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**UNIVERSITY OF SOUTHAMPTON**

Faculty of Environmental and Life Sciences
School of Ocean and Earth Science

# Faults, Fluids and Geohazards at Subduction Zones

*by*

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*A thesis for the degree of*
*Doctor of Philosophy*

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Abstract

Faculty of Environmental and Life Sciences
School of Ocean and Earth Science

Doctor of Philosophy

**Faults, Fluids and Geohazards at Subduction Zones**

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Large magnitude earthquakes ($M_W$ 7.5) that occur within marine settings have the potential to generate powerful tsunamis with devastating consequences for coastal communities. The 2004 $M_W$ 9.2 Aceh-Andaman megathrust earthquake and 2012 $M_W$ >8.0 intraplate earthquakes offshore North Sumatra served to highlight our limited understanding of the extent of seismogenesis associated with accretionary subduction zone margins, and the kinematics of oceanic intraplate deformation.

International Ocean Discovery Program (IODP) drilling into the sediments on the incoming plate, in the Wharton Basin offshore North Sumatra in 2016 has provided new information on sediment depositional history, lithology and physical properties. These new data have been coupled with seismic reflection profiles and swath bathymetry to make detailed analyses of intraplate fault activity and to generate thermal models of the subduction zone input sediments. Pervasive faulting through the input sediments is the result of Riedel shearing in response to deeper left-lateral strike-slip deformation in the oceanic crust and upper mantle, which agrees with the complex rupture pattern of the 2012 earthquakes. Deformation and the timing of fault activity in the Wharton Basin is contemporaneous with activity throughout a diffuse plate boundary hosted by the wider Indian Ocean. The deformation is likely driven, and intensified, by an acceleration of the convergence between the Indian and Australian sub-plates around 7-9 Ma. Thermal models of the subduction zone input sediments indicate that high rates of accumulation and deep burial generates high temperatures, in excess of the requirements for thermally driven mineral dehydration, within the basal sediments before they reach the point of accretion/subduction. Fluids are liberated from the basal sediments since 1 Ma by the smectite-illite transformation in sufficient quantities to explain seismic reflection polarity reversal on the North Sumatra pre-décollement. The interplay between lithology, sediment dehydration, faulting and hydrogeology within the incoming sediments control the position at which the décollement forms.

The southern Hikurangi margin is analogous to North Sumatra as an accretionary margin, exhibiting along-strike changes in: input sediment thickness, obliquity of convergence, and subduction velocity; with related variations in accretionary prism morphology and fault structure. Using widely spaced seismic reflection profiles (including new

data acquired in 2017 by the Seismogenesis at Hikurangi Integrated Research Experiment (SHIRE) processed in this study) that image the southern Hikurangi incoming plate and accretionary prism integrated with high resolution bathymetry, structural variations and their relationship to along-strike changes in the subduction zone configuration are described. The southern Hikurangi margin can be divided into three segments with contrasting characteristics: an eastern segment characterised by a moderately wide ( 70 km), low taper ( 5°) prism with rapid outward growth since 1 Ma, where seismically resolvable deformation accounts for 20% of crustal shortening, comparable with the central Hikurangi margin; a central segment characterised by a narrow ( 40 km), moderate taper ( 8°) prism, with similar shortening to the eastern segment, but taking place earlier (between 2 and 0.6 Ma), and where outward growth has effectively ceased; and a western segment where the prism widens (to 50 km) via rapid outward migration of the deformation front and utilisation of pre-existing incoming fault structure, accommodating very little to no crustal shortening since 2 Ma. These contrasting prism characteristics are interpreted to relate primarily to the westward increasing proximity of the Chatham Rise ancient accretionary complex, changing subduction velocities, and relative prominence of strike-slip faulting in the prism interior.

# Contents

List of Figures ix

1 Introduction 1
1.1 Earthquake Mechanics and the Seismo- and Tsunami-genic Potential of Subduction Zones . . . 3
1.2 Accretionary Margins . . . 8
1.3 Deformation, Composition and Seismicity of the Incoming Plate . . . 9
1.4 Along-strike variation of outer subduction zone structure . . . 12
1.5 Research Aims . . . 14
1.6 Thesis Overview . . . 16

2 A Complete Structural Model and Kinematic History for Distributed Deformation in the Wharton Basin 19
2.1 Introduction . . . 20
2.2 Methods . . . 24
2.2.1 MCS data and interpretation . . . 24
2.2.2 Integration with IODP Expedition 362 borehole data . . . 25
2.2.3 Multibeam processing and interpretation . . . 26
2.2.4 Vertical offset measurements . . . 26
2.3 Results . . . 27
2.3.1 Biostratigraphy (from IODP) . . . 27
2.3.2 Fault geometry . . . 27
2.3.3 Class A faults . . . 29
2.3.4 Class B faults . . . 30
2.3.5 Class C faults . . . 30
2.3.6 Class D faults . . . 32
2.3.7 Fault displacement and dip direction variations from the NER to the subduction zone . . . 34
2.3.8 Relationships between seismic and core-scale faulting and sediment material properties . . . 34
2.4 Discussion . . . 36
2.4.1 Fault sets across the northern Wharton Basin . . . 36
2.4.2 Activity history of intraplate strike-slip faulting in the Wharton Basin . . . 36
2.4.3 Compressional deformation in the Wharton Basin . . . 39
2.4.4 Strike-slip related fault patterns and mechanisms . . . 40
2.4.5 Changes in deformation from the Ninety East Ridge to the Sunda subduction zone . . . 42

2.4.6 Relationship between fault displacement patterns at a range of scales and material properties . . . 43
2.5 Conclusions . . . 43

3 Evolution of the Thermal and Dehydration State of Sediments Entering the North Sumatra Subduction Zone 45
3.1 Introduction . . . 46
3.1.1 Tectonic Setting . . . 48
3.2 Methods . . . 52
3.2.1 Datasets . . . 52
3.2.2 Sedimentary Parameters . . . 52
3.2.3 Regional Heat Flow . . . 53
3.2.4 Thermal Modeling and Input Parameters: Time-Depth Conversion, Thermal Rejuvenation and Radiogenic Heat Production . . 54
3.2.5 Dehydration Calculations . . . 57
3.3 Results . . . 59
3.3.1 Initial Thermal Model Tests . . . 59
3.3.2 Burial History . . . 61
3.3.3 Temperature Structure and Progression of Mineral Dehydration Through the Sediment Column . . . 62
3.4 Discussion . . . 65
3.4.1 Heat Flux Across the Incoming Section . . . 65
3.4.2 Mechanisms for Trapping of Diagenetic Fluids and Explanation of Seismic Properties . . . 67
3.4.3 Thermal Structure In uence on D écollement Development and Seismic Behavior . . . 69
3.5 Conclusions . . . 72

4 Variations in Structure on the Southern Hikurangi Subduction Zone 75
4.1 Introduction . . . 76
4.2 Tectonic Setting . . . 79
4.3 Methods . . . 82
4.3.1 Data . . . 82
4.3.2 Stratigraphic interpretation and depth conversion . . . 82
4.3.3 Progressive Retro-Deformation . . . 84
4.4 Results . . . 86
4.4.1 Incoming plate structure . . . 87
4.4.1.1 Segment A (eastern segment) . . . 88
4.4.1.2 Segment B (central segment) . . . 89
4.4.1.3 Segment C (western segment) . . . 92
4.4.2 Prism structure . . . 92
4.4.2.1 Segment A Prism . . . 92
4.4.2.2 Segment B Prism . . . 95
4.4.2.3 Segment C Prism . . . 96
4.4.3 Reconstruction of Restored Transects and Fault Activity . . . 98
4.4.3.1 Pro le A Restoration . . . 98
4.4.3.2 Pro le B Restoration . . . 100
4.4.3.3 Pro le C Restoration . . . 102

4.5 Discussion . . . . . 103
4.5.1 Potential controls of structural change along the southern Hikurangi margin . . . . . 104
4.5.2 Global signi cance and comparison to other subduction zone margins . . . . . 107
4.6 Conclusions . . . . . 111

5 Synthesis and Concluding Remarks 113
5.1 Summary . . . . . 113
5.2 Key Results . . . . . 116
5.3 Discussion . . . . . 117
5.4 Limitations . . . . . 121
5.5 Future Work . . . . . 123

Appendix A Supporting Information for Chapter 2 125

Appendix B Supporting Information for Chapter 3 133
Appendix B.1 Stress Periods . . . . . 134
Appendix B.2 Bulk thermal conductivity analysis . . . . . 134
Appendix B.3 Radiogenic heat production implemented in models . . . . . 135
Appendix B.4 Initial nodes . . . . . 135
Appendix B.5 General model parameters without data constraints . . . . . 136

Appendix C Supporting Information for Chapter 4 155

References 157

# List of Figures

1.1 Schematic diagrams of the two end member subduction zone margin types. . . . . 2
1.2 Slip on a fault caused by elastic rebound. . . . . 4
1.3 Schematic diagrams explaining the critical slip distance L. . . . . 5
1.4 Plan view schematic of contact asperities on a fault plane demonstrating the control on the state variable q. . . . . 7
1.5 Sketch showing the main components of the marine forearc at accretionary margins, as well as the locations of key processes. . . . . 10
1.6 Rupture segments from historical megathrust earthquakes along the Sunda subduction zone. . . . . 13

2.1 Regional Tectonic Con guration, Seismicity and Datasets . . . . . 21
2.2 Regional seismic pro le B and position of IODP Site U1481 . . . . . 24
2.3 Interpolation of re ector ages from core-seismic integration . . . . . 25
2.4 Distribution of sea oor faulting interpreted from multibeam bathymetry 28
2.5 Example Class A fault on NW-SE trending seismic pro le C . . . . . 29
2.6 Analysis of Class B and C faulting . . . . . 31
2.7 Examples of fault displacement pro les for different fault classes plotted with depth/sediment age alongside strain of faults west of the NER and plate motion history through time. . . . . 33
2.8 Example Class D reverse fault structure on NW-SE trending pro le C . . 35
2.9 Along-pro le B fault displacement analysis . . . . . 38
2.10 Compiled fault map and structural model of the northern Wharton Basin 41

3.1 Regional tectonics, plate age, and sea oor heat ux. . . . . 49
3.2 Interpretation of seismic pro le 1, with Site U1480, re ectors R1-14 used for modeling, and major lithostratigraphic units. . . . . 50
3.3 Age information at Site U1480 and extrapolation to deformation front (DF). 51
3.4 Natural Gamma Ray logging data for uncased intervals of sites U1480 and U1481 and conversion to radiogenic heat production based on the method of Bücker and Rybach (1996). . . . . 56
3.5 Burial history, thermal evolution, and uid release rate at re ector R11 and top basement at Site U1480, Pseudo site 1, and Pseudo Site 2, over the last 9.3 Myr (deposition of Nicobar Fan and trench wedge). . . . . 62
3.6 Thermal structure evolution of the incoming sediment column across seismic pro le 1, and corresponding modeled sea oor heat ux. . . . . 64
3.7 Smectite dehydration state of the incoming sediment column across seismic pro le 1, and corresponding uid thickness generated based on the Pytte and Reynolds (1989) kinetic. . . . . 68

3.8 Cumulative total uid expelled from sediments underlying re ector R11 along seismic pro le 1 broken down into constituent sources. . . . . 70

4.1 Tectonic setting of the study area within the wider Hikurangi subduction zone. . . . . 80
4.2 Stratigraphic interpretation and velocity structure. . . . . 83
4.3 Structure and segmentation of the southern Hikurangi margin. . . . . 88
4.4 Seismic Pro le A (MH44). . . . . 90
4.5 Seismic Pro le B (PEG09-017). . . . . 91
4.6 Seismic pro le C (PEG09-009). . . . . 93
4.7 Retro-deformation restoration results for depth converted transect of seismic pro le A. . . . . 99
4.8 Retro-deformation restoration results for depth converted transect of seismic pro le B. . . . . 101
4.9 Retro-deformation restoration results for depth converted transect of seismic pro le C. . . . . 103

Appendix A.1 Seismic Pro le A from Figure 2.1 . . . . . 126
Appendix A.2 Expanded un-interpreted- (X1) and interpreted- (X1') seismic section indicated on Figure 2.2 . . . . . 127
Appendix A.3 Expanded un-interpreted- (X2) and interpreted- (X2') seismic section indicated on Figure 2.2 . . . . . 128
Appendix A.4 Seismic Pro le C from Figure 2.1 . . . . . 129
Appendix A.5 Expanded multibeam bathymetry map showing example surface trace associated with Class C faults. . . . . 130
Appendix A.6 Comparison of fault throw/displacement with depth measurements with data from IODP expedition 362 borehole U1480 data. . . 131

Appendix B.1 Seismic pro le 1 (composite of continuous multichannel seismic lines BGR06-101 and 102). . . . . 137
Appendix B.2 Seismic pro le 2 (composite of continuous multichannel seismic lines BGR06-103, 104 and 105). . . . . 138
Appendix B.3 Porosity-depth relationship at drill site U1480. . . . . 139
Appendix B.4 Analysis of Site U1480 core thermal conductivity measurements. 139
Appendix B.5 The effects of sedimentation on sur cial (sea oor) heat ux at Pseudo Site localities. . . . . 140
Appendix B.6 Downhole and lithological gamma ray variation at U1480. . . 141
Appendix B.7 Evolution of the silica dehydration uid source rate in the sediment column along seismic pro le 1. . . . . 142
Appendix B.8 Evolution of the silica dehydration uid source rate in the sediment column along seismic pro le 1. . . . . 143
Appendix B.9 Thermal structure evolution of the incoming sediment column across seismic pro le 1, and corresponding modelled sea oor heat ux when +15% seismic velocities are applied in depth conversion. . . . 144
Appendix B.10 Thermal structure evolution of the incoming sediment column across seismic pro le 1, and corresponding modelled sea oor heat ux when -15% seismic velocities are applied in depth conversion. . . . 145

Appendix B.11 Smectite dehydration state of the incoming sediment column across seismic pro le 1 after +15% seismic velocities are applied in depth conversion, and corresponding uid thickness generated based on the Pytte and Reynolds (1989) kinetic. . . . . 146
Appendix B.12 Smectite dehydration state of the incoming sediment column across seismic pro le 1 after -15% seismic velocities are applied in depth conversion, and corresponding uid thickness generated based on the Pytte and Reynolds (1989) kinetic. . . . . 147
Appendix B.13 Smectite dehydration state of the incoming sediment column across seismic pro le 1 after +15% seismic velocities are applied in depth conversion, and corresponding uid thickness generated based on the Huang et al. (1993) kinetic. . . . . 148
Appendix B.14 Smectite dehydration state of the incoming sediment column across seismic pro le 1 after -15% seismic velocities are applied in depth conversion, and corresponding uid thickness generated based on the Huang et al. (1993) kinetic. . . . . 149
Appendix B.15 Silica dehydration state of the incoming sediment column across seismic pro le 1 after +15% seismic velocities are applied in depth conversion, and corresponding uid thickness generated. . . . . 150
Appendix B.16 Silica dehydration state of the incoming sediment column across seismic pro le 1 after -15% seismic velocities are applied in depth conversion, and corresponding uid thickness generated. . . . . 151
Appendix B.17 Cumulative total uid expelled from sediments underlying re ector R11 along seismic pro le 1 broken down into constituent sources when +15% seismic velocities are applied in depth conversion. . . . . 152
Appendix B.18 Cumulative total uid expelled from sediments underlying re ector R11 along seismic pro le 1 broken down into constituent sources when -15% seismic velocities are applied in depth conversion. . . . . 153

Appendix C.1 Bedding-perpendicular thickness of units R0–R3, R3–R4, R4–R5, and R5–R5B in depth converted transects of pro les A, B and C . . . . . 156

# Declaration

I, Duncan Eliott Stevens, declare that this thesis and the work presented in it are my own and has been generated by me as the result of my own original research.

Faults, Fluids and Geohazards at Subduction Zones

I con rm that:

- This work was done wholly or mainly while in candidature for a research degree at this University;
- Where any part of this thesis has previously been submitted for a degree or any other quali cation at this University or any other institution, this has been clearly stated;
- Where I have consulted the published work of others, this is always clearly attributed;
- Where I have quoted from the work of others, the source is always given. With the exception of such quotations, this thesis is entirely my own work;
- I have acknowledged all main sources of help;
- Where the thesis is based on work done by myself jointly with others, I have made clear exactly what was done by others and what I have contributed myself;
- Part of this work has been or will be published in peer reviewed scienti c journals:
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  - Stevens, Duncan E., Henstock, T. J., and McNeill, L. C. (2021). Evolution of the thermal and dehydration state of sediments entering the North Sumatra Subduction Zone. Geochemistry, Geophysics, Geosystems.

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# Chapter 1

# Introduction

Large magnitude earthquakes that occur in marine settings have proven to be one of the most damaging natural geohazards facing our civilisation. Where large magnitude earthquakes occur within marine settings they may generate powerful tsunamis with potentially devastating consequences for coastal communities. More than 3.5 billion people live along or within 150 km of a coastline, and the percentage of the global population living in coastal communities is ever growing. This growth is emphasised in the developing world where natural disasters in ict disproportionately higher levels of casualties and property loss (in Southeast Asia for example) (Dilek et al., 2021). The December 2004 Indian Ocean tsunami, generated by the $M_W$ 9.2 Aceh-Andaman (North Sumatra) earthquake (Shearer and Bürgmann, 2010), and the March 2011 Japan tsunami, generated by the $M_W$ 9.0 Tohoku-oki earthquake (Simons et al., 2011), claimed a combined total of over 270 000 lives, displaced many more and devastated the economies of the communities affected. The impact of such extremely damaging events extends beyond local humanitarian crises: in our ever more interconnected society the results of these disasters can lead to geopolitical consequences for the global economy, for example disruption to trade and human displacement (Dilek et al., 2021). These are issues that are likely to become more serious as coastal communities are increasingly affected by the results of climate change in terms of sea level rise. Sea level rise could compound the risks posed by tsunamis, by increasing the likelihood of tsunami ooding (Nagai et al., 2020). Therefore, and due to the recurrence of marine earthquakes, the motivation and impetus for improving our understanding of the mechanics of the environments in which they are generated has never been stronger.

Subduction zones are convergent boundaries formed where a cold, dense oceanic lithospheric plate is thrust below more buoyant lithospheric material, such as a younger, hotter oceanic plate or at continental margins (Figure 1.1). The downgoing plate is subducted into the mantle at rates that vary from 20-120 mm yr $^{1}$ (McCaffrey, 1994). Though there are mechanisms by which subduction is accommodated aseismically, when the subducting and overriding plates are strongly coupled together at the interface, the convergence of the two plates results in the accumulation of elastic strain (interseismic phase), and when coupling of the plate interface fails, the strain is released through seismic slip during an earthquake (co-seismic phase). The two plates are composed of different materials, where the subducting plate is being introduced continually into the system. The top of the subducting plate on one side of interface consists of ocean oor sediments. The degree to which the material being introduced to the subduction zone is consolidated and metamorphosed has a signi cant in uence on the frictional behaviour of the plate interface (Scholz, 2019). The interface between the two plates may be referred to as the `décollement', `megathrust' or `plate boundary fault'.

FIGURE 1.1: Schematic diagrams of the two end member subduction zone margin types: (a) accretionary and (b) erosive (after Clift and Vannucchi, 2004).

Understanding the mechanics of how the subduction process operates and its geohazard potential is an endeavour that requires an international, interdisciplinary effort utilising many different avenues of methodologic approach. For example: passive source/earthquake seismology to study the focal mechanisms and locations of active faulting in four dimensions over recent time periods (e.g., Tichelaar and Ruff, 1993; Du et al., 2004; Tilmann et al., 2010); geodetic analysis of plate motion using GPS measurements to determine convergence rates and obliquity, subduction velocities and the strength of plate interface coupling (over recent time periods) (e.g., Ito et al., 1999; Wallace et al., 2004; Simons et al., 2007, 2011); active source tomography (including recent advances in full seismic waveform inversion) to determine the physical properties structure of the subduction zone environment by analysing subsurface seismic velocities (e.g., Henrys et al., 2013, 2020; Gray et al., 2019; Arnulf et al., 2021); geological mapping of subaerial or exhumed accretionary complexes to examine ancient structures as analogues for modern subduction zones (e.g., Kondo et al., 2005; Rowe et al., 2005; Meneghini and Moore, 2007; Bender et al., 2018); experimental and analogue sandbox modelling to determine possible ways in which observed fault structure at accretionary margins is formed (e.g., Lallemand et al., 1992; Wang and Davis, 1996; Dominguez et al., 2000); and the use of multichannel seismic re ection data and high-resolution bathymetry, sometimes in combination with ocean drilling (for sediment coring and downhole geophysical logging) to map the lithological, physical properties and active fault structure of the upper crust at contemporary subduction zones (e.g., Gulick et al., 2004; Hutnak and Fisher, 2007; Moore et al., 2009; Saffer et al., 2012; Cook et al., 2014; Hupers et al., 2017). The latter approach is the one applied in this thesis.

## 1.1 Earthquake Mechanics and the Seismo- and Tsunami-genic Potential of Subduction Zones

Earthquake magnitude ( $M_W$) is proportional to seismic moment ( $M_0$) (Equations (1.1) and (1.2)), which is a linear measure of the amount of energy released. Fundamentally, seismic moment is de ned by the product of the shear modulus of the rupturing material, the amount of slip across the rupture plane and the rupture area (Equation (1.1)) (Scholz, 1998, 2019).

$$M_0 = GuA \tag{1.1}$$

Where G is the shear modulus of the material, u is the average slip across the fault plane and A is the rupture area.

$$M_W = \frac{2}{3}\log M_0 \quad 10.7 \tag{1.2}$$

It follows that the larger the area ruptured during an earthquake, the greater the amount of energy released, with the biggest earthquakes occurring where large amounts of slip takes place across very extensive rupture areas. Subduction zone margins are extensive and the décollement presents a potential rupture plane 100s to 1000s of km long and up to 200 km wide. For this reason, subduction zones account for the major proportion of the world's seismic moment release and are responsible for the largest magnitude earthquakes (Scholz, 2019). The potential destructive power of such earthquakes, and the associated tsunami risk, depends more precisely on the extents within which seismic slip can be nucleated and/or the extent to which it can be propagated, that is the width of the `seismogenic zone'. The wider the seismogenic zone, the greater the seismic moment potential (Hanks and Kanamori, 1979; Scholz, 1998). The farther this zone extends updip along the décollement to shallower depths within the crust and seaward to beneath deeper water, the larger the magnitude of the earthquake and the greater the potential for movement of the sea oor during an earthquake to trigger a large tsunami.

FIGURE 1.2: Slip on a fault caused by elastic rebound. The rock surrounding a fault is relaxed at the beginning of a seismic cycle, elastic strain is accumulated until the resultant shear stress exceeds the frictional strength of the fault causing seismic slip (after Fagereng and Toy, 2011)

Understanding why parts of a subduction interface may be seismogenic, relies on our understanding of the friction laws that govern the mechanics of earthquake generation. Earthquakes are the result of sudden movement (in the order of metres per second) along a fault plane (or along the plate boundary décollement in the case of megathrust subduction earthquakes), which occurs when elastic strain accumulated in the surrounding rock is released (Brace and Byerlee, 1966) (Figure 1.2). Earthquakes are thus a frictional phenomenon (rather than the result of brittle fracture), and the result of stick-slip frictional instability (Brace and Byerlee, 1966; Scholz, 1998, 2019). Rupture of the slip surface occurs when the ratio of shear stress `$\tau$' to effective normal stress (the difference between normal stress `$\sigma_n$' and fluid pressure `$P_f$') reaches the static coefficient of friction `$\mu_s$' (Scholz, 1998) in Amontons' law (Equation (1.3)):

$$\mu_s = \frac{\tau \quad C_0}{\sigma_n \quad P_f} \tag{1.3}$$

FIGURE 1.3: Schematic diagrams explaining the critical slip distance L. A) The contact between two sides of a fault plane (solid black lines) is made up of a number of asperities (indicated as contact junctions). When slip velocity V is rapidly increased during a stick-slip event, friction does not evolve to a new dynamic value until the contact junctions are renewed over the critical slip distance, which is equal to the mean diameter of the contact junctions (i.e. when slip (u) = L). B) The relationship between V and mfor velocity- weakening and strengthening materials, the steady-state relationship is represented by the dotted lines. The solid lines indicate the path of mduring a stick-slip event. (After Scholz, 2019)

Where $C_0$ is cohesive strength (Fagereng and Toy, 2011). However, this provides no indication of the rate at which slip will occur, which is the main determinant of seismicity. Experimental laboratory studies of arti cial faults in largely homogeneous materials have yielded a complete constitutive law for rate- and state- variable friction (Rabinowicz, 1958; Dieterich, 1972, 1979; Scholz, 1998). It has been shown that, when measured in the steady-state sliding regime, there is a relationship between the dynamic coef cient of friction ` $m_d$' and slip velocity ` V', which may either be positive or negative. In clearer terms, for any value of V there is a corresponding value of $m_d$; and as V increases, friction ($m$) will either increase or decrease, depending on material properties. When the slip velocity is suddenly (effectively instantaneously) increased, as is the case for earthquake rupture, the change in friction `Dm' does not occur in the same instantaneous manner. Instead, friction will retain the same value over a `critical slip distance' ' L'. Speci cally, when the time interval ` Dt' between slip velocities $V_0$ (initial/reference) and $V_1$ approaches zero (instantaneous slip), there must be a certain amount of movement on the fault surface before friction will evolve to a new value (the rate dependence) (Rabinowicz, 1958; Dieterich, 1972; Scholz, 2019).

Friction is dependent on the real area of contact between the two sides of a fault/sliding surface. The critical slip distance is the amount of slip required for contact renewal, i.e. the motion required to remove all junctions of contact that existed before slip, and to establish completely new junctions of contact (Scholz, 2019) (Figure 1.3). The amount of contact between each side of the sliding surface, and therefore the coef cient of (static) friction, changes with time, prior to seismogenic slip (Dieterich, 1972). During periods when no slip is occurring, the surface is subject to a constant normal load, over time this increases the area of real contact between each side of the surface, increasing static friction $m_s$ (the state-dependence) (Dieterich,

1979). Therefore, for longer `inter-seismic' periods a greater amount of shear stress is required for slip to occur (Equation (1.3)).

The rate- and state-dependencies of friction are related to each other as follows (in Equations (1.4) to (1.7)) (Brace and Byerlee, 1966; Scholz, 1998; Ohmura and Kawamura, 2007; Fagereng and Toy, 2011):

$$m = m_0 + a \ln \frac{V}{V_0} + b \ln \frac{qV_0}{L} \tag{1.4}$$

$$\frac{dq}{dt} = 1 \quad \frac{qV}{L} \tag{1.5}$$

$$t = \quad m_0 + ( a \quad b) \ln \frac{V}{V_0} \quad s_e \tag{1.6}$$

$$(a \quad b) = \frac{\P m_s}{\P[\ln V]} \tag{1.7}$$

Where $m_0$ is a reference friction, $V_0$ is a reference slip velocity and a and b are constants that relate to material properties and are empirically derived (Beeler et al., 1994). `q' is the state variable, which evolves with time (at steady state) as the area of real contact `$A_r$' on the surface increases, and is related to the critical slip distance in Equation (1.5) (Figure 1.4) (Scholz, 2019). Equation (1.4) is simpli ed to Equation (1.6) for steady state, where `$s_e$' is effective stress.

Frictional stability is controlled by the two friction rate parameters L and (a - b) (which is de ned in Equation (1.7)). If ( a - b) 0, m(friction) and t (shear stress) (in Equation (1.4) and Equation (1.6)) will be positive values and hence increase with increases in V, this is known as velocity-strengthening. Velocity weakening refers to cases where (a - b)< 0 since m and t are likely to be negative values (and more negative with increases in V). The resultant values are effectively the change that occurs during slip (Dieterich, 1972; Scholz, 1998). Stable sliding will always occur if (a - b) 0, but for the velocity-weakening regime friction can be either always unstable for large negative values of (a - b) or conditionally stable for small negative values (Scholz, 1998). In the conditionally stable eld, sliding is stable below a critical effective normal stress that is dependent on the friction rate parameters (a - b) and L (and the `stiffness' of the system `k') in Equations (1.8) and (1.9) (Rabinowicz, 1958; Dieterich, 1972):

$$s_c = \frac{kL}{(a \quad b)} \tag{1.8}$$

FIGURE 1.4: Plan view schematic of contact asperities on a fault plane demonstrating the control on the state variable q. The area of a fault plane is often assumed as equivalent to $A_r + A_p$, however friction relies on $A_r$ – the area of real contact. As the two surfaces either side of a fault plane are held in stationary contact under constant normal stress s, $A_r$ increases. This has an effect on the critical slip distance in Figure 1.3 and the static coef cient of friction $m_s$ (After Scholz, 2019).

$$z = (a \quad b)s_e \tag{1.9}$$

Where $s_c$ is a critical value of effective normal stress at which the bifurcation between unstable and conditionally stable regimes occurs. Suf cient dynamic loading or a suf ciently large jump in slip velocity may cause a fault in the conditionally stable eld to become unstable. Earthquakes can only nucleate within regions of a fault/rupture plane which are frictionally unstable, but can propagate into conditionally stable regions if the dynamic stresses continue to produce a large enough velocity jump during rupture (Scholz, 1998). From Equation (1.6), a simplistic frictional stability can be de ned as z (Equation (1.9)) (Scholz, 1998). Based on Equation (1.9), the three stability zones can be de ned as where z 0 – intrinsically stable, k < z < 0 – conditionally stable and z < k – intrinsically unstable.

In a geological sense, parameters (a - b) and L contain information about the mineralogy, uid pressure, dominant deformation mechanism and rock structure, and pressure-temperature environment surrounding a fault. Since the constants a and b are empirically derived from laboratory experiments on individual materials the effects of geological conditions upon them are still poorly understood (Fagereng and Toy, 2011). However, the velocity-strengthening and weakening regimes de ne which portions of a fault will exhibit aseismic creep and seismic slip, respectively.

For subduction zones, we must endeavour to understand how geological conditions affect the limits to the potential extents of seismic slip in order to evaluate the risks. Limits to the extent of the seismogenic zone are often regarded to be controlled by the ambient pressure and temperature at the décollement (e.g. Hyndman et al., 1997). The

deep, downdip limit is generally accepted to be the high temperature (350-450°C) transition in the style of quartzo-feldspathic rock deformation from `brittle' to `ductile' (Byrne et al., 1988; Hyndman et al., 1997), beyond which aseismic plastic ow accommodates shearing forces (Marone and Scholz, 1988). This brittle-ductile transition typically occurs along the subduction zone megathrust at around 40 km depth (Tichelaar and Ruff, 1993).

The updip limit of seismogenesis can be more important than the downdip limit from a geohazards perspective. It is apparently more complex in that it relies on the interplay between a multitude of interconnected geological factors, but due to its shallower depth within the crust it is more accessible for investigation using current technology. Traditionally, the shallowest part of the d écollement at subduction zones had been thought to behave aseismically (Byrne et al., 1988; Wang et al., 2010), because of a view that the cold ( < 150°C), wet sediments which comprise the environment of the shallow part of the subduction zone are cohesively weak and inhibit the propagation of seismic slip during an earthquake (Scholz, 1998). However, this assumption has been challenged by the recent north Sumatra 2004 (and following events between 2005 and 2007) and Japan 2011 earthquakes, where co-seismic slip propagated to, or nucleated within, unexpectedly shallow portions of the d écollement, potentially reaching the shallowest d écollement at the subduction trench (Simons et al., 2011; Lay, 2012). Furthermore, these two earthquakes occurred at subduction zones that represent end member types in terms of sediment thickness on the incoming plate and with big differences in the age of the subducting plate. The Japan margin is a partly `erosional' type with limited accretion, with thin sediment and an old oceanic plate, and the north Sumatran margin is an `accretionary' type with thick sediment and a relatively young oceanic plate (e.g., von Huene and Scholl, 1991; Clift and Vannucchi, 2004). This suggests that the mechanisms controlling shallow rupture in the two cases may be different.

## 1.2 Accretionary Margins

At accretionary margins, relatively thick sediment on the incoming plate ( > 1 km) and/or slow convergence ( < 76 mm yr $^{1}$) (Clift and Vannucchi, 2004) (Figure 1.1), results in the build-up of an accretionary wedge (or prism) comprised of sediments scraped off the subducting plate. In contrast, erosional margins occur where sediment cover on the incoming oceanic plate is thin ( < 1 km), convergence rates are relatively high ( > 60 mm yr $^{1}$), most of the sediment is subducted, and the base of the overriding plate is partially eroded and entrained by the subducting plate. Many margins include sections of accretion and erosion and these modes change with time. Though large magnitude earthquakes, such as the $M_W$ 9.0 Tohoku-oki event, have been recorded at erosional margins, most instrumentally-recorded large magnitude

earthquakes ($M_W$ 7.5) have occurred at accretionary margins (Ruff, 1989; Heuret et al., 2012; Scholl et al., 2015).

On a large scale, the structure of both margin types is similar (Figure 1.1), but at accretionary margins it typically consists of 6 main components (Figure 1.5): (1) an outer trench rise (or `swell', or `high'), the point on the downgoing plate where it begins to ex in response to subduction; (2) the trench, where exure of the downgoing plate creates a depression (in lled with sediment); (3) the deformation front marking the seawardmost extent of the d écollement (usually in the subsurface), the onset of the accretionary wedge and the outermost thrust fault (frontal thrust) that propagates up from the d écollement; (4) the accretionary prism, a wedge of deformed and deforming oceanic and terrigenous sediments, that have been incorporated onto the overriding plate by thrust faults propagating up from the d écollement; (5) the forearc basin, where axial sediments, or those derived from the landward edge of the subduction zone are deposited in the depression formed between the accretionary prism and the massif of the overriding plate; and (6) a volcanic arc forming the massif of the overriding plate, generated by magmatic activity in the wedge of mantle material between the top of the subducted portion of the downgoing plate and the base of the overriding plate (Karig, 1974; Stern, 2002). In addition, some subduction zones will have a back arc basin formed behind the volcanic arc, potentially by rollback of the subducting slab.

Geohazards, volcanic or seismic, may occur throughout an entire subduction zone. However, the outer part of the subduction zone presents the most potent tsunami risk. Slip on the décollement leads to movement of the sea oor – this is the primary cause of tsunami generation (e.g. Rosenau et al., 2010). In addition, slip on thrust faults within the accretionary prism may also slip and generate additional movement of the sea oor at seismic rates if rupture of the d écollement extends to their location during an earthquake (Oleskevich et al., 1999; Gulick et al., 2011). In our attempts to understand the hazard potential at subduction zones, we must carefully examine the properties of the incoming plate sediments that may in uence the updip limit of décollement seismogenesis, and the fault structure of the accretionary prism. These aspects are the focus of this thesis.

## 1.3 Deformation, Composition and Seismicity of the Incoming Plate

One of the factors that is often overlooked when examining the subduction process is the tectonic evolution of the incoming plate. Deformation of the incoming plate due to tectonic forces, potentially interlinked to pull of the slab that has been subducted, alters the structural geology and hydrology through the material that is being

FIGURE 1.5: Sketch showing the main components of the marine forearc at accretionary margins, as well as the locations of key processes. (1) the outer rise; (2) the trench; (3) the deformation front; (4) the accretionary prism; (5) the forearc basin; (the volcanic arc (6) occurs in the terrestrial realm and is not pictured). (After Kopp, 2013).

introduced to the subduction zone. The nature and timing of such deformation has important implications for the development of the accretionary wedge and the physical properties of the materials from which it is made. Speci cally, fault structure that pre-exists within the incoming sediments may present preferential lines of weakness that can be exploited or reactivated as thrust faults propagating from the décollement; faulting of the incoming plate may affect permeability and uid ow through the incoming sediment column (e.g. Geersen et al., 2013), in uencing the uid pressures and physical stresses experienced by the subducting/accreting materials; and the timing of the incoming plate deformation may correlate to varying structure within the accretionary prism.

Faulting of oceanic plates via normal and strike-slip mechanics (including transforms and fracture zones) at their point of origin along mid ocean ridges is a well-understood key process affecting the structure of the plate. Deformation within oceanic plates generated away from the mid ocean ridge, however, is still poorly understood. Such `intraplate' deformation does not occur everywhere and the forces that drive it remain unclear, but undoubtedly it has important implications for the mechanics of subduction where it does occur. It can also be a signi cant source of seismicity (e.g. Deplus, 2001) and therefore important from a geohazards perspective in its own right. Oceanic intraplate seismicity has indicated that large magnitude

marine earthquakes are not limited to the forearc of subduction zones. Offshore the Sunda subduction zone, the largest magnitude ($M_W$ 8.6 and $M_W$ 8.2), and most complex, strike-slip earthquakes ever recorded occurred within the incoming plate in the Wharton Basin (Duputel et al., 2012; Wei et al., 2013; Hill et al., 2015). Understanding the kinematics of what controls this style of deformation and how it manifests in terms of the structure of the oceanic plate as it approaches subduction zones, and how it may impact on the geological structure of the upper plate, is key to our ability to assess the risks posed and the interaction with the subduction process.

The subduction process is also affected by the properties of the sediments on the incoming oceanic plate. During deposition and subsequent burial, the geothermal gradient raises the temperature of sediments altering their properties through thermally-controlled diagenetic and low-grade metamorphic reactions. These alterations also change the frictional properties of the sediments, potentially changing the lithological environment from velocity-strengthening to velocity weakening (e.g. Vrolijk, 1990). Reactions that may occur during the burial history of the incoming plate sediments include the smectite to illite transition (e.g. Saffer et al., 2008); the opal to quartz transition (Spinelli and Underwood, 2004); carbonate and zeolite cementation; and hydrocarbon maturation (Moore and Saffer, 2001). These processes not only alter the intrinsic frictional strength of sediment grains, but consolidate the bulk sediment and expel mineral-bound fluids, increasing the cohesive strength of the bulk sediment and its ability to accumulate elastic strain. The temperature interval over which the sediment altering reactions are estimated to take place is between 50 and 175°C (e.g., Ernst, 1990; Mizutani, 1970; Pytte and Reynolds, 1989).

Sediment temperatures depend on their burial history, generally with deeper burial leading to higher temperatures. In most subduction zones, the critical temperatures required for the advanced stages of diagenesis are only reached some distance landward of the deformation front, below the forearc. However, where the sediment column is sufficiently thick on the incoming plate, basal sediments have been proposed to reach the temperatures of diagenetic and low-grade-metamorphic reactions, before sediments enter the subduction zone (e.g., Spinelli and Underwood, 2004; Ike et al., 2008; Smith et al., 2013; Hüpers et al., 2017). This has led to the hypothesis that thick input sediments can bring the change from velocity-strengthening (stable-sliding) to velocity-weakening (seismic slip) (the updip limit of seismogenesis) further updip, by drawing the onset of mineral transformations and associated cementation seaward and increasing the potential seismogenic width and thus earthquake magnitude (e.g., Moore and Saffer, 2001; Saffer et al., 2008; Dean et al., 2010; Smith et al., 2012).

In addition, dehydration of sediments through total completion of the diagenetic reactions outlined above may release volumes of fluid up to 29% of the original mineral volume for opal and 40% for smectite (Hüpers et al., 2017). When and where

this uid is released during the subduction history of the incoming sediments has implications for seismic behaviour. In subduction zones where dehydration of the sediments occurs well after they are accreted or subducted, the escaped uids may be focussed along faults acting as conduits (Saffer and Bekins, 2002), or may be trapped within the accretionary wedge close to the level of the d écollement, generating high pore- uid pressures (Moore and Saffer, 2001). Such high pore uid pressure reduces effective stress, which reduces the magnitude of the frictional stability parameter z in Equation (1.8) (Scholz, 2019). Conversely, if the sediments around the décollement level undergo dehydration prior to their subduction, and the expelled uid is ef ciently drained, then uid pressures along the d écollement will be relatively reduced and effective normal stress increased. Where the frictional stability parameter z is already negative, early (pre-subduction) dehydration would act to increase frictional instability along shallow parts of the d écollement, drawing the limit of potential earthquake generation updip to shallower depths (Moore and Saffer, 2001). Therefore, evaluating the thermal evolution and progress of sediment dehydration at accretionary subduction zones with thick incoming sediments is crucial in determining the updip limit of seismogenesis.

## 1.4 Along-strike variation of outer subduction zone structure

In our considerations of the potential magnitude of megathrust earthquakes at subduction zones, seismogenic width up- and down-dip along the d écollement is one factor, and the distance rupture can propagate along the length of a margin is another. Megathrust earthquakes do not often rupture the d écollement along the entire length of a margin, but instead, rupture is usually limited to particular segments (e.g., Tilmann et al., 2010; Meltzner et al., 2012, Figure 1.6). Differences in the structural characteristics of the forearc/accretionary prism along margin strike may be indicative of segments of a margin that will exhibit different seismic behaviour or have different seismogenic and tsunamigenic potentials (Kopp, 2013). Comparisons of observations of along-strike structural heterogeneity within the forearc and accretionary prism with changes in the patterns of seismicity, suggest that changes in structure and termination of rupture propagation during a megathrust earthquake may have a common cause (e.g., Sparkes et al., 2010; Cook et al., 2014). The answer to what controls these structural differences has been investigated on many margins but conclusive answers remain elusive, and addressing the problem is complicated by the timescales over which the subduction process operates. However, we can make correlations between observed structural variation and along-strike changes in tectonic dynamics and con guration of the incoming plate. Some of the factors which may be in play include, for example: topography of the incoming oceanic basement basement (e.g., McNeill and Henstock, 2014; Pedley et al., 2010); thickness of the

FIGURE 1.6: Rupture segments from historical megathrust earthquakes along the Sunda subduction zone. Toothed black line indicates the Sunda subduction front; coloured areas indicate rupture areas; black lines are faults; gray lines are fracture zones; blue arrows show relative plate motion (after Meltzner et al., 2012).

incoming sediment (e.g., Cook et al., 2014; Heuret et al., 2012; von Huene and Scholl, 1991; McNeill and Henstock, 2014); temporal variations in how much sediment is either accreted, subducted or underplated (Clift and Vannucchi, 2004); the degree of plate convergence obliquity and the rate of orthogonal convergence (e.g., Clift and Vannucchi, 2004; McNeill and Henstock, 2014); and fault structure of the incoming plate (e.g., Contreras-Reyes et al., 2008, this study).

These various elements are interlinked, for example, sediment thickness on the incoming plate is largely controlled by the topography of the top of the underlying

oceanic basement (Westbrook et al., 1982) and in turn sediment thickness and/or physical properties may exert control on the stratigraphic position of the d écollement, which controls mass transfer (accretion vs subduction) (e.g., Shipley and Moore, 1986; Moore et al., 1988). The interconnectedness of these elements highlights the complexity and multidisciplinary nature of the problem of trying to unravel the rst order controls on shallow subduction zone mechanics.

## 1.5 Research Aims

The primary goal of this thesis is to understand how the structure of the upper plate at subduction zones, and their hazard potential, is affected by processes within the incoming oceanic plate and overlying sediments. There are three main areas of research focus: (1) fault deformation within the incoming plate; (2) the thermal evolution of the sediments deposited on the incoming plate; and (3) spatial variation in the fault structure and morphology of the accretionary prism and how it relates to differences in properties of the incoming plate along-margin-strike.

The rst two aspects will focus on the incoming plate offshore of the North Sumatra subduction zone. Shortly before the commencement of this project, International Ocean Discovery Program (IODP) Expedition 362 ocean drilling sampled the entire sediment column on the oceanic plate in the Wharton Basin region offshore of the 2004 Aceh-Andaman megathrust earthquake (McNeill et al., 2017a). The data collected by that expedition has provided an opportunity, not only to investigate the properties of the subduction zone's incoming sediments and their relationship to the 2004 rupture, but also to robustly analyse fault deformation, which is pervasive across the oceanic plate and likely related to the 2012 intraplate earthquakes and the wider diffuse deformation zone hosted by the Indian Ocean. By combining the drilling results with multichannel seismic re ection data and swath bathymetry I aim to:

- Describe the styles of intraplate faulting occurring within the Wharton Basin (Chapter 2)
- Assess the timing of fault activity occurring within the Wharton Basin (Chapter 2)
- Determine if fault activity in the Wharton Basin can be linked to intraplate fault activity in the wider Indian Ocean (Chapter 2)
- Assess the potential forcing mechanisms that may be generating intraplate deformation in the Wharton Basin (Chapter 2)

- Determine the relationship between the observed fault styles in the Wharton Basin and the rupture pattern observed for the 2012 intraplate earthquakes (Chapter 2)
- Assess the links between faulting of the incoming sediments and their physical properties (Chapters 2 and 3)
- Determine how the deposition of a thick incoming sediment column affects the temperatures of those sediments, and speci cally of the sediments at the locus of décollement initiation at the North Sumatra subduction zone deformation front (Chapter 3)
- Evaluate the dehydration state of the incoming sediments offshore North Sumatra based on their temperature evolution (Chapter 3)
- Assess how sediment dehydration affects the hydrology of the incoming plate offshore North Sumatra, what control it exerts on the stratigraphic position of the décollement and the implications for seismogenesis (Chapter 3)

The third aspect, on the relationship of the incoming plate to accretionary prism structure, will focus on the southern Hikurangi subduction zone margin, New Zealand. Due to tectonic rotation of New Zealand's North Island, the southern Hikurangi margin hosts an along-strike transition from approximately orthogonal plate convergence to strike slip tectonics (Wallace et al., 2012a). This results, within a relatively small distance, a signi cant increase in obliquity, and decrease in subduction velocity from east to west, which is accompanied by increasing sediment thickness on the incoming plate and changes in the structural con guration of the incoming plate. Because of these characteristics, the southern Hikurangi margin is a prime setting for examination of the structural response of the accretionary prism. During the early stages of this project new, deep penetrating, multichannel seismic re ection data were acquired along- and across-strike of the central and northern Hikurangi margin. I will utilise some of this new data, which images the incoming plate and accretionary prism of the south-central margin (eastern-southern margin), with pre-existing data acquired by industry that images the central and western parts of the southern Hikurangi margin. By combining these data with high-resolution bathymetry, and information on the con guration of the subduction zone from literature I aim to:

- Evaluate the geometry, spacing and along-strike continuity of thrust faulting within the accretionary prism at the southern Hikurangi margin (Chapter 4)
- Describe the morphology of the southern Hikurangi accretionary prism (Chapter 4)

- Describe the activity history of thrust faulting within the outer parts of the prism at the southern Hikurangi margin, and its contribution to crustal shortening (Chapter 4)
- Determine the stratigraphic level at which the subduction zone d écollement is initiated along the southern Hikurangi margin (Chapter 4)
- Assess possible linkages between southern Hikurangi outer prism structure and incoming plate characteristics (Chapter 4)
- Compare the links between outer prism structure and incoming plate characteristics at southern Hikurangi with observations from other subduction zone margins (Chapter 4)

## 1.6 Thesis Overview

- Chapter 1 presents an introduction to the background of subduction zones, including their structure, properties of the incoming plate and the nature and controls of natural hazards, namely earthquakes and tsunami. This chapter also provides an outline of the thesis and its aims.
- Chapter 2 presents a novel analysis using seismic re ection pro les and swath bathymetry of the structural styles of faulting in the Wharton basin offshore the north Sumatran segment of the Sunda subduction zone. This area is the locus of the large magnitude intraplate marine strike-slip earthquakes, and is part of a wider diffuse deformation zone that stretches across the eastern Indian Ocean, possibly forming a nascent plate boundary. I evaluate how fault activity observed in the upper plate of the Wharton basin, which is incoming to subduction zone, relates to the kinematics of the intraplate earthquakes and the wider deformation zone. The fault structure of the oceanic plate is relevant to evolution of the incoming sedimentary materials which then affect forearc structure and morphology and d écollement properties.
- Chapter 3 presents an analysis using seismic re ection pro les and IODP drilling data of how an exceptionally thick sediment column on the incoming plate offshore the north Sumatra subduction zone affects the thermal evolution of those sediments. From this analysis the progress of thermally controlled diagenetic dehydration reactions can be estimated. Using such estimates, this chapter aims to determine the interaction between sediment dehydration and the stratigraphic position at which the contemporary d écollement is forming, and the properties of the d écollement.
- Chapter 4 presents an analysis of variation in the structure of the accretionary prism at the southern Hikurangi subduction zone margin. Rapid along-strike

changes in various aspects of the con guration of the subduction zone (e.g., obliquity, sediment thickness, incoming basement topography) occur along this part of the Hikurangi margin, and correlations between these aspects and structural changes can be attempted. This chapter aims to describe the structural variations and determine what correlations can be made, and assess how the results compare with other `similar' subduction zone margins (speci cally the Sumatra-Andaman).

- Chapter 5 summarises the scienti c  ndings of Chapters 2-4, discusses the links between them and their various implications for structural controls and geohazards at subduction zones. The potential for future avenues of research within the focus of the topics outlined in this thesis are also discussed.

# Chapter 2

# A Complete Structural Model and Kinematic History for Distributed Deformation in the Wharton Basin

This chapter forms a paper that has been published in Earth and Planetary Science Lettersas a research article:

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J. M. Bull: Supervision

## Abstract

The equatorial eastern Indian Ocean hosts a diffuse plate boundary, where widespread deformation accommodates the relative motion between the Indian, Australian and Capricorn sub-plates. We integrate IODP Expedition 362 borehole data, which for the rst time provides an accurate, ground-truthed chronostratigraphy of the sedimentary sequence east of the Ninety East Ridge (NER), with 2D seismic re ection pro les and multibeam bathymetry to assess the styles of faulting between the NER and the Sunda subduction zone, timing of activity and comparison with physical and rheological properties. We identify four distinct fault sets east of the NER in the northern Wharton Basin. N-S (350-010°) orientated faults, associated with the N-S fracture zones formed at the now extinct Wharton spreading centre, are still active and have been continuously active since at least 10 Ma. NNE- and WNW-trending fault fabrics develop between the fracture zones. The orientations and likely sense of displacement on these three sets of faults de nes a Riedel shear system responding to NNE-SSW left-lateral strike-slip activity at depth, demonstrated by the recent 2012 great intraplate earthquakes. We also nd evidence of NE-SW reverse faults, similar in style to E-W reverse faults observed west of the NER, where reverse faulting is more dominant. We show that the activity of this strike-slip system increased ca. 7-9 Ma, contemporaneous with reverse faulting and intraplate deformation west of the NER.

## 2.1 Introduction

The oceanic lithosphere of the equatorial Indian Ocean hosts a 3000 km wide (West to East) zone of complex deformation representing a diffuse plate boundary between the Indian, Capricorn and Australian plates (DeMets et al., 1990; Stein and Okal, 1978; Sykes, 1970; Wiens et al., 1986). The relative motion between the three plates, and the transfer of stress from plate boundaries surrounding the Indian Ocean, results in a variation in sense and style of deformation across the diffuse plate boundary (Figure 2.1) (DeMets et al., 1990; Gordon et al., 1998; Wiens et al., 1986). The Ninetyeast Ridge (NER), a north-south aseismic ridge, that stretches from 34°S to 10°N ( 90°E), seems to play a role in the west to east variation in deformation. West of the NER, deformation is characterised by crustal shortening (Bergman and Solomon, 1985; Bull, 1990; Stein and Okal, 1978; Wiens et al., 1986), whereas in the equatorial Indian Ocean, at the northern end of the NER and east of the NER (the northern Wharton Basin, Figure 2.1), primarily left-lateral strike-slip motion dominates (Deplus, 2001; Deplus et al., 1998; Sager et al., 2013). The change in deformation is likely due to the increased proximity and in uence of the Sunda subduction zone.

FIGURE 2.1: A) Map showing regional tectonic con guration. Red dashed line indicates extent of diffuse deformation (Gordon et al., 1998). Red toothed lines show schematically the area affected by prominent reverse faults (Bull and Scrutton, 1992; Deplus, 2001). Earthquake focal mechanisms within > 2 degrees of narrow plate boundaries, scaled based on magnitude: brown = strike-slip; orange = thrust; blue-grey = normal (Gordon and Houseman (2015), events 5.3 < $M_W$ < 8.6, 1976-2014). Light blue box shows enlarged in part B (and approximate area shown in Figure 2.10). NER – Ninety East Ridge; CR – Carlsberg Ridge; CIR – Central Indian Ridge; SWIR – South West Indian Ridge; RTJ – Rodriguez Triple Junction; SEIR – South East Indian Ridge; CAP – Capricorn plate. Bathymetry contoured at 2000 m intervals. B). Yellow dots indicate IODP drill sites. Black lines are multichannel seismic pro les A-C, sections highlighted in yellow indicate the locations of seismic data depicted in Figures 2.5, 2.6 and 2.8. Light blue shaded area shows multibeam bathymetry coverage used in this study (presented in Figure 2.4) Red, dashed lines are fracture zones (Jacob et al., 2014) (nomenclature from Singh et al. (2011)). Focal mechanisms for April 2012 events in red, data from Global centroid moment tensor project [www.globalcmt .org]. Orange shaded area shows rupture plane model for 2012 earthquakes from Wei et al. (2013). SSZ – Sumatra Subduction Zone.

The northern Wharton basin (north of 10°S) includes sediments of the Nicobar Fan, part of the Bengal-Nicobar Fan system (McNeill et al., 2017c). The complete sedimentary section was sampled by International Ocean Discovery Program (IODP) Expedition 362 to basement at 1415 m below sea oor (mbsf) (McNeill et al., 2017b). Fan sediments are dominated by siliclastic sediment gravity- ow deposits (e.g.,

turbidity currents and debris ows), that range from clay to silty clay to ne-grained sand. These are underlain by pelagic and tuffaceous sediments overlying ocean basement (McNeill et al., 2017b). Sediments are mostly unlithi ed, with lithi ed sediments only encountered in the deepest intervals.

Compressional deformation west of the NER, between 6°N and 8°S, is well documented (e.g. Bergman and Solomon, 1985; Bull et al., 2010; Bull and Scrutton, 1990, 1992; Chamot-Rooke, 1993; Krishna et al., 2001, 2009; Stein et al., 1989; Stein and Weissel, 1990; Stein and Okal, 1978) by the integration of seismic re ection data with information from ocean drilling sites (ODP Leg 116, DSDP 215, 218). Deformation takes place along E-W trending reverse faults, i.e. N-S compression, interpreted as reactivating the original ridge-parallel spreading fabric due to stress transfer from the N-S continental collision between India and Eurasia (e.g. Bull and Scrutton, 1992, 1990; Chamot-Rooke, 1993). Fault reactivation here started at 14-15.5 Ma (Krishna et al., 2009), with suggestions of increased activity at 7-8, 4-5 and 0.8 Ma (e.g. Bull et al., 2010; Krishna et al., 2001). Bull et al. (2010) showed that increased compressional deformation at 7-8 Ma (e.g. Krishna et al., 2009) coincides with the acceleration of Indian-Capricorn rotation and relative convergence from plate motion models (DeMets et al., 2010, 2005). Many have suggested that the timing of pulses in distal fault activity are related to uplift of the Himalayas and Tibetan Plateau (Bull et al., 2010; Gordon, 2009; Merkouriev and DeMets, 2006; Molnar and Stock, 2009), buckling the Indo-Australian lithosphere (Copley et al., 2010; Weissel et al., 1980). However, studies of the Himalayas and Tibet have generated a range of timings for activity and accelerated deformation over the last 20 Ma (e.g. Clift et al., 2008; Molnar et al., 1993; Molnar and Stock, 2009), therefore correlating discrete collisional/uplift episodes with phases of distal compressional deformation is challenging. Alternatively, Iaffaldano et al. (2018) have suggested that an acceleration in compressional deformation at 7-8 Ma within the Indo-Australian plate may be linked to increased asthenospheric ow resulting from the re-emergence of volcanism along the Rodrigues Ridge at 11 Ma.

Until now the timing of intraplate deformation in the Indo-Australian plate has not been well-constrained east of the NER. There are fewer studies of deformation (e.g. Deplus et al., 1998; Geersen et al., 2015; Singh et al., 2017; Hananto et al., 2018), and until recently there has been no direct age control. Seismicity is generally characterised by left-lateral strike-slip motion on reactivated N-S trending fracture zones that offset the E-W trending Wharton Ridge, a fossil spreading centre (Deplus, 2001; Deplus et al., 1998). Spreading at the Wharton Ridge was active at least as early as 84 Ma, separating India from Australia, but ceased around 42 Ma (along with related transform fault/fracture zone activity), with the Indian and Australian plates

becoming a single plate (Liu et al., 1983). The same tectonic driving mechanism responsible for compressional deformation west of the NER may also have reactivated the Wharton Ridge fracture zones but the reactivation timing of the latter is unclear. The difference in deformation style on either side of the NER may be due to variation in the stress eld within the Indo-Australian plate with the Sunda subduction zone causing the principal stress in the Wharton basin to be NNW-SSE, compared to N-S west of the NER (Gordon and Houseman, 2015).

In April 2012, the largest and most complex strike-slip earthquakes ever recorded ruptured a set of WNW-ESE and NNE-SSW fault planes oblique to the N-S fracture zones. The initial mainshock ruptured along a WNW trending fault with a centroid depth of 30 km with a $M_W$ of 8.6, that initiated multiple ruptures to the north and south on NNE trending faults at similar depths, this was then followed by another $M_W$ 8.2 earthquake that ruptured a WNW trending fault 180 km to the south (Fig.1; e.g. Duputel et al., 2012; Hill et al., 2015; Wei et al., 2013). Recent studies have identi ed faults/lineations from bathymetry data with similar orientation to faults modelled to have ruptured (e.g. Geersen et al., 2015; Hananto et al., 2018; Singh et al., 2017). The various orientations observed have been interpreted as a Riedel shear system related to the reactivated N-S fracture zones (Geersen et al., 2015; Hananto et al., 2018). Geersen et al. (2015) suggested these Riedel shear fabrics developed around 20 Ma, with fracture zones reactivated since 40 Ma. However, these timings were based on pre-drilling chronostratigraphic estimates.

Data from IODP Expedition 362 boreholes provides the rst ever core-stratigraphic ages for the sedimentary sequence east of the NER (McNeill et al., 2017b). These data indicate that the chronostratigraphy of the sedimentary sequence is signi cantly different to that previously assumed. Based on this new information, we update here the activity timings of all types of fault in the northern Wharton Basin, east of the NER. In addition, we have conducted a thorough integration between the seismic and bathymetry data in the study area, to derive fault orientations and fault types. Together, these new integrated data enable us to perform a more complete fault analysis. We also directly compare our results, east of the NER, with previous studies west of the NER, for the rst time with chronostratigraphy in both locations, to correlate and examine existing discrepancies between deformation events and fault activity timing across the eastern Indian Ocean and discuss potential deformation forcing mechanisms. We also use the IODP results to test relationships between faulting and the lithological and physical properties of the oceanic plate sediments.

FIGURE 2.2: Regional seismic pro le B and position of IODP Site U1481 (vertical pink line). (A) Uninterpreted and (B) interpreted, with faults (black), horizons/re ectors (coloured) R1 to R14 and channels (red). Dashed blue boxes indicate the locations of expanded sections (Figure 2.6E (see below) and examples X1 and X2 (see Appendix A Figures A.2 and A.3). NC = Nicobar Channel; NER = Ninety East Ridge. Figure 2.9 shows along pro le analysis for this pro le.

## 2.2 Methods

### 2.2.1 MCS data and interpretation

We have re-interpreted multichannel seismic re ection (MCS) data from the northern Wharton Basin (e.g. Figure 2.2) (building on Dean et al., 2010; Geersen et al., 2015, 2013; McNeill et al., 2017c). The seismic data are two composite SW-NE pro les A(BGR06 101-102) and B (BGR06 103-104-105) from Gaedicke et al. (2006, Cruise SO186 on FS Sonne) which extend from the NER to the Sunda subduction zone at North Sumatra (Figure 2.1), and NW-SE pro le C (MD116-ANDAMAN84) from Chamot-Rooke (2000, Cruise MD116 on the R/V Marion Dufresne). The seismic data have a vertical resolution of 10-15 m. We interpret all faults with maximum vertical offsets > 10 ms of two-way-travel-time (TWT) (approx. 10 m), including blind faults (some obvious blind faults were interpreted that have < 10 ms vertical separation). We interpret and use roughly evenly spaced seismic horizons R1-R14 that can be

FIGURE 2.3: Interpolation of re ector ages from core-seismic integration. Coloured crosses are seismic re ectors (Figure 2.2). The age of re ectors is estimated by time-depth conversion and linear sedimentation accumulation rates between biostratigraphic tie points (McNeill et al., 2017b,c). Errors of re ector ages are 5% due to velocity uncertainty and the age ranges of tie points.

continuously correlated. Some of these re ectors are equivalent to those of Geersen et al. (2015), while R11 corresponds to the `High-Amplitude-Negative-Polarity' (HANP) pre-d écollement re ector of Dean et al. (2010). We locally interpret additional re ectors to allow detailed analysis of selected faults.

### 2.2.2 Integration with IODP Expedition 362 borehole data

All three seismic pro les intersect IODP Expedition 362 borehole sites U1480 and U1481 where sonic log and core-log seismic integration (McNeill et al., 2017c) allow us to determine the depth intervals of re ectors R3 to R12, and R14 and correlate borehole data to the seismic stratigraphy. Age control is based primarily on calcareous nannofossils, and we note that the age markers are approximate (see Figure. 2.3; McNeill et al., 2017b,c). The section is dominated by Nicobar Fan sedimentation since 9-10 Ma in contrast to the older pre-drilling predicted ages used by Geersen et al. (2015). We also compare fault displacement with other parameters in the IODP boreholes, including lithology, physical properties (e.g., porosity, velocity), and fracture intensity.

### 2.2.3 Multibeam processing and interpretation

We reprocessed a subset of the multibeam bathymetry data from cruise SO186 collected using the SIMRAD EM120 Multibeam System. We applied an updated and improved sound velocity pro le correction, a ship roll correction and removed spikes from individual pings, to improve the signal to noise ratio of the true sea oor topography. We then re-interpreted this data and re-interpreted multibeam data from cruise MD116.

### 2.2.4 Vertical offset measurements

We measured the vertical offset/throws (hereafter referred to as displacements) for re ectors R2 to R12 across all interpreted faults on the NE-SW seismic pro les. We developed a semi-automated method which relies on the precise and consistent interpretation of the re ectors across each fault to calculate offset. Seismic horizons were picked in common-depth-point (CDP) – two-way-time (TWT) (ms) space. CDP spacing is 6.25 m. Seismic horizons were picked so horizons terminate at the fault-re ector intersection, leaving a horizontal gap across each normal-offset fault. The vertical gap is the fault throw in TWT. Where fault drag is apparent, horizons were picked so that the maximum vertical offset of the re ector across a fault is maintained (i.e. extrapolation from either side of a fault unaffected by drag). We are therefore con dent that our interpretation represents true vertical separation. The seismic horizons provide a database of TWT picks at every CDP where picks were made.

We convert the fault displacement TWT measurements to metres using seismic interval velocities picked at  1000 CDP spacing from the original processing of the MCS re ection data by BGR. This is an additional improvement on Geersen et al. (2015), which used a single set of estimated interval velocities throughout the seismic lines.

Around faults of particular interest, we increase the number of interpreted horizons and grid all of the observed vertical separations as a heat map (see Figure 2.6 C and G).

## 2.3 Results

### 2.3.1 Biostratigraphy (from IODP)

Biostratigraphic tie points from the IODP (Figure 2.3) borehole data show two distinct periods of sediment deposition in our study area occurring at different rates (McNeill et al., 2017b,c). The age of the oceanic basement is 68 Ma. Basal materials, dominantly pre-fan pelagic sediments, are 150 m thick at Site U1480 (Exp. 362 Units III-V) and accumulated at an average rate of 2.5 m Myr $^{1}$ over the 60 Myr period from 68 to 9 Ma (rate excluding hiatuses). The overlying Nicobar fan sediments (Exp. 362 Units I-II) are 1250 m thick at borehole U1480 and accumulated over 9 Myr at an average rate of 139 m Myr $^{1}$. Sediment thickness overlaying the oceanic basement increases to 4 km towards the subduction zone due to the in uence of plate exure and lling of the subduction zone trench (SU1, Figure 2.2). Most of our isochrones (laterally continuous seismic re ectors) are within the thick fan sediments.

### 2.3.2 Fault geometry

Seismic pro les (Figures 2.1 and 2.2) show pervasive fault deformation across an approximately 300 km wide area between the NER and the Sunda subduction zone in the northern part of the Wharton basin. The bathymetric data image the same structures in plan view and indicate a range of orientations, which vary across the oceanic plate. The large number of faults imaged on pro les A and B are roughly evenly distributed from west to east. The fault dip is between 60° and 75° with both landward and seaward dip directions (along-pro le). Pro le C images sub-vertical faults, which are more heterogeneously distributed from north to south.

We observe numerous lineaments in the multibeam bathymetry data (Figure 2.4). We are con dent that these lineaments are the surface expression of faults, since they can be tied to faults in the seismic data that displace the sea oor. Lineaments show three distinct orientations between the NER and the Sunda subduction zone that are geographically distinct from each other. We group these lineaments and associated faults into three Classes. Between the NER and 91° 45'E Class A faults are the dominant fabric, trending NNE-SSW (020-030°); between 91° 45'E and 92° 20'E Class B WNW-ESE (100-120°) lineaments dominate; and between 92° 20'E and 93° Class C N-S (350-010°) dominate. It is unclear whether the Class A and B fabrics overprint each other, but in the area of Class B faulting, very few NNE-SSW faults are observed in the bathymetry suggesting that there is a genuine change from one orientation to the other. In the area of Class C faulting, Class A and B fault

FIGURE 2.4: Distribution of sea oor faulting interpreted from multibeam bathymetry. A) uninterpreted slope map of multibeam data showing intraplate earthquake focal mechanisms (as shown in Figure 2.1). B) Interpreted lineations for each fault class (colour coded) and corresponding rose diagrams. Thick red dotted lines represent fracture zones (nomenclature from Singh et al., 2011). Light blue lines indicate seismic pro les and the positions of Figures 2.5, 2.6 and 2.8 are highlighted in pink.

FIGURE 2.5: Example Class A fault on NW-SE trending seismic pro le C, (A) uninterpreted and (B) interpreted. Note V-shape of re ectors at the position of the fault and apparent layer thickness changes across the fault. (C) displacement analysis of example fault. Cumulative line shows the total net vertical displacement accumulated by each horizon since deposition, back-stripped line indicates how vertical displacement has accumulated with time i.e. how much displacement occurred ca the time of deposition of each horizon. Vertical separation attributed positive values for normal and negative for reverse assuming that the footwall is to the SE.

orientations are effectively absent. The N-S trending faults (Class C) are in good agreement with both the position and orientation of previously mapped Wharton Ridge fossil fracture zones (e.g. Singh et al., 2011).

### 2.3.3 Class A faults

Class A faults are only observed on seismic pro le C in the west of our study area (Figure 2.1). They are sub-vertical, propagate through the entire sedimentary section and deform the oceanic basement. The sense of displacement is unclear, sedimentary layers frequently show a v-shaped pattern that is expressed at the sea oor (Figure 2.5A, B). We also note differences in sedimentary layer thickness across these faults and complex displacement patterns that indicate both apparent normal and reverse displacement (Figure 2.5C). These features are very similar to interpreted strike-slip faults at the NER (Sager et al., 2013) and strike-slip faults in general. The complex displacement pro les for these faults limits our ability to constrain the timing of their formation and subsequent activity.

### 2.3.4 Class B faults

Class B faults are imaged by seismic pro les A and B, making up the majority of faults on these pro les. Class B faults form structures of conjugate normal faults that converge at depth within the sedimentary section, showing little to no offset at or below re ector R12 (e.g. Figure 2.6, E and F). Typically, a central pair of faults that reach the seabed meet at around 1.5 s TWT below sea oor ( 1800-2000 mbsf) in the vicinity of re ector R10 (8.6 Ma). One or both of these central faults may penetrate the entire sediment section, but die out before or very shortly after reaching the top of the oceanic basement. The central pair of faults are surrounded by multiple, typically 10-20, seismically-resolvable faults, with a roughly even distribution between seaward- and landward-dipping geometries. These minor faults include multiple blind faults that branch from the central pair (hard-linked) and faults that are just spatially related (soft-linked). The number of blind faults increases below re ector R5 (2.6 Ma), but then decreases below re ector R10. Gridded displacement for Class B conjugate fault groups (example in Figure 2.6 G) shows that the more vertically extensive faults dominate, but that maximum displacement is concentrated in the middle of the sediment section, decreasing up- and down-section. This depth-displacement relationship is maintained when all the faults in the group are summed together (Figure 2.6 H). We suggest that this is the result of purely strike-slip motion on a sub-vertical plane at depth expressed as extension in the surface cover, therefore requiring vertical displacement to decrease downwards, and that the upward decrease indicates continuous activity started at where maximum displacement is observed (Figure 2.6 I).

### 2.3.5 Class C faults

Class C faults are only observed in the eastern part of our study area (on seismic pro les A and B) and are spatially related to signi cant basement topography. Anomalous basement highs - long-wavelength ( 20 km wide), topographic changes of 0.6-0.8 s of TWT (approx. 900-1200 m) - are also coincident with the locations of Wharton Ridge fracture zones (e.g. Jacob et al., 2014). In the seismic data these basement anomalies show a different character to seamounts; seamounts are shorter wavelength ( 5 km wide), more acoustically transparent, and show up-warping of the adjacent and overlaying sedimentary layers (e.g. at 160 km along pro le, Figure A.1). We also note that across the basement anomalies, we observed the same, strongly re ective seismic character as at the top of the oceanic basement, whereas the top of seamounts are less re ective.

FIGURE 2.6: A-D Example Class C (fracture zone) faulting. (A) uninterpreted and (B) interpreted seismic sections. Note long-wavelength basement topography. (C) Heat map of fault throw, warmer colours indicate greater amounts of throw (see text for method explanation). (D) Fault throw summed across the whole fault zone against depth in two-way-travel-time (also plotted as example F7N in Figure 2.7B). (E-H) The same analysis but for Class B faulting. Panel D also plotted as example pro le B in Figure 2.7A. In B and F subhorizontal coloured lines are used to create plots C and D, and G and H, respectively (where brightly coloured lines are re ectors R1-12, and dull blue lines are additional re ectors interpreted for detailed analysis). TOB – Top of Oceanic Basement. (I) Schematic interpretation of Class B faulting. A purely strike-slip shear fault at depth, within the crust and upper mantle creates extension in the sedimentary cover (see text). spB – Seismic Pro le B.

We attribute all Class C faults observed in the seismic data within the extents of basement anomalies to be structurally related to the major fracture zones. Within these zones there are 3-6 sea oor to basement faults, with up to 50 minor blind faults. The number of blind faults increases with depth in the section, with a marked increase at re ector R5 (2.6 Ma). Unlike Class B faults, the number of blind faults does not decrease below re ector R10 (8.6 Ma). Displacement across Class C faults (e.g. Figure 2.6 C, top panel) shows that individually, blind faults have minor vertical offset, and that sea oor-to-basement faults dominate. When all faults within each Class C structure are summed, the displacement increases with depth to at least re ector R11 (9.3 Ma) (e.g. Figure 2.6 D). However, this trend does not necessarily continue down to basement, and displacement appears to be less at re ector R12 (14.4 Ma) for the basement anomalies that relate to fracture zone F7 (Figure 2.7 B).

The bathymetric trace above and extrapolated north-south from the seismic Class C faults (see Appendix A Figure A.5) shows an overall N-S trend, but is composed of en-echelon NW-SE normal faults. In places, NW-SE faults form small pull-apart basins, indicating left-lateral motion.

### 2.3.6 Class D faults

On NW-SE pro le C (Figure 2.1), we nd evidence of reverse faulting and associated fault-propagation folding. These Class D faults are buried by the youngest sediments with no surface trace in the bathymetry data, therefore we cannot con rm their orientation. Since we do not image these features on the WSW-ENE seismic pro les A and B (Figure 2.1), we infer they strike broadly. NE-SW to E-W, in agreement with Abercrombie et al. (2003). Maximum compression in the Wharton basin is NW-SE (Gordon and Houseman, 2015), and Hananto et al. (2018) observed ENE-WSW bathymetric features associated with thrust faults to the south of our study area. This is a clear contrast with the dominant E-W trending reverse faults west of the NER (between 78° E and 82° E) (Bull, 1990; Bull and Scrutton, 1992, 1990; Krishna et al., 2001). In addition, we observe buried, high-angle (65-70°) normal faults in pro le C that offset the oceanic basement (not observed on pro les A and B).

The warping of sedimentary horizons due to reverse folding (Figure 2.8) affects most/all of the sedimentary section, starting at the top of the oceanic basement, and decreasing in amplitude upward, suggesting recent activity. We observe growth strata in the hanging wall onlapping onto the crest of the fault-propagation folds. The occurrence of onlap is not continuous, but is concentrated onto re ector R8 (7.2 Ma) and onto a horizon in the interval between re ectors R7 (5.8 Ma) and R6 (3.5 Ma)

FIGURE 2.7: Examples of fault displacement pro les for different fault classes plotted with depth/sediment age alongside strain of faults west of the NER and plate motion history through time. (A) Class B fault displacement with depth/re ector age. Coloured lines are three fault examples; black line is overall average for all Class B faults. (B) Displacement pro les with depth/re ector age for major fracture zones F6 and F7 (the latter from pro les A (F7N) and B (F7S)). The frequency of fractures detected in core at site U1480 is adjusted to age and superimposed on both (A) and (B). (C) Analysis by Krishna et al. (2009) for strain budget for faults west of the NER from fault throw, plotted against age. Strain is normalized by deformation extent on each seismic pro le used (longitude of N-S pro les indicated). D) India-Capricorn nite rotation angles with time from plate kinematic reconstructions (Bull et al., 2010), showing acceleration around 8 Ma. Used here as a proxy for India-Australia relative motion.

(arrows, Figure 2.8). Onlap/growth strata are visible deeper than re ector R8, i.e. older than 7.2 Ma, however, our ability to constrain earlier activity is limited by the quality and resolution of the seismic data, and basement topography between the IODP drill sites and the locations of Class D faults obscuring sedimentary horizons. Above re ector R8 where we observe no onlap, thickness is generally maintained suggesting activity was discontinuous.

### 2.3.7 Fault displacement and dip direction variations from the NER to the subduction zone

Overall trends in fault displacement were analysed along seismic pro les A and B to determine its variation with proximity to the subduction zone. We split displacement measurements into landward- and seaward-dipping fault polarities for each re ector, with positive values for landward-dipping fault displacements and negative for seaward-dipping. The polarity of cumulative displacement therefore re ects the dominance of landward- (positive) versus seaward-dipping (negative) faults. For each re ector analysed (R2-12), this differential cumulative displacement (e.g., Figure 2.9) shows an overall positive increase towards the subduction zone hence a net dominance of landward-dipping faults. However, there is variation between re ectors. Re ectors R7-9 ( 6–8 Ma) show the greatest increase in displacement towards the subduction zone, on both landward- and seaward-dipping geometries, and are the same re ectors where maximum displacement is most commonly observed on Class B faults.

Since faults are roughly evenly distributed west to east, we can group faults and sum displacement for each re ector (R2-12) for overlapping 20 km segments along pro le to show how displacement with depth varies across the Wharton Basin (Figure 2.9 B). This shows peak displacement close to the NER is in the middle of the section, increasing roughly linearly with depth towards the east, approaching the subduction zone. However, fracture zone structures in the eastern part of the section, with speci c displacement patterns, may bias this result. We observe no signi cant difference in the depth-displacement pro les of landward- versus seaward-dipping faults.

### 2.3.8 Relationships between seismic and core-scale faulting and sediment material properties

For each 20 km segment on seismic pro les A and B, we normalise the depth-displacement pro le against the corresponding maximum displacement (black line, Figure 2.7A). This allows us to directly compare the pattern of displacement with depth for each 20 km segment with the IODP borehole data. By adjusting the depths of the seismic re ectors (R2-12) to Site U1480, we can plot fault displacement against various datasets from the borehole (i.e. lithology, sand fraction, age-depth relationship, fracture intensity, seismic velocity and porosity, see Appendix A Figure A.6). Figure 2.7A shows the overall average displacement of the 20 km segments (black line) and examples (coloured lines) from detailed analyses (e.g. as shown in Figure 2.6 E-H). There is a linear increase in displacement with depth down to

FIGURE 2.8: Example Class D reverse fault structure on NW-SE trending pro le C. Addi- tional dashed horizons show onlap of growth strata (GS) onto the crest of the fault- propagation fold (indicated by arrows). Also shown is a spreading-ridge-normal- fault (SRNF) that has been re-activated through Class A strike-slip faulting.

re ector R7 (5.8 Ma), however the patterns are bell-shaped, showing maximum displacement in the middle of the sediment section, mostly between re ectors R7 and R10 (5.8 and 8.6 Ma, 600-1100 mbsf in borehole U1480). The overall average shows maximum displacement at re ector R8 (7.2 Ma), at borehole depth 800 mbsf close to the lithological IIB-IIC subunit boundary, which represents a general downhole reduction in sand content (McNeill et al., 2017b).

There is some correlation between the increased frequency of faults in cores with the mid-section zone of increased seismic-scale fault displacement (Figure 2.7A and B). No faults were observed in cores shallower than 400 mbsf (re ector R6), but a signi cant number were observed between re ectors R6 (3.5 Ma) and R10 (8.6 Ma) (400-1100 mbsf). Between re ector R10 (8.6 Ma) and R12 (14.4 Ma) (1100 and 1300 mbsf), no core faults were observed. We observe no other correlation between seismic-scale fault structure/displacement and other material properties, e.g., velocity, porosity.