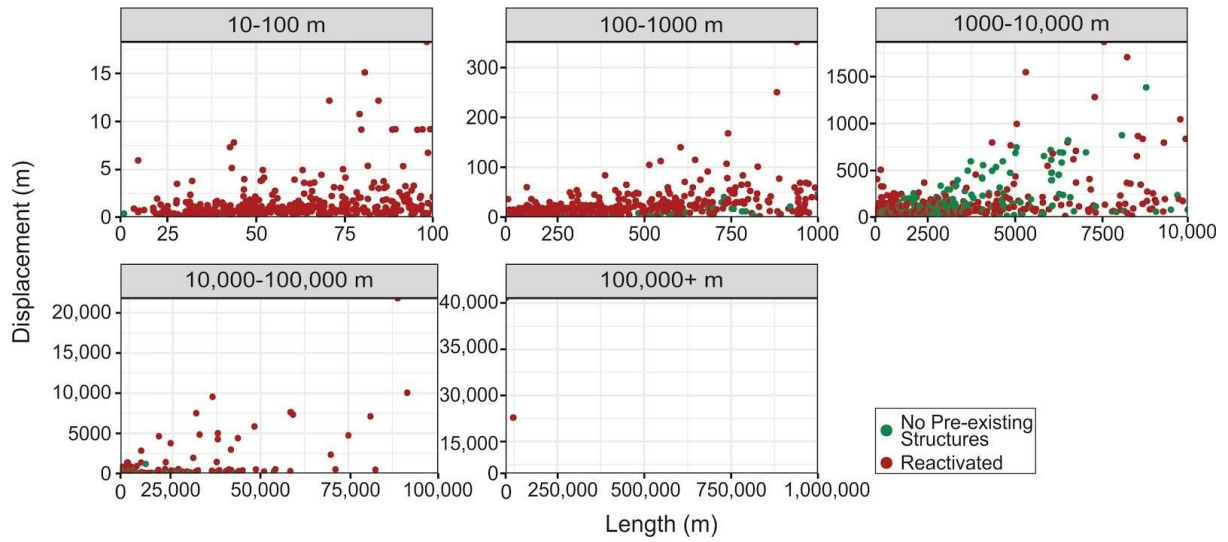


Figure 3. 8. Plots showing normal fault D/L data from reactivated faults and faults with no pre-existing structures, separated by order of magnitude

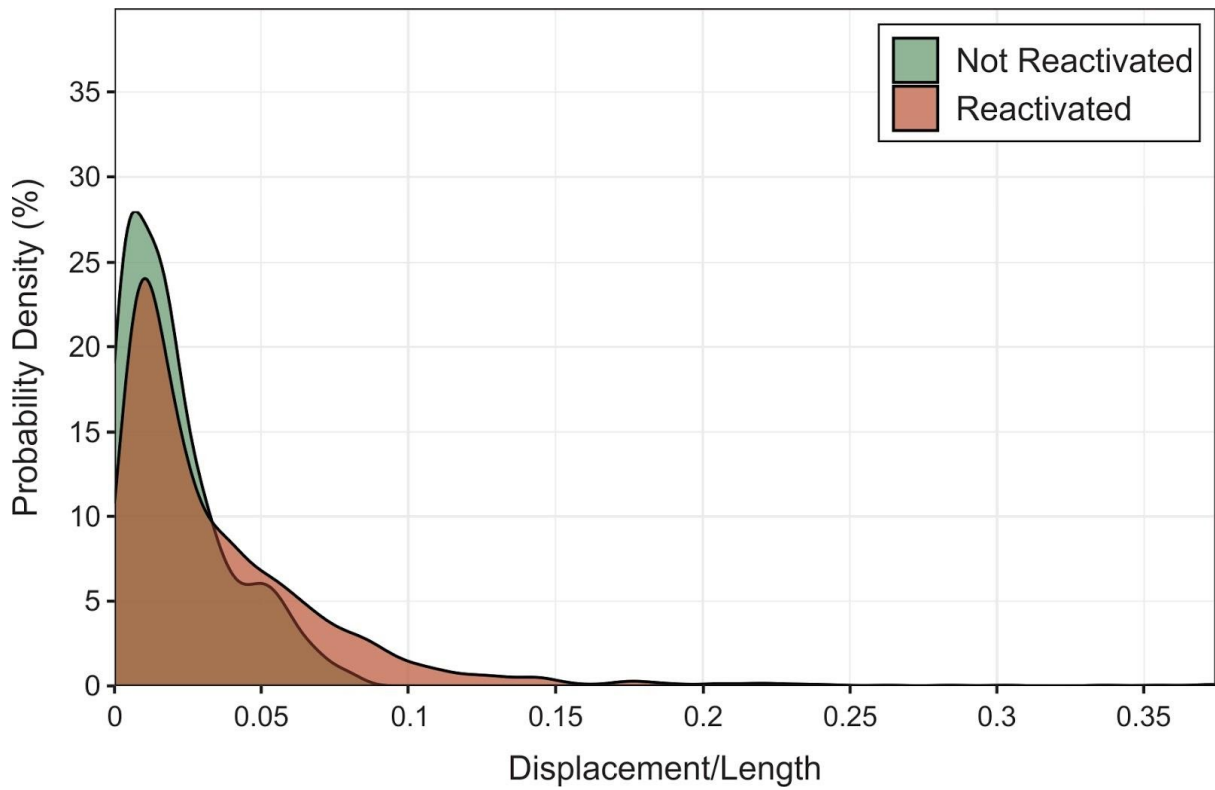


Figure 3. 9. Density estimates of the D/L value of faults that have and have not been reactivated in our dataset. Peaks in the density plot are at the D/L values with the highest probability.

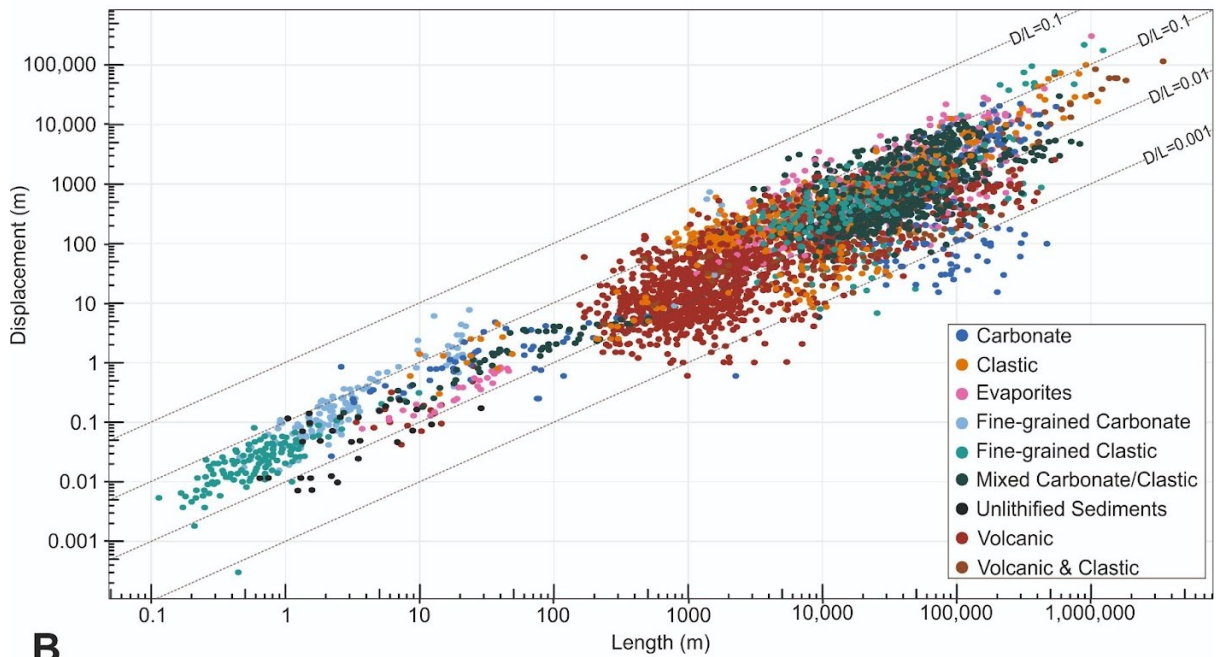
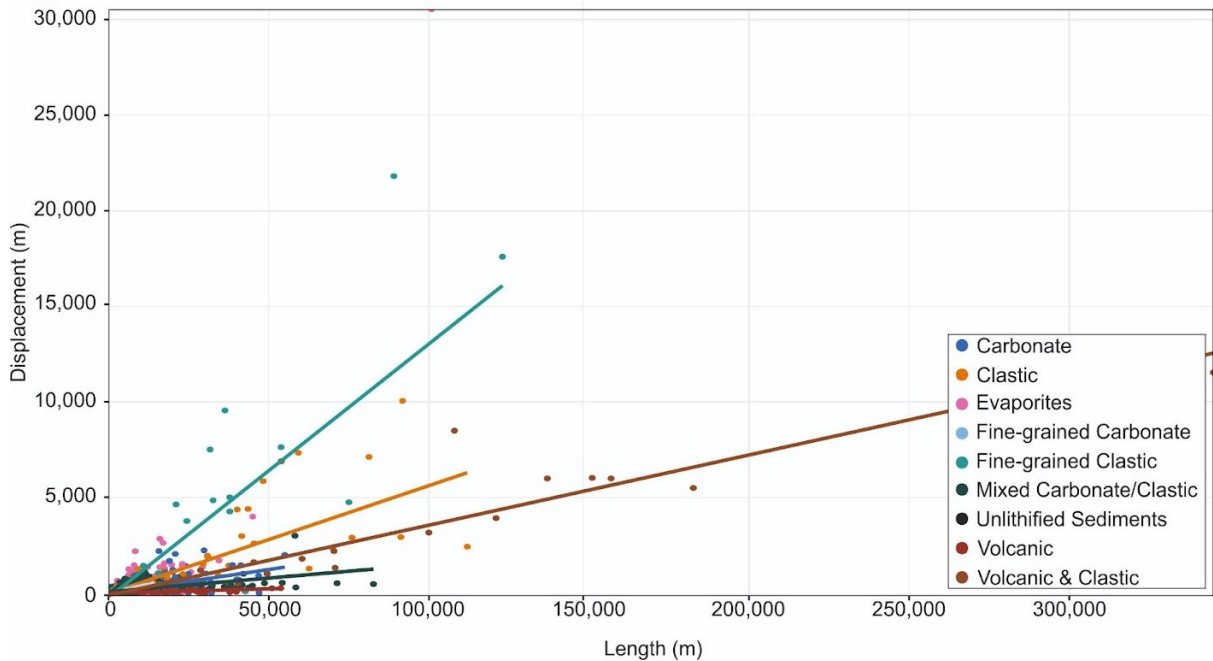
A**B**

Figure 3. 10. Plots showing fault length vs displacement for normal faults with host rocks of different lithologies included in our dataset, including carbonate, clastic, evaporites, fine-grained carbonate, fine-grained clastic, mixed carbonate/clastic, unlithified sediments, volcanic, and volcanic/clastic. A) Data in log-log space. B) Data not in log-log space.

3.5.4. Lithology

A power-law trendline was calculated for each lithology sub-category, with n values ranging from 0.007 to 1.1 and c values from 0.79 to 1 (Table 3.3). There is some variation in trendlines, however, the confidence intervals for n of faults in fine-grained carbonate, fine-grained clastic, mixed carbonate/clastic, and sedimentary with volcanic rocks overlap and other host rocks do not (Table 3.3). Faults with clastic sedimentary and fine-grained clastic

sedimentary host rocks tend to have a higher D/L ratio (i.e., they are over-displaced) compared to the other lithologies (Figures 3.10 and 3.11). Faults with volcanic sedimentary and clastic with volcanic sedimentary host rocks tend to have a lower D/L ratio compared to the other lithologies (Figures 3.10 and 3.11).

According to density plots (Figure 3.12), clastic sedimentary rocks have the highest probability of high D/L values compared to the other lithologies, i.e., there is a ~13% probability of a fault in a clastic host rock having a D/L value of ~0.09, i.e., faults with clastic sedimentary host rocks in this dataset are most likely to have a displacement that is 9% of faults length. Faults with volcanic and clastic with volcanic host rocks have a higher probability of low D/L values than other lithologies; for volcanic and clastic with volcanic host rocks, there is a ~32% and ~35% probability respectively of a D/L value of ~0.01, i.e., faults with volcanic or volcanic/clastic host rocks are most likely to have a displacement that is 1% of length.

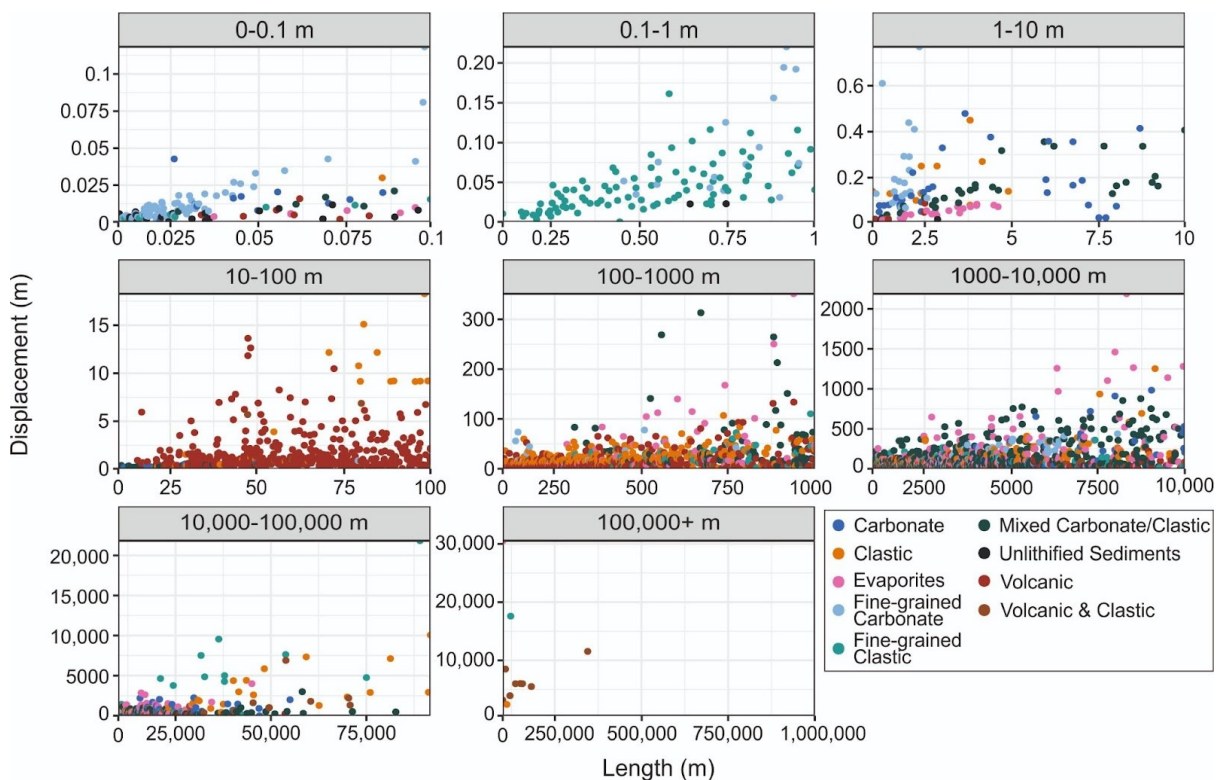


Figure 3. 11. Plots showing normal fault D/L data for normal faults with host rocks of different lithologies included in our dataset, including carbonate, clastic, evaporites, fine-grained carbonate, fine-grained clastic, mixed carbonate/clastic, unlithified sediments, volcanic, and volcanic/clastic, separated by order of magnitude.

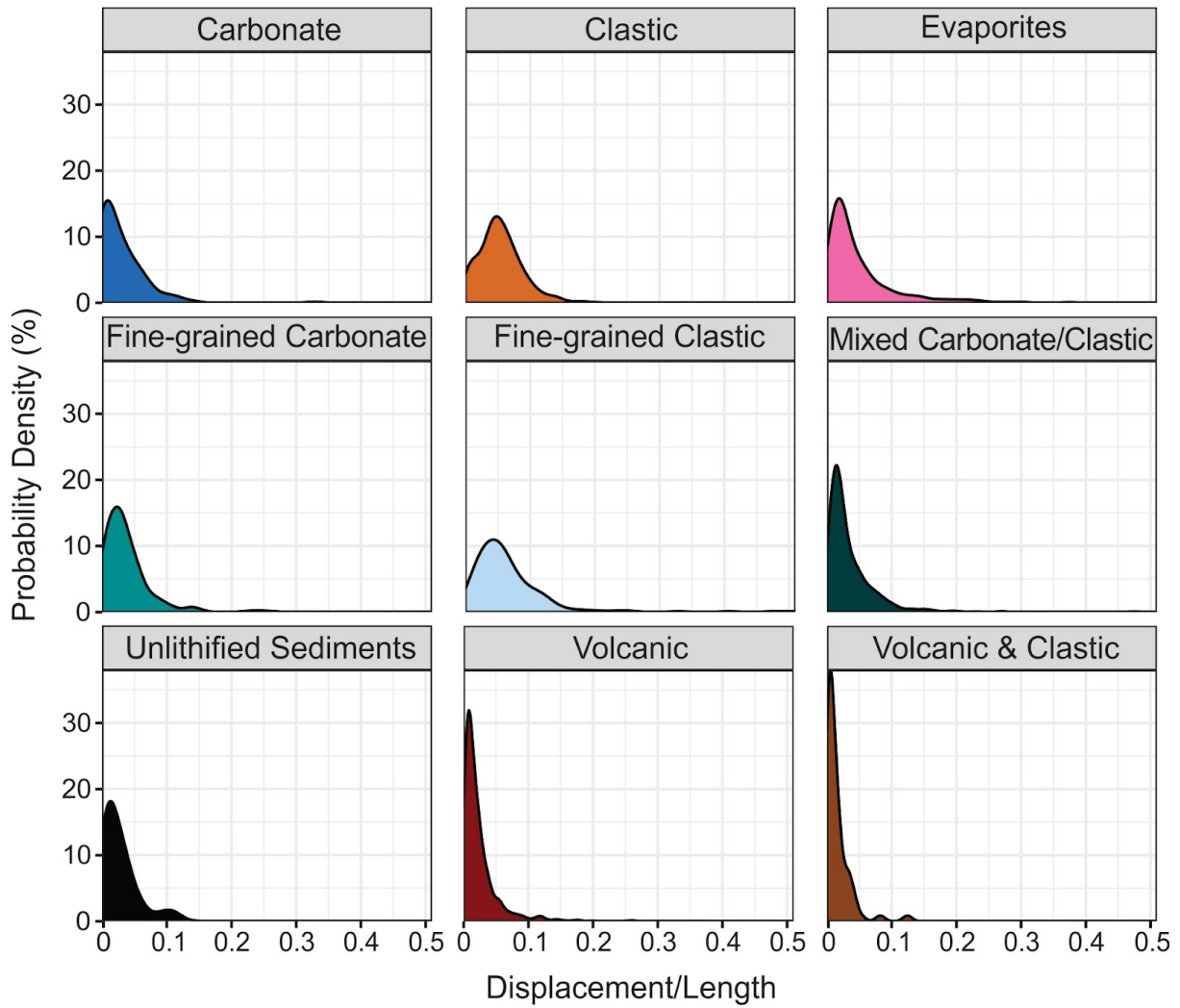


Figure 3. 12. Density estimates of the D/L value of faults with host rocks of different lithologies included in our dataset, including carbonate, clastic, evaporites, fine-grained carbonate, fine-grained clastic, mixed carbonate/clastic, unlithified sediments, volcanic, and volcanic/clastic. Peaks in the density plot are at the D/L values with the highest probability.

3.5.5. D/L through time

37 faults from six different sources were included in a D and L through time dataset (Figure 3.13). 24 natural faults imaged in 3D seismic reflection data were included, with these faults ranging from 1.9-42 km long. Six faults generated in from physical analogue models and three from numerical models were also included. The D/L trajectories of these faults are shown against the global D/L database (Figure 3.13a) and in normalised D-time and L-time plots (Figure 3.13b).

There is a wide range of displacement trajectories in the studied faults for example, in the first 25% of the faults' lives, some faults had only accumulated 6% of total displacement, while others had reached up to 75% of maximum displacement (Figure 13b); on average, faults accumulate displacement at a constant gradient, however on a fault-to-fault basis, there is more variability. 26 of the 37 (70%) faults attain >75% of their maximum length within the

first 25% of their lives, and 35 of 37 (95%) faults reach their lengths within the first half of their lives (Figure 3.13b). Faults then either maintain their maximum length or decrease in active trace length until they become inactive. On average, faults reach their maximum length within the first 30% of their lives and then decrease in length by 5-10%. 23 of the 37 (62%) faults experience late-stage lateral tip retreat, where their tips become inactive in the later stages of the faults' lives.

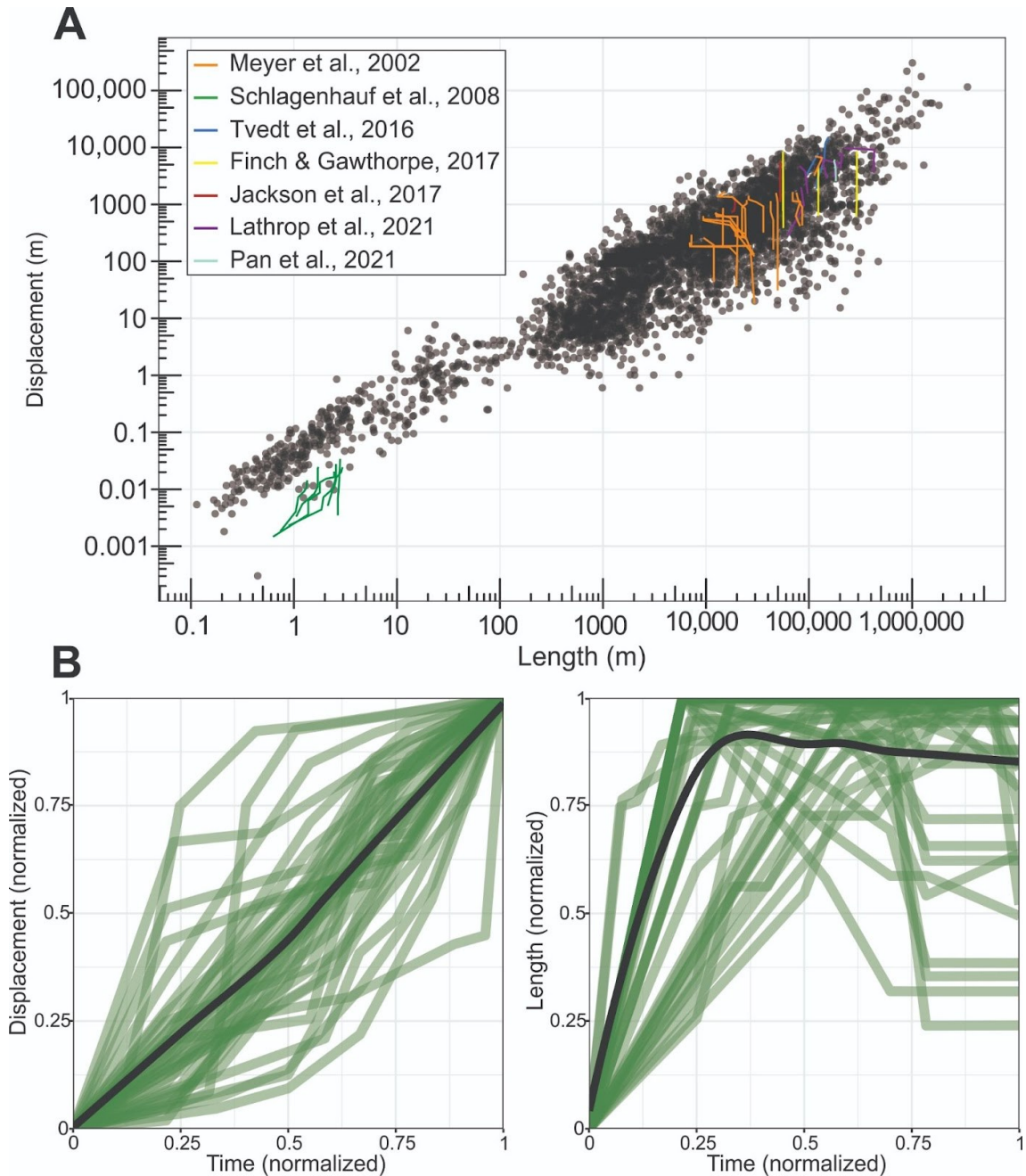


Figure 3. 13. Figures showing fault growth through time (with data extracted from Meyer et al., 2002; Schlagenhauf et al., 2008; Tvedt et al., 2016; Finch & Gawthorpe, 2017; Jackson et al., 2017; Lathrop et al., 2021/Chapter 4; Pan et al., 2021). A) Global D/L dataset (black) for normal faults in log-log space with D/L through time data in colour above it. B) Displacement and length through time, normalised in green, average values in black.

3.6. Discussion

We here summarise some key observations regarding the relationship between D and L and fault size, activity, tectonic history, and lithology, and then use specific, well-constrained case studies to indicate how the various parameters control fault growth and associated scaling relationships. We then discuss D/L changes through time, fault growth models, and the processes that control the upper limits of the D/L scaling relationship.

3.6.1. Size

There is little consensus in the literature about how fault size affects the relationship between D and L. Schlische et al. (1996) found no relationship between fault size and D/L ratio. In contrast, Cowie & Scholz (1992a) found that very large faults (>1 km) were over-displaced compared to smaller faults. Torabi & Berg (2011) showed that small faults (<1 m displacement) and large faults (>1 km displacement) have a higher displacement/length ratio than medium faults (between 1 m and 1 km), suggesting both small and large faults are over-displaced. They explained that the low D/L ratio of medium-sized faults is likely due to faults of this size being in the process of overlapping, interacting, and linking, i.e., they will eventually become larger and accrue more displacement (Torabi & Berg, 2011). However, the low D/L ratio of these faults could be due to sampling biases, i.e., there is a scarcity of published medium-sized faults included in their database.

Our results show that large and medium-sized faults have similar displacement/length ratios, but small faults (<1 m) tend to be relatively over-displaced (Figure 3.3). Assuming a constant-length growth model (e.g., Walsh et al., 2002; Jackson et al., 2017; Rotevatn et al., 2019), faults reach their maximum length quickly, and then accumulate displacement. Medium and large faults are active for a longer period, and under a constant-length model, they are likely to have reached their maximum length and to be in some stage of displacement accrual and thus under-displaced. Under-displaced medium-to-large faults could either be still active and in the displacement accrual stage or became inactive before they reached their maximum displacement potential (e.g., due to rheological changes from mechanical stratigraphy, kinematic interactions between faults, strain partitioning onto more optimally positioned faults). Small faults are active for a shorter period, so faults can lengthen and accumulate a relatively high amount of displacement without and are less likely to become inactive before reaching their maximum possible displacement.

Duration of faulting may also explain scatter in the global D/L plot; the displacement on large faults, which presumably have been active for longer, span up to 4 orders of magnitude,

whereas small faults only span 1-1.5 orders of magnitude (Figure 3.2). Scatter for large faults represents faults that have become inactive prematurely, and the lack of scatter for small faults represents fault growth stages not being seen at a detectable resolution.

3.6.2. Maturity

Fault length and displacement accumulation tend to be strongly partitioned in time (Figure 3.13b) (e.g., Walsh et al., 2002; Tvedt et al., 2016; Rotevatn et al., 2019). Thus, if the maximum displacement had (in the case of an inactive fault) or has (in the case of a still-active fault) been measured part-way through a fault's life rather than at the end, it would plot as under-displaced. When estimating fault scaling, it is important to keep in mind if the faults are active, and if so, how mature they are. However, there is still a huge amount of scatter among both active and inactive faults; the scaling laws between inactive and active faults have n values with overlapping confidence intervals.

Active faults tend to be younger and have been active for less time compared to inactive faults; they are, therefore, comparatively under-displaced. This aligns with our understanding of fault growth under a “constant length” or “hybrid growth” model (Walsh et al., 2002, 2003; Nicol et al., 2005, 2017; Jackson & Rotevatn, 2013; Henstra et al., 2015; Fossen & Rotevatn, 2016; Hemelsdaël & Ford, 2016; Tvedt et al., 2016; Childs et al., 2017b). Under a constant-length or hybrid growth model, faults reach their maximum length in the first 20-30% (or less) of their life. Active faults are likely generally under-displaced because they have reached their maximum length but are still accruing displacement.

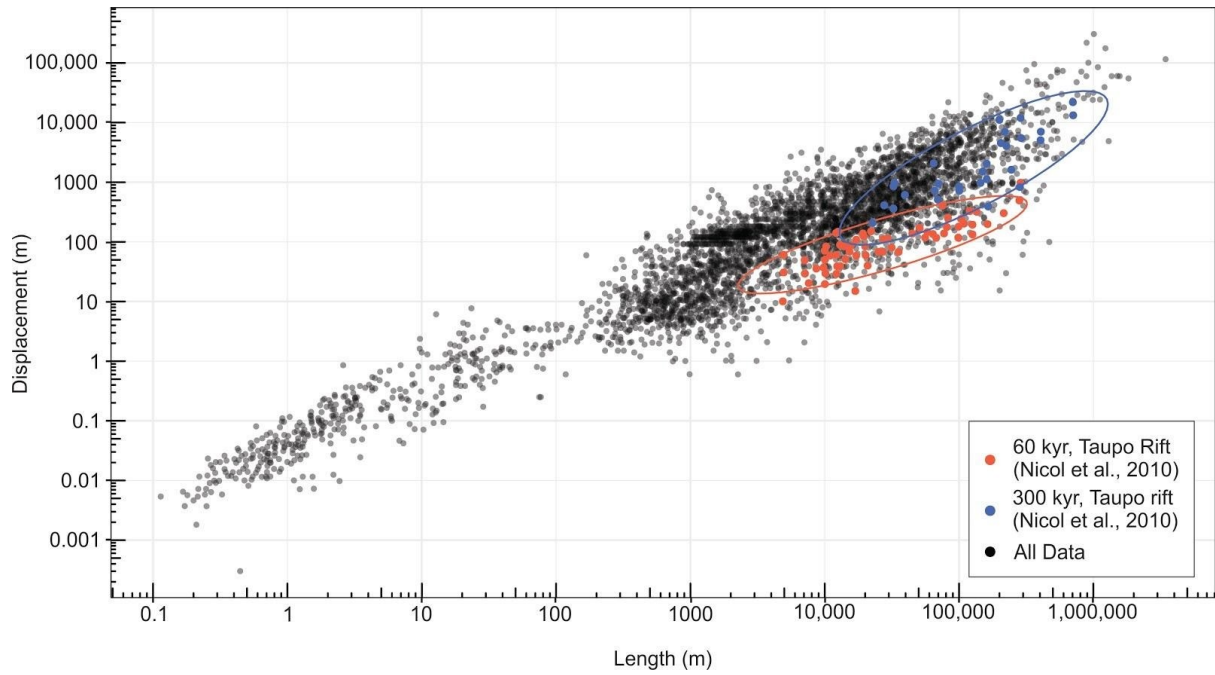


Figure 3. 14. Global D/L dataset (black) for normal faults in log-log space with faults of different maturities highlighted: 60 kyr faults from the Taupo Rift in orange and 300 kyr faults from the Taupo Rift in purple (from Nicol et al., 2010).

One example from the database of under-displaced immature normal faults come from the Taupo rift on the central North Island of New Zealand (Nicol et al., 2010; Figure 3.14). Rifting began 1-2 Ma, with the studied faults having been active for 60 kyr and 300 kyr. The area is tectonically active, and the faults accommodate 15 mm/yr of extension. The older faults, which have been active for 300 kyr, are 2.3 km to 70.7 km long and have displacements ranging between 20.7 m and 2198 m. $D_{\max}/L$ is between 0.002-0.06 (average 0.017) (Nicol et al., 2010; Figure 3.14). In contrast, the younger faults, which have been active for only 60 kyr, are 487 m to 28.7 km long and have displacements ranging between 1 m and 97.9 m. $D_{\max}/L$ between 0.0009-0.01 (average 0.004) (Nicol et al., 2010; Figure 3.14). It is often difficult to deduce whether a fault is under-displaced due to fault maturity or lithology (see section 3); however, in the Taupo case, given that these faults formed in the same host rock, it is likely these still-active faults are under-displaced solely due to fault maturity.

The Taupo Rift faults are under-displaced compared to a set of inactive faults of similar length from the Exmouth Plateau, NW Shelf of Australia (Pan et al., 2021). Faults on the Exmouth Plateau were active from the Early Jurassic-Early Cretaceous (85.5 kyr), are 307 m to 181.2 km long, and have displacements ranging between 18.2 m and 857.6 m (Pan et al., 2021). $D_{\max}/L$ is between 0.006-0.5 (average of 0.06). These faults grew in accordance with

and support the constant-length model, reaching their final length in less than 7.2 myr (8% of their total lifespan) before accruing displacement (Pan et al., 2021). **3.6.3. Tectonic history**

Faults that have been formed due to the reactivation of a pre-existing structure tend to be slightly under-displaced compared to faults in areas that have no reported pre-existing faulting (Figure 3.7b). There is also a slightly higher proportion of under-displaced, non-reactivated faults compared to reactivated faults (Figure 3.9). In previous literature, reactivated faults tended to have a higher displacement to length ratio because the maximum length of the fault is generally established in the first phase of faulting (Vétel et al., 2005; Baudon & Cartwright, 2008). However, we believe that the amount that the reactivation of older structures and thus pre-extensional tectonic history affects D/L ratios is strongly dependent on how long the fault has been active, since newly-formed faults will tend to be under-displaced, according to both the constant-length and hybrid fault growth models (Rotevatn et al., 2019).

One example of reactivated normal faults is from the tectonically active Turkana Rift, Northern Kenya (Vétel et al., 2005; Figure 3.15 in pink). Faults range are 208 m-29.5 km long, have displacements ranging from 82.5 m to 101 m, and have been active for <3 Myr (Vétel et al., 2005). Faults are thought to have reactivated Proterozoic basement faults, or possibly utilise basement metamorphic foliation, and the area currently extended with a strain rate of ~0.1 mm/yr (Vétel et al., 2005). Fault arrays were able to reach lengths of ~40 km in a very short period of time, despite these relatively low strain rates, likely due to them exploiting and activating pre-existing weaknesses (Vétel et al., 2005). The average D/L ratio is 0.007, (displacement is 0.7% of length). These faults are thus under-displaced, which is likely due to them having lengthened rapidly by exploiting intra-basement weaknesses; these faults are thus likely still at the beginning of their displacement accumulation stage.

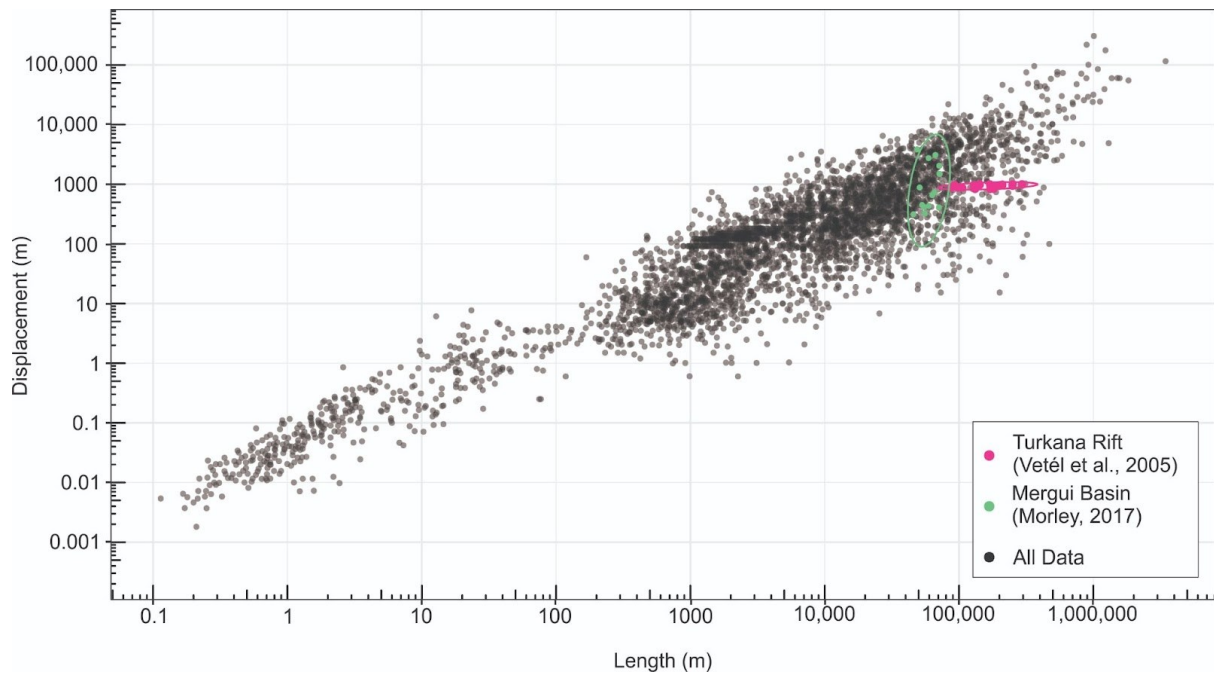


Figure 3. 15. Global D/L dataset (black) for normal faults in log-log space with reactivated faults highlighted: faults from the Turkana Rift in pink (from Vétel et al., 2005) and the Mergui Basin in green (from Morley, 2017).

An example of more mature, but still-active reactivated faults in the database are from the Mergui Basin, Thailand (Morley, 2017; Figure 3.15, in green), which unlike the Turkana Rift faults (Vétel et al., 2005) are relatively over-displaced. The area has a complicated tectonic history: the Mergui basin experienced Triassic-Early Jurassic, followed by Early Cenozoic strike-slip faulting, which were then reactivated by normal faults (Morley, 2017). These faults have been active since the Early Eocene to Late Miocene, are 20.9-123 km long, and have 458 m-21.8 km displacement (Morley, 2017). They are over-displaced, with a D/L average of 0.14 (displacement is 14% of length), with D/L ratios as high as 0.26. These faults are over-displaced because they were able to establish their maximum length quickly by exploiting and reactivating pre-existing weaknesses inherited from previous faulting, and then accommodate strain by accruing displacement. These faults are still-active, but are very mature (i.e., they have been active since the Early Eocene); as a result, they have been able to attain high D/L ratios.

In summary, reactivated faults are, on average, over-displaced, and this should be considered when using D/L scaling laws to estimate faults length or displacement. However, it should be kept in mind that relatively young, still-active faults, such as the ones in the Turkana Rift (Vétel et al., 2005), will be under-displaced as they reached maximum length quickly but are still accruing displacement. When assessing reactivated faults, it is important to know how long the faults have been active.



Lithology	Young's Modulus (average)	Young's Modulus (range)	Average D/L ratio
Sedimentary w/ Evaporites	4-64 GPa (Evaporites)	23 GPa (Evaporites)	0.05
	0.04-67 GPa (Sedimentary rocks)	24 GPa (Sedimentary rocks)	
Fine-Grained Clastic Sedimentary	0.04-36 GPa	14 GPa	0.04
Clastic Sedimentary	6-67 GPa	25 GPa	0.06
Volcanic w/ Clastic Sedimentary	5-99 GPa (Volcanic)	49 GPa (Volcanic)	0.15
	6-67 GPa (Clastic Sedimentary)	25 GPa (Clastic Sedimentary)	
Carbonate	24-66 GPa	45 GPa	0.03
Mixed Clastic & Carbonate	24-66 GPa (Carbonate)	45 GPa (Carbonate)	0.04
	6-67 GPa (Clastic Sedimentary)	25 GPa (Clastic Sedimentary)	
Metamorphic	15.9-109 GPa	42 GPa	0.03
Volcanic	5-99 GPa	49 GPa	0.02

Figure 3. 16. Figure showing how the Young's Modulus of different lithologies relates to D/L. The range of Young's Modulus, average Young's modulus, and average D/L for fault with host rocks of each lithology is shown. Generally, as Young's Modulus increases, D/L decreases. Our Young's Modulus data and sources can be accessed here: https://figshare.com/articles/dataset/Young_s_Modulus/17087342

3.6.4. Lithology

Host rock lithology has been known to influence the relationship between fault length and displacement due to the stiffness of different lithologies, often described by host rock shear modulus (Walsh et al., 1988, 1989; Cowie & Scholz, 1992a; Wibberley et al., 1999). There is an inverse relationship between host rock shear modulus and D/L; fault in host rocks with a high shear modulus (stiffer rocks, for example, a granite) are under-displaced compared faults with a high shear modulus (softer rocks, for example, a mudstone) (Walsh et al., 1988, 1989; Cowie & Scholz, 1992a; Wibberley et al., 1999; Gudmundsson, 2004; Childs et al., 2017a).

The stiffness of rocks relates to their elastic properties, also expressed by the Young's Modulus and the Poisson ratio (Roche et al., 2013). Fault length and displacement have been related to rock stiffness in the following equation from crack models:

$$\frac{L}{D_{max}} = \frac{E}{2\Delta\tau(1 + \nu)}$$

Where E is Young's Modulus, ν is Poisson's ratio, and $\Delta\tau$ is the shear stress driving the fault (Roche et al., 2014). Poisson ratio can fall between 0.05 and 0.4, although values usually range between 0.3 and 0.4 (Gereck, 2007). There is generally little variation in Poisson's ratio between different lithologies, compared to Young's Modulus (Gudmundsson, 2004). Young's Modulus has a high amount of variation, ranging between 0.05 GPa and 100 GPa (Roche et

al., 2013). Fault displacement is inversely proportional to the Young's modulus of the rock (Wibberly et al., 1999, 2000a; Gudmundsson, 2004) which suggests that stiffer rocks, such as volcanic and metamorphic rocks, are more likely to be under-displaced than pyroclastic or sedimentary rocks. Factors such as increasing temperature, increasing porosity, and water content can decrease Young's Modulus. Highly fractured rocks have a low Young's Modulus; the breccia of a faults core has a low Young's Modulus, like that of a weak clay or pyroclastic tuff (Gudmundsson, 2004).

Our data appears to have a relationship between the D/L ratio of normal faults and host rock lithology, with faults with a low Young's Modulus tending to have a higher D/L ratio (i.e., they are over-displaced) (Figure 3.16). Evaporite-bearing sedimentary host rocks tend to be over-displaced compared to other lithologies, which could be due to the softness of the rocks making tip propagation difficult. Both evaporites and sedimentary rocks have a relatively low Young's Modulus, with sedimentary rocks ranging between 0.04-67 GPa (24 GPa average; Figure 3.16) and evaporites ranging between 4-64 GPa (23 GPa average; Figure 3.16). Faults within either fine- and coarse-grained clastic host rocks, are also relatively over-displaced, with Young's Modulus estimated between 0.04-36 GPa (14 GPa average; Figure 3.16) for fine-grained clastic sedimentary rocks, and 6-67 (25 GPa average) for sandstones and conglomerates (Figure 3.16). Faults in carbonates and mixed clastic/carbonates tend to lie in the middle of the various D/L trendlines, with carbonates having an estimated Young's Moduli between 24-66 GPa (45 GPa average; Figure 3.16). Faults within volcanic host rocks are significantly under-displaced (Figure 3.10B), which is possibly in part due to the stiffness of volcanic rocks; volcanic rocks have the highest estimated Young's Modulus, between 5-99 GPa (49 GPa average; Figure 3.16).

One example of under-displaced faults in host rocks with a high Young's Modulus are in the East African Rift (Figure 3.17; Williams et al., 2021). Here, normal faults are forming in a metamorphic host rock. Fault ages are not well constrained, but they are estimated to be roughly Pliocene in age and they are demonstrably still active (Scholz et al., 2020). Faults are 13 km to 130 km long and have displacements ranging from 122 m to 2.5 km. D_{max}/L is very low, between 0.003-0.03 (average of 0.01), indicating the faults are relatively under-displaced (Williams et al., 2021). It should be noted that these faults are active, which as discussed in section 2 could result in them being under-displaced. Additionally, some of the faults in the East African Rift have reacted foliation.

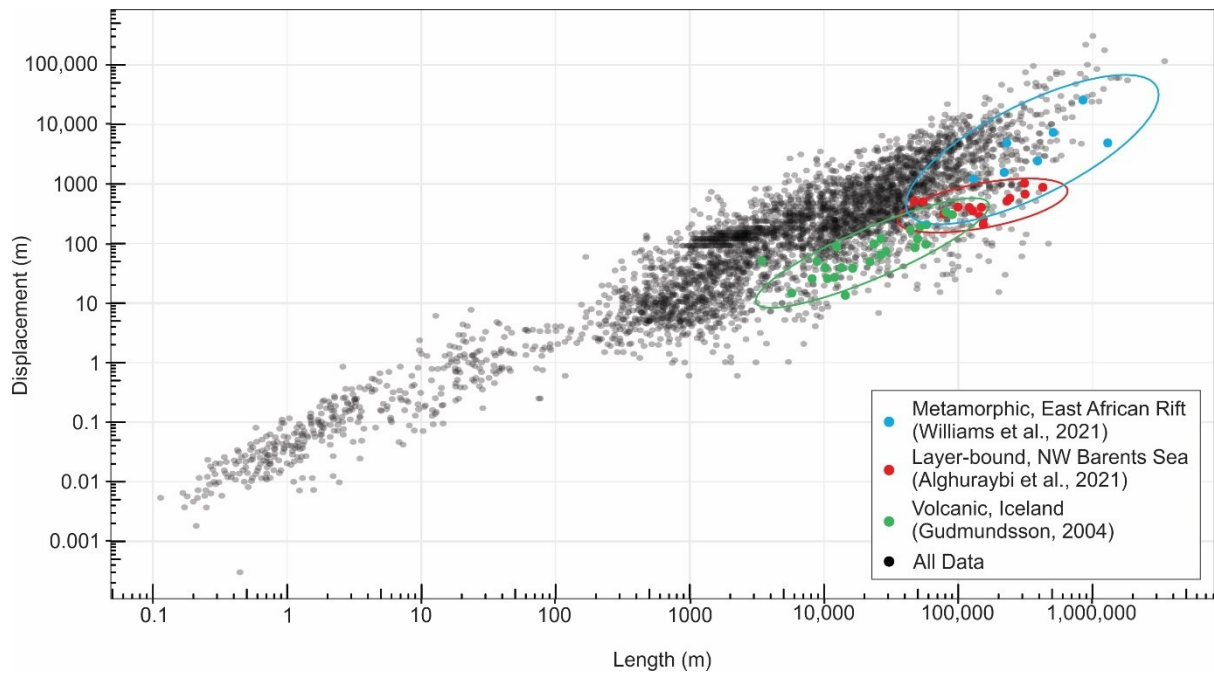


Figure 3. 17. Global D/L dataset (black) for normal faults in log-log space with faults from different host rocks highlighted: metamorphic host rocks from the East African Rift (from Williams et al., 2021), layer-bound faults from the NW Barents Sea in red (from Alghuraybi et al., 2021), and volcanic host rocks from Iceland (from Gudmundsson, 2004).

Another example of under-displaced normal faults within stiff host rocks come from an active rift zone in Iceland (Figure 3.17; Gudmundsson, 2004). The faults are Holocene ($<10,000$ years old) and cut through basaltic pahoehoe lava flows with an estimated Young's Modulus of 30-60 GPa, and possibly as high as 100 GPa. Faults range from 345 m to 9 km long and have displacements ranging from 1.3 m to 33 m. D_{max}/L is between 0.0009-0.01 (average of 0.004), meaning the faults are under-displaced (Gudmundsson, 2004). We expect the stiff host rock lithology has contributed to these faults being under-displaced; however, these are active faults, so according to a constant-length fault growth model, they have possibly reached their maximum length, but not their maximum displacement.

Differences in mechanical stratigraphy between lithological units can create vertical barriers that inhibit fault growth, which can cause faults to be under-displaced (Peacock & Sanderson, 1992; Wilkins & Gross, 2002; Welch et al., 2009; Roche et al., 2014). There is a relationship between rock stiffness and fault displacement gradient (i.e., the displacement variation per unit length across a fault), with these gradients tending to be higher in rock units with lower Young's Modulus (Roche et al., 2014). Mechanical stratigraphy can restrict faults from propagating vertically, causing faults to have a high aspect ratio (fault height/length; height is the fault dimension along dip) (Nicol et al., 1996; Scholz & Fossen, 2002; Soliva et al., 2006; Roche et al., 2013; Alghuraybi et al., 2021). In a numerical modelling analogue study by

Roche et al. (2013), aspect ratios for faults in homogeneous rock properties not bounded by mechanical stratigraphy are typically >2 , whereas aspect ratios of faults in limestone-clay sequences are, on average, 13, and even as high as 50 (Roche et al., 2013). However, no aspect ratios >20 have been reported in natural studies (Torabi et al., 2019). If faults have a high aspect (height-length) ratio, it stands that they would likely also have a high displacement-length ratio. However, this is likely only applicable to relatively small faults, or possibly large faults cutting through thick layers (e.g., a fault with 1 m of displacement offsetting a 20 cm-thick mudstone package, vs. a 1 km displacement fault offsetting a 100 m-thick mudstone package).

One example of under-displaced faults with high aspect ratios included in our database are layer-bound, thin-skinned normal faults from the NW Barents Sea (Figure 3.17; Alghuraybi et al., 2021). The faults in this study were only active in the Late Jurassic and they occur in a fine-grained clastic host rock. Faults are 4.7 km to 42.7 km long, have displacements ranging from 21 m to 103 m, and their $D_{\max}/L$ is between 0.001-0.009 (average of 0.003). They have aspect ratios as high as 19, compatible with aspect ratios found in the numerical models of Roche et al. (2013). The faults from the NW Barents Sea are interpreted to have reached their final length quickly (i.e., they grew in accordance with the constant-length model) and were not able to reach their full potential displacement, likely due to the mechanical layering.

In summary, host rock lithology influences D/L ratios; softer rocks (such as sedimentary rocks) tend to be over-displaced, and stiffer rocks (such as volcanic rocks) tend to be under-displaced. Mechanical stratigraphy also causes rocks to be vertically restricted and causes them to be under-displaced.

3.6.5. How useful are D/L scaling laws?

Our overall scaling relationship between normal fault length and maximum displacement for all data in our revised database is $D=0.03L^{0.92\pm0.01}$. This agrees with previous literature that estimated $n=1$ (Cowie & Scholz, 1992a; Dawers et al., 1993; Scholz et al., 1993; Clark & Cox, 1996; Schlische et al., 1996; Kim & Sanderson, 2005; Xu et al., 2006). Our database has thousands of faults that span eight orders of magnitude in terms of fault length, thus we believe that we can confidently say that, overall, $n=1$, and that our equation could be used to reliably estimate D or L within 1-2 orders of magnitude in most cases.

There is so much variability in our plots ($r^2=0.85$) when all D/L data is considered that it could be questioned whether a single global scaling law should be used at all. Even after

conducting a detailed quality check of the data and removing data for which we believe there are errors/inconsistencies, there is still significant variation in the D/L scaling relationship for normal faults. Within each of our trendlines, there is overlap within the confidence intervals, making some of them statistically insignificant, so fault D/L should not be decided based on one scaling law or even based on a singular geological trait, unless it is a one of the established under-displaced outliers (e.g., volcanic rocks, new rift zones). We believe that there *is* value in being able to estimate D from L, and that more specialised scaling relationships considering size, lithology, tectonic history, and fault maturity are thus warranted. For example, using our global scaling law ($D_{max}=0.03L^{0.92}$), we would estimate that a 3 km long normal fault within a sandstone host rock in a tectonically active area would have a displacement of c. 47 m. In contrast, if we use the ‘clastic sedimentary’ D/L equation ($D_{max}=0.11L^{0.84}$), we would estimate a displacement of c. 277 m; by using the ‘inactive’ D/L equation ($D_{max}=0.05L^{0.93}$) we would estimate a displacement of c. 126 m. Both values are likely more accurate than the global estimate, which may have implications for situations which require estimating fault displacement, such as understanding fault sealing, possible CO₂ leakage, how large an earthquake could be. For the most accurate estimations, we would suggest either 1) calculating an average of the applicable equations, in this example, an average between the value for a ‘clastic sedimentary’ and ‘inactive’, or to have an even more accurate estimation, 2) use our database to combine faults with similar factors to make a bespoke equation for that area.

The relationship between fault D and L is also dynamic, changing throughout a fault’s life. It is important to practice caution when working with D/L ratios. As shown in Figure 3.13, the relationship between D and L evolves through time, thus using static data to infer a dynamic relationship can be problematic. Plotting data only in log-log space can hide variability and statistical spread, as shown by Rotevatn et al. (2019). For example, different stages of fault growth will likely be masked in a large log-log plot, as the fault lies within the global scatter at every stage of fault growth. This shows that fault growth cannot be inferred from global D/L plots, and that plotting D and L *through* time (Figure 3.13) is important to understanding fault growth.

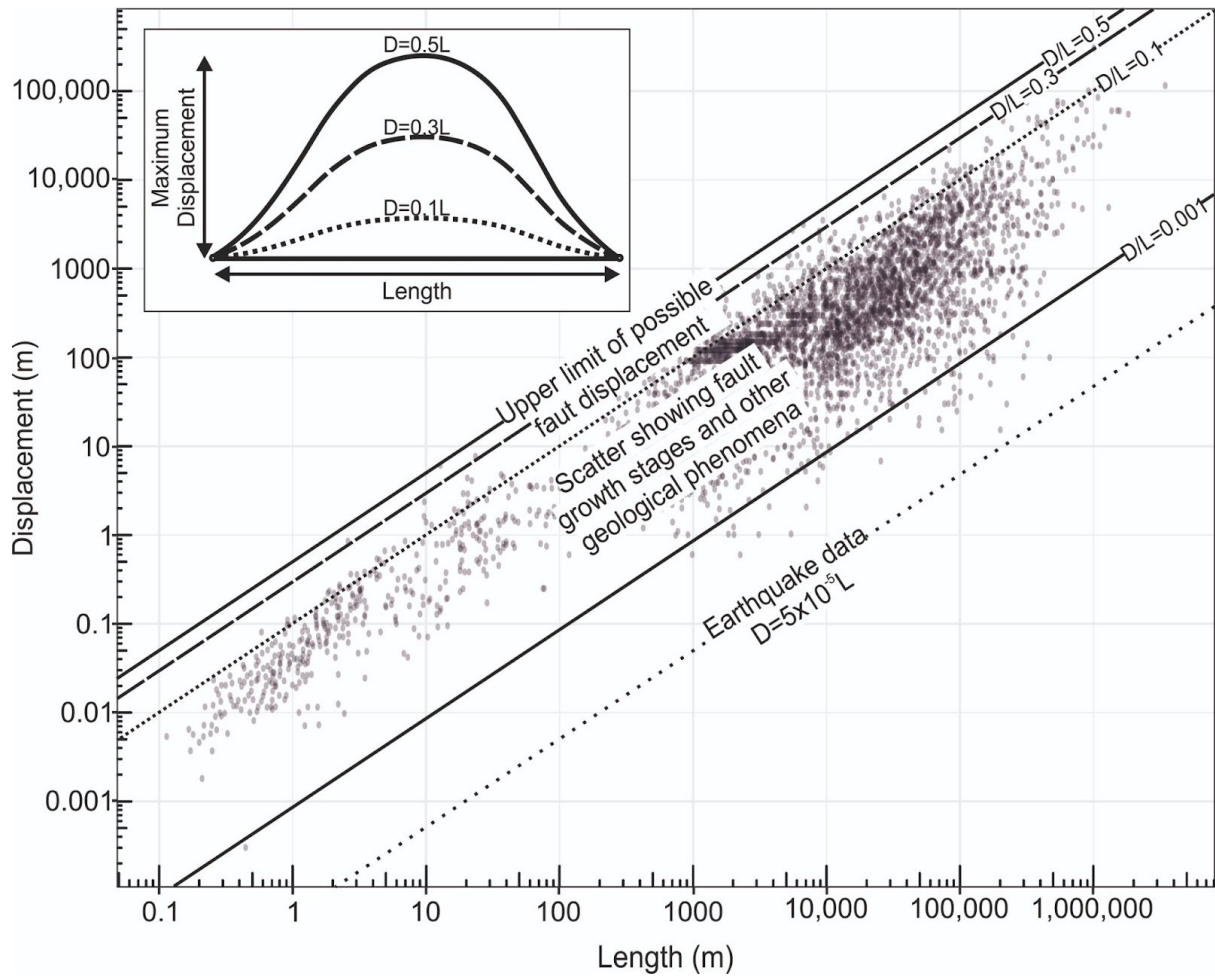


Figure 3. 18. Global D/L dataset for normal faults with our suggested upper limit of D/L ($D/L=0.1$). The average D/L value of a single earthquake is also shown (Wells & Coppersmith, 1992).

3.6.6. Upper bounds of displacement

In contrast to the lower limits of the D/L scaling dataset, which shows significant scatter likely reflecting the process of fault growth, there appears to be an upper limit of maximum displacement, with a significant amount of variation on the lower bounds of displacement (Figure 3.18). The absolute upper bound is the upper limit of $D_{max}/L=0.5$ (i.e., at max, faults displacement can be $\frac{1}{2}$ of length), however very few faults have a D/L value that high. 99.7% of the data falls below $D/L = 0.3$, and 94% of the data falls below $D/L=0.1$.

There is a great amount of scatter in the lower bounds of D/L ratios and a hard upper limit. Most faults follow a constant-length or hybrid fault growth model; they will reach max length quickly and then can become inactive before reaching maximum displacement potential, creating scatter in the lower bounds of fault displacement (Figure 3.18). The D/L upper-limit seen in our global dataset is related to an overarching rule of fault mechanics in which faults cannot accommodate a certain amount of displacement without additional propagation or linkage with another fault. The wall-rock that borders the fault tips can accommodate a finite

amount of shear stress, and beyond that the rock will fail, resulting in additional fault tip propagation (Freeman et al., 2010). Upper D/L limits could also be due to isostatic restoring forces due to the topography generated in the hanging wall and footwall blocks of the fault (Cowie & Scholz, 1992a).

3.7. Conclusions

We here present a new normal fault database that presents fault length and displacement along with host rock lithology, fault maturity, and tectonic history that will now be available to the public. In our interrogation of our global normal fault database, we found that 1) for the complete dataset $n=0.92$ in terms of the standard equation $D_{max}=cL^n$, but there is a lot of scatter in D/L in the global dataset when faults of all lithologies, maturities, and tectonic histories are grouped together, 2) small faults (> 1 m) tend to be over-displaced, 3) stiffer rocks tend to be under-displaced, and softer rocks tend to be over-displaced, 4) active faults tend to be over-displaced compared to inactive faults, and 5) reactivated faults are over-displaced compared to faults in previously undeformed settings, unless the reactivated faults are still active. We also collected normal fault D/L through time data and found that faults grow via a constant length-to-hybrid fault growth model. Since D/L ratios are changing throughout a fault's life, it is important to express caution when looking at static D/L data.

Chapter 4

Normal fault kinematics and the role of lateral tip retreat: an example from offshore NW Australia

The work presented in this chapter has been published in **Tectonics**:

Lathrop, B. A., Jackson, C. A. L., Bell, R. E., & Rotevatn, A. (2021). Normal Fault Kinematics and the Role of Lateral Tip Retreat: An Example from Offshore NW Australia. *Tectonics*, 40(5). <https://doi.org/10.1029/2020TC006631>

Acknowledgements

We would like to thank Geoscience Australia for making all of the data used in this study publicly available. We thank Schlumberger for providing access to Petrel software. We also thank the Imperial College Basins Research Group (BRG) for their input and help throughout this research, and many colleagues with whom we have had beneficial conversations with at EGU and TSG conferences. Finally, we thank Telemaco Tesei and two anonymous reviewers for their constructive feedback, which helped improve the final version of this manuscript.

Data availability statement

The Glencoe 3D seismic data set and associated wells can be downloaded from <https://www.ga.gov.au/nopims> by searching for the survey “Glencoe” in the data access search engine.

Abstract

Understanding how normal faults grow is key to determining the tectono-stratigraphic evolution of rifts. According to recent studies, normal faults tend to grow in two temporally distinct stages: a lengthening stage, followed by a throw/displacement accumulation stage. However, this model is still debated and not widely supported by many additional studies. Relatively few studies have investigated what happens to a fault as it becomes inactive, that is, does it abruptly die, or does the at-surface trace-length progressively shorten by so-called tip retreat? We, here, use a 3D seismic reflection data set from the Exmouth Plateau, offshore Australia, to develop a three-stage fault growth model for seven normal faults of various sizes and to show how the throw-length scaling relationship changes as a fault dies. We show that during the lengthening stage, which lasted <30% of the faults' lives, faults reached their near-maximum lengths, yet accumulated only 10%–20% of their total throw. During the throw/displacement accumulation stage, which accounts for c. 30%–75% of the faults' lives, throw continued to accumulate along the entire length of the faults. All of the studied faults also underwent a stage of lateral tip-retreat (last c. 25% of the faults' lives), where the active at-surface trace-length decreased by up to 25%. The results of our study may have broader implications for fault growth models, slip rate variability during fault growth, and the way in which faults die, in particular the role of lateral tip-retreat.

4.1. Introduction

Normal fault growth models have been widely debated over the past c. 20 years. The propagating fault model, also referred to as the isolated fault model (Walsh et al., 2003), suggests that normal faults grow via a synchronous increase in length and displacement, that is, that when faults lengthen, they also accumulate displacement. Faults can also lengthen via tip propagation and linkage of these individual segments (e.g., Morley et al., 1990; Dawers et al., 1993; Cartwright et al., 1995; Walsh & Watterson, 1988; Walsh et al., 2003). The constant-length model instead suggests that normal faults reach their near-final lengths relatively rapidly and spend the rest of their lives accruing displacement without further significant lengthening (Nicol et al., 2005, 2017; Jackson & Rotevatn, 2013; Henstra et al., 2015; Fossen & Rotevatn, 2016; Hemelsdaël & Ford, 2016; Tvedt et al., 2016; Walsh et al., 2002, 2003; Childs et al., 2017b; see also Cowie & Shipton, 1998). More recently, Jackson et al. (2017) and Rotevatn et al. (2019) used 3D seismic reflection data and physical analog models to propose a third model, the so-called the “hybrid growth model.” This model states that the propagating fault model and the constant-length models may not in fact be mutually exclusive, end-member models, but instead represent discrete kinematic phases in the life of a single fault: that is, an initial lengthening stage (propagating fault stage) is followed by a later displacement accumulation stage (constant-length stage) (Jackson et al., 2017; Rotevatn et al., 2019). During the lengthening stage, which encompasses c. 20%–30% of the duration of a fault’s life, faults reach their near-final length via the propagation and linkage of relatively small, discrete segments; during this time, the fault accumulates 10%–60% of its total displacement (Jackson et al., 2017; Rotevatn et al., 2019). During the displacement accrual stage, which takes place during the latter 70%–80% of the fault’s life, the fault accumulates 40%–90% of its total displacement (Jackson et al., 2017; Rotevatn et al., 2019).

Whereas many studies have investigated how normal faults initiate and grow (see above), few have considered what happens at the end of a fault’s life. These few studies propose that faults die in two general ways: the entire trace-length of the fault remains active before slip ceases, or that as the fault dies, activity is focused onto the center of the fault, leading to a progressively shorter active fault trace-length (Meyer et al., 2002; Childs et al., 2003). In the latter case, normal faults experience a stage of fault tip retreat, that is, the lateral tip regions do not accumulate further displacement or throw as strain is localized near the fault center (Figure 4.1; Meyer et al., 2002). In 3D seismic reflection data, tip retreat can be observed by identifying packages of growth strata that are deposited over progressively shorter along-

strike lengths as the fault reaches the end of its life (Meyer et al., 2002). Tip retreat has also been interpreted as a result of relay breaching during segment linkage (Childs et al., 2003); however, this is what we would classify as a stage of fault growth and not, strictly speaking, lateral tip retreat.

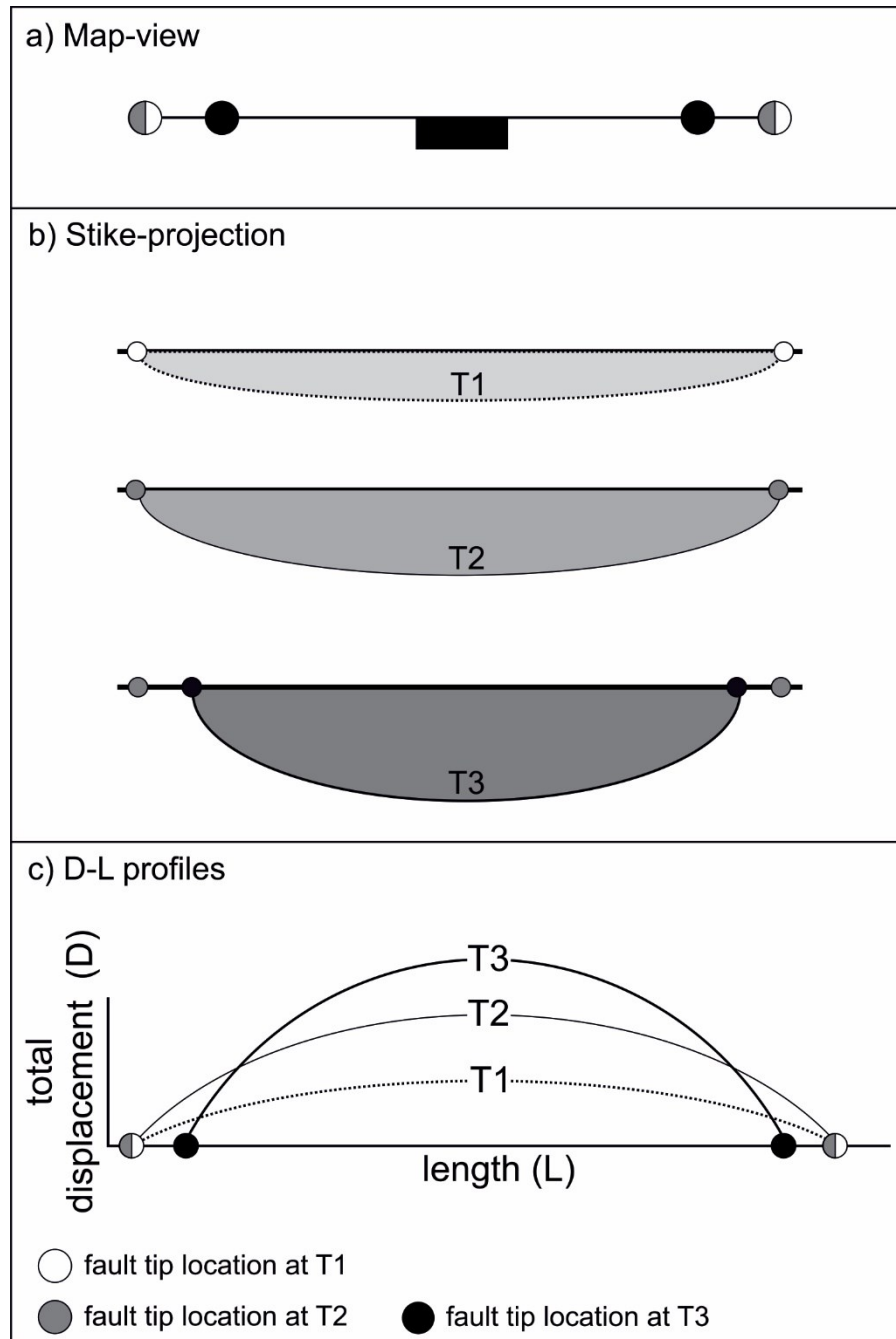


Figure 4. 1. Conceptual models for the development of normal faults following a “hybrid fault model” (Rotevatn et al., 2019) with a stage of tip retreat. Time 1 (T1) represents the lengthening/propagating fault model stage, Time 2 (T2) represents the displacement accumulation/constant-length model stage, and Time 3 (T3) represents a phase of fault tip-line retreat. A) map view of the active fault trace line at T1-3. Note that the fault reaches its maximum length at T1 and has a shortened active trace line at T3. B) Along-strike projection of throw at T1-3. An increasing amount of displacement is accumulated at each stage. C) Displacement/length profile at T1-3.

Relatively few studies have discussed the role tip retreat plays in the evolution of normal faults (Meyer et al., 2002; Morley, 2002; Childs et al., 2003; Freitag et al., 2017; Nicol et al., 2020), and it is therefore not usually included in fault growth models. This likely reflects the fact it is very difficult or sometimes impossible to constrain the kinematics of normal faults, for example in cases where growth strata are absent and/ or only locally preserved. To the best of our knowledge, tip retreat has also not yet been the focus of or identified in, physical or numerical models. Freitag et al. (2017) show an example of tip retreat in the Columbus Basin, offshore Trinidad; these are, however, thin-skinned, gravity-driven faults, and it is not clear if the kinematics would apply to thick-skinned faults offsetting crystalline basement. Morley (2002) also show an example of possible tip retreat in the East African Rift, but since this is a sediment-starved (i.e., underfilled) basin, it is difficult to tell if the fault really experienced tip retreat, or whether the observed geometries simply reflect post-fault death passive filling of hanging wall accommodation. Motivated by the lack of examples that highlight the potentially important role of tip retreat, we here provide a well-constrained example of tip retreat occurring on basement-involved, tectonically (i.e., plate-motion) driven normal faults, as well as guidance on how to identify this important process in the rock record.

In this paper, we use 3D seismic reflection and borehole data from the Exmouth Plateau, offshore Australia to study the kinematics of thick-skinned tectonic normal faults that offset crystalline basement. More specifically we: (1) constrain the temporal relationship between fault lengthening and throw and (2) investigate the role of tip retreat as faults become inactive. This is an excellent place to study this process because syn-sedimentary normal faults are well-preserved, age-constrained, and well-imaged in excellent-quality, open-source, 3D seismic reflection data. The rift basin was also overfilled for much of the duration of faulting, meaning the faults are flanked by well-developed growth (syn-tectonic) strata.

4.2. Geologic Setting of the Exmouth Plateau

Our study area is located on the Exmouth Plateau, North Carnarvon Basin, offshore NW Australia (Figure 4.2). The North Carnarvon Basin formed due to rifting in the Late Carboniferous-Permian as a result of the breakup of Pangea, and the Exmouth Plateau formed as a result of rifting between Greater India and Australia, creating NE-trending blocks (Stagg & Colwell, 1994; Longley et al., 2002; Gibbons et al., 2012). The Exmouth Plateau is located in the northern part of the North Carnarvon Basin, bounded by the continental shelf to the southeast, and the Curvier, Gascoyne, and Argo abyssal plains to the SW, SW, and NE,

respectively (Longley et al., 2002). The Exmouth Plateau is a block of thin crystalline crust, and based on geophysical evidence, it has been suggested that the Exmouth Plateau basement is continental crust, however this has not been confirmed by direct sampling (Stagg et al., 2004). The crystalline basement is overlain by a thick pre-rift succession, consisting of the fluvial-deltaic to marginal marine, Mungaroo Formation (Triassic) (Longley et al., 2002; Stagg et al., 2004).

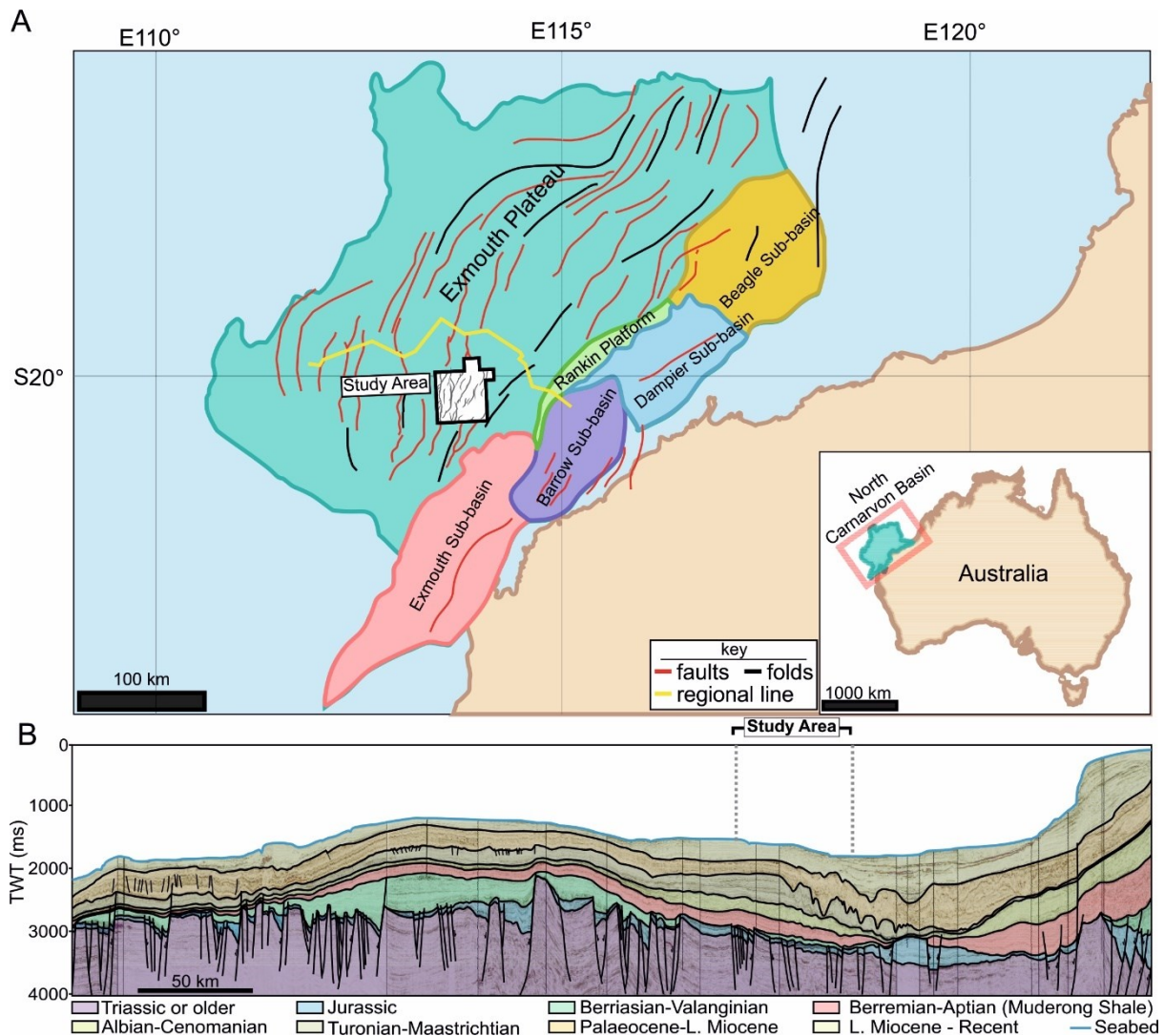


Figure 4. 2. Study area. A) Location of the Exmouth Plateau in the North Carnarvon Basin, offshore Australia (fault locations modified from Pan et al., 2020), B) Regional 2D seismic line across the study area, modified from Nugraha et al. (2019). The location of this regional line is labeled on Figure 2a (yellow).

The syn-rift extension began in the Late Triassic (Rhetian) until Late Jurassic (Oxfordian), during which time the Murat and Athnol siltstones were deposited in a sediment-starved basin (Figure 4.3) (Longley et al., 2002; Tindale et al., 1998). After a short period of tectonic quiescence in the Late Jurassic, rifting continued in the Early Cretaceous in an over-filled basin environment, during which time marine claystones (Dingo Claystone) and coarser-

grained, deltaic clastics (Barrow Group) were deposited (Longley et al., 2002). Rifting in the Exmouth Plateau ceased in the Hauterivian, and the area became a passive margin (Gibbons et al., 2012; Longley et al., 2002). In this paper, we focus on the Jurassic-Early Cretaceous, syn-sedimentary normal faults which are generally trending N-NE.

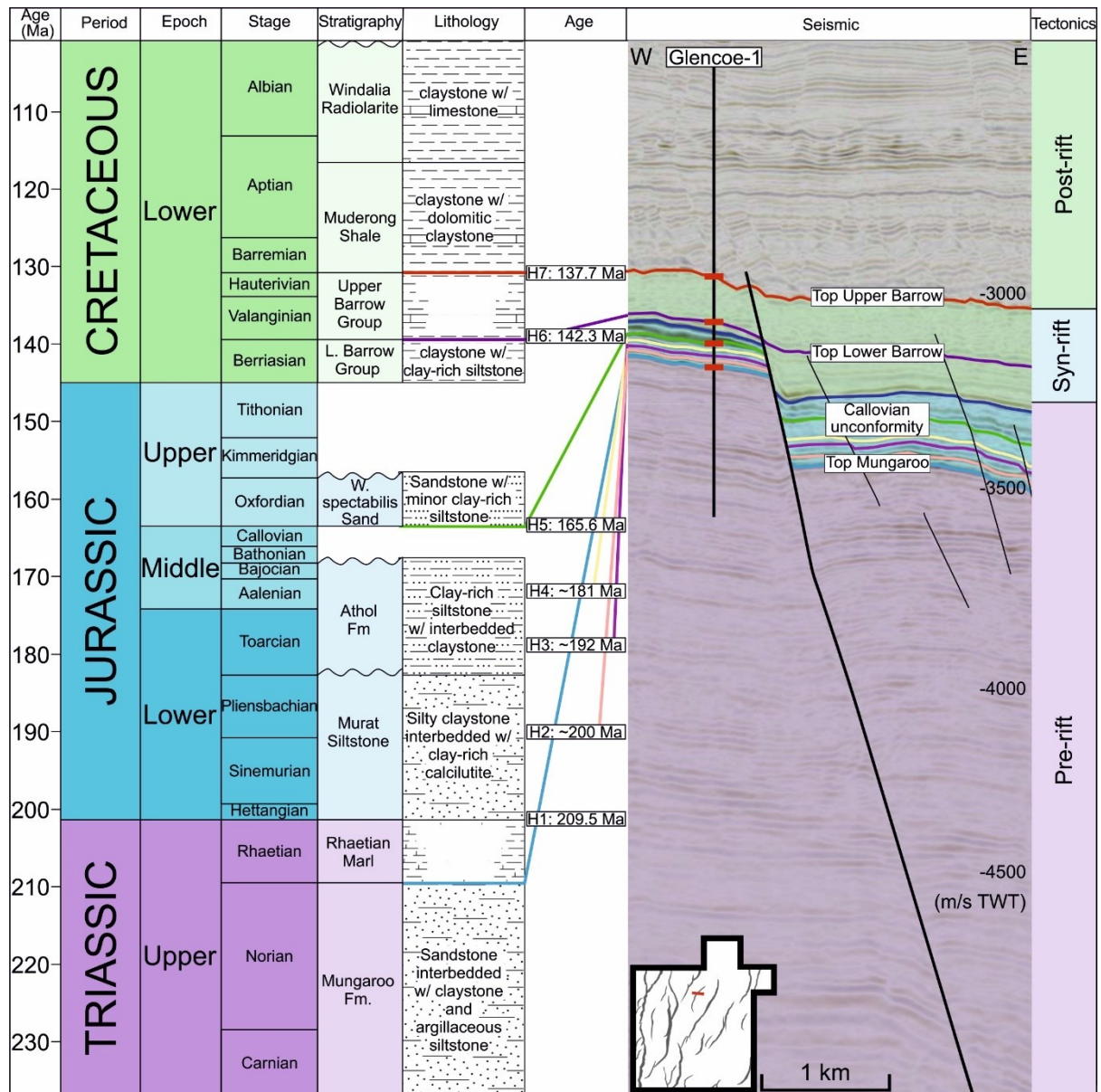


Figure 4.3. Stratigraphic framework showing the key interpreted seismic horizons, their ages, and the tectonic evolution of the Exmouth Plateau (H = horizon). Ages for H1, H5, H6, and H7 were taken from Marshal and Lang (2013), and H2, H3, and H4 are relative dates assuming constant sedimentation. Information on the tectonostratigraphic framework are from Bilal et al. (2018) and Geoscience Australia.

4.3. Data and methodology

4.3.1. Data

The Glencoe data set is a 3D time-migrated seismic reflection survey that encompasses approximately 3,900 km² of the Kangaroo syncline in the Exmouth Plateau (Figure 4.4). It

has a bin spacing of 25 m and a record length of 8 s two-way time (TWT). The vertical and horizontal resolution are approximated by measuring the dominant wavelength in the interval of interest ($\lambda = 26.3$ m) and calculating $\lambda/4$ (where λ is the seismic wavelength), yielding c. 6.6 m within the syn-rift sequence (Brown, 2011).

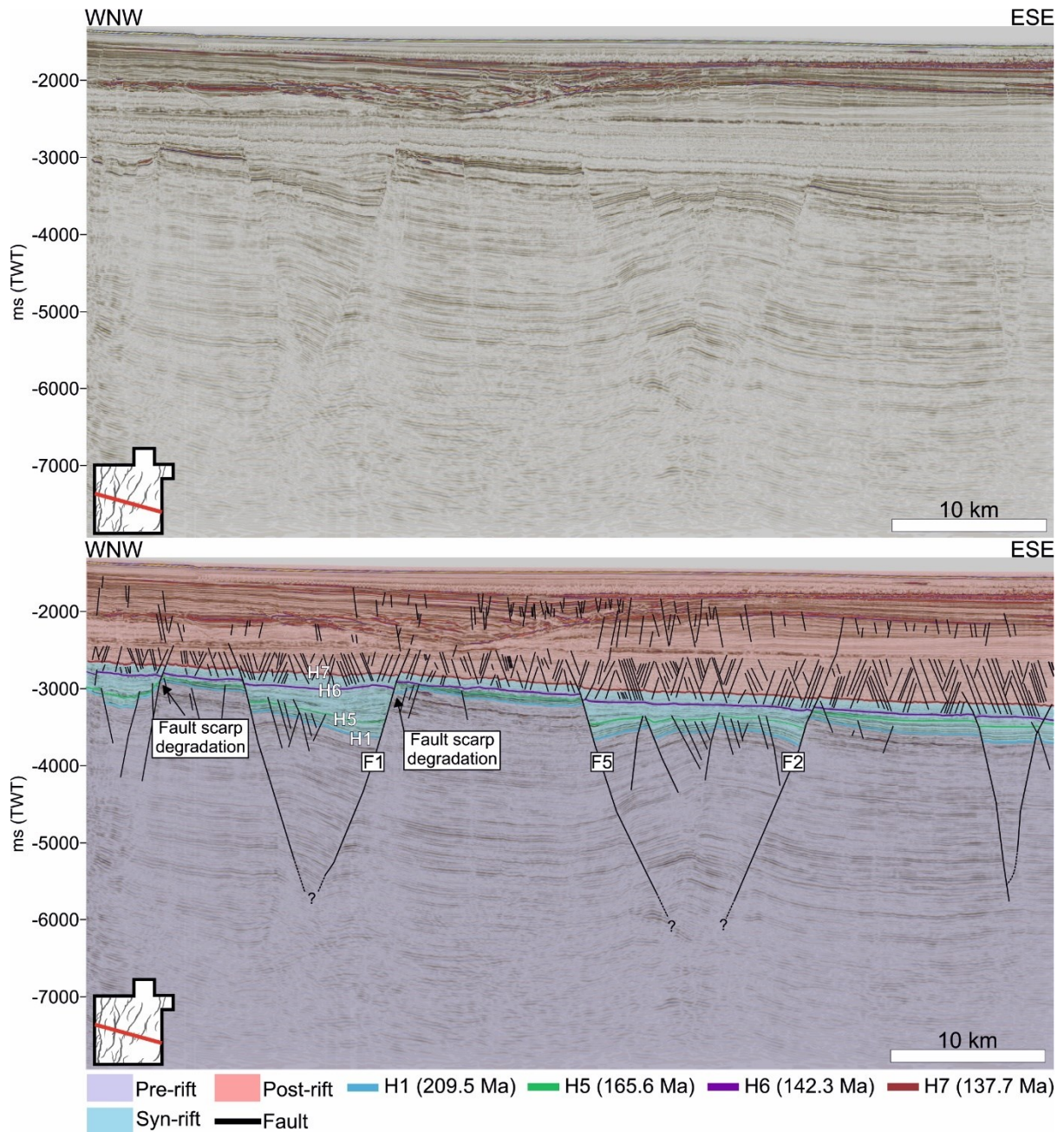


Figure 4.4. Representative seismic line in TWT along the central section of the 3D data set, across strike of the studied faults. Data is shown with and without interpretation.

Seismic sections are displayed with normal polarity (SEG European Convention; Brown, 2011), where increase in acoustic impedance is represented by a peak (red), and a decrease by a trough (black). Seismic inlines are orientated WNW-ESE and the survey is tied to four wells (Glencoe-1, Nimblefoot-1, Warrior-1, and Breseis-1). Well-logs, formation tops, and

biostratigraphic ages were provided with the wells. All seismic and well data are open-access and available from Geoscience Australia.

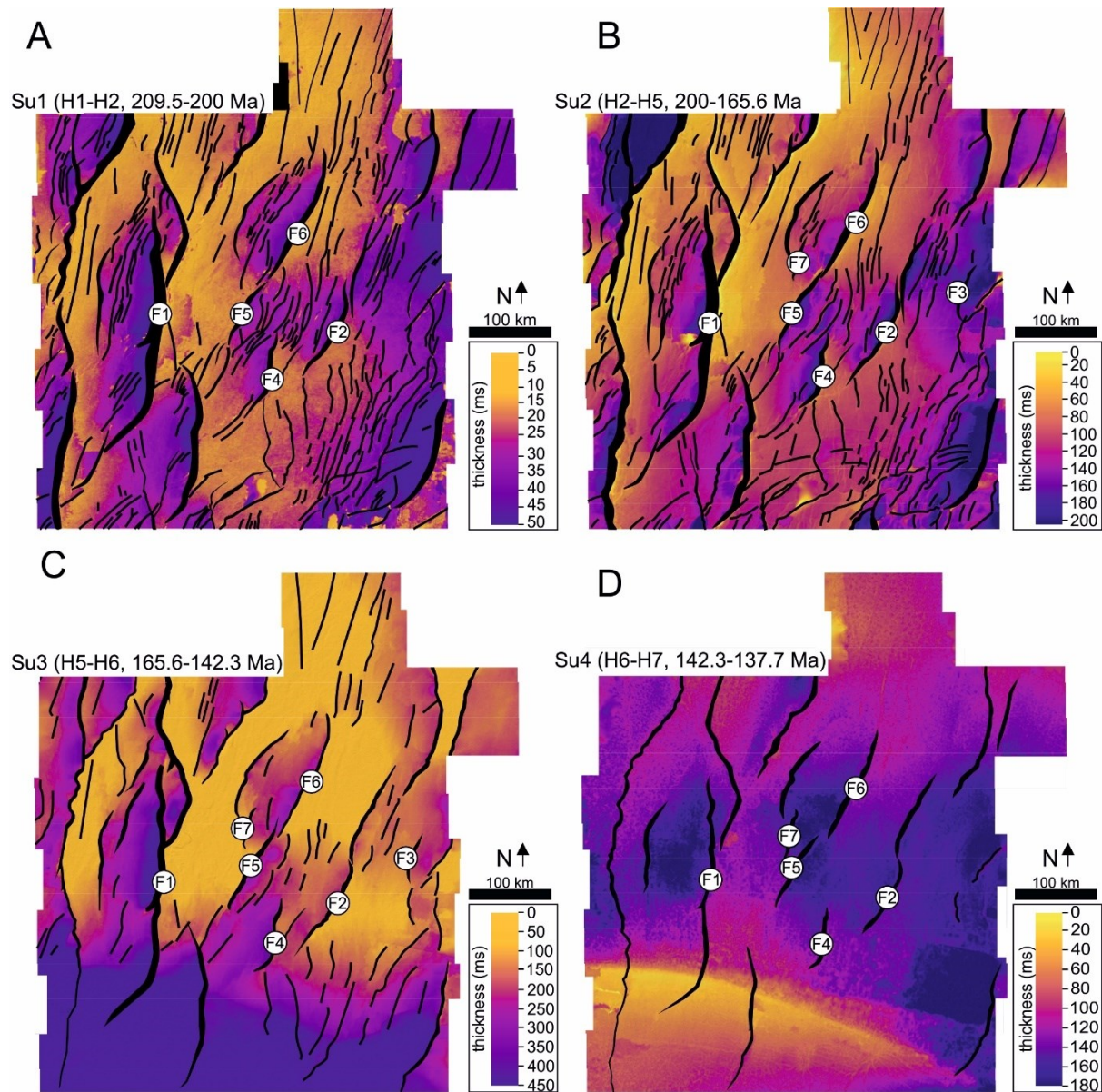


Figure 4. 5. Representative isochrons of the study area with interpreted major and minor faults labeled. A) Seismic Unit 1: Horizon 1–2, 209.5–200 Ma, B) Seismic Unit 2: Horizon 2–5, 200–165.6 Ma, C) Seismic Unit 3, Horizon 5–6, 165.6–142.3 Ma, D) Seismic Unit 4, Horizon 6–7, 142.3–137.7 Ma.

We have mapped seven regionally extensive seismic horizons (H1–7); H1, H5, H6, and H7 are age-constrained well-tied horizons with ages from well reports, as well as ages obtained by Marshall and Lang (2013) using biostratigraphy from 1,500 wells around the North Carnarvon Basin (Figure 4.3). We lack direct age-constraints for H2–4, thus we estimated their ages by assuming a constant sedimentation rate between horizons of known ages (Figure 4.3). We also locally picked additional horizons within the syn-sedimentary deposits (e.g., H5.5) that are not continuous across the entire data set; we estimated their ages based on an

assumption of constant sedimentation rates between overlying and underlying, age-constrained horizons. We mapped and analyzed seven faults of varying sizes (8.8–42 km long, with 165–680 m of throw) to show how faults of different sizes grow in the area, and to see if the styles of fault growth are scale dependent (see Figure 4.5 for fault locations).

4.3.2. Methodology

In this study we used three different methods to quantify fault growth: isochron analysis, throw backstripping, and expansion index (EI) analysis (see review by Jackson et al., 2017). First, we created time-thickness (isochron) maps of key stratigraphic intervals, which illustrate variations in sediment thickness. This highlights across-fault hanging wall thickening, which can reveal the growth history of a fault (e.g., Jackson & Rotevatn, 2013). Isochron analysis was done first in order to establish the general style of fault growth (i.e., a propagating, constant-length, or hybrid fault growth model), and then we conducted throw backstripping to be able to see exact fault throw and length through time in the faults' lives (Jackson et al., 2017). To begin throw backstripping, we created throw-length (T-x) plots by picking the hanging wall and footwall cutoffs for every chosen horizon across the length of the faults (appendix figures 4. 1–4.3). In the case of folding or erosion (Figure 4.4), horizons used to calculate throw were projected across the fold or eroded fault scarp (e.g., Wilson et al., 2013). Throw backstripping involves subtracting the throw of a shallower horizon directly from the throw of a deeper horizon at the same along-strike position, with this being repeated for successively deeper horizons (Chapman & Meneilly, 1991; Peterson et al., 1992). We opted to use the “original method” of throw backstripping, where throw across different horizons is simply subtracted, as we did not want to make any assumptions about the style of fault growth (see Jackson et al., 2017 for more details on fault displacement backstripping methods). Finally, we used EI analysis to measure variations in stratal thickness across the fault by dividing the thickness of hanging wall stratal unit by that of the equivalent unit in the footwall (Thorsen, 1963; Cartwright et al., 1998; Bouroullec et al., 2004; Jackson et al., 2017) (See appendix Figure 4.A4). This technique shows the formation and growth of depocenters, and therefore how the faults lengthened (Jackson & Rotevatn, 2013).

We also calculated vertical throw gradients by dividing the change in throw by the change in depth of the shallowest two horizons offset across the fault. We calculated upper-tip throw gradients in order to demonstrate that the top of the fault was interacting with the free surface rather than acting as a blind fault; this is important when trying to understand if faults experienced real tip-line retreat or not (Walsh & Watterson, 1988; Meyer et al., 2002; Childs

et al., 2003). Finally, we calculated fault slip rates by dividing displacement for a particular time period by the duration of that time period; this was done in order to investigate whether slip rates varied between the different stages of fault growth. It is important to note that we plotted total throw and length through time using data derived from: (i) all seven of our seismic-stratigraphically defined horizons, four of which were directly age-constrained by well data, and three for which the ages were only estimated (see above); and; (ii) only our four age-constrained seismic horizons. We plotted slip rate through time using only age-constrained horizons. This allowed us to constrain a range of rates for time-variable parameters, which future additional well data may help refine.

Since the basin was sediment-starved from the Early Jurassic until the Late Jurassic, as evidenced by the fault scarp degradation until the deposition of H5 (Figures 4.4, 4.6a, and 4.8a), our fault lengthening calculations are upper limit estimations. For example, if active faulting created hanging wall accommodation but the basin was sediment starved, this accommodation would have remained unfilled. Thus, what looks like tip propagation could just be prolonged filling of the hanging wall of an inactive normal fault (see Jackson et al., 2017). It is therefore possible that the faults reached their maximum lengths even quicker than what we estimate.

There is a level of uncertainty when attempting to map fault tip positions in 3D seismic, even with high-quality data (Pickering et al., 1997). Our seismic data set has a vertical and horizontal resolution of c. 6.6 m; this means faults smaller (i.e., shorter and with less displacement) than this value are not imaged, and that the tips of otherwise larger faults will also not be imaged. Because of this, it is likely we are underestimating fault lengths by a few hundred meters (see Pickering et al., 1997). However, this seismic imaging resolution issue applies at all stages of fault growth and therefore does not impact our key observations (i.e., that EI values fall below 1 near the fault tips during the later stages of their lives) and related interpretations (i.e., that the active fault-trace length shortens as the fault dies).

We use checkshot (velocity) data from our four wells to convert throw values from milliseconds two-way time (m s TWT) to depth (m) (see appendix Figure A1). Throw values are presented in meters. Burial-related compaction of sedimentary rocks can result in throw calculations being underestimated, especially when rocks have a high shale content or are deeply buried (>2 km; see Taylor et al., 2008). Decompaction typically decreases throw

estimates by <20% (Taylor et al., 2008), so we here give all throw and slip rate values an error to account for maximum of 20% decompaction.

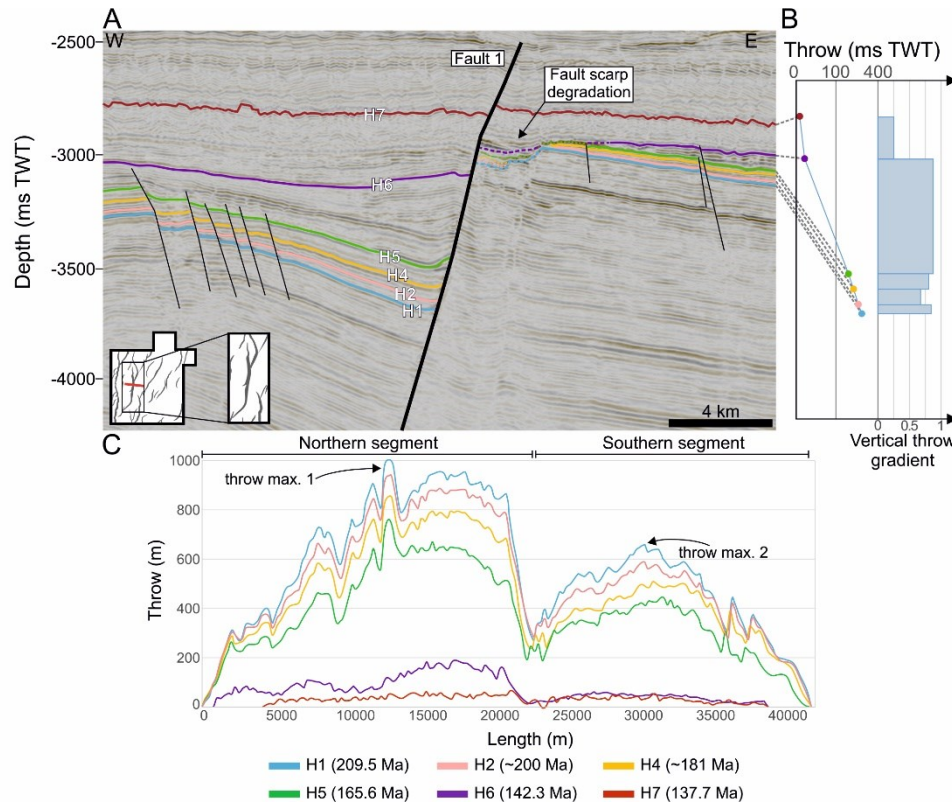


Figure 4. 6. A) Seismic profile illustrating F1 at its point of highest throw and its correlated throws and throw gradients, B) Vertical throw gradients for each horizon, C) Throw-distance plot illustrating the lateral variations in throw across each seismic unit. All throw values could be underestimated up to 20% due to post-depositional compaction of faulted strata (Taylor et al., 2008).

4.4. Results

We have completed a comprehensive geometric and kinematic analysis of seven faults of various sizes (appendix Figures A2–A7). We first provide a detailed description of the geometry of three faults (and their related growth strata) that are representative of the various fault sizes identified in the study area, before describing their kinematics. Fault 1 (F1) represents the largest studied fault, Fault 2 (F2) represents a mid-sized fault, and Fault 3 (F3) represents the smallest studied fault in the data set. We then present and discuss the results for all of the studied faults.

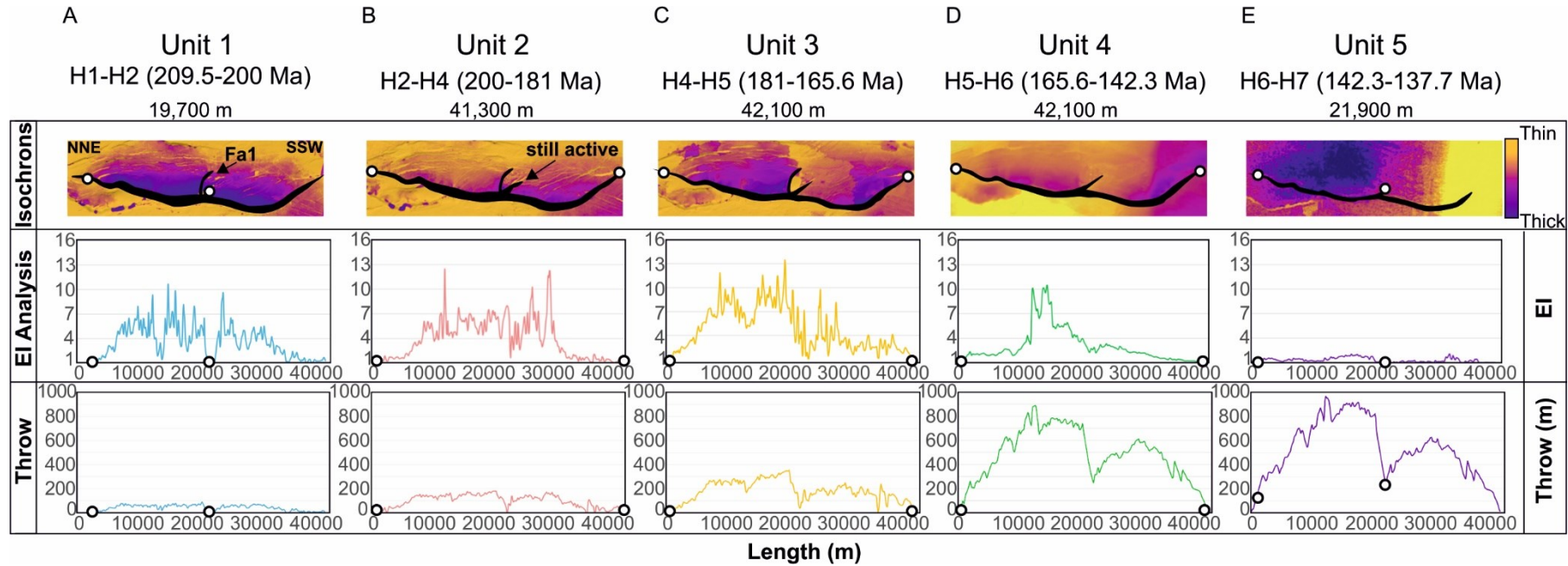


Figure 4. 7. Isochrons, expansion index analysis, and throw throughout different stages of the life of F1. The throw through time values are taken from throw backstripping, which can be seen in detail in the appendix figures. White dots indicate the length of the fault at the specified interval. A) Isochron showing the thickness between H1 (209.5 Ma) and H2 (200 Ma), maximum throw is 78 m, and length is 19,700 m, B) Isochron showing the thickness between H2 (200 Ma) and H4 (181 Ma), maximum throw is 191 m and length is 41,300 m, C) Isochron showing the thickness between H4 (181 Ma) and H5 (165.6 Ma), maximum throw is 438 m and length is 41,100, D) Isochron showing the thickness between H5 (165.6 Ma) and H6 (142.3 Ma), maximum throw is 993 m and length is 42,100, E) Isochron showing the thickness between H6 (142.3 Ma) and H7 (137.7 Ma), maximum throw is 1,098 and length is 21,900 m. All throw values could be underestimated up to 20% due to post-depositional compaction of faulted strata (Taylor et al., 2008).

4.4.1. Fault 1

4.4.1.1. Observations

Fault 1 (F1), the largest fault in the data set that has both of its tips imaged, is ~42 km long, strikes N-S, and dips to the E. Based on along-strike changes in strike and throw (Figure 4.6c), we split F1 into a 24 km- long northern segment and an ~18 km-long southern segment (Figures 4.5 and 4.6). The upper tip-line of F1 is located in Lower Cretaceous strata, where it physically links to a tier of polygonal faults (Velayatham et al., 2019) (Figure 4.4), and its lower tip-line is difficult to locate due to poor seismic imaging in the pre-rift, but F1 appears to tip out deep in the study area or into the basement (Figure 4.4). F1 shows two clear throw maxima; a northern maximum ($680 + 136$ m s TWT or $1,000 + 200$ m at H1; error shows possible decompaction) near the center of the northern fault segment, and a southern maximum ($433 + 87$ m s TWT, $658 + 132$ m at H1) near the center of the southern fault segment. Both segments are generally characterized by approximately bell-shaped throw distributions, the peak of which is skewed away from the center due to the related throw maxima being offset from the fault segment center (Figure 4.6c). There are clear wedge-shaped stratigraphic packages between H1 and H7 in the hanging wall; these thicken toward F1. In contrast, pre-H1 and post-H7 strata are isopachous (Figure 4.6a). EI plots show across-fault thickening (i.e., values ≥ 1) in Unit 1 along the central parts of the northern and southern segments; the unit is, however, isopachous where the two segments link (Figure 4.7a). In contrast, EI values in Unit 2 are ≥ 1 across the link between the two segments (Figure 4.7b). EI values are ≥ 1 across a progressively longer portion of the fault in Units 2–5 (Figures 4.7b–7d), until the youngest interval, Unit 6 (Figure 4.7e), where the upper tip of the fault is associated with EI values < 1 .

4.4.1.2. Interpretations

We see across-fault thickening in the hanging wall between H1 and H7 (Units 1–5) in cross section (Figure 4.6a) and in isochron thickness maps (Figures 4.7a–7e), suggesting F1 was active from 209.5 to 137.7 Ma (Early Jurassic–Early Cretaceous). In detail, however, the EI plots show that different parts of the fault were active at different times. The fact that F1 is associated with two discrete throw maxima (and two associated bell-shaped throw distributions), as well as an EI of < 1 in the middle of the fault in the first time interval (Unit 1; Figure 4.7a), suggests it formed by the linkage of two, initially separate segments. Linkage likely occurred sometime between the deposition of H2 and H4, based on EI values of > 1 only occurring in units above H1. Often, when faults link, their paleo-tip-lines become

inactive (Childs et al., 2003). In this case, however, F1 is a footwall-breached relay and the tip of the northern segment continued to accrue displacement on portions of the fault tips bounding the now-breached relay ramp (Figure 4.7e). The lack of throw in the middle of the fault is likely due to the still-active northern segment paleo-fault tip accommodating strain in the middle of the fault, as well as a minor E-W fault (labelled F1a in Figure 4.7a) that cuts perpendicularly across F1. F1 reached its maximum length by the deposition of H5 (Unit 3), or possibly sooner, based on the observation of EI values ≥ 1 across its length for this interval (Figure 4.7c). During Unit 5, the lateral ends of the fault have an EI value of <1 , which suggests that the fault tips became inactive at this time. Additionally, during this last stage of fault growth, the breached relay ramp between the northern and southern segment had an EI value of <1 , which suggests that the fault along the previously active relay ramp between the two fault segments became inactive (Figure 4.6e).

In summary, according to throw backstripping and EI analysis, F1 initiated after the deposition of H1 (c. 209.5 ma), and within c. 9.4 Myr (13% of its total life) consisted of two separate segments that were 19.7 and 19.2 km long. During this first phase of activity, it accumulated only 7% of its total throw. Approximately 18.8 Myr later, the two segments linked and the outermost tips of the newly formed fault system had propagated slightly, meaning it was now 41.3 km long. The fault had therefore reached c. 98% of its maximum length and accrued 19% of its total throw by this point (i.e., 39% of its life). Its maximum length was reached 15.7 Myr later, by which time it had accumulated 36% of its total throw. During the last 4.6 Myr of the fault's life, the remaining throw was accrued, and the northernmost 0.9 km and southernmost 2.9 km became inactive.

4.4.2. Fault 2

4.4.2.1. Observations

Fault 2 (F2) is a 19 km long, NNE-SSW-striking, WNW-dipping normal fault. F2 comprises a long (11 km) central segment that is linked at each end via abrupt bends, to shorter (3–5 km) segments (Figures 4.8 and 4.9). The upper tip-line of F2 is located in Lower Cretaceous strata, and its lower tip-line is difficult to locate due to poor seismic imaging in the pre-rift, but F2 appears to tip out deep in the study area or into the basement (Figure 4.4). F2 presently has three local throw maxima; a central maxima (380 + 76 m s TWT, 620 + 124 m at H1) on the main, central segment, a southern maxima (348 + 70 m s TWT, 560 + 112 m at H1) that is located along the southern fault segment, and a smaller, northern maxima (94 + 19 m s TWT, 150 + 30 m at H1) on the northern fault segment (Figure 4.8b). The throw maxima are

separated by two throw minima that coincide with the abrupt bends in the map-view trace of F2, where the northern and southern segments connect with the central segment (Figure 4.9). The main segment has an overall symmetrical throw distribution, and the northern and southern segments are skewed to the south and north respectively (Figure 4.8b).

There are clear wedge-shaped stratigraphic packages between H1 and H7 in the hanging wall, which thicken toward F2. Pre-H1 and post-H7 strata are isopachous (Figure 4.8a). EI plots show values ≥ 1 along a progressively longer portion of the fault from the oldest to the second youngest stratigraphic intervals (Units 1–4; Figures 4.9a–9d). In Unit 5, the lateral tips of the fault have an EI value that is < 1 (Figure 4.9e).

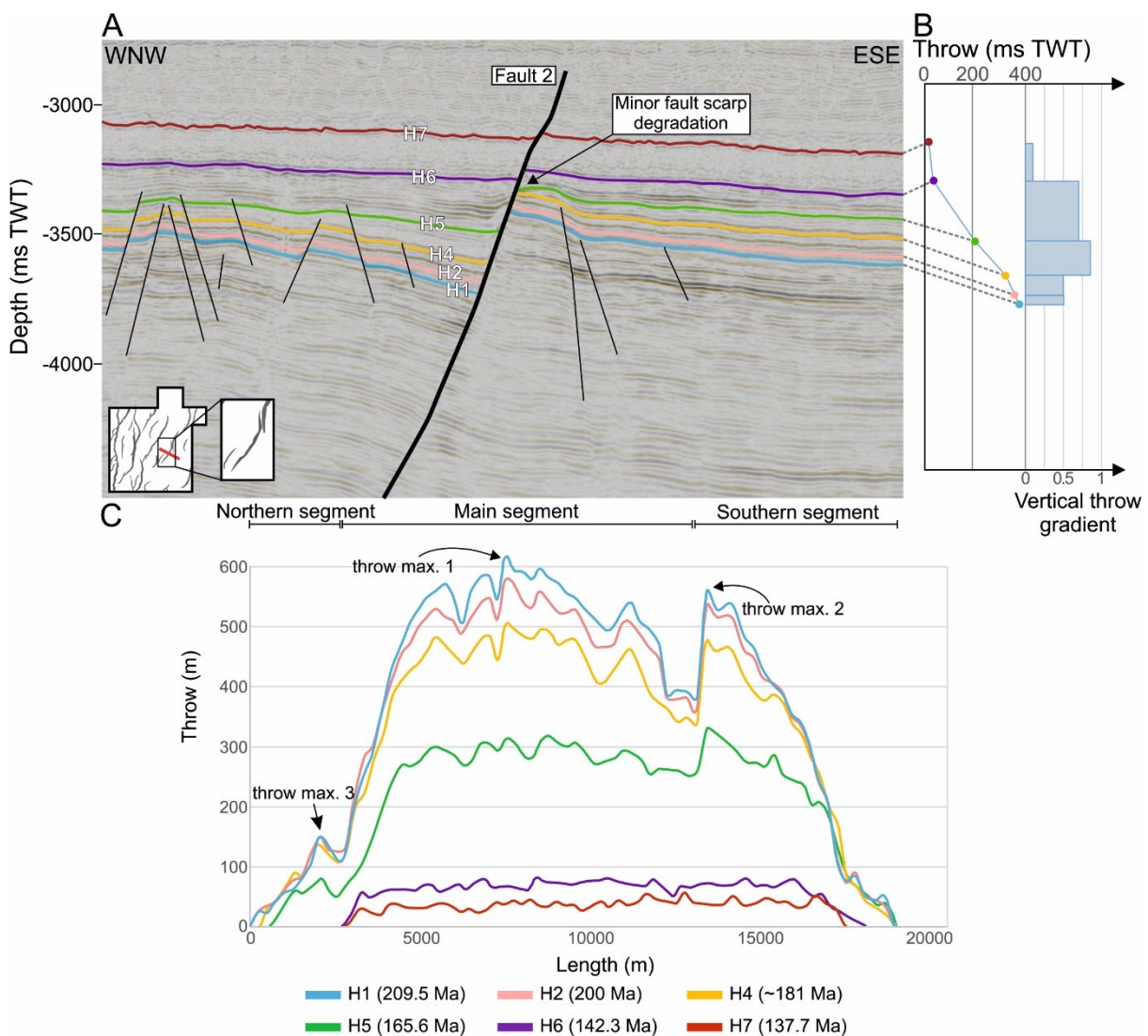


Figure 4. 8. A) Seismic profile illustrating F2 at its point of highest throw and its correlated throws and throw gradients, B) Vertical throw gradients for each horizon, C) Throw-distance plot illustrating the lateral variations in throw across each seismic unit. All throw values could be underestimated up to 20% due to post-depositional compaction of faulted strata (Taylor et al., 2008).

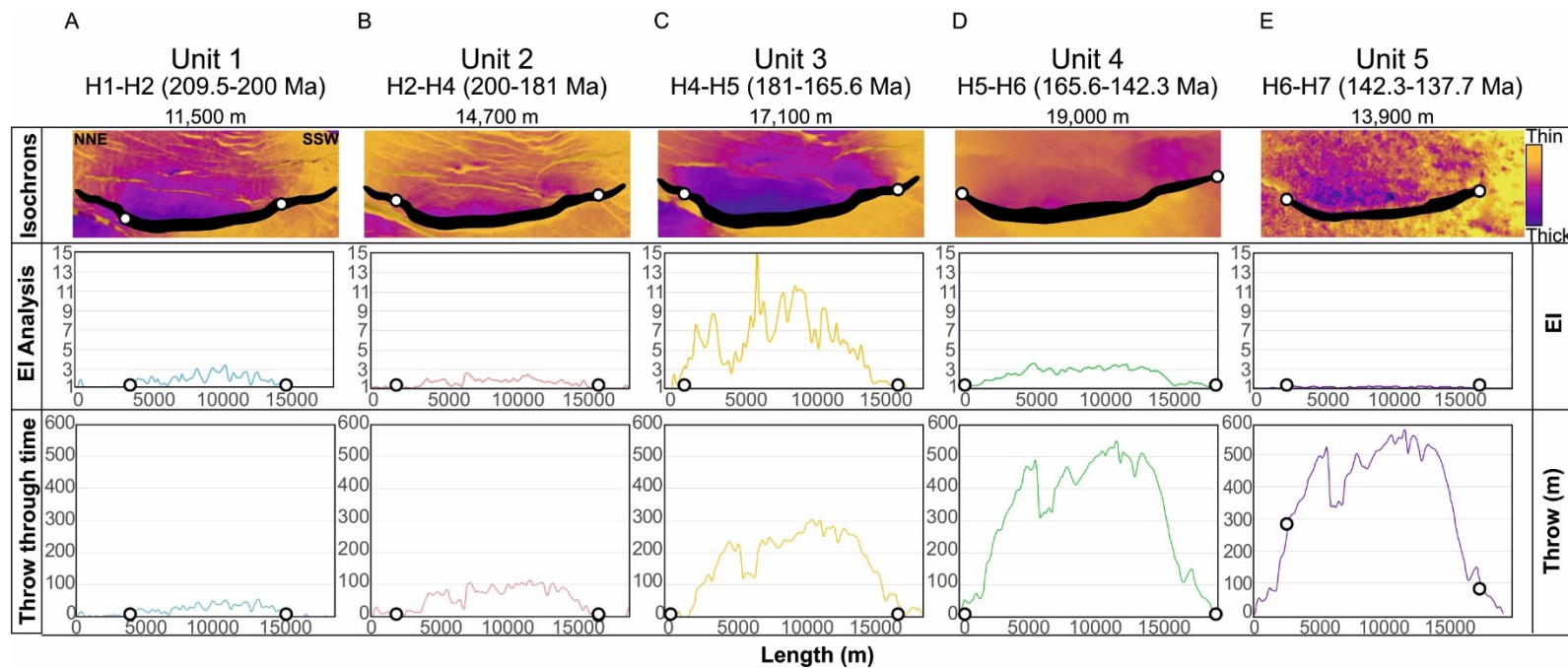


Figure 4. 9. Isochrons, expansion index analysis, and throw throughout different stages of the life of F2. The throw through time values are taken from throw backstripping, which can be seen in detail in the appendix figures. White dots indicate the length of the fault at the specified interval. A) Isochron showing the thickness between H1 (209.5 Ma) and H2 (200 Ma), maximum throw is 55 m, and length is 11,500 m, B) Isochron showing the thickness between H2 (200 Ma) and H4 (181 Ma), maximum throw is 116 m and length is 14,700 m, C) Isochron showing the thickness between H4 (181 Ma) and H5 (165.6 Ma), maximum throw is 303 m and length is 17,100, D) Isochron showing the thickness between H5 (165.6 Ma) and H6 (142.3 Ma), maximum throw is 550 m and length is 19,000, E) Isochron showing the thickness between H6 (142.3 Ma) and H7 (137.7 Ma), maximum throw is 617 and length is 13,900 m. All throw values could be underestimated up to 20% due to post-depositional compaction of faulted strata (Taylor et al., 2008).

4.4.2.2. Interpretations

We see across-fault thickening between H1 and H7 (Units 1–5) in cross-section (Figure 4.8a) and in isochron thickness maps (Figures 4.9a–9e), suggesting F2 was active from 209.5 to 137.7 Ma (Early Jurassic–Early Cretaceous). The fact that EI values ≥ 1 are limited to the central segment of the fault in Unit 1 (Figure 4.9a) suggests that F2 initiated here, an interpretation that is supported by the symmetry of the throw distribution on this segment (Figure 4.8b). The shorter southern segment was clearly present and active by Unit 3 (Figure 4.9c) and possibly already by Unit 2 (Figure 4.9b) times, as evidenced by EI values ≥ 1 along these segments in the corresponding interval. The throw maxima on the southern segment is skewed toward the NNE (Figure 4.8b), which is interpreted as a result of the mechanical interaction of the southern section with the already-existing central segment (Wilkinson et al., 2015). The northern segment was present and active by Unit 3 times (c. 28.2 Myr), based on the observation of EI values ≥ 1 along the segment (Figure 4.9c). This northern segment may have simply formed due to lateral (i.e., north-northeastward) propagation of the northern tip of the central segment. However, our preferred interpretation is that it initiated as a separate segment, based on: (i) the observed EI distribution within Unit 4 (i.e., the EI peak is located centrally along the SSW segment; Figure 4.9c); (ii) the fact the throw maximum is offset to the SW of the center of the mapped trace of the northern segment (Figure 4.8b); and (iii) the pronounced bend between the central and northern segments, which we infer reflects a now-breached relay ramp (Peacock & Sanderson, 1994; Walsh et al., 1999). F2 reached its maximum length by the deposition of H5, or possibly sooner, as evidenced by EI ≥ 1 across its length during this interval. EI values drop below 1 on the lateral tips of F2 in Unit 5, which we interpret as the outer 2–2.5 km of the fault becoming inactive (Figure 4.9e).

In summary, according to throw backstripping and EI analysis, F2 initiated after the deposition of H1 (c. 209.5 ma), and within c. 9.4 Myr (13% of its total life) was 11.5 km long (60.5% of maximum length). During this first phase of activity, F2 only accumulated c. 13% of its total throw. Approximately 18.8 Myr later, F2 had grown via tip propagation and possibly segment linkage to be 14.7 km long. At this time, the fault had reached 77.4% of maximum length and only 20% of total throw by this point (39.3% of the faults life). C. 15.7 Myr later (i.e., 50.2% of the total life), the central segment of F2 grew via segment linkage to be 17.1 km long, and accumulated c. 52.4% of total throw. Its maximum length was reached within the next 23.3 Myr, by which time it had accumulated c. 95.1% of its total throw.

During the last 4.6 Myr of the fault's life, the remaining 4.9% of throw was accrued, and the northernmost 2.6 km and southernmost 2.4 km of the fault became inactive.

4.4.3. Fault 3

4.4.3.1. Observations

Fault 3 (F3) is an 8.8 km long, NNE-SSW-striking, ESE-dipping normal fault. Its plan-view geometry consists of a slightly curved, convex-into-the-footwall segment with a small (1 km) fault branch near its northern tip (Figures 4.10 and 4.11). The upper tip-line of F3 is located in Lower Cretaceous strata, and its lower tip-line is difficult to locate due to poor seismic imaging in the pre-rift, but F3 appears to tip out deep in the study area (Figure 4.10). The present-day throw distribution for F3 shows two throw maxima; the main maxima (165 + 33 m, 83 + 17 m s TWT at H4) is located in the center of the main segment, with another, more minor maxima (59 + 12 m, 37 + 7 m s TWT at H4) being associated with a possible northern segment (Figure 4.10b).

There are wedge-shaped stratigraphic packages between H4 and H6 in the hanging wall, which thicken toward F3. In contrast, pre-H4 and post-H6 strata are isopachous (Figure 4.10a). EI values are ≥ 1 across a progressively longer portion of the fault from Unit 1–2 (Figures 4.11a and 4.11b), and in Unit 3, the outer tips of the fault have EI values < 1 (Figure 4.11c).

4.4.3.2. Interpretations

We see across-fault hanging wall thickening between H4 and H6 (Units 1–3) in cross section (Figure 4.10a) and in isochron thickness maps (Figures 4.11a–11c), suggesting F3 was active from c. 181 Ma to 142.3 Ma (Early Jurassic–Early Cretaceous). F3 likely initiated along its central segment during Unit 1 and reached its maximum length by the time of deposition of Unit 2 (Figure 4.11b); this is clearly evidenced by EI values ≥ 1 along the fault's entire trace-length. Together with the overall bell-shaped (present) distribution of throw, these EI data (Figure 4.11) suggest F3 grew as a single fault segment, or possibly as one large fault segment that linked with a very small segment at its northern tip. During Unit 3, EI values were < 1 on the northern-most part of the fault, suggesting that the F3's northern tip became inactive (Figure 4.11c).

In summary, according to throw backstripping and EI analysis, F3 initiated after the deposition of H4 (c. 181 Ma), and within c. 15.7 Myr (40% of its total life) was c. 6.2 km long (70.5% of maximum length). During this first phase of activity, it accumulated only 18.9% of its total throw. Approximately 13.6 Myr later, F3 propagated to its maximum length

of 8.8 km. The fault had therefore reached its maximum length and accrued 67% of its total throw by this point (i.e., 75% of its life). During the last c. 9.7 Myr of the fault's life, the remaining 33% of throw was accrued, and the length of the fault shortened by 600 m on the NNE tip of the fault.

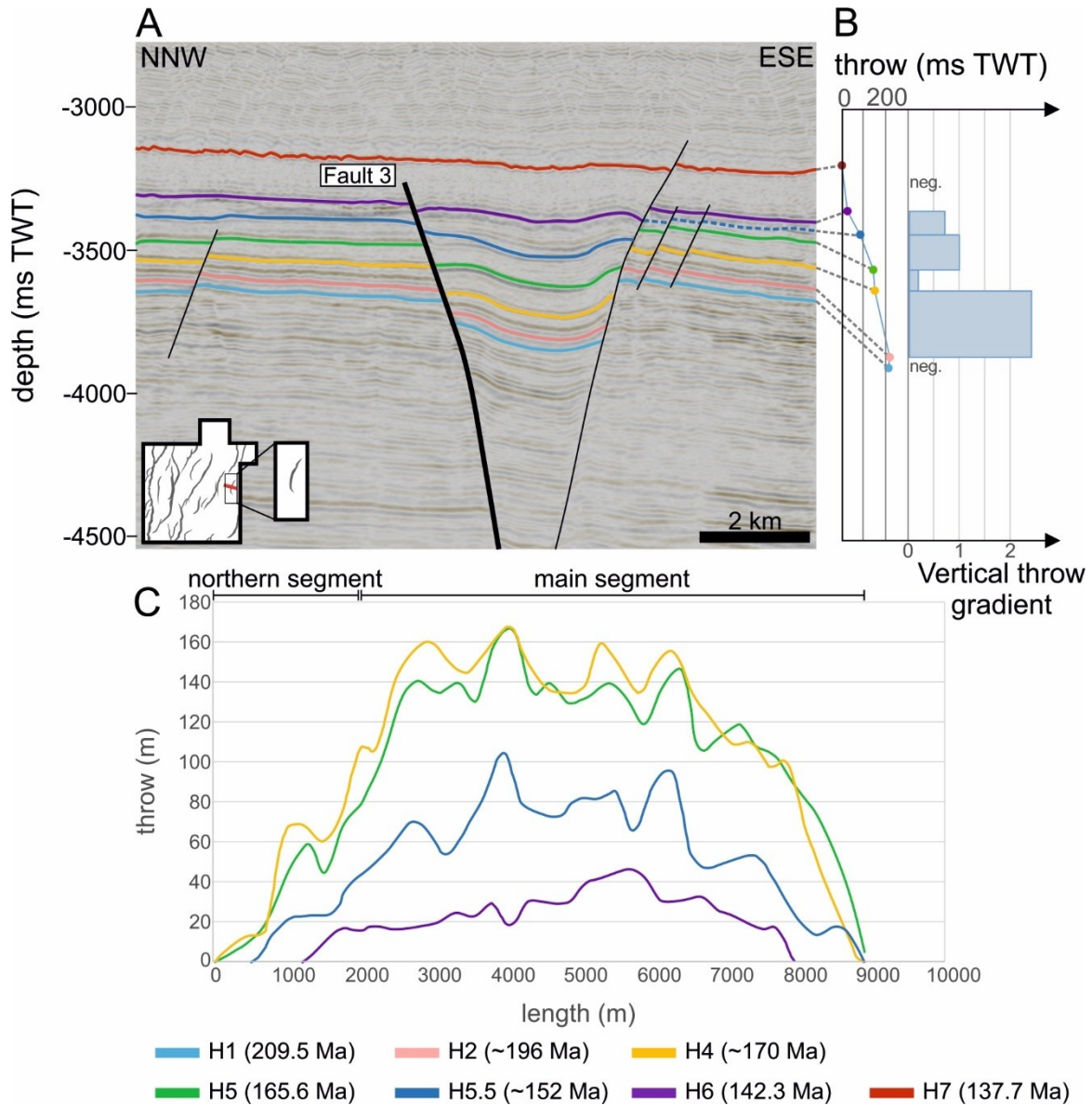


Figure 4. 10. A) Seismic profile illustrating F3 at its point of highest throw and its correlated throws and throw gradients, B) Vertical throw gradients for each horizon, C) Throw-distance plot illustrating the lateral variations in throw across each seismic unit. All throw values could be underestimated up to 20% due to post-depositional compaction of faulted strata (Taylor et al., 2008).

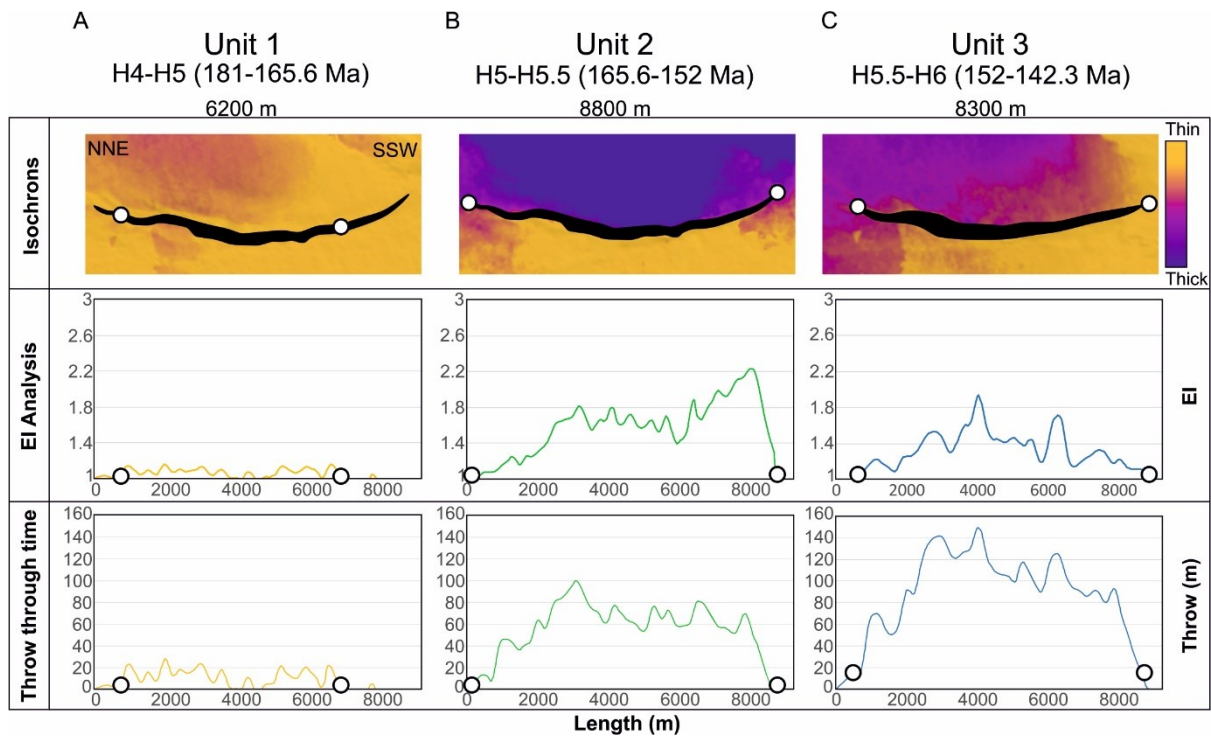


Figure 4. 11. Isochrons, expansion index analysis, and throw throughout different stages of the life of F3. The throw through time values are taken from throw backstripping, which can be seen in detail in the supplementary figures. White dots indicate the length of the fault at the specified interval. A) Isochron showing the thickness between H4 (181 Ma) and H5 (165.6 Ma), maximum throw is 28 m, and length is 6,200 m, B) Isochron showing the thickness between H5 (165.6 Ma) and H5.5 (152 Ma), maximum throw is 100 m and length is 8,800 m, C) Isochron showing the thickness between H5.5 (152 Ma) and H5 (142.3 Ma), maximum throw is 149 m and length is 8,300 m. All throw values could be underestimated up to 20% due to post-depositional compaction of faulted strata (Taylor et al., 2008).

4.4.4. Temporal Evolution of Throw and Length

All of the seven studied faults seem to have grown in three distinct stages: a lengthening stage, a throw accumulation stage, and a tip retreat stage. All of the faults had an early (i.e., first 20%–30% of their lives) relatively rapid lengthening phase, during which time they reached 60%–95% of their maximum length (see Figure 4.5a for isochrons across the study area, Figure 4.12a for values). Fault tips then grew slowly via tip propagation or segment linkage, reaching their maximum lengths after 57%–93% of their lives (Figures 4.5b and 4.5c). After they reached their maximum length, all the faults experienced a stage during which their overall at-surface trace-lengths reduced by up to 2.5 km (up to 25% of their total length) (Figures 4.12a and 4.12b). Three-stage fault growth can also be seen in additional faults in Figure 4.5.

Because of these shared kinematics, all of the faults studied displayed similar temporal changes in their throw-length scaling relationship (Figures 4.12b and 4.12c). The three-stage kinematics identified above also correlated with changes in throw (Figures 4.12b and 4.12d) and slip rate (Figure 4.12e). For example, during the lengthening stage, slip rate was

relatively low ($1.8 + 0.51\text{--}9.7 + 2.0$ m/Myr). During the subsequent throw accumulation stage, there was an abrupt increase in slip rate (to $5.3 + 1.0\text{--}23 + 4.6$ m/Myr). Slip rate decreased slightly ($6.5 + 1.2\text{--}20.7 + 4.8$ m/Myr) as the faults died. It should be noted that the slip rate during the first half of the fault's life (until deposition of H5) is likely underestimated due to the basin being somewhat sediment starved during this period. A higher slip rate would mean the time difference between the lengthening and throw accumulation stages is smaller.

4.5. Discussion

4.5.1. Implications for fault growth models

Our study identifies three key stages of fault growth on the Exmouth Plateau, offshore NW Australia. First, there was an initial lengthening stage; all of the faults reached 60%–95% of their maximum length within the first 20%–30% of their lives (Figure 4.13a). Maximum length was later reached via tip propagation or segment linkage. Second, there was then a throw accumulation stage that lasted from ~30%–75% of the faults' lives; during this time, faults lengthened very little and experienced an increased slip rate (Figure 4.13b). Third, the tip retreat stage, which lasted for the final 25% of the faults' lives, and during which the faults experienced tip-line retreat and throw was partitioned toward the center of the fault (Figures 4.13c and 4.13d). Our findings are generally consistent with the model of Rotevatn et al. (2019), with two exceptions. First, the fault maximum length is not always reached during the initial lengthening stage; that is, our results demonstrate that, while the bulk of lengthening happens relatively quickly, 5%–40% can subsequently occur during lateral tip propagation and/or segment linkage. Variations in when a fault reached its maximum length was likely controlled by whether a fault links with a nearby segment or not, a process perhaps dictated by the ability of the faults to breach intervening relays. Second, there was a stage of tip retreat, a behavior characterizing the end of life of all the studied faults (Figures 4.13c and 4.13d). Further work on normal faults imaged in 3D seismic reflection data may reveal this is a more common aspect of normal fault behavior than currently thought, meaning this stage of fault growth or more precisely, death, could be included in general fault evolution models (Nicol et al., 2020).

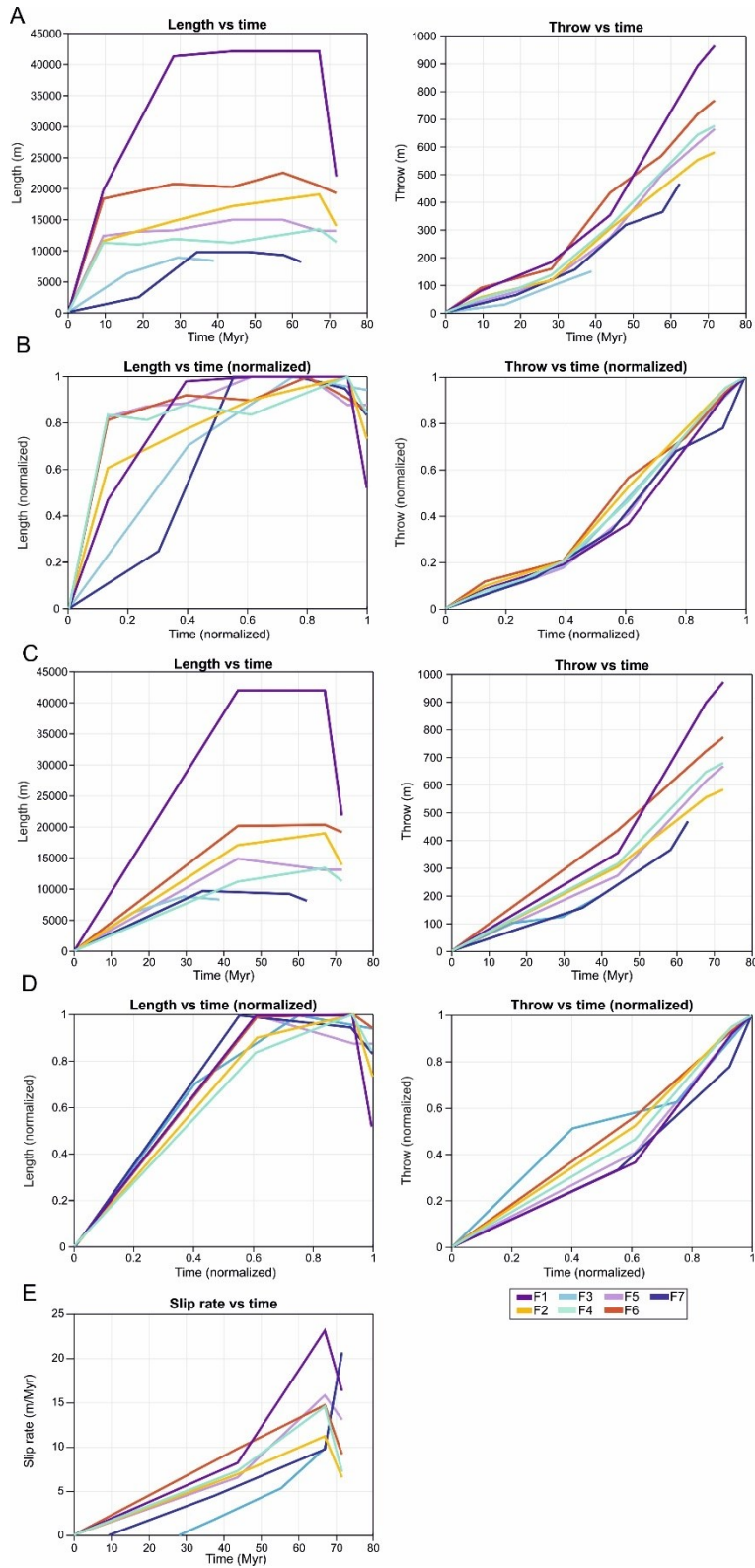


Figure 4. 12. Throw and lengthening through time for F1-7 in time and normalized. The lengthening, throw/displacement, and tip retreat stages of faulting are labeled in the normalized graphs. (a–b) Throw and lengthening stage; that is, our results demonstrate that, while the bulk of length through time, including all studied horizons, including H2-4 which are not age constrained. (c–f) Throw, length, and slip rate through time, only including horizons that have been directly age-constrained. All throw values could be underestimated up to 20% due to post-depositional compaction of faulted strata (Taylor et al., 2008).

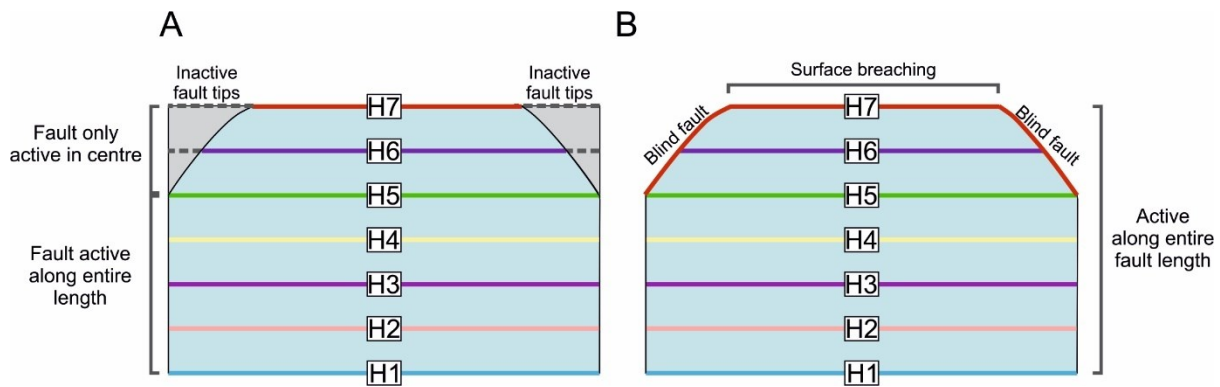


Figure 4. 13. Schematic models showing the phases of fault growth. A) Fault lengthening stage, when faults establish their near-maximum length within the first 30% of their lives B) Fault throw accrual stage, where faults have already reached their maximum length, and throw rate increases C) the beginning of the tip retreat stage, where the tips of the fault become inactive and D) the continuation of the tip retreat stage, where the active fault trace line is progressively shorter.

4.5.2. The role of tip retreat

Fault tip retreat was present on one or both tips of all of the faults in our study. This process has, however, only very rarely been described. Meyer et al. (2002) note a stage of tip retreat on Tertiary normal faults in the Vulcan Sub-basin, NW Shelf, Australia. The reasons why fault tips might retreat may have been overlooked because of a historical focus on how normal faults grow as opposed to how they die, and/or because high-quality, age-constrained seismic reflection data, with numerous mappable horizons within fault-related growth strata, are not available.

All of the studied faults decreased in length (by up to 2.5 km, or 25% of their trace-length) during the last 14 million years (25%) of their lives by retreat of one or both of their lateral tips. This could be explained by two hypotheses. The first hypothesis is that the faults were experiencing late-stage true tip retreat; that is, strain became localized near the fault center, leading to progressively shorter surface trace. We would expect that fault surface ruptures shortened as the fault gets closer to death (Figure 4.14a). An alternate hypothesis is that tip-line retreat was only apparent and was related to the faults having an elliptical geometry during the later stages of faulting, due to it having a plunging upper tip-line (i.e., during deposition of H5-7; Figure 4.14b). In this scenario, the fault would have intersected the free-surface along progressively shorter trace-lengths, with the fault tips being blind (Figure 4.13b). An increase in sediment accumulation rate relative to fault slip rate could drive this progression. The fault geometries associated with both hypotheses would look similar in seismic data (Figures 4.14a and 4.14b). We argue that we are seeing true fault tip retreat because if retreat was only apparent and related to the faults elliptical shape, we would expect vertical throw gradients across the horizons to be similar to those encountered on blind

normal faults (Childs et al., 2003; Meyer et al., 2002; Walsh & Watterson, 1988). For a blind fault, only modest strain can be accommodated by the rock volume without upward tip propagation.

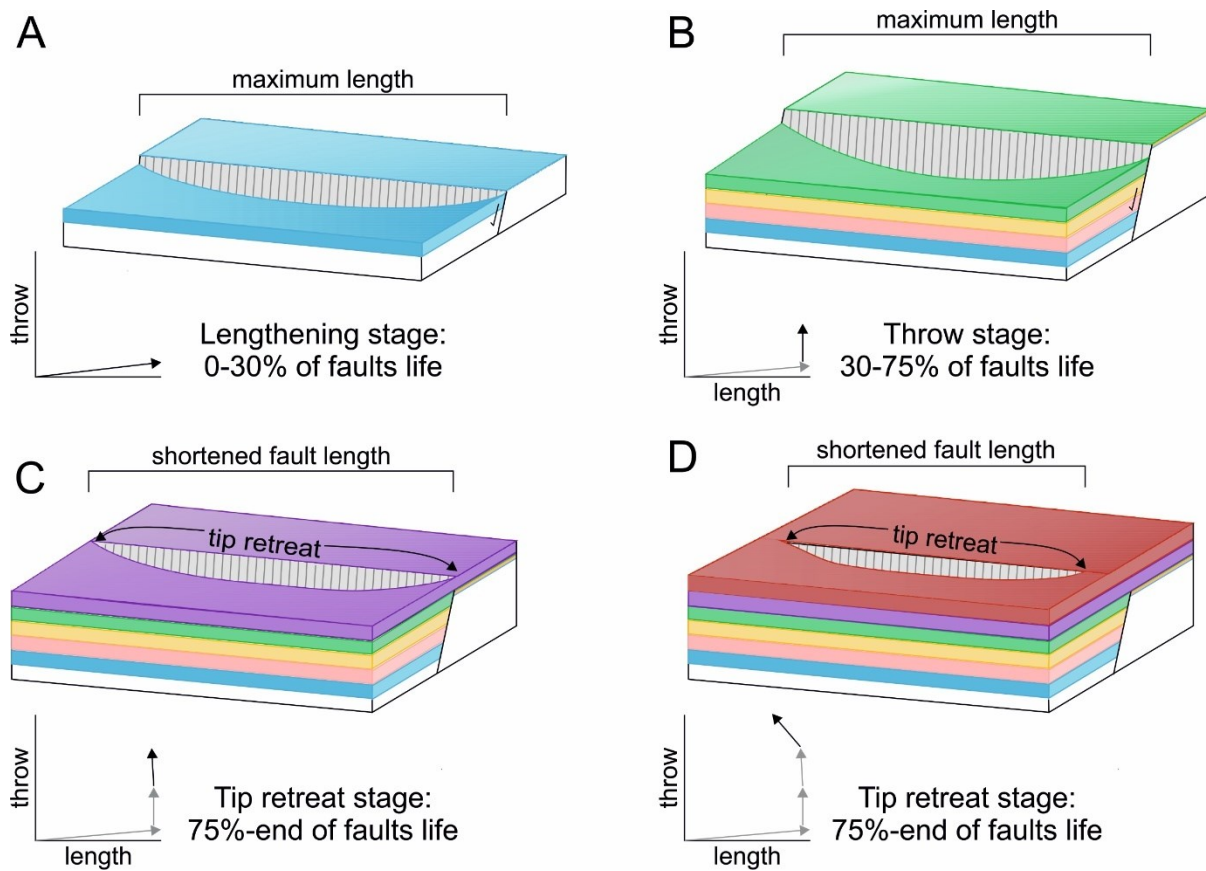


Figure 4. 14. Schematics showing two possibilities for the apparent tip line retreat in this study. Fault planes along dip are shown, and colored lines indicate the active length of the fault at the time of the deposition of the associated horizons (H1-7). A) Fault length remains constant from H1-5 and becomes active a progressively shorter distances across H6 and H7, which can be interpreted as lateral tip retreat. B) The fault remains active across the entire length of the fault, but in the later stages of faulting (H6-7) the fault only breaches the surface in the center of the fault, and the fault tips remain active at depth, acting as blind faults.

In the case of the upper tip of a blind fault, a maximum vertical displacement gradient of <0.1 is typical (Walsh & Watterson, 1988; Meyer et al., 2002; Childs et al., 2003; Baudon & Cartwright, 2008). We measured the maximum vertical displacement gradients between H6 and H7 in the center of the faults and found values between 0.08-4.5 (Figures 4.6, 4.8, and 4.10b), which is higher than that typically found for blind faults (Walsh & Watterson, 1988) (Figure 4.13). Such high vertical displacement gradients suggest that the entire length of the upper tip-line must have intersected the free-surface (Meyer et al., 2002; Childs et al., 2003; Baudon & Cartwright, 2008), and that a portion of the fault had become inactive during the later stages of the fault's life, before the deposition of H6 and H7.

When verifying that the tip retreat we see is real, it is important to ensure that the faults are not sediment-starved during the latter stages of their development (post-Jurassic). A reported example of possible tip retreat comes from the East African Rift (Morley, 2002); however, in this case it is possible that the faults became inactive earlier than assumed, and what appears to be tip retreat is only (passive) sediment filling of a starved basin. In our study, we argue the faults were not sediment-starved at the end of their lives (Berriasian-Hauterivian), given appreciable amounts of sediment were deposited in the footwall of the faults during deposition of H5 to H7 (Figures 4.6a, 4.7a, and 4.9a).

4.5.3. Slip rates and slip rate variability

We document a distinct, order-of-magnitude increase in slip rates from as little as 4 m/Myr to as much as 23 m/Myr, some 40–50 Mya after the studied faults initiated, around the time of the deposition of H5 (c. 165.6 Ma). These rates fall toward the lower end of long-term slip rates determined from the analysis of seismic reflection data imaging other natural extensional basins in a range of geodynamic setting (e.g., 4–1,000 m/Myr; Nicol et al., 1997). However, these rates are broadly consistent with rates calculated over the relatively long timescales (i.e., >40 Myr) considered here (i.e., 25 m/Myr; North Sea example in Figure 4.2 in Nicol et al., 1997). We note that the increase in slip rates in our NW Shelf examples temporally correspond to a time when many of the minor faults became inactive (see minor faults that tip-out below H5; Figures 4.4 and 4.5a–5c), when strain localized onto the large faults presently defining the basin structure. Similar relationships between increasing fault slip rates (or related subsidence rates) and strain localization are documented in several other natural rifts (e.g., Gawthorpe et al., 2003) and are reproduced in numerical models (e.g., Cowie & Roberts, 2001; Gupta & Scholz, 1998), with this relationship thought to reflect stress-feedback interactions during fault system growth. Long-term fluctuations in slip rate could also reflect changes in regional strain rate, related to the fundamental plate-driven processes driving deformation (Nicol et al., 1997; Mouslopoulou et al., 2009). More specifically, the marked increase in slip rates could reflect an increase in regional strain rate associated with rifting and, ultimately, continental break-up. However, we cannot independently constrain the rate of plate boundary processes during the time interval considered here (i.e., Middle Jurassic to Early Cretaceous), principally because the time-equivalent margin facing the NW Shelf is not preserved.

4.6. Conclusions

We use 3D seismic data from offshore NW Australia to study normal fault growth through time. We show that the majority of the studied faults had three distinct stages of fault evolution. During the first stage, the “lengthening stage,” the faults accumulated at least between 60%-95% of their final length and accrued between 10%–20% of throw. This stage lasted for up to 30% of the faults’ lives. The second stage, termed the “throw stage,” fault slip rate increased, and the remainder of maximum fault length was reached. We also suggest that these faults had a third stage of fault growth, the “tip retreat stage,” where the active trace line of the fault decreases by up to 25% and throw continues to be accrued. More evidence is needed to determine how prevalent tip retreat is, but it could be an important part of late-stage fault growth and possibly should be included in future fault growth models.

Chapter 5

Insights into normal fault kinematics using 4D analogue models

We are planning to submit this chapter to **Solid Earth** in the near future.

Lathrop, B. A., Zwaan, F., Schmid, T., Jackson, C. A. L., Bell, R. E., Rotevatn, A., Schreurs, G. (in prep). Insights into normal fault kinematics using 4D analogue models.

Acknowledgements

We thank the AAPG for funding used to carry out this research. The physical analogue modelling carried out in this chapter was conducted at the University of Bern, Switzerland. I thank Frank Zwaan, Timothy Schmid, and Guido Schreurs for the use of their facilities and their help and guidance throughout. I also thank Jean-Marie Mengus (IFP Energies Nouvelles) for converting my CT scan files to .SGY files.

Abstract

An increasing amount of research has gone into understanding how normal faults grow through time. The kinematic growth of normal faults is commonly described by two end-member models: 1) the propagating model, and 2) the constant-length model. The propagating model claims a sympathetic increase between fault lengthening and displacement. The constant-length model states that faults reach near-final length before accumulating significant displacement. A hybrid model of the two has also been suggested. We have observed faults growing via the constant-length model in previous studies using high-quality 3D seismic data, however we were unable to track early-stage fault growth (i.e., the first 30% of a fault's life) due to the limited resolution of seismic data.

Physical analogue models complement field and 3D seismic reflection-based studies, as they allow us to track fault growth in relatively homogenous materials. However, the analysis of such models is traditionally limited to visual inspection of overhead “plan-view” images of the surface, thus fault geometries at depth cannot be determined. We present results from a new, innovative modelling set-up that allows us to track fault growth in 4D. We used a 30x80 cm sandbox apparatus with a fixed base plate and two moving sidewalls controlled by precise computer-guided stepper motors. We utilised a brittle-ductile experiment set-up, which allowed strain to be localised as is observed in the upper crust in nature. Models were imaged at every 2 mm of extension with a medical-grade CT scanner as well as stereoscopic top-view time-lapse photography to quantify deformation by means of 3D Digital Image Correlation (3D DIC) throughout extension. We were able to see how fault segments interact and link via relay breaching, as well as how fault length and throw changes through time.

Our study showed that faults established their full vertical height quickly and grew via a cyclical growth pattern where faults alternate between quick lateral lengthening stages due to segment linkage and prolonged displacement stages. This step-wise growth pattern continued until the model was halted after 4 cm of extension. Faults that were optimally positioned to adjacent fault segments (i.e., synthetic boundaries with the adjacent fault and aligned along strike) linked and then continued to accrue displacement, while faults that were in stress shadows (i.e., places that are not optimal for additional growth via segment linkage) and faults that form conjugate boundaries at depth (i.e., faults that interact with opposite dips at depth) became inactive. This is the first time that fault length and displacement has been tracked in 3D at such close time intervals, made possible by the unique use of a CT scanner.

5.1. Introduction

Understanding how normal faults grow through time is important for understanding the distribution of extensional strain (e.g., Gupta et al., 1998; Cowie et al., 2000; Meyer et al., 2002; Walsh et al., 2003) and the tectonostratigraphic development of rift basins (e.g., Gawthorpe & Leeder, 2000; McLeod et al., 2002; Ge et al., 2017; Henstra et al., 2017; Jackson et al., 2017; Rotevatn et al., 2019). Physical analogue models offer a unique insight into normal fault growth and the development of rift basins, given they allow us to: (i) directly observe incremental changes in their geometry (e.g., displacement/throw and length) and kinematics (e.g., initiation, slip rate, and death); and (ii) assess the impact of various parameters on fault and rift development, including extension obliquity and the role of pre-existing structures (e.g., Mansfield & Cartwright, 2001; Hus et al., 2006; Filbrandt et al., 2007; Schlagenhauf et al., 2008; Whipp et al., 2016; Zwaan et al., 2016, 2019; Rotevatn et al., 2019). However, the analysis of physical analogue models is traditionally undertaken using overhead (i.e., plan-view) images of the model top surface (e.g., Schlagenhauf et al., 2008) meaning fault geometries and evolution at depth, temporal changes in the subsurface geometry of the fault network or individual faults, and how this might relate to their surficial expression (e.g., segmentation) cannot be addressed.

The new and innovative models presented herein allow us to track fault growth in three-dimensional space *and* through time (4D). We focus on how fault length and throw changes through time, and how these changes in geometric parameters relate to fault segment interaction, relay ramp development, and segment linkage. This is the first time that changes in fault length and throw have been tracked at such close time intervals, and in three dimensions; this is only made possible by the use of the CT scanner and overhead image analysis.

These models allowed us to identify that faults follow a growth pattern that alternates between lengthening and throw accrual. Initially, there are many densely spaced faults. As extension continues, optimally spaced and orientated faults grow via segment linkage and faults in stress shadows or with double conjugate boundaries become inactive. Growth by segment linkage is followed by an extended period of displacement accrual. This lengthening-throw pattern continues throughout extension.

5.1.1. Fault growth and fault growth models

The growth of normal faults has historically been described by two end-member models: the propagating fault model and the constant-length model. According to the propagating fault model, faults grow via a sympathetic increase in displacement/throw and length (e.g., Walsh & Watterson, 1988; Dawers et al., 1993; Cartwright et al., 1995; Walsh et al., 2003), and according to the constant-length model, faults reach their near-maximum length relatively quickly and then spend the remainder of their active lives accruing displacement/throw (e.g., Walsh et al., 2002, 2003; Nicol et al., 2005, 2020; Jackson & Rotevatn, 2013; Tvedt et al., 2016; Rotevatn et al., 2019). Several recent studies (Jackson et al., 2017; Rotevatn et al., 2019; Nicol et al., 2020), suggest that faults likely follow a hybrid model of the two, adhering to the propagating fault model for the first 20-30% of the fault's life and the constant-length model for the remainder (see also Finch and Gawthorpe, 2017). A step-wise fault growth pattern has also been suggested (Filbrandt et al., 2007; Pan et al., 2021), whereby faults grow via alternating phases of lengthening and displacement/throw, with faults in strain shadows becoming inactive as stress localises onto optimally placed faults (e.g., Cowie et al., 1995; Nicol et al., 1997; Cowie, 1998; Gupta & Scholz, 1998; Cowie et al., 2000; Walsh & Watterson, 2003; Pan et al., 2021).

Our use of physical analogue models complements work done on the natural faults in chapters 3-4 of this thesis and in previous works on normal fault kinematics. There are some limitations associated with using field and subsurface based data to characterise normal fault growth. Outcrops generally capture a snapshot of a fault's life, frozen in time. With 3D seismic studies, fault kinematics can be determined if a very specific set of conditions are met; normal growth faults must be imaged by high-quality seismic data and must be age-constrained by well data. The study area must not be sediment-starved in order to determine an accurate fault growth history. Even with ideal conditions in seismic, faults can only be imaged when they are large enough to be resolvable in the data, which is typically from ~10 meters of displacement for high-resolution data (Pickering et al., 1997), and as a result only the later stages of fault growth can be imaged. Our physical analogue models can see more stages of fault growth than that seen in natural faults.

5.1.2. Relay ramp evolution and geometric analysis

As fault segments grow, they will invariably approach and interact with adjacent fault segments (e.g., Peacock and Sanderson, 1994; Crider and Pollard, 1998; Cowie et al., 2000; Gupta & Scholz, 2000; Fossen & Rotevatn, 2016; Peacock et al., 2016). The development of

overlapping lateral fault tips is an important stage of fault growth as it often occurs during fault growth via segment linkage. Fault tips can be described as underlapping, non-overlapping, and overlapping, defined according to their map-view location (Figure 5.1a; e.g., Crider & Pollard, 1998; Gupta & Scholz, 2000; Hus et al., 2005; Peacock et al., 2016; Childs et al., 2017a). The tips of underlapping fault segments under-step (Figure 5.1a). Non-overlapping fault tips approach each other and are aligned perpendicular to the strike of the fault, but do not overlap (Figure 5.1a). Overlapping fault tips overlap perpendicular to strike, often forming relay ramps (Figure 5.1a). The boundaries between fault segments can be described as either synthetic boundaries, conjugate boundaries, or free tips (Figure 5.1b). Synthetic boundaries describe interacting fault segments with the same sense of dip direction; depending on how the segments relate to each other spatially, they could interact and link (Figure 5.1b; e.g., Morley et al., 1990; Gawthorpe & Hurst, 1993; Childs et al., 1995; Whipp et al., 2016). Conjugate (or antithetic) boundaries describe stepping fault segments with opposite dip directions; these types of fault interactions typically block fault growth by not allowing additional tip propagation or segment linkage (Figure 5.1b; e.g., Morley et al., 1990; Gawthorpe & Hurst, 1993; Childs et al., 1995; Whipp et al., 2016). Free tips are fault tips that are not interacting physically or mechanically with another fault and thus are free to propagate laterally (Figure 5.1b; e.g., Morley et al., 1990; Gawthorpe & Hurst, 1993; Whipp et al., 2016).

Relay ramp geometry is quantified by measurement of fault tip overlap (i.e., the space between the two faults along strike), spacing (i.e., the space between the two fault segments along dip), and aspect ratio (overlap/spacing) (Figure 5.1a; Crider & Pollard, 1998; Mansfield & Cartwright, 2001; Soliva et al., 2006; Long & Imber, 2011; Whipp et al., 2016; Childs et al., 2017a). Aspect ratios of relay ramps are known to range from values of 0.2 to 22, with an average value of 4.2, according to a global dataset of aspect ratios of normal faults across eight orders of magnitude compiled by Long & Imber (2011). Aspect ratios have been shown to be scale dependent and not affected by host rock lithology as long as it is homogenous; relay ramps in mechanical stratigraphy tends to result in higher aspect ratios (Long & Imber, 2011). Relay ramps associated with pre-existing basement structures also tend to have higher aspect ratios (Childs et al., 2017a). Relay ramps with high aspect ratios are taken to indicate that faults are connected at depth, which has implications for greater magnitude earthquake hazards (Soliva et al., 2008), thus understanding how at-surface relay ramp aspect ratio relates to subsurface fault geometry is vital.

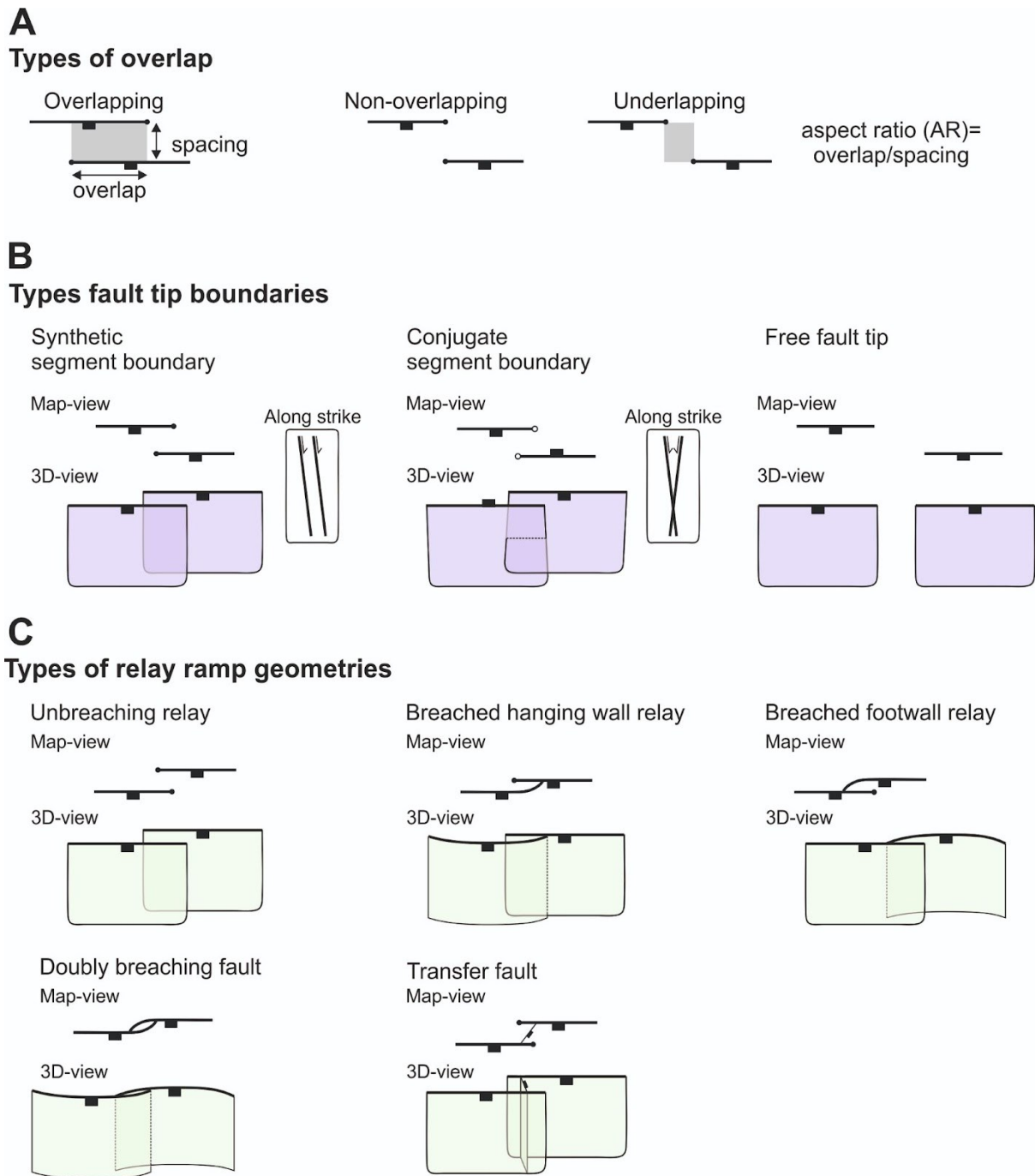


Figure 5. 1. Outline of different relay ramp geometries. A) Types of lateral fault tip overlap including overlapping, non-overlapping, and underlapping. B) Types of fault tip boundaries including synthetic segment boundaries, conjugate segment boundaries, and free fault tips, shown in map-view, 3D-view and along strike.

Relay ramps are zones of kinematic linkage between overlapping, geometrically uncoupled, sub-parallel faults, where strain is relayed from one fault to the other (Larsen, 1988; Peacock & Sanderson, 1991; Peacock et al., 2016). Relay ramp formation and subsequent breaching is the most efficient way for faults to lengthen (e.g., Soliva et al., 2004; Nicol et al., 2010; Fossen & Rotevatn, 2016). Faults can either be hard-linked or soft-linked (e.g., Walsh & Watterson, 1991; Trudgill & Cartwright, 1994; Childs et al., 1995; Fossen & Hesthammer,

1997; Kim & Sanderson, 2005; Fossen & Rotevatn, 2016; Nicol et al., 2020). Hard-linked faults are overlapping synthetic faults that are physically linked by a breached relay ramp (e.g., Walsh & Watterson, 1991; Trudgill & Cartwright, 1994; Fossen & Hesthammer, 1997; Childs et al., 1995; Fossen & Rotevatn, 2016; Whipp et al., 2016; Nicol et al., 2020). The relay ramps can be hanging-wall breached, footwall breached, double breached (i.e., both the hanging-wall and footwall breach) or transfer-fault breached (i.e., a fault forms between the two tips, physically connecting them) (Figure 5.1c; Trudgill & Cartwright, 1994; Cartwright et al., 1995; Walsh et al., 1999; Whipp et al., 2016). Soft-linked faults are overlapping synthetic faults with an unbreached relay ramp; that are kinematically linked but not physically linked (Figure 5.1c; e.g., Walsh & Watterson, 1991; Trudgill & Cartwright, 1994; Childs et al., 1995; Fossen & Hesthammer, 1997; Fossen & Rotevatn, 2016; Whipp et al., 2016; Nicol et al., 2020). Mansfield & Cartwright (2001) found that the optimum aspect ratio for relay ramps to breach is ~ 4 , however others have found that there is no relationship between aspect ratio and relay ramp breaching (Whipp et al., 2016).

Exp.	Type	Base	Dimensions	Sand thickness	Sidewalls	Extension rate	Extension interval
808	brittle only	8 cm foam	30 cm x 120 cm	4 cm	No	8 mm/hr	run straight through
809	brittle only	8 cm foam	30 cm x 120 cm	6 cm	No	8 mm/hr	run straight through
815	brittle-viscous	glass/foam	30 cm x 80 cm	6 cm	Yes	800 mm/hr	paused every 10 mm ext.
816	brittle-viscous	glass/foam	30 cm x 80 cm	4 cm	Yes	400 mm/hr	paused every 10 mm ext.

Table 5. 1. Table showing details of the 4 different model set-ups.

5.2. Methodology

5.2.1. Methodology for all models

We ran four experiments in order to see what set-up best replicated normal fault growth in nature (Table 1). Models were either single- or double-layer (two-layer), simulating a brittle upper crust or a brittle-ductile crust, respectively, following the methodology of Zwaan et al. (2019) (Table 5.1). Each of the brittle and brittle-ductile experiments were run twice; once with a 4 cm thick layer of fine-grained quartz sand and one with a 6 cm thick layer of fine-grained quartz sand (Table 5.1). The viscous layer, which models the viscous lower crust, is a 4 mm-thick, near-Newtonian viscous layer (viscosity η ; ca. 1.5×10^5 Pa s; stress exponent $n=1.05$), mixed at a 1:1 ratio of SGM-36 polymethylsiloxane (PDMS) silicone and corundum sand (ρ specific = 3950 kg^{-3} ; Panien et al., 2006; Zwaan et al., 2016, 2019). This silicone/corundum sand mixture has a similar density to that of quartz sand ($\sim 1560 \text{ kgm}^{-3}$), which prevents the viscous layer from buoyantly rising (Zwaan et al., 2016, 2019). This

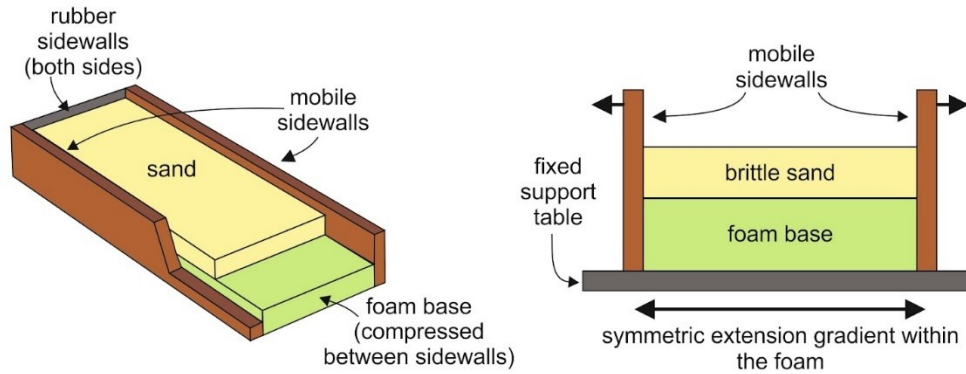
would happen if pure PDMS was used, as this has a lower density than sand ($\rho = 965 \text{ kgm}^{-3}$; Weijermars, 1986). Sand was sieved from ~30 cm above the model surface in order to guarantee a density of $\sim 1560 \text{ kgm}^{-3}$ and the sand was flattened with scraper at every 1 cm of thickness to create layering (Zwaan et al., 2019). We added a thin dusting of corundum sand every 1 cm marker to make the layering stand out enough that it that could be imaged by the CT scanner. The angles of internal peak and stable friction were 36.1° and 31.4° , respectively, for quartz sand and 37° and 32° for corundum sand (Panien et al., 2006; Klinkmüller, 2016; Zwaan et al., 2016); both types of sand have very similar mechanical properties, thus adding layers of corundum sand as a marker did not significantly alter the model development of more specifically, the fault geometry and kinematics (Zwaan et al., 2016).

The experiment apparatus used for these models was the same used in Zwaan et al. (2019) (Figure 5.2). The set-up consisted of a fixed base plate with two sidewalls that moved independently of each other, with the motion of the sidewalls controlled by a precise computer-guided stepper motor (Zwaan et al., 2019). The models had a relatively high width-to-length ratio (see below), which diminished the influence of boundary effects at the sidewalls (Zwaan et al., 2019).

In the brittle models (Figure 5.2a), the base of the model was an 8 cm-thick, RG 50 polyurethane foam base. The foam was first laterally compressed by 4 cm before the sand was set on top, thus when the sidewalls moved apart, the expanding foam caused uniform deformation in the overlying sandpack. The brittle experiments were 120 cm long and had an initial width of 30 cm (initial length-to-width ratio of 4). In the ductile models (Figure 5.2b), the base of the model consisted of alternating 1 cm strips of RG 50 polyurethane foam and 1 cm thick plastic strips. The silicone layer made the model heavy, so the plastic strips were needed to make sure the model did not sag in the middle. The brittle-ductile models had a length of 80 cm and an initial width of 30 cm (initial length-to-width ratio of 2.67). No seed was placed in either of these model set-ups to initiate rifting, as has been done in previous experiments using this same apparatus (Zwaan et al., 2016, 2019, 2021), and in other physical analogue models (Filbrandt et al., 2007; Liang et al., 2021). Models were extended 4 cm, giving the models a final measurement of 34 cm wide. The brittle models were extended at a rate of 8 mm/hr (i.e., 4 mm/hr on each sidewall) without stopping and scanned at every 2 mm of extension (Table 1). The viscous-brittle models were extended at higher rates (400 mm/hr and 800 mm/hr) because we found this was needed to form any structures within the

sandpack. Extension was briefly stopped every 2 mm of extension for the model to be scanned.

1. Foam base set-up



2. Foam+viscous layer set-up

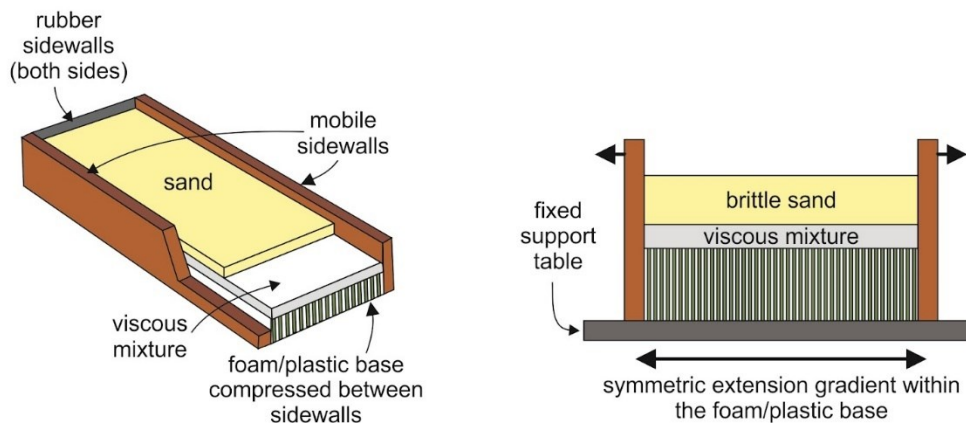


Figure 5. 2. Model set-ups for the two types of analogues. The upper figure shows foam base set-ups, and the lower figure shows the foam+viscous layer set-up.

The models were scanned in a medical grade CT-scanner, which provided 3D imaging of the models. The CT scans were 512 x 512 pixels with a resolution of ~0.6mm/pixel. The scans were converted into .SEG-Y files in order to load and view the models in Petrel seismic interpretation software and to analyse them as one would a 3D seismic reflection volume (i.e., with inlines, crosslines and depth slices). Fault throw and length were measured on the top model surface at every time step (every 2 mm of extension), starting from 22 mm of extension when the faults had high enough throw to be tracked and thus measured in the 3D CT scans. Relay ramp aspect ratios were measured at the elevation where overlap was highest for every 2 mm of extension for selected faults. Aspect ratios are used to see how overlapped fault segments are before, during, and after their relay ramps mechanically breach. Fault throw gradients were measured by dividing a throw maximum by the throw minimum at the

tip of the fault. Fault throw gradients were used to show how conjugate boundaries between faults can change fault throw-length profiles, as throw gradient increases when a fault tip is pinned (e.g., Peacock & Sanderson, 1991; Childs et al., 1995; Wilkins & Gross, 2002).

Cameras were positioned over the models to take time-lapse photos of the model surface throughout extension in order to conduct particle image velocimetry (PIV) analysis, which tracks particle movement on the top surface of the model throughout the experiment (Zwaan et al., 2019). We conducted this surface analysis because we are able to get higher-resolution images from the top of the model than from the CT scans. The overhead images were 8929 x 5286 pixels, with a resolution of ~ 0.12 mm/pixel. For the sake of simplicity, we use compass directions to describe the models, i.e., the top of the scanned model is north, and the bottom of the scanned model is south.

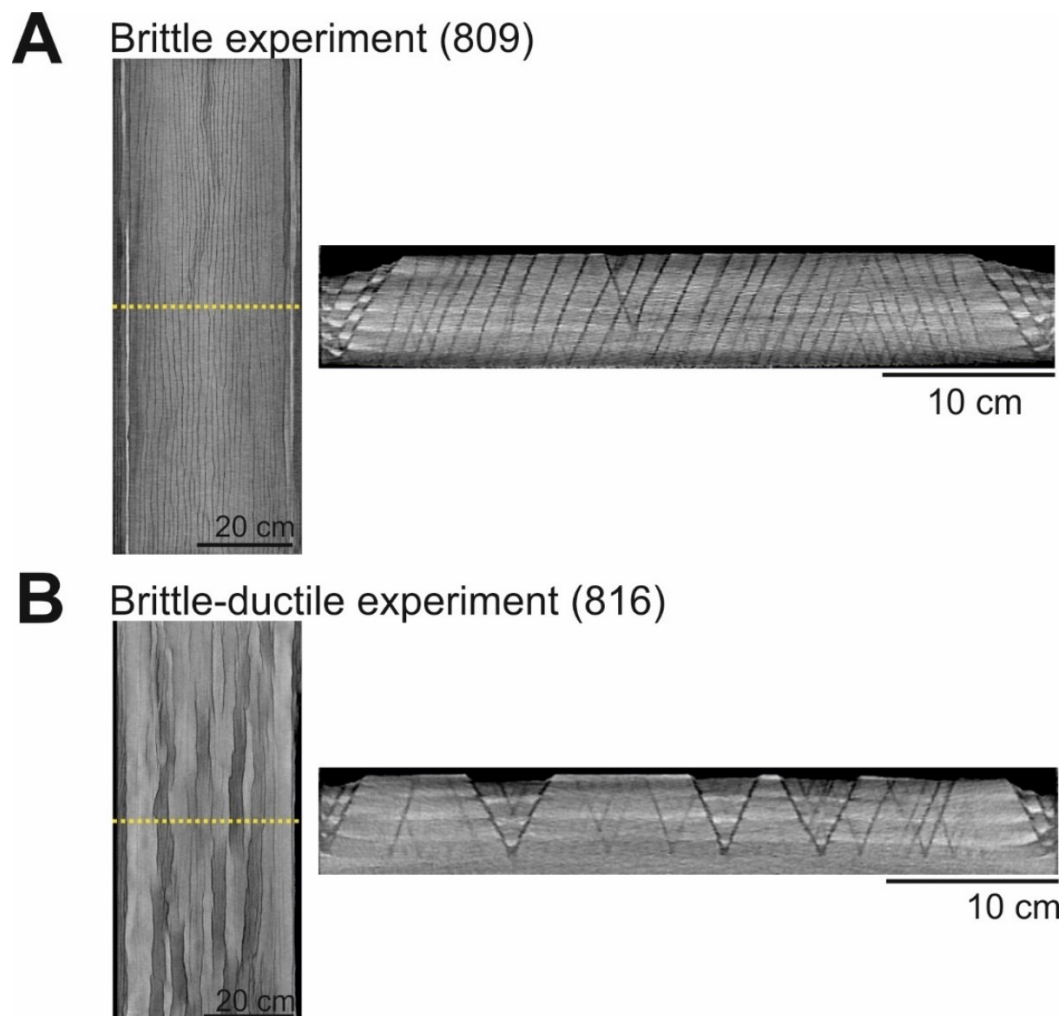


Figure 5. 3 Model set-ups for the two types of analogues. The upper figure shows foam base set-ups, and the lower figure shows the foam+viscous layer set-up

Experiment 816

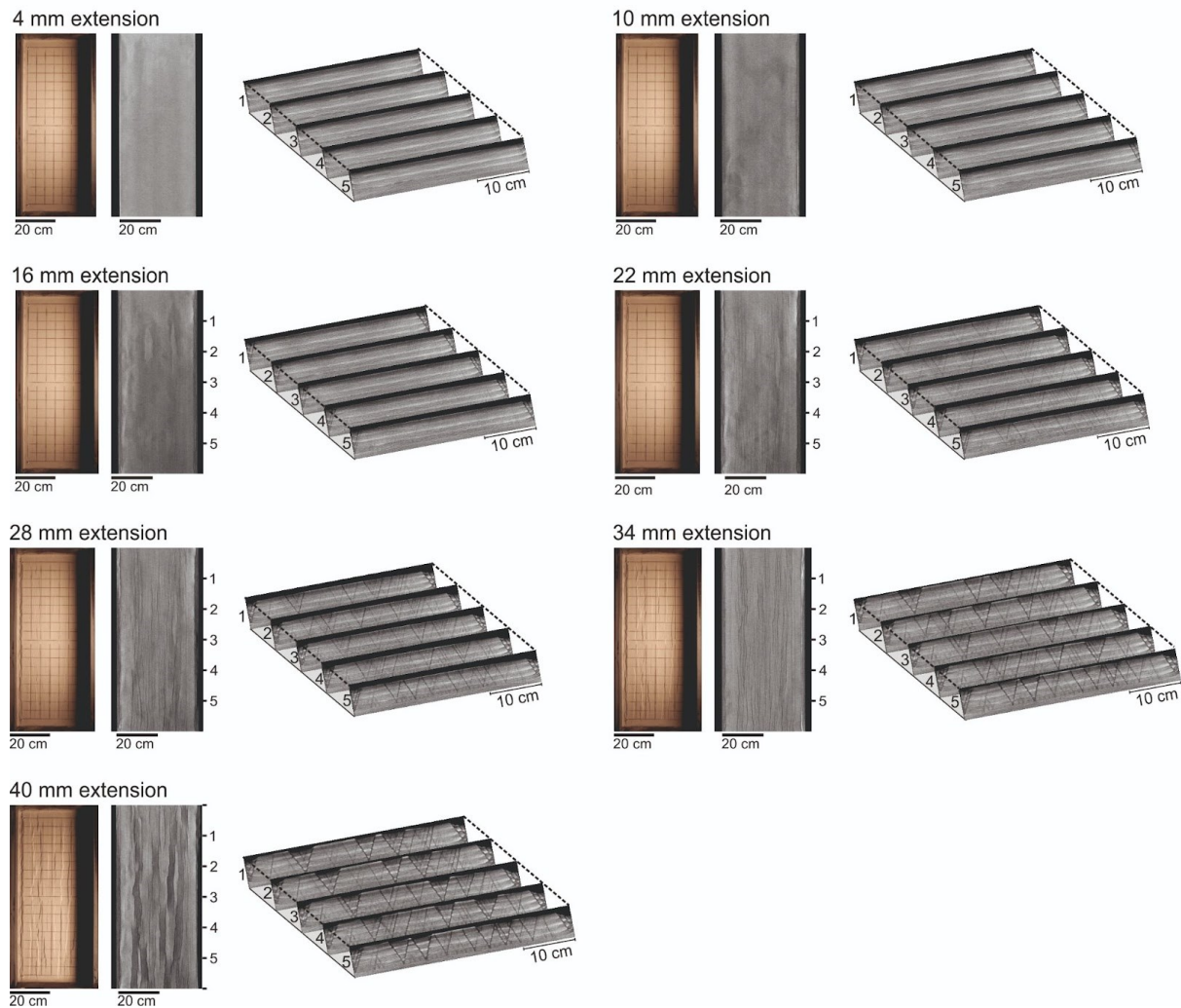


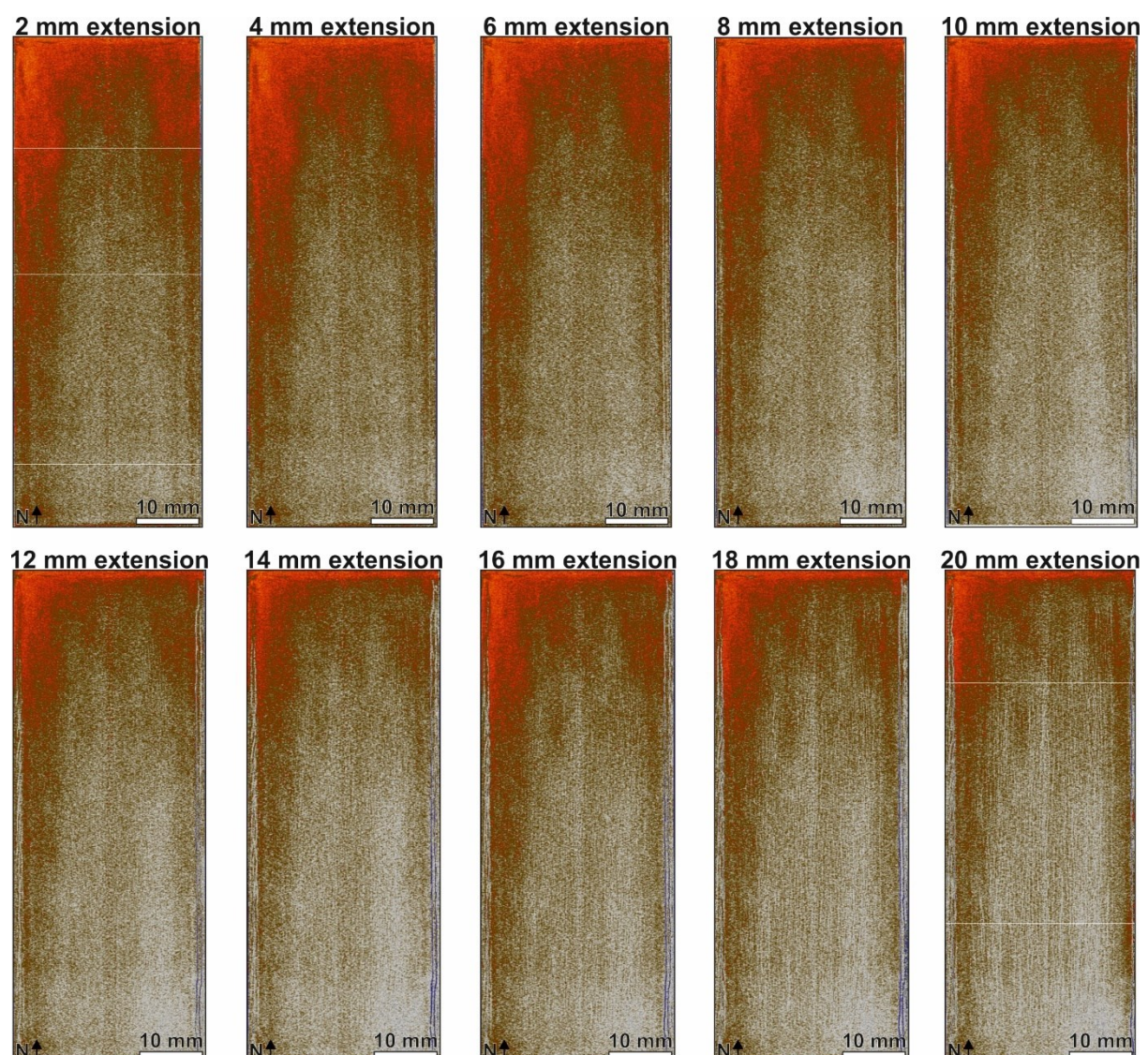
Figure 5. 4. Progression of experiment 816 at 4, 10, 16, 22, 28, 34, and 40 mm extension. Photographs of the top view, top of the model in X-Ray view, and cross sections across the model are shown. Details on model 808 can be found on Table 5.1.

We chose these specific model set-ups because the apparatus allows stress to be equally applied across the model, creating an idealised, yet realistic fault network. We used a variety of different set-ups, varying the speed of extension, thickness of the sand, and thickness of the ductile base layer before committing to our 4 chosen models (Table 5.1). The aim of this work was to quantify normal fault growth at a higher resolution than can be tracked in 3D seismic and running the models inside a CT scanner at every 2 mm of extension allowed us to do this.

5.2.2 Rationale for chosen set-up for analysis

We carried out 4 different model set-ups inside the CT scanner (Table 5.1) and we opted to carry out analysis only on model 816: a brittle-viscous model with a 4 cm sandpack. We opted to only carry out analysis on one model due to time constraints, as analysing faults

across 20 time-steps takes several months. Model 816 was chosen because it yielded the most realistic structures of the 4 experiments that we ran. The brittle-ductile model set-up proved to create the most realistic faults because the base-layer allowed strain to be partitioned onto different faults as we see in nature (Figure 5.3). In the brittle-brittle set up, faults were very tightly spaced with more or less equal amounts of displacement across each fault because without a ductile layer, strain had to be accommodated directly above the expanding foam (Figure 5.3a). The ductile layer allowed strain to be partitioned across the model, which permitted the experiment to have realistic geological phenomena, such as strain localisation onto larger faults and horst and graben formation (Figure 5.3b). We the brittle-ductile model with a 4 cm sandpack as opposed to the 6 cm sandpack as it was the easiest to track faults, however both models would have been acceptable choices. Figures that show the additional three models that were run inside the CT scanner can be found in the appendix of this thesis (appendix figure 5.1-3).



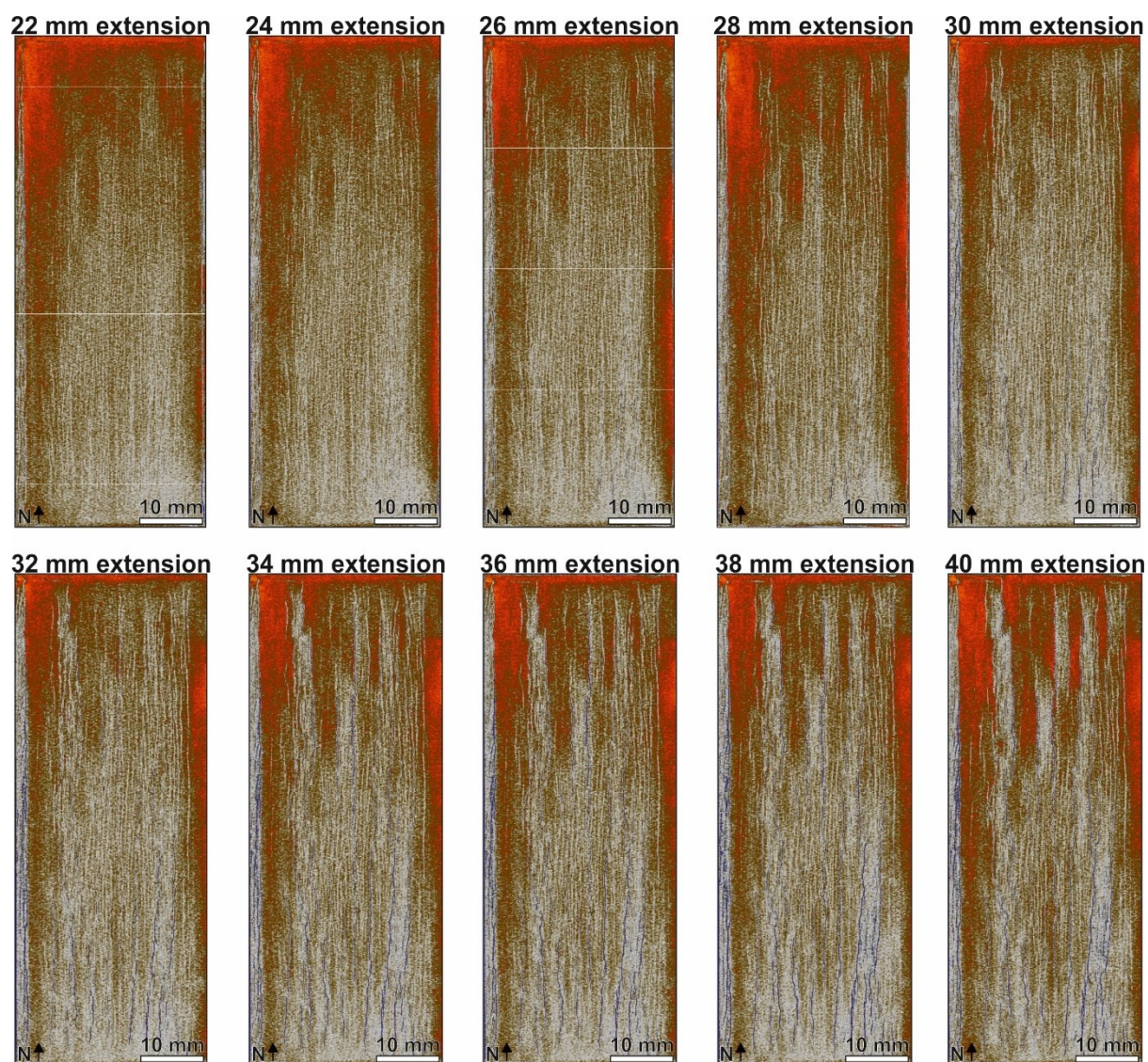
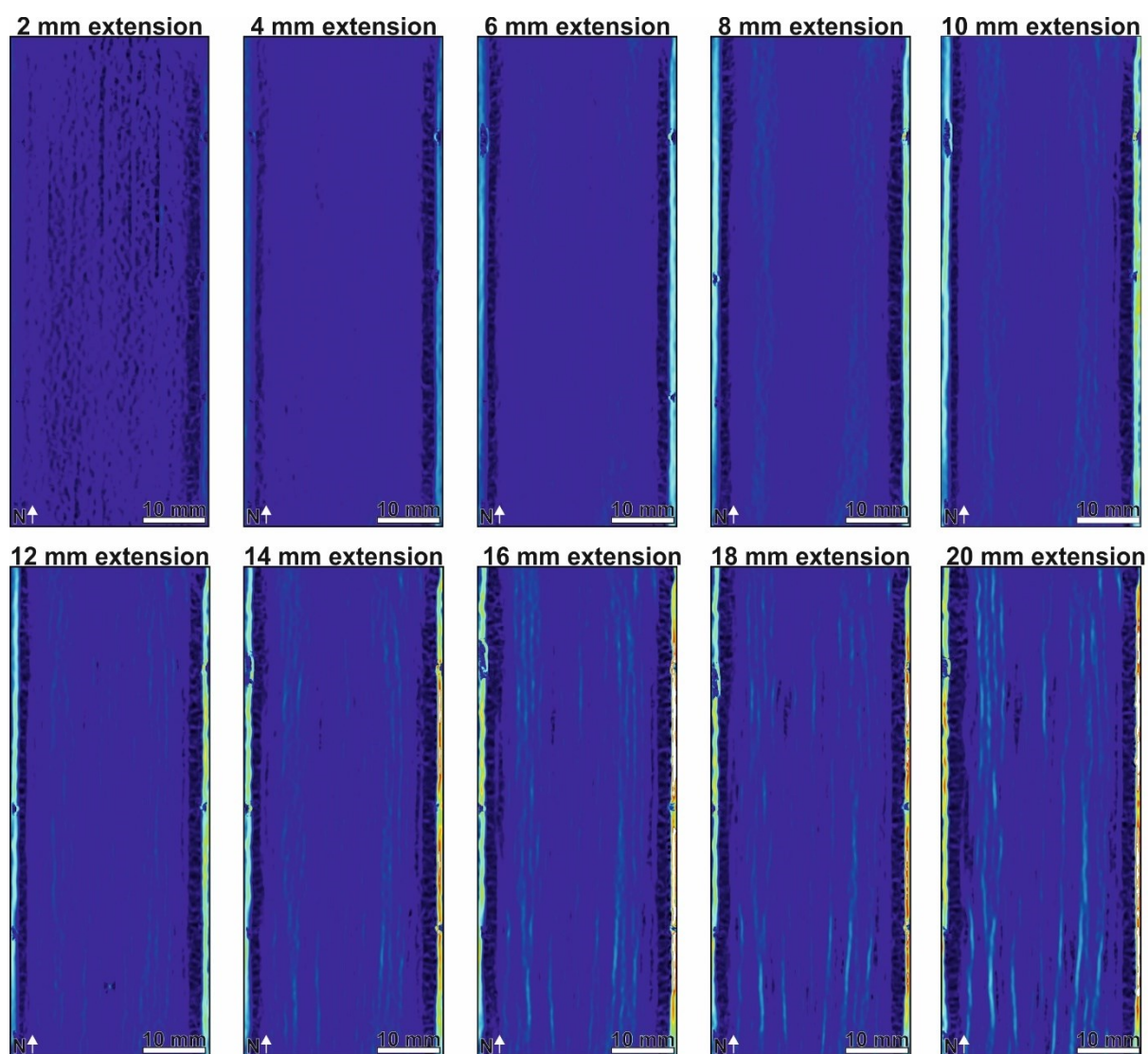


Figure 5. 5. CT scan images of model 816 at every 2 mm of extension. Images are showing the upper part of the model (~0.5 cm from the top). Some fault faulting can be seen at ~10 mm extension.



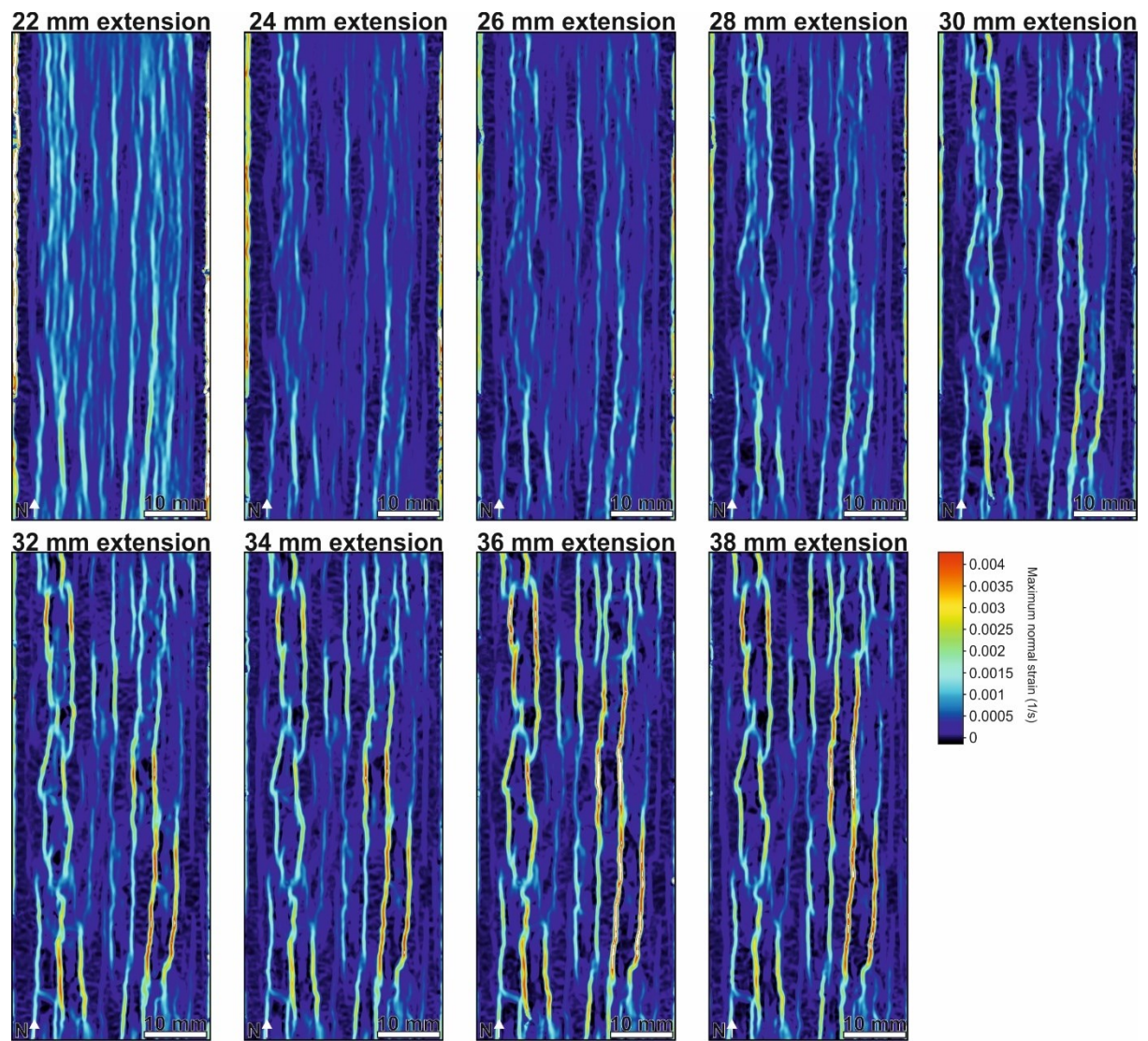


Figure 5. 6. Images from overhead PIV analysis of model 816 at every 2 mm of extension. Some fault faulting can be seen at ~6-8 mm extension.

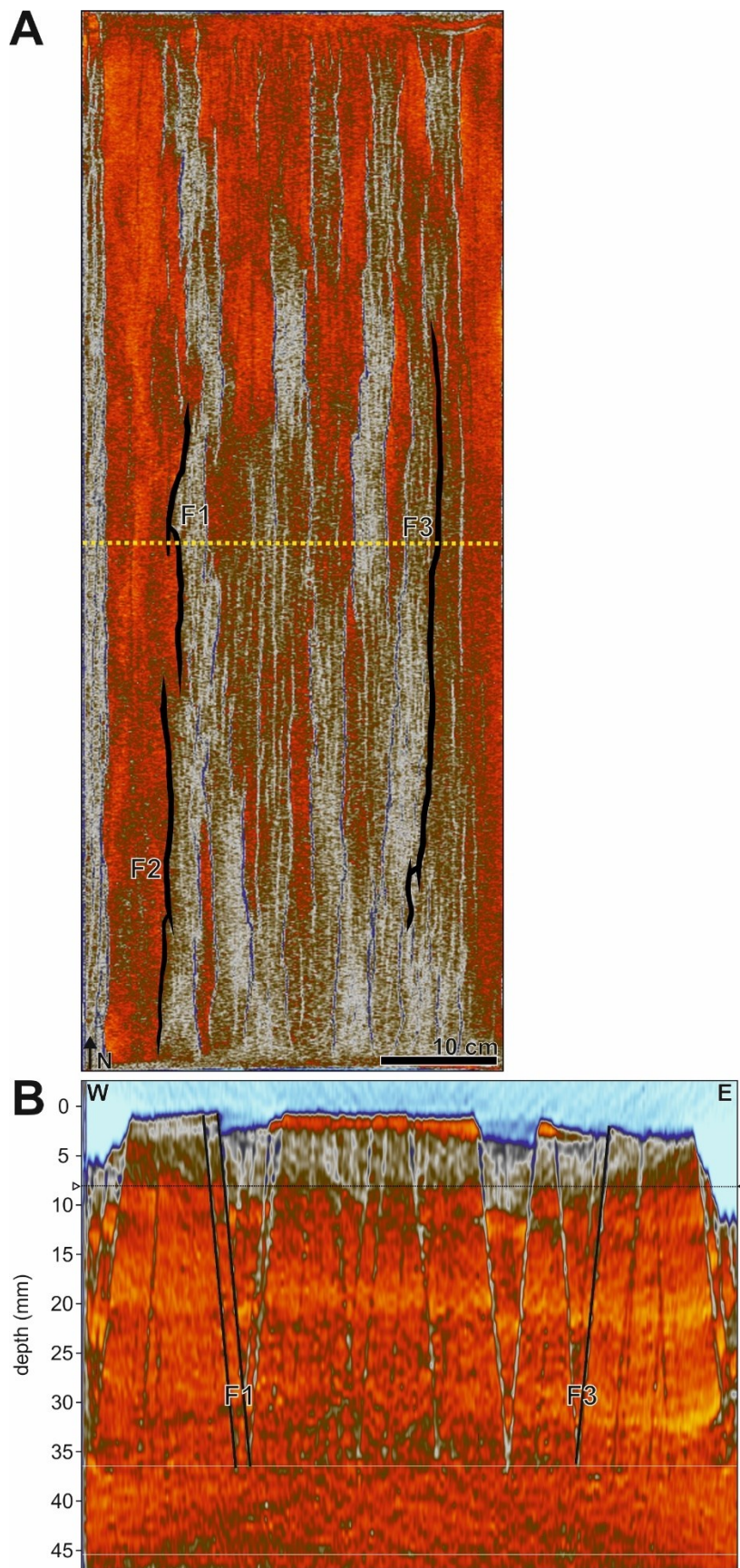


Figure 5. 7. Overview and cross-section of Model 816. A) Overhead view of Model 816 at 40 mm extension, showing the overall structures of the model. F1, F3, and F3 are labelled. Depth of image is shown in figure B. B) Cross-section view. Location of cross section is labelled in A.

5.3. Results

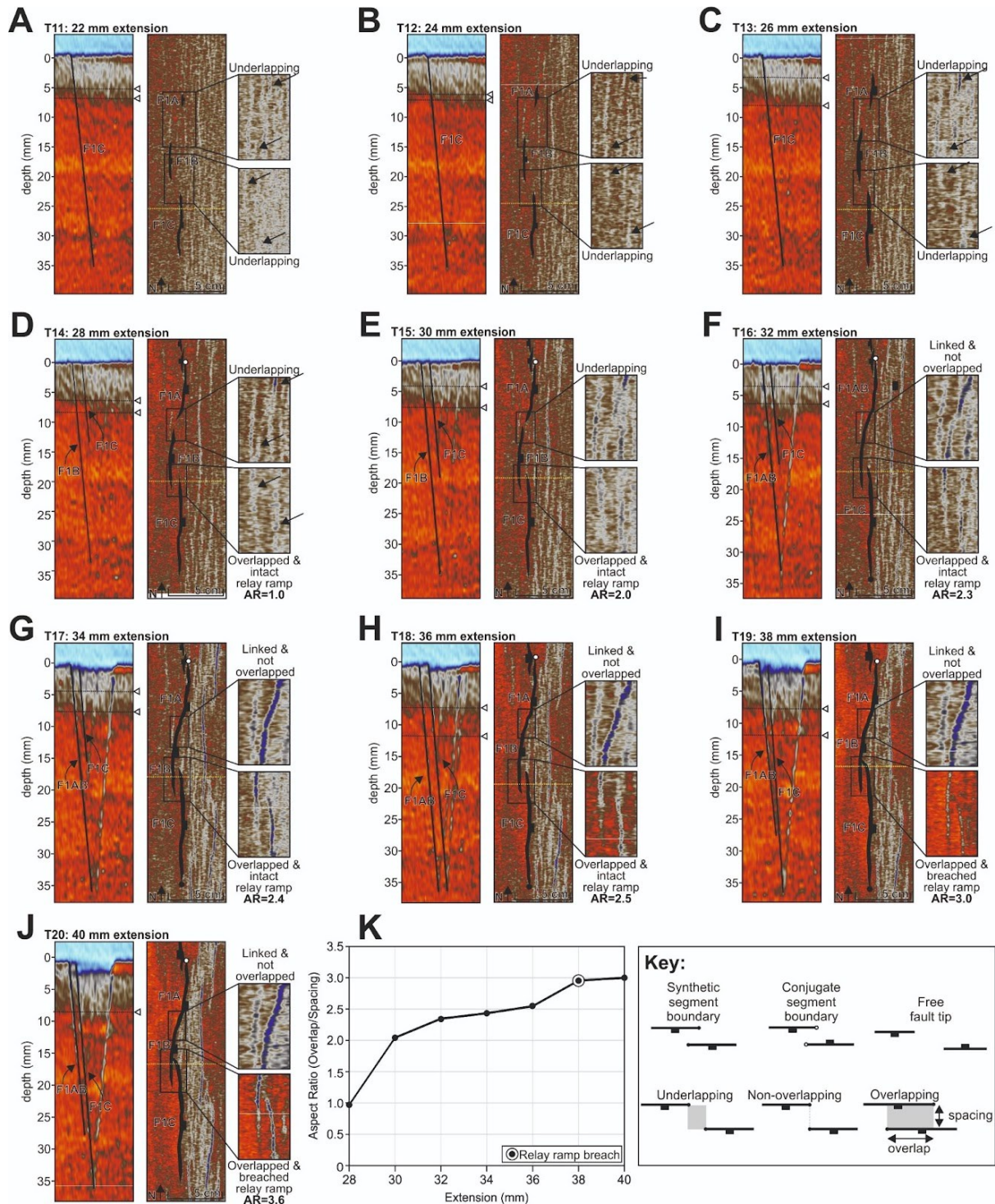
In this section we first take a qualitative look at the overall structural development in model 816, from the initiation to the cessation of extension (Figures 5.5 and 5.6). We then focus on three segmented faults and their associated relays, in order to quantify how fault segments develop and link (faults labelled in Figure 5.7). We focus on fault development between 22 mm of extension, i.e., when throw is high enough for faults to be measured in the 3D data, and 40 mm of extension, i.e., the end of extension.

5.3.1. Overview of faults

Faulting at the edges of the model could first be seen after 6 mm of extension (Figures 5.5 and 5.6), although these are considered edge effects (see Zwaan et al., 2016). Due to the higher resolution of the top-view images compared to the CT scans, the overhead particle analysis was able to image fault activity earlier than the CT scans; for example, some incipient faulting across the middle of the model could be seen at around 6-8 mm extension in the overhead images (Figure 5.6) and at around 10 mm extension in the CT scan (Figure 5.5). From 10-14 mm extension, there were many closely spaced, seemingly low-throw faults observed in both the particle analysis images and CT scans (Figures 5.5 and 5.6). Between 20-22 mm extension (Figure 5.5 and 5.6), some faults appeared to become more active, or at least look as though they have a more well-defined fault surface. In the later stages of the model (post-26 mm extension), longer and better-developed faults started to form grabens. Small and likely inactive faults could be seen between the larger structures (Figures 5.5 and 5.6).

No matter how closely spaced synthetic faults were in the models, they maintained a constant dip angle throughout and thus did not overlap or vertically link to form branching faults as we tend to see in nature (e.g., Soliva et al., 2008). This was likely a result of the model set-up, so in regard to this aspect of fault geometry, the models are not perfect analogues to nature. The reason that fault dip angle does not change could be a result of a number of factors relating to the model. Quartz sand is considered to be nearly cohesionless (Zwaan et al., 2019), which is likely different to rocks in nature. In this model, extension comes from the expanding foam at the bottom of the model, but in nature it could come from the side as well, so this could cause our faults to behave differently. Additionally, it is possible that experiments with thicker sand layers are needed to experience changes in fault dip. There have not been many examples of extensional physical analogue models that have been scanned in 3D, however, the faults in Filbrandt et al. (2007) also appear to have a constant dip angle. Branched faults have been

shown in numerical models (Finch & Gawthorpe, 2016) and it is unclear if they have been modelled in clay-based physical analogue models, as clay-based models have not yet been scanned in 3D.



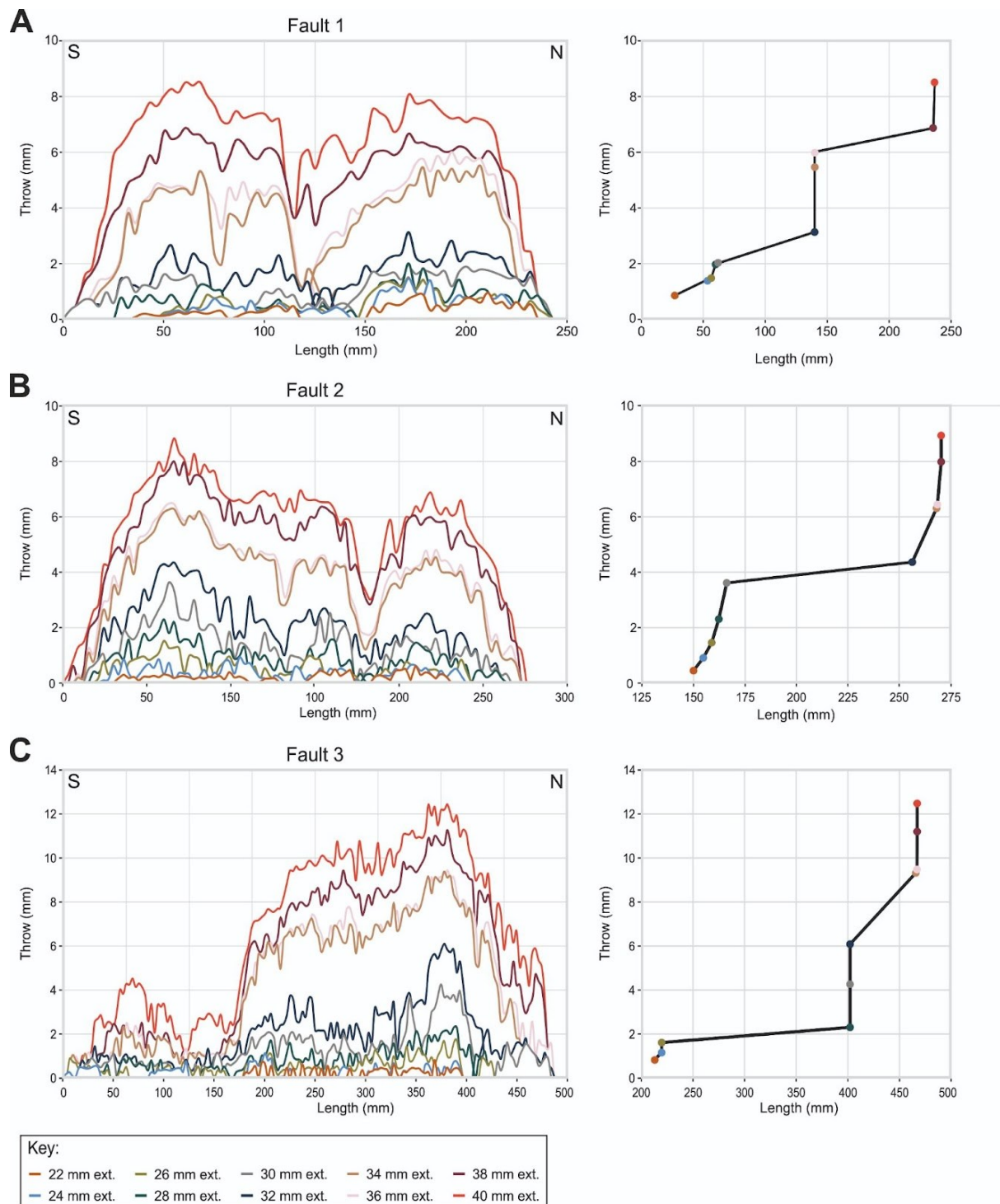


Figure 5. 9. Throw-length plots for every 2 mm extension, from 22-40 mm extension (left) and throw and length tracked at every 2 mm of extension. Each point represents 2 mm of extension. A) Fault 1, B) Fault 3, and C) Fault 3.

5.3.1.1. Fault 1

Having looked at the very general patterns and styles of fault network development, we will now focus on three specific segmented faults, tracking temporal changes in their length and throw, as well as the geometry and evolution of their related relay ramps. We chose these three faults because they had a high enough throw to be imaged relatively early in the

experiment, and because they showcased different relay ramp geometries and different types of interactions with adjacent faults. These three faults are labelled in CT scan images in Figure 5.7.

F1 was 200 mm long and had a maximum throw of 9 mm at the end of the experiment. It has N-S strike and dipped to the E (Figure 5.7). F1 had a conjugate fault boundary at its northern tip and a synthetic boundary at its southern tip (Figure 5.7 and 5.8). F1 consisted of three segments (referred to as F1A-C) that were initially separate and eventually linked via propagation and segment linkage (Figure 5.8).

The segments that make up F1 first had high enough throw to be mapped at 22 mm extension, and at this time, the three segments were underlapping (Figure 5.8a). The length of F1C was first measured with a length of a 22 mm and 0.9 mm throw at 22 mm extension (Figure 5.9a). Between 22 mm and 28 mm of extension, the tips of F1A-C propagated, such that the tips of F1B and F1C overlapped; at this time, they were kinematically but not physically linked (Figure 5.8d). F1A then physically linked with F1B at 32 mm extension (Figure 5.8f), bringing the length and throw of the linked segments of 118 mm and a throw of 3.2 mm (Figure 5.9a). Before F1A and F1B had physically linked; they had not overlapped before this point, suggesting tip-to-tip linkage (Figure 5.8f). Between 32-36 mm extension, the fault maintained its length 118 mm and accrued a total of 6.1 mm of throw (Figure 5.9a). At 38 mm extension, the relay between F1A/B and F1C breached (Fig. 5.8i), bringing its total length to 199 mm and maximum throw of 7 mm (Figure 5.9a). The length of F1 then remains relatively stable (up to 200 mm length) while maximum throw increased to 8.9 mm by the end of extension (Figure 5.9a).

The aspect ratios of the relay ramps of F1B-F1C ranged from 1 to 3 and the relay ramp between the two breached at an aspect ratio of 2.9 (Figure 5.8j). The resolvable tips of F1A and F1B were 1.6 cm apart at 30 cm extension and connected at 32 cm extension (Figures 5.8e and 5.8f). F1A and F1B physically linked without having overlapping tips before or after linking, so aspect ratio cannot be measured (or aspect ratio=0; Figure 5.8). F1B and F1C had aspect ratios ranging between 1.0 to 3.6 and the relay ramp breached at 3.0 (Figure 5.8k). F1A-F1B had hanging wall breached relays; the southern tip of F1A (the footwall) remained present and active and the northern tip of F1B (the hanging wall) propagated and breached the hanging wall (Figure 5.8).

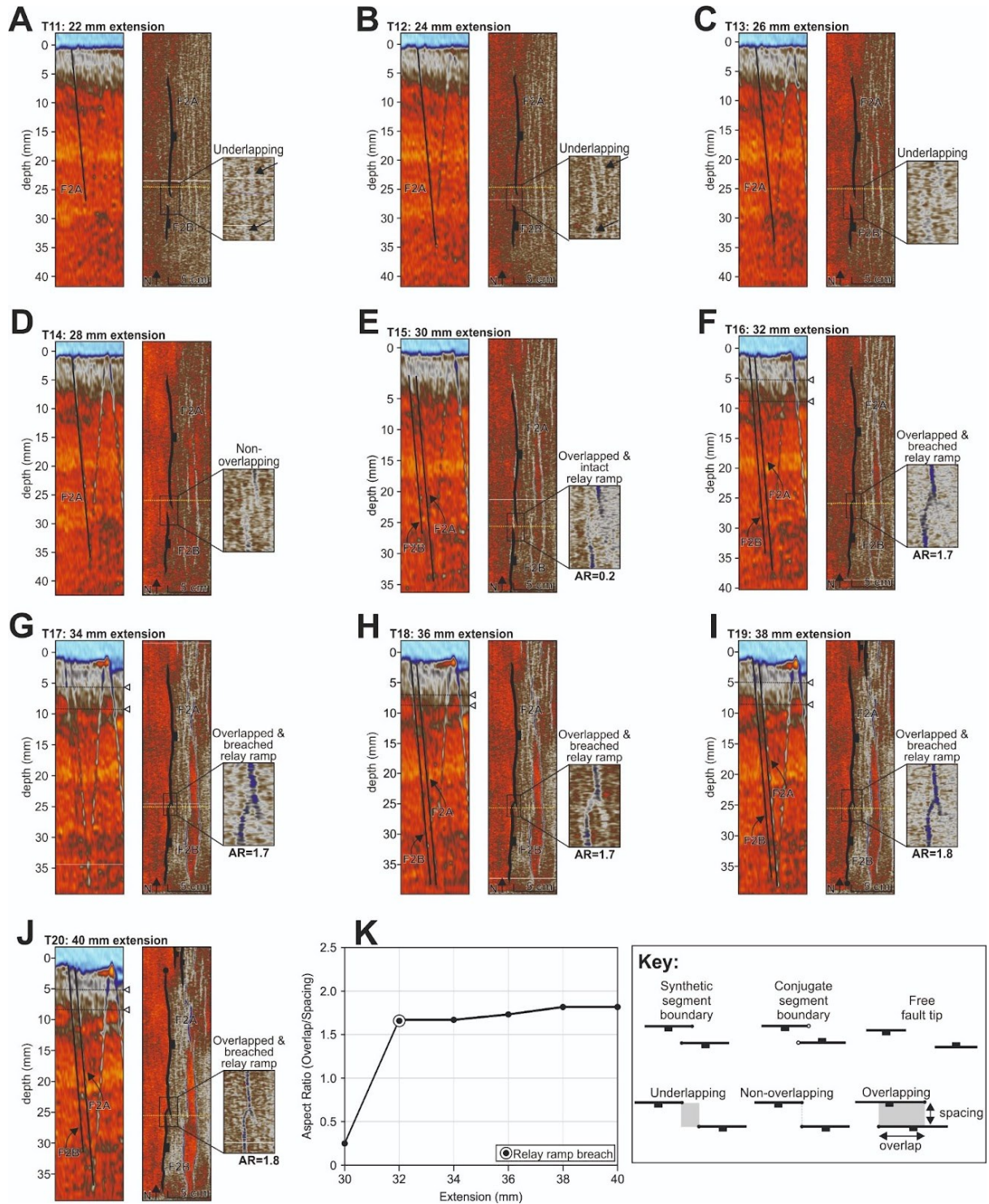


Figure 5.10. Figure showing the segment linkage for Fault 2 from 22 mm extension to 40 mm extension. Faults are shown in cross-section and overhead view, as well as close-up views of the fault tips approaching each other. Depth of the overhead image and the fault tip images are shown on the cross-section view. Aspect ratios are given for each relay ramp zone, and tracked in figure K.

5.3.1.2. Fault 2

At maximum extension, F2 was 271 mm long with a maximum throw of 9 mm, N-S strike, and dipped to the E (Figure 5.7). It had a synthetic fault boundary at its northern tip and a free southern tip, however the southern boundary of F2 was close to the bottom of the model

(Figure 5.7 and 5.10). F2 consisted of two separate fault segments (referred to as F2A-B) that were initially kinematically and physically separate, but eventually linked (Figure 5.10).

F2A and F2B first had a large enough throw to be mapped at 22 mm extension, and F2A had a measurement of 150 mm and 0.6 mm of throw at that time (Figure 5.9b & 5.10a). At this time there were many small, low-throw faults surrounding F2 (Figure 5.10a). Between 22 and 30 mm of extension, F2A grew slightly by tip propagation from 150 mm to 168 mm and accrued a total of 3.8 mm throw (Figure 5.9b). The tips of F2A and F2B overlapped at 30 mm extension (Figure 5.10e) and at 32 mm extension (Figure 5.10f) the relay ramp between the two fault segments breached and they became mechanically linked, bringing the total fault length to 257 mm and a throw of 4.5 mm (Figure 5.9b). Between 32 and 40 mm extension, F2 lengthened slightly to a final length of 271 mm with 9.3 mm throw (Figure 5.9b).

The aspect ratio of F2A-F2B ranged from 0.2 to 1.8 and the relay ramp between the two breached at an aspect ratio of 1.7 (Figure 5.10k). F2 had a hanging wall breached relay; southern tip of F2A (the footwall) remained present and active throughout the fault's life and the northern tip of F2B (the hanging wall) propagated through and breached the relay ramp (Figure 5.10).

5.3.1.3. Fault 3

At maximum extension, F3 was 467 mm long with a maximum throw of 12 mm, N-S strike, and dipped to the W (Figure 5.7 and 5.11). Both the northern and the southern tips of F3 had conjugate boundaries with adjacent fault segments (Figure 5.7 and 5.11). F3 consisted of three separate fault segments (referred to as F3A-C) that were initially kinematically and physically separate, but eventually linked (Figure 5.11).

F3B first had a high enough throw to be mapped at 22 mm extension and had a length of 212 mm and a throw of 0.8 mm at that time (Figures 5.9c & 5.11a). F3B and F3C first had a high enough throw to be mapped at 24 and 26 mm extension respectively (Figures 5.11b and 5.11c), and all three segments were underlapping up to this point (Figure 5.11). Between 22 mm and 26 mm extension F3B grew slightly to 218 mm and accrued a total of 1.6 mm throw (Figure 5.9c). At 28 mm extension, F3A and F3B overlapped and the relay ramp between them breached, bringing the total length to 404 mm with a throw of 2.2 mm (Figures 5.9c & 5.11d); the tips of F3B and F3C overlapped and remained unlinked (Figure 5.11d). Between 28 and 32 mm extension, F3 maintained a length of 404 mm and accumulated 5.9 mm of total throw (Figure 5.9c). The relay ramp between F3B and F3C was breached and the entirety of

F3 was physically linked by 34 mm extension (Figure 5.11g), bringing the total length of F3 to 465 mm and 8.8 mm of total throw (Figure 5.9c). Between 34 and 40 mm extension, the length of F3 remained relatively stable, measuring 467 mm long and 12 mm throw at the end of extension (Figure 5.9c).

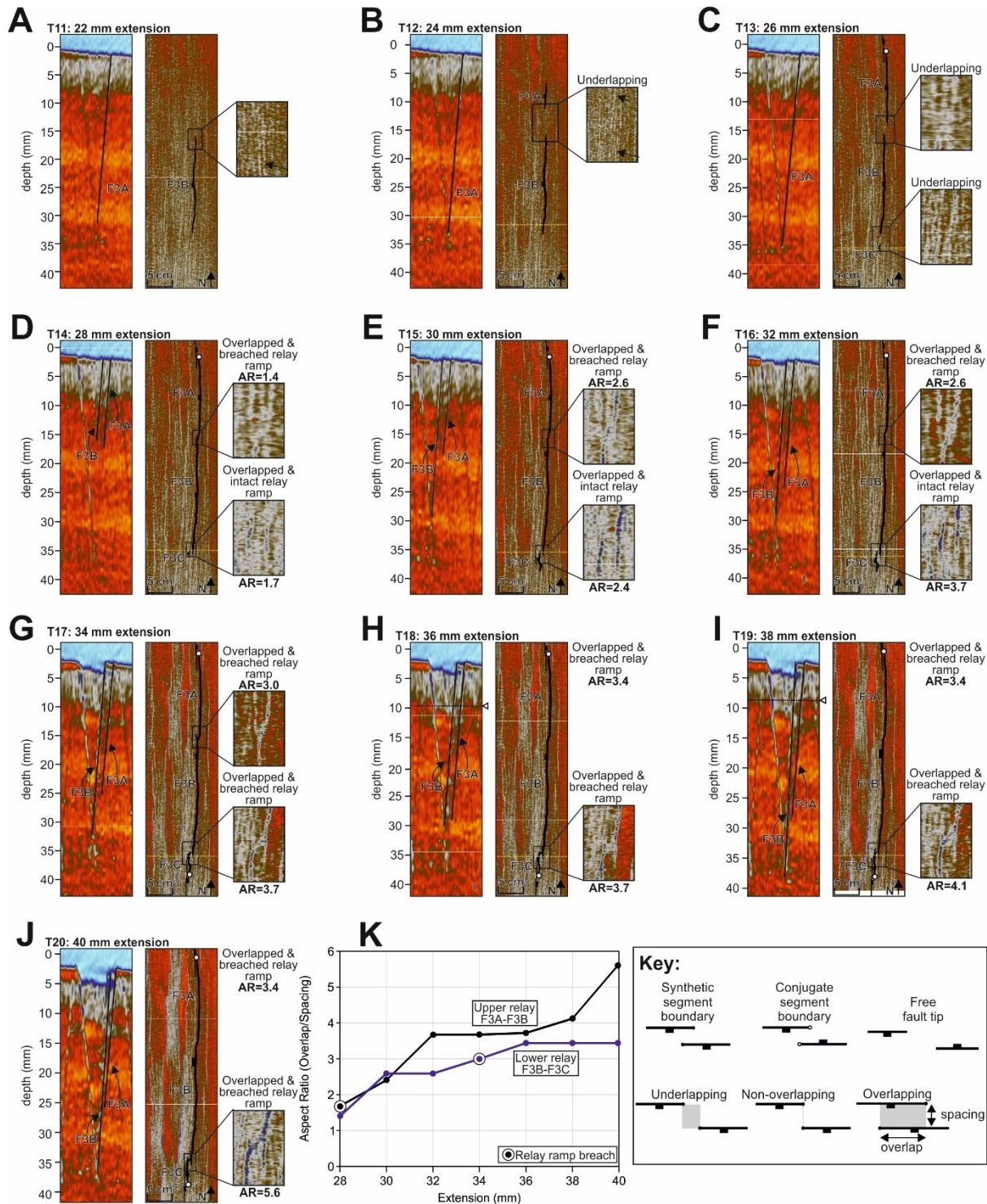
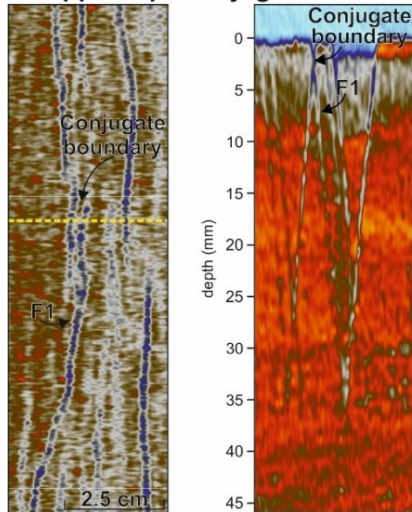
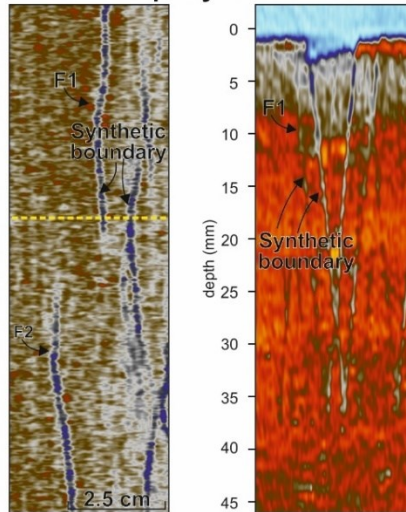


Figure 5. 11. Figure showing the segment linkage for Fault 3 from 22 mm extension to 40 mm extension. Faults are shown in cross-section and overhead view, as well as close-up views of the fault tips approaching each other. Depth of the overhead image and the fault tip images are shown on the cross-section view. Aspect ratios are given for each relay ramp zone, and tracked in figure K.

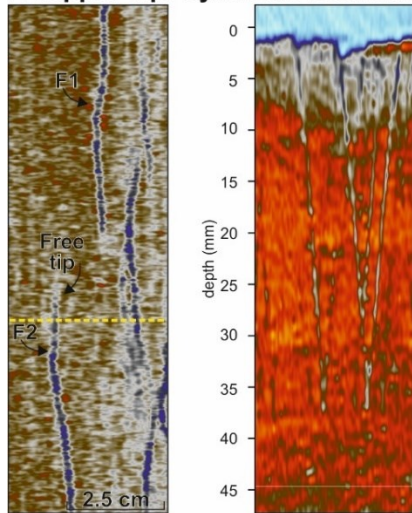
F1 Upper Tip: Conjugate Boundary



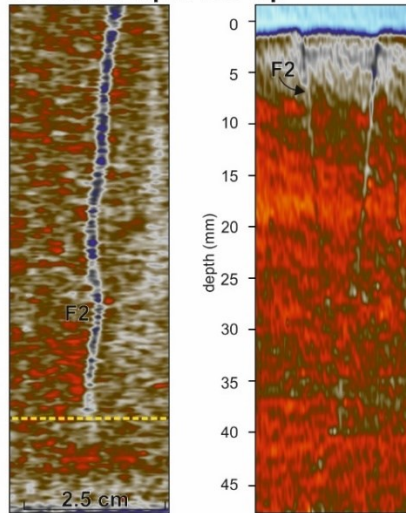
F1 Lower Tip: Synthetic Boundary



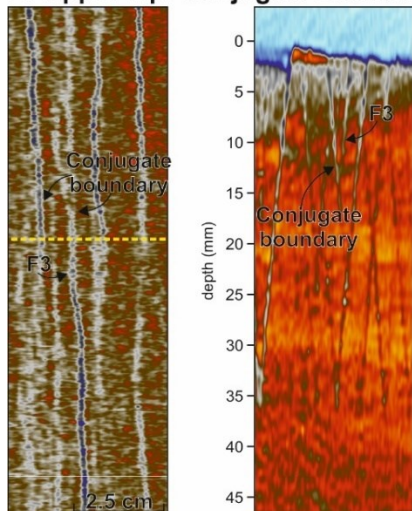
F2 Upper Tip: Synthetic Boundary



F2 Lower Tip: Free Tip



F3 Upper Tip: Conjugate Boundary



F3 Lower Tip: Conjugate Boundary

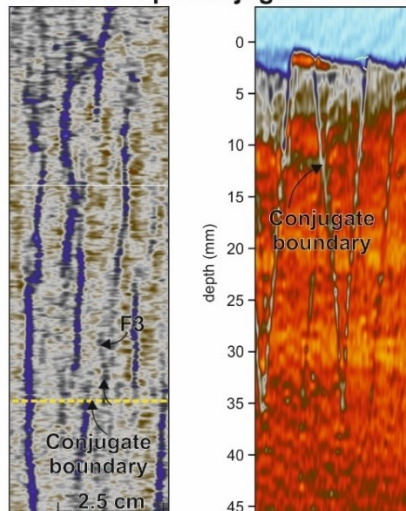


Figure 5. 12. Figures showing faults 1-3 and their boundaries with adjacent fault segments on their upper and lower tips

The aspect ratio of F3A-F3B ranged from 1.4 to 3.4 (Figure 5.11k) and they linked without first forming a relay ramp; the resolvable tips of F3A and F3B were 1.3 cm apart at 26 mm extension and mechanically linked at 28 mm extension (Figures 5.11c and 5.11d). The aspect ratio of F3B and F3C ranged from 1.7 to 5.6 and the relay ramp between the two breached with an aspect ratio of 3.7 (Figure 5.11k). F3B-F3C had a transfer breached relay ramp: both the southern tip of F3B and the northern tip of F3B were present and active throughout, and a separate fault segment formed between them, breaching the relay ramp and physically linking the two fault segments (Figure 5.11). Both the hanging wall and footwall tips remained active and propagating even after the relay ramp breached (Figure 5.11).

5.3.2. Aspect ratios and relay ramp development

Relay ramp aspect ratios of the three studied faults ranged from 0.2 to 5.6 and relay ramps breached at an aspect ratio of 3.0, 1.7, 1.5, and 3.7 (Figures 5.8, 5.10 & 5.11). Relay ramp aspect ratios increased at every stage of extension, both before and after the ramp was breached. In the faults that we look at in detail, there was not a distinct relationship between relay ramp aspect ratio and relay ramp breaching; some fault segments linked without overlapping, some linked via relay ramp breaching with a low aspect ratio, and some linked via relay ramp breaching with a high aspect ratio.

There are three types of segment linkages observed for the three faults that were studied in detail including hanging wall breached relay ramps, transfer relay ramps, and fault segments that have linked without forming relays. Faults that had relay ramps that formed before breaching were likely mechanically and kinematically linked before becoming physically linked. Fault segments that linked without first forming a relay were likely situated in a position where their tips naturally propagated towards each other, and they linked without overlapping. They were likely not mechanically and kinematically linked before becoming physically linked.

5.3.3. Interactions with adjacent faults and its impact on throw-length plots

The northern tip of F1 and both the northern and southern tips of F3 had conjugate fault boundaries, i.e., they interacted with an adjacent fault that was dipping in the opposite direction (Figure 5.12). The tips of the faults overlapped slightly in map view and throw quickly tapered off (Figure 5.12). The tips of F1 and F3 and their adjacent fault segments were not interacting at the surface but interacted at depth (Figure 5.12). The throw gradient (length/throw) of the throw-length plots of the northern tip of F1 (Figure 5.11a) is 3.6 and the tips of F3 (Figure 5.11c) were 2.65 and 12.89 respectively. The throw gradients of northern

tip of F1 and F3 (Figure 5.11a) were high due to the conjugate boundary preventing the fault tip from propagating (c.f. Whipp et al., 2016; Nicol et al., 2017a). The southern tip of F3 had a relatively low gradient despite having a conjugate boundary, however we believe this was due to the southern tip having an overall low throw and not being very active.

The southern tip of F1 and the northern tip of F2 had synthetic segment boundaries with adjacent structures (Figure 5.12). The faults are likely too far away to be optimally placed to have linked, but there is likely some strain being accommodated by the adjacent fault, which could result in a lower throw gradient. The throw gradient of the southern tip of F1 (Figure 5.11a) was 3.4 and the northern tip of F2 (Figure 5.11b) had a throw gradient of 4.5. These gradients were high from strain at the ends of the tips being partitioned onto the adjacent faults, but not as high as the previously mentioned faults with conjugate boundaries.

The southern tip of F2 was a free tip, i.e., there were no nearby fault segments and the fault tip was able to propagate freely, however, the southern tip of F2 was near the southern end of the model and thus there was no room for major additional faults below (Figure 5.12). The throw gradient of the southern tip of F2 (Figure 5.11b) was 5.9, which is lower than the previously mentioned conjugate and synthetic faults.

5.4. Discussion

5.4.1 Fault growth patterns

The studied faults all followed a similar growth trajectory. Segments lengthened quickly and then maintained a relatively stable length while accruing throw (Figure 5.13). The fault segment then linked with an adjacent synthetic segment, which caused the total fault length to increase dramatically (Figure 5.13). The newly linked fault segments then maintained a relatively stable length whilst accruing throw. In theory, this pattern could continue until either 1) extension stops, or 2) the fault comes into contact with a barrier that prevents further segment linkage, such as a conjugate boundary with an adjacent fault, or 3) the fault becomes inactive due to stress being partitioned onto a more optimally situated fault. This could be tested in the future with physical analogue models with extended extension (i.e., more than 4 cm extension). The exception to this growth pattern was the increase in length in F1 from 22 to 24 mm extension, however this was at least partially due to the throw being too low to clearly see F1 at 22 mm extension (Figure 5.8a).

Faults had alternating stages of abrupt lengthening and steady throw accumulation and changes in fault length did not correspond to changes in fault throw (Figure 5.13). Throw and

length ratios were constantly in flux throughout the life of a fault and a constant slip rate or tip propagation rate (as suggested by the propagating fault model) cannot be assumed.

In the fault growth model proposed by Pan et al. (2021), a cyclical growth model was suggested where faults alternate between stages of rapid lengthening and slow displacement/throw accrual. Pan et al. (2021) suggested that faults grew in this step-wise manner across all scales, from individual fault arrays to full fault networks, however their dataset only captured one stage of fault growth in km-scale normal faults. Observations from our analogue models are consistent with their predictions, with faults showing several stages of step-wise growth (Figure 5.13).

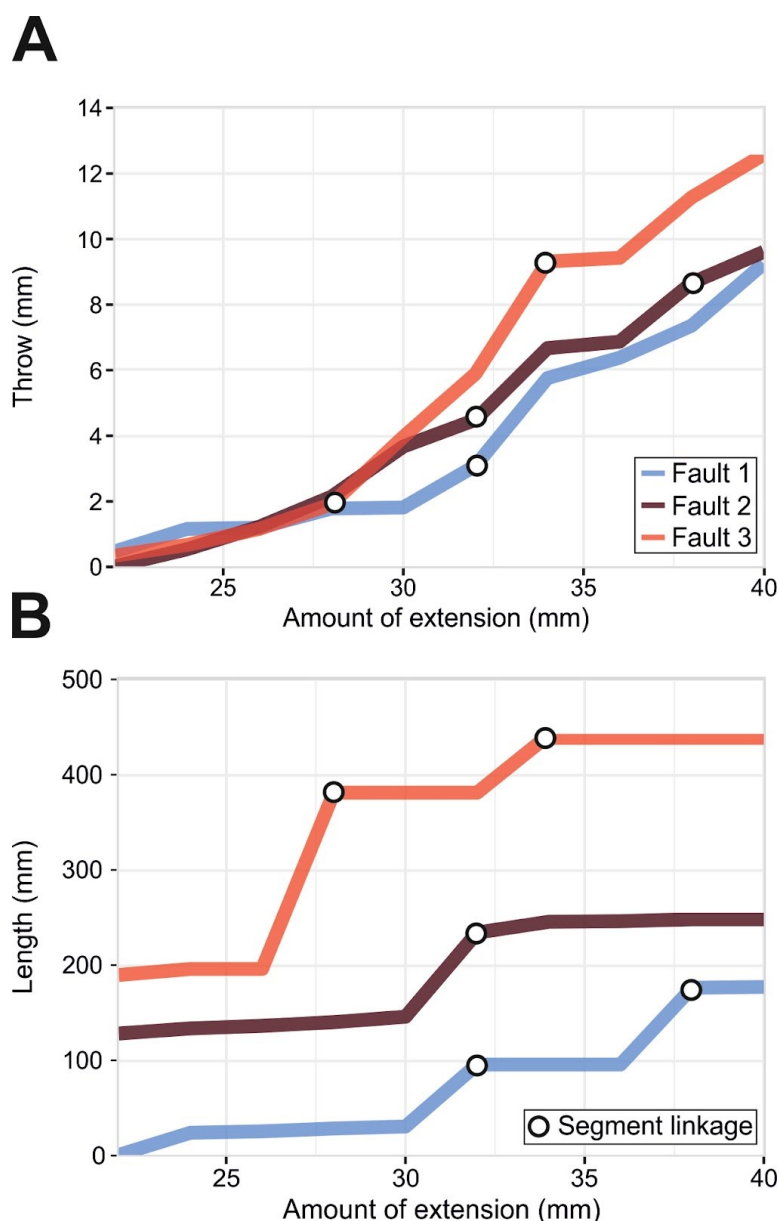


Figure 5. 13. Figure tracking fault growth through time of Faults 1-3 in model 816. A) Throw through extension and B) Length through extension. Stages of segment linkage is indicated on each fault.

5.4.2. Strain localisation

As faults grow from the segment/array scale to the network scale, strain is partitioned onto select optimally positioned faults, while many faults become inactive in the early stages of extension (Cowie, 1998; Gawthorpe & Leeder, 2000; Cowie & Roberts, 2001). Strain localisation is also seen in our experiments. Unfortunately, we are not able to measure fault length and throw until 22 mm extension because the throw is too low for the fault to be mapped confidently. However, we can still get some qualitative insights into the early stages of fault growth and strain localisation by inspecting the initial stages of the models. We can see that faults formed in the models as early as ~8 mm extension in the overhead PIV images (Figure 5.6) and as early as ~12 mm extension in the CT images (Figure 5.5). Faults started off very closely spaced and appear to have had very low throws. Many of those faults became inactive/stop accruing throw early, and the faults that continued growing were the ones that we are able to observe and measure past 20 mm extension (Figures 5.5 and 5.6). As extension continued, optimally positioned faults (i.e. faults with synthetic boundaries) continued to grow via segment linkage, while the low-throw faults present in earlier stages of fault growth became inactive and died (Figures 5.5 and 5.6).

5.4.2. Relay ramp development

The faults that were overlapped prior to physical linkage breached with aspect ratios of 3.0, 1.7, 1.5, and 3.7 for F1, F2, and the northern and southern segments of F3, respectively (Figures 5.8k, 5.10k, and 5.11k). The average aspect ratio for a relay ramp to breach is typically reported to be <4 (Mansfield & Cartwright, 2001), with an average between 3-3.5 (Acoella et al., 2000; Hus et al., 2005; Fossen & Rotevatn, 2016). The aspect ratio needed to breach a relay ramp was smaller in some of these studied models than previously reported. This is possibly due to the high temporal resolution nature of these data, i.e., with our models, we are able to pinpoint the exact time at which the relay ramp physically breaches. After relay ramp breaching occurs, the fault tips can continue to propagate, and aspect ratios continue to increase on all of the studied faults. However, when measuring a relay ramp in outcrop, only the end stage is seen (if the fault is inactive), and the ramp could have breached earlier/at a lower aspect ratio. This is in contrast to several other studies that suggest fault tips become inactive and are abandoned after their adjacent relay ramps become breached (e.g., Peacock & Sanderson, 1991, 1994), however, we observed fault tips that continued to accumulate throw post relay-ramp breaching in our previous work in the Exmouth Plateau (Fault 1, Figure 4.6), as well as in a previous sandbox analogue model (Hus et al., 2005). This is possibly because immediately after relay breaching, the newly formed connection is not

able to accommodate all of the stress immediately, thus the tips can remain active for some time after breaching. However, in the Exmouth Plateau example, the fault tip of the breached fault remained active for an additional ~58 million years after the relay ramp was breached, so perhaps this is a phenomenon that could occur across longer time scales (Chapter 4; Lathrop et al., 2021).

Fault linkage by relay ramp formation and breaching is a common way for faults to link, however some faults in this model linked without needing to form a relay ramp. F1A-F1B and F3A-F3B are non-overlapping synthetic faults (Schlische, 1995), also referred to as collinear fault tips (Morley et al., 1990) (Figures 5.8 and 5.11). They were able to do this because their across strike distance was relatively small, such that as their non-overlapping synthetic fault tips approached each other, zones of stress enhancement at their fault tips merged, causing the segment strikes to slightly change, and their tips to ultimately link without overlapping to form a relay ramp (Soliva et al., 2006). This is likely especially common in the early stages of fault growth, when there are many small closely spaced faults.

5.4.3. Interactions with adjacent faults

In our models, the northern tip of F1 and both the northern and southern tips of F3 (Figure 5.12) have conjugate segments boundaries with adjacent structures. The tips of these faults are also characterised by high throw gradients, as their fault tips were pinned by the adjacent fault but throw continued to accumulate (Figure 5.11). Fault segments become pinned relatively early, which restricts further fault growth via segment linkage or tip propagation. For example, the northern tip of F1A was pinned by a conjugate fault by 28 mm extension, which prevented it from growing laterally for the rest of extension (Figure 5.8). This is likely one of the reasons that faults stop lengthening quickly; once faults start to grow and interact with a conjugate fault, they are mechanically blocked from lengthening any further, and this is very common across the model.

5.5. Conclusion

We use sandbox analogue modelling to study normal fault growth by tracking relay ramp geometries and length and throw through time. Several types of relay ramps are seen, and some fault tips remain active even after relay ramps are breached. If closely spaced synthetic fault tips approach each other, they can link without forming a relay ramp. Overall, faults show a step-wise growth pattern with alternating stages of lengthening and displacement. Fault segments lengthen quickly, usually interacting with adjacent fault segments and then accumulate displacement. Faults then grow via linking with an adjacent synthetic fault

segment, and then remain at that length while accumulating throw. Only optimally placed fault segments will remain active, and faults in stress shadows will become inactive early and strain is partitioned onto larger faults.

The work presented in this chapter builds upon the work in chapters 3-4 of this thesis. Fault growth patterns in chapters 3-4 show one stage of fault growth with a short lengthening and quick displacement stage, and at times a stage of tip retreat. By using physical analogue models we were able to resolve three cycles of these lengthening-displacement stages, which provides a fuller picture of normal fault growth at the array scale. These models do not show a stage of lateral tip retreat, but this is likely because we did not slow down the rate of extension at the end of the model run which would mimic extension cessation

Chapter 6

Discussion

The research presented in this thesis provides new insights into how normal faults grow and how different geological factors can affect normal fault growth. I used a multipronged approach to gain these new insights by first collecting and critiquing pre-existing data (Chapter 3), then presenting a new seismic-based case study from Offshore NW Australia (Chapter 4), and finally by simulating normal fault growth in high-resolution physical analogue models (Chapter 5).

The key implications drawn from the studies presented in this thesis are described in this chapter, with the specific aim of answering the key questions raised in Chapter 1, namely (i) is there a single universal relationship between normal fault length and displacement?, (ii) how do factors, such as host rock lithology, fault size, tectonic setting, and maturity affect D/L scaling?, and (iii) how does the relationship between fault length and displacement change through time?, and (iv) how do normal faults die? I end this chapter with possibilities for future work.

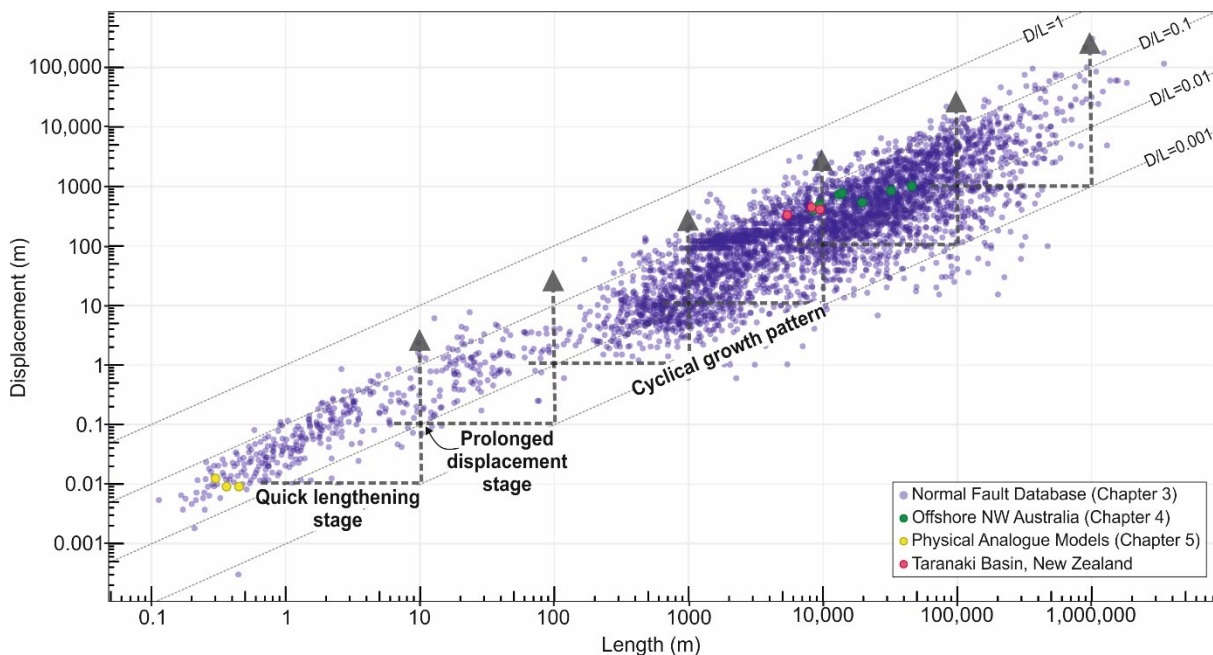


Figure 6. 1. Graph showing normal fault displacement and length data as compiled in Chapter 3 compared with the normal faults studied in Offshore NW Australia in Chapter 4 (green), the physical analogue models in Chapter 5 (yellow), and the Taranaki Basin (red). The trajectory suggested by Pan et al. (2021) is shown where faults experience cyclical stages of rapid lengthening and prolonged displacement.

6.1. Is there a single universal relationship between normal fault length and displacement?

Universal scaling laws are not precise indicators of fault length and displacement and there is no universal D/L relationship linking normal fault length and displacement within several orders of magnitude. The relationship between fault length and displacement has historically been described by the power-law equation $D_{\max} = cL^n$, and values of n between 0.5 and 2.0 have been suggested (Watterson, 1986; Walsh & Watterson, 1988; Marrett & Allmendinger, 1991; Gillespie et al., 1992; Cowie & Scholz, 1992a; Dawers et al., 1993; Scholz et al., 1993; Clark & Cox, 1996; Schlische et al., 1996; Fossen & Hesthammer, 1997; Kim & Sanderson 2005; Xu et al., 2006). In my database of 4046 normal faults presented in Chapter 3, the average power-law relationship between normal fault length and displacement is $D = 0.3L^{0.92}$. My findings broadly agree with previous estimates (i.e., $n=1$; Cowie & Scholza, 1992; Dawers et al., 1993; Scholz et al., 1993; Clark & Cox, 1996; Schlische et al., 1996; Kim & Sanderson, 2005; Xu et al., 2006). The average value for a normal fault is $D = 0.3L^{0.92}$, but using this equation to estimate the length or displacement of a fault will likely yield an imprecise value due to the amount of scatter in the data. For a given length, displacement spreads across several orders of magnitude; in fact, all of the faults that were studied in seismic data and models in Chapters 4-5 fit within the scatter (Figure 6.1). One-size-fits-all scaling laws could be used to estimate fault length or displacement within 1-3 (or even more) orders of magnitude. However, bespoke scaling laws derived from my normal fault database can be more accurate, especially in geological areas that have proven to have anomalously high or low D/L ratios, like new rifts and volcanic rocks (Chapter 3). My normal fault database records displacement and length, as well as lithology, tectonic setting, and fault maturity, and will enable researchers to create bespoke scaling relationships for the specific area that they are looking at. The data is available open access online so that it can be downloaded, tested, and, ideally, added to. However, as I have demonstrated throughout this thesis, the relationship between fault length and displacement is dynamic, changing throughout a fault's life. D/L measurements capture only a momentary snapshot of a fault's final geometry (see section 6.3). The average D/L ratio from Chapter 3 was $D = 0.3L^{0.92 \pm 0.01}$, however it should not be used to estimate fault length or displacement.

6.2. How do factors, such as host rock lithology, fault size, tectonic setting, and maturity affect D/L scaling?

In Chapter 3 I discovered that host rock lithology, tectonic setting, fault size, and maturity can all affect the relationship between normal fault length and displacement, however, often not enough to be statistically significant when using best-fit trendlines. Small faults (<1 m

length) tended to be over-displaced compared to larger faults (Figure 3.3). This is likely because small faults reach their maximum length and displacement quickly, possible in 1-2 slip events, whereas larger faults can be disproportionately long if they have recently lengthened via segment linkage and are still in the displacement-accrual stage of fault growth, highlighting the importance of fault maturity. Faults with stiffer host rocks, such as volcanics, tend to be under-displaced compared to faults with softer host rocks, such as clastic sedimentary rocks (Figure 3.4). This is at least partially due to fault maturity, as faults in volcanic host rocks often occur in active rift zones and are thus relatively immature, whereas those in other areas are often more mature or even extinct. Reactivated faults tend to be over-displaced compared to faults that form in previously undeformed or only weakly deformed settings (Figure 3.7), but I found that this was heavily influenced by the maturity of the reactivated faults. Reactivated faults can have an especially quick lengthening stage due to them exploiting pre-existing weaknesses, such that mature reactivated faults may be over-displaced whereas young, reactivated faults may be under-displaced. Host rock lithology, fault size, and tectonic setting can affect the relationship between normal fault length and displacement, however there is so much scatter within each sub-category that caution is needed when using a singular scaling law will not precisely estimate displacement or length.

In Chapter 3, I discovered that fault maturity has the most significant impact on normal fault D/L scaling relationships (see Figure 3.4 and 3.6). Throughout my thesis I have continually demonstrated (see chapters 3, 4 and 5) that fault D/L ratios change throughout a fault's lifetime, thus the age of a fault will affect whether it may be considered over or under-displaced. The D/L database (Chapter 3) revealed that faults in active regions tend to be under-displaced. In the Offshore NW Australia case study (Chapter 4), faults experienced lengthening before accruing significant displacement, thus D/L ratios change during fault growth. My physical analogue models (Chapter 5) also showed that faults have differing phases of lengthening and displacement. Overall, younger faults tend to be under-displaced because they will have already reached a near-maximum length and are still accruing displacement, whereas more mature faults will be comparatively over-displaced.

6.3. How does the relationship between fault length and displacement change through time?

The propagating fault model and constant-length models have been debated amongst the structural geology community for many years. In reality, they are end-member models, and most faults probably behave in a hybrid mix of the two (Rotevatn et al., 2019; Nicol et al.,

2020). A hybrid fault growth model with clear stages of lengthening and displacement were observed in natural and analogue data in Chapters 3, 4, and 5 of my thesis. In Chapter 3, I compiled, to the best of my knowledge, all the existing published displacement and length data from natural and analogue models. The vast majority of the examples from previously published papers (Walsh et al., 2002; Schlagenhauf et al., 2008; Tvedt et al., 2016; Finch & Gawthorpe, 2017; Jackson et al., 2017; Pan et al., 2021) show that faults have a quick lengthening stage: 26 of the 37 (70%) of the faults included in the database reached >75% of their near-maximum length within the first 25% of their lives, and 35 of 37 (95%) of faults reached near-maximum length within the first 50% of their lives. During the lengthening stage, between 5-75% of total displacement was accrued. Faults then experienced a slow displacement stage where faults maintained a constant length while accruing displacement. 23 of the 37 (62%) of the faults in the global database experienced a stage of lateral tip retreat in the last 25% of their lives. These faults can be categorised as following either a hybrid fault growth model or a constant-length fault growth model (Figure 3.11).

Active faults in the global D/L dataset are relatively under-displaced compared to inactive faults, which is consistent with the constant-length or hybrid fault growth models in that fault lengthening precedes significant displacement accumulation. In Chapter 4, I tracked syn-sedimentary normal fault growth through time in the Exmouth Plateau, offshore NW Australia; all seven of the studied faults reached their near-maximum lengths relatively quickly (within ~9.5 Myr), and then spent the rest of their active lives accruing displacement without significant lengthening (Figure 4.12). In addition to the seven studied faults for which length and displacement were tracked through time, the rest of the faults in the dataset also showed similar growth patterns, as can be seen from the isochrons of the study area (Figure 4.5). The faults studied in the physical analogue models presented in Chapter 5 also follow a hybrid fault growth model, in which they experience an initial stage of rapid lengthening by relay ramp formation and segment linkage followed by a more protracted stage of displacement accumulation (Figure 5.13); this pattern of lengthening and displacement occurs several times in concordance with a step-wise fault growth model (Pan et al., 2021).

According to the cyclical growth pattern suggested by Pan et al. (2021), faults could experience many alternating stages of lengthening and displacement as arrays develop. At the start of extension, there are many closely spaced, low-displacement faults, and as time

continues, faults that are optimally positioned will accommodate most of the strain, while smaller, low-displacement faults in stress shadows become inactive (see Pan et al., 2021).

A step-wise fault growth pattern, defined by alternating phases of quick lengthening and extended displacement accrual, was distinctly observed in the physical analogue models (Figure 6.1; Chapter 5). In the early stages of the model run (~16 mm extension; Figures 5.6 and 5.7) many small faults were visible and active. As extension continued, fault segments that were in an advantageous location to adjacent fault segments (i.e., physically close, along strike of each other, dipping the same direction) would link and an extended period of displacement accrual would follow. At the same time, smaller faults that could not physically link to adjacent segments became inactive (Figures 5.6 and 5.7). This growth pattern continued until the end of extension (Figures 5.6 and 5.7).

One stage of this fault growth pattern can be observed in seismic reflection data from Offshore NW Australia (Chapter 4); during early rifting (209.5-200 Ma) there were many faults present (Figure 4.5), but as time went on, the smaller faults stopped accruing displacement while the larger faults remained active. This documents one of the cyclical growth patterns described by Pan et al., 2021 (Figure 6.1) and seen in the physical analogue models (Chapter 5). There were likely several earlier stages of step-wise fault growth between the time of fault initiation until the faults were large enough to be resolvable in seismic data.

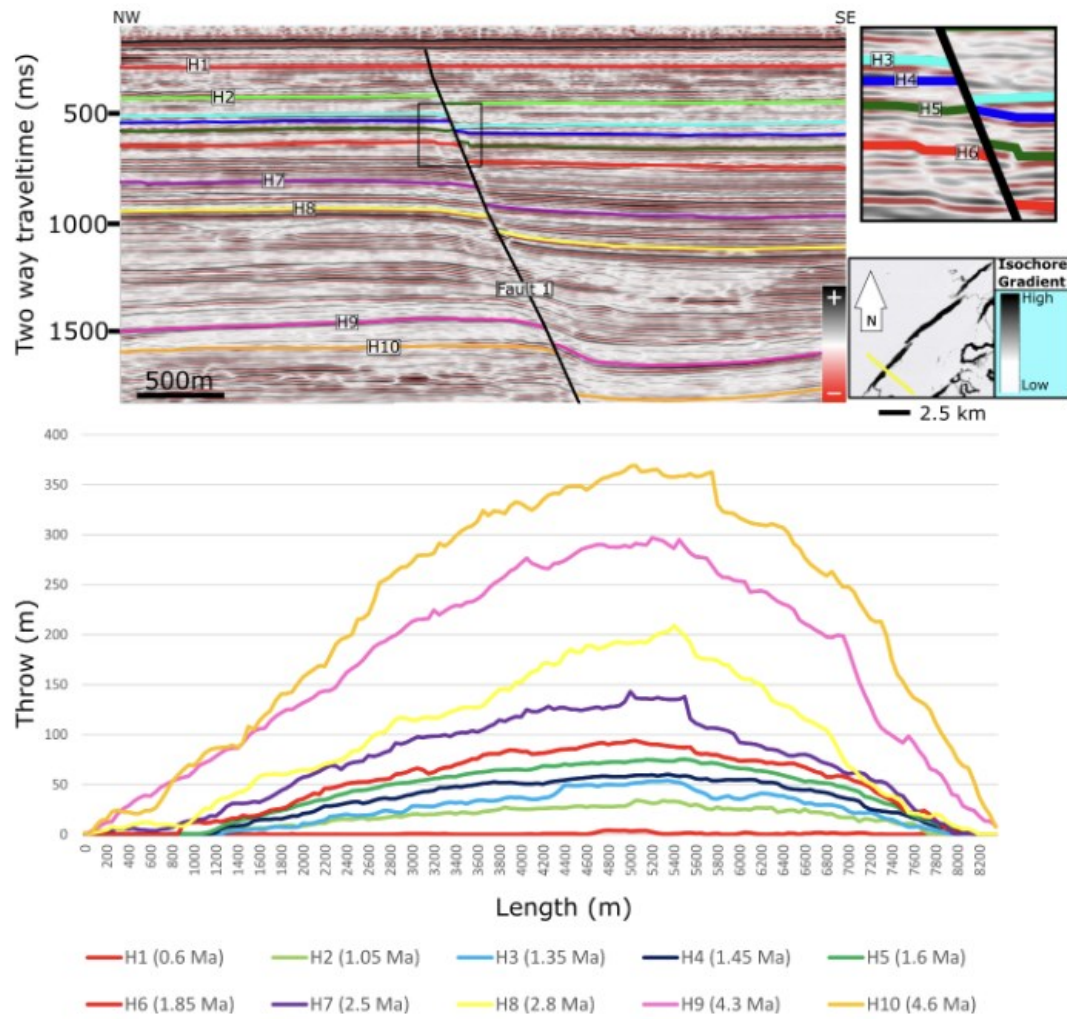


Figure 6. 2. Top: Seismic profile showing Fault 1 at its point of maximum throw, with horizons 1-10 (H1-10) marked. Bottom: Throw-distance plot displaying the lateral variations in throw across each horizon. From Kwong, 2021

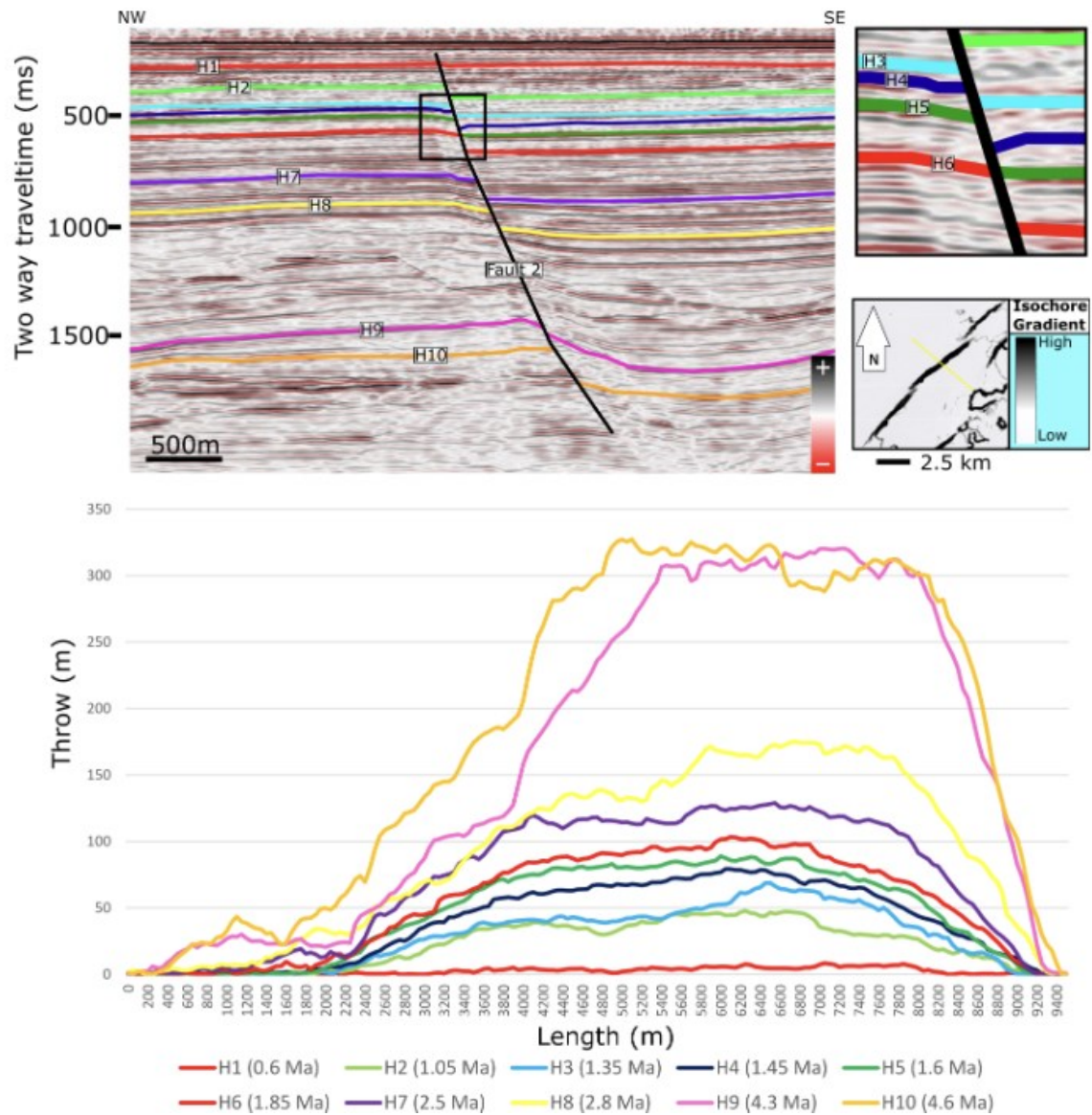


Figure 6. 3. Top: Seismic profile showing Fault 2 at its point of maximum throw, with horizons 1-10 (H1-10) marked. Bottom: Throw-distance plot displaying the lateral variations in throw across each horizon. From Kwong, 2021

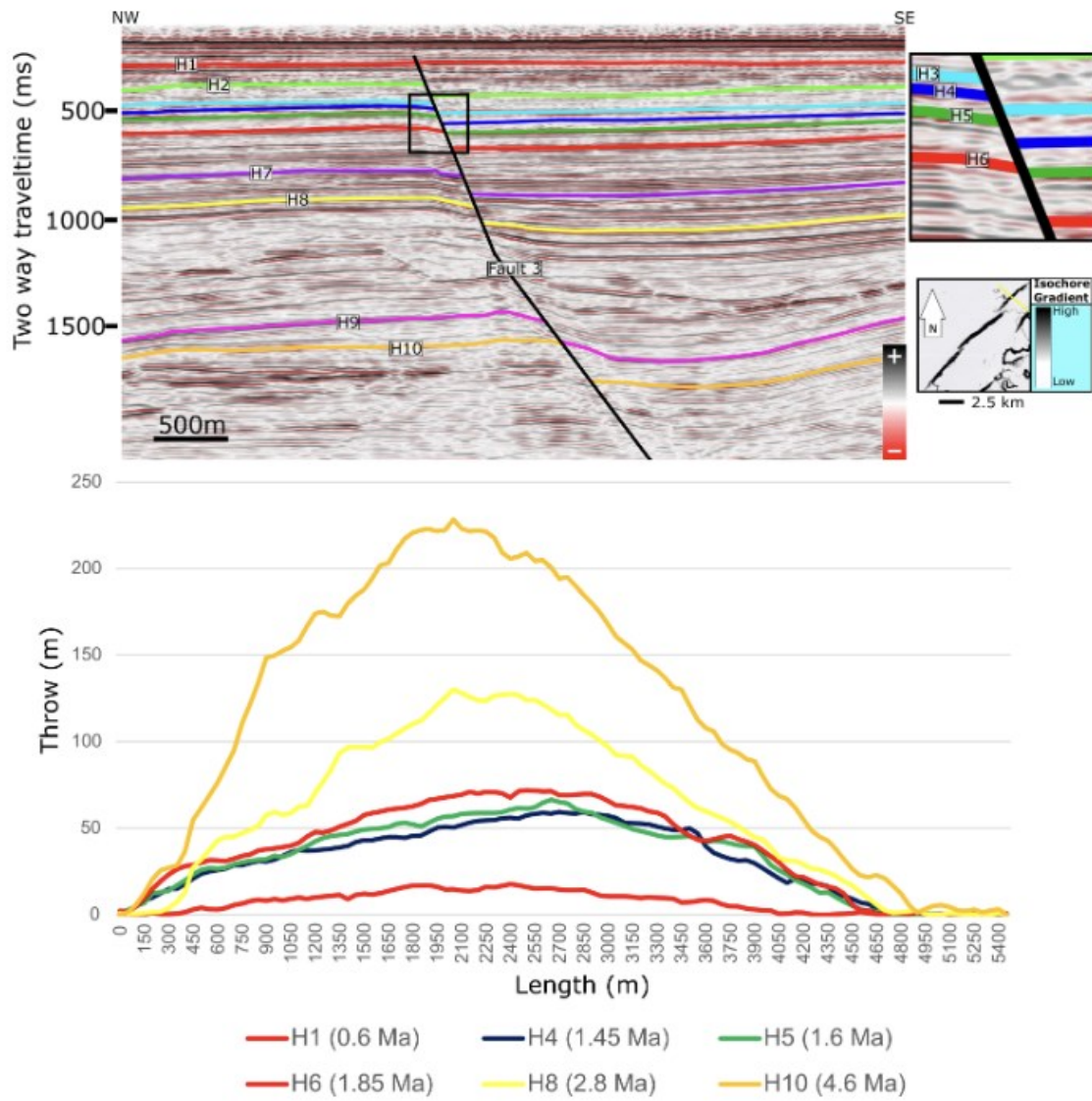


Figure 6. 4. Top: Seismic profile showing Fault 3 at its point of maximum throw, with horizons 1-10 (H1-10) marked. Bottom: Throw-distance plot displaying the lateral variations in throw across each horizon. From Kwong, 2021

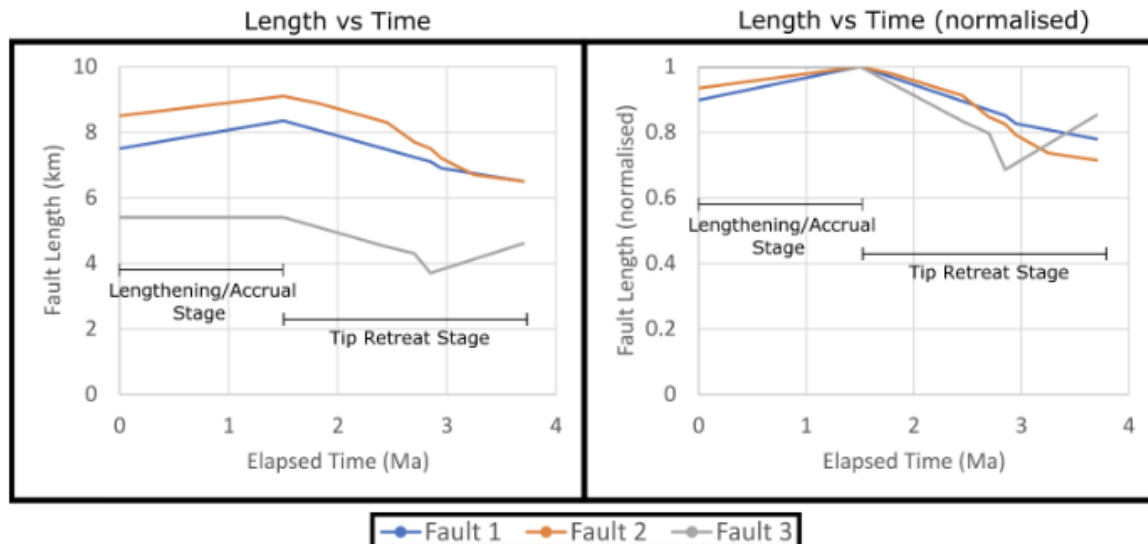


Figure 6. 5. Fault length through time for faults 1-3 calculated using results from throw backstripping. True fault length is shown on the left and normalised fault length is shown on the right. Fault experienced a quick lengthening stage followed by a stage of lateral tip retreat. From Kwong, 2021

6.4. How do normal faults die?

Normal fault nucleation and growth has been widely researched. However, how normal faults become inactive has received significantly less attention. In my Exmouth Plateau, offshore NW Australia case study (Chapter 4), as faults became inactive, strain was partitioned onto the middle of the faults and up to 2.5 km of the lateral tips became inactive, a process we refer to as lateral tip retreat. The process of tip retreat has been referred to previously, but not investigated in detail (see Meyer et al., 2002; Morley, 2002; Freitag et al., 2017; Nicol et al., 2020). In response to this, I created a detailed methodology that can help future workers identify lateral tip-line retreat (Chapter 4).

In addition to my thesis research, I designed and led two masters projects identifying additional examples of fault tip-line retreat using the Parihaka 3D seismic dataset, Taranaki Basin, Offshore New Zealand. This information is included in this thesis because there are very few examples of lateral tip-line retreat in the literature, and this provides additional insight into the later stages of fault growth. The studied faults were reactivated normal faults that experienced two distinct phases of rifting, in the Late Cretaceous-Early Miocene and then in the Late Miocene-Recent (Giba et al., 2012). Under my close supervision and instruction, students used 3D seismic data to track fault growth through time, focusing on fault displacement and length from the time of reactivation (~3.7 Ma) until the faults became inactive. Three faults were studied in detail (Figures 6.2-4), which showed the faults reaching their near-maximum length geologically instantaneously, lengthening only an additional 1-2 km during the subsequent 1.5 million years (Figure 6.5Kwong, 2021). The faults then

experienced a stage of lateral tip retreat from ~2.2 Ma-Recent, with their active traces decreasing in length by up to 3 km (Figure 6.5; Kwong, 2021). More research and additional examples of late-stage fault growth are needed, but the Exmouth Plateau, offshore NW Australia and Taranaki Basin examples provided some insights into late-stage fault growth and death.

6.5. Wider implications

The themes discussed in this thesis have important implications as society embraces the energy transition; understanding fault geometries and fault growth are important for geothermal energy exploration and CO₂ storage. In order to keep up with energy needs, known geothermal systems will need to be improved and new systems will have to be developed. Faults are important conduits in geothermal systems and can serve as pathways for upwellings (Moeck, 2014; Moek & Beardsmore, 2014; Siler et al., 2019). Appropriate fault scaling laws will be needed in geothermal exploration, especially when 3D seismic reflection and borehole, and other geophysical and geological data are not available for the area. Understanding the fault geometries of prospective CO₂ or nuclear waste storage locations will be vital to prevent unwanted leakage. CO₂ can travel along faults at a rate of 1 km/yr, which can lead to leakages into the atmosphere or water aquifers of 15000t/yr (Nicol et al., 2017). Assessing if faulting has affected cap rock integrity of storage sites will also be important (Faulkner et al., 2013; Nicol et al., 2017).

The findings in this thesis also have implications in earthquake geohazards. Currently, there is a disconnect between research that tracks fault growth and displacement on the million-year scale, thousand-year scale, and seconds/minute scale (i.e., earthquakes). Understanding how faults grow on the million-year time scale is helpful for understanding geohazards. One important area of research is understanding changes in fault slip-rate of active tectonic zones. Changes in slip-rate can have geohazard implications as earthquakes can cluster through time (van der Woerd et al., 2006; Dolan et al., 2007; Kozaci et al., 2009). There were examples of changes in fault slip-rate through time in this thesis, both in the Exmouth Plateau, offshore NW Australia faults (Chapter 3), as well as in the physical analogue models (Chapter 5). Faults can experience high rates of slip rate clusters of multiple earthquakes within a few hundred years, followed by several thousand years of low slip rates (Dolan et al., 2007). Slip-rate fluctuations on active faults are a particularly dangerous seismic hazard because it can be difficult to discern if a fault is active or inactive. A fault that is experiencing a period of low slip could be interpreted to have a low seismic hazard or be completely inactive, when in

reality the fault is experiencing a period of extended quiescence. This could result in risk to human populations, especially if buildings in the area are not prepared for damaging earthquakes or if there is a nuclear power center near a quiescent fault (Chapman et al., 2014).

6.6. Future work

My thesis has identified a number of possible future research opportunities that may help in our understanding of fault growth.

6.6.1. How does your improved knowledge of normal fault growth translate to reverse or strike-slip faulting?

Historically, normal fault kinematics have been more widely studied than reverse or strike-slip faults. More research time and money has previously been done on normal faults because normal faults form sedimentary basins, which are of great economic value. Most seismic data that is used in an academic setting was originally obtained by oil and gas companies, thus more seismic data from extensional basins exists. Additionally, normal growth faults are often well-preserved; as the hanging wall of a normal fault moves downwards, accommodation space is created for sedimentation and these sediments are less likely to be eroded (Figure 6.8). These packages can be chronostratigraphically dated and fault growth through time can be tracked. Growth packages from reverse faults are not easily preserved; as the hanging wall is thrust over the footwall, the growth package is at a topographic high and susceptible to erosion, thus making it harder to date (Figure 6.6).

Understanding reverse and strike-slip fault growth is vitally important from a geohazards perspective as these faults can be significantly larger and host more devastating earthquakes than normal faults. It is often assumed that reverse faults and normal faults have similar kinematics (briefly mentioned in Kim & Sanderson, 2005 and Torabi & Berg, 2011), and some studies on reverse fault kinematics have been done (e.g., Higgins et al., 2009; Landgraf et al., 2009; Ghassemi et al., 2014; Pizzi et al., 2020) but the comparison between reverse and normal fault kinematics has not been studied in detail.

The next step in my research could be to create reverse and strike-slip fault D/L databases following the methodology used in Chapter 3. I would see if different geological factors affect D/L relationships and compare them to normal faults. I would also compile reverse fault D/L through time data where it exists to understand if reverse faults follow a similar growth trajectory to normal faults. Additionally, new research could be carried out tracking

reverse fault growth through time using 3D seismic data, as done with normal faults in Chapter 4.

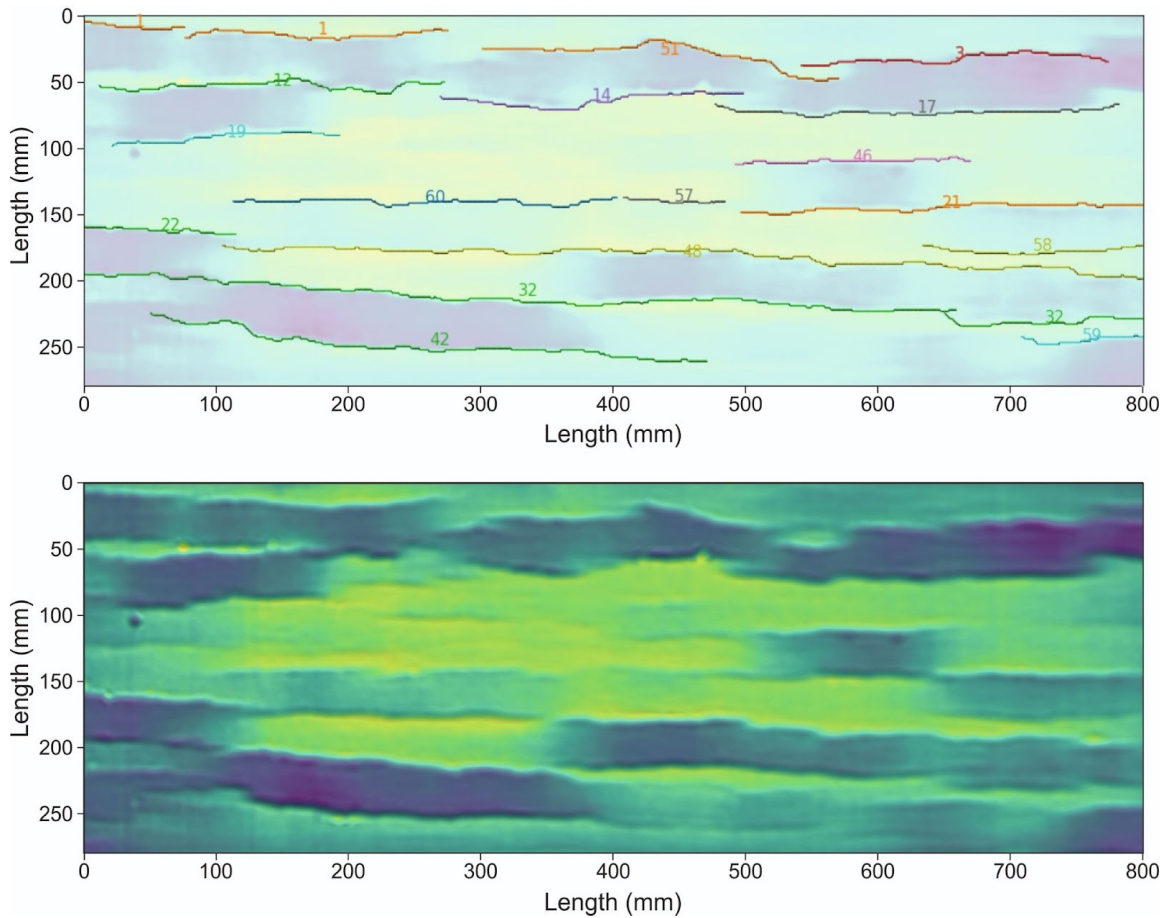


Figure 6. 6. Model 816 at 40 mm of extension imaged using a code created by Dr Nic Molnar (see Chapter 5 for details on the experiment). Data was obtained from top-view images of the model. Faults are numbered in the figure above and imaged in the figure below. A scale is not yet available due to issues with the code, however, lower areas are shown in blue and higher areas are shown in yellow.

6.6.2. Making fault analysis more efficient

Having tested various automated fault picking software made by Petrel and Traptester, it became clear that manual fault picking is currently the only reliable way to show accurate fault growth through time; however, this method is very time intensive. For example, quantifying the kinematics of one fault in my Exmouth Plateau, offshore NW Australia 3D seismic dataset can take up to a week per fault, depending on its size and the desired resolution. There are dozens of other faults in my seismic data and in my physical analogue models that could have been analysed in detail if time allowed. There are several codes being developed that could carry out fault picking, length and displacement measurement, and displacement backstripping to show fault growth through time. If fault analysis was made more efficient, we would be able to complete a detailed kinematic analysis of more faults,

which would allow us to have a better idea of fault growth than if only a few are being studied.

A code that uses our analogue models to pick faults, measure maximum length and displacement, and track fault growth and segment linkage of faults through time, including when segments link based on our analogue models is currently being created by Dr. Nic Molnar (Aachen University) with my input. Unfortunately, early-stage fault growth could not be quantified using the tools that I have available now, as the resolution was not high enough to measure and track low-displacement faults in the 3D models created by the CT scans. However, we are able to trace faults from the surface of the models using elevation data that was obtained by PIV analysis (see Chapter 5) and Dr. Molnar's code (Figures 6.5 and 6.6). The code is not yet good enough to track faults through time as is wanted, but in the future this should be possible. We are working on this with my analogue models, but eventually this could be modified to be used to conduct automated displacement backstripping of syn-sedimentary normal faults in 3D seismic data.

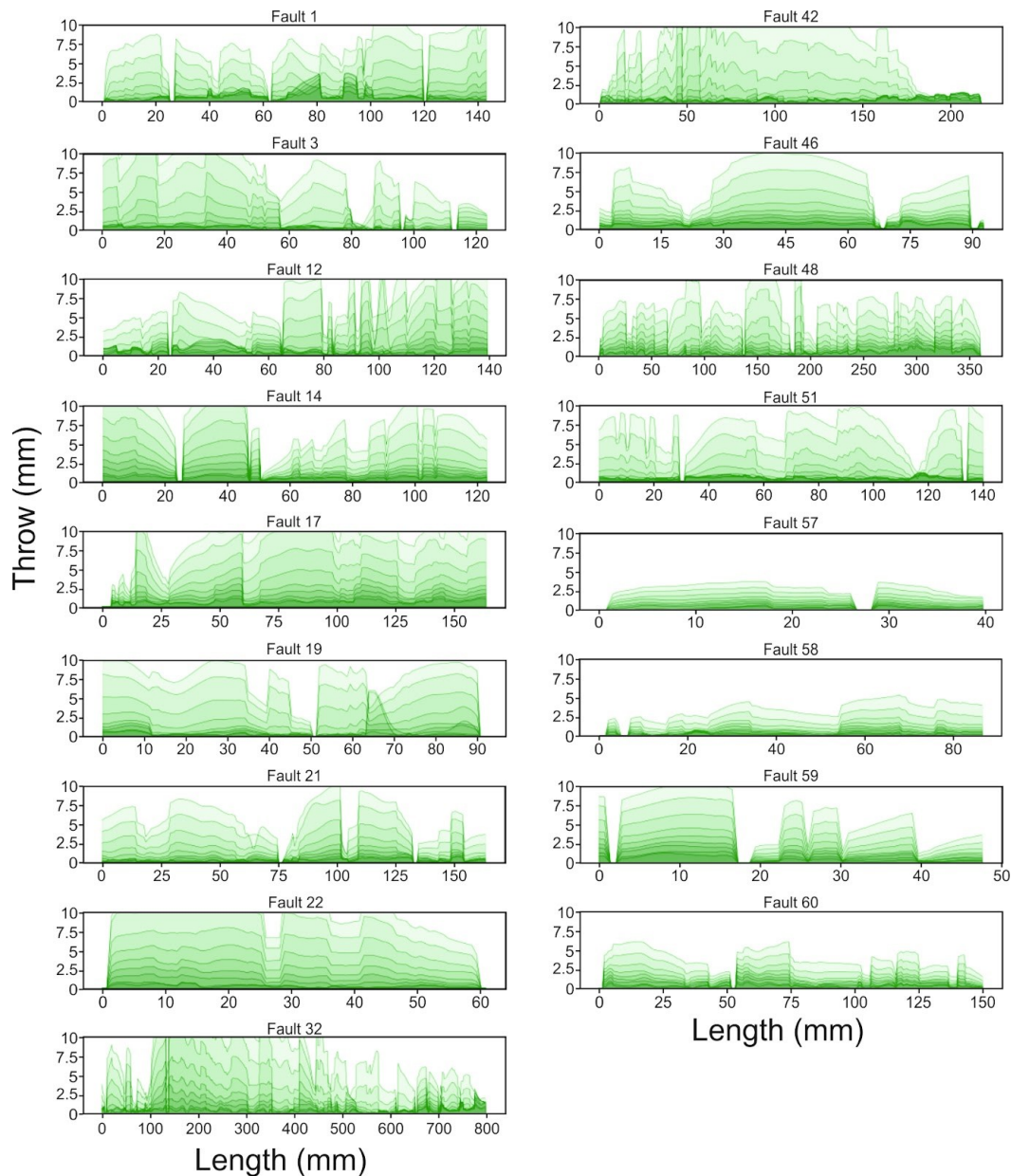


Figure 6. 7. Fault throw and length through time from the major folds in Experiment 816 obtained using the code from Dr Nic Molnar. Throw and length at every 2 mm of extension is shown for each fault. The fault numbers correlate to the numbers listed in Figure 6.6. There are still some problems with this interpretation and perfecting the code is an ongoing project.

6.7 Conclusions

Overall, I have concluded that individual normal faults follow the following growth pattern: a lengthening stage, displacement stage, and possible tip retreat stage. During the lengthening stage (10-30% of a fault's active life), faults initiate and quickly establish the bulk of their length quickly (over 90%). During this time, some displacement is accrued (put number here). Some faults can reach their near-maximum length in less than 10% of their active life. Faults then enter the fault displacement stage (30-75% of a fault's active life). Average fault displacement is constant throughout, however changes in displacement rate are common on

an area-by-area basis. Some faults experience a phase of lateral tip-line retreat in the last 25% of fault activity as stress is localised onto the middle of the fault and displacement is no longer accommodated on one or both fault tips. On a fault array scale, faults also follow a stepwise growth pattern, where faults alternate between the lengthening and displacement stages. Faults that are optimally positioned to adjacent fault segments (i.e., synthetic boundaries with the adjacent fault and aligned along strike) will link and then continue to accrue displacement, while faults that are in stress shadows or in places that are not optimal for additional growth via segment linkage will become inactive.

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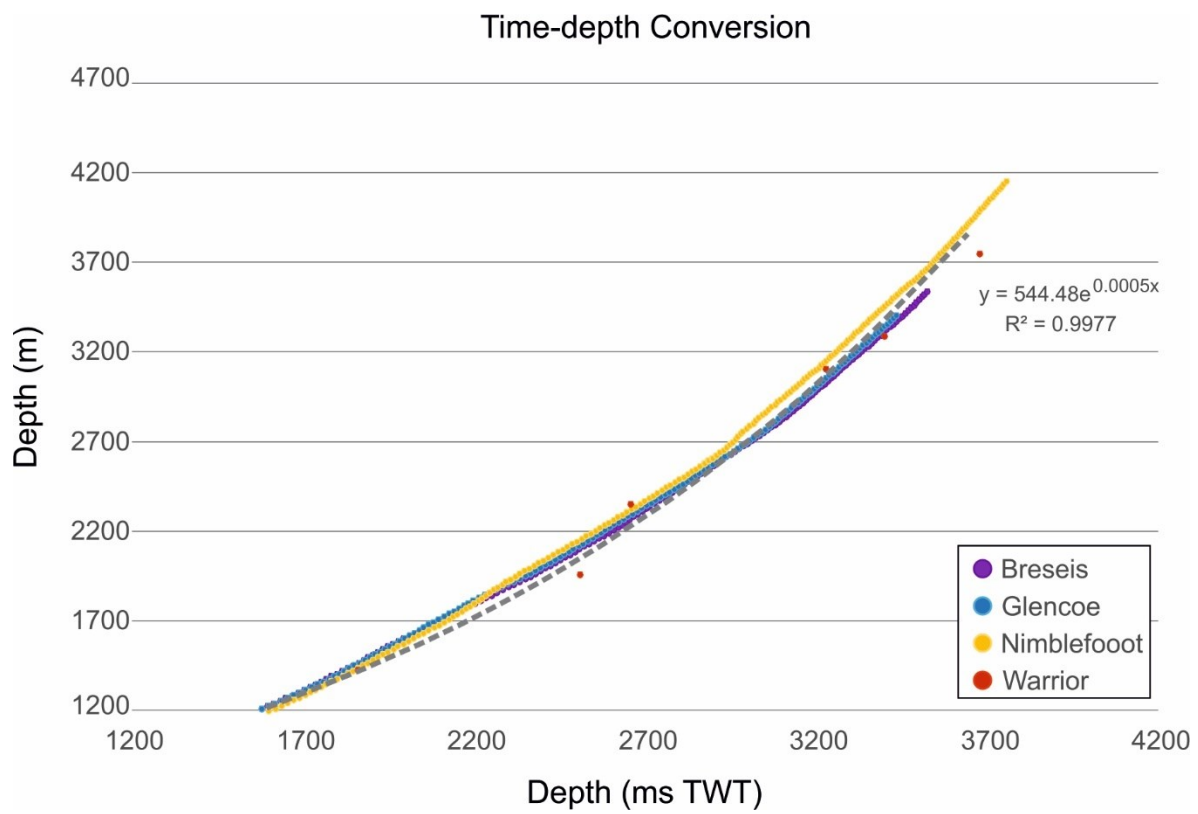
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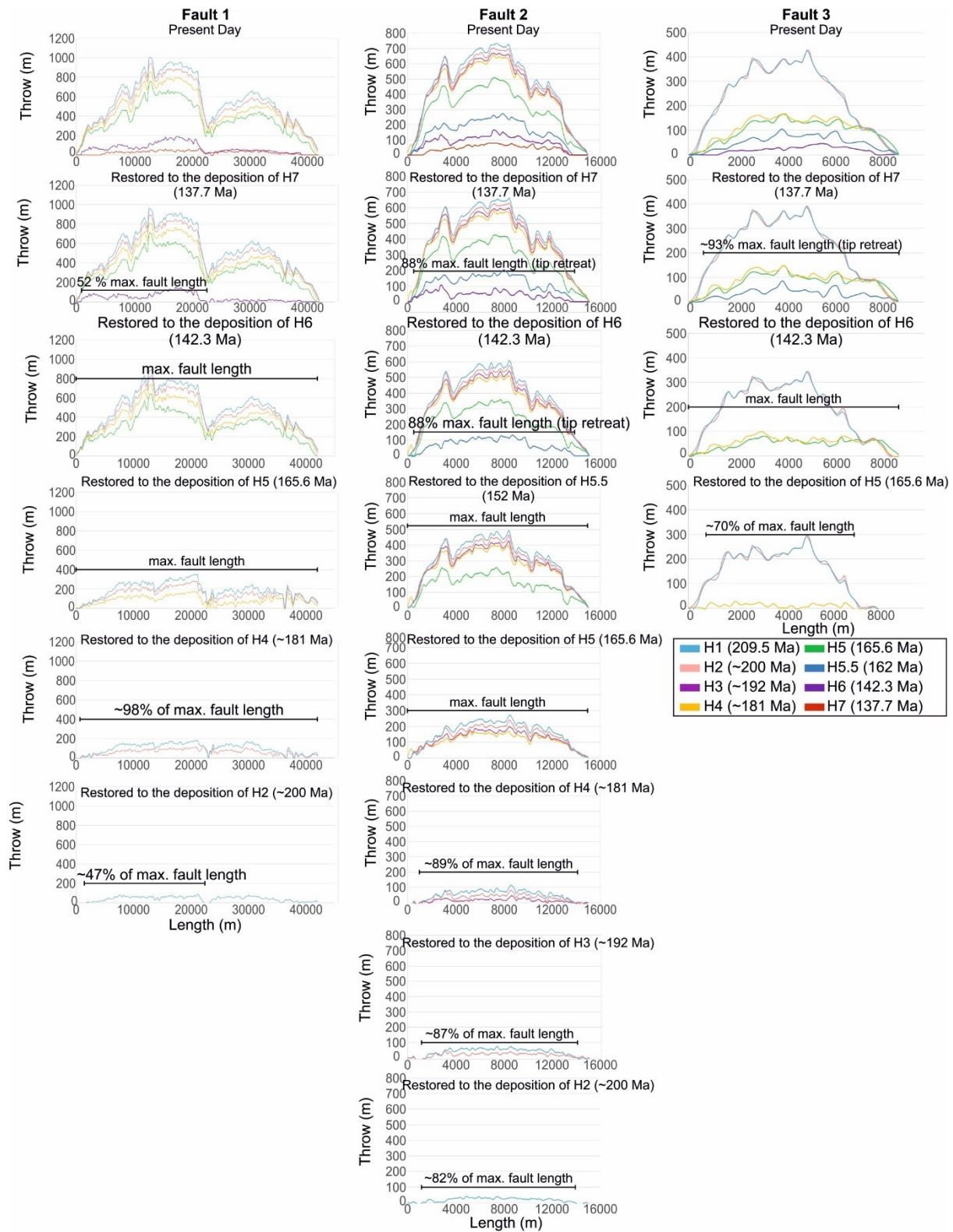
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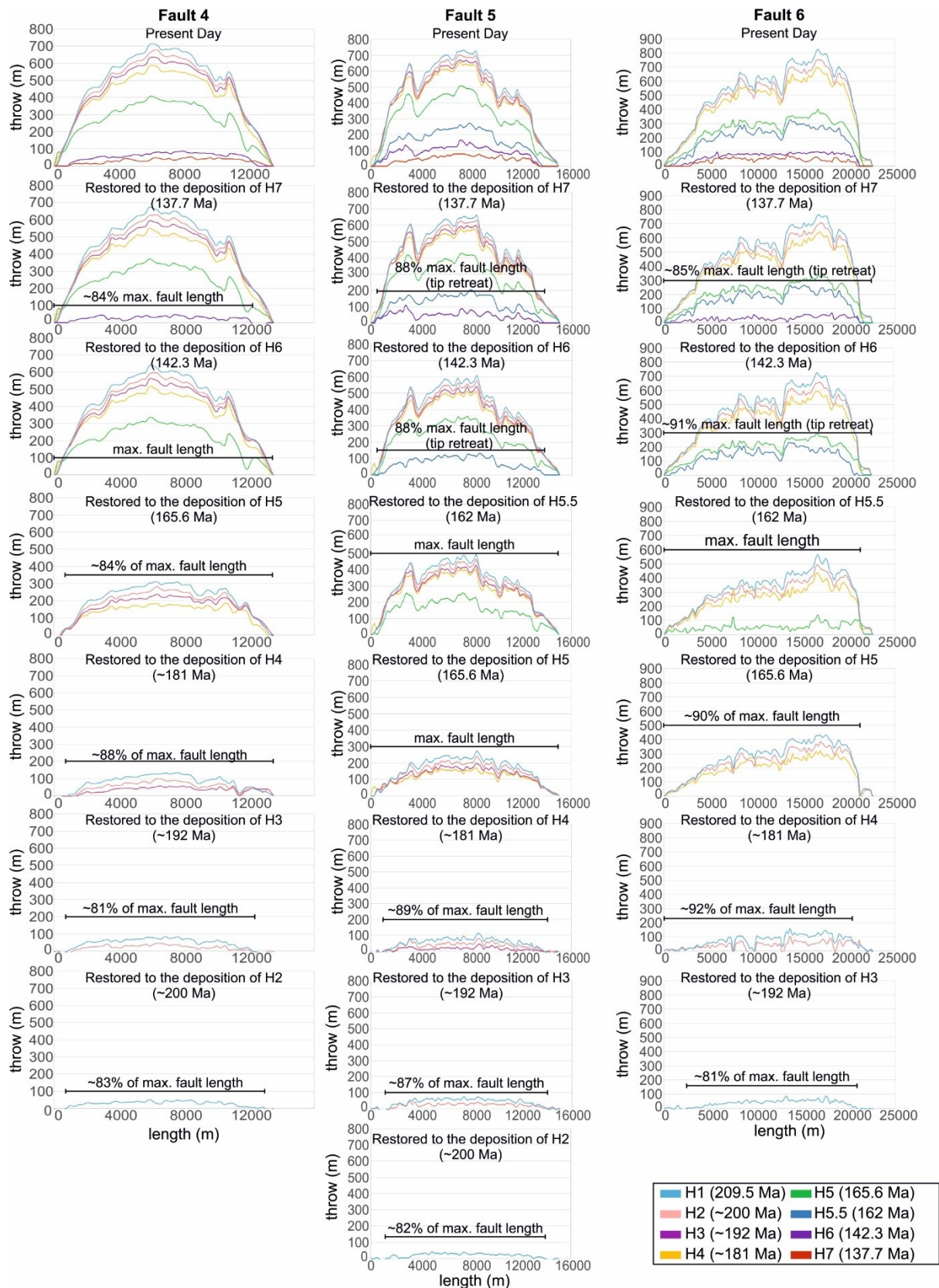
Appendix



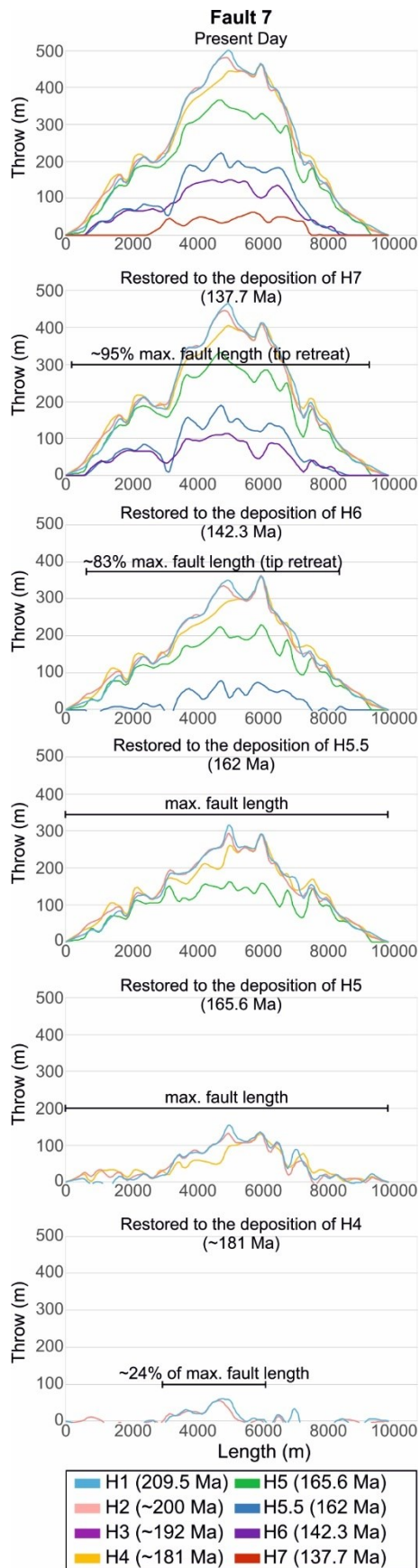
Appendix figure 4. 1. Time-depth conversion for the 4 four wells in the study area: Breseis-1, Glencoe-1, Nimblefoot-1, and Warrior-1.



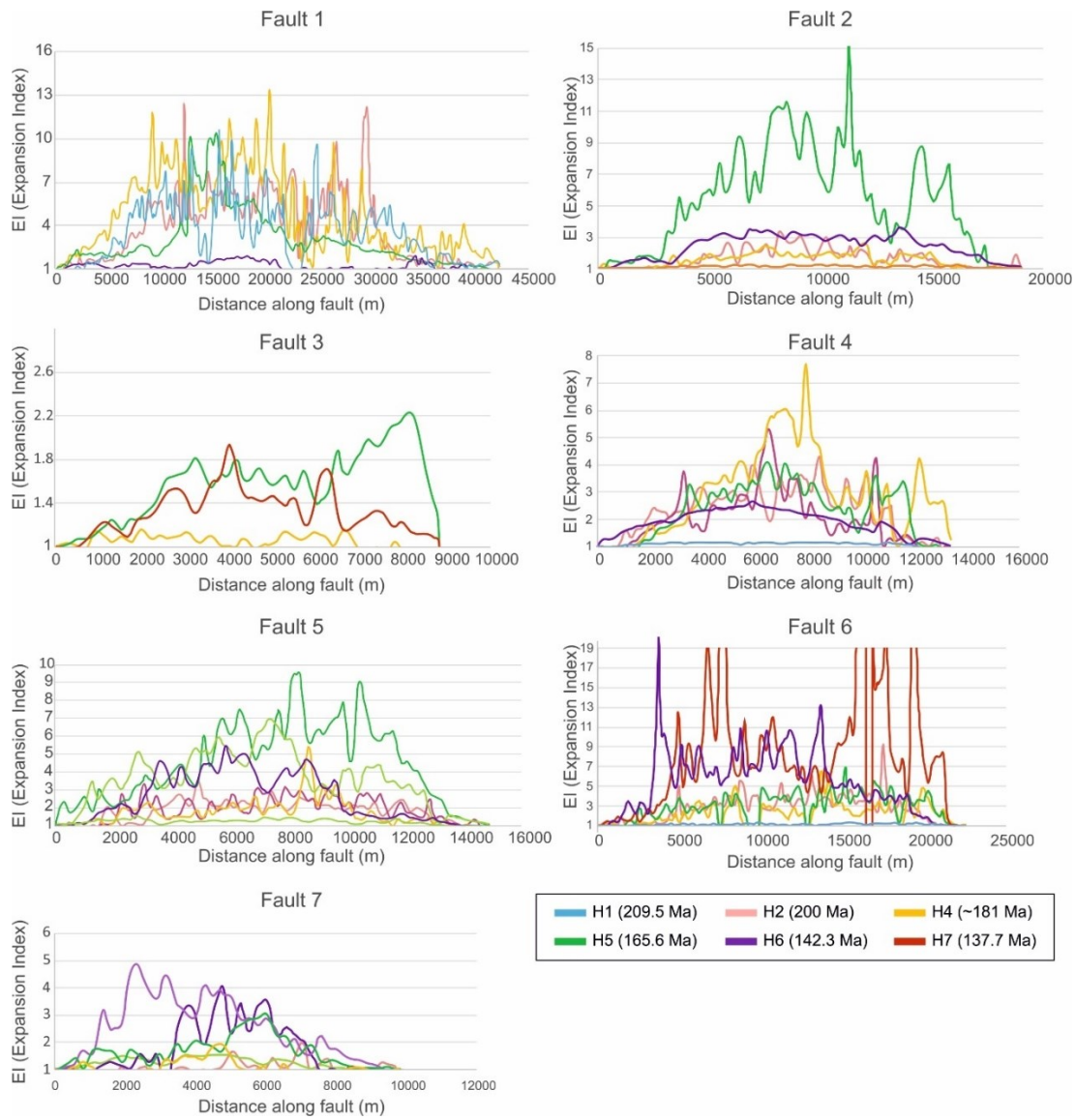
Appendix figure 4. 2. Throw backstripping faults 1-3.



Appendix figure 4. 3. Throw backstripping faults 4-6.



Appendix figure 4. 4. Throw backstripping fault 7.



Appendix figure 4. 5. EI Analysis for faults 1-7.

Fault	Horizon	Time (Ma)	Time Accrued (Myr)	Length (m)	Throw (m)	Slip rate (m/myr)
1	H1 (start)	209.5	0	0	0	0
	H2	200.1	9.4	19,700	78	9.3
	H4	181.3	28.2	41,300	181	5.1
	H5	165.6	43.9	42,100	351	11.2
	H6	142.3	67.2	42,100	890	23.1
	H7	137.7	71.8	21,900	963	16.3
2	H1 (start)	209.5	0	0	0	0
	H2	200.1	9.4	11,500	55	6.0
	H4	181.3	28.2	14,700	116	3.4
	H5	165.6	43.9	17,100	303	12.4
	H6	142.3	67.2	19,000	550	10.8
	H7	137.7	71.8	13,900	578	6.5
3	H4 (start)	181.3	0	0	0	0
	H5	165.6	15.7	6200	28	2.3
	H5.5	152	23.9	8800	100	6.3
	H6	142.3	39	8300	148.5	7.5
4	H1 (start)	209.5	0	0	0	0
	H2	200.1	9.4	11,200	50	5.3
	H3	190.7	18.8	10,900	84	3.6
	H4	181.3	28.2	11,800	135	5.4
	H5	165.6	43.9	11,200	313	11.3
	H6	142.3	67.2	13,400	641	14.1
	H7	137.7	71.8	11,300	673	7.0

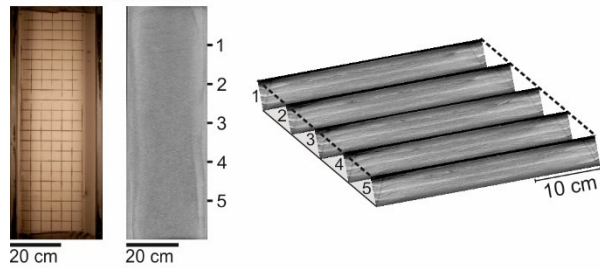
Appendix figure 4. 6. Table showing fault horizon age, length, throw, and slip rate for each horizon of faults 1-4.

Fault	Horizon	Time (Ma)	Time Accrued (Myr)	Length (m)	Throw (m)	Slip rate (m/myr)
5	H1 (start)	209.5	0	0	0	0
	H2	200.1	9.4	12,300	40	4.4
	H3	190.7	18.8	13,000	74	3.6
	H4	181.3	28.2	13,200	115	4.8
	H5	165.6	43.9	14,900	270	10.6
	H5.5	152	57.5	14,900	495	17.9
	H6	142.3	67.2	13,100	609	12.9
	H7	137.7	71.8	13,100	662	13.0
6	H1 (start)	209.5	0	0	0	0
	H2	200.1	9.4	18,300	88	9.9
	H4	181.3	28.2	20,700	157	3.8
	H5	165.6	43.9	20,200	432	16.8
	H5.5	152	57.5	22,500	565	12.4
	H6	142.3	67.2	20,400	716	17.9
	H7	137.7	71.8	19,200	766	9.1
7	H2 (start)	200.1	0	0	0	0
	H4	181.3	18.8	2400	62	3.2
	H5	165.6	34.5	9700	155	5.9
	H5.5	152	48.1	9700	316	12.2
	H6	142.3	57.8	9200	362	6.2
	H7	137.7	62.4	8100	465	20.1

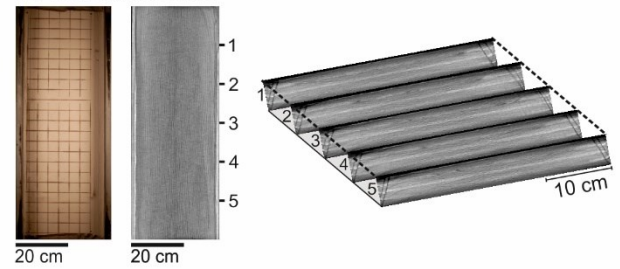
Appendix figure 4. 7. Table showing fault horizon age, length, throw, and slip rate for each horizon of faults 5-7.

Experiment 808

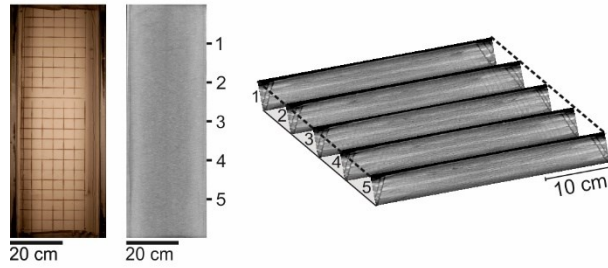
4 mm extension



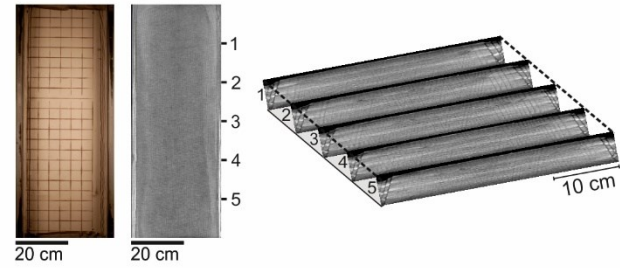
10 mm extension



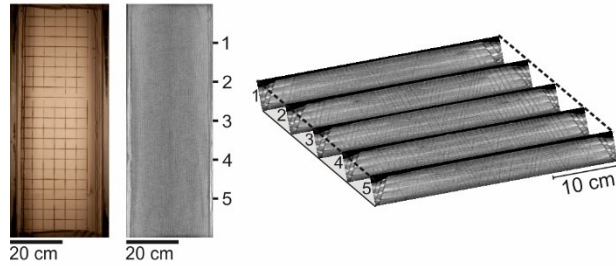
16 mm extension



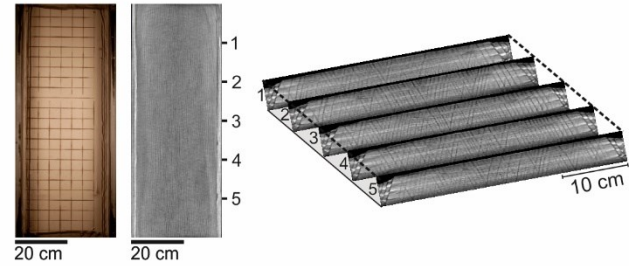
22 mm extension



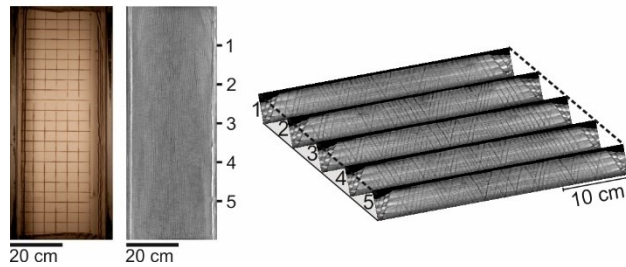
28 mm extension



34 mm extension



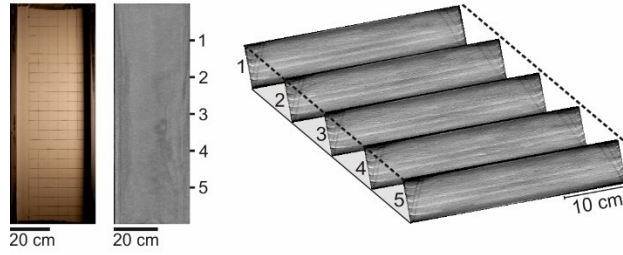
40 mm extension



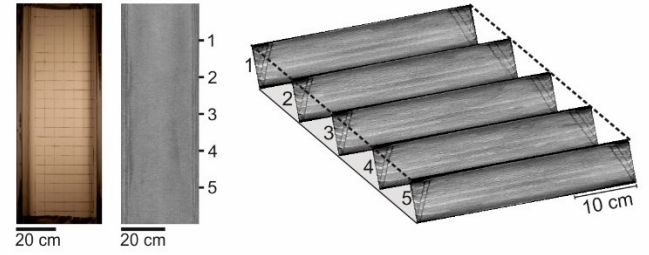
Appendix figure 5. 1. Progression of experiment 808 at 4, 10, 16, 22, 28, 34, and 40 mm extension. Photographs of the top view, top of the model in X-Ray view, and cross sections across the model are shown. Details on model 808 can be found on Table 5.1.

Experiment 809

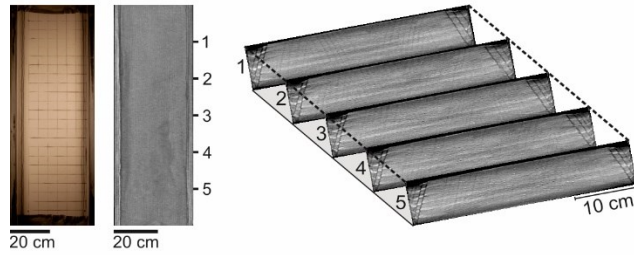
4 mm extension



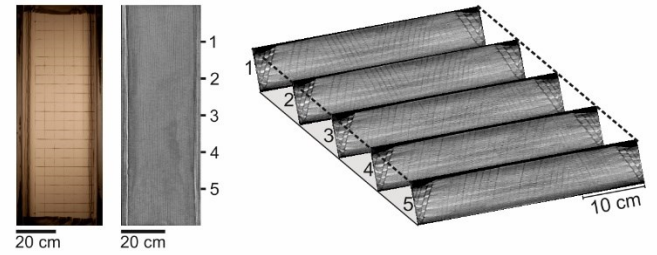
10 mm extension



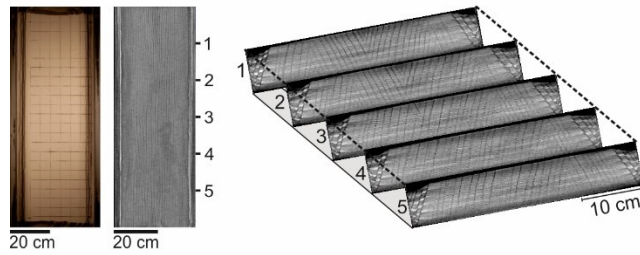
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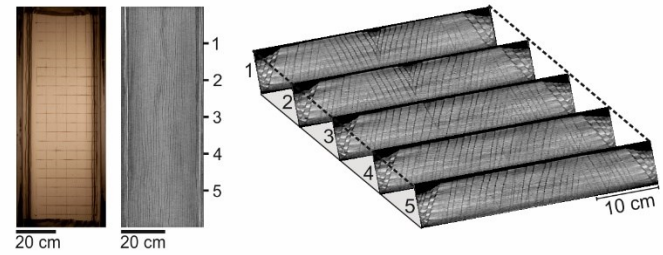
22 mm extension



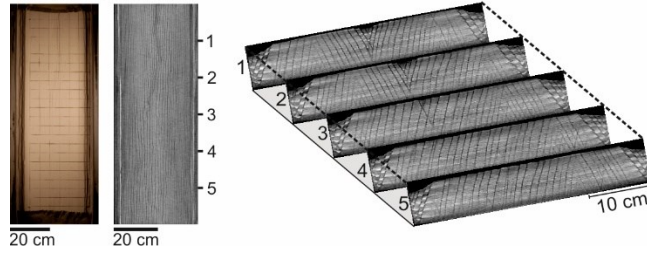
28 mm extension



34 mm extension



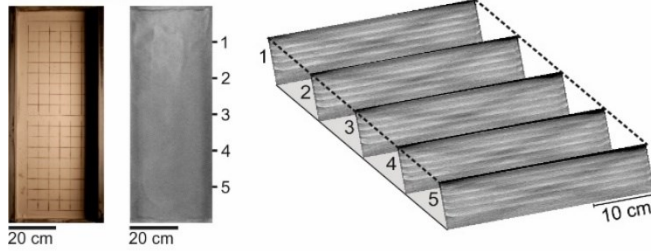
40 mm extension



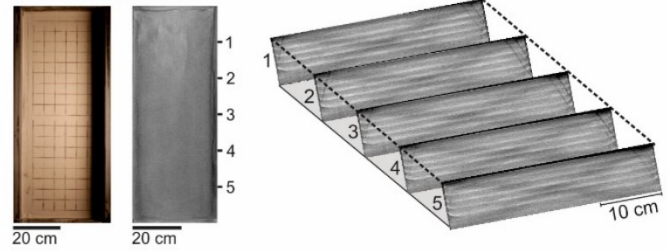
Appendix figure 5. 2. Progression of experiment 809 at 4, 10, 16, 22, 28, 34, and 40 mm extension. Photographs of the top view, top of the model in X-Ray view, and cross sections across the model are shown. Details on model 808 can be found on Table 5.1.

Experiment 815

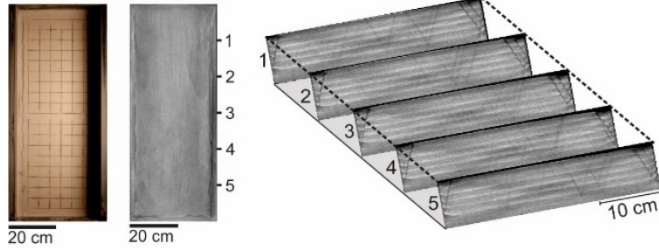
4 mm extension



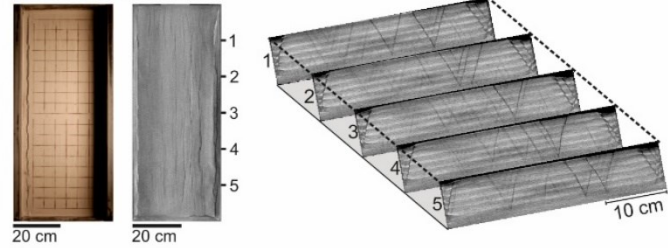
10 mm extension



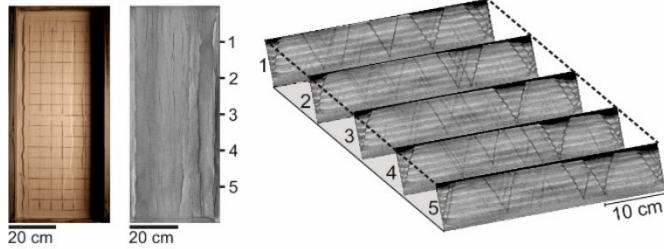
16 mm extension



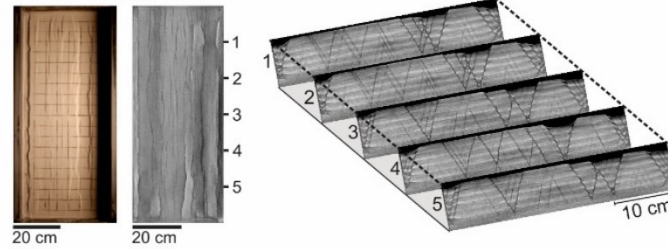
22 mm extension



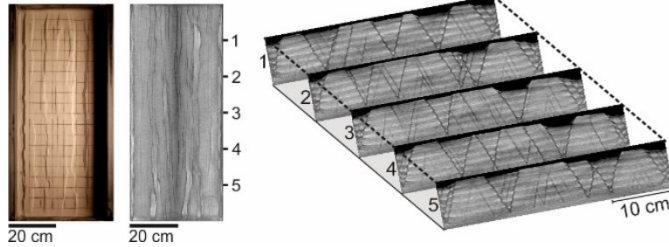
28 mm extension



34 mm extension



40 mm extension



Appendix figure 5. 3. Progression of experiment 815 at 4, 10, 16, 22, 28, 34, and 40 mm extension. Photographs of the top view, top of the model in X-Ray view, and cross sections across the model are shown. Details on model 808 can be found on Table 5.1.