The Thinning of Continental Lithosphere Leading to Continental Breakup and the Initiation of Seafloor Spreading

UNIVERSITY OF LIVERPOOL

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Abstract

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This thesis investigates how continental lithosphere thins leading to continental breakup and the initiation of a seafloor-spreading centre. The formation of an intra-continental rift basin can be explained assuming depth-uniform continental lithospheric thinning; however, this model breaks down when applied to rifted continental margins and failed propagating tip basins, as there is a difference between the observed extension in the upper crust and the extension of the rest of the lithosphere. Rifted continental margins form by a process of depth-dependent lithospheric thinning, how this deformation mode manifests is poorly understood. The Woodlark Basin provides the opportunity to observe continental lithospheric thinning at the present time. Pre-, syn- and post-breakup processes are all observable allowing for the formation of an analogue of continental lithospheric thinning through time leading to continental breakup and the initiation of seafloor spreading. Thinning factors have been determined, at the rifting to spreading transition in the Woodlark Basin, for the whole lithosphere from subsidence analysis, whole continental crustal from gravity inversions, and upper crust from fault analysis. Subsidence analysis using flexural backstripping of 2D cross-sections near the Moresby Seamount in the region of pre-breakup continental lithosphere thinning gives whole lithosphere thinning factors 0.5 to 0.8 increasing eastwards towards the propagating tip of sea-floor spreading, and indicates that subsidence requires substantial crustal thinning. Gravity inversion has been used to determine Moho depth, crustal thickness and thinning factors; predicted thinning factors from gravity inversion are similar to those obtained from flexural backstripping and Moho depths from gravity inversion are consistent with those from receiver function analysis. Fault analysis of seismic reflection data show upper crustal thinning factors of between 0.1 to 0.4 for the vicinity of the Moresby Seamount, substantially lower than thinning factors predicted for the whole lithosphere and continental crust, indicating depth-dependent lithospheric stretching and thinning. The observation of depth-dependent lithospheric thinning prior to continental breakup shows that it is a syn-rift, and not a post-breakup process. A comparative study has been undertaken on a long composite seismic line across the South China Sea. Here the rifted continental margins are also shown to have undergone depth-dependent lithospheric thinning prior to continental breakup. The breakup history of the paleo Papuan Peninsula is shown to have varying degrees of volcanism, both regionally and locally. This variation is to such a degree that the classification scheme of rifted continental margins struggles to label the Woodlark Basin as a single rift type. How the initiation of a seafloor spreading centre manifests is poorly understood. In the Woodlark Basin, the initiation seafloor spreading does not always lead to the formation of a mature seafloor-spreading centre. Young spreading centres can become abandoned, leading to a ridge jump, and the formation of a rifted continental ribbon.

Acknowledgments

Without a discussion, over a cup of coffee halfway up Mt. Teide, I would never have thought about doing a PhD. I have to thank Nick Kusznir for the most persuasive and energising two-hour discussion about the Woodlark Basin and continental lithospheric thinning. For your time, intelligence and, most of all, belief in me, thank you.

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My family has always supported me. Mum and Dad, no one has inspired me as much as you two have. To Chris, thanks for the occasional transaction and Liz, thank you for taking six years to go through university, my nine does not look too excessive now. I hope I can always be there for you all as you have been for me.

Finally, and most importantly, Sheena – you are everything, and without you, I would break.

The first question which you will ask and which I must try to answer is this, "What is the use of climbing Mount Everest?" and my answer must at once be, "It is no use." There is not the slightest prospect of any gain whatsoever. Oh, we may learn a little about the behaviour of the human body at high altitudes, and possibly medical men may turn our observation to some account for the purposes of aviation. But otherwise nothing will come of it. We shall not bring back a single bit of gold or silver, not a gem, nor any coal or iron. We shall not find a single foot of earth that can be planted with crops to raise food. It's no use. So, if you cannot understand that there is something in man which responds to the challenge of this mountain and goes out to meet it, that the struggle is the struggle of life itself upward and forever upward, then you won't see why we go. What we get from this adventure is just sheer joy. And joy is, after all, the end of life. We do not live to eat and make money. We eat and make money to be able to enjoy life. That is what life means and what life is for.

George Leigh Mallory, 1922

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Chapter 1

Introduction

1.1 Motivation

The thinning of continental lithosphere and crust, leading to continental breakup and seafloor-spreading initiation, is a fundamental process in the Wilson plate tectonic cycle. At present, this process is poorly understood. The understanding of how the continental lithosphere thins, prior to continental breakup, has consequences for our knowledge of lithosphere deformation leading to continental lithospheric rupture, the formation and structure of rifted continental margins and the formation of the first oceanic crust that forms after continental breakup has occurred. The Woodlark Basin, east of Papua New Guinea, provides an ideal natural laboratory to study the thinning of continental lithosphere leading to continental breakup and seafloor-spreading initiation. The understanding of how the continental lithosphere thins and how rifted continental margins form is not only of academic interest; furthering our knowledge will aid petroleum exploration of these frontier deepwater regions and improve our future energy security.

Recent discoveries from observations of continental rifted margins, such as the exhumation of continental lithospheric mantle (Manatschal and Bernoulli, 1999, Pickup et al., 1996, Whitmarsh et al., 2001, Boillot et al., 1987, Pinheiro et al., 1992), and a discrepancy between the amount of thinning of the whole lithosphere and crust than implied by upper crustal fault extension (Davis and Kusznir, 2004, Driscoll and Karner, 1998, Kusznir and Karner, 2007, Roberts et al., 1997, Royden and Keen, 1980), have led us to question our current knowledge of how the lithosphere thins leading to continental breakup and seafloor spreading. The McKenzie pure shear model (McKenzie, 1978) has been used successfully for the past three decades to describe lithospheric thinning at intra-continental rifts, however it fails to predict depth-dependent lithosphere thinning and the exhumation of mantle observed at many rifted continental margins. This has led to the development of other numerical models (Huismans and Beaumont, 2002, Kusznir and Karner, 2007, Lavier and Manatschal, 2006) that try to explain the formation of rifted continental margins.

The majority of rifted continental margins that are studied have lithosphere that is old and cold, and this lithosphere has thermally equilibrated since continental breakup occurred. Many margins are covered in thick sediment sequences, deposited since their formation, making observations of the formation processes that occurred to form them difficult to determine. The Woodlark Basin provides an opportunity to study the breakup of a continent and the formation of an ocean basin that is ongoing (figure 1.1). The study of the Woodlark Basin has advantages over ancient margins, as it is possible to observe continental lithosphere that is presently thinning in a process that will lead to continental breakup and the initiation of seafloor spreading. As it is a young system with few surrounding landmasses, the Woodlark Basin is sediment starved enabling the observations of the formation of rifted continental margins to be made with greater clarity. The Woodlark Basin, not only allows for the observation of post-breakup systems, but also presents the opportunity to observe the pre-rift and syn-rift processes that lead up to continental lithospheric rupture.

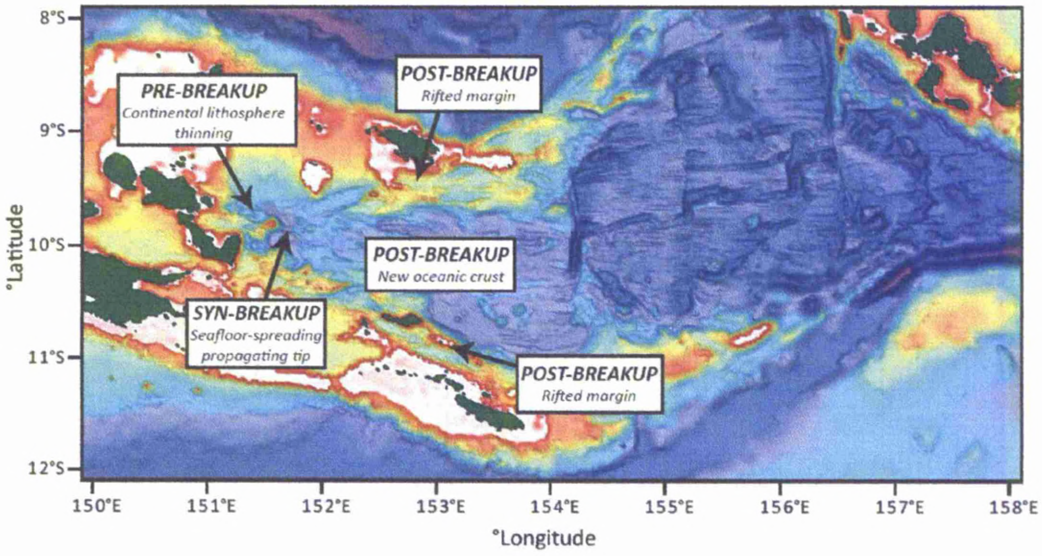

Figure 1.1. Bathymetric map of the Woodlark Basin. The Woodlark Basin is a good case study of continental breakup as it exhibits regions of pre, syn and post-breakup, these regions are identified.

1.2 Approach

The thesis investigates how a continent thins and breaks up, leading to the onset of seafloor spreading, using primarily observational data to determine thinning and stretching of the lithosphere at the level of the whole lithosphere, the whole crust and the upper crust (figure 1.2). Thinning for the whole lithosphere is determined from flexural back-stripping and the McKenzie continental lithosphere extension model (McKenzie, 1978), crustal basement thinning is determined using satellite gravity inversion and upper crustal stretching is determined from fault analysis.

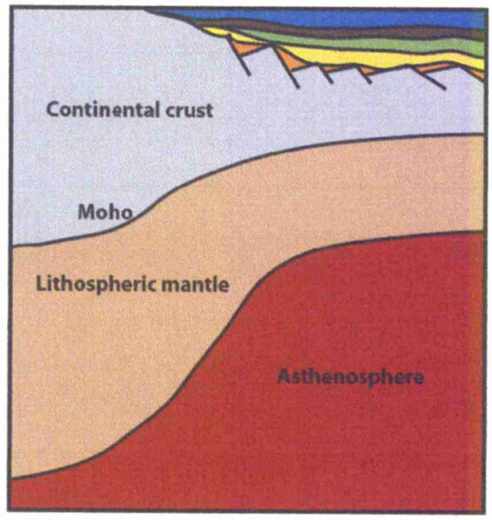

Figure 1.2 The three levels of continental lithosphere, the whole lithosphere, the whole crust and the brittle upper crust, at which it is possible to determine the degree of thinning that has occurred. Modified from Kusznir & Karner 2007.

For the Woodlark Basin, and in a shorter comparative study on the South China Sea (SCS), the following has been determined:

i. Whole lithosphere thinning factors using a 1D McKenzie pure shear model (McKenzie, 1978). Subsidence analysis of known pre, syn and post-rift sediments is used to determine water-loaded subsidence values. These values of water-loaded subsidence are subsequently used within a 1D McKenzie pure shear model to determine whole lithospheric thinning factors.

ii. Crustal basement thinning is derived from a satellite gravity inversion, incorporating a correction for the lithospheric thermal gravity anomaly, which determines Moho depth. Crustal thickness is determined from the vertical distance from the top basement to the Moho. This technique is used to determine continent-ocean transition zones and crustal basement thinning.

Initial oceanic crust thicknesses, derived from the crustal thickness maps, are used to determine the volcanism of the Woodlark Basin breakup history.

iii. Upper crustal extension, due to faulting of the brittle upper crust, is determined by fault heave analysis.

Differential lithospheric thinning is determined from a comparison of results from methods i, ii and iii. For depth-uniform thinning of the lithosphere during continental thinning, leading to continental breakup and the initiation of seafloor spreading, results for all three levels should be similar. If there are discrepancies between results, then the region is assumed to have undergone differential lithospheric thinning prior to continental lithospheric rupture. Forward modelling of the continental lithosphere is undertaken, from initial pre-rift conditions through to continental breakup and formation of new oceanic crust. This technique allows for the trial of both old and new models of continental lithospheric thinning and investigates whether they can explain the formation of either the Woodlark Basin and SCS, using constraints derived from methods i, ii & iii.

1.3 Thesis Structure

The core of this thesis, chapters 4, 5, 6 and 8, take the form of a series of journal papers. Due to this format, some background material and key concepts are repeated and reintroduced in each chapter.

Chapter 2 provides an overview of models currently used to explain the thinning of continental lithosphere leading to the formation of either an intra-continental sedimentary basin or a rifted continental margin. This chapter critically reviews current models and explains why further research is necessary.

Chapter 3 discusses the geological setting and tectonic history of the Woodlark Basin. The data sets, which are used in this study, are reviewed and their acquisition discussed.

Chapter 4 is presented in the form of a GJI journal paper, yet to be submitted, entitled *"Evidence for Depth-Dependent Lithosphere Thinning at the Rifting to Spreading Transition in the Woodlark Ocean Basin, Western Pacific."* This paper uses observational data from the Woodlark Basin to determine whole lithosphere, whole crustal and upper crustal extension in the region immediately ahead of the propagating tip of seafloor spreading. Thinning factors for the three different levels of the lithosphere are compared and extension, prior to continental breakup, is shown to occur depth-dependently and not depth-uniformly.

Chapter 5 takes the form of a paper, to be submitted to GJI, entitled *"The Distribution of Crustal Basement Thinning in the Final Stages of Continental Breakup: Evidence from the Woodlark Basin derived from Satellite Gravity Inversion."* This paper uses a gravity inversion, that incorporates a lithospheric thermal gravity anomaly correction (Chappell and Kusznir, 2008, Greenhalgh and Kusznir, 2007), to produce crustal thickness and whole crustal thinning factor maps of the Woodlark Basin. Sensitivities to breakup age, crustal reference thickness, sediment thickness and volcanic addition are shown. The concept of rifted continental ribbons, and how they form, is described using evidence from the gravity inversion of the Woodlark Basin.

Chapter 6 is a short paper, to be submitted to Geology, entitled *"Oceanic Crustal Thickness Variation in the Woodlark Basin: Evidence from Satellite Gravity Inversion."* This paper shows a change in oceanic crustal thickness across the Moresby Transform and a regional variation in oceanic crustal thicknesses in the Woodlark Basin. The observation of this variation in initial oceanic crustal thickness shows that the Woodlark Basin does not behave as a classic magma poor basin as previously reported. The variability of initial oceanic crustal thickness means the Woodlark Basin does not conform to current end member models of volcanism during continental breakup and a new system of classification is needed to incorporate basins that experience average amounts of volcanism during breakup.

Chapter 7 synthesises the work in chapters 4, 5 and 6 on the Woodlark Basin. It discusses the implications of depth-dependent lithospheric thinning, prior to continental breakup, for the Woodlark Basin. Models of the formation of the present rifting to drifting to transition are shown and compared; these models are calibrated using thinning factors results described in chapter 4. The issue of a possible divergence velocity discrepancy within the lithosphere, observed in the Woodlark Basin, is raised and its implications for the use of a single Euler pole model for regions prior to and after continental breakup are discussed.

Chapter 8 is a paper, to be submitted to Tectonophysics, entitled *"Depth-Dependent Lithospheric Thinning During the Formation of the South China Sea Derived From a Long Composite Seismic Reflection Line"* and presents results from a comparative study in the South China Sea. The South China Sea is an ocean basin with many similarities with the Woodlark Basin in terms of shape and the propagation of seafloor spreading. This paper shows the translation of techniques used previously on the Woodlark Basin to the South China Sea. Depth-dependent lithosphere thinning is shown to occur prior to the initiation of seafloor spreading.

Chapter 9 is a short chapter examining the nature of the basement in the Phu Khanh Basin, offshore Vietnam in the South China Sea and is modified from the paper entitled *"Rifting of the South China Sea: new perspectives"* of which Simon Gozzard is a co-author (Cullen et al., 2010). The nature of basement of the Phu Khanh Basin is unclear since there is no publically accessible literature of any drilling in the region. Whilst this work assumes depth-uniform thinning only, it presents the application of lithospheric thinning factors as a test to investigate the nature of crust and as a tool to determine OCT and COB location. The results speculate whether the basement is continental crust, highly thinned continental crust with volcanics or oceanic crust.

Chapter 10 is a summary overview of how the continental lithosphere thins, prior to lithosphere rupture and the initiation of seafloor spreading. Results and observations are discussed and areas that need further investigation are highlighted. The implications of depth-dependent lithospheric thinning at both the Woodlark Basin and the South China Sea are used to explain possible mechanisms of how a continent breaks up leading to seafloor spreading. The author's views on future work on the subject of continental lithospheric thinning prior to continental breakup and the study of the Woodlark Basin are summarised.

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Chapter 2

The Thinning and Breakup of Continental Lithosphere Leading to the Formation of a Rifted Continental Margin

This chapter reviews existing models of continental thinning leading to breakup and the initiation of seafloor-spreading system.

2.1 Historical background

The breakup up of continents and formation of new ocean basins has been a field of scientific research ever since the 1500s when it was observed that the continental edges of South America and Africa looked like they would fit together like pieces in a jigsaw. However, the assumption at the time that the Earth was static, made various proposals of how they would have broken apart hard to explain. It was Alfred Wegener who first proposed the theory that the continents had drifted apart from each other (Wegener, 1912). However, without detailed evidence and a force sufficient to drive the movement of the continents, the theory was not generally accepted: the Earth might have a solid crust and mantle and a liquid core, but there seemed to be no way that portions of the crust could move around.

It was not until the 1960s, when Harry Hess proposed seafloor spreading (Hess, 1962) and the discovery and understanding of the parallel magnetization of the oceanic crust (Mason and Raff, 1961, Vine and Matthews, 1963), that it was realised that tectonic plates not only existed but moved, converged and diverged over time, destroying old crust and forming new crust, respectively, in the process (Wilson, 1966). This led to the general acceptance of the theory of plate tectonics by almost all earth scientists as the cause for continents to breakup.

This study will look at observational evidence from the Woodlark Basin and the South China Sea to try to further our understanding of the mechanics of how lithosphere stretching and thinning can lead to continental breakup and the formation of a new ocean basin.

Several models have been developed over the past 30 years to explain how the continental lithosphere thins. These models, whilst theoretical, are based on and tested against observations. They do not explain all the observations at rifted continental margins. Therefore, they do not fully explain the thinning of continental lithosphere that leads to continental breakup and the initiation of seafloor spreading.

All models that are introduced in this chapter are 2D, they do not imply any flow in or out of the page.

2.2 The McKenzie pure shear model

The McKenzie pure shear model (McKenzie, 1978), as shown in figure 2.1, proposes that the whole continental lithosphere thins uniformly with depth. The mechanical thinning of the continental crust causes a subsidence at the surface of the crust. The stretching and thinning of the lithosphere has another effect; it causes a shallowing of hot lithospheric material and therefore increases the geothermal gradient. This increase in the geothermal gradient causes uplift known as *thermal uplift*. The thermal uplift is a smaller magnitude than that of the subsidence associated with the crustal thinning, hence immediately after the thinning of the lithosphere there is a net subsidence, known as the *initial subsidence*. Over time, the elevated geothermal gradient in the stretched continental lithosphere restores to thermal equilibrium and the lithosphere becomes denser causing a secondary stage of subsidence known as *post-rift thermal subsidence*. Total subsidence due to the pure shear model is the combination of the initial subsidence and the amount of post-rift thermal subsidence; this is shown in figure 2.2.

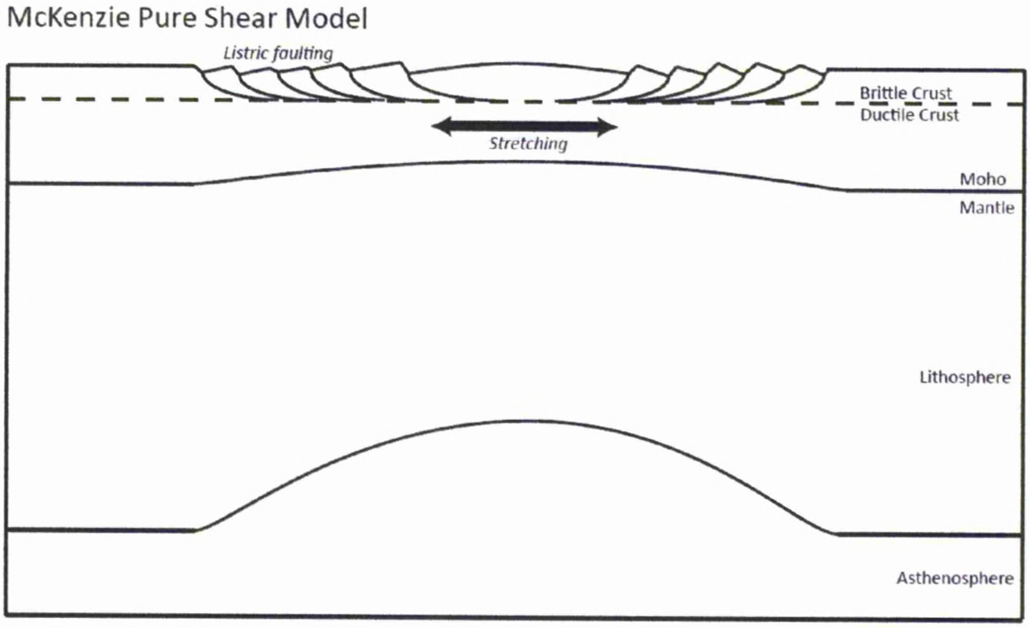

Figure 2.1 The pure shear depth-uniform mode of lithospheric extension as proposed by Dan McKenzie in 1978. The whole lithosphere thins depth-uniformly and the volume of lithosphere is conserved.

The McKenzie pure shear model assumes depth-uniform stretching and thinning of the continental crust and lithosphere. This model is a simplification since the seismogenic upper crust deforms by brittle faulting, and not by a pure shear ductile mechanism. Assuming only pure shear stretching and thinning of the continental lithosphere, it is difficult to model continental lithospheric rupture and breakup, because the lithosphere will thin to an infinitely thin layer without breaking apart. The formation of a rifted continental margin requires continental lithospheric rupture to occur, thus the McKenzie model cannot model the breakup of continental lithosphere.

The McKenzie pure shear model (McKenzie, 1978) assumes instantaneous rifting. Observations of intra-continental rift basins and rifted continental margins show that stretching and thinning of the continental lithosphere can occur for tens of millions of years prior to either continental breakup or the failure of the rift. Throughout this time, hot asthenosphere will constantly be upwelling underneath the thinning continental lithosphere; however, the heat brought into the system will also be dissipating at a rate whereby at the end of rifting, the continental lithosphere will not be as hot as predicted by the instantaneous McKenzie model.

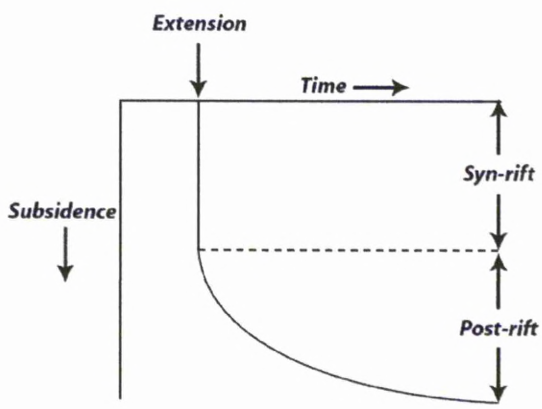

Figure 2.2 The components of total subsidence due to McKenzie type stretching and thinning of continental lithosphere. Syn-rift thinning of the lithosphere is assumed to occur instantaneously whereas the amount of post-rift subsidence due to the cooling of the lithosphere is dependent on the time since rifting. Total subsidence is the sum of syn-rift and post-rift subsidence.

2.3 The extension discrepancy

The McKenzie pure shear model (McKenzie, 1978) provided the first simple quantitative framework for predicting the subsidence in an intra-continental rift basin or a rifted continental margin due to lithospheric extension assuming depth-uniform stretching and thinning of the lithosphere. The McKenzie model has been successfully applied to numerous intra-continental rift basins (Marsden et al., 1990, Roberts et al., 1993, White, 1990).

When extension was measured at rifted continental margins, in contrast to intra-continental rift basins, it was observed that the amount of faulting in the upper crust could not explain the observed subsidence if depth-uniform stretching and thinning of the continental lithosphere was assumed (Baxter et al., 1999, Davis and Kusznir, 2004, Driscoll and Karner, 1998, Kusznir and Karner, 2007, Roberts et al., 1997). This led to the phrase *extension discrepancy* being coined. The extension discrepancy has been described for many rifted continental margins and they are listed below:

- Voring Basin, Norway (Kusznir et al., 2005, Roberts et al., 1997)
- Exmouth Plateau (Driscoll and Karner, 1998)
- Pattani Basin, Thailand (Watcharanantakul and Morley, 2000)
- South China Sea (Clift and Lin, 2001, Cullen et al., 2010)
- Black Sea (Meredith and Egan, 2002)
- Goban Spur (Davis and Kusznir, 2004)
- Galicia Interior Basin (Davis and Kusznir, 2004)
- Lower Congo Basin (Contrucci et al., 2004).

When the thinning of the whole lithosphere is directly compared to the thinning of the upper crust, derived from fault extension and assuming depth-uniform lithospheric thinning, at intra-continental margins the degrees to which both have thinned are the same. However, at rifted margins the whole lithosphere has thinned to a greater degree than what is observed in the upper crust (figure 2.3). An additional 40% of sub-seismic upper crustal extension (Walsh et al., 1991) is insufficient to account for the difference in the amount of thinning between the whole lithosphere and the upper crust.

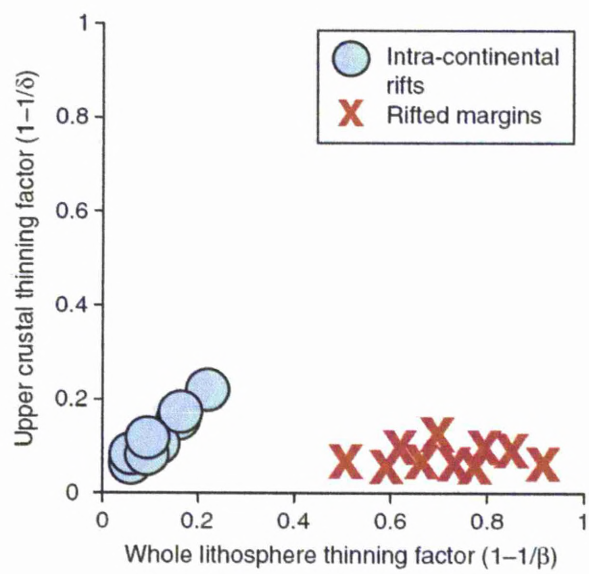

Figure 2.3 Cross-plot of upper crustal thinning factors for intraplate continental rifts and rifted continental margins. Intraplate continental rifts show depth-uniform lithospheric thinning, while rifted continental margins show depth-dependent lithospheric thinning. Taken from Kusznir & Karner 2007.

These observations of the extension discrepancy lead to two conclusions; either that the lithosphere is not thinned uniformly and that depth-dependent stretching is occurring during rifted margin formation or that not all the extension of the upper crust is observed from seismically visible fault extension (Reston, 2009). It is plausible that multiple phases of faulting occur during the stretching of the upper crust prior to continental breakup. Only the last phase of faulting, prior to continental breakup, would be observed in seismic since earlier faults would be sub-horizontal and hard to distinguish in seismic reflection data (Reston, 2009). Therefore the measurement of extension in the upper crust, from fault heaves, would be grossly underestimated, leading to the observation of the extension discrepancy.

An extreme example of depth-dependent lithospheric stretching during the formation of a rifted continental margin occurs on the Møre segment of the Norwegian rifted continental margin. Here a stretching factor of 4 is needed to account for the water loaded subsidence, whilst faulting is negligible and less than 1.02. Here, it is clear that no multi-phase faulting has occurred during rifted margin formation (Kusznir and Karner, 2007).

2.4 A 2D decoupled pure shear model

Based on the McKenzie pure shear model, a 2D decoupled pure shear model (figure 2.4) assumes uniform stretching in a vertical column between defined depths. The difference between this model and the McKenzie model is that different depths can experience different degrees of stretching. Within the model, two different levels, normally the top 15km of the crust and the rest of the lithosphere, exhibit different distributions of stretching to each other but with identical total extension. Within these defined depths, stretching is depth-uniform but for the whole lithosphere stretching is depth-dependent.

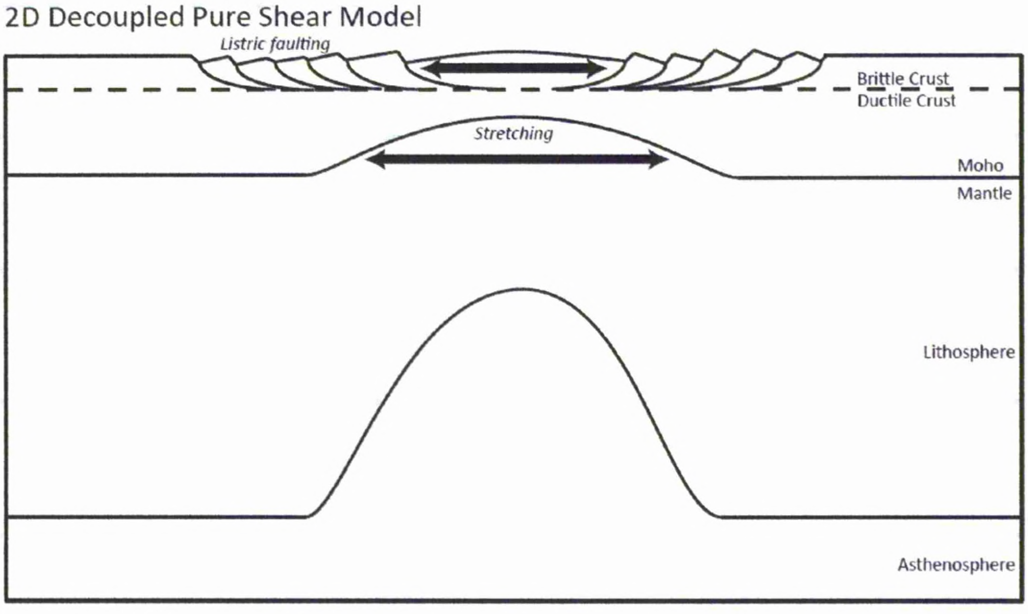

Figure 2.4 Diagram showing a 2D decoupled model. It works similarly to the McKenzie pure shear model. It differs by separating the lithosphere into 2 layers where different degrees of thinning can occur invoking depth dependency.

A 2D decoupled model requires a weak decoupling horizon at a depth within the lithosphere to accommodate the differential thinning (Lavier and Manatschal, 2006). Where and how this operates is unclear.

2.5 The Wernicke low angle detachment model

The Wernicke model (Wernicke, 1985), shown in figure 2.5, uses a large low angle detachment fault that reaches from the upper crust to the lower lithosphere to accommodate regional extension. The extension along this detachment causes the formation of an asymmetrical rift basin or asymmetrical conjugate rifted margins since one side exhibits the footwall, exposing lower crustal rocks, whilst the other side is the hanging wall with only upper crustal rocks (Lister et al., 1991). Upwelling material from the mantle is not directly underneath the zone of upper crustal extension, as in the McKenzie pure shear model. This means that the observed rift basin exhibits little thermal subsidence, whilst a distal region on the rifted margin that was also the hanging wall to the detachment fault, has initial thermal uplift followed by subsequent thermal subsidence. The Wernicke model exhibits depth-dependent lithosphere thinning.

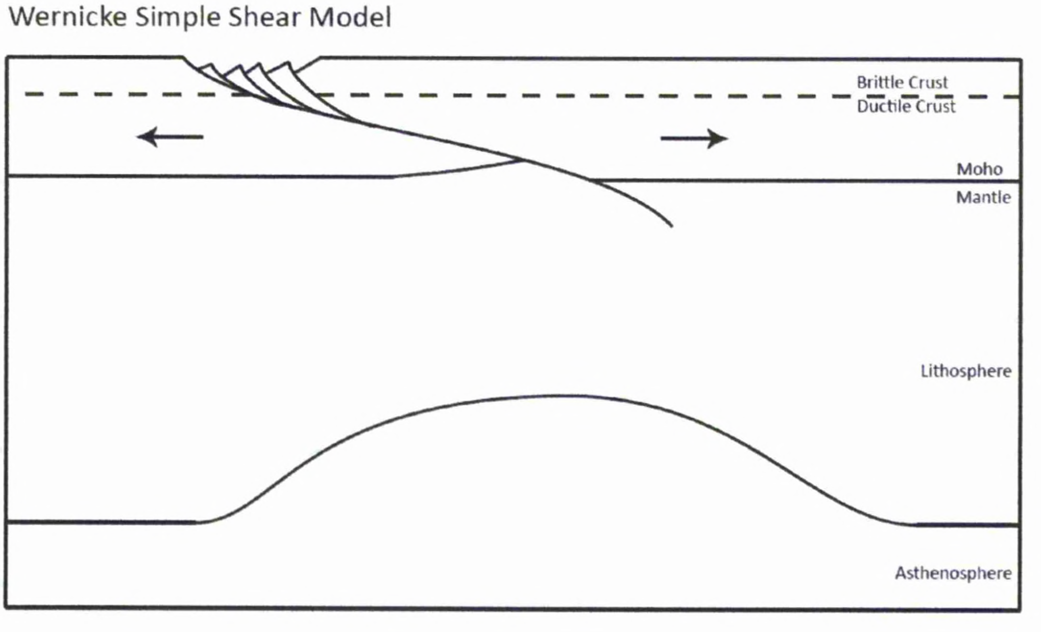

Figure 2.5 A diagram showing the Wernicke simple shear model of lithospheric extension. A large detachment fault is needed in order to thin the crust and lithosphere. The upwelling mantle is not directly beneath the observed extension in the upper crust. The volume of the lithosphere and crust is conserved and whilst thinning is depth-dependent, the depths at which the majority of thinning is accommodated varies between conjugate margins.

2.6 The Upper Plate Paradox

Studies of intracontinental rift basins (Marsden et al., 1990, Roberts et al., 1993, White, 1990) show that the McKenzie depth-uniform model (McKenzie, 1978) can be successfully applied to model their formation. However, at rifted continental margins, studies show that the lithosphere thins depth-dependently (Baxter et al., 1999, Davis and Kusznir, 2004, Driscoll and Karner, 1998, Kusznir and Karner, 2007, Roberts et al., 1997) and that the application of the McKenzie pure shear model cannot explain their formation. The Wernicke detachment model of lithosphere thinning exhibits depth-dependent lithospheric thinning (Wernicke and Burchfiel, 1982); however, in studies of rifted continental margins, thinning increases with depth. The Wernicke model only explains increased thinning of the lithosphere with depth for one of the two margins formed, as on the conjugate, thinning is predicted to be greater in the upper crust than the whole lithosphere. This is not observed, as studies of conjugate rifted continental margins have shown that both exhibit greater thinning at depth. This is known as the" upper plate paradox" (Driscoll and Karner, 1998) and shows the failure of the Wernicke model to be successfully applied to the modelling of the formation of conjugate rifted continental margins.

2.7 Pre-breakup upwelling divergent flow models

The extension discrepancy is not observed at intracontinental rift basins, such as the North Sea, however these rift basins are not associated with the onset of seafloor spreading. It is possible that depth-dependent lithospheric stretching, observed at rifted continental margins, is a process uniquely related to rifts that eventually became oceans.

The pre-breakup upwelling divergent flow model, shown in figure 2.6, assumes that stretching and thinning of the continental lithosphere leading to continental breakup and the onset of seafloor spreading occurs in response to pure shear upper crustal stretching and amplified by a buoyancy induced upwelling in the form of a small convection cell (Kusznir and Karner, 2007). The upwelling divergent flow field is based upon divergent mantle flow models that have been successfully applied to ocean ridges (Buck, 1991, Morgan, 1987, Spiegelman and McKenzie, 1987, Spiegelman and Reynolds, 1999). It accounts for depth-dependent lithosphere thinning, with greater thinning occurring at depth. Upwelling divergent flow fields have successfully been applied to model subsidence at rifted continental margins where there is an observation of the depth discrepancy between extension determined from upper crustal faulting and whole lithosphere thinning (Kusznir and Karner, 2007).

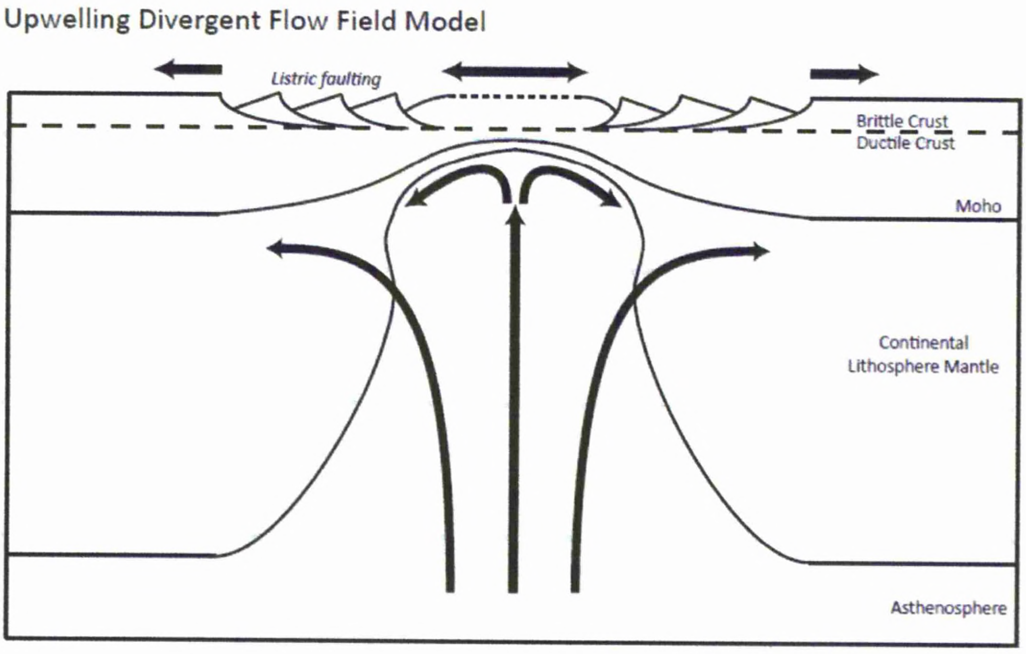

Figure 2.6 A diagram showing thinning of the lithosphere and crust due to an upwelling divergent flow field. The divergence of the flow pulls the lithosphere and crust apart as well as "eroding" the lithosphere from beneath. This model assumes depth-dependent lithosphere thinning.

2.8 Continental lithosphere and crust conservation in depth-dependent lithospheric thinning models

Models that exhibit depth-dependent thinning of the continental lithosphere and lower crust, displace the lower lithosphere, relative to the upper crust, by either exhuming it oceanwards, producing a lower crustal extrusion or by displacing it landwards, thickening the lithosphere causing hinterland uplift. This is problematic since no such passive margin mountains or lower crustal extrusions have been observed (Reston, 2009). The Woodlark Basin is more problematic since its rifted margins are bounded by oceanic plates to the north and south, making the displacement of lower crust and lithosphere, in a narrow region, harder to explain.

2.9 Structure of rifted continental margins

The structure of continental rifted margins varies significantly. Whilst individual margins are unique; they have similarities depending on the geological setting in which they form. Once continental breakup has occurred and a fully mature seafloor-spreading system has evolved, the structure of a rifted continental margin does not change significantly (Taylor et al., 2009), therefore it is the process by which a continent breaks up that predominantly defines the structure of a rifted continental margin.

The ocean continental transition zone (OCT) is assumed as the region where the continental lithosphere is intensely thinned and where complex tectonics, variable magmatism and possible continental lithospheric mantle exhumation make the identification of the continent-ocean boundary (COB) difficult. The width of the OCT can vary greatly depending on the amount of volcanism associated with the continental breakup. Magma-rich continental rifted margins, figure 2.7, are typically narrow and exhibit high amounts of magmatism prior to final breakup (Mutter et al., 1982, Barton and White, 1997, Hopper et al., 1992) and abnormally thick initial oceanic crust (Roberts et al., 1984, White and McKenzie, 1989, White et al., 1987, Geoffroy, 2005). In contrast, the OCT at magma-poor margins are usually wide and exhibit the exhumation of lower crustal and mantle rocks prior to the onset of seafloor spreading (Dean et al., 2000, Manatschal and Bernoulli, 1999, Pickup et al., 1996, Whitmarsh et al., 2001). The COB is the location where continental lithosphere or crust is ruptured by the initiation of seafloor spreading and the formation of oceanic crust. Margins with average amounts of volcanism during continental stretching and thinning are currently poorly described and classified; this issue is addressed in chapter 6.

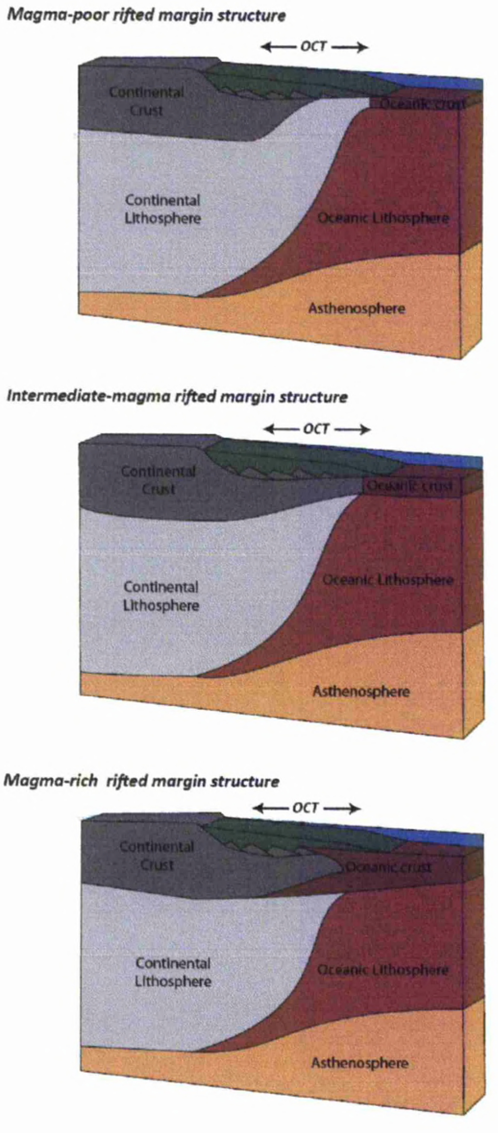

Figure 2.7 Schematic cartoons of magma-poor, intermediate-magma and magma-rich rifted margins after Hamsi (2011). Magma-poor margins exhibit thin initial oceanic crust and often have exhumed serpentinized mantle. Intermediate-magma margins have initial oceanic crust of a thickness that is similar to the global average. Magma-rich margins exhibit large volumes of volcanism resulting in thick initial oceanic crust and magma underplating.

The different types of rifted continental margins have led to the development of continental lithospheric stretching and thinning models to explain specific types of margins. Using field observations and geodynamic modelling, Lavier and Manatschal (2006) have suggested a complex lithospheric stretching and thinning history for the formation of magma-poor rifted continental margins, shown in figure 2.8. A sequence of different modes of extension is used to describe structures observed in both the Iberian and Alpine margins. The first stage of deformation is a broadly distributed stretching phase and is followed by strain localisation and crustal thinning along upper crustal and mantle ductile shear zones decoupled along a mid-crustal decollement. The thinning phase is localised by a weak mid-crustal rheology. This thinning can be accommodated by a limited amount of rift flank uplift and subsidence in the hanging wall. The final stage, prior to the onset of seafloor spreading is an exhumation phase where crustal embrittlement and continued extension leads to the formation of crustal-scale detachments along downward concave faults that are favourable to mantle exhumation (Lavier and Manatschal, 2006). This model, whilst possibly explaining the formation of two magma-poor margins, does not explain the formation of all magma-poor margins. It is likely that, whilst the formation of an individual margin is unique, rifted continental margins form due to a variation of a common process that is not yet fully understood.

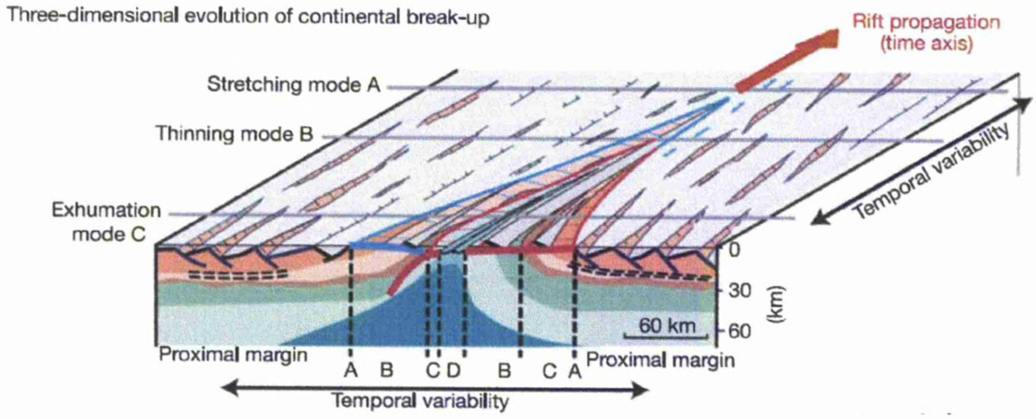

Figure 2.8 Summary diagram from Lavier & Manatschal (2006) showing the formation of a magma-poor rifted continental margin through different phases of stretching and thinning. The stages of continental stretching and thinning leading to continental breakup are a stretching phase, followed by a thinning phase, then an exhumation mode and finally the initiation of seafloor spreading.

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Chapter 3

Geological Overview of the Woodlark Basin

This chapter summarises the geology of the Woodlark Basin, which is located in the Southwest Pacific, its geological past and the data from which observations are made. It explores the possible reasons why the region is currently undergoing extension leading to the formation of the Woodlark Basin.

3.1 Location

The Woodlark Basin is a small ocean basin located east of the Papuan Peninsula in the southwest Pacific. It is surrounded by Papua New Guinea to the east and the Solomon Islands in the west, and by the continental rises of the Woodlark Rise and the Pocklington Rise to the north and south, respectively. The Woodlark is a v-shaped oceanic basin, with a seafloor-spreading centre propagating westwards into the Papuan Peninsula; it is approximately 700 km in length and 300 km at its widest in the east (figure 3.1).

The Woodlark Basin began rifting during the late Miocene (Taylor and Huchon, 2002), whilst the earliest seafloor spreading initiated at approximately 6 Ma in the east (Taylor et al., 1999). Seafloor spreading has propagated westwards at a rate of approximately 140 mm/yr, stepping across the Simbo Transform and the Moresby

Transform at 4 and 1.9 Ma respectively and at the present day it is interpreted to be immediately east of the Moresby Seamount.

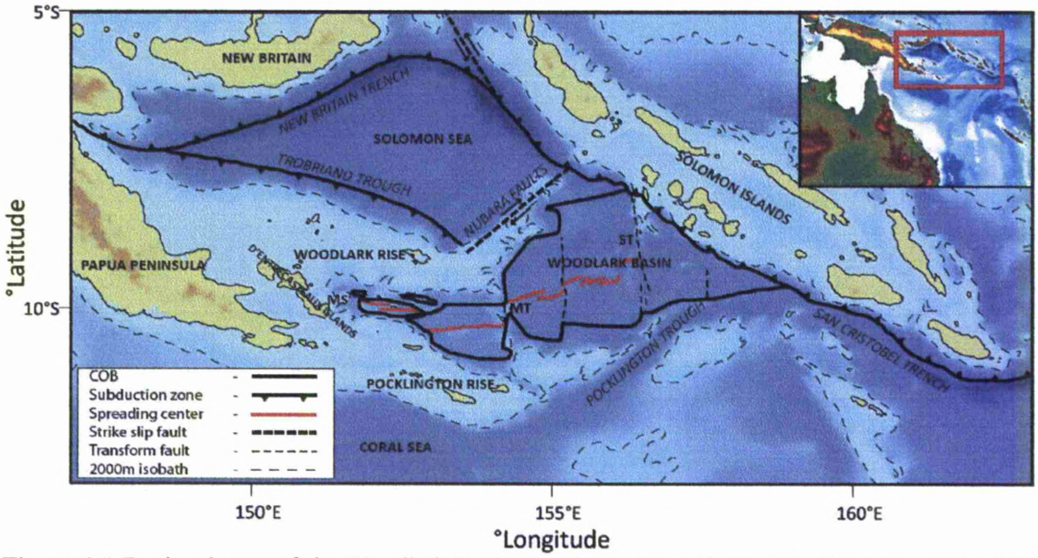

Figure 3.1 Regional map of the Woodlark Basin showing surrounding subduction zones, transform faults and active seafloor spreading centres. MS (Moresby Seamount), MT (Moresby Transform), ST (Simbo Transform) & COB (continent-ocean boundary) (Taylor et al., 1995). Illuminated bathymetry from GEBCO (Intergovernmental Oceanographic Commission, 2003)

3.2 Regional history of the Woodlark Basin and surrounding area

The palaeoextension of the Papuan Peninsula, that now makes up the conjugate margins of the Woodlark Basin, was formed in a setting which involved subduction and arc construction followed by collision and suturing of continental and arc terrains (Martinez et al., 1999). North of the Woodlark Basin lies the Trobriand Trough, a subduction zone where the Solomon Plate is being subducted. In unpublished seismic reflection data, it is possible to observe the subducting slab under the Woodlark Rise. Both the Solomon and Woodlark plates are being subducted in the east at the San Cristobel Subduction Zone. This subduction has led to the formation of the Solomon Islands, a volcanic island arc, and the destruction of the oldest seafloor in the Woodlark Basin. The numerous subduction zones in the region have led to a large component of differential plate movement; this is accommodated by the Nubara fault, a large dextral transform fault that penetrates the Woodlark Rise (Weissel et al., 1982, Taylor and Exon, 1987).

The Pocklington Rise and Papuan Peninsula formed when continental blocks rifted from Australia during formation of the Coral Sea (62-56 Ma). They moved northwards and eventually collided with arc terrains at the subduction zone where the continental blocks were partially underthrust by the Palaeogene arc (Weissel and Watts, 1979). The Pocklington Rise, a continental rise that makes up the southern-rifted margin of the Woodlark Basin, has contrasting features either side of the Moresby Transform. West of the Moresby Transform, the underthrusting of continental crust, occurring during the collision that formed the palaeoextension of the Papuan Peninsula, eliminated the remnants of the oceanic trench of the ancient subduction zone and led to the thickening of the Pocklington Rise. However, east of the Moresby Transform, the ancient oceanic trench is still preserved and is known as the Pocklington Trough. Here, the crust of the Pocklington Rise is thinner and comprises of the extinct arc associated with the ancient subduction zone (Karig, 1972, Martinez et al., 1999).

3.3 The D'Entrecasteaux Islands

The D'Entrecasteaux Islands are situated approximately 80 km ahead of the propagating tip of seafloor spreading in the Woodlark Basin and lie in a region of continental lithosphere that is undergoing extension. Rocks of high metamorphic grade are found on these islands, they exhibit eclogite facies rocks (figure 3.2), that were at depths of ~75 km at 4.3 Ma ± 0.3 Ma and have subsequently been exhumed at tectonic rates of centimetres per year (Baldwin et al., 2004). The nature of the exhumation of the rocks found on the D'Entrecasteaux Islands is not fully understood and is contested within the literature (Martinez et al., 2001, Webb et al., 2008). One model suggests subduction inversion for the exhumation of the eclogites (Webb et al., 2008) whereas another model suggests diapirism of subducted continental material (Martinez et al., 2001). The ancient subduction and collisional history of the tectonics creating the Papuan Peninsula and surrounding continental crust has led to a preferential geological system where these islands can form now that regional extension is taking place. However, it is unlikely that similar systems are a predominant feature in most regions that experience high degrees of continental extension leading to breakup.

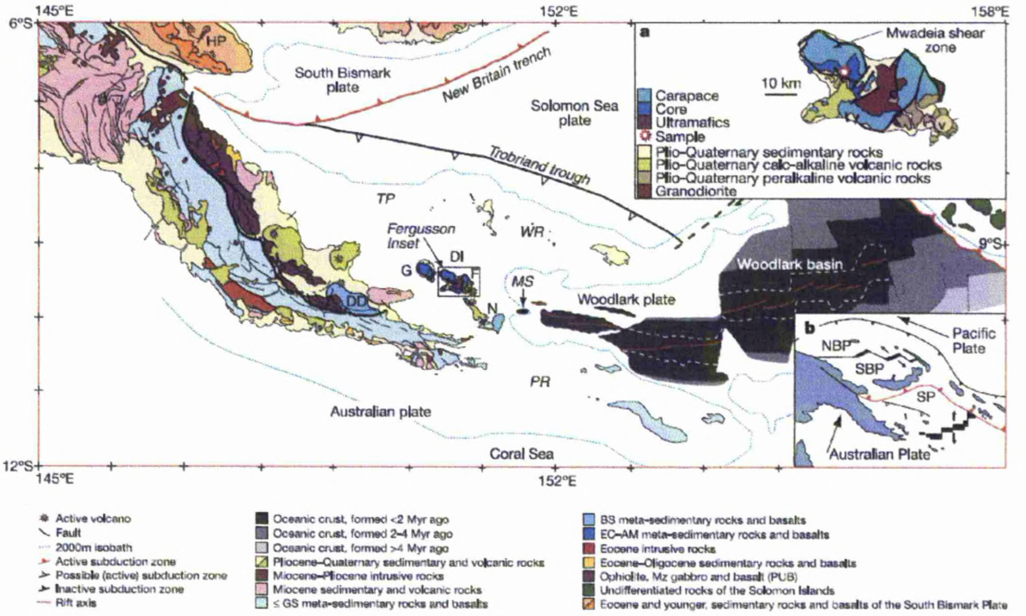

Figure 3.2 Eastern Papua New Guinea tectonic and geological map (Baldwin et al., 2004). The D'Entrecasteaux Islands exhibit rocks of eclogite facies alongside volcanic rocks.

3.4 Comparison of the eastern and western Woodlark basins

The Woodlark Basin is divided into two sub-basins by the Moresby Transform fault; these basins are a younger and smaller western basin and the larger and older eastern basin. The onset of rifting is thought to be similar for both basins at 8.4 Ma. However, the eastern basin is more distal to the Euler pole of spreading and therefore had a higher divergence rate that led to the initiation of seafloor spreading in the eastern basin at approximately 6 Ma. The seafloor spreading centre has propagated westwards at a rate of approximately 140 km per million years. In the western Woodlark the oldest oceanic crust is less than 2 million years old (Goodliffe, 1998).

The eastern and western basins have several distinct differences. The eastern basin is more distal to the Euler pole of spreading hence has a higher spreading rate of 53 to 67 mm/yr. The western basin has a spreading rate of less than 53 mm/yr and at the propagating tip of seafloor spreading the divergence rate is 36 mm/yr (Taylor et al., 1995). Seafloor spreading centres that have different divergence velocities tend to exhibit different geophysical and geological characteristics. The western basin, when compared to the eastern basin, exhibits ~500 m shallower seafloor, >30 MGals lower Bouguer anomalies, higher magnetic anomaly amplitudes, spreading centres with axial highs compared to the spreading centre valleys in the eastern basin and finally non-transform spreading centre offsets (Martinez et al., 1999). These characteristics alone would suggest that the western basin is forming at a faster rate than the eastern basin; however, it is actually diverging at a slower rate.

3.5 Surface volcanism in the western Woodlark Basin

The onset of seafloor spreading may be defined as a distinct difference between the continental crust and the formation of new crust consisting of mid ocean ridge basalt (MORB). The transition between continental crust and oceanic crust can occur over a distance ranging from a few kilometres to tens of kilometres, and this region is known as the *ocean continent transition zone* (OCT). The OCT of a rifted margin can vary greatly depending on initial continental crustal and lithospheric thickness, rate of extension, regional geotherm and composition of the crust, lithosphere and upwelling mantle.

Previous studies of the Woodlark Basin have used a combination of high-resolution bathymetry, magnetic, acoustic sounding and seismic reflection to locate the continent ocean boundary (COB). As continental lithosphere thins, the geotherm becomes elevated and during the formation of some rifted margins, there is increased regional volcanism that can produce volcanic extrusives at the surface. Using these observational techniques to decipher the COB location, cannot clearly distinguish between oceanic crust and highly thinned continental crust with basaltic extrusives on the sea floor; however, these techniques can be successfully used to locate where volcanism has reached the surface.

The magnetization of the western Woodlark Basin (figure 3.3) clearly shows areas of recent volcanism. The last magnetic reversal occurred at 780 ka (Cande and Kent, 1995), when the propagating tip of seafloor spreading was approximately 110 km east of where it currently is. Areas surrounding the current location of the spreading centres have a magnetic anomaly of +20 Am^{-1}. This is, predominantly, new oceanic crust formed since the time of the last magnetic reversal. The edges of the magnetic anomalies are mapped as the COB (Taylor et al., 1995). It is plausible that they are not the limit of oceanic crust, but part of the OCT where there are regions of highly thinned continental crust with large amounts of volcanic intrusive and extrusives.

The latest region of magnetized crust, to the north east of the Moresby Seamount, corresponds with two large ring dykes that are visible in the bathymetry. These ring dykes are thought to be the beginnings of a seafloor spreading system (Goodliffe and Taylor, 2007) located to the north west of the previous spreading centre due to a non transform spreading centre offset, however it can be argued that ring dyke extrusives are only evidence of volcanism in a region of highly thinned continental crust.

There is no recent magnetization of the crust in either the north or south Moresby Grabens suggesting that, at least since the beginning of the Brunhes chron, this region, whilst experiencing a high degree of continental thinning, has experienced no significant volcanism. Using magnetic data alone to locate the COB is further complicated by recent studies (Sibuet et al., 2007) that show that exhumed mantle at rifted continental margins can exhibit magnetization similar to that of oceanic crust. However, there is no evidence for exhumed mantle in the western Woodlark despite not being significantly volcanic; therefore, the onset of seafloor spreading should be more easily recognised.

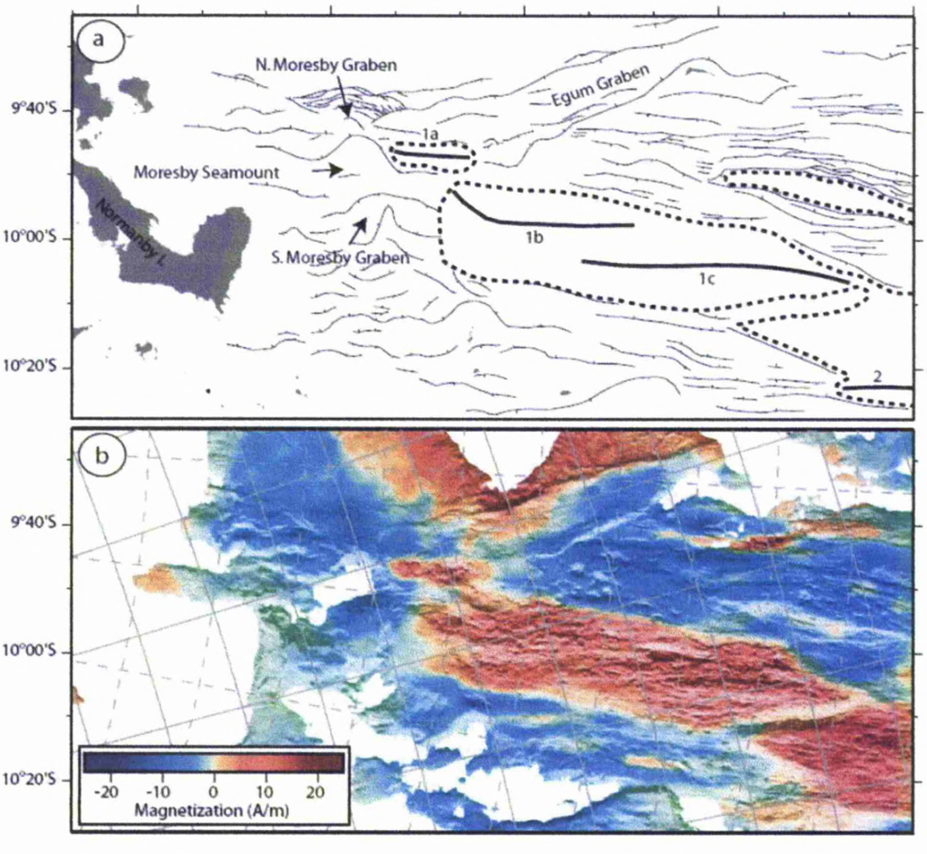

Figure 3.3 a. Major fault locations and interpreted COB location, b. Magnetization of the western Woodlark seafloor with bathymetry illuminated from the north, from Goodliffe and Taylor (2007).

3.6 Stratigraphy of the western Woodlark Basin

The region around the Moresby Seamount is covered in sediment. Since there are no major land masses, sedimentation rates have been slow, and the sediment thicknesses are thin when compared to other rifted margins. ODP Leg 180 drilled several wells around the Moresby Seamount (figure 3.4) and the lithostratigraphy has been determined from the cores of these wells. Sites 1118, 1109 and 1115 are all situated to the north of the North Moresby Fault, figure 3.4, (Taylor and Huchon, 2002).

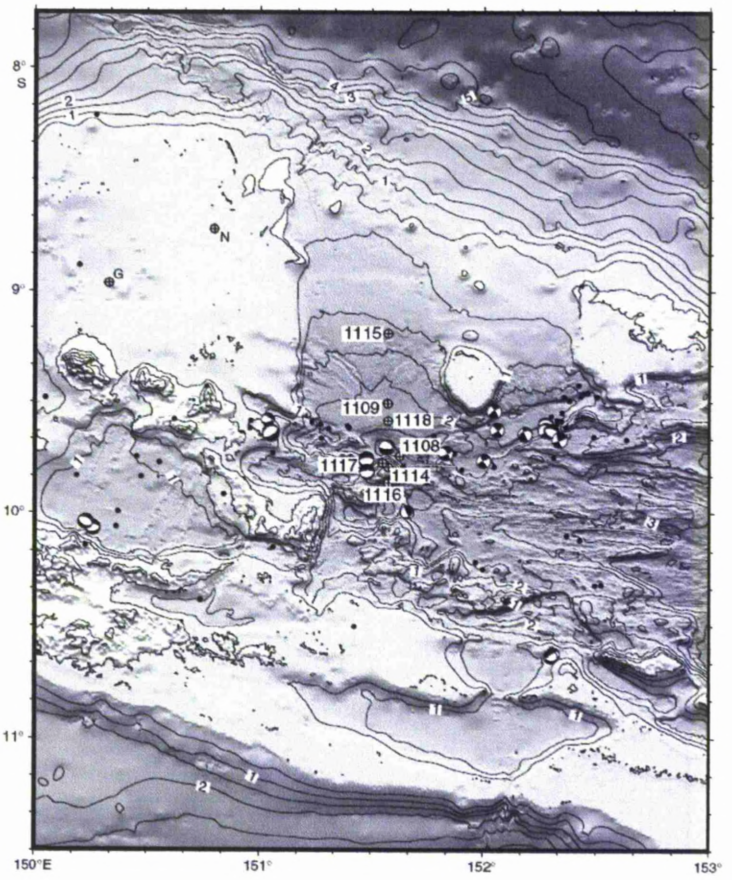

Figure 3.4 Bathymetry, illuminated from the north, of the western Woodlark Basin. Locations of ODP Leg 180 well sites are shown. Taken from Taylor and Huchon (2002).

These 3 wells show an unconformity where the sediment above is less than 8.4 Ma and the sediments below are aged 59 Ma (Monteleone et al., 2001). The cores from these wells (figure 3.5) predominately show a mixture of claystone, siltstone and a little amount of sandstone. Detailed analysis of site 1109 is shown here only since it enables a good insight into the palaeobathymetry of the region. Site 1109 is located on the Woodlark Rise, 11km north of the North Moresby Fault. The well was drilled to a depth of 375 metres below the seafloor, where it cut across the regional unconformity and entered a dolerite unit aged 59 Ma (Monteleone et al., 2001). Above the unconformity, dated at 8.4Ma, is a conglomerate sequence. Above the conglomerate, a 50m thick siltstone unit containing wood deposits and peralic coals is found. These peralic coals have been used within this study as a sea level palaeobathymetry indicator. There is a significant amount of water and sediment above the peralic coals showing that the region has undergone subsidence since they formed at sea level. Since the late Miocene, when the region was at sea level, it has gone through a lagoonal phase, shown by a limestone and calcareous sandstone unit, until it became progressively deeper, first transcending shallow marine facies until finally becoming the deep marine setting found today.

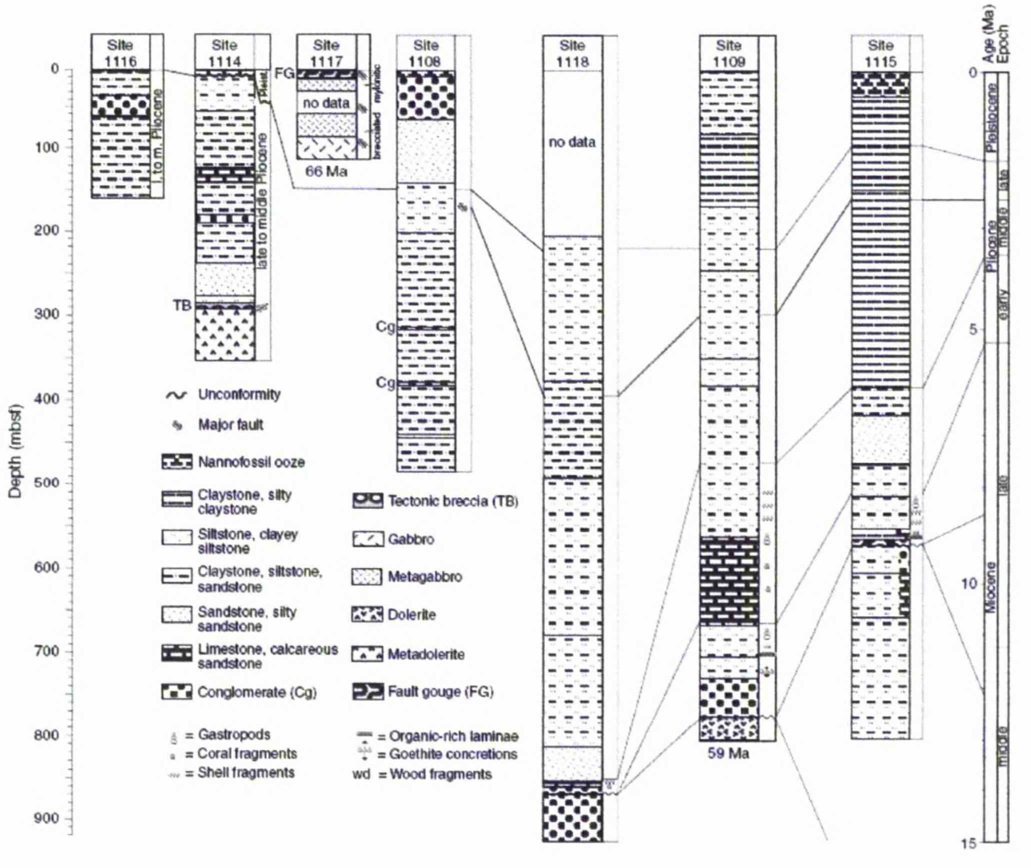

Figure 3.5 Lithostratigraphy of Leg 180 sites, taken from Taylor and Huchon (2002). Basement ages from (Monteleone et al., 2001).

3.7 Geophysical data sets for the Woodlark Basin

Bathymetry, free air gravity, magnetic, seismic and acoustic sounding data sets have all been used within this study. Satellite derived and multi-beam ship track bathymetry is publically available online (Becker et al., 2009, Smith and Sandwell, 1997) and has been used extensively within this work. Conventional ship track data for the globe is sparsely distributed and some of the data have large navigational errors, in the order of kilometres, associated with it. The Woodlark Basin has neither political nor economical restrictions on bathymetric data since no hydrocarbon system exists. Bathymetry, free air gravity and ocean floor age of the Woodlark Basin are shown in figure 3.7.

The magnetization of the seafloor has been used to determine the age of the oceanic crust and thus it is possible to map out the breakup ages of the Woodlark Basin using the most distal magnetization of the crust from the seafloor-spreading centre from which it formed. The oldest crust is approximately 6 Ma in the west, determined from magnetic reversal 3 (Cande and Kent, 1995). In the eastern Woodlark, east of the Simbo Transform, there is no seafloor-spreading centre forming new oceanic crust. The spreading centre here has likely been subducted at the San Cristobel Subduction Zone. The magnetisation of the crust has been mapped and interpreted in terms of magnetic reversals (figure 3.6) (Goodliffe, 1998).

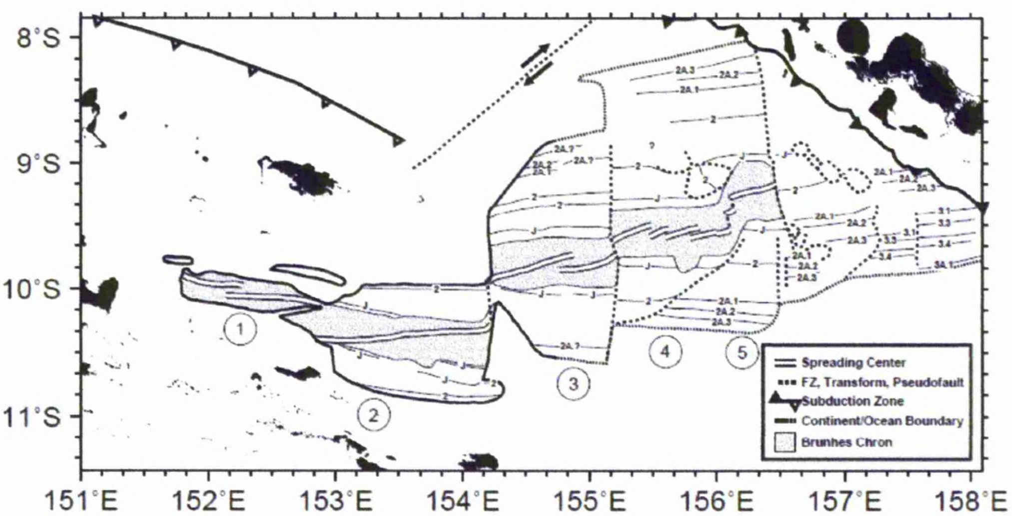

Figure 3.6 Interpreted magnetization lineations. Normal polarity magnetization chrons are labelled following the convention of Cande and Kent (1995), except for chron 1r.1 (Jaramillo) which is labelled as J. Taken from Goodliffe (1998).

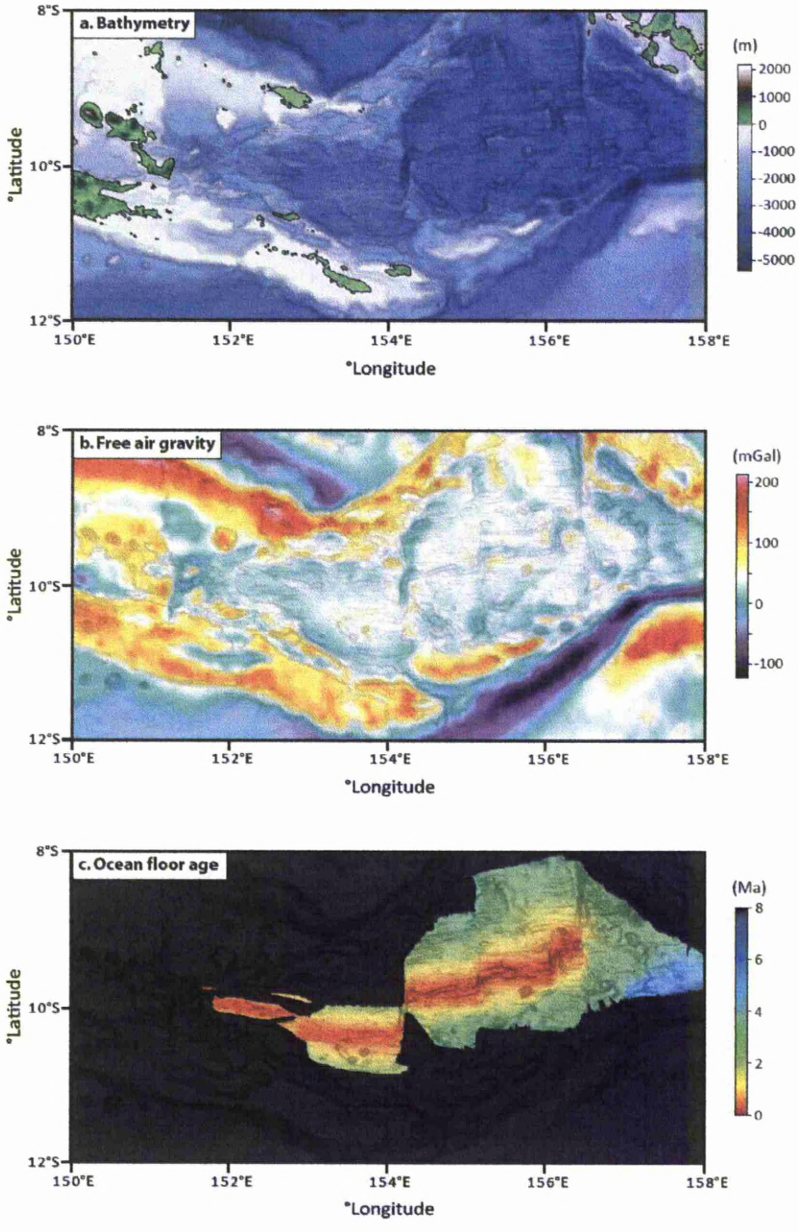

Figure 3.7 Bathymetry (Becker et al., 2009), free air gravity (Sandwell and Smith, 1997) and ocean floor (Goodliffe, 1998) data sets for the Woodlark Basin, shown with bathymetry illuminated from above.

There are numerous seismic reflection lines across the Woodlark Basin; these vary in quality but are mostly poor when compare to industry data. The seismic reflection lines (figure 3.8) represent the best-available seismic sections across the rifting to spreading transition in the Woodlark Basin. In all three profiles, MW9304_30, MW9304_50 AND MW9304_70, there is significantly less faulting observed in the northern margin than in the southern margin. The northern margin has a thicker sequence of syn-rift sediment than the southern margin; however, this could be due to poorer seismic imaging at depth in the south, making interpretation of the base of the syn-rift unreliable.

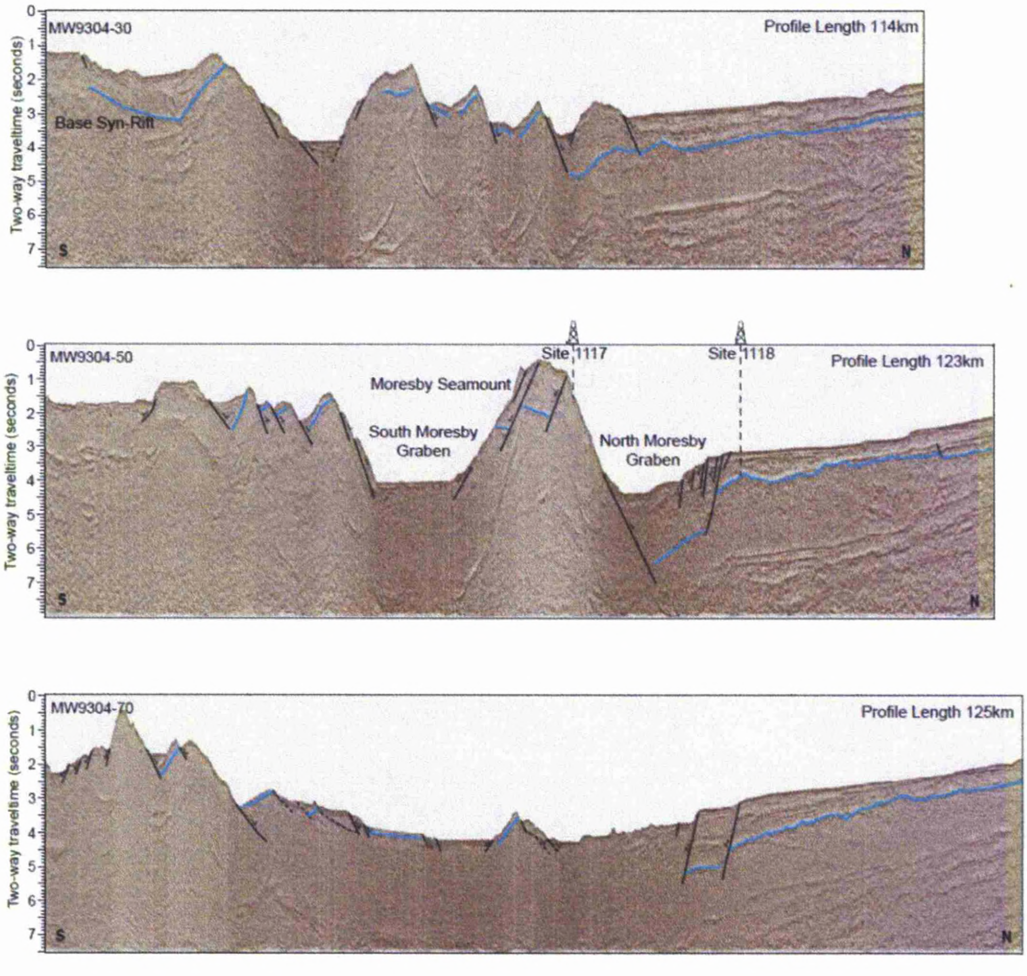

Figure 3.8 6 channel seismic cross sections of the rifting to spreading transition of the Woodlark Basin. Blue line shows base syn-rift sediments where observable. Well locations for ODP Sites 180, 117 and 118 are shown on MW9304_50. Modified from Goodliffe and Taylor (2007).

3.8 Why is regional extension occurring in the Papuan Peninsula?

The thinning of the palaeo-Papuan Peninsula and the formation of the Woodlark Basin are poorly understood. There are many arguments for possible causes for the regional divergence, yet none fully explains all the observations.

The Solomon Sea Plate is being subducted to the north of the western Woodlark Basin; it is likely that a back arc tensile stress is being applied due to subduction at the Trobriand Trough. The argument that the Woodlark Basin, being an oceanic seafloor-spreading system, is due solely to a back arc tensile stress is doubtful. East of the Moresby Transform the eastern Woodlark Basin is not in a back arc system since the Nubara Fault lies to the north and this is a transform fault, not a subduction system. Also asymmetrical spreading systems are associated with back arc basins (Deschamps and Fujiwara, 2003) and no asymmetry is observed in the Woodlark Basin.

Orogenic collapse of the Papuan Peninsula is a possible driving force of the present regional extension. The peninsula has topography due to the collision of the continental blocks with arc terrains when the subduction system closed. The Papuan Peninsula could possibly undergo orogenic collapse, causing similar extension as seen in the Basin and Range system in the USA. Unlike the Basin and Range system, the extension here leads to continental breakup and the initiation of seafloor spreading. It is not currently known how an orogenic collapse alone could lead to continental breakup. After the initial loss of topography, there would be no driving force to continue thinning the continental lithosphere.

Obduction of the ancient subducted slab is an alternative model for the Pliocene exhumation of D'Entrecasteaux Islands, and this exhumation has been occurring at rates of centimetres per year (Webb et al., 2008). It is unclear whether obduction of the slab is a driving force for the formation of the D'Entrecasteaux Islands, or if regional divergence has created the space for an obducting slab. The obduction of a slab alone would be insufficient to cause the major regional divergence of tectonic plates that is observed. It is important to note that the axis of rifting is parallel to that of the suture zone found on the Papuan Peninsula and indicating that the upper crust in this ancient collisional zone might be more conducive to thinning in a certain direction.

It is likely that a number of causes influence the regional extension and collectively lead to the thinning of the Papuan Peninsula and formation of the Woodlark Basin. This work will investigate how the continental lithosphere thins prior to continental lithospheric rupture and the initiation of seafloor spreading.

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Chapter 4

Evidence for Depth-Dependent Lithosphere Thinning at the Rifting to Spreading Transition in the Woodlark Ocean Basin, Western Pacific

This chapter is in the form of a paper to be submitted to Geophysical Journal International. Co-authors include Nick Kusznir and Andy Goodliffe. It investigates how the continental lithosphere thins leading to continental breakup and the initiation of seafloor spreading in the Woodlark Basin.

Abstract

Understanding how continental lithosphere thins at the transition from continental rifting to sea-floor spreading is key to understanding the continental breakup process. In this study, we have determined thinning at the level of the whole lithosphere, whole crust, and upper crust at the rifting to sea-floor spreading transition of the Woodlark Basin, a young, late Miocene to Present, westward propagating, ocean basin in the western Pacific. Whole lithosphere thinning (γ) has been determined from subsidence analysis, continental crustal thinning from gravity inversion, and

upper crustal stretching (β) from fault heave summation. For three 2D seismic reflection profiles, adjacent to the Moresby Seamount, whole lithosphere thinning have been determined from the water loaded subsidence using flexural back-stripping and the McKenzie continental lithosphere extension model, modified to include magmatic addition at high thinning factors. Whole lithosphere thinning factors (1−1/β) increase from an average of 0.5 to 0.8 across the Moresby Seamount eastwards towards the propagating tip. Thermal subsidence alone cannot account for the observed water loaded subsidence, and that additional initial subsidence from crustal thinning is needed. Gravity inversion, incorporating a lithosphere thermal gravity anomaly correction and sediment thickness data, have been used to determine Moho depth, continental crustal thickness and thinning at the propagating tip in the Woodlark Basin. Crustal thinning factors near the Moresby Seamount are similar to those observed for the whole lithosphere derived from subsidence analysis. Fault analysis of the three 2D profiles near the Moresby Seamount has been used to determine upper crustal extension and stretching. Conversion of upper crustal fault extension, assuming depth-uniform stretching (pure-shear), gives thinning factors between 0.1 to 0.4 that are substantially lower than thinning factors predicted for the whole lithosphere and continental crust. The measured distribution of lithosphere stretching and thinning with depth is inconsistent with depth-uniform (pure-shear) lithosphere deformation and implies depth-dependent continental lithosphere stretching and thinning prior to breakup at the propagating tip of Woodlark Basin sea-floor spreading.

4.1 Introduction

The McKenzie model, which assumes depth-uniform thinning and stretching of continental lithosphere, (McKenzie, 1978) has been successfully applied to intracontinental rift basins. However, when applied to numerous passive margins there is a discrepancy between the amount of thinning needed to accommodate the observed subsidence and the amount of thinning calculated from observed faulting in the upper crust. From studies of numerous rifted margins (Driscoll and Karner, 1998, Baxter et al., 1999, Roberts et al., 1997, Davis and Kusznir, 2004), depth-dependent lithosphere thinning has been proposed to explain the subsidence at rifted margins where there is insufficient faulting in the upper crust to account for the whole lithosphere extension needed to accommodate such subsidence. The Woodlark Basin provides the opportunity to study an area where continental lithosphere is actively being stretched and thinned leading to continental breakup and the onset of seafloor spreading. With limited sediment thickness and good palaeobathymetry indicators, the Woodlark Basin provides an ideal natural laboratory in which to observe the mechanism of how the lithosphere thins prior to breakup.

4.1.1 Regional Setting

The Woodlark Basin (figure 4.1) is a young ocean basin located east of the Papuan Peninsula in the southwest Pacific. Rifting began in the late Miocene and the oldest seafloor is approximately 6 Ma, identified from seafloor magnetization (chron 3A.1), in the eastern Woodlark (Taylor et al., 1995, Cande and Kent, 1995). The Moresby

Transform, a large transform fault, divides the Woodlark Basin into two sub basins, an eastern basin and a western basin. The seafloor spreading system is propagating westwards, across the Moresby Transform, and with less than 2 million years of seafloor spreading the western basin is younger than the eastern basin where seafloor spreading started approximately 6 million years ago. The spreading ridge has been propagating westwards into continental lithosphere at a rate of 140 km per million years (Taylor and Exon, 1987) and the propagating tip is currently located 10 km northeast of the Moresby Seamount. Two distinct types of seafloor spreading initiation have been identified in the Woodlark Basin; propagation of a spreading tip into rifting continent and simultaneous nucleation of seafloor spreading along >100 km rift segments (Taylor et al., 1999).

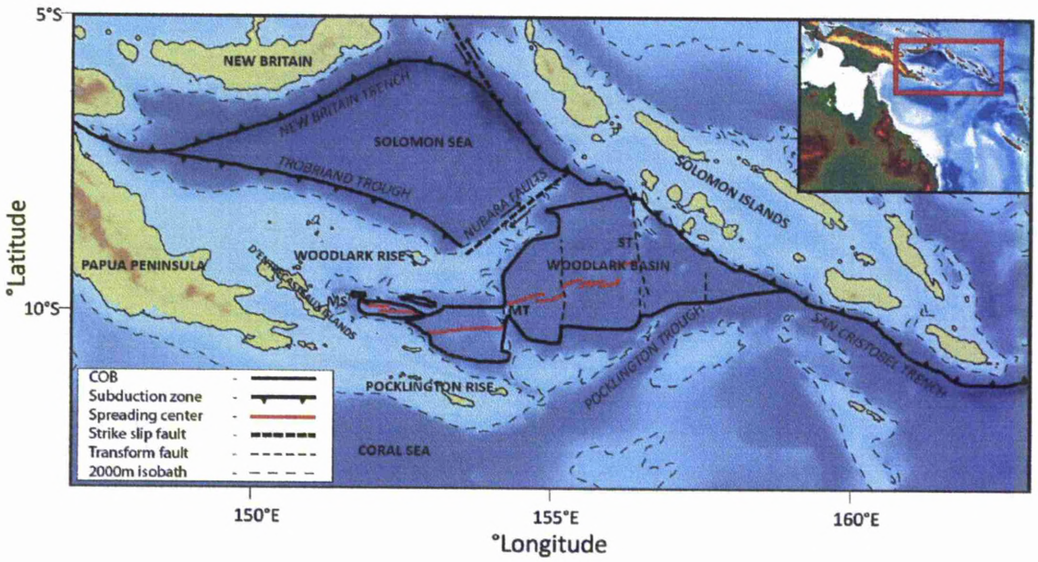

Figure 4.1 Regional map of the Woodlark Basin showing surrounding subduction zones, transform faults and active seafloor spreading centres. MS (Moresby Seamount), MT (Moresby Transform) & COB (continent-ocean boundary (Taylor et al., 1995)).

This region of the southwest Pacific has a complicated geologic history. The prerift evolution of the conjugate margins of the Woodlark Basin involve subduction and arc construction followed by collision and suturing of continental and arc regimes (Honza et al., 1987, Taylor and Huchon, 2002). North of the Woodlark Rise is the Trobriand Trough, a region where the Solomon Sea Plate is being subducted under the Woodlark Plate. This process potentially adds a back arc stress to the youngest part of the basin. It is likely that the formation of the Woodlark Basin is more complex than a typical back arc system, as seafloor spreading originated outside of the back arc domain in the eastern Woodlark Basin. Orogenic collapse of the Papuan Peninsula could possibly explain the continental rifting; however how an orogenic collapse could lead to seafloor spreading is currently unknown. It is more plausible that a combination of several events and processes has led to this region undergoing rifting and breakup, and consequently a simple model of continental breakup is unlikely to be sufficient to account for all the observations. The Woodlark Basin provides an ideal natural laboratory in which to observe, determine and test how the crust and lithosphere thins prior to continental breakup and subsequent seafloor spreading initiation.

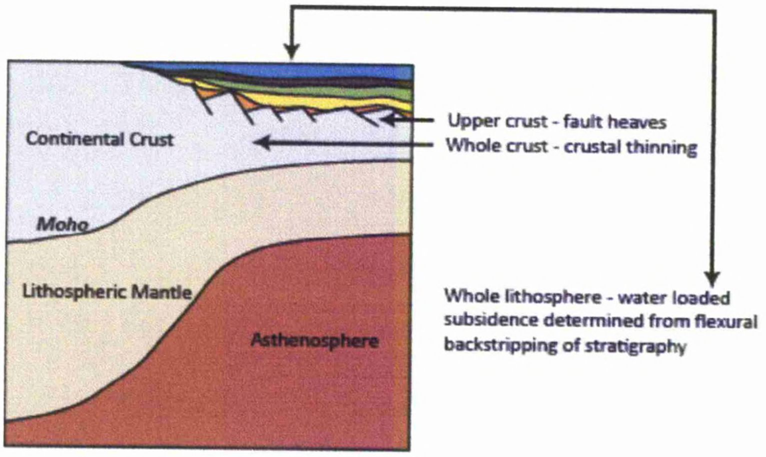

Figure 4.2 Stretching and thinning of continental lithosphere at rifted margins is measured for the upper crust from fault heaves, the whole crust from crustal thinning and the whole lithosphere from subsidence analysis. Modified from Kusznir and Karner (2007).

Depth-dependent lithosphere thinning has been observed during the formation of numerous rifted margins (Davis and Kusznir, 2004, Kusznir and Karner, 2007). The purpose of this study is to measure lithosphere thinning at the levels of whole lithosphere, the whole crust and the upper crust assuming depth-uniform thinning for each layer. Thinning is measured in three ways: 1) for the whole lithosphere from water loaded subsidence using flexural back-stripped stratigraphy (figure. 4.2); 2) whole crustal thinning from gravity inversion; and 3) extension of the upper crust estimated from fault heave summation from three seismic reflection lines perpendicular to the direction of seafloor spreading propagation (figure. 4.3).

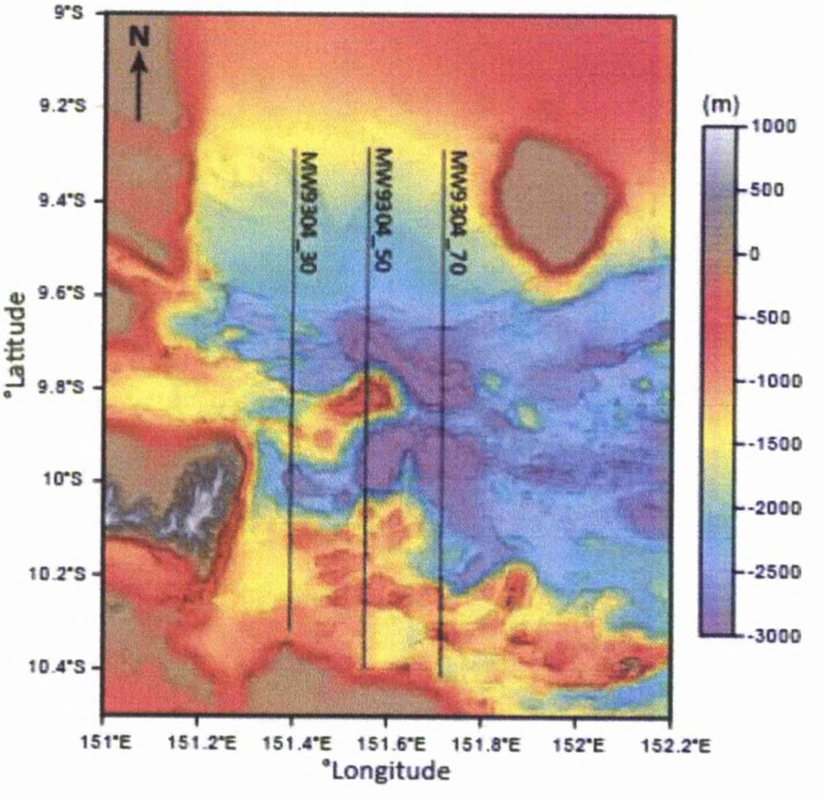

Figure 4.3 Illuminated bathymetry map of the rifting to spreading transition zone in the Woodlark Basin. Locations of the multi-channel seismic reflection lines used within the modelling are shown.

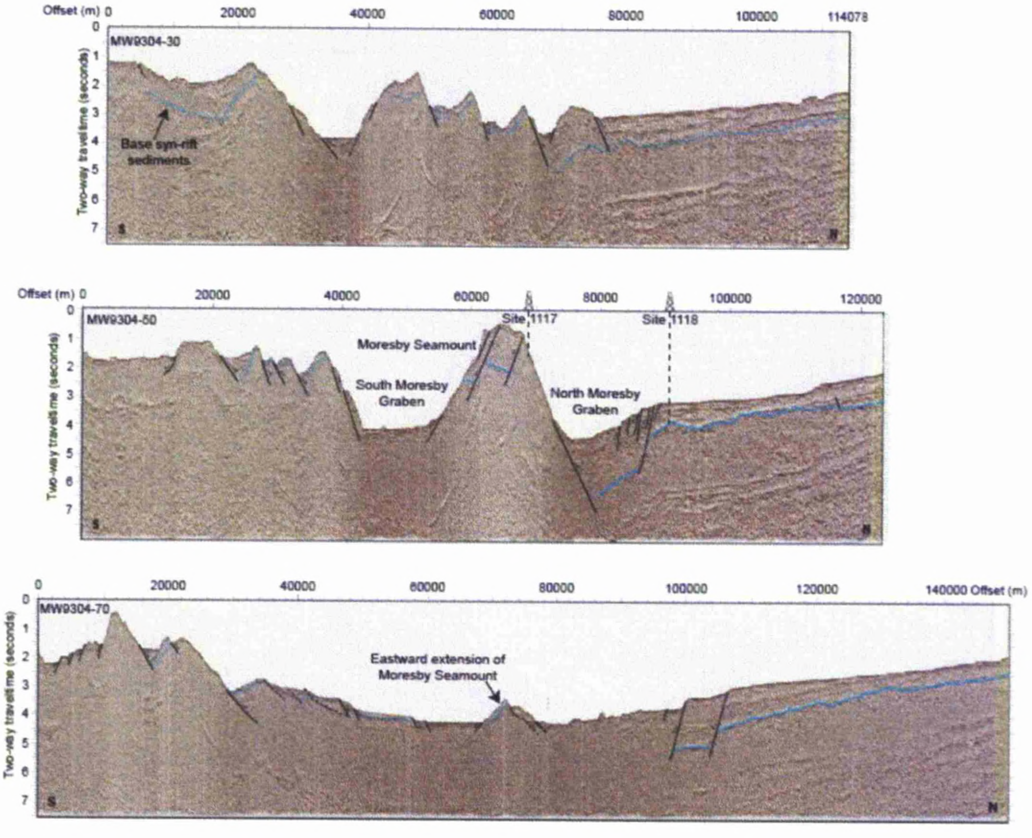

Figure 4.4 Six-channel reflection seismic lines MW9304-30 (top), MW9304-50 (middle), and MW9304-70 (bottom) across the continental margins in the vicinity of the rifting to spreading transition. The location of the SCS lines is shown in Figure 2. There is significantly more normal faulting in the southern margin than in the northern margin visible in all 3 lines, whereas thicker sediment sequences are visible within the seismic in the north.

4.2 Measuring Continental Lithosphere Thinning Prior to Breakup and Seafloor Spreading in the Woodlark Basin

4.2.1 Whole lithosphere thinning from subsidence analysis

We determine the whole lithosphere thinning factor, γ, using a modified McKenzie pure shear model that incorporates magmatic addition (McKenzie, 1978) (figure 4.5).

Observed water loaded subsidence, $S_i + S_t$, is used as a target within the McKenzie model to determine the whole lithosphere thinning factor; where S_i is the subsidence due to the initial lithosphere thinning and S_t is the subsidence due to the post rift cooling of the thinned lithosphere. Dynamic topography, due to mantle dynamics, can also be a cause of regional subsidence or uplift; however, this is ignored in this study.

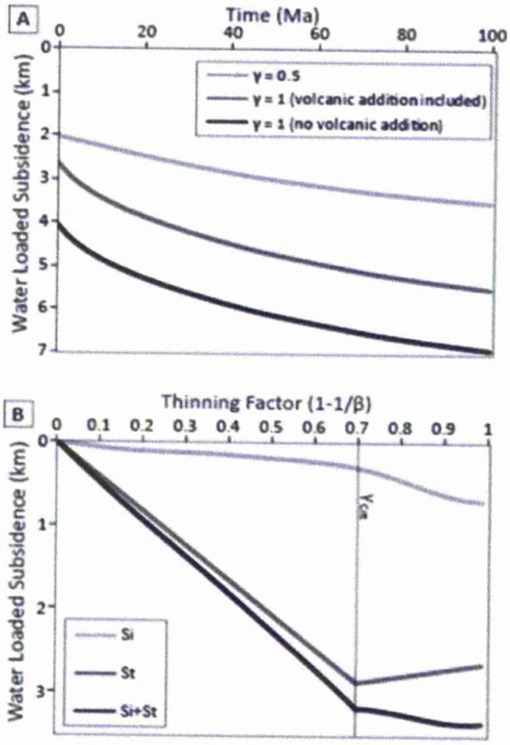

Figure 4.5 A) The relationship between time and water loaded subsidence within modified McKenzie pure shear model showing the difference between the inclusion of magmatic addition after a critical thinning factor (0.7γ) and a model where magmatic addition is ignored. Initial crustal thickness of both models is 35 km; maximum magmatic addition is 7 km. B) Water loaded subsidence versus thinning factors showing the amount of subsidence, determine from a modified McKenzie model to include magmatic addition, for initial subsidence (S_i), thermal subsidence (S_t) and both combined ($S_i + S_t$) using parameters suitable for the Woodlark Basin. Initial crustal thickness = 40 km; $\gamma_{crit} = 0.7$; maximum magmatic addition = 7 km and 8.4 Ma rift age.

The determination of the water-loaded subsidence requires a palaeobathymetry indicator within the stratigraphy at the time of the onset of rifting. Since the crust has a flexural isostatic response to the loading of these sediments upon it (Watts and Ryan, 1976), flexural back-stripping is used. Flexural back-stripping of sediments allows for the isostatic removal of sediment loads to determine sediment corrected bathymetry. A palaeobathymetry indicator is required to constrain the original depositional bathymetry. Isostatically removing the sediments above, and decompacting the sediments below the palaeobathymetry indicator, should restore it to its original bathymetry if no thinning of the lithosphere has occurred. However, where the lithosphere has thinned, the palaeobathymetry indicator does not restore to its original depositional bathymetry. The difference between where it isostatically restores to and the original depositional bathymetry is the water loaded subsidence (WLS). This water loaded subsidence is used as the target to determine whole lithosphere thinning factors within a modified McKenzie pure shear model that includes a magmatic addition prediction (McKenzie, 1978).

In the study area, wood deposits in the form of peralic coals have been found in cores from wells 1109 and 1115 from ODP Leg 180 (Taylor and Huchon, 2002). These peralic coals are situated less than 50m above an erosional unconformity that has been dated at least 8.4Ma (Taylor and Huchon, 2002). Peralic coals are formed at sea level in coastal plain environments; therefore, it can be assumed that approximately 8.4 million years ago the region around the Moresby Seamount was approximately at sea level. Flexurally back-stripping the sediments to the coal seams and decompacting the sediment below has determined the water loaded subsidence profiles shown in figure 4.6 for the region ahead of the propagating tip of seafloor spreading. For regions where there was no well data, the base of the syn-rift sediment is assumed to have been deposited at sea level. Global eustasy for sea level change (Haq et al., 1987, Harland et al., 1989) have been included when determining the water loaded subsidence.

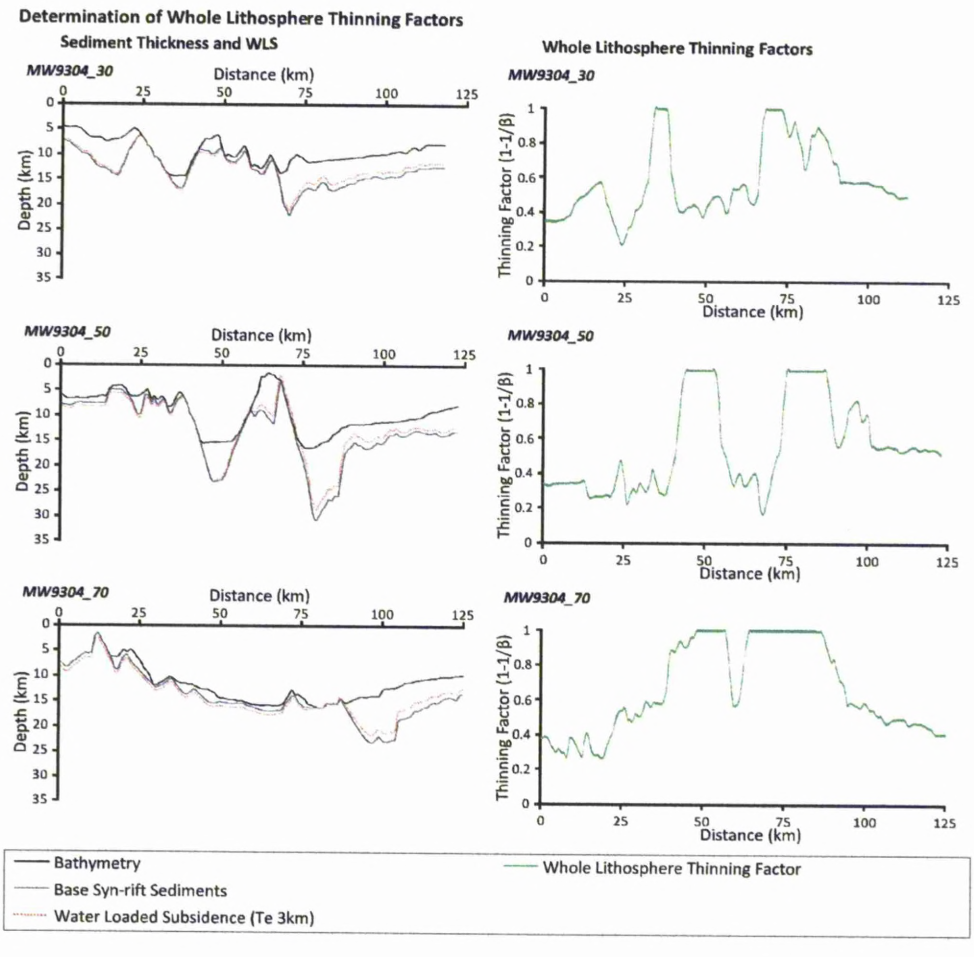

Figure 4.6 Water loaded subsidence (WLS) profiles for the three seismic reflection lines. The sediments have subsequently been back-stripped to the peralic coals immediately above the erosional unconformity, basement is assumed to be immediately beneath the unconformity hence there is no decompaction of lower sediments. $T_e = 3$ km. Whole lithosphere thinning factors determined from water loaded subsidence are shown for each profile. Thinning factors range between 0.3 and 1.

The McKenzie model predicts subsidence due to stretching and thinning of the continental lithosphere. In regions of greater thinning magma is generated by decompression melting during stretching and thinning. This melt adds to the thickness of the observed crust and modifies subsidence isostatically. We have modified the McKenzie model to incorporate magmatic addition, parameterised by decompression melting predictions (McKenzie and Bickle, 1988, White and McKenzie, 1989). Magmatic addition is included within the model after the lithosphere has thinned more than the critical thinning factor (γ_{crit}) for the onset of volcanism. Including magmatic addition assumes that a proportion of the thickness of the crust is magmatic and predicts higher thinning factors than if magmatic addition is ignored and the present crustal thickness is assumed solely original material. Magmatic addition affects the amount of subsidence that can be predicted for regions that have experienced thinning greater than the critical thinning factor; for a McKenzie model including magmatic addition the total water loaded subsidence when complete breakup of the continental lithosphere has occurred, where $\gamma = 1$, is 1.5 km less than when no magmatic addition assumed as shown in figure 4.5a. For a model applicable for the Woodlark Basin, a critical thinning factor (γ_{crit}) of 0.7 is applied and a maximum magmatic addition of 7 km is used since previous isostatic studies of oceanic crustal thicknesses predict 7 km as a typical thickness of the oceanic crust (Martinez et al., 1999). Until the critical thinning factor is reached within the model, it is assumed that there is no magmatic addition. After this critical thinning factor, magmatic addition starts to be calculated as a component of the crustal thickness, decreasing the thickness of the crust that is residual continental crust, until breakup has occurred and no continental crust remains when $\gamma = 1$.

Figure 4.6 shows whole lithosphere thinning factors (γ) for all 3 seismic lines range between 0.3 and 1 determined from subsidence analysis. At the ends of the profiles, where it is expected that the continental lithosphere is thickest, the thinning factors are lowest. The highest thinning factors are found, as expected, towards the centre of the profiles, near the rift axis. Thinning factors of 1 in profiles MW9304_30 and MW9304_50 to the north and south of the Moresby Seamount imply that the continental lithosphere has thinned to zero thickness and has ruptured. Though very close to breakup, the magnetic data (Goodliffe and Taylor, 2007) imply that this is not the case. More likely, this is a by-product of fault structural topography that the methodology used cannot resolve. Whole lithosphere thinning factors, from subsidence analysis, for MW9304_70 show a 40 km wide section where thinning factors are greater than 0.95. Coupled with seismic and magnetic evidence (Goodliffe, 1998, Goodliffe and Taylor, 2007) it is likely that this profile transects the youngest oceanic regime of the Woodlark Basin and that rupture of the continental lithosphere has occurred.

4.2.2 Whole crustal thinning from gravity inversion

To determine whole crustal thinning factors, an initial uniform crustal thickness (tc_0) is assumed and then divided by the present crustal thickness (tc_{now}) minus the thickness of the magmatic addition (tc_{mag}) in order to determine the stretching factor (β) that the crust has experienced.

$$\beta = tc_0/(tc_{now} - tc_{mag})$$

Crustal thickness has been determined for a gravity inversion, incorporating a lithospheric thermal gravity anomaly correction (Greenhalgh and Kusznir, 2007, Chappell and Kusznir, 2008), has been used to determine Moho topography using the scheme of Parker (1972).

$$\Delta g_{mra} = \Delta g_{fag} + \Delta g_b + \Delta g_t$$

where

$$F[\Delta g_{mra}] = 2\pi G \Delta \rho e^{-|k|z_0} \sum_{n=1}^{\infty} \frac{(|k|)^{n-1}}{n!} F[\Delta r^n]$$

Δg_{fag} is the observed free air gravity anomaly, Δg_b is the gravity anomaly from bathymetry, Δg_t is the lithosphere thermal gravity anomaly correction and Δg_s is the gravity anomaly from sediment thickness; Z_0 is the mean Moho depth, G = 6.67 × 10^{-11} $m^3kg^{-1}s^{-2}$, $\Delta \rho = \rho_m - \rho_c$, F denotes a Fourier transform and k is the wave number. Δg_{mra} was filtered before the inversion to remove high frequency components that come from crustal and sediment heterogeneities, using a Butterworth low-pass filter with a cut-off wavelength of 100 km. A Butterworth low-pass filter of 75 km was used to determine residual continental crust thickness. The assumption is made that

Δg_{mra} is caused solely by variations in Moho depth. Densities for seawater ρ_w, crust ρ_c and mantle ρ_m used in the inversion are 1039 kgm^{-3}, 2850 kgm^{-3} and 3300 kgm^{-3} respectively. Sediment density assumes normal compaction (Sclater and Christie, 1980).

This method has been applied to the Woodlark Basin in order to determine Moho topography and crustal thickness. Figure 4.7 shows crustal thickness for the rifting to spreading transition, around the Moresby Seamount, with and without sediments included within the gravity inversion. Including sediments within the inversion reduces the crustal thickness and thus predicts higher whole crustal thinning factors; where sediments are not used, the crustal thickness determined is an upper bound (figure 4.8). Magmatic addition is used within the gravity inversion using a maximum magmatic addition of 7km when γ is equal to 1; a critical thinning factor (γ_{crit}) of 0.7 is applied and this is the minimum amount of thinning required for the onset of volcanism to occur within the crust. Ignoring magmatic addition within the gravity inversion gives results where the Moho depth is overestimated and hence predicted whole crustal thinning factors are underestimated (figure 4.9).

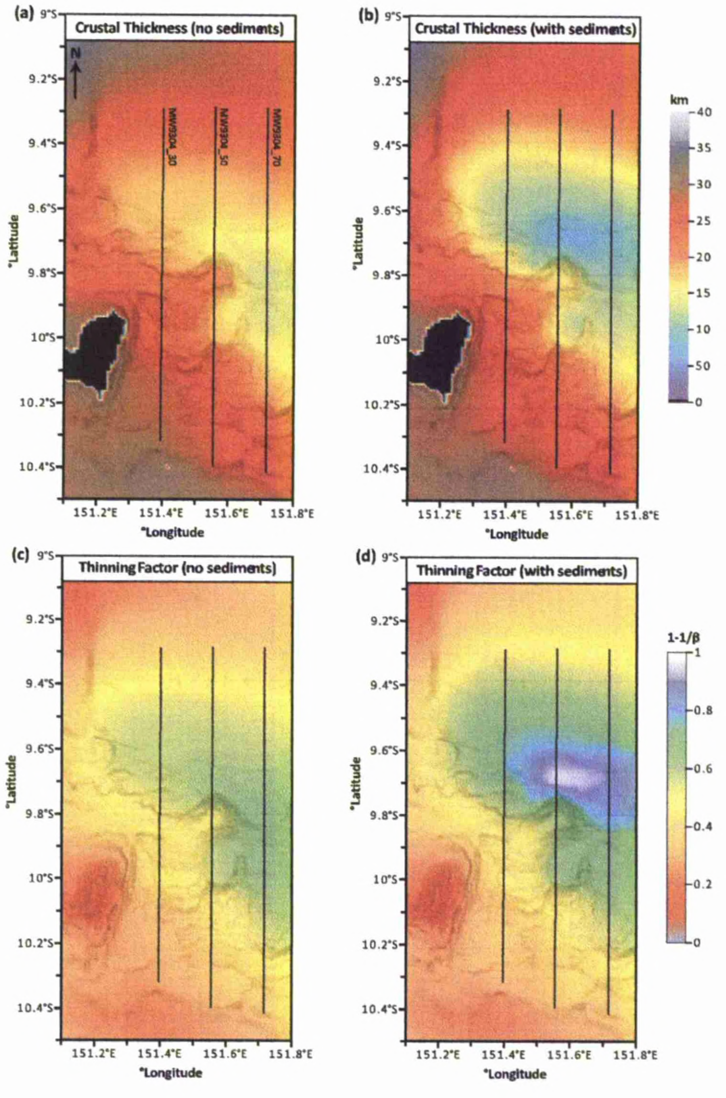

Figure 4.7 Crustal thickness and whole crustal thinning factors predicted by the gravity inversion (a and c) without sediment thickness included within the inversion, b & d are with sediments included within the gravity inversion. All gravity inversions assume a 0 Ma breakup age. Including sediments within the gravity inversion significantly predicts thinner crust and hence higher thinning factors. Sediment thickness derived from depth conversion of the seismic reflection lines.

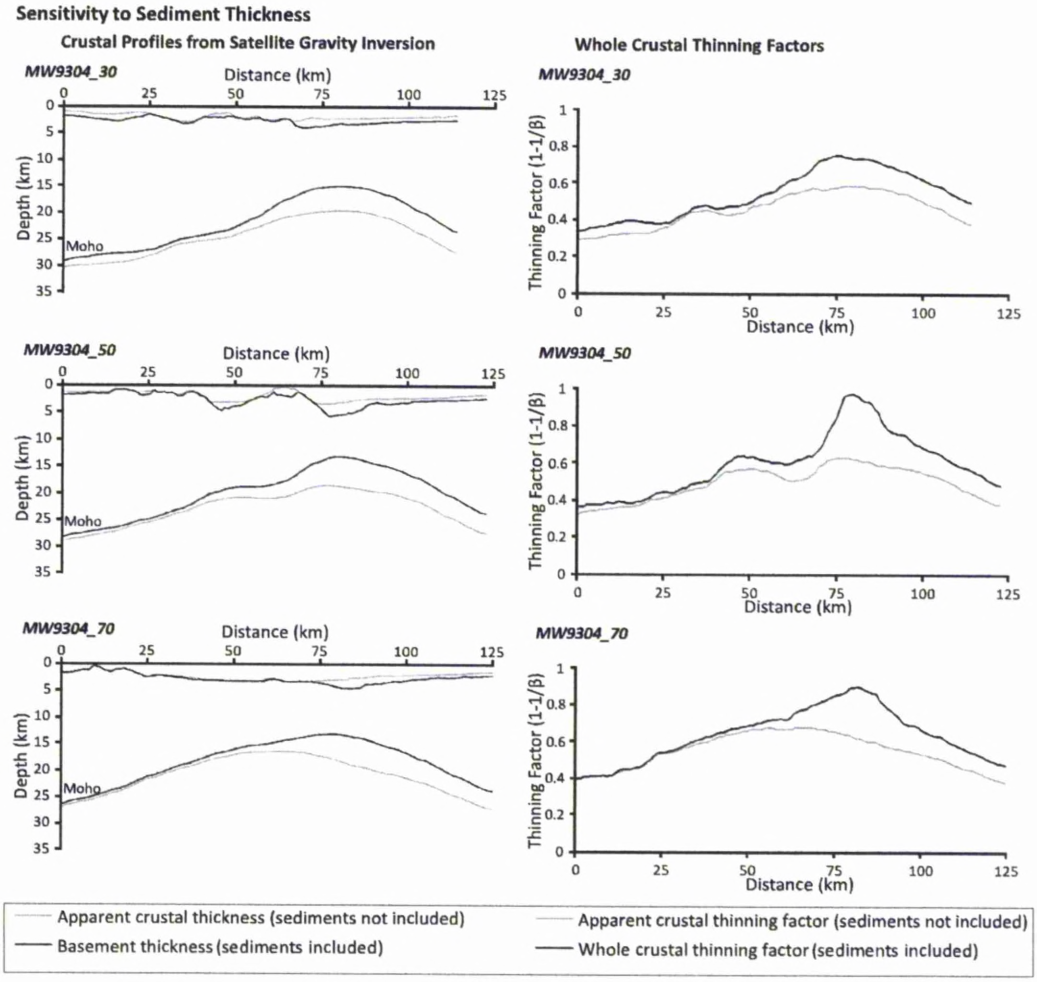

Figure 4.8 Crustal cross sections and whole crustal thinning factors predicted from gravity inversion showing sensitivity to sediment thickness. Including sediments gives thinner continental crust and higher thinning factors.

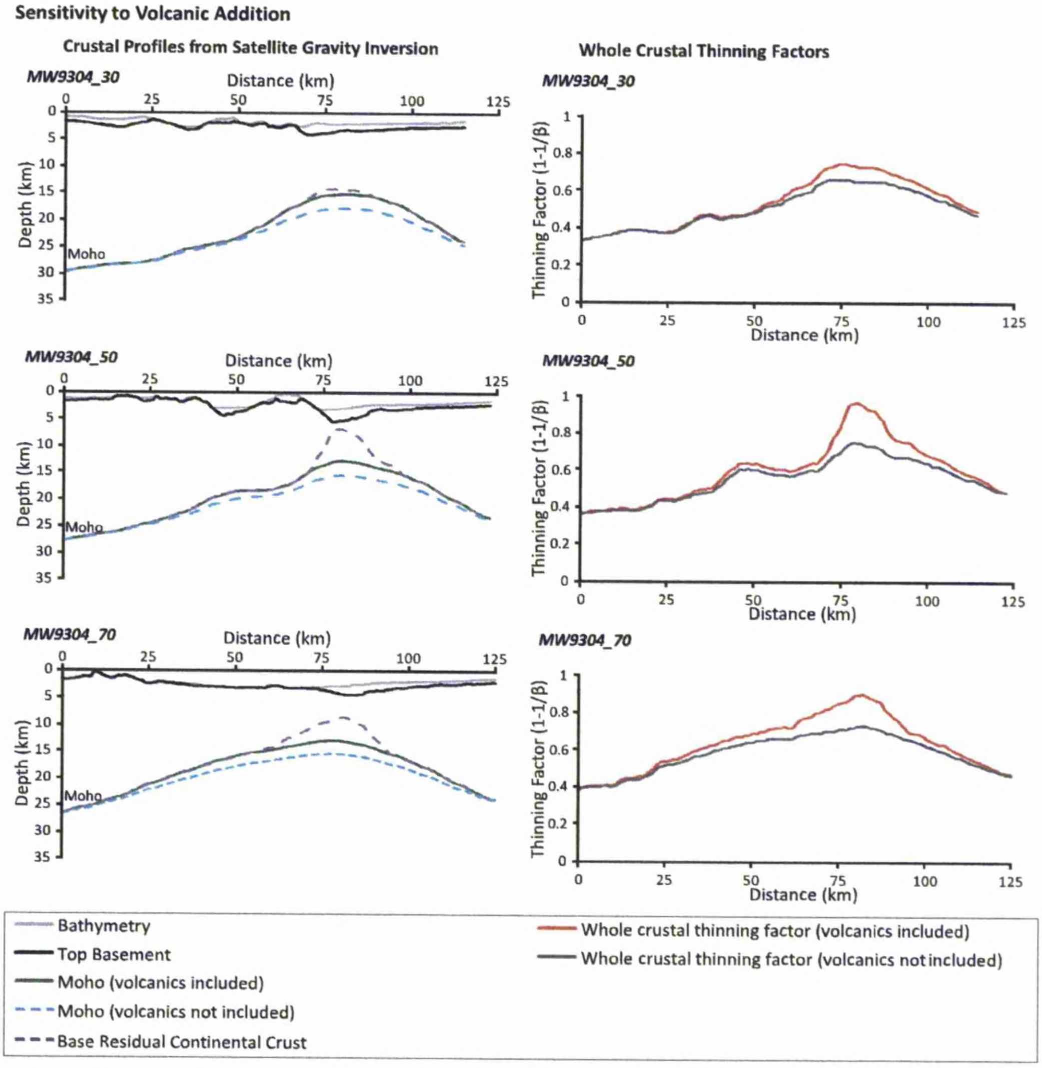

Figure 4.9 Crustal cross sections and whole crustal thinning factors from gravity inversion the sensitivity to magmatic addition. Including magmatic addition predicts thinner continental crust and hence thinning factors determined are higher. Crustal cross sections show areas where the onset of magmatic addition is predicted to occur.

Figures 4.7 and 4.8 show 2 sets of results for whole crustal thinning factors, calculated with and without sediment thickness data included within the gravity inversion. Sediment thickness has been derived from seismic lines in the vicinity that has subsequently been depth converted using a velocity function derived from well data from ODP Leg 180 (Goodliffe et al., 2002). There is six channel seismic reflection coverage of the region; however it is of poor quality. Several areas, especially the Southern Moresby Graben, are poorly imaged, possibly due to mafic flows and here sediment thickness is a poor estimation. The sediment coverage of the area varies greatly; a thickness of 3 km is observed in the North Moresby Graben. To the north of the Moresby Seamount the sediments vary between 500 and 1250 m thickness whereas to the south there is an apparent lack of sediment. It is possible that top basement there is not being imaged in the seismic reflection data; however, it is assumed that the sediment thickness here is predominantly thinner than that to the north of the Moresby Seamount. The effect of including sediment thickness within the gravity inversion is evident in the whole crustal thinning factor results; where there is a large accumulation of sediments in the North Moresby Graben the difference between the thinning factors, calculated with and without sediment thickness data, is greatest. Whole crustal thinning factors for all 3 profiles range between 0.3 and 0.95 suggesting that in some locations the crust has been thinned to near breakup.

A low pass Butterworth filter of 100 km has been applied to remove high frequency gravity signals from heterogeneities in the crust and sediment. The thinning factors derived by gravity inversion for the whole crust are similar to those determined for the whole lithosphere using the modified McKenzie model; lower amounts of thinning on the profile ends increasing towards the rift axis.

4.2.3 Upper crustal thinning from fault heave summation

The upper 10 to 15 km of the crust is assumed to be brittle and when stretched (Jackson and White, 1989), the extension is accommodated by the brittle failure of this layer and the formation of faults. Upper crustal extension has been measured from fault heave summations. The amount of lateral displacement caused by normal faulting is measured for each fault and then the sum of individual displacements give the total amount of brittle extension across a profile. Using the value of upper crustal extension, we determine an upper crustal stretching factor using:

$$\beta_{uc} = \frac{L_{pr}}{L_{pr} - X}$$

Where β_{uc} is the upper crustal stretching factor, L_{pr} is the present day length of the profile and X is the total extension measured across all the faults, results are shown in table 1. The faults are not evenly distributed across all 3 profiles, the southern margin has significantly more faults than the northern margin, and this is evident in figure 4.

The stretching factors have also been estimated from faults in a forward syn-rift model. To determine a stretching factor profile of upper crustal thinning rather than an average value, individual fault extensions and geometries were forward modelled to produce a stretching factor profile. These profiles, which have been converted into thinning factors assuming depth-uniform thinning of the upper crust and are shown in figure 10 and table 1. The thinning factors show where the extension in the upper crust is occurring along the profiles. Within the forward modelling of the faulting, the upper crust is assumed to be brittle, beneath each fault, in the ductile regime, the extension is distributed across a pure shear width of 100 km.

This method of determining upper crustal thinning factors assumes that all extension is measured from the faults visible in the seismic data; however this is unlikely and results shown are minimum upper crustal thinning factors. Sub-seismic resolution faulting, where faults are too small to be observed in seismic data (Walsh et al., 1991), could be a cause of an under estimation of thinning factors. It has been argued that not all of the extension that the upper crust has undergone can be determined by summing fault heaves. It is possible that there has been multi-phase faulting and that earlier faults cannot be observed in the seismic (Reston, 2009).

Profile	Extension (km)	Initial Length (km)	Final Length (km)	Stretching Factor (β)	Thinning Factor (γ)
MW9304_30	15.08	102.92	118	1.15	0.13
MW9304_50	37.7	85.3	123	1.44	0.31
MW9304_70	37.67	87.33	125	1.43	0.30

Table 1 The determination of upper crustal stretching and thinning factors from fault heave summation.

For all 3 profiles (figure 4.10) upper crustal thinning, γ, is limited to a maximum of 0.45 and at the ends of the profiles the thinning is substantially less. The majority of the normal faulting associated with the rifting of the continental crust ahead of the propagating tip occurs in the southern margin. There are relatively few faults in the northern margin and this is reflected in the thinning factor distribution determined for the upper crust. MW9304_30 is furthest away from the propagating tip of seafloor spreading and therefore is at an earlier stage of breakup than the profiles further east. This profile shows the lowest thinning factors. Further east, in profiles MW9304_50 and MW9304_70, upper crustal faulting has increased, suggesting that immediately prior to break up the plate divergence is accommodated by faulting and magmatic dykes before seafloor spreading initiates.

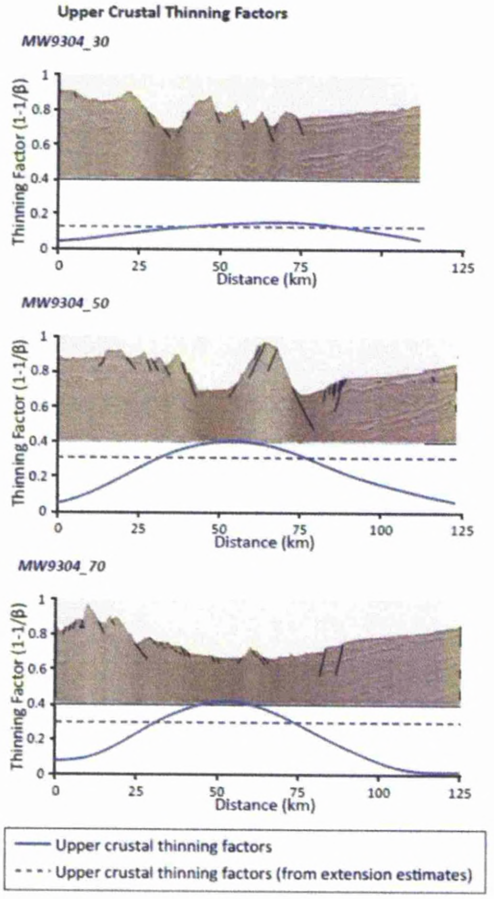

Figure 4.10 Fault heave summations of seismically visible faults to determine upper crustal thinning factors. Two sets of results are presented, an average value of upper crustal thinning determined from total extension observed across the seismic profiles and a profile determined from individual fault heave estimates.

4.3 Discussion & Summary

4.3.1 Depth-dependent thinning of continental lithosphere prior to continental lithospheric rupture

The nature of continental lithospheric thinning during the formation of rifted continental margins is poorly understood. The Woodlark Basin allows for observations of continental lithospheric thinning prior to continental breakup and the initiation of seafloor spreading. Whole lithosphere, whole crustal and upper crustal thinning factors have been determined independently of one another, assuming depth-uniform thinning, for three profiles in the rifting to spreading transition. Figure 4.11 shows whole lithosphere, whole crustal thinning and upper crustal thinning factors compared against each other. Whole lithosphere and whole crustal thinning factors both greatly exceed the thinning factors determined for the upper crust. Although the upper crustal thinning factors determined are a minimum prediction, they exhibit substantially less extension than that of the whole lithosphere and whole crust and it is unlikely that sub-seismic resolution faulting could account for such a difference. This work shows evidence for depth-dependent thinning in the continental lithosphere prior to breakup and the onset of seafloor spreading during the formation of the Woodlark Basin.

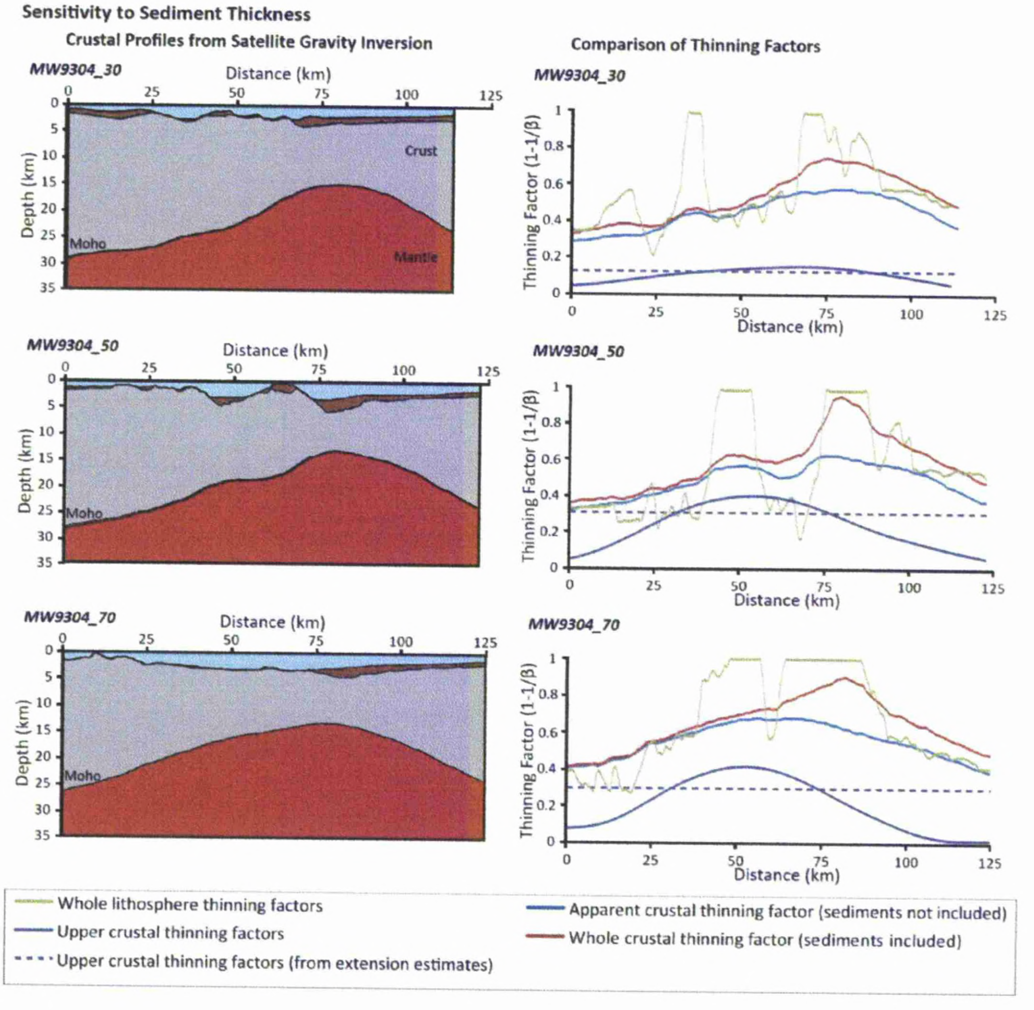

Figure 4.11 Crustal cross sections, determined from gravity inversion, are shown for each seismic line. Comparison of whole lithosphere, whole crust and upper crustal thinning factors assuming depth-uniform stretching and thinning. Whole lithosphere and whole crustal thinning factors are similar and both greatly exceed the thinning factors determined for the upper crust; hence depth-dependent lithosphere thinning is occurring.

The upper crust exhibits thinning asymmetry across the rift axis, with greater thinning of the Pocklington Rise than in the Woodlark Rise to the north. This thinning is attributed to a series of normal faults in the southern margin that are observed in the seismic reflection data. Whole lithosphere and whole crustal thinning factors do not exhibit the asymmetry of thinning that is observed in the upper crust. It is plausible that an asymmetrical process is occurring in the upper crust, possibly due to the reactivation of pre-existing faults formed during the continental collision that formed the Papuan Peninsula.

4.3.2 Upper Crustal Plate Divergence Velocity Discrepancy

The onset of rifting of the Papuan Peninsula, determined from numerous studies (Taylor et al., 1999, Tjhin, 1976, Francis et al., 1987, Pinchin and Bembrick, 1985) is regarded as synchronous over an east-west region of 1000 km encompassing the Woodlark Basin. From Euler pole estimates of rifting rates at the propagating tip derived from seafloor spreading velocities in the study area, 220 km of lithospheric extension is expected since 6 Ma (Taylor et al., 1999). The current N-S width of continental crust in the region around the Moresby Seamount is approximately 320 km, hence the initial width of the pre rifted Papuan Peninsula, if a stretching factor (β) of 2 had been applied to the upper crust, would be 110 km. Maximum values of observed brittle extension from seismically visible faults are 16, 38 and 38 km for SCS lines MW9304_30, MW9304_50 and MW9304_70 respectively; however the entire width of the continent plateau has not been imaged by seismic reflection. Other studies (Kington and Goodliffe, 2008, Fang, 2000) have calculated a maximum of 105 km of extension, including a correction for sub-seismic faults, in the upper crust across the entire continental width, significantly less than what would be expected from Euler pole motions.

If the upper crust is thinning, to a lesser extent than the lower crust, this implies that there would have to be a zone within the lithosphere where this difference in extension rates would need to be accommodated. This could be in the form of a weak decoupled horizon; at present, there is no evidence for such a structure in the Woodlark Basin.

The extension that is expected in the upper crust in the region prior to continental breakup, from a Euler pole model derived from seafloor spreading velocities, is not (Kington and Goodliffe, 2008). It is plausible that the divergence velocity of the upper crust and the rest of the lithosphere, prior to breakup, is different and that it is only after continental lithospheric rupture that they diverge at the same rate. Only after continental breakup and the initiation of seafloor spreading is true plate motion observed, at the surface of the upper crust, in the Woodlark Basin. It is currently unknown why the pre-breakup upper crust apparently extends and diverges slower than after breakup has occurred if Euler plate rotation operates. If the propagating tip of seafloor spreading is the point where the difference in relative surface expressions of plate motion change, then there should be a structure for the accommodation of the different movements of the upper crust such as a transform fault; however no obvious structure is seen. A difference in divergence velocity between pre and post-breakup regions has to be accommodated. The Pocklington Rise, to the south, has no obvious structure that could accommodate the stress in the upper crust caused by local differences in plate velocity, and is therefore assumed to be rigid. However, the Woodlark Rise could accommodate the difference in divergence rate. The Trobriand Subduction Zone and the large strike slip Nubara Fault, which seemingly extends to the propagating tip of seafloor spreading in the Woodlark Basin, are structures that could allow for the difference in velocity.

The true plate motion at the extremities of the continental platform of the Papuan Peninsula is unknown. It is possible that the divergence expected from the Euler pole model here is true. This requires the whole extension of the upper crust, prior to continental breakup, to be much wider than that of the lower crust and lithosphere. This would create a zone where the upper crust diverges at a different velocity to the rest of the lithosphere and hence is "decoupled". Only on the extremities of continental crust would the upper crust and the rest of the lithosphere diverge at the same velocity. In post-breakup settings, it is assumed that the divergence of the plates is accommodated by the formation of new oceanic crust and that all of the continental lithosphere moves uniformly.

Depth-dependent lithospheric thinning has been observed at other rifted continental margins (Baxter et al., 1999, Davis and Kusznir, 2004, Driscoll and Karner, 1998, Kusznir and Karner, 2007, Roberts et al., 1997) however, the Woodlark Basin is a region where continental lithosphere can be observed in the thinning stage before continental breakup and seafloor-spreading initiation has occurred. The observation of depth-dependent lithospheric thinning, before continental lithospheric rupture, and the divergence velocity discrepancy between the upper crust and the rest of the lithosphere that is observed prior the breakup of the Papuan Peninsula may hold the key to understanding the processes that occurs prior to continental breakup and seafloor spreading initiation.

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Chapter 5

The Distribution of Crustal Basement Thinning in the Final Stages of Continental Breakup: Evidence From the Woodlark Basin Derived From Satellite Gravity Inversion

This chapter is a paper to be submitted to GJI. Co-authors include Nick Kusznir, Andy Goodliffe and Jorden Hayes.

Abstract

The Woodlark Basin is an example of present day continental breakup. The area surrounding the Moresby Seamount is one of the few places where the final stages of continental lithospheric thinning and rupture can be observed, prior to the initiation of seafloor spreading. Using satellite gravity inversion and incorporating a lithospheric thermal gravity anomaly correction, crustal thickness, lithospheric thinning factor and residual continental crust variations for the Woodlark Basin are shown. Crustal thickness and thinning factors are shown for the region undergoing the final stages of crustal thinning, prior to continental breakup and seafloor spreading initiation, from a gravity inversion incorporating sediment thickness data. Immediately to the east of the Moresby Seamount, crustal thickness is greater than

expected for initial oceanic crust produced in a magma-poor rifted setting. It is inconclusive from gravity inversion alone whether this region is thick oceanic crust or thinner oceanic crust under a blanket of sediment. The breakup of the palaeo-Papuan Peninsula has occurred over the past 8 Ma, with the Moresby Transform dividing the Woodlark Basin into two basins with contrasting characteristics and different breakup histories. The difference in initial oceanic crustal thickness between the western and eastern basins are examined, concluding that neither basin formation falls into the extremities of magma starved or magma-rich and that the difference in volcanism is subtle. Failed rift zones and isolated slithers of continental crust within the oceanic regime of the Woodlark Basin are observed and it is proposed that the Moresby Seamount is an analogue for the formation of an isolated ribbon of continental crust.

5.1 Introduction

The distribution of crustal thinning during the formation of a rifted continental margin is poorly understood. The majority of rifted continental margins studied are old, cold and have thermally equilibrated since their formation. Due to the time elapsed since forming, many are covered in thick sediment sequences, making understanding the processes that occurred to form them difficult to determine from observations alone. The sediment-starved Woodlark Basin is one of the few places where continental lithosphere, which is thinning leading to continental breakup and the formation of a new ocean basin, can be observed. Pre, syn and post-breakup processes are all observed at the present day, thus the Woodlark Basin allows the study of the distribution of crustal thinning before continental lithospheric rupture and the formation of oceanic crust has occurred.

5.2 Geological Setting

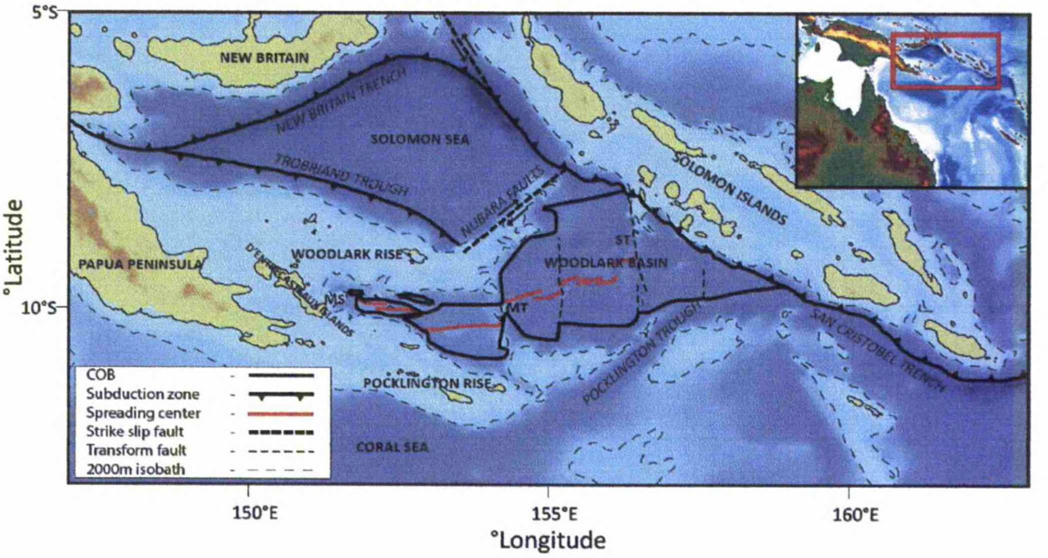

Figure 5.1 Location map showing tectonic features of the Woodlark Basin and surroundings; MT is Moresby Transform, ST is the Simbo Transform, MS is Moresby Seamount. The continent-ocean boundary of the Woodlark Basin is shown. Insert map is of the regional bathymetry. Modified from Taylor et al. (1995)

The Woodlark Basin is a small ocean basin located east of the Papuan Peninsula in the southwest Pacific (figure 5.1). The basin is bounded by the Woodlark Rise to the north and the Pocklington Rise to the south. These rises represent the conjugate margins that were once joined and formed an eastward extension of the continental Papuan Peninsula. The prerift evolution of these margins involved subduction and arc construction followed by collision and suturing of continental and arc regimes

(Davies et al., 1984). North of the Woodlark Rise is the Trobriand Trough where the Solomon Sea Plate is being subducted under the Woodlark Plate. The Eastern Woodlark Basin is being subducted at the San Cristobal Trench (Yoneshima et al., 2005). East of the Trobriand Trough lies the Nubara Fault, a large dextral transform fault that accommodates the difference in subduction between the Woodlark and the Solomon Sea plates at the New Britain Trench (Taylor and Exon, 1987, Weissel et al., 1982).

The southern margin of the Woodlark Basin, the Pocklington Rise, was once part of the Palaeo-Papuan Peninsula. This formed by an underthrust of continental crust extending from the present day Papuan Peninsula to Rossel Island. This emplacement of continental crust underneath the Pocklington Rise has meant that it has eliminated the trench in this location. Further east, past Rossel Island, no such continental fragment was underthrust underneath the Pocklington Rise which has meant the survival of the Pocklington Trough (Karig, 1972).

The palaeo-Papuan Peninsula began rifting during the late Miocene, whilst the earliest seafloor spreading initiated at approximately 6Ma in the east (Taylor et al., 1995). Seafloor spreading has propagated westwards at a rate of approximately 140 km/yr, stepping across the Simbo Transform and the Moresby Transform at 4 Ma and 1.9 Ma respectively and at the present day it is interpreted to be immediately east of the Moresby Seamount.

Bathymetry, free air gravity and seafloor age data have all been used within this study (figure 5.2). Satellite derived and multi-beam ship track bathymetry and free air gravity is publically available online (Becker et al., 2009, Smith and Sandwell, 1997, Sandwell and Smith, 1997), whilst ocean floor age has been determined from seafloor magnetization modelling (Cande and Kent, 1995, Goodliffe, 1998).

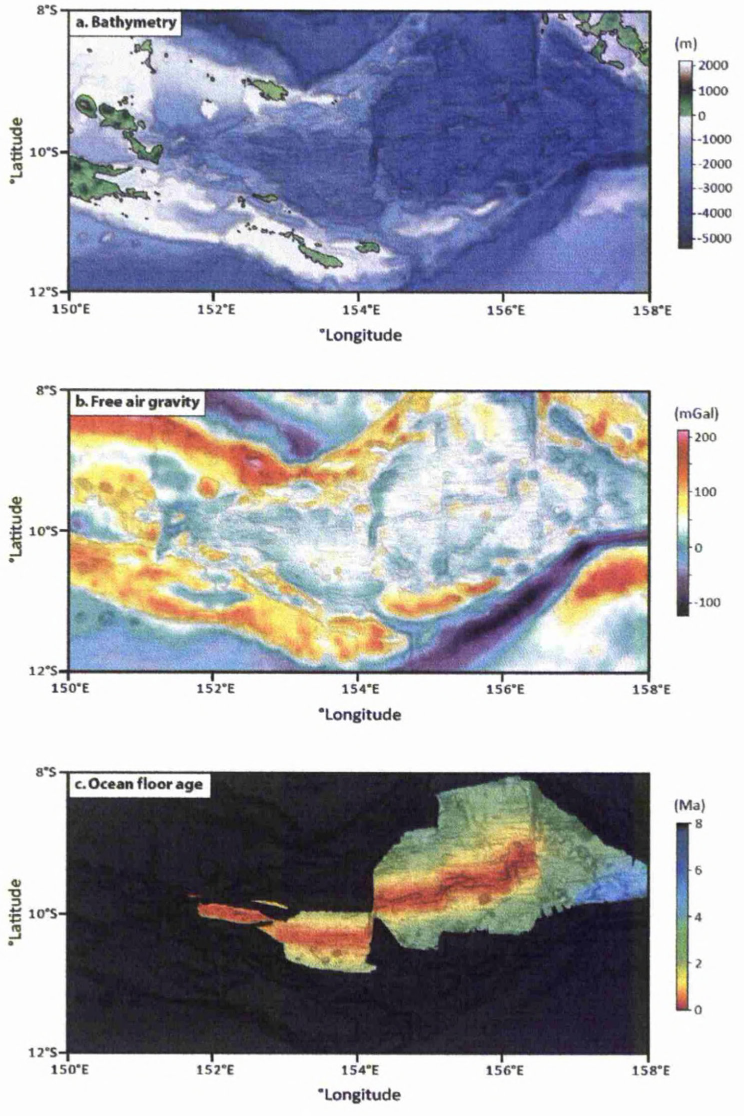

Figure 5.2 (a) Bathymetry (Smith and Sandwell, 1997), (b) satellite free air gravity (Sandwell and Smith, 1997) and (c) ocean floor age (Goodliffe, 1998) showing the primary data used within the gravity inversion. Data sets are overlaid on illuminated bathymetry.

5.3 Crustal Thickness Determination from Gravity Inversion with a Lithosphere Thermal Gravity Anomaly Correction

Satellite gravity inversion, using a lithospheric thermal gravity anomaly, has been used to determine the Moho depth and crustal thinning for the Woodlark Basin. A thermal gravity anomaly correction has been applied due to lateral density variations caused by an elevated geotherm in regions of continental lithospheric thinning or ocean crust formation. These lateral density variations produce a significant negative lithosphere thermal gravity anomaly that can be as much as -300 mGal, for which a correction must be applied in order to determine Moho depth accurately from gravity inversion. The gravity inversion is described in (Chappell and Kusznir, 2008, Greenhalgh and Kusznir, 2007) and uses bathymetry, satellite free air gravity and ocean floor age (figure 2) to determine Moho topography.

Moho topography Δr has been calculated from the residual gravity anomaly Δg_{mra} using the scheme of (Parker, 1972).

$$F[\Delta g_{mra}] = 2\pi G \Delta \rho e^{-|k|z_0} \sum_{n=1}^{\infty} \frac{(|k|)^{n-1}}{n!} F[\Delta r^n]$$

where

$$\Delta g_{mra} = \Delta g_{fag} + \Delta g_b + \Delta g_t + \Delta g_s$$

Δg_{fag} is the observed free air gravity anomaly, Δg_b is the gravity anomaly from bathymetry, Δg_t is the lithosphere thermal gravity anomaly correction and Δg_s is the gravity anomaly from sediment thickness; Z_0 is the mean Moho depth, G = 6.67 ×

10^{-11} $m^3kg^{-1}s^{-2}$, $\Delta\rho = \rho_m - \rho_c$, F denotes a Fourier transform and k is the wave number. Δg_{mra} was filtered before the inversion to remove high frequency components that come from crustal and sediment heterogeneities, using a Butterworth low-pass filter with a cut-off wavelength of 100 km. A Butterworth low-pass filter of 75 km was used to determine residual continental crust thickness. The assumption is made that Δg_{mra} is caused solely by variations in Moho depth. Densities for seawater ρ_w, crust ρ_c and mantle ρ_m used in the inversion are 1039 kgm^{-3}, 2850 kgm^{-3} and 3300 kgm^{-3} respectively. Sediment density assumes normal compaction (Sclater and Christie, 1980).

Crustal thickness ct was calculated from Δr, where d is Moho depth, d_{ref} is Moho reference depth, b is bathymetry and s is sediment thickness.

$$d = d_{ref} + \Delta r$$

$$ct = d - b - s$$

d_{ref} has been determined by calibration using estimates of Moho depth from seismic tomography studies in the area (Ferris et al., 2006, Zelt et al., 2001) and corresponds to a thickness of crust that has a zero bathymetry. A value of 41 km was used in the gravity inversion for the study area; d_{ref} is likely to be higher here than the global average due to surrounding complicated tectonics and numerous subduction zones.

Volcanic addition, from the sea-floor spreading process, results in the formation of oceanic crust and thickening of the continental crust adjacent to the ocean-continent transition. Volcanic addition is estimated from the lithosphere thinning factor γ, where $\gamma = 1 - 1/\beta$, using a parameterisation of decompression melt generation model predictions (Bown and White, 1994, White and McKenzie, 1989). For this, a critical thinning factor is defined for the initiation of oceanic crust production, and a maximum oceanic crustal thickness; for this study area, values of 0.7 and 7 km were used respectively, consistent with melt production at margins with average volcanism (Chappell and Kusznir, 2008).

The Woodlark Basin is a young ocean basin; therefore, assuming a single breakup age is erroneous. Different values for the breakup age (τ), dependent on when breakup occurred, are used. Values of 0, 0.78, 1.95, 4.18 and 5.89 Ma, determined from the oldest magnetic anomaly in each section of the inversion, are used for τ (see **table 1**) (Goodliffe, 1998). The magnitude of the anomaly is governed by β and t. For oceanic lithosphere, $\beta = \infty$ and t is the age of the oceanic lithosphere, defined by isochrons. For continental margin lithosphere,

$$\beta = \frac{ct_{ref}}{ct_{now} - ct_{va}}$$

the ratio of the assumed initial thickness of continental crust ct_{ref} to the present continental crustal thickness ct_{now}, including a correction for volcanic addition thickness ct_{va}, derived from gravity anomaly inversion; t is the time since continental breakup. Lithosphere thinning is assumed equivalent to crustal thinning.

In the absence of oceanic ages from isochrons, or where isochrons are unreliable, an alternative strategy is used to condition the lithosphere thermal model used to define Δg_t. For this, the whole region is treated as continental lithosphere with values for β and t as above. However, this approach fails to predict an increasing thermal anomaly towards the ridge; therefore, it over-predicts the crustal thickness in these regions. To overcome this, a combination of the two methods is used, with isochrons defining t in areas of oceanic crust close to the ridge axis and using a uniform breakup age for t nearer the margin. This method gives an independent prediction of the OCT location and marginal crustal thickness.

Area (Spreading Segment)	Latitude (°S)	Longitude (°E)	Critical Thinning Factor (γ)	Volcanic Addition (km)	Breakup Age (Ma)
Pre-breakup to MS	8 – 12	150 - 151.8	0.7	7	0
1a, 1b & 1c	8 – 12	151.8 - 152.95	0.7	7	0.78
2 to MT	8 – 12	152.95 - 154.2	0.7	7	1.95
MT to 3a & 3b	8 – 12	154.2 - 155.18	0.7	7	4.18
4a, 4b, 4c, 4d, 4e, 5a & 5b	8 – 12	155.18 – 158	0.7	7	5.89

Table 1. Gravity inversion parameters for areas with different breakup ages. Spreading segments from (Goodliffe, 1998). MT is the Moresby Transform, MS is the Moresby Seamount.

An iterative cycle of gravity inversion to predict crustal thickness, β stretching factor, volcanic addition and lithosphere thermal gravity anomaly is used and rapidly converges. Where no sediment thickness data is used within the gravity inversion, the scheme used in this paper produces an upper bound of crustal thickness and a lower bound of lithosphere thinning factor. Where sediment thickness data has been used, crustal basement thickness is determined.

5.4 Crustal Thickness, Lithosphere Thinning Factor and Residual Continental Crust Distribution Predicted by Gravity Anomaly Inversion in the Woodlark Basin

Figure 5.3 shows apparent crustal thickness, continental lithosphere thinning factors and residual continental crust determined for the whole Woodlark Basin from satellite gravity inversion. Previous crustal thickness maps of the Woodlark Basin (Martinez et al., 1999) have shown differences in oceanic crustal thickness between the east and west basins, and these differences are also observed in the crustal thickness map (figure 5.3a). The western basin has oceanic crustal thickness predominantly greater than 8 km whereas in the eastern basin, the oceanic crust is typically less than 7 km. In both the eastern and western basins, there are locally thicker areas of oceanic crust; these are due to seamounts in both basins.

The crustal thickness of the rifted continental margins, the Woodlark and Pocklington Rises, has been determined from the gravity inversion (figure 5.3a). West of the Moresby Transform, the Woodlark Rise crust ranges between 20 and 30 km thickness, whilst the Pocklington Rise varies between 25 and 32 km thickness. Both margins, east of the Moresby Transform, decrease in crustal thickness. The formation of the Woodlark Basin has not equally divided the palaeo-Papuan Peninsula; the Pocklington Rise is generally thicker than the conjugate Woodlark Rise. This is evident in the eastern Woodlark Basin where the continental volume of the Woodlark Rise is significantly less. The continental crust, of the Woodlark Rise, also has a greater variation of crustal thickness than the Pocklington Rise. A crustal thickness difference is apparent either side of the Nubara Fault; this is evident east of Egum Atoll where there is a sudden increase in thinned continental crustal thickness to the north of the Nubara Fault from 14 km to more than 17 km.

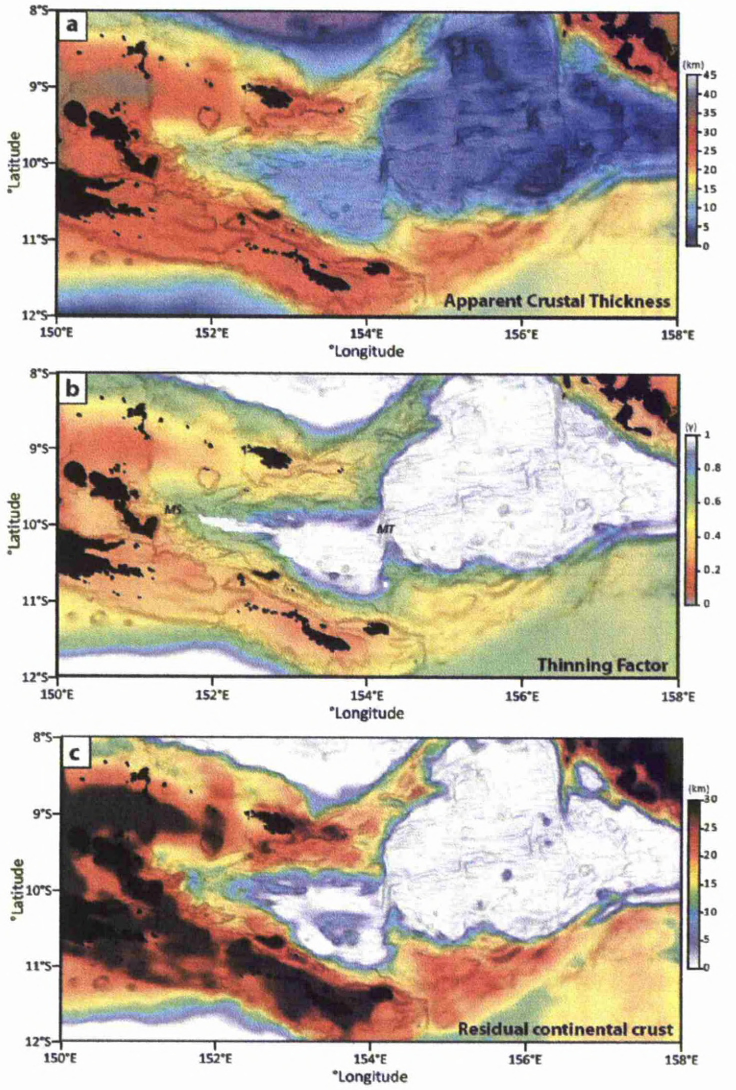

Figure 5.3 Crustal thickness predicted by gravity inversion (a) with a lithosphere thermal gravity correction and corresponding predicted lithosphere thinning factor (b) for the Woodlark Basin. Densities of 1000, 2850 and 3300 kg/m3 are used for the water, crust and mantle respectively. A volcanic addition of 7 km has been used for the whole Woodlark Basin. Residual continental crust thickness (c) is determined from a gravity inversion using low-pass Butterworth filter of 75 km, (a) and (b) use a filter of 100 km.

Continental lithospheric thinning factor maps have been produced for the Woodlark Basin from the satellite gravity inversion (figure 5.3b). This map shows the distribution of crustal thinning. The Woodlark Rise has a minimum thinning factor of 0.4; whereas the Pocklington Rise has regions that correspond to a thinning factor of 0.25. Both margins exhibit higher degree factors in the eastern basin. Ahead of the propagating tip of seafloor spreading, most of the thinning of the crust is to the northwest of the Moresby Seamount where the crust is between 15 and 20 km thick and has a thinning factor of 0.5 to 0.6. This thinning corresponds with an area of few upper crustal faults. South of the Moresby Seamount, with the exception of the Southern Moresby Graben, the crust, despite being intensely faulted, rapidly thickens westwards towards Normanby Island.

Residual continental crust is determined by the subtraction of the volcanic addition predicted thickness from the total thickness of the crust derived from gravity inversion and it allows for further interpretation of the ocean continent transition zone. Residual continental crustal thickness for the Woodlark Basin, derived by gravity inversion, is shown in figure 5.3c. In the northern margin of the western Woodlark Basin, here the OCT is wide and has variable continental crustal thickness. This is distinctly different to both the margins in the eastern basin and the southern margin in the western basin, where the OCT is relatively narrow exhibiting a rapid decrease from continental crustal thickness of 15 km to 0 km in less than 20 km.

5.5 The breakup of the palaeo-Papuan Peninsula and initiation of seafloor spreading

The residual continental crustal thickness and thinning factor maps of the Woodlark Basin (figure 5.3) show that in the eastern basin, the breakup of continental crust has occurred after significant thinning of the crust and continental lithosphere. However, when breakup occurred and seafloor-spreading centres initiated, these spreading centres matured and formed the bulk of the oceanic crust that is observed today in the eastern basin. The only significant change in their formation since breakup is a re-organisation of the spreading centres at 80 ka (Goodliffe et al., 1997). Between the Moresby Transform and immediately east of Misima Island, in the western basin, the initial rupturing of the continental crust was similar to in the east. Once breakup of the continental crust had occurred, seafloor spreading initiated and these spreading centres still form new oceanic crust today. Single continental lithospheric and crustal rupturing has not always occurred further west and breakup here is more complicated.

Figure 5.4 shows (a) the ocean floor magnetization, (b) residual continental crust and (c) two crust cross sections in the western Woodlark Basin between Woodlark Island on the Woodlark Rise and Misima Island on the Pocklington Rise. The breakup of the palaeo-Papuan Peninsula between Woodlark Island on the Woodlark Rise and Misima Island on the Pocklington Rise was not a single rupturing of the crust. The magnetization of the ocean floor shows two oceanic crust bodies (Goodliffe, 1998); the most recent seafloor spreading that has occurred during the Brunhes epoch and an earlier formation that has a negative magnetic polarity to the north of the main oceanic crust. Gravity inversion predicts that between these two magnetic anomalies lies a slither of continental crust approximately 10 km thick (figure 5.4).

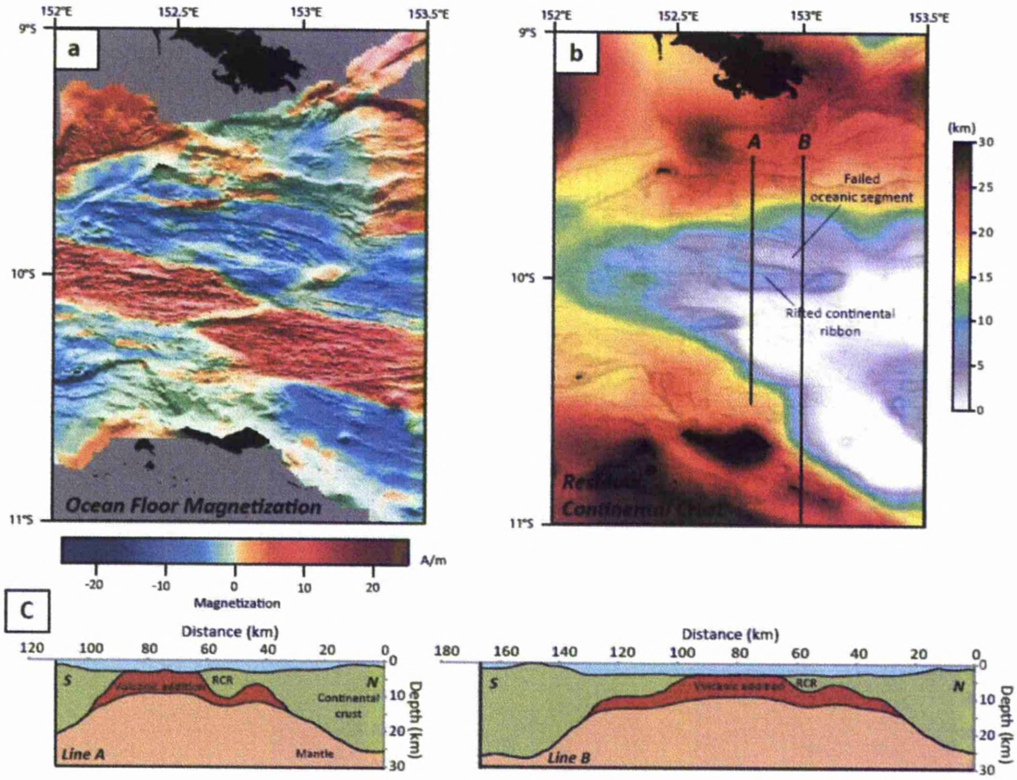

Figure 5.4 Ocean floor magnetization (A) and residual continental crust (B) predicted by satellite gravity inversion. Oceanic crust formed since the start of the Brunhes chron (0.78-0 Ma) has a positive magnetization. Cross sections derived from a gravity inversion assuming volcanic addition of 7 km and a critical thinning factor of 0.7 show predictions of Moho depth and volcanic addition for two lines across a failed oceanic segment and a rifted continental ribbon.

Here, as the crust and lithosphere thinned, continental lithospheric rupture initially occurred to the north of the slither of continental crust. This rupturing of the lithosphere did not develop into a seafloor-spreading centre of any longevity; instead thinning of the crust continued or jumped approximately 25 km to the south, where the crust and continental lithosphere once again ruptured, and this time forming the seafloor-spreading centre that is still active today. The slither of thicker crust is predicted not to be solely continental crust, but this slither of crust has thinned to an extent where a degree of volcanic addition has occurred. The timing of the volcanic addition to this slither of continental crust is unknown, but could have been associated with one, if not both, of the continental lithospheric thinning and rupturing events. It could be that the thinning of the crust to rupture it to the north was sufficient to produce volcanics here, however it is equally plausible that the second thinning phase, leading to the formation of the present seafloor spreading centre further thinned this continental crustal slither and that it had either a primary or secondary magmatic phase then.

The formation of a continental slither is not unique to this one location. These structures should be classified separately from micro-continents, as they are significantly smaller, and are thus termed *rifted continental ribbons* (RCR). A RCR is a thin continental fragment that is surrounded by oceanic crust on both sides or with thinner continental crust with a higher degree of volcanism on one of the sides. They either are located within the ocean continental transition of a rifted continental margin or are surrounded by oceanic crust.

Rifted continental ribbons can form in one of two ways (figure 5.5). A type 1 RCR forms due to a secondary rupturing of the continental lithosphere that then becomes the dominant seafloor-spreading centre. The first spreading centre that forms may not be in the location where the process driving seafloor spreading is directly centred underneath and due to the difference in actual seafloor spreading location and preferred spreading location, a slither of highly thinned continental crust can be formed when the spreading centre jumps by several ten's of kilometres.

A type 2 RCR, as shown in figure 5.5, forms when the crust ruptures in two locations along a profile perpendicular to the Euler pole of opening. Two spreading centres can form when the propagation of seafloor spreading is in stages of sequentially offset spreading centres; these stages can initially overlap by several kilometres (Goodliffe, 1998, Taylor et al., 2009). The two spreading centres, simultaneously forming new oceanic crust, cannot co-exist in close proximity to each other and eventually one spreading centres fails whilst the other becomes the main centre of oceanic crust formation. The crust that lies between the two spreading centres will be a highly thinned rifted continental ribbon. A type 2 RCR has formed in the western Woodlark Basin where two spreading centres have overlapped (figure 5.6). Due to the difference in spreading rates along the seafloor spreading centres, the RCR has not only separated away from the rifted continental margin, but has also rotated anticlockwise by 30°. The rotation of this RCR only began upon the onset of seafloor spreading and the detachment of it from the continental crust of the Pocklington Rise at 0.65 Ma.

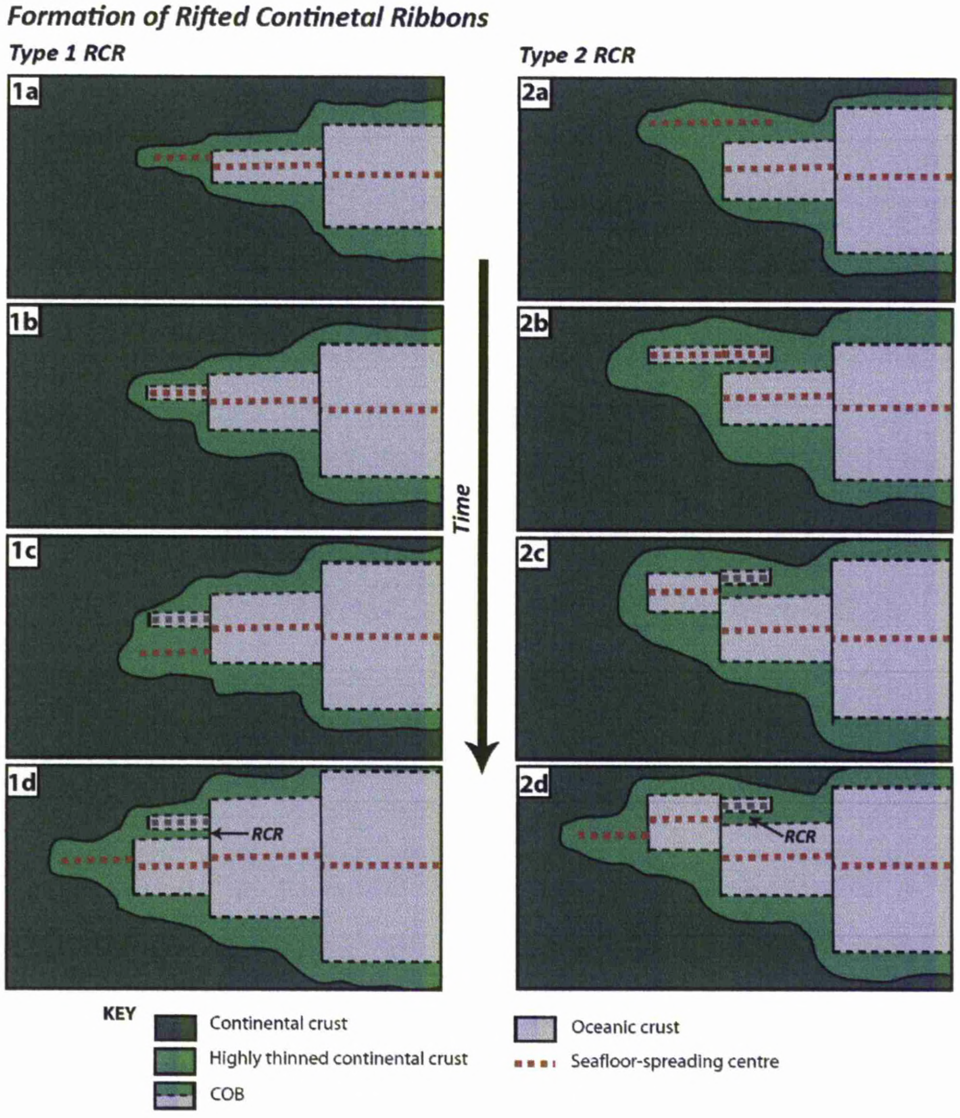

Figure 5.5 Diagram not to scale, showing two ways of forming a rifted continental ribbon (RCR). (1a) A type 1 RCR forms when a new offset seafloor-spreading segment begins to form, (1b) oceanic crust is formed at this new spreading segment, (1c) a ridge jump to the south, ending the previous seafloor-spreading segment, this ridge jump isolates a slither of highly thin continental crust (a rifted continental ribbon). Finally (1d) the seafloor-spreading segment develops into the main spreading centre for that section of oceanic crust and the spreading centre continues to propagate into highly thinned continental crust. (2a) A type 2 RCR forms when a new spreading centre overlaps with the previous spreading centre. (2b) The new spreading centre develops and forms new oceanic crust. (2c) Part of the spreading centre fails separating a slither of continental crust from the rifted continental margin between oceanic crust. (2d) The propagation of seafloor spreading continues.

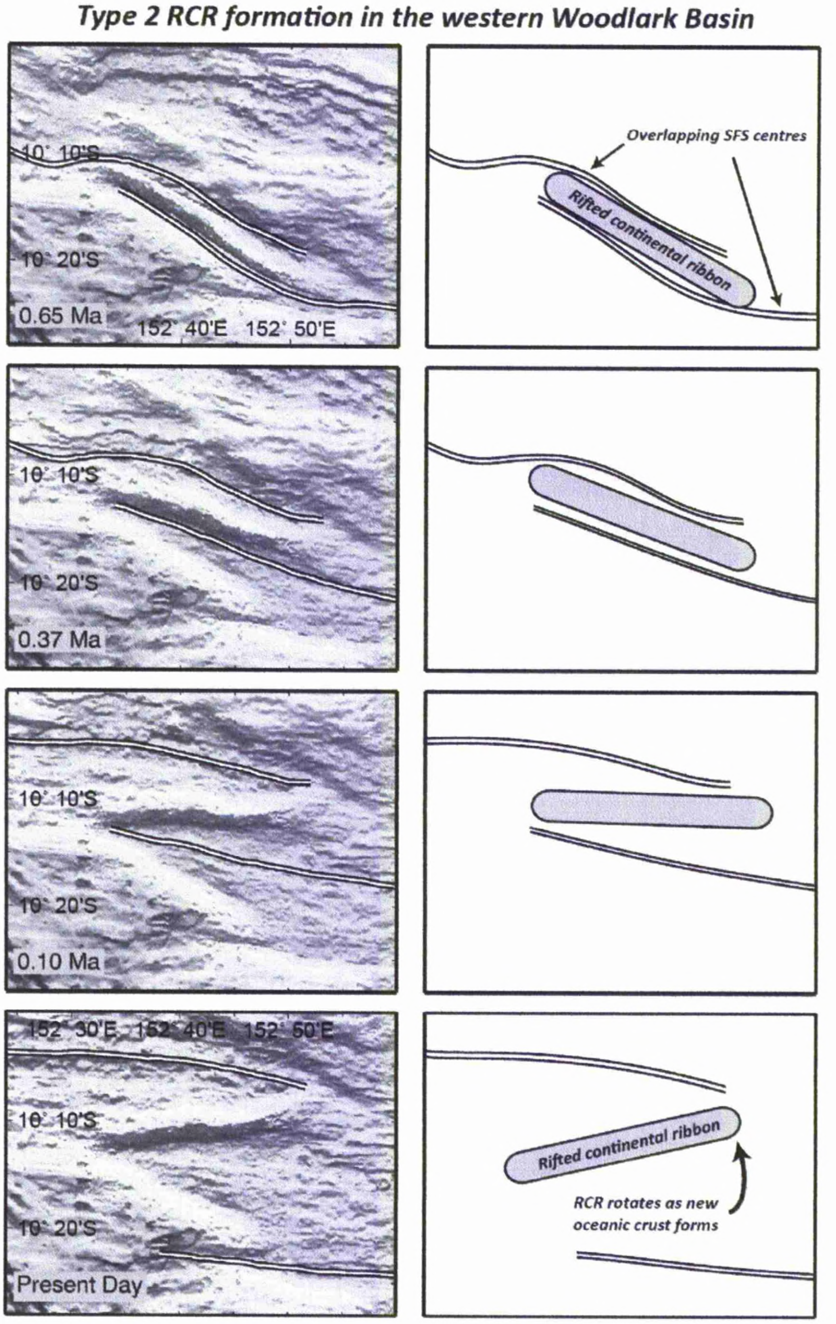

Figure 5.6 Diagram showing the formation of a type 2 RCR in the western Woodlark Basin overlain on illuminated bathymetry. At 0.65Ma seafloor spreading initiates detaching the slither of continental crust away from the Pocklington Rise. This RCR rotates anticlockwise as new oceanic crust is formed around it.

5.6 Crustal basement thinning at the present rifting to spreading transition

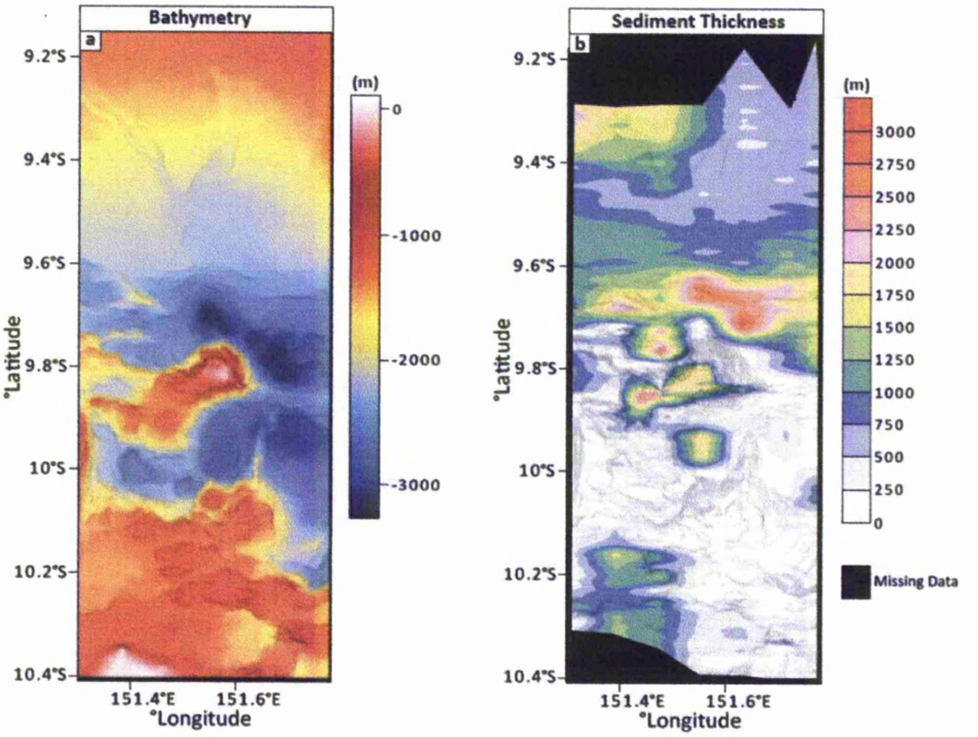

Figure 5.7 Sediment thickness map produced by depth converting a series of multi channel seismic lines. Black areas correspond to areas where there is no seismic coverage.

Figure 5.7 shows the sediment distribution for the region around the Moresby Seamount. The inclusion of sediment thickness data into the gravity inversion to determine crustal basement thickness allows for a further, more detailed insight of how the crust thins prior to continental breakup. A new sediment thickness map has been determined (figure 5.7), using multi-channel seismic reflection lines, for the area surrounding the Moresby Seamount. These lines have been depth converted using a equation derived from well log velocities taken from ODP Leg 180

(Goodliffe et al., 2002) and it has been tested using well locations where depth to basement is known. The region has good 2D seismic reflection coverage and therefore an approximate sediment thickness map can be produced for the region.

In the North and South Moresby Grabens, where sediments are thickest, the error in sediment thickness is estimated to be in the order of ±250 m since the depth conversion has been determined using thinner sediment sequences. Picks of top basement are clear in most seismic lines, except across the South Moresby Graben. Here sediment thickness has been estimated; it is possible that there are basaltic layers within the sediment and that they affect the quality of the seismic imaging.

There is little to no volcanic addition predicted in either the northern or southern Moresby Grabens (figure 5.3, 5.8 and 5.9) from gravity inversions that do not incorporate sediment thickness data; however, if sediments are included within the gravity inversion several kilometres of volcanics are now predicted and this is shown in the crustal cross sections in figure 5.9. Well, acoustic sounding and magnetization data show that there is little or no volcanic material within the sediment in either graben. It is possible that the volcanic addition does not reach the surface and precipitates itself as magmatic underplating. It is also feasible that there is no magmatism in this region and that the prediction of volcanic addition for the grabens is incorrect. The first evidence of surface volcanism presents itself in the form of ring dykes (Goodliffe and Taylor, 2007) approximately 10 km east of Moresby Seamount.

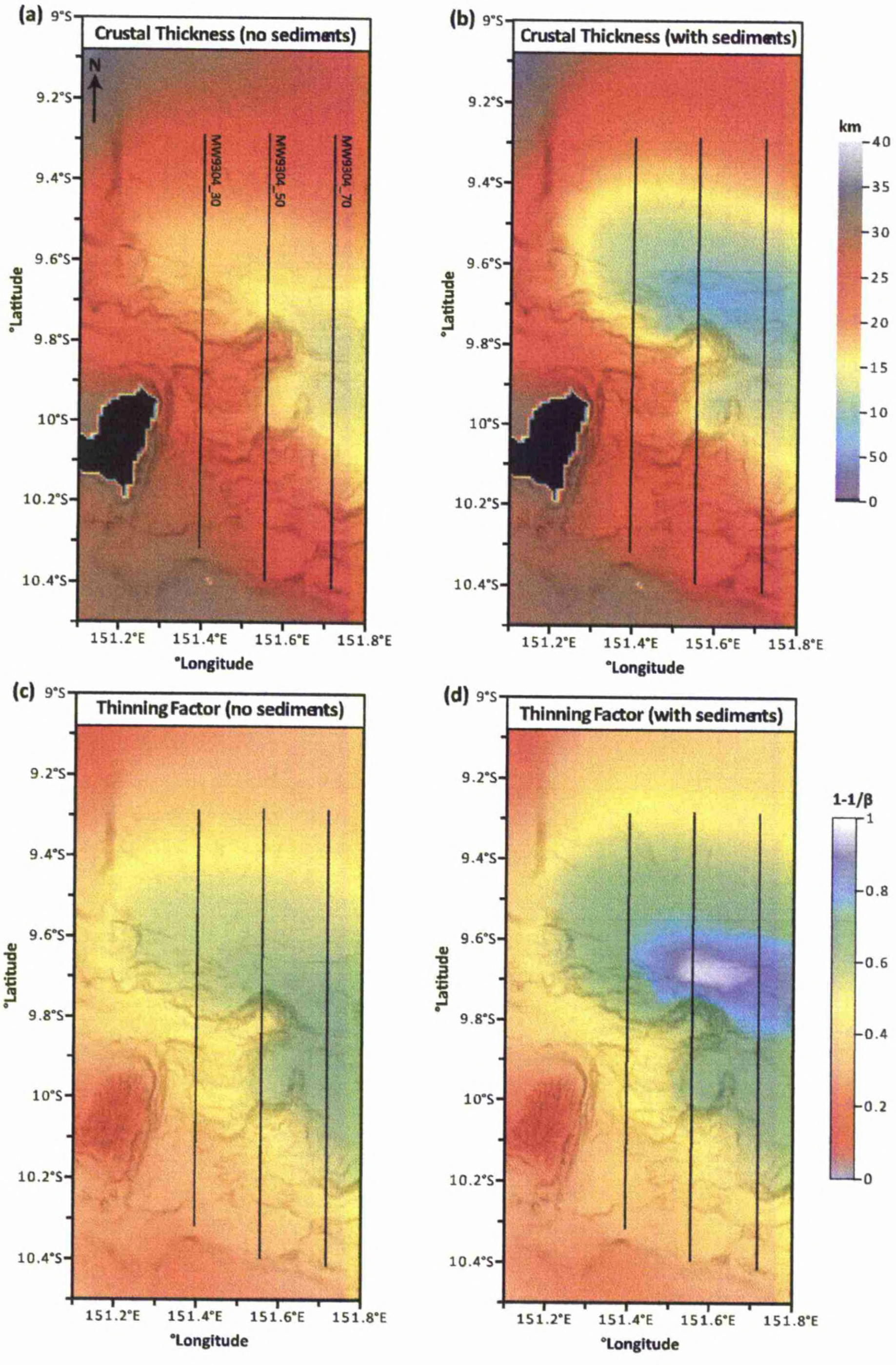

Figure 5.8 Crustal thickness and thinning factor maps for the region around the Moresby Seamount produced from gravity inversions with and without sediment thickness data included. With sediment thickness data included, thinning factors reach almost 1 predicting the onset of oceanic crust formation to the north east of Moresby Seamount.

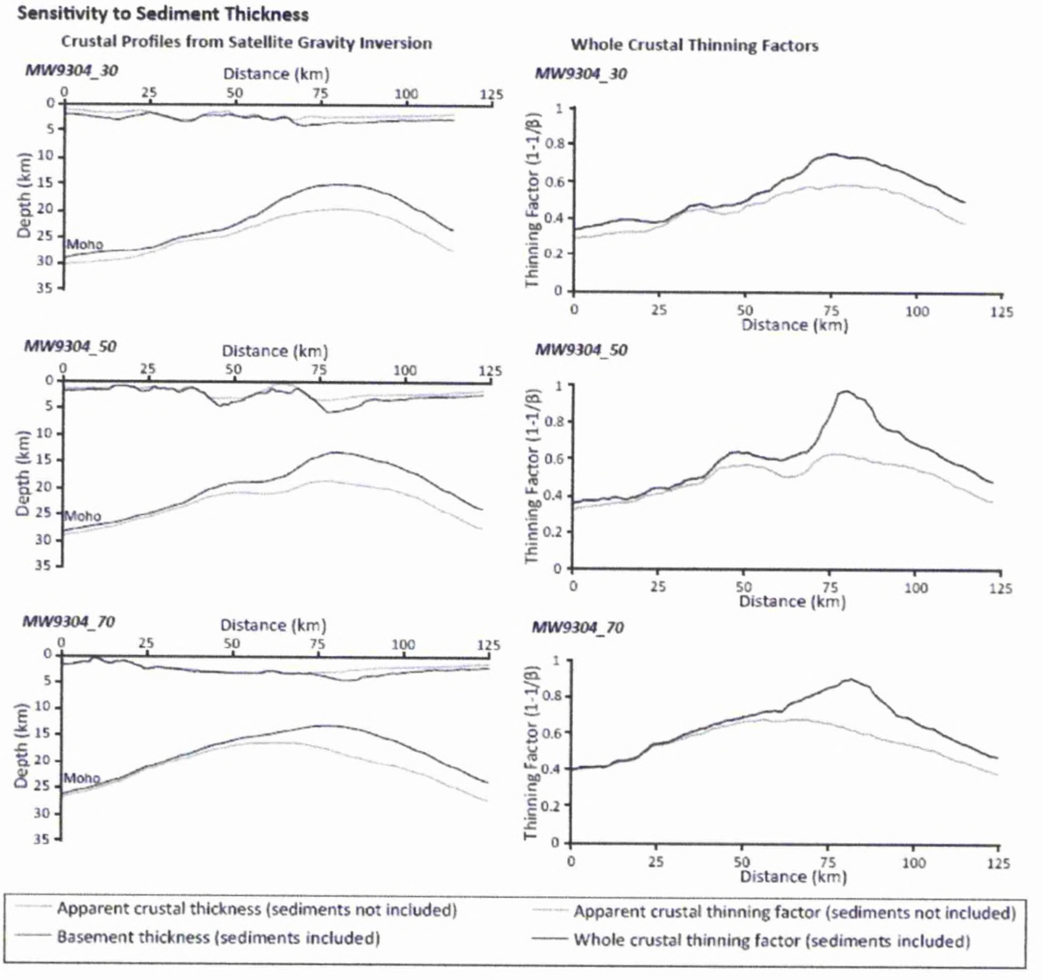

Figure 5.9 Crustal cross sections and whole crustal thinning factors showing sensitivity of gravity inversion to sediment thickness. Including sediments predicts thinner continental crust and hence determines higher thinning factors.

The Woodlark Basin initially struggles to produce volcanics during the breakup process; however, once breakup has occurred, the gravity inversion predicts that the most recent section of oceanic crust is thick, ranging between 11 and 13 km. This thick initial oceanic crust is shortly followed by production of more normal thicknesses of oceanic crust after several hundred thousand years. This production of thicker crust is consistent with observations from magma-rich margins (Mutter et al., 1982, Barton and White, 1997, Hopper et al., 1992) however; these margins exhibit much more volcanism prior to breakup than the Woodlark. The region 20-60 km east of Moresby Seamount exhibits shallower bathymetry than the grabens around the Moresby Seamount suggesting thicker crust in this region. This region, if true oceanic crust, is the thickest oceanic crust observed within the Woodlark Basin. There is seemingly more volcanism here, the presence of the Franklin and Cheshire Seamounts in this area confirm this. It is unclear whether this location suffered a magma-rich episode in the breakup of the Woodlark Basin. The crustal thickness prediction for this area is from a gravity inversion that does not include sediment thickness data hence crustal thicknesses are possibly over estimated. It is believed that there is little sediment in this region of oceanic crust; however, the possibility cannot be ignored that the sediment here is of substantial thickness to affect results. If there were more sediments than previously known, this would reduce this predicted crustal thickness derived from the gravity inversion.

Further west, in the region where the continental crust and lithosphere is thinning but has not yet ruptured, the inclusion of sediment thickness into the gravity inversion does not affect the overall interpretation. The crust is predicted to be thinner than the previous gravity inversion that did not include sediment thickness data; however, the distribution of thinning is still to the north of the rift axis location around Moresby Seamount. The focus of crustal thinning to the north of Moresby Seamount, as predicted by gravity inversion, is contradicted by heat-flow measurements. Heat-flow values in the North Moresby Graben reach 100 mWm^{-2}, however these are superseded by measurements of up to 280 mWm^{-2} in the South Moresby Graben

(Goodliffe et al., 2000). The breakup history of the western Woodlark Basin has shown that the initial breakup of the crust does not mean that a stable seafloor-spreading centre will fully develop and that continental breakup in another location and the formation of a rifted continental ribbon can occur. The contradicting evidence as to where continental breakup will occur, raises the question of where will the continental lithosphere rupture around Moresby Seamount (figure 5.10). Heat-flow measurements suggest breakup will occur to the south, whilst gravity inversion predicts greater thinning of the crust to the north. It is possible that multiple breakup events will occur and that the Moresby Seamount will, in the future, become a rifted continental ribbon.

The Moresby Seamount is poorly named geologically as seamounts have become synonymous with volcanic edifices. The Moresby Seamount is a large fault horst of continental crust (Taylor and Huchon, 2002). The bulk of its composition is continental crust that is 18-25 km thick. It is bounded by two large grabens to the north and south; the North Moresby Graben (NMG) and the South Moresby Graben (SMG).

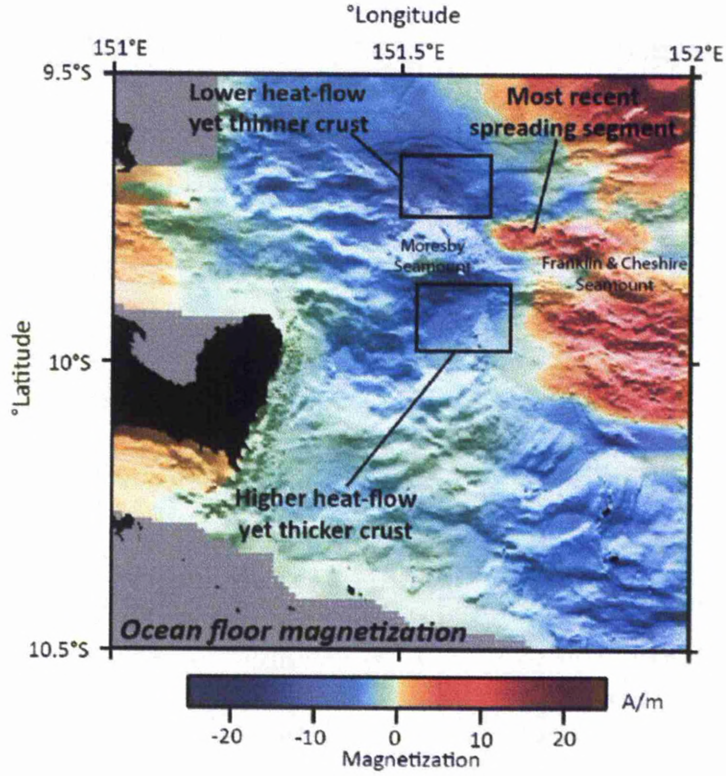

Figure 5.10 Seafloor magnetization illuminated from the north by bathymetry. Oceanic crust formed since the start of the Brunhes chron (0.78-0 Ma) has a positive magnetization. The gravity inversion predicts that breakup will occur to the north of the Moresby Seamount since this has thinned more than in the south; this prediction is contradicted by heat-flow measurements that show that the SMG is significantly hotter than the NMG. It is possible that the Moresby Seamount, in the future, will become a rifted continental ribbon with highly thinned continental crust with volcanics on one side and oceanic crust on the other.

The most recent segment of seafloor-spreading to form is to the immediate northwest of the Moresby Seamount (Goodliffe and Taylor, 2007) in alignment with the NMG suggesting that continental breakup will occur here. This is contradicted by the heat-flow measurements and it is possible that the latest spreading centre will jump to the south and the main seafloor-spreading ridge will develop here. Thus, the Moresby Seamount would become a rifted continental ribbon bounded by a slither of either highly thinned continental crust with numerous volcanics or oceanic crust to the north and a main oceanic system to the south.

5.7 Summary

The palaeo-Papuan Peninsula underwent significant crustal thinning prior to breakup and the initiation of seafloor spreading. Evidence from the western Woodlark Basin has shown that the first continental crustal rupture has not necessarily led to the present seafloor-spreading centre, and that a rifted continental ribbon can form due to spreading centre relocation.

At the rifting to spreading transition, the crust has thinned significantly prior to the first observed volcanism at the ocean floor. Crustal thinning is predicted, from gravity inversion, to be distributed primarily to the north of the rift axis in the continental lithosphere and crust that is currently thinning. It is unclear, due to higher heat-flow measurements south of Moresby Seamount, where the final rupturing of the lithosphere will occur around the Moresby Seamount. It is plausible that the continental crust of the Moresby Seamount will become isolated from both the Woodlark and Pocklington Rises and become a rifted continental ribbon.

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Chapter 6

Regional and Local Variations in Continental Rifted Margin Type: Evidence From Crustal Thickness Mapping of the Woodlark Basin Derived From Satellite Gravity Inversion

This chapter is a short paper to be submitted to Geology. Co-authors include Nick Kusznir and Andy Goodliffe.

Abstract

Crustal thickness and lithosphere thinning factors have been mapped in the Woodlark Basin using a satellite gravity inversion incorporating a lithospheric thermal gravity correction. Oceanic crustal thickness in the Eastern Woodlark Basin are predicted to be between 4 and 7 km whereas in the Western Woodlark oceanic crustal thicknesses range between 6 and 13 km. The two sub-basins are divided by the Moresby Transform, across which the difference in oceanic crustal thickness is sharp with thicker oceanic crust in the west. The observation of variation in crustal thickness across the Moresby Transform has implications for the classification of a rifted margin in terms of its breakup volcanism. Previous studies have stated that different pre-rift lithosphere thicknesses between the eastern and western basins could result in

differences in oceanic crustal thickness; however, it is possible that the mantle in the Western Woodlark is more hydrated due to surrounding subduction zones supplying the seafloor spreading system with a greater amount of adiabatic melt. It is concluded that there are regional and local variations in initial oceanic crustal thickness, and thus the Woodlark Basin cannot be classified solely as a magma poor margin. A rifted continental margin may need labelling different classifications for different parts since the initial thickness of oceanic crust can vary abruptly and a simple system of classifying rifted margins locally in terms of initial oceanic crustal thickness is proposed.

6.1 Introduction

Rifted continental margins are classified on the basis of the amount of volcanism that occurs during their formation. Initial oceanic crustal thicknesses at the continental ocean boundary are good indicators of how volcanic the rupture of the continental lithosphere was. Initial oceanic crustal thicknesses can vary by up to 10-20km globally (Coffin and Eldholm, 1994) and can be influenced by a number of different factors; mantle temperature (White and McKenzie, 1989, McKenzie and Bickle, 1988), mantle composition, rate and volume of mantle through the zone of partial melting and initial seafloor spreading rate. Continental rifted margins have been classified on their magmatic addition and initial oceanic crustal thickness between the two end members, magma poor and magma rich. Magma poor rifted margins are typically characterised by anomalously thin initial oceanic crust and more recently, magma poor margins have become associated with the exhumation of serpentinised mantle (Dean et al., 2000, Manatschal and Bernoulli, 1999, Pickup et al., 1996, Whitmarsh et al., 2001). Magma rich margins exhibit inner and outer seaward-dipping reflectors (Mutter et al., 1982, Barton and White, 1997, Hopper et al., 1992) and abnormally thick initial oceanic crust (Roberts et al., 1984, White and McKenzie, 1989, White et al., 1987, Geoffroy, 2005). A problem with labelling margins as either magma poor or volcanic is that a margin can become stereotyped incorrectly since there is a sampling bias towards certain rifted continental margins as the "classic" model of end member margins.

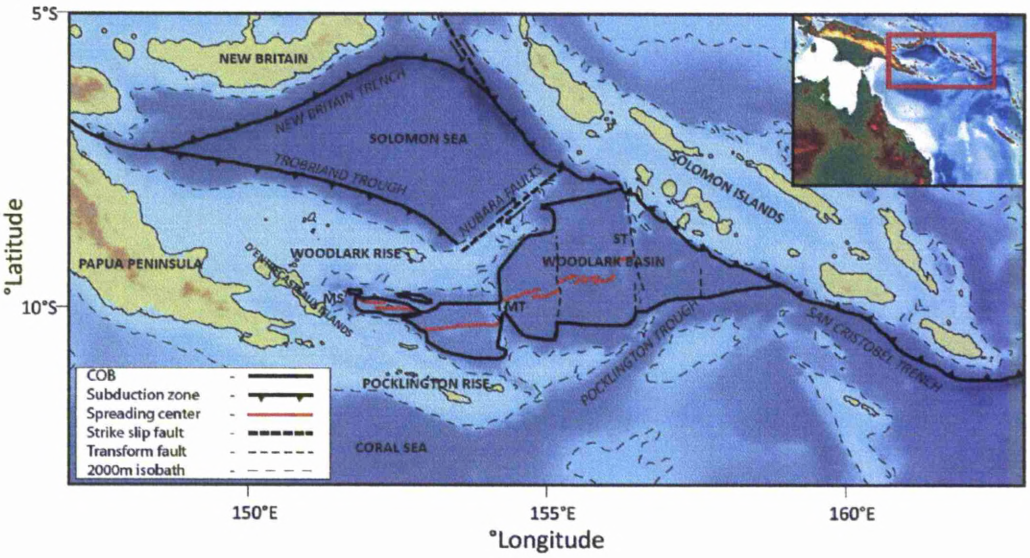

Figure 6.1 Location map showing tectonic features of the Woodlark Basin and surroundings; MT is Moresby Transform, MS is Moresby Seamount. (Modified from Taylor et al. 1995)

Oceanic crustal thickness varies in thickness along the continent ocean boundary of a rifted continental margin; however, significant differences in oceanic crustal thicknesses are typically over distances >10 km ranging up to 100's of kilometres. Recent studies in the Black Sea (Shillington et al., 2009) have shown that oceanic crustal thickness can also vary abruptly, with differences in the thickness of oceanic crust across individual transform faults.

The Woodlark Basin (figure 6.1) is a young ocean basin in the western Pacific where seafloor spreading commenced at ~6 Ma and has been propagating westwards into the continental crust and lithosphere of the Papuan Peninsula at a rate of 140km every million years (Taylor et al., 1995, Taylor and Exon, 1982, Weissel et al., 1982). Previous studies of oceanic crustal thickness (Martinez et al., 1999) using isostatic arguments have proposed an oceanic crustal thickness change, coinciding with a difference in bathymetry, across the Moresby Transform, a large transform fault that separates the western and eastern basins of the Woodlark Basin. The western basin exhibits the thicker oceanic crust and a shallower seafloor. The oceanic crust in the western basin is greater than 7km thickness whereas in the eastern basin thicknesses are typically less than 7km (Martinez et al., 1999). Despite this variation in oceanic crustal thickness, the Woodlark Basin is regarded as a magma poor basin due to the lack of large amounts of volcanism at the surface prior to continental breakup and the onset of seafloor spreading (Goodliffe and Taylor, 2007). There are small volcanic edifices on the rift margin south of Egum Atoll; however, the volumes of magmatism are small. The seafloor spreading centres are offset by approximately 40 km across the Moresby Transform, which is the youngest transform fault in the spreading system. The oceanic crust close to the spreading centre in both the east and west basins is the same age; the difference is that in the western basin there are no transforms where the spreading centre has nucleated most recently (Taylor et al., 2009, Taylor et al., 1995, Taylor et al., 1999).

6.2 Crustal Thickness Determination from Gravity Inversion with a Lithosphere Thermal Gravity Anomaly Correction

A satellite gravity inversion using a lithospheric thermal gravity anomaly (Chappell and Kusznir, 2008, Greenhalgh and Kusznir, 2007) has been used to determine Moho depth for the Woodlark Basin. Due to lateral density variations caused by the elevated geotherm in thinned continental margins and adjacent ocean basin lithosphere, a lithosphere thermal gravity anomaly correction has been applied. These lateral variations in lithosphere temperature and density produce a significant negative gravity anomaly that can be in excess of -300 mGal, therefore a correction must be applied in order to determine Moho depth accurately from gravity inversion. The gravity inversion uses bathymetry (Smith and Sandwell, 1997, Becker et al., 2009), free air gravity (Sandwell and Smith, 1997) and ocean age data sets (Goodliffe, 1998), (figure 6.2) in the determination of Moho depth and crustal thickness.

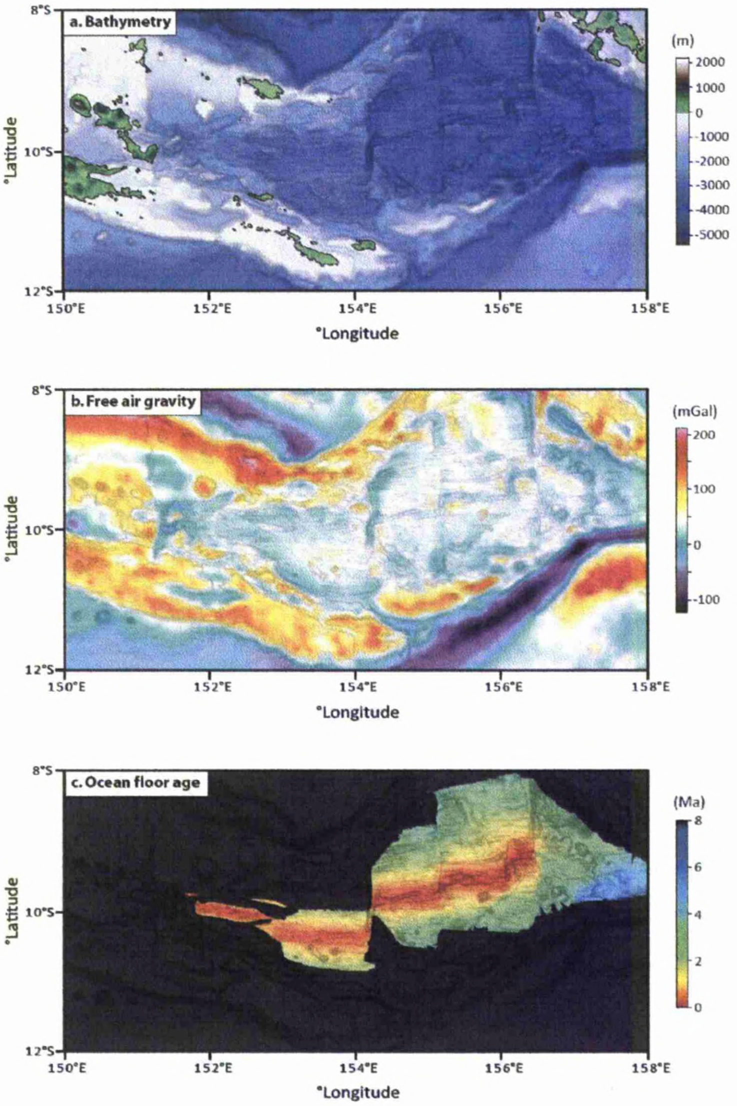

Figure 6.2 (a) Bathymetry (Smith and Sandwell, 1997), (b) satellite gravity (Sandwell and Smith, 1997) and (c) ocean age (Taylor et al., 1995) showing the primary data used within the gravity inversion.

Moho topography Δr was calculated from the residual gravity anomaly Δg_{mra} using the scheme of Parker (1972).

$$F[\Delta g_{mra}] = 2\pi G\Delta\rho e^{-|k|z_0} \sum_{n=1}^{\infty} \frac{(|k|)^{n-1}}{n!} F[\Delta r^n]$$

where

$$\Delta g_{mra} = \Delta g_{fag} + \Delta g_b + \Delta g_t$$

Δg_{fag} is the observed free air gravity anomaly, Δg_b is the gravity anomaly from bathymetry, Δg_{mra} is the mantle residual gravity anomaly and Δg_t is the lithosphere thermal gravity anomaly correction; z_0 is the mean Moho depth, G = 6.67 × 10^{-11} $m^3kg^{-1}s^{-2}$, $\Delta\rho = \rho_m - \rho_c$, F denotes a Fourier transform and k is wave number. Δg_{mra} was filtered before the inversion to remove high frequency components, using a Butterworth low-pass filter with a cut-off wavelength of 100 km. A Butterworth low-pass filter with a cut-off of less than 100km allows high frequency components to be included into the inversion resulting in oscillations in the determination of Moho depth. The assumption is made that Δg_{mra} is caused solely by variations in Moho depth. Densities for crust ρ_c and mantle ρ_m used in the inversion are 2850 kgm^{-3} and 3300 kgm^{-3} respectively. Crustal thickness *ct* was calculated from Δr, using:

$$ct = d - b$$

$$d = d_{ref} + \Delta r$$

where d is Moho depth, d_{ref} is Moho reference depth and b is bathymetry.

The Moho depth, derived from gravity inversion, has been compared to values determined from seismic tomography studies in the area (Zelt et al., 2001, Ferris et al., 2006) in order to determine d_{ref}. A value of 41 km was used for d_{ref} in the gravity inversion for the study area.

The observed thermal gravity anomaly Δg_t at ocean ridges may be as much as −300 mGal (Greenhalgh and Kusznir, 2007), with lower, but still substantial values observed away from the ridge and within rifted continental margin lithosphere. The initial perturbation of the geotherm is described by the lithosphere stretching factor β (McKenzie, 1978). Δg_t is caused by the density contrast $\Delta\rho_t$ arising from lateral variations in lithosphere temperature, and has been calculated using $\Delta\rho_t = \rho\alpha\Delta T$. Lithosphere temperature may be calculated using the lithosphere thinning model by (McKenzie, 1978) and used to predict the lithosphere thermal gravity anomaly correction (Greenhalgh and Kusznir, 2007).

$$\Delta g_t = \frac{8G\alpha\rho a T_m}{\pi}\sum_{m=0}^{\infty}\frac{1}{(2m+1)^2}\left[\frac{\beta}{(2m+1)\pi}\sin\frac{(2m+1)\pi}{\beta}\right]\exp\left(-\frac{(2m+1)^2 t}{\tau}\right)$$

Where: a, the lithosphere thickness = 125 km; α, the coefficient of thermal expansion = 3.28 × 10^{-5} °C^{-1}; ρ, the lithosphere density = 3300 kgm^{-3}; T_m, the base lithosphere temperature = 1300°C and t is the lithosphere thermal equilibration time (Ma); τ, the lithosphere cooling thermal decay constant is dependent on location within the basin.

Since the Woodlark Basin is a young ocean basin, a single breakup time cannot be assumed as eastern and western basins have different thermal gravity anomalies, and because of this, different values for τ dependent on when breakup occurred for that sector are used (see table 1). The magnitude of the anomaly is governed by β and t. For oceanic lithosphere, $\beta = \infty$ and t is the age of the oceanic lithosphere, defined by isochrons (Goodliffe, 1998). For continental margin lithosphere,

$$\beta = \frac{ct_{ref}}{ct_{now} - ct_{va}}$$

The ratio of the initial thickness of continental crust ct_0 to the present continental crustal thickness ct_{now} minus the thickness of volcanic addition due to decompression melting ct_{va} derived from gravity anomaly inversion; t is the time since continental breakup. Lithosphere thinning is assumed equivalent to crustal thinning.

Area (Spreading Segment)	Latitude (°S)	Longitude (°E)	Critical Thinning Factor (γ)	Volcanic Addition (km)	Breakup Age (Ma)
Pre-breakup to MS	8 – 12	150 - 151.8	0.7	7	0
1a, 1b & 1c	8 – 12	151.8 - 152.95	0.7	7	0.78
2 to MT	8 – 12	152.95 - 154.2	0.7	7	1.95
MT to 3a & 3b	8 – 12	154.2 - 155.18	0.7	7	4.18
4a, 4b, 4c, 4d, 4e, 5a & 5b	8 – 12	155.18 – 158	0.7	7	5.89

Table 1. Gravity inversion parameters for areas with different breakup ages. Spreading segments from (Goodliffe, 1998).

In the absence of oceanic ages from isochrons, or where isochrons are unreliable, an alternative strategy may be used to condition the lithosphere thermal model used to define Δg_t. For this, the whole region is treated as continental lithosphere with values for β and t as above. However, this approach fails to predict an increasing thermal anomaly towards the ridge so it over-predicts the crustal thickness in these regions. To overcome this, a combination of the two methods was used, with isochrons defining t in areas of oceanic crust close to the ridge axis and using a uniform breakup age for t nearer the margin. This method gives an independent prediction of the OCT location and marginal crustal thicknesses (table 1).

Volcanic addition (*va)* may be estimated from the lithosphere thinning factor γ, where $\gamma = 1 - 1/\beta$, using the adiabatic decompression melt generation model predictions of McKenzie &

Bickle (1988) and Bown & White (1995). For this, a critical thinning factor is defined for the initiation of oceanic crust production, and an initial oceanic crustal thickness; for this study area, values of 0.7 and 7 km were used respectively, consistent with melt production at margins with normal quantities of volcanism. An iterative cycle of gravity inversion to predict crustal thickness, β stretching factor, volcanic addition and lithosphere thermal gravity anomaly is used and rapidly converges (Chappell and Kusznir, 2008). When no sediment thickness data is used within the gravity inversion, the scheme used in this paper produces an upper bound of crustal thickness and a lower bound of lithosphere thinning factor. Where sediment thickness data has been used, basement thickness, defined as the thickness of the crust below the sediment, is determined.

6.3 Crustal Thickness and Lithosphere Thinning Factor Distribution Predicted by Gravity Inversion

Figures 6.3 & 6.4 show predicted oceanic crustal thicknesses of 6 and 13 km for the whole of the western Woodlark from gravity inversion. Across the Moresby Transform in the eastern basin, oceanic crust is predicted to have a thickness ranging between 4 and 7km, several kilometres thinner than that in the western basin. The change in crustal thickness is abrupt across the transform fault. Several inversions using varying ocean age and breakup parameters were processed (table 1). Older ocean isochrons and breakup ages are used in the east since breakup of this part of the Woodlark Basin occurred earlier and seafloor spreading has been active for longer.

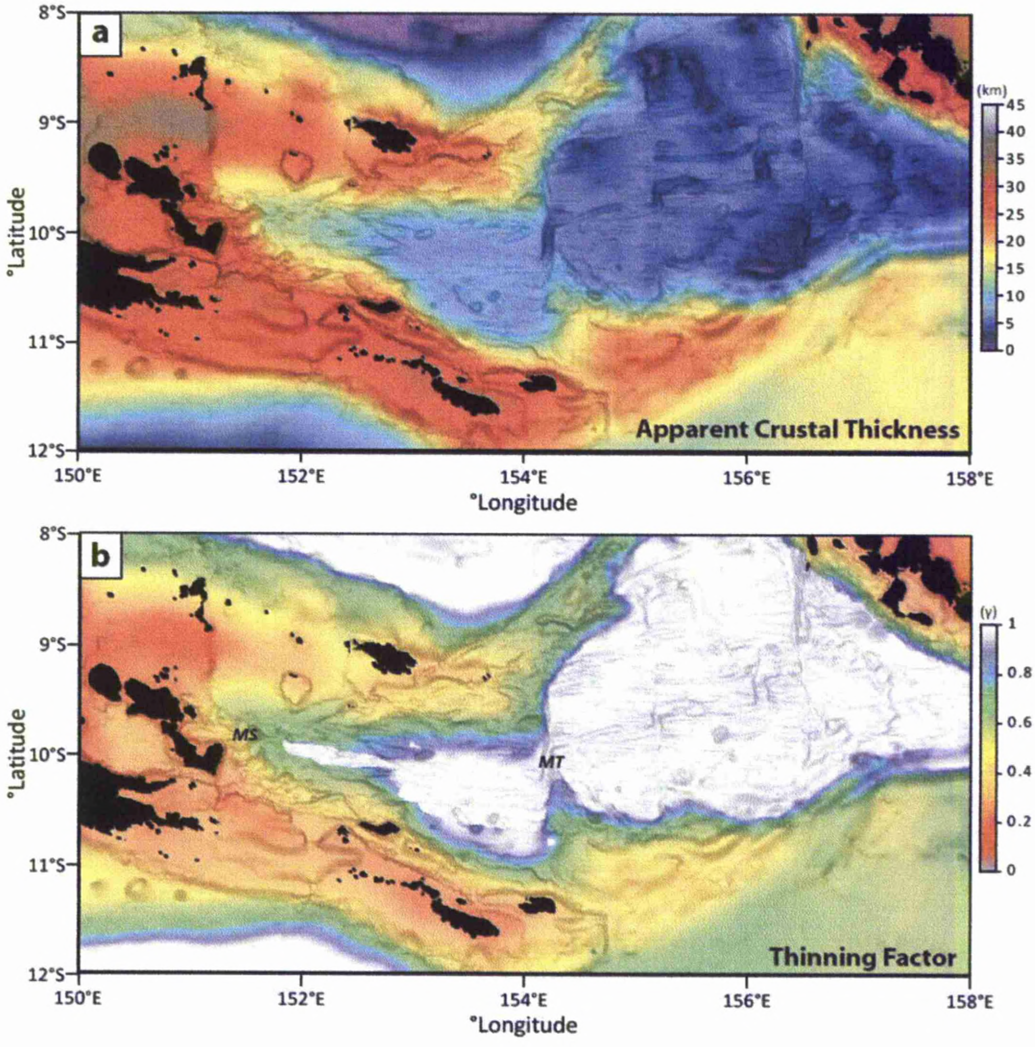

Figure 6.3 (a) Crustal thickness predicted by gravity inversion with a lithosphere thermal gravity correction and (b) corresponding predicted lithosphere thinning factors for the Woodlark Basin. A volcanic addition of 7km has been used for the whole Woodlark Basin. MS is the Moresby Seamount, MT is the Moresby Transform.

Predicted crustal thicknesses of 20 km, approximately 40 km to the west of the propagating tip of seafloor spreading, from the gravity inversion with the lithosphere thermal gravity anomaly correction used are similar to those determined from seismic receiver functions (Ferris et al., 2006) and regionally in agreement with other studies (Abers et al., 2002, Finlayson et al., 1977). The pattern of oceanic crustal thicknesses of the Woodlark Basin, shown in figure 3a, is similar to those determined from isostasy studies (Martinez et al., 1999), however a lack of seismically observed Moho depth prevents the ideal calibration of the gravity inversion. Imaging the Moho in this location has been problematic for all reflection studies in this area. Since the Woodlark is a young and hot system is it possible that a seismically observable Moho has not yet developed meaning that the gravity inversion has not been able to be calibrated against seismic reflection data, only seismic tomography.

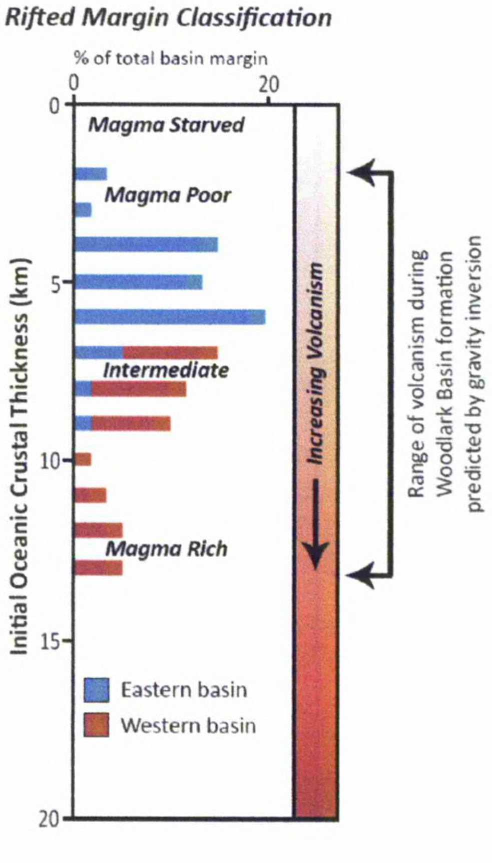

Figure 6.4 Histogram of initial oceanic crustal thickness. Oceanic crustal thickness, derived from satellite gravity inversion, measured every 0.1° longitude approximately 5 km towards the spreading centre from the continental ocean transition from both margins. New nomenclature is used to describe margins that exhibit average thicknesses of initial oceanic crust.

Figure 6.4 shows a histogram of initial oceanic crustal thicknesses derived from gravity inversion. These measurements were spot checked every 0.1° longitude, 5 km towards the spreading centre from a previously determined continental ocean boundary and measurements were taken from both margins. The initial oceanic crust, forming immediately after the rupturing of the continental crust and lithosphere, in the Woodlark Basin, range between 2 to 13 km in thickness (figure 6.4). The eastern basin, separated from the western basin by the Moresby Transform, shows significantly thinner oceanic crust than that predicted in the western basin. Typically, the eastern basin would be considered a *magma poor* volcanic margin since the initial oceanic crust is predominantly 7km or less yet there is no observed exhumed mantle. In the western basin the thickness of first oceanic crust formed, ranging between 7 and 13 km is typical of *intermediate* to *magma rich* rifted margins despite having little magmatism observed prior to continental breakup. The thick initial oceanic crustal thickness in the western basin occurs nearer the present day location of the propagating tip of seafloor spreading. It is evident that the majority of the western basin has thicker initial oceanic crust than seen in the eastern basin, and that this difference is not a gradual change but a sharp difference across the Moresby Transform.

Continental lithosphere thinning factors, shown in figure 6.3b, vary for the conjugate margins either side of the Moresby Transform. In the eastern basin, the margins have thinning factors of approximately 0.6 whereas in the west these are lower, approximately 0.3-0.4. The inversion assumes that the lithosphere was the same thickness prior to extension and subsequent breakup, however it has been proposed that the lithosphere was initially thinner in the east (Martinez et al., 1999), hence these thinning factors could be an overestimation in the eastern basin.

6.4 Discussion and Summary

Labelling the margins of the Woodlark Basin as forming magma poor, suggests that thin initial oceanic crust is expected once the continental lithosphere has ruptured; however, the western Woodlark Basin has thicker oceanic crust than the eastern Woodlark Basin. The cause for this difference in oceanic crustal thickness is unclear. Greater thinning of thicker initial lithosphere in the west (Martinez et al., 1999) has been suggested. Instantaneous rifting, assumed by the McKenzie model (McKenzie, 1978), of different continental lithosphere thicknesses to the same final thickness, would result in the thicker initial lithosphere thinning to a greater degree. The thicker initial lithosphere would undergo a higher degree of thinning and hence the geotherm would be higher than in the east (Martinez et al., 1999). This would provide higher melt production and result in thicker initial oceanic crust in the west; however, it is unclear whether the prerift continental lithosphere of the Pocklington and Woodlark Rises varied along strike.

The role of the surrounding subduction zones could be important. The Trobriand Trough, to the north of the western Woodlark is an active subduction zone (Davies et al., 1984). The true nature and extent of the Solomon plate subducting slab is unknown but it is likely that a subduction zone 200 km to the north of a mid ocean ridge will have a significant effect on the mantle processes that underlie it. There is no evidence to suggest that an actively subducting slab has been beneath the eastern basin during the rifting phase of the formation of the Woodlark Basin; however much earlier the New Britain/San Cristobel subduction zone did have the opposite polarity. As a slab descends into the mantle, it will dehydrate and subsequently hydrate the mantle above (Abers et al., 2006, Karato and Jung, 1998, Karato and Wu, 1993, Tatsumi and Hanyu, 2003). This would decrease the solidus of upwelling mantle and produce more melt. Whilst a subducting slab would have the effect of cooling the surrounding mantle, the hydration of the mantle would still mean an increase of melt that could be generated, enabling the generation of thicker oceanic crust. The proximity of a slab to a spreading centre system will affect the thickness of oceanic crust. In the Lau Basin, thick crust is produced where the spreading centre is close to the subduction zone and then thinner crust is produced further from the slab (Martinez and Taylor, 2003). As in the Woodlark Basin, this change in oceanic thickness is an abrupt transition. Closer to the subduction zone and dehydrating slab, the spreading centre can tap into a hydrated mantle, where the same temperatures will produce more decompression melt than average mantle, whilst further away the spreading centre can no longer access this source of melt and results in the production of thinner oceanic crust.

The formation of the Woodlark Basin is typically thought of a magma poor process, as magnetic evidence from the location of the present day propagating tip of seafloor spreading shows that there is little surface volcanism prior to breakup. The first major evidence of any volcanism are small volcanic cones south of Egum Atoll and two topographic features, interpreted as ring dykes, to the northeast of the Moresby

Seamount (Goodliffe and Taylor, 2007). This surface volcanism has been inferred as the onset of a seafloor spreading system after breakup of the continental lithosphere has occurred. Magnetic, sidescan and seismic data all point to this being the onset of seafloor spreading; however, this region could just as easily be defined as an area of highly thinned continental crust where igneous activity associated with continental breakup has resulted in dykes reaching the surface of the seafloor. The gravity inversion, predicts a crustal thickness of approximately 13 km here, and would classify the present day breakup as marginally magma rich if this is truly the initial oceanic crust; however, sediments are not included within the gravity inversion and their inclusion would predict thinner crust than without sediments included (Gozzard et al., 2009), thus 13 km is a maximum prediction for total crustal thickness. It is possible that these values are not truly representative of the initial oceanic crust since this region could still have a component of remnant highly thinned continental crust within the volcanics. This initial oceanic crust, thicker than 7 km, is typical for the western Woodlark but not for that of the eastern Woodlark where the whole process of rifting, the breakup of the continental lithosphere and the subsequent onset of seafloor spreading has been less volcanic.

The breakup of the palaeo-Papuan Peninsula and subsequent formation of the Woodlark Basin is difficult to classify in terms of margin characteristics, as it does not conform regionally to either the currently accepted views of magma poor or magma rich rifted margin models. However when viewed in terms of margin volcanism and initial oceanic crustal thickness the eastern Woodlark Basin would classify as magma poor and the western Woodlark Basin as slightly more volcanic but not to the extent of being magma rich. Since neither the western nor the eastern basins of the Woodlark Basin conform to the either currently accepted magma poor or magma rich models, a modification to the nonclementure of the classification of rifted margins is devised and used in figure 6.4. Current magma poor margins exhibiting exhumed serpentinised mantle are termed magma starved since the continental crust has ruptured without the production of oceanic crust. The term magma poor is used for a margin where, upon the rupture of continental crust, the formation of thin oceanic crust occurs immediately. In this new system, the western Woodlark would predominantly be considered as an intermediate basin due its initial oceanic crustal thickness, whereas the eastern Woodlark would be viewed as a magma poor basin since thin initial oceanic crust is observed without the exhumation of serpentinised mantle during continental breakup.

6.5 References

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Chapter 7

The Breakup of the Papuan Peninsula and Formation of the Woodlark Basin

This chapter summarises observations from previous chapters and discusses how they aid our understanding of the formation of the Woodlark Basin and the thinning of continental lithosphere leading to continental breakup. It discusses the problems that have been addressed and the new questions that have arisen from this body of work.

7.1 Introduction

The thinning continental lithosphere and crust, leading to continental breakup and seafloor spreading initiation, is a fundamental process in the Wilson plate tectonic cycle. At present, it is not fully understood. The Woodlark Basin provides a natural laboratory where pre, syn and post-breakup processes are all observable at the present time. Studies on depth-dependent lithospheric thinning at the rifting to spreading transition, crustal thickness maps derived from gravity inversion and variations in the volcanism of continental lithospheric rupture have been described in the chapters 4, 5 and 6. The implications of these studies, tied in with other work are now discussed.

7.2 Observations from the Woodlark Basin

The breakup of the Papuan Peninsula and subsequent formation of the Woodlark Basin is poorly understood. The previous chapters have described observations of the following:

i. Evidence for depth-dependent lithosphere thinning prior to continental breakup and the initiation of seafloor spreading.
ii. Observations of regional and local variations in the volcanism of continental breakup during the formation of the Woodlark Basin
iii. The final stages of continental lithospheric rupture and the propagation of the seafloor-spreading centre.
iv. The formation of rifted continental ribbons.

7.2.1 Depth-dependent lithosphere thinning

There is a discrepancy between observations of the stretching of the upper crust and the thinning of the rest of the lithosphere if the thinning of continental lithosphere leading to continental breakup in the Woodlark Basin is assumed to occur depth-uniformly. The thinning factors derived for the whole lithosphere from stratigraphic analyses and the whole crust from gravity inversion are similar, and both greatly exceed the extension that is observed from faulting in the upper crust. This has led to the conclusion that depth-dependent lithospheric thinning, not depth-uniform thinning, is occurring in the continental lithosphere leading to continental breakup. This is different to observations made at intracontinental rift basins where the lithosphere thins depth-uniformly (Kusznir and Karner, 2007, Marsden et al., 1990).

The formation of the Woodlark Basin is different to the majority of other studied rift basins in that the original continental crust, that thinned leading to continental breakup, was a narrow accretion of volcanic arcs, rather than a large continental plate. The study of the stretching and thinning of continental lithosphere at the majority of rifted continental margins, formed by the breakup of a large continental plate, can lead to assumptions about displacement of lower lithosphere that perhaps cannot be simply applied to the Woodlark Basin. These assumptions include the extrusion of lower crustal material or the formation of a landward mountain range. The accommodation of the lower lithosphere, during the thinning of continental lithosphere in the Woodlark Basin, by either of these two modes is hard to justify since it is bounded by oceanic plates to the north and south. Thinned lithosphere and lower crust cannot be displaced beneath this oceanic crust.

Slab steepening of the Solomon Plate, which is being subducted at the Trobriand Trough to the north of the region of continental thinning, has been argued as a mechanism that would accommodate the lower crust and lithosphere when depth-dependent lithospheric thinning occurs (figure 7.1) (Kington and Goodliffe, 2008). A steepening of a slab creates space for the lithospheric material to move into, without the requirement of a horizontal detachment, of infinite length, in the lithosphere. This explanation, when applied to the Woodlark Basin is problematic. A slab can only steepen to a certain degree so a limited amount of space is available. Kington and Goodliffe (2008) state that a change of 45° in the dip of the slab since 8.4 Ma is needed to accommodate the difference in thinning of the upper crust with the rest of the lithosphere, and this would appear to be an excessive change in slab dip. Also, since the whole lithospheric extends at a similar rate to the formation of new oceanic crust at a spreading ridge once breakup has occurred, the question should not be where is the lithospheric thinning being accommodated at depth, but why is there so little upper crust stretching.

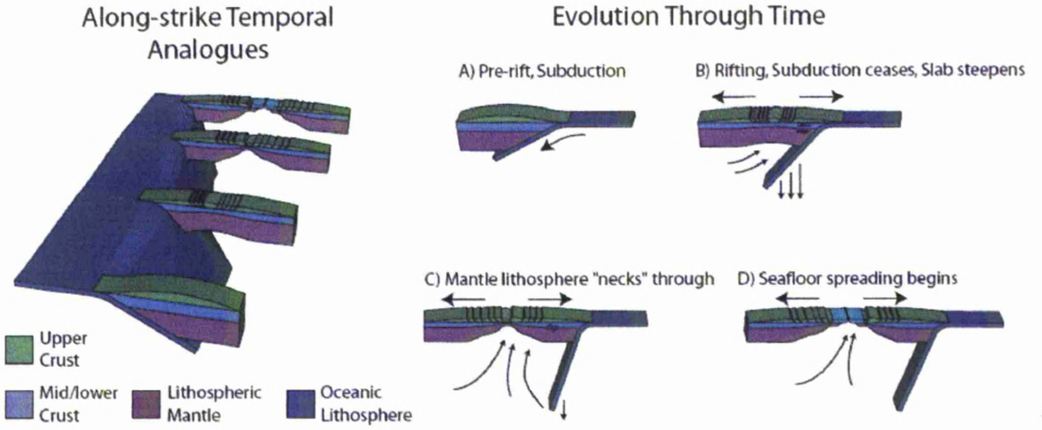

Figure 7.1 (Kington and Goodliffe, 2008) Cartoon illustrating accommodation of preferential extension of the mantle by a steepening in the angle of the Trobriand Slab. Extending the lower lithosphere a by a greater amount than the upper lithosphere requires and explanation for what occurs at the boundaries. This summary figure shows how a steepening of the subducting slab can create the space required to allow a preferential thinning of the lithosphere. A change in slab dip of 45° is required to accommodate the thinning of the lower lithosphere around Moresby Seamount.

Slab steepening could possibly account for depth-dependent lithospheric thinning in the western Woodlark Basin; where there is the Trobriand Subduction Zone to the north, but it cannot explain why depth-dependent lithospheric thinning is observed at other rifted continental margins where there is not a slab to steepen. If it were proven that slab steepening is the mechanism that is driving the preferential thinning of the continental lithosphere, it would imply that the Woodlark Basin is not a suitable region to test other models that explain depth-dependent lithospheric thinning.

7.2.2 Timing of depth-dependent lithospheric thinning in the Woodlark Basin

Observations from the Woodlark Basin allow the opportunity to study continental lithospheric thinning from a pre-rift setting through to breakup and the initiation of seafloor spreading. Most other rifted continental margins, that have been studied, allow only for post-breakup observations. The analysis of the three profiles across the Moresby Seamount (chapter 4) indicates that depth-dependent lithospheric thinning occurs in the thinning of the continental lithosphere, prior to continental lithospheric rupture and the onset of seafloor spreading. Peak upper crustal thinning factors increases towards the propagating tip of seafloor spreading suggesting that in the final stages of continental thinning, immediately prior to continental breakup, the extension of the upper crust becomes focused on the faults close to the rift axis.

A lack of data to the west of the latest spreading centre to form, has meant that only continental crustal lithospheric thinning factors have been determined. Here, the quality of the seismic reflection data is too poor for upper crustal thinning factors to be determined. Therefore, due to a lack of evidence, it is unknown if these margins experienced depth-dependent or depth-uniform continental lithospheric thinning prior to continental breakup.

The D'Entrecasteaux Islands lie ahead of the propagating tip of seafloor spreading in a region of continental lithosphere that would otherwise represent a region where early syn-rift thinning is occurring. The emplacement of eclogite facies rocks in the upper crust within the last few million years has meant that understanding early syn-rift processes in the Woodlark Basin is difficult from present day observations. The thinning of the crust and lithosphere around the D'Entrecasteaux Islands is not suitable as an analogue for the early thinning phase that the region around the Moresby Seamount as the pre-rift lithospheric conditions were different.

7.2.3 Local and regional variation in volcanism during continental breakup

In the Woodlark Basin, the breakup of the continental lithosphere and initiation of a seafloor-spreading centre has exhibited various amounts of volcanism through its history.

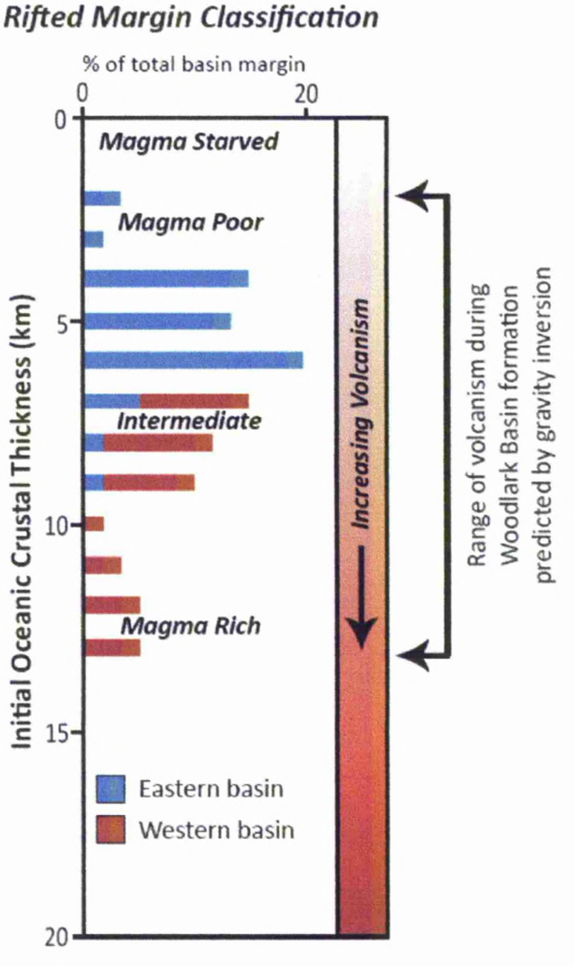

Figure 7.1 Rifted margin classification (RMC) histogram of initial oceanic crustal thickness. Oceanic crustal thickness, derived from satellite gravity inversion, measured every 0.1 degrees longitude approximately 5km towards the spreading centre from the continental ocean transition from both margins.

The eastern Woodlark Basin

The breakup of the eastern Woodlark Basin had little volcanism associated with it. The continental lithosphere was highly thinned prior to any volcanic activity, and in the final stages of continental lithospheric rupture, large grabens formed and were eventually cut by the initiation of seafloor-spreading volcanism at the ocean floor (Taylor et al., 1995). The initial thickness of the first oceanic crust is thin, ranging between 2 and 9 km thickness but predominantly less than typical values of an intermediately magmatic margin. Although labelled as a magma-poor system (Goodliffe and Taylor, 2007, Taylor et al., 1995), neither rifted continental margin is known to exhibit exhumed continental lithospheric mantle, a structure typically associated with such margins.

The western Woodlark Basin

In the western Woodlark Basin, across the Moresby Transform, the breakup of the palaeo-Papuan Peninsula has been significantly different to that of the eastern Woodlark Basin. Here, the Papuan Peninsula was initially thicker than further east, thus it has had to thin to a greater degree prior to continental breakup (Martinez et al., 1999). The Trobriand Trough lies to the north, a subduction zone, which has possibly had an effect on the mantle composition underneath the western Woodlark Basin. The surrounding subduction zones have possibly hydrated the mantle source of the seafloor spreading centres allowing for the generation of more melt at the same temperature than if it was not hydrated. Nonetheless, there is a westward increase in initial oceanic crustal thickness of 2-3 km across the Moresby Transform.

The transform represents the boundary between the eastern Woodlark that is a magma-poor system and the western Woodlark that has had a more volcanic breakup history. The initial oceanic crustal thicknesses range between 7 and 13 km, ranging from average volcanism into thicknesses expected of margins of a more volcanic nature. Where high values of initial oceanic crust have been predicted, it is possible that this value is an overestimation as sediment thickness data was not included in gravity inversion that these values were taken from. Including sediment thickness within the gravity inversion would decrease the predicted thickness of oceanic crust. The caveat to the observation of thicker than average initial oceanic crust to this is that, prior to the onset of this volcanism, there is relatively little volcanism observed in the rifted continental margins, the Pocklington and Woodlark Rises, compared to typical magma-rich rifted continental margins. The initial oceanic crust being formed at the present propagating tip of seafloor spreading is thicker than the average for the rest of the Woodlark Basin, yet immediately west, in the large grabens to the north and south of the Moresby Seamount there is no evidence of any volcanism (Goodliffe and Taylor, 2007).

The Moresby Seamount is a poorly named structure as it is not a volcanic edifice, but rather a large fault horst with a large fault to the north dipping at 28° (Floyd et al., 2001, Goodliffe and Taylor, 2007) and also most likely a large fault bounding it to the south forming the Southern Moresby Graben. This region is in the last stage of continental lithospheric thinning prior to breakup and continental lithospheric rupture.

7.2.4 The rupture of continental lithosphere and propagation of the seafloor-spreading centres in the Woodlark Basin

Proto-ocean basins

The offsets in seafloor spreading centres in the eastern and western basins of the Woodlark Basin are accommodated by different mechanism. In the eastern Woodlark Basin, where continental breakup and the initiation of seafloor spreading occurred over 2 myr ago, the offsets in seafloor spreading centres are accommodated by transform faults. These transform faults have formed to accommodate the relative difference in plate motion with respect to each other. At 80 ka, there has been a reorganization of the spreading centres in the Woodlark Basin due to a migration of the Euler pole of spreading (Goodliffe et al., 1997). Since 80 ka, the change in orientation of the spreading has led to the formation of several new echelon seafloor-spreading centres at the location of the older spreading centres in the eastern Woodlark Basin.

In the western Woodlark Basin, where continental breakup has occurred more recently, the spreading centres are younger. Here there are no transform faults accommodating the offsets in the spreading centres (Taylor et al., 2009). It is unclear how the local differences in plate motion are accommodated. It is possible that the oceanic crust is too weak for a single transform to occur and that the differential motion is accommodated across a series of smaller structures. It is equally plausible that the oceanic crust is too strong for transform faults to have developed and that there is a build up of stress that will eventually result in the brittle failure of the oceanic crust and formation of a transform fault. Whilst the axis of a new spreading centre can be offset to the previous spreading centre when propagating into continental lithosphere, in the western Woodlark Basin transform faults do not develop at the same time as continental lithospheric rupture.

Formation of rifted continental ribbons

The propagation of seafloor spreading centres into continental lithosphere has not been a continuous unzipping of the Papuan Peninsula, but a series of individual spreading centres propagating sequentially. These spreading centres are offset, yet in the western Woodlark, the centres have not yet formed transform faults.

A new seafloor-spreading centre can be abandoned after the initiation of seafloor spreading and leading to a ridge jump and the formation of a new spreading centre. This has occurred at least once in the Western Woodlark Basin and has led to the formation of a rifted continental ribbon, a ribbon of continental crust sandwiched between two strips of oceanic crust. The process of RCR formation can occur when a seafloor-spreading ridge has failed soon after continental rupture and has relocated to the side of it due to the locality of the upwelling divergent flow field or when two spreading centres have overlapped after the propagation of a new spreading centre leading ultimately to the abandonment of one of them.

How does the initiation of seafloor spreading manifest?

In the most eastern part of the North Moresby Graben, the first evidence of volcanism associated with the final stages of continental rupture appears in the form of ring dykes. Figure 7.2 shows a ring dyke that has formed in the North Moresby Graben. It has cut through the low angle Moresby Fault and extruded volcanics on the surface that are observed in the magnetization of the seafloor. Whether these ring dykes are truly the first rupture of the continental lithosphere by a seafloor-spreading centre or volcanism that occurs in the final stage of continental thinning prior to breakup is unknown.

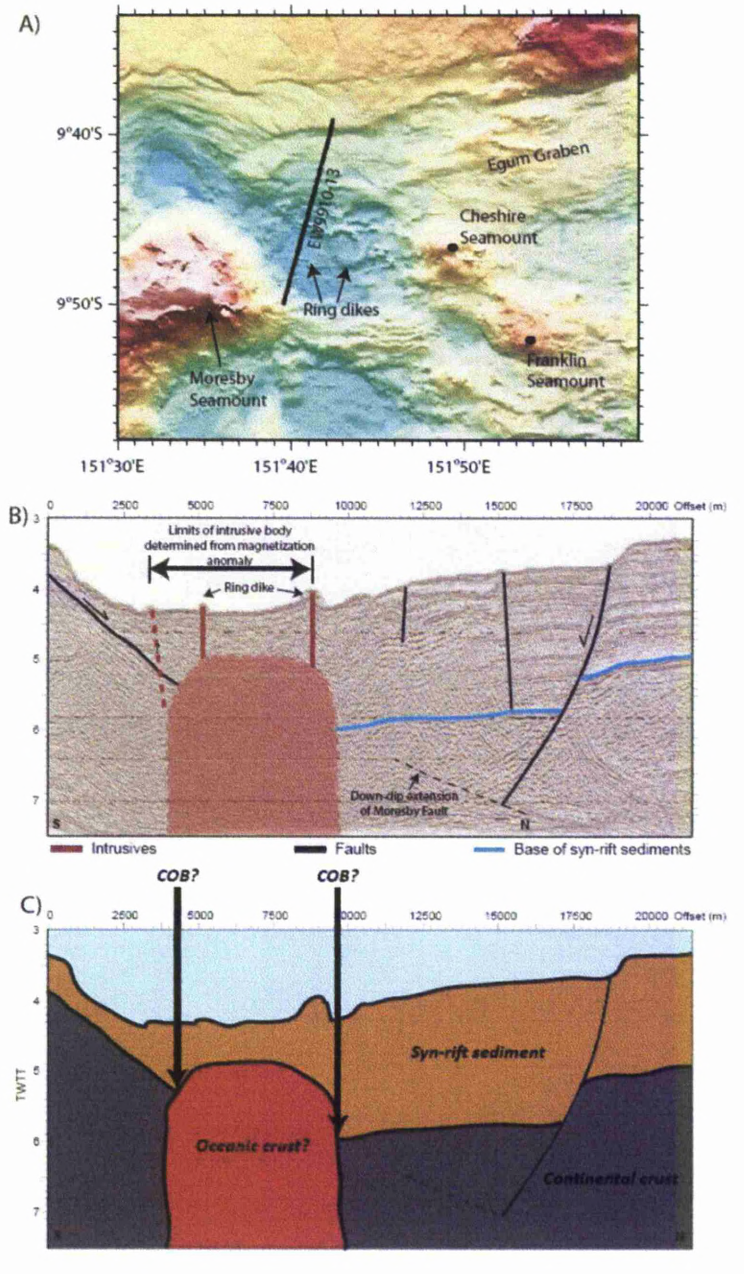

Figure 7.2 (Goodliffe and Taylor, 2007) A) Location of seismic reflection line EW9910-13 to the northwest of Moresby Seamount. B) Interpretation of the seismic line. The currently forming continent-ocean boundary involves igneous intrusions penetrating the shallow-angle normal fault of an asymmetric rift graben. Moresby

Fault is curt by an approximately 5km wide intrusive body that is overlain by subhorizontal sediments that have been cross-cut by a ring dyke. C) Cartoon interpretation of seismic reflection line EW9910-13 showing a volcanic intrusion dividing the continental crust. This is possibly the formation of the continent-ocean transition at the propagating tip of seafloor spreading. It is unknown if this volcanism is the first stage of the development of a true ocean spreading centre or just volcanism associated with the thinning of the lithosphere prior to continental lithospheric rupture.

When a continental breaks up, two continental ocean boundaries are formed, one on each margin. The initiation of a spreading centre separates the highly thinned continental crust with oceanic crust, thus creating two continent-ocean boundaries parallel to each other separated only by a narrow strip of volcanism (figure 7.2c).

It is unclear whether the volcanism in the eastern part of the North Moresby Graben is the initiation of a new spreading centre as heat-flow measurements do not suggest that this is the area of highest lithospheric thinning, as the Southern Moresby graben is three times hotter than the northern graben (Goodliffe et al., 2000). It is feasible that continental breakup could occur to the north of Moresby Seamount, fail and jump south forming a new rifted continental ribbon (figure 7.3). The other case is that the surface volcanism here is solely volcanism, due to the regional thinning of the lithosphere, leading to a surface extrusion and is not the first stage of a developing spreading ridge.

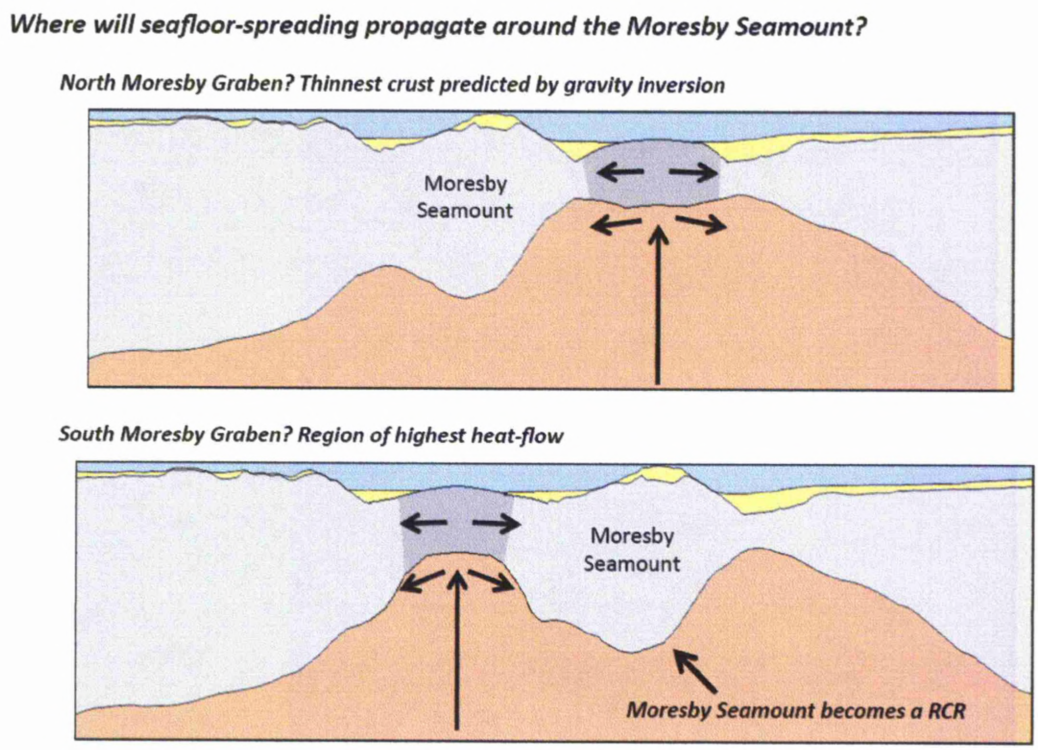

Figure 7.3 Diagram showing the possible locations of the formation of a seafloor-spreading centre in either the NMG or the SMG. The thinnest crust is predicted to be in the NMG from gravity inversion yet the highest heat-flow measurements are in the the SMG. If breakup occurs in the SMG, the Moresby Seamount becomes a rifted continental ribbon.

7.3 A lithospheric velocity discrepancy?

A possible divergence velocity discrepancy between the upper crust and the lower lithosphere, as discussed in the latter parts of chapter 4, has implications for how continental plates diverge as they undergo lithospheric thinning and continental breakup. The nature of depth-dependent lithospheric thinning requires different levels of the lithosphere to move relative to each other, however in distal regions, away from the rift axis the lithosphere moves uniformly. Within models that explain depth-dependent lithospheric thinning, there can be no horizontal detachment extending to infinity, as this is geologically implausible.

Pole of rotation

The Woodlark Basin is a v-shaped ocean basin that is opening around a pole of rotation at approximately 144°E, 9°S (figure 7.4). The further away from this pole of rotation, the greater the velocity of plate divergence. In the eastern Woodlark Basin at 156°E oceanic crust forms at the rate of 67 km per myr. Closer to the Euler pole at the Moresby Transform, oceanic crust forms at the rate of 53 km per myr and closer still, immediately east of the Moresby Seamount oceanic crust will form at the rate of 36 km per myr. The location of the Euler pole of rotation is derived from seafloor-spreading centre geometry and relative velocity. This work assumes that the Euler pole has not significantly changed location.

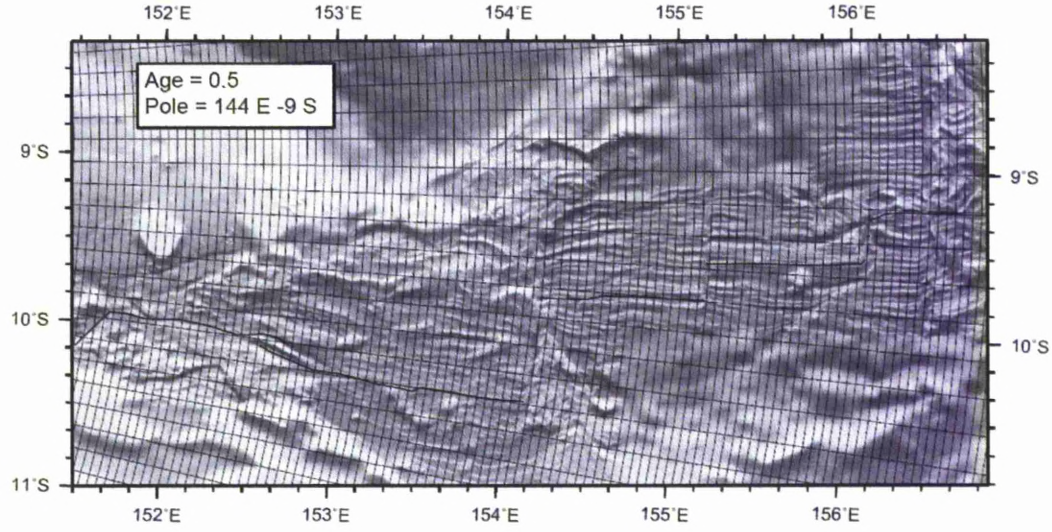

Figure 7.4 The location of the Euler pole of rotation by which the opening of the Woodlark Basin conforms to Goodliffe (1998).

Observed extension discrepancy

The Palaeo Papuan Peninsula was formed from collisional tectonics (Davies et al., 1984) and its continental lithosphere would have had a finite width in the order of 100-200 km. The present extension of the Papuan Peninsula at the rifting to spreading transition is approximately 300 km wide (figure 7.5). Euler pole estimates of plate divergence, derived from seafloor spreading velocities in the Woodlark Basin, estimate a plate divergence of 36 km per million years at the Moresby Transform (Goodliffe, 1998, Kington and Goodliffe, 2008, Taylor et al., 1995). Seafloor spreading rates in the Woodlark Basin have been seemingly constant over the past 6 million years. At that rate, assuming that the divergence of the Pocklington and Woodlark Rises has also been constant since the onset of rifting at 8.4Ma, the rifted continental crust would have been extended by approximately 300 km. This is a discrepancy of between 100 and 200 km of that observed from the width of the continental crust and a discrepancy of approximately 200 km to what is observed from fault extension in the upper crust across the whole of the Papuan Peninsula.

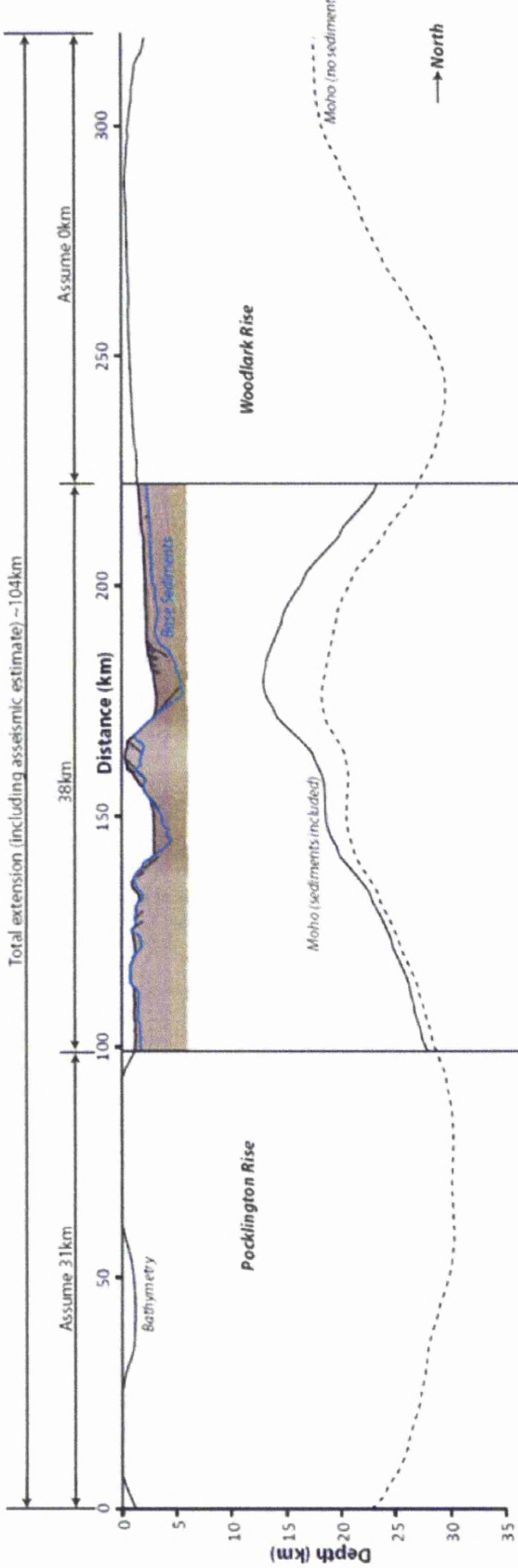
Total extension (including asseismic estimate) ~104km
Assume 0km
38km
Assume 31km
Distance (km)
Depth (km)
Woodlark Rise
Pocklington Rise
Moho (no sediments)
Moho (sediments included)
Bathymetry
Base Sediments
Thinning Factor
Whole lithosphere
Whole crust
Upper crust
North

Figure 7.5 Summary diagram for the MW9304_50 profile with extended bathymetry including the Pocklington and Woodlark Rises. Seismic reflection data, where available is shown with base sediment and fault interpretations. A Moho derived from a satellite gravity inversion is shown, and includes a correction for sediment thickness where available. Total extension from fault heave analysis is shown and estimated for the margins using values from Kington and Goodliffe 2007. Aseismic fault extension is accounted for by including an extra 40% of extension for the upper crustal estimate. Thinning factors for MW9304_50 derived in chapter 4 are shown. The total width of the profile is 320 km, divergence estimates from Euler pole model derived seafloor spreading rates at the Moresby Seamount are over 300 km since 8.4 Ma, thus highlighting the discrepancy between upper crustal extension and whole lithospheric extension.

There is an observed discrepancy between the divergence velocity of pre-breakup continental crust and the divergence velocity post-breakup. It is possible that this is not a true observation as:

i. *The determination of the seafloor-spreading rates could be incorrect.* This is unlikely since the Woodlark Basin is still forming new oceanic crust thus the magnetic modelling of anomalies is straightforward and the rate of production clearly deduced (Taylor et al., 1995).

ii. *The spreading rate, thus divergence velocity could have increased with time.* There is no evidence for this, since the magnetic anomalies are known, spreading velocities are also known for the past 6 million years and there is no evidence to suggest a difference in the spreading rate over time (Goodliffe, 1998).

iii. *The upwelling divergent flow field associated with seafloor spreading has a different divergence rate than that of the upper crust before continental breakup occurs.* This model assumes that seafloor-spreading rates are representative of the divergence of the plates. This model could also explain the observation of depth-dependent lithospheric thinning prior to breakup.

The model introduced in figure 7.6, at first glance, contradicts both the Euler pole and rigid plate models (Bullard et al., 1965). This model assumes that, before breakup and seafloor spreading have occurred, the upper crust diverges at a different rate to the rate at which it diverges after breakup. If the divergence rates were different pre and post breakup, then the Euler pole model would not work simultaneously for the both regions of pre-breakup and post-breakup crust. Instead, whilst the pole at which divergence revolves would remain the same for both areas, the divergence rate of both could not be deduced from one another. The only relationship, from observations, that holds true is:

$$D_{pre} \leq D_{post}$$

Where D_{pre} is the divergence velocity of the two plates prior to breakup and D_{post} is the divergence velocity of the two plates after breakup. For all post-breakup regions in the same system, a Euler pole model can estimate the velocity of seafloor spreading at any point in the system if the location of the Euler pole is assumed fixed and the velocity of seafloor spreading known at a single point at a known distance away from the pole. It is plausible that the Euler pole model, when determined from seafloor spreading velocities does not apply to regions of thinning continental crust, ahead of the propagating tip of seafloor spreading.

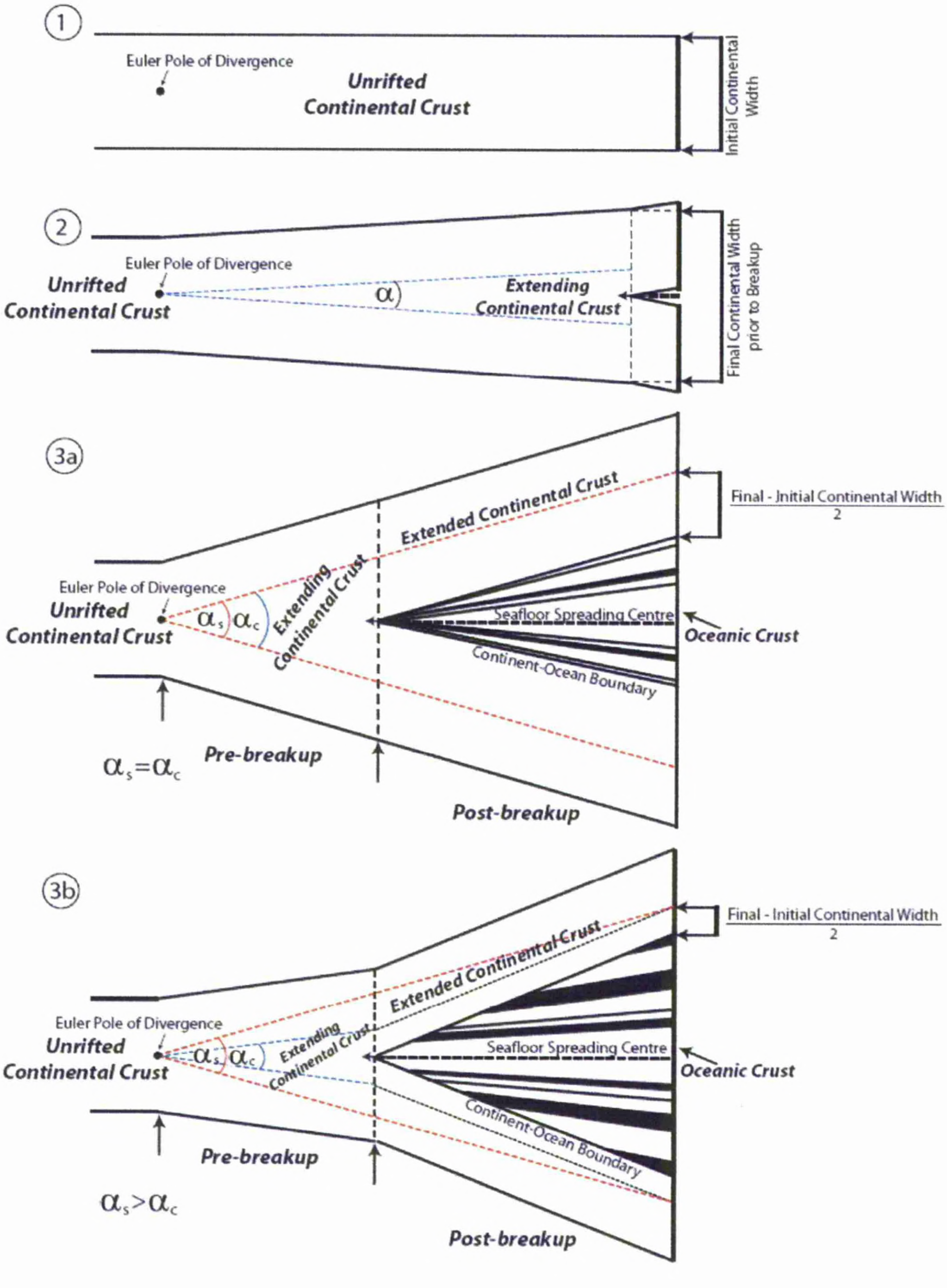

Figure 7.6 Diagram showing the implications to the Euler pole model if there is a discrepancy between pre and post-breakup plate divergence. 1. Pre-rift finite width of continental crust. 2. The rifting of continental crust to its final width where continental crust is continuous, but highly thinning, immediately prior to breakup and the onset of seafloor spreading. 3a. This model does not have a divergence discrepancy and the velocity of a point remains constant before and after breakup has occurred. To match the amount of divergence that occurs in model 3b, the

width of rifted continental crust is wider. 3b. This model exhibits the divergence discrepancy with an increase in divergence rate after continental breakup has occurred. A single Euler pole model does not work for the whole system. Both models 3a and 3b assume that continental breakup occurs after certain amount of thinning. The degree of thinning is higher in 3a in order to conform to a single Euler pole model.

A divergence velocity discrepancy has to be accommodated in the surrounding plates, as they are assumed rigid. The tip of the propagating seafloor spreading centre should be the point at which the divergence velocity changes, the Nubara Transform Fault seemingly extends into the North Moresby Graben. It is feasible that this fault accommodates the different divergence velocities. There is no obvious accommodating structure in the Pocklington Rise, it is reasonable to assume this rigid, and that all the divergence discrepancy is accommodated in the Woodlark Rise through the Nubara Fault and the Trobriand Subduction Zone. It is unknown whether the Nubara Fault and Trobriand Subduction Zone accommodate this divergence velocity discrepancy, and at present, it is purely speculation.

A difference in divergence velocity and the response of the upper crust is shown in the model in figure 7.6 Assuming depth-uniform thinning of the lithosphere, the divergence velocity of the upper crust will be the same as that of the rest of the lithosphere and is shown in stage 3a (figure 7.6). However, if the divergence velocity of the lithosphere is greater than that of the upper crust, the lithosphere will be thin depth-dependently, and upon continental breakup, the plate divergence velocity will increase locally, this is shown in stage 3b (figure 7.6). This model requires the assumption that lithospheric divergence velocity during continental thinning is the same as the formation of new oceanic crust after continental breakup has occurred. The Euler pole model, whilst practical in the oceanic domain, cannot be applied to determine divergence velocity in pre-breakup continental crust in the Woodlark Basin.

A change in the local divergence velocity upon the rupturing of continental crust and initiation of seafloor spreading

When thinning continental crust and lithosphere eventually ruptures and seafloor spreading initiates, the mode by which plate divergence is accommodated changes (figure 7.7). In thinning crust and lithosphere, the divergence of the plates is accommodated across a pure shear width where inside the lithosphere is extending and outside the pure shear width it is not. Only at the edge of the pure shear width will the velocity equal that of the plate divergence velocity. Once seafloor spreading has initiated, all divergence is accommodated by the formation of new oceanic crust and not by further extension of the lithosphere. It is unclear how the separation of the plates is accommodated immediately prior to continental breakup, it is plausible that the pure shear width gets progressively narrower until seafloor spreading initiates. This model, to show the difference in divergence accommodation, assumes that there is no divergence velocity discrepancy (i.e. plate divergence velocity is the same pre- and post breakup).

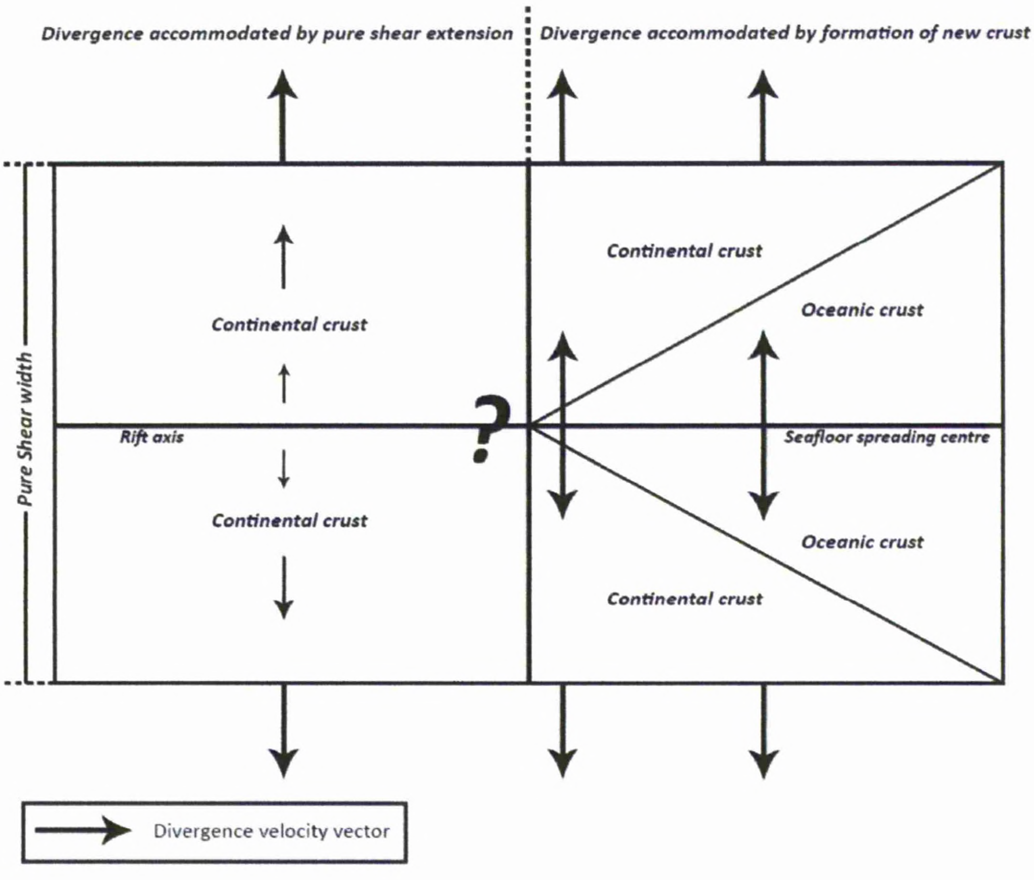

Figure 7.7 Model of how the plate divergence velocity is accommodated in pre-breakup and post-breakup settings. Pre-breakup the divergence velocity is accommodated by extension of the lithosphere across a pure shear width, post-breakup the plate divergence is accommodated by the formation of new oceanic crust and not by extension of the lithosphere. The question mark represents the region immediately prior to continental breakup and the initiation of seafloor spreading; here, the mechanism by which the accommodation mode changes from extension to the formation of new crust is at present unknown.

7.4 How does the continental lithosphere of the Papuan Peninsula thin prior to breakup and the formation of the Woodlark Basin?

7.4.1 Introduction

The nature of continental thinning, leading to continental breakup, is poorly understood. Depth-uniform thinning is known to occur at intra-continental rift basins, but these rift systems do not thin to the degree where the continental lithosphere ruptures and seafloor spreading initiates. At numerous rifted continental margins, including the Woodlark Basin, there is a discrepancy between the amount of extension observed from faulting in the upper crust and the amount of thinning predicted from subsidence analysis for the whole lithosphere (Baxter et al., 1999, Davis and Kusznir, 2004, Driscoll and Karner, 1998, Kusznir and Karner, 2007, Roberts et al., 1997). This has led to the hypothesis that the lithosphere thins depth-dependently prior to continental lithospheric rupture. Several mechanisms could account for the differential thinning of the continental lithosphere prior to continental breakup during the formation of the Woodlark Basin. These include:

- An upwelling divergent "*ocean-ridge*" flow operating at depth underneath the thinning lithosphere and crust (figure 7.8)
- A buoyancy driven thinning of the lower lithosphere
- Slab steepening at the Trobriand Subduction Zone
- Preferential pure shear thinning of the lower crust and lithosphere

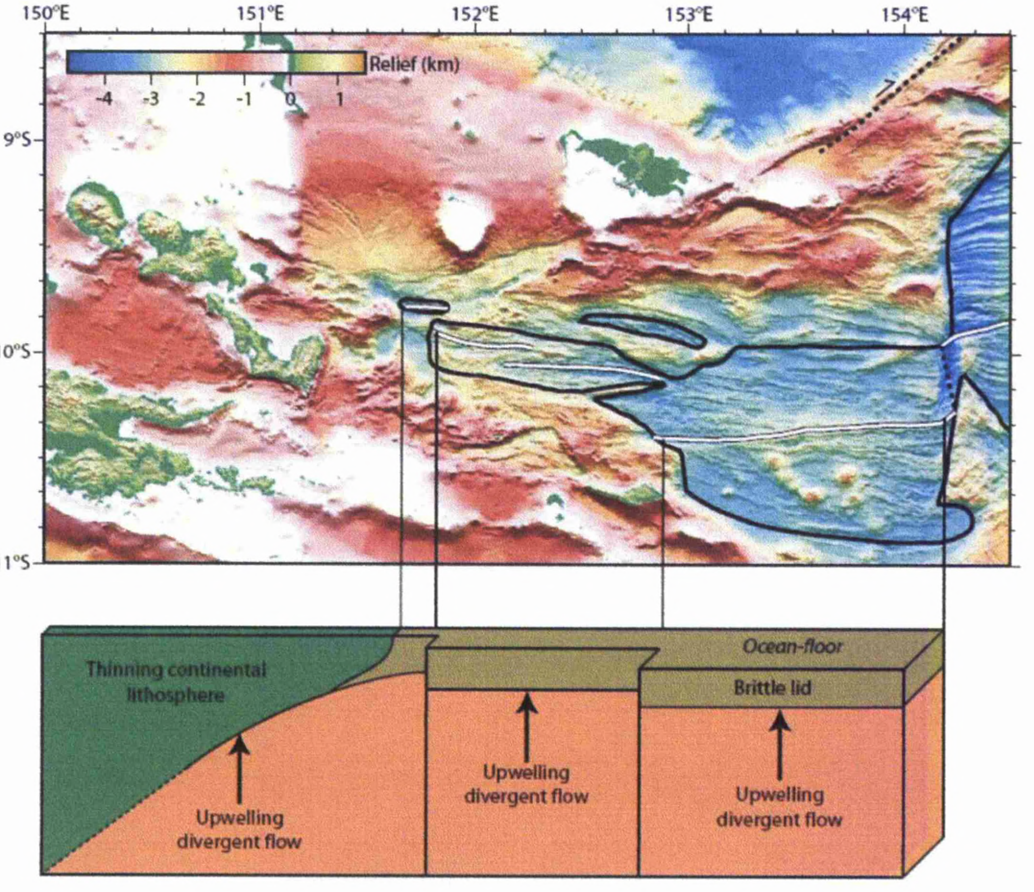

Figure 7.8 Cartoon illustrating a dip line of how upwelling divergent flow might occur in a propagating seafloor spreading system prior to and after the rupture of continental lithosphere. The black line represents the previously determined COB (Taylor et al., 1995).

Forward modelling of different lithospheric deformation modes has been undertaken to try to understand how the lithosphere is thinning during the formation of the Woodlark Basin. These models have been run using a programme called SfMargin, a forward modelling programme that allows the testing of various different lithospheric thinning modes against observational data. Four different deformation modes have been examined, they are:

i. Pure shear *"McKenzie style"* thinning

ii. Decoupled pure shear thinning

iii. Pure shear and buoyancy induced thinning

iv. Pure shear and a *"seafloor-spreading"* upwelling divergent flow

Various models of lithospheric deformation and thinning have been trialled to explain the observed water loaded subsidence observed at the rifting to spreading transition in the Woodlark Basin. 2D models were calibrated against a north-south profile of bathymetry and water loaded subsidence where available, coincident with seismic line MW9304_70, and extended further to the north and south to include the Woodlark and Pocklington Rises (figure 7.9). Water loaded subsidence was determined from the flexural backstripping of stratigraphy to a known palaeobathymetry indicator (chapter 4). The Pocklington Rise possibly has carbonate build ups, in the form of coral reefs; the modelling of basin geometry assumes that the most southern high (at 120km, not 200 km – figure 7.9) is the edge of the basin.

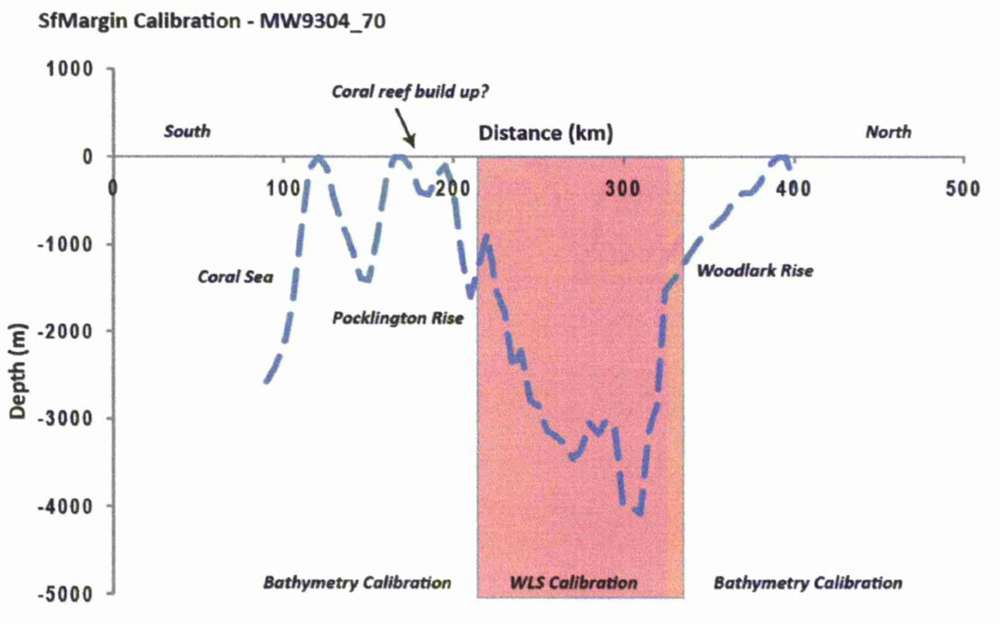

Figure 7.9 Water-loaded subsidence profile of seismic line MW9304_70 and bathymetric extensions to the north and south. Models of continental thinning to form the Woodlark Basin are calibrated against this profile. In the southern margin, coral reefs have built up and are assumed not to be part of the original continental crust.

Model parameters

The various models of different lithospheric thinning modes all assume initial crustal thickness of 40 km and a whole lithosphere thickness of 125 km. No volcanic addition has been assumed. They assume that each layer of the lithosphere is homogenous and that the pre-rift lithosphere extends to infinity. Where sensitivities have been used, these are clearly shown.

7.4.2 Pure shear, depth-uniform thinning models

Intra-continental style, pure shear lithospheric extension has been tested to see if it can explain the observed water loaded subsidence. Two different models were trialled; the first model (figure 7.10) assumed that the total observed extension in the upper crust is representative of the extension of the whole lithosphere. 38km of extension has been measured from fault heave summation across seismic reflection line MW9304_70 (chapter 4). The profile has been extended to the north and south, using estimates of upper crustal extension, including a 40% estimate of sub-seismic faulting, from Kington and Goodliffe (2008). Across the whole profile, 104 km of upper crustal extension is assumed (figure 7.5) and is applied over a pure shear width of 300km. This extension takes place over 8myr at a constant rate.

Results

This model cannot account for the observed water loaded subsidence profile (figure 7.10). The basin geometry is too wide and too shallow. The lithosphere has not thinned to the point where continental breakup and seafloor-spreading initiation is imminent. Since only pure shear thinning of the lithosphere is applied, the thinning factors for all levels are the same and peak at 0.5 γ; this is lower than predicted thinning factors for the whole crust and whole lithosphere derived in the chapter 4.

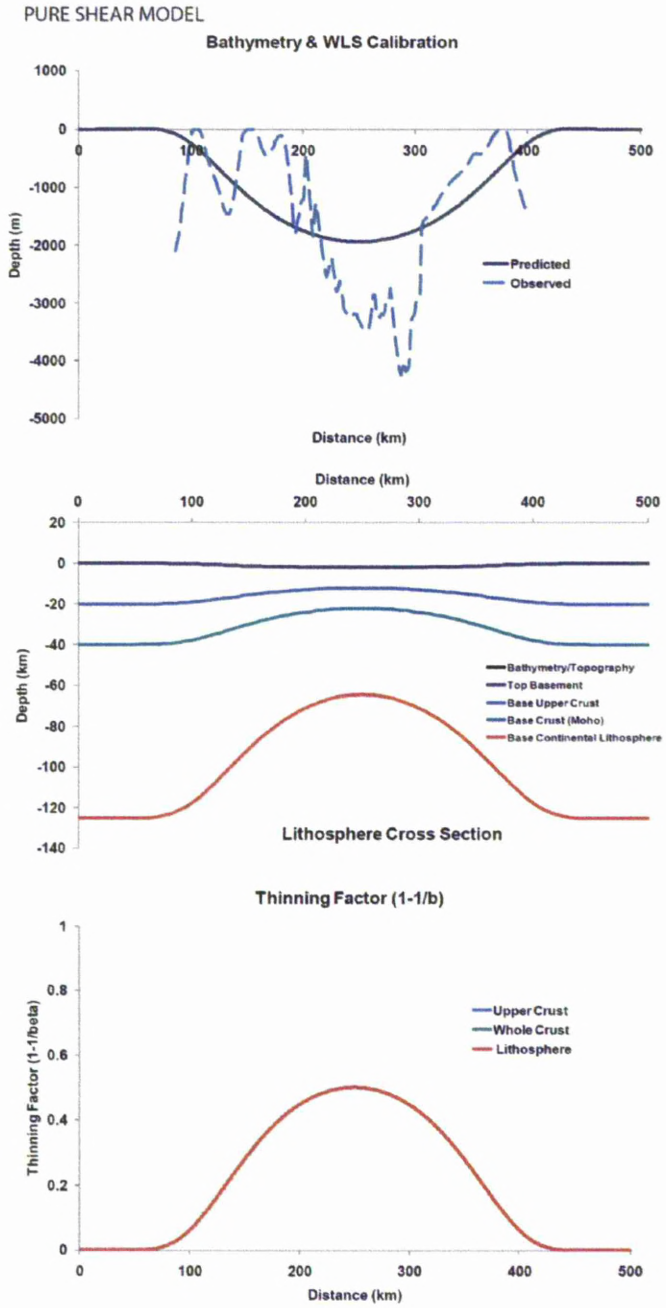

Figure 7.10 A model of uniform pure shear thinning of the lithosphere. Assuming 104 km extension across a 300 km pure shear width, the model does not calibrate against observations from the Woodlark Basin.

A second model of pure shear lithospheric extension is the best-fit pure shear extension model of basin geometry to water-loaded subsidence observations (figure 7.11). In order to gain an approximate fit, two deformation events are required. From 8 Ma to 3.5 Ma, the model extends the lithosphere by 72 km over a pure shear width of 210 km. At 3.5 Ma, the rift axis is then shifted to the north by 15 km and the pure shear width focused to 50 km. In the final 3.5 myr, the 50 km pure shear width is extended by 33 km. A total of 105 km extension has been applied since the onset of continental lithospheric thinning.

Results

This model approximately calibrates against the water loaded subsidence profile of MW9304_70 (figure 7.11). The focused second event of pure shear lithospheric deformation reaches similar WLS depths to what is observed in the Woodlark Basin; however, it is not as wide as the observed deep WLS target. Whilst, it can be interpreted that the WLS profile of line MW9304_70 can be explained through depth-uniform thinning of the lithosphere alone, in order to form the focused deep in the profile, over 60 km of extension is needed to form this geometry. This is over 200% more than what is observed from seismic data in chapter 4.

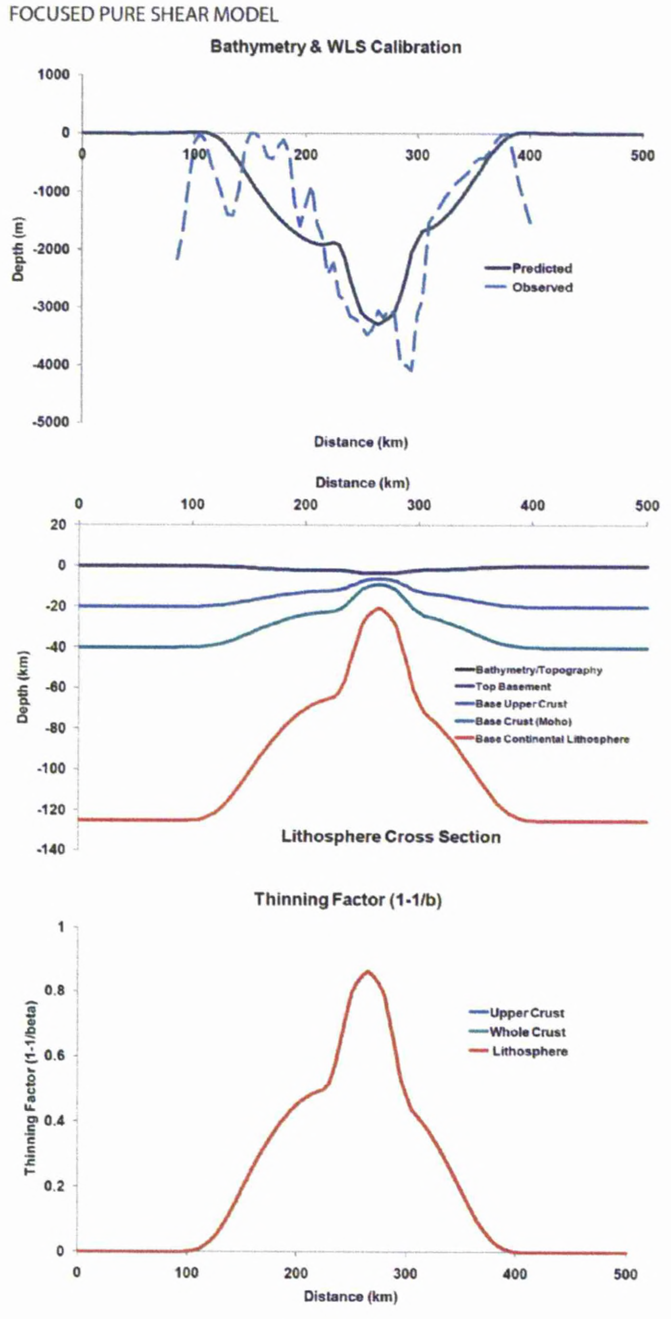

Figure 7.11 A model of pure shear lithospheric deformation. This model has two deformation events, and initial wide pure shear event followed by a more focused event. Whilst this model calibrates against water loaded subsidence, over 60 km extension occurs in the deepest region and this is not observed in seismic.

7.4.3 Decoupled pure shear model

A decoupled pure shear mode of lithospheric thinning was trialled assuming a total extension of 105 km in 8 myr (figure 7.12). The model assumes a one constant event for the 8 myr duration. The continental lithosphere is decoupled at 10 km. The top 10 km thins over a pure shear width of 300 km and extends by 104 km. Below 10 km in the lithosphere, the lithosphere also extends by 104 km in 8 myr; however, this thinning is accommodated over a pure shear width of only 50 km.

Results

The basin geometry formed by the model is a poor fit with the observed WLS profile (figure 7.12). The predicted WLS, whilst reaching the observed depths of WLS, is too wide at depth. There is also a component of thermal uplift. This is due to the thinning of the upper crust and the amount of thermal perturbation at depth due to the high amount of extension in the lower lithosphere. It is unclear how thermal uplift in the Woodlark Basin is accommodated, however, no such topographic feature is observed. This model exhibits depth-dependent lithospheric thinning of the continental lithosphere. The thinning factors for the upper crust, the whole crust and the whole lithosphere are 0.6, 0.8 and 0.95 γ respectively.

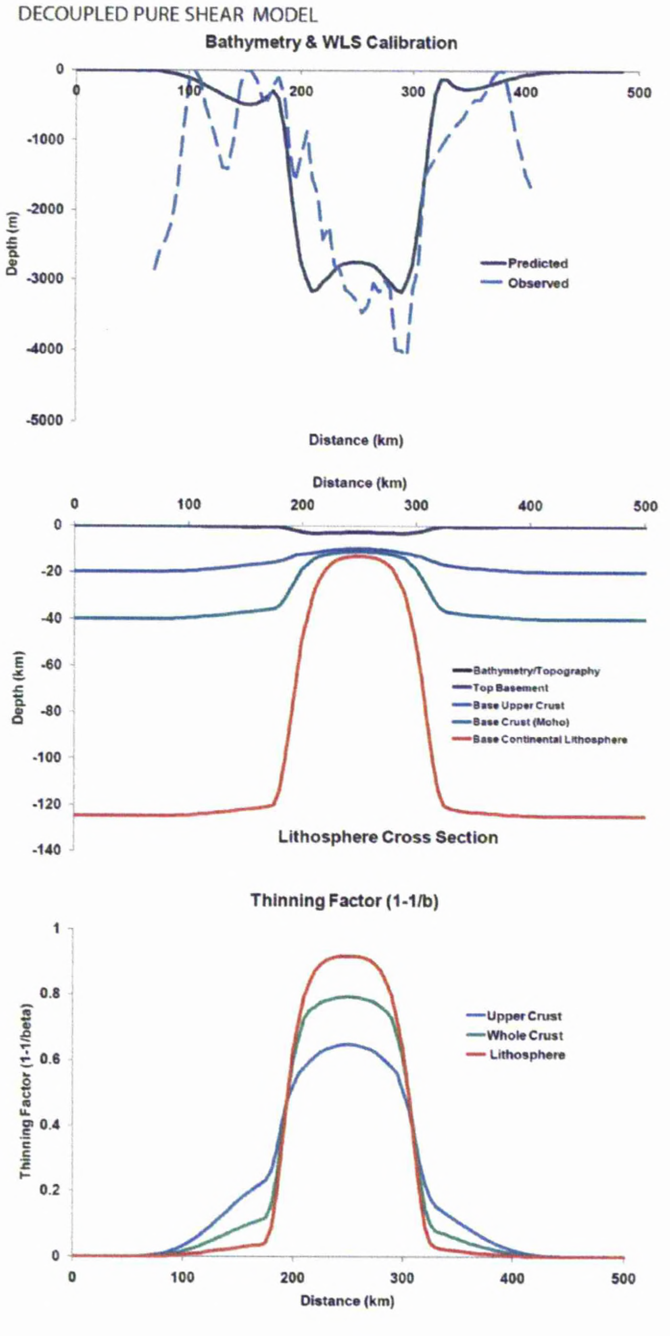

Figure 7.12 A decoupled pure shear model of the formation of the Woodlark Basin. 104 km of extension is accommodated in the top 10 km over a pure shear width of 300 km but only over 50 km below a depth of10 km.

7.4.4 Buoyancy induced thinning model

This model assumes that, after an initial 6 million years of pure shear lithospheric extension, the lithosphere thins due to a buoyancy (Stokes flow) induced thinning of the lithosphere (figure 7.13). The initial 6myr of extension is depth-uniform pure shear extension, similar to observations from intra-continental rifts. 90 km of extension is accommodated over a pure width of 300 km. The final 2 myr of lithosphere deformation and thinning is modelled as the top 15 km of the crust deforming by 15 km of pure shear extension with a more focused pure shear width of 100 km. Below 15 km, the lithosphere is modelled as thinning due to a upwelling buoyant sphere that has a cut off depth of 15 km.

Results

This model generates a poor fit of predicted bathymetry and WLS. The whole lithosphere, shown in figure 7.13, has not thinned to the point of rupture; however, thinning factors in excess of 0.9 γ suggest that lithospheric rupture would be imminent if continental lithospheric extension continues. This model exhibits depth-dependent lithospheric thinning of the continental lithosphere. The upper crust thins less than the rest of the lithosphere due to the different deformation mode of thinning that it undergoes.

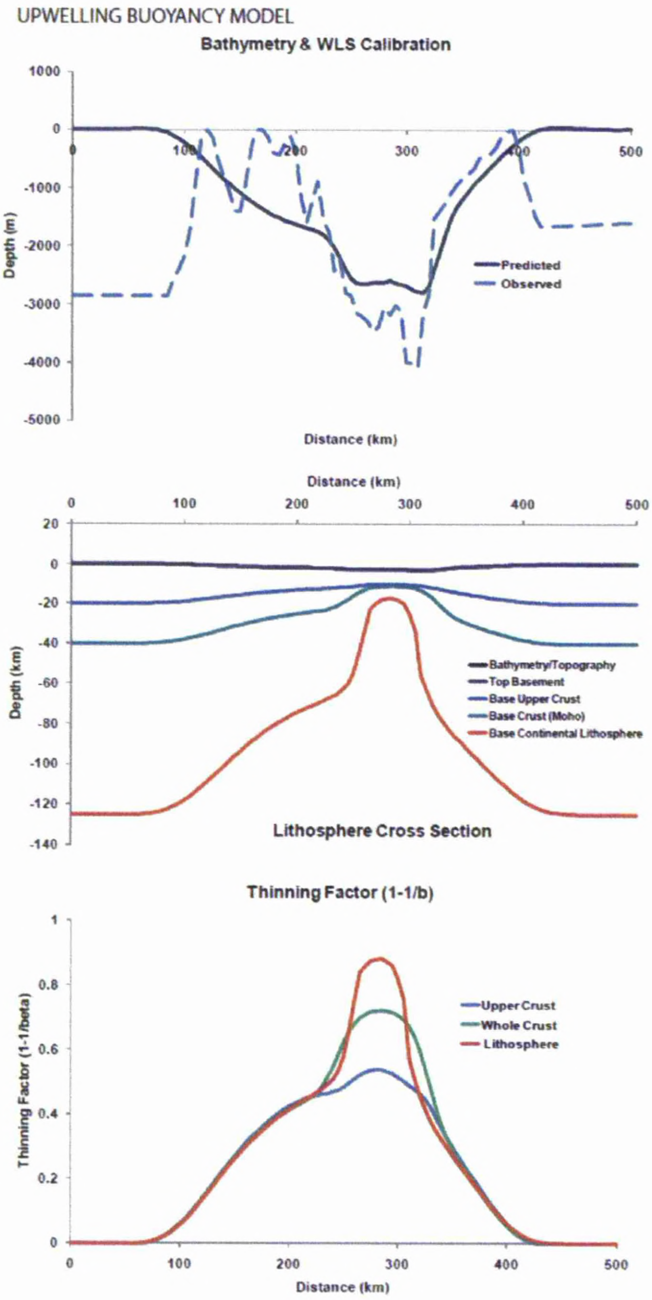

Figure 7.13 An upwelling buoyancy model of lithospheric thinning. An initial pure shear event thins the lithosphere depth-uniformly, but is subsequently followed by an upwelling of lower lithospheric material that induces depth-dependent lithospheric thinning..

7.4.5 Upwelling divergent flow field model

After an initial 6 million years of pure shear thinning of the continental lithosphere where 75 km of extension is accommodated over a width of 200 km, an upwelling divergent flow field is modelled as thinning the lower lithosphere. This flow has a rift axis 15 km to the north of the axis of the earlier pure shear event. This is required to produce slight asymmetry of the margins. This upwelling divergent flow is modelled as having a cut-off depth of 8 km, similar to that expected at ocean ridges where the top 8 km is assumed to deform in a pure shear manner, consistent with models and observations of oceanic crust formed at slow spreading ocean ridges (Cannat, 1996). With upwelling divergent flow occurring at depth, the upper crust thins due to pure shear deformation. Over the whole duration of continental lithospheric thinning, the upper crust extends by 100 km, similar to observations in figure 7.5.

Results

In order to gain a good fit with the WLS profile, this model needs to thin the lithosphere to the degree where seafloor spreading is occurring under a brittle lid of upper crust. This model exhibits depth-dependent lithosphere thinning with thinning factors peaking at 0.6, 0.8 and 0.95 γ for the upper crust, the whole crust and the whole lithosphere respectively.

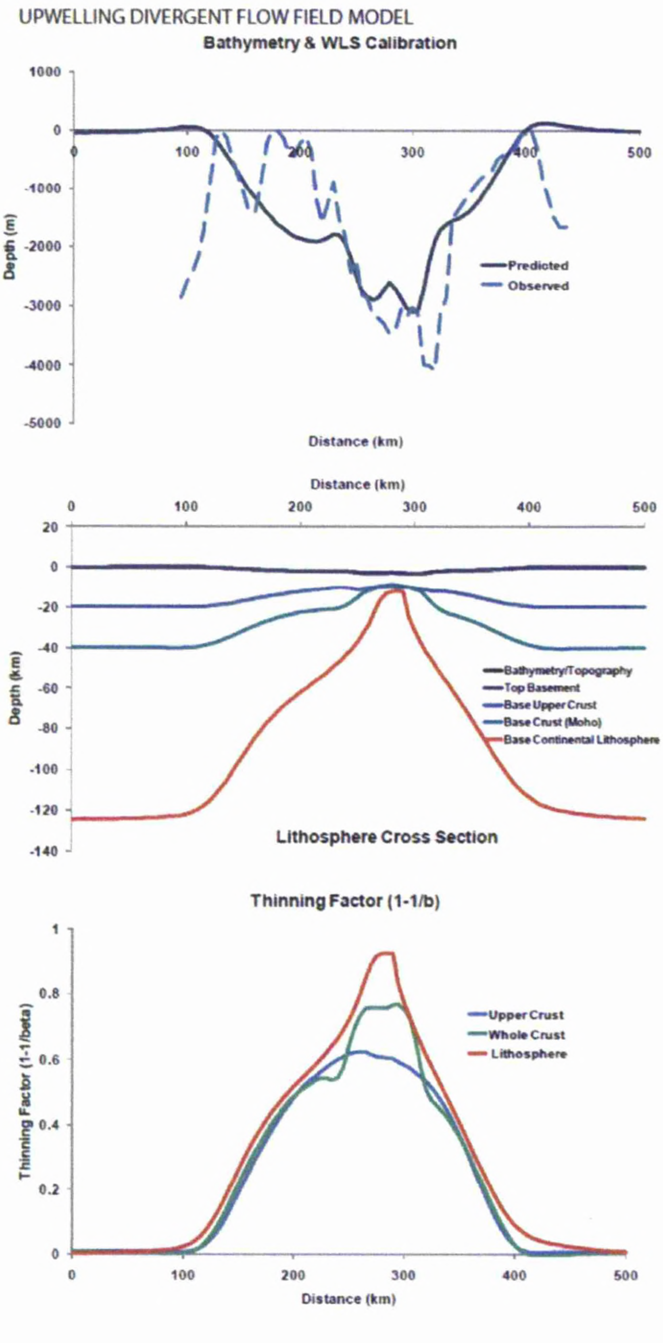

Figure 7.14 An upwelling "seafloor-spreading" divergent flow field model of the formation of the Woodlark Basin. Initially the continental lithosphere thins via a pure shear deformation mode. At 2 Ma an upwelling divergent flow field initiates at depth and starts to thin the lower lithosphere.

Review of lithospheric thinning models when applied to the Woodlark Basin

Several of the models of depth-dependent continental lithospheric thinning approximately fit the observational data; however, depth-uniform lithospheric thinning of the continental lithosphere alone cannot model the observed water-loaded subsidence near Moresby Seamount unless it is focused across a narrow width. The amount of extension required, across a narrow width, to gain a fit is not observed in upper crustal faulting. The continental lithosphere of the Papuan Peninsula thins differently to an intracontinental rift. There is a fundamental difference between rift basins that subsequently cease and those that breakup to form a new oceanic basin. It is possible that depth-dependent lithosphere thinning only happens in the latter stages of continental thinning in rift systems that lead to breakup and that the initial thinning of the continental lithosphere is depth-uniform.

The palaeo Papuan Peninsula, prior to continental thinning, was initially sub aerial, yet both the Pocklington and Woodlark Rises are now predominately under water. This suggests that all the initial continental lithosphere and crust that constituted the palaeo Papuan Peninsula has thinned. Evidence from gravity inversion (chapter 5), suggest that the bulk of the margins have experienced a thinning factor between 0.5 and 0.8 γ; however this might be an overestimate as the initial thickness of the crust further east is poorly constrained and likely to be less than the initial pre-rift crust around the Moresby Seamount. The thinning of all the initial continental lithosphere is a major difference to other rifted margins. Other rifted margins, such as the Brazilian - sub-Sahara African margins, were once part of a large continental plate where only a finite width of the continental lithosphere experience thinning leading to continental breakup. This may make the formation of the Woodlark Basin uniquely different and therefore a poor analogue to apply to the formation of other margins.

There is no known hinterland uplift or exhumed sub-continental mantle, in the Woodlark Basin, to accommodate depth-dependent thinning of the continental lithosphere. It is, at present, unknown how depth-dependent lithospheric thinning in the Woodlark Basin is accommodated.

The continental lithospheric thinning models, discussed in this thesis, are all 2D with the movement of crust and lithosphere perpendicular to the rift axis. All current models of preferential thinning of continental lithosphere, when applied to the Woodlark Basin, struggle to explain how depth-dependent lithospheric thinning can be accommodated. It seems likely that the current models are all insufficient to model the breakup of the palaeo Papuan Peninsula and that the Woodlark Basin requires a 3D model in order to understand how the thinning of continental lithosphere occurs. It is plausible that material ahead of the propagating tip of seafloor spreading, in the region of depth-dependent thinning, can not only move both outwards, but also eastward, towards the propagating seafloor-spreading centres.

7.5 References

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Chapter 8

Depth-Dependent Lithospheric Thinning During the Formation of the South China Sea Derived From a Long Composite Seismic Reflection Line

This chapter describes a comparative study of continental lithospheric thinning leading to continental breakup undertaken on a long seismic reflection line across the South China Sea. Co-authors include Dieter Franke, Gijs Henstra, Nick Kusznir, Paul Reemst and Andrew Cullen.

Abstract

A seismic reflection profile across the conjugate margins of the South China Sea provides an opportunity to compare brittle extension, crustal thinning and whole lithosphere thinning prior to continental breakup and the onset of seafloor spreading for a weakly magmatic rifted continental margin. Whole lithosphere thinning ($1-1/\beta$) has been determined from subsidence analysis, continental crustal thinning from gravity inversion, and upper crustal stretching (β) from fault heave summation. Whole lithosphere thinning and whole crustal thinning factors are similar and range between 0.2 and 1. Conversion of upper crustal fault extension, assuming depth-uniform stretching (pure-shear), gives thinning factors between 0.1 to 0.35. Thus, upper crustal thinning factors are substantially lower than thinning factors predicted for the whole lithosphere and continental crust. The calcuated distribution of

lithosphere stretching and thinning is inconsistent with depth-uniform (pure-shear) lithosphere deformation that implies that the continental lithosphere of both conjugate margins underwent depth-dependent stretching prior to breakup and formation of the South China Sea. A deeply subsided part of the Macclesfield Bank continental block is modelled as highly thinned continental crust and lithosphere with oceanic crust to the north and the oceanic crust of the South China Sea to the south.

8.1 Introduction

The thinning of continental lithosphere during continental breakup is a fundamental process in the Wilson plate tectonic cycle (Wilson, 1966). Our present understanding of the rifting of continental margins is incomplete; primarily derived from seismic experiments and scientific drilling along the continental margins of the Atlantic Ocean where the transition from continental basement to oceanic crust is deeply buried under thick Late Jurassic to Recent sedimentary sequences. Despite having some of the highest terrigenous inputs in the world (Millman and Meade, 1983), the basins adjacent to the SCS have largely captured this sediment load (Figure 8.1) which accounts for a relatively thin sedimentary cover the Continent-Ocean-Transition of the SCS. This thin sediment cover coupled with geological constraints from industry and ODP drilling creates the opportunity for detailed study of the formation of the conjugate rifted continental margins of the SCS. Determining whether thinning of the continental lithosphere prior to continental breakup is depth-uniform or depth-dependent is fundamental to the development of models of rifted margin formation. To address this issue we have compared the stretching factors calculated for three methods divisions of the lithosphere (Figure 8.1):

1) Thinning for the whole lithosphere is determined from flexural back stripping of post and syn-rift sediments and the McKenzie continental lithosphere extension model (McKenzie, 1978).
2) Crustal basement thinning is determined using satellite gravity inversion
3) Upper crustal stretching is determined from fault analysis

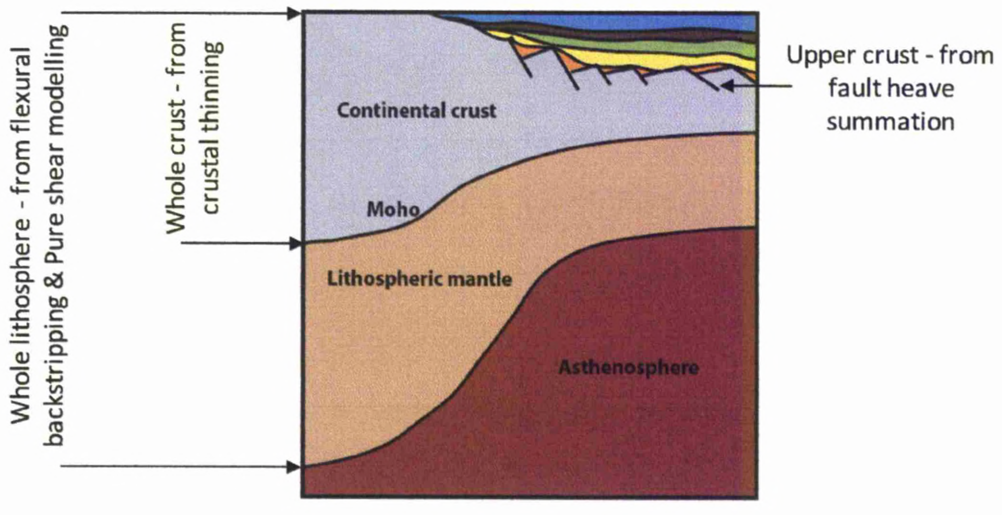

Figure 8.1 Schematic diagram illustrating the three independent methods used to determine extension and stretching at depth within the lithosphere. Modified from Kusznir and Karner (2007).

8.1.1 Geological formation and setting

The pre-breakup phase of continental lithospheric thinning, subsequently resulting in the formation of the South China Sea (figure 8.2), started with initial uplift of the rift shoulders and widespread erosion in the latest Cretaceous to Early Paleocene (Schluter et al., 1996, Taylor and Hayes, 1980, Pigott and Ru, 1994). The post-breakup phase, with the initiation of seafloor-spreading centres and the formation of oceanic crust, started from the end of lower Oligocene to the early Middle Miocene (32-15.5 Ma) (Briais et al., 1993), or from ~31 to 20.5 Ma (Barckhausen and Roeser, 2004). Rifting appears to have occurred in several episodes. Fault analysis along the northern conjugate margin of the South China Sea in the Pearl River Mouth Basin suggests at least two episodes of rifting (Ru and Pigott, 1986). The earlier Cretaceous-Paleocene episode resulted in northeast-southwest oriented extensional faults. The second rift Late Eocene to Early Oligocene episode is expressed as east-west oriented normal faults. According to Pigott and Ru (1994), regional volcanic activity is indicative of a third episode of rifting in the middle to late Miocene.

The history of rifting of the southern conjugate margin is poorly understood. Continental crust dredged from fault scarps in the Spratley Islands (Kudrass, 1985; Hutchison 2010) shows that the region northwest of Palawan is a continental block that separated from mainland Asia during opening of the South China Sea. The Manila Trench is a distinct bathymetric feature from about 20°N to about 13°N at which oceanic lithosphere of the SCS is presently being subducted (Figure 8.2). The termination of this trench terminates in a southern direction is poorly defined (Hayes and Lewis, 1984).

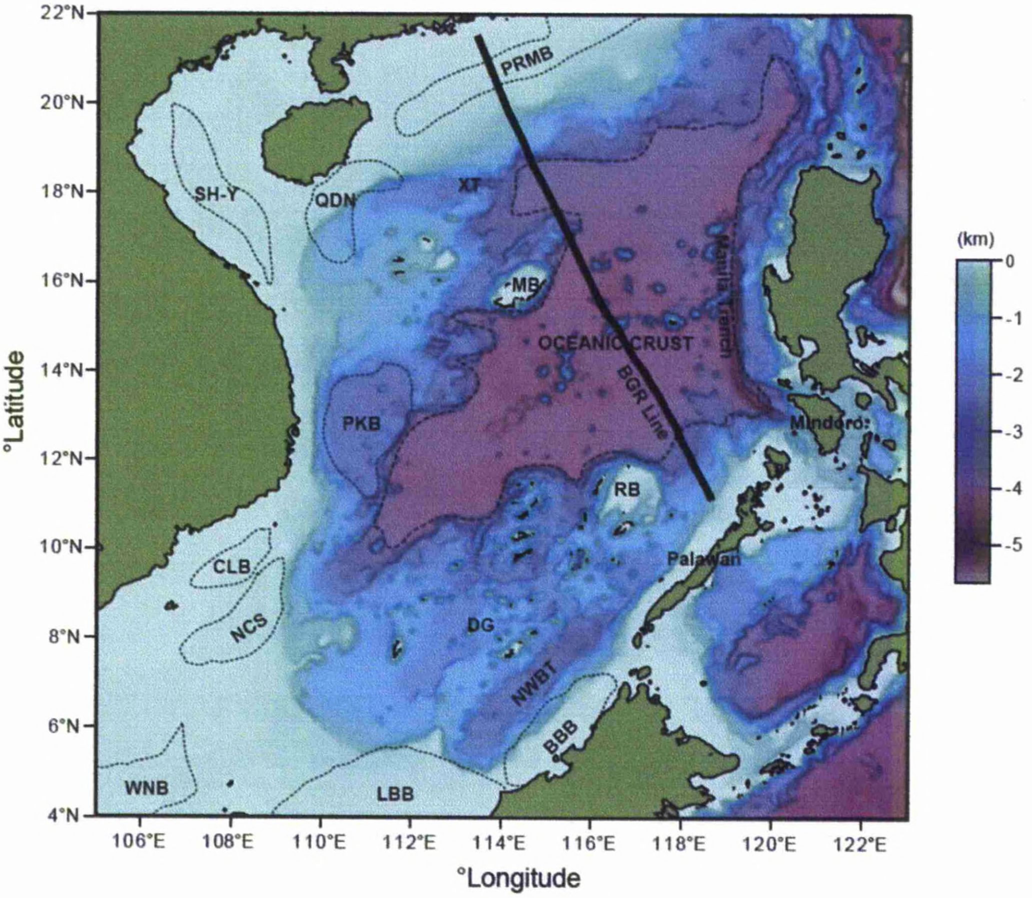

Figure 8.2 Regional setting with GEBCO bathymetry. Purple regions are predominantly oceanic crust. Lightest blue areas correspond to the continental shelf and upper slope. The location of the BGR line is shown. Offshore major sedimentary basins outlined in black dashed lines: BBB, Baram–Balabac Basin; CLB, Cuu Long Basin; LBB, Luconia-Balingian Basin; NCS, Nam Con Son; Basin; PKB, Phu Khanh; PRMB, Pearl River Mouth Basin; QDN, Quiondongnan Basin; SH-Y Song Hong-Yengeahai Basin; WNB, West Natuna Basin. Features in SCS rift: DG, Dangerous Grounds; MB, Macclesfield Bank; NWBT, NW Borneo Trough; RB, Reed Bank; XT, Xisha Trough.

East of the Manila Trench the Philippines consist of an assemblage of terranes of uncertain origin (Hall, 2002) and the extent of continental crust having a mainland Asia affinity is not established. Continental crust is found on the islands of Mindoro (Mitchell and Leach, 1991) and Panay (Gabo et al. 2009). Although initial suturing of these continental fragments may have occurred in the Late Eocene to

Early Oligocene during Sarawak Orogeny, in the Early Miocene they collided with an earlier subduction zone either along the Cagayan Ridge in the Sulu Sea (Hinz et al., 1995) or with the Philippine Arc further east (Yamul et al. (2009).

8.2 Crustal thickness and thinning factors determined from satellite gravity inversion incorporating a lithospheric thermal gravity anomaly correction

A satellite gravity inversion using a lithospheric thermal gravity anomaly correction has been used to determine Moho depth for the South China Sea and the immediate surrounding area (figure 8.3). Owing to the young age of rifting, the elevated geotherm produces a significant lithosphere thermal gravity anomaly that can be in excess of 350mGal, for which a correction is applied in order to determine Moho depth accurately from gravity inversion. The gravity inversion methodology is described in Greenhalgh & Kusznir (2007) and Chappell & Kusznir (2008). The data used in the gravity inversion for the determination of Moho depth are bathymetry (Smith and Sandwell, 1997, Becker et al., 2009), free air gravity (Sandwell and Smith, 1997), ocean age (Barckhausen and Roeser, 2004) and sediment thickness (Divins, 2003).

The continental crust of the margins surrounding the oceanic crust in the South China Sea is predicted to vary greatly in thickness from gravity inversion results (figure 8.3a). A rift limb extension of the first seafloor-spreading centre to the north of the Macclesfield Bank is shown ahead of the fossilized propagating tip, and has resulted in the formation of the Xisha Trough.

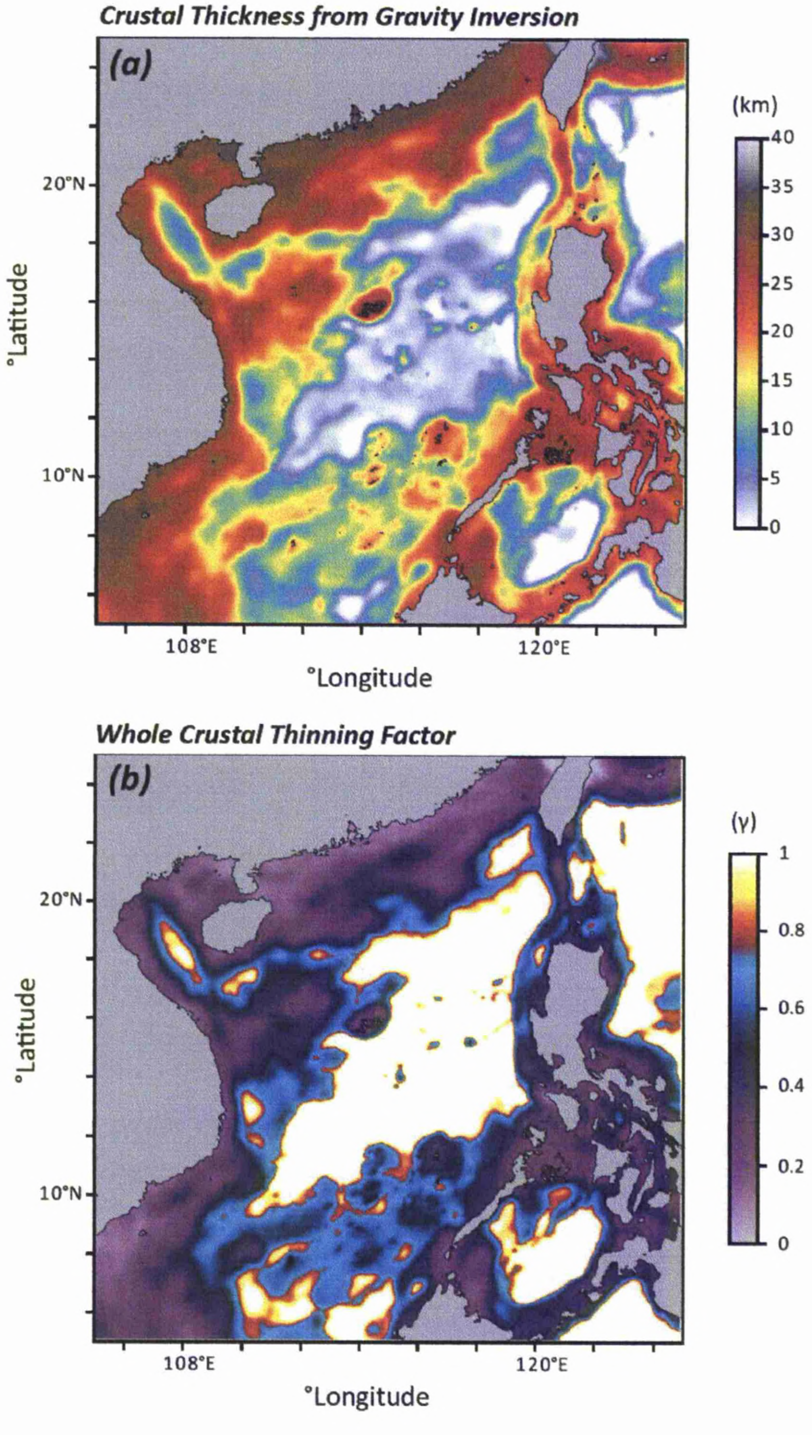

Figure 8.3: a) Crustal thickness predicted by gravity inversion with a lithosphere thermal gravity anomaly correction (Chappell and Kusznir, 2008, Greenhalgh and Kusznir, 2007). NOAA sediment thickness data is used within the gravity inversion. (b) Corresponding continental lithosphere thinning factor (1-1/β) for the South China Sea and surrounding area.

Further south, highly thinned continental crust is observed in the Phu Khanh Basin. The thinning factors, predicted from gravity inversion, reach oceanic values (~0.95-1) and are consistent with the work of Cullen et al. (2010) that the basement here is possibly exhumed serpentinised mantle (figure 8.3b). Ahead of the failed propagating tip of seafloor spreading, offshore southern Vietnam, thinned crust is predicted for the Cuu Long and Nam Con Son Basins. These basins formed ahead of a propagating tip of seafloor spreading and experienced depth-dependent lithospheric thinning similar to other fossilized propagating tip spreading basins such as the Faroe-Shetland Basin (Fletcher, 2009).

The southern margin of the SCS consists of fragmented blocks of continental crust separated by regions of higher degrees of thinning. The Reed Bank, once part of the same continental block as the Macclesfield Bank, is continental crust of 20-25 km thickness. The Dangerous Grounds, further west, is predicted to be continental crust that has been thinned to a higher degree than the Reed Bank and continental crust here is between 10-20 km thick.

8.3 A composite regional seismic reflection profile

8.3.1 Seismic acquisition and processing

The seismic reflection data (used in this study were primarily acquired by the Bundesanstalt für Geowissenschaften und Rohstoffe (BGR) in 1987 and 2008. The BGR line (figure 4) is a composite line using data from both surveys. The 1987 data were acquired using research vessel SONNE equipped with a 48 channel, 2400 m streamer (AMG 37-43). Seismic signals were generated with a 25.6 l airgun array consisting of 10 airguns and operated at 140 bar. The shot interval was 50 m and data were recorded at 4 ms sample interval up to 12 s (TWT, two-way-time). The BGR08 data were also acquired with RV SONNE, equipped with a 312 channels SEAL streamer 3,900 m active length. The source was a G-Gun array that was subdivided into two sub-arrays with eight guns each. The total volume used was 3,100 in^3 (50.8 l). Shot-intervals were 18 s ± 0.3 s, resulting in 50 m shot intervals at a speed of 5.4 kn. The data were sampled at 2 ms and recorded up to 14 seconds. The working pressure was 2,100 psi (145 bar). The BGR line is 1280 km long and runs from the Pearl River Mouth Basin off China across the South China Sea to the continental rise off NW Palawan.

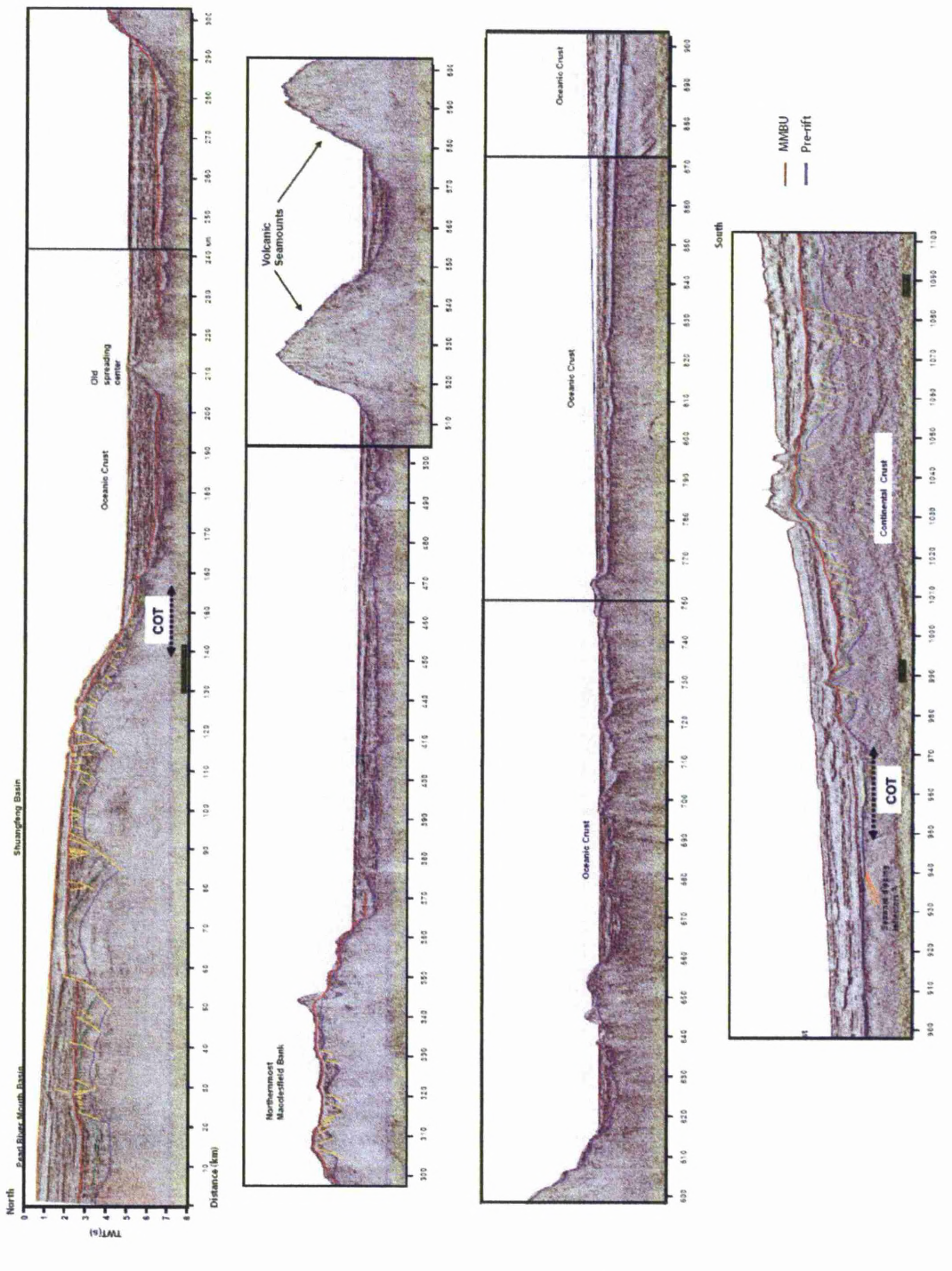

Figure 8.4 The composite BGR seismic reflection line from offshore China to offshore Palawan. The seismic line traverses across the Pearl River Mouth Basin, the Shuangfeng Basin, the Xisha Trough, the northernmost extent of Macclesfield Bank and the oceanic crust of the South China Sea. Volcanic seamounts are labelled in the seismic. The red line is the breakup unconformity and thus the base of the post-rift sediments. MMBU is the Mid-Miocene breakup unconformity.

8.3.2 Seismic interpretation

Offshore of Palawan second to the draping Mio-Pliocene pelagic sediments, the most prominent stratigraphic sequence is a widespread Early to Late Oligocene to late Early Miocene carbonate platform sequence that is expressed seismically as subparallel reflections of high continuity and low frequency content. Locally, shallow marine carbonates reef complexes have developed on top of the platform sequence. Differential subsidence has resulted in thickness variations as carbonate deposition and reef growth continued on topographic highs until middle Miocene or even recent times. In the deepwater area carbonate development stopped in response to the subsidence of the north-western part of the NW Palawan continental block during the latest Miocene (Fulthorpe and Schlanger, 1989). On top of the carbonate platforms, rest Middle Miocene to Pliocene clastic sediments. Seismic imaging beneath these carbonates is poor. Because of a small or weak impedance contrast between the 'pre-rift basement' and the basal graben fill sequences, imaging of this contact is difficult. The rift-onset unconformity at the boundary between latest Cretaceous to beginning of early Paleocene is mostly a high amplitude reflection with moderate continuity that locally defines the base of a complex system of half-grabens (Franke et al., 2011). Beneath the rift-onset unconformity, a high reflective pattern is dominant indicating Mesozoic sediments that in placers have affinity to Triassic sequences in Vietnam (Hutchison, 2010).

At both conjugate margins, seismic interpreters consider the breakup unconformity as being formed during the mid-Oligocene. It likely represents a diachronous event (Ru et al., 1994). From the IODP Leg 184, close to the northern end of our line an age of ~31.5 to ~31.8 Ma is reported for breakup unconformity that matches the age of the first seafloor spreading anomaly in the central SCS as derived from magnetic data (~ 31 Ma (Barckhausen and Roeser, 2004)). However, this is contradicted by data from ODP site 1148 on the Chinese continental slope of the South China Sea. This site provides a reasonable estimate of the end of extension (Shipboard Scientific Party, 2000). In this area, seismic reflection data and the sediment accumulation rate suggest that active extension was completed by about 28 Ma (Wang et al., 2000) or by 25 Ma according to Clift & Lin (2001).

8.4 Whole lithospheric thinning factors determined from subsidence analysis using flexural backstripping of stratigraphy

8.4.1 Determination of whole lithosphere thinning factors

The extent to which the whole lithosphere thinned, during rifting leading to continental breakup and the initiation of seafloor spreading, is determined using the McKenzie pure shear model (McKenzie, 1978). A thinning factor for the whole lithosphere can be determined if the total amount of subsidence, due to the initial mechanical thinning of the lithosphere and post-rift thermal subsidence, is known; therefore the total amount of subsidence has to be determined.

At the time of the onset of rifting, the margins were in a coastal marine setting, as determined from stratigraphy. The sediments deposited in this coastal marine setting now lie under several kilometres of post and syn-rift sediment, and at several kilometres depth because of the subsidence due to the rifting of the continental lithosphere prior to the onset of seafloor spreading in the South China Sea (figure 8.5).

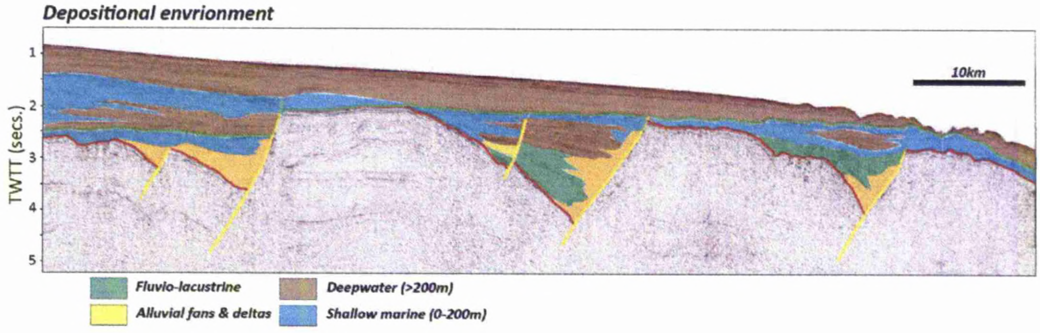

Figure 8.5 Depositional environment of the syn- and post-rift sediments in the Shuangfeng Basin, offshore China. The region has undergone subsidence associated with the thinning of continental lithosphere and formation of the South China Sea. Palaeo-bathymetry of the red line has been assumed to be approximately at sea level.

The removal of the post and syn-rift sediment, taking into consideration the flexural and isostatic response of the lithosphere and the decompaction of older pre-rift sediments, should, if no thinning of the continental lithosphere has occurred, restore these sediments to their original depositional level. However, since rifting and thinning of the lithosphere has taken place since the time of their deposition, the coastal marine sediments do not restore to a coastal marine level and this difference in where the sediment restores to, and their original depositional level, is the water loaded subsidence. Global eustasy for sea level change since the onset of rifting (Haq et al., 1987, Harland et al., 1989) have been included when determining the water loaded subsidence.

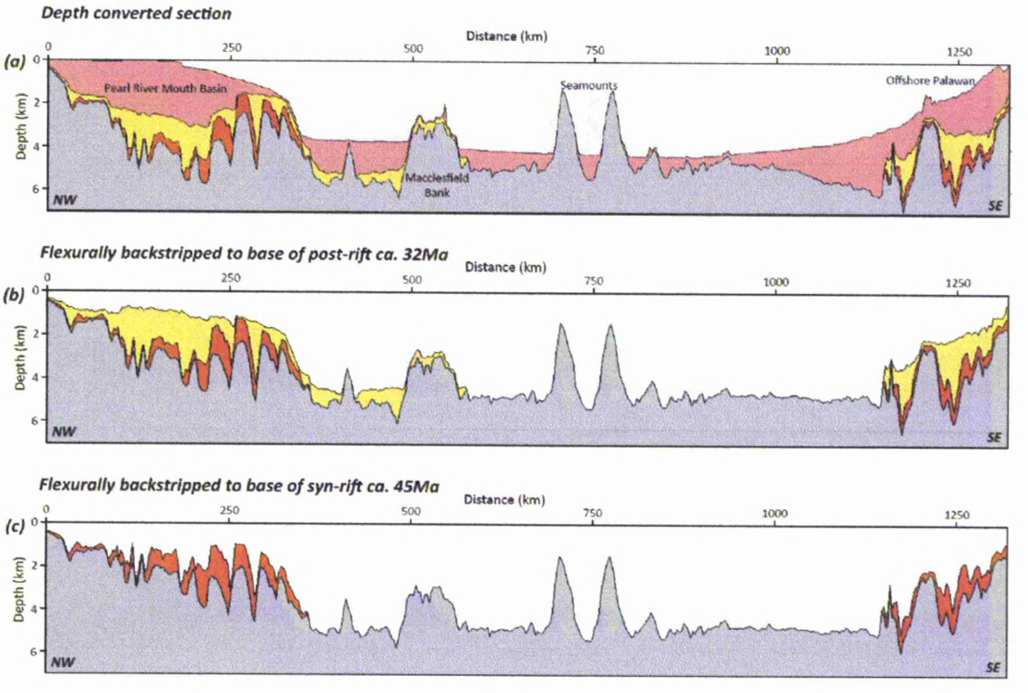

Figure 8.6 (a) Depth converted cross section of the BGR line showing post-rift, syn-rift and pre-rift sediment sequences. Post-rift sediments are pink, syn-rift sediments are yellow and pre-rift sediments are orange. (b) Cross section after the removal of post-rift sequence by flexural back-stripping and decompaction of the pre-rift and syn-rift sediments. Flexural isostatic response of the lithosphere assumes Te=3 km. (c) Cross section after the removal of post-rift and syn-rift sequences by flexural backstripping and decompaction of the pre-rift sediments.

To determine the water loaded subsidence profile for the BGR line, the line was first depth converted using a velocity function based on interval velocities derived from stacking velocities. Within the flexural backstripping modelling, the post, syn and pre-rift sediment sequences were assigned different densities and compaction decay constants depending on the average lithology in that sequence (figure 8.6). The post-rift sediment sequence is predominantly shale, the syn-rift sediments are a mixture of sands and shales and the pre-rift is a mixture of conglomeratic sandstones with intervals of siltstone including some coaly fragments.

The unloading of sediments off the lithosphere produces local and regional isostatic uplift. The regional flexural response to the unloading of sediments is controlled by an effective elastic thickness (T_e), if $T_e = 0$km then only Airy isostasy is assumed. A T_e of 3 km is assumed for this profile in order to consider the flexural response of the lithosphere.

Initial seafloor spreading occurs at approximately 32 Ma (Barckhausen and Roeser, 2004) to the north of Macclesfield Bank, however rifting would have taken place for several millions of years prior to the continental breakup. The McKenzie pure shear model (McKenzie, 1978) assumes instantaneous rifting, the 1D McKenzie model used in the determination of whole lithosphere thinning factors for the composite seismic line uses a breakup age of 32 Ma. This time signifying the end of rifting and transition to divergence of the margins being predominately accommodated by the formation of oceanic crust.

8.4.2 Whole lithosphere thinning factors for the BGR profile

Offshore China

Offshore China encompasses the Pearl River Mouth Basin out to the oceanic crust north of Macclesfield Bank. Whole lithosphere thinning factors range between 0.1 γ in the most proximal part of the margin to 1γ in the oceanic crust to the north of Macclesfield Bank. Whole lithosphere thinning factors were found to be predominantly between 0.2 and 0.6 γ. In general, the northern margin does not thin enough for volcanics to be produced due to the thinning of the lithosphere, and it is only in the final stages of continental breakup that volcanism may occur (south of 350 km in figure 8.7) and this transition rapidly reaches oceanic crust. This means that whilst offshore China is a wide region, +300 km, of thinned continental lithosphere, the actual transition of thinned continental lithosphere to oceanic crust is close and that the final stages of continental breakup was localised.

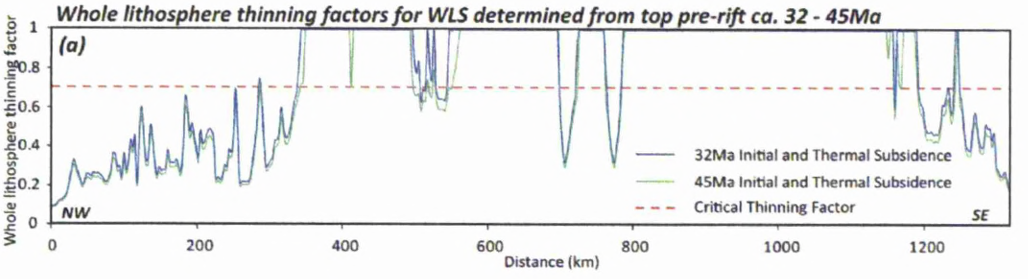

Figure 8.7 Whole lithosphere thinning factors determined from water loaded subsidence (WLS) of the top pre-rift (figure 5c). Sensitivities to rifting ages of 32 Ma and 45 Ma are shown. Whole lithosphere thinning factors assume total subsidence due to initial (syn-rift) lithosphere thinning and post rift thermal subsidence. Northern margin, offshore China, has thinned less than the southern margin, whilst the Macclesfield Bank is highly thinned. The onset of volcanic addition, due to decompressive melting, starts at the critical thinning factor $\gamma = 0.7$ ($\beta \sim 3$)

The deeply subsided part of Macclesfield Bank

Northeast of Macclesfield Bank a continental fragment was traversed by the seismic line. Average whole lithosphere thinning factors for this subsided part of Macclesfield Bank are higher than either margin, ranging between 0.6 and 0.8 γ. These thinning factors suggest that the crust has thinned to an extent where decompressive melting might be expected within the lithosphere.

Offshore Palawan

The southern margin, offshore Palawan, is narrower than the northern margin at approximately 160 km long. Whole lithospheric thinning range between 0.3 and 0.8. The lithosphere here has thinned to a lesser degree than predicted for the lithosphere of the deeply subsided part of Macclesfield Bank. Offshore Palawan is predominately highly thinned continental lithosphere, with magmatism, due to thinning of the lithosphere, being possible only close to the oceanic crust of the South China Sea and in the final stages of continental breakup.

8.5 Determination of upper crustal stretching from fault heave summation

8.5.1 Methodology

The upper crust is considered the top 15 km of the crust and is assumed to extend through brittle deformation only. Upper crustal extension is determined by summing the extension across individual faults observed in the seismic profile. Using this value of upper crustal extension, we determine an upper crustal stretching factor using:

$$\beta_{uc} = \frac{L_{pr}}{L_{pr} - X}$$

Where β_{uc} is the upper crustal stretching factor, L_{pr} is the present day length of the profile and X is the total extension measured across all the faults. Assuming depth-uniform thinning, this stretching factor can be converted to a thinning factor allowing a direct comparison with whole crustal and whole lithospheric thinning factors.

8.5.2 Results

Location	Extension (km)	Margin Length (km)	Initial Length (km)	Stretching Factor (β)	Thinning Factor (γ)
Offshore China	50	335	286	1.17	0.15
Macclesfield Bank	9	27	18	1.48	0.325
Offshore Palawan	20	160	140	1.14	0.125

Table 1. Table showing the determination of upper crustal thinning factors from fault heave summations for the BGR line. Total fault heave extension is used to determine the upper crustal β stretching factor. The stretching factor (β) is converted into a thinning factor (γ) assuming depth-uniform stretching and thinning.

The upper crust offshore China and Palawan has thinned to a similar degree (table 1), approximately 15% longer than its original length in both cases. This is significantly less than the thinning factors determined for the whole crust and the whole lithosphere. The upper crust of the Macclesfield Bank has extended by 48% of its original length. This is significantly more than either of the two other margins and is likely to be due the small nature of the remnant continental block and that it has experienced thinning to continental breakup and the onset of seafloor spreading on either side.

8.6 Depth-dependent lithosphere thinning leading to continental breakup

8.6.1 Whole crustal thinning factors

Whole crustal thinning factors are determined from gravity inversion (figure 8.3). Sediment thicknesses derived from the depth conversion of the seismic data is used along the BGR line rather than NDGC data. From a crustal cross section of the BGR line, crustal thicknesses can be described for each margin from the predictions of the gravity inversion (figure 8.8). Offshore China is predicted to be continental crust of ~30 km thick that thins towards the oceanic crust north of Macclesfield Bank. Continental crust at Macclesfield Bank is thinner than either of the other two rifted margins at approximately 15-20 km thickness. The oceanic crustal thickness of the SCS is typical of average oceanic crust predominantly varying between 5 and 7 km and it thins prior to the continental crust offshore Palawan.

8.6.2 Depth-dependent lithosphere thinning

Stretching factors cannot be directly compared against thinning factors, since the stretching of the upper crust is observed; the upper crustal stretching factors are converted to thinning factors assuming depth-uniform thinning. A comparison of the thinning factors derived for the whole crust and those of the whole lithosphere shows that they are similar to each other, and that they both greatly exceed those predicted for the upper crust from fault heave summation (figure 8.8). All the rifted continental margins, offshore China and Palawan and the rifted continental fragment of the Macclesfield Bank, that the seismic profile crosses exhibit preferential thinning of the lower lithosphere than that of the upper crust. Depth-dependent lithospheric thinning during continental breakup is not a new observation. It has been observed at many other rifted margins (Baxter et al., 1999, Davis and Kusznir, 2004, Driscoll and Karner, 1998, Kusznir and Karner, 2007, Roberts et al., 1997, yet intra-continental rifts seemingly exhibit depth-uniform thinning (Kusznir and Karner, 2007, Marsden et al., 1990). Why rifts that eventually lead to continental breakup and the initiation of a seafloor-spreading centre exhibit depth-dependent thinning whilst rifts that subsequently fail only exhibit depth-uniform thinning is not fully understood. It is plausible that depth-dependent lithospheric thinning only occurs in the late stages of continental thinning and hence those regions that do breakup exhibit it. It is equally possible that rifts that breakup and lead to oceanic crust formation are fundamentally different to intra-continental rifts in terms of their mechanism of deformation.

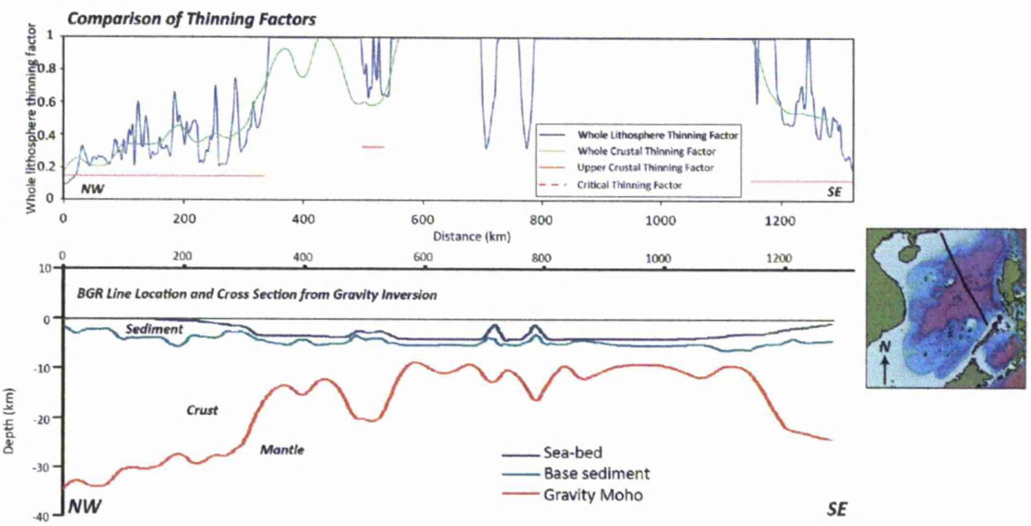

Figure 8.8 Comparison of whole lithosphere thinning factors derived from subsidence analysis, whole crustal thinning from gravity inversion and upper crustal thinning estimated from fault heave summation. Whole lithosphere and crustal thinning are similar and greatly exceed that of the upper crust implying depth-dependent continental lithosphere stretching prior to continental breakup during the formation of the South China Sea. A

location map and crustal cross section for the BGR line with Moho determined from the gravity inversion is shown.

8.6.3 The thinning of the continental lithosphere leading to continental breakup and the formation of the SCS

A schematic diagram of the formation of the SCS has been developed (figure 8.9). An initial pure shear event has been applied to pre-rift continental lithosphere (figure 8.9 i); whilst this is similar to what is observed at an intra-continental rift (Marsden et al., 1990, McKenzie, 1978), there are many different explanations for this rifting event. As rifting proceeds the deformation mode of the whole lithosphere changes. As pure shear deformation continues in the upper crust, simple shear occurs at depth as buoyancy induced upwelling, thins the lithosphere (Figure 8.9 ii). The rift axis of continental lithospheric thinning shifts northwards and buoyant upwelling flow starts to diverge, similar to what is observed at ocean-ridge systems (Cannat, 1996). , and eventually ruptures the Eventually the continental lithosphere to the north of the Macclesfield Bank ruptured (Figures 8.9 iii and iv) forming new oceanic crust at 32 Ma (Barckhausen and Roeser, 2004). The early asthenospheric upwelling (*ca.* 45 Ma) in our model suggests that initial rifting in the South China Sea was driven more by deep-seated "local" processes than by far-field events related to extrusion tectonics (Briais et al. 1993).

Upon the initiation of seafloor spreading, the stress field in the continental lithosphere is reduced and plate divergence is mainly accommodated by the formation of new oceanic crust at the spreading ridge. The seafloor spreading, to the north of Macclesfield Bank, is short lived and is subsequently abandoned. The upwelling divergent flow driving the seafloor-spreading subsequently migrates 80 km to the south. In the time between the failure of the spreading ridge to the north and the rupturing of the crust further south, the divergence of the plates is accommodated by the further thinning of the continental lithosphere and crust. This further thinning of the already thinned continental lithosphere is exhibited in the high thinning factor values in the profile across the Macclesfield Bank, where thinning of the lithosphere and crust has been more extensive than either offshore China or offshore Palawan. Once continental breakup and the initiation of the new spreading centre occurs, the continental crust once again relaxes as seafloor spreading and the formation of new oceanic crust once again accommodates the divergence of the plates.

Had seafloor spreading not jumped from north of the Macclesfield Bank to the south at 24.7 Ma the geography of the area would have been very different. As it happens, the BGR line transects a series half graben along the rifted continental margin that exhibits major asymmetry. The seafloor spreading ridge jump meant the formation of four ocean continent transitions rather than just the two typically associated with one breakup event and no subsequent ridge jump. The OCTs are located in offshore China to the oceanic crust north of the Macclesfield Bank, the oceanic crust north of the Macclesfield Bank to Macclesfield Bank itself, Macclesfield Bank to the SCS and finally the SCS to offshore Palawan. Offshore China, incorporating the Pearl River Mouth Basin is significantly wider than offshore Palawan despite not having thinned to the same degree.

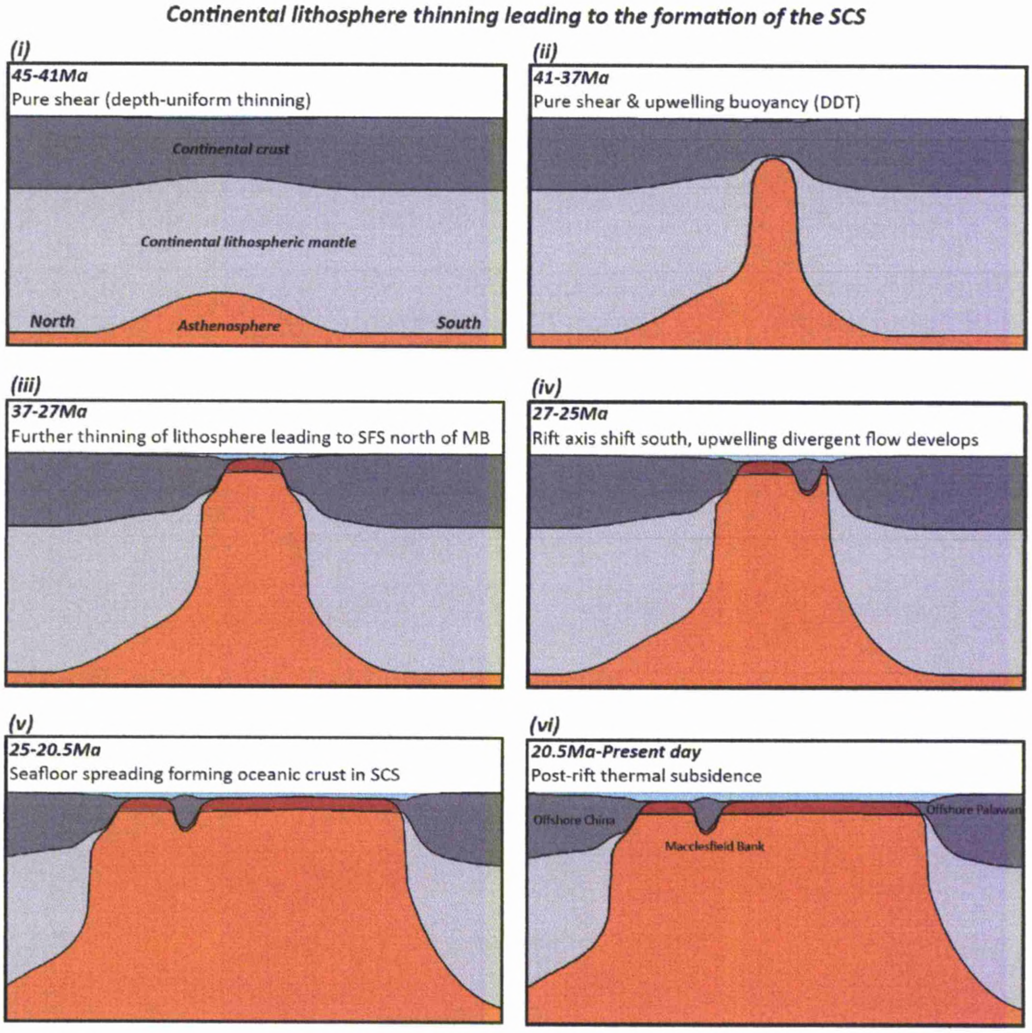

Figure 8.9 Schematic cartoon showing the evolution of the South China Sea along the BGR line. Oceanic crust formation timing, including ridge jumps, taken from (Barckhausen and Roeser, 2004). (i) Pure shear thinning of the lithosphere. (ii) Pure shear of the upper crust with asthenosphere buoyancy. (iii) Further thinning of the continental lithosphere leading to the initiation of seafloor spreading to the north of Macclesfield Bank. (iv) Rift axis moves southwards and buoyancy induced upwelling starts to diverge leading to seafloor spreading initiation to the south of Macclesfield Bank. (v) Continuation of seafloor spreading in the South China Sea leading to spreading centre abandonment at 20.5 Ma (vi) Post rift thermal subsidence of the region. MB is Macclesfield Bank, DDT is depth-dependent thinning, SFS is seafloor spreading.

The ridge jump at 24 Ma, and subsequent formation of the main embayment of oceanic crust, has led to the geography of the SCS of today. The southern margin, as observed in the gravity inversion (figure 8.3), gets progressively wider westwards of the BGR line towards and past the Dangerous Grounds . The early stages of rifting in the South China Sea are unknown due to the subduction in the east of the oldest parts of the oceanic SCS. The ridge jump, at 24.7 Ma coincides with a noticeable change in the spreading azimuth of the SCS. This suggests that the ridge jump is a response driven by changes in the regional stress field (Cullen et al., 2010). The change in location of the spreading centre does not alone account for the formation of the Dangerous Grounds and the Baram Basin. Most of the rifting associated with the formation of the large region of highly thinned continental lithosphere that makes up the Dangerous Grounds pre-dates opening of the South China Sea (Theis et al, 2005).

Stratigraphic, gravity inversion and fault extension analyses have revealed that the margins of the South China Sea underwent depth-dependent lithospheric thinning prior to the rupture of the crust and initiation of seafloor spreading. The Macclesfield Bank was more extensively thinned, than either offshore China or offshore Palawan, due to its close proximity to both breakup events. At 20.5 Ma, the seafloor spreading centres in the South China Sea were abandoned (Barckhausen and Roeser, 2004), since then, no new oceanic crust has formed and the rifted continental margins and the oceanic crust have thermally subsided.

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Chapter 9

Application of Analytical Techniques to Determine the Nature of Basement Crust in the Phu Khanh Basin, South China Sea

This short chapter shows the application of a 1D McKenzie model, incorporating volcanic addition, to determine the nature of the basement in the Phu Khanh Basin. It is modified from part of Cullen et al. (2010) and is my contribution to that paper.

9.1 The South China Sea

The history of the South China Sea's formation encompasses rifting, seafloor spreading, subduction, terrane collision and large-scale continental strike-slip faulting. The South China Sea (SCS) is tectonically complex and the reasons for the formation of this large oceanic basin and the subsequent failure of the oceanic spreading centres are not fully understood. Similar to the Woodlark Basin, the SCS is expressed as an irregular triangular wedge of oceanic crust; however, the SCS is substantially larger than the Woodlark Basin. Seafloor spreading began at approximately 37 Ma until 25.5 Ma at which stage there was an oceanic ridge jump 50 km to the south (Barckhausen and Roeser, 2004) . This new spreading centre continued to form oceanic crust, propagating into the continental lithosphere to the west until it ceased at 20.5 Ma.

The rifted margins of the South China Sea hold numerous rift basins; these include the Cuu Long Basin, the Nam Con Son Basin, the Phu Khanh Basin, Qiong Dong Nan Basin, the Pearl River Mouth Basin, the Xisha Trough, the Hoang Sa Basin and the Song Hong-Yengehai Basin. This thesis will use these names for these basins, however it is recognised that ownership of several parts of the South China Sea is disputed by surrounding countries and that basins tend to have one or more different names in the literature. The complexity of the region means that each basin has its own history and by understanding individual basins, the formation of the region as a whole can be further deciphered.

In the western SCS, large-scale continental strike slip faulting is the dominant tectonic process occurring in the present day. The Red River Fault, a large transform of 100s km of later displacement, is responsible for the formation of the Song Hong-Yengehai Basin and it is possible that it plays a major role in the recent histories of the Phu Khanh and Nam Con Son Basins.

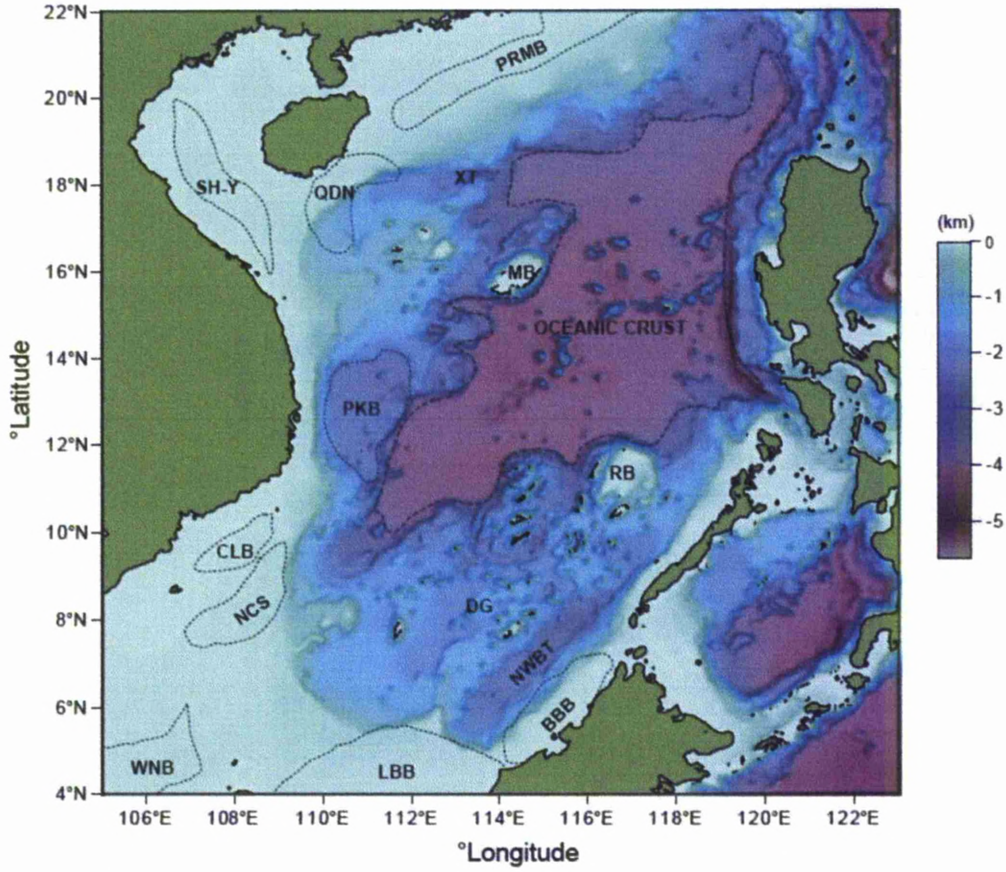

Fig. 9.1. Regional setting and features map with GEBCO bathymetry and illuminated bathymetry as an underlay. Purple regions are predominantly oceanic crust. Lightest blue areas correspond to the continental shelf and upper slope. Offshore major sedimentary basins outlined in black dashed lines: BBB, Baram–Balabac Basin; CLB, Cuu Long Basin; LBB, Luconia-Balingian Basin; NCS, Nam Con Son; Basin; PKB, Phu Khanh; PRMB, Pearl River Mouth Basin; QDN, Quiondongnan Basin; SH-Y Song Hong-Yengeahai Basin; WNB, West Natuna Basin. Features in SCS rift: DG, Dangerous Grounds; MB, Macclesfield Bank; NWBT, NW Borneo Trough; RB, Reed Bank; XT, Xisha Trough.

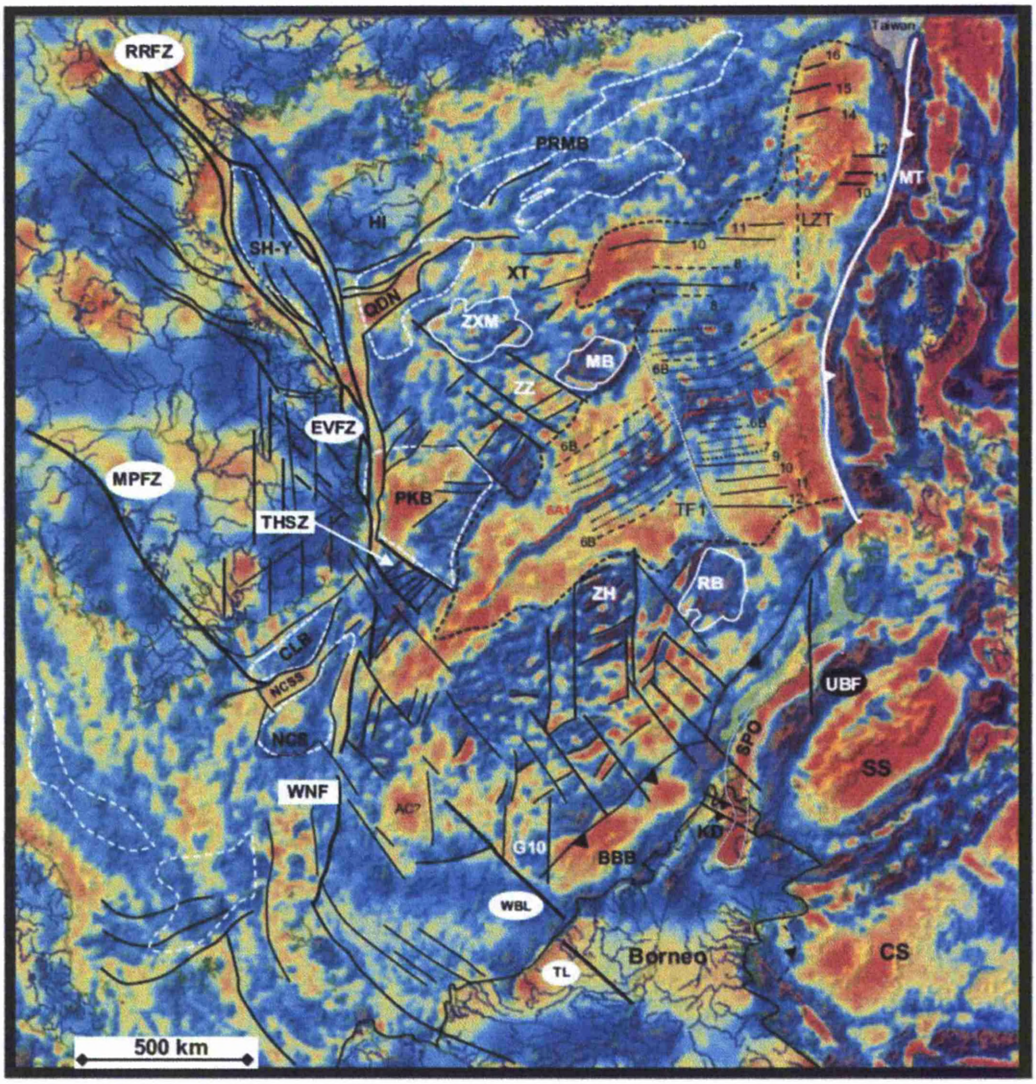

Figure. 9.2. Taken from (Cullen et al., 2010). 150km high pass Bouguer gravity map draped over illuminated 500km high pass Bouguer gravity map. Dashed line around SCS marks approximate limit of oceanic crust. Magnetic anomalies are from (Barckhausen and Roeser, 2004). Solid black lines show faults and fault relays are from Liu et al. (2004), Morley (2002), Rangin et al. (1995), Tongkul (1994), Zhu et al. (2009). EVFZ, East Vietnam Fault Zone; KD, Kudat Peninsula; LZT, Luzon Transform; MPFZ, Mae Ping Fault Zone; MY, Manila Trench; NCSS, Nam Con Son Swell; RRFZ, Red River Fault Zone; SPO, Sabah-Palawan Ophiolite; THSZ, Tuy Huy Shear Zone; TL, Tinjar Line; TF1, transform fault; WBL, West Baram Line; UBF, Ulugan Bay Fault; WNF, Wan-Na Fault; ZZ, Zonghsah Zone. Open circles show Vietnam hot springs with waters > 50°C (Qui 1998). Fine black line with filled triangle marks modern NW Borneo-Palawan collision zone.

9.2 The Phu Khanh Basin

The Phu Khanh Basin is different to the other surrounding basins of the SCS, such as the Nam Con Son Basin, as the timing of its formation is not clearly understood. The basin lies to the northwest of the ancient spreading centre of the SCS and its relationship to the formation of the SCS is not clear. It is bounded to the west by the Red River Fault; this creates a sharp bathymetric contrast on the western edge of the PKB. The eastern PKB borders the oceanic crust of the South China Sea. This oceanic crust formed at approximately 20.5 Ma (Barckhausen and Roeser, 2004), and across the SCS, the conjugate margin to the PKB is the Dangerous Grounds.

The Phu Khanh Basin has greater bathymetry than the other marginal basins. The Nam Con Son Basin has been infilled with sediment whereas the PKB has 2500 to 3000 m of bathymetry, similar to that expected of young oceanic crust. Figure 9.2 shows the distinct gravity signature of the PKB, similar to that of the oceanic crust in the SCS, and since well data about crustal basement type is not available in the literature, it raises questions about the nature of the crust in the PKB.

(The following is taken from Cullen et al. (2010) of which I am the author of this part of the paper.)

We tested the hypothesis regarding extreme crustal thinning in the Phu Khanh Basin by determining whole lithosphere thinning factors along a published cross-section (Fig. 9.3 is modified from Fyhn et al. (2009), fig. 3). We assumed an initial crustal thickness of 35 km (Clift et al., 2002; Holt, 1998). The cross section was flexurally backstripped to the base of the Middle Miocene sediments (model A) and to the base of the Middle to Upper Eocene sediments (model B) to determine water loaded subsidence profiles using the lithosphere extension model of (McKenzie, 1978) modified to include volcanic addition at high thinning factors (White and McKenzie, 1989). Models A and B, figure 9.3, both show post-rift thermal subsidence (St-only) rapidly reaching a value of 1, e.g. a beta factor approaching infinity. The numerical solution in both models requires whole lithospheric thinning factors exceeding one in order to account for the observed whole lithosphere subsidence. Such solutions are physically impossible and are not plotted. Therefore, additional initial (syn-rift) subsidence (Si) is required. Model A, assuming Si + St, predicts that the whole lithosphere has thinned to approximately half of its original thickness in the east since the Lower Miocene; less thinning has occurred in the west. Model B addresses the thinning required to account for accommodation of space represented by the 3–4 km of Oligocene sediments following Middle to Late Eocene rifting. The lithosphere has thinned considerably. Assuming volcanism occurs when the thinning factor exceeds 0.7, there is volcanic addition of 6.5 km. Model B suggests that the crust and lithosphere have thinned substantially and the crustal basement under the eastern part of the Phu Khanh Basin is predominately highly thinned continental crust with a voluminous igneous component. Extreme Palaeogene crustal thinning in Model B is consistent with 2D gravity modelling along the line that shows rapid eastward thinning of the crust to a thickness of 2–6 km (Fyhn et al., 2009).

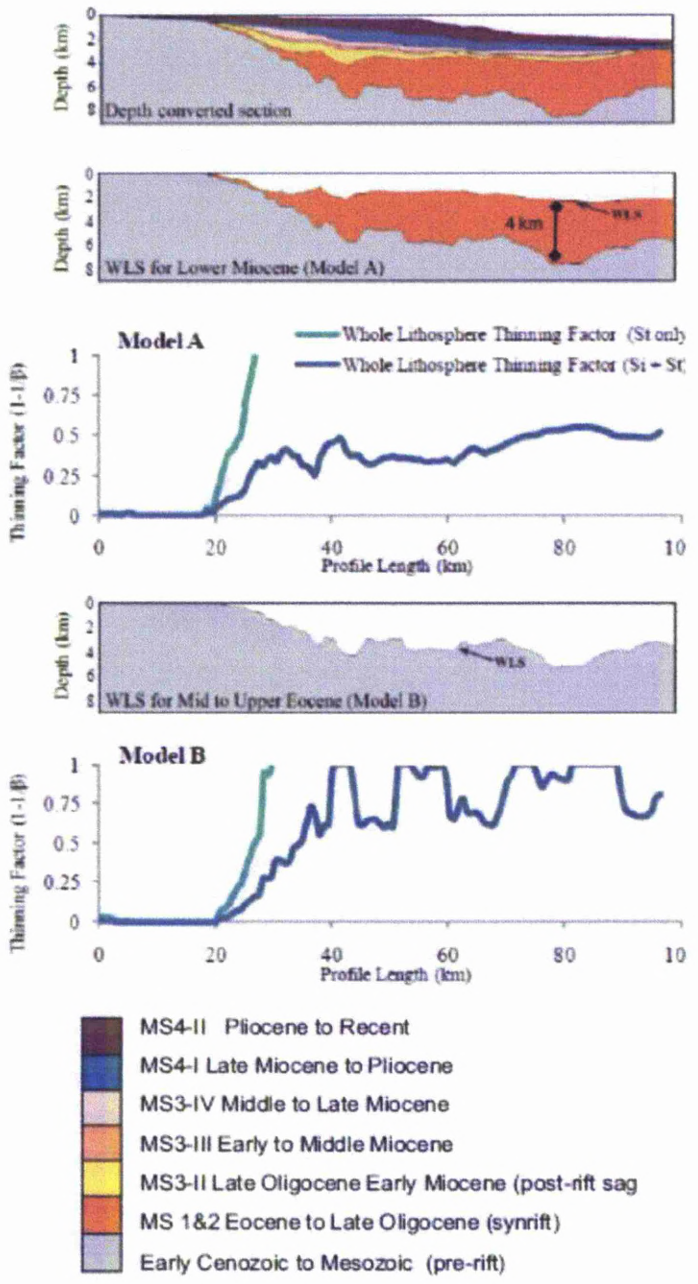

Fig. 9.3 Depth converted model with sediments flexurally backstripped to produce water-loaded subsidence profiles. Model A & B results are from a McKenzie lithosphere extension model, modified to include volcanic addition at high thinning factors. Whole thinning factors for the lithosphere have been determined from the water loaded subsidence. Model A assumes a rift event in the Lower Miocene; model B assumes a rift event in the mid to Upper Eocene. Post rift thermal subsidence alone cannot account for the water loaded subsidence observed in

both models; and results from model B suggests that the Phu Khanh basin consists of highly thinned continental crust with a large component of volcanic addition.

Recently available long-offset 2D seismic data in the ultra deep water Phu Khanh Basin appear to confirm the results of our subsidence modelling (figure 9.3). We believe that part of the Phu Khanh Basin east of the East Vietnam Fault has Eocene to Lower Oligocene syn-rift sediments resting directly on oceanic crust or exhumed subcontinental mantle (figure 9.4).

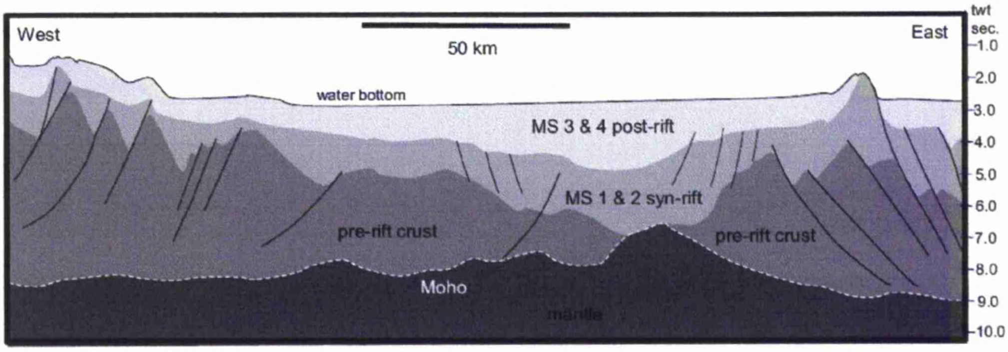

Fig. 9.4 Line drawing of interpreted 2D seismic line traversing eastern PKB. Sequence abbreviations are those used Figure 6. In central part of line pre-rift crust is completely attenuated and syn-rift sediments rest directly on the Moho.

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Chapter 10

Discussion and Conclusions

10.1 Introduction

The aim of this thesis was to investigate the mechanism behind the thinning of the continental lithosphere, prior to continental breakup and the initiation of a seafloor-spreading centre, which ultimately leads to the formation of a rifted continental margin. The final stages of continental lithospheric rupture and the initiation of seafloor spreading have also been investigated.

10.2 Depth-dependent lithospheric thinning at the Woodlark Basin

In chapters 3-6, the breakup of the Papuan Peninsula and subsequent formation of the Woodlark Basin was investigated. The Woodlark Basin is a region where continental breakup can be observed. It can also be used as an analogue for the breakup of a continent through time as pre-, syn- and post-breakup processes are all observed in the present day.

Ahead of the propagating tip of seafloor spreading in the Woodlark Basin, peralic coals deposited at sea level at approximately 8.4 Ma now lie some 3 km below the sea level as the region has undergone significant subsidence associated with the thinning of the continental lithosphere (Goodliffe et al., 2002). The continental lithosphere, through the analyses of chapter 4, is shown to thin depth-dependently, with the upper crust stretching less than the rest of the lithosphere. The modelling in chapter 7 shows that the observed bathymetry and water-loaded subsidence of the Woodlark Basin cannot be modelled assuming depth-uniform thinning of the continental lithosphere alone.

Depth-dependent lithospheric thinning of the lithosphere during continental breakup is not a new observation. It has been observed at many other rifted margins (Baxter et al., 1999, Davis and Kusznir, 2004, Driscoll and Karner, 1998, Kusznir and Karner, 2007, Roberts et al., 1997), yet intra-continental rifts (i.e.rifts that do not lead to continental breakup) seemingly exhibit depth-uniform thinning (Kusznir and Karner, 2007, Marsden et al., 1990).

The Woodlark Basin provides the opportunity to study the continental lithosphere in the final stages of continental thinning prior to seafloor spreading initiation. Unfortunately, due to the complexity of the region surrounding the D'Entrecasteaux Islands, early syn-rift processes are harder to understand. The observation of depth-dependent lithospheric thinning, prior to seafloor spreading, shows that, at least in the case of the Woodlark Basin, it is a syn-rift process rather than a post-rift process. This means that a basin can exhibit depth-dependent lithospheric thinning without there ever being continental lithospheric rupture and the initiation of seafloor spreading, such as in the cases of the Nam Con Son Basin or the Faroe-Shetland Basin (Fletcher, 2009).

The observation of depth-dependent lithospheric thinning at rifted continental margins means that the modelling of subsidence and heat-flow, assuming depth-uniform thinning of the lithosphere, will be incorrect. If the continental lithosphere thins to a greater degree than the upper crust, heat-flow peaks and subsequently melt generation predictions will be greater than expected if upper crustal thinning factors are assumed to represent the thinning of the whole lithosphere.

It is possible that the discrepancy between stretching of the upper crust and thinning of the whole lithosphere is not as profound as reported in chapter 4. The quality of the seismic data from which upper crustal stretching estimations are derived is poor; however, the extension on the major faults is easily determinable. If upper crustal faulting is underestimated, then depth-dependent lithospheric thinning is overestimated.

The geological nature of the Woodlark Basin means that it might not be an ideal setting to determine a model of continental thinning prior to continental lithospheric rupture. The palaeo Papuan Peninsula formed due to collisional tectonics and the accretion of volcanic arcs; hence, the continental structure is highly heterogeneous. Most rifted margins that are studied were once originally part of a large continental plate that has broken up, such as the margins of the Southern Atlantic; however, the Woodlark Basin formed from a collisional terrane adjacent to oceanic crust and subduction zones. It is at present, unclear if the difference in pre-rift conditions causes a bias towards a certain preferential mode of continental lithospheric thinning.

10.2.1 Causes of depth-dependent lithospheric thinning in the Woodlark Basin

The cause of depth-dependent lithospheric thinning in the Woodlark Basin is unclear. Several explanations for depth-dependent lithospheric thinning have been proposed:

i. Steepening of the Trobriand slab (Kington and Goodliffe, 2008)

ii. A *"seafloor spreading"* upwelling divergent flow, ahead of existing seafloor spreading, operating at depth prior to continental lithospheric rupture

iii. Buoyant upwelling of the asthenosphere

iv. Small scale convection due to a wet continental lithospheric mantle (Fletcher, 2009)

The determination of depth-dependent lithospheric thinning, in continental lithosphere leading to continental breakup, has thrown up more questions than answers. They are suggestions for further work and include:

i. Where does the lower continental lithospheric material go when there is no evidence of hinterland uplift or exhumed subcontinental lithospheric mantle?

ii. Can depth-dependent lithospheric thinning be explained in a 2D model or does it require 3D modelling?

iii. Can the lower lithosphere, in the final stages of thinning prior to breakup, move out towards a propagating seafloor-spreading centre?

iv. Is the upper crust extension a response to the thinning of the lower lithosphere? (I.e. is continental lithospheric thinning driven from depth?)

v. Can an upwelling *"seafloor spreading system"* operate below highly thinned continental lithosphere and crust prior to continental lithospheric rupture?

10.3 Depth-dependent lithospheric thinning in the South China Sea

In chapter 8, the South China Sea was investigated as a comparative study to the Woodlark Basin. Both are v-shaped propagating ocean basins; however, the SCS is older and larger than the Woodlark Basin. Unlike the Woodlark Basin, it no longer has active seafloor-spreading centres. A long composite seismic line provided the opportunity to study continental lithospheric thinning across the margins. The degree of thinning that the whole lithosphere and the whole crust have undergone is similar and greatly exceeds the observed extension in the upper crust. This has led to the conclusion that the continental lithosphere at the rifted margins of the SCS underwent depth-dependent, and not depth-uniform, continental lithospheric thinning prior to continental breakup.

The South China Sea experienced a major ridge jump from the north of Macclesfield Bank to 80km further south. This ridge jump has meant that the Macclesfield Bank experienced a higher degree of thinning than either offshore China or offshore Palawan due to two breakup events in close proximity. Whilst the analysis of the thinning of the Macclesfield Bank does not allow for the determination of the amount of thinning or mode of thinning for the individual thinning phases, the total thinning of the Macclesfield Bank still exhibits a discrepancy between the upper crustal fault extension and the thinning of the rest of the lithosphere if depth-uniform thinning is assumed.

10.4 Regional and local variations in initial oceanic crustal thickness

Rifted continental margins are typically classified from the volume of volcanic addition that accompanies continental breakup. They can be labelled as magma-starved, magma-poor, intermediate-magmatic or magma-rich without consideration of local variations in volcanism. The variation in volcanism can be such that a single rifted margin classification for a margin may be locally incorrect.

The Woodlark Basin exhibits both regional and local variations in volcanism. There is a westward trend of increasing volcanism of the continental breakup. The Eastern Woodlark Basin can and should be classified as a magma-poor rifted margin since it had little to no volcanism during continental thinning and exhibits thin initial oceanic crust. The current labelling of a rifted continental margin is not only based on volcanic content, but its structure as well. Labelling the eastern Woodlark Basin as having a magma-poor breakup at present suggests that it is likely that there is exhumed serpentinised mantle in the OCT of the margins, which there is not. The link between label and expected structure is due to the influence and volume of literature on the Iberia-Newfoundland margins, the "classic" magma-poor margins.

The Western Woodlark Basin is more complicated to classify as it falls into no one single rifted margin classification. It exhibits initial oceanic crust ranging from 7 to 12km. Whilst this is more volcanic than a magma-poor margin, the conundrum is that it exhibits little volcanism prior to continental lithospheric rupture. It can be argued that the western Woodlark Basin fits several classifications:

i. **Magma-poor:** very little volcanism observed prior to continental lithospheric rupture and the initiation of seafloor spreading. The eastern Woodlark Basin falls into this category.

ii. **Intermediate-magmatic:** most of the initial oceanic crustal thickness lies within this range

iii. **Magma-rich:** near the propagating tip, initial oceanic crustal thickness can be up to 12km. This is thicker than average oceanic crust. The caveat to this is that little to no volcanism, prior to breakup, is observed.

The current rifted margin classification scheme means that it is hard to apply to rifted continental margins that exhibit regional and local variations in the volcanism of their formation. A classification should give an indication of structures and the volume of volcanics expected. At present, rifted continental margin classifications are too generalized and do not apply to the whole length of a margin.

10.5 Lithosphere velocity discrepancies during the thinning of continental lithosphere

Plate motions can be determined by either GPS data (a data set that is not available in this region) or by seafloor spreading anomalies. Both data sets determine the plate motion at the surface of the lithosphere and deeper lithosphere velocities are assumed but not measured. The conundrum is that when continental breakup occurs in the Woodlark Basin, seafloor-spreading rates are greater than the divergence velocity determined from the total amount of upper crustal fault extension since the onset of rifting. This could be explained as a major underestimation of the total upper crustal fault extension if it was not for the observation that the total width of continental crust is narrower than the expected total divergence derived assuming a constant rifting velocity that is the same as post-breakup seafloor-spreading rates.

An explanation for this phenomenon is that the upper crust does not extend at the same rate as the rest of the lithosphere – *depth-dependent lithospheric thinning*. The problem is that, since no hinterland uplift or exhumed subcontinental lithospheric mantle is observed in this region, where is the extension of the lower lithosphere accommodated (figure 10.1).

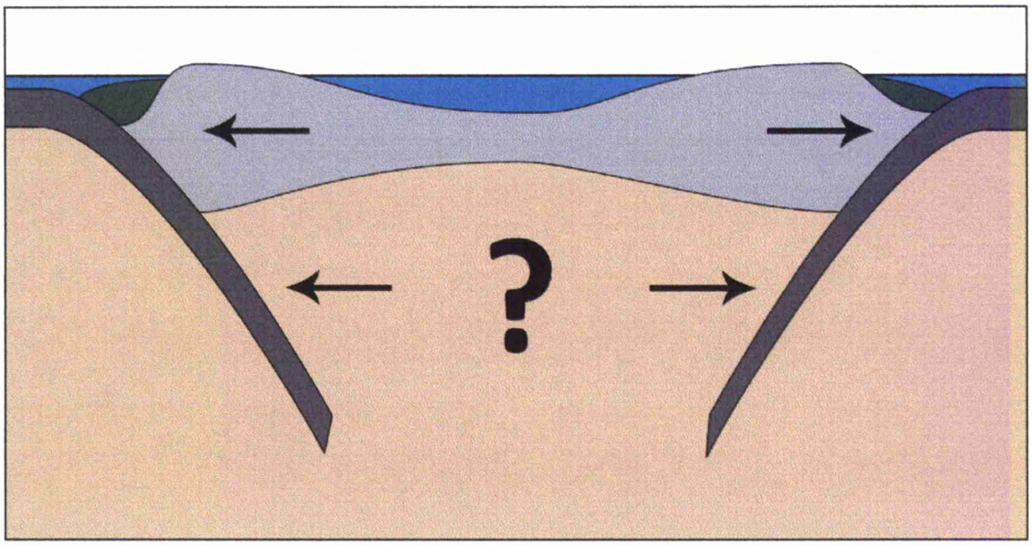

Figure 10.1 Diagram showing the problem of the accommodation of the lower lithosphere where it is being preferentially thinned. If the subducting slabs of the Coral Sea and Solomon Plate remain rigid, it becomes hard to explain where the lower lithosphere can move to.

The observation of a discrepancy between the present width of the continental margins and the expected width determined from Euler pole models derived from seafloor-spreading velocities may hold the key to understanding continental lithospheric thinning in the final stages of continental breakup and should be further investigated. It should be recognised that this observation is derived from the thinning of an ancient volcanic arc terrane and that the pre-rift setting involving numerous subduction zones is inherently different to the pre-rift settings of many other rifted continental margins.

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