OBLIQUE LITHOSPHERIC EXTENSION: 
A Comparative Analysis of the East African Rift System and some Australian Margins

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The work described in this thesis was carried out while I was a full-time scholar at the Research School of Earth Sciences, at the Australian National University, between July 1990 and February 1994. During this time I was partially supported by an ANU Scholarship. Except where referenced otherwise, the research described here is my own, and is believed to be original. No part of the thesis has been submitted to any other university or similar institution.

Deborah L. Scott

February 1994
This thesis is Dedicated to:

Mom, in memory, a promise kept

and Dad

Thanks for teaching me the world is my oyster, and the pearl awaits those who strive to reach their peak.....and for believing in me.
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....then anyone who leaves behind him a written manual, and likewise anyone who receives it, in the belief that such writing will be clear and certain, must be exceedingly simple-minded.....

Plato, *Phaedrus*
ABSTRACT

This study focusses on constraining the three dimensional geometries and kinematics of lithospheric extension and establishing the relationship of extensional systems with pre-existing structure in both the intracontinental and passive margin settings. Case studies of the intracontinental, linked Lake Tanganyika, Lake Malawi and Lake Rukwa rift zones of the Western Branch of the East African Rift System (EARS) are presented. The Queensland Trough of the Northeast margin of Australia, the Ceduna and Eyre Sub-basins of the Great Australian Bight and the western edge of the Exmouth Plateau and the Enderby Half graben of the Northwest Shelf of Australia comprise case studies from the passive margin setting. The results of the case studies generally cannot be fitted to traditional, ideal models of extension. Although each case study is unique, there are common themes that recur in non-ideal (i.e. oblique) extensional settings. The combined results of the case studies have implications for plate tectonic and deformation models, as well as, basin analysis techniques. Oblique extension is common to all of the case studies. That is, the tectonic transport direction of blocks in the deformed zone is oblique to the rift axis. Rift zone scale geometry and the intra-rift fault geometries are commonly different.

Rift zone scale geometry, defined by the rift axis trend, bounding faults and segmentation, is influenced by pre-existing deep-seated, large-scale crustal structures and lateral heterogeneities in the lithosphere. Lithosphere heterogeneity, which corresponds broadly to variations in crustal type (e.g. Archean versus Proterozoic and younger crust) controls the location and trend of the rift axis. Pre-existing deep-seated, large-scale structures (e.g. shear zones and terrane boundaries) control the segmentation and sometimes the bounding fault trends. Repeated reactivation in the same locality suggests that these features extend through the lithosphere and are intimately coupled with deep lithosphere processes. The repeated deformation in the same locality results in maximum structural complexity. The influence of pre-existing structures on basin development is documented in all of the case studies and suggests that tectonic inheritance is fundamental in intracontinental deformation and plate tectonic processes.

The alignment of a large-scale, deep-seated tectonic boundary is not critical to it's reactivation during later deformation pulses. However, the results of the case studies herein, particularly the Rukwa rift zone and the Ceduna Sub-basin, suggest that when the tectonic boundary is aligned with the tectonic transport direction it commonly evolves into a transform fault with a distinctive bend at its
continentward extent, which is apparent in gravity signatures. The strike of the bent ends are a better indicator of the pre-breakup extension direction than fault geometries of isolated rift zones, or subsequent seafloor spreading azimuths, because the fault geometry of the rift zones is dependent on pre-existing heterogeneities and may not accurately reflect the regional tectonic transport direction. These large-scale transcurrent structures may also act as strain guides for adjacent blocks of lithosphere in a similar fashion to the smaller-scale influence of pre-existing shear zones on adjacent half graben geometries within rift zones.

Intra-rift geometry comprises normal and transverse faults, both of which can, and frequently do, trend obliquely to the rift axis. Transverse faults are of two types: reactivated structures which can trend at any angle to the rift zone axis and intra-rift transfer faults which are generally subparallel to the tectonic transport direction, but can be locally influenced by pre-existing large-scale transverse structures which appear to act as strain guides where they cross-cut rift zones. Empirical evidence from the EARS suggests that the tectonic transport direction of the deformed blocks remains consistent throughout the extensional system even when fault geometries vary from one segment of a rift zone to the next or one rift zone to the other. Because fault geometries can be quite variable, the tectonic transport direction appears to be best constrained by dip analysis, a method developed here to measure accurate hanging wall dips from seismic data.

Understanding the three dimensional geometries and kinematics of each of the rift basins studied has been improved by the use of dip analysis combined with traditional structural interpretation and basin analysis techniques. Dip analysis can be a useful tool in several aspects of basin analysis. In oblique extension it appears to be the most consistent kinematic indicator, given a large, widely distributed dataset. In repeated intracontinental extension that does not lead to successful continental break-up, normal faults are frequently composed of more than one segment. The trend of one or more of the segments is usually controlled by pre-existing planes of weakness. The resulting configuration results in a "corner" fault plane geometry that brackets the tectonic transport direction. Even when the tectonic transport direction changes from one pulse of deformation to the next, these structures are frequently reactivated. Changes in tectonic transport direction through time can be recognized by conducting dip analysis at key stratigraphic horizons. Determination of accurate dips can help to separate the rift zone geometry (i.e. bounding faults and sementation) from the intra-rift (i.e. half graben) geometry. It constrains normal fault correlations and kinematics. It is especially useful for identifying required lateral offsets in normal faults and the location of
intra-rift transfer faults. Dip analysis can also help identify seismic artifacts and constrain bulk deformation models for the hangingwall.
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1.0 INTRODUCTION

Then the Sorcerer Science entered,
and where e'er he waved his wand
fresh wonders and fresh mysteries rose on every hand
—Frances P. Cobbe

1.1 Objectives

The primary objective of this dissertation is to determine the geometry and kinematics of non-ideal extension, with particular emphasis on the evolution of intracontinental rifts into rifted passive margin systems. The principal aims of the study are:

• To determine the three dimensional fault geometries in a range of extensional systems to improve the largely two dimensional framework that currently exists.

• To determine, from the three dimensional fault geometries, the importance of oblique extension in rift systems and to assess processes and lithosphere characteristics responsible for oblique divergence.

• To investigate the implications of oblique extension in the upper crust for both lithospheric scale tectonic processes and basin analysis in extensional settings.

In order to accomplish the above objectives, a range of extensional basins have been analyzed to determine fault geometries common to various stages of extension. Non-ideal (oblique extension) geometries are found to be common and the causes of oblique extension are investigated. The obliquity of rift axes to the direction of extension and the geometries of oblique extensional systems is integrated with regional pre-existing structural relationships and found to be intimately tied to pre-existing lithosphere heterogeneities.

The inherent ambiguity in correlating and mapping fault traces between seismic profiles, the most common data type used to assess extensional basin-forming structures, requires that an independent and rigorous technique be developed to determine the overall fault geometry and kinematics in this setting. To this end, a method to assess the kinematics of non-ideal extensional basins where primarily seismic reflection data are available and traditional geological kinematic indicators are unavailable was sought and developed (Chapter 2). The method, termed dip analysis, can be performed independently of mapped fault geometries, helps constrain fault correlations, constrains models of bulk
deformation of hangingwalls in extensional terranes and contributes to an interpretation that is balanced in the third dimension. In addition to traditional structural interpretations, dip analysis was performed where basement was discernible (Chapters 3 and 4) and in each case provided additional constraints on structural and kinematic interpretations at both the local (basinal) and regional scales.

1.1.1 Definitions

Non-ideal extension is often referred to as transtension. The term transtension was originally coined to describe crustal deformation in oblique zones of ocean spreading that consist of stepped transform faults (Harland 1971). Later usage broadened to describe development of normal fault systems obliquely inclined to the strike in strike slip systems (Ramsay & Huber 1987). The association with oceanic systems or strike slip systems does not fully or accurately describe the non-ideal extension documented in several intracontinental rifts and described in later sections, so the term 'oblique extension' is preferred. *Oblique extension is defined as extension which is neither parallel nor perpendicular to the rift basin elongation.* Extension is generally accommodated by oblique slip fault pairs, or sometimes three distinct fault sets. Both faults of a pair can trend obliquely to the regional least compressive stress although one set of the pair is generally subparallel (within 30°) to this direction. In this dissertation, a *dip line* is defined as a rift axis perpendicular profile and a *strike line* is defined as a rift axis parallel profile.

Dip analysis, as defined in Chapter 2, can yield a consistent azimuth even in rift zones where bounding fault trends are highly variable. The direction defined by the most prominent dip direction is termed the *tectonic transport direction.* It is argued that this direction is equivalent to the *maximum horizontal extension direction,* $e_h$, in oblique extension systems. *Dip domains* are segments of rift zones wherein the computed dip directions of the hangingwalls do not vary more than 5°, but are distinguishable from adjacent clusters of dip directions.

*Figure 1.1* Location map of rift basins analyzed. A) The East African Rift System (EARS): T = Lake Tanganyika rift zone, R = Lake Rukwa rift zone, M = Lake Malawi rift zone B) Australian Passive Margins: Northwest Shelf, B/D = Barrow-Dampier Sub-basins, EN = Enderby Half graben, EX = Exmouth Plateau; Great Australian Bight, E = Eyre Sub-basin, C = Ceduna Sub-basin; Northeast Australian Margin, QP = Queensland Plateau, QT = Queensland Trough.
The terms *balanceable* and *retrodeformable* are used to describe structural interpretations (cross sections and plan view maps) that do not require significant internal bulk strain or non-elastic deformation of fault blocks. That is, the pre-deformation configuration is obtainable by reversing the interpreted movement along fault planes. In discussions of the pre-existing structure of the crust and lithosphere, the term *non-uniform* refers to differences in the bulk properties which are generally transitional (e.g. lithosphere thickness or bulk composition), whereas the term *heterogeneity* is used where material properties change across some fairly distinct boundary (e.g. a large scale shear zone).

1.1.2 Case Studies

In order to assess structural geometries of various types and stages of extensional systems both intracontinental and passive margin basins have been investigated in this study (Figure 1.1). These include:

- Lake Tanganyika, Lake Malawi and Lake Rukwa rift zones of the East African Rift System (EARS);
- Queensland Trough of the NE Australian margin;
- Ceduna and Eyre Sub-basins of the Great Australian Bight;
- Enderby Half graben and extensional structures bounding the western edge of the Exmouth Plateau of the NW Shelf of Australia.

Representing the juvenile intracontinental stage, the Lakes Tanganyika, Malawi and Rukwa rift zones have a variety of axial trends and thus provide a spectrum of rift geometries produced by the same regional extension. The Queensland Trough, combined with the connected Townsville trough, provides a similar crustal-scale geometry as the Tanganyika/Rukwa/Malawi depressions around a craton and represent the "failed rift" category. The Ceduna and Eyre Sub-basins and the Enderby Half graben are stranded mature rift basins from "successful" extension leading to seafloor spreading. The western edge of the Exmouth Plateau is an example of extensional structures adjacent to oceanic crust. The passive margin case studies provide examples of rift elongation at a variety of orientations with respect to the eventual spreading axis, both with and without marginal plateaux.

Primarily seismic data, supplemented by well information and other geophysical data as available, were examined for each of the rifted basins. Analysis
and interpretation of each area comprise case studies presented in Chapters 3 and 4 that emphasize the regional tectonic architecture, the geometry within and the kinematic development of the rift zones.

Some of the results presented in this thesis have been published or are in press, although all of the work was done specifically as doctoral research. Particular sections and their published references follow:

- the dip analysis of the Lake Tanganyika and Lake Malawi rift zones (Section 3.2.2) was presented at an American Geophysical Union Fall Meeting (Scott et al. 1990) and the case studies (Section 3.2) were published in Scott et al. (1992);

- the development of dip analysis (Chapter 2) will be published in Scott et al. (in press);

- the case study of the Queensland Trough (Section 4.2) is published in Scott (1993).

Figure 1.2 Comparison of predicted fault geometries in the ideal A) orthogonal extension model and B) strike slip "pull-apart" model. Heavy arrows indicate regional tectonic transport directions. A-A' = rift-axis-parallel (strike) transect. B-B' = rift-axis-perpendicular (dip) transect.
1.2 Extensional Basin Forming Mechanisms

Since the widespread acceptance of plate tectonic theory there have been continuing attempts to link surficial geological phenomena to lithospheric scale mechanisms. This led earlier investigators (e.g. McKenzie 1978) to propose an essentially one dimensional extensional basin-forming model wherein large regions of continental lithosphere are stretched by pure shear. Subsequent recognition that:

- extensional basins are commonly asymmetric and are characterized structurally by half graben (e.g. Bally 1981, 1982, Gibbs 1984, Etheridge et al. 1984, 1985, Gibbs 1985, Rosendahl & Livingstone 1983, Rosendahl et al. 1986, Rosendahl 1987), rather than the symmetric graben predicted by earlier models, and

- large, low-angle normal faults play an important role in achieving extension of the continental lithosphere

has led to the proposal of a variety of more complex two dimensional models. Indeed, entire volumes have been dedicated to the description of rift zone fault systems (e.g. Ramberg & Neumann 1978, Coward et al. 1987, Rogers et al. 1989, Roberts et al. 1991, Ziegler 1992a). However, the three dimensional geometries in these deformed zones and the details of their evolution into successful spreading systems are still controversial topics. The two dimensional models invoke mechanisms that range from whole lithosphere simple shear (e.g. Wernicke & Burchfiel 1982, Lister et al. 1986) to combinations of upper crustal simple shear, crustal delamination and lower crustal pure shear (e.g. Davis & Hardy 1981, Etheridge et al. 1989, Lister et al. 1991).

Despite many significant advances, thinking is still dominated by a largely two dimensional view. That is, kinematic models of extensional basins are generally considered to be one of two end members--orthogonal extension or strike slip "pull-apart". Both end member models predict a simple relationship between the rift elongation, the geometry of the principal fault sets and the regional kinematics or tectonic transport direction (Figure 1.2). The orthogonal extension model predicts normal faults that strike parallel to the rift axis linked to perpendicular strike to oblique slip faults, termed transfer faults (e.g. Gibbs 1984, 1985, Lister et al., 1986). Strike-slip models predict "pull-apart" basins wherein the strike slip faults are parallel to the rift elongation and normal faults range from
perpendicular to 30\(^\circ\) from perpendicular to the rift axis creating rhombo-chasms (e.g. Aydin & Nur 1982, Mann \textit{et al.} 1983, Sylvester 1988).

1.2.1 \textit{Juvenile extensional basin geometries}

Juvenile, or low-extension, basin geometries are generally derived from observations of intracontinental rift zones. As more data become available it is apparent that many juvenile basin geometries do not fit into either the orthogonal extension or strike slip pull-apart models but are characterized by oblique extension or transtension. Seismic reflection data from the EARS basins demonstrate that many aspects of the orthogonal extension model, including the rarity of full-graben morphologies and the polarity switching of half graben along the axis of the rift, typify the intracontinental rift setting and particularly the EARS basins (Rosendahl & Livingstone 1983, Rosendahl \textit{et al.} 1986, Rosendahl 1987, Reynolds 1984, Sander 1986, Specht 1987). However, mapped fault geometries are frequently incompatible with an orthogonal extension model so a new model, referred to as the EARS model was proposed (Rosendahl & Livingstone 1983, Rosendahl \textit{et al.} 1986, Rosendahl 1987).

Instead of the two primary fault sets predicted by the orthogonal extension model, the EARS model proposes that rifts are composed of arcuate bounding faults defining half graben units. Juxtaposition of half graben units of opposing polarity led Reynolds (1984) to propose that the alternating arcuate bounding faults of the EARS formed a roughly sinusoidal deformation pattern (Figure 1.3). He investigated the work involved in stretching the crust and concluded that while an orthogonal, rectilinear strain pattern was preferred, the next "least work" scenario was best described by a sinusoidal curve. A conceptual analogy is a walk in the forest. It takes less work to follow a sinusoidal path around the trees than to progress in a straight line through the trees. He applied the sinusoidal concept to a variety of intracontinental rifts and concluded that each rift zone studied could be described by a sine curve. In particular, there was consistency in the dimensions of half graben units as defined by the wavelength and amplitude of the sine curve suggesting that continental crust does deform in this manner.

\textbf{Figure 1.3} A) Idealized sinusoidal intracontinental rift zone fault model as derived from multifold seismic data from the Lake Tanganyika rift zone (after Reynolds, 1984). B) Idealized passive margin rifted basin fault model (after Lister \textit{et al.}, 1986). Both models incorporate half graben polarity switches across transverse structures. However, the transfer fault (T) in the passive margin model is a predicted consequence of the implied kinematics, whereas the accommodation zone (AZ) in the intracontinental model apparently bears no genetic relationship to the kinematics indicated and is probably related to pre-existing heterogeneities.
Sander (1986) argued that because the maximum subsidence in the Lake Tanganyika rift zone generally occurs at the peaks and troughs of a sine curve describing the half graben units, the line perpendicular to the tangent to these apices defined the direction of extension. In the Lake Tanganyika case, this predicts E-W extension, compatible with the N-S elongation of the rift zone. But, as no rift perpendicular structures could be identified, neither orthogonal extension or strike slip pull-apart models appeared capable of accounting for the observed faulting within the EARS.

EARS mapped geometries are not strictly balanceable in the third dimension; that is, they require a substantial amount of non-elastic strain in adjacent unfaulted regions and within fault blocks. Also, the rift zone is compartmentalized by accommodation zones that appear to have no consistent trend or relationship to the direction of extension or the rift axis (Versfelt & Rosendahl 1989) as do the transfer faults of the orthogonal extension model. The primary problem with the sinusoidal fault geometry is that it predicts accommodation zones of particular position and trend relative to half graben, and these positions and trends are not supported by the mapped accommodation zones in the EARS, especially in the Lake Malawi rift zone. Further, with no structures developed parallel to the direction of extension it is difficult to forward model this system into a successful spreading system without creating additional structures. Understanding the accommodation zones seems key to understanding the structural and kinematic history of the rifts. With this in mind, an investigation of some of the accommodation zones in the EARS was undertaken, supplementing the multifold seismic data with single channel and echo sounder records. The results of this investigation are presented in Section 3.2.4. The interpretations suggest that although accommodation zones may be complexly deformed they can be described by the interaction of pre-existing heterogeneities and rift-zone-consistent fault patterns when viewed at small enough scales.

1.2.2 Mature extensional basin geometries

Mature, or high-extension, basin geometries are generally found in passive margin settings adjacent to successful spreading systems. The orthogonal extension model described above has the advantage that it links small, basin-scale deformation patterns into the global tectonic framework. In the orthogonal extension case, it has been proposed that the normal faults are listric and sole into a detachment (Figure 1.3). Listric fault geometry models predict the basin asymmetry noted above. Further, since normal faults generally do not extend the entire length of a rift zone axis, cross-cutting subvertical transfer faults that trend perpendicular
to the rift axis and parallel to the direction of extension are a required element of this geometry (Gibbs 1984). Transfer faults compartmentalize the rift and allow for differential motion on the normal fault segments and polarity or sense of asymmetry switching along the strike of the rift axis. The normal/transfer fault geometry predicted by orthogonal extension results in rectilinear fault patterns in map view. In Chapter 4 several rift zones from passive margins are analyzed. None of the rift zones analyzed have precisely orthogonal geometries, although the Eyre Sub-basin and Enderby Half graben are well constrained examples of near rectilinear fault geometries. The repetition of oblique rifting in all of the case studies indicates that it is an important mechanism during intracontinental deformation and continental fragmentation.

1.3 Global Implications and Basin Analysis

In Chapters 5 and 6 the conclusions from the individual case studies are summarized and integrated to investigate the importance of oblique extension on global and basin analysis scales, respectively. The summarized results argue that oblique extension, rather than being a rarity, is a common consequence of deforming the highly heterogeneous lithosphere.

Consistent patterns of deformation and reactivation are found to occur in a variety rift systems and settings. Implications of this consistency at basin and tectonic scales are discussed. One of the primary implications of this study is that tectonic inheritance has a fundamental role in determining the geometry of intracontinental deformation and continental fragmentation. Consequences for passive margin architecture and plate tectonic reconstructions and processes are explored.

When the available data consists primarily of two dimensional reflection seismic data, dip analysis can greatly enhance structural interpretations, especially where data are sparse and can aid in the design of subsequent data acquisition. Improvements to traditional basin analysis techniques, particularly the interpretation of reflection seismic data, are suggested. Characteristics common to oblique extensional systems are outlined.
Chapter 2

Dip Analysis

2.0 Dip Analysis as a Tool for Estimating Regional Kinematics in Extensional Terranes

A mind once stretched by a new idea never regains its original dimension.
Oliver Wendell Holmes

2.1 Introduction

In basin analyses that rely almost solely on seismic reflection data to deduce the underlying structure, it has become common practice to use extensional fault geometries to infer kinematic histories. This procedure relies on the existence of fault geometries predicted by ideal orthogonal extension and assumes the structure is well constrained. In areas where data are sparse or of equivocal quality, this often leads to circular analysis. That is, ideal orthogonal extension is assumed to have been operative, then under-constrained fault geometries are mapped according to model predictions. Fault geometries so derived are subsequently used as evidence that orthogonal extension was operative and that the direction of extension is determinable. In this chapter, an alternative tool for inferring kinematics which can be performed independently of mapped fault geometries is developed. The method of analysis is hereafter termed "dip analysis". Dip analysis is based on the relationship between the dips of hangingwalls and the geometry of the bounding faults and is supported by empirical evidence that the dip of the hangingwall(s) is systematic and predictable in extensional systems, even when fault geometries are non-ideal (Chapter 3). Although not without some limitations, dip analysis can provide additional constraints for structural and kinematic interpretations, especially where seismic reflection data are the main source of information and dip domains define discrete crustal blocks.

In recent years a plethora of seismic reflection data have become available for the structural analysis of basins, especially submerged basins with hydrocarbon prospectivity. These data are often the only source of information for the interpretation of structures that formed the basin. Fault geometries thus derived are frequently ambiguous because fault trace correlations are made between widely spaced profiles, but are nonetheless used to infer regional tectonic kinematics, particularly in extensional terrains. Geometric models which imply different kinematic histories range between the end-member models of orthogonal extension and strike slip "pull-apart" basins as described in Section 1.2. Each end-member model predicts a simple relationship between the rift or basin elongation, the geometry of the principal fault sets and the regional kinematics or tectonic transport direction (Figure 1.2). However, in the absence of geological kinematic
indicators, estimating tectonic kinematics from fault geometries is only as valid as the structural interpretation is constrained. Additionally, the models are "ideal" fault geometries that are predicted to form in homogeneous media, which is virtually non-existent in the modern lithosphere.

It is common practice when evaluating a basin for resources to produce depth-to-horizon structure contour maps. These maps are, in essence, a picture of the dip of the surface of interest. Apparent dip directions as measured perpendicular to contours on the surface may not reflect the actual dip direction as measured at a point of control because to contour is to smooth. An interpreter effectively eliminates the direction of the dip at control points, which occur at profile intersections where two apparent dips are measurable, by imposing contour shapes that are commonly influenced by the interpreters perceptions of fault geometries at the level in question.

In the following sections, it is argued that the rotation or tilting of infra-basinal basement blocks determined by dip analysis at intersections of seismic profiles can be used to establish a tectonic transport direction. Dip analysis can be performed independent of a priori knowledge of fault geometries, thus providing corroborative evidence for interpreted fault patterns. Further, empirical evidence (presented in Chapters 3 and 4; see also Scott et al. 1992) and analog experimental and analytical modelling (Braun et al. in press) suggest that the method can be used in areas where deformation is characterized by oblique slip faults, even when extension-accommodating fault shapes and orientations vary throughout the system.

To develop the method of dip analysis, the geometric concepts relating fault geometries to tectonic kinematics in ideal model terms are described. The relationships between fault geometries imaged in seismic profiles, block rotations or tilting, and direction of tectonic transport on the system scale are considered. Individual block rotations and the significance of the dip of the block are evaluated to develop dip analysis. The validity of the method in the ideal orthogonal extension case is demonstrated. To explain observational data, the geometric relationships in the ideal, orthogonal deformation scenario are hypothetically extrapolated to situations where extension is non-ideal and is accommodated by oblique slip faults and some possible rationales are explored. Finally, the method is applied to some reflection seismic data to demonstrate the strengths and limitations of dip interpretation in basin analysis.
2.2 Extensional (Rift) Systems

Currently accepted models of ideal orthogonal extension in the brittle upper crust predict two distinct fault sets which are mutually perpendicular and lead to simple fault/basement-dip geometries in cross sections aligned perpendicular to the fault sets (Profiles E and F, Figure 2.1). These ideal models suggest that rift elongation is most commonly perpendicular to the direction of horizontal extension, $e_h$, mimicking geometries at oceanic spreading centers. "Dip lines" parallel to $e_h$ or the tectonic transport direction will show a series of similarly tilted fault blocks bounded by normal faults (e.g. Profile E, Figure 2.1).

Deformation is balanceable or retrodeformable (Suppe 1985) in the plane of the profile. "Strike lines" parallel to the traces of normal faults will image flat-lying subsided blocks and vertical faults (e.g. Profile F, Figure 2.1). "Strike line" geometry may appear reconstructible at the basement surface by purely vertical motion along the faults, but this is not a correct reversal of movement on these fault planes and normally the overlying synrift beds will not retrodeform by this motion. Profiles in any other orientation will image unbalanceable geometries in the profile plane (e.g. Profiles A - D, Figure 2.1). That is, gaps or overlaps appear when blocks are moved back along faults and movement in or out of the profile plane is not allowed. The "dip" line orientation will image the fault(s) at its steepest and the true dip of the hangingwall block. Profiles at any other angle will image seemingly shallower fault and hangingwall dips. A profile oriented obliquely to a transfer fault may image an over-rotated hangingwall block compared to the steepness of the transfer fault or the hangingwall block may even dip away from the transfer fault.

2.2.1 Hangingwall dips--orthogonal extension

Movement on individual listric or planar extensional faults within rift systems invariably gives rise to rotation or tilting of adjacent fault blocks overlain by synrift sediments. The rotation, tilt or dip of the subsided blocks is measurable at seismic profile intersections. As is true of the profiles discussed above, the direction of the dips of an individual block within an orthogonally extending system is systematically related to the bounding fault geometry, the tectonic transport direction (TT) and $e_h$ (Figure 2.2a). That is, hangingwall dip direction is exactly opposite to the dip of the bounding fault if movement on the fault is pure dip slip. If a fault surface is not planar or circular some internal deformation of the hangingwall will occur. When substantial non-elastic strain within the hangingwall block modifies its geometry, hangingwall block surface dips will vary depending on the mechanism of deformation. In the following discussion, it is
assumed that no volume change or distortion of fault blocks occurs. Despite the
simplifying assumptions, empirical evidence from several rift settings (e.g. Scott et
database on a regional scale can enhance structural and kinematic interpretations.

Consider a single normal fault which accommodates extension perpendicular to it (fault 2 in Figure 2.2a). The magnitude of the dip of the hangingwall block is determined by the curvature of the normal fault, the amount of displacement on it and any rotation of the fault due to movements on other faults in its footwall. However, the direction of the dip of the hangingwall block is perpendicular to the strike of the fault and is unaffected by the aforementioned parameters. A depth converted seismic line shot exactly perpendicular to the fault will image a recoverable or balanceable geometry. In this geometry, the dip direction of the hangingwall block is parallel to the transfer fault (1) and perpendicular to the normal fault (2). In plan view, the true dip direction (TD) of the hangingwall block is parallel but opposite to the trends of the net slip vectors of the transfer fault (NS1) and normal fault (NS2). The azimuth of NS and TD is the horizontal motion or tectonic transport (TT) direction of the subsided or rotated block and is equivalent to $\varepsilon_h$. In the same system, a fault of the opposite polarity will have mirror image hangingwall dip, net slip and tectonic transport directions. The azimuth of NS, TD and TT is parallel to the trace of the intersection of the fault planes, which in this case, is the direction of the transfer faults. Analytical and experimental models support these relationships for the case of orthogonal extension (Withjack & Jamison 1986).

2.2.2 Hangingwall dips—oblique extension

When extension is accommodated by movement on non-ideally aligned fault surfaces, producing oblique slip deformation, the relationship between fault geometry, the direction of maximum extension and the tectonic transport direction observed in orthogonal extension is less predictable. Indeed, depending on fault
orientation it is arguable which faults are acting as transfer faults and which are the extension-accommodating normal faults. Likewise, the dips of the hangingwall are not as simply related to the fault geometry. For instance, in some primarily strike slip terranes the hangingwall dip adjacent to the bounding normal fault of an oblique pull-apart basin is directly toward the normal fault, rather than aligned with the trend of the strike slip fault sets, which corresponds to the tectonic transport direction (Walker & Wernicke 1986). In contrast, strike slip segments of the extensional EARS (e.g. the Lake Rukwa rift, Section 3.3) have dips that are consistent with the extensional segments (e.g. the Lake Tanganyika and Lake Malawi rifts, Section 3.2). Experimental analog observations also indicate that hangingwalls directly adjacent to bounding oblique slip faults dip toward the faults (Braun et al. in press). However, the majority of dips recorded over the entire hangingwall surface are aligned with the tectonic transport direction (op. cit.), emphasizing the need to perform dip analysis over a broad area before inferences about kinematics are made.

Two oblique slip geometries are presented in Figures 2.2b and 2.2c. The first oblique geometry mimics analog and analytical models that produce a large proportion of hangingwall dips parallel to the direction of extension (Braun et al. in press), rather than directly toward non-ideally aligned fault surfaces. The second geometry describes a smaller deviation from the orthogonal case which is proposed as a common geometry in rifted terranes.

Figure 2.2b shows mutually perpendicular faults that both strike at 45° to the direction of extension. The fault dips and curvature are also equal. The faults will have equal amounts of dip slip, as well as equal components of strike slip. As in the orthogonal extension fault geometry case (Figure 2.2a), the trends of the net slip vectors (NS1 and NS2) of two linked oblique slip faults, responding to the same extension, are parallel to each other and to the trace of the intersection of the two fault surfaces. The trend of the net slip vectors still gives the tectonic transport direction or displacement vector of the block as in the ideal orthogonal extension case (Figure 2.2a).

An entire spectrum of fault set trends may theoretically exist, ranging from the ideal orthogonal to symmetrically oblique end members discussed above. The possible geometries are limited only by whether movement will or will not occur on a non-ideally aligned fault surface. Scotti et al. (1991) have provided convincing theoretical arguments that a fault surface may be rotated up to 75° from ideal alignment and still experience motion. The variation of the oblique slip block geometry presented in Figure 2.2c was originally derived from recent
interpretations of EARS basins (Chapter 3; see also Scott et al. 1992) and recurs in several rifted basins on Australian (Chapter 4) and South American passive margins that have been analyzed previously using proprietary industry data.

In the geometry of Figure 2.2c, fault 1 is substantially steeper than fault 2. Also, fault 1 is at a low angle and fault 2 is at a higher angle to the direction of extension. In other words, fault geometries deviate only slightly from the rectilinear patterns predicted by the orthogonal extension model. This geometry results in a higher component of dip slip motion on fault 2 and a higher component of strike slip motion on fault 1, yet the trends of the net slip vectors remain parallel to each other. The net slip vectors of both fault surfaces are parallel to the trace of the line of intersection of the fault surfaces which, as outlined below, describes the tectonic transport direction of the block. In the EARS, the consistency of the dip measurements (see Chapter 3) suggests that the tectonic transport direction is generally the same in all of the structural domains, regardless of their bounding fault orientation and whether or not extension is aligned perpendicular to rift elongation or to the main bounding faults. Below, a possible rationale for the relationship of non-ideal extensional fault surfaces to a consistent regional dip of hangingwalls is outlined. The relationship between fault geometry and the hangingwall dip direction in non-ideal extension is also investigated.

2.2.3 Tectonic Transport Direction

The relationship of the dip of the upper surface of the hangingwall block to the direction of tectonic transport in the oblique slip cases (e.g. Figures 2.2b and 2.2c) depends upon the shape of the fault surfaces and the mechanism of internal deformation. In Figure 2.2, block diagrams and plan views illustrating the relationship between fault geometries, hangingwall dip direction and tectonic transport direction. Heavy arrows labeled TT = direction of tectonic transport, mid-weight arrow labeled TD = direction of true dip of tilted/rotated block, light-weight arrows labeled NS1 or 2 = net slip vector of faults with corresponding numbers in block diagram projected to the horizontal. Block diagrams are drawn with straight listric faults, truncated above the detachment, for clarity. The same principles apply to listric faults that sole at depth (Figure 2.3). A) Orthogonal/rectilinear deformation with a normal (dip slip) fault perpendicular to and a transfer (oblique to strike slip) fault parallel to the direction of extension (TT). B) Orthogonal faults each 45° from TT and with equal dips and curvature. C) A common variation of fault orientations and dips. Fault 1 has minimal curvature and is misaligned with TT by no more than 30°. Fault 2 is subperpendicular to TT, but may be highly variable depending on the dip and orientation of fault 1 and basement fabric. Possible orientations of inclined shear planes that would produce the depicted dips of the hangingwall blocks are indicated in A and B.
deformation of the block. Extensional deformation often leads to the formation of low-angle, listric normal faults which sole out into a detachment (e.g. Dula 1991, Etheridge et al. 1989, Forsyth 1992, Lister et al. 1986, 1991, Gibbs 1984, Roberts et al. 1991). If both fault surfaces have constant curvature, it can be shown that the intersection of the two fault surfaces lies in a vertical plane and the strike of this plane is the direction of motion or horizontal displacement (tectonic transport direction) of the hangingwall.

Consider the two adjoining listric fault surfaces F1 and F2 shown in Figure 2.3. Fault surfaces can be described by:

\[
\begin{align*}
F1: \quad z_1^T & = -z_0 \left(1 - e^{-d_1/\lambda_1}\right) \\
F2: \quad z_2^T & = -z_0 \left(1 - e^{-d_2/\lambda_2}\right)
\end{align*}
\]

where \(z_1^T\) is the depth of a point on the fault surface, \(z_0\) is the depth to detachment, \(\lambda_1\) is a constant that characterizes the curvature of the fault plane, and \(d_1\) is the horizontal distance from the surface trace of the fault to the horizontal projection of any point \((x, y)\) on the fault(s) surface in a direction perpendicular to the fault strike. For \(\alpha_1, \alpha_2, \) and \(\beta\) as defined on Figure 2.3:

\[
\begin{align*}
\text{(3)} \quad d_1 & = \sqrt{x^2 + y^2 \sin(\beta - \alpha_1)} \\
\text{(4)} \quad d_2 & = \sqrt{x^2 + y^2 \sin(\alpha_2 - \beta)}
\end{align*}
\]

By the identity expansion of the sine of the difference of two angles and by substitution of right triangle identities into equations (3) and (4) they become:

\[
\begin{align*}
\text{(5)} \quad d_1 & = y \cos \alpha_1 - x \sin \alpha_1 \\
\text{(6)} \quad d_2 & = x \sin \alpha_2 - y \cos \alpha_2
\end{align*}
\]

It is apparent from equations (1) and (2) that the fault surfaces intersect at:

\[
\text{(7)} \quad \frac{d_1}{\lambda_1} = \frac{d_2}{\lambda_2}
\]

Substituting (5) and (6) into (7) yields:

\[
\lambda_2 (y \cos \alpha_1 - x \sin \alpha_1) = \lambda_1(x \sin \alpha_2 - y \cos \alpha_2).
\]
Rearranging in terms of $x$ and $y$ gives the relationship:

$$x (\lambda_1 \sin \alpha_2 + \lambda_2 \sin \alpha_1) = y (\lambda_1 \cos \alpha_2 + \lambda_2 \cos \alpha_1) \quad (8).$$

The direction of extension is equivalent to the maximum horizontal displacement vector or the direction of the intersection of the two fault planes and thus can be given as:

$$\varepsilon_h = \beta = \arctan \left( \frac{\lambda_1 \sin \alpha_2 + \lambda_2 \sin \alpha_1}{\lambda_1 \cos \alpha_2 + \lambda_2 \cos \alpha_1} \right).$$

The relationship of the coordinates of a point on the trace of the intersection of two listric fault surfaces is given in equation (8), where $\alpha_1$ is the angle which the strike of the fault, $F_1$, makes with with the $x$-axis and $\lambda_i$ characterizes the curvature of a fault as depicted in Figure 2.3. If contact is to be maintained across the fault surfaces, motion of the hangingwall block is constrained to follow the "tectonic track" defined by the intersection of the two planes. The relationship in equation (8) is clearly linear and establishes the kinematic path of the block and the direction of maximum horizontal extension. This relationship holds true for any combination of curvatures for the two faults and regardless of their orientation with respect to each other. It applies particularly to local "corners" made by two linked faults and where rotation or tilting of adjacent blocks into the faults occurs. However, if one fault is planar and not orthogonal to the other fault, or the trace of the track is curvilinear, or the ratio of the curvatures changes (e.g. ramp and flat morphologies within a non-orthogonal extensional system) the intersection of the fault surfaces is not linear. Thus, local changes in dip direction may provide clues to fault geometries at depth.

**Figure 2.3** Idealized block diagram of the "corner" of two listric fault surfaces ($F_1$ and $F_2$) and a plan view labelled with the various angular and distance relationships of the faults. $Z_0 =$ depth to detachment. $\lambda$'s = constants characterizing the curvature of the fault surfaces. $\alpha$'s = the angles of the fault traces with the X axis. d's = the perpendicular distance from the fault traces to the horizontal projection of an arbitrary point $(x, y)$ on a fault surface. $\beta =$ the computed angle of the horizontal trace of the intersection of the two fault surfaces with the X axis. $\delta u =$ translation of the hangingwall from an arbitrary point $(x_a, y_a)$ to $(x_b, y_b)$. $s =$ a random direction of dip of the hangingwall that makes an angle, $\theta$, with the X axis. See text for a development of the relationships between variables.
2.2.4 Internal Deformation of Hangingwall--Vertical Shear

In the more general case where the fault surfaces are neither planar nor circular arcs, the hangingwall block must undergo some internal deformation in order to maintain contact across the fault plane. How the dip of the upper surface relates to the "tectonic track" depends upon the internal deformation of the block, even in the simplest linear tectonic track case described in Figure 2.3. Different deformation mechanisms produce different hangingwall block surface morphologies. However, it is still possible to use dip analysis to infer regional kinematics in most geologically reasonable situations.

For example, consider the simple case of vertical shear. Particles in the hangingwall block are constrained to subside in vertical columns to fill the gap caused by the displacement of the hangingwall away from the fault surfaces. A random point in Figure 2.3, \((x_a, y_a)\), on the surface \((z^T_S = 0)\) of the undeformed hangingwall is translated in the direction of the tectonic track and is located at \((x_b, y_b)\) in the deformed state. Translation along the track or in the direction of \(\varepsilon_h\) as derived in the previous section ensures that a point originally overlying one fault will remain above that same fault after deformation. If the block undergoes vertical shear, a surface point's new depth, \(z^S\), is given by:

\[
z^S = z^T_a - z^T_b
\]

where \(z^T_i\) is the depth to the fault plane before and after translation. These depths are given by:

\[
z^T_a = -z_0 (1 - e^{-da/\lambda_1}) \quad \text{and} \quad z^T_b = -z_0 (1 - e^{-db/\lambda_1}).
\]

Substituting the pre- and post-deformation relationships above into (9) gives the new depth as:

\[
z^S = -z_0 (e^{-da/\lambda_1} - e^{-db/\lambda_1})
\]

(10).

The translation motion, \(\delta u\), is in the direction of \(\beta\) so that:

\[
x_a = x_b - \delta u \cos \beta \quad \text{and} \quad y_a = y_b - \delta u \sin \beta
\]

(11).

Since \(d_1\) in (10) is given by the relationship in (5) as:

\[
da = y_a \cos \alpha_1 - x_a \sin \alpha_1 \quad \text{and} \quad db = y_b \cos \alpha_1 - x_b \sin \alpha_1
\]
solving for $d_a$ in terms of $\beta$ and $d_b$:

$$
\begin{align*}
    d_a &= (y_b - \delta u \sin\beta)\cos\alpha_1 - (x_b - \delta u \cos\beta)\sin\alpha_1 \\
    &= y_b \cos\alpha_1 - x_b \sin\alpha_1 + \delta u \cos\beta\sin\alpha_1 - \delta u \sin\beta\cos\alpha_1 \\
    &= d_b + \delta u (\cos\beta\sin\alpha_1 - \sin\beta\cos\alpha_1) \\
    &= d_b + \delta u \sin(\alpha_1 - \beta)
\end{align*}
$$

(12).

Substituting (12) into (10) yields:

$$
Z^S = -Z_0 (e^{-\delta u \sin(\beta - \alpha_1)/\lambda_1} - 1)
$$

(13).

To simplify our calculation of the dip of the hangingwall surface we define $Z_0^D$ as:

$$
Z_0^D = Z_0 (e^{\delta u \sin(\beta - \alpha_1)/\lambda_1} - 1)
$$

a quantity that is independent of $d_b$. Substituting it into (13) the surface of the block is given by:

$$
Z^S = -Z_0^D e^{-\delta u \sin(\beta - \alpha_1)/\lambda_1}
$$

(14)

in the region over $F_1$. The same result may be derived for a point in the region over $F_2$:

$$
Z^S = -Z_0^D e^{-\delta u \sin(\beta - \alpha_1)/\lambda_2}
$$

Note that $Z_0^D$ is the same in both regions as:

$$
\frac{\sin(\beta - \alpha_1)}{\lambda_1} = \frac{\sin(\alpha_2 - \beta)}{\lambda_2}. 
$$

The dip of the hangingwall surface, $\tan\delta$, in a random direction $s$ which makes an angle $\theta$ with the $x$-axis (Figure 2.3) is given by the derivative of the surface function (14) with respect to $s$:
Chapter 2 Dip Analysis

$$\tan \delta = \frac{dz}{ds} = \frac{\partial z}{\partial x} \frac{\partial x}{\partial s} + \frac{\partial z}{\partial y} \frac{\partial y}{\partial s} = \frac{\partial z}{\partial x} \cos \theta + \frac{\partial z}{\partial y} \sin \theta.$$ 

The dip of the hangingwall surface over F1 can then be written as:

$$\tan \delta = \frac{-z_o}{\lambda_1} \left( \frac{\partial d_1}{\partial x} \cos \theta + \frac{\partial d_1}{\partial y} \sin \theta \right) e^{-d_1/\lambda_1}$$

$$= \frac{-z_o}{\lambda_1} (-\sin \alpha_1 \cos \theta + \cos \alpha_1 \sin \theta) e^{-d_1/\lambda_1}$$

$$= \frac{-z_o}{\lambda_1} \sin(\theta - \alpha_1) e^{-d_1/\lambda_1}$$

(15).

Since $\frac{-z_o}{\lambda_1}$ and $e^{-d_1/\lambda_1}$ are constants, the maximum dip direction of the hangingwall surface is when $\theta = \alpha_1 + \frac{\pi}{2}$ or perpendicular to F1. By a similar derivation the dip of the hangingwall over F2 is given by:

$$\tan \delta = \frac{-z_o}{\lambda_2} \sin(\alpha_2 - \theta) e^{-d_2/\lambda_2}$$

again indicating a maximum dip direction perpendicular to the fault (F2) when $\theta = \alpha_2 - \frac{\pi}{2}$. Dip directions into F1 and F2 will create a ridge on the hangingwall surface over the tectonic track in the direction of $\varepsilon_h$ or $\beta$ in a three dimensional extensional listric fault system that undergoes vertical shear. In this case, dip directions on the top of the hangingwall will trend perpendicular to each of the bounding faults on either side of the tectonic track. Dip domains defined by fault compartments would be subdivided by the ridge aligned with and on top of the tectonic track. The morphology of the ridge yields information of the relative curvatures of the fault surfaces and their strikes. For example, if it bisects the two fault traces symmetrically a linked pair as shown in Figure 2.2b may be inferred.

2.2.5 Internal Deformation of Hangingwall—Inclined Shear

Various two dimensional geometric models of rollover (or dip of the upper hangingwall surface in orthogonal extension) and fault surfaces which consider a
variety of internal deformation mechanisms have been investigated (e.g. Jackson et al. 1988, Groshong 1989, King & Cisternas 1991, Walker & Wernicke 1986, Walsh et al. 1991, Waltham 1989, Westaway 1992, White et al. 1986, White & Yielding 1991, Williams & Vann 1987). The various internal deformation mechanisms alter the depth to detachment and magnitude of dip of the upper surface of the hangingwall slightly, but do not affect the dip direction, which is still aligned with the tectonic transport direction in these two dimensional cases. However, Dula (1991) concludes that "the inclined shear (planar slip surfaces) and the related constant-displacement (curved slip surfaces) mechanism was the most successful in predicting observed clay-model and earth examples revealed in seismic data". White and Yielding (1991) in a similar evaluation of geometric methods independently conclude that "hangingwalls deform by bulk, antithetically inclined, simple shear." Given the geometry of two surfaces in the hangingwall, generalized three dimensional modelling successfully predicted simple, scalloped-shaped non-planar faults surfaces used in sandbox modelling by assuming bulk, antithetically inclined shear deformation in the hangingwalls (Kerr et al. 1993).

In the two dimensional case, inclined shear planes trend parallel to the normal faults or perpendicular to the tectonic transport direction (Figure 2.2a). If we invoke vertical or inclined shear planes aligned perpendicular to the tectonic transport direction in the three dimensional geometry described by Figures 2.2b and 2.3, the dip of the upper surface of the hangingwall block would align perpendicular to the shear planes or parallel to the tectonic transport direction. In other words, the shear plane or "sheet" adjacent to the intersection of the two fault surfaces (i.e. the corner) would subside the most with each successive "sheet" subsiding less away from the "corner" made by the adjoining faults. Modifications to this direction within each "sheet" might be caused by increased subsidence adjacent to the faults as in the vertical shear column case discussed above. The effect of such a modification would depend primarily on the degree of curvature of the "transfer" fault of the bounding fault pair and what angle the fault pair makes. It is obvious that as the angle of the faults deviate from rectilinear the system passes into the realm of strike slip deformation.

Observational data from the EARS (Chapter 3; see also Scott et al. 1992) suggests that the deviation of the "transfer" fault from the direction of tectonic transport does not exceed 30° and is more commonly less than 15°. In contrast, the "normal" faults vary widely from one dip domain to the next. The inclined-shear-oriented-perpendicular to tectonic transport internal deformation may explain the consistency of the dips measured in the EARS, given that major extensional fault trends appear to vary substantially along the rift. Alternatively, the deformation
and subsequent hangingwall dips recorded in analog experimental models of oblique slip geometries is accurately reproduced by a recently developed analytical kinematic model in which lines within the hangingwall block that are normal to the fault surface remain normal after deformation (Braun et al. in press). To date, observational dip data from natural rift systems does not provide any reason to prefer either inclined shear or "rule of the normal" internal deformation, although it strongly supports one or the other of these models over dips predicted by the vertical shear deformation model as described above.

2.2.6 Computing Dips from Seismic Reflection Data

Dip analysis is particularly suited to seismic reflection data because profile intersections are a point where two apparent dips of the underlying block can be measured along well constrained azimuths. The apparent dips can be used to determine the direction of the true (maximum) dip of the upper surface of the rotated or tilted block. Although simple computer programs can easily and quickly compute the real dip from two apparent dips, measurement of the apparent dips is a very time-consuming task, particularly in the large datasets required for statistically valid determinations. It may be possible to create computer algorithms that can identify coherent reflections at a particular depth and automatically measure the dip of the reflection. However, great care must be taken that the same horizon on both profiles is being measured. It is also necessary that block boundaries (i.e. fault picks) have been made so that the correct apparent dips are measured. The latter determination introduces a measure of subjective interpretation into the measurement.

If accurate velocity data are available, depth converted sections are preferred. However, the velocity-depth function at the point of intersection is identical on both profiles and reflections will be moved identically on depth converted sections. If lateral facies changes away from that point in the direction of each of the profiles are radically different, the direction of the true dip computed from time or depth sections will be different. Sediments in rift lake settings are commonly characterized by rapid lateral facies changes, but the distribution of facies is quite predictable and so can be taken into account (e.g. Flannery 1988, Scott 1988, Scott et al. 1991). In most cases, the direction of the dip remains consistent in both time and depth converted data forms (e.g. Hoffman & Reston 1992, Stone 1991). Likewise, the direction remains largely unaffected by the migration process, although the magnitude of the dip will vary in each of these processed data forms. Because reflection seismic data are normally acquired in a grid, intersections tend to be abundant, thus providing a sufficient database to provide statistical credence.
to the dip direction determination. However, regional dips in the pre-deformation setting and depositional dips may yield an apparent consistency which is not due to tectonic processes. As with all observational methods, care in discerning the possibilities for error is crucial.

*Figure 2.4* Computed dip directions of blocks underlying seismic profile intersections in the Lake Malawi rift zone of the East African Rift. There are 108 control points plotted and used in the rose diagram. The rift zone is compartmentalized by dip domains which are internally consistent within 5°-10°. Small grayed circle at about 11°45' S indicates the intersection of profiles shown in Figure 2.5.
2.3 An Application of Dip Analysis

The apparent dips of any two intersecting profiles over a rotated block are measurable and can be used to compute the direction of the true dip of the block. The computed direction of true dips of 170 intra-rift blocks in the Lakes Malawi (Figure 2.4) and Tanganyika rift zones (Chapter 3) of the EARS are remarkably consistent. Dips are consistent along more than 1000 km of the rift system, even though they are derived from a variety of structural domains and defined by composite half graben with bounding faults of varying trends (Rosendahl 1987). The profiles shown in Figure 2.5 are intersecting profiles from the N-S trending Lake Malawi rift zone (Figure 2.4) of the EARS.

The data and interpretations presented in Figure 2.5 are taken from Scholz et al. (1989) and demonstrate the use of dip analysis as well as some of the limitations. The segment of Line 815 (top panels) images a series of synthetic tilted fault blocks whose profile geometry is that of a "classic" dip line. The bold arrow locates where the left tilt block is intersected by Line 938 (bottom panels). This portion of Line 938 has been interpreted to be a single westward tilting block, cut by two minor intra-block antithetic faults. The overall apparent dip of the basement is to the west on both profiles, suggesting correlation of the faults labelled F and a true dip in the direction that bisects the trend of the profiles, or about 280°. This qualitative approach is the common mode of seismic data interpretation. Note, however, that the base synrift reflection at the profile tie in the uninterpreted section of Line 938 is practically flat-lying. Further, the reflection retains this dip until a disruption of the reflection to the left (Figure 2.5, "D"), where its dip steepens before reaching the interpreted bounding fault. Also, the rotation in the footwall of fault F is quite different in each profile. Using the measurements displayed in Figure 2.5, the dip of the block underlying the profile intersection trends 318°. Dips determined for adjacent blocks along this profile, determined at the intersections of this profile with other northeast trending profiles are within 1° of the 318° trend. This direction is in agreement with the regionally determined tectonic transport direction (Chapter 3, Scott et al. 1992). The "classic dip geometry" suggested by Line 815 (Figure 2.1, Profile E) is due to the profile's near perfect alignment with the direction of tectonic transport, even though it is oblique to the rift axis. These data support a NW-SE direction of extension as determined by onshore fault slip data (Chorowicz & Sorlien 1992), rather than the E-W direction inferred from the rift elongation and earlier interpretations of fault geometries (Sander 1986, Morley et al. 1989, 1990).
Dip analysis can also help the interpreter with individual fault correlations. Previous interpretations (Scholz et al. 1989) correlated faults F in Figure 2.5, which produces a structure striking nearly N-S at about 015°. However, the computed dip of the block underlying the profile intersection suggests a normal fault striking perpendicular to the dip at 048°. Ideal orthogonal extension models would then predict transfer faults parallel to the dip direction. Perhaps the disruption of the reflection on Line 938 (Figure 2.5, D) is a transfer structure that offsets an appropriately aligned (that is, perpendicular to the dip of the hangingwall) normal fault at F on both profiles. If there are unambiguous fault correlations which do not align in these directions, then an interpreter is signalled to develop an interpretation which incorporates pairs of faults which can produce this direction of dip or to consider interpretations with considerable internal deformation of the blocks.

It should be noted that a small error in the measurement of the apparent dips in the example in Figure 2.5 can produce a significant change in the determined dip direction, emphasizing the need for a large data set which yields a consistent result. Care must be taken to account for anomalies in the seismic reflection method, such as velocity anomalies near fault planes or lateral velocity changes in time sections. Distortions within the synrift package due to differential compaction, reverse drag or folding along normal faults needs to be considered, especially where acoustic basement is not clearly prerift basement. Depositional dip or prerift regional dip may introduce systematic errors into the analysis. This type of analysis on rifted passive margins is limited by knowledge of post rift deformation, such as thermal subsidence or postrift fault reactivation. However, if these factors are regionally consistent it may be possible to "back strip" their effect with good result.

Figure 2.5  Sections of unmigrated multifold seismic lines 815 and 938 from Lake Malawi, East Africa. Interpretations are from Scholz et al. (1989). Bold, unlabeled arrows indicate the location of the intersection of the two profiles. The strike of the profile and the apparent dip of acoustic basement at the point of intersection are indicated above each interpreted section. D on the uninterpreted Line 938 profile indicates a disruption and change in apparent dip of the acoustic basement reflection as discussed in text. For location of profiles see Figure 3.3.
2.4 Discussion and Conclusions

As any reflection seismic data interpreter will attest, interpretation of these data is commonly ambiguous. Spacing of the acquisition grid often is the controlling factor in precision of the fault correlations (i.e. if spacing between profiles is two or more times the spacing of faults, a significant potential for incorrect correlation exists). Subtleties such as the dip change in Line 938 (Figure 2.5) are often missed. If the structure is complex or the seismic data are poor, the problem is exacerbated. In most cases, dip of the acoustic basement and/or base of the synrift sequence is much more easily determined. Where dip analysis provides a consistent result with a sufficiently large dataset, it may be used to constrain tectonic kinematics, internal deformation mechanisms and fault geometries, as only a few combinations will produce this consistency. Locally, dip analysis can provide guidance for fault correlations between profiles where they are not well constrained by data. Where dips are locally inconsistent with the regional set the interpreter is signalled to look for non-ideally aligned faults or changes in the fault surfaces at depth.

It is also apparent that, depending on the deviation of the fault geometries from rectilinear, areas of polarity reversals (i.e., sense of asymmetry) or changes in adjacent half graben bounding fault geometries will be areas of "non-ideal" oblique slip deformation and may have anomalous dips due to local wrenching or even compression. Finally, convincing kinematic interpretations derived from dip analysis rely heavily on having a high quality, large, areally extensive, statistically credible dataset. That is, acoustic horizons are clearly discernible for precise measurement of apparent dips and navigation during acquisition was accurate, and therefore profile intersections are well constrained. Inferences about kinematics are restricted to the scale of the areal coverage (e.g. block, half graben, rift zone). To make inferences about regional kinematic histories, dip directions from a minimum of an entire rift zone, incorporating multiple dip domains and preferably several rift zones, are needed. Obviously, the greater the number of control points, the more confidence can be placed in the significance of the statistically determined tectonic transport direction.

Despite numerous limitations, it is proposed that in the absence of geological kinematic indicators, the dip method is a more robust tool for constraining regional kinematics than fault trends and, when combined with careful fault correlations, adds to the confidence of inferred kinematic histories. The method is particularly useful in extensional terrains where seismic reflection
data are the primary source of information. Furthermore, dip analysis allows for the identification of dip domains which must be separated by faults that decouple dissimilar basement dips, thereby identifying accommodation or transfer structures even where the faults are not directly imaged. Although there are various possible fault geometries in oblique slip deformation models, in the simple cases considered here, true dip direction of the hangingwall block aligns with the tectonic transport direction. Furthermore, dip analysis is a reliable kinematic indicator if sufficient measurements are available over a regionally consistent tectonic terrain, regardless of the strike or dip of the faults that bound the block(s).

Figure 3.1 Map of East Africa locating Lakes Tanganyika, Rukwa and Malawi, highlighted in black. Major pre-existing transcontinental dislocations, the Aswa lineament and the Rukwa trend (dotted lines) are proposed Cenozoic intracontinental transforms.
3.0 THE EAST AFRICAN RIFT SYSTEM

All my life through,
the new sights of Nature
made me rejoice like a child.
Marie Curie

3.1 Introduction

The East African Rift System (EARS) is widely considered to be the archetypal continental rift. Whether or not the current (Cenozoic to Recent) rift zones will result in continental fragmentation, a requirement inherent in proposals that the EARS rift zones are analogues for the early stages of evolution of passive continental margins, remains the topic of lively debate. Regardless of its eventual fate, there is no doubt that the interior of the African continent has been repeatedly deformed by extension throughout the Phanerozoic. In particular, a Karroo (Permo-Triassic) rift system is well documented in equatorial to south Africa by preserved structural troughs whose axes trend generally NW-SE and NE-SW, cross-cutting the Cenozoic rifts. And, an extensive Mesozoic (Late Jurassic-mid-Cretaceous) rift system is well developed in northern equatorial Africa, associated with the well known Benue Trough system of west Africa.

Because the rift systems span the entire continent and are widely dispersed in geological time, there exists a vast body of literature on intracontinental deformation in the African continent, a large portion of which pertains to the EARS, and so no attempt is made here to provide a complete review. A fairly thorough and recent listing of the literature pertinent to the EARS rift zones considered in this chapter (i.e. the Lake Tanganyika, Lake Malawi and Lake Rukwa rift zones of the Western Branch of the EARS) is provided in Rosendahl (1987).

Although each of the rift zones have been previously studied and the available data have been interpreted using standard techniques (see references in the following sections), there exists a gap in our knowledge with regards the three dimensional geometry of the individual rift zones, the kinematics associated with the Cenozoic deformation and the influence of pre-existing heterogeneities on Cenozoic rifting. The objectives of the case studies in this chapter are to refine our understanding of the three dimensional deformation of the three linked rift zones listed above at both local and regional scales and to determine how this geometry relates to the regional kinematic history of the EARS.
To accomplish these objectives, dip analysis is performed in each rift zone in addition to traditional methods of structural interpretation of reflection seismic data. The regional geometry of these three linked rift zones allows us to draw some conclusions as to the configuration of the continental lithosphere and an in-depth study of a few of selected accommodation zones provide insight into the internal geometry of linked half graben within an actively deforming rift zone. Investigation of both scales of geometries helps to constrain the relationship between oblique versus orthogonal extension and pre-existing heterogeneity.

In the following sections, the generally N-S Lake Tanganyika and Lake Malawi rift zone geometries are assessed and compared, as are the results of dip analysis in each lake. Then, the connecting NW-SE Lake Rukwa rift zone is investigated. Here, dip analysis is applied to several stratigraphic horizons in order to track the deformation in this rift zone through time. The results derived from the three rift zones are integrated to establish an overview of the regional deformation that has occurred in the EARS.

Figure 3.2 Generalized tectonic maps of the Lake Tanganyika and Lake Malawi rift zones (modified from Versfelt and Rosendahl, 1989). Major pre-existing dislocations discussed in text are labeled Rukwa and N. and S. Chamaliro, other smaller-scale pre-existing fabrics are schematic. Heavy ticked lines within the lakes indicate major bounding faults of the rift zone. A-D refer to major polarity switching accommodation zones (AZ): A = Kavala Island Ridge AZ; B = Burton's Bay Peninsula AZ; C = Chilumba AZ; D = Likoma AZ. Darker shading within the lakes indicate the "deeps" created by composite half graben subsidence.
3.2 Oblique slip Deformation in Extensional Terrains -- A case study of the Lakes Tanganyika and Malawi Rift Zones

3.2.1 Introduction

The analysis of seismic data collected from the EARS lakes (Figure 3.1) has documented several features predicted by orthogonal extension, including the rarity of full-graben morphologies and the polarity switching of half graben along the axis of the rift (e.g. Reynolds 1984, Rosendahl & Livingstone 1983, Rosendahl et al. 1986, 1988, Rosendahl 1987, Sander 1986, Sander & Rosendahl 1989, Specht 1987, Specht & Rosendahl 1989, Scholz et al. 1989). However, previously interpreted mapped fault geometries are incompatible with the rectilinear fault geometries of orthogonal extension and clear-cut rhombo-chasm geometries expected in strike slip terrains have not been mapped either (Figure 3.2).

Instead of the two primary fault sets predicted by orthogonal extension, the EARS model (Rosendahl 1987) proposes that rifts are composed of quasi-arcuate bounding faults and compartmentalized by accommodation zones (Figure 1.3a). The bounding faults are considered elongate along the rift axis (N-S) and the tangent to their point of maximum subsidence coincident with bounding faults of the orthogonal extension model. Sander & Rosendahl's (1989) and Morley's (1988) interpretations suggest extension orthogonal to the rift axis, or generally E-W. However, the cross-cutting structures or accommodation zones separating half graben of opposing polarity appear to have no consistent trend or relationship to the direction of extension or the rift axes (Versfelt & Rosendahl 1989).

Surface structural mapping of the northernmost border fault in Lake Malawi has documented slickenslides plunging 34° at 298° (Wheeler 1989). Wheeler (1989) infers WNW-ESE extension from the structural data, but also interprets NE-SW trending "transfer faults" which offset the border fault. None of the interpretations identify rift perpendicular structures. EARS arcuate geometries are not strictly balanceable or retrodeformable in the third dimension; that is, they require a substantial amount of nonelastic strain in adjacent, unfaulted regions and within fault blocks. The enigmatic fault geometries of the EARS model suggest that neither orthogonal extension nor strike slip pull-apart mechanisms are capable of accounting for the observed faulting within the EARS.

In the following sections, dip analysis as described in Chapter 2 is used for inferring kinematics independently from mapped fault geometries within EARS.
rift zones. Based on arguments in the preceding chapter, kinematics are inferred on a half graben (i.e. dip domain), rift zone, and rift system scale from computations of the true dip direction of subsided/rotated blocks. The distribution of control points through several connected rift zones allows us to compare the tectonic transport direction to measured dips. Dip analysis is applied to seismic data from both the Lake Tanganyika and Lake Malawi rift zones, where presumably the apparent acoustic basement dips have not been modified by postrift deformation or burial. The results of the dip analysis and an alternative structural interpretation of the same data used by previous workers (Rosendahl 1987, Rosendahl et al. 1988, Flannery 1988, Morley 1988, Sander 1986, Sander & Rosendahl 1989, Scholz et al. 1989, Specht 1987, Specht & Rosendahl 1989, Versfelt 1988, Wheeler 1989), complemented by 28-kHz echosounding data, are used to support the conclusion that these rifts are the product of oblique extension, that is, rifting accomplished by extension at an oblique angle to the rift axis or elongation.

3.2.2 Application of Dip Analysis to the Lakes Tanganyika and Malawi Rift Zones of the East African Rift System

Figure 3.3 shows the distribution of multifold seismic (MFS) profiles and 28-kHz data used in this study. It also locates 170 MFS profile intersections which provide control points for dip analysis within Lakes Tanganyika and Malawi. The 108 control points in Lake Malawi are more evenly distributed throughout the rift zone and more numerous than those from Lake Tanganyika. At each intersection, the apparent dip of the base synrift reflection along the strike of both profiles has been measured. Although this is a time-consuming and tedious task, this is the only way to determine the actual dips of rotated or dipping intra-rift blocks. As mentioned in Chapter 2, a procedure to compute the dips at particular horizons automatically could and should be developed to aid in balancing structural interpretations in the third dimension. The direction of the maximum or "true" dip of the acoustic basement block underlying the synrift section has been

Figure 3.3 Trackline map of seismic data used in this study. Heavy lines locate multifold seismic coverage. Lighter lines are single channel and 28-kHz traverses. Intersections of heavy lines locate points where dip measurements and calculations plotted in Figure 3.4 were made. Numbered and dashed lines show location of data shown in Figures 2.5, 3.5, 3.6, 3.9, 3.10, 3.11 3.14 and 3.15. Interpretations of areas presented in Figures 3.7, 3.8 and 3.12 are outlined by boxes.

Figure 3.4 (Next page) Rose diagrams and areal distribution of directions of true dips of subsided blocks underlying multifold seismic profile intersections. These data divide the rift zone into dip domains within which true dip directions are consistent within 5°. "Likoma Data" refers to data points that lie within the interpreted area in Figure 3.7. See text for further discussion.
computed from the apparent dips. The directions of the true dips of blocks at each profile intersection are plotted in Figure 3.4. The rose diagram labeled ALL DATA demonstrates that the dip data are regionally consistent and indicate a tectonic transport direction of NW-SE to WNW-ESE for the underlying blocks (Figure 3.4). These data provide strong evidence that NW-SE regional extension has created these rift zones.

The N-S elongation and long, linear shorelines in some sub-basins of these lakes have been used to argue for E-W extension (e.g. Morley 1988), which should produce E-W synrift dip directions. However, dips of tilted blocks in the Lakes Tanganyika and Malawi rift zones are predominantly NW-SE. When data from each rift zone are plotted separately, the Tanganyika data record a more NW-SE dip direction at N50°-60° W compared to the Malawi data, which are clustered at N60°-70° W. This is somewhat surprising given the concentration of data points in the most N-S trending portion of the Lake Tanganyika rift zone.

In both rift zones there are concentrations of dip directions in the N70°-90° W range (Figure 3.4). For example, WNW-ESE trends occur at the very southern end of Lake Tanganyika and adjacent to the NNE-SSW trending peninsula at about 4°S. There are only two E-W block dips in Lake Tanganyika; they lie in the NW-SE trending section of the rift at about 5°15' S and lie along the same E-W profile which may indicate a navigational problem (e.g. a kink in the ship track). In Lake Malawi data, the WNW-ESE trend is concentrated in the extreme north and south of the rift zone. An isolated E-W dip direction occurs at about 11°S. A NW trend (N40°-50°W) trend is recognizable in both data sets.

There is an obvious compartmentalization of both lakes into dip domains dominated by SE-ESE or NW-WNW dips in Figure 3.4. Computed dips in each compartment, or dip domain, are internally consistent within 5°-10°. Consideration of some specific examples of computed dip information shows how the use of the dip analysis helps to clarify the interpretation of structures and tectonic processes from these data.

Line 815 trends NW-SE and appears to image a series of "balanceable" tilt blocks, as described in Chapter 2 (Figures 3.5 and 2.5). Closer examination of the largest block reveals subvertical disruptions of synrift reflectors that separate changes in the dip of the acoustic basement reflector. Computation of the true dip direction of the blocks underlying profile intersections yields tectonic transport directions of N38°W, N42°W and N43°W. If this were an orthogonally extending system normal faults should trend approximately N50°E with transfer faults.
trending parallel to the direction of tectonic transport, approximately N40°W. However, fault correlations very close to profile ties constrain the orientations of the imaged faults to N20°E and N30°W, requiring oblique slip on both fault sets to accommodate the transport direction of the blocks.

An example of a limitation of dip analysis is the tie of Lines 908 and 8311 from Lake Tanganyika (Figure 3.6). In this case, Line 908 which trends 080° has an apparent acoustic basement dip that is flat lying or possibly to the west directly under the tie. This reflection geometry appears to be due to internal deformation of the synrift section by faulting at D. Other possibilities include "reverse drag" or folding next to the normal fault, differential compaction or velocity anomalies near the fault plane. Line 8311, trending at 184° has a base synrift reflection dip of 12.5° S. These relationships yield a computed dip to the S or SSW. However, the overall dip of acoustic basement on Line 908 is definitely easterly and measures about 9° from fault E to fault C. Dip direction data so derived indicate a tectonic transport direction for this block of S41°E consistent with the overall predominance of this trend.

Note that in both the Lake Malawi and Lake Tanganyika rift zones, where two apparent dips are measurable, computed true dip directions consistently trend between N40°-70°W or S40°-70°E, regardless of mapped fault trends. The two examples presented above are from good ties with deep (base) synrift reflection correlations which are convincingly top of basement. Block dips are aligned within 5° of each other, even though these measurements are from opposite polarity composite half graben nearly 1000 km away from each other along the rift. Data points that are inconsistent with the NW-SE trend usually can be correlated with dubious navigation evident from misties at water bottom levels, areas with strong pre-existing anisotropies (see discussion below), or mismatched reflectors in areas of shallow acoustic basement.

**Figure 3.5** Multifold seismic Line 815 and coincident 28-kHz Profile 41 from Lake Malawi. Profiles are displayed at the same horizontal scale. Ticks on bottom of profile 41 indicate where other multifold seismic data tie with this line and dip computations were made. Vertical exaggeration (V.E.) is computed with respect to the lakefloor. Refer to Figure 3.3 for location of profile.

**Figure 3.6** (Next page) Migrated multifold seismic Lines 8311 and 908 from Lake Tanganyika. Inset interpretations indicate faults and acoustic basement (heavy lines) and the water/sediment interface (dashed lines). Letters refer to faults discussed in text. Vertical exaggeration (V.E.) is computed with respect to the lakefloor. Refer to Figure 3.3 for location of profiles.
Chapter 3

East African Rift System

This study supports the contention that quantitative dip analysis, both on the domain and individual control point levels, is a robust method of constraining regional kinematics that can help to constrain fault geometries in extensional terrains and provides an important tool for determining structural patterns where data are sparse. For instance, changes in dip domains (Figure 3.4) from westerly to easterly trends document the half graben polarity switches noted by previous interpreters (e.g. Reynolds 1984, Rosendahl et al. 1986, Rosendahl 1987, Sander 1986, Sander & Rosendahl 1989, Specht 1987, Specht & Rosendahl 1989, Ebinger et al. 1987, Versfelt 1988, Morley 1988, Morley et al. 1990). Dip domain analysis identifies where faults are required to decouple basement dips even where no specific data are available.

In the study area a NW-SE tectonic transport direction is indicated which suggests that the major bounding faults are oblique slip. Given the apparent nonrectilinear fault geometries in the Lakes Tanganyika and Malawi rift zones, the consistency of the dip direction determinations requires that the internal bulk deformation mechanism be able to accommodate variably trending faults. Two possibilities: the inclined shear and "rule-of-the-normal" model recently developed by Braun et al. (in press) have been identified in Chapter 2. In the next section, the results of the dip analysis are used to reinterpret the deformation patterns within portions of the Lakes Tanganyika and Malawi rift zones.

3.2.3 A Reinterpretation of the Structure of Parts of the Lakes Tanganyika and Malawi Rift Zones

MFS data (Figure 3.3) collected by Project PROBE in Lakes Tanganyika and Malawi show a wide variety of fault geometries which commonly are not balanceable in the plane of the profile despite the fact that many lines from Lake Tanganyika are oriented close to the conventional dip and strike alignments with respect to the rift axis. It is probable that most faults in the EARS setting are oblique slip (Daly et al. 1989, Versfelt 1988, Wheeler 1989) as data herein suggest. Referring again to the schematic fault and profile geometries from an ideal orthogonally extending system in Figure 2.1 wherein the extension is directed NW-SE: profiles A - E are aligned in the plan view as the EARS data profiles, from which they are derived, are oriented in real space. This suggests that all of these profile geometries conceivably could derive from a NW-SE extension as depicted. However, profile location within the rift system and with respect to each other is not realistic, nor is the overall geometry of the schematic rift system comparable to natural systems. Therein lies the problem of applying orthogonal extension model concepts and
fault geometries to these rift zones: to interpret specific line geometries within a regionally consistent tectonic setting.

A segment of Line 815 (Figure 3.5) from central Lake Malawi images classic "domino-style" tilted fault blocks. The trend of the profile lies about midway between the "dip" and "strike" directions expected from the N-S elongation of the rift system but is close to the overall WNW transport direction determined from dip data (Figure 3.4). The faults in Line 815 dip at approximately 50° on a true-scale, depth converted section. Extension is calculated to be approximately 13% across these blocks. Although a relatively low extensional strain, this is higher than the average from central Lake Tanganyika (7.4%; Morley 1988) and higher than similar calculations done on other lines from Lake Malawi (6-7%; Specht 1987). Basement dips of these blocks are some of the highest recorded in both lakes. True block dips and the balanceable geometry of this profile are suggestive of a true "dip" line (e.g. Profile E; Figure 2.1).

Consideration of ideal extensional geometries above would argue for a direction of extension subparallel to the segment of Line 815 shown in Figure 3.5. The NW trend of Line 815 is at an oblique angle to the overall N-S rift elongation, not perpendicular to it. In general, seismic lines shot in Lake Tanganyika trend N-S or E-W to ENE-WSW, so it is not surprising that the classic tilted fault block extensional geometry of Line 815 is not generally imaged there. Furthermore, the irregular and widely spaced grid make fault correlations tentative at best. In contrast, the denser and more uniformly spaced MFS grid in Lake Malawi (Figure 3.3) contains a number of profiles which are aligned subparallel to the above proposed maximum extension direction. The Lake Malawi herringbone grid was designed by Rosendahl partially to address the question of "dip direction," yet this
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section of Line 815 is the only section in the entire data set which shows this apparently balanceable geometry.

Lines adjacent and subparallel to Line 815 image tilted blocks, as do NE-SW oriented cross-tying lines (e.g. Figure 2.5, Line 938), but none are strictly balanceable. That is, relatively flat to over-rotated synrift reflectors abut steep faults, synrift reflectors dip and diverge into or away from each other or are otherwise not balanceable in the plane of the profile, as in profiles A-D, Figure 2.1. This implies one or all of the following:

- Profiles are not true "dip" sections in an orthogonally extending system or
- More than one set of faults with different strikes intersect the profile or
- There is some strike slip motion on the imaged faults.

These complex and "unbalanceable" geometries can be found in virtually every profile in the EARS data set. In other words, no consistent relationship between geometry and profile orientation is evident, although sections of NW-SE oriented lines adjacent to the segment of Line 815 described above come the closest to being balanceable "dip lines".

The inability of an orthogonal extension model (Figure 1.3) to adequately predict geometries encountered in these data, and remain consistent with the general N-S elongation of the rift system, argues that this is not a simple extensional system. It is possible that the rift zones are the response to an "extension + shear" mechanism as modeled by Withjack and Jamison (1986), in which case, fault geometries are most consistent with a right-lateral shear mechanism (their Figure 9). Alternatively, the tectonic transport direction may have changed with time (Pollitz 1991, Ring et al. 1992, Strecker & Bosworth 1991) producing the oblique slip, complex fault geometries in these rift zones. However, the remarkable consistency of the dip data suggest that the linked fault system that comprises the rift formed in response to a regionally consistent finite strain distribution and a simply definable tectonic transport direction.

The following reinterpretation of the MFS data used by previous workers from selected areas within the Lake Tanganyika and Lake Malawi rift zones suggests that NW-SE to WNW-ESE directed extension, possibly in two distinct pulses, created the present day structural geometry. Reactivation of pre-existing planes of weakness controls local variations of fault trends, which form imperfect
rhombs that define the dip domains. The imperfect development creates "corners" which, as demonstrated for the variations of adjacent structural domains in the strike slip regime of the Transverse Ranges (Scotti et al. 1991), can be reactivated by a broader range of stress configurations than fault segments of a single trend. The reinterpretations are kinematically consistent with the tectonic framework derived from the dip analysis.

3.2.3.1 Central Lake Malawi

Figure 3.3 illustrates the multifold data set used in the interpretation of central Lake Malawi, where data density and quality are the highest of the areas discussed in this study. In addition, 28-kHz data (e.g. Profile 41, Figure 3.5) collected continuously during MFS profile acquisition and during "dead head" runs between, were used to help constrain the strike and extent of fault segments. A major half graben polarity switch in Lake Malawi occurs at the largest kink or bend in the lake (labeled D in Figure 3.2 and C in Figure 3.7). This area is associated with the southern branch of a major continental scale dislocation, the Chamaliro zone (Versfelt & Rosendahl 1989). The structural patterns interpreted from the data in this region are shown in Figure 3.7. The overall pattern is rectangular to rhombohedral or trapezoidal blocks. Blocks tend to be elongate NW-SE except where major basement structures intersect the rift. Faults that have a northwesterly trend appear to be more continuous.

Within the NW-SE fault set, a N30°W trend prevails over a subordinate trend of N45°W south of latitude 11°25'S (Figure 3.7). North of this latitude this relationship is reversed with the N45°W trend becoming more prominent. The northeasterly fault set is less continuous and more variable, with trends ranging from N17°E to N63°E. This mapping emphasizes the existence of a linked fault system consisting of two principal, but somewhat irregular, fault sets with NE-SW and NW-SE trends. The two fault sets define approximately rectangular blocks with dimensions of about 2 by 6 km.

Dips of synrift reflectors in this region argue for a tectonic transport direction of NW-SE (Figure 3.4, Likoma Data rose diagram). Within this tectonic
setting, extensional models predict that the NE trending fault set should experience a higher component of dip slip motion. Referring again to line 815 (Figure 3.5), tilted fault blocks are overlain by consistently diverging reflectors and similar block rotations against the imaged faults. Apparent measured dips of blocks on this profile are consistently greater than 10°, whereas those on NE trending tie lines range between 4° and 10°. Orientations of the mapped faults are N20°E and N30°W. The nearly balanceable geometry of this profile is consistent with a NW-SE extension direction. The compartment mapped from this profile and those which tie it is labeled A in Figure 3.7. Two fault sets bounding trapezoidal to rhombohedral blocks are indicated.

A flat bottomed-deep labeled B in Figure 3.7 at the northern extent of the study area has been represented previously as a rectangular feature with its long axis trending N-S (e.g. Yairi 1977, Specht 1987, Scholz et al. 1989). This is the deepest point in the lake (Figure 3.2) and is underlain by 3 to 4 km of sediment. Opening a deep with a N-S elongate geometry seems more plausible if one assumes a N-S strike slip or an E-W extension deformation mechanism. One set of MFS profiles that cross this deep trend nearly E-W (Figure 3.3) and image flat lying reflectors truncated by subvertical faults. NW-SE profiles show growth against the bounding faults. Hangingwall block dips in this region trend NW-SE (Figure 3.4). Computed true dips are some of the highest in the entire data set. Dip values (in time) cluster around 18°. The added control provided by the 28-kHz data allow a more precise determination of fault correlations forming several of the walls of the deep, which indicate the morphology of the north end of this deep is probably rhombohedral with sidewall trends of N30°W and N35°E (Figure 3.7). Again, a fault system consisting of two linked sets is indicated. One fault set apparently trends at a low angle and obliquely to the direction of tectonic transport, whilst the other is aligned subperpendicular to the extension direction. This may indicate that the bounding fault is a series of en echelon relay faults, rather than the long, linear N-S trending structure previously mapped.

The 370 m and 110 m isobaths along the western shoreline are indicated by the heavy solid lines in Figure 3.7. These isobaths are associated with steep scarps or scoured channels on the 28-kHz data which can be correlated with deeper structures in the MFS data (e.g. Figure 3.5). Dense 28-kHz coverage in this area has allowed a refinement of contours here. Control points require that the isobaths "jog" along N30°W and N30-40°E trends. The bathymetric trends defined by the 370 m isobath lose definition where the southern branch of the Chamaliro dislocation crosses the lake. The undulating nature of the bathymetric contours
supports the contention that deformation in Central Lake Malawi is occurring on NE-SW and NW-SE, rather than N-S, fault sets.

3.2.3.2 North Central Lake Tanganyika

Problematic, unbalanceable geometries are prevalent on all profiles in Lake Tanganyika. In the northernmost sub-basin of Lake Tanganyika, interpretation of MFS data is hampered by severe ringing caused by shallow water depths and coarse-grained sediment input from the axial Ruzizi river. In general, data are sparser throughout Lake Tanganyika than in Lake Malawi, so the fault geometry is not as well constrained. Also, 28-kHz and single-channel data do not significantly increase the data density in key areas (Figure 3.3). The reinterpretation of the Tanganyika data set (Rosendahl et al. 1988) consists of identical fault picks as previous interpretations of the profiles (e.g. Morley 1988, Reynolds 1984, Sander 1986, Sander & Rosendahl 1989). In addition, the available computed block dip information (Figure 3.4) is considered and individual reflector/fault geometries (e.g. Figure 3.6) and bathymetric subtleties are examined to determine likely fault trends. The consistency of the dip analysis throughout both rift zones has also led to the use of the structural patterns observed in central Lake Malawi as an interpretation template.

Dip data compiled from the lake as a whole support NW-SE extension (Figure 3.4). The structural pattern in northern Lake Tanganyika in Figure 3.8 suggests that the N30°W trend predominates among the NW-SE faults, whereas the NE-SW trending structures are quite variable. This interpretation is presented as an alternative to previous models that does not contradict the existing data and

**Figure 3.9** 28-kHz Profiles 51 and 23 from the Chilumba accommodation zone, Lake Malawi. Both profiles are coincident with multifold seismic (MFS) lines and the observed scarps can be identified as the surface expressions of basement faults. Vertical exaggeration helps delineate changes in bathymetric gradient which define faults D through G. Blow-up at D shows how recognition of acoustic character changes can also define block boundaries in these data. Previous interpretations using only the multifold seismic data correlated fault E in Profile 51 with fault F in Profile 23. Horizontal lines in all of the 28-kHz data are at 30 foot intervals on these analog paper records. Vertical exaggeration (V.E.) is computed with respect to the lakefloor. See Figures 3.3 and 3.13 for location of profiles.

**Figure 3.10** (Next page) 28-kHz Profiles 21 and 161 from the Chilumba accommodation zone, Lake Malawi. These profiles show a variety of features including channels, scarps, gradient changes and acoustic character changes which are used to define lineaments in these data. A - F mark features mapped in Figure 3.13. Vertical exaggeration (V.E.) is computed with respect to the lakefloor. See Figures 3.3 and 3.13 for location of profiles.
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is geometrically consistent, both internally and with the inferred tectonic transport direction. Structures within Lake Tanganyika are not likely to be as orthogonal, regular, or linear as depicted here; but, existing data are not capable of resolving the internal geometries or individual blocks. In this reinterpretation, compartments of two oblique slip fault sets comprise the larger composite half graben dip domains. The trends of these faults vary along the strike of the rift but seem to be relatively consistent within each dip domain. It is easy to visualize a sinusoidal sub-basin pattern as syn-sedimentary fill obscures small-scale steps in faults.

3.2.4 Structure of Selected Accommodation Zones in the Lakes Tanganyika and Malawi Rift Zones

Since dips and therefore block tilts or rotations are consistently aligned NW-SE, accommodation zones of different orientations must have different internal geometries. Where adjacent half graben fault orientations are reasonably consistent with each other, appropriately aligned accommodation zones may be relatively simple structures (i.e. a transfer fault). If the intervening accommodation zone is inappropriately aligned, its geometry still should be fairly simple, consisting of fault pairs which mimic the fault orientations of the juxtaposed half graben in an en echelon fashion along the zone. However, where neighboring half graben have significantly different fault pair orientations, accommodation zones will display a plethora of fault orientations, including wrench and compressional features similar to restraining bends in strike slip regimes.

The multifold seismic grids collected in Lakes Tanganyika and Malawi have line spacings that range from about 5 km to 20 km in worst cases (Figure 3.3). Structural interpretations using only this data set (Morley 1988, Morley et al. 1989, 1990, Reynolds 1984, Rosendahl et al. 1986; 1988, Rosendahl 1987, Sander 1986, Sander & Rosendahl 1989, Scholz et al. 1989, Specht 1987, Specht & Rosendahl 1989) are often poorly constrained because fault correlations between control points are widely spaced. However, the polarity switches labelled A-D in Figure 3.2 have been recognized by all workers and are herein termed accommodation zones (Rosendahl et al. 1986, 1987). Versfelt (1988) and Versfelt and Rosendahl (1989) proposed that these polarity switches were genetically related to major pre-existing anisotropies, but they do not discuss the internal geometries of the polarity reversals. Morley et al. (1990) described the internal geometry of some of the polarity switches in Lake Tanganyika, and coined the term "transfer zones" for structures that connect major bounding faults. However, he did not relate them to
the regional tectonic regime which he believed is best described by E-W extension (Morley 1988).

Polarity switching accommodation zones are key to understanding the evolution of rifting on a regional scale. Increased data density using 28-kHz echosounder data has elucidated the internal geometry of portions of some accommodation zones in the EARS. The assumption is made that bathymetric lineaments and deformation of the uppermost reflections have structural significance in this synrift setting. That is, synrift structure is modified by synrift sedimentation but not masked by any overlying postrift sedimentation. There is some evidence from seismic stratigraphic analysis of these data that extension has temporarily ceased in these lakes; for instance, the upsection flattening of synrift reflectors in profiles 908 and 8311 (Figure 3.6). Draping of modern reflectors over tilted fault blocks in Profile 41 (Figure 3.5) may be indicative of a recent period of quiescence in some parts of these rift zones. However, the 28-kHz profile shown in Figure 3.5 still clearly records the faults imaged in the MFS profile. Also, the evaluation of 12,000 km of echosounder data and comparison with 3300 km of coincident MFS data from Lake Malawi (Scott 1988, Scott et al. 1991) indicates that structural features can be distinguished from wave and current induced features. The intense vertical exaggeration enhances recognition of these features in many cases (e.g. Figures 3.9, 3.10 and 3.11). In the following sections, the 28-kHz data are included in an interpretation of deformation within some accommodation zones in the Lakes Tanganyika and Malawi rift zones.

Figure 3.11 A segment of 28-kHz Profile 35 from the Chilumba accommodation zone, Lake Malawi. This profile is subparallel but approximately 1 km away from Profiles 21 and 161 (Figure 3.10). Correlation of features (e.g. B and C) on these three profiles definitively constrain lineament trends in this area. Acoustic reflections diverge to the north (upslope) on all three blocks. Vertical exaggeration (V.E.) is computed with respect to the lakefloor. See Figures 3.3 and 3.13 for location of profile.

Figure 3.12 (Next page) Bathymetric lineaments and recent structural patterns of the Chilumba accommodation zone, Lake Malawi based on both multifold seismic and 28-kHz data and mapped at 1:100,000. In the SW corner of the map area lineaments are aligned N30°W and N70°E. Profiles shown in Figures 3.9 - 3.11 are located within the inset area. Detailed interpretation of the inset area is provided in Figure 3.13. Bathymetric lineaments are controlled by a delta system to the east and so are not as clearly tied to underlying structure, although the delta lobe front clearly is influenced by the mapped lineaments. To the north, lineaments are aligned N45°W and N45°E and may indicate an increasing influence on modern deformation by the pre-existing Rukwa trend (Figure 3.2).
PROFILE 35

1 KM

V.E. 50 : 1

N

S

240 m
250 m
260 m
270 m
280 m
290 m
300 m
310 m
320 m
3.2.4.1 Chilumba Accommodation Zone

In northern Lake Malawi, the Chilumba accommodation zone (C in Figure 3.2) separates a deep-to-the-east basin to the north from a deep-to-the-west basin to the south and marks a major change in trend of the rift axis, from NW-SE to nearly N-S. According to Versfelt and Rosendahl (1989) this neck in the rift can be correlated with the northern branch of the Chamaliro dislocation, a major NE trending transcontinental lineament of Pan-African age, which also separates basement fabric terrains.

Figures 3.9, 3.10 and 3.11 illustrate examples of feature correlation used in this analysis. The profiles in Figure 3.9 are coincident with MFS data where faults D through G can be seen to continue to acoustic basement. Dense 28-kHz echosounder data coverage (Figure 3.3) allows precise determination of modern bathymetric trends on the northwestern side of the accommodation zone. Bathymetric "scarps" or lineaments were mapped at a scale of 1:100,000 (Figure 3.12). Although structures continue across the lake to the east (see profiles 816 and 805 in Scholz et al. 1989), coarse grained sediments from a major drainage system, the Ruhuhu River, entering the lake on the east inhibit penetration in the 28-kHz data and also obscure resolution in the MFS data. Furthermore, bathymetric lineaments are heavily influenced by delta deposition, making their connection to underlying structure ambiguous. The extent of the delta lobe is mappable from a change in acoustic character in the 28-kHz data (from profiles with sub-lakefloor reflectors to those with no sub-lakefloor reflectors). The lobe front follows mapped NE-SW and NW-SE lineaments, giving it an undulating morphology. In the southwest corner of the mapped area in Figure 3.12, where data coverage is best, mapped trends are dominated by N30°W and N70°E. To the northwest, lineaments are oriented N45°W and N45°E.

Locations of the profiles shown in Figures 3.9, 3.10 and 3.11 and a plan view of the correlated bathymetric lineaments is provided in Figure 3.13. Although channeling along fault B is apparent and gives the impression of relief towards the north, growth into fault B from the south is recorded by diverging reflectors in profiles 161, 21 and 35 (Figures 3.10 and 3.11) and clarifies the sense of throw. The closely spaced and variously oriented profiles used to produce the map in Figure 3.13 allow a high degree of confidence in the orientation and length of mapped fault segments. This interpretation strongly favors a two-fault-set system aligned approximately at N30°W and N70°E. Block size is as small as 1 km. The N70°E trend appears to be truncated by the more extensive and throughgoing N30°W trend. Previous interpretations based on MFS data correlated fault G
(Figures 3.9 and 3.13) in a similar fashion, but correlated faults D and E in Profile 51 with fault E and F in Profile 23, respectively, so that NW trending structures remained unrecognised.

The recognition of two linked fault sets at a high angle to each other is not consistent with previous interpretations. The highly schematic, vertically exaggerated block diagram in Figure 3.13 suggests a rectilinear fault pattern. However, the plan view clearly shows that the two trends are not strictly orthogonal. In addition, there is a change in orientation of the NE trending fault set from N50°E at fault G to N70°E at fault A, giving these blocks a distinctly trapezoidal to triangular aspect. Block dips of the three control locations on the western side of the accommodation zone are oriented N65°W, S60°E and S62°E (Figure 3.4). The dip data suggest a tectonic transport direction that neither fault trend is perpendicular or parallel to. Within the context of the oblique slip deformation model presented in Figure 2.2, the NE trending fault set should have a higher component of dip slip, as it accommodates the majority of the extension, than the NW trend which serves to decouple adjacent blocks along subvertical structures.

3.2.4.2 Likoma Accommodation Zone

Another major polarity switch in Lake Malawi occurs at the largest kink or bend in the lake (D in Figure 3.2). This area is associated with the southern branch of the Chamaliro zone (Versfelt & Rosendahl 1989). The structural patterns of adjacent areas have been discussed in Section 3.2.3.1 and are represented in Figure 3.7. The overall pattern is that of NW-elongate rectangular to rhombohedral or trapezoidal blocks. Analogous to the Chilumba accommodation zone, structural trends of N30°W prevail over a subordinate trend of approximately N45°W close to the actual polarity switch, while further away the N45°W trend becomes more prominent.

The location of the actual polarity switch of major bounding faults (C in Figure 3.7) appears to be a subtle depression which approximates a NW-SE elongate rectangle or rhomb. The N30°W walls of the depression are apparently truncated to
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the south by a N45°E fault, which parallels the trend of the southern branch of the Chamaliro dislocation, and is bounded to the north by a more NNE fault. The polarity switch apparently occurs as a scissoring along an extensive NW trending structure that deviates approximately 30° from the tectonic transport direction of blocks in this area (Likoma Data rose diagram, Figure 3.4).

3.2.4.3 Kavala Island Ridge and Burton’s Bay Peninsula Accommodation Zones

Two major polarity switches (Figure 3.2) have been recognized in northern Lake Tanganyika: A) the Kavala Island Ridge Accommodation zone and B) the Burton’s Bay Peninsula Accommodation Zone. A horst separates compartments of opposite polarity in both zones (e.g. Figure 3.14). This contrasts with the style of polarity switching in Lake Malawi which seems to occur at bathymetric saddles or subtle depressions.

Comparison of the horst in adjacent profiles (e.g. Lines 224 and 214 in Figure 3.14 and Profiles 214 and 216 in Figure 3.15) suggests that the horst is not a simple, undeformed ridge. It changes along strike both in outcrop morphology and in seismic expression. A reinterpretation of the adjacent half graben helps explain these changes. Profile 214 is coincident with MFS Line 214 (Figure 3.14) and profile 216 is also coincident with MFS data (not shown). Previous interpretations have correlated faults trending nearly perpendicular to these two profiles, or subparallel to the western edge of the horst, across a distance of more than 20 km. However, the morphology of each of the profiles is quite different: in Profile 214 a relatively steep gradient merges directly into a stepped, double scarp onto the horst, whereas Profile 216 shows a steep gradient merging into a flat-lying deep adjacent to a single major scarp. Also, the sense of dip atop the horst is opposite on the two profiles.

Profile T1000 is nearly perpendicular to Profiles 214 and 216 (Figure 3.15). Tilted fault block morphology is readily apparent in this profile. Comparison of these three profiles suggests that active present day faults strike ENE-WSW and have a significant component of dip slip motion on them. That is, there exist structures that trend subparallel to Profiles 214 and 216 and at a high angle to Profile T1000. Long, linear shorelines with the ENE-WSW trend on both sides of the kink in the lake also supports the interpretation of faults of this orientation. The previous mapping of this area correlated faults of a NNW-SSE trend (Morley 1988, Reynolds 1984, Rosendahl et al. 1988, Sander 1986, Sander & Rosendahl 1989) but recognized no cross-cutting structures. The mis-match in rotation of adjacent blocks on Line 214 suggests that at least two trends of faults cross this profile (Figure 3.14).
A lack of profile intersections means that only apparent dip information is available through this area. However, dip data in other segments of the lake support NW-SE extension (Figure 3.4). Northeasterly striking normal faults suggested for this half graben are consistent with this regional kinematic interpretation. In any event, the Kavala Island Ridge is most certainly not a continuous, unbroken horst, but rather a series of NW-elongate blocks which are most likely offset by NE trending normal faults and NW trending oblique to strike slip faults. In contrast, the Burton's Bay Peninsula which trends NNE and is expressed as a bathymetric feature well into the lake, is composed of more equidimensional blocks (Figures 3.2 and 3.8)

The polarity switching accommodation zones studied herein all appear to be associated with some pre-existing heterogeneity. The Lake Tanganyika examples appear to coincide with a resistant zone where extensional forces were insufficient to continue deformation in the manner found in adjacent half graben, creating ridges or horsts. The Lake Malawi examples are spatially related to branches of the Chamaliro Dislocation, which seems to have acted as strain guides allowing deformation to transfer across the rift axis to a half graben with an oppositely dipping bounding fault.

3.2.5 Discussion

Several PROBE authors refer to the existence of transfer faults in the EARS rift zones (e.g. Versfelt 1988, Wheeler 1989), but they confine this term to structures which offset segments of major bounding faults of the same polarity. These interpretations label only NE trending structures "transfer faults" which, according
to orthogonal extension models implies NE-SW extension, contrary to the slickenslide data (see Section 3.2.1) on a major bounding fault. I am aware of no other evidence to support the NE-SW direction of extension and consider it untenable. The data and interpretations herein suggest a linked fault system comprising two principal sets and other minor sets. Significant oblique slip occurs on both sets of faults. Oblique slip on major faults also has been documented by onshore mapping of several of the major faults in the area (e.g. Daly et al. 1989, Wheeler 1989). The data also suggest that the NE trending faults are actually the fault segments which should show the highest component of dip slip. The magnitude of dip slip movement will vary depending on the orientation of the fault sets with relation to the regional tectonic transport direction.

Accommodation zones as previously mapped show no consistent trend (Rosendahl et al. 1988, Versfelt & Rosendahl 1989, Scholz et al. 1989), although polarity reversing accommodation zones commonly cross the rift axis at -45° (Figure 3.2). Of seven mapped in Lake Tanganyika, three trend NNE (05°-15°) and four trend NW (295°-330°); all five in Lake Malawi (or three, depending on the significance of one with three apparent trends) trend N to NE (0°-42°) (Versfelt & Rosendahl 1989, their Figure 3). Because of this variability in trend they cannot be considered analogous either geometrically or kinematically to transfer faults, although internal structures within them may be. Morley et al. (1990) provides a detailed description and classification of structures from the Lake Tanganyika rift zone termed transfer zones, a hybrid of transfer faults and the accommodation zones of Rosendahl (1987). The transfer zones vary widely in their orientation and apparent function. However, the internal structure of accommodation zones mapped in this study (Figures 3.7, 3.8, 3.12 and 3.13) seems to be characterized consistently by two primary fault sets.

Within the EARS rift zones, one fault set is aligned subparallel to the tectonic transport direction and the other is at a high angle to it. Each fault pair bounds a block that has a base synrift dip direction which is consistent, within about 5°, in a larger compartment defined by dip domains. Regionally, the average block dips within each domain are broadly consistent, within about 30°, and are believed to correspond to a regional scale tectonic transport or extension direction. The wider spread of dips regionally is most probably due to the effects of local influences of reactivation of non-ideally aligned pre-existing structures. Although both fault sets can be oblique slip, one member of a block bounding pair is commonly mostly dip slip, while the other fault set has a higher component of strike slip. It is proposed that these fault trends are kinematically analogous to the normal and transfer faults described by others (e.g. Gibbs 1984, Lister et al. 1986,
Etheridge et al. (1989), albeit within an obliquely extending system and modified by pre-existing heterogeneities in the brittle upper crust. Certainly, arcuate faults exist (e.g. anastomosing strike slip faults) and they appear to be common in the EARS. However, the accommodation zone studies in the preceding sections suggest that fault geometry is described by two subplanar, nearly rectilinear, fault sets at increasing degrees of resolution, even in an extending system dominated by oblique slip deformation. Accommodation zones then reflect the mismatch between two adjacent half graben fault pair geometries and their location when accommodating a half graben polarity switch is dependent on major transcontinental dislocations as proposed by Versfelt (1988) and Versfelt & Rosendahl (1989). The greater the mismatch the more complex the internal deformation of the accommodation zone.

Versfelt and Rosendahl (1989) have proposed a significant correlation between the location and orientation of polarity reversals and intersecting pre-existing structure. The interpretation herein demonstrates that fault set trends change across polarity switching accommodation zones, which seems to reflect the change in influence of intersecting basement structures. Versfelt's (1988) study of the relationship between pre-existing geology and structure and Cenozoic rifting led to the proposal of a hierarchy of control exerted on the architecture of the Tanganyika and Malawi rift zones by pre-existing structure. Various antecedent anisotropies affect rift zone fault geometries, depending on their scale. In general, rift zones follow mobile belts between cratons. Where steep, continental scale strike slip dislocations intersect these belts, kinks or changes in the orientation of the rift axis occur. The steep continental scale strike slip dislocations (e.g. the Rukwa and Chamaliro Dislocations, Figure 3.2) correlate to polarity reversing accommodation zones. Local smaller-scale dislocations are associated with offsets in border fault systems of the same polarity (Versfelt & Rosendahl 1989).

Versfelt and Rosendahl (1989) also have proposed a model for the early stages of continental rifting, in which rifting propagates from the intersections of the transcontinental dislocations and mobile belts and follows the mobile belts in the sinusoidal fashion proposed by Reynolds (1984) until it connects two dislocations. In support of this hypothesis, the deepest points (Figure 3.2) and therefore probably the most extended areas in Lakes Tanganyika and Malawi lie directly to the NW of the central polarity switches in both lakes. The second deepest point in Lake Malawi lies just north of the northern kink in the rift axis which is also a polarity switch. Sub-basins away from the deep nuclei are certainly shallower and may well have more chaotic structural patterns. As noted by Morley (1988), strain may not be evenly distributed throughout the rift zone, which fits
into the nucleated rift propagation scheme and the observed structural and dip variations. The dip domain analysis (Figure 3.4) suggests there is a concentration of anomalous block dip directions away from the major polarity switches. Where extension is well developed (i.e. the nuclei), a "corner" is well developed by a fault pair and dips are consistent over a broad area. Away from the nuclei, extension has not yet created sufficient structural relief to produce this consistency and dips are controlled by local, perhaps unlinked, faults.

Under a scheme of pre-existing structural control, nonvertical crustal weaknesses whose surfaces traces are aligned approximately orthogonal to the direction of extension would be favored for reactivation as faults with dip slip motion, those trending subparallel, with subvertical dips, would be preferred for strike and oblique slip motion (Etheridge 1986). However, non-ideal alignment does not preclude slip on pre-existing structures (e.g. Scotti et al. 1991). Sub-basins adjacent to the nuclei would have a higher probability of reflecting the precise regional tectonic extension (e.g. Line 815, Figure 3.5) than ones further away where strain is diminished. Also, if this hypothesis is correct, rift structures that appear to be coincident with and reactivated along these transcontinental dislocations, and are most closely aligned with the direction of tectonic transport, will most likely evolve into transfer/transform/lines of flow features in a successful birth of an oceanic spreading ridge.

In other words, the N50°W trending central Lake Tanganyika accommodation zone will probably evolve into a major transform if extension continues along the present direction. The N30°W trending central Lake Malawi accommodation zone is more likely to evolve into a transform than the NE trending north central Lake Tanganyika and north central Lake Malawi polarity switches (Figure 3.2). Accommodation zones with trends at high angles to the rift will most likely be preserved as interbasinal highs in rifted passive margin sequences.

Although there are various opinions about the eventual evolution of the EARS, structures within this setting compare with structures on rifted passive margins (e.g. Bally 1984, Skilbeck & Lennox 1984). There is strong evidence that indicates that the overall regional tectonic regime of the Western Branch of the EARS can be described by NW-SE extension, oblique to the generally N-S rift axes. In all cases, where control is sufficient to delineate blocks, they have been bounded by faults which vary in trend, but in general form a NW and NE trending pair. Segments of the rift that lie along the NW-SE trending dislocations will fit closely to pull-apart, strike slip models. Rift segments that lie along appropriately aligned
mobile belts or are underlain by significant fabric which trend nearly perpendicular to the direction of extension, should be preserved as nearly orthogonal rectilinear systems. The remaining rift segments will comprise a variety of oblique slip geometries depending on the orientation of activated fault pairs. The fault geometries of the Lakes Tanganyika and Malawi rift zones are varied, but can be best described by oblique slip deformation rather than ideal orthogonal extensional deformation, indicating a prevalence of non-ideally oriented weaknesses. Nonetheless, all structural domains or compartments within the rift zone are consistent with regional tectonics if evaluated at a small enough scale and with the appropriate methodology. This study indicates that the oblique slip geometries may, in fact, be the rule rather than the exception during intracontinental extension. If so, dip domain analysis can aid in constraining fault geometries. Also, in the oblique slip case, fault geometries cannot be used to deduce regional kinematics, but block dip analysis, when it yields a regionally consistent result, can constrain the tectonic transport direction.

3.2.6 Conclusions

Analysis of fault and synrift bedding geometries within Lakes Malawi and Tanganyika has led to the conclusion that the deformation in the western branch of the East African rift system is the result of approximately NW-SE extension, significantly oblique and not perpendicular to the elongation direction of the rift zones that underlie the lakes. Both the obliquity of the rift elongation to the tectonic transport direction and some of the internal structural complexity of individual lakes are interpreted to have resulted from the deformation being localized by a network of pre-existing basement shear/fault zones (Versfelt & Rosendahl 1989).

Because the tectonic transport direction is at 20° to 45° to the strike of the lakes and their border faults, the latter must have a significant component of strike slip movement. Two principal fault trends have been identified beneath the lake floor. The dominant fault set is oriented generally NW-SE, is subvertical and is aligned subparallel (within 15°-30°) to the tectonic transport direction. In the studied rift zones, the NW-SE trend is regionally more consistent and through going, dominated by N30°W and N45°W strikes. Although the NW-SE fault set must have a substantial component of dip slip because it is not perfectly aligned with the tectonic transport direction, it acts primarily as accommodation structures between blocks that have extended in the transport direction, in much the same fashion as the transfer faults recognized in orthogonally extended rifts (Gibbs 1984, Etheridge et al. 1989). The other fault set strikes at a high angle to the tectonic
transport direction, albeit with a somewhat larger variation in strike. The latter fault set generally has shallower fault plane dips and greater rotation of the hangingwalls toward them. This fault set is predominantly dip slip and has accomplished most of the extension in the tectonic transport direction.

The variation in strike within both fault sets and in their distribution beneath the lake floors seems to be due principally to reactivation of basement shear/fault zones. Polarity switches are also at least partly controlled by the location of major pre-existing dislocations. Accommodation zones are of two main types:

- those subparallel to the tectonic transport direction comprise clusters of the steep faults bounding blocks that are elongate subparallel to the tectonic transport direction and are akin to transfer fault zones;

- those at high angles to the tectonic transport direction comprise complex, semi-equidimensional tilt blocks bounded by two linked fault sets, one of which is subparallel to the tectonic transport direction.

Zones of compression and non-rectilinear geometries commonly occur within some of these accommodation zones, especially at dip domain transitions between adjacent half graben with highly dissimilar or opposing fault geometries.
3.3 Dip Domain Analysis of the Lake Rukwa Rift Zone—An intracontinental transform?

3.3.1 Introduction

The Lake Rukwa rift zone is a NW trending segment of the EARS that connects the N-S Lake Tanganyika and Lake Malawi rift zones (Figure 3.1). Geological investigations of the Rukwa area began early in this century when Grantham (1932) named the NW striking, SW dipping Lupa fault, which bounds the NE margin of the rift valley and defines its axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend. Geological descriptions for the next few decades were primarily concerned with axial trend.

In the 1980's, increased petroleum industry interest in East Africa led to geophysical surveys over the area, including aeromagnetic, gravity (Peirce & Lipkov 1988) and reflection seismic data acquisition (Morley et al. 1989, Morley et al. 1992, Kilembe 1990, Kilembe & Rosendahl 1992, Rosendahl et al. 1992a). In addition to interpretation of these geophysical data sets, structural analyses have included interpretation of satellite images and outcrop studies (Chorowicz et al. 1982, 1987, Chorowicz 1983) and some recent outcrop work reported by Wheeler (1992). Two industry wells tied to the reflection seismic data indicate that the sedimentary fill in the rift zone includes three depositional sequences, from bottom to top: the Karroo Sequence, the Red Sandstone Group and the Lake Bed Sequence. Traditionally, these stratigraphic packages have been considered, Permo-Triassic, Jurassic/Cretaceous and Tertiary, respectively (e.g. Dypvik et al. 1990 and references therein). Recent biostratigraphic analysis of cuttings from two wells suggests that the Tertiary package is much thicker and probably includes the Red Bed Sequence (Wescott et al. 1991).

In this case study, selected reflection seismic data are re-evaluated and previously unrecognized structural trends are presented. The re-evaluation suggests that previous interpretations of the extent of the Karroo and Red Bed Sequences needs modification. This study attempts to track the kinematic history of the Rukwa rift zone using dip domain analysis, with particular emphasis on Cenozoic deformation (Figures 3.16 and 3.17). The analyses demonstrate the modifying effect of pre-existing structures on rift geometry over time, a key element frequently overlooked in traditional structural interpretations. The changing morphology of surfaces through time, which can be seen in the dip...
analysis of selected horizons, is crucial to understanding the geological and deformational histories of basins.

In conjunction with the analysis of the Lake Tanganyika and Lake Malawi rift zones presented in Section 3.2, the dip data provide a regionally consistent Cenozoic kinematic history in East Africa (see also Scott et al. 1990, 1992). Changes in dip domain boundaries through time support a two-stage, variably directed extensional history. The results of the current study support the intracontinental transform role of the Rukwa lineament during the Tertiary (Chorowicz et al. 1982, 1987, Chorowicz 1983, Daly et al. 1989, Asfaw 1992). Some implications of the role of intracontinental transforms to the kinematics of large scale continental deformation are explored.

**Figure 3.16** Schematic kinematic history of the Rukwa rift zone. A) Basement geology consists of the Archean Tanzanian Craton with adjacent "sutured" Proterozoic (Ubendian) terranes. B) A possible structural interpretation explaining the distribution of the Karroo Sequence. See Figure 3.19 for clarification of isochrons. Two large arrows indicate the NNW trend of the Karroo Trough. Strike slip arrows indicate "stepping" of deeps. C) Red Bed Sequence deformation deduced from dip analysis and examination of growth of the sequence in differently oriented profiles. Small arrows indicate tectonic transport direction of hangingwalls, which is accomplished by oblique slip on the Lupa Fault and adjoining splays (Figure 3.18). D) Late Tertiary to Present structural setting of the Rukwa Rift zone. Small arrows indicate tectonic transport direction of hangingwalls as in C. Larger onshore arrows are measured Neogene slip vectors on exposed faults (Chorowicz et al. 1987, Chorowicz 1989, Daly et al. 1989). Onshore structures are omitted in B and C for clarity, although most of the structures are inherited terrane boundaries (A) and have experienced several reactivations. The strike slip fault in the NW corner of D, labelled KIH, continues to the NW and connects with the fault bounding the west side of the Kavala Island horst in Lake Tanganyika (Figure 3.14).

**Figure 3.17** (Next page) Plotted dip directions computed from apparent dips measured at reflection seismic profile intersections. T1 - T3 = progressively younging Tertiary reflection horizons as shown in Figure 3.20, 3.22, 3.23 and 3.24. D1-D4 = dip domains. Hashed bands indicate approximate location and trend of dip domain boundaries. Location of seismic profiles shown in Figures 3.20 and 3.22 are labelled TVZ-# and TXZ-#, respectively. A-D = location of seismic profiles shown in Figure 3.23. "Strike" profile (Line 23) shown in Figure 3.24 is located in the T2 map. The long, linear NE lakeshore corresponds approximately with the bounding Lupa fault (Figure 3.18).

Strike/dip symbols with arrows in T1 map show sense of deviation from predicted dip direction of hangingwall if purely normal motion is assumed for the bounding Lupa fault. Larger black arrows in T2 and T3 maps show general sense of dip within domains. The deviation from NE trending dips is apparent in the bimodal rose diagram plot of computed T1 and T2 dips. A greater proportion of dips have become oriented NW-SE at T3 levels. S1 and S2 = 300° trends indicated by arrows in Figure 3.22. E and F = NNE trends of structures in profiles A-B and C-D, respectively, shown in Figure 3.23.
3.3.2 Structural Framework

Several workers have studied the Lake Rukwa rift zone and its tectonic relationship with the N-S Lake Tanganyika and Lake Malawi rift zones (e.g. Delvaux 1992 and references therein). A few of the investigations have been concerned primarily with how these Cenozoic rifts relate to reactivation of major crustal/lithospheric structures (e.g. Daly et al. 1989, Chorowicz et al. 1987, 1982, 1979). It has been hypothesized that the Rukwa lineament, including the segment of the rift zone studied herein, is acting as an intracontinental transform (Chorowicz 1989, Chorowicz et al. 1987, 1979).

That Cenozoic rifting in East Africa has followed pre-existing lines of weakness is not a new idea, and the nucleation sites in the Lakes Tanganyika and Malawi rift zones appear to be associated with pre-existing structures (e.g. Versfelt & Rosendahl 1989 and references therein). The NW striking Lupa fault (Figure 3.18) is a normal to oblique slip extensional fault, coincident with the NE shore of the lake that appears to bound the entire rift zone. The Rukwa rift zone is underlain by rocks of the early Proterozoic Ubendian Belt (Figure 3.16). The Ubendian belt comprises a number of fault- or shear zone-bounded blocks or terranes (Daly et al. 1989). The Lupa fault parallels the ancestral (Ubendian) Rukwa trend defined by the shear zone boundaries between the terranes. The Lupa fault coincides precisely with the northeastern boundary of the Mbozi terrane. However, foliation in the Mbozi terrane dips to the NE, so the SW dip of the Lupa fault is not precisely coincident with pre-existing fabric (Kilembe 1990, Versfelt 1988). Slip vectors on Neogene NW-SE fault planes in the Rukwa area are commonly subhorizontal, indicating dextral strike slip displacements, whereas N-S and NE-SW trending faults in the adjacent Tanganyika and Malawi rift zones display oblique to dip slip displacement (Daly et al. 1989, Chorowicz & Sorlien 1992, Chorowicz et al. 1987, Ring et al. 1992, Wheeler & Karson 1992, Wheeler 1989).

A previous interpretation of the basement structure (Kilembe 1990, Kilembe & Rosendahl 1992) using a reflection seismic data grid of approximately 4 km spacing has been slightly modified here (Figure 3.18). For example, minor differences in Kilembe & Rosendahl's (1992) interpretation and an interpretation of the same data set by Morley et al. (1992) have been reconciled. Also, in some places, contours contradicted sense-of-throw on faults in the earlier interpretations. Since only half the dataset were available for this study, these modifications were made only where data were available to correct fault/contour relationships. Although the Lupa fault is shown as a linear, continuous feature in this interpretation, it's
exact position is not well constrained by the seismic data, which often does not image the upper section of the fault or the footwall. Local lateral offsets in the fault at basement levels are suggested by the variation in position and fault geometry at depth on adjacent seismic profiles (Rosendahl et al. 1992a, their Fig. 3). There is also some indication that it changes from an overall 305° trend west of 32°30'E to a more northwesterly trend of 315° to the east, where the linear lakeshore also changes trend (Morley et al. 1992, Daly et al. 1989).

The intra-rift deformation appears to be dominated by NNW (330°-350°) faults. Many of the interpreted faults are antithetic to the bounding Lupa fault. Basement structures, as mapped, can be grouped into three zones approximately separated by the 32°00'E and 32°30'E meridians. East of 32°30'E, intra-basinal faulting breaks the basement into roughly rhombohedral blocks. Sides of the rhombs are oriented subparallel to the Lupa fault linked to ~335° trending structures. Basement reaches depths of 5.4 seconds or about 7 km, using reasonable sediment velocities and gravity modelling (Peirce & Lipkov 1988). Several well constrained normal offset WNW-ESE to nearly E-W faults are mapped in the southern part of this area. Structural contour orientation in the eastern section of the rift varies from subparallel or oblique to trends at high angles to the major normal offset structures.

Mapped structures in the area between 32°00'E and 32°30'E are variable and complex (Figure 3.18). Apparently, large normal offset faults mapped in the northeastern sector of the central section trend NNW at their eastern extent, but curve around to trend WNW or almost E-W farther to the west, resulting in a most improbable extensional fault configuration, wherein the fault traces are convex toward the hangingwall. The complexity of this group of structures contrasts with the simpler rhombic pattern in the western portion of the central section of the rift zone. Across a "no data zone" (NDZ) in the south of the central section, structures are mapped consistently NNW-NW. Although no cross-cutting structures are interpreted, structural contours in this section of the rift zone are

![Figure 3.18](Image) Depth-to-acoustic basement structure map. Note high-angle of structural contours to mapped faults. Lupa border fault position is not well constrained by the available seismic data, but outcrop data indicates that it is relatively linear west of 32°30' E, where it bends southward. Compiled and modified from Kilembe (1990) and Kilembe & Rosendahl (1992).

![Figure 3.19](Image) (Next page) Karroo Sequence isochron (thickness) map. Prominent trends are indicated. Note that the apparent depocenter axis is not parallel to the bounding Lupa fault. Compiled from Kilembe (1990) and Kilembe & Rosendahl (1992).
Depth-to-Acoustic Basement Structure in LAKE RUKWA
KARROO Sequence Isochrons in LAKE RUKWA
generally at high angles to the mapped structures. Depth-to-basement is shallower than to the east, reaching depths of 4.2 seconds or about 5 km.

In the westernmost portion of the rift, acoustic basement is generally shallower. Intra-rift faulting trends are predominantly NNW, but are distinctly more NW trending than structures in the northern half of the area to the west, appearing to align with the structures in the southern part of the central section of the rift. Definite correlation of the western structures to the south-central faults is not possible due to the lack of intervening data. Although there appear to be no structures at high angles to the rift axis in the basement structure map (Figure 3.18), basement contours generally trend across the axis of the rift, indicating that hangingwalls do not dip toward the major normal faults. This is a common feature of depth-to-horizon structure maps generated by traditional seismic reflection data interpretation procedures (e.g. the Eyre Sub-basin, Chapter 4; Figures 4.14 and 4.15). The inconsistent relationship between mapped normal faults and hangingwall dips is common in oblique extensional systems (e.g. the Lake Tanganyika and Lake Malawi rift zones, Section 3.2).

The Karroo Sequence isochron map in Figure 3.19 is reproduced from Kilembe & Rosendahl (1992a; their Figure 7). The isochrons appear to define a trough that trends oblique to, rather than parallel to, the Lupa Fault. Although the Karroo Sequence thickens dramatically into the Lupa Fault (Figure 3.20), Karroo Sequence thickness contours do not trend rift-axis-parallel, as one might expect, but trend at about 330° or 15°-20° from the trend of the Lupa fault. The Karroo Sequence thicknesses have a similar tripartite configuration as defined by changes in structural style in the depth-to-acoustic basement map (Figure 3.18). According to this interpretation of the seismic stratigraphy, thick accumulations of Karroo sediments occur only in the SE section, with moderate accumulations present in the central section and a very thin Karroo Sequence in the northwest section. The thickness of the Karroo Sequence is two to four times thicker in the southern depocenters than the preserved Karroo section in the north. The Red Bed sequence is also thickest in these locations (Kilembe 1990, Kilembe & Rosendahl 1992). The central section, between 32°00'E and 32°30'E, is characterized by small pockets of moderately thick Karroo near the Lupa fault to very thin or no Karroo section in the south. The northwest portion of the rift is marked by a generally very thin Karroo section. A consistent isochron trend of 330° is evident along the entire rift zone, but "complementary" isochron trends are N-S in the southeast, significantly different from the NNE complementary trends in the northwest.
Overlaying the depth-to-basement structural contours on the Karroo Sequence (Figure 3.21) demonstrates a relatively poor correlation between accumulation of Paleozoic section and mapped basement structure depth. The correlation between isochrons and basement structure is the most reasonable in the SE where the thickest accumulations of Karroo section occur; that is, basement structural contours are generally subparallel to faults which are subparallel to isochron trends. However, the thickest depocenters have unusual relationships with the mapped structure. Undulations in the NE boundary of the southeastern depocenter are not explained by the linear fault passing through it. Although the thickest portion of the other major depocenter overlies the deepest basement, mapped structures in this area cannot explain the rhombic outline of this accumulation. Also, in this area basement dips as defined by contours appear to be subparallel to the Lupa fault rather than toward it as one might expect if motion on this structure was purely normal slip.

The peculiarity of basement dips oriented subparallel to mapped structures is more common in the central and NW sections of the rift zone. Structural contours rather than structures align with Karroo isochrons and are commonly at high angles to the mapped structures. The Karroo Sequence distribution may be an artifact of preservation rather than a true indication of depositional configuration. However, although the Karroo/Red Bed Sequence boundary is recognized as an erosional unconformity (e.g. Figure 3.20), areas of truncated reflections are restricted to the SE where the correlation between basement structure and Karroo distribution is the best. In an attempt to reconcile this issue a re-examination of approximately half of the original reflection seismic data grid (i.e. $\approx 8$ km spacing) was undertaken to re-evaluate the seismic stratigraphy. Although there are no wells to definitively date sediments in the NW section of the rift zone, on the basis of a re-evaluation of the seismic stratigraphy manifest in the seismic reflection profiles, it is proposed that acoustic basement may be overlain directly by Tertiary sediments in the far NW or that only a very thin (one or two reflection cycles) Paleozoic section exists here (Figures 3.22 and 3.23). If this is true, the change of

**Figure 3.20** Two "dip" profiles from the southern end of the rift zone. Seismic stratigraphy is indicated, including the Permo-Triassic (P-TR) Karroo Sequence, the ?Lower Tertiary (LT) Red Bed Sequence and the Upper Tertiary (T) Lake Bed Sequence. Note the steepening of the Lupa fault up section. Dip analyses of the T1-T3 reflection horizons is presented in Figure 3.17. For location see Figure 3.17, T1.

**Figure 3.21** (Next page) Overlay of basement structure and Karroo isochron maps (Figures 3.18 and 3.19). Note the relatively poor correlation between the distribution of the Karroo Sequence and basement structures that worsens to the NW. Compiled from Kilembe (1990) and Kilembe & Rosendahl (1992).
Acoustic Basement Structure Superimposed on KARROO Sequence Isochrons in LAKE RUKWA
structural style and isochron trends between the SE and the NW may be explicable in terms of variable timing of deformation. On the basis of the isochron trends in Figure 3.19, Karroo-age intra-rift faults must have included NNW faults. In Morley et al.'s (1992) interpretation of the Rukwa dataset, which attempted to distinguish between Karroo-age intra-rift normal faults and Cenozoic faulting, Karroo normal faults strike 330°, supporting the interpretation that NNW faults are required by the isochron trends presented in Figure 3.19.

3.3.3 Dip Domain Analysis

It is common practice within the petroleum industry to produce depth/time-to-horizon maps from reflection seismic data (Fig 3.18). These maps are, in essence, topographic maps of the desired surface and yield information about the dip of that surface. However, because of the subjective nature of contouring and the uncertainties introduced by correlating structures between relatively widely-spaced and arbitrarily oriented profiles, they often can be misleading. As outlined in Chapter 2, seismic profile intersections where a true dip can be computed from the two apparent dips of a tied reflection horizon on each profile, provide the only truly constrained information of the dip of the surface of interest. Measurements of apparent dip have been made and the true dip computed at profile intersections to produce the dip maps in Figure 3.17. The apparent dips were measured at 111 intersections of the seismic profiles used to re-evaluate stratigraphic surfaces.

As discussed in Chapters 1 and 2, the true dip of a rotated or tilted block is expected to be opposite to the dips of the normal faults in ideal extensional deformation. It also has been argued that in non-ideal extension or transtension settings, a statistically consistent regional dip direction indicates the tectonic transport or extension direction. The case studies of the Lakes Tanganyika and Malawi rift zones demonstrate a consistent NW-SE to WNW-ESE intra-rift basement dip for these rift zones (Section 3.2). The measured dip azimuth mode is oblique to the generally N-S strike of these two rift zones, but is consistent over a 1000 km extent along the rift system. Using dip analyses, the following question is addressed: Does the connecting NW-SE elongated Lake Rukwa rift zone show any evidence of the inferred NW-SE regional tectonic transport direction in Lakes Tanganyika and Malawi?

Dips azimuths computed for acoustic basement, the base of the Red Bed Sequence and the base of the Lake Bed Sequence (Figure 3.17, T1) surfaces within the Rukwa rift zone are basically indistinguishable and trend mainly NE as
predicted by normal movement on the Lupa fault. However, there are significant differences between dip geometries of T1, T2, and T3 reflections (Figure 3.17), implying different kinematics at different times throughout the Tertiary.

The T1 and T2 horizons (Figure 3.20) were chosen for their distinctive high amplitude character and regional extent and coincide with horizons A and B of Morley et al. (1992) who report that wells intersecting these horizons document that they record a change from lake beds above, to interbedded red brown sandstones and mudstones between, to increased red mudstones below. For convenience, the T1 horizon is referred to as the base of the Lake Bed Sequence. None of the base Lake Bed Sequence (T1) dips trend exactly toward (i.e. perpendicular) the Lupa fault, as would be predicted if the Lupa fault was a purely extensional normal fault controlling the hangingwall dip. Instead, the dips can be grouped into domains based on whether the majority of dips diverge from perpendicular to the Lupa Fault to the north or to the south, producing a bimodal distribution of dips. Using this criterion the rift zone can be divided into four domains (Figure 3.17).

The great thicknesses of sediment adjacent to the Lupa fault, and its listric nature (Figure 3.20), indicate that it has experienced a normal component of movement at several stages in its history (Kilembe & Rosendahl 1992, Rosendahl et al. 1992a). Morley et al. (1992) measured up to 10 km of extension at the base of the Red Bed Sequence across the Lupa fault on one seismic profile in the far SE of the rift zone, but reports more common values of 1.7 - 2.0 km in the NW, increasing to 3 - 4 km in the SE. However, since dips do not trend perpendicular to the Lupa fault on the base Lake Bed Sequence (T1) surface, some oblique to strike slip motion must have occurred on it by that time. The abrupt change in dip of the Lupa fault at the T1 horizon, from a shallow dip below to near vertical above,

Figure 3.22 Portions of "dip" seismic profiles from northern Lake Rukwa. High amplitude reflections labelled T1 - T3 as in Figures 3.17 and 3.20. S1 and S2 structures indicated by arrows are mapped in Figure 3.17, T3. Location shown in Figure 3.17, T1.

Figure 3.23 (Next page) Portions of "strike" seismic profiles from northern Lake Rukwa. E and F = flat-lying zones whose trends are mapped in Figure 3.17, T3 and coincide with minor deeps in Kilembe & Rosendahl's (1992) Karroo isochron map (Figure 3.19). Re-evaluation of the seismic data suggests that little to no Karroo exists here and that the "deeps" (E and F) shown in Figure 3.19 are post-Karroo features. Compare the higher magnitudes of the dips on these "strike" profiles, to the left of the arrows, with dips imaged on the left side of Figure 3.22 "dip" profiles. Location shown in Figure 3.17, T1.
Line TXZ-16

047°

Line TXZ-15

10 km
Line TXZ-49(A)

Line TXZ-20(B)

Line TXZ-50(C)

Line TXZ-21(D)
Chapter 3  East African Rift System

supports the hypothesis that little extension has been accommodated on the Lupa fault in post Red Bed Sequence times (Figure 3.20).

The trend of the two southern dip domain boundaries in Figure 3.17, T1, is not well constrained. The northern boundary has a reasonably well constrained orientation of 280°-290°, or 25°-35° from (oblique to) the NW trend of the Lupa fault. A distinctive set of apparently large-offset faults have this orientation in the SE of the rift zone (Figure 3.18) and correspond with the southernmost domain boundary in Figure 3.17, T1. If these structures are extended to the west, they align with a large WNW-ESE offset in the onshore, southwest bounding fault of the rift zone (Figure 3.16, D) (Peirce & Lipkov 1988). Although there are apparently no structures with a WNW-ESE orientation mapped in the northern portion of the rift zone by Kilembe & Rosendahl (1992), a previously mentioned bending of major NNW structures to this trend coincides with the middle domain boundary. Discontinuity of minor internal structures occurs at the northern domain boundary. Additionally, structural contours at high angles to the mapped faults (Kilembe 1990) suggest that unidentified structures may be contributing to the dip of the acoustic basement. Finally, 280°-300° is a prominent trend of intra-rift block dips in the Lakes Tanganyika and Malawi rift zones (Section 3.2.2, Scott et al. 1992). It is suggested that these domain boundaries represent transfer faults or accommodation zones of the Red Bed Sequence extensional phase. The structures effectively decoupled adjacent blocks allowing them to rotate semi-independently and produce the variation of dips across the boundaries. The apparent inconsistency of dips measured at Tertiary levels (T1-T3) and basement isochrons can be explained if the younger faults sole out high in the section, as is indicated on many profiles (e.g. Figure 3.20).

Dips computed on the T2 horizon in the NW portion of the rift zone, are almost exclusively NW and SE in contrast to the ESE and ENE T1 dip trends (Figure 3.17). In the far north, T2 dips adjacent to the bounding (Lupa) fault are still to the NE (Figure 3.17). Farther away from the bounding structure in the northern domain, the dips become subparallel to the strike of the Lupa fault, as would be predicted in a strike slip terrain or in an area where normal faults trend NE-SW. Several NE-SW normal faults that dip to the SE are mapped onshore to the NW of the lake (Chorowicz et al. 1987, Daly et al. 1989) (Figure 3.16, D). The dip variation seen between Domains 1 and 2 at T1 appears more pronounced at the T2 level, where the dip direction reverses across the boundary. An ENE-WSW boundary separates contrasting dips in the northern zone from the southern two-thirds of the rift zone and corresponds to a relatively linear shoreline. This boundary is not evident in dips measured on the younger T3 horizon. T2 dips south of the
boundary largely mimic those shown for the T1 horizon; indeed, the bimodal
distribution of dips is even more pronounced at this level. Although dips at T1
and T2 levels are quite large for load induced subsidence the vertical fault geometry
and increasingly up-section flattening of Tertiary reflections suggest that very little
extension has taken place on the Lupa fault during the deposition of the Lake Bed
Sequence (Figure 3.20). The general northeasterly trend of dips, especially in the
southern portion of the rift zone and adjacent to the Lupa fault in the north, may
be due to remnant rift topography created by previous extension.

It has been argued previously that the Karroo and Red Bed sequences are
relatively thin to absent in the NW as indicated in two seismic lines from this area
(Figure 3.22). Both profiles can be divided into three segments by the structures
indicated at the arrows. An elevated acoustic basement on the left drops to a broad,
largely unbroken block between the S1 and S2 arrows, which passes into a chaotic,
complexly deformed zone on the right. Correlation of the structures indicated by
the arrows between the profiles yields strikes of about 300° (Figure 3.17, T3).
Although there is normal offset on both structures, dips on acoustic basement are
away from these steep (vertical?) faults to the right, arguing that these are not
normal faults. Indeed, actual dips on both T2 and T3 horizons in this part of the rift
zone are to the SE, suggesting that NE trending normal faults cause the tilting in
this area. Previous mapping of the reflection seismic data (Kilembe 1990, Kilembe
& Rosendahl 1992, Morley et al. 1992) recognized no NE striking structures,
although structural contours on acoustic basement of this orientation are evident
(Figure 3.18) and correlate with isochron trends for the sequence directly overlying
basement (Figure 3.19).

The S2 structure separates northeasterly dips from the southeasterly dips at
both T2 and T3 levels. The dip relationship suggests that this structure is a strike
slip fault and that dips to the northeast of it are still reflecting substantial control by
subsidence or infill adjacent to the pre-existing Lupa fault, whereas dips to the
southwest of it are reflecting NW-SE extension. The complexity of structures on
the right end of the profiles relative to the simpler deformation on the left (Figure
3.22) suggests that the pre-existing Lupa Fault has acted as the boundary to Cenozoic
deformation in the Rukwa rift zone (Rosendahl et al. 1992a).

Figure 3.24 "Strike" profile Line 23 which parallels the Lupa fault in the central
and southern portions of the Rukwa Rift zone. A broad antiform and synform are
evident in these data. N = nappe-like imbrications of T1 and T2 horizons. U =
location of truncated upper Tertiary reflections that indicate the crest of the
antiform has been eroded. F1-F3 are major faults and asterisk (*) indicates the
hangingwall cutoff of the F3 fault discussed in text. For location see Figure 3.17, T2.
Rukwa Line 23

NW

10km

SE
Dips computed at the T3 horizon are consistently NW-SE (Figure 3.17, rose diagram). NW-SE dips extend over most of the rift zone, although dips in the extreme southeast are still to the northeast. The three T1 dip domain boundaries are evident at the T3 level, separating zones of opposite dip directions. The few "anomalous", NE dips in the northern dip domain coincide with the S1 and S2 structures (Figure 3.17, T3). It is possible that the T3 dips in the NW are primary depositional dips (e.g. if sedimentation throughout the rift zone was due exclusively to axial deltaic deposition). However, the NW and SE dips extend through much of the rift zone, crossing significant structural boundaries, which most likely were sedimentation barriers (e.g. the D2/D3 boundary). The polarity reversal across the northern domain (D1/D2) boundary also argues against depositional dip being the controlling process.

There is ample evidence of structurally controlled steeply dipping reflections on rift-axis-parallel, or "strike" lines, in the Rukwa rift zone (Figures 3.23 and 3.24). It is therefore not surprising that computed dips at profile intersections usually do not align perpendicular to the Lupa bounding fault. The dipping sections often merge into adjacent, nearly flat-lying or shallowly and oppositely dipping sections without obvious normal offset faults (e.g. Figure 3.23a). The change in dip is commonly across a zone of discontinuous or chaotic reflections that resembles the fault-folds described in Hills (1963) or the extensional fault-bend folds of Groshong (1989). Fault surfaces in these models are planar with distinct bends (e.g. the bounding fault in Figure 3.20, Line TVZ 28), rather than curvilinear with a continuously decaying dip (e.g. listric or exponential). As the hangingwall moves over the fault segments, the bends in the fault create axial surfaces within it. The axial surfaces separate distinct dip domains within the hangingwall, as do axial surfaces in fault-bend folds in compressional terranes.

Correlation of the changes in dip observed in the strike lines in Figure 3.23 yields NE trending structures offset from each other across the NW trending structures mapped on dip lines in a right lateral sense (Figure 3.17, T3, E and F). This sense of motion is consistent with that described for the adjacent Livingstone strike slip basin described by Wheeler (1989) and Neogene slip vectors measured onshore in the Rukwa area (Figure 3.16, D), which suggests that these domain boundaries are strike slip structures.

The other cause of significant rift-axis parallel dips is demonstrated in Figure 3.24. The interpretation of this profile is not constrained by well information and is not unique, but the interpretation of a major synform and antiform cannot be doubted. How these folds are faulted is less clear. For instance,
the large fault on the northwest (F1, Figure 3.24) might be interpreted as a normal fault at base Red Bed Sequence, rather than a reverse fault as shown. However, the T1 and T2 reflections are convincingly offset in a reverse sense along both arms of this fault. The converse appears to be true on the F2 fault. That is, the base Red Bed Sequence shows reverse offset, while T1 and T2 appear to be offset in a normal sense. An apparent change in offset sense is a common feature of strike slip faults. A series of imbricated T2 reflections at the NW end of the profile support a component of compression (N, Figure 3.24). T3 reflections and those above it are truncated in an angular unconformity (U, Figure 3.24) indicating that the crest of the antiform has been eroded. The F3 fault is best mapped as a NNE splay from the Lupa fault that terminates at the southernmost dip domain boundary (Figure 3.16, D and 3.17, T1). The location of the hangingwall cutoff of the splay corresponds exactly with the deepest and similarly aligned trough in the depth-to-basement map (Figure 3.18). The geometry of the the master and splay faults, combined with the southward deviation of dips (from perpendicular to the Lupa Fault) in the D3 domain (Figure 3.17) argue for a tectonic transport direction out of the corner formed by the fault segments in a manner as described in Chapter 2 (Figure 2.3).

The profile in Figure 3.24 is parallel to the strike of the Lupa fault and within 7 km of its shoreline outcrop. Morley et al. (1992; their Figure 4) show the adjacent parallel profile 5 km lakeward. The adjacent profile shows a broad arch without appreciable offset of horizons, but the normally coherent and continuous sequence boundaries and T1-T3 reflections are broken and disrupted over the crest of the arch. The imbrication of the T1 and T2 horizons and the tilted, relatively undisturbed reflections at the SE end of the profile are similar to those seen in Line 23 (Figure 3.24). Correlation of the axes of the disrupted zone on the adjacent profile and the antiform in Figure 3.24 coincides with the domain boundary between D2 and D3 (Figure 3.17). These data support a broad wrench geometry for the 280° domain boundaries.

The late Tertiary deformation evident in seismic profiles suggests that normal offsets at the T1 level, where it is separated by a thick sequence from acoustic basement, frequently do not offset basement substantially and appear to sole out high in the section (e.g. Figure 3.20). This observation was also made by Morley et al. (1992). The Tertiary section of the fault is often much steeper than the pre-existing structure over which it is positioned, sometimes with no convincing evidence of a connection. In contrast, where T1 is near acoustic basement, many small offset structures are the rule at basement levels, normally accompanied by a broader zone of disrupted reflections above the T3 horizon. (Figures 3.21, 3.22 and 3.23). A myriad of complex flower structures (Harding 1985) on both strike and dip
profiles attest to the transcurrent nature of Cenozoic deformation. In sum, the above observations argue for a significant component of wrenching and transpression during Tertiary deformation and extension accommodated by oblique slip fault pairs during Red Bed Sequence times.

3.3.4 Discussion

Several previously unrecognized fault trends are revealed by the dip domain analysis of the Rukwa rift zone and subsequently supported by subtle features in the reflection seismic data. In particular, the 280° and 300° domain boundaries and NE trending structures were recognized by dip changes across them. Several workers have suggested a "pull-apart" basin formation mechanism for the Rukwa rift zone (e.g. Kilembe & Rosendahl, 1992, Rosendahl et al. 1992a), similar in style to that presented by Aydin & Nur (1982). Although there is convincing evidence for strike slip deformation in the data, the lack of major normal offset of Tertiary horizons in many areas, especially on NE trending structures, argues against this model. Close inspection of Figure 3.18 reveals that the relationship of Karroo Sequence thicknesses to sense of throw on structures is reversed locally and could not be modified by re-evaluation of the data. That is, the thicker accumulation is associated with the footwall of a normal basement fault. This relationship is common in areas of inversion and suggests that there may have been periods of compression or at least reverse motion on some of the basement faults.

The Rukwa rift zone has been the locus for intracontinental deformation throughout its history (Figure 3.16). Four main deformations can be deciphered from the presently available data and are discussed below. They are:

• Proterozoic "suturing" of terranes bounded by NW-SE, steep shear zones;

• NE-SW extension (or possibly NNE-SSW strike slip) in Permo-Triassic times resulting in deposition of the Karroo Sequence in rhomb shaped deeps bounded by the Lupa fault and NNE splay faults along a NNW-SSE trough;

• WNW (280°) extension in the ?lower Tertiary reactivating both the NNE striking splay faults and the Lupa fault by oblique slip and initiating the 280° dip domain boundaries;

• Extension has become NW by T3 time (?Pliocene), the NNE splays are accommodating most of the extension with primarily strike slip reactivation of the Lupa fault. The 280° domain boundaries are wrenched to create broad antiforms and synforms. NW oriented strike slip structures are initiated.
The accretion of Proterozoic (Ubendian) terranes against the Archean Tanzanian craton established deep-rooted NW-SE shear zones and a distinct heterogeneity in the crust (Figure 3.16, A). The basin lies wholly within the Mbozi terrane (Daly et al. 1989). Since Karroo sediment accumulations are not aligned to the strike of the Lupa fault, normal reactivation of the terrane boundary shear zone to create the Lupa Fault may have preceded the Karroo trough-forming deformation. Subsequent transcurrent movement during Karroo deformation, created a trough whose axis is oblique to the Lupa fault by ~15°, marked by right-stepping, en echelon rhomb-shaped deeps (Figures 3.16, B and 3.19). If recent biostratigraphic analysis is correct (Wescott et al. 1991), Karroo deformation was then followed by a long period of quiescence through the Mesozoic.

Multiple-phase deformation in the Tertiary is hypothesized on the basis of the dip analysis of the Rukwa rift zone (Figures 3.16 and 3.17). An early (Red Bed Sequence), nearly E-W (280°) extensional phase would have been appropriately oriented to reactivate the NW Lupa fault, depending on its dip, with a fairly high component of dip slip (Figure 3.16, C). The extension was also accommodated on NNE splay faults of the Lupa Fault, which may have reactivated Karroo "transfer" faults. Comparable growth of the Red Bed Sequence into faults on both strike and dip lines (cf. Figures 3.20 and 3.24) supports tectonic transport of the blocks out of a corner made by the Lupa fault and adjoining splay faults, similar to that described in Chapter 2, along some direction between the profile strikes. Analog experimental and theoretical modelling of deformation of hangingwalls extending from corners of fault pairs indicate that dips over most of the hangingwall should align with the tectonic transport direction (Braun et al. in press). The dips in Domain 1 and adjacent to the Lupa fault in the northwest corner of Domain 3 at T1 times support this sense of slip (Figure 3.17). The 280° orientation of the T1 domain boundaries acting as transfer zones during this early Tertiary deformation is consistent with oblique slip reactivation of the Lupa fault paired with NNE trending oblique slip faults (e.g. F3, Figure 3.24). Etheridge (1986) has argued that it is relatively easier to initiate strike slip rather than normal reactivation of steeply dipping zones of weakness and this seems to be supported by the repeated oblique slip reactivation of the Lupa fault during later deformation events.

It may be that high sedimentation causing a significant load adjacent to the Lupa fault has enhanced the dip slip component on it during deposition of the Red Bed Sequence. The long period of quiescence through the Mesozoic provides ample time for the filling in of the Karroo trough, given that adjacent highlands were present to erode, but pre-existing dip on the top of the Karroo Sequence may
be the cause of large NE dips on the base and, smaller but still significant, NE dips on the top of the Red Bed Sequence. The 280° domain boundaries may have nucleated along "steps" in the Karroo structural trough, as is suggested by comparing the distribution of the Karroo deeps and the location of the boundaries (Figure 3.16). A 280° tectonic transport direction at this time would also be consistent with the formation of generally N-S trending rift zone segments such as the Lake Tanganyika and Lake Malawi rift zones.

By T3 time, extension was oriented NW-SE so that primarily strike slip reactivation of the Lupa Fault occurred. Overall, dips at the T3 level are consistent with regional NW-SE extension in the late Cenozoic. This is supported by field measurements of slickenslides which are subhorizontal and indicate dextral, strike slip Neogene and later displacement on NW trending faults in the Rukwa rift zone (Figure 3.16, D). At this time, the S1 and S2 dip domain boundaries (Figure 3.17) were initiated as strike slip faults. The 300° structures (S1 and S2) are aligned with the tectonic transport direction determined from dip analysis for the Lake Tanganyika and Lake Malawi rift zones (Section 3.2, see also Scott et al. 1992) and also align with the southwestern continuation of the western bounding fault of the Kavala Island Ridge horst in Lake Tanganyika (Figures 3.14 and 3.16, D). Combined with the extension accommodated by NE-SW normal to oblique slip onshore faults to the NW of the lake (Chorowicz et al. 1987, Daly et al. 1989), the many small throw faults seen at Tertiary levels in the Lake Rukwa rift zone are consistent with the low amount of extension computed for the Lake Tanganyika and Lake Malawi rift zones (Specht 1987, Specht & Rosendahl 1989, Morley 1988, et al. 1992).

Tracking the change in dips of seismic reflection data horizons in the Rukwa rift zone documents the changing influence of pre-existing structure on subsequent deformational phases. Subtle changes in dips have led to the recognition of structural domains that divide the rift and allow for changes in structural style along rift-axis strike at the acoustic basement level. The domain boundaries may not coincide with particular structures, but rather define transfer or accommodation zones with complex deformation patterns, but whose trends are consistent with regional kinematics as long as oblique slip on the adjoining faults is considered.

The change in domain boundaries between T1 and T3 times may be evidence within the western branch of the EARS of a change in extension direction from an E-W to NW-SE direction. A two phase kinematic scenario has been hypothesized by Strecker and Bosworth (1991) and the change from one phase to the other dated by Pollitz (1991) at between 4 and 8 my for the eastern branch of the
4.0 SOME RIFTED PASSIVE MARGINS OF AUSTRALIA

There is something fascinating about science. One gets such wholesale returns of conjecture out of such a trifling of fact.
Mark Twain

4.1 Introduction

The results from the EARS (Chapter 3) demonstrate that intracontinental areas (sensu stricto) undergo repeated activation that is localized by pre-existing structures and crustal heterogeneities. These heterogeneities must extend through the entire lithosphere if they are to be part of successful continental fragmentation. The case studies in Chapter 3 suggest that inherited structure affects the patterns of intracontinental rifting. Which structures, if any, are active throughout the fragmentation process? Although some of the implications of the results from Chapter 3 have been projected to the problem of continental breakup, there is still considerable debate about whether that system will evolve to successful spreading and the creation of a new oceanic basin. In order to test some of the hypotheses put forward in Chapter 3, we need to identify which structural patterns from the intracontinental setting are preserved on passive margins. Once identified, how do these structural patterns fit into inferences about pre-breakup and breakup tectonic processes?

In order to assess the answers to the above questions, several Australian passive margins have been analyzed with the aim of constraining the kinematics and fault geometry of pre-breakup continental deformation and its relationship to continental fragmentation. Evaluating the relationship between break-up deformation and pre-existing heterogeneities is also emphasized. Structural interpretations are supported by evidence from my interpretation of over 30,000 km of seismic data from these margins. Some published and unpublished interpretations are also presented and are referenced accordingly. However, detailed local structural interpretations are not the emphasis of this chapter. Rather, structural interpretations of individual rifted sub-basins are used to obtain an integrated regional tectonic and kinematic interpretation for each margin in order to evaluate the role of pre-existing structures in subsequent deformation.

Three sections of Australia's passive margins are considered (Figure 1.1). Margins were chosen on the basis of data availability and apparent similarities in gross crustal configuration with the EARS. If we assume that the EARS will eventually fragment the Somali plate (i.e. continental crust to the east of the
Eastern Branch of the EARS) from the African plate (i.e. continental crust to the west of the Western Branch of the EAR) and that the Aswa and Rukwa lineaments (Figure 3.1) are actively involved in the process, at least two final configurations are possible.

First, the Eastern Branch may rupture. The passive margin formed will juxtapose the extended continental crust of the Eastern Branch with newly formed oceanic crust. The Tanzanian craton, between the Eastern and Western Branches and the two aforementioned lineaments, will likely subside to form an apparent continental indentation underlain by continental crust, a configuration represented by the Great Australian Bight. Second, the Western Branch may rupture. The Tanzanian craton would then be attached to the eastern continental fragment and bounded to the north and south by transform margins (i.e. the Aswa and Rukwa lineaments) creating a continental promontory. The promontory will be bounded towards the continent by rift zones and will most likely subside. The Queensland Trough and Northwest Shelf areas both have extended continental crust both inboard and seaward of submerged continental plateaux, suggesting a similar regional configuration as the Western and Eastern Branches bounding the Tanzanian Craton of the EARS. A documented shear margin bounding the southern end of the Northwest Shelf plateau (Exmouth) may be considered an analogous structure to the Rukwa and Aswa lineaments acting as intracontinental transforms and evolving into oceanic transforms. In the event that both branches rupture, the Tanzanian craton would become a "stranded" continental fragment surrounded by oceanic crust.

A study of the NE Australian margin included analysis of previously unpublished data from the Queensland Trough, bounding the continentward margin of the Queensland Plateau (Scott 1993). The large-scale crustal structure is characterized by a central craton bounded by thinned, rifted crust. Data from this area were made available by the Australian Geological Survey Organization (AGSO). Unfortunately, recently collected data from the adjoining Townsville Trough is still being processed and detailed interpretations are just now being conducted (P. Symonds, pers. comm.). However, preliminary results from these interpretations have been presented at various symposia (e.g. Symonds et al. 1987, Symonds & Davies 1988) and my informal interaction with AGSO workers has ensured that the results presented herein are consistent with interpretations of the new Townsville Trough data.

The initial interpretations of the Northwest Shelf and the Great Australian Bight (GAB) were undertaken under the auspices of Marathon Petroleum Ltd., in
conjunction with Mr. Greg Irwin. All data were provided in paper form and the interpretations of these data have been presented as technical reports and are housed in Marathon's (Perth) private library (Marathon 1991, Marathon 1990). Parts of the sections dealing with the GAB and NW Shelf are excerpted from these reports, but are my original work except where noted. The sheer volume of data considered from these two margins precludes presentation of all the profiles, and only a few seismic profiles are shown to highlight key points. The details of the interpretation of the Northwest Shelf data were considered proprietary up to a year after the submission of the technical report (Marathon 1991). However, the pertinent aspects of the NW Shelf study are summarized in Section 4.4.
Figure 4.1 Location map of the Queensland Trough (heavy box) on the northeast margin of Australia and major physiographic features. Contours are in 1000's of meters. GI = Georgetown Inlier. BRE = Broken River Embayment.
4.2 Architecture of the Queensland Trough

4.2.1 Introduction

The Queensland Trough is a bathymetric deep located just seaward of the Great Barrier Reef of northeast Australia (Figure 4.1). Its present-day axis trends NW-SE at 155°. The trough defines the boundary between the continental shelf and the submerged Queensland Plateau, considered to be continental crust (Ewing et al. 1970). The deeper northern end of the trough has a maximum depth of 2800 m, and is linked to the Coral Sea Basin to the north by the Osprey Embayment, which appears to be underlain in part by oceanic crust (Symonds et al. 1984). The southern terminus of the trough shallows to 800 m at the intersection with the approximately E-W trending Townsville Trough, separating the Queensland and Marion Plateaux.

The Queensland Trough can be considered a "typical" rift zone in that it is a long, narrow (i.e. < 100 km wide, ~ 500 km long, similar in scale to the Lake Tanganyika and Lake Malawi rift zones) apparently linear structure. Classical models of rifting, as discussed in Chapters 1 and 2, would predict normal rift bounding faults parallel to the trough's elongation and transfer faults perpendicular to that trend. The interpretations presented in this section suggest that this fault geometry does not adequately describe the basin forming structures within the trough. Previous interpretations of the Queensland Trough invoked either Early to Late Cretaceous extension associated with the creation of the Tasman Sea spreading ridge to the south, or Paleocene to Eocene extension associated with the Coral Sea spreading ridge to the north, as mechanisms for trough formation (e.g. Falvey & Taylor 1974, Falvey & Mutter 1981, Falvey et al. 1990, Gardner 1970, GSI 1980, Karig 1971, Pinchin & Hudspeth 1975, Mutter 1977, Mutter & Karner 1978, Taylor & Falvey 1977, Symonds et al. 1983, 1984, Symonds & Davies 1988). Seismic refraction data (Ewing et al. 1970, Falvey & Taylor 1974) suggest a sediment package at least 4 km thick in the trough. However, no reliable dating of the deep sediments is available, so age constraints on the basin forming structures are lacking.

The present study complements other work (largely Australian Geological Society Organization (AGSO) sponsored projects) that addresses the evolution of Australia's northeast margin. The Queensland Trough is a little known major depocenter, whose evolution and architecture are critical to understanding the tectonic evolution of this region. The data are presented as digitized line drawings
of seismic profiles as only paper copies were available for reproduction. Profile-
specific geometries demonstrate that both "dip" and "strike" lines exhibit
substantial rotation of extended basement blocks.

The wide grid spacing of the presently available data makes it impossible to
precisely constrain the structural geometry of the Queensland Trough. However,
two extension models can be fitted to the observed structures, providing insights
into the underlying rift architecture. Both alternative fault patterns, combined
with measured basement rotation directions, indicate that the Queensland Trough
formed by oblique rifting. The proposed direction of extension appears to be
incompatible with plate motion vectors involved in the seafloor-spreading phases
responsible for opening either the Coral and Tasman Seas. Kinematic analysis of
the structures mapped in the trough, and in the adjoining Townsville Trough,
demonstrates that basin-forming structures were not associated with either of the
aforementioned seafloor-spreading systems. It is suggested that the extension
which produced the Queensland Trough pre-dates the extension episodes
responsible for opening both the Coral and Tasman Seas, as suggested by
earlier workers (Mutter & Karner 1980, Symonds et al. 1987) who propose that both
troughs were part of a ?Paleozoic triple junction.

4.2.2 The Seismic Reflection Database

Figure 4.2a shows the location of approximately 3700 km of multi-channel
seismic data collected by GSI (1979, 1974) and provided in paper form by AGSO. The
1979 rectilinear grid over the trough has a spacing of approximately 50 km. The
1974 grid was shot in a zigzag pattern, thus providing line spacing as close as a few

![Figure 4.2](next page) An example of the seismic data from the Queensland
Trough. Two megasequences (referred to as the post and synrift) overlying acoustic
basement (prerift) are recognizable throughout the data set. Dotted line represents
the boundary between flat-lying continuous reflections of the postrift section and
the dipping, semi-continuous reflections, separated by chaotic zones and
diffractions of the synrift section. Solid line separates the base of the synrift from
the chaotic, discontinuous reflections of the prerift basement. Note the variable
rotation of tilted fault blocks on this NE trending "dip" line. For location see
Figures 4.2A and 4.4B.
LINE 7 s.p. 4500-5700

Line 111

054°

10 km

POST

SYN

PRE

-1.0 s
-2.0 s
-3.0 s
-4.0 s
-5.0 s
-6.0 s
-7.0 s
kilometers locally. The seismic data were processed using standard industry parameters of the time and range in quality from fair to poor. However, postrift and synrift sequences and acoustic basement are discernible on most profiles. The interpretation presented herein is supplemented by gravity and magnetic data collected during the same surveys (GSI 1980). Profiles perpendicular to the trough axis are conventionally referred to as "dip" lines and profiles parallel to the trough axis as "strike" lines and this usage is retained herein. However, as has been shown in the EARS case studies, these terms are misleading in the context of oblique rifting.

Two acousto-stratigraphic megasequences can be recognized in the seismic data (Figure 4.3). The postrift section comprises flat-lying, very continuous reflections and extends up to two seconds (two-way-travel-time, TWT) below the water bottom. The syn rift section is characterized by semi-continuous, moderately dipping, diverging reflections that are separated by zones of chaotic reflections and diffractions. The megasequence boundary separating the syn- and postrift packages is distinct, characterized by onlap above and angular truncation below. The prerift or basement section is commonly seismically chaotic or incoherent, but steeply dipping, discontinuous reflections are recognizable locally, especially in areas of high-standing basement within the rift and on the adjacent plateau. The boundary between the synrift and underlying basement is sometimes quite distinct, but is more commonly defined by a gradual loss of seismic character and/or a velocity jump. No wells have penetrated the synrift package, but 1000 m of the postrift section were cored in the central Queensland Trough during Leg 133 of the Ocean Drilling Program (Davies et al. 1991). The hole was drilled near the intersection of Lines 7 and 111 (Figure 4.2a). Its total depth corresponds to 3.2 seconds TWT, well above the interpreted boundary between the post- and synrift megasequences interpreted herein (Figure 4.3), and the oldest sediments intersected have been dated at Early Miocene.

Digitized line drawings of the "dip" lines are shown in Figure 4.4. In each case, the dashed line is the seabed. The majority of these profiles trend northeasterly (Figure 4.2), but profile trends range from northeast, to due east, to southeast. The dotted line is the boundary between the flat-lying reflections of the postrift megasequence and the dipping reflections of the synrift megasequence. The synrift reflection configuration can vary substantially. Synrift reflections are sometimes quite continuous, as in the half graben on the right end of the profile in Figure 4.3, but also can be semi-continuous dipping reflections or diffractions, as in the half graben imaged on the left. As the focus of this study is the synrift
megasequence and its tectonic significance, the flat-lying postrift reflections have been left out of the line drawings for clarity.

As noted earlier, water depths increase to the north along the axis of the Queensland Trough. However, there is no similar systematic increase in the thickness of the postrift sediments; rather, they thicken over areas underlain by the dipping reflections of the synrift section and thin over basement highs. This may indicate either sedimentation into remnant rift topography, increased subsidence due to greater sediment loads adjacent to major faults or later fault movement in some places (i.e. basement faults propagate through the Tertiary to Recent section in some localities). Where there is no evidence of an underlying synrift section or where basement shallows, profile sections have been omitted from the line drawings.

4.2.3 Structural Style

A striking feature of the profiles is the predominance of half graben morphology along the entire rift. The majority of the half graben deepen to the east and are bounded by west dipping faults. Exceptions are the isolated half graben deepening westward on the right end of Line 1122 (Figure 4.4a) and a system of likewise westward deepening half graben in the southern-central portion of the rift on the right end of Lines 1124, 1125 and 7 (Figure 4.4b). Profile characteristics vary down the axis of the rift and are separated in the following discussion into the northern, central and southern zones. The northern profiles (Lines 1 and 2, Figure 4.4a) image distinctly more chaotic and less coherent reflections below the postrift megasequence. To the south on Lines 1121, 1122 and 3 (Figure 4.4a), the deep structure of the trough is characterized by one or two half graben, separated by apparently undeformed high-standing basement blocks. Farther south still, most profiles show a series of tilt-blocks that define larger composite half graben (Lines 1123, 4, 5, 6, 1124, 1125, 7, and 8, Figures 4.4a and b). Finally, in the southernmost profile (Line 10, Figure 4.4b), extensional strain is manifested by two widely separated half graben which lie on both sides of the bathymetric trough. In the next

Figure 4.4 A) Line drawings of the northern "dip" lines in the Queensland Trough. B) (Next page) Line drawings of the southern "dip" lines in the Queensland Trough. Orientations of profiles are indicated by arrows and azimuths. Profiles are "hung" on the strike Line 110 and view is looking NW (335°) along the trough axis. Intersections with strike lines are indicated (Figures 4.5 through 4.7). Boxes indicate portions of seismic profiles shown in Figures 4.3, 4.8 and 4.10 for comparison to line drawings. Dashed lines represents the water/sediment interface. Dotted lines represent the boundary between the post- and synrift megasequences. See Figure 4.2a for locations.
Figure 4.10

QUEENSLAND TROUGH

LINE 1
070°

LINE 2
068°

LINE 1121
090°

LINE 1122
052°

LINE 3
057°

LINE 1123
108°

LINE 4
058°
sections, each of the three zones is examined in more detail, progressing from north to south. "Strike" or rift axis parallel profiles are also included to determine the overall geometry of the synrift structure.

4.2.3.1 Northern Zone

The three most northern dip lines (Figure 4.4a) are tied by strike lines shown in Figure 4.5. Similarly to the two most northern profiles (Lines 1 and 2), the northern ends of the strike profiles have very discontinuous and chaotic reflection character beneath the postrift megasequence. The sense of half graben polarity and the presence of tilted fault blocks is ambiguous, although some dipping reflections do extend deep into the section. The next line to the south (1121) records a clear, albeit isolated, half graben that is manifested on the strike profiles (Lines 110 and 111, Figure 4.5) by a thickening of the postrift section in this location.

Although it is not clear from these line drawings, the northern synrift reflections tend to be of higher amplitude than those to the south. The highly diffractive character of the reflections suggests the presence of volcanics; this is supported by the proximity to interpreted oceanic crust to the north in the Osprey Embayment (Symonds et al. 1984). Although no wells have penetrated the very high-amplitude events within the postrift section, they are commonly interpreted to be flows or sills (e.g. Figure 4.4a, labelled with "V"s on the right end of Line 1 between the intersections with Lines 110 and 109.)

Dip Lines 1121 and 1122 (Figure 4.4a) show well-defined half graben, isolated from each other by seemingly undeformed, planated basement. These isolated half graben average 10-15 km in width. On Line 3, there is some evidence for a series of tilted fault blocks joining the two half graben on either end of the profile. On the eastern end of Line 1122 on the Queensland Platform, the half graben which deepens to the west does not extend to either adjacent profile and is apparently totally isolated from the main trough. To the south on Line 3, no half graben is present, and plateau basement to the east appears to be undeformed so is not shown in the line drawings. An isolated half graben with the opposite polarity is imaged on Line 1121 to the north, suggesting that an unrecognized structure (?transfer fault) must be present between the two profiles to allow for the different basement dip directions.

Strike profiles (Figure 4.6) that tie dip Lines 1122 and 3 record dipping reflections, but individual half graben are not well defined. On Line 110, a full-
graben morphology is suggested by inward dipping faults and outward dipping synrift reflections at either end of the profile. These reflection/fault geometries are consistent with our traditional view of "dip" versus "strike" lines. That is, the data record higher rotation of synrift reflections, and more clearly imaged extensional features, such as tilted fault blocks, on dip profiles rather than on strike profiles. This relationship is predicted if extension has been aligned NE-SW, approximately perpendicular to the rift axis. However, this relationship is not consistent throughout the rift zone.

4.2.3.2 Central Zone

In the central part of the trough, dip Lines 1123, 4 and 5 (Figures 4.4a and b) image a series of tilted fault blocks combining to form composite half graben rather than the single isolated half graben that characterized the northern part of the trough. The composite half graben morphology is typical of rifted basins elsewhere (e.g. Florensov 1969, de Charpal et al. 1978, Ramberg & Neumann 1978, Chenet et al. 1982, Logatchev et al. 1983, Bally 1984, Ramberg & Morgan 1984, Skilbeck & Lennox 1984, Etheridge et al. 1985, Rosendahl 1987). The average size of tilted fault blocks has decreased to 5-10 km. The postrift section thins and thickens over each fault block indicating probable later reactivation of the rift faults. The only strike line through this part of the trough with recognizable synrift section is the one that runs down the axis of the trough (Line 110, Figure 4.7a). It has all the characteristics one expects to see: that is, reflection packages with a variety of dips separated by subvertical reflection disruptions. The postrift section undulates near the ties with Lines 4 and 1123, but is fairly constant in thickness and depth south of the tie with Line 5. The thickness of the postrift section along this strike line does not vary as systematically as on the dip lines, where changes in thickness reflect the asymmetry of deeper structures.

Seismic profiles over the half graben on the western end of Line 5 and the section of a strike line that it crosses are shown in Figure 4.8. Although basement dips toward a west dipping fault on the western end of Line 5, there is clearly as much, if not more, basement dip and higher subsidence to the south along Line 111. Likewise, the relationship of the postrift thickness and underlying structure noted above is clearer on the strike line (111) than on the dip line (5). This example
QUEENSLAND TROUGH (North-Central)

LINE 110

LINE 111

20 km
Chapter 4 Australian Passive Margins

highlights two key observations in constraining the fault geometry of the basin­
forming structures in the Queensland Trough. That is, seismic profiles of all ori­
entations in the data set reveal tilted basement fault blocks, and there is not a con­
sistent relationship between basement rotation and orientation of the profiles with respect to the rift zone axis as is expected in simple extensional regimes (Chapters 1 and 2).

4.2.3.3 Southern Zone

In the next "dip" profile to the south (Line 6, Figure 4.4b), tilted fault blocks are locally as narrow as 1 to 3 km across. Lines 1124, 1125 and 7 to the south image the only observed linked polarity switch along the rift axis. On these three adjacent profiles, a basement horst separates opposing polarity half graben, both deepening toward the horst. The westernmost strike profile (Line 111, Figure 4.7b) also images both tilted fault blocks and diverging synrift reflections, in contrast to what is expected from a "strike" line in simple rift-perpendicular extension. In fact, the half graben imaged at the tie with Line 1124 is clearly tilted on the strike profile (Line 111). The easternmost strike line (Line 109, Figure 4.7b) sits almost entirely on the western edge of the Queensland Plateau. However, where the synrift section is imaged to the right of the tie with the dip profile (Line 1125), it also has a significant dip. The middle strike profile is not shown, because it images only the apparently undeformed basement horst seen in the "dip" lines. Although the horst appears to be a continuous feature, crossing Lines 1125 and 7 near their intersection and continuing north a distance of approximately 40 km to Line 1124, it is not seen on Line 6 farther north. Likewise, the horst is not imaged on Line 8 to the south. On the western end of Line 8 (Figure 4.4b), westward dipping synrift reflections and eastward dipping fault plane reflections, continue through the intersection of Line 110 where one expects to see the continuation of the horst. Eastward of the expected location of the horst a platform begins and continues apparently undeformed to the eastern end of the line, where one might expect to see evidence of the half graben imaged on Line 7, or the opposite polarity half graben on Line 10.

On Line 10 (the southernmost profile, Figure 4.4b), a large block of apparently undeformed basement is substantially broader than the horst described above. The absence of either the horst or this broad expanse of undeformed basement on Line 8 is evidence that they are not continuous. Well-defined half graben on either side of the undeformed block on Line 10 are of the same polarity, in contrast to the opposite polarity half graben adjacent to the horst to the north. The half graben are widely separated, suggesting that the extensional strain has...
been more widely distributed. At this location, the trough is merging with the nearly E-W Townsville Trough, and the half graben on the right end of the profile may be continuous with this system.

Although too few seismic profile intersections exist to rely on dip analysis, computed dips do constrain the block directly underlying the intersection. There are 22 profile ties where underlying basement rotation can be measured with reasonable confidence, 15 yield a true dip direction of WNW-ESE implying that tectonic transport direction or extension direction is aligned along this azimuth. Five of the seven NE-SW dips are found in the north, where the presence of volcanics obscures rift-related structures. The other two are adjacent to WNW-ESE faults near their intersection with NNE-SSW faults. As discussed previously, dips measured directly adjacent to non-ideally aligned faults, may not reflect the overall dip of the half graben.

The above descriptions of profile geometries and basement morphology in the Queensland Trough lead to the following conclusions:

• the trough is divisible into three zones, each with a characteristic deformation style;

• seismic profiles of all orientations in the data set reveal tilted basement fault blocks;

• there is not a consistent relationship between basement rotation and orientation of the profiles with respect to the rift zone axis;

• computed basement dips, where available, are consistent with oblique, not orthogonal, extension.

**Figure 4.7** Line drawings of the A) central and B) south-central "strike" lines of the Queensland Trough. Notations and locations as in Figure 4.4.

**Figure 4.8** (Next page) Tying seismic profiles from the Queensland Trough. Line 5 is a "dip", or trough perpendicular profile. Line 111 is a "strike", or trough parallel profile. Apparent rotation of the half graben near the profiles' intersection is greater on the "strike" line than on the "dip" line. Note also the thicker post-rift section on the "strike" line indicating greater subsidence adjacent to the bounding fault. Notations and locations as in Figure 4.3.
4.2.4 Architectural Models of the Queensland Trough

Because the available data are widely gridded, fault correlations from profile to profile are poorly constrained, except where fault picks are imaged on intersecting profiles and the fault lies close to the intersection. Indeed, the variation in rift morphology described in the profile descriptions above make continuation of any of the structures along the axis of the rift suspect. In the following sections, fault picks are extrapolated with the aid of rift geometry models (i.e. the EARS curvilinear model and rectilinear orthogonal extension model, Figure 1.3). These interpretations are tested for compatibility with the additional information that can be gleaned from the available computed basement dips, the thickness of the synrift sequence (Figure 4.2) and careful analysis of dip relationships on the profiles themselves.

4.2.4.1 The Curvilinear Interpretation

The data show that seismic profiles of all orientations contain tilted basement fault blocks. Many bounding faults are clearly listric (Figure 4.3), but there seems to be no consistent "dip" versus "strike" line relationship. This inconsistency is analogous to the data from the EARS (Chapter 3), where it has been proposed that half graben are bounded by arcuate or curvilinear faults connected by various types of accommodation zones into a sinusoidal fault system (Rosendahl 1987). In fact, the crustal scale analogy between the Tanganyika/Rukwa/Malawi rift zones bounding the Tanzanian Craton and the Queensland/Townsville Trough system bounding the continental block of the Queensland Plateau was the principal motivation for this study. As discussed in Chapter 3, field measurements of slip on rift faults (Chorowicz & Sorlien 1992) and the dip analysis in the EARS, indicate that extension is oblique to the rift axis (see also Scott et al. 1990, 1992). In addition, the geometry of the horst and adjacent half graben in the south central part of the Queensland Trough (Lines 1124, 1125 and 7, Figure 4.4b) is that described as an isolation accommodation zone by Rosendahl (1987) (cf. Line 214, Figure 3.14 and Line 7, Figure 4.4b).

With the EARS sinusoidal geometry as a model, the curvilinear fault interpretation shown in Figure 4.9a was derived from the available data. Fault picks near profile ties constrain fault strikes locally, but model concepts were used to correlate structures across these widely spaced data within the trough. The half graben bounding faults form a series of synrift depocenters with up to 5 km of fill, elongated along the axis of the trough at 155°-160°, as shown in the isopach map of
the synrift section (Figure 4.2b). This apparent structural grain suggests rift parallel bounding faults and orthogonal extension. However, the available data yield no evidence for major linear structures parallel or perpendicular to the trough axis. The series of arcuate half graben north of 14°S in the Queensland Trough (Figure 4.9a) are elongate more or less parallel to the trough trend; that is, the tangent to their point of maximum curvature is parallel to the rift trend. However, farther south fault patterns become more rhombohedral, suggesting that a component of strike slip deformation was involved in the formation of the half graben. Note also that most of the synrift depocenters are composed of two or more smaller "deeps" (Figure 4.2b) that cannot be explained by the curvilinear fault model interpretation presented in Figure 4.9a.

The main criticism of the sinusoidal rift model (Reynolds 1984, Rosendahl 1987) proposed for the EARS is that it requires quite large internal deformation of fault blocks. Furthermore, the model does not include structures (i.e. transfer faults) that allow inferences about regional kinematics or direction of extension and so has no predictive value in tectonic reconstructions. In other words, linking structures aligned approximately parallel to the extension direction are required to "release" blocks bounded by variably oriented faults along the rift-axis strike to allow tectonic transport of the blocks. The sinusoidal model predicts at least two directions for half graben linking structures (accommodation zones), which do not have a consistent relationship with the direction of extension. Scott et al. (1992) and Chorowicz & Sorlien (1992) addressed this problem by analyzing the kinematics of the EARS from two different data sets. Both investigations concluded that transverse structures must exist to explain fault striae and structural dips. It is possible that the structures are ill-defined in intracontinental rifts that have a small overall extension and so have remained unrecognized in the seismic data previously. However, sufficient evidence exists in both the EARS and the Queensland Trough to conclude that the structures are present and do provide a consistent kinematic history for the rift zones.

Figure 4.9  A) Curvilinear fault map of the Queensland Trough using model concepts from the East African Rift (Rosendahl 1987). Shaded strip indicates a possible zone of structures and their trend, that are required if rift-perpendicular extension is assumed, to decouple mapped normal faults of varying orientation or opposing polarity. Structures of this orientation are not recognized in the available data. AZ = accommodation zone. B) Rectilinear fault map of the Queensland Trough after concepts of the model presented by Lister et al. (1986). Note that in both interpretations fault correlations adjacent to profile intersections are oblique to the axial trend of the rift.
4.2.4.2 The Rectilinear Interpretation

In a simple, ideal orthogonal extension model, a "dip" line shot parallel to the direction of extension images a series of tilted fault blocks with similar rotation and recoverable fault/base synrift relationships (Figure 2.1). Likewise, a "strike line" images flat-lying to shallowly dipping synrift reflections and vertical faults. Profiles in any other orientation are predicted to image complex geometries which cannot be reconstructed in the plane of the profile simply by movement along the imaged faults. The rift axis will align perpendicular to the direction of extension, although deeps may shift their lateral position from the axis or switch polarity across extension parallel, rift perpendicular transfer faults.

As outlined in Section 4.2.2.3 basement rotation dip azimuths suggest that rotated basement block dip direction is NW-SE within the trough (Figure 4.2a). An interpretation based on the rectilinear fault geometry model is consistent with the limited dip information (Figure 4.9b). Line 1123 (Figure 4.10) trends southeast (108°) rather than northeast, the trend of the majority of the trough-perpendicular "dip" lines (Figure 4.2a). Line 1123 most closely resembles the series of "balanceable" tilt blocks within a single profile that is predicted to be on a dip line in an orthogonal extension system. Note the morphology of the base of the synrift section. Block dips are consistent with each other. Line 4, which intersects Line 1123, also images tilted fault blocks, but they are not as clear, and there is considerable disruption of synrift reflections within each apparent fault block (Figure 4.4a). Changes in the thickness of the postrift section indicate that both profiles consist of three tilted fault blocks starting at their intersection and proceeding eastward. However, correlation of the bounding faults does not produce parallel structures. The easternmost block on Line 1123 appears to be smaller than the middle block, but the reverse relationship seems to be the case in Line 4. Any attempt at correlating faults on these two intersecting profiles requires at least two fault trends. The strike line that intersects these profiles (Line 110, Figure 4.7a) images flat to dipping reflectors, which are truncated on a 1 to 3 km scale versus the 5 to 10 km scale of blocks in the "dip" lines. Interpretation of these profiles suggests that the normal faults accommodating extension in the rift trough are aligned roughly perpendicular to Line 1123, or at about 018°, and that tilted blocks are separated from each other by "transfer" or oblique to strike slip faults aligned subparallel to the profile. This interpretation can accommodate all the observed faults as well as explain the chaotic nature of reflections within the half graben imaged in Line 4.

It is also instructive to examine the geometry of the two opposing half graben and the horst located on Lines 1124, 1125 and 7 (Figure 4.4b) interpreted as
an isolation accommodation zone in the previous section. Line 110 (not shown) is trough axis-parallel and images continuous elevated basement from just south of Line 7 to just north of Line 1124. Given the similarity in profile morphology, and the proximity and orientations of the lines, the interpretation of a trough-parallel elongate horst shown in the curvilinear fault map (Figure 4.9a) seems reasonable. However, attempting to correlate the rest of the faults within either of the adjacent half graben leads to a variety of interpreted fault trends, as was the case for structures imaged on Lines 4 and 1123. Also, the opposing polarity half graben on Line 6 (Figure 4.4b) requires an intervening structure between the profiles to allow the asymmetry reversal.

Close inspection of the morphology of the horst (Figure 4.4b) reveals that basement dips atop the horst vary from profile to profile. In Line 7, the apex of the horst is east of the intersection with Line 110, with a wider westward dipping surface. On Line 1125, the apex has shifted to the west of the intersection with Line 110 and has a distinctly wider eastward dipping surface. The relationship of the apex to Line 110, and between the widths of the dipping surfaces, again reverses on Line 1124. In addition, a strong diffraction to the west of the intersection of Line 1124 with Line 110 suggests a possible structure(s) running through the horst at an angle with Line 110 rather than parallel to it. Correlation of these horst features yields a zigzag pattern whose segment trends are both oblique to the rift axis and suggests that this basement high is not a continuous unbroken feature but may be cut by subvertical strike slip faults.

Where half graben or tilted fault blocks are discernable, and where correlations of faults very near the profile ties is possible, faults trend NNE-SSW, striking between 010°-025°, or WNW-ESE, with strikes between 100° and 120° (Figure 4.9b). These strikes are compatible with those predicted by the reflector dip/fault geometry interpretation done on Lines 1123, 4 and 110 (Section 4.2.3.2, Figures 4.4a and 4.7a) based on orthogonal extension models, even though these trends are both oblique to the rift axis trend (Figure 4.9b).

4.2.5 Discussion

The oblique relationship between the fault trends and the rift axis in the rectilinear fault map (Figure 4.9b) is nearly identical to that in the interpretation of
the Lake Tanganyika rift zone in the Western Branch of the EARS (Chapter 3, Figure 3.8). In the EARS basins, determination of the tectonic transport direction based on dip analysis confirms that oblique rifting has been operative there. This conclusion is supported by landbased fault slip analysis by Chorowicz & Sorlien (1992). A similar dip analysis on the Queensland Trough data is limited due to the small number of profile intersections (Figure 4.2a). However, the computed dips in the Queensland Trough are consistent with oblique, rather than orthogonal extension.

It is likely that the structural pattern in the Queensland Trough is neither as regular and rectilinear, nor as perfectly arcuate and unlinked, as that shown in the two alternative interpretations in Figure 4.9. Rather, it is more likely that the actual structural style lies somewhere in between as modelled for the EAR by Chorowicz & Sorlien (1992), wherein extensional rift-bounding normal faults are arcuate and are compartmentalized by transfer faults aligned approximately in the direction of extension, which is oblique to the rift axis (Figure 4.11).

By incorporating arguments based on predicted orthogonal extension geometries and observations of geometries within the data, it is possible to construct an interpretation which provides a linked, balanceable fault system with a consistent extension direction (Figure 4.11). The underlying synrift extension of the Queensland Trough appears to have been accommodated by somewhat variably trending NNE-SSW normal, possibly curvilinear, faults. The normal faults are truncated by pervasive WNW-ESE oblique to strike slip, steeply dipping faults in this interpretation. The structural architecture presented in Figure 4.11 also fits with the compartmentalization of the synrift isopach cells into smaller "deeps". Further, the interpretation is consistent with our current understanding of strain distribution in extensional systems, with the qualification that both normal and transfer faults are aligned oblique to the trough axis.

In Figure 4.11, there is a noticeable change in the trend of the normal faults at 15° 15'S from NNE-SSW in the north to N-S in the south, reverting to a NNE-SSW trend south of Cairns. The changes are reflected in slight changes of the computed dips (Figure 4.2a) and appear to be associated with larger scale crustal structures. That is, the NW-SE structures that separate the variable normal fault trends are approximately aligned with changes in the continental shoreline trend from N-S to NW-SE. Also, N-S trending residual negative Bouger gravity anomalies onshore are truncated by NW-SE trending relative highs that align with the NW-SE structures within the trough (Murray et al. 1989; their Figure 1). The change in fault trends is therefore attributed to major crustal shear zones that have...
partitioned the structural style along the axis of the Queensland Trough, as discussed in the next section. This hypothesis could be tested by improved gravity or magnetic data coverage offshore, which would verify (or negate) the continuation of crustal features from the continent into the trough.

4.2.6 Tectonic Implications

The coastal regions of northeastern Queensland are underlain by rocks of the Paleozoic Tasman Fold Belt. The width of the Tasman Fold Belt here is much less than the width of the age-correlated Lachlan and New England Fold Belts to the south, suggesting that the Queensland Plateau may represent the seaward continuation of Tasman Fold Belt (Ewing et al. 1970). Acoustic basement on the Queensland Plateau was penetrated by rotary drilling at Sites 824 and 825 during ODP Leg 133. The basement rocks are described as "altered and deformed metasediments" and "undeformed intermediate dikes" similar to the Ordovician to Devonian lithologies of the onshore Hodgkinson Province (Symonds 1992, Feary et al. in press) (Figure 4.1).

It has been suggested that the mapped transverse faults in the Townsville Trough are utilizing a prominent NNE structural grain evident in the onshore Tasman Fold Belt (P. Symonds, pers. comm.). The bend in the shoreline mentioned in the previous section aligns with the Palmerville Fault Zone, a proposed Proterozoic suture (Munson 1986). The Hodgkinson Province, lying between the Palmerville Fault Zone and the coast, has inspired a wide variety of tectonic interpretations, from arc-trench accretionary complex models to a thrust belt on the margin of an intracontinental basin. Recent detailed structural analysis favors the latter interpretation (Hammond 1986). However, extension of onshore continental lineaments to the Queensland Trough remains enigmatic. Crustal models generally suggest normal thick continental crust underlying the Great Barrier Reef (Figure 4.12) (Mutter & Karner 1980). Only sparse geopotential data are available to constrain the extent of lateral heterogeneities offshore and very little new data have been acquired to assess the boundary between normal and thinned crust in the Queensland Trough due to its proximity to the Great Barrier Reef.

Figure 4.11 Interpreted tectonic architecture of the Queensland Trough. Grayed contours are the synrift isopachs shown in Figure 4.2B. Note the change in the orientation of the normal faults just south of 15° S latitude. This interpretation is consistent with the synrift isopach data, the dips of basement rotations (Figure 4.2A), profile geometries discussed in the text and correlated structural trends close to profile intersections. XXX = westward extent of relatively highstanding basement blocks. V = volcanic seismic signature.
The structural interpretation of the Queensland Trough suggests that extension was directed WNW-ESE at an oblique angle to the NW-SE (155°) trend of the rift axis (Figure 4.11). The axis of the trough may be so aligned in response to the regional grain, or perhaps a boundary of the Paleozoic Tasman Fold Belt (Mutter 1977). The structural pattern of NNE normal faults and WNW transfer faults in the Queensland Trough is broadly consistent with interpretations from the Townsville Trough (Falvey et al. 1990, Symonds et al. 1987, 1988). Analysis of the Townsville Trough data suggests NW to NNW extension resulting in a ENE set of normal faults, with NW to NNW transfer faults, which appear to align with the fabric of the onshore Tasman Fold Belt (P. Symonds, pers. comm.). As an alternative to the WNW regional extension direction proposed above, it may be that NW to NNW extension indicated by Townsville Trough structural interpretations exploited a WNW crustal anisotropy suggested by the strain partitioning along the Queensland Trough and onshore gravity signatures (Murray et al. 1989). Whichever the case, extension was generally directed NW-SE throughout the system (Figure 4.12). How can we fit this kinematic interpretation logically into the overall tectonic history?

Infinitesimal strain theory for rotational movements of crustal blocks on a sphere predicts that rift segments whose axes do not trend along great circles that intersect the pole of rotation will experience oblique opening (Smith & Durney 1992, Smith 1993). The structural angle, β, between the normal rift bounding faults and the rift axis trend, is theoretically related to the divergence angle, α, between the displacement vector and the rift axis (Figure 4.12b and c) (Sanderson & Marchini 1984, McCoss 1986, Withjack & Jamison 1986, Tron & Brun 1991, Smith & Durney 1992). The trend of the rift axis and the distance of the segment from the pole determine the degree of obliquity (Figure 4.12c). It is possible to determine the approximate location of the instantaneous pole of rotation at the time of formation of the normal, "en echelon" structures within a rift segment from these relationships, provided the structures themselves have not been subsequently rotated. Obviously, the more structural orientation measurements, the higher the accuracy of the determination. Well constrained bounding structure orientations from synchronous rift segments with different axial trends adds to the validity of the determination. Figure 4.12 depicts the predicted pole of rotation for the Queensland and Townsville Troughs using the best constrained structure from each rift zone (i.e. the interpretation of the Queensland Trough presented herein and P. Symonds interpretation of the Townsville Trough, pers. comm.). Although no claim is made for the precision of the pole, it is quite certain that the Queensland/Townsville Trough formation pole was to the SE of the Queensland Plateau, and not coincident or even broadly compatible with either the
Coral or Tasman Sea poles. Given the general structural relationships in both troughs as outlined above, we can also say that the pole was close to the rift zones, which define the boundaries of the Queensland Plateau crustal block. Close proximity of a pole to the boundary of a crustal block may imply a higher rate of rotation than if the pole were located at a greater distance (e.g. the "roller-bearing" model of Schouten et al. for oceanic microplates). If the Queensland Plateau has undergone a rotation about a vertical axis located within itself, shearing along all its margins is to be expected and provides a mechanism for the the component of strike slip seen in both troughs.

As mentioned earlier, the regional crustal morphology of rift zones wrapped around a craton describes both the Tanganyika/Rukwa/Malawi Rift and the Queensland/Townsville Trough. However, the large-scale similarity ends there. The Tanganyika and Malawi rift zone axes are subparallel and do not intersect; rather, they are joined by the strike slip Rukwa rift zone (e.g. Rosendahl et al. 1992a). The Queensland and Townsville Troughs intersect at their southern and western termini, respectively. Further, the troughs’ axes strike radially from their point of intersection at an angle of approximately 115° to each other. This configuration is very suggestive of an incipient triple junction (McKenzie & Morgan 1969). There is some evidence that a third arm of the system extends inboard of the Marion Plateau. White (1961, 1965) first proposed that NE trending structures bounding the Broken River Embayment (Figure 4.1) represent an intracontinental rift. Mutter & Karner (1980) also hypothesize a possible Paleozoic three-branch system of which only the Queensland and Townsville branches were reactivated in the Mesozoic. As evidence they cite prominent structural lineaments of the appropriate trend in the Broken River Embayment and the Georgetown Inlier (Figure 4.1). Arnold & Fawkner (1980) argued that these half graben structures were active during deposition of Ordovician and Silurian

**Figure 4.12**  A) Kinematic analysis of trough-forming extension, offshore NE Australia. B) The angular relationships between the displacement direction (α) and normal "en echelon" rift structures (β) with the rift trend (RT) during oblique divergence is theoretically predictable from the infinitesimal kinematics of rotational rifting on a sphere and closely reproduced in analog experiments (Smith 1993, Smith & Durney 1992 and references therein). C) Assuming structural trends are preserved, the location of the instantaneous pole of rotation during the earliest phases of continental rifting is related to these angles, which vary for different segments of a rift zone, depending on the orientation and distance from the pole of the rift segment. Structures in the Queensland and Townsville trough are used to predict an approximate pole for the earliest rifting, $P_0$, which is inconsistent with both the Coral Sea or Tasman Sea poles (Veevers et al. 1991). Note the close agreement between the predicted displacement direction, basement dip directions (Figure 4.2A) and interpreted "transfer fault" orientations in the Queensland Trough (Figure 4.11).
Normal, thick continental crust

Thinned continental crust

Normal en echelon fault

Ridge

Transform

**A.** Crustal Model from Mutter & Karner, 1980

*B. and C. from (Smith, 1993; Smith & Dumey, 1992)*
sediments. However, their Figure 4 suggests that the structures existed in pre-Silurian time, suggesting that these structures formed even earlier. If this is true, then southeastward movement of the Queensland Plateau and relatively more southeastward movement of the Marion Plateau is consistent with the fault geometries in the troughs. Indeed, it is difficult to accommodate the extension suggested by the structural interpretations from both troughs without opening a third arm. Given the geometric relationship between the Queensland and Townsville Troughs, a triple junction seems to be a plausible explanation for the creation of these troughs, but evidence for the existence of a plume or other mechanism for creation of the triple junction is still lacking.

The structural interpretation presented herein differs radically from previous structural interpretations of the northeast margin of Australia with regards to implications for large-scale tectonic evolution. The NW or WNW extension suggested by the structural interpretations of the Queensland Trough (herein) and Townsville Trough (Symonds et al. 1987, 1988) implies that extension at basement levels in this system was not kinematically consistent with either the spreading direction of the Coral Sea to the north or the Tasman sea to the south, as is commonly stated in other studies. Because we have no age constraints on the extension for the Queensland/Townsville Trough system, it is probable that extension here predates tectonism in both of the adjacent spreading centers. A (?pre-) Paleozoic triple junction, reactivated in the Mesozoic, is supported by the kinematics determined from the structural interpretation of the Queensland Trough presented above.
Figure 4.13 Regional map of the Great Australian Bight. Contours are kilometers to basement as mapped from magnetic data by Stagg et al. (1989) and Etheridge et al. (1989). Insets show locations of the Eyre Sub-basin, E = Figure 4.14 and the Ceduna Sub-basin, C = Figure 4.19 and P = the Polda Trough, discussed in text. Generalized continental geology is indicated by dotted boundaries (AGSO 1980) and includes the Archean to Proterozoic (A-P) Gawler Block, the late PreCambrian to Paleozoic (PreC-Pz) Adelaide and Kanmantoo Fold belts and the Cenozoic Eucla Basin.
Chapter 4 Australian Passive Margins

4.3 The Great Australian Bight

In this section, two rift basins from the Great Australian Bight (GAB) are evaluated. They are the Ceduna Sub-basin to the east and the Eyre Sub-basin on the west (Figure 4.13). The GAB traditionally has been thought of as part of a classical "Atlantic-type" passive margin between Australia and Antarctica (Sproll & Dietz 1969, Smith & Hallam 1970, Griffiths 1971, Boeuf & Doust 1975). That is, the margin underwent a simple evolution from Mesozoic rifting to Tertiary drifting or seafloor spreading. Apparent seaward dipping orthogonal extension structures and geophysical signatures in the Eyre Sub-basin led Etheridge et al. (1989) and Lister et al. (1991) to propose that the GAB portion of the southern Australian margin is a "lower plate" margin following detachment model concepts. The GAB margin represents a "successful" rifted margin without any marginal plateaux, although the continental shelf incorporates a large "magnetic quiet zone" which includes an "outer rise" seaward of both the Eyre and Ceduna depocenters. Recognition of the geometric complexity of structures along the margin with improved geophysical data sets has led to the postulation of several pulses of pre-drift deformation (Willcox & Stagg 1990) and oblique rifting (Stagg et al. 1989, Willcox 1990) along the margin, hypotheses this study endorses.

The following sections begin with a review of the data and interpretations from the Eyre Sub-basin. Dip analysis is applied to a subset of that data to test kinematic inferences made by previous authors based on interpreted fault geometries. An analysis of the tectonic framework of the Ceduna Sub-basin by interpretation of open-file seismic data is presented and used to integrate this portion of the GAB with the more clearly imaged western rift segments. Finally, a model for the evolution of the GAB, with particular emphasis on the earliest phases of deformation, is presented. Observed structures within the GAB and the conjugate margin in Antarctica are integrated to support the model.

4.3.1 The Eyre Sub-basin

The Eyre Sub-basin extends approximately 150 km westward from the Western Australia/South Australia border at 129°00'E (inset E, Figure 4.13) and is underlain by Proterozoic metamorphics and granite (Clark & Alley 1993). Although NE to ENE (~60°) half graben bounding faults have been mapped within the sub-basin (Bein & Taylor 1981 and Figure 4.14), the axis of the trough formed by the half graben actually runs nearly E-W. The trough aligns with the Paleozoic/Proterozoic Polda trough (P, Figure 4.13) on the eastern side of the GAB.
that also has an E-W axial trend. Interpretations of seismic data show the northern margins of both the Polda (Nelson et al. 1986) and Eyre troughs (Etheridge et al. 1989, Willcox & Stagg 1990) consist of a series of en echelon jogs. An en echelon basin bounding fault configuration is supported by interpreters of both gravity (Willcox 1990) and magnetic data (Stagg et al. 1989). However, the gravity data interpretations showing offsets in the bounding fault system of the Eyre trough (Willcox 1990) are not strictly en echelon, exhibiting a variable sense of offset (also see offset of bounding fault in Figure 4.15).

The structural interpretation of the Eyre trough at basement levels by Bein & Taylor (1981) is reproduced in Figure 4.14, wherein offsets in the major bounding fault are apparent. Etheridge et al. (1989; their Figure 7) suggest ESE trending transfer faults, but the orientation of the transfer structures becomes nearly N-S over the the Ceduna area. Modification of the Bein & Taylor (1981) interpretation by Willcox & Stagg (1990) suggests 110° trending transfer faults within the Eyre trough, but in their interpretation transfer faults trend 135°-145° over the Polda and Ceduna areas (their Figures 1, 6 and 7). Transfer faults at 140° are mapped along the entire margin, from the Naturaliste Plateau in the west to the Otway Basin in the east, on the basis of regional gravity data, but structural trends interpreted from these data are referred to as "tectonic fabric" and may vary considerably locally (Willcox 1990). In this section, an attempt is made to reconcile the apparent incompatibility of transfer fault trends along this presumed simple "Atlantic-type" margin and present evidence for a well constrained transfer fault in the Eyre trough.

A reinterpretation of a subset of the Eyre data done by G. Irwin of Marathon Petroleum is concentrated around the Jerboa and Wombat prospects (Figure 4.15). The interpretation supports a northeasterly (~060°) trend for the major bounding fault. The two structures (petroleum prospects) lie on either side of a major offset in the bounding fault, at both the basement and "top synrift" levels. The interpreted picks of the main bounding faults were confirmed and true dips computed on basement to try to discern the tectonic transport direction in the early

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**Figure 4.14** Depth to basement in the Eyre Sub-basin after Bein & Taylor (1981). Contour values in seconds of two-way-travel-time. Triangles on contours point into deeps. Heavy lines indicate major faults, with ticks indicating downthrown side. Gray dashed lines are proposed transfer fault orientations (110°) after Willcox & Stagg (1990). Solid grayed lines are proposed transfer fault trends from this study. Eastern transfer fault is imaged in seismic profiles (T) in Figures 4.17 and 4.18. Dashed rectangular inset is the location of Figures 4.15 and 4.16. Jerboa and Wombat drilled prospects are labelled. See Figure 4.13 for location.
EYRE SUB-BASIN
DEPTH-TO-ACOUSTIC BASEMENT ISOCHRONS
stages of rifting (Figure 4.16). The 1970's N-S and E-W reflection seismic grid has approximately 4 km spacing, with a few diagonal profiles shot over the Jerboa prospect. Data quality ranges from quite good to poor, but acoustic basement is imaged on most profiles.

In the basement interpretation, structural contours are at high angles to the faults, especially in the region of the Jerboa prospect (Figure 4.15b). At top synrift levels (Figure 4.15a) contours are much less continuous with saddles appearing to correspond to contours at high-angles to faults at basement levels. The dip analysis (Figure 4.16) demonstrates that hangingwall dip directions are 320°-325° in the Jerboa area. This dip orientation suggests that the tectonic transport direction in the western GAB was oriented along a SE trend (140°-145°), rather than the ESE (110°) trend of previous interpretations. The interpretation of basement dips in the EARS (Chapter 3) suggests that transfer, or rift transverse faults, are commonly within 15° of the tectonic transport direction. Interpretation of regional gravity data from the GAB (Willcox 1990) supports a more southeasterly (135°-140°), rather than the ESE (110°) trend originally suggested by Willcox & Stagg (1990). The non-orthogonal geometry that results from NW-SE trending transfer faults implies oblique slip on the main bounding fault. However, post-breakup thermal relaxation of the margin has most likely modified basement dips. Present day bathymetry suggest thermal subsidence contours are generally E-W, so that the presently recorded NW-SE dips would have been originally more NNW-SSE, or more orthogonal to the 060° trend of the main bounding faults.

Evidence of transverse structures in the Eyre Sub-basin is provided by jogs in the bounding fault, dip analysis and structural contours at high angles to the bounding fault. Figures 4.17 and 4.18 are seismic profiles over the area extending seaward from the eastern jog in the bounding fault between the Jerboa and Wombat structures. The NW-SE Line 119 (Figure 4.17a) is subparallel to dip azimuths computed at the profile intersections (Figure 4.16) and has slightly higher measured basement dips than the E-W Line 6 and N-S Line 19 profiles (Figures 4.17b and 4.18, respectively). The NE-SW Line 104 images nearly flat-lying basement (Figure 4.18a). Reflection offsets above the fault separating the two tilted blocks on Line 119 document continued normal-sense reactivation of this moderately dipping structure during synrift deposition, but reflections over the blocks are generally undisturbed. Identical dips on the two tilted basement blocks suggest this is a "dip" line in the traditional sense.

A near vertical disruption of synrift reflectors near the bounding fault on the N-S Line 19 (Figure 4.18, T), separates opposing dips within the synrift, but does
not appear to offset reflections. Although basement relationships are obscured by the expression of this structure higher in the section and diffractions off of the shoulder on the left hand side of the profile, this structure is interpreted as a transfer fault separating offset segments of the bounding fault. Just east of the tie with Line 119 on the E-W Line 6, the coherent westward dipping basement reflection is apparently normally offset across a steeply dipping structure to an elevated chaotic basement character (Figure 4.17b, T). The transition in basement coincides with a change in the synrift seismic character. The NE-SW Line 104, with subhorizontal basement dips, images a relatively elevated basement block to the left of T (Figure 4.18a).

Although the features on the N-S, E-W and NE-SW lines discussed above may not traditionally be correlated in seismic interpretation (e.g. Figures 4.14 and 4.15), the lineament created by correlation of the structures coincides with the high-angle contours in plan view. Furthermore, the lineament overlays the steepest gradient of the high-angle structure contours. The offset of the elevated basement to the left on Line 104 (Figure 4.18a) correlates with the western jog in the main bounding fault (Figure 4.15b) along the same azimuth (~155°) as the proposed transfer fault. Thus, evidence from seismic data support the interpretation that transverse structures in the Eyre basin do not trend ESE, but at ~155°. This trend is nearly, but not quite orthogonal to the bounding fault trend, consequently, it is concluded that recognizable, but fairly limited oblique slip has occurred on the main bounding faults and that tectonic transport of the blocks within the sub-basin was in a NW-SE to NNW-SSE direction. The previous inference that transfer faults trend 110° in the Eyre Sub-basin (Willcox & Stagg 1990) was most likely based on the recognition of a bounding fault segment of this trend in the Bein & Taylor (1981) interpretation (Figure 4.14). However, it is obvious that bounding fault
trends change from NE to E-W across this structure. It is suggested that this segment represents a reactivated pre-existing dislocation that segments the rift zone in a manner similar to the Rukwa and Chamaliro dislocations in the EARS. This analysis of the Eyre Sub-basin emphasizes the need to balance structural interpretations in the third dimension by incorporating both dip and fault offset features in seismic data before any inferences about the kinematics can be made. The consistency between careful three dimensional structural analysis and the dip domain data in this case study demonstrates the usefulness of this approach.
Figure 4.17  A) NW-SE Line 119 from the Eyre Sub-basin that trends parallel to the proposed transfer faults from this study (Figure 4.14). The two tilted basement blocks have near identical dips. B) Part of E-W Line 6 from the Eyre Sub-basin. Coherent westward dipping basement reflections are truncated and vertically offset at T, interpreted to be the continuation of the transfer fault on Lines 19 and 104 (Figure 4.18). Location of the intersections of the profiles and those shown in Figure 4.18 are indicated. See Figure 4.16 for locations of profiles.

Figure 4.18 (Next page) A) NE-SW Line 104 from the Eyre Sub-basin. The offset of basement at T is interpreted to be a transfer fault. The structure is correlated to the transfer fault (T) interpreted on Line 19 (below) and Line 6 (Figure 4.17) to explain the contours at high angles to the main bounding fault (Figure 4.15B). B) The northern end of N-S Line 19 from the Eyre Sub-basin images the bounding fault. Extreme disruption (T) of synrift reflectors, just south of the rift shoulder, is interpreted to be a transfer fault separating two offset segments of the ~060° bounding fault. Location of the intersections of these profiles and those shown in Figure 4.17 are indicated. See Figure 4.16 for location of profiles.
Eyre Sub-basin

5 kms

A) Line 119

B) Line 6
4.3.2 The Ceduna Sub-basin

The Ceduna Sub-basin is located at the eastern edge of the Great Australian Bight (GAB) from 129°00'E to 133°20'E and from 32°45'S to 36°00'S (inset C, Figure 4.13). The axis of the depocenter trends NW-SE (~130°) wherein depth to basement reaches 8-14 km (Figure 4.13) (Etheridge et al. 1989, Stagg et al. 1989). Approximately 17,000 km of seismic data, ranging in vintage from 1969 to 1982, were provided by Marathon Petroleum to analyze the tectonic history of the area. The Ceduna Sub-basin was considered likely to display oblique geometries because the axis of basin elongation is at a high angle to the eventual spreading ridge trend of approximately E-W.

The structural interpretation of the underlying tectonic framework (Figure 4.19) in the Ceduna Sub-Basin is hampered by the shallow penetration of most of the early vintage seismic data and limited seismic data that actually image synrift or basement structures. Deformation in the shallow section of the study area can be characterized by fault segments which trend generally NW-SE and NE-SW, but not necessarily orthogonally. Segments of the NW-SE fault set exhibit normal offset, but are highly variable in trend, ranging from 105° to 165°. The variability of the NW-SE trend seems to be divisible into three broad zones (from west to east: A, B, C, Figure 4.19). In area A, normal faults are nearly perpendicular to the NE-SW structures. To the southwest, in areas B and C, the normal faults become increasingly complex and have a variety of trends relative to the NE-SW structures, corresponding to an increase in wrench and flower structures apparent in the seismic data.

The NE-SW structures consistently trend around 060° and are characterized in the seismic data by zones of disrupted reflections, both with and without apparent vertical offset. Vertical offset is more common in the seismic data crossing the southwest extent of the structures. Besides direct seismic profile evidence for the NE-SW trending fault set, careful mapping of the hingeline (i.e. where shallow basement steepens or drops suddenly) identifies large offsets that coincide with sharp bends in contours at all seismic horizon levels in contrast to the apparent linearity of a basement scarp in this vicinity as determined by magnetic data (Figure 4.13) (Stagg et al. 1989). However, there is no consistent sense of offset, as in an en echelon relay system. There are apparently two "wavelengths" of hinge offset which correlate with the magnitude of lateral offset of the hingeline. Smaller offsets of 1 to 5 km appear to occur at a periodicity of 5 to 10 km. Larger offsets of 5 km or more occur at ~50 km intervals. The two scales of
hingeline offset may suggest different levels of detachment of normal faults. That is, the smaller scale relates to a shallow detachment (probably involved in Cretaceous deformation), whereas the larger scale "wavelength" may indicate the location of crustal to lithosphere scale pre-existing structures. In support of this hypothesis, the larger scale offsets do correspond to the location of the deeps recorded in the magnetic data further seaward and with the location of Tertiary magmatism.

The hinge is often expressed on seismic profiles as a simple monocline implying that if deep-seated dip slip faults control rotation of basement seaward, they must lie basinward and dip toward the continent in contrast to the Eyre trough's apparent seaward dipping basement faults. Where faulted, basement is often offset across a subvertical fault and the basement reflection in the hangingwall has a flat or seaward dip. Landward dipping bounding faults for these segments of rotated basement have been recorded, but appear to be rare. The seaward dipping faults affect the postrift section and occasionally sole out into the upper synrift section, suggesting that the predominance of seaward dipping faults is the result of later gravity faulting rather than early rift related faulting. However, in many other cases, seaward dips extend into the the deepest sections recorded (i.e. 6 seconds or ~ 8-12 km below sea level). Rollover into the faulted hinge is absent on available seismic data with one exception; a 160° trending profile which intersects a proposed junction of the two fault sets at the Potoroo well. The location of the well marks a major bend in the hingeline trend from NW-SE to nearly E-W (Figure 4.19).

Although the NW-SE fault segments appear to be more commonly truncated and offset by the NE-SW trend, there is also local evidence that the converse is true. Considering the study area from west to east, some general observations can be made. Western areas, between the Potoroo well and the major jog in the hingeline between zones A and B (Figure 4.19), appear to have been less intensely deformed even though a slightly denser seismic data grid was available. This may be a consequence of very poor penetration to synrift levels, but it is supported by the lack of major jogs in the well constrained hingeline.

Figure 4.19 Tectonic architecture of the Ceduna Sub-basin. Seismic stratigraphy is controlled by only one well, the Potoroo (P), which is perched above the hingeline (H). Regions seaward of A, B, & C exhibit a different style of deformation within the sub-basin whose boundaries coincide with major offsets in the hingeline (indicated by heavy arrows). Insets 1 and 2 are areas with tightly spaced data coverage and are enlarged in B & C on the following page. V = intense magmatic activity.
Chapter 4 Australian Passive Margins

The synrift is best imaged in a small area in deep water at about 131°15'E and 34°29'S (inset 1, Figure 4.19b). NW-SE seaward dipping faults, trending between 120°-150°, characterize the deformation in this area. The NW-SE striking structures are truncated and offset by apparently through-going ~060° trending structures. There is vertical offset recorded on the NE-SW trending faults, but it is commonly less developed than on the ~060° fault segments. The NE-SW structures are often imaged in seismic profiles as near vertical, discontinuous and chaotic zones, in contrast to the generally more distinct, lower angle NW-SE trending faults. In the rare instances where measurements of rollover, rotation and fault dip in more than one direction are possible, true dip direction of the synrift section suggests that there is significant oblique slip on both fault sets. Because there is no clear acoustic basement imaged in the reflection seismic data in most cases, a comprehensive dip analysis was not done.

Landward dipping faults create a NW-SE trending trough (T, inset 1, Figure 4.19b) that is dextrally offset by an 060° structure. These two trough segments and two well-developed 120° normal faults located just landward are offset in the same sense as the hingeline segments (H, Figure 4.19a) that would be within the same compartments if the NE-SW structures are extrapolated to the NE. Reflection disruption or offset on near vertical planes on intervening seismic profiles support the interpretation of a continuous feature.

In the southwest, the best seismic coverage, both with regards to the density of the grid and modern vintage data, covers a rectangular area that is elongate along a NW-SE axis (inset 2, Figure 4.19c). Here, a zone of complex wrench and flower structures can be mapped at postrift horizons. These structures frequently obscure direct observation of synrift or basement structures. Offsets in the hingeline, where constrained, can be projected into the wrench system along the regional 060°-065° trend to connect with NE trending faults that apparently compartmentalize the flower structures. A major (>300 ms throw) seaward dipping fault defines the SW edge of the wrench system within at least two of the compartments. This structure also changes trend and is offset by the NE trending structures, but seismic data seaward of this fault show that the NE trending structures do not extend far beyond it. The sense of lateral offset on the NE trending structures within zones B & C (Figure 4.19) at the hinge appears to have the same sense of lateral motion as the offset of the main normal faults, but appears to have the opposite sense between the regional zones (i.e. A, B and C, Figure 4.19).
A schematic representation of the interplay of various fault sets mapped in inset 2 (Figure 4.19c) is shown in Figure 4.20. Flower structures appear to lie within discrete fault blocks, although rare structures associated with them sometimes appear to cross the NE trending structures. The westernmost flower structure (inset 2, Figure 4.19c) lies within a grossly rhomb-shaped block between two major NW-SE trending normal to oblique slip faults and adjacent to a NE trending structure which coincides with a large (>10 km) offset in the hingeline. In the next compartment to the SE, a flower structure has been mapped with a NE-SW axis and lies within a roughly triangular shaped block. This flower appears to be less complex and its axis aligns with the regional northeasterly trend of the hinge offsetting structures.

Further to the southwest, structures become very complex and seismic data in this area is characterized by large chaotic and transparent zones. A large anticlinal (positive) flower structure manifest in the postrift section attests to late-stage wrench deformation (Figure 4.19c). Deeper structures are further obscured by significant ?Tertiary igneous activity. Seismic profiles document widespread extrusive and intrusive igneous activity affecting the southern end of the largest of the flower structures (V, Figure 4.19a), which corresponds to a large offset in the hinge along an 060° projection.

The interpretation of NE trending transfers implies extension essentially orthogonal to published interpretations of the Eyre Sub-Basin to the west (e.g. Etheridge et al. 1989, Willcox & Stagg 1990). However, the inferred NE-SW extension suggested by the fault geometries in Figure 4.19 is broadly consistent with recent interpretations of the extension within the Otway, Sorell, Bass, and Gippsland basins to the east. The "transfer fault" trends mapped in the Ceduna Sub-basin (060°) diverge approximately 30° from published trends (~030°) of transfer faults of the eastern sub-basins (e.g. Etheridge et al. 1984, 1985, Willcox 1990).

The timing of extension of the eastern versus western sub-basins along the Australian southern margin is relatively well constrained. The Jerboa-1 well

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**Figure 4.20** Schematic block diagram of the deformation in the Ceduna Sub-basin. A major lower crustal shear zone has focussed later deformations. The first phase included significant strike slip movement and shearing along the zone, creating the deep basins recognized in magnetic data (Figure 4.13), probably as isolated pull-aparts. Later Cretaceous extension created through-going NE-SW transfer faults which terminate at the seaward boundary of the lower crustal shear zone. Extensive ?Tertiary igneous activity is focused at the intersections of the transfer faults and the shear zone.
Ceduna Sub-basin
SCHEMATIC REACTIVATION RELATIONSHIPS

Lower Crustal shear zone
intersects Middle to Late Jurassic synrift sediments in the Eyre Sub-basin in the west, while numerous wells in the eastern Bass, Gippsland and Otway basins suggest an Early to Middle Cretaceous age for basin-forming structures (Falvey et al. 1990). The timing of deformation in the Ceduna Sub-basin mapped in Figure 4.19 relies solely on seismic stratigraphy projected from the only well in the study area. The Potoroo-1 sits atop the hingeline and so has only inferred ties to the deeper sections seaward of the hingeline. On the basis of geometry and without unambiguous stratigraphic information, the extensional structures observed in the upper section of the Ceduna Sub-basin appear to be Cretaceous in age. However, none of the structures observed appear to account for the great depth to basement as determined from magnetic and sonobuoy refraction results, which range from 8 to 14 km (Figure 4.13) (Willcox & Stagg 1990, Stagg et al. 1989, Etheridge et al. 1989).

Published interpretations of basins along the GAB margin invoke orthogonal extension geometries with a variety of preferred orientations of extension (e.g. Etheridge et al. 1984, 1985, 1989, Moore & Eittreim 1987). Willcox and Stagg (1990) recognize "oblique extension" along the margin, but by this they mean orthogonal extension along a NW-SE trend which is oblique to the eventual spreading direction of approximately N-S. Under this scenario, fault patterns are still classified into primary dip slip and strike slip categories, which align generally perpendicular and parallel, respectively, to the initial rift extension direction.

The implication is that extension was directed NW-SE in the earlier western phase and nearly orthogonal to that trend in the later eastern phase. Willcox & Stagg (1990) have recognized the inconsistency of the implied kinematics from either end of this margin and invoke an "overprinting" of deformation phases. However, they give no rationale for the changing extension directions, nor the change to N-S drifting. Attention is called to the fact that all interpretations, regardless of the vintage of data, the author's model bias, or inferred direction of extension, recognize NW-SE and NE-SW fault trends rather than N-S and E-W trends along most of the Southern Australian Margin which might be expected from the "steady state" spreading geometries in the Southern Ocean.

It is possible that a regionally consistent extension direction has operated throughout the pre-drift extension of Australia and Antarctica. A N-S directed extension producing the NE-SW and NW-SE fault pairs that we now observe in basins of a variety of ages along the southern margin of Australia would suggest that virtually all of the NW-SE and/or NE-SW "normal" faults contain some element of oblique slip. The analyses herein provide some evidence for minor oblique slip in the Eyre and nearly strike slip deformation in the Ceduna Sub-basin.
during a Jurassic deformation event. On-going work and new data in the Bass Strait basins apparently indicate that Cretaceous basin-forming structures there also contain oblique slip movements (J.B. Willcox, pers. comm.). Even so, the evidence is compelling that the later and temporally distinct Cretaceous extension was along a primarily northeasterly azimuth, whereas the earlier Jurassic extension in the west had a northwesterly azimuth. If a consistent N-S directed force has instigated extension along the entire margin through a period spanning some 100 m.y., why are the basins asynchronous rather than time transgressive, and why do they show a geometric tendency to nearly orthogonal extension directions during different phases of deformation?

The termination of the NE structures in the Ceduna Sub-basin at or near the most seaward normal faults suggests that this line defines the western limit of Cretaceous NE-SW continental extension and the eastern limit of Jurassic NW-SE oriented continental extension (with the possible exception of the Polda Trough). The zone also served to focus Tertiary igneous activity at its intersections with Cretaceous transfer faults. The great depth to basement found in the Ceduna Sub-basin argues for either 1) rapid and voluminous sedimentation if deep-seated basin-forming structures are Cretaceous in age or 2) basin forming structures are older (i.e. pre-Cretaceous) than those observed and were subsequently re-activated during renewed tectonic activity in the Cretaceous.

Could the axis of the Ceduna Sub-basin represent a long-lived lithospheric shear zone, involved in the earliest phase of extension between Australia and Antarctica? Did it act as a intracontinental transform fault even prior to the initiation of continental fragmentation, in the sense of the Lake Rukwa rift described in Section 3.3? Is there evidence that it is a tectonic boundary that was repeatedly reactivated? Did this zone act as a "strain guide" for the earliest phases of extension, influencing NW-SE directed extension in the lithospheric block containing the Eyre Sub-basin? The deeps along its trend are similar to the deeps that punctuate the Rukwa rift zone and it's trend is aligned with the tectonic transport direction established for the Eyre Sub-basin, establishing a similar regional configuration as that described in the EARS. The idea that the axis of the Ceduna Sub-basin is a tectonic boundary receives additional support from Willcox & Stagg (1990) who describe a "deep-seated transfer fault that defines the southwest margin of Ceduna Terrace" (p.274). In the next section, dealing with some aspects of continental separation of Australia and Antarctica and the configuration of Gondwanaland, evidence is presented in support the hypothesis that the Ceduna lineament acted as an intracontinental shear zone or transform.
Chapter 4 Australian Passive Margins

4.3.3 The Separation of Australia and Antarctica


All of the published reconstructions seem to show significant overlap of continental crust in the Ceduna Sub-basin region, with the exception of Konig’s (1987) reconstruction. In Konig’s (1987) reconstruction, the overlap is avoided by using the seaward edge of the magnetic quiet zone as the continent/ocean boundary (COB), rather than the 2000 m isobath, as the location of the pre-drift suture. Konig’s (1987) reconstruction produces a pre-drift rift zone 250 km wide, positioned symmetrically about the eventual rupture, but the configuration and the kinematics of the continental rifting are not addressed, Paleozoic terranes do not align, and no prerift configuration is presented. The reconstruction also implies the presence of symmetric belts of extended continental crust on either margin, which is not supported by current understanding of the margins' configurations deduced from geophysical data.

4.3.3.1 Correlation of Continental Geology of Australia and Antarctica

In this and following sections, I attempt to establish the age and scale of the lineament that corresponds to the axial trend of the Ceduna Sub-basin. A brief review of the continental geology of the Australian (Figure 4.13) and Antarctic (Figure 4.21) margins highlights some of the controversies surrounding the reconstructions. Both continents are characterized by Archean to Proterozoic
cratons in the west, with either Mesozoic and/or Cenozoic accreted and intrusive terranes in the east, joined by a series of Precambrian and Paleozoic fold belts. On the Australian margin, the Cenozoic Eucla basin overlies the Archean-Proterozoic blocks on either side, obscuring basement relationships and limiting insight into how structures in the GAB are related to pre-existing structure. Magnetic (Stagg et al. 1989) and gravity data (Willcox 1990) indicate that the submerged southwest edge of the Gawler block may coincide with the trend of the Ceduna Sub-basin (130°-140°) and the onshore boundary between the Yilgarn and Albany-Fraser provinces trends generally NE, as do the bounding faults in the Eyre Sub-basin.

The tripartite crustal configuration of Antarctica was recognized early on (Adie 1964, Voronov 1964) and Hamilton (1964) proposed that West Antarctica is made up of accreted terranes (Figure 4.21). The continent comprises the Archean and Proterozoic cratonic crust of East Antarctica, the late PreCambrian and Paleozoic fold belts of the Transantarctic Mountains, which span the entire continent from the extensional Ross and Weddell Seas on the east, and the Mesozoic and Cenozoic "accreted terranes" of West Antarctica. Geochronological work ties Archean rocks from the Adelie Coast (~144°E) of Antarctica with rocks of the Gawler Block in Australia (V. Bennett and M. Fanning, pers. comm.).

The gross similarities of older cratonic elements with juxtaposed Paleozoic and Mesozoic belts in Antarctica and Australia led early workers to propose a variety of terrane correlations. In particular, speculation that the Ross and Adelaide Orogens, corresponding to the Transantarctic Mountains (Northern

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Figure 4.21 Schematic geological map of Antarctica. West Antarctica is composed of Mesozoic and Cenozoic (Mz-Cz) accreted terranes. The Transantarctic Mountains consist of Late PreCambrian to Paleozoic (PreC-Pz) fold belts intruded by Jurassic dolerite dikes indicating extension. Asterisk (*) at Horn's Bluff (HB) at approximately 150°E indicates the known "eastern" extent of these dikes along the continental margin. East Antarctica comprises Archean to Proterozoic (A-P) cratonic rocks. NVL = Northern Victoria Land which is divided into terranes by major oblique slip faults, the Lanterman (L) and Rennick (R) faults and separated from East Antarctica by the Exiles Thrust (EX). BI = Balleny Islands. Striped and cross-hatched patterns indicate gravity gradient trends after Gahagan et al. (1988) that can be separated into three zones between 156°E and 96°E. Note the sharp bend and oblique trend of gravity gradients south of the George V Fault Zone (GVFZ). WB = Wilkes Basin. A major scarp (dashed line crossing 132° E) bounds the Wilkes Basin. Extension of this trend connects with a polarity reversal in the gravity gradients offshore to oceanic crust, where it connects with the transform fault which defines the eastern boundary of the Australian-Antarctic Discordance (AAD, Figure 4.22). Inferred extension direction (double arrows) in the ice-covered Ross and Weddell Seas aligns perpendicular to gravity gradients (lines). A bend in the gravity gradients under the Ross Sea is in the same sense as the change seen at the south end of the George V Fault Zone.
Victoria Land) and the Adelaide Fold belt, respectively, were at one time continuous. This led to geological (Craddock 1972, 1982, Oliver 1972) and chemical and mineralogical comparisons (Ravich 1982), which have stood up under more recent analyses. Finally, Quilty (1986) documents the similarity of non-marine micorfloras between the Wilkes Basin of Antarctica (Figure 4.21) and the Otway basin of Australia, although significant differences in the marine microfloras leave a depositional connection with it equivocal.

4.3.3.2 Mesozoic deformation in Antarctica

Paleomagnetic evidence suggests that East and West Antarctica were at one stage two separate plates, whose sutured boundary lies within the Transantarctic Mountains (Scharnberger & Scharon 1972), but the existence of two separate plates is still being debated (Schmidt & Rowley 1986). The field and paleomagnetic data, along with difficulties in aligning the Paleozoic terranes, including those in Tasmania, led to a lively debate about whether there were large transcurrent movements after the deposition of the Paleozoic belts, as well as, whether the Paleozoic and Mesozoic terranes were emplaced by strike slip or convergent mechanisms (Dietz et al. 1972, Crawford & Campbell 1973, Daily et al. 1973, Harrington et al. 1973, Cooper 1975, Laird et al. 1977, Craddock 1982, Grikurov 1982, Findlay 1983, Grindley & Oliver 1983, Weaver et al. 1984, Gibson & Wright 1985, Schmidt & Rowley 1986, Kleinschmidt & Tessensohn 1987). Modern geochemical studies and continuing field studies have refined correlations of various terranes but have not yet resolved the controversy (Stump et al. 1983, 1986, Borg et al. 1987, Borg & Stump 1987, Borg & DePaolo 1990, 1991, Wilson 1990, Storey 1991, Bradshaw 1991, Foster & Gleadow 1992, 1993). However, a thorough review of the evidence in the literature suggests that it is likely that transcurrent or oblique motions have played a significant role in both convergent and divergent phases of tectonism in Antarctica. Paleomagnetic and other evidence from the Weddell Sea, at the other end of the Transantarctic Mountains, support this view (e.g. Grunow et al. 1987, Garrett 1991, Kristoffersen & Hinz 1991).

Although the vast majority of the Antarctic continental geology is obscured by a permanent ice cap, good exposures are found along the Transantarctic Mountains and Northern Victoria Land (Figure 4.21)—the area juxtaposed to the Ceduna Sub-basin in most reconstructions. Flottmann et al. (1993) provides convincing kinematic field evidence for the correlation of the Wilson terrane, between the Rennick Fault and the Archean craton of East Antarctica, with the Adelaide Fold belt. He speculates that the Exiles Thrust is the boundary between the ice covered Archean craton and the Wilson terrane, implying a deep seated
suture between the two terranes (Figure 4.20). In support of this hypothesis, aeromagnetic data show a change in anomaly pattern, interpreted to reflect the change from "platform" to "shield", running along a NW-SE line at 73°S and ~158°E (Roland 1991), which appears to be continuous with a similar anomaly pattern observed offshore by early Russian geophysical surveys (Ushakov 1960). More recently, offshore marine surveys reported a typical "transform margin type morphology from 147°E to 144°E" and although glacial scouring is evident, scouring has apparently been "focused along major structural features" also aligned NW-SE (Domack & Anderson 1983).

Field studies document oblique slip on most of the major faults (e.g. the Lanterman, Leap Year and Rennick faults) in Northern Victoria Land (Gair 1964, Bradshaw et al. 1982, Katz 1982) that all strike ~130°, subparallel to the trend of the Ceduna Sub-basin. Jurassic subsidence of the Rennick Graben has been documented by field studies (Grindley & Oliver 1983). Occurrence of Jurassic dolerite dikes in Northern Victoria Land indicate early Mesozoic extension in this region and offset of some of these dikes by the main faults provide evidence for reactivation of the main structures in post-Jurassic times (Bradshaw et al. 1982). The above discussion highlights the repeated activity on the main faults in Northern Victoria Land spanning the entire Mesozoic.

The presence of Jurassic dolerite dikes at Horn's Bluff (Figure 4.21) argues that continental crust at this longitude was involved in Mesozoic extension. A series of "horst and graben features", offsetting interpreted Paleozoic section, to the west of the Transantarctic Mountains in the ice-covered Wilkes Land (Figure 4.21) was revealed by seismic refraction data (Steed 1983) and magnetic data (Kadmina & Kurinin 1983) ending in a scarp at ~130°E. The scarp projects along strike into the offshore boundary between changing gravity gradients (Gahagan et al. 1988) and changing character of the magnetic quiet zone (Royer & Sandwell 1989, Veevers 1987), and finally connects with the transform fault which defines the eastern boundary of the Australian-Antarctic Discordance (AAD, Figure 4.22).
4.3.3.3 Deep structure beneath the Transantarctic Mountains and the Antarctic margin

Geophysical studies indicate a large gravity gradient at the eastern edge of the Transantarctic Mountains (Groushinsky & Sazhina 1982a, 1982b, Kadmina & Kurinin 1983). Seismic evidence does not support an explanation of the anomaly by a large glacio-isostatic imbalance or an abnormally thick crust, implying that the gravity anomaly stems from a mantle source (Bentley 1983). Trace element and isotopic comparative studies of intrusives in Marie Byrd Land in West Antarctic and the Transantarctic Mountains suggest that differences in geochemistry stem from a foundering slab that has acted as an "effective barrier within the mantle" between the two areas (Brewer & Clarkson 1991). Apatite fission track data, demonstrating that uplift of the Transantarctic Mountains is coeval with extension in the adjacent Ross Sea, led Fitzgerald et al. (1986) to present a model of asymmetric extension in the Ross Sea along a penetrative detachment and to suggest that "the localization and asymmetry of this detachment and its unusually deep level expression are attributed to a profound crustal anisotropy inherited from an early Paleozoic collision along the present site of the mountain range."

The offshore geology of Oates and Wilkes Lands, Antarctica (~160°E-130°E) is known largely from a 1984 USGS geophysical cruise of the S. P. Lee (Eittreim & Smith 1987). Deep reflection seismic data, with approximately 100 km spacing between profiles, were used to describe the seismic stratigraphy, to pinpoint the continent/ocean boundary (COB) and to describe the geometry of the Moho as it crossed the COB (Eittreim & Smith 1987). The Moho is described as "rough and undulant" under the oceanic crust and generally smoother, with local parallel/laminated reflections, broken by offsets that "may be faults or shear zones" under continental crust (Eittreim & Smith 1987). Crustal models derived from sonobuoy, gravity and magnetic data indicate "extreme thinning for approximately 35 km landward of the COB, underlain by a body with density intermediate between crust and mantle rock", which may be the result of underplating (Childs & Stagg 1987).

The release of satellite altimetry data provides a major contribution to understanding the earliest phases of deformation along this boundary (Gahagan et al. 1988, Royer & Sandwell 1989). Gravity gradients clearly delineate transform faults in the Southern Ocean for the first time (Figure 4.22). A striking feature is the sharp bend in the trend of gravity gradients at the southern end of the Spencer and George V Fault Zones, in contrast to the smoothly curving Tasman and
anomalous mantle exists, and has existed since at least the Jurassic, on both sides of the Transantarctic Mountains.

4.3.3.4 Break-up and the Connection of the Ceduna Sub-basin to Antarctica

Seafloor magnetic anomalies (Cande & Mutter 1982, Weisel & Hayes 1972) indicate a simple N-S drift pattern for the Southern Ocean. Subsidence history studies of basins along the Australian margin suggest a time transgressive (west to east) opening for the basin (Mutter et al. 1985), although synchronous ridge-jumps, leaving portions of newly created ocean floor of one plate stranded on the other, have also been proposed to explain these data (Veevers 1987, Veevers et al. 1991). The oldest magnetic lineations from the oceanic crust between the two continents were first modelled at A22, or early Eocene (55 Ma) (Weisel & Hayes 1972). Cande & Mutter (1982) revised the interpretation of the oldest anomaly to A34 (mid-Cretaceous) and suggested a slow initial spreading phase that changed to a fast spreading phase in the Eocene. The A34 interpretation was subsequently confirmed and refined by Veevers (1986) to a 95 ± 5 Ma age of continental rupture.

The COB has been defined along the Australian side by seismic and magnetic data (Veevers et al. 1990). The COB trends ENE from western Australia to ~130°E, where it turns to the SE and jogs back to the NE (Figure 4.22). East of ~135°E spacing of control data is approximately 100 km so the details of the COB location are not well constrained where it trends generally SE (Veevers et al. 1990). At ~147°E the COB bends sharply to the south where it comes on strike with the Tasman FZ. The long wavelength gravity gradients (Gahagan et al. 1988) are interpreted to overly continental crust in the Ceduna Sub-basin and eastward. They are subparallel to the COB where it trends SE, but appear to cross to the seaward side of it locally. Further to the east (~142°E), the broad gravity gradients are seaward of, and trend at an oblique angle to, both the SW and S trending COB. The character of the gravity gradients is similar in character to those south of the George V FZ and this signature is interpreted herein to overlie continental crust in both areas, requiring a shift in published locations of the COB.

Closing the Southern Ocean along the Spencer and George V Fault Zones to the extent of clearly defined N-S gravity gradients (Gahagan et al. 1988) results in a reconstruction at the A13 (35.5 Ma) isochron (Figure 4.23a). In this reconstruction the alignment of the Ceduna Sub-basin with the westernmost gravity gradient from the Antarctic side is nearly perfect and there is little overlap. Although Veevers et al. (1991) push the A34 anomaly east to the Tasman FZ in schematic reconstructions, the companion paper (Veevers & Li 1991) dealing with the
magnetic anomaly data extends it only to ~130°E. On the basis of the interpretation of the gravity gradients on either end of the George V FZ representing deformed continental crust, it is suggested that the anomalies older than A13 do not extend eastward past the Spencer FZ (Figure 4.22). Instead, the long, linear belt defined by the gravity gradients at A13 time probably represents a shear zone or transform boundary, much like the Rukwa lineament, within continental crust. A continental connection at this late stage is consistent with the large time gap between continental extension and the initiation of rapid seafloor spreading at A13 time. The hypothesis of an A13 time continental connection eliminates continental overlaps in the Ceduna area. It also eliminates the need for the change in spreading direction at A20 (45Ma), which creates apparent loss of oceanic crust shown in the schematic reconstructions of Veevers et al. (1991). The Ceduna lineament's late stage continental connection (A13, 35.5Ma) which appears to have existed for approximately 60 million years after initiation of seafloor spreading in the west (A34, 95.5 Ma), indicates that it played a fundamental role in inhibiting continental rupture to the east of it, and suggests an intimate coupling of this feature with the processes responsible for break-up.

Magnetic data indicate a truncation of isochrons A34-A20 at approximately 132°30'E and a major bend in the magnetic trough at about this same location (Konig 1987). Recognition of the COB and the magnetic trough in the magnetic data is tentative to the east of the same longitude and are not discernible eastward of 135°E. It is notable that the magnetic trough passes through the Ceduna Sub-basin rather than along its continentward edge, unlike other GAB basins, which are located between the magnetic trough and the COB (Veevers 1987). Several lines of evidence, discussed above, suggest that the Ceduna Sub-basin and the lineament defined by its extension to similar features on the Antarctic margin mark a major tectonic boundary during break-up.

The connection of the Ceduna Sub-basin with Jurassic deformation in Northern Victoria Land, and the enigmatic depth to basement indicated by magnetic data (Figure 4.13), suggest that deformation occurred here during the earliest phases of continental break-up. The reconstructions presented in Figure 4.23 imply that there was still a continental crust connection between the Ceduna Sub-basin and Northern Victoria Land at A13 time. Several comparable features of the Ceduna and Rukwa lineaments suggest an analogy between their roles as tectonic boundaries during intracontinental deformation:

- the lineaments are of comparable scale (i.e. transcontinental);
the lineaments are long-lived;

- the lineaments are punctuated by deeps that appear to be the loci for repeated reactivation during temporally separated deformation phases;

- the lineaments align with the regional tectonic transport direction determined from adjacent sub-basins;

- the lineaments act as a boundary for the transmission of strain through the continental lithosphere.
Figure 4.24 Schematic tectonic architecture of the Northwest Shelf, Australia (bottom). Inset defines limits of data coverage used in this study. The main structural elements identified from interpretation of the regional seismic data grid and the location of a deep reflection seismic line (Line 3185, Figure 4.29) are displayed in the blowup of the inset (upper right). Normal faults on the western edge of the Exmouth Plateau (E) indicate a different extension azimuth than faults mapped in the Enderby Half graben (EN). Measured dips on rotated basement from the Exmouth Plateau (rose diagram, upper left) suggest that tectonic transport of the deformed continental blocks was nearly orthogonal to the mapped fault trends, but apparently oblique to the resulting spreading azimuth as indicated by the trend of magnetic isochrons. A set of ESE crustal shear zones appear to segment and change strikes of structures of a wide range of ages.
Chapter 4

Australian Passive Margins

4.4 The Northwest Shelf of Australia

4.4.1 Introduction

The preceding interpretations of passive margins are limited by widely spaced and early vintage seismic data and the lack of well data to constrain the seismic stratigraphy and to precisely date deformational events. In contrast, extensive hydrocarbon exploration and production on the Northwest Shelf has resulted in the drilling of numerous wells and extensive seismic data coverage. An opportunity to evaluate some of the data from this passive margin was provided by an invitation to participate in a study group, funded by Marathon, whose brief was to identify structural characteristics common to individual phases within an overall tectonic evolution (Marathon 1991). In particular, we attempted to identify locations with good data coverage, where the structural relationship between individual phases of deformation might be determined by a more in depth, local study to be undertaken in later phases of the project. Some of the results of the evaluation of the Carnarvon basin, comprising the southern part of the shelf, are presented below.

Approximately 8600 km of seismic data were provided by Marathon and Curtin University from the southern portion of the Northwest Shelf. The project data provided a regional grid that included 18 NW-SE trending ("dip") composite profiles and 11 NE-SW approximately shore parallel ("strike") composite profiles at 20 to 50 km spacing and extending to the seaward edge of the Exmouth Plateau (Figure 4.24). The data were obtained from at least 20 different surveys and were consequently of varying quality, ranging from good to very poor. The study area includes the Barrow-Dampier Sub-basin, which comprises several rift zones or half graben and troughs bounded seaward by the Exmouth Plateau. The configuration of tectonic elements is similar to the gross crustal setting on the NE margin (i.e. the Queensland Trough and Plateau), except that a large region of antiforms and synforms is also incorporated in the terrane inboard of the plateau (Figure 4.24). The morphotectonic elements of the Northwest Shelf have been compared previously to the EARS (Veevers & Cotterill 1976, Veevers & Cotterill 1978).

In addition to the regional grid, 1500 km of seismic data from the Enderby Half graben, a series of tilted extensional blocks within the Barrow-Dampier Sub-basin (Figures 4.25, 4.26 & 4.27), with a grid spacing of 1-2 km, were provided by Marathon and analyzed with the aim to precisely constrain fault geometries therein. The following sections summarize the results of my interpretation of the
regional data grid (Marathon 1991) and describe the results of my structural interpretation of the extensional Enderby Half graben to establish similarities in the morphotectonic elements of this passive margin and continental deformation revealed in the other case studies from both the passive margin and intracontinental settings.

Several tectonic and paleogeographic histories have been developed for this passive margin (e.g. Barber 1982, Bradshaw et al. 1988, Etheridge et al. 1991, Veevers 1988). At least four main stages of deformation have been previously recognized. The main phases of deformation proposed by Etheridge et al. 1991 are:

- (Early) Permian crustal extension producing NE normal faults and NW transfer faults;
- Late Triassic to Early Jurassic left-lateral strike slip or transtensional reactivation of the Permian normal faults associated with a regional N-S compression;
- Middle to Late Cretaceous left-lateral wrenching; and
- Miocene (right-lateral) wrenching in response to the collision of Australia with Timor to the north (Stagg & Brassil 1991).

My interpretation of the regional grid confirms the polyphase deformation of the Northwest Shelf. The results of this study differ from the above in regards to particular structural trends and their kinematic implications (Marathon 1991). For instance, a deep pre-Mesozoic (Permian) structural lineation trending 026° has been interpreted on the northwest corner and the eastern boundary of the Exmouth Plateau (Figure 4.24). These NNE faults may be the "NE normal faults" proposed by Etheridge et al. (1991). An ESE (~100°) set of structures or lineaments, which appear to span the entire shelf, have been recognized in this study which have not been proposed or identified previously. They appear to be associated with the pre-Mesozoic NNE trends both in the NW corner of and inboard of the Exmouth Plateau and also with pre-rupture Cretaceous normal faults on the western margin of the Exmouth Plateau (Figure 4.24).

Timing constraints on mid- to late Mesozoic deformation are fair to very good, but the earlier periods of deformation are not well constrained by the regional data and rarely penetrated by wells. Likewise, later Tertiary event structures are evident in the regional data, but are concentrated in sub-basins to the north and south of the study area and are better constrained there. Thus, the following section concentrates primarily on the well constrained basin forming
structures of the extensional Enderby Half graben and the sub-basin's relation to the tectonic evolution of the margin as a whole.

4.4.2 Regional structural patterns of the Northwest Shelf

Recent aeromagnetic data have been used to identify magnetic isochrons in the Argo, Gascoyne and Cuvier Abyssal plains seaward of the Exmouth Plateau (Figure 4.24) (Fullerton et al. 1989). A complete set of isochrons, trending 070°, ranging from M26 (158Ma) to M16 (142Ma) have been identified in the Argo Abyssal Plain and a 150° spreading azimuth has been inferred (Veevers & Li 1991). Anomalies M10 (130Ma) to M0 (118Ma) are identified, trending about 030°, in the Gascoyne and Cuvier Abyssal Plains and a 120° spreading azimuth inferred (op. cit.). Additionally, a 10° clockwise rotation of the spreading azimuth has been proposed to explain the change of isochron trends at M4 (~125Ma) in the Cuvier Abyssal Plain (Fullerton et al. 1989). The relationships recorded in the magnetic data indicate at least two distinct periods of Mesozoic extension characterized by NNW-SSE Late Jurassic (M26: Oxfordian) extension, followed by WNW-ESE extension in the Early Cretaceous (M10: Mid-Neocomian) swinging more northwesterly in the Late Neocomian (M4: Barremian).

The Mid-Neocomian extension preceding seafloor spreading along the western margin of the Exmouth Plateau, was accommodated by N to NNE normal faults which can be mapped from the seaward edge of the plateau eastward to 114° E (Figure 4.24). The faults terminate against lineaments (?transfer faults), identified from seismic data, which trend WNW-ESE (~100°). Dips of tilted basement blocks on the seaward edge of the Exmouth Plateau trend essentially E-W (rose diagram, Figure 4.24), indicating that pre-breakup extension here was nearly orthogonal to the northerly bounding normal faults, but oblique (by ~30°) to the eventual spreading azimuth (120°) cited above. The 100° lineaments are well constrained at seismic profile intersections on the seaward edge of the Exmouth Plateau, where they segment the normal faults, but are less defined across the plateau, where their offset relationship to other structures is sometimes ambiguous.

There is no reason to expect transfer faults associated with extension on the seaward side of the Exmouth Plateau to extend through the plateau to inboard areas, but projections towards the continent of these structures along strike correspond to major changes or deflections of axial trends of features such as the Kangaroo Syncline and Rankin Platform (Figure 4.24). Many hydrocarbon producing wells in the study area lie along projections of these lineaments. The
lineaments are often spatially associated with late stage (?Tertiary) igneous activity (e.g. the Enderby Well, Appendix I). Bouguer gravity anomalies on the southern Northwest Shelf clearly indicate lineaments of an ESE trend (100°-120°), extending at least to the present continental shoreline (Stagg & Brassil 1991). The boundary between the Hamersley Basin and the northern Pilbara block, essentially defined by the 500 km long Fortescue River, also runs along a WNW-ESE trend. Finally, recent mapping in the Pilbara Craton documents large-scale deep-seated ESE and NNE lineaments that have controlled the location, size and shape of sedimentary basins and focussed igneous activity from at least the late Archean by being repeatedly reactivated (Krapez et al. 1990). The combined evidence suggests that the ESE lineaments represent a deep-seated, regional set of crustal-scale shear or fault zones, which were repeatedly involved in subsequent deformation events.

4.4.2.1 Structure of the Enderby Half graben

The Enderby Half graben is part of a series of extensional structures which form a rift zone that is perched on the continental shelf of the Barrow-Dampier Sub-basin. Intervening synforms and antiforms separate these rift zones from the Exmouth Plateau and extensional structures on its western edge (Figure 4.24). The structure of the Enderby Half graben is characterized by a shore-parallel (042°-047°), SE dipping bounding fault and synthetic intra-basinal faults defining two and in some places three, tilted fault blocks (Figure 4.25). The normal faults are clearly listric and sole out at shallow depths (~3.0 s TWT, D, Figure 4.25) even though "rollover" is rare on the apparent base synrift reflection. Although there are coherent reflections below a prominent reflection labelled BES in Figures 4.25 and 4.26, this horizon is referred to as the "base Enderby synrift" in the following discussion.

The dipping, continuous reflections of the Enderby synrift package have been penetrated by several wells which provide the following age constraints (Figure 4.25) (Marathon 1991). Sediments overlying and underlying the angular
unconformity defining the top of the synrift (TS) have been dated as Lower Cretaceous and Upper Jurassic, respectively. Prominent reflections within the synrift package are dated as unconformities separating Lower Jurassic/Upper Triassic and Middle/Lower Triassic sequences, so that the base Enderby synrift is at least Lower Triassic (BES, Figure 4.25). Regionally, the late Permian to late Triassic sequence is a nearly parallel-bedded sequence that thickens gently seaward and the detachment level of the Enderby structures is probably a shaley unit within the thick pile of sediments accumulated during this pre-extension period. Although there is “growth” of the overall synrift package towards the faults, a careful look at reflection relationships shows that the package is characterized by diverging reflections, usually one or two cycles near the prominent reflectors listed above, interspersed with larger packages of parallel reflections. This relationship suggests that there were distinct pulses of movement on the faults, separated by periods of quiescence from the Triassic through the Jurassic. Fault movement culminated in the Late Jurassic, significantly prior to, and not reactivated by, the Cretaceous extension (M10 or M4). The angular unconformity separating the Upper Jurassic/Lower Cretaceous sections documents significant uplift and erosion of the synrift stratigraphic package prior to the deposition of Cretaceous sediments.

According to orthogonal extension models, the apparent flat lying synrift on strike lines (e.g. Figure 4.26) suggests that transfer faults associated with the normal faults should trend perpendicular to the trend of the normal structures. Although the structure is termed a half graben (which implies a single bounding fault), and the pattern of two to three tilted blocks extends the length of the half graben and into adjacent extensional structures to the north and south, the width of the intra-basinal blocks varies along strike, as recorded by by offsets or jogs in the normal fault plan view (Figure 4.27). The jogs are best correlated by 120° lineaments, oblique to the normal fault trends.

Measured dips on the base synrift reflection are dominantly northwest, as expected in hangingwalls of NE trending, SE dipping normal faults (rose diagram, Figure 4.25). However, many of the dips are not strictly orthogonal to the faults, varying 10°-30° to the north from the expected 313°-317° azimuth, implying that some oblique slip has occurred on the mapped normal faults. It is interesting to note that the dip direction varies toward the direction of tectonic transport direction or spreading azimuth inferred from magnetic lineations of the same age. Modification of the hangingwall dips by later deformation, if it occurred, would have been prior to the deposition of lower Cretaceous sediments as there is no evidence for reactivation of any of the structures in the Enderby Half graben in the overlying Cretaceous package. Minor oblique slip on these faults is also supported
by small dips and offsets of synrift reflections on strike lines (e.g. the right end of Figure 4.26).

The mapped transfer faults intersect Line 117 (T1 to T5 in Figure 4.26). Standard interpretation of this line, without the use of precise dip information, would probably not include these five structures as specific rift faults. That is, the character of a single transfer fault can be very different on adjacent strike lines and so correlation of them has been traditionally overlooked, as was demonstrated in the Eyre Sub-basin interpretation. However, the synrift package reflection character in each of the compartments defined by these structures is distinctive. To the left of T1, the synrift package is largely undeformed and dips slightly toward the right. Base Enderby synrift (BES) is offset across T1. Between T1 and T2 the synrift package dips slightly to the left and is slightly less coherent than the synrift package to the left of T1. Between T2 and T4 the synrift is characterized by undulating reflections with well defined reflection terminations, which correspond to at least one additional transverse structure, T3 (Figure 4.27). Between T4 and T5 the synrift package dips to the left, before being offset across T5 to a more flat lying geometry.

The transfer faults do not affect reflections above the Upper Jurassic/Lower Cretaceous unconformity, indicating that Cretaceous and Tertiary deformation has

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**Figure 4.26** NE-SW Line 117 from the Enderby Half graben. Top synrift (TS), base Enderby synrift (BES) and the detachment surface (D) reflections are indicated. Although generally flat-lying, the synrift package has local relief and is obviously deformed. Dash-dot lines indicate locations of proposed transfer faults associated with the extension accommodated by the normal faults imaged in Line 120, Figure 4.25. The transfer faults separate the synrift package into compartments with different reflection character. See text for further discussion. Dipping reflections below BES are probably out of plane reflections of the subvertical transfer faults (T1 and T2). Location of the intersection of this profile with Line 120 is indicated. See Figure 4.27 for location of profile.

**Figure 4.27** (Next page) A 1-2 km spaced grid of seismic reflection data covering the Enderby Half graben, north of the Enderby Well, reveals two seemingly continuous SE dipping normal faults (e.g. Line 120, Figure 4.25). Jogs in the fault traces, causing the intervening tilted blocks to vary in size along the strike of the half graben, correlate with subtle features identified in strike lines (e.g. Line 117, Figure 4.26) that are interpreted to be transfer faults. A pre-Mesozoic NNE fault (e.g. Line 1016, Figure 4.28) is also mappable from the data in the southwest. Both Mesozoic and earlier structures appear to terminate against a 100° lineament recognized in the interpretation of the regional grid. A NW dipping fault of unknown age, which soles out at about 3.0 s and with approximately the same trend as the Enderby Half graben normal faults is mapped in the extreme southwest of the study area. The Enderby Well is located at the intersection of the four structures. Cuttings from the well include igneous rocks that can be seen to disturb Tertiary section in the seismic data.
not reactivated these structures. The T1 and T2 structures mark the down section termination of the shallowly dipping reflections below the base synrift (Figure 4.26). It is probable that these coherent reflections, which appear to sole out at the same level as the normal faults on the dip line (D, Figures 4.25 and 4.26), represent out-of-plane reflections of energy returned from the subvertical structures themselves.

The transfer faults trend 120°, not 150°, the trend of spreading in the Argo Abyssal Plain, with which they are correlated by well constrained stratigraphic data. They are closely aligned, but clearly not parallel with, the ESE (100°) lineaments mapped from the regional grid. However, it is not clear that they affect section below the detachment surface. Two possible relationships between the transfer faults of the Enderby Half graben and the regional ESE lineaments could explain the structural geometry presented in Figure 4.27. First, the ESE lineaments appear to bound the Enderby crustal block along strike of the half graben and may have acted as strain guides during the Jurassic extension event. In this scenario, the boundaries (and possibly a pervasive fabric within the block) may constrain the transfer faults to a more SE orientation than the extension direction predicted by the 150° spreading azimuth. Alternatively, if the blocks were "pinned" to the boundaries (see discussion of rotation models in Chapter 5), the interior blocks may have rotated to produce the rhomb-shaped geometries, similar to that proposed for other areas of distributed continental deformation (McKenzie & Jackson, 1986). Either model would be consistent with the proposed transtensional nature of the Triassic/Jurassic deformation in other parts of the margin (Etheridge et al. 1991, Stagg & Brassil 1991).

The transfer structures within the Enderby Half graben were not reactivated in the later Cretaceous pulse, that had essentially the same mechanism (as proposed by others, op. cit.). There are some coherent reflection segments below the base Enderby synrift (e.g. below Line 117 and Line 120 intersection at about 4.0 s) but on these data they are not continuous enough to map over wide areas and it is difficult to assess to what depth the transfer structures extend. The transfer faults may affect only the synrift section (i.e. not the underlying section) so that later deformation was accommodated by reactivation of basement structures rather than the detached structures of the Enderby Half graben.

In the southwest corner of the Enderby Half graben, to the northwest of the Enderby Well (Figure 4.27), reflections below the base Enderby synrift are clear in the data (Figure 4.28). The fault plane causing rotation of this sequence is also clearly visible. The fault is mappable and extends about 15 km along a nearly N-S
azimuth, with data control in both the dip and strike direction at 1-1.5 km spacing. The structure soles out at about 4.0 s, deeper than the normal faults in Figure 4.25 and dips in the opposite direction. To the south, the structure merges with the NE normal faults of the Enderby Half graben at the Enderby Well (Figure 4.27). In plan view, both faults are truncated by a 100° lineament at the Enderby Well. A NE trending, NW dipping normal fault continues the shore-parallel basin-forming structures to the south of the 100° lineament. The fault dips in the same direction as the seaward dipping bounding faults of the Barrow Sub-basin to the south, whose extensional deformation as constrained by well data is synchronous with the Enderby half graben. The fault plane is identical, although oppositely dipping, to the Enderby structures (i.e. it is strongly listric, soles out at about 3.0 s). The ESE lineament apparently has served as a polarity reversing accommodation zone during the formation of these synchronous linked half graben, similar to the relationship the Chamaliro and Rukwa dislocations have with the EARS rift zones (Chapter 3, Figure 3.2).

The evidence provided by these data, and data from elsewhere in the regional study area, suggests that the pre-Mesozoic (?)Permian) extension is characterized by N to NNE normal faults linked to ESE transfer structures inboard of the Exmouth Plateau (Figure 4.24). There is minor normal offset of the base Enderby synrift on Line 1016 (Figure 4.28), but the Lower Jurassic/Upper Triassic reflector continues across the structure at approximately the same level. Deformation above this reflector is characterized by anastomosing disruptions or chaotic reflections above the deeper structure, suggesting that normal reactivation of this structure was minor during the main Late Jurassic Enderby Half graben forming event, although significant strike slip may have occurred.

The detailed mapping of the Enderby Half-graben clearly shows that the bounding fault is not linear, but offset laterally at small (5-10 km) intervals. This fault geometry is consistent with the hypothesized small lateral offsets in the Lake Tanganyika bounding faults (Figure 3.8). The evidence from the study of accommodation zone fault geometries also suggests fault blocks on this scale, much smaller than the half graben usually considered to define rift zone geometries. In

**Figure 4.28** NE-SW Line 1016 from the southwestern corner of the Enderby Half graben. The profile intersects a normal fault plane obliquely. Steeper reflections below the BES reflection on this profile indicate that motion on this fault pre-dates the extension that created the tilted fault blocks which characterize the half graben further to the north. This fault segment dips to the WNW, in contrast to SE dipping normal faults of the Enderby Half graben, and soles out at a much deeper level (4.0 s). See Figure 4.27 for location of profile.
Line 1016, NW Shelf
all three examples the fault blocks are bounded by nearly orthogonal, but nonetheless oblique, faults. This study highlights the need to constrain the intra-rift geometry separately from the apparent bounding fault trends, especially where seismic profiles are sparse. Dip analysis provides a fairly rigorous tool for determining anomalous (with respect to the interpreted bounding fault trend and dip) tilts in hangingwalls.

**4.4.2.2 Deep Structure across the margin**

A deep reflection seismic line across the margin just to the north of the Enderby Half graben images the northward continuation of the two to three tilted block deformation style and ties the tilt blocks to the large-scale synforms and antiforms inboard of the Exmouth Plateau (Figures 4.24 and 4.29 a-d). Stratigraphic control is provided primarily by wells situated on "highs" such as the Rankin Platform and the Dampier-Madeleine Trend. The stratigraphy presented in the interpretations has been confirmed by the well data on the highs and by ties with other deep seismic lines shot by AGSO that are tied to wells elsewhere (H.M.J. Stagg, pers. comm.). Nonetheless, the location of the top Triassic (except on the Rankin Platform where it has been drilled) and deeper reflections are speculative. Synforms are the most continuous structures recognized regionally, although where data are dense the axial trend is clearly not everywhere continuous, but appears to be deflected across the proposed 100° lineaments. The antiforms on the other hand, comprise a variety of forms and are not as continuous along strike. These are described variously as "platforms", "arches", "terraces", "trends" and anticlines (e.g. Rankin, Alpha, Candace, Legendre-Rosemary, Bambra, respectively). The antiforms are truncated, offset or change trend at their intersections with transverse structures interpreted from the regional grid (Figure 4.24; also see the right-lateral offset of the Bambra Anticline to the Legendre-Rosemary Trend in Kopsen & McGann 1985, their Figure 2). The transverse structures are coincident with projected 100° lineaments from the western edge of the Exmouth Plateau.

The tilted blocks on the shoreward end of the profile (Figure 4.29a) are perched high on the shelf. The block bounding faults sole out at between 3.0 s and 4.0 s as in the Enderby Half graben. Below the detachment, fairly continuous high amplitude reflections between 4.0 s and 6.0 s dip to the left (SW). The tilted fault blocks pass to the NW (left) across a zone of chaotic, discontinuous reflections known as the Legendre-Rosemary Trend (hereafter referred to simply as the Legendre Trend). Deeper in the section, a band of short, but high amplitude reflections is characteristic of lower crust reflectivity. This band appears to dip to the SE, suggesting that the crust is thickening toward the continent as expected.
Lower crust high reflectivity appears to terminate under the Legendre Trend, although this may be due to scattering of energy higher in the section.

The NW (left) boundary of the chaotic zone is marked by a major NW dipping structure that appears to truncate the highly continuous reflections of the adjacent synform known as the Lewis Trough (LT, Figure 4.24). There are several possible interpretations of the fault geometries and their relationships through the chaotic zone. At some levels, reflections appear to be offset in reverse sense, whereas there are equally good examples of normal offset. Truncation of reflections below the top Jurassic reflector suggest that this area was elevated and eroded. It is difficult to confidently state whether the syncline bounding structure cross-cuts the bounding fault of the tilted blocks or the converse is true, but inconsistent reflection and offset relationships across the chaotic zone suggest several reactivations were focussed here.

The Legendre Trend antiform or high (Figure 4.29a) is frequently mapped as a continuous NE trending structure up to 80 km long. However, mapping of data near this deep seismic line, suggests that in this location, the high is triangular in plan view (Figure 4.24, upper right). The synform bounding fault merges to the south with the shelf tilt blocks. Where the two structures merge they are truncated by a 100° lineament, similar to the structural configuration mapped in the southern Enderby Half graben. The interpretation of the regional grid suggests that the axial trends of both the Legendre Trend and Lewis Trough, is in fact, deflected or offset across 100° lineaments, again suggesting that these lineaments have exerted an influence on the patterns of reactivation along the shelf.

The Lewis Trough synform rises to the left onto a high termed the the Dampier-Madeleine Trend (Figure 4.29b). This high coincides with a 100° lineament where it intersects with this profile. Normal offset across the structure of the top Jurassic reflector is confirmed by well data (H.M.J. Stagg, AGSO, pers. comm.). However, reverse offset of a distinctive two cycle, high amplitude reflection lower in the section is also apparent in the data. An angular unconformity at the top Jurassic reflector on the downthrown side of the the main structure suggests significant uplift of the antiform before the end of the Jurassic.
Interpretation of Line 3185, Northwest Shelf (shoreward)
Interpretation of the Line 3185, Northwest Shelf (seaward)
Line 3185, (shoreward) NW Shelf
To the NW (left) of the 100° lineament, the Mesozoic section thickens dramatically toward the fault bounded Rankin Platform. Reflection offsets across the platform bounding structure are all normal, generally decreasing upward. Well data confirm that the top Triassic reflector sub-crops in an angular unconformity beneath the top Jurassic reflector indicating a substantial loss of Jurassic section. Reflections below the Mesozoic section, between the Rankin Platform and Dampier-Madeleine Trend, dip fairly steeply to the left, and the apparent large offset of the top Triassic (and Permian?) reflector compared to offsets higher in the section suggests that a large-throw ESE-SE dipping normal fault existed prior to the mid to late Mesozoic deformation events. The repeated reactivation of this structure has created a broad "flower" structure typical of wrench tectonics, suggesting that the thickening of the Mesozoic section was caused by normal movement on a fault not intersected by this profile (i.e. subparallel to it) and that Mesozoic reactivation of this structure was largely strike slip to transtensional.

Many of the producing fields along the Rankin Platform trend, lie within triangular shaped highs, bounded on the east by a NE normal fault and a more northerly west dipping normal fault on the west (e.g. North Rankin, Goodwyn and West Tryall Rocks, Figure 4.24; see also Kopsen & McGann 1985, their Figure 2). This configuration is very similar to the pattern of the N-S and NE-SW normal faults near the Enderby Well and further to the north on the shelf, described above. The interpretation herein suggests that the intersections of pre-existing NNE-N normal faults and the 100° lineaments were sites where oppositely dipping NE normal faults during Jurassic deformation events initiated. Further, the relationships described in the deep reflection line suggest that the 100° lineaments were reactivated in a transpressive manner prior to the Jurassic and later extensional events, whereas the NNE pre-existing normal faults were transtensionally reactivated later in the Mesozoic. There is quite good evidence of a Permo-Carboniferous extensional event that produced NE-SW trending rift zones resulting in the pattern of shore parallel sediment filled troughs and eroded highs (Figure 4.29). There is also evidence that the N-S and 100° trends may be inherited from Precambrian heterogeneities (see discussion in Section 4.4.2 and M. Etheridge pers. comm.). Where the Permo-Triassic is imaged, tilted blocks are commonly bounded by the same structures that were active in later extensional phases. This style of reactivation is reminiscent of the relationship of the pre-existing large-scale structures with the Rukwa rift zone and the deeps in the Lake Tanganyika and Lake Malawi rift zones (Chapter 3). That is, the intersections of the 100° structures and their N-NNE conjugates have acted as loci for later
deformation, even though stress orientations have been substantially different in each pulse.

Features in the Enderby Half graben and adjacent shelf areas describe an oblique slip fault system that accommodated NW-SE extension in the late Jurassic. The direction of extension in this sub-basin as determined from the interpreted transfer faults and dip analysis is approximately 30° oblique to the direction of spreading described by magnetic anomalies in the Argo Abyssal Plain (Figure 4.24), to which the faulting is tied by well data. As discussed in Section 4.1, Smith (1993) has shown that rift zones that are not aligned with the pole of rotation will normally diverge obliquely. It has been a recurring theme in the case studies of passive margins herein that rift zones initiate in non-ideal locations and orientations due to pre-existing zones of weakness. However, there is no convincing corroborating evidence at present for a rift axis parallel suture zone or lithospheric scale anisotropy under the Enderby Half graben. However, the non-ideal orientation of the rift axis and the oblique geometries mapped therein suggest that one may exist.

The extension recorded in the upper crust in the Enderby Half graben must have had a counterpart in the lower crust. Ideal detachment plus pure shear models of lithospheric extension predict that lower crustal extension does not underlie the upper crust extension, but is offset in the direction of normal fault dips (e.g. Lister et al. 1986). Thus, the dip direction of the normal faults in the Enderby Half graben suggests that lower crustal extension (thinning) would be located continentward of the sub-basin. There is no indication of lower crustal thinning toward the continent in gravity data (Stagg & Brassil 1991), nor does lower crustal reflectivity suggest it (Figure 4.29). Thickening of Jurassic and Triassic sequences underlying a present day bathymetric trough at the axes of the synclines seaward of the tilt blocks, suggest that these have been areas of active subsidence, without obvious normal faulting and thus are areas of very thin crust. This thinning must have begun prior to the deposition of the Mesozoic sequences. Depth converted, deep seismic reflection data indicate that Mesozoic sediments extend as deep as 10 km in the synclines (Stagg & Brassil 1991), again suggesting extreme lithospheric (lower crust) thinning below them or perhaps flexure of the whole lithosphere. Thus, type models of the connection of the lower and upper crustal extension appear inadequate to describe the relationships interpreted from the extensive data of the Northwest Shelf.

Vink et al. (1984) has quantitatively evaluated the tensile strength of the lithosphere, incorporating geotherm and pore pressure parameters. The strength
of the lithosphere is highly dependent on these parameters, but Vink et al.'s (1984) results suggest that variations in these parameters do not strongly affect the comparative ratio of strengths for different types of lithosphere. For example, "a 10 km increase in crustal thickness from 35 to 45 km, reduces the total strength of the lithosphere by about one half " (Vink et al. 1984, p. 10075). It is possible that whatever mechanism was responsible for the apparent extreme thinning of the crust below the syncline created a pre-existing "lithospheric boudinage" as proposed by Dunbar and Sandwell (1988). If so, the adjacent thickened area would underlie the Enderby Half graben, making this lithosphere weaker which may explain the subsequent localization of upper crustal extension here, although this concept remains largely speculative based on the data available for this study.

The Northwest Shelf has been the site of repeated deformation both before and after the initiation of seafloor spreading, making the term "passive margin" largely inappropriate. Reactivation of pre-existing tectonic fabric, or crustal shear zones, is supported by the use of the ~100° lineaments during several deformation pulses. Some consistent reactivation patterns have been suggested by this study of distributed deformation. The large quantity of data available for the margin allow the reactivation hypotheses put forward by this preliminary analysis to be tested by further identification of similar patterns of reactivation. Recently acquired deep seismic data (Stagg & Brassil 1991) and continued petroleum prospectivity resulting in additional high quality geophysical data promises to greatly enhance our understanding of the nature of the coupling between the upper and lower crust and the nature of lithosphere heterogeneity.
4.5 Concluding remarks, Australian passive margins

The fault geometries of rift zones and large-scale crustal structure of various Australian passive margins have been outlined for comparison with features in the intracontinental rifts of East Africa. With the exception of the extensional structures on the western margin of the Exmouth Plateau, all of the rift zones studied lie well inboard of any inferred COB so are reasonably considered to be the result of intracontinental deformation. This fact points to an important concept often ignored in plate reconstructions and models of passive margin development. That is, rifted basins on passive margins frequently do not represent the locus of extension directly tied to lithospheric rupture, but are "far-field" expressions of the tectonic forces that initiate continental fragmentation. The extension recorded in them, depending on the timing, needs to be removed for accurate plate reconstructions. Further, since their location and geometries appear to be controlled by pre-existing heterogeneities, it is not surprising that the kinematics inferred from structures within them is frequently oblique to the eventual kinematics of seafloor spreading.

The morphotectonic elements of the NE and NW Australian margins with their plateaux surrounded by extensional sub-basins are similar to the Tanzanian Craton bounded by the Eastern and Western rifts of East Africa. The apparently common occurrence of this juxtaposition of crustal features suggests that this is a common mode of continental fragmentation resulting from tectonic forces. The observed configurations suggest that the location of stress accumulation in the lithosphere is dependent upon large scale lithospheric heterogeneities, which do not have to be appropriately (in terms of traditional strain models) aligned. Upper crust deformation is thought to reactivate pre-existing anisotropies only when they are appropriately aligned with the plate motion/tectonic transport direction. However, the case studies herein suggest that the range of alignments which can be reactivated is much broader than traditionally assumed.

Entire rift zones, as in the case of the NE margin of Australia can be reactivated during widely separated periods of deformation (i.e. Paleozoic basin-forming faulting, probable activity in Cretaceous episodes associated with the opening of the Tasman and Coral Seas, and clear reactivation of some of these faults in the known Tertiary sequence). Smaller scale heterogeneities, possibly limited to the upper crust but probably extending deeper, can also be repeatedly reactivated (e.g. the presumed Precambrian 100° lineaments on the NW Shelf involved in both the ?Permian and Mesozoic extension events). The apparently
distinct pulses of deformation on long-lived "passive" margins, makes the term a misnomer. Better understanding of the deformation stages and their interaction on these continental fringes is critical to constraining models of lithosphere dynamics.

Several anomalous features of the Ceduna Sub-basin including the magnetic trough running through it, rather than bounding it, the presence of the unusually deep half graben (Figure 4.13) and repeated activity including extensive igneous activity, suggests that some lithospheric, regionally extensive heterogeneity has localized of deformation and argues for significant tectonic inheritance in the break-up of Australia and Antarctica. The association of the proposed lower-crustal shear zone with evidence of two anomalous mantle areas on either side of it, suggests an intimate coupling between base lower crust topography and perturbations of mantle flux.

The long-lived and repeated deformation along the Ceduna and Rukwa lineaments suggest a focussing of stress accumulations within them compared to the apparently instantaneous, geologically speaking, creation of the "extensional gash" sub-basins (e.g. the Eyre and Polda and Tanganyika and Malawi, respectively). The possibility that pre-existing lineaments and pervasive crustal heterogeneities act as strain guides in the early phases of continental break-up casts doubt on the use of structural trends from isolated sub-basins to determine early kinematic histories of passive margins. Indeed, even if the fault geometry is fairly well constrained, and from the example of the Enderby Half graben it is apparent that this requires a very tight data grid, kinematics inferred from them can be misleading. The results of these case studies emphasize the need to consider regional structural relationships, including the identification of pre-existing structures and lithospheric heterogeneities, in such exercises. Perhaps more importantly, it is clear that intra-rift and bounding fault geometries may vary considerably. Dip analysis provides an additional test for fault correlations and kinematic inferences based on bounding faults.
5.0 OBLIQUE EXTENSION: IMPLICATIONS FOR DEFORMATION MODELS AND PLATE TECTONIC MODELS

Basic research is what I am doing when
I don't know what I'm doing.
Werner von Braun

5.1 Introduction

A series of case studies of extensional basins from both juvenile intracontinental (Chapter 3) and mature passive margin (Chapter 4) settings have been undertaken to constrain the geometry and kinematics of intracontinental deformation that precedes continental breakup. In this chapter, deformation models which are consistent with the observational data are reviewed. It is found that none can adequately explain the diversity and complexity of deformation that is recorded in the case studies. As has been suggested in earlier discussion sections, pre-existing structure and lithospheric heterogeneity and anisotropy are the cause of anomalous structural relationships and the reactivation patterns that are recognized in oblique extension. Thus, a primary conclusion of this research is that tectonic inheritance plays a fundamental role in intracontinental deformation and continental fragmentation.

Although the data used for the interpretations in the previous chapters are primarily dealing with the upper crust, a better understanding of these systems and how they are integrated with continental scale structure should help to constrain theoretical models of lithospheric extensional deformation and processes between the elements of the whole lithosphere (i.e. upper and lower crust and sub-lithospheric mantle). Table 5.1 summarizes the relationships of the case study rift zone geometries with what is known about the regional pre-existing crustal structure and lithospheric heterogeneity.

The case study interpretations are not necessarily unique and so deriving global implications from them needs to be undertaken with care. Considered together, the case studies indicate that certain reactivation structural patterns are repeated. In the following sections, a comparison of the common structural styles with deformation models is undertaken. A review of the effect of pre-existing lithospheric heterogeneity, including transcontinental structures and whole lithosphere anisotropy or non-uniformity, on the studied rift zone's geometries is presented. The implications of tectonic inheritance, as suggested by the case studies, with regard to plate tectonic models are explored.
## Table 5.1: Regional Geometry

<table>
<thead>
<tr>
<th>Rift Zone</th>
<th>Prerift Geological Setting</th>
<th>Rift Axis Alignment</th>
<th>Rocks Deformed</th>
<th>Segmentation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lake Tanganyika</td>
<td>deformed and metamorphosed Proterozoic belt between Archean cratons</td>
<td>variable: cuts pre-existing terrane boundaries</td>
<td>Precambrian folded metamorphic terranes cut by Paleozoic / Mesozoic rifts</td>
<td>associated with pre-existing transcontinental dislocations</td>
</tr>
<tr>
<td>Lake Malawi</td>
<td>deformed and metamorphosed Proterozoic belt between Archean cratons</td>
<td>variable: cuts pre-existing terrane boundaries</td>
<td>Precambrian folded metamorphic terranes cut by Paleozoic / Mesozoic rifts</td>
<td>associated with pre-existing transcontinental dislocations</td>
</tr>
<tr>
<td>Lake Rukwa</td>
<td>Proterozoic terrane boundary</td>
<td>linear, aligned with pre-existing terrane boundary</td>
<td>Proterozoic terrane boundary (shear zone) / Paleozoic trough</td>
<td>associated with intersection of 3 Proterozoic terranes</td>
</tr>
<tr>
<td>Queensland Trough</td>
<td>Paleozoic Tasman Fold Belt (?)</td>
<td>linear, aligned with fold belt structural trends</td>
<td>pre-existing Paleozoic rift / folded Paleozoic metasediments (?)</td>
<td>can be projected along strike to connect with onshore basement gravity trends</td>
</tr>
<tr>
<td>Eyre Sub-basin</td>
<td>Proterozoic Albany-Fraser Fold Belt</td>
<td>linear, aligned with fold belt structural trends</td>
<td>platform sediments (?) overlying Proterozoic metasediments</td>
<td>change in bounding fault trend across a possible pre-existing deep shear zone</td>
</tr>
<tr>
<td>Ceduna Sub-basin</td>
<td>craton / Paleozoic fold-belt or terrane boundary (?)</td>
<td>linear, aligned with pre-existing terrane boundary; bracketed by anomalous mantle</td>
<td>basement rock type unknown</td>
<td>major offsets in hingeline associated with pre-Cretaceous deeps and Tertiary magmatic centers</td>
</tr>
<tr>
<td>Enderby Half graben</td>
<td>thick Paleozoic / Mesozoic platform sediments</td>
<td>linear, aligned with regional structural trends in flexed / boudinaged (?) lithosphere</td>
<td>thick Paleozoic / Mesozoic platform sediments</td>
<td>half graben bounded along strike by pre-existing crustal shear zones</td>
</tr>
<tr>
<td>West Exmouth Plateau</td>
<td>unknown</td>
<td>offset-linear (?)</td>
<td>thick platform sediments (?)</td>
<td>reactivated pre-existing crustal shear zones (100° lineaments)</td>
</tr>
</tbody>
</table>
5.2 Experimental and Numerical Models of Rifting

A detailed examination of selected accommodation zones in the Lakes Tanganyika and Malawi rift zones (Section 3.2.4) has demonstrated that at small scales in modern, relatively isotropic sediments, deformation patterns are approximately conformable to ideal orthogonal extensional strain fault systems as proposed by others (e.g. Etheridge et al. 1984, 1985, 1989, Gibbs 1984, 1985, Lister et al. 1986, 1991). However, structures are orthogonal to the tectonic transport direction and thus, oriented obliquely to the rift axis. The Enderby Half graben and Eyre Sub-basin fault geometries are near orthogonal. In the Enderby Half graben, the nearly orthogonal geometry has been attributed to near isotropic bulk properties of the package being deformed and it's detachment from lower sequences. However, in both faults systems major bounding faults are oblique to the rift axis, especially the Eyre Sub-basin, so they must still be considered to be the product of oblique extension.

The pervasive oblique slip on major bounding faults in Lakes Tanganyika and Malawi (Section 3.2) and the apparently common occurrence of oblique rifting (Chapters 3 and 4 and references therein) argue that the ideal orthogonal models are not adequate to describe the deformation occurring during intracontinental extension, at either the half graben and dip domain scales or the regional scale. The complexity of deformation in all of the systems studied suggest that they are the products of transtensional (i.e. triaxial) deformation (e.g. Sanderson & Marchini 1984).

The concept of oblique rifting has received increasing support in the literature describing natural systems (e.g. Wu & Bruhn 1990, Willcox & Stagg 1990, Scott et al. 1992, Bartley & Glazner 1991, Ring et al. 1992, Chorowicz & Sorlien 1992). All of the rift zones studied herein show a degree of oblique divergence and thus successful deformation models must be able to accommodate oblique slip deformation. The case studies herein represent a wide variety of scales and tectonic settings and document the prevalence of oblique slip faults which cannot be explained by traditional strain models (Anderson, 1951) which predict a conjugate set of normal faults whose traces are perpendicular to the least compressive stress (Figure 5.1a). The Anderson (1951) model has received wide support in experimental deformation studies, due largely to the fact that experimental rigs are generally axisymmetric and generally do not include truly triaxial strain.
A theoretical model of faulting in three dimensions, developed to explain the occurrence of synchronous multiple fault sets observed in natural settings (Reches 1978), suggested that three or four sets of oblique slip faults are necessary to accommodate three dimensional strain (Figure 5.1b). Reches & Dietrich (1983) devised a servo-controlled rig that allowed independent application of strain along three axes and ran a series of experiments on Berea Sandstone, Sierra-White and Westerly granites and Candoro and Solnhofen limestones. The results of these experiments indicate that for conditions of plane strain, slip on two conjugate faults is sufficient to accommodate the applied strain, but for the more general three dimensional strain case, three or four fault sets with orthorhombic symmetry develop (Figure 5.1b). Reches (1983) further developed his three dimensional model to describe the results of the experimental data cited above and found good agreement between predicted and observed fault orientations in the sedimentary samples, whereas faults in the granites deviate from the predicted orientations. Good agreement is also found between predicted and observed stress in all experiments. The common occurrence of oblique slip faults in nature makes it clear that assumptions of plane strain or coaxial strain, in general, are not valid for most real Earth systems.

Rosendahl (1987) suggested that incomplete development of the orthorhombic pattern could be used to explain his "fundamental rift unit", consisting of three fault segments (Figure 1.3). However, the development of a dip slip fault perpendicular to the least compressive stress is not supported by the orthorhombic model. Faults produced in the orthorhombic model can clearly result in a "corner" where two oblique slip faults meet. Horizontal extension is aligned along a bisector of the angle made by the fault segments (Figure 5.1b), similar to the models presented in Chapter 2. It is suggested here that this is a common fault geometry in intracontinental extension where total extension is small and may result from underdevelopment of one of the pairs of faults compared to the other in the orthorhombic model. The cause of such unequal development is that deformation in natural systems is rarely affecting homogeneous crust. The "corner" fault geometry provides a system that is more susceptible to reactivation than faults of a uniform orientation (Figure 5.2), because there are more plane-of-weakness orientations (i.e. at least two) available to reactivate.
A. Anderson (1951) Model

B. Reches (1983) Model
Experimental analog and theoretical models of oblique divergence on a rift zone scale have reproduced a range of fault patterns, which suggest either a distinct boundary between strike slip and orthogonal basin forming mechanisms (Smith & Durney 1992, Withjack & Jamison 1986, McCoss 1986) or a broad transition between these two types of deformation (Tron & Brun 1991). Fault patterns derived from an individual run of the models can usually be qualitatively applied to a particular natural system. Given the variety of fault geometries produced, it is clear that oblique slip dominates fault motion in these models. Indeed, Tron and Brun (1991) conclude that in oblique rifting, faults are never perpendicular to the stretching direction.

A weakness of the above models is the assumption that divergence is translational rather than rotational. That is, they considered the effect of deformation on a planar, rather than a spherical, surface. Smith (1993) addressed the problem of translational versus rotational motion by noting that, in natural rift systems, horizontal movements are on a sphere and therefore must be considered to be rotations about some pole even in the earliest stages of deformation. Smith (1993) developed an infinitesimal kinematic model of rifting that incorporates both the oblique and rotational aspects of divergent motion on a sphere. In this model, the orientation of extensional structures in a rift depends on both the distance to the pole of rotation and the offset of the rift trend from the pole of rotation (Figure 4.12). If the trend of the rift does not lie along a great circle through the pole of rotation, the divergence will be oblique. Why a rift zone might initiate in a particular location or orientation is not addressed, but the case studies herein argue that tectonic inheritance and lithospheric anisotropies may be the key. These topics are discussed further in the following sections.

The degree of divergence obliquity at a specific point on the rift in Smith's (1993) model is predictable from the ratio between the offset of the pole from the rift trend and the distance from the point on the rift to the rotational pole (Figure 4.12). The model predicts that the strike of en echelon structures bounding the rift will vary, as will the divergence direction, for differently trending rift segments and even along long straight rift segments since different points on the rift segment will be at different distances from the rotational pole. This effect is more pronounced for rift zones close to the pole. Dip analysis in the EARS (Chapter 3) appears to contradict this prediction, if it is indeed an accurate reflection of the tectonic transport direction as argued in Chapter 2. If the pole of rotation for the EARS is located at a great distance from these rift zones, differences in divergence direction would be minimized and may not be readily apparent. Or, the location and trend of rift axes may be fortuitously arranged such that apparent divergence is
Deformation models and Plate Tectonic models

the same. The model does allow for initiation of rift segments of varying orientations and locations, but does not address the effects of pre-existing heterogeneities at different locations nor why rifting would be initiated in these "non-ideal" locations with "non-ideal" rift axis trends.

Bounding structures in the EARS are not strictly en echelon and have a range of trends, and yet hangingwall dips are consistent between widely separated rift zones (e.g. the Lake Tanganyika and Lake Malawi rift zones). The structural complexity recorded in the rift zones is attributed to extension operating in the highly heterogeneous crust of the African continent. Indeed, the underlying weakness with all of the attempts to describe oblique divergence cited above is the use of isotropic starting configurations, a rarity in natural settings. In the following section, reactivation and tectonic inheritance relationships in the case studies herein, and from other regions documented in the literature, are elucidated and highlight the problems of assuming a homogeneous isotropic lithosphere.

**Figure 5.2** A plane of weakness is most likely to be reactivated when the shearing stress, \(\tau\), is maximized. \(\tau\) is maximized when the angle of the normal to the plane of weakness, \(\theta\), is oriented 45° or 135° to the stress (graph). A schematic cartoon shows a vertical fault in plan view. In this situation, the plane of weakness is "ideally" oriented to be reactivated for two stress orientations, both horizontal. If a fault plane is composed of two fault segments, as is suggested for the "corner" structures discussed in Chapter 2, there are four stress orientations which have a high probability of reactivating the fault. Similar cartoons depicting faults composed of three segments and a curvilinear surface (i.e. infinite fault segments) suggest that some portion of these complex surfaces will be more probable to reactivate under a broader range of subsequent stresses than faults of a single trend.

Dip of the fault has not been considered in these simple cartoons, but the principles remain valid, except that a dipping fault will never be in ideal orientation for reactivation by horizontal stress as is thought to exist in the lithosphere. This is probably why large scale transcurrent shear zones, which are generally subvertical structures, are frequently the locus for reactivation during subsequent regional intracontinental deformation.
Single Fault Segment

Corner Fault

Three Fault Segments

Curvilinear

\( \theta \), angle of \( \perp \) to fault surface to stress

\( \tau_{\text{max}} \)

\( \tau \)

\( \pi/4 \), \( \pi/2 \), \( 3\pi/4 \), \( \pi \)
5.3 Reactivation and Tectonic Inheritance

5.3.1 Reactivation of Large-scale Continental Structures

Reactivation of pre-existing structures is frequently invoked in areas where the deformation cannot be explained by traditionally accepted mechanisms, but rarely are the particulars of the reactivation elucidated. This study has demonstrated that oblique extension is probably more common than conventional orthogonal extension and that structural complexity and multiple fault orientations are the norm in Phanerozoic rift zones. This is a direct result of the fact that inherited structures and lithosphere anisotropy or lateral heterogeneities play major roles in localizing rifting and determining the degree and style of oblique extension (Tables 5.1). A summary of the details of the rift zone specific geometries is presented in the next chapter (see also Table 6.1), where their implications for basin analysis methodologies are discussed. Reactivation at the smaller, more local scale is also discussed. The discussion below concentrates on the effects of pre-existing structures on the larger scale, whole rift zone geometry, comprised of rift axis alignment, border faults and segmentation, and the associated kinematics of oblique extension.

Many of the bounding faults of the rift zones studied here have documented oblique slip on them. Within a single rift zone, bounding fault trends are frequently highly variable. The variability of trends of these presumably synchronous structures is attributed to the reactivation of pre-existing structures. This type of reactivation is well documented in northern Lake Malawi. There, the NW-SE Livingstone bounding fault reactivates the same Proterozoic terrane boundary reactivated in the Rukwa rift zone as the Lupa bounding fault. The adjacent N-S Usisya bounding fault, to the south, reactivates a Cretaceous extensional structure. It is clear from these examples that simple inferences about regional, or even half graben scale, kinematics based on rift zone bounding fault trends are potentially misleading and that assumptions of deformation occurring in isotropic crust are inadequate to describe natural systems.

Dip domains define a deformation partitioning within most of the studied rift zones. That is, adjacent half graben may not have the same geometries, but faults within them are normally similar in trend to each other. Large scale pre-existing structures also seem to control deformation partitioning in an oblique convergence setting in the Alps (e.g. Polinski & Eisbacker 1992). Likewise, in areas of distributed strike slip deformation pre-existing structures, even though
inappropriately aligned, are thought to be reactivated as the boundaries of structural style partitioning (Jackson & Molnar 1990, Scotti et al. 1991).

Dip domain boundaries, especially where the dip directions of adjacent half graben are substantially different or define opposing polarity half graben, appear to correspond to the high angle intersections of the rift zone with steep transcontinental shear zones (Figures 3.2 and 3.4)(Versfelt & Rosendahl 1989, Courtillot 1982). A partitioning of normal fault trends in the Queensland Trough is related to deep basement trends revealed in gravity data onshore. A similar influence by a PreCambrian shear zone is postulated for changes in rift geometry in the Red Sea (Dixon et al. 1987) and to control the zig-zag fault pattern in the Red Sea and Gulf of Suez (Jarrige et al. 1986). Crustal heterogeneity in the Kenya rift, consisting of a series of late Proterozoic continental scale ductile/brittle shear zones, is thought to control the location and geometry of graben structures, the location of intrusion of magma, which corresponds to transfer faults, and changes in rift trend and major bounding fault polarity (Smith & Mosley 1993).

The segmentation of rift zones is also the result of the influence of pre-existing structures on rift zone geometry (Table 5.1). Crustal shear zones control the segmentation of rift zones in the Northwest Shelf and Queensland Trough systems. On the Northwest Shelf, the ~100° lineaments may be associated with an earlier Permian extensional event, but they could be older. Etheridge (1986) has argued that reactivation of extensional fault systems may be mechanically favored, but will only occur when weak zones are in near ideal orientations. However, the theoretical modelling of distributed deformation in the Transverse Ranges of California by Scotti et al. (1991) suggests that movement can occur on pre-existing planes of weakness that are up to 75° out of ideal orientation and suggests that paleomagnetically determined rotations do not necessarily indicate a rotation of the stress field. The Late Jurassic extension on the seaward side of the Exmouth Plateau appears to have been directed approximately E-W, so that the 100° lineaments were indeed in near perfect orientation to be reactivated as transfer faults. The Enderby Half graben axis is offset in map view from synchronous half graben axes to the north and south across the 100° lineaments. However, within the Enderby Half graben, where extension was oriented NW-SE, these lineaments were not used to segment the normal faults. Instead, 120° transfer faults developed.

Detachment of hangingwall blocks high in the basin stratigraphy is recorded in the Enderby Half graben of the Northwest Shelf and suggested for some of the Tertiary faults in the Lake Rukwa rift zone. Detachment of the crust at shallow
levels allows for the initiation of a new set of faults which may form somewhat independently of pre-existing structures (e.g. the Enderby Half graben). However, the evidence in the Rukwa rift zone suggests that new faults initiate over underlying structures, especially where "corner" or splay faults off of a major boundary are well developed.

These corner structures may also occur where bounding faults consist of fault segments that are reactivated local pre-existing structures joined with fault segments that are not influenced by a pre-existing structure. If the relative components of dip versus strike movement on a fault in a given pair reverses from one pulse of deformation to the next, rotation of the hangingwall block about a vertical axis is predicted.

As stated above, the 100° lineaments appear to bracket the Enderby Half graben as a whole (Figure 4.24) and thus may have exerted some control on the deformation geometry of the crustal block they bound. For instance, if the Enderby blocks were "pinned" to these boundaries (as discussed below, see Figure 5.3), NW-SE directed Jurassic extension would reactivate the boundaries in an oblique slip sense, and rotation of the Enderby block(s) would occur and could explain the rhombohedral structural style mapped there. Thus, reactivation of non-ideally aligned pre-existing structures by oblique slip seems to cause rotation of crustal blocks on a variety of scales. Apart from the ubiquitous oblique slip documented in the rift zones studied here, evidence is largely lacking for the hypothesized rotation. However, evidence for rotation phenomenon from other areas of distributed continental deformation and models to describe it may be fitted successfully to the some of the observational data from these case studies.

Freund (1970) was one of the first to propose rotations of large continental blocks, and their bounding faults, in the Zagros Thrust Zone near Sistern in southeastern Iran. With increasing use and understanding of paleomagnetic measurements, other studies invoked rotation of continental blocks, especially in strike slip terranes (e.g. Garfunkel 1974, Ron et al. 1984, Garfunkel & Ron 1985). Discordant paleomagnetic poles led Beck (1976) to postulate shear and subsequent rotation of large continental blocks in the western Cordillera of North America. Block rotations of tens of degrees in areas of distributed continental deformation are well documented in southern California in a variety of domains adjacent to the San Andreas Fault (e.g. Luyendyk et al. 1985, Terres & Luyendyk 1985) and in the northern Dead Sea transform (Heiman & Ron 1993). Eyal et al. (1986), studying the well exposed Bir Zreir rhomb-shaped graben in Eastern Sinai, state that standard "pull-apart" models predict 3 to 6 times more vertical separation on the bounding
faults than is evident and suggest that rotation of a block formed in an earlier phase of deformation and documented oblique slip on the pre-existing bounding "strike slip" faults better explain the present configuration.

Some early attempts to model the kinematics of rotation during transcurrent faulting predict extrusion of rotating blocks (Freund 1970, Freund 1974, Garfunkel 1974, Garfunkel & Ron 1985, Ron et al. 1984). Noting that rotated terranes in southern California are not being thrust over the San Andreas Fault, and so are not being extruded, led to the development of a model in which blocks are "pinned" to adjacent terranes (Figure 5.3) (McKenzie & Jackson 1983, 1986). The mechanism of rotation is similar to that proposed for microplates in the oceanic setting as discussed below (Schouten et al. 1993), in that the rotation is a response to forces acting on the lateral boundaries of the blocks (Figure 5.3).

Jackson & Molnar (1990) have assessed and integrated the geometry of faulting, slip vectors of major earthquakes and the results of very long baseline interferometry data from the Transverse Ranges of California which suggest that rates of rotation are on the order of 3° to 6° per million years. They note that rotation rates of blocks responding to viscous deformation below them are likely to be different (slower) than those responding to rotations caused by forces acting on the blocks boundaries. The observed rates are more consistent with the former mechanism and lead the authors to propose that blocks are "floating" on a continuously deforming substrate. Convincing evidence that the Enderby Half

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*Figure 5.3* A series of rotation models. Twiss et al.'s (1993) model uses micropolar continuum theory that considers material to be essentially granular, taking into account two separate scales of motion: a large-scale average motion of the material (i.e. a macrostrain rate) and an associated macrospin; and a local motion, the microspin, describing the average rotation rate of grains or fault blocks in the material. The Twiss model can produce a wide variety of fault geometries, including the "pinned" block model of Mckenzie & Jackson (1986), simple shear with no rotation of the deformed blocks, and simple shear wherein the microspin = macrospin and others (see original article). The various fault geometries produced depend on the variables, \( \beta \), the angle that the relative velocity vector between the plates makes with the plate boundary and, \( \alpha \), the angle a material line (p) within the deformed zone makes with the plate boundary.

Schouten et al. (1993) has developed the "Roller Bearing Model" for microplate rotation based on his studies of the Easter and Galapagos microplates. In this model, rotation is the result of edge-driven shear rather than by a shear flow of mantle underneath. The instantaneous rotation axes (small circles) describing the microplate's motion relative to the bounding major plates are located close to its margins with those plates and close to the tips of propagating rifts. A single gear or double gear mechanism is possible. This model may also be appropriate to describe the rotation of large blocks of continental lithosphere, especially in the development of marginal plateaux on passive margins.
Pinned block model (after McKenzie & Jackson, 1986)

Simple shear with microspin = 0

General micropolar continuum theory (after Twiss et al., 1993)

 Roller bearing model for microplate rotation (after Schouten et al., 1993)
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graben deformation was detached from its substrate during Jurassic deformation suggests that this model may be applicable there (Figure 4.25).

Twiss et al. (1993) compare distributed brittle deformation to granular material and perceive that classical continuum mechanics doesn’t adequately describe such deformation. They develop "micropolar continuum theory" which separates large scale (plate/macrospin) average motion of the material and smaller scale (block/microspin) local motion (Figure 5.3). The theory predicts a variety of fault patterns including conjugate shears, rotated blocks on uni-directional shears, anastomosing faults and the "pinned block" model depending on the relative velocities of the two bounding plates (or blocks) and the relative microspin of the blocks within the deformed zone. All of the above models for block rotation predict a prevalence of oblique slip on block bounding faults.

While many paleomagnetic studies seem to establish the reality of block rotations, paleomagnetic data were used to argue against previously proposed large rotations in the Ameca Tectonic Depression, Mexico (Nieto-Obregon et al. 1992). These data suggest that tectonic transport direction was out of the "corner" of a bent, but apparently continuous listric fault, identical to the geometry suggested in Chapter 2 (Figure 2.3) and demonstrated by dip analysis for the "corner" splay faults in the Lake Rukwa rift zone (Section 3.3, Figure 3.16). The kinematic history proposed for the Rukwa rift zone suggests that non-ideally aligned intra-rift block bounding faults, composed of a fault segment which is coincident with and a fault segment at a high angle to the Lupa fault, have been repeatedly reactivated by oblique slip regardless of the inferred stress orientations. As suggested earlier, the configuration of two joined fault segments, oriented at a high angle to each other, provides a zone of weakness that can be reactivated by a wider range of stress orientations than a single fault of a particular orientation (Figure 5.2).

Etheridge (1986) cautions that some fault zones may be strong, rather than weak, zones because of annealing or silicification, for example. When a fault zone is strain hardened, there is nonetheless a significant lateral strength gradient and the adjacent weak zone may be favored for deformation so that crustal heterogeneity still may control the location and orientations of later deformation. For example, rift zones aligned at a high angle to the tectonic transport direction seem to nucleate at the intersection of non-cratonic terranes and other transcontinental scale dislocations (e.g. Tanganyika and Malawi rift zones). Extension of the upper crust (as inferred from subsidence) appears to be greatest adjacent to the transcontinental dislocations, but the dislocations are not necessarily reactivated in the sense of later motion on them. Indeed, the Kavala
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Island Ridge (Figures 3.2 & 3.13), a continuation of the Rukwa lineament through the Lake Tanganyika rift zone, appears to have been a zone of resistance to the extension recorded on either side of it, but the lineament has had an obvious influence on the geometry of structures in adjacent half graben, as discussed below in Section 5.2.2.

The intersections of transcontinental dislocations with the rift zones are commonly associated with dip reversals along the strike of the rift zones and with deflections of the rift axis (Table 5.1). For example, the N. Chamaliro dislocation separates the oppositely dipping NW-SE Livingstone and N-S Usisya half graben in the Lake Malawi rift zone and a 100° lineament separates the Enderby Half graben and half graben in the Barrow Sub-basin, which have opposite senses of asymmetry, on the Northwest Shelf. In each case, the transcontinental shear zone or dislocation affects the geometry of linked half graben or linked rift zones.

The evidence accumulated in the Rukwa and Ceduna case studies suggest that repeatedly re-activated zones experience a significant component of oblique slip or transcurrent motion. The orientation of the reactivated structures subparallel to a regional tectonic transport direction supports this contention. Depending on the deformation along the other boundaries of the adjacent lithosphere, transcurrent motion along these large scale trends could result in rotation of the bounded lithospheric block in a manner similar to that proposed for oceanic microplates and outlined below.

Rotation is a natural consequence of moving rigid plates or crustal blocks around on a sphere. Several oceanic microplates have been shown to undergo rapid rotation about a nearby vertical axes as indicated by fanning magnetic lineations. Extensively documented examples include the Easter and Juan Fernandez microplates (Searle et al. 1993). These microplates commonly have angular velocities that are more than an order of magnitude higher than those of the bounding larger plates suggesting that their motion is not controlled by the same balance of forces controlling larger scale plate tectonic motions. The differences in motion between micro-plates and larger plates led Schouten et al. (1993) to develop a kinematic model, termed the "roller-bearing" model, wherein edge-driven shear induces vorticity of the plate, increasing its angular velocity. Microplates in the oceanic setting are normally surrounded by propagating rifts and are thought to be substantially decoupled from the underlying mantle so that edge-driven shear is an attractive mechanism for inducing rotation.
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Block rotations on a continental scale are proposed to explain strike slip faulting in eastern Tibet (England & Molnar 1990). Wilson et al. (1993) presents evidence that the Zambezi Belt was deformed intracontinentally and suggests that rotation of the Kalahari craton could explain the kinematics indicated by mineral lineations. Kamp (1986) requires a 25° rotation of the continental Campbell Plateau in his tectonic reconstructions of the New Zealand, Australia and Antarctica. Thus, a growing body of literature documents rotations of micro-continental blocks, on the basis of paleomagnetic data and other kinematic indicators, in a variety of settings and scales including western North America, Greece, the Aegean Sea, Alaska, Wales, the Pyrenees, Cyprus, the Phillipine Sea, Norway, New Zealand, Ecuador and northern Peru (Kissel & Laj 1989 and references therein). It is suggested that the above mechanism may also be applicable in the intracontinental setting, although decoupling of the upper crust, lower crust and mantle is not well-established.

Although none of the case studies herein conclusively document rotation of large micro-continental blocks, the inferred pole for the Queensland and Townsville Troughs is relatively close to the rift zones, in line with the "roller-bearing" model discussed above. On the Northwest Shelf, the Exmouth Plateau is bounded to the south by the Cape Range Fracture Zone (Figure 4.24), separating oceanic and continental crust (Lorenzo et al. 1991). It's role as a transform fault in the early stages of oceanic crust development and it's extension into the continental crust makes it reasonable to expect some shearing along it's trend during continental deformation and suggests that some rotation of the Exmouth Plateau may also have occurred.

The above discussion suggests that rotations of continental blocks about local vertical axes are common, although the exact mechanism for the rotation is still under debate. The scale of the blocks can range from a few kilometers to hundreds of kilometers (e.g. Jackson & Molnar 1990, England & Molnar 1990, Beck Jr. 1976). In many of these studies, reactivation by oblique slip of pre-existing structures is invoked to set up the boundaries of the blocks (e.g. Scotti et al. 1991, England & Molnar 1990, Jackson & Molnar 1990, Eyal et al. 1986, McKenzie & Jackson 1986). The hypothesis that micro-continent sized cratons may experience rotation about a vertical pole located near their margins or within them, as hypothesized for the Tanzanian craton, surrounded by two rifts characterized by oblique slip (Chapter 3), is receiving ever-growing support in studies of distributed deformation in the continental setting.
5.2.2 Lithospheric Anisotropies and Lateral Heterogeneities

Oblique slip on major bounding faults and oblique slip reactivation of pre-existing structures has been a recurrent theme throughout this thesis. Rotation of continental blocks is a direct result of the prevalence of strike- to oblique slip faulting in the continental setting. Can this kinematic style in the upper crust tell us anything about the geodynamics of deformation of the whole lithosphere in the continental setting? The increasing recognition of oblique slip faulting and rotation of upper crustal blocks during reactivation of extensive continental lineaments (e.g. Ben-Avraham 1992) suggests that tectonic inheritance is a common phenomena and may provide a link between surficial deformation and deep lithospheric processes.

On a continental scale, Phanerozoic extensional strain appears to have avoided Archean cratons and is concentrated the intervening Proterozoic and Paleozoic fold belts. At the low temperatures and pressures of the upper crust, rock strength is thought to be controlled by frictional resistance to brittle failure (Byerlee's Law). However, because the range of strength variability is small compared to the strength of the whole lithosphere, differences within the upper crust should not result in any significant variation of whole lithosphere strength. Thus, if anisotropies or heterogeneities are to have an effect on whole lithosphere extension they must extend well into the lower lithosphere. The repeated focussing of deformation in Proterozoic and Paleozoic fold belts implies that they comprise weaker lithosphere and thus yield evidence for lithosphere anisotropy and lateral heterogeneity. That is, there is a fundamental difference in bulk lithospheric properties associated with particular types of continental crust.

Strain also appears to be focused repeatedly along pre-existing Proterozoic terrane boundaries. In the Lake Rukwa rift zone, the pre-existing structure is oriented subparallel to the Late Tertiary tectonic transport direction. Strain focussing along a similar tectonic boundary is also argued for to explain distinctive features of the Ceduna Sub-basin and its conjugate margin in Antarctica. The Ceduna Sub-basin axial trend is also subparallel to the regional pre-breakup tectonic transport direction, as documented in the Eyre Sub-basin. The Rukwa trend was repeatedly reactivated throughout the Phanerozoic even when extension was oblique to the trend, without continental fragmentation. Taken together these two case studies suggest that the subparallel relationship between a major tectonic boundary and the tectonic transport direction is a necessary precursor to successful continental fragmentation. Although there are large segments of passive margins
that do not appear to include such a tectonic boundary that extends from an oceanic transform into continental crust, there are several other examples (e.g. see listing in Section 5.3) and more may be revealed as altimetry and other geophysical data reveal more of the large-scale geometry of passive margins.

The pervasive intra-rift transverse faults in the Lake Tanganyika and Lake Malawi rift zones trend generally NNW (330°) (Figures 3.7 and 3.8). Geometries of intra-rift half graben adjacent to where the NW (315°) Rukwa trend cuts the Lake Tanganyika rift zone (i.e. the Kavala Island Ridge) and the Malawi Rift Zone (i.e. the Livingstone half graben) are distinctive in that rift transverse faults trend parallel to the Rukwa trend, rather than at the more common 330° trend. The influence the pre-existing structure exerts on adjacent half graben geometry suggests that these tectonic boundaries may act as strain guides for adjacent crustal blocks. On a larger scale, the Eyre Sub-basin fault geometries may result from the Ceduna tectonic boundary acting as a strain guide.

Both the Ceduna and Rukwa trends are punctuated by deeps along their axial strikes. The repeated reactivation of the Rukwa rift zone during distinct extensional events of variably oriented stress regimes argues that the orientation of the boundary is not, in itself, sufficient to explain its reactivation. However, in the Ceduna case of successful fragmentation of the continent, the boundary involved in the final pre-breakup stage was subparallel to the regional tectonic transport direction. And, as stated earlier, this may be a pre-condition of successful fragmentation. Thus, pre-Tertiary deformation in the Rukwa rift zone was unsuccessful, but it's present subparallel alignment with the regional tectonic transport direction may result in continental break-up if the present day stress regime is long-lived. Conversely, the repeated activity along the boundaries may make them so favorable for use (or non-use) in a strength sense that they act as strain guides or barriers to deformation of adjacent lithospheric blocks as is suggested by the Ceduna Sub-basin's apparent role as a boundary between NW-SE Jurassic and NE-SW Cretaceous extension on the southern Australian margin. Confirmation of the hypothesis that these boundaries may act as strain guides for adjacent lithospheric blocks awaits further testing in basins where both the upper crustal regional structural patterns and lateral lithospheric anisotropies or heterogeneities can be clearly identified and related.

The role of tectonic inheritance in lithospheric deformation depends upon the lateral heterogeneity thought to exist in the lithosphere and the depth to which various surface features extend into the lithosphere (Figure 5.4). There is tremendous controversy and debate about the deep structure of the continental
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lithosphere (e.g. Silver et al. 1988). Several recent compilations emphasize the rapidly evolving nature of our understanding of the lower crust and upper mantle components of the lithosphere (e.g. Drummond 1991, Giese et al. 1991, Fountain et al. 1992, Salisbury & Fountain 1990). The expanding fields of seismic tomography and deep reflection seismic data, and advances in isotopic geochemistry and geobarometry, are beginning to allow resolution of some of the salient features of the lower crust and upper mantle. Some further observations pertinent to the question of whole lithosphere heterogeneities are summarized below.

Sutton & Watson (1986) speculated that lineaments which originated as intra-plate transcurrent shear zones or rift systems (e.g. the Rukwa and Ceduna lineaments), with their steep attitudes, probably extend down through the base of the crust into lithospheric mantle. In contrast, lineaments that originated as sutures welding crustal blocks may be oriented at a low angle to horizontal and not connected with mantle structures. Evidence bearing on how deep lateral heterogeneities extend is beginning to accumulate. Deep reflection seismic data indicate that the lower crust and sub-lithospheric mantle are indeed highly heterogeneous (e.g. Drummond 1991, Fountain et al. 1992 and references therein,).

A reinterpretation of gravity data from the boundary of the Mozambique Belt and the Tanzanian Craton suggests a deep crustal root of that suture (Nyblade & Pollack 1992). Seismic and other geophysical data confirm the presence of an asymmetric lithospheric root beneath the Alps, extending under the Po Basin (Giese et al. 1991), adding to the evidence that some extensive lineaments extend deeply into the lithosphere and modify the sub-crustal morphology.

Early seismic wave velocity data indicated lateral density heterogeneity in the lower crust and upper mantle and suggested that continents may have roots up to 400 km deep (Jordan 1975). This idea was challenged by Anderson (1987) who argued that thick continental roots and deep slab penetration were not supported.

Figure 5.4 Plate tectonic model (above, after Woodcock 1986) showing the continental crust characterized by large scale strike slip faults. The "indentor" continent adjacent to the oceanic crust is broadly modelled after the Australian continent. The role of pre-existing strike slip faults in the development of the passive margin rift zones studied are suggested and placed approximately in their real geographic positions. Q = Queensland Trough, C = Ceduna Sub-basin, E = Eyre Sub-basin, Ex = Exmouth Plateau The linked EARS system is imbedded in the interior of the continent. T = Lake Tanganyika rift zone, R = Rukwa rift zone, M=Malawi rift zone.

A hypothetical cross section of the lithosphere based on the relationship of the various rift zones studied herein with pre-existing lithosphere scale heterogeneities. Depths of lower crust reflectivity and base lower crust from Fountain et al. (1992) and references therein. See text for further discussion.
Continental crust back-arc basin (after Woodcock, 1986)

Archean craton
Proterozoic and Paleozoic Fold belts
10-20 km
25-35 km
20-25 km
sub-lithospheric mantle perturbation
35-50 km
80-150 km
20-25 km
35-50 km
250-400 km
Topography on base lower crust or base lithosphere
perturbation in mantle convection cells

(after Woodcock, 1986)
by seismic and geoid data when isobaric phase changes are included in the shear wave velocity analysis. However, as seismic tomography methods have been refined, the relationship of high-velocity anomalies under the continents, implying a deep "keel" has received increasing support (Jordan 1988, Lay 1988, Lerner-Lam & Jordan 1987, Montagner & Tanimoto 1991, Dziewonski & Anderson 1984, Anderson & Dziewonski 1984).

As more data are compiled the lateral heterogeneity of the lower lithosphere has been resolved into smaller domains (Jordan 1988, Montagner & Tanimoto 1991, Anderson & Dziewonski 1984, Dziewonski & Anderson 1984, Anderson et al. 1992, Green et al. 1991). For example, seismic tomography indicates that mantle underlying back-arc basins and continental extension provinces have generally slower velocities than hotspot mantle, possibly reflecting partially molten and/or hydrous mantle, and reinforcing the fact that the mantle is not isothermal (Anderson et al. 1992). Silver and Chan (1988) studied the seismic anisotropy in the Canadian Shield province, and discovered strong shear wave splitting beneath the craton, indicating a "fabric" that appears to align with mineral orientations. The alignment of the rock and seismic fabrics led these workers to hypothesize that this root has been stable for 2.5 Ga.

Seismic tomography beneath the Kenya rift indicates that the lower crust is typified by an axial high velocity zone that varies in width and magnitude, suggesting that crustal modification also varies along the length of the rift (Green et al. 1991). Green et al. (1991) argue that the variation of the lower crust, and a steep-sided low velocity body in the upper mantle suggest that asthenospheric diapirs cause the segmentation seen in the rifts rather than lateral heterogeneities in the upper crust. This model agrees well with a similar proposal for asthenospheric instabilities causing the segmentation of the obliquely spreading Reykjanes Ridge (Murton & Parson 1993). However, the common association of pre-existing structures and segmentation of rift zones, often coinciding with the location of magmatic activity, and the documented anomalous mantle on either side of the Ceduna lineament suggests that heterogeneities or structure of the lower crust may have some influence on the initiation of sub-lithospheric mantle instabilities. Additional support for this idea comes from the unequivocal evidence that the disposition of repeated magmatic activity, with both upper mantle and crustally contaminated isotopic signatures, is controlled by pre-existing lithosphere structures throughout the African continental plate (Bailey 1992, Wilson & Guiraud 1992).
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Many studies, particularly those investigating Archean and Proterozoic systems, have noted the persistent reactivation of major structures which extend laterally from a few hundred to a thousand or more kilometers (e.g. Reading et al. 1986 and references therein). White et al. (1986) suggest that the continental crust, especially PreCambrian basement, is so fractured that the orientation requirements are commonly met and cite examples of near vertical strike slip zones reactivating as oblique reverse-slip (the Alpine fault of New Zealand) and as inclined normal faults (the Darling Mobile Zone) without creating a new fault trace. However, because the fault plane must depart from the inherited structure over most of its down-dip extent in these examples, the relationship between reactivation of structures at shallow levels with deeper processes is not clear. Nonetheless, the apparent widespread occurrence of oblique rifting and the role of pre-existing structures and lithospheric heterogeneities in localizing deformation and influencing rift zone geometries argues strongly for extensive structuring deep in the continental lithosphere.

The concept of tectonic inheritance, that is, reactivation of pre-existing tectonic boundaries in the intracontinental setting and repeated deformation of particular lithosphere types (e.g. Proterozoic and Paleozoic fold belts), is supported by the long-lived and repeated deformation suggested by the EARS, Ceduna and Queensland Trough case studies. The case studies herein support the proposal that the location of stress accumulation is dependent upon large scale whole lithosphere anisotropies and deep-rooted transcontinental structures and that the style of rifting is largely controlled by the type of pre-existing weakness in the lithosphere (e.g. Dunbar & Sawyer 1988, 1989). In all cases, extension seems to avoid cratons and affect the pre-deformed edges of cratons or intervening terranes, particularly terrane boundaries. The pre-existing large scale structures and anisotropies are generally not located nor aligned appropriately to the eventual pole of rifting/spreading and are therefore predicted to be deformed by oblique extension (Smith 1993). The repeated use of non-ideally aligned lithospheric anisotropies and the reactivation of continental scale structures in later pulses of deformation:

- explains the preponderance of oblique extension suggested by the results of this dissertation and
- predicts maximum structural complexity.
5.3 Plate Tectonic Models

One of the major deficiencies of the current plate tectonic theory is that it does not adequately explain intracontinental deformation (Chatterjee & Hotton 1992). Evidence for repeated and geologically prolonged reactivation in the exclusively intracontinental Rukwa rift zone has been presented. Interpretation of recently released satellite altimetry data, and its bearing on the breakup of Australia and Antarctica, has led to the proposal that the Ceduna Terrace deeps are a remnant analog of the Rukwa system in the passive margin setting of the Great Australian Bight. The "bent", as opposed to curvilinear, configuration of the ends of other transform faults which appear to extend into continental crust, such as the Dampier Fault Zone in the Tasman Sea, the Atlantis and Tyro fault zones in the North Atlantic and the Ascension and Bode Verde Fault Zones in the South Atlantic (Gahagan et al. 1988), suggest that the influence of tectonic boundaries during later deformation is fundamental. In the Queensland Trough, Cretaceous reactivation of a probable Paleozoic rift, formed in a Late Proterozoic to Early Precambrian fold belt, suggests that localization of deformation is dependent on whole lithosphere anisotropies or heterogeneities. Variation in lithosphere thickness, resulting in lateral gradients in lithosphere strength, has been hypothesized for the disposition of the rift zone that includes the Enderby Half graben. Clearly, the interaction between sub-lithospheric mantle processes and crustal variation is key to understanding intracontinental deformation.

Coupling between the three elements (i.e. the upper crust, lower crust and upper mantle) of the lithosphere depends on the vertical heterogeneity of the lithosphere which results from the poorly understood, complex interactions of compositional and phase changes and thermal and mechanical boundaries. However, evidence is beginning to accumulate which suggests that there is a significant connection between upper crust and mantle features. For example, Reston (1993) interprets reflectivity in the uppermost mantle, in reflection seismic data from offshore Britain, to be mantle shear zones. The data demonstrate a connection of the mantle reflectivity with surface faults. Structure at Moho levels and deeper is observed in nearly all deep reflection seismic data that have been acquired (e.g. Rosendahl et al. 1992b, see also references relating to BIRPS, MOIST etc., LITHOPROBE, EUROPROBE and COCORP data). And, there appears to be a correspondence between large lateral gradients in upper mantle seismic velocities and transition in continental crustal type (Jordan 1988, Fountain et al. 1992).
The question of upper mantle convection has been an "enduring paradox" (Silver et al. 1988) and well beyond the scope of this study. It is difficult to argue with the increasing evidence for heterogeneity in the upper mantle, but as yet we have no clear picture of the flow, flux or motion therein, nor is there a good understanding of how that is coupled to the crustal components of the lithosphere. There are arguments for and against small scale convection (cf. Parsons & McKenzie 1978, Davies & Richards 1992). Upper mantle convection, if it exists, is thought to be driven by density gradients between spreading ridges and subduction zones (Officer & Drake 1983) or by channel flow of the upper mantle avoiding the continental "keels" (Alvarez 1982, 1990).

Determining the coupling of any inferred flow of the upper mantle lithosphere to the crust requires an understanding of plate driving mechanisms. Determining plate driving mechanisms has also proved elusive (Wilson 1993). Plate motion models change and are still evolving (e.g. the NUVEL-1 model of DeMets et al. (1990) rotated some plate motion vectors by 10°-15° from previously accepted models). The geochemical and geobarometric evidence that not all hotspots originate at the same depth calls into question their use as a framework for absolute motion (Wilson 1993). Further, inferences made about plate driving forces from kinematics, which result from the integrated effect of all the forces acting on it, are suspect because the relative contribution of each force is still unknown. Forces that are thought to play a role in plate motion include the plate boundary forces, the so-called slab-pull, trench suction and ridge push, and plate-wide shear traction exerted by an asthenospheric "flow". Although early attempts to model the relative contributions of various forces seemed to yield definitive ratios that were consistent with what was known of plate kinematics at the time (Forsyth & Uyeda 1975, Backus et al. 1981), opinions regarding the relative contributions of each force still vary depending on the approach to the problem.

The lack of correlation between plate velocities and surface area (e.g. the relatively fast movement of India and slow Caribbean movement) have been used to argue against a significant contribution by shear traction (Gripp & Gordan 1990). Likewise, an attempt to estimate plate torques for the Cenozoic Era assuming a dynamical balance between active torques (slab-pull and ridge-push) and passive torque (plate drag) suggests no shear-traction is necessary to drive the plates (Jurdy & Stefanick 1988). However, a misfit of the balanced active torque and the drag torque occurs, increasing systematically with age, which may be ameliorated by including shear traction in the model. Modelling of the resistive and driving forces of the South American plate assuming it is in mechanical equilibrium led Meijer and Wortel (1992) to conclude that the concept of a basal shear stress actively
driving plate motion cannot be discarded. The recent compilation of the state of stress in the lithosphere has shown that the intraplate regions (primarily continental as data from oceanic crust is exceedingly sparse) are characterized by a compressional stress regime (Zoback et al. 1989, Zoback 1992). Maximum horizontal stress is subparallel to the direction of absolute plate motion suggesting that forces driving the plates also dominate the stress distribution (op. cit.). Thus, investigators whose emphasis is consideration of Phanerozoic plate movements, particularly the assembly and break-up of Pangaea, seem to favor shear traction as an important, possibly dominant plate moving mechanism (Ziegler 1992b, 1993, Pavoni 1992, 1993).

Finite element modelling of the deviatoric stress produced in the strong upper layer by sub-lithospheric loading caused by density contrasts in the asthenosphere suggest that slab-pull is two to four times greater than ridge push (Whittaker et al. 1992) and that plate boundary forces originate where such stressed regions are intersected by zones of weakness (Bott 1992, 1993). This is a particularly interesting result in that it requires a lithospheric heterogeneity to initiate plate boundary forces. On the other hand, a similar analysis of tectonic stress in subduction zones led Richardson (1992) to propose that the slab forces are largely balanced within the subducted slab itself so that it has little effect on intra-plate deformation, elevating the importance of ridge-push. Thus, we are still plagued by a confusing array of arguments for the relative contribution of each force to plate motion.

The extent of the Rukwa rift zone, and its repeated reactivation, suggests that this feature is likely to extend deep into the lithosphere. Sheared zones are commonly associated with thickened crust, which implies a weaker lithosphere (Vink et al. 1984) and perhaps a more intimate coupling with deep lithospheric processes. It is proposed that these zones are the foci of stress accumulations in the continental lithosphere, act as strain guides and play a crucial role in modifying upper mantle processes, initiating intraplate deformation and influencing upper crustal deformation geometry (Figure 5.4). This in turn, supports the idea of an upper mantle flux actively contributing to plate motion. The results of the case studies cannot differentiate between the two currently proposed models of upper mantle processes discussed above (i.e. small-scale convection versus channel flow). However, the anomalous mantle bracketing the Ceduna trend may be an indication of an intimate coupling between these deep-seated structures and sub-lithospheric mantle processes. Lower crust morphology may set up small-scale convection cells, or may act as conduits or barriers to "channel" flow as proposed by Alvarez (1982, 1990).
The results of this research and growing evidence from other sources argue strongly that tectonic inheritance exerts an important influence during intracontinental deformation. Studies refining our understanding of reactivation of major lithospheric lineaments can provide important constraints on continental lithosphere geodynamics, including deformation and plate tectonic models.
Chapter 6

6.0 CASE STUDY CONCLUSIONS AND APPLICATION OF
RESULTS TO BASIN ANALYSIS

Nothing in science has any value...
if it is not communicated.
Anne Roe

Well, I've gotten to the end of the subject---
of the page---
of your patience and my time.
Alice B. Toklas

6.1 Summary of Case Study Conclusions

The primary conclusion to be drawn from the case studies in Chapters 3 and 4 is that oblique extension of continental crust is common. As broadly defined in Chapter 1, oblique extension is divergence of continental crust oblique to the axis of a rift zone. This research has attempted to quantify the structural and regional kinematic relationships within oblique extension rift zones. The results refine the definition of oblique extension by relating detailed structural patterns to kinematic interpretations of individual rift zones and by integrating oblique rift zone geometries with regional structural relationships. The previous chapter establishes the fundamental role of pre-existing crustal structures and lateral heterogeneity of the lithosphere in oblique extension. Results of the case studies provide observational constraints on intra-basin structural and kinematic relationships that have implications for several aspects of basin analysis.

The intra-rift structural relationships from each of the rift zones studied during this research are summarized in Table 6.1. Fault geometries within obliquely diverging rift zones are characterized by two primary fault sets, here termed rift bounding faults (BF) and rift transverse faults (TF), which are commonly not orthogonal to each other. Both rift bounding faults and rift transverse faults strike obliquely to the regional rift axis (RA). Transverse faults can be grouped into two categories: 1) intra-basin transfer faults that trend subparallel to the tectonic transport direction and 2) reactivated transcontinental dislocations (possible sites of rift zone nucleation termed a Transverse Graben Zone by Versfelt, 1988).

Rift bounding fault trends can be highly variable within a single rift zone (e.g. the Lake Tanganyika and Lake Malawi rift zones and the Queensland Trough). Bounding fault trend variability is highest when rift zones are initiated in highly folded, metamorphosed "mobile belts" (usually Proterozoic or Paleozoic) that trend
at high angles to the regional tectonic transport. The complexity of rift zone structures in this setting is a direct result of the variability of structures in adjacent basement terranes. Often the rift zone "axis" is also quite variable in this setting (RA in Table 6.1). In this case, the trend of bounding faults of individual dip domains may be controlled by local pre-existing structures or fabric (e.g. the Usisya and Livingstone Border Faults, Lake Malawi, Table 6.1) so that an en echelon pattern is not apparent and offsets of the bounding fault in plan view due

**Table 6.1 Rift Zone Geometry**

<table>
<thead>
<tr>
<th>Rift Zone</th>
<th>Bounding Faults (angle to RA)</th>
<th>Transverse Faults (angle to RA)</th>
<th>Hanging wall Dips (dip angle to TF)</th>
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</thead>
<tbody>
<tr>
<td>Lake Tanganyika</td>
<td>(variable, majority oblique)</td>
<td>(30°-45°) KIR parallel to Rukwa lineament</td>
<td>(5°-38°1'-30°) oblique to RA and to most bounding faults</td>
</tr>
<tr>
<td>(000°, variable)</td>
<td>(variable, majority oblique)</td>
<td>(15°-45°) 45° trend associated with N &amp; S branches of Chamaloro dislocation</td>
<td>(4°-28°1'-30°) oblique to RA and to most bounding faults</td>
</tr>
<tr>
<td>Lake Malawi</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(000°, variable)</td>
<td>(variable, majority oblique)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lake Rukwa</td>
<td>(0°) linear, reactivates</td>
<td>(25°) E. Tertiary</td>
<td>(4°-48°10°) oblique to RA (bimodal?)</td>
</tr>
<tr>
<td>(135°)</td>
<td>Proterozoic terrane boundary</td>
<td>(0°) L. Tertiary</td>
<td>in E. Tertiary, (4°-26°10°) parallel to RA in L. Tertiary</td>
</tr>
<tr>
<td>Queensland Trough</td>
<td>(variable) possibly reactivates Paleozoic Tasman Fold Belt structural trends</td>
<td>(45°) parallel to major onshore gravity trends</td>
<td>(12°-38°10°') oblique to RA</td>
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<tr>
<td>(155°)</td>
<td></td>
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<td></td>
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<tr>
<td>Eyre Sub-basin</td>
<td>(35°) aligned with mid-Paleozoic Albany - Fraser Fold Belt structural trend</td>
<td>(-80°) - perpendicular to most normal faults</td>
<td>(8°-38°10°-25°) oblique to RA - perpendicular to most normal faults</td>
</tr>
<tr>
<td>(090°)</td>
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<tr>
<td>Ceduna Sub-basin</td>
<td>(0°-20°) reactivation of major tectonic boundary</td>
<td>(70°) Cretaceous transfer faults - perpendicular to RA</td>
<td>dip analysis not performed because basement rarely imaged</td>
</tr>
<tr>
<td>(135°)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Enderby Half-graben</td>
<td>(0°-5°) aligned with regional compressive structural trends</td>
<td>- perpendicular to normal faults, oblique to pre-existing crustal shear zones</td>
<td>(7°-38°10°-30°) majority slightly oblique to RA and to normal faults</td>
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<tr>
<td>(045°)</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>West Exmouth Plateau</td>
<td>(0°?) oblique to adjacent magnetic lineaments</td>
<td>perpendicular to normal faults, reactivates pre-existing crustal shear zones (100° lineaments)</td>
<td>(3°-38°1&lt;10°) perpendicular to RA and to normal faults</td>
</tr>
</tbody>
</table>

* \* computed from apparent dips on time sections
RA, Rift axis TF, Transverse fault
UBF, Usisya Border Fault LBF, Livingstone Border Fault KIR, Kavala Island Ridge
to transverse faults may be interpreted as an apparently curvilinear fault trace, especially if data are sparse.

In rift zones where bounding fault trends are consistent, but offset by the transverse faults in a regular, but not necessarily en echelon (sensu stricto), fashion in plan view, the bounding faults' obliquity to the rift axis reflects non-alignment of the rift axis with a great circle passing through the pole of rotation (e.g. the Enderby Half graben and the Eyre Sub-basin). In both the variable and consistent bounding fault trend scenarios, intra-rift normal faults may trend obliquely to the bounding fault trend.

When a rift zone is localized along a pre-existing terrane boundary or shear zone that trends parallel to the regional tectonic transport direction (e.g. the Rukwa rift zone and Ceduna Sub-basin) the bounding fault reactivates the pre-existing, usually planar, structure and consequently is itself quite planar and may not be offset in plan view. In such cases, intra-rift normal faults often comprise two sets. One set is subparallel to the bounding fault while the other set's trend is dependent on the relationship of the tectonic transport direction to the bounding fault trend and may be at a high angle to it. The latter set may occur as splays off of the bounding fault (e.g. the Rukwa rift zone). Whether or not one fault segment is coincident with the reactivated shear zone or just subparallel to the trend, a pair of intra-rift linked normal fault segments often form a "corner" that constrains the horizontal movement of the hangingwall along some bisector of the angle made by the fault pair. When extension is limited (i.e. lithospheric rupture does not occur), these "corner structures" are preserved and may be reactivated in later pulses of deformation even if the tectonic transport direction is different than that during formation of the corner structures. In fact, in repeated extensional phases, they are more likely to be reactivated than fault systems of a single trend (Figure 5.2). Thus, these "corner" segments of rift zones are long-lived and are characterized by semi-isolated deeps due to repeated reactivation of intra-basin corner faults.

Rift transverse faults that reactivate pre-existing dislocations are normally associated with deeps, magmatic centers, and changes in the rift axis trend. They are the major cause of rift segmentation (Table 5.1). They commonly group bounding fault segments into rift zone segments having broadly consistent bounding fault trends and control the geometry of adjacent half graben. All known half graben polarity reversals occur across these structures. The common occurrence of the greatest subsidence in half graben adjacent to these transverse structures suggests that deformation of the upper crust initiates here. An effect of pre-existing structures is to give rise to triaxial rather than plane strains on the
scale of the upper crustal rift zone, leading to additional sets of faults and more complex fault geometries (Figure 5.1).

Intra-rift transverse faults, or transfer faults, are generally subparallel to the inferred tectonic transport direction and have relatively consistent trends throughout the rift zone, regardless of the bounding fault trend of individual half graben. The transfer fault trend may be influenced locally by pre-existing dislocations that trend subparallel to the tectonic transport direction, but cut dislocations that are at a high angle to the tectonic transport direction (e.g. see Figure 3.8).

The direction of hangingwall dips is quite consistent throughout obliquely extending rift zones even though the bounding fault trends can be highly variable. Dips tend to be low (e.g. 5°-25° as computed from time sections, Table 6.1). Combining dip analysis with traditional methods of structural interpretation provides a better understanding of the three dimensional relationship between deformation and kinematics.
6.2 Basin Analysis

The relationships described in the previous section and in Chapter 5 have several important consequences with regard to basin analysis techniques. Reflection seismic data are the most common data type used in basin analysis. Normally, data grids are laid out orthogonally, with one profile set trending parallel to and the other trending perpendicular to the basin elongation. As these case studies document, rarely will these profile trends provide "dip" and "strike" structural cross sections. To determine whether a basin is the product of oblique extension an interpretation should start by asking the following questions:

• Can rift-axis-perpendicular profiles be balanced in two dimensions?
• Are hangingwall dips consistent with bounding fault dips?
• Are adjacent intra-rift blocks imaged on a profile perpendicular to the rift axis tilted by the same amount?

If the answer to any of the above questions is no, it is highly likely that the basin-forming structures are the product of oblique extension.

It is also common practice to produce depth-to-horizon structural contour maps on the basis of reflection seismic data. As the case studies of passive margins have highlighted, this type of interpretation often masks the true dips of individual hangingwalls and can be misleading. If horizon contours trend at high angles to the interpreted fault, it is likely that unidentified structures exist. Dip analysis (Chapter 2) can be a useful tool in the following aspects of basin analysis:

• constraining regional kinematics (tectonic transport direction);
• planning future data acquisition;
• constraining normal fault correlations;
• identifying transfer faults or lateral offsets in normal faults;
• identifying seismic artifacts;
• constraining kinematic histories in basins subject to polyphase deformation;
• constraining bulk deformation mechanisms of the hangingwall;
• identifying the location of major pre-existing structures.
In oblique extension, bounding faults are poor indicators of tectonic transport direction, whereas hangingwall dips are generally subparallel to it. Thus, if an original reconnaissance data grid is available, dip analysis at profile intersections can provide constraints on appropriate alignment of future profiles, to provide true structural dip and strike cross-sections of the intra-rift deformation.

When a seismic grid has been shot in the traditional manner, dip analysis can provide important constraints on fault correlations. Intra-rift hangingwall dips that do not trend directly toward bounding faults indicate that oblique slip has occurred on them. An interpreter is signalled to look for intra-rift normal fault trends that are more perpendicular to the trend of reflection dips. In widely spaced data, fault correlations of this trend may not be apparent so that the interpretation may require lateral offsets in the normal fault traces at frequently unrecognized intra-rift transfer faults.

In the seemingly rare instances that orthogonal, rectilinear extension (or nearly so) occurs in a package whose bulk properties are near isotropic as shown in the accommodation zone studies (Section 3.2.4), transfer faults are nonetheless exceedingly difficult to identify and correlate as specific structures in seismic data (e.g. the Eyre Sub-basin and Enderby Half graben). The incorporation of precise dip information provides a method for predicting their location and should be used as a standard tool in basin analysis, especially when "two dimensional" reflection seismic data are the primary source of information. Transfer faults in seismic data are commonly characterized by vertical disruptions of reflections across which reflection dips change (Figure 6.1). Sometimes vertical offset of reflections is evident but dips of adjacent reflections are not consistent with the structure. That is, they often dip towards or away from each other at the structure or have highly different tilts.

Dip analysis can also help to identify seismic artifacts. For example, in the analysis of the Enderby Half graben, it is tempting to correlate the dipping reflections between BES and D reflectors on the strike line (Figure 4.26) with fault plane reflectors on the dip lines (Figure 4.25). The dip analysis clearly shows that the hangingwall reflections do not always dip directly towards the normal faults. However, the deviation of hangingwall dips is to the north, in the opposite sense to that predicted by the geometry of the fault correlation made if these reflections are interpreted as fault planes.
Chapter 6

Conclusions and Basin Analysis

The corner structures formed by obliquely linked fault pairs can be reactivated repeatedly by a wider range of stress orientations than single faults of a particular orientation (e.g. the Rukwa rift zone and Figure 5.2). Dip analysis on a series of key stratigraphic horizons should be included when modelling basin development, because repeated activity on such fault pairs, responding to varying stress orientations through time, is reflected in changing hangingwall dips at different stratigraphic levels. This is especially relevant for determining pre-breakup kinematics in passive margin rifted basins where repeated deformation prior to and after break-up may occur. Dip analysis of a series of stratigraphic horizons can constrain the timing and changes in the direction of extension when inferring the kinematic history of rift basins. Establishing the kinematic history also has important consequences for the thermal history of the basin.

Most major boundary faults are probably best described by an irregular surface, given the role of pre-existing structures and that triaxial strain is most probable in natural settings. The dip data in this thesis provides evidence that the majority of the bounding faults can best be described by some variation of the corner structures as described in Chapter 2. As noted previously, if the relative components of dip versus strike movement on a fault in a given pair reverses from one pulse of deformation to the next, rotation of the hangingwall block about a vertical axis is predicted. Detailed dip analysis at different stratigraphic horizons may be combined with paleomagnetic data, if available, to constrain rotation phenomena in the basin setting.

The empirical models developed in this dissertation have been tested by scaled experimental analog models of extension accommodated on non-ideally oriented faults (Braun et al. in press). These experiments have reproduced the dip patterns recorded in natural settings and have led to the development of a new, simple mathematical description of the bulk internal strain of hangingwalls in extensional settings (op. cit.). Continued accumulation of precise dip data and information on fault patterns can further constrain the quantitative descriptions of three dimensional bulk strain in these settings.

The case studies herein establish that pre-existing lithospheric structures and heterogeneities exert a substantial control on the geometry of rift zones and the basins within them. Dip analysis clearly defines dip domains. Recognition of the pre-existing large scale structures is crucial in basin analysis, especially when analyzing a series of linked sub-basins. Detailed dip analysis of upper crust extension can help identify them, providing pointers for interpretations of geophysical data that can confirm their deep-rooted nature. The fact that the
intersections of rift zones and transcontinental dislocations are frequently associated with magmatic activity suggests that identification of them can also aid in reconstructing the thermal histories of basins. It is clear from the examples studied that any realistic attempt to model rift basin development must include pre-existing regional structural relationships and the effect of lateral lithospheric heterogeneity.

**Figure 6.1** A series of idealized fault/reflection dip geometries across transfer faults (A-F) and bounding faults (G-H). The existence of any dip on reflection horizons adjacent to a vertical structure is inconsistent or "anomalous". Reflection offsets are arbitrary in these cartoons.

A) The ideal strike profile geometry. Transfer faults are vertical. Reflection dips are horizontal because hangingwall dips are directly toward normal faults which are orthogonal to the transfer faults (e.g. S2 structure in Line TXZ-16, Figure 3.22; Line 117, Figure 4.26). Even in this setting, if a profile crosses a transfer fault obliquely, it may not appear vertical. This is also true in all of the following cases, but in all cases, the dip of the hangingwall will most commonly not be consistent with the apparent dip of the fault.

In natural settings transfer faults may not be strictly vertical, nor orthogonal to normal faults. B) A "V" geometry occurs when a transfer fault separates hangingwalls bounded by different normal fault segments with different geometries, such as opposite dip direction or faults with substantially different dips (steepness) or strikes. Tilted hangingwalls adjacent to vertical structures usually indicate the profile is crossing a transfer fault obliquely, but can also occur due to differences in normal fault plane segment geometries. Examples from the seismic data presented in previous chapters include: Fault A, Line 8311, Figure 3.6; Lines 224 and 214, Figure 3.14; Line 19, Figure 4.18.

C) and D) are hybrid geometries which may represent a variation on the "V" geometry (B) or unequal tilt geometry (E). Which class such a geometry represents can usually be determined by balancing local fault geometries and precise hangingwall dips in the third dimension with the imaged structure. Examples: C) Lines 214 and 224, Figure 3.14; Line TXZ-49, Figure 3.23; D) Line 6, Figure 4.17; Line 104, Figure 4.18; Line 117, Figure 4.26.

E) Unequal tilts of the same sense on adjacent blocks are common when a profile crosses both normal and transfer faults (e.g. SW end of Line 214, Figure 3.14; Line 7, Figure 4.3).

F) Tilts away from a transfer structure can also occur depending on the orientation of the profile relative to adjacent normal faults. The inverted "V" geometry commonly occurs across a half graben polarity reversal and gives rise to apparent full graben profile geometries.

G) Hanging wall dips adjacent to bounding faults can help constrain the dip of the fault if the tectonic transport direction is known. Horizontal reflections adjacent to faults with large normal offset usually indicate oblique to strike slip motion on a subvertical fault which is likely to be a connecting "jog" in the bounding fault (e.g. Line 908, Figure 3.6).

H) Unusual reflection disruptions and tilts adjacent to bounding faults often characterize the "corners" where transfer faults link with bounding faults (e.g. Line 19, Figure 4.18).
Fault/reflection-dip geometries

a. 

b. 

c. 

d. 

e. 

f. 

g. 

h.
6.3 Conclusion

This study of oblique lithospheric extension has provided as detailed a structural interpretation as is possible with current data from a variety of rifted basins (Table 6.1). The geometries revealed add to the evidence that oblique extension is a common phenomenon in the continental crust and place constraints on models of oblique divergence. Pre-existing structure is the principal cause of oblique extension and exerts a substantial control on rift zone geometries (Tables 5.1 and 6.1) and must be taken into account when tying rift zone geometry to regional kinematics. Dip analysis is an effective tool, when combined with traditional structural analysis, for constraining regional kinematics and for identifying separate pulses of deformation. Pre-existing structure (i.e. transcontinental transcurrent faults, terrane boundaries) can focus strain and control rift axis trends, which tend to be quite linear because their bounding faults reactivate planar structures. Lithospheric anisotropies (e.g. Proterozoic fold belts between Archean cratons) can also focus strain and in this case rift axes can be variable within a single rift zone. In either case, the rift axis will most likely not be perpendicular to the direction of extension. This is also true of the bounding faults in both cases. Intra-basin geometry can be significantly different than the gross bounding fault geometries, and is usually a better indicator of the extension direction. Separating the two geometries requires knowledge of the rift zone geometry in the third dimension, the pre-existing crustal heterogeneities and lithosphere anisotropies.
APPENDIX I (Chapter 4)

Petrographic report

Petrographic report on cuttings from two wells drilled in the Enderby Half graben, Northwest Shelf, Australia (Figure 4.27).

ENDERBY-1

This sample is a fine-grained, aphyric volcanic or a high level intrusive of andisitic or basaltic composition with "trachytic texture" defined by plagioclase lathes. The largest grainsize approaches 0.5 mm and crystals tend to have elongate rather than equant habits. Free quartz is absent. Relict plagioclase and proxene (opx + cpx?) are heavily altered to clays, including sericite and chlorite, and opaque minerals (e.g. magnetite). Possible zircon identified. The thin section contains a thin (approximately 0.5-1.0 mm wide) vein consisting of quartz, feldspar and chlorite.

HAMPTON-1 (2540-2545m)

This sample contains chips of what appear to be related volcanic material. All chips consist of a very fine-grained interlocking network of remnant pyroxene and plagioclase which have broken down to a variety of clays, opaque minerals, and iron oxyhydroxides. The mineral present in many chips as apple-green clots is probably epidote.

Textures vary considerably between different fragments, and quenching is evident in some by the presence of spherulites, acicular crystal form, and devitrified glass. These grade to more even-grained textures, probably indicating a history involving somewhat slower cooling.

A number of chips contain large (up to 1mm diameter) blocky crystals surrounded by dark, fine-grained reaction halos. The low relief, high clarity, and low birefringence suggests that these are quartz phenocrysts (xenocrysts?). One chip contains a once-hexagonal grain showing strong embayment. Fine "hair-like" veining in some chips is dominated by quartz of highly undulose extinction. Groundmass is altered to chlorite and sericite or clays and carbonate.

Majority of fragments appear to be either volcanic or subvolcanic, NOT chert as originally described, as evidenced by:

i) porphyritic texture
ii) high-T quartz phenocrysts
iii) embayed quartz due to rapid cooling
iv) fine feldspar dominated "trachytic texture" matrix.
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Oblique Lithospheric Extension


DATA ARCHIVE LOCATIONS

All seismic data used in this thesis was provided free of charge by outside institutions on condition that the data, including my interpreted sections, be returned to the lending institution. Interested parties are directed to those institutions to obtain copies or to view interpreted sections as listed in the following table:

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