The Structural Style Of Intraplate Deformation, Central Indian Ocean Basin.

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Declaration

I declare that this thesis is my own work except where otherwise stated.

J. M. Bull
Abstract

The intraplate deformation in the Central Indian Ocean Basin is a well known example of a deviation from an axiom of plate tectonics: that of rigid plates with deformation concentrated at plate boundaries. In this thesis a range of geophysical data collected during CHARLES DARWIN Cruise 28, together with several different geodynamic modelling techniques, is used to investigate the structural style of intraplate deformation.

The deformation occurs in lithosphere of age 65 - 90 Ma that formed at a fast spreading-rate (~60 mm yr\(^{-1}\)) at the south-east Indian Ridge. It manifests itself as a diffuse zone of compressional and strike-slip earthquakes, high localised heat-flow, geoid anomalies and tectonic deformation. The tectonic deformation can be considered to be occurring on two spatial scales: the first order is represented by long wavelength (100 - 300 km), large amplitude (1 - 2 km) undulations of oceanic basement and overlying sediments; and the second order by unusual high-angle reverse faults with associated folds in the basement cover.

The reduction and inversion of disposable sonobuoy data revealed that the velocity-depth structure of the oceanic crust is not unusual. The sediment velocity increases from 1.6 - 1.7 km s\(^{-1}\) in the near surface to 3.4 - 3.5 km s\(^{-1}\) immediately above basement with a velocity gradient of 0.75 s\(^{-1}\). A velocity for the top of the oceanic layer 2 of 4.1 km s\(^{-1}\) was identified as layer 2A. An estimate for the thickness of the crust could not be made from the sonobuoys and other refraction work is contradictory.

A study of structural style from single-channel seismic reflection profiles revealed that the reverse fault fabric, which has a strike (090° - 100°E) perpendicular to that of fracture zones (005° - 010°E) developed in this area, resulted from the reactivation under compression of two sets of spreading-centre formed normal faults. In the survey area the first and second orders of deformation are discontinuous across fracture zones. A transpressive model is developed for the survey area to explain the discontinuity of the axes of the undulations, other basement trends, and regional seismicity studies.
Multichannel seismic profiles collected during the CHARLES DARWIN cruise were commercially processed to stack and migration. These profiles imaged the reverse faults penetrating down to the expected level of the oceanic Moho and raised the possibility that original spreading-centre formed inward-dipping faults have higher dip (40 - 45°) than outward-dipping faults (30 - 40°). The average dip of ~40° is broadly in agreement with seismicity studies along mid-ocean ridges. The recognition that the lower part of the fault planes may not have been formed at the spreading-centre, but during the later deformation, leads to the proposition of nucleation at the brittle/ductile transition, propagation upwards and the reactivation of the original spreading-centre formed fabric in the upper crust. Hydrothermal circulation may have been an important component in controlling reflector impedance. Estimates of the tectonic shortening rate from the multichannel profiles suggest that the long wavelength undulations contributed little (<0.2 mm yr⁻¹) whereas reverse faulting is estimated to have produced an average shortening rate of 3.6 ± 1.2 mm yr⁻¹. The total shortening rate of 3.8 ± 1.2 mm yr⁻¹ is consistent with the most recent plate motion study predictions for the Central Indian Ocean Basin.

Previously, two hypotheses had been advanced for the formation of the long wavelength undulations: buckling and inverse boudinage of the lithosphere. This study used three modelling techniques to determine the deformation mode, two of which decisively favoured buckling. Physical modelling using the sandbox technique, in which the oceanic lithosphere was modelled with a two layer (brittle/viscous) rheology, suggested that buckling of the entire brittle lithosphere was responsible for the formation of the long wavelength undulations. Two and three-dimensional gravity modelling supports buckling of at least the oceanic crust. However, numerical modelling was unsuccessful in modelling the long wavelength undulations.

The transpressive model for the Central Indian Ocean Basin, in which buckling forms within fracture zone compartments, and strike-slip motion occurs both along fracture zones (left-lateral) and along the reactivated ridge-parallel fabric, may be extended eastwards to the Wharton Basin. Compressive forces with anticlockwise rotation are seen as a consequence of a dramatic change in plate boundary activity in the north-eastern Indian Ocean from 7 Ma ago, with subduction occurring faster at the Sunda Trench than continental collision at the Himalayas.
Acknowledgements

I would like to thank my supervisor Roger Scrutton for all his advice and encouragement during my PhD, and pay tribute to his vision in creating the project and to his powers as a motivator.

I thank the Master, officers and crew of the CHARLES DARWIN and the engineers of RVS (Barry) without whose expertise at sea this project would not have been possible. Thanks also to Pat Condon and Richard Hillis for acting as able watchmen during the Central Indian Ocean Basin cruise.

During my three years in Edinburgh I have had useful discussions with, and would like to thank, Mike Johnson, Bob Cheeney, Roger Hipkin, Ian Main and Richard Hindmarsh. Without Richard’s help in particular much of my rheological modelling and analytical work would have been thwarted by mathematical intricacies and apparent peculiarities. I thank Maya Tolstoy for helping with the early reduction of the sonobuoys as part of her undergraduate project.

An essential element of the project was the integration of pre-existing data into the CHARLES DARWIN dataset. I thank Jeff Weissel and Jim Cochran for allowing access to the Lamont-Doherty Geological Observatory profiles and for their hospitality during Roger’s and my brief visit.

One of the most awkward and cumbersome processes during my PhD was the conversion of multichannel seismic data from analogue to digital form. This was made substantially easier by the expertise of Charlie Fyfe of the Global Seismology Group of the British Geological Survey (BGS), Murchison House, Edinburgh. The facilities for the data conversion were provided by BGS which was greatly appreciated.

I am very grateful to my French colleagues Joseph Martinod, Phillipe Davy and Peter Cobbold for making my time in Rennes successful, and I am indebted to Isabelle Coat and Marc Audebert for their friendship and excellent hospitality.
I thank Martin Bott for allowing access to his finite element program and for his advice on how to produce buckling. His honesty in admitting puzzlement at our lack of success was appreciated.

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# Contents

<table>
<thead>
<tr>
<th>Declaration</th>
<th>I</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abstract</td>
<td>II</td>
</tr>
<tr>
<td>Acknowledgements</td>
<td>IV</td>
</tr>
</tbody>
</table>

## Chapter 1 - Introduction

1.1 Plate Tectonics, Rigid Plates and Intraplate Deformation | 1  
1.2 Central Indian Ocean Basin | 3 
1.3 Intraplate Deformation in Geophysical Datasets  
1.3.1 Anomalous Seismicity | 5 
1.3.2 Scales of Tectonic Deformation Revealed by Seismic Data | 9 
1.3.3 Regional Heat Flow and Localised High Values | 11 
1.3.4 Geoid and Gravity Anomalies | 12 
1.4 Afanasy Nikitin Seamount | 12 
1.5 History of the Indian Ocean and Ninetyeast Ridge  
1.5.1 Evolution of the Indian Ocean | 14 
1.5.2 Ninetyeast Ridge | 17 
1.6 O.D.P. Leg 116 Principal Results | 18 
1.7 Estimates of Shortening | 20 
1.8 CHARLES DARWIN Cruise 28, Project Rationale and Thesis Plan | 21 

## Chapter 2 - Data Sources

2.1 CHARLES DARWIN Cruise 28 - Central Indian Ocean Basin  
2.1.1 Cruise Summary | 24 
2.1.2 Cruise Plan | 25 
2.1.3 Data Collection | 25 
2.2 Lamont-Doherty Geological Observatory Data | 29 
2.3 Other Sources | 31 

## Chapter 3 - Sediment Velocities, Fault Block and Deep Structure at the Leg 116 Sites

3.1 Introduction | 32 
3.2 Data Reduction and Results | 32 
3.3 Discussion  
3.3.1 Sediment Velocity Structure | 39 
3.3.2 Basement Topography and Structure | 40 
3.3.3 Anatomy of a Fault Block | 43 
3.4 Summary | 45 

## Chapter 4 - Structural Style of Intraplate Deformation

4.1 Introduction | 49 
4.2 Fracture Zone - Undulation Relationship | 49 
4.3 Fault Analysis and the Fault - Undulation Relationship | 56 
4.4 Emplacement of the Afanasy Nikitin Seamount | 66 

VI
<table>
<thead>
<tr>
<th>Chapter 5 - Oceanic Crustal Structure from Multichannel Seismic Profiles</th>
</tr>
</thead>
<tbody>
<tr>
<td>5.1 Introduction</td>
</tr>
<tr>
<td>5.2 Seismic Processing</td>
</tr>
<tr>
<td>5.3 Observations</td>
</tr>
<tr>
<td>5.3.1 Northward-dipping basement faults</td>
</tr>
<tr>
<td>5.3.2 Southward-dipping basement faults</td>
</tr>
<tr>
<td>5.3.3 Northward-dipping basement reflectors</td>
</tr>
<tr>
<td>5.3.4 Sub-horizontal basement reflectors and non-resolution of the Moho</td>
</tr>
<tr>
<td>5.4 Comparison with Fault Structures Imaged in the Western North Atlantic</td>
</tr>
<tr>
<td>5.5 Comparison of Fault Dip with Seismicity at Present Day</td>
</tr>
<tr>
<td>5.6 Implications for Fault Propagation</td>
</tr>
<tr>
<td>5.7 Reverse Faulting and Long Wavelength Undulations</td>
</tr>
<tr>
<td>5.8 A Strike-Slip Component to Faulting</td>
</tr>
<tr>
<td>5.9 Estimates of Shortening from the Multichannel Profiles</td>
</tr>
<tr>
<td>5.9.1 From the long wavelength undulations</td>
</tr>
<tr>
<td>5.9.2 From the reverse faulting</td>
</tr>
<tr>
<td>5.10 Summary</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Chapter 6 - Analogue and Numerical Models of Intraplate Deformation</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.1 Introduction</td>
</tr>
<tr>
<td>6.2 Rheological Models of the Oceanic Lithosphere</td>
</tr>
<tr>
<td>6.2.1 Introduction: Experimental evidence for rheology</td>
</tr>
<tr>
<td>6.2.2 Elastic Models and Intraplate Deformation</td>
</tr>
<tr>
<td>6.2.2.1 Inapplicability of Elastic Model for Quantitative Studies</td>
</tr>
<tr>
<td>6.2.2.2 Elastic Models and Geodesy</td>
</tr>
<tr>
<td>6.2.3 Layered Rheological Models and Intraplate Deformation</td>
</tr>
<tr>
<td>6.3 Analogue Modelling</td>
</tr>
<tr>
<td>6.3.1 Introduction</td>
</tr>
<tr>
<td>6.3.2 Scaling of the Natural and Analogue Systems</td>
</tr>
<tr>
<td>6.3.2.1 Depth to Brittle/Ductile Transition</td>
</tr>
<tr>
<td>6.3.2.2 Ductile Layer Thickness</td>
</tr>
<tr>
<td>6.3.3 Experimental Design and Methodology</td>
</tr>
<tr>
<td>6.3.4 Results</td>
</tr>
<tr>
<td>6.3.5 Discussion</td>
</tr>
<tr>
<td>6.3.6 Summary of Analogue Modelling</td>
</tr>
<tr>
<td>6.4 Numerical Modelling using the Finite Element Technique</td>
</tr>
<tr>
<td>6.5 Summary</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Chapter 7 - Geodesy and Gravity Modelling</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.1 Introduction</td>
</tr>
<tr>
<td>7.2 Geodetic Data</td>
</tr>
<tr>
<td>7.2.1 Definition of the Geoid</td>
</tr>
<tr>
<td>7.2.2 The Relationship between the Geoid and Bathymetric Features</td>
</tr>
<tr>
<td>7.2.3 SEASAT-derived F.A.A. Gravity Field</td>
</tr>
<tr>
<td>7.3 Ship-collected F.A.A. Gravity Field</td>
</tr>
<tr>
<td>7.4 Modelling Rationale</td>
</tr>
</tbody>
</table>
# 7.5 Two-dimensional Modelling
- 7.5.1 Line A
- 7.5.2 Line B
- 7.5.3 Line C
- 7.5.4 Line D

# 7.6 Three-dimensional Modelling

# 7.7 The Indian Ocean Geoid Low, Convection and Intraplate Deformation

# 7.8 Summary

## Chapter 8 - General Discussion

8.1 General Introduction

8.2 Fault Fabric in the Central Indian Ocean Basin and Spreading-Centres
- 8.2.1 Introduction
- 8.2.2 Evolution of Spreading-Centre Fabric and Inward and Outward Dipping Faults
- 8.2.3 Abyssal Hills and Fault Reactivation
- 8.2.4 Comparison with the Gorda Plate

8.3 Oceanic Crustal Structure

8.4 Buckling/Faulting and Rheology
- 8.4.1 Discussion for Oceanic Lithosphere
- 8.4.2 Buckling and the Continental Lithosphere

8.5 A Diffuse Plate Boundary for the Northern Indian Ocean
- 8.5.1 Introduction
- 8.5.2 Review of Kinematic Models for a Diffuse Plate Boundary
- 8.5.3 Estimates of Lithospheric Shortening
- 8.5.4 Transpressive Tectonics of the Diffuse Plate Boundary
- 8.5.5 Current Plate Motions and Diffuse Plate Boundaries

8.6 Speculations on Reasons for the Timing of the Onset of Deformation

8.7 Speculations on Subduction and the Future Development of the Intraplate Deformation

8.8 Recommendations for Future Work

## References

## Appendix A - Multichannel Seismic Processing

## Appendix B - Derivation of Shortening from Long Wavelength Undulations

## Appendix C - Two-dimensional Elastic Plate Bending Theory

## Appendix D - Analogue Modelling Experimental Details and Results

## Appendix E - Numerical Modelling Experiments

## Appendix F - SEASAT coverage

## Published Paper

Enclosure 1 - North-South Multichannel Seismic Profile Through the O.D.P. Leg 116 Sites

Enclosure 2 - Sonobuoy Record (S/b 4)

Enclosure 3 - Example of 3.5 kHz Record Showing Active Faulting at the Sea-floor
Chapter 1 - Introduction

1.1 Plate Tectonics, Rigid Plates and Intraplate Deformation

The Plate Tectonic theory explains the outer parts of the dynamic earth in terms of independent lithospheric plates overriding a viscous asthenosphere and moving relative to each other over the earth's surface. One of the central premises of this theory is that plates should behave rigidly with deformation only at their margins. Certainly it is manifestly true that most of the large scale topography on the earth's surface, whether mountain ranges, deep sea trenches or transform fault zones, is concentrated at these plate boundaries.

The first suggestion of a deviation from the rigid plate premise is given by detailed analysis of intraplate earthquakes. Deciding what is intraplate seismicity in the continents is difficult: a subjective decision is required as to whether the seismicity represents some form of plate boundary (or micro-plate boundary) or if it is truly intraplate in nature. In the continents intraplate seismicity is often associated with pre-existing lines of weakness caused by old plate boundary features, now incorporated into the continental plates. With the comparative tectonic simplicity of the ocean basins it is much easier to ascertain areas of intraplate deformation.

There are a significant number of oceanic intraplate earthquakes as shown in Figure 1.1. Bergman (1986) recognised that they could be roughly divided into two types. A large proportion of the seismicity occurs in oceanic lithosphere less than 35 Ma old and is a poor indicator of the state of regional stress because it is dominated by stresses related to the early thermal evolution of the lithosphere. The second type of seismicity occurs in lithosphere older than 35 Ma old and is characterised by thrust or strike-slip faulting indicating an intraplate stress field dominated by horizontal compression. The focal mechanisms associated with this type accurately reflect the regional state of stress.

The seismicity in old oceanic lithosphere is not randomly distributed throughout the oceans, but is associated with a few areas of intraplate deformation (Bergman, 1986). These areas of deformation have also been called diffuse plate boundaries (Wiens et al., 1985). One of them lies between the Caribbean Arc and the Mid-Atlantic Ridge.
Figure 1.1 Oceanic intraplate earthquakes for the period 1964 - 1983 (from Bergman, 1986). Note the strong concentration of seismicity in the northern Indian Ocean. See text for discussion.
and may be associated with the relative motions between the North and South American plates (Ball and Harrison, 1970; Bergman, 1986; DeMets et al., 1990). Another is in the Gorda Plate in the north-east Pacific as a consequence of compression between the Pacific and Gorda Plates (Masson et al., 1988). However, the best developed of these deformation zones is centred on the Central Indian Ocean Basin (Bergman and Solomon, 1985; Bergman, 1986) and is the subject of this thesis.

The aim of this thesis is to examine the structural style of intraplate deformation in the Central Indian Ocean Basin. During October 1987 RRS CHARLES DARWIN Cruise 28 visited part of the deformation area (Figure 1.2) and provided most of the data for this thesis. In the rest of this Chapter an overview is given of the state of knowledge prior to this project (Sections 1.2 - 1.7) and outstanding problems requiring resolution. Section 1.8 details the cruise and project rationale and describes the thesis plan.

1.2 Central Indian Ocean Basin

This intraplate deformation area centred on the Central Indian Ocean Basin (Figure 1.2) is sandwiched between the Chagos-Laccadive Ridge and the Ninetyeast Ridge and extends from 5°N to 10°S, although there is some deformation in the Wharton Basin to the east of the Ninetyeast Ridge. It is assumed to have resulted from north-south compression associated with the collision of India with Asia (Weissel et al., 1980) and manifests itself in a number of geophysical data sets (Section 1.3) including seismic reflection, gravity and geodetic data, heatflow measurements as well as prominent seismicity.

The oceanic lithosphere in this area was formed at the South-East Indian Ridge and is 65 - 90 Ma old. The evolution of this Basin (see Section 1.5) has resulted in a number of near north-south fracture zones (Figure 1.2). Covering much of the area is a thick sediment cover deposited from the distal channels of the Bengal Fan. The Ocean Drilling Program (O.D.P. Leg 116) recently visited the area and provided important information on the sediment types and tectonic history (Section 1.6). Prior to this leg only one DSDP site (Leg 22, Site 218) had been drilled in the Bengal Fan and this was more than 1000 km north-east of the Leg 116 sites (08° 00'N, 86° 17'E). This site recorded four main pulses of sandy turbidites, one in the middle Miocene, two in the late Miocene-Pliocene and one in the Pleistocene.
Figure 1.2 The position of the study area and CHARLES DARWIN 28 cruise in the Central Indian Ocean Basin. Also indicated are the bathymetry, principal fracture zones and the Afanasy Nikitin Seamount (A.N.).
In the centre of the intraplate deformation in the Central Indian Ocean Basin is the Afanasy Nikitin Seamount (Section 1.4). It has been suggested that this seamount may have influenced the position and development of the deformation (Section 6.2.2.1).

1.3 Intraplate Deformation in Geophysical Datasets

In this section evidence for intraplate deformation in the northern Indian Ocean, from geophysical datasets available prior to this thesis, is documented. The deformation is revealed by anomalous seismicity (Section 1.3.1), seismic reflection profiles (Section 1.3.2), heat-flow anomalies (Section 1.3.3), and geoid and gravity anomalies (Section 1.3.4).

1.3.1 Anomalous Seismicity

As already discussed there is a large amount of anomalous seismicity associated with the northern Indian Ocean (Stein and Okal, 1978; Bergman and Solomon, 1985; Bergman, 1986; Petroy and Wiens, 1989) and it was this unusually high level of intraplate activity that first attracted geophysicists attention (Sykes, 1970). The seismicity recorded during the last three decades in the area has been dominated by thrust faulting on EW striking faults and strike-slip motion on faults striking NS (Stein and Okal, 1978; Bergman and Solomon, 1985; Wiens and Stein, 1983; Petroy and Wiens, 1989).

Sykes (1970) interpreted the seismicity pattern of the northern Indian Ocean for 1950 - 1966 as evidence for initialisation of subduction. However, the distribution of epicentres since 1964 is more N - S than the NW - SE of Sykes's data and no other evidence has been found to support his suggestion (Stein and Okal, 1978; Bergman and Solomon, 1985). Researchers nevertheless still look to the area as a potential site of subduction initiation.

Stein and Okal (1978) recognized that the northern end of the Ninetyeast Ridge was seismically active in a left-lateral strike-slip sense and deduced a horizontal slip rate of 22 mm yr⁻¹. They suggested that this may represent the decoupling of the Indo-Australian plate on either side of the ridge possibly because of the stress pattern due to the plate boundary configuration. To the east, oceanic lithosphere subducts readily at the Sunda trench, whilst to the west continental shortening by collision at the Himalayas retards the motion of the plate. However, a recent study (Petroy and
Wiens, 1989) while agreeing with the interpretation of left-lateral strike-slip motion, found that the Ninetyeast Ridge had a rate of motion of at most 3 mm yr\(^{-1}\). Petroy and Wiens (op. cit.) also concluded that the seismicity was not solely concentrated along the old transform fault along the eastern side of the Ninetyeast Ridge, but that significant seismicity occurs on its western flank.

Although the northern Indian Ocean is dominated by compressional and strike-slip events, there is a region centred on the Chagos Bank (see Figure 1.2 for position) where tensional earthquakes occur as well (Wiens, 1986). The majority of the earthquakes for this area are characterised by N - S extension and approximately equal values of horizontal (E - W) and vertical compression, resulting in variable components of strike-slip and normal faulting. Wiens (1986) concluded that the high seismicity is unlikely to be related to thermo-elastic stresses (the presence of a magnitude 7.2 thrust earthquake supports this) and more likely to be due to intraplate deformation.

Petroy and Wiens (1989) noted that strike-slip faulting occurs along the outer-rise of the Sumatran Trench and, because strike-slip faulting is uncommon in other outer-rise settings, could be associated with the intraplate deformation. If this seismicity is due to intraplate deformation then there is more intraplate seismicity east of the Ninetyeast Ridge than to the west. However, as discussed in the next sections, tectonic deformation and geoid anomalies are apparently best developed in the Central Indian Ocean Basin to the west.

Bergman and Solomon (1985) noted that during the last few decades of world-wide earthquake monitoring, despite the large amount of low-level seismicity there have only been 11 intraplate earthquakes in the northern Indian Ocean of size sufficient for body-waveform modelling. They looked at these large earthquakes, whose position is shown in Figure 1.3, and determined the best-fitting double-couple point source from an inversion of long period P and SH waveforms recorded at stations of the World-Wide Standard Seismograph Network (WWSSN) and the Global Digital Seismic Network (GDSN). From this inversion they obtained estimates of the centroid depth, double-couple orientation, seismic moment and source time function. The results of this analysis are given in Table 1.1. They noted that this small subset of the total seismicity of the area may have focal mechanisms which are unrepresentative of the long term average seismogenic lithospheric deformation in the northern Indian Ocean. Petroy and Wiens (1989) have recently used historical records to add a few more
Figure 1.3 Seismicity and bathymetry of the northern Indian Ocean from Bergman and Solomon (1985). The Indrani (at ~79°E) and Indira (at ~85°E) fracture zones are marked by dashed lines. Open circles indicated epicentres of earthquakes which have occurred between 1906 and 1963 with larger symbols for events with M or M_s greater than 6. Solid circles indicate epicentres of earthquakes since 1964 with larger symbols for events with m_b greater than 5.4.
events to the compilation of Bergman and Solomon (1985) and these conform to the
trend illustrated in Figure 1.3 with additional strike-slip earthquakes on the northern
end of the Ninetyeast Ridge and in the Wharton Basin.

Table 1 - Epicentral Data and Source Parameters for Earthquakes in the
Northern Indian Ocean (From Bergman and Solomon, 1985)

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<th>Long°E</th>
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<td>5.5</td>
<td>5.9</td>
<td>261/53/090</td>
<td>39</td>
</tr>
<tr>
<td>2/12/81</td>
<td>-15.76</td>
<td>88.39</td>
<td>5.7</td>
<td>5.5</td>
<td>5.8</td>
<td>222/44/078</td>
<td>23</td>
</tr>
</tbody>
</table>

Source: International Seismological Centre (ISC), except for surface wave magnitude Ms for May 25,
1964 (Stein and Okal, 1978) and October 31, 1965 (Wiens and Stein, 1983).

Seismic moment in units of $10^{17}$ Nm

Focal Mechanism specified in degrees as strike, dip, slip direction

Centroid Depth measured from the top of the crust (or top of the sediment layer if used)

a- Denotes events modelled with more than one point source
More regionally, in the Indo-Australian plate, various indicators of stress delineate the orientation of the greatest horizontal compressive stress. This is orientated roughly N-S in India, swings to NW-SE in the northern Indian Ocean and is nearly E-W in Australia. (Weissel et al., 1980; Bergman and Solomon, 1985; Cloetingh and Wortel, 1986; Petroy and Wiens, 1989).

1.3.2 Scales of Tectonic Deformation revealed from Seismic Data.

The intraplate deformation is perhaps most spectacularly revealed by seismic reflection data as a consequence of which the tectonic deformation can be thought of as occurring on two spatial scales. A first order of the deformation is represented by roughly east-west trending, long wavelength (100 - 300 km) undulations of oceanic basement and overlying oceanic and early Bengal Fan sediments with an amplitude of 1 - 2 km. Geller et al. (1983) using a sparse dataset mapped the orientations of the long wavelength undulations as shown in Figure 1.4.

Three principal models have been proposed to explain the long wavelength undulations: firstly, uniform crustal folding (buckling) with the Moho paralleling basement (Weissel et al., 1980; McAdoo and Sandwell, 1985); secondly, inverse boudinage with thickening beneath the basement highs (Zuber, 1987); and finally some form of block faulting (Neprochov et al., 1988). The three hypotheses will be examined throughout the rest of the thesis, but in particular detail in Chapter 6. Seismic refraction studies carried out to detect the presence or absence of boudinage have been contradictory, with Leger (1987) supporting the crustal thickening model while Neprochov et al. (1988) found uniform crustal folding.

Superimposed on this first order of deformation is a second order represented by faulted blocks 5 - 20 km in width bounded by steep faults which offset oceanic basement by up to 600m. (Eittreim and Ewing, 1972; Weissel et al., 1980; Geller et al., 1983; Stein, 1984; Neprochov et al., 1988). Weissel et al. (1980) found that the strike of the faults was 100°E and interpreted them to have resulted from the reactivation of pre-existing features. However, prior to this project seismic reflection data was often insufficiently good to reveal whether the high-angle faults were predominantly normal, reverse or a combination of the two. The vast majority of them appeared to be reverse. Examples of seismic profiles illustrating the two scales of deformation are given in Chapters 4 and 5.
Figure 1.4 Acoustic basement highs and lows (from Geller et al., 1983) with plus signs representing basement highs and minus signs relative lows. Note that because the spatial coverage of seismic profiles is poor in the western part (west of 78°E), the undulations may continue further to the west than indicated here. Bathymetry in kilometres.
Although tectonic deformation is best developed in the Central Indian Ocean Basin there is also some deformation in the Wharton Basin (Weissel et al., 1980; Geller et al., 1983) where analysis of the magnitude and extent of deformation is hampered by poor coverage of seismic profiles.

A study of the relationship between the two scales of tectonic deformation in the Central Indian Ocean Basin is undertaken in Chapter 4 together with a detailed analysis of the structural style of the deformation. In Chapter 5 new multichannel seismic data yields important information on the nature of the high-angle faults.

1.3.3 Regional Heat Flow and Localised High Values.

During 1986 Lamont-Doherty cruises RC2706 and RC2707 obtained heat flow measurements, using digital multiple penetrating instruments, over a variety of sites in the Central Indian Ocean Basin in order to understand the relationship between the deformation and previously recorded high heat flow values (Weissel et al., 1980; Geller et al., 1983). Geller et al. (1983) found the mean of 118 individual measurements to be 64.7 mWm\(^{-2}\) with a standard deviation of 7.4 and a range of 47 - 84 mWm\(^{-2}\). The mean value is not significantly above that expected for lithosphere of 65-70 Ma - 58 mWm\(^{-2}\). They found no difference between the crests and flanks of the long wavelength undulations. However, the range of values in an area so uniformly sediment covered is surprising. This range, and the non-linear (convex-up) temperature-depth profiles present in the area, they interpreted as indicating upward flow of fluid through the sediments.

The O.D.P. Leg 116 took 41 measurements in a very detailed area and found a mean heat flow of 83.7 mWm\(^{-2}\) with a standard deviation of 21.7 and a total range of 44 - 166 mWm\(^{-2}\) (Shipboard Scientific Party, 1989). This localised mean value is much higher than that expected for lithosphere age. The O.D.P. sites were located over two prominent fault blocks and the range and high values can be explained in terms of hydrothermal convection of fluids within the fault blocks (Shipboard Scientific Party, 1989). This is consistent with the earlier suggestion of Geller et al. (1983).

Therefore, although there is localised high heat flow associated with the faulted crustal blocks, regional surface heat flow is not necessarily anomalous. This, together with other factors including residual depth anomalies and depth of seismicity, led Stein and Weissel (1990) to conclude that lithospheric temperatures in the Central Indian Ocean Basin were not significantly different from those expected for its age.
Stein et al. (1988) looked at heat-flow in the Wharton Basin and found that the heat-flow was not characterised by unusually high-values. From this it was concluded that the Wharton Basin has undergone a lower level of deformation. However, it should be noted again that there is a lack of spatial coverage in the Wharton Basin, with much less information than for the Central Indian Ocean Basin.

1.3.4 Geoid and Gravity Anomalies

Linear geoid anomalies in SEASAT altimeter data have been interpreted as resulting from the folding of the lithosphere in the deformation area (Weissel and Haxby, 1982; Weissel and Haxby, 1984; McAdoo and Sandwell, 1985). The east-west trending geoid undulations in the Central Indian Ocean Basin (Figure 1.5) have wavelengths ranging from 130 to 250 km with a mean of around 190 - 200 km (McAdoo and Sandwell, 1985, Zuber, 1987). Zuber (1987) found that the mean wavelength increased from south to north in the deformation area and interpreted it as due to the load effect of sediments in the north of the area but not in the south. Alternatively, as discussed in Chapter 6, this could be at least partly due to the increase in lithospheric age northwards, resulting in a progressively thicker plate bending with longer wavelength.

Large amplitude (30 - 80 mGal) free-air anomalies are also associated with the deformation (Weissel et al., 1980). If the tectonic fabric within the Central Indian Ocean Basin was two-dimensional it would have been possible, on the basis of consideration of amplitude and topography, to conclude whether inverse boudinage or folding of the crust was likely to be the origin of the free-air anomalies. However, as will become clear later, this area cannot be considered to be two-dimensional, and therefore such an analysis would be too simplistic. Gravity modelling in two and three dimensions is undertaken in Chapter 7 to determine between the hypotheses of inverse boudinage and buckling. A detailed analysis of the relationship between the geoid and gravity fields and basement features is also given in that Chapter.

1.4 Afanasy Nikitin Seamount

This seamount lies towards the centre of the intraplate deformation (see Figure 1.2) and extends about 300 km in a north-south direction from 2.0 to 5.0°S centred on 82.5°E. A complex northern peak at 3°S rises to 3000m above the surrounding ocean basin, while a more subdued southern plateau rises to 1500m above the surrounding topography (Paul et al., 1990; Karner and Weissel, in press). The surrounding crust
Figure 1.5 Grey scale image depicting N-S gradients of SEASAT-derived gravity anomalies in the northern Indian Ocean. Slopes down to the north are light, slopes up to the north are dark. The east-west trending gravity anomalies are clear in the upper left part of the image (from Shipboard Scientific Party, 1989).
was formed at anomaly 33 time (R.A. Scrutton, pers. com.) close to the Indira Fracture Zone (Sclater and Fisher, 1974) which runs along the western side of the seamount. In general the western flank of the seamount has a much steeper gradient than the eastern one.

Admittance studies (Paul et al., 1990) indicate that the seamount load was emplaced on young, thin (with an elastic thickness of 2 - 5km) and weak oceanic lithosphere. Paul et al. (1990) conclude that the seamount was formed at the junction of a spreading axis and the Indira Fracture Zone. They are, however, unclear as to the date of formation, perhaps because of their erroneous determination of crustal ages from characteristic magnetic anomaly studies.

Karner and Weissel (in press) investigate the interaction of the seamount load with the later deformation, by using simple one and two dimensional models for the compression of a thin elastic plate containing an initial perturbation overlying a inviscid fluid to deduce an elastic thickness of 10 - 15 km. They suggest an emplacement in the Late Cretaceous or Early Tertiary from analysis of sediment onlap patterns onto the flanks of the seamount and from dredge samples (R.A. Scrutton, pers. com.) giving biostratigraphic ages from chalks intercalated with basalts. In addition they suggest that the later compression amplified suitable wavelength components of the seamount deflection. This hypothesis that the seamount may have influenced the position and development of the deformation will be discussed in detail in Chapter 6.

1.5 History of the Indian Ocean and Ninetyeast Ridge

In this section a brief overview is given of the evolution of the Indian Ocean and Ninetyeast Ridge with particular reference to the Central Indian Ocean Basin up to the onset of intraplate deformation.

1.5.1 Evolution of the Indian Ocean

Accurate reconstruction of the Indian Ocean back to the time of Gondwanaland is possible because of the absence of significant trenches (with the exception of the Scotia and Sunda Arcs) so that much of the seafloor created during the evolution of the ocean is still in existence. Many authors have worked on the evolutionary history
Figure 1.6 Evolution of the Indian Ocean from Gondwanaland to present (After Curray et al., 1982). See main text for discussion.
and there are several recent summary papers (Norton and Sclater, 1979; Curray et al., 1982; Tarling, 1988; Molnar et al., 1988). The general consensus regarding the evolution is shown in Figure 1.6.

Prior to 125 Ma (A) India separated from Antarctica-Australia with separation parallel to the south-east coast of India (B).

At around 80 - 90 Ma India separated from Madagascar causing a major re-organisation of spreading direction between India and Antarctica-Australia (C). Spreading proceeded in a ~N - S direction parallel to a transform fault on the east-side of the Ninetyeast Ridge. During the period 90 - 53 Ma India made its rapid northward flight.

It was during this period that the oceanic lithosphere of the Central Indian Ocean Basin was formed. McKenzie and Sclater (1971) showed that it was impossible to make continuous east-west correlation of distinctive magnetic anomalies across this basin. On magnetic grounds they suggested that the ocean floor was cut by at least two fracture zones parallel to the Ninetyeast Ridge - they called these the Indrani (at 79°E) and the Indira (at 83°E). Later Sclater and Fisher (1974) postulated the existence of another fracture zone at 86°E. (The positions of these fracture zones are shown in Figure 1.2). By comparison with the magnetic anomaly time scale of Berggren (1972), Sclater and Fisher (1974) found a half-spreading rate of 120 mm yr⁻¹ for anomalies 27 - 30 and 57 mm yr⁻¹ for anomalies 30 - 33.

The end of this phase of spreading occurred at around anomaly 22 (53 Ma) (D) perhaps due to initial contact of India with Asia. Also around this time Antarctica and Australia started separating significantly. Spreading then occurred in a NE - SW direction along the South-East Indian Ridge between Australia and Antarctica (D). This last phase of spreading has continued to the present with the development of the important change in tectonic style across the northern end of the Ninetyeast Ridge, with continental collision to the west and subduction to the east at the Sunda Trench (F).

Continental collision between India and Asia about 53 Ma ago has resulted in the formation of the Himalayas, but the age of onset of mountain building is very uncertain. However, denudation of the mountain range has resulted in the largest subaerial delta in the world resulting from the confluent Ganges and Brahmaputra rivers filling the Bengal Basin. Associated with this are the Bengal and Nicobar deep-
sea fans. Copeland and Harrison (1990) have shown that a significant portion of the material in the Bengal Fan is first cycle detritus from the Himalayas. The deep sea fan complex is the largest in the world extending over 3000 km to ~7°S (Curray et al., 1982) and is divided into two lobes by the Ninetyeast Ridge: the main Bengal Fan, west of the Ninetyeast Ridge, and the Nicobar Fan lying to the east. Most of the intraplate deformation in the Central Indian Ocean Basin is covered by the main Bengal Fan. The sedimentary history and development of the deformation was investigated by O.D.P. Leg 116 as described in Section 1.6.

The onset of intraplate deformation in the Central Indian Ocean Basin started ~7Ma (see Section 1.6). Reasons for the timing of the onset of the deformation are unclear, although as discussed in Chapter 8, may be linked with the evolution of the Tibetan Plateau.

1.5.2 Ninetyeast Ridge

The near meridional Ninetyeast Ridge has a width of 150 - 250 km and extends ~5000 km from 31°S to 17°N, but north of 10°N is buried beneath the sediments of the Bengal Fan (Cury et al., 1982). North of about 10°S the ridge has a blocky en echelon structure whereas the southern part is relatively straight and flat-topped. Evidence for isostatic compensation can be obtained from gravity data (Bowin, 1973) which has been interpreted as resulting from the emplacement of relatively less dense material in the upper mantle beneath a fairly normal oceanic crust structure, a fact confirmed by seismic refraction (Francis and Raitt, 1967). This together with subsidence studies indicates that the Ninetyeast Ridge has long been an integral part of the Indian plate (Detrick et al., 1977).

Many theories have been expressed for the evolution of the Ninetyeast Ridge, although the generally accepted one now is the passage of the Indian plate over a hot spot. The most convincing evidence has come from a recent O.D.P. leg (121) which shows a northward increase in basement age consistent with the Ninetyeast Ridge being the trace on the Indian plate of the Kerguelen/Ninetyeast hotspot (Peirce et al., 1988).
1.6 Initialisation and Development of Intraplate Deformation and the Sedimentary History of the Bengal Fan - O.D.P. Leg 116 Principal Results

This leg of the Ocean Drilling Program, whose position is shown in Figure 2.1, was designed to investigate the timing and development of the intraplate deformation as well as the sedimentary history of the Bengal Fan. A seismic section showing the relative positions of the drill sites is shown in Figure 1.7 together with the significant seismic horizons. The principal results of the leg (Shipboard Scientific Party, 1989) are discussed here.

The first of the objectives was met by drilling a pair of companion sites on one of the fault blocks in the deformation area. One of these sites, 717, was in the thickest part of the axis of a syncline, while site 719 was partway up the block where syn-deformational sediments were thinner. Sediments at both sites are fan turbidites with distinctive turbidite layers correlatable between the two sites. Reduction of the section between the two sites has occurred by pinching out of beds and thinning of individual turbidites. From this reduction and dating studies using calcareous nanofossils it was concluded that movement on the fault had been gradual and fairly constant, perhaps with a slight increase with time. The onset of deformation (reflector A in Figure 1.7, see also Figure 4.7), which is not a lithologic boundary, is 7 Ma, which with the 350 m of uplift across the fault gives an average rate of motion of 50 m/Ma.

The two sites yield a record of sedimentation on the distal Bengal Fan since the late Miocene. A thin layer of mud and clayey ooze covers a sequence dominated by micaceous silty turbidites of late Pleistocene age (unit II). This formation constitutes a distinctive seismic stratigraphic unit on top of reflector B (Figure 1.7, see also Figure 4.7) that truncates lower reflectors. These relatively coarse grained, rapidly deposited turbidites probably reflect the Quaternary uplift of the Himalayas. Below this unconformity units III and IV comprize mainly mud turbidites of late Miocene to Pleistocene age, while unit V is composed of mainly silt and silty mud turbidites of early to middle Miocene age.

The second objective, the sedimentary history of the distal Bengal Fan, was met by drilling site 718 on the next block south (Figure 1.7) where the post-Miocene section is attenuated, allowing penetration of older sediments. At the base of the hole fan sediments of early Miocene age (17 Ma) were still being penetrated with no evidence of reaching the base of the Fan. Almost the entire Miocene section consists of silt and silty mud turbidites. The fact that the Bengal Fan was well established in this region
Figure 1.7 Conrad 2706 single-channel profile running north-south through O.D.P. Leg 116 sites. See text for discussion of lithostratigraphic units drilled (denoted by roman numerals) and reflectors A and B (from Shipboard Scientific Party, 1989).
2500 km from the Ganges delta suggests that the main uplift of the Himalayas or Tibet occurred earlier than is generally assumed. Some studies have suggested that the major uplift of the Himalayas occurred in the late Miocene (e.g. Powell and Conaghan, 1973) whereas Leg 116 evidence indicates substantial uplift had taken place by the early Miocene or even pre-Miocene.

Site 718 is in an area of high heat flow (see Section 1.3.3). At this site there is excellent evidence for vigorous hydrothermal circulation in the form of temperature inversion. The silty turbidites of unit II have temperatures actually higher than the upper part of the underlying clay turbidites of unit III. This was interpreted as resulting from warm water rising up the fault to the north of site 718 and spreading laterally through permeable layers in the silty turbidites. Meanwhile, from the outcrop of clayey turbidites a few kilometre to the south, cooler seawater must be flowing downwards. This interpretation is supported by geochemical studies (Shipboard Scientific Party, 1989).

From these results it was concluded (Shipboard Scientific Party, 1989) that there were several controls on the sedimentation in the distal Bengal Fan. The uplift history of the Himalayas is one of these controls with the Quarternary pulse of coarse silty turbidites following a significant pulse of mountain building. Another control is variations in sea level with a hypothesised sharp rise near the Miocene-Pliocene boundary contributing to the trend from silty to more muddy turbidites. Channel migration and lobe switching would also have controlled sedimentation as for any delta. Finally the tectonic activity due to the intraplate deformation has had a considerable effect on sedimentation with the movement of fault blocks altering distribution channels.

1.7 Estimates of Shortening

One of the most interesting problems associated with the intraplate deformation area is that there is a large discrepancy between estimates for the rate and amount of shortening across the zone of deformation. Estimates from plate-motion models suggest north-south shortening of ~ 4 -5 mm yr\(^1\) (Royer and Chang, submitted; Gordon \textit{et al.}, 1990) whereas previous estimates from the tectonic properties of the deformation are ~ 1 mm yr\(^1\) (Weissel and Geller, 1981; Geller \textit{et al.},1983). Over the 7Ma since the onset of deformation these rates give ~30 km and 7 km of shortening respectively. Weissel and Geller (1981) found that the long wavelength undulations contributed little to the shortening and concluded that most of the shortening was
taken up by the high-angle faults. In Chapter 5 a new estimate of shortening is obtained from the CHARLES DARWIN multichannel data. A comparison between this estimate and plate motion studies estimates and reasons for any discrepancy is given in Chapter 8. It is important to resolve any discrepancy because, if plate motion studies continue to predict more shortening than is observed from present geophysical datasets, then this could support the inverse boudinage hypothesis of Zuber (1987), with shortening being taken up by lithospheric thickening.

1.8 CHARLES DARWIN 28, Project Rationale and Thesis Plan

The foregoing discussion has indicated that understanding of the non-rigid behaviour of the conventional Indo-Australian Plate is far from complete and as summarised in the following sections there are many unresolved questions. During October 1987 CHARLES DARWIN Cruise 28 took place in the Central Indian Ocean Basin with the aim of generally increasing understanding of the style of intraplate deformation. The principal objectives of this cruise and other components of this project are outlined below.

1.8.1 Since previously collected datasets have been too sparse for any detailed analysis of the deformation, a principal objective of the cruise was to collect data so that a detailed examination of structural style could be undertaken, from which it was hoped that a number of questions could be answered. What is the real continuity of the E - W long wavelength undulations, hitherto interpreted as continuous across fracture zones? What is the relationship between the high-angle faults and the long-wavelength undulations? What is the geometry of the high-angle faults at depth? Are estimates of tectonic shortening across the area accurate? By answering these and other questions which would emerge in the course of the investigation it was hoped that a tectonic model could be proposed for the Central Indian Ocean Basin.
This study of structural style required a grid of seismic profiles, making use of pre-existing profiles (collected by the Lamont-Doherty Geological Observatory), close enough to map fault blocks but covering enough area to include undulations and fracture zones. A mixture of single and multichannel profiles (the first multichannel profiles to be acquired in the Central Indian Ocean Basin) were collected as well as bathymetric, gravity, and magnetic data, the latter being particularly useful for determining the location of fracture zones.

1.8.2 In order to investigate the deep structure of the O.D.P. Leg 116 sites and improve the significance of the Leg's results a multichannel seismic profile was shot as close as possible to the O.D.P. sites. Perpendicular to this profile a series of parallel single channel profiles were collected together with concomitant wide-angle sonobuoy profiles to determine the velocity-depth structure around the O.D.P. sites. Velocities determined from the sonobuoys could also be used in the multichannel seismic processing and for depth conversion.

1.8.3 Clearly one of the biggest questions associated with the intraplate deformation is the mode of formation of the long wavelength undulations. Is it buckling, inverse boudinage or block faulting? The geometry of the Moho could help to resolve this and one of the main reasons for collecting multichannel seismic profiles was to elucidate the mode of deformation. Additionally, and an important part of the overall project, it was planned to use various modelling techniques, including gravity, analogue and numerical modelling, to resolve this issue. Furthermore, it should be noted that our understanding of non-elastic deformations in the oceans lags behind that of the continents. By studying the intraplate deformation in the Central Indian Ocean Basin it was hoped to redress the balance and yield significant information on the nature of oceanic lithosphere rheology.
1.8.4 Another aim of the cruise, although outside the brief of this thesis, was an investigation of the Afanasy Nikitin Seamount. The main purpose of this investigation being to establish its age and potential role in the initialisation of lithospheric deformation. The survey over the seamount included single channel seismic profiling and the collection of gravity, magnetic and bathymetric data as well as dredging and coring. Although the findings concerning the seamount will be principally written up elsewhere, the results will be discussed as necessary for a feature so close to the survey area.

The foregoing was the rationale for the cruise and the project; the following is an explanation of the thesis layout. Chapter 2 will describe data sources, and include a summary of data collection during CHARLES DARWIN cruise 28. The velocity-depth structure for the O.D.P. sites will be presented in Chapter 3, while the investigation of structural style from single-channel seismic profiles, magnetics and earthquake seismicity is reported in Chapter 4, together with the development of a tectonic model. In Chapter 5 the multichannel results are presented, the findings incorporated into the structural style and the wider framework, and a mechanism for the propagation of the deformation proposed. An Appendix (A) provides details of the multichannel processing. Chapter 6 comprises a review of oceanic lithosphere rheology and details numerical modelling attempted and physical modelling using the sandbox technique. Gravity modelling is the principal element of Chapter 7 together with a discussion of the relationships between gravity-geoid anomalies and basement topography. Finally Chapter 8 discusses the findings of the previous chapters, extends them into the wider framework and presents an explanation of the deformation in the Central Indian Ocean Basin in terms of a diffuse plate boundary.
CHAPTER 2 - Data Sources

This project was based on data collected by an Edinburgh University cruise aboard the RRS CHARLES DARWIN (CD 28) in the Central Indian Ocean Basin during October 1987. The cruise was designed to integrate with pre-existing profiles collected by the Lamont-Doherty Geological Observatory (Section 2.2) which were made available to the author. Small amounts of data from other sources were also used to complement the Edinburgh and Lamont data as described in Section 2.3.

2.1 RRS CHARLES DARWIN Cruise 28 - Central Indian Ocean Basin

2.1.1 Cruise Summary

CHARLES DARWIN cruise 28 (CD 28) departed from Muscat, Oman on September 27th 1987 where the start-of-cruise gravity reading was taken. The scientific personnel disembarked on the 1st of November in Mauritius and the end-of-cruise gravity reading was taken in Durban when the ship arrived there on the 6th of November. Weather conditions throughout the cruise were generally good, with only a few squalls when coring was being attempted over the Afanasy Nikitin seamount. The cruise was a success in that all the principal scientific objectives were met. Data quality was mixed with the seismic profiles highly variable in quality, although many of the final multichannel profiles were of high quality. During the course of the 18.5 days working time the following quantities of data were collected.

- Bathymetric profile: 4116km
- Magnetic profile: 4116km
- Gravity profile: 4116km
- 3.5kHz sediment profile: 3091km
- SCS reflection profile: 2220km
- MCS reflection profile: 1300km
- Total reflection profile: 3520km
- 10 wide-angle reflection profiles with sonobuoys
- 4 dredge stations
- 2 piston core stations.
2.1.2 - Cruise Plan

The positions of the CD 28 profiles were designed to fulfill the four principal objectives of the cruise. These objectives, as described in more detail in Section 1.8, were: to obtain an understanding of the structural setting of the O.D.P. Leg 116 sites; to study the structural style of the intraplate deformation in an area containing a substantial number of pre-existing Lamont profiles; to determine the mode of formation of the long wavelength undulations; and to investigate the Afanasy Nikitin Seamount Group. A long north-south multichannel seismic line, line A in Figure 2.1, was designed to link the study areas. It was also hoped that this line would determine the geometry of the Moho and hence crustal thickness variations. A track chart of the CD 28 cruise shown in Figure 2.1 illustrates the relative positions of the study areas.

2.1.3 - Data Collection

Navigation throughout the cruise was by SATNAV and GPS position fixing with direct-reckoning during periods between them. Quality of navigation was good although there was usually jumps in position fixes between SATNAV and GPS. Navigation was automatically inputed into the digital data logging and processing system. This three level (A, B and C) computer system greatly reduced the amount of time spent logging and processing gravity, magnetic, bathymetric and navigation data. Level A of the system dealt with equipment interface for the gravity, magnetic and navigation data, with all readings being recorded alongside a time base. These data were then collected in level B and written onto magnetic tape. Level C processed the data and controlled the final presentation. It was at this level that the gravity readings were reduced to Free-Air and Bouguer Anomalies (including Eotvos and latitude corrections) and that the magnetic readings were reduced to total field magnetic anomalies. Level C also controlled the production of navigation plots at any scale as well as along track magnetic, gravity and bathymetric profiles. Bathymetric measurements from the Precision Echo Sounder were the only data which had to be typed manually into level C. Once in level C the bathymetry was corrected according to Matthew's Areas tables which correct for the variation of acoustic velocity in seawater.

In addition to a Precision Echo Sounder (PES) there was also on board a 3.5 kHz bathymetry and sediment profiler. The 3.5 kHz fish was useful for determining possible dredge and core sites and providing an insight into the near-surface sediment deformation. Unfortunately the 3.5 kHz fish was lost midway through the cruise.
Figure 2.1 Track chart for RRS CHARLES DARWIN Cruise 28 indicating the relative positions of the principal study areas and seismic profiles illustrated in this thesis.
when its corroded tow cable gave way, probably after having snagged on fishing nets.
Throughout the survey the uncorrected bathymetric data (calculated for a velocity of seawater of 1500 ms\(^{-1}\)) from the PES was typed into level C. The data from the PES was preferred because a small discrepancy between the two echo sounders (only 5-8 m) was found to increase with ship velocity, suggesting that the 3.5 kHz fish was giving slightly inaccurate depths.

Gravity measurements were made with a Lacoste and Romberg Air-Sea gravimeter which performed well throughout the cruise with very low drift and a mean cross-over error of less than 2 mGal with a standard error of 0.3. Gravity base station tie-ins were made at Muscat and Durban. The very low level of drift meant that underway processing of the gravity data was accepted and no post-cruise reprocessing was thought necessary.

Magnetic data was collected using a Barringer proton-precession magnetometer which measured the total field. Underway processing reduced the total field measurements to anomalies with respect to the International Geomagnetic Reference Field. Cross-over errors ranged from zero to greater than 100 nT. These errors were found to be heading dependent and a simple cosine function was used to reduce the cross-over errors everywhere to less than 40 nT. Two sets of magnetic anomaly were produced from level C, one uncorrected for any errors, the other with the heading error correction. The processed gravity and magnetic data were recorded both digitally via the computer system on magnetic tape and on paper records.

Throughout the survey the acoustic signal for the seismic profiles was produced by an airgun array operating at approximately 1800 psi (0.53 kgm\(^{-2}\)). For the multi-channel lines a large source array comprising 160 ci (2.62 dm\(^{3}\)) + 300 ci (4.92 dm\(^{3}\)) + 466 ci (7.64 dm\(^{3}\)) + 700 ci (11.47 dm\(^{3}\)) airguns were used, to produce the predominance of low frequencies necessary for deep penetration, firing every 18 seconds. For the single-channel profiles either two 160 ci (2.62 dm\(^{3}\)) airguns firing every 10 seconds or two 40 ci (0.67 dm\(^{3}\)) airguns firing every 9 seconds were used to obtain the high frequencies needed for resolution of the shallow structures. When a wide-angle profile (see below) was carried out concomitantly with a single-channel profile a 160 ci (2.62 dm\(^{3}\)) and a 300 ci (4.92 dm\(^{3}\)) airgun with a wave shaper gave the most suitable source signature. Although the wave-kit significantly reduced the energy of the array the signature produced was much more compact (higher primary to secondary ratio) giving higher resolution.
For the single-channel seismic profiles a two-channel Geomechanique hydrophone array was used comprising two active and two passive sections each of length 50m. The signals detected by the array were then recorded on a Racal Store 4 tape recorder along with shot instant and time. The two channels from the array were summed (normal moveout is negligible for only two channels in deep water) and displayed on two EPC recorders. EPC recorder 1 displayed the data input using a 4 second sweep after being passed through a 40 - 200Hz filter. EPC recorder 2 displayed the data copy quality (read after write) using an 8 second sweep and a 30 - 150Hz filter. For both the EPC recorders the delay until the sweep was triggered varied, depending on the water depth.

For the multichannel seismic profiles a 12-channel analogue recording system based on a Geomechanique array and a Racal Store 14 tape recorder was used. The Geomechanique array was composed of 12 sets of active and passive sections, with each active section separated by a passive. Each of these sections was 50 m long and contained 48 piezoelectric crystals. This gave an active section spacing of 100 m, which with a shot spacing of 50 m (ships speed 5.4 knots) would give 12 fold coverage and a common depth point (CDP) spacing of 50 m. The distance between the first active section and the airgun array was approximately 150 m with a rope length and spring section decoupling the vibrations of the boat from the hydrophone. The extra two channels of the Store 14 were for recording the shot trigger and a time base. Unfortunately the failure of channel 14 on the Store 14 and the inability to generate a suitable time base on another channel meant that no time base was recorded during the survey. This may have allowed errors to develop in the multichannel processing due to different tape drives operating at different speeds. Fortuitously this was not a significant problem during processing. The first two active sections of the array were connected to the two EPC recorders in the same way as for the 2-channel recording system.

The wide-angle profiles were recorded using Ultra disposable sonobuoys and analogue recording on a Store 4 tape recorder. As already mentioned, concomitant single channel seismic profiles were also collected using a 2-channel hydrophone array and displayed on EPC recorder 1 after being summed and passed through a 40 - 200 Hz filter. A couple of sonobuoys were also recorded concomitantly with multichannel seismic profiling. The signal detected by the sonobuoy hydrophone was relayed back to the ship via a radio transmitter. It was passed through a 300 Hz low pass filter, recorded on the Store 4, reread and amplified, and then passed through another filter (which varied but was typically 4-50 Hz) before being displayed on
EPC recorder 2. The wide-angle profiles collected were generally of reasonable quality when using the older Ultra sonobuoys. However, the new type of buoy, model SSQ 904B, required a strong high-cut filter on the demodulated signal, and consequently gave a poorer final record.

Four dredge and two piston core stations were carried out over the Afanasy Nikitin Seamount. Of the four dredge sites, one was extremely successful with the recovery of large pieces (up to 0.3 m) of pillow basalt. The other dredge sites were less successful. Both the piston cores achieved a good recovery with a 3 m and a 4 m core. Neither the dredge nor piston core data is discussed at length in this thesis, although the principal results from the seamount survey are tied into the wider framework.

2.2 - Lamont-Doherty Geological Observatory Data

In the past, the research institution which has had most cruises in this area is the Lamont-Doherty Geological Observatory. A brief visit was made to Lamont in September 1988 in order to study the seismic reflection profiles from these cruises. Within the study area single-channel profiles from the following cruises were inspected and analysed:

<table>
<thead>
<tr>
<th>Cruise</th>
<th>km of seismic reflection profile</th>
</tr>
</thead>
<tbody>
<tr>
<td>VEMA 2901</td>
<td>345</td>
</tr>
<tr>
<td>VEMA 2902</td>
<td>530</td>
</tr>
<tr>
<td>VEMA 3305</td>
<td>340</td>
</tr>
<tr>
<td>VEMA 3616</td>
<td>800</td>
</tr>
<tr>
<td>ROBERT CONRAD 1708</td>
<td>535</td>
</tr>
<tr>
<td>ROBERT CONRAD 2706</td>
<td>1350</td>
</tr>
<tr>
<td>ROBERT CONRAD 2707</td>
<td>330</td>
</tr>
<tr>
<td>Total</td>
<td>4230</td>
</tr>
</tbody>
</table>

Seismic reflection profiles from VEMA 1909 were also inspected. However, these early records were of too poor quality to be of use with only limited sediment penetration. The total data coverage of the combined EU and Lamont profiles, shown in Figure 2.2, amounts to 7750 km of seismic reflection profile, and this coverage represents the greatest density of profiles available for any substantial area of the intraplate deformation at the present time.
Figure 2.2 Track chart of seismic profiles which were analysed as part of this project. Bold lines in the track chart indicate the position of RRS CHARLES DARWIN multichannel lines.
The magnetic anomaly profiles from these cruises (and also from non-Lamont cruises) have been studied recently by Shipboard Scientific Party (1989) and the EU magnetic anomaly profiles could be interpreted with close reference to this study.

Gravity data from VEMA 2902 and CONRAD 1708, amounting to 1055 km of data, was obtained from the National Geophysical Data Centre at Boulder, Colorado via the British Oceanographic Data Centre at Liverpool.

2.3 - Other Sources

Gravity data from the French cruise, JEAN CHARCOT 840004111 (230 km of data), and the Russian cruise DM MENDELEYEV E07 (340 km) were also obtained from Boulder. Data from these two cruises, the positions of which are shown in Figure 7.3, together with the two Lamont cruises were obtained specifically to investigate the structure of a particular fracture zone, the Indrani. This data was included in the contour map of the Free-Air Anomaly. Cross-over errors between surveys were generally less than 3mGal with a mean value of 3.2 mGal and a standard error of 1.8 mGal.

Some data were also gleaned from a recent summary paper (Neprochov et al., 1988) of Russian cruises in part of the study area which were used to help define the position of the Indrani fracture zone.
Chapter 3 - Sediment Velocities, Fault Block and Deep Structure at the Leg 116 Sites

3.1 Introduction

One of the principal objectives of the RRS CHARLES DARWIN cruise was an investigation of the deformation in the vicinity of O.D.P. Leg 116 sites where stratigraphic control and other data sets are excellent. This was considered to be the best area in which to study the detailed anatomy of an individual fault block. During the survey around the drill sites, vertical incidence single and multichannel seismic reflection, gravity, magnetic, bathymetric, and wide-angle reflection data were collected. The results from the wide-angle reflection survey are presented in this Chapter with the aim of describing the velocity-depth structure in sediments and basement around the O.D.P. sites, seeking any lateral variations of velocity within the fault blocks related to tectonic history, and discussing the geological and tectonic implications of these features.

The wide-angle reflection data were collected concomitantly with vertical incidence reflection profiles, using disposable sonobuoys on east-west tracks parallel to the structural grain (Figure 3.1). Details of the methods of collection and acquisition parameters are given in Chapter 2 (2.1.3). Data quality on the six sonobuoy profiles was highly variable, and the picking of wide-angle events was made difficult by the fairly monotonous sediment sequence of silts and sands that gave few discrete strong reflections that could be traced for far along profile. For the same reason it was also difficult to correlate velocity changes selected from these profiles with any lithologic variations noted from the O.D.P. data. Some details of the velocity structure within the fault blocks were resolved, however, and the top of oceanic layer 2 produced strong reflected and refracted arrivals on most of the sonobuoys, which facilitated a consistent measurement of a mean sediment velocity above basement. Good depth conversions for basement followed from this sediment velocity data.

3.2 Data Reduction and Results

The sonobuoy data were digitized and then inverted using a least squares T²-X² program to give r.m.s. velocities and then interval velocities and layer thicknesses. An example of a digitised sonobuoy record is shown in Figure 3.2. The structure produced was
Figure 3.1 The local setting of the sonobuoy profiles. Bold straight lines indicate sonobuoy profiles, dotted lines the positions of the CHARLES DARWIN 28 track lines, dashed lines the trace of the principal faults and the thin solid lines the bathymetry. Line A is drawn to show the line of the depth-converted section shown in Figure 3.3. Also indicated are the positions of seismic profiles illustrated in this Chapter.
Figure 3.2 Digitised travel times from the wide-angle reflection record section of sonobuoy 4. Reflection arrivals are marked as continuous line for the seafloor (R1) and three sediment horizons (S1, S2, S3) and a sub-basement horizon (B1). The interval velocities determined from wide-angle reflections are shown in brackets. The two principal refraction arrivals (from top of oceanic layer 3) and B1 are also marked as dashed lines. The original record for this sonobuoy is shown as Enclosure 2.
confirmed by forward modelling using a ray-tracing program to iteratively adjust the calculated traveltimes until agreement was reached with the observed travel times. Velocity gradients rather than constant velocities within layers in the sediments could also be investigated in this way. For a more detailed discussion of sonobuoy data reduction and sediment velocity determination see Le Pichon et al. (1968), Houtz et al. (1970) and White (1979). A summary of the results from each of the sonobuoys is given in Table 3.1.

Of the six sonobuoy profiles those in the centre of the fault blocks (1,4,8) gave the best records. Those shot closer to the uplifted edges of the fault blocks (3,5,6) may have been degraded by the combination of steep out-of-plane dip and increased fracturing up block to give poorer, less decipherable profiles. However, despite these problems, there is a suggestion from mean sediment velocities above basement that velocities decrease slightly southwards in each fault block, perhaps reflecting the increased fracturing towards the principal faults where anticlinal deformation occurs (see Section 3.3.3 and Figure 3.7).

Using the fact that the calculated interval velocities are actually the mean velocities within the layers, a crude estimate of the average sediment velocity gradient above basement was obtained by plotting the interval velocities at a depth corresponding to the mid-points of the layers and then fitting a least-squares first degree polynomial to the data. The straight line is of form $V = V_0 + aH$ where $V$ is the instantaneous velocity at depth $H$, $V_0$ is the sediment surface velocity and $a$ is the average sediment velocity gradient. For the interval velocities determined from the six sonobuoys, $V_0 = 1.73 \text{ km s}^{-1}$ and $a = 0.75 \text{ s}^{-1}$. This gives an estimate of the sediment velocity immediately above basement of $3.4 - 3.5 \text{ km s}^{-1}$, which agrees well with estimates of the base-sediment velocity from nearby published sonobuoy results (Curray et al., 1982; Stein and Weissen, 1990). Using this simple velocity function, the north - south single-channel reflection profile shot by the CHARLES DARWIN through the O.D.P. sites (Line A, Figure 3.1) was depth converted (Figure 3.3).

Two principal refracted arrivals were recorded in the sonobuoy data, one of velocity $4.0 - 4.1 \text{ km s}^{-1}$ from the top of basement and another of $6.8 - 6.9 \text{ km s}^{-1}$, presumably from the top of layer 3. No refracted arrivals were recorded from within layer 2 or from below the top of layer 3. A composite crustal velocity structure incorporating reflection, refraction, and O.D.P. well data (J. Cochran, pers. com.) is shown in Figure 3.4. The fitted line is biased toward the well data close to the surface, although the two sets of results are not in great disagreement. There is, nevertheless, frequently some lack of agreement between
<table>
<thead>
<tr>
<th>Layer</th>
<th>Layer</th>
<th>Interval Velocity (km s(^{-1}))</th>
<th>Thickness (km)</th>
<th>Two-Way Travel Times (s)</th>
<th>Standard Deviation of Travel Times</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1</td>
<td>1.8 0.12 6.4 0.02</td>
<td>2.1 0.42 6.8 0.14</td>
<td>2.55 1.78 8.13 0.04</td>
<td>4.1 0.24 8.25 0.13</td>
</tr>
<tr>
<td>3</td>
<td>3</td>
<td>2.2 1.6 7.75 0.10</td>
<td></td>
<td></td>
<td>No velocity determined</td>
</tr>
<tr>
<td>4</td>
<td>4</td>
<td>2.0 0.35 6.69 0.09</td>
<td>2.5 0.9 7.5 0.03</td>
<td>3.3 0.8 7.98 0.11</td>
<td>4.1 0.41 8.18 0.46</td>
</tr>
<tr>
<td>5</td>
<td>5</td>
<td>2.2 1.88 7.83 0.24</td>
<td></td>
<td></td>
<td>4.1 0.34 8.19 0.40</td>
</tr>
<tr>
<td>6</td>
<td>6</td>
<td>2.2 1.78 7.89 0.11</td>
<td></td>
<td></td>
<td>No velocity determined</td>
</tr>
<tr>
<td>8</td>
<td>8</td>
<td>1.7 0.33 6.86 0.01</td>
<td>2.1 0.15 7.0 0.02</td>
<td>2.5 0.88 7.5 0.01</td>
<td>No velocity determined</td>
</tr>
</tbody>
</table>

Table 3.1. Summary of interval velocities and thicknesses determined from the wide-angle profiles. Within each layer the numbers refer, left to right, to the interval velocity (km s\(^{-1}\)), the thickness (km), the two-way travel-times (s) below the seafloor to the bottom of the layer and the standard deviation of the travel times from the T\(^2\)-X\(^2\) plot.
Figure 3.3. Depth converted section along line A in Figure 3.1 close to the positions of the O.D.P. drill sites. The time section was depth converted using a simple velocity function for instantaneous velocity \( V \) at a depth \( H \): \( V = 1.73 + 0.75H \). The positions of the sonobuoy stations and drill sites are shown together with the velocity structure deduced from three of the wide angle profiles.
Figure 3.4. A composite velocity structure for the area around the O.D.P. sites incorporating well (dots), reflection (circles) and refraction data.
velocity data calculated from well logs and from sonobuoy profiles. It seems likely that sonobuoy reduction tends to overestimate velocities in the near surface (1.7 - 1.8 \text{ km s}^{-1} to 1.6 - 1.7 \text{ km s}^{-1} here) where the sonobuoy data invariably average over the top few hundred meters.

3.3 Discussion

3.3.1 Sediment Velocity Structure

The sediment velocities obtained here are in reasonable agreement with the limited amount of sonobuoy data already published from this area (Naini and Leyden, 1973; Hamilton et al., 1974; Curey et al., 1982; Stein and Weissel, 1990). Although generally consistent with a gradually consolidating turbiditic lithology to basement, it is likely that prior to the large-scale development of the Bengal Fan (pre-mid Miocene - the 'O' sediments of Curry and Moore, 1971) a greater proportion of the sediment was pelagic chalks and clays. The jump in interval velocity from 2.5 to 3.3 \text{ km s}^{-1} at 1250 \text{ mbsf} on sonobuoy 4 may represent the presence of these pelagics towards the bottom of the sediment sequence. As sonobuoy 4 gave the best quality record, and given the uniform lateral character on the vertical incidence reflection profiles of the pre-deformational sediment sequence, it is likely that data quality rather than lateral variability prevented this event from being seen on the other records. The depth to the base of the fan sediments is, however, uncertain as there are no discrete reflectors present on the wide-angle profiles below about 1250 \text{ mbsf}.

The average sediment velocity gradient above basement of 0.75 \text{ s}^{-1} compares well with velocity gradients in deep ocean sediment sequences measured elsewhere. Houtz et al. (1970) using results from several hundred sonobuoy stations in the North Pacific have found average velocity gradients of 0.6 - 0.8 \text{ s}^{-1}. Hamilton et al. (1974) present a regression equation based on variable angle reflections from sonobuoys in the Bay of Bengal, Bering Sea, Japan Sea and North Pacific which predicts a decrease in average linear velocity gradient from 1.3 \text{ s}^{-1} at the surface to 0.8 \text{ s}^{-1} at a depth of 1000 \text{ mbsf}. Bachman et al. (1983) give a near surface velocity gradient for the southern Bengal Fan of 1.18 \text{ s}^{-1}. The O.D.P. sonic log data suggest a near surface gradient of about 1.2 \text{ s}^{-1}, in excellent agreement with this. Our average gradient must incorporate these high values but also low values close to the basement, possibly no more than 0.25 \text{ s}^{-1} at 1500 - 2000 \text{ mbsf}. Again, these low values may be more representative of sediments dominated by pelagic components.
3.3.2 Basement Topography and Structure

An interesting question in this area of intraplate deformation concerns the scale of topography on the oceanic basement: to what extent is it original and to what extent is it deformation related? The velocity data now available allow us to calculate basement depth. The amount of pre-deformation topography has been estimated along line A (Figure 3.1) by calculating the thickness of pre-deformation sediment and assuming that the pre-deformation seafloor was flat. The simple velocity gradient used suggests that the thickness of pre-deformation sediment varies within fault blocks by 400 m and by up to 400 m between fault blocks. As mentioned earlier, no significant velocity contrast was detected across the onset of deformation between syn and pre-deformation sediments - as would be expected as there is no gross change in lithology - so variations in syn-deformation sediment thickness should not affect the estimate of original topography. However, some of the variation in sediment thickness may arise from deviation from the simple velocity gradient used, furthermore it is possible that by, say, 1500 mbsf the velocity gradient is significantly lower than the $0.753 \text{s}^{-1}$ used in the depth conversion. This would tend to reduce our estimate of pre-deformation topography to something closer to the abyssal hill topography of 100 - 300 m amplitude and 5 - 10 km wavelength commonly observed in the Pacific (Luyendyk, 1970; Macdonald and Luyendyk, 1985). The Pacific was chosen for comparison because both oceanic crusts were produced at relatively high spreading rates of about 60 mm yr$^{-1}$. Indeed, a short section of depth converted multichannel profile (Chapter 5) suggests an original topography of 100 - 200m. We conclude from this that the topography observed is entirely consistent with deformation having affected an oceanic basement of normal topographic characteristics, in agreement with Shipboard Scientific Party (1989).

A weak reflector just below the top of oceanic layer 2 allowed an interval velocity of 4.1 km s$^{-1}$ to be calculated for the top layer of oceanic layer 2 for sonobuoys 4 and 5. This result is confirmed by refracted arrivals of 4.0 - 4.1 km s$^{-1}$ from this region. At first sight these appear to be low values for oceanic basement velocities, especially as in adjacent areas to the north values in the range of 4.4 - 5.1 km s$^{-1}$ have been reported (Curry et al., 1982; Neprochnov et al., 1988). However, these results are in good agreement with those of Houtz and Ewing (1976) and the sonobuoy results of White (1979) from other areas of similar basement age and we agree with their prognosis of an origin in layer 2A. On the CD 28 wide-angle profiles the layer 2A/2B interface was not defined by refractions and we have no evidence of the velocity at the top of layer 2B.
It is however possible that the interval velocity of 4.0 - 4.1 km s\(^{-1}\) attributed to the top of oceanic basement may actually represent manganiferous-rich mudstones or umbers, possibly interbedded with pillow lavas. These hydrothermal deposits form adjacent to faults near the axial rift zone of a spreading-centre (Rona, 1978). Additional support for this interpretation is given by the multichannel seismic sections (for example Figure 5.2) which show the reflector at the top of basement as fairly smooth. Normally the top of oceanic basement has a rugged appearance. Interbedding with and ponding between pillow lavas would give a mudstone of irregular thickness (Reading, 1986) and could account for the relatively smooth appearance of the top of oceanic layer 2. Without drill samples it is impossible to conclude whether the velocity of 4.0 - 4.1 km s\(^{-1}\) represents manganiferous-rich mudstones, layer 2A, or a mixture of both.

Refractions from the top of layer 3 with velocity 6.8 - 6.9 km s\(^{-1}\) were present from most of the sonobuys. Typically layer 3 refractions will appear as first arrivals, but their strong presence, even on sonobuoy records of modest quality, suggests that the layer 2 / layer 3 boundary is well defined in this area. To calculate the thickness of layer 2 and the depth to the top of layer 3, it has been assumed that the layer 2A velocity of about 4.1 km s\(^{-1}\) is underlain by a velocity gradient in layer 2. The velocity gradient is poorly constrained in our data, but by assuming a gradient of 0.7 s\(^{-1}\) (White, 1979; White and Matthews, 1980) an estimate for the thickness of layer 2 of about 1.5 km can be obtained. The only other estimates of layer 2 thickness in this area are those of Neprochnov et al. (1988), who record values of 1.3 - 1.6 km, and Stein (1984) who state a thickness of 2 km. These thicknesses of 1.3 - 2.0 km are fairly typical for oceanic layer 2 world-wide (Bott, 1982) and appear in no way anomalous as a result or cause of the intraplate deformation. Neprochnov et al. (1988) suggest that beneath the crest of one of the deformation highs in this area oceanic layer 2 is anomalously thin.

We have no direct information on the total thickness of the crust in the area of the drill sites. However, there is some indirect evidence that the crust might be thinner than normal. If the effects of sediment loading in the vicinity of the Leg 116 sites are removed by backstripping according to the methods of Sclater and Christie (1980) and Crough (1983) it is found that unloaded basement is generally deeper than that predicted by Parsons and Sclater's plate cooling model (1977) (assuming a crustal age of 78 Ma for the drill sites). The Crough method predicts a residual depth anomaly ~200 m shallower than the Sclater and Christie method (Figure 3.5). The former is likely to be more accurate because it is based empirically on oceanic data, whereas the Sclater and Christie method is biased towards continental sediments. For the Crough method there is an average residual depth anomaly of ~250 m which increases southwards to as much as ~450 m at 4
Figure 3.5. Backstripped section for the long CHARLES DARWIN N - S profile (Line A in Figure 2.1). The crossed ornamented line is the seafloor while the diamond ornamented line with vertical lines below is the present basement depth. The plus ornamented line is the unloaded basement using the method of Sclater and Christie (1980), while the circle ornamented line immediately above is the unloaded basement predicted using the method of Crough (1983). The gently sloping line with square ornament is the theoretical basement predicted by Parsons and Sclater (1979) assuming a 78Ma age for the northern end of the profile and a spreading rate of 60 mm yr⁻¹. All the above were calculated using a water velocity of 1.50 km s⁻¹ and a sediment velocity of 2.15 km s⁻¹.
- 5°S. There may be a small contribution to this negative depth anomaly from the flexural response of the Indo-Australian plate to loading by the Bengal Fan to the north, however this is likely to be at least an order of magnitude less than the observed anomaly. This would seem to leave two possible explanations: either it is associated with the intraplate deformation or it can be explained in terms of simple Airy isostatic compensation of a thin crust. Since the intraplate deformation is compressive in nature, one would intuitively expect shortening to produce a positive depth anomaly - the opposite of what is observed. The latter hypothesis is therefore favoured. White and McKenzie (1989) infer variations in oceanic crustal thickness due to random fluctuations in ambient asthenospheric potential temperature at the spreading centre. Using crustal and mantle densities of 2800 and 3300 kg m\(^{-3}\) respectively and a normal crustal thickness of 6 km, a residual depth anomaly of -250 m requires a thinning of about 1.2 km. Thus, we may suggest on the basis of residual depth anomalies that the crust is thin, but specifically that layer 3 is thinner than normal since above it was commented on that layer 2 appears to have a normal thickness.

3.3.3 The Anatomy of a Fault Block

The large amount of data at the O.D.P. sites makes this an ideal area to describe some features of the detailed anatomy of a fault block. In previous sections the velocity-depth structure (Section 3.2) and original basement topography (Section 3.3) have been discussed. In this section other characteristics of fault block structure are briefly described including fault orientations, strike-line versus dip-line differences, minor cover fault patterns, onlap patterns indicating pulsed fault activity, and evidence for fault block twisting.

The major reverse faults within the O.D.P. area have an orientation of 90° - 95°E (Figure 3.1) which is only slightly oblique to the axis of the crest of a long wavelength undulation (Figure 1.4) which approximately runs through the O.D.P. sites. This coincidence of the strike of the two spatial scales of tectonic deformation described in Section 1.3.2 means that there is very little basement topography on E - W ("strike line") seismic profiles through the O.D.P. area, as illustrated by two profiles in Figure 3.6. Note also that there is relatively little deformation - in the sediments just minor faulting, for example. Because it is impossible to follow these faults between the very closely spaced profiles they must have a fault length of less than 2 - 3 km. These minor faults may be caused as a consequence of stresses generated in the sediments by the major, basement rooted, reverse faulting.
Figure 3.6 Two east-west ("strike-line") single-channel seismic profiles within the O.D.P. area. The positions of these lines are indicated in Figure 3.1. Note the relatively flat basement and undeformed cover sequence. Minor faulting is restricted to the cover. Vertical lines mark 10 km intervals.
In Section 3.2 the suggestion was made that sediment velocity decreases southwards towards the principal faults. This decrease was explained in terms of increased fracturing in the same direction. As shown in Figure 3.7, this is manifestly true with a large number of minor faults in the hanging wall anticline associated with the reverse fault. The observation of this fracturing is common (e.g. see Figures 5.2 and 5.3) but unlike the major reverse faults (Section 4.3) it does not offset the basement/cover interface.

The cover sequence can be divided into syn and pre-deformational units as discussed in Section 1.7. The pinching out of reflectors within the upper unit suggests that the deformation took the form of the rotation of the fault blocks with net uplift of their southern boundaries. Although the Shipboard Scientific Party (1989) concluded that the motion on the fault blocks has been gradual and has been occurring at a fairly constant rate for the past 7 Ma, analysis of the seismic stratigraphy (Figure 3.8) suggests that motion may have been more erratic. Sediment onlap patterns shown in Figure 3.8, which indicate periods of major fault motion and block rotation, are distinctly episodic. The suggestion that the rate of fault motion may vary through time has implications for seismicity studies (Petroy and Wiens, 1989), referred to again in Chapter 8, which use the level of present day seismicity to predict the amount of shortening since the onset of deformation.

Within the O.D.P. area (Figure 3.1) there is a marked variation in throw of the reverse faults along strike (Shipboard Scientific Party, 1989). The throw on the faults between sites 719 and 718 decrease rapidly from 0.4 s in the east to 0.1 s in the west, while the next fault to the south has a throw of 0.15 s in the east increasing to 0.5 s (Figure 3.7) in the west. This has resulted in "twisting" of the fault blocks as well as rotation. The implications of this twisting for the tectonic regime are discussed in Section 4.6.

The fault blocks within the O.D.P. area are typical of those occurring in the intraplate deformation area, and therefore the anatomy described and inferences made above should be widely applicable throughout the deforming region.

3.4 Summary

1. The crustal velocity structure around the Leg 116 sites is broadly similar to that found elsewhere in the Central Indian Ocean Basin. We find an average velocity gradient for the sediments of 0.753 s⁻¹, although the gradient close to the surface (<400 m), as revealed by the O.D.P. velocity logs, is much higher - up to 1.2 s⁻¹ - and near basement it
Figure 3.7 Single-channel seismic profile, whose position is shown in Figure 3.1 crossing the southern reverse fault in the O.D.P. area. Fracturing is well developed in the hanging wall anticline. This seismic profile crosses the same fault illustrated in Figures 4.7 and 5.2 which is 15 km to the east. The approximate fault throw in this profile is 0.35 s TWT compared to 0.5 s TWT in Figures 4.7 and 5.2, illustrating the variability of fault throw discussed in Section 3.3.3. Generally it is not possible to interpret the onset of deformation across large faults outside the O.D.P. area.
Figure 3.8 Line drawing of part of the multichannel profile illustrated in Figure 5.2 through the fault block containing Sites 717 and 719. This shows that onlap patterns within the sedimentary cover are not uniformly developed as would be expected for gradual and uniform fault motion since the onset of deformation. Four pulses of intense fault block rotation (and hence fault motion) are indicated.
is probably much lower - about 0.25 s\(^{-1}\). Near surface the sediment velocity is about 1.6 km s\(^{-1}\), increasing to 3.4 - 3.5 km s\(^{-1}\) immediately above basement. These higher velocity sediments may have a significant pelagic component.

2. After depth conversion of a seismic reflection profile through the drill sites it is shown that there was pre-deformation topography on basement reminiscent of the normal abyssal hill topography found developed in the Pacific.

3. A compressional wave velocity of 4.1 km s\(^{-1}\), typical of layer 2A, was found for the top of oceanic layer 2. It is however possible that this velocity may have resulted from manganiferous-rich mudstones or umbers immediately overlying or interbedded with pillow lavas. With the assumption of a velocity gradient of 0.7 s\(^{-1}\) for oceanic layer 2 an estimate of 1.5 km for the thickness of this layer was obtained, which is normal.

4. Good quality refractions were obtained from the top of layer 3. Although there is no direct evidence of the thickness of layer 3, residual depth anomalies are consistent with it being anomalously thin.
Chapter 4 - Structural Style Of Intraplate Deformation

4.1 Introduction

One of the major aims of the research project in the Central Indian Ocean Basin was an investigation of the structural style of intraplate deformation which, as discussed in Section 1.8, would involve analysis of the relationships between the faults, long wavelength undulations and fracture zones, with the aim of determining a tectonic model. In this chapter single-channel normal-incidence seismic profiles, magnetic anomaly profiles and pre-existing studies of natural seismicity are used to fulfil this objective. For this study some pre-existing single-channel normal-incidence seismic profiles (collected by the Lamont-Doherty Geological Observatory) were used as well as profiles collected during CHARLES DARWIN cruise 28. The spatial coverage of normal-incidence seismic profiles is shown in Figure 2.2.

It is perhaps pertinent to reintroduce (see Chapter 1.3.2) the two spatial scales of tectonic deformation. The two scales are illustrated by the seismic profile (Line A, Figure 2.1) shown in Figure 4.1. Long wavelength (100-300 km) roughly east-west trending undulations of oceanic basement and overlying sediments with an amplitude of 1-3 km represent the first scale. The spatial distribution of these undulations in the Central Indian Ocean Basin is shown in Figure 1.4. The second scale is represented by the 5 - 20 km spaced faulted blocks.

4.2 Fracture Zones - Undulation Relationship

Offsets in magnetic anomalies may be used to determine the approximate positions of fracture zones (Figure 4.2). Within the study area are two fracture zones, both striking ~005°E to 010°E, a previously documented one called the Indrani (Sclater and Fisher, 1974) at 79°E, and another one at 80.5°E. Characteristic magnetic anomaly studies (Sclater and Fisher, 1974; LaBrecque et al., 1977; Shipboard Scientific Party, 1989) facilitated the determination of oceanic lithosphere age within each block between fracture zones: from ~65 Ma (anomaly 30) in the south, west of the Indrani, to ~78 Ma (anomaly 33R) in the north-east around the O.D.P. Leg 116 sites. The age
Figure 4.1. Seismic profile (Line A, Figure 2.1) showing the two orders of the deformation - the long wavelength undulations and the high-angle faults.
Figure 4.2 Magnetic anomalies along profile and the positions of the fracture zones - the oceanic lithosphere youngs to the west across the fracture zones. Lithospheric age ranges from ~65 Ma (anomaly 30) in the south-west to ~78 Ma (anomaly 33R) in the north-east.
contrast between oceanic lithosphere on either side of the fracture zones within the study area is small, only ~5 Ma.

On the seismic profiles, the fracture zones appear as a rise in oceanic basement towards the younger crust on their western sides, as plate-cooling models predict (Parsons and Sclater, 1977), with little topographic relief as expected for the relatively small lithospheric age contrasts across them. Sediments onlap towards the younger lithosphere. Because earthquake focal mechanism solutions from the intraplate deformation area suggest fracture zone reactivation in a strike-slip sense some recent fault activity in the fracture zones was expected. However no deformation could be detected in the sediments likely to have been caused by reactivation of the fracture zones at any time since their formation (Figure 4.3). Neprochnov et al. (1988, Figure 5) show another crossing of the Indrani in which there is also no evidence for reactivation. However, a fracture zone on the west flank of the Afanasy Nikitin Seamount, the Indira (Sclater and Fisher, 1974), does show recent reactivation (Figure 4.4) and therefore it appears that fracture zone reactivation during deformation has been selective, perhaps being concentrated on the weakest mechanical fractures.

Although data quality on the older seismic profiles is highly variable, oceanic basement is nearly always clearly visible. This enabled the production of a contour map of basement topography, with reasonable spatial data coverage (Figure 4.5), showing the position of the Indrani fracture zone as a clear, linear, 005°E - 010° striking feature. The fracture zone at 80.5°E is far less well defined and its position is, in places, indeterminate.

Within the survey area, the wavelength of the undulations is 150-200 km with an amplitude of 1-2 km. The long wavelength undulations strike ~070°E on average but are discontinuous across fracture zones, with the frequent juxtaposition of highs and lows. This is broadly consistent with the tectonic chart (Figure 1.4) of Geller et al. (1983) which, while not indicating or discussing their positions, shows the axis of the undulations discontinuous across some of the fracture zones (80.5°E, 82.5°E and 86°E). The sparse nature of the seismic profiles available to Geller et al. (1983) led to the interpretation of the apparent continuity of undulations across 79°E, the position of the Indrani fracture zone. This study (Figure 4.5) and that of Neprochnov et al. (1988) show this is clearly not the case. The manner in which the fracture zones striking 005°E - 010°E break up the continuity of the ~070°E undulations is reflected in the patchy character of SEASAT free-air gravity anomaly contours (Weissel, pers.
Figure 4.3 The Indrani fracture zone profiled obliquely during CHARLES DARWIN 28 cruise. There is no evidence for reactivation of the fracture zone in the sediments or the basement. Position of the profile is shown in Figure 2.1.
Figure 4.4 Reactivated fracture zone on the west flank of the Afanasy Nikitin Seamount observed on a single-channel seismic profile and interpretation below (see Figure 2.1 for position). This fracture zone could be mapped on DARWIN profiles further to the north (Scrutton pers com.) and has an orientation of 10°N, consistent with other fracture zones in the area (Figure 4.5). The decrease in sedimentary layer thickness towards the seamount, with onlap onto basement, indicates that the seamount did not cause a significant flexure on the lithosphere during sedimentation and deformation. Note also the lack of faulting for most of the profile line. Presumably most of the deformation has been taken up by motion along the fracture zone.
Figure 4.5 Contour map of oceanic basement in TWT (seconds). The Indrani fracture zone can be observed as a clear, linear, 10°E striking feature at 79°E while the position of the other fracture zone is less clear. The ~70°E striking basement undulations are clearly offset across the fracture zones.
com.). In detail, these anomalies are rarely continuous for more than ~150 km in the east-west direction, roughly the spacing of the main fracture zones (see Chapter 7 for more discussion). Not only are the undulations discontinuous across fracture zones, there is a suggestion that their axes rotate slightly within the fracture zone blocks, from roughly 085°E west of the Indrani to 065°E east of the 080.5°E fracture zone. Strong NE or ENE and NW basement trends (Figure 4.5), the discontinuity of undulations across fracture zones, and the rotation of axes indicate a complex stress regime, which is discussed in detail in Sections 4.5 and 4.6.

The possibility that some of the basement undulations are original features of the oceanic basement topography has been considered. From close examination of the seismic profiles for areas of sediment onlap onto basement and by the production of a total sediment thickness contour map (Figure 4.6) it is possible to make some comments about the original topography. The region west of the Indrani at 4.0 -5.0°S that includes two undulation crests and an intervening trough was originally an elevated area (as revealed by sediment onlap) which was later reinforced by the subsequent deformation to become a basement high. The area in the easternmost fracture zone block around 5.0°S was also an elevated area, although this was not reinforced. Elsewhere, for example around the O.D.P. Leg 116 sites at 1.0°S 80.2°E (Bull and Scrutton, 1990b; Shipboard Scientific Party, 1989), there is evidence for the abyssal hill topography similar to that widely found in the Pacific today, an area of similar sea floor spreading rate. The two elevated areas mentioned are perhaps features too large to be abyssal hills, which typically have relief of 100 - 200 m (Luyendyk, 1970; Macdonald and Luyendyk, 1985), and are best described as low plateaus with basement elevations of ~800 - 900 m above their surroundings. The origin of these plateaus may be closely related to that of the Afanasy Nikitin Seamount which lies close to the eastern edge of the survey area. In general, it appears that the original basement topography was either shorter wavelength (abyssal hills) or longer wavelength (plateaus) than the undulations and did not significantly affect, nor mask, the deformation style and pattern.

4.3 Fault Analysis and the Fault-Undulation Relationship

The second order of deformation represented by the high-angle reverse faults is highly variable in nature. The amplitude, throw direction (sense) and spacing of the faults appear, on preliminary inspection of the seismic profiles, to be complex. However, nearly all the faults have two common characteristics: oceanic basement is offset in a reverse sense; and the faulted sediments can be divided into a lower pre-
Figure 4.6 Isopach map of total sediment thickness. The thickness of sediments reduces from north to south with increasing distance from their source. Arrows indicate the principal areas of sediment onlap onto oceanic basement indicating areas of upstanding basement prior to deformation. The axes of the underlying basement undulations are indicated.
deformational unit and an upper syn-deformational unit, with the onset of deformation dated as approximately 7 Ma (Cochran et al., 1987). Figure 4.7 (see also Figures 1.7 and 5.2 and Enclosure 1) shows an annotated seismic section, shot north-south through the ODP Leg 116 sites, showing two typical faults.

Where the density of Edinburgh University lines was high - for example around the O.D.P. Leg 116 sites and around 81.0°E, 5.0°S (see Figure 4.8) - it was possible to follow a few of the high-angle faults along strike for as far as 40 km. However, more generally, it was not possible to trace the majority of faults between lines only 10 km apart, suggesting a mean fault length of less than 10 km. Where faults could be followed their strike was found to be roughly 090°E to 100°E (for example see Fig 4.8), in agreement with Weissel et al. (1980). A histogram of fault spacing taken from the DARWIN north-south profiles is shown in Figure 4.9A with a mean and median spacing of 6.6 ± 0.3 km and 4.8 km respectively. Statistical details are given in Table 4.1. Fault spacing was measured consistently from the offset of the sediment-basement interface. After allowing for end of profile effects caused by line changes (which probably add a few erroneous small spacings) the trimmed mean (which sorts the data and removes the top and bottom 5%) of 5.9 km is probably a more realistic value. In seeking structural analogies in the oceanic domain, it is significant that this sort of spacing is similar to the wavelength of fault generated topography of 4 - 8 km found on the East Pacific Rise today (Searle, 1984; Macdonald and Luyendyk, 1985), which has a spreading-rate comparable to the approximate 60 mm yr⁻¹ rate developed in the Central Indian Ocean 65-78 Ma (Anomalies 30-33R of Sclater and Fisher, 1974).

The orientation of the faults, 90° - 100°E, which is perpendicular to the strike of the fracture zones, the manner in which they offset oceanic basement, and their short length resulting in an en echelon pattern, are all characteristic of the fault fabric found parallel to spreading-centres. Therefore, it seems likely that the faults are the result of the reactivation of the pre-existing spreadingcentre fabric.

The Edinburgh University profiles (~3520 km of data) were analysed using simple non-parametric statistical techniques to see if any association could be detected between the high-angle faults and the first order of deformation, the long wavelength undulations. In particular to see if the long wavelength topography is produced by faulting.
Figure 4.7 Seismic profile illustrating two of the high-angle faults. The profile is part of Line A (Figure 2.1 for position and 4.1 for whole profile) through the O.D.P. Leg 116 sites (see Figure 3.1). The high-angle faults always offset basement and the sediments can be divided into a pre-deformational and syn-deformational units, separated by an unconformity at 7 Ma (Cochran et al., 1987). See also Figure 5.2 and Enclosure 1 for multichannel profile.
Figure 4.8 The planform of reverse faulting in two areas of the intraplate study area. Top left illustrates the relative positions of the two areas: area A around the ODP Leg 116 sites and area B around 81°E 5°S. The position and orientation of two fracture zones (FZ) developed in the area are also shown. In the detailed maps of area A (top right) and area B (bottom) reverse faults are indicated by dashed lines with a black triangle pointing in the direction of the hanging wall. DARWIN tracks are continuous lines while in area A the dotted lines are additional profiles collected during ODP site surveys and used to constrain the positions of the faults. In area B the dashed lines with the black diamond ornament show the strike of characteristic hanging wall anticlines which could be used to tie between lines. It should be noted that, in the interest of clarity, not every fault has been annotated in area B; those shown are only the larger ones. In both areas the strike of the reverse faults is 90°E - 100°E, roughly perpendicular to the strike of fracture zones (005°E - 010°E) developed nearby. The general short fault length (usually <10 km) and resultant en echelon pattern displayed in area B is similar to the fabric developed close to active spreading centres and is consistent with the reverse faults resulting from the reactivation of the pre-existing spreading-centre formed fabric.

A 3.5 kHz record illustrating Fault S is shown in Enclosure 2.
Within the survey area, it was found that an approximately equal number of faults downthrow towards the spreading centre (to the south: inward) as downthrow away from it (outward), in contradiction of earlier studies which suggested that the vast majority of faults downthrow towards the spreading-centre. Here only a very small majority downthrow towards the spreading centre. When the trimmed mean spacing between faults downthrowing in the same direction are compared, we found that the spacing for the outward facing is 11.0 km while that for the inward facing is 8.0 km (see Figure 4.9, B & C). The corresponding median spacings for all the data are 7.8 km for the outward facing and 5.0 km for the inward facing faults. It is possible to test whether or not these median spacings arise from different populations by using the Mann-Whitney test. When this is done it is found, with 95% certainty, that the fault spacing medians come from different populations. This raises the possibility that prior to reactivation there were two fault sets with different spacing characteristics. On the East Pacific Rise, Searle (1984) distinguished two normal fault sets with different mean fault spacings, one set inward-facing with a mean spacing of 1.7 ± 0.07 km and another set, more widely spaced and outward facing, with a mean spacing of 2.55 ± 0.24 km. These mean spacings are much smaller than those described above from the survey area. The larger mean fault spacing in the survey area may suggest that approximately one in four or five of the faults generated at the spreading-centre has been reactivated to an extent detectable on the seismic profiles by the subsequent deformation.

The magnitude of the fault throw varies from an offset of basement of ~600 m to that so small as to be undetectable on the seismic sections. No relationship could be found between the magnitude of fault throw and position relative to the long wavelength
Figure 4.9 Histograms of fault spacing for north-south profiles
undulations. Examples of the 'Deformation Fronts' noted by Weisel et al. (1980) are found where there are zones of intense fault activity. However, these could not be correlated with the longer wavelength phenomena. A complementary question is to what degree has the second order of deformation contributed to the first order? That is, is the topography represented by the long wavelength undulations fault generated? To answer this it is necessary to investigate the relationship between the sense of faulting (direction of downthrow) and the direction of regional slope of the underlying oceanic basement. Faults occurring on the CD 28 lines were classified according to the sense of throw (south or north) and the direction upslope of the underlying oceanic basement (south or north). This gave four different classes (Table 4.2) which were tested for association via the chi-squared test. For the entire survey, there is an association between sense of faulting and slope of basement such that faults tend to downthrow downslope accentuating the crests of the undulations. However, if certain parts of the survey were tested for association, for example the long CD 28 north-south line at 81.5°E (Line A in Figure 2.1), no relationship was suggested. In summary there is a weak association between the faults and long wavelength undulations and where it is best developed it has accentuated the crests of the undulations.

Although, as mentioned previously, the planform of the sense of faulting (direction of downthrow) is complex, it is possible to map areas in which the majority of faults have the same sense. In Figure 4.10 the sense of faulting is shown overlying the contour map of oceanic basement. Two important conclusions can be drawn from this.

Firstly, there is a change in the sense of throw across the well-formed crests, with faults downthrowing downslope on either side, as suggested by the statistical analysis (e.g., west of the Indrani at 3°S and in the centre block between fracture zones at 4°S). More commonly, however, there is no relationship between the sense of faulting and the underlying basement topography at less well-formed crests and through the troughs (for example in the centre fracture zone block at 3.5°S and for the long DARWIN N-S line). It would seem that although the first order of deformation is not formed by the second order, it is modified by it as manifested by the accentuation of some of the crests. This has produced the 'sharp crest-broad trough' topography sometimes seen on the seismic profiles from this area. The lack of a general
Figure 4.10 The general relationship between fracture zones, basement topography and fault sense.
relationship between the high angle faulting and the long wavelength undulations supports an origin that is fundamentally flexural (including buckling) for the undulations.

Table 4.2

Statistical study of the fault/undulation relationship

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<td>32</td>
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</tbody>
</table>

N is the total number of faults.

T1 is the number of faults where basement rises to the south that downthrow to the north.

T2 is the number of faults where basement rises to the south that downthrow to the south.

T3 is the number of faults where basement rises to the north that downthrow to the north.

T4 is the number of faults where basement rises to the north that downthrow to the south.

S1 is the total number of faults which downthrow to the north.

S2 is the total number of faults which downthrow to the south.

The number of faults which occur on flat basement is given by (N-T1-T2-T3-T4).
Secondly, the fracture zones clearly offset regions of the same fault sense and is perhaps the most convincing evidence to support the hypothesis that the faults are the result of the reactivation of the pre-existing spreading-centre formed fabric. Small variations in the environment of formation of the fault fabric at the spreading centre could have allowed the offset in fault sense observed across the fracture zones (see also Bull, 1990)

4.4 Emplacement of the Afanasy Nikitin Seamount

As discussed in the Section 1.4, Admittance studies (Paul et al., 1990) suggest that the Afanasy Nikitin Seamount was emplaced on young, thin and weak lithosphere, close to the spreading-centre. Such an early emplacement would be expected to lead to effective Airy isostatic compensation of the seamount load.

Analysis of the distribution and onlap patterns of sediment onto basement displays no indication of the seamount causing lithospheric flexure: there is no sediment moat surrounding the seamount, only gradual sediment onlap onto the flanks (Figure 4.4 and 4.11). This is consistent with early emplacement.

Full analysis of the seamount is beyond the scope of this project. However, the principal results, pertinent to the discussion of emplacement, of this ongoing research are now briefly described. Biostratigraphic dating of chalks intercalated with basalts supports an emplacement in Late Cretaceous or Early Tertiary time. Magnetic anomaly reconstructions suggest that the Afanasy Nikitin Seamount was formed at anomaly 33 time and that it may have originally been joined to the Marion Dufresne Seamount, now in the south-west Indian Ocean, at the original spreading-centre. In addition geochemical evidence supports a near ridge-axis emplacement (R.A. Scrutton pers com.).

All the evidence outlined above is consistent with the Afanasy Nikitin Seamount being formed at or very close to a spreading-centre. This has led to Airy isostatic compensation with the seamount, although causing a very large local deflection of the plate, producing a very small wavelength of flexure.
Figure 4.11 Backstripped profile (Figure 4.4) showing gradual decrease of sediment thickness towards the seamount (see also Figure 4.4). Note the absence of a depression around the seamount and that all topography has been smoothed. The position of the profile is shown in Figure 2.1.
4.5 Natural Seismicity and Fracture Zone Reactivation

The large amount of intraplate seismicity recorded in the northern part of the Indo-Australian plate has been dominated by thrust faulting on ~E-W striking faults and left lateral strike-slip motion on faults striking ~N-S (Stein and Okal, 1978; Bergman and Solomon, 1985). Stein and Okal (1978) suggested that the left lateral strike-slip motion along the northern part of the Ninetyeast Ridge marked the decoupling between the Indian and Australian halves of the plate as a consequence of changes in plate boundary on either side of the ridge. There is more resistance to northward plate motion west of the ridge where India collides with Asia than to the east where oceanic crust subducts at the Sunda Trench. However left-lateral strike-slip motion extends well to the west of the Ninetyeast Ridge, as evidenced by focal mechanism solutions of several earthquakes showing nodal planes orientated parallel to the fracture zones. While no earthquakes have been recorded to date along two of the fracture zones (79°E and 80.5°E) in the survey area, the fracture zone immediately to the east (the Indiri of Sclater and Fisher, 1974) is seismically (Bergman and Solomon, 1985; Neprochov et al., 1986) and tectonically (Figure 4.4) active.

For much of the study area, the apparent contradiction between the discontinuity of the undulations across fracture zones and the lack of surface expression of fracture zone reactivation needs to be addressed. Although, as discussed earlier, significant compressional deformation occurs at high crustal and possibly upper mantle levels, this has to be reconciled with the absence of evidence for high level strike-slip motion along the fracture zones. Two possible explanations of this are that either the reactivation is not detectable within the resolution of the seismic profile data or, more likely, rheology controls the tectonic behaviour.

Sandwell and Schubert (1982) argue that fracture zones are not inherent zones of weakness since they support the bending stresses imposed by juxtaposed cooling lithosphere of different ages. They argue that below the fractured surface layer lies coherent oceanic lithosphere that is not significantly weaker than the surrounding lithosphere. However, in this area of the intraplate deformation, the strain rates and stress levels are much higher than those studied by Sandwell and Schubert (1982). The approximate N-S compression acts on 005°E - 010°E striking fracture zones, and earthquake evidence is strong for a left lateral strike-slip regime controlled by fracture zones. The lack of surface expression of reactivated fracture zones suggests a rheological control, such that at high crustal levels the fracture zones are not weaker than the surrounding crust. The surrounding crust is likely to have a relatively low
brittle strength as evidenced by the absence of high magnitude shallow seismicity (both compressional and strike-slip) in the intraplate deformation area (Bergman and Solomon, 1985; Zuber, 1987) which allows very limited transmission of significant stresses away from their source (Bergman, 1986). This agrees with the experimental work of Richard and Krantz (in press) who suggest that not only does a previously faulted zone control the deformation of a strike-slip area in sand-box experiments, but that reactivation of faults in a strike-slip sense can occur at depth without localised surface deformation. They attribute this difference in behaviour to the increasing importance of mechanical cohesion with decreasing depth. Alternatively this implied reduction in localised strike-slip offset with decreasing depth may simply be due to the reactivated slip not having penetrated to the surface yet. The relatively deep mantle seismicity (25 - 40 km) described by Bergman (1986) suggests a localised high strain rate, possibly due to buckling or inverse boudinage and consequently a deeper brittle-ductile transition than normal. The dynamic faulting has yet to reach the surface from such a deep nucleation point.

4.6 Plate Boundary Models and a Transpressive Model for the Survey Area

Following a description of the structural style of deformation consideration must be given to the wider stress system initiating the deformation. Qualitative discussion has already taken place concerning factors influencing the character of seismicity. For quantitative information plate boundary modelling is required, in this instance, of the Indo-Australian plate. Modelling of this type should incorporate slab pull, ridge push, estimates of resistance at the collision zones and basal lithosphere drag. For the Indo-Australian Plate Cloetingh and Wortel (1986) included these factors using the finite-element technique over a uniform elastic plate. Their modelling found a concentration of compressive stresses of several kbar in the Ninetyeast Ridge area, as a consequence of resistance associated with the Himalayan collision and the subduction of young oceanic lithosphere at the northern part of the Sunda Trench. The calculated regional stress field for the Ninetyeast Ridge area and the Bay of Bengal is shown in Figure 4.12A. This stress field closely matches the orientations of maximum horizontal compressive stress (Figure 4.12B) found by the focal mechanism study of Bergman and Solomon (1985). Note that in the survey area the predicted orientation of maximum horizontal compressive stress is somewhat oblique to north-south, 160° - 170°N. A major simplification of their model is that they assume the Indo-Australian plate to be homogeneous. They make no allowance for strike-slip motion along the
Figure 4.12 Regional stress field in the Ninetyeast Ridge area and the Bay of Bengal. (A) - Plate Boundary Calculated Stress Field (Cloetingh and Wortel, 1986). Arrow in bold corresponds to the survey area. (B) - The orientation of maximum horizontal compressive stress inferred from the focal mechanism study of Bergman and Solomon (1985). Redrawn from Cloetingh and Wortel (1986).
northern end of the Ninetyeast Ridge. It is possible that this omission may significantly alter the calculated magnitude of compressive stress, particularly in the Wharton Basin. This is discussed in Chapter 8.

The compressive stresses deduced from earthquake focal mechanism studies (Stein and Okal, 1978; Bergman and Solomon, 1985) and derived from plate boundary forces (Cloetingh and Wortel, 1986) would give a far simpler tectonic fabric than that actually observed. The lack of continuity of the undulations across the fracture zones, the orientation of the axes of the undulations, the strong oblique NE or ENE and NW basement trends, and the earthquake source mechanisms suggest that the fracture zone fabric has substantially modified the stress regime. An apparent change in strike of the undulations towards the Afanasy Nikitin Seamount suggests that this feature may also have modified the stress system. The selective reactivation of the fracture zones may have provided little effective stress linkage across the fracture zones in some areas, with the formation of the long wavelength undulations taking place independently in each block.

One of the conclusions of the O.D.P. Shipboard Scientific Party (1989) was that the drilled fault blocks had been twisted. A consequence of this could be that the reverse faults have a strike-slip as well as a dip-slip component. In the following Chapter an additional reason for interpreting the faults as oblique-slip is outlined. Furthermore, the orientation of the faults is suitable for reactivation in a strike-slip sense. Instead of two Riedel fractures developing, the presence of a pre-existing weakness has led to reactivation of existing structures and has not necessitated the development of a second set of fractures.

The combination of the calculated regional stress field with all the tectonic observations and ideas leads to a tectonic model for the survey area. This model (Figure 4.13) implies a transpressive environment, in which long wavelength undulations of basement form within fracture zone compartments and strike-slip motion occurs both along fracture zones and along the original spreading-centre formed faults. This transpressive regime, which can be thought of as compression with anti-clockwise rotation is a consequence of the plate boundary configuration, with less resistance to motion at the Sunda Trench than the Himalayas.
Figure 4.13 Tectonic environment in the survey area. (Top) Long wavelength undulations form in fracture zone compartments (from Figure 4.5) and are offset across fracture zones (F) which have been reactivated in a left-lateral strike-slip sense. The sense of faulting is also offset as shown in Figure 4.9. (Bottom) Schematic transpressive environment within each fracture zone compartment. The reverse faults are shown as having a right-lateral strike-slip component consistent with the motions of the fracture zones. The transpressive regime is caused by the plate boundary configuration (stresses P as modelled by Cloetingh and Wortel, 1986) and the relative positions of the fracture zones (F) and the original spreading-centre fault fabric and has resulted in anticlockwise rotation.
Although the overall direction of rotation is anticlockwise, as indicated in Figure 4.13 (bottom), this is scale dependent. Within a fracture zone compartment rotation is clockwise, while on an even smaller scale, within a reverse fault-bounded block the motion is anticlockwise. These scale dependent changes in rotation direction are inherent to any complex transpressive model.

The tectonic model will be extended to a diffuse plate boundary across the northern Indian Ocean for the decoupling of two separate Indian and Australian plates in Chapter 8. Resolution of any discrepancy between observable estimates of shortening (from the multichannel seismic profiles, Chapter 5) and plate motion studies estimates, with the recognition of a strike-slip component to the deformation is also included in this final Chapter.

4.7 Summary

From this study of the structural style of intraplate deformation the following principal conclusions can be made:

1. The spacing, orientation (perpendicular to the fracture zones), length along strike, manner in which they offset basement, and the change in fault sense across fracture zones is consistent with the high-angle reverse faults resulting from the reactivation of the pre-existing spreading-centre formed fabric. Furthermore, there is evidence that the complex fault pattern is the result of the reactivation of both sets (inward and outward facing) of pre-existing normal faults.

2. Symmetrical faulting on either side of some of the crests has led to accentuation of these highs. The lack of any general relationship between the first and second orders of deformation supports a flexural origin (including buckling) for the undulations.

3. In the survey area the first and second orders of deformation are discontinuous across fracture zones. The orientation of the axes of the undulations and other basement trends, and regional seismicity studies suggest that the survey area has experienced not only compression but also left lateral strike-slip deformation. The latter developed from the interaction of plate boundary forces with the fracture zone fabric developed in this area. This has led to the development of a transpressive model for the survey area.
Chapter 5 - Oceanic Crustal Structure From Multichannel Seismic Profiles

5.1 Introduction

This Chapter will examine the multichannel seismic profiles acquired during CHARLES DARWIN Cruise 28. The profiles, the first multichannel data to be collected over the intraplate deformation area, are of high quality and yield exciting new images of faults extending throughout the deforming oceanic crust. Hydrothermal circulation and alteration may have been an important control in determining reflection character. The oceanic Moho was not, however, observed. Using the conclusions from Chapter 4, these images are considered in the context of seismicity-depth relations at present day spreading-centres. From this a hypothesis is proposed for the nucleation and propagation of faulting in the Central Indian Ocean Basin. Furthermore, the multichannel profiles facilitate a new estimation of tectonic shortening across the deformation area.

5.2 Seismic Processing

The CHARLES DARWIN cruise produced 1300 km of 12 fold analogue multichannel data. Details of seismic acquisition are given in Section 2.1.3. After taking advice from various sources it was decided that the best way to process the data was to firstly use the facilities of the Global Seismology Unit at the British Geological Survey (BGS) at Murchison House, Edinburgh to convert the data into digital form. Thereafter, and within financial constraints, the GECO Geophysical Company completed the rest of the processing sequence including migration of the final stacked data under the control of the author. A summary of the final processing sequence is shown in Figure 5.1. To date 763.9 km of the multichannel data has been processed to final stack and 311.2 km of this has been migrated. A small panel of data was also depth converted. A full description of the processing sequence is included in Appendix A.
DEMULTIPLEX & REFORMAT

SPHERICAL DIVERGENCE CORRECTION

NORMAL MOVEOUT CORRECTION

PRESTACK EQUALISATION

DECONVOLUTION BEFORE STACK 360/36 (LENGTH/GAP)

COMMON MIDPOINT STACK

SHOT & STREAMER STATICS

TIME VARIANT BANDPASS FILTERING

TIME VARIANT BANDPASS FILTERING

WAVE EQUATION MIGRATION

EQUALISATION

DISPLAY

TIME VARIANT BANDPASS FILTERING

EQUALISATION

DISPLAY

Figure 5.1 Final processing sequence selected for the 12 fold multichannel data: route 1 was used for the final stacked sections, while route 2 was used for those sections that were migrated. See Appendix A for a full description of the processing sequence.
5.3 Observations

The multichannel profiles improved the resolution of the sedimentary reflectors, removed multiples and better imaged anastomosing near vertical faults in the sediments. However, apart from the anastomosing faults, the significance of which are discussed in Section 5.8, the multichannel profiles did not really add any additional information to that gained from the single channel normal-incidence profiles, discussed in Chapter 4, for the sedimentary sequence. New information was gained on the nature of the top basement reflector, and on sub-basement structure.

The most exciting observations are three types of intra-basement reflector with significant dip (> 20°) and strong sub-horizontal reflectors in the top of basement with lower dips. The steeper dipping reflectors can be divided into three types depending on their direction of dip and the degree of offset in the basement/cover interface: i) most spectacular are northward-dipping events which offset the basement/cover interface; ii) rare southward-dipping events which also offset the basement/cover interface; iii) northward dipping events that do not offset the basement/cover interface. Clearly, with the offset of the basement/cover interface and the ubiquitous faulted overlying sediment sequence, the first two types of reflector are faults and will be referred to as northward and southward-dipping basement faults respectively. The third type could be either low-offset fractures or compositional boundaries and will henceforth be referred to as northward-dipping basement reflectors. The non-resolution of the fault planes in the sedimentary cover suggests that the faults have far higher dip through the cover than in basement.

5.3.1 Northward-Dipping Basement Faults

This is the most common type of intra-basement reflector. In the vast majority of cases the basement/cover interface is offset in a reverse sense with characteristic hanging-wall anticlines in the overlying cover. There is no clear example of a normal offset. Fully migrated profiles (Figures 5.2 and 5.3) show typical examples of these northward-dipping reverse faults.

Figure 5.2 illustrates three examples in close proximity. Two of the faults have large basement/cover offset, while the third, probably a less mature fault, has only a negligible offset. All show well-developed hanging-wall anticlines in the overlying sediments. The dip of the faults in basement ranges from 30 - 40° in this example. The multichannel profiles are nearly perpendicular to the strike of the faults and
Sub-horizontal basement reflectors

Northward dipping reverse faults in basement and overlying cover

Figure 5.2 A section of fully processed and migrated 12 channel, 12 fold multichannel profile from north (left) to south (right) through the ODP Leg 116 Sites and perpendicular to the structural grain (located A in Figure 3.1). Vertical exaggeration is about 3:1 in the sediments and 1.5:1 to 2:1 in basement (the top of which is marked by the strong reflector at around 8 seconds). Details of the high-angle fault planes in the sediments are evident: some anastomose and show curvature, some are just developing and some are mature. The fault planes on which the displacement has taken place dip at 30° to 40° in basement and can be traced to depths at around 10s TWT (about Moho depth). The multichannel profiles are nearly perpendicular to the strike of the faults and therefore this apparent dip should be close to the true dip. These faults are interpreted as forming at the mid-ocean ridge to the south in which case they must have been outward-dipping faults. Also illustrated are sub-horizontal basement reflectors which are interpreted as hydrothermal alteration fronts. Note how the front intersects one of the reverse faults. This short profile (Shot points 850 - 1350) is part of a longer section included in Enclosure 1.
Figure 5.3 A section of fully processed and migrated 12 channel, 12 fold multichannel profile from north (left) to south (right), whose position is shown in Figure 2.1, illustrating an isolated large northward-dipping reverse fault, associated minor faulting and a small southward-dipping reverse fault. The northward-dipping fault has a dip of 37° in basement and is interpreted as a reactivated outward-dipping spreading-centre formed normal fault. Note the complex fault planes in the sedimentary cover.
therefore this apparent dip should be close to the true dip. The dip of the faults in sediments is not resolved. Resolution of the faults is poor for the first 0.3 - 0.4 seconds TWT beneath basement, which could be due to an increase in fault dip close to the top of basement, while one of the faults extends down to 10 seconds TWT. Parallelism of basal sediments and top of basement suggests basement was originally quite planar and has since been displaced during the intraplate deformation.

As discussed in Chapter 4, the spacing of faults is irregular, and large amplitude examples of these northward-dipping reverse faults can occur as isolated examples as shown in Figure 5.3. Characteristic features are present here, with hanging-wall anticline, a fault plane dip of 37° and very large (~600 m) reverse offset of the basement/cover interface. The fault is not resolved for the first 0.5 seconds below basement and then is present down to 8.6 seconds.

Following depth-conversion on a selected panel of data the faults appear to be perfectly planar in basement.

5.3.2 Southward-Dipping Basement Faults

The resolution of this type of reflector is much rarer within the survey area and where the faults are resolved they are substantially less clear. In a similar manner to the northward dipping faults, the basement/cover interface is offset in a reverse sense and faulting extends into the cover with the creation of hanging-wall anticlines.

Within the survey area the clearest example of a southward-dipping fault is illustrated in Figure 5.4. The basement reflector is offset in a reverse sense. While the fault is clearly resolved between 9.0 and 10.1 seconds, it is poorly resolved in basement between 8.0 and 9.0 seconds, perhaps because of the large amount of multiple energy present in this part of the section. Because this section was not migrated it is difficult to estimate true fault dip. However, with comparison to other migrated faults, the fault is likely to have dip between 35 and 40°.

More generally, the faults of this group appear to have dips of above 40 - 45°. In comparison with the basement/cover offset for the northward-dipping faults, the offset appears, while remaining reverse, to be more vertical for this class of fault. The more steeply-dipping nature of the southward-dipping faults may explain their relatively poor resolution in basement.
Figure 5.4 A section of fully stacked 12 channel, 12 fold multichannel profile from north (left) to south (right), whose position is shown in Figure 2.1, illustrating a southward-dipping reverse fault which is interpreted as a reactivated spreading-centre formed inward-dipping normal fault. The fault plane in basement is likely to have a dip between 35 and 40°.
The conclusion, from Chapter 4, that the complex fault pattern that has developed is the result of the reactivation of two sets of original spreading-centre inward and outward facing normal faults has clear implications for the multichannel observations. Faults dipping at 30 - 40° towards the north represent reactivated spreading-centre formed outward-dipping normal faults while faults dipping at 40 - 45°+ towards the south are reactivated original inward-dipping normal faults.

Another conclusion from Chapter 4 was that approximately equal numbers of inward and outward-dipping faults are present in the survey area. The resolution in basement, of far more of the outward-dipping faults, and their greater clarity relative to the inward-dipping set, could be because of their lower dip. This is consistent with the observation of generally steeper reverse offset of the basement/cover interface for the inward-dipping faults.

5.3.3 Northward-Dipping Basement Reflectors

Even in areas where there is little tectonic deformation there are regularly spaced northward-dipping strong sub-basement reflectors. These reflectors (Figure 5.5) typically start 0.4 - 0.5 seconds TWT beneath the top of basement and continue for 0.4 seconds. With dips of 20 - 30° (estimated from migrated and stacked sections) they generally have slightly lower dip than the northward-dipping basement faults. The recognition of northward-dipping reflectors which do not offset the basement/cover interface is interesting. There would seem to be two possible interpretations. Either they are original outward-dipping normal faults on which reactivation is just initiating, or they are some form of compositional boundary within the crust, such as the top of the spreading-centre magma chamber. The regular spacing (5 - 6 km in Figure 5.5) of these reflectors is consistent with both of these hypotheses. However, the observation of the flexure in the top of basement in close proximity to some of the dipping reflectors, suggesting fault reactivation, and their similar reflection character to fully developed faults (transparent in the uppermost crust) strongly supports the view that they are original spreading-centre formed outward-dipping normal faults. Estimation of average dip from a mixture of migrated and unmigrated profiles is difficult, although the difference between 20° - 30° and 30° - 40° between the low offset faults and their fully reactivated counterparts is likely to be real. If the difference is real then it suggests that either originally steeper dipping outward-dipping faults are reactivated preferentially or steepening occurs with reactivation.
Figure 5.5 A section of fully processed and migrated 12 channel, 12 fold multichannel profile from north (left) to south (right), whose position is shown in Figure 2.1, illustrating regular northward-dipping reflectors that do not offset the basement/cover interface. These are interpreted as unreactivated spreading-centre formed outward-dipping faults. The examples here dip at the 20 - 30°. Note the unfaulted cover sequence above.
5.3.4 Sub-Horizontal Basement Reflectors and Non-Resolution of the Moho

Where the quality of the multichannel profiles is particularly high (e.g. Figure 5.2) sub-horizontal basement reflections are observed within 0.3 - 0.7 seconds of the top of oceanic basement.

The strong reflectors are usually undulatory and often intersect with either northward or southward-dipping reverse faults. Illustrated in Figure 5.2 is an example, with the depth to the basement reflector from the top of oceanic layer 2 varying from 0.35 to 0.7 seconds TWT. Note that this reflector has a geometry dissimilar to any of the overlying strata, thus removing any possibility that it represents a multiple. Much of the energy beneath basement in Figure 5.2 does however represent multiples. This reflector appears to be truncated at one end by faulting.

There are two possible explanations for the sub-horizontal basement reflections. Either they represent the magmatic boundaries between the dykes and gabbros (McCarthy et al. 1988) or subsequently imposed hydrothermal alteration fronts (White et al. 1990).

Hydrothermal circulation along fault planes has been proposed to explain the non-linear temperature gradient recorded down one of the ODP Leg 116 Holes (Shipboard Scientific Party, 1989) and to explain the localised high heat-flow in the intraplate area (Geller et al., 1983). Hydrothermal circulation is known to penetrate 1 - 3 km into the crust at spreading-centres and to result in a thin transition zone between unaltered and hydrothermally altered layers (Campbell et al., 1988). The observation of fault planes truncating the sub-horizontal reflectors, the depth of the reflectors ( < 2 km beneath basement), and the presence of vigorous hydrothermal circulation strongly supports the sub-horizontal reflectors representing hydrothermal alteration fronts.

Nowhere along the multichannel profiles is the oceanic Moho resolved. This is perhaps surprising when faults extending down to 10 seconds TWT, roughly the expected level of the oceanic Moho, are clearly resolved. Another question is why are the imaged faults such prominent reflectors? The answers to these questions are interlinked.
For the faults to have such prominent reflection character there must be a large contrast in acoustic impedance across them: implying a large change in either density or seismic velocity or both. A possible explanation of this could be the presence of fluids along fault planes. This explanation is supported by the evidence for hydrothermal circulation as discussed above.

Although the seismic source aboard the CHARLES DARWIN was sufficiently large to resolve the faults with their enhanced acoustic impedance contrast it was not sufficient to resolve the Oceanic Moho, which has apparently lower acoustic impedance contrast, at the same depth.

5.4 Comparison with Fault Structures Imaged in the Western North Atlantic

Images of the internal structure of the oceanic crust have also been recently obtained by White et al. (1990) using a two-ship multichannel seismic survey in the western north Atlantic. White et al. (1990) divide images obtained perpendicular to the ridge axis (flowline profiles) into five classes. Figure 5.6 illustrates the five classes, the most prominent images being planar reflectors dipping at 20 - 40°+ in the lower crust (1 in Figure 5.6). Two-thirds of these dip toward the spreading-centre and a third away. Inward-dipping reflectors tend to have a dip 5 - 10° steeper than reflectors that dipped outwards.

The other classes were: steeper faults (average dip of 35°) cutting throughout the crust (class 2); bright sub-horizontal reflectors 1 - 3 km beneath the top of basement (class 3); horizontal reflectors at Moho depths (class 4); and finally short reflectors in the uppermost crust that dip in either direction with highly variable dips (class 5).

On ridge-parallel or isochron profiles the most striking features are low-angle planar reflectors cutting through the entire crust (class 6 in Figure 5.6) once every 20 km, with a mean dip of 20°.

For the dipping reflectors White et al. (1990) conclude that they have imaged two different structures, one striking parallel to the ridge crest and dipping 30° - 40° towards the ridge crest and the other dipping 20 - 30° to the south. All the whole crustal reflectors imaged on flowline profiles are interpreted as inward-facing normal faults originally formed at the spreading-centre with fracturing, hydrothermal alteration, and steepening of the faults in the upper crust cited as reasons for resolution of some of the faults only in the lower crust. They also speculate that some
Figure 5.6 Structures observed in the western North Atlantic by White et al. (1990). The various classes of reflector are discussed in the text. (From White et al., 1990).
of the lower crustal reflectors may have magmatic origin. The shallower dipping reflectors imaged on the isochron profiles are interpreted as later compressional failures, generated by thermo-elastic stresses as the oceanic lithosphere cools. Note though that tensional structures would be expected on flowline profiles, rather than compressive, as a result of cooling induced thermo-elastic stresses.

In comparison with the DARWIN Indian Ocean profiles the dominant fault set that is imaged is the inward-dipping set as opposed to the outward-dipping set. However, it should be noted, and this was not stressed by White et al. (1990), that a third of the lower crustal dipping reflectors on flowline profiles were dipping away from the spreading-centre, and could be interpreted as original outward-dipping normal faults. Interestingly, these possible outward-dipping faults have lower dip than the inward-dipping faults, which is consistent with the observations from the Indian Ocean. In general, the fault sets in the Atlantic oceanic crust have lower dips than those imaged in the Indian oceanic crust.

Although in the DARWIN multichannel profiles usually only the outward-dipping faults are imaged below basement, statistical analysis of the Indian Ocean fault sets suggests that approximately equal numbers of inward and outward-dipping faults (Chapter 4) have been reactivated. In the Atlantic two-thirds of the faults imaged appear to be inward-dipping. This difference, although the database is pretty thin, can be interpreted as being due to differing acquisition systems, preferential reactivation of one set or, more likely in terms of spreading rate. The spreading rate that formed the Atlantic crust (~20 mm yr⁻¹) is approximately three times slower than that which formed the Indian crust (~60 mm yr⁻¹). A possible mechanism for generating this difference in fault style dependent on spreading-rate is discussed in Chapter 8.

In the Atlantic crust, White et al. (1990) propose an origin for the sub-horizontal reflectors 1 - 3 km beneath the top of basement in terms of a hydrothermal alteration front. This is in agreement with the origin suggested above for similar phenomenon observed in the Indian Ocean.

White et al. (1990) suggest that the horizontal strong reflectors observed within the lower crust and upper mantle may be caused by sill intrusion or by variations in fluid content. This layering towards the bottom of the crust is not observed in the Indian crust. A combination of two factors could explain this non-resolution. In the Atlantic a more powerful seismic source was used and so it is possible that the layering was
not imaged because of an inadequate acoustic source. Furthermore, the disruption of the Indian crust by the deformation, and large amounts of fluid penetration, may prevent the imaging of the lower crustal reflectors because of the greater reflection of seismic energy at higher levels and by the fault planes.

5.5 Comparison of Fault Dip with Seismicity at present day Spreading-Centres

The character of reactivated spreading-centre formed inward and outward-dipping normal faults can be compared to results of earthquake studies and hence faulting along active mid-ocean ridges. Huang and Solomon (1988) found that large earthquakes along the axes of slow spreading mid-ocean ridges have mechanisms indicating normal faulting on planes dipping at 40 - 55°. This is broadly consistent with the dips of faults observed in the Central Indian Ocean Basin of 30 - 45°+ (Bull and Scrutton, 1990a). Huang and Solomon (1988) suggest that much of the fault activity is likely to be concentrated on inward-dipping faults given that the topographic relief of the median valleys of slow-spreading ridges is predominantly formed by this fault set (Macdonald and Luyendyk, 1977). Additionally seismic moments and source durations at active spreading-centres are found to be consistent with typical fault lengths of 10 km. It will be remembered that one of the conclusions of Chapter 4 was that the mean length of reactivated faults in the intraplate deformation area was less than 10 km. However, while Huang and Solomon's (1988) study is useful for crude comparison with observations in the Central Indian Ocean Basin, their work was on slow spreading ridges. Little analogous work has been carried out on fast spreading ridges (barring the study of Riedesal et al. (1982) discussed below) similar to the one that produced the Central Indian Ocean Basin lithosphere.

Earthquake seismology studies along mid-ocean ridges are particularly useful in determining the depth extent of faulting and hence giving important information on the depth to the brittle-ductile transition in the axial region. Using the assumption that the centroid depth marks the mean depth of fault slip, it appears that faulting, for half spreading-rates of 20 mm yr\(^{-1}\) or less, extends from 2 - 10 km beneath the seafloor - into the upper mantle (Huang and Solomon, 1988). Additional evidence for the whole crustal extent of faulting at these spreading rates of spreading-centre faulting is given by the whole crustal faults imaged by White et al. (1990) in the western North Atlantic and discussed in Section 5.4.
The dependency of the thickness of the brittle layer on spreading-rate is well illustrated by the study of Huang and Solomon (1988). Figure 5.7, taken from their paper, shows that for slow-spreading ridges the greatest centroid depths shallow with increasing spreading rate. That is, brittle thickness decreases with increasing spreading-rate. Note that for half spreading-rates of 25 mm yr\(^{-1}\) the brittle layer may be as little as 3 km thick. Although, as already mentioned, little analogous work has been carried out on fast-spreading ridges, the thickness of the brittle layer would be expected to decrease further. One local study (Riedesal \textit{et al.}, 1982) of microseismicity on the East Pacific Rise at 21°N, where the half spreading-rate is 36 mm yr\(^{-1}\) (MacDonald and Luyendyk, 1985), concluded that fast-spreading ridges are characterised by only small magnitude earthquakes, and that the brittle layer was only 2 - 3 km thick.

From these studies it is possible to conclude that for the half spreading-rate that formed the Central Indian Ocean lithosphere (~60 mm yr\(^{-1}\)) the thickness of brittle deformation would only have been a few kilometres. Whole-crustal faulting at the original spreading-centre is unlikely. Therefore, the deepest portions of the faults imaged on the multichannel profiles, which extend to lower crustal and possible mantle depths, may not have originated at the spreading-centre (see also Bull and Scrutton, 1990a). This has important implications for fault propagation and the mode of deformation in the intraplate deformation area.

5.6 Implications for Fault Propagation

The recognition that the deepest parts of the reverse faults on the multichannel profiles are unlikely to have been formed at the spreading-centre clearly suggests that they were formed during the later deformation. Despite the clear evidence for faulting there is a lack of recorded earthquakes associated with the crustal deformation in the Central Indian Ocean Basin. This could be due to the brittle strength of the oceanic crust and uppermost mantle being too low to sustain significant stresses (Bergman, 1986) resulting in low magnitude seismicity that is below the level of detection on the world-wide seismograph network. The focal depths of large magnitude earthquakes recorded in this area have a range of 10 - 40 km, but are concentrated at depths between 29 - 39 km (Bergman, 1986) extending down to the expected depth of the brittle/ductile transition. It is interesting that this may be compared with the pattern of continental seismicity in the seismogenic upper continental crust (e.g. Jackson, 1987) with the largest earthquakes nucleating at or near the base of the brittle layer. By analogy, recent brittle deformation in the Central Indian Ocean Basin may nucleate at
Figure 5.7 Centroid depth versus half-spreading rate predicted by the plate velocity model of Minster and Jordan (1978). There is a clear indication of a decrease in centroid depth, and hence brittle thickness, with increasing spreading-rate. (From Huang and Solomon, 1988).
upper mantle depths and propagate into and through the crust, in the upper part by the reactivation of pre-existing spreading-centre formed faults. This hypothesis would suggest that the fault structures may extend much deeper than the images presently observed on the multichannel seismic profiles. The loss of reflection strength with depth may reflect the loss of fluid circulation in the faults at deeper levels.

5.7 Reverse Faulting and Long Wavelength Undulations

One of the major conclusions of the study of structural style (Chapter 4) was that the second spatial scale of deformation, the reverse faulting, had not formed the first spatial scale, the long wavelength undulations. There is an underlying flexural cause for the undulations. However, there is evidence that the two scales are interrelated. Reverse faulting has modified the long wavelength undulations leading to accentuation of some of the crests (Chapter 4). Resolution of the relationship between the two spatial scales of deformation is a fascinating problem and will be addressed in Chapter 8.

As described in Chapter 1 two theories have been advanced for the formation of the long-wavelength undulations: buckling and inverse boudinage. Could either of these deformation styles be the cause of the underlying flexural behaviour? To my mind it is possible to envisage the core of the lithosphere driving the reverse faulting from below by either of these mechanisms. Geometry alone, with the core of the lithosphere behaving in either an elastic or plastic manner does not preclude either of these hypotheses (Figure 5.8). Further consideration of lithosphere rheology (Chapter 6) and modelling (Chapters 7 and 8) is needed to determine the mode of flexural control.

5.8 A Strike-Slip Component to Faulting

As discussed in Section 4.6 the reverse faults are in the correct orientation for reactivation with a strike-slip component. Additional evidence for these faults having oblique-slip is given by the twisting of fault blocks observed by Shipboard Scientific Party (1989) around the Leg 116 Sites.
Figure 5.8 Cartoon illustrating deformation propagation and possible driving mechanisms. (1) Although fault fabric accentuates the crests relative to the troughs, the undulations are not formed by faulting: there is an underlying flexural mechanism (2 - arrows). The driving mechanism, which could be buckling (3 - dashes) or inverse boudinage (4 - dots), causes stress concentration and the initialisation of faulting (5) in the region of the brittle/ductile transition. Brittle deformation then propagates upwards (5) reactivating the original spreading-centre fabric in the crust. Modelling (see Chapters 6 and 7) is required to determine the driving mechanism.
Recent laboratory studies by Richard and Krantz (1990) suggest that one reason for a reactivated fault to significantly increase its dip in the cover is the presence of a strike-slip component as well as dip-slip. The reverse faults in the deformation area change their dip from 30 - 45° in basement to 65°+ in the cover. While some of this change in dip may be due to the change in lithology, with stress-drop in the cover, it is possible that an additional reason is the presence of a strike-slip component.

While it is not possible to prove that the 'reverse' faults have a strike-slip component their inclusion as oblique-slip faults in the transpressive regime proposed in Chapter 4 for the survey area seems sensible.

5.9 Estimates of Shortening from the Multichannel Profiles

The multichannel profiles can be used to make an estimate of the amount of shortening caused by the intraplate deformation. With knowledge of the spatial extent of the deformation and the timespan of activity an estimate of the rate of shortening can be obtained.

As already discussed the tectonic deformation can be divided into two spatial scales represented by the long wavelength undulations and reverse faults respectively. Estimates of shortening will be made for each of the spatial scales in turn.

5.9.1 From the Long-Wavelength Undulations

Estimates of the amount of shortening were obtained by two methods. Firstly, the top of oceanic basement along the long DARWIN N-S multichannel profile (Line A in Figure 2.1) was coarsely digitised and depth-converted. The digitising interval was greater than 10 km to ensure that no contribution was included from the reverse faulting or from small-scale original topography. The approximate length of the basement surface was found by simply summing all the perpendicular distances between adjacent points. The percentage shortening was then found using the profile length. This was found to be 0.02 %.
Figure 5.9 Schematic diagram illustrating shortening estimate methodology for the two spatial scales of the deformations.

(A) - The long wavelength undulations are assumed to approximate to a cosine wave. The cosine wave is integrated over all the line increments dl to find the restored length. Appendix B details the mathematical derivation of the horizontal shortening over a fold train of length F. Note that (A) has enormous vertical exaggeration.

(B) - For the reverse faults the method of Weissel and Geller (1981) was used. The construction (B) is redrawn from Weissel and Geller (1981).
An alternative method is to assume that the long wavelength undulations can be approximated by a cosine curve (Figure 5.9A) of wavelength $\lambda$, amplitude $A$ and length $F$. The restored length $I$ can be found by evaluating the line integral over the fold train length $F$

$$I = \int_0^F (B \sin^2 kx + 1)^{0.5} dx$$

where $B = \left(\frac{A}{\lambda}2\pi\right)^2$; $k = \frac{2\pi}{\lambda}$

Which for $\lambda>>A$ can be approximated to

$$I = F + \frac{A^2 \pi^2 F}{\lambda^2}$$

The full derivation of this result is given in Appendix B. Using a fold length of 1500 km (the total N-S spatial coverage of the deformation), an average wavelength of 200 km and an average amplitude of 1.5 km (3) yields a restored length of 1500.83 km. A shortening of only 0.83 km over 1500 km corresponds to a percentage shortening of 0.05%. If the deformation is assumed to have begun at 7 Ma this means a shortening rate of 0.12 mm yr$^{-1}$. Even for extreme values (a wavelength of 150 km and an amplitude of 2 km) this only yields a spatial shortening of 2.6 km, a percentage shortening of 0.17% and a shortening rate of 0.37 mm yr$^{-1}$.

From both of the above methods it seem unlikely that the deformation reflected in the long wavelength undulations has caused any more than 0.1% shortening. This result agrees with the conclusion of Weissel and Geller (1981) that the long wavelength undulations have contributed little to the total shortening. Since the deformation extends over 1500 km north to south this suggests that at most \sim1.5 km of shortening could be due to the undulations. Over the 7 Ma since the onset of deformation this corresponds to a maximum shortening rate of 0.2 mm yr$^{-1}$.

5.9.2 From the Reverse Faulting

Previous estimates of shortening due to reverse faulting have been hampered by the quality of single-channel seismic profiles available (Weissel and Geller, 1981). The multichannel seismic profiles should facilitate a better estimate of shortening.
Weissel and Geller (1981) proposed that the construction shown in Figure 5.9B could be used to determine the amount of faulting. This construction was used on both migrated and depth-converted sections to try and deduce the average horizontal shortening due to reverse faulting.

Even for the best-imaged faults (e.g. Figures 5.2 and 5.3) it is difficult to estimate the amount of crustal overlap. On the largest faults the crustal overlap could be as much as 0.5 km. This typically results in 7 - 8 % shortening for values of \( \alpha \) (top of fault block dip) of 6 - 7° for these faulted blocks. However, the average crustal overlap is much less than 0.5 km and an average crustal value of 0.1 km with an average dip for \( \alpha \) of 3° was estimated from the multichannel profiles. Using a mean fault spacing of ~6 km (from Chapter 4) these values give a shortening of 1.7 %. The errors on this estimate are large, perhaps \( \pm \) 1 %. This gives, over the 1500 km deformation area, 25.5 (\( \pm \) 15.5) km of shortening and, for the 7 Ma period of deformation, a shortening rate of 3.6 (\( \pm \) 1.2) mm yr\(^{-1}\). This shortening rate estimate is rather more than Weissel and Geller's (1981) estimate of 1 mm yr\(^{-1}\) made from single-channel profiles.

5.10 Summary

1. Perhaps the most important result of the work presented in this Chapter is the proposition that deformation nucleates at the brittle/ductile transition and propagates up through the upper mantle and crust in part by the reactivation of original spreading-centre faults in the upper crust. In this model the thrust faulting recorded by earthquakes at 25 - 40 km depths relates directly with the crustal faulting. At least the upper part of the lithosphere is behaving in a brittle manner.

2. Another important conclusion is that outward-dipping faults formed at spreading-centres are likely to have a significantly lower dip than their inward-dipping counterparts (30 - 40° as opposed to 40 - 45° +). The average dip of ~40° is broadly consistent with seismicity studies along mid-ocean ridges.

3. Hydrothermal circulation has had considerable influence on the reflection character of events beneath basement. The presence of fluids along fault planes may explain their excellent resolution down to Moho depths. The seismic source aboard the DARWIN was adequate to resolve the faults at these depths but not the Moho, which may have a lower acoustic impedance contrast.
4. The change in dip of the fault planes between basement and cover can be interpreted as suggesting that the faults have a strike-slip component. This would be consistent with the transpressive model proposed in Chapter 4 for the survey area.

5. Estimates of the shortening rate made from the multichannel profiles suggest that the long wavelength undulations contributed little to the average shortening rate (< 0.2 mm yr\(^{-1}\)), whereas reverse faulting is estimated to have produced an average shortening rate of 3.6 (± 1.2) mm yr\(^{-1}\). In total this gives a maximum shortening rate of 3.8 (± 1.2) mm yr\(^{-1}\). This estimate will be compared with plate motion models estimates in Chapter 8.

6. Presently available geophysical datasets are inadequate for observationally determining the mode of formation of the long wavelength undulations: modelling is needed.
Chapter 6 Analogue and Numerical Models of the Intraplate Deformation

6.1 Introduction

In Chapter 5 it was concluded that the mode of deformation in the core of the lithosphere, forming the long wavelength undulations and driving the reactivation of fault fabric, could not be determined without modelling. This Chapter describes attempts to model lithosphere deformation in the Central Indian Ocean Basin using the analogue (physical or sandbox) technique and numerical methods.

The analogue modelling was successful and favoured decisively one of the hypotheses for the generation of the long wavelength undulations. In addition, growth rates of undulations observed in the model can be closely correlated with the natural system. Numerical modelling using the Finite Element software of Professor Martin Bott, University of Durham, was however, unsuccessful. Preparatory work is described here, including a review of rheological models of the oceanic lithosphere.

6.2 Rheological Models of the Oceanic Lithosphere

6.2.1 Introduction: Experimental evidence for rheology

The concept of rigid plates (constituting the lithosphere) overriding a viscous asthenosphere has been successfully applied to the ocean basins. From observations of flexure over seamounts and deep-sea trenches it is clear that the oceanic lithosphere is capable of supporting large differential stresses over long time periods. Yield stress envelopes constructed from experimental work (Goetze and Evans, 1979; Brace and Kohlstedt, 1980) confirm that the strength of the oceanic lithosphere is such that for normal intraplate stresses non-elastic deformation will not occur.
Early models of flexural features used a simple elastic plate model for the oceanic lithosphere (see Section 6.2.2.1). These have been used to define plate bending resistance in terms of flexural rigidity and elastic thickness. Watts (1978) found, by calculating flexural rigidities and effective plate thicknesses for oceanic lithosphere of differing ages at the time of loading, that flexural strength increases with age.

Since the oceanic crust is only 5 - 6 km thick, it can be assumed that the mechanical properties of the oceanic lithosphere in the ocean basins are likely to be controlled by the materials of the upper mantle. In recent years a wealth of results have been published from experimental work on the determinations of the frictional, fracture and flow properties of peridotite and olivine, which are the dominant rock and mineral phase respectively of the upper mantle (see Kirby, 1983; Kirby and Kronenberg, 1987 for reviews). In simplest terms these studies have resulted in a rheological model in which, for upper regions of the lithosphere, the stresses may be predicted by Coulomb laws for fracture and frictional sliding along pre-existing weaknesses. This brittle layer obeys the relation between the shear stress $\tau$ on a sliding plane and the normal stress along the plane $\sigma_m$ obtained in laboratory studies by Byerlee (1978)

$$\tau = 50 \text{ MPa} + 0.6\sigma_m \quad (\text{if } \sigma_m > 200 \text{ MPa for } z_m > 4 \text{ km})$$

The stress difference necessary to deform the upper brittle lithosphere increases linearly with depth. However, Ord and Hobbs (1989) using the results of laboratory experiments suggest that the Byerlee relation only holds for the uppermost part of the lithosphere. At moderate depths (>10 - 15 km) the shear stress necessary for failure is lower than Byerlee relation predictions. Nevertheless, at these depths the rocks are still "localising materials", materials in which the shear stress necessary to activate a deformed zone decreases with the velocity of the slip rate. At these levels, faulted zones and nonhomogeneous perturbations develop preferentially to homogeneous deformations. Additionally it should be noted that the presence of fluids at depth within the oceanic crust (Chapter 5) may lead to significant reductions in strength within the brittle field. Confirmation of the obvious brittle behaviour of at least the oceanic crust in the Central Indian Ocean Basin is shown by the recognition of faults penetrating throughout the oceanic crust (Chapter 5).

Brittle behaviour does not extend to the lower parts of the lithosphere where, at higher temperatures, ductile creep predominates. Ductile creep of materials of the upper mantle is well known due to many laboratory measurements (e.g. Carter and Tsenn, 1987).
In summary the oceanic lithosphere can be divided into two parts: an upper one including the crust and upper mantle, is made of localising brittle materials and a lower one of non-localising viscous material. The boundary between the two domains for a given surface heat flow can be calculated, although with substantial errors.

Table 6.1 - Definitions and Values of Parameters and Functions

<table>
<thead>
<tr>
<th>Parameters/Functions</th>
<th>Definition</th>
<th>Value/Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Amplitude of Buckling</td>
<td>m</td>
</tr>
<tr>
<td>A₀</td>
<td>Initial Amplitude</td>
<td>m</td>
</tr>
<tr>
<td>D</td>
<td>Flexural Rigidity</td>
<td>Nm</td>
</tr>
<tr>
<td>E</td>
<td>Youngs modulus (relaxed)</td>
<td>6.5x10^{10} Nm⁻²</td>
</tr>
<tr>
<td>ε</td>
<td>Strain Rate (natural)</td>
<td>10⁻¹⁶ s⁻¹</td>
</tr>
<tr>
<td>g</td>
<td>Acceleration of gravity</td>
<td>9.82 m s⁻²</td>
</tr>
<tr>
<td>h</td>
<td>Elastic thickness</td>
<td>m</td>
</tr>
<tr>
<td>h₁₀</td>
<td>Analogue Brittle Layer Thickness</td>
<td>m</td>
</tr>
<tr>
<td>h₁₁</td>
<td>Natural Brittle Layer Thickness</td>
<td>m</td>
</tr>
<tr>
<td>h₂₀</td>
<td>Analogue Ductile Layer Thickness</td>
<td>m</td>
</tr>
<tr>
<td>K</td>
<td>Coefficient of Thermal Conductivity</td>
<td>3.0 (+0.5) Wm⁻¹ K⁻¹</td>
</tr>
<tr>
<td>l₀</td>
<td>Natural Folding Wavelength</td>
<td>m</td>
</tr>
<tr>
<td>l₁</td>
<td>Buckling Wavelength</td>
<td>m</td>
</tr>
<tr>
<td>L₀</td>
<td>Length of analogue system</td>
<td>m</td>
</tr>
<tr>
<td>L₁</td>
<td>Length of natural system</td>
<td>m</td>
</tr>
<tr>
<td>μₐ</td>
<td>Viscosity of the analogue system</td>
<td>Pa s</td>
</tr>
<tr>
<td>μₐ</td>
<td>Viscosity of the natural system</td>
<td>Pa s</td>
</tr>
<tr>
<td>μₐ₀</td>
<td>Viscosity of the Brittle Layer</td>
<td>Pa s</td>
</tr>
<tr>
<td>μₐ₁</td>
<td>Viscosity of the Ductile Layer</td>
<td>Pa s</td>
</tr>
<tr>
<td>q</td>
<td>Growth rate factor</td>
<td></td>
</tr>
<tr>
<td>ρₐ</td>
<td>Density of analogue system (Heavy water)</td>
<td>1400 kg m⁻³</td>
</tr>
<tr>
<td>ρₐ</td>
<td>Density of natural system (Asthenosphere/Mantle)</td>
<td>3300 kg m⁻³</td>
</tr>
<tr>
<td>ρₐ</td>
<td>Density of crust and sedimentary infill</td>
<td>2600 kg m⁻³</td>
</tr>
<tr>
<td>ρₐ</td>
<td>Density of Sediments</td>
<td>2300 kg m⁻³</td>
</tr>
<tr>
<td>R</td>
<td>Gas Constant</td>
<td>1.98 cal/K mol</td>
</tr>
<tr>
<td>σₐ</td>
<td>Critical buckling stress</td>
<td>Pa</td>
</tr>
<tr>
<td>σₐ</td>
<td>Normal stress</td>
<td>Pa</td>
</tr>
<tr>
<td>tₐ</td>
<td>Time scale in model</td>
<td>s</td>
</tr>
<tr>
<td>tₐ</td>
<td>Time scale in nature</td>
<td>s</td>
</tr>
<tr>
<td>T</td>
<td>Temperature at a depth z</td>
<td>°K</td>
</tr>
<tr>
<td>τ</td>
<td>Shear stress</td>
<td>Pa</td>
</tr>
<tr>
<td>νₐ</td>
<td>Poisson's Ratio</td>
<td>0.25</td>
</tr>
<tr>
<td>w</td>
<td>Flexural topography</td>
<td>m</td>
</tr>
<tr>
<td>z</td>
<td>Depth</td>
<td>m</td>
</tr>
</tbody>
</table>
6.2.2 Elastic Models and Intraplate Deformation

6.2.2.1 Inapplicability of Elastic Models For Quantitative Studies

Although laboratory studies suggest that the rheology of the lithosphere is considerably more complicated than a simple elastic plate model, it is worthwhile investigating its applicability to the intraplate deformation.

If a force \( P \) is applied to a pinned elastic plate of thickness \( h \) (Figure 6.1), it is possible to use two-dimensional elastic plate bending theory (given in Appendix C) to show that the critical stress for buckling is given by

\[
\sigma_c = \left[ \frac{Eh}{3(1 - \nu^2)} \right]^{0.5} \left( \frac{P_m - P_w}{g} \right)
\]

Explanation of symbols and values of constants are given in Table 6.1.

Using an elastic thickness of 38 km, corresponding to a 70 Ma lithosphere at the onset of compressive deformation in the Central Indian Ocean Basin (Bodine et al., 1981), the lithosphere buckles with a wavelength of 385 km with an average compressive stress of 4.5 GPa. This wavelength is twice the wavelength seen in the deformation area and the stress is unreasonably large. The lithosphere would fail by faulting before buckling could take place. For buckling of wavelength 200 km an elastic thickness of 16 km is required with a buckling stress of 2.9 GPa which is still very large. More geologically reasonable stresses of < 1 GPa require an elastic thickness of < 1.9 km and produce a buckling wavelength of < 40 km. Clearly, two-dimensional elastic plate bending theory is inadequate, non-elastic processes must contribute to the formation of the long wavelength undulations.
Figure 6.1 Deflection of an elastic pinned beam. See main text for discussion and Appendix C for two-dimensional elastic plate bending theory.
The realisation that the elastic model is inappropriate casts doubt on the modelling by Karner and Weissel (in press) in which they contend that critical wavelength components in the deflection, caused by the emplacement of the Afanasy Nikitin seamount in late Cretaceous or Early Tertiary time, were preferentially amplified when north-south directed compression occurred around 7 Ma. They develop one and two-dimensional models for the compression of a thin elastic plate containing an initial deflection and find a best fit for an applied horizontal compressive stress of 1. -2 GPa (which is still very large) and an effective elastic thickness of 10 - 15 km. They suggest that this thin elastic thickness (compared to plate cooling model estimates of 38 km) could be due to plastic yielding at the top and bottom of the lithosphere.

Although elastic models are inappropriate, they can be used to investigate the origin of the small residual depth anomaly of ~250 m associated with the intraplate deformation (Section 3.3.2). To the north of the intraplate area is the enormous load on the Indian plate caused by the Bengal Fan. By extending simple elastic bending theory, it is possible to test if the residual depth anomaly in the Central Indian Ocean Basin is a result of flexure due to this Fan. Turcotte and Schubert (1982) showed that if

\[ \lambda >> 2\pi(D)^{0.25} \rho \varepsilon \delta \]

where \( \lambda \) is the wavelength of the load and other terms are defined in Table 6.1, then local isostatic compensation is dominant. For a reasonable range of elastic thicknesses for the Indo-Australian plate the wavelength for effective isostatic compensation is certainly less than 20% of the wavelength of the Bengal Fan. Thus the Bengal Fan is unlikely to have produced the residual depth anomaly and the anomaly is more likely to be due to variations at the original spreading-centre (Section 3.3.2).
Associated with the long wavelength undulations are undulations of the geoid. In order to ascertain if undulation wavelength varied across the region of the deformation, Zuber (1987) measured the spacings of maxima and minima of deflections of the vertical, calculated from SEASAT-derived geoid anomalies which could be correlated with undulations in basement. Figure 6.2 shows histograms of the spacing for all latitudes, and also for latitude bins. Overall, Zuber found a range of wavelengths between 100 and 300 km with a mean of 186 km. From Figure 6.2 it is clear that the wavelength of deformation increases towards the north, with an increase in the mean wavelength of 53.8 km.

This increase in wavelength towards the north is interpreted by Zuber (1987) as being due to the presence of sedimentary loading in the north but not in the south at the onset of deformation. Seismic data tied to results from ODP Leg 116 (Shipboard Scientific Party, 1989) suggest that sediments covered much of the Central Indian Ocean Basin prior to deformation so Zuber's hypothesis is unlikely to be perfectly correct. However, the presence of more sediments towards the north may partly explain the observed increase in wavelength. An additional component can be derived by thinking in terms of an elastic plate model. Flexural studies have shown (Watts, 1978) that there is a relationship between age and elastic thickness, such that elastic thickness increases approximately proportionately with the square root of lithospheric age. Clearly, from elastic bending theory, thicker elastic plates bend with longer wavelength. In the Central Indian Ocean Basin, for an increase in lithosphere age of 25 Ma from south to north at the onset of compression, the elastic thickness increases by ~8 km in the same direction (using the equations of Bodine et al., 1981) which corresponds to an increase in buckling wavelength of ~50 km (assuming 3 is valid). This is similar to the observed increase in wavelength (Figure 6.2) and therefore, the increase in geoid wavelength, and hence deformation wavelength towards the north, may be due to increases in elastic thickness with increasing lithospheric age, as well as increases in sedimentary thickness.
Figure 6.2 Spacing of Seasat derived geoid anomalies, associated with basement deformation between 75°E and 85°E, broken into three latitude bins. After Zuber (1987). Note that the mean spacing increases towards the north.
6.2.3 Layered Rheological Models and Intraplate Deformation

From these studies it is clear that a purely elastic rheological model for the oceanic lithosphere is too simplistic: the upper part of the lithosphere behaves in a brittle manner and the lower part behaves ductilely. Only the central core of the lithosphere can be considered to be behaving elastically. Therefore, it is not surprising that layered rheological models (McAdoo et al., 1985; Bodine et al., 1981) achieve a better fit for the bathymetry and gravity of outer-rise trench systems for instance. In addition these models predict lithospheric thicknesses close to that associated with the maximum depths of oceanic intraplate seismicity (Wiens and Stein, 1983). When compared to plate cooling models (Parsons and Sclater, 1977) it appears that the base of the mechanically-defined lithosphere corresponds to an isotherm of between 700 and 800°C.

Layered intraplate models have been applied to the intraplate deformation in the Central Indian Ocean Basin. An elastic-plastic model, using the experimental rock mechanics results of Goetze and Evans (1979), was developed by McAdoo and Sandwell (1985). In this model, yield stress is a function of depth (Figure 6.3). The lithosphere is divided into three rheological regions: a brittle layer, with failure stress increasing with depth; a lower ductile region, in which the yield stress decreases exponentially with increasing pressure and temperature, and an elastic layer between these two layers. Using the yield stress envelope (Figure 6.3) McAdoo and Sandwell (1985) showed that with thinning of the elastic core due to horizontal compression by yielding at the top and bottom of the lithosphere, the lithosphere buckles prior to whole lithosphere failure. Their model predicts a wavelength of the right order, 160 - 240 km, with reasonable average compressive stresses of 600 MPa (which are of the same order as those predicted by Cloetingh and Wortel, 1986). Another result of their modelling was that sedimentary loading produced longer wavelength folding (Figure 6.4). The model of McAdoo and Sandwell (1985) does contain one simplification - they assume that for buckling all the compressive stress is constrained in the elastic layer. In reality some of the stress will be transmitted through the ductile layer. This said, the elastic-plastic model shows that the long wavelength undulations could be the result of the buckling of the elastic core under reasonable compressive stresses.

Alternative rheological models were developed by Zuber (1987), who treated the lithosphere as a viscous or plastic layer of uniform strength, above a viscous half-space in which strength decreases exponentially with depth. She used models in which a dominant wavelength developed in response to flexural buckling and by the
Figure 6.3 Idealised yield strength versus depth for elastic-plastic lithosphere of age 55 Ma. Typical bending stress profile is shown (arrows) for state of net axial compression. Redrawn from McAdoo and Sandwell (1985).

Figure 6.4 Buckling wavelength versus age for elastic-plastic lithosphere with (dotted curve) and without (solid curve) sediment loading and for fully elastic lithosphere. Redrawn from McAdoo and Sandwell (1985).
hydrodynamical growth of instabilities. For the viscous flexural model uniform folding or buckling occurred. However, for the hydrodynamical approach the deformation style depended on the assumed strength of the lithosphere. For a strong lithosphere flexural folding occurred, while for a weaker lithosphere inverse boudinage was the mode of deformation, with thickening below the highs. If a plastic rheology was assumed the deformation mode was by inverse boudinage.

The two possible modes for the deformation predicted by Zuber (1987) are shown in Figure 6.5. Zuber (1987) concluded that both modes are consistent with the geophysical datasets. Another conclusion was that the presence of sediments reduced, by a factor of three, the lithosphere compressive strength. A range of model results suggested a compressive stress in the Central Indian Ocean Basin of several hundred MPa. This is in broad agreement with the McAdoo and Sandwell (1985) figure of 600 MPa and with plate boundary modelling estimates (Cloetingh and Wortel, 1986).

6.3 Analogue Modelling

6.3.1 Introduction

This next section describes the use of the Sandbox technique to model oceanic lithosphere rheology under compression, and determine the mode of the deformation. An analogue rheological model of a brittle layer overlying a viscous layer is used, scaled to the natural system. The modelling was carried out at the Institut de Geologie, Rennes, France.

6.3.2 Scaling of the Natural and Analogue Systems

The theory of scaling has been applied to the Earth Sciences by Hubbert (1937) and Ramberg (1981): to scale an analogue model all the dimensionless values should be similar in nature and in the experiment. Davy (1986) and Davy and Cobbold (1990) describe in detail the scaling of the analogue modelling performed in Rennes.

In the models the lithosphere is made of dry sand and silicone putty. The sand is pure quartz Fontainebleau sand. It models the localising brittle rocks of the upper part of the oceanic lithosphere. Sand is a brittle material with a negligible cohesion, and an internal angle of friction of 30°. This sand is thus an excellent analogue of rocks that obey the Byerlee relation. Silicone putty (Gomme 7007, manufactured by Rhone-Poulenc, France) is a Newtonian viscous material and models the viscous lower
Figure 6.5 Figure illustrating the two hypotheses for the mode of formation of the long wavelength undulations: buckling and inverse boudinage.
lithosphere. Underlying these two layers we used either pure Acacia honey or heavy (dense) water, to represent the viscous asthenosphere as a weak fluid and to give isostatic support. The scaling of the natural and analogue systems is shown schematically in Figure 6.6.

For failure at 30° the fracture strength of the brittle layer can be written as

\[
\sigma_B = (\sigma_{xx} - \sigma_{zz}) \equiv 2.0 \rho g z
\]

such that failure occurs when the right-hand side is exceeded (Byerlee, 1978). See Table 6.1 for explanation of symbols. From Bassi and Bonin (1988) the effective viscosity can be defined by

\[
\sigma_{xx} - \sigma_{zz} = 4\mu \varepsilon_{xx}
\]

For the ductile layer, assuming olivine rheology, the strain rate for yielding (Bassi and Bonin, 1988) can be expressed as

\[
\varepsilon_{xx} = P(\sigma_{xx} - \sigma_{zz})^3 \exp[-Q/RT]
\]

Where \( P = 7 \times 10^4 \text{ MPa}^{-3}\text{s}^{-1} \) and \( Q = 520 \text{ KJ mol}^{-1} \)

6.3.2.1 Depth to Brittle/Ductile Transition

In the Central Indian Ocean Basin, which contains lithosphere of 65 - 80 Ma, a heat flux of 60 mWm\(^{-2}\) (± 5) is the theoretical value expected. Stein and Weissel (1990) concluded that, despite the presence of localised heat flow anomalies, on the basis of the absence of a bathymetric anomaly (although as discussed in Section 3.3.2 there is a small residual depth anomaly of ~250 m) and the presence of deep seismicity, that lithospheric temperatures in the Central Indian Ocean Basin are not significantly different from those expected for its age. With a coefficient of thermal conductivity of 3 (± 0.5) \text{ Wm}^{-1}\text{K}^{-1} \) and Fourier's Law, it is simple to deduce a linear temperature
Oceanic Lithosphere Analogue Modelling

Figure 6.6 Comparison of schematic Yield Strength Envelopes for typical oceanic lithosphere and that used in the analogue experiments. $h_1$ is the depth of the brittle/ductile transition. Thickness $h_1 - h_2$ is the thickness of the ductile silicone layer found by equating areas A and B. This is discussed in the main text.
gradient through the lithosphere (8). The depth to the brittle/ductile transition was found by equating the viscosity of the brittle layer to that in the ductile layer (9) and by substituting in the temperature gradient and solving iteratively.

\[ T = T_0 + 0.02z + 273.1 \text{ K} \]  

\[ \mu_B = \mu_D \]  

The depth to the brittle/ductile transition was found to be 30 (± 5) km.

### 6.3.2.2 Ductile Layer Thickness

In the natural system the ductile part of the lithosphere is composed of an infinite number of layers each with a different viscosity, the viscosity decreasing with depth. However, it is only practicable to model the lithosphere using a single ductile layer of finite thickness and constant viscosity. The approximation made, was to integrate with respect to depth over the ductile part of the yield strength envelope (area A in Figure 6.6) to obtain the viscosity-thickness product ('strength') (10). This product was then equated to the product of the thickness of a single layer \( t_s \) and a constant viscosity \( \mu_s \) (area B in Figure 6.6). In this way the choice of thickness \( t_s \) determined the viscosity \( \mu_s \) and vice versa.

\[
\int \mu(z)dz = \int \frac{[(\varepsilon/P)\exp(Q/RT(z))]^{0.33}}{4\varepsilon} \, dz
\]

\[ = t_s \mu_s = 1.94 \times 10^{28} \text{ Pa m}^{-1} \]

The model was appropriately scaled for time, length, density and viscosity. An example of the scaling between the oceanic lithosphere and an analogue model is given in Table 6.2 and Table 6.1 defines all parameters and functions used.
### Table 6.2

Scaling of Natural and Analogue Systems

<table>
<thead>
<tr>
<th></th>
<th>L (m)</th>
<th>t (s)</th>
<th>( \mu ) (Pa s)</th>
<th>( \rho ) (kg m(^{-3}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Natural System</td>
<td>( 3 \times 10^4 )</td>
<td>( 10^{16} ) (1)</td>
<td>( 1.08 \times 10^{24} ) (4)</td>
<td>( 3.3 \times 10^3 ) (2)</td>
</tr>
<tr>
<td>Analogue System</td>
<td>( 2 \times 10^{-2} ) (3)</td>
<td>( 1.08 \times 10^4 ) (5)</td>
<td>( 3.3 \times 10^4 )</td>
<td>( 1.4 \times 10^3 )</td>
</tr>
</tbody>
</table>

1. \( t_N = 1/e \)
2. \( \rho_N \) and \( \rho_M \), here of the asthenosphere and heavy water respectively
3. Brittle Layer thickness in model
4. Assume ductile thickness is 18 km. \( \mu_N \) from equation 10
5. By equating analogue and natural viscosities -

\[
\frac{t_m}{t_n} = \frac{\mu_m \rho_n I_n}{\mu_n \rho_m I_m}
\]

6.3.3 Experimental Design and Methodology

The three layer model was placed in a sandbox with piston and motor as shown in Figure 6.7. Above the model was a laser ranging device which could be used to measure the amplitude of the topography developed in the model. Additional equipment used in running the experiments is shown in Figure 6.8. These were the first experiments to be run in Rennes using a laser, and controlling software was developed during the course of the experiments. For this reason no laser profiles are available for the first experiment and are poor for the next two. A camera was also placed above the model to photograph changes in a grid of white plastic powder (ethyl cellulose) set up on top of each model.

Before the piston was started the initial topography was measured several times using the laser so that the final profiles could be corrected for any pre-existing topography. A constant compressive force could then be applied through the model by the piston. Initially identical models were run using different strain rates; strain rate was varied by using different motor speeds. The topography developed was the same in all experiments and hence we had verified that the deformation was independent of strain rate.
Figure 6.7 The experimental apparatus. The three layer model (sand, silicone, honey/heavy water) is placed in a plastic box, towards one end of which is a piston that is driven, at a constant rate, by a motor. Above the model is a laser ranging device and camera (not shown) which record the topography developed throughout the experiment. A regular grid of white dots (ethyl cellulose) was placed on the top of the model to aid photographic analysis.

Figure 6.8 Additional experimental equipment. The micro on the left runs the software which controls the motor driving the laser ranging device. The collection of points by the laser is controlled by the micro on the right, while the micro on the far right controls the motor that drives the piston. In the centre of the photo is the camera stand (the model is hidden by the laser ranging device).
Clearly only realistic amounts of shortening should be allowed in the model. In the Central Indian Ocean Basin there is some controversy as to the amount of shortening that has taken place during the deformation. Gordon et al. (1990), from inversion of present-day spreading rates and directions from plate boundaries in the Indian Ocean, suggest a shortening rate of 1 - 7 mm yr\(^{-1}\) in the Central Indian Ocean Basin with the rate increasing eastwards away from the predicted pole of rotation. However, estimates made directly from the properties of the deformation suggest shortening rates of 3.8 ± 1.2 mm yr\(^{-1}\) (Chapter 5) or 1 mm yr\(^{-1}\) (Weissel and Geller, 1981). These estimates correspond to a range of compressive strain rates between 10\(^{-1}\) and 10\(^{-2}\) for the approximate 100 km of N-S shortening that has occurred in the deformation area. Therefore in our models we focussed on the topography that developed for only a few percent horizontal shortening.

Eight successful experiments were run in total, using a variety of brittle layer thicknesses and box widths. Parameters used for each model are listed in Appendix D. The topography developed was measured by the laser at regular periods during compression for experiments 2 to 8. Following the experiment the initial topography was removed from all the measured profiles. Spectral analysis was then undertaken on these corrected profiles to determine the principal wavelengths present. The wavelength was estimated, purely from photographs for experiment 1, and for experiments 2 and 3, from both photographs and spectral analyses. Laser profiles were sufficiently good for experiments 4 to 8 that wavelengths were taken from the spectral analyses, although the corresponding photographs were checked for corroboration.

In experiments where heavy water was used for the asthenosphere, the model was frozen after completion. When completely frozen these experiments were cut into cross-sections to determine the mode of deformation in the model.

Finally, a ninth experiment comprising a four layer continental rheology was run. A resultant profile and spectral analysis are shown in Appendix D.

6.3.4 Results

In all the experiments clear undulations could be observed in the top of the brittle layer (Figures 6.9 and 6.10) before the appearance of the first reverse faults. Cuts of the experiments (Figure 6.11A+B), after the end of compression and for reasonable
Figure 6.9 Undulations in the top of the brittle layer in experiment 1. The 4 cm grid spacing corresponds to 60 km in the natural system. A range of wavelengths are present, with an average value of around 200 km.

Figure 6.10 Undulations in the top of the brittle layer in experiment 2. The 4 cm grid spacing corresponds to 60 km in the natural system. A range of undulation wavelengths are present with an average value of around 200 km. Note that the position of the light source at the right hand-side masks a wavelength visually documented. In general it was far easier to see wavelengths present in the laser topography than in the photographs.
Figure 6.11A Photograph showing cross-sections produced by cutting a frozen model in half using a circular saw (in the background). Clear uniform folding (buckling) of the brittle layer (grey layer) is present. A close up of one half of the model is shown in Figure 6.11B. Photograph courtesy of J. Martinod.

Figure 6.11B Close up of cross-section showing folding of the brittle layer (grey sand). The black layer overlying the grey sand (brittle upper lithosphere) is additional sand which was carefully added to maintain the topography after the cessation of compression and prior to freezing. The pink layer is made of viscous silicone while the violet layer is frozen heavy (dense) water. The lines within the purple layer are due to the circular saw. Note that there is no topography at the boundary between the pink silicone layer (viscous lower lithosphere) and the purple heavy water (asthenosphere). In this model the brittle (sand) layer corresponds to 30 km in the natural system. The buckling wavelength is hence 220 km, and is ~7 times the thickness of the brittle layer. Photograph courtesy of J. Martinod.
amounts of shortening, show that these undulations are the result of buckling of the whole of the brittle layer. The ductile part of the lithosphere only accommodates the buckling in the upper domain, and the lithosphere/asthenosphere (silicone/heavy water) interface does not show any vertical deformation.

The appearance and evolution of the undulations have been recorded using the laser (Figure 6.12). The topography of the models has only been registered on cross-sections parallel to the direction of compression, and situated in the middle of the box. These cross-sections are representative of the deformation of the whole model because the undulations were continuous along strike.

Spectral analysis of the successive topographies have been performed (Figures 6.13 and 6.14 and Appendix D). They all show that, at least at the beginning of the appearance of vertical movements, only one wavelength is present. It is the wavelength of lithosphere buckling.

The results of the experiments and spectral analysis are given in Table 6.3. Listed for each experiment is the brittle thickness (analogue and natural equivalent), buckling wavelength developed (analogue and natural equivalent), and the ratio of buckling wavelength to brittle thickness. There is always a natural equivalent buckling wavelength of about 200 (± 40) km - the average wavelength that has been observed in the Indian Ocean for buckling. The ratio of buckling wavelength to brittle layer thickness (Table 6.3) ranges between 6.4 and 8.0 with an average value of 6.9.

<table>
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<tr>
<th>Experiment Number</th>
<th>hLm (cm)</th>
<th>hLb (km)</th>
<th>λn (km)</th>
<th>Ratio λn/hLm</th>
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<td>30</td>
<td>200</td>
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</tr>
<tr>
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<td>7.0</td>
</tr>
<tr>
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<td>30</td>
<td>190</td>
<td>6.3</td>
</tr>
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</tr>
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<td>8</td>
<td>1.0</td>
<td>30</td>
<td>240</td>
<td>8.0</td>
</tr>
</tbody>
</table>
Figure 6.12 Growth of folds in experiment 8. All profiles have been corrected for any initial topography and rescaled. Note how the wavelength of folding is unchanged at ~240 km. See also Figure 6.13.
Figure 6.13 Corrected topography and spectral analysis of experiment 8 for 2.7% shortening (profile F). A single dominant wavelength of folding is developed equal to 240 km. Five wavelengths of folding are present.
Figure 6.14 Corrected topography and spectral analysis of experiment 7 for 2.8 % percent shortening (profile E). A single dominant wavelength of folding is developed equal to 220 km. Two to three wavelengths of folding are present in this experiments which was undertaken in a smaller box than experiment 8 (see Figure 6.13).
Observation of amplification of the undulations (Figure 6.12), leads to the conclusion that the positions of crests and troughs remains the same and therefore the wavelength constant. Through analysis of spectral power of the main wavelength as a function of time (Figure 6.15), it is possible to note that the amplification of the magnitude of the folds is exponential during the first stages of deformation. That is, for a few percent shortening, Figure 6.15 is a straight line. This can be compared to the theory developed by Biot (1961), Fletcher and Hallet (1983) and Ricard and Friodevaux (1986) for visco-plastic materials. This theory predicts that the small perturbations develop exponentially with respect to time and can be expressed as

\[ A(t) = A_0 \exp(q \varepsilon \cdot t) \]  

where \( A \) is the amplitude of the folds \( A_0 \) is the initial amplitude, \( \varepsilon \cdot \) is the homogeneous horizontal strain rate and \( q \) is the dimensionless growth rate factor of the deformation. Therefore, the straight line in Figure 6.15 has a gradient equal to the product of the growth rate factor and the strain rate (\( q \varepsilon \cdot \)). Although the parameters cannot be measured with much precision, for strain rates in the models of \( \sim 5 \times 10^{-5} \), values of \( q \) vary from experiment to experiment between 40 and 300 for the first few percent shortening. An average growth rate factor is about 100.

With knowledge of growth rate in the experiments it is interesting to compare this to the natural system. If the initial perturbations were of the order of the amplitude of the pre-deformational abyssal hill topography, then an estimate of 100 - 200 m is reasonable for the initial perturbation (Chapter 3, Chapter 4). Amplitudes of folding observed in the Indian Ocean today (Chapter 4) range between 1 - 2 km. Using equation 15 this order of magnitude increase in amplitude gives

\[ \frac{A}{A_0} = \ln(10) = q \varepsilon \cdot t \]  

which for a growth rate factor of 100 and a strain rate of \( 10^{-16} \text{ s}^{-1} \) leads to an estimate of the time since the beginning of amplification of \( \sim 7 \text{ Ma} \). This result is discussed in the next section. Beyond a few percent shortening the amplification of folds is no longer exponential and decreases rapidly. This is likely to be because further amplitude growth under the influence of gravity is energetically unfavourable. Instead it is easier for the whole lithosphere to fail as illustrated by the development of reverse faulting. These faults generally develop at the inflexion points of the undulations.
Figure 6.15 Spectral power versus amount of shortening and time since the onset of deformation for the dominant wavelength present (240 km see Figures 6.12 and 6.13) in experiment 8. For small amounts of shortening (<4.7 %) this log-normal graph is a straightline suggesting exponential growth of the principal wavelength. Using visco-plastic theory, for a given strain rate in the model, a growth rate factor $q$ can be estimated. See main text for discussion.
6.3.5 Discussion

One of the most characteristic features of the intraplate deformation are the ~E-W trending geoid anomalies. Zuber (1987) analysed these anomalies and found that the mean of the principle wavelength present ranged between 168.5 km and 222.3 km, increasing to the north, as would be expected with increasing lithospheric age. The equivalent wavelengths obtained from the models (Table 6.2) are very similar, between 190 km and 240 km. This agreement suggests that we have adequately modelled the oceanic lithosphere rheology with the simple two layer model.

The observation of buckling in the analogue experiments gives strong support to the hypothesis of buckling for the mode of deformation of the oceanic lithosphere in the Central Indian Ocean Basin. It should be noted that we observe buckling of the whole of the brittle layer. Because this layer is 30 km thick it implies that not only the crust, but also a substantial amount of upper mantle is uniformly folded.

Although the analogue experiments successfully model the long wavelength features, they are less satisfactory in modelling the shorter wavelength features represented by reverse faulting. In the experiments, buckling precedes faulting and when faulting occurs it appears preferentially at the inflexion points of the undulations. The appearance of faults at inflexion points is not surprising, because at these positions the bending stress is a maximum. By contrast, in the natural system faults do not preferentially occur at the inflexion points (Chapter 4), and motion along the faults appears to have been steady since the onset of deformation (Shipboard Scientific Party, 1989). This discrepancy suggests that the modelling procedure was inadequate for the shorter wavelength features. A possible reason for the latter discrepancy is that faulting was present from the onset of deformation, but below the level of detection on the laser profiles. Also, pre-existing fault weaknesses were present throughout the natural system that may have allowed their early reactivation.

Simplifications in the modelling procedure included the absence of sedimentation. If sedimentation had been included, it is likely that it would have speeded up the growth of the undulations (see Section 6.2.3). Another simplification is the absence of a weak negative density gradient at the base of the lithosphere. In the models, for obvious practical reasons, it was necessary to build a lithosphere very weakly less dense than the asthenosphere. This positive density contrast prevents the lithosphere sinking into
the asthenosphere as might be observed after the appearance of faults that break the entire brittle domain. Thus subduction zones cannot occur in the models, as might be expected for large amounts of shortening and extreme buckling.

Using a growth rate factor of 100; assuming initial perturbations of 100 - 200 m and exponential growth of buckling in the natural system, it was shown in the last section that the time for amplification of pre-existing perturbations to the present 1 - 2 km folds would be ~7 Ma. With consideration of all the errors and assumptions involved, the observations of Leg 116 (Shipboard Scientific Party, 1989) of 7 Ma since the onset of deformation is probably coincidental. In fact, with the inclusion of sedimentation it would be expected that this time would be significantly reduced. The only weight the author would like to see put on this analysis of growth rate, is that this agreement further supports the supposition that the models are reasonable analogues of the natural system for a few percent shortening.

It is difficult to know what the significance is of the relationship between the brittle thickness and the wavelength of buckling. In the experiments, the buckling wavelength is ~7x the brittle thickness. At present there are no numerical models available to explain this relationship.

Substantive discussion of the continental example is beyond the scope of this thesis. However, the resultant topography was far more complex than that produced by the oceanic models. The possibility of continental buckling is discussed in Chapter 8.5.2

6.3.6 Summary of Analogue Modelling

The principal conclusions of the analogue modelling are:

1. That buckling is likely to be the mode of deformation in the Central Indian Ocean Basin, with the uniform folding of the whole of the brittle layer (~30 km

2. An oceanic lithosphere model, based on experimentally determined yield stress envelopes, in which a brittle layer overlies a ductile layer, is valid.

3. There is an association such that the buckling wavelength is typically 7 times the brittle layer thickness. The reasons for this association are poorly understood and are the subject of proceeding work.
6.4 Numerical Modelling using the Finite Element Technique

In the Earth Sciences, the two most commonly used numerical modelling techniques are Finite Difference and Finite Element modelling. For intraplate deformation, in an attempt to determine the mode of the deformation, the fixed node finite element approach was used.

Finite Elements is a technique which has only recently become widespread in science and engineering, because it relies heavily on large amounts of computer capacity that has only been available since the 1960's. In the last 30 years it has been applied to a number of lithospheric scale problems including subduction zones (Bott et al., 1990) and grabens (Mithen, 1980). However, it has not previously been applied to the intraplate deformation to determine the mode of deformation.

In general, the finite element method can be used to solve differential equations over the domain of a complex shape. The domain is represented by a large number of finite elements of similar shape. These finite elements are described by nodal points, the larger the number of nodes per element the more complicated the element. However, it should be noted that the computer time to solve the problem is the cube of the increase in the number of nodes. The computer time is taken up by solving by matrix methods the simultaneous equations that result at all the interelement boundaries.

Because of the great complexity of the finite element technique it was not practicable to write a program specifically designed for modelling the oceanic lithosphere in compression as part of this project. Instead a pre-existing program designed for large-scale lithospheric problems was used. This program was run under the direction of Professor Martin Bott at Durham and is the result of 15 years work and a host of PhDs. The program has been used previously to model grabens (Mithen, 1980), subduction zones (Bott et al., 1990), and presently spreading-centres and the effects of plumes (Bott in prep.).

The program can handle various surface loads as well as horizontal tension and compression. Its major constraint is that it can only use two types of rheology: elastic and visco-elastic. Whereas for comparison with McAdoo and Sandwell (1985) the oceanic lithosphere starting model would have comprised an elastic layer overlying a plastic one.
As a starting model (Figure 6.16A) a lithosphere of 70 km thickness was used, comprising an upper 10 km thick elastic layer overlying a 60 km thick visco-elastic layer. Using this model over the time of loading, viscous relaxation within the lower lithosphere would be expected to lead to reduction in lower lithosphere stresses and the amplification of stresses within the upper lithosphere (Kuznir, 1982; Bott and Kuznir, 1984). The oceanic crust was not modelled separately as it was assumed that the strength contrast between the crust and mantle would be relatively small. Horizontal compression could then be applied to one end and the other end was constrained in the horizontal direction but not the vertical. An isostatic restoring force was applied to the upper surface of the model, dependent on the deflection from the horizontal reference axis such that

\[ R = (\rho_m - \rho_s)gdx = 1000gdx \]  - 13

To perturb the model initially and to stimulate deformation a relatively small (compared to the horizontal compression) sinusoidal oscillation was applied to the upper surface.

\[ S = A_0 \sin \frac{2\pi x}{\lambda} \]  - 14

Various models were run with different applied forces, elastic thicknesses, viscosities, boundary and initial conditions. The resulting profiles and parameters used from these models are shown in Appendix E. Unfortunately none of the model geometries or conditions produced flexure of the upper surface of the model similar to that observed in nature.

Following this the model was simplified and flexural theory applied to check if the model could handle buckling. A 5 km thick elastic layer was subjected to horizontal compression with the restoring forces and sinusoidal oscillation as above. This model is shown in Figure 6.16B. Values predicted from simple elastic bending theory (Appendix C) were applied. Buckling was not observed (Figures, Appendix E). All parameters in the model were varied to try and produce buckling. Unfortunately the model could not be persuaded to buckle.
Figure 6.16 Finite Element meshes used to try and model the long wavelength undulations with (A) elastic/visco-elastic rheology and (B) purely elastic rheology.
This disappointing result could be because of possible problems with the program. These could include the possibility that the quadratic formulation of each individual element results in the program being incapable of handling whole model buckling or that the condition at critical buckling leads to a singular matrix (M.H.P. Bott, pers. com.). Due to time limitations it was not possible to follow this modelling technique further. However, in the future, the author intends to continue modelling the deformation using this technique and using plastic rheologies.

6.5 Summary

1. The most important result of this Chapter is the strong support that analogue modelling gives for the buckling of the whole of the brittle lithosphere. Furthermore growth rates observed in the analogue system are closely similar to those in the natural system.

2. Numerical modelling has to date been unsuccessful in modelling the long wavelength undulations.
Chapter 7 - Geodesy and Gravity Modelling

7.1 Introduction

One of the principal conclusions of Chapter 5 was that the observed tectonic deformation could be explained in terms of the brittle behaviour of the upper part of the oceanic lithosphere being driven by an underlying flexural mechanism. With the present dataset a purely observational approach (Chapter 4 and 5) is inadequate for determining the exact mechanism - modelling is required. An investigation of crustal thickness variations with spatial respect to the long wavelength undulations may give an indication of the mechanism. If the crust is of uniform thickness, and the Moho parallels the top of oceanic basement, this would favour the buckling hypothesis for the driving mechanism. Alternatively, if thickening of the crust under crests occurs, this would support the hypothesis of inverse boudinage proposed by Zuber (1987). The principal aim of this Chapter is to resolve the geometry of the oceanic Moho by gravity modelling and hence determine the underlying mode of deformation.

Shipboard Free-Air Anomaly (F.A.A.) gravity data collected during CHARLES DARWIN Cruise 28 (Section 2.1) is combined with F.A.A. gravity data from other sources (Section 2.3). This dataset is used in this Chapter not only for two and three-dimensional gravity modelling, but also for qualitative discussion of the tectonic fabric.

Geodetic data can be used to derive the F.A.A. gravity field. The first part of this Chapter deals with geodetic data, its uses and the derivation of the F.A.A. gravity field. Following this derived gravity anomalies are compared with the underlying topography.

7.2 Geodetic Data

7.2.1 Definition of the Geoid

Polar flattening, a consequence of the rotation of the earth, means that the surface of the earth is not a sphere but an ellipsoid of revolution. As stated by Bott (1982)
"the theory of the shape of the earth can be divided into two parts: (i) determining the shape and dimensions of the ellipsoid which gives the best fit to the sea-level surface (this ellipsoid is called the spheroid); and (ii) determining deviations of the sea-level surface (which is called the geoid) from the spheroid."

One of the principal aims of the NASA satellite SEASAT was to document the second part of this theory. During three months of 1978 SEASAT collected information on the height of the sea surface as well as other parameters. SEASAT used a radar altimeter to measure the altitude above the sea surface. After various corrections (Gahagan et al., 1988) this distance is subtracted from the distance between the satellite and the reference ellipsoid. This difference is the height of the sea surface above the reference ellipsoid. Geoid maps can then be made by fitting a minimum curvature surface to the sea surface height measurements.

7.2.2 The Relationship Between the Geoid and Bathymetric Features

The shape of the geoid (and therefore the sea surface which closely approximates the geoid) are caused by lateral density variations within the earth (Gahagan et al., 1988). The geoid can be interpreted broadly in terms of long wavelength (>1000 km) anomalies thought to be caused by mantle convection and relatively shorter wavelength (<200 km) anomalies caused by density variations close to the surface of the lithosphere. These density variations can be positive, as for an ocean ridge or seamount resulting in a geoid high, or negative for a trench or deep fracture zone valley causing a geoid low. (Gahagan et al., 1988; NASA report, 1986). It is clear that the shorter wavelength contributions to the geoid are highly correlated with seafloor topography (Haxby et al., 1983). This has led to identification and location of bathymetric features from the geoid anomalies (Gahagan et al., 1988).

In the Central Indian Ocean Basin, as already discussed in Chapter 1 (Figure 1.5) undulations trending ~E-W with wavelength of 100 - 300 km are clearly visible in geoid maps. For looking at relatively long wavelength features, and for obtaining truer information about the deep structure of the lithosphere geoid maps are ideal. However, for an investigation of the structural style of deformation, the geoid maps have the disadvantage of emphasizing relatively longer wavelength features. Figure 7.1 illustrates the SEASAT-derived gravity field over the survey area which is more sensitive to short wavelengths. The SEASAT orbital coverage is shown in Appendix F. The north-eastern part of the survey area is well covered, while the south-western part is less well covered, with some points 100 km from the nearest SEASAT profile.
Figure 7.1 SEASAT-derived gravity anomalies using Strange (left) and Haxby (right) algorithms. The bright anomaly centred at 82.5°E, -2.0 - -5.0°S is the Afanasy Nikitin Seamount Group. Contours in mGals.
7.2.3 SEASAT-Derived F.A.A. gravity field

Various processing routes for Fourier transforming both the geoid and geoid gradient have been used to obtain the F.A.A. gravity field from the original geoid data (Haxby et al., 1983; Freedman and Parsons, 1986). The coverage in Figure 7.1 was obtained by the reduction of geoid data using Haxby and Strange algorithms by Parsons and Strange at Oxford University. The basic difference between the two algorithms is that the Strange emphasizes the high frequency (short wavelength) features better than the original Haxby using the least-squares collocation method of Rapp (1986). The resultant gravity maps are, however, very similar.

The bright feature in the upper tonal images of Figure 7.1 centred at 2.0 - 5.0°S 82.5 - 83.5°E is the Afanasy Nikitin Seamount Group. This seamount has gravity maxima of ~10 mGals and is surrounded by lows of -20 - -30 mGals. Because the Strange algorithm handles the high frequency data better the numerically contoured Strange image (Figure 7.1, bottom left) was inspected in detail, as shown in Figure 7.2.

Within the detailed study area (Figure 7.2) gravity values range between -70 mGals and -10 mGals. The prominent crest at -4.0°S 80.0°E closely mirrors the shape and extent of the basement high in Figure 4.5. There is a striking similarity between the SEASAT-derived gravity anomaly and basement topography. Although there is some seafloor topography, the major component to the gravity anomaly is likely to result from the sediment/basement interface with any Moho topography generating a lower amplitude broader anomaly. Generally the troughs in the gravity field are less well defined than the crests. This seems to be consistent with the seismic/statistical observation (Section 4.3) of sharp crests and broader troughs in the basement topography.

Gravity anomaly trends appear to be discontinuous across fracture zones developed at 79.0°E and 80.5°E. This agrees with the conclusions from mapping the basement undulations (Section 4.2). However, this conclusion cannot be made from the geoid data which tend to emphasize the longer wavelength features. It was by using the geoid data that Haxby and Weissel (1986) and Stein et al. (1989) missed this discontinuity across fracture zones.

The fracture zone orientations cannot be clearly discerned in the data. This is likely to be due to the nearly N-S orientation of these structures. Bathymetric features running in a N-S orientation are generally far harder to identify than E-W features because of
Figure 7.2 Detailed SEASAT-derived gravity anomaly using the Strange algorithm. The prominent high at -4.0°S, 80.0°E correlates with a basement high in the same position (Figure 4.5). Contours in mGals.
the ~SSE-NNW and NNE-SSW satellite orbits. However, after allowing for the effects of the long wavelength undulations the gravity anomalies become less negative westwards across the fracture zones, consistent with the decrease in crustal age, and shallowing of basement depth, in the same direction (Section 4.2).

7.3 Ship-Collected F.A.A. Gravity Field

In the study area the network of lines shown in Figure 7.3, when supplemented with the Lamont, French and Russian profiles detailed in Sections 2.2 and 2.3, gives a good spatial coverage of ship-collected F.A.A. with some 5200 km of data. The contour map of the F.A.A. (Figure 7.4) is a compilation of all the available data.

Values of the F.A.A. range between +5 and -94 mGals. Figure 7.4 closely mirrors the contour maps of oceanic basement and SEASAT-derived gravity with gravity highs (typically -30 mGals) over the crests of the undulations and gravity lows (typically -50 - -60 mGals) over the troughs with discontinuity of the highs and lows across the fracture zones. Deviations between satellite-derived (e.g. Figure 7.2) and ship-derived contour maps towards the northern end of the survey area could be due to poor ship data coverage. In general the ship-derived contour map delineates the pronounced crests in the southern part of the survey area as well as the satellite-derived contour maps, although there is a lack of detail in the ship-derived contour map compared to the satellite-derived map with its greater number of tracks. The suggestion of a ENE - WSW trend to the ship-derived contours is likely to be an artefact of the position of ship tracks.

7.4 Modelling Rationale

Using a simple approximation to the velocity-depth structure elucidated in Chapter 3, it was possible to depth convert the single-channel seismic reflection profiles described in Chapter 4. The depth conversion used a mean sediment velocity of 2.15 km s⁻¹ and a water velocity of 1.50 km s⁻¹ and gave the thicknesses of the sediment and water layers to be used for modelling. For gravity modelling it is necessary to know not only the thicknesses of the layered model, but also the constituent densities of the model. The mean sediment velocity was compared with published velocity-density curves (Nafe and Drake, 1963; Hamilton, 1978) assuming that the mixed turbiditic lithology described by Shipboard Scientific Party (1989) continued down to basement. A mean sediment density of 2200 ± 100 kg m⁻³ was deduced. As discussed in Chapter 3 evidence for pelagics is not convincing within the survey area.
Figure 7.3 Track chart showing the positions of ship-derived F.A.A. profiles. Bold lines, A - D, are discussed in the text (Section 7.3) and illustrated in Figures 7.5 to 7.8.
Figure 7.4 Ship-derived F.A.A. contour map using tracks shown in Figure 7.3. Contours in mGals.
and it is not possible to pick on the normal incidence seismic profiles a horizon which might represent the top of pelagic sedimentation. It was hoped that any errors in the approximation of the sediment column to one density, due to compaction, or the presence of a pelagic layer towards the base of the cover sequence, would be small. Hamilton (1978) showed that the velocity-density gradient for a turbiditic lithology is high. That is, the density changes little for a wide range of velocities (150 kg m$^{-3}$ for a 1.0 km s$^{-1}$ change in interval velocity). Densities for seawater, crust and mantle of 1030 kg m$^{-3}$, 2800 kg m$^{-3}$ and 3300 kg m$^{-3}$ were assumed respectively for the initial models (Bott, 1982). These densities give a density contrast across the Moho of 500 kg m$^{-3}$, across the sediment/basement interface of 600 kg m$^{-3}$ and 1170 kg m$^{-3}$ for the seafloor.

With the densities of all the layers constrained, and the thicknesses of the sediment and water layers determined, the only unknown is the thickness of the crust - the position of the Moho.

In the first instance a uniform thickness crust was assumed for two-dimensional modelling by adding a constant thickness to the basement depth. Crustal thickness and the density of the constituent model layers could then be varied within allowable degrees of freedom. Two-dimensional modelling was undertaken on a representative sample of gravity profiles within the survey area. The typical crustal structure deduced from this modelling was then used as a starting model for three-dimensional modelling.

For the survey area three-dimensional modelling as well as two-dimensional modelling is necessary because of the absence of two-dimensionality. If there is variation perpendicular to strike within two - three times the depth of the structure then two-dimensionality is violated (Nettleton, 1979). The discontinuity of undulations across fracture zones revealed in the contour map of oceanic basement means that an interpretation based solely on two-dimensional modelling would have been inadequate.

Leger (1989) found from seismic refraction that, over one of the crests in the survey area at -4.0°S 80.0°E the crust thickens by 30 % relative to neighbouring troughs. To test the effect of thickening under the crests, an average Moho depth for the troughs of the undulations for each profile was found. This average Moho depth was then used along the whole profile line. The effect of this flat Moho model was to produce ∼10 - 20 % thickening under the crests relative to the troughs. Alongside the observed
and best-fitting anomalies, the anomaly produced by a flat Moho is also illustrated. The effect of 30% thickening is also shown for one of the profiles (line B) across the crest investigated by Leger (1989). Generally, because of the clear inadequacy of this fit for 30% thickening, for graphical comparison with the best-fitting anomaly a flat Moho was used to emphasize the component of the anomaly due to Moho topography.

7.5. Two-Dimensional Modelling

In this section four gravity profiles are modelled using two-dimensional techniques. The positions of the four profiles are shown in Figure 7.3. A, B and C are "dip" profiles while D is oblique.

7.5.1. Line A

The longest profile available for gravity modelling is the CHARLES DARWIN N-S profile A. This line is some 500 km long and runs nearly perpendicular to the axes of the undulations (see Chapter 4). Knowledge of the lateral extent of the long wavelength undulations is poor, except at the southern end, making estimation of errors caused by end effects along strike impossible.

Along profile A (see Figure 7.5) the oceanic crust youngs southwards from 78 Ma around the O.D.P. Leg 116 sites to 74 Ma and, as predicted by plate cooling theories (Parsons and Sclater, 1977), basement depth decreases towards the southern end. Superimposed on this general decrease are the long wavelength undulations. Sediment thickness also decreases in a southerly direction as the distance from the sediment source, the Bengal Fan increases.

Profile A could be two-dimensionally modelled using a uniform, 5.0 km thick crust, except at the northernmost end. At this end a thicker crust up to 5.9 km thick was necessary to obtain a reasonable fit. A flat Moho, which in effect leads to the relative thickening of the crust below crests and thinning below troughs resulted in an anomaly of insufficient amplitude. No reasonable changes in densities could change this fact.

The apparent thickening towards the northern end could be due to variations in the crustal thicknesses formed at the original spreading-centre. White and McKenzie (1990) suggest that small variations in the potential temperature of the mantle can...
Figure 7.5 Two-dimensional gravity modelling for Line A (see Figure 7.3). Densities for water, sediment, crust and mantle are 1030, 2200, 2800 and 3300 kg m$^{-3}$ respectively. Stippled area indicates the sediment layer. Continuous line is the observed anomaly and the dashed line is the anomaly produced by the underlying model. This model comprises a uniform thickness crust of 5.0 km except at the northern end where thickening up to 5.9 km occurs. Dotted line is the anomaly associated with a flat Moho, while the dot-dashed line at the northern end shows the effect of a continuous thickness crust.
have a large effect on crustal thickness. However, it is also possible that the crustal thickening is due to tectonic deformation. In this area faults are imaged on multichannel profiles (Chapter 5) penetrating towards the base of the crust.

Although it is significant that for most of the profile a uniform thickness crust is present, with thickening at one end, the absolute crustal thicknesses determined from the modelling cannot be established. Because we know neither the density nor the thickness of the crust accurately there is considerable ambiguity between these two unknowns. That is, if a higher density had been used this would have resulted in a thinner crust. Although there is some evidence from residual depth anomalies (Chapter 3) which suggest a thinner crust than normal, the gravity method, with its inherent ambiguity cannot be used to support this suggestion.

7.5.2 Line B

This line, some 300 km long, runs from north to south within the centre fracture zone block, perpendicularly over the best developed undulation crest in the survey area. Because this line is sited in the centre of the survey area reasonable estimates can be made of the end effects along strike.

The oceanic crust decreases in age from ~73 Ma to 69 Ma towards the south. However the corresponding decrease in depth to basement is masked by the prominent crest, which has nearly a kilometre of relief along the profile. Sediments thin to the south and appreciably over the basement high.

Using the same model as for Line A, with a uniform crustal thickness of 5.0 km, an excellent fit (Figure 7.6) was obtained with the observed anomaly. A flat moho, which gave relative thickening under the basement high gave an anomaly of insufficient amplitude. The anomaly produced by 30 % thickening under the crest produces an even worse fit.

The recognition of a uniformly folded crust under this prominent high contradicts refraction work undertaken by Leger (1989). Leger found 30 % thickening of the crust under the same feature relative to the adjacent troughs. The gravity work is, however, in agreement with Neprochnov et al. (1988) who, using refraction data, concluded that the crust was of uniform thickness under the basement highs and lows.
Figure 7.6 Two-dimensional gravity modelling for Line B (see Figure 7.3). Densities as for Line A. Stippled area indicates the sediment layer. Continuous line is the observed anomaly and the dashed line is the anomaly produced by the underlying model which has a uniform 5.0 km thick crust. The dotted line shows the anomaly for a flat Moho while the dot-dashed line is the anomaly for ~30% thickening under the basement crest.
Using simple equations to calculate the end effect along strike caused by the absence of lateral continuity of basement structures perpendicular to the profile line (Nettleton, 1976), an estimate of the maximum error in the modelled anomaly of less than 10% was obtained. The closeness of the fit obtained (Figure 7.6) suggests that the error is appreciably less than this, or that some of it is taken up by the density contrast for the sediment/water boundary.

7.5.3 Line C

Towards the western edge of the survey area, this N-S profile (Figure 7.3), which is roughly 280 km long, lies on the western side of the Indrani fracture zone. Relief to the east of this line is adequately known and should not result in appreciable end effects along strike. However, relief to the west is poorly known, although mapping of the axes of the undulations (Figure 1.4) suggest continuity along strike.

Crustal ages range from 68 Ma at the northern end of the profile to ~65 Ma at the southern end. As would be expected (Parsons and Sclater, 1977) the depth to basement at the northern end is greater than that at the southern. However, complicating this picture, in the centre of the line, from ~70 to 240 km (Figure 7.7) is a basement high. On the basis of onlap patterns (Section 4.2) in the pre-deformational sediment package, this appears to have been a pre-existing high which has been reinforced by the subsequent deformation.

This line is best modelled using a uniform crustal thickness of 5.0 km for the southern (<70 km) and northern (>240 km) ends. In the middle of the profile a slightly variable crustal thickness of 5.6 - 5.9 km best fitted the observed anomaly. For this profile neither uniform crustal thickness nor a flat Moho adequately modelled the observations (Figure 7.7).

The recognition of crustal thickening under a basement high could be interpreted as evidence for inverse boudinage. However, the well developed pre-deformational sediment onlap relations on the flanks of this feature suggest that it was a pre-existing high. This thickening of the crust could be due to variations in the spreading-centre environment as suggested for the northern part of line A. These variations could also be associated with the emplacement of the Afanasy Nikitin Seamount Group. Simple
Figure 7.7 Two-dimensional gravity modelling for Line C (see Figure 7.3). Densities as for Line A. Stippled area indicates the sediment layer. Continuous line is the observed anomaly and the dashed line is the anomaly produced by the underlying model. This model comprises a uniform 5.0 km thick crust at the southern and northern ends and a crust of variable thickness (5.6 - 5.9 km) in the centre. The dotted line shows the effect of a flat Moho while the dot-dashed line is for a uniformly folded 5.0 km thick crust.
Airy isostasy could then have caused the observed sediment onlap. Concerning the subsequent deformation, it is clear that while it may have reinforced (or "flexed") this isostatic high, the crustal thickness variations along this profile have not been caused by it.

7.5.4 Line D

This Vema 2902 line runs ~SW - NE across the survey area (Figure 7.3) crossing both the fracture zone at 80.5°E and the Indrani at 79°E. The general decrease in depth to basement westwards (Figure 7.8) reflects the decrease in crustal age within each fracture zone block.

As would be expected from the models proposed for lines A, B and C this line can be modelled as a uniform crustal thickness of 5.0 km east of the Incirarn fracture zone and on both sides of the 80.5°E fracture zone, while to the west of the Indrani the crustal thickness increases (to 6.1 km).

The anomalies directly associated with the fracture zones are small, < 10 mGals, and partly obscured by the long wavelength features. Because of this very little information could be gained from the gravity modelling along this profile to elucidate the deep structure of fracture zones. As already discussed in Chapters 4 and 5 the fracture zones have little surface expression, although various features (basement trends and earthquakes for example) suggest that they may have been reactivated in a left-lateral sense. The gravity modelling neither supports nor negates the hypothesis of fracture zone reactivation.

7.6 Three-Dimensional Gravity Modelling

In a similar manner to the two-dimensional models, the three-dimensional program required a 3 layer model. A requirement for this is a contour map of each constituent interface: seafloor, top of basement and Moho. Only the southern part of the survey area, centred on the prominent crest modelled in Section 7.3.2, contained enough data to give sufficiently detailed maps.

The seafloor and top of basement maps (Figure 7.9) were contoured from depth-converted profiles. The Moho contour map was found by adding on a constant thickness to the basement contour map.
Figure 7.8 Two-dimensional gravity modelling for Line D (see Figure 7.3). Densities as for Line A. Stippled area indicates the sediment layer. The positions of two fracture zones are indicated. Continuous line is the observed anomaly and the dashed line is the anomaly produced by the underlying model. This model comprises 5.0 km uniformly folded crust east of the 79°E fracture zone increasing to a 6.1 km thickness crust to the west of it.
Figure 7.9 Contour maps from depth-converted profiles of (A) seafloor and (B) the top of oceanic basement. Contours in metres.
Each of the contour maps was digitised using the AUTOCAD facility in the Grant Institute and transferred to the Mainframe (EMAS). The three-dimensional program (LAMINA) computed the gravity effect of each surface in turn for a specified density contrast. The gravity anomalies from the three surfaces were then added to give the total modelled anomaly (Figure 7.10). The modelled anomaly could then be compared to the observed anomaly along line AA'.

The computed three-dimensional F.A.A. contour map is very similar to the observed F.A.A. contour map (Figure 7.4) and the SEASAT-derived contour map (Figure 7.2). A uniform thickness crust gave a modelled anomaly that closely fitted the observed (section AA' in Figure 7.10). The deviations in the fit, particularly towards the northern end of the Line AA', could be due to poor control in the contour map of basement.

A flat Moho producing relative thickening under the crests (Figure 7.10) gave an anomaly of insufficient amplitude. As already discussed greater thickening would diminish the amplitude further. Both the three-dimensional modelling and the two-dimensional modelling strongly favour the interpretation that the oceanic crust is uniformly folded over the prominent basement high. This is in agreement with the suggestion of Weissel and Haxby (1984) that the geoid undulations are consistent with crustal buckling. It agrees with the refraction work of Neprochnov et al. (1988) but contradicts the recent refraction study of Leger (1989).

7.7 The Indian Ocean Geoid Low, Small-scale Convection and Intraplate Deformation

The largest geoid anomaly world-wide is centred in the north-west of the Central Indian Ocean Basin, close to the Chagos-Laccadive Ridge. Although the Darwin study area is on the flanks of the geoid low it is possible that a component of the residual depth anomaly noted in Section 3.3.2. is associated with this feature. However, in a regional study Cochran and Talwani (1977) note that Central Indian Ocean Basin depths are close to those that would be expected for their lithospheric age.

Stark and Forsyth (1983) measured differential travel times for various S wave phases. They found that these travel times varied so that there was a clear increase in travel time with distance from the centre of the geoid low. They also found variations in the Central Indian Basin which were periodic with a wavelength of ~640 km and
Figure 7.10

(Top) Comparison of observed anomaly (solid line) with three-dimensional modelled anomaly for a uniform thickness crust (dashed line) and a flat Moho (dotted line). Underlying model is taken from Figure 7.6, not from the depth converted contour maps.

(Bottom) Three-dimensional computed gravity anomaly map for a three layer model with a 5.0 km thick crust. Densities for water, sediment, crust and mantle are 1030, 2100, 2800 and 3300 kg m\(^{-3}\) respectively. Absolute values of gravity are not meaningful - however relative values are. Line AA' is a cross-section across the prominent crest.
with a NNE trend. This agrees approximately with the trend that would be expected for linear, convective rolls in the upper mantle. As they note themselves the intraplate deformation described in this thesis has the wrong strike and wavelength to be associated with the convection rolls. They postulate that the intraplate deformation may mask the surface effects of convection.

7.8 Summary

1. Gravity modelling is consistent with a uniformly folded crust throughout much of the survey area. The basement high around 80°E 4°S was modelled in detail using both two and three-dimensional techniques and was found to have uniform thickness crust. This contradicts Leger (1989) who suggested thickening under this crest. If folding of the crust if representative of the behaviour of the brittle upper oceanic lithosphere this would support the buckling hypothesis.

2. Using a density of 2800 kg m\(^{-3}\) for the oceanic crust, crustal thicknesses of between 5.0 and 6.1 km were found within the survey area. This range is likely to be due to variations in the spreading-centre environment at formation and could be associated with the emplacement of the Afanasy Nikitin Seamount Group.

3. The SEASAT-derived geoid and ship-derived gravity contour maps are consistent with the undulations being discontinuous across fracture zones.

4. Little information could be gained from gravity modelling on fracture zone geometry. The gravity modelling neither supports nor negates the hypothesis of fracture zone reactivation.
8.1 General Introduction

To a large extent the format of this thesis is representative of the evolution of ideas during the project. That is, the first substantive Chapter (3) represents the earliest work, with the rationale and interpretation of each subsequent Chapter strongly influenced by the conclusions of the preceding ones. This final Chapter continues the trend with its aim being to discuss and fully combine the conclusions of the previous Chapters and explore the wider implications of these findings.

8.2 Fault Fabric in the Central Indian Ocean Basin and Spreading-Centres.

8.2.1 Introduction

The conclusion that the complex fault fabric developed in response to intraplate compression was the result of the reactivation of two sets of spreading-centre formed normal faults was made from the study of structural style (Chapter 4). One of these fault sets dips northwards, and was the original spreading-centre formed outward-dipping normal fault set, while the other, dipping southwards, corresponds to the inward-dipping normal fault set. By comparison with fault fabrics developed at similar spreading rates on the East Pacific Rise, it is suggested that every fourth or fifth fault was reactivated by the later deformation. New information on basement structure was provided by multichannel normal incidence seismic profiles (Chapter 5) which imaged the dip of faults throughout the deforming oceanic crust: it is thought that the original inward-dipping set may have a higher dip angle (35° - 45°+) than the outward-dipping set (30° - 40°). The aim of this section is to show the relationship between the principal conclusions outlined above for the fault fabric in the Central Indian Ocean Basin, and the fault fabrics formed by spreading-centres generally.
8.2.2 Evolution of Spreading-Centre Fabric and Inward and Outward-Dipping Faults

Fine scale studies of seafloor morphology have shown that the oceanic crust at mid-ocean ridges is highly fractured and that normal faulting accounts for the vast majority of ridge relief. Considerable evidence now exists for several phases of tectonism at different distances from spreading-centres.

Deep-tow and submersible studies of spreading-centres and direct observation of subaerial spreading suggests that the first tectonism of newly generated oceanic lithosphere everywhere consists of pervasive fissuring (Lonsdale, 1977; MacDonald and Luyendyk; 1977; Searle, 1984). With time these fissures generate vertical offsets, and small scale relief occurs within 1 - 2 km of the axis (Lonsdale, 1977). Following this, about 2 km from the axis, major inward-dipping normal faults evolve parallel to the spreading-centre. At this distance tectonic character varies markedly from spreading-centre to spreading-centre, mainly as a consequence of spreading-rate. The consistent onset of inward-dipping normal faults on all spreading-centres suggests two things: a universal extensional environment and (from polarity) a thinner weaker lithosphere towards the centre of the rift. Searle (1984) suggests that the uniformity of fault spacing of inward-dipping faults in the axis area is due to the presence of an isothermal layer controlled by strong hydrothermal circulation. On the East Pacific Rise, for example, Searle (1984) found the mean spacing of inward-dipping faults to be 1.7 km and their formation to be within 2 km of the axis.

Topography and fault fabric developed away from the axial region is, however, not universally the same and is strongly controlled by spreading-rate. On the fast-spreading East Pacific Rise, spreading at a rate similar to the one that formed the Central Indian Ocean Basin lithosphere, the topography is characterised by a rise 5 - 15 km wide and a few hundred metres high (Figure 8.1). By contrast, generally on slow-spreading ridges (Mid-Atlantic Ridge) the topography is characterised by 1.5 - 3.0 km deep axial rift valley and flanking rift mountains (Sempere and MacDonald, 1987; MacDonald and Atwater, 1978).

On the fast-spreading East Pacific Rise the next phase of tectonism usually occurs at between 5 and 8 km from the axis (Figure 8.1) with the formation of outward-dipping faults. However, Bicknell et al. (1988) find that on an ultrafast part of the East Pacific Rise at 19.5°S both sets develop simultaneously. Searle (1984) found that the spacing of this set of faults to be 2.6 km and suggested that their formation at this distance could be related to the crust moving off a sub-axial magma chamber. In general there
Figure 8.1 Deep Tow profile across the East Pacific Rise at 3° 25'S from Lonsdale (1977). Areas of thick pelagic sedimentation are shown in bold. The orientations of all faults in basement are deduced from fault scarps. Regions of inward and outward-dipping fault formation are inferred by the author from GLORIA observations along the same profile by Searle (1984). Pervasive fissuring occurs within the ~2 km wide axial shield volcano and significant offset faulting develops outside this zone.
appear to be approximately equal numbers of inward and outward-dipping normal faults (Searle, 1984; Bicknell et al., 1988). A combination of inward and outward-dipping normal faulting with back tilting results in abyssal hill formation (MacDonald and Luyendyk, 1985; Bicknell et al., 1988).

Following the formation of the rift mountains on slow-spreading ridges by normal faulting on inward-dipping faults, the transformation of the inward sloping fault slope into the undulating relief of the rift mountains occurs (Harrison and Steiltjes, 1977; MacDonald and Atwater, 1978). It is now generally accepted that this transformation takes place through a combination of back tilting of fault blocks and the formation of outward-dipping faults. Although outward-dipping faults are now well documented on slow-spreading ridges (MacDonald and Atwater, 1978; Searle, 1984; Kong et al., 1989) they only occur in significant numbers on fast-spreading rises (Searle, 1984; Bicknell et al., 1988; Carbotte and MacDonald, 1990).

An explanation for the predominance of inward-dipping faulting on slow spreading ridges and the more equal occurrence of inward and outward-dipping faults on fast-spreading ridges has been recently proposed by Carbotte and MacDonald (1990). They propose that the rapid increase in lithosphere thickness with distance from the axis on slow-spreading ridges makes it unfavourable for a second outward-dipping fault set to form (Figure 8.2). On fast-spreading ridges, the rate of increase with distance is slower and little extra work is necessary for the formation of the outward-dipping fault set. This is consistent with the maximum depth of seismicity as discussed in Section 5.5. In any event, the observation of approximately equal numbers of original inward and outward-dipping faults in the Central Indian Ocean Basin is wholly consistent with their formation at an original fast spreading-centre.

8.2.3 Abyssal Hills and Fault Reactivation

From an analysis of sediment onlap patterns onto basement and depth conversion of seismic profiles (Chapter 3), it was concluded (Chapter 4; Shipboard Scientific Party, 1989) that the original topography, prior to reactivation, was likely to have been similar to the abyssal hill topography found on the flanks of the East Pacific Rise today. Due to the complex nature of the fabric developed after deformation it is impossible to deduce if the original abyssal hill topography was controlled more by one set of faults than the other. Analysis of the East Pacific Rise suggests that the average throws of inward and outward-dipping faults are very similar (Searle, 1984; Bicknell et al., 1988) although Bicknell et al. (1988) find that the inward-dipping
Figure 8.2 Schematic figure from Carbotte and MacDonald (1990) explaining the greater occurrence of outward-dipping faults on fast-spreading ridges, by illustrating the relative difference in fault length between inward and outward-dipping faults inclined at 45°, at slow and fast spreading rates, assuming the base of the brittle layer is marked by the 600° isotherm. They suggest that it will take more work to break the lithosphere at slow spreading rates along an outward-dipping fault than an inward-dipping fault due to the difference in fault length, and hence the dominance of the inward-dipping faults. However, at faster rates the difference in length between the two fault sets diminishes, and so the work necessary to form outward-dipping faults is not significantly greater than that needed to form inward-dipping faults, explaining their more equal occurrence.
faults have slightly larger average throws. Bicknell et al. (1988) conclude that the abyssal hill topography on the East Pacific Rise is formed by inward and outward-dipping faults which form horst and graben structures (Figure 8.3) of wavelength 2 - 8 km, while topography of wavelength < 2 km is likely to be volcanic in origin. This differs from the mechanism for abyssal hill formation at slow and intermediate ridges which is dominated by back tilting of blocks bounded by inward-dipping normal faults (MacDonald and Luyendyk, 1985). In the Central Indian Ocean Basin the faults forming the abyssal hill topography, similar to Figure 8.3, have been reactivated with an average spacing of 6.6 km, close to the wavelength of the topography on the East Pacific Rise. Large irregularities in fault spacing at original spreading-centres may contribute to the irregular fault fabric in the Central Indian Ocean Basin.

The difference in dip between inward and outward-dipping faults, deduced from the multichannel profiles, suggests that there may have been some asymmetry to the abyssal hills prior to deformation in the Central Indian Ocean Basin. Assuming approximate parity in numbers of faults and similar average throws, the multichannel results suggests that the abyssal hills may have been more steeply dipping towards the spreading-centre than away (Figure 8.4). It is difficult to judge from published profiles (e.g. Figure 8.3) whether this asymmetry exists. Analysis of the original profiles is required before the hypothesis of asymmetry of abyssal hill topography on fast-spreading ridges can be confirmed or rejected.

8.2.4 Comparison with the Gorda Plate

Although the intraplate deformation in the Central Indian Ocean Basin is the most well-known and best developed example of intraplate deformation in the oceans, as mentioned in Section 1.1, it is not unique. Masson et al. (1990) report internal deformation of the Gorda plate with reactivation of ridge-parallel fault fabrics. The aim of this section is to compare and contrast the reactivation of ridge-parallel fault fabrics in the Gorda plate and the Central Indian Ocean Basin.

Deformation of the Gorda plate, associated with southward decreasing spreading-rates along the southern Gorda Ridge, is accommodated by clockwise rotation of Gorda plate crust, coupled with left-lateral motion (inferred from seismicity) on the normal faults of the oceanic crust as illustrated in Figure 8.5 (A+B) (Masson et al., op. cit.). In addition seismic reflection profiles suggest a small component of reverse faulting.
Figure 8.3 Line tracing of Deep Tow bathymetry over the ultrafast East Pacific Rise at 19° 30'S from Bicknell et al. (1988) with depths in metres. Fault bounded horsts (abyssal hills) are well developed. Note that the abyssal hills are formed by inward and outward-dipping faults and that back-tilting only has a minor role.
Figure 8.4 Schematic diagram of predicted abyssal hill topography formed on the flanks of a fast spreading-centre. Although the topography is irregular, with marked variations in fault spacing, there is a marked asymmetry (dashed lines) to the abyssal hills with a steeper gradient towards the spreading-axis due to the greater dip of the inward-facing faults than the outward-facing faults.
Figure 8.5 Comparison between Gorda Plate and Central Indian Ocean Basin internal deformation style (A) Schematic model to show how ~N - S shortening of a crustal block can be achieved by a combination of rotation and strike-slip faulting along the original ridge-parallel fabric. Dashed area shows potential overlap created by rotation which is eliminated by strike-slip faulting. Note that the angle between the shortening direction and the original ridge-parallel fabric is less than 30°. (After Masson et al., 1988). (B) Schematic model of Gorda Plate deformation (From Masson et al., 1988). Note that anomaly lengths (a-a' and b-b') remain constant during deformation. (C) Transpressive regime in the Central Indian Ocean Basin as developed in Chapter 4 comprising left-lateral reactivation of fracture zones, predominantly reverse (but with a component of right-lateral) reactivation of the original ridge-parallel fabric, and buckling within fracture zone compartments. Note that the angle between the shortening direction and the original ridge-parallel fabric is ~75°.
A transpressive model which was proposed for the Central Indian Ocean Basin in Chapter 4 is illustrated in Figure 8.5 (C), with left-lateral reactivation of fracture zones, buckling within fracture zone compartments and reactivation of the ridge-parallel fabric in a dominantly reverse sense, but with a right-lateral component.

Clearly in both areas the original ridge-parallel fabric has been reactivated in response to intraplate compression. However, in the Gorda plate the reactivation is predominantly in a strike-slip sense, whereas in the Central Indian Ocean Basin the reactivation is dominantly dip-slip. This contrast in tectonic style can be explained in terms of the orientation of the ridge-parallel fabric relative to the shortening direction: in the Gorda plate the fabric is orientated at less than 30° (Figure 8.5A), suitable for predominantly strike-slip reactivation; while in the Central Indian Ocean Basin the fabric is orientated at ~75° (Figure 8.5C), appropriate for mainly dip-slip reactivation.

8.3 Oceanic Crustal Structure

The aim of this section is to combine all the discussion of oceanic crustal structure from other Chapters into a coherent summary.

Results from wide-angle reflection profiles (Chapter 3) indicate that the crustal velocity structure in the Central Indian Ocean Basin is not anomalous. Sedimentary velocities are consistent with a consolidating turbiditic lithology, although higher velocities closer to basement may indicate a pelagic component. Oceanic layer 2 was found to have a normal thickness of 1.5 km. The thickness of layer 3 could not be determined from the sonobuoys.

Estimates of total crustal thicknesses in the Central Indian Ocean Basin are sparse and contradictory. A large study by Neprochov et al. (1988) found an average crustal thickness of 6 km. For the basement high at 80°E 4°S they found a thickness of 5.5 km. However, Leger (1989) for the same high found a crustal thickness of 9 km and for the adjacent troughs a thickness of 7 km. Leger's experiment was hampered by the loss of several instruments during collection. Gravity modelling (Chapter 7), while not giving direct evidence on crustal thickness, suggests a largely uniform crustal thickness with local variations which are more likely to be due to spreading-centre temperature variations than caused by the later deformation. While multichannel seismic profiles image faults penetrating to the expected level of the Moho (Chapter 5), the Moho was not imaged and crustal thickness not determined.
As discussed in Chapter 3 there is a small residual depth anomaly in the Central Indian Ocean Basin. The significance of this is unclear, although it could be due to random fluctuations of crustal thickness caused by variations in spreading-centre conditions at the time of crustal formation. In Chapter 6 it was concluded that the residual depth anomaly is unlikely to be caused by the load of the Bengal Fan.

In conclusion, the contradictory nature of present refraction evidence make it difficult to make a definite statement on absolute crustal thickness in the Central Indian Ocean Basin. However, from gravity modelling, it is clear that crustal thickness is fairly uniform, and variations are consistent with random fluctuations of conditions at the original spreading-centre. The later deformation has not significantly altered crustal thickness. This conclusion supports the buckling hypothesis for the formation of the long wavelength undulation and is discussed in more detail in the next section.

8.4 Buckling, Faulting and Rheology

8.4.1 Discussion for Oceanic Lithosphere

Both gravity modelling (Chapter 7) and analogue modelling (Chapter 6) strongly favour buckling as the mode of formation of the long wavelength undulations, with not only the crust but the entire brittle upper lithosphere folded. Elastic-plastic models (McAdoo and Sandwell, 1985) have shown that buckling is possible with thinning of the elastic core. Inverse boudinage, representing a 'weak' oceanic lithosphere (Zuber, 1987), is not favoured. The strong buckling mode is certainly more consistent with observations of the strength of the lithosphere in response to long term geologic loads.

Conclusions from various parts of this thesis on buckling and faulting of the oceanic lithosphere need to be combined into a coherent statement on the nature of oceanic lithosphere deformation and reconciled with possible rheological models.

As a result of the combination of the analysis of structural style (Chapter 4) with the interpretation of the multichannel seismic profiles, it was concluded that while the reverse faults had modified the long wavelength undulations, they had not formed them and therefore there was an underlying flexural control. It was also suggested that the flexural mechanism led to the propagation of faults from the region of the brittle/ductile transition, and the reactivation of the pre-existing spreading-centre
formed fabric in the crust. This has to be reconciled with the observations from analogue modelling that not only the crust, but the entire brittle thickness of the lithosphere is uniformly folded or buckled. Two possible rheologies consistent with these observations are discussed.

One interpretation would be that folding of the brittle layer occurred for the first few percent of shortening before stress increased, leading to the initialisation of faulting at the brittle/ductile transition. This faulting then propagated upwards and modified the earlier topography, with the observed association between the faults and the undulations presumably because of spatial variations in the level of the brittle/ductile transition. In favour of this interpretation is the observation of folding prior to faulting in the analogue models. Against are the results of ODP Leg 116 (Shipboard Scientific Party, 1989) which suggest that the faults have been consistently active from the onset of deformation, and the (admittedly tentative) estimates of growth rates from the analogue modelling. The long wavelength nature of the folding makes it impossible to use seismic profiles to determine if faulting and folding were/are synchronous.

An alternative explanation is to treat with caution the analogue modelling and suggest that the brittle layer is in fact behaving, at least for a few percent shortening, quasi-elastically. Thus no faulting is observed initially. Another explanation could be that faulting may be present in the analogue models, but below the level of detection of the laser. Thus faulting and folding may develop simultaneously.

An alternative rheology to the brittle/ductile model and one giving similar predictions is the elastic-plastic model in which the elastic core of the lithosphere thins due to brittle failure in the upper lithosphere and ductile creep in the lower lithosphere. For this interpretation the schematic Figure 8.6 is directly applicable, with the elastic core of the lithosphere driving the faulting. Here faulting and folding would develop simultaneously.

It is not possible to favour either of the models (elastic/plastic or brittle/ductile), they may be equally applicable. However it is clear that a layered rheology is required for the oceanic lithosphere to explain the observed long wavelength undulations, in agreement with the discussion of Section 6.3.2.
Figure 8.6 Cartoon illustrating the relationship between the first and second orders of the deformation. (1) In the survey area the crust is uniformly folded (as shown by gravity modelling in Chapter 7) and pervasively faulted (Chapter 5). One of the main conclusions of the analogue modelling (Chapter 6) was that not only the crust but the entire brittle lithosphere was folded (2) while the lithosphere/asthenosphere boundary was unperturbed (3). Brittle deformation nucleates at the brittle/ductile transition (6) and propagates upwards (4) reactivating the pre-existing spreading-centre formed fabric in the crust and uppermost mantle. The accentuation of the undulation crests by faulting could be due to variations in the depth to the brittle/ductile transition. Alternatively the deformation can be thought of as being driven by a thin elastic core (5) extending upwards and downwards from the brittle/ductile transition (6).
8.4.2 Buckling and the Continental Lithosphere

The recognition of buckling in the strong oceanic lithosphere raises the question as to whether such a phenomenon occurs in the rheologically more complex continental lithosphere. Compared with the simple two-layer brittle/ductile yield strength envelope, the continental lithosphere can be divided into three and four layer rheological models. (Figure 8.7).

Stephenson and Cloetingh (in press) show that the values of the total integrated, compressional strength of the lithosphere are greater for the oceanic domain than for the continental for all ages. They also show, from finite element modelling, that the critical in-plane buckling stress for young continental lithosphere could be as low as 100 MPa. From this Stephenson and Cloetingh (in press) conclude that folding of the continental lithosphere is likely to be an important mechanism of large scale continental deformation. Three examples of continental buckling are given: a Proterozoic example (Hoffman et al., 1988); a Palaeozoic example (Lambeck, 1983) and a Tertiary example (Stephenson and Ricketts, 1989). However, it should be emphasised that none of these examples seem, to the author, to be convincing, and there is certainly less good evidence than in the northern Indian Ocean.

The ocean basins are generally far more laterally homogenous than the continents. Even in the homogenous ocean basins, this study has shown (Chapter 4) that pre-existing structures or weaknesses (fracture zones) complicate simple buckling. In the continents there are very many pre-existing weaknesses. It is the author's view that large scale folding of the continental lithosphere is unlikely, due to failure of pre-existing weaknesses prior to the critical stress. It is, however, possible that in areas of unusual tectonic simplicity continental folding may be of significance.

Consequences of rheological structure for folding in the continents are not considered by Stephenson and Cloetingh (in press). With decoupling in the ductile lower crust, buckling is likely to occur on two wavelengths: a shorter wavelength will represent upper crustal folding, while a longer wavelength may represent whole lithosphere folding. If the relationship between brittle thickness and folding is the same as for the oceanic lithosphere (folding at ~7 times the brittle thickness) crustal folding would be expected at wavelengths of 70 - 105 km for typical upper continental crust of brittle thickness 10 - 15 km (Jackson and White, 1989). In regions of compression with tectonic simplicity, analyses of satellite imagery and topographic relief may be of use in looking for crustal folding. Such analyses are currently being undertaken at the
Figure 8.7 Yield Strength envelopes for various geotherms illustrating the greater complexity of continental lithosphere rheology. A Oceanic lithosphere; B Quartz-dolerite-olivine continental lithosphere; C Quartz-diorite-olivine continental lithosphere. All envelopes based on the experimental work of Goetze and Evans (1979) and Brace and Kohlstedt (1980).
University of Rennes (Institut de Geologie) by Peter Cobbold, Phillip Davy and Joseph Martinod. An additional complication is that the ductile lower crust may not completely decouple the crustal and whole lithosphere folding, producing interference of wavelengths and very complex deformations (J.Martinod, pers com.)

In summary, if the continental lithosphere was homogenous, then by comparison of compressive yield strength envelopes with the oceanic domain, buckling should be possible. However, the rheological structure of the continental lithosphere, means that two wavelengths of folding (crustal and lithospheric) may occur which may not be independent. Finally the inherent heterogeneity of the continental lithosphere may inhibit or make the deformation even more complex. For these reasons, even if folding of the continental lithosphere is an important mechanism of large scale compressive deformation, it is unlikely to be observed except in areas of unusual tectonic simplicity.

8.5 A Diffuse Plate Boundary in the Northern Indian Ocean

8.5.1 Introduction

The evolution of ideas for a diffuse plate boundary in the northern Indian Ocean are documented in this section, and the most recent models discussed in connection with the transpressive model proposed in Chapter 4 for the Central Indian Ocean Basin. Estimates of tectonic shortening from the multichannel profiles (Chapter 5) are compared with predictions of the plate boundary models. Finally the discussion is widened to include the current plate motion compilation of DeMets et al. (1990) and other hypothesised diffuse plate boundaries world-wide.

8.5.2 Review of Kinematic Models for a Diffuse Plate Boundary

The first attempt to explain the anomalous seismicity in the northern Indian Ocean, within the newly developed plate tectonic theory was made by Sykes (1970), who suggested that this seismicity might represent incipient subduction. Sykes (1970) proposed an incipient island arc from Sri Lanka to Australia, perhaps resulting from the westward migration of the Indonesian arc (Figure 8.8A). However, subsequent studies of seismicity (Stein and Okal, 1978; Bergman and Solomon, 1985) have shown that the spatial extent and significant strike-slip faulting make this hypothesis unlikely.
Figure 8.8 Diagram illustrating the historical development of early tectonic models for the northern Indian Ocean (after Wiens et al., 1986). See text for discussion.
Evidence for north-south shortening in the Central Indian Ocean can be found by the inversion of relative plate motion data from the ridge systems of the Indian Ocean. Lack of velocity closure at the triple junction between the African, Antarctic and Indian plates, and the misfitting of spreading-rates and directions at the Indian Ocean ridge systems, led Minster and Jordan (1978) to suggest that a better fit could be obtained by dividing the Indo-Australian plates into two plates. Minster and Jordan (1978) suggested that this boundary could extend from the Sumatran Trench to the Southeast Indian Ridge along the Ninetyeast Ridge. They predicted 10 mm yr\(^{-1}\) of north-west directed convergence along this proposed boundary between the Indian and Australian plates.

This hypothesis was partly supported by Stein and Okal (1978) who found that the northern part of the Ninetyeast Ridge (north of 10°S) is seismically active. They suggested left-lateral strike-slip motion along the northern part of the Ninetyeast Ridge at a rate of > 20 mm yr\(^{-1}\) with a small component of north-west directed convergence. Therefore, Stein and Okal (1978) supported the idea of the Ninetyeast Ridge acting as a plate boundary (Figure 8.8B), but only north of 10°S. Interestingly this latitude marks a change in the morphology of the Ninetyeast ridge from a straight flat-topped ridge to the south, to a series of discrete en echelon blocks to the north.

Following the observations of distributed intraplate deformation in the Central Indian Ocean Basin (Weissel \textit{et al.}, 1980; Geller \textit{et al.}, 1983), and the absence of seismicity south of 10°S along the Ninetyeast Ridge, it became clear that a number of problems were not explained by the model of Minster and Jordan (1978). These problems were resolved by Wiens \textit{et al.} (1985) who found that the best fit to plate motion data could be obtained with a plate containing India and Arabia and a plate containing Australia with a diffuse plate boundary extending across the equatorial Indian Ocean from the Central Indian Ridge to the Ninetyeast Ridge and north along the Ninetyeast Ridge to the Sumatran Trench (Figure 8.8C). This reconciled earthquake activity, the compressional deformation in the Central Indian Ocean Basin and plate motion data then available. The NUVEL-1 relative motion dataset (Stein and Gordon, 1984) was used to calculate a rotation pole between the Indo-Arabian and Australian plates at 1.5°S 69.6°E with a rotation rate of 0.48°Ma\(^{-1}\) and a lithospheric shortening of 11 mm yr\(^{-1}\) at 81°E.

Wiens (1986) interpreted the complex seismicity around the Chagos Bank (Section 1.3.1) in terms of the Wiens \textit{et al.} (1985) configuration, with the proximity of the pole explaining the complex seismicity.
The next study was made by Demets et al. (1988) who obtained a more consistent database for Indian Ocean plate kinematics by remodelling all the magnetic anomaly data along the spreading ridges of the Indian Ocean. This study found that non-closure of the triple junction was insignificant when deformation was concentrated along a diffuse plate boundary, with a pole of rotation to the east of the Chagos Bank, as suggested by Wiens et al. (1985).

Mainly from reanalysis of seismicity, Petroy and Wiens (1989) postulated the extension of the diffuse plate boundary eastwards across the Ninetyeast Ridge, with convergence in a considerable part of the Wharton Basin (Figure 8.9). They argue that the Ninetyeast Ridge only takes up a small fraction (~2.8 of ~7 mm yr⁻¹) of the relative plate motion and that east of the Ninetyeast Ridge the diffuse plate boundary widens, with seismicity and linear gravity anomalies indicating lithospheric flexure extending further south. In addition they suggest that the diffuse deformation affects the style of faulting along the outer-rise of the Sumatran Trench, which the boundary intersects over a zone several thousand kilometres long. Petroy and Wiens (1989) explain the distribution of strike-slip and compressive earthquakes by noting that strike-slip events occur preferentially along large features, such as the Ninetyeast Ridge and 86°E fracture zones, while thrust faulting occurs preferentially between these features.

In a subsequent study of all available Indian Ocean and Gulf of Aden plate motion data (comprising about four times as much data as used by Demets et al., 1988), Gordon et al. (1990) find a Euler vector of 0.313°Ma⁻¹ about 5°S 78°E, for the motion between India and Australia. With this expanded dataset the best fit is found for India and Arabia as separate plates, divided in part along the Owen Fracture Zone (Figure 8.10). This model (Figure 8.11) predicts ~N-S extension of 6 ± 2 mm yr⁻¹ at 68°E and ~N-S compression of 4 ± 3 mm yr⁻¹ at 85°E, only 30% as fast as Wiens et al. (1985), and broadly consistent with earthquake focal mechanisms (Bergman and Solomon, 1985; Wiens, 1986; Petroy and Wiens, 1989).

The most recent study, using an alternative approach, has been completed by Royer and Chang (submitted). They test the hypothesis for a diffuse plate boundary between the Indian and Australian plates by using magnetic anomaly data and fracture zone traces along the three mid-oceanic ridges of the Indian Ocean, to obtain plate reconstructions from 10 Ma back to 65 Ma. They found an averaged India/Australia Euler vector at 11.1°S 78.0°E with a rotation of 0.354°Ma⁻¹. This analysis predicts shortening of 4.6 ± 5.2 mm yr⁻¹ at 85°E (solution 1 - 10 Ma) or, in their favoured
Figure 8.9 Petroy and Wiens (1989) interpretation of the tectonics of the diffuse deformation in the northeast Indian Ocean. Diamonds denote linear gravity highs associated with the folding of the lithosphere. As discussed elsewhere in this thesis neither the lithospheric folds nor the gravity anomalies are continuous across fracture zones so this interpretation is simplistic. Small circles denote well located earthquakes, and the dotted line is the approximate geographic extent of the zone of intense deformation as suggested by seismicity and gravity data. In this interpretation the plate boundary extends across the Ninetyeast Ridge and intersects the Sumatran Trench along a diffuse zone several thousand km long.
Figure 8.10 Plate geometries that have been proposed for the Indian Ocean (From Gordon et al., 1990). A) The conventional plate boundary with the single Indo-Australian plate. B) India and Australia lie on distinct plates separated by a diffuse plate boundary. In this model, proposed by Wiens et al. (1985), there is no motion along the Owen Fracture Zone and so India and Arabia lie on one plate (the Indo-Arabian plate). C) The present model (Gordon et al., 1990; Royer and Chang, submitted) with India, Australia and Arabia lying on distinct plates: India and Australia are separated by the diffuse plate boundary; motion between India and Arabia occurs along the Owen Fracture Zone.
Figure 8.11 The diffuse boundary (white region) separating the Indian and Australian plates as proposed by Gordon et al. (1990). The Australian plate moves anti-clockwise relative to the Indian plate; the thick arrows on the Australian plate's northern boundary show the motion of the Australian plate relative to an arbitrarily fixed Indian plate. The solid triangle is the Euler pole of Wiens et al. (1985), and the open circle is the Euler pole determined assuming India and Arabia lie on the same plate (From Gordon et al., 1990).
solution, of $5.8 + 3.5 \text{ mm yr}^{-1}$ (solution 3 - 21 Ma) at the same longitude. The latter solution is preferred because of varying Euler vectors at different times of reconstruction. Royer and Chang (submitted) suggest deformation started sometime after 21 Ma and suggest that the limited evidence for the onset of deformation at 7 Ma may be because the deformation was diachronous.

In summary, the most recent studies (Gordon et al., 1990; Royer and Chang, submitted) which use the most up to date databases, although differing in methods, find a pole of rotation between India and Australia which is broadly consistent with observations of seismicity throughout the northern Indian Ocean and of compression in the Central Indian Ocean Basin.

8.5.3 Estimates of Lithospheric Shortening

An estimate was obtained in Chapter 5 from the multichannel profiles of total tectonic shortening of $3.8 \pm 1.2 \text{ mm yr}^{-1}$, with reverse faulting contributing more than 20 times as much as the long wavelength undulations. This estimate was obtained for profiles lying between latitudes 79.5°E and 81.5°E and assuming an onset of deformation at 7 Ma. Comparison can be made with the predictions of the studies of Gordon et al. (1990) and Royer and Chang (submitted) for the same longitude (~80°E). These predict a shortening of $1 \pm 3 \text{ mm yr}^{-1}$ and $3.0 \pm 2.8 \text{ mm yr}^{-1}$ respectively. Although the estimates agree within errors, this agreement does not prove the validity of the assumptions used in plate motion studies.

A prediction of both the studies of Royer and Chang (submitted) and Gordon et al. (1990) is that shortening should decrease towards the pole of rotation. Although the longitude spread of the CHARLES DARWIN profiles is limited (78.3°E - 81.5°E) there is no indication of a decrease in fault activity (and hence shortening) westwards: Indeed, a seismic profile lying very close to the pole of Gordon et al. (1990) at 78°E 5°S shows considerable shortening. It seems, to the author, to be highly questionable as to whether the development of the complex deformation zone through time can be represented by a single Euler pole. This said, the estimates of tectonic shortening from the multichannel profiles are broadly in agreement with the best available estimates from plate motion studies.
In Chapter 4 it was concluded that the compressive stresses had interacted with the pre-existing oceanic fabric to contribute to a transpressive environment in the survey area, with a significant component of strike-slip motion. However, at present it is impossible to determine the amount of strike-slip motion because although estimates of tectonic shortening agree within errors, these errors are so large that it is likely that the contribution due to strike-slip motion has been hidden.

8.5.4 Transpressive Tectonics of the Diffuse Plate Boundary

Perhaps the major conclusion of the study of structural style (Chapter 4) was the extent to which the pre-existing fabric of the ocean floor (fracture zone and original ridge-parallel) controlled the development of the deformation. A transpressive model was also proposed in which undulations form within fracture zone compartments, fracture zones are reactivated in a left-lateral strike-slip sense, and original ridge-parallel faults become reactivated as reverse faults with a strike-slip component.

The transpressive model indicates that motion within at least part of the diffuse plate boundary has been accommodated by en echelon fracture zone blocks. The complex undulation plan both locally (Figure 4.5) and within the Central Indian Ocean Basin (Figure 1.4) is primarily because of fracture zone control. In many cases (Stein et al., 1989; Petroy and Wiens, 1989) interpretation of SEASAT profiles has been too simplistic and often the spacing of the profiles is not tight enough to allow unique profile to profile correlation of gravity highs and lows (Royer and Chang; submitted). In detail it is clear (Chapter 7) that the continuity of SEASAT gravity undulations is broken up by fracture zones.

Some authors have argued (Stein et al., 1989; Stein et al., in press) that the detailed changes in orientation of the axes of the undulations are consistent with the intraplate stress configuration predicted from numerical modelling of plate boundary forces (Cloetingh and Wortel, 1986). Although very large changes in orientation, broadly NE-SW east of the Ninetyeast Ridge to more E-W to the west, must be due to changes in the orientation of the axes of maximum compression (see Figure 4.12), this is unlikely to be true for more local (even on the scale of the Central Indian Ocean Basin) variations. The author strongly argues that the style of deformation, including changes in the orientations of the undulations, has been controlled by the
pre-existing structural fabric. The use of the predictions of numerical modelling of plate boundary forces (Cloetingh and Wortel, 1986) must be viewed with caution as the Indo-Australian plate is neither homogenous on a large (Ninetyeast Ridge) nor small scale (e.g. fracture zones).

Plate motion studies (Royer and Chang, submitted; Gordon et al., 1990), numerical modelling (Cloetingh and Wortel, 1986) and the most recent seismicity study (Petry and Wiens, 1989) are all consistent in predicting that the stress should increase eastwards across the Ninetyeast Ridge into the Wharton Basin, and that deformation should be best developed in this area. However, several authors have stated that deformation is less well developed in the Wharton Basin (Stein et al., 1989; Karner and Weissel, in press). There are a couple of points that should be made concerning this disagreement: firstly, there is a general lack of data from the Wharton Basin; and secondly, it is likely that with transpressive tectonics the deformation planform would be considerably more complex than in the Central Indian Ocean Basin, including greater amounts of strike-slip motion and fracture zone reactivation, because with increasing distance from the rotation pole greater amounts of shortening have to be accommodated. Greater fracture zone reactivation would tend to disrupt the continuity of the SEASAT gravity undulations and perhaps make the deformation look less intense.

In conclusion, the deformation in the Indian Ocean occurs in a broad zone, which plate boundary motion studies define as a diffuse plate boundary between the Indian and Australian plates. When considering the transressive regime in the Central Indian Ocean Basin and northeastern Indian Ocean it is perhaps more useful to think of the deformation as compression with anticlockwise rotation (see Section 4.6) within a series of en echelon fracture zone compartments (Figure 8.12). The driving mechanism for deformation being provided by the focussing of stresses due to dramatic changes in plate boundary configuration, with subduction occurring faster at the Sunda Trench than continental collision at the Himalayas.

8.5.5 Current Plate Motions and Diffuse Plate Boundaries

Diffuse plate boundaries within the oceans are now being incorporated in plate motion models. DeMets et al. (1990) have recently produced a revised model of plate motions, NUVEL-1, which describes the relative motions over the last 3 Ma of the 12 major plates. NUVEL-1 resolves many of the problems associated with earlier models by incorporating about three times as much data. The most significant
Figure 8.12 (Top) The diffuse plate boundary area (delimited by dotted lines) deforms by left-lateral strike-slip along fracture zones (dashed lines) with buckling and reverse faulting (not shown) within the fracture zone compartments. Positions of the fracture zones from Royer (unpublished work). (Bottom) Schematic diagram showing the overall result is not only shortening, but anticlockwise rotation of the deforming zone.
changes are the inclusion of a diffuse plate boundary between separate Indian and Australian plates and between the North and South American plates (Figure 8.13). Predictions of the NUVEL-1 data set for shortening rates in the northern Indian Ocean are described in detail by Gordon et al. (1990) and are discussed in Section 8.5.3.

A diffuse plate boundary between the North and South American plates is invoked because the strikes of well-mapped transform faults along the Mid-Atlantic Ridge have small, systematic, significant departures from the predictions of a model with a discrete North America-South America plate boundary (DeMets et al., op. cit.). In addition it is suggested that remaining systematic misfits to azimuth data along the Southeast Indian Ridge and along the western oceanic part of the Antarctic-South America plate boundary may be small diffuse deformations of the oceanic lithosphere.

In summary the premise of plate tectonics discussed in Section 1.1 of rigid plates divided by discrete boundaries continues to be useful (DeMets et al., 1990). However it is clear that diffuse plate boundary zones of variable width can separate rigid plates in the oceans as well as in the continents, where they have long been recognised.

8.6 Speculations on Reasons for the Timing of the Onset of Deformation

Results from ODP Leg 116 (Shipboard Scientific Party, 1990), discussed in Section 1.6, give a date for the onset of deformation of 7 Ma. Although it is difficult to interpret the unconformity representing the onset of the deformation continuously along seismic profiles, it is clearly present throughout the DARWIN survey area and is seen on other profiles throughout the intraplate deformation area (Weissel et al., 1980; Curray and Munasinghe, 1989). With the present database it is impossible to know if the onset of deformation was diachronous throughout the Indian Ocean. However the universal relative shallowness of the unconformity in the sedimentary sequence attests to its recent creation. In this section two possible (and speculative) reasons are given for the onset of deformation at 7 Ma.

Many authors (Weissel et al., 1980; Bergman and Solomon, 1985; McAdoo and Sandwell, 1985) have appealed to unspecified processes in the Himalayas as causing the onset of deformation. Weissel et al., (1980) suggested that the intra-plate deformation might be associated with the phase of Himalayan uplift that resulted from the collision of India with Asia during the late Miocene. However, O.D.P. Leg
Figure 8.13 Data locations and plate geometries assumed for NUVEL-1. Regions with vertical lines mark diffuse plate boundaries between North and South America and between India and Australia. Within each of these diffuse boundaries a dashed line shows the discrete boundary assumed in NUVEL-1. Squares show the locations of spreading rates, circles show locations of transform azimuths, and triangles show earthquake locations for slip vectors (except those along transform faults offsetting mid-ocean ridges, which are omitted for clarity). Plate name abbreviations: Cocos (CO), Caribbean (CA), Indian (IN), Arabian (AR), Philippine (PH), and Juan de Fuca (JF). From DeMets et al. (1990).
116 evidence (Cochran et al., 1987) suggests that the Himalayan uplift had already begun by the early Miocene (17 Ma), while the onset of deformation is at 7 Ma. Continuing studies of the timing and magnitude of events in the Himalayas shed light on this discrepancy. For example, a recent study of the Tibetan plateau suggests that much of the uplift and extension there may have occurred as recently as the past ~5 Ma (Dewey et al., 1988; Shackleton et al., 1988). The uplift of the Tibetan plateau has been modelled by England and Houseman (1988) as a consequence of delamination of a thickened lithospheric root. If lithosphere shortening and thickening ceased when delamination and uplift started, then continued convergence between India and Asia will have forced shortening to be accommodated in other areas. A renewal of deformation in the Himalayas (Zeitler, 1985) may mark this accommodation as well as the onset of deformation in areas as widely separated as the Tien Shan (Windley et al., 1990) and the Central Indian Ocean.

An alternative explanation is given by Wiens et al. (1986), who suggest that the onset of deformation and convergence in the northern Indian Ocean may be related to a process of regional plate boundary reorganisation contemporaneous with the separation of Arabia from Somalia and the cessation of motion on the Owen Fracture Zone. They note that seafloor spreading began in the Gulf of Aden about 10 Ma.

8.7 Speculations on Subduction and the Future Development of the Intraplate Deformation

It is clear from the foregoing that shortening between India and Australia is occurring over a wide zone (~1500 km), predominantly by reverse faulting, but also by buckling of the oceanic lithosphere and strike-slip motions. There is no evidence within the deformation zone of any structure representing incipient subduction, however with continued convergence, as presently shown by anomalous seismicity and neotectonic faulting, there are several possible ways in which the deformation could develop.

Firstly, the deformation could increase in north-south extent, propagating further to the south and further north under the Bengal Fan. Secondly, or after the development of the first suggestion, the amplitudes of the buckles could increase. Analogue modelling has shown (Chapter 6; J. Martinod pers. com.) that, with continued shortening, buckling amplitude increases until a whole lithosphere fault forms, thereafter the amplitude of folding decreases away from the fault zone, and all shortening is taken up by a thrust fault (Figure 8.14). Over the time interval since the
Figure 8.14 Cartoon illustrating how the diffuse deformation may develop into a diffuse plate boundary zone and how, with continued shortening, it may develop into a discrete plate boundary with the initialisation of subduction. A For small amounts of shortening complex low amplitude topography results from the interaction of the pre-existing topography with a large range of perturbation wavelengths. B With continued shortening a dominant wavelength (that of buckling) emerges. C As shown by the analogue experiments this dominant wavelength is amplified to produce the phenomena observed in fracture zone compartments today in the northern Indian Ocean. During the development of the deformation over \(~7\) Ma elastic strain energy has been stored in the buckles. D In the future with continued shortening, further amplification and increase in bending stress, the oceanic lithosphere may fail by the development of a fault. In the analogue experiments this is observed at the inflexion point of the buckles, where the shear stress is a minimum. E With the formation of the whole lithosphere fault, or subduction zone, all the elastic strain energy stored within the fold train can be dissipated by driving the downgoing slab into the asthenosphere. This process is directly analogous to an elastic spring being compressed and then released. The stored energy overcomes the initial viscous resistance and bending moment necessary until the subduction zone becomes self-driven. With energy dissipation the topography away from the subduction zone is attenuated, as observed in the analogue experiments.
onset of intraplate deformation elastic strain energy has been stored in the buckles. With whole lithospheric failure this energy is dissipated doing the work necessary (e.g. overcoming the viscous resistance, formation of the enormous bending moment) to make the subduction zone self-perpetuating, i.e. when enough dense lithosphere has been subducted to pull more lithosphere down without the need for an external driving force. The recognition of the driving force represented by the elastic strain energy stored in the undulations may make the initialisation and early continuation of subduction mechanically easier than has been previously thought (e.g. McKenzie, 1977). Thus a diffuse plate boundary could still become a discrete plate boundary, with the deformation in the northern Indian Ocean representing an integral part of the cycle of plate tectonic activity (Wilson Cycle). At present the initialisation of subduction is the most poorly understood part of this cycle.

A third possibility for the future development of the deformation is an increase in the strike-slip component, with greater reactivation of pre-existing structures, leading to an even more complex tectonic regime. Strike-slip motion would have to be concentrated on one fracture zone if this was to develop into a discrete plate boundary.

8.8 Recommendations for Future Work

1. Although good evidence has been presented for buckling as the mode of formation of the long wavelength undulations, it is not conclusive. Resolution of the Moho by multichannel seismics is an obvious next step, and while the CHARLES DARWIN MCS was unsuccessful, more powerful systems could be expected to routinely image the Moho. Not surprisingly there are two MCS cruise proposals to visit the area in the near future: an American cruise (Principal Scientists, Jim Cochran and Jeff Weissel at the Lamont-Doherty Geological Observatory) and a French Cruise (Principal Scientist, Beatrice de Vogd at the Ecole Normale Superieure in Paris), with the latter provisionally planned for August 1991. The northward extent of the intraplate deformation under the Bengal Fan is unknown and it would be interesting to collect an MCS profile as far north as the top of basement could be imaged.
2. CHARLES DARWIN multichannel profiles imaged reverse faults to the bottom of the oceanic crust. Earthquake seismicity is concentrated at depths of 29 - 40 km (Bergman and Solomon, 1985) and thus the question is raised as to the depth extent of faulting. Another unknown is the component of strike-slip associated with these reactivated features. The best way of resolving these uncertainties would be to carry out an Ocean Bottom Seismometer (OBS) experiment around one of the active faults. The major problem is to locate a seismically active fault. While not revealing the location, Neprochnov et al. (1988) reports that more than 100 small earthquakes were registered at two sites during 10 days of continuous seismological recording. To collaborate with Soviet scientists and revisit these sites to undertake detailed OBS experiments would yield valuable results.

3. This thesis has mainly concentrated on the larger and deeper structure of the deformation. There has been relatively little analysis of fine-scale sedimentological and structural features associated with the reverse faults as revealed by MCS profiles and 3.5 kHz records. It is hoped to use these two sources of data from CHARLES DARWIN 28 to carry out a sedimentological/structural investigation in the near future.

4. Finite element modelling of the oceanic lithosphere under compression, while unsuccessful in this study, could, with further work, be a useful tool for understanding the intraplate deformation and oceanic lithosphere rheology.

5. One of the conclusions of this thesis, reached by comparison of fault geometry with seismicity studies at present day spreading-centres, was that the deeper parts of the reverse faults may have been formed during the phase of deformation. An alternative explanation could be that the whole fault planes were already present, and that the deeper parts of faults at spreading-centres form by aseismic creep, or with a level of seismicity beneath present levels of detectability.

A way of checking the original depth extent of faulting at spreading-centres may be by detailed mapping of ophiolites. Wide shear zones, dipping at 45°, have been observed in the upper gabbros of the Troodos Ophiolite in Cyprus, and have been interpreted as original spreading-centre fabric (G. Jones pers com.). The author floats the idea that detailed mapping of the original fault fabric in ophiolites (depth extent, orientation and spatial regularity) may yield significant information on the mechanics of active spreading-centres.
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Appendix A - Multichannel Seismic Processing

**Digitising**

After the failure of the time base on channel 14 of the Store 14 tape recorder aboard ship the rest of the processing sequence was dependent on the replay of the analogue tapes being at the same speed as that at the time of recording. Remarkably, any errors introduced appear to have been negligible. The 14 data channels were sampled every 4 ms, equivalent to a Nyquist frequency of 125 Hz. Because no facility was available at the British Geological Survey to provide an anti-alias filter prior to digitisation there was a danger that frequencies would be misrepresented or aliased. Each analogue tape produced seven 1600 bpi digital tapes so that 105 digital tapes were produced in total. An overlap of three shots was allowed between each tape to ensure that no data was missed. Each digital tape contained 80 minutes of data and as the digitising could be undertaken at twice the recorded speed, the digitisation of each analogue tape took a full working day. After a few initial problems getting the system running (which were solved by the considerable expertise of Mr C. Fyfe) this digitising stage took the author about four weeks.

During the survey six multichannel lines of varying length were shot (Figure 2.1). The line numbers and corresponding digital tape numbers are detailed below.

<table>
<thead>
<tr>
<th>Line Number</th>
<th>Digital Tape Numbers</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1 - 42</td>
</tr>
<tr>
<td>2</td>
<td>43 - 58</td>
</tr>
<tr>
<td>3</td>
<td>59 - 68</td>
</tr>
<tr>
<td>4</td>
<td>69 - 88</td>
</tr>
<tr>
<td>5</td>
<td>89 - 94</td>
</tr>
<tr>
<td>6</td>
<td>95 - 105</td>
</tr>
</tbody>
</table>

Financial constraints were such that not all the digitised tapes could be processed by GECO. Eventually tapes 1-48 and 56-72 were processed to stack, corresponding to the whole of line 1, most of line 2, line 3 and the start of line 4. Additionally, tapes 1-18 and 43-48 were processed to migration.
Data Recovery and Reformatting

The BGS digitised format, although sensibly simple, was not in the industry standard SEG Y format which GECO's processing system demanded. Digitisation at BGS into SEG Y format would have been prohibitive, both in terms of cost and time. More economically, a computer program was written by GECO to convert the BGS format into the desired 6250 bpi SEG-Y format.

Some problems were experienced by GECO in reading the digitised tapes, however this was overcome by 'cleaning' the tapes prior to recovering the data. Once this was done the data were quickly reformatted and gain recovery applied.

As the water bottom had a TWT of ~6 seconds, it was decided to only process the first twelve seconds, as below this the primary signal would be greatly masked by the water bottom multiple.

Spherical Divergence Corrections

Trials were undertaken to determine the best Spherical Divergence Correction to apply. The Correction in the form below was judged the most suitable to compensate for amplitude loss with increasing TWT.

<table>
<thead>
<tr>
<th>Velocity (m/s)</th>
<th>Applied until (seconds)</th>
</tr>
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<tbody>
<tr>
<td>1480</td>
<td>6.2</td>
</tr>
<tr>
<td>1750</td>
<td>8.2</td>
</tr>
<tr>
<td>3500</td>
<td>12.0</td>
</tr>
</tbody>
</table>

CDP Gathers and Whole Trace Equalisation

After the loss of a time base on the Store 14 the processing sequence is dependent on the shot interval being constant. When the data was gathered it became clear that in a few places along the profiles some shots were missing. Overall the effect of this was small in the final stacked sections. However areas with shots missing tended to migrate poorly.

The traces were equalised such that all traces had equal amplitude within a gather.
Deconvolution Before Stack

This operator was designed to attenuate some of the short period peg-leg multiples in the sedimentary sequence. Following trials a operator of length 360 ms and gap 36 ms was selected. The long gap length was necessary to take account of the length and variability of the source signature.

Signature Deconvolution - not applied

Originally it was envisaged that signature deconvolution would be carried out in the processing sequence. However, after advice from GECO and checking with people who had previous experience of processing data from NERC ships, it was decided that, without a recorded source signature and taking into account the likely variability of this signature along the profiles, signature deconvolution would simply not be worthwhile. It was likely that, in order that the data was not degraded, a separated operator would have to be 'data derived' every few shots. This would be an expensive process, beyond our budget, and therefore not applied.

Stacking Velocity Determination

In the deep water of the Central Indian Ocean Basin (4.7 km) normal moveout along the 1 km long array was small. This made stacking velocity determination difficult - particularly in the deeper sediments and crust.

Doubts were expressed by GECO as to whether their full VELSEIS analysis (which is relatively expensive) would be worthwhile. Velocities were picked by both the author and GECO using firstly just a Semblance and Amplitude plot and then using the full VELSEIS analysis, which includes stacked panels, for the same selection of CDP numbers. On comparison of the two picked velocity spectra it was found that they were practically identical except in the deeper crust where neither technique had any control. It was agreed that the rest of the profiles should be picked using Semblance and Amplitude plots. What the VELSEIS stacked sections did illustrate, however, was that in the lower sediments and crust large variations in stacking velocities did not change the quality of the stacked section.
It is perhaps worthwhile noting that velocities picked for stacking are those that give the best stacked section. Often they have little relation to physically meaningful (as determined by the Wide-Angle profiles) velocities. These stacking velocities should not though be used unchanged for migration purposes as discussed below.

Contour plots of stacking velocity along the profiles illustrate how sensitive the stacking velocities are to structural change, especially in the near surface. Velocity spectra were determined every 10 km, and additionally on either side of major fault blocks and in areas of structural significance.

**Stack**

By interpolating between picked velocity spectra the sections were stacked to give 12 fold stacked sections.

**Shot and Streamer Statics**

A shot and streamer static of 13 ms was applied to bring all datum to sea-level.

**Deconvolution After Stack (DAS) - not applied**

Deconvolution after stack is traditionally used mainly to suppress the water-bottom multiple. With this Indian Ocean data the seabed was sufficiently deep that it's multiple was out of the zone of interest (below 12 seconds). Tests were undertaken to see if various DAS operators improved the quality of the section by removing other long period multiples. No improvement was detected and so DAS was not applied.
Time Variant Bandpass Filter

Because the earth acts as a high-cut filter a time varying filter is applied to select the frequency interval with the best signal to noise ratio for each part of the time section. For this Indian Ocean data a balance had to be reached between the desire to remove the effects of higher frequency multiples (including some of the effect of Overshooting discussed below) and the wish to retain the higher frequencies to obtain the best possible resolution. On the basis of both narrow and broad band filter tests the following time varying filter was chosen.

<table>
<thead>
<tr>
<th>Low cut (Hz)</th>
<th>Roll off (db/octave)</th>
<th>High cut (Hz)</th>
<th>Roll off (db/octave)</th>
<th>Applied down to (seconds)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>22</td>
<td>80</td>
<td>72</td>
<td>6.0</td>
</tr>
<tr>
<td>5</td>
<td>22</td>
<td>70</td>
<td>72</td>
<td>6.5</td>
</tr>
<tr>
<td>5</td>
<td>22</td>
<td>55</td>
<td>72</td>
<td>7.5</td>
</tr>
<tr>
<td>5</td>
<td>22</td>
<td>40</td>
<td>72</td>
<td>9.0</td>
</tr>
<tr>
<td>5</td>
<td>22</td>
<td>28</td>
<td>72</td>
<td>12.0</td>
</tr>
</tbody>
</table>

Equalisation

On the basis of tests a 500 ms balance scalar with a 50 percent overlap was chosen to be applied down the entire time window as this was thought to give the optimum data stand out.

Migration

Prior to migration our specified time varying filter was applied with 5 Hz added to the high cut. Selected lengths of the profiles were migrated using a finite difference wave equation algorithm which permits dips of up to 50° to be migrated without significant distortion. Around half of line 1 (tapes 1-18) and some of line 2 (tapes 43-48) was migrated using this algorithm.
The most important variable in any migration is the velocity-depth function used. Stacking velocities determined were generally some 10 - 15 % lower than our 'physically realistic' velocities determined from wide-angle reflections (see Chapter 3). Migration tests were undertaken using velocity-depth functions which were defined by various percentages of the stacking velocities, as well as using a function that was defined purely by wide-angle reflection results. It was found that the wide-angle reflection data gave an overmigrated section. For most of the profiles 105 % of the stacking velocities gave the best migrated sections. However, locally a rather lower percentage was necessary to avoid overmigration.

Steep dip migration, accommodating rapid lateral velocity variation was also performed on a selected panel of data using an XF algorithm. This allowed correct migration of dips utilizing a variable velocity field.

Display

A 10 % positive bias was applied as it gave the most interpretable section. The profiles were annotated by CDP number as well as occasionally by Julian Day and time. The final sections were displayed with scales of 10 traces per cm, 10 cms per second and a datum at sea-level.

Post-Migration Depth Conversion

Depth conversion is a useful process as it reveals true geometries and prevents incorrect conclusions being drawn from the time sections. For example, faults on a time section may appear listric because of the increase of interval velocity with depth, whereas after depth conversion using the appropriate velocity-depth function the faults reveal themselves to be in fact planar.

A small panel of data, through the ODP sites, was depth converted using the following processing sequence. A time varying filter was applied to the existing raw migrated time data using previously defined parameters. After this, time to depth conversion utilizing the migration velocity field was applied. The depth section, corresponding to the input time window of 6.0 - 12.0 seconds, was displayed at the usual 1:50000 scale.
This processing sequence gave only an approximation to the depth section. The largest assumption is that of a constant interval velocity within a specified zone. After studying this depth section it was decided that depth conversion of more data was not worthwhile. This decision was reached because the depth conversion did not significantly alter the interpretation of the time sections and for this reason was not economically justified.

**Concatenation of Final Stacks and Migrations onto tape**

The final migrations and final stacks without the time varying filter were concatenated in SEG Y form onto magnetic tape. The final stacks had no TVF in order that, at some future date, someone could use the SKS package to do additional or alternative migration.

**Additional Comments On Processing**

1. The most serious and certainly time consuming problem encountered by GECO was one of editing bad records. However, the final edited sections were of high quality.

2. It was noticed that on parts of the survey (particularly on southern end of line 1) overshooting was a problem. Overshooting occurs where the response from the previous shot overlaps with the desired response from the present shot. In places at around 9 s TWT (a second after basement) the whole section repeated itself. This was presumably a multiple from the previous shot or shots. There is no simple way of removing this effect - the only (inexpensive) way to reduce it is to use a relatively harsh time-varying filter.

3. With a fold of coverage of only 12 it was important not to carry out any processing step which might involve the reduction of the signal to noise ratio. For this reason no near trace mute was applied. Where the fold of coverage is large a near trace mute can be usefully applied to attenuate multiples.

4. A test was undertaken to see if the application of a pre-migrate two-dimensional dip filter improved the quality of the final migration. Unsurprisingly, it was found that the application of any dip filter had a disastrous effect on the continuity of sub-basement dipping reflectors. In the final processing sequence no dip-filter of any kind was applied.
Appendix B - Derivation Of Shortening From Long Wavelength Undulations

Assume undulations can be approximated by a cosine curve (Figure 5.9A) of amplitude A and wavelength \( \lambda \).

\[
z = ACos\left(\frac{2\pi x}{\lambda}\right)
\]

Line Increment (Figure 5.9A) \( dl = (dx^2 + dz^2)^{0.5} \) \( \text{ where } \)

\[
dz = \left(\frac{A2\pi}{\lambda}\right)Sin\left(\frac{2\pi x}{\lambda}\right)dx
\]

\[
dl = \left(\frac{(A2\pi)^2Sin^2\left(\frac{2\pi x}{\lambda}\right)}{\lambda}\right)dx^2 + dx^2)^{0.5}
\]

Simplifying 4

\[
dl = (BSin^2kx + 1)^{0.5}dx
\]

where \( B = \left(\frac{A2\pi}{\lambda}\right)^2 \); \( k = \frac{2\pi}{\lambda} \)

Find the horizontal shortening over length of the curve by evaluating the line integral

\[
I = \int_{0}^{F} (BSin^2kx + 1)^{0.5}dx
\]

where \( F \) is the length of the fold train. From the Binomial theorem

\[
(1 + BSin^2kx)^{0.5} = 1 + BSin^2kx - \frac{B^2Sin^4kx}{2} + \frac{B^3Sin^6kx}{8} + \frac{B^4Sin^8kx}{16} + ....
\]
Solve 6 by approximating to Binomial theorem and solving integrals for each term separately:

\[ \int \sin^2kx\,dx = \frac{(\sin kx \cos kx - kx)}{2k} \quad 8a \]

\[ \int \sin^4kx\,dx = -\frac{(2\sin^3kx \cos kx + 3\sin kx \cos kx - 3kx)}{8k} \quad 8b \]

\[ \int \sin^6kx\,dx = -\frac{(8\sin^5kx \cos kx + 10\sin^3kx \cos kx + 15\sin kx \cos kx - 15kx)}{48k} \quad 8c \]

Combining 8a, b, c with 6 and integrating between \( F \) and zero many terms become zero and the result is a series:

\[ I = F + BF - 3B^2F + 15B^3F \quad \frac{4}{64} \quad \frac{768}{} \quad 9 \]

Substituting for \( B = A \frac{2\pi^2}{\lambda^2} \)

\[ I = F + A \frac{2\pi^2F}{\lambda^2} - 3A^4 \frac{\pi^4F}{4\lambda^4} + 5A^6 \frac{\pi^6F}{4\lambda^6} \quad \frac{}{} \quad 10 \]

Which for \( \lambda \gg A \) can be approximated to

\[ I = F + A \frac{2\pi^2F}{\lambda^2} \quad 11 \]

The line integral \( I \) can be thought of as the restored length. Using a fold length of 1500 km (the total N - S spatial coverage of the deformation), an average wavelength of 200 km and an average amplitude of 1.5 km yields a restored length of 1500.83 km. A shortening of only 0.83 km over 1500 km corresponds to a percentage shortening of 0.05%. If the deformation is assumed to have begun at 7 Ma this means a shortening rate of 0.12 mm yr\(^{-1}\). Even for extreme values (a wavelength of 150 km and an amplitude of 2 km) this only yields a spatial shortening of 2.6 km, a percentage shortening of 0.17% and a shortening rate of 0.37 mm yr\(^{-1}\). This result is discussed in Section 5.9.
Appendix C - Two-dimensional Elastic Plate Bending Theory

A load $V_0$ is applied to a thin elastic plate of length $L$ and thickness $h$ which is pinned at both ends (Figure 6.1). The load produces two restoring forces $V_0/2$. Following the method of Turcotte and Schubert (1982), when a horizontal force $P$ is applied to the plate, the deflection $w$ of the plate can be expressed as a fourth order differential equation

$$\frac{Dd^4w}{dx^2} + \frac{Pd^2w}{dx^2} = 0 \quad (1)$$

providing that $h<<L$ and $w<<L$ and where the flexural rigidity is given by:

$$D = \frac{Eh^3}{12(1-\mu^2)} \quad (2)$$

For explanation of symbols and values of constants used see Table 6.1. At present I does not include a term for the hydrostatic restoring forces. The net force per unit area acting on the elastic plate is (see Figure 6.1)

$$q = q_0 - (\rho_m - \rho_w)gw \quad (3)$$

Therefore the full equation for the deflection of the elastic oceanic lithosphere becomes

$$\frac{Dd^4w}{dx^2} + \frac{Pd^2w}{dx^2} + (\rho_m - \rho_w)gw = q_0(x) \quad (4)$$

Setting $q_0 = 0$, the stability of the earth's lithosphere under a horizontal line load can be investigated. The solution of 4 can be assumed to be of form

$$w = w_0 \sin \left( \frac{2\pi x}{\lambda} \right) \quad (5)$$

and substituting gives

$$\frac{D(2\pi)^4}{\lambda} - \frac{P(2\pi)^2}{\lambda} + (\rho_m - \rho_w)g = 0 \quad (6)$$
This quadratic equation has solution

\[
\frac{(2\pi)^2}{\lambda} = P + \frac{P^2 - 4(p_m - p_w)gD}{2D} \quad 7
\]

Since the wavelength of the deformed lithosphere must be real there can only be a solution if \( P \) exceeds a certain value:

\[
P_c = (4Dg(p_m - p_w))^{0.5} \quad 8
\]

If \( P > P_c \) buckling occurs. Substituting for \( D \) from 2

\[
P_c = \left[ \frac{Eh^3(p_m - p_w)g}{3(1 - \nu^2)} \right]^{0.5} = \sigma_c h \quad 9
\]

Where \( \sigma_c \) is the critical stress associated with force \( P_c \)

\[
\sigma_c = \left[ \frac{Eh(p_m - p_w)g}{3(1 - \nu^2)} \right]^{0.5} \quad 10
\]

It is now possible to solve 7 for \( P = P_c \)

\[
\lambda_c = \frac{2\pi(2D)^{0.5}}{P_c} \quad 11
\]

and substituting

\[
\lambda_c = \frac{2\pi(\frac{Eh^3}{12(1 - \nu^2)(p_m - p_w)g})^{0.25}}{12(1 - \nu^2)(p_m - p_w)g} \quad 12
\]

Equations 10 and 12 are used and discussed in Section 6.2.2.1.
Appendix D - Analogue Modelling Experimental Details and Results

All buckling wavelengths are ± 30 km and are rescaled assuming an oceanic lithosphere brittle thickness of 30km. The first part of this Appendix concerns experimental details while the second part documents examples of laser profiles and spectral analyses not shown in Chapter 6 (e.g. Figure 6.13 and 6.14).

Experiment 1
Thicknes Brittle Layer (Sand) - 2 cm  Motor Speed - 1 cm hr\(^{-1}\)
Thicknes Ductile Layer (Silicone) - 1.2 cm  Box Length - 70 cm
Asthenosphere Analogue - Heavy Water  Box Width - 46 cm
Laser? - No  Total Shortening 0/o - 5.7
Comments - Buckling developed with wavelength of ~ 200km. Difficult to estimate without laser. Problem with water rising up the side of the model.

Experiment 2
Thicknes Brittle Layer (Sand) - 2 cm  Motor Speed - 11 cm hr\(^{-1}\)
Thicknes Ductile Layer (Silicone) - 1.2 cm  Box Length - 70 cm
Asthenosphere Analogue - Honey  Box Width - 46 cm
Laser? - Yes  Total Shortening 0/o - 5.2
Comments - Buckling developed with wavelength of 210 km. Good experiment.

Experiment 3
Thicknes Brittle Layer (Sand) - 2 cm  Motor Speed - 23 cm hr\(^{-1}\)
Thicknes Ductile Layer (Silicone) - 1.2 cm  Box Length - 71 cm
Asthenosphere Analogue - Honey  Box Width - 46 cm
Laser? - Yes  Total Shortening 0/o - 10.7
Comments - Buckling developed with wavelength of 190 km. Good experiment.

Experiment 4
Thicknes Brittle Layer (Sand) - 2 cm  Motor Speed - 1 cm hr\(^{-1}\)
Thicknes Ductile Layer (Silicone) - 1.2 cm  Box Length - 66.5 cm
Asthenosphere Analogue - Honey  Box Width - 46 cm
Laser? - Yes  Total Shortening 0/o - 3.1
Comments - Buckling for small amounts of shortening with wavelength of 200 km and then rapid development of faulting.
Experiment 5
Thickness Brittle Layer (Sand) - 2 cm  Motor Speed - 20 cm hr⁻¹
Thickness Ductile Layer (Silicone) - 1.2 cm  Box Length - 41.5 cm
Asthenosphere Analogue - Honey  Box Width - 20 cm
Laser? - Yes  Total Shortening 0% - 16.1
Comments - Using this small box good buckling of wavelength of 190 km. Towards the end reverse faults developed.

Experiment 6
Thickness Brittle Layer (Sand) - 1.33 cm  Motor Speed - 7 cm hr⁻¹
Thickness Ductile Layer (Silicone) - 0.8 cm  Box Length - 66 cm
Asthenosphere Analogue - Honey  Box Width - 46 cm
Laser? - Yes  Total Shortening 0% - 5.3
Comments - Buckling of wavelength 200 km. Possible that initial perturbations due to problems with the motor may have been preferentially amplified.

Experiment 7
Thickness Brittle Layer (Sand) - 2 cm  Motor Speed - 9 cm hr⁻¹
Thickness Ductile Layer (Silicone) - 1.2 cm  Box Length - 52 cm
Asthenosphere Analogue - Honey  Box Width - 46 cm
Laser? - Yes  Total Shortening 0% - 5.8
Comments - Buckling wavelength of 220 km. Good experiment

Experiment 8
Thickness Brittle Layer (Sand) - 1 cm  Motor Speed - 9 cm hr⁻¹
Thickness Ductile Layer (Silicone) - 0.6 cm  Box Length - 53 cm
Asthenosphere Analogue - Honey  Box Width - 46 cm
Laser? - Yes  Total Shortening 0% - 1.0
Comments - Buckling wavelength of 240 km. Good experiment.
Figure D.1 Corrected topography and spectral analysis of experiment 2 for 1.9 % shortening (profile A). The spectral analysis suggests a dominant wavelength of folding equal to 210 km. This is not obvious from the laser profile, but was confirmed by analysing photographs taken during the experiment. Note that for this experiment and experiment 3, the laser software was not fully developed
Figure D.2 Corrected topography and spectral analysis of experiment 3 for 4.1% percent shortening (profile H). The spectral analysis suggests a dominant wavelength of folding equal to 190 km. This is not immediately obvious from the laser profile, but was confirmed by analysing photographs taken during the experiment. Note that for this experiment and experiment 2, the laser software was not fully developed.
Figure D.3 Corrected topography and spectral analysis of experiment 4 for 1.4 % shortening (profile E). A single dominant wavelength of folding is present equal to 200 km. Four wavelengths of folding are present. The prominent peak marked by $\lambda_s$, and the sharp topography, is due to the laser inadvertently recording the plastic grid that was left on the top of the model to aid photographic analysis. The 4cm experimental grid with a scaling of $1\text{cm} = 15 \text{km}$, corresponds to the prominent peak of wavelength $\sim 60 \text{km}$. The second sharp peak, $\lambda_{hc}$ is a harmonic of $\lambda_s$. The folding wavelength was confirmed by analysis of the photographs.
Figure D.4 Corrected topography and spectral analysis of experiment 5 for 2.4 % percent shortening (profile B). A single dominant wavelength of folding is present equal to 190 km. Two to three wavelengths of folding are developed in this experiment which was undertaken in a small box. Photographs confirmed the wavelength of folding. Reverse faulting developed very quickly after this profile was recorded.
Figure D.5 Corrected topography and spectral analysis of experiment 6 for 1.3% shortening (profile C). A problem with the motor (producing extension rather than compression) at the beginning of the experiment may have contributed to the generation of two peaks in the spectral analysis. The longer wavelength peak $\lambda_s$ was interpreted as being due to the experimental problems, while the shorter wavelength peak $\lambda_c$, equal to 200 km, was interpreted as being due to folding.
Figure D.6 Corrected topography and spectral analysis of experiment 9, utilizing a four layer continental lithology, for 17.4 % percent shortening (profile E). Note that the topography and spectral analyses are more complex than for the oceanic lithosphere rheology.
Appendix E - Numerical Modelling Experiments

This Appendix contains figures of numerical models attempted to model the long wavelength undulations. Each model is annotated to show the parameters used. This failure is highlighted by the unlikely or non-existent topography generated in many cases.
Figure E.1 Initial Finite Element model comprising a 5 km thick elastic layer overlying a 70 km thick visco-elastic layer (see Figure 6.16A for the design of the elemental grid). A pressure of $0.6 \times 10^9$ Pa was applied to the left hand margin of the elastic layer and the model was run for 10 iterations each of length $0.5 \times 10^5$ years, with a viscosity of the visco-elastic layer of $0.1 \times 10^{24}$ Pa s and a Young's modulus of $0.18 \times 10^{12}$ Pa. A Young's modulus of $0.65 \times 10^{11}$ Pa s was used for the elastic layer. A density contrast of $1000$ kg m$^{-3}$ was used in the isostatic restoring force applied along the top of the model while a value of 0.27 was used for Poisson's ratio throughout. A sinusoidal perturbing force was applied as discussed in the Section 6.4. The model was constrained so that no horizontal motion was permitted along the right-hand side of the model.
Figure E.2 Model with the same parameters as in Figure E.1 except that 32 iterations were used.
Figure E.3 Model with the same parameters as for Figure E.1 except that 100 iterations were used.
Figure E.4 Model with the same parameters as for Figure E.1 except that 300 iterations were used.
Figure E.5 Model with the same parameters as for Figure E.3 except that a pressure of $0.4 \times 10^8$ Pa was applied along the entire left hand side of the model.
Figure E.6 Model with the same parameters as E.3 except that a larger pressure $0.6 \times 10^{10}$ Pa was applied to the left hand side of the elastic layer.
Figure E.7 Model comprising a 5 km thick layer perturbed by the sinusoidal force as in Figure E.1. The elemental mesh is shown in Figure 6.16B. A compressive pressure of $0.11 \times 10^{10}$ Pa was applied to the left hand side of the model. All other parameters as for Figure E.1.
Figure E.8 Model with the same parameters as used in Figure E.7 except that a larger compressive force per unit length of $0.11 \times 10^{11}$ Pa was applied to the left hand side of the model.
Appendix F - SEASAT coverage

Figure F.1 SEASAT data points used in producing Free-Air Anomaly contour maps illustrated in Figure 7.1 and 7.2.
FIG. 2 A short section of fully processed and migrated 12-channel, 12-fold multichannel profile from north (left) to south (right) through the ODP Leg 116 Sites and perpendicular to the structural grain (located A in Fig. 1). Vertical exaggeration is about 3:1 in the sediments and 15:1 to 2:1 in basement (the top of which is marked by the strong reflector at ~8-s two-way travel time). Details of the high-angle fault planes in the sediments are evident: some anastomose and show curvature, some are just developing and some are mature. The parallelism of basal sediments and basement suggests basement was originally quite planar and has since been displaced during the intraplate deformation. The fault planes on which the displacement has taken place dip at 30°-40° in basement and can be traced to depths at ~10-s two-way travel time (about Moho depth). The multichannel profiles are nearly perpendicular to the strike of the faults and therefore this apparent dip should be close to the true dip. Originally these faults may have been formed at the mid-ocean ridge to the south in which case they must have been outward-dipping faults.

result of the reactivation with reverse movement of two sets of pre-existing spreading-centre-formed normal faults. One of these sets of basement faults dips to the north and therefore faces outwards from the original spreading centre, whereas the other set dips to the south and is the more commonly envisaged inward-dipping set. We have imaged the outward-dipping set with the multichannel seismic profiles, dipping at 35°-40° throughout the oceanic crust. The inward-facing set, dipping to the south, which is not resolved in the multichannel data, we infer above is likely to have a higher dip in basement.

FIG. 3 Plan views of reverse faulting in two areas of the intraplate study area. Top left illustrates the relative positions of the two areas: area A around the ODP Leg 116 sites and area B around 81°E 5°S. The position and orientation of two fracture zones (FZ) developed in the area are also shown. In the detailed maps of area A (top right) and area B (bottom) reverse faults are indicated by dashed lines with a black triangle pointing in the direction of the hanging wall. Darwin tracks are continuous lines; in area A the dotted lines are other profiles collected during ODP site surveys and used to constrain the positions of the faults. The position of Fig. 2 is also indicated in area A. In area B the dashed lines with the black diamond ornament show the strike of characteristic hanging-wall anticlines which could be used to tie between lines. It should be noted that, for clarity, not every fault has been annotated in area B: those shown are only the larger ones. In both areas the strike of the reverse faults is 90°E-100°E, roughly perpendicular to the strike of fracture zones (005°E-010°E) developed nearby. The general short fault length (usually <10 km) and resultant en echelon pattern displayed in area B is similar to the fabric developed close to active spreading centres and is consistent with the reverse faults resulting from the reactivation of the pre-existing spreading-centre-formed fabric.

An important tectonic implication of these observations is the obvious brittle behaviour of the oceanic lithosphere. The resolution of faults penetrating to at least 10-s two-way travel time indicates that the whole crust and possibly the uppermost mantle is deforming under intraplate compression in a brittle manner. Studies of mid-ocean ridge earthquakes have revealed that normal faulting within the median valley occurs on planes that dip at ~45°. Using the assumption that the centroid depth marks
Fault reactivation in the central Indian Ocean and the rheology of oceanic lithosphere

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The intraplate deformation in the central Indian Ocean basin is a well-known example of a deviation from an axiom of plate tectonics: that of rigid plates with deformation concentrated at plate boundaries. Here we present multichannel seismic reflection profiles which show that high-angle reverse faults in the sediments of the central Indian Ocean extend through the crust and possibly into the uppermost mantle. The dip of these faults, which we believe result from the reactivation of pre-existing faults formed at the spreading centre, is ~40° in the basement, which is consistent with the distribution and focal mechanisms of earthquakes on faults now forming at spreading centres. This style of deformation, coupled with the observation of large earthquakes in the mantle lithosphere, indicates that brittle failure of the oceanic lithosphere may nucleate in the vicinity of the brittle/ductile transition and propagate through the crust.

The intraplate deformation occurs in lithosphere of age 60–90 Myr that formed at a fast spreading rate (6 cm yr⁻¹) at the southeast Indian Ocean ridge. It manifests itself as a diffuse zone of compressional and strike-slip earthquakes, high localized heat-flow, geoid anomalies and tectonic deformation. The tectonic deformation can be considered to be occurring on two spatial scales: the first order is represented by long-wavelength (100–300 km), large-amplitude (1–2 km) undulations of oceanic basement and overlying sediments; and the second order by unusual high-angle reverse faults and associated folds in the basement cover. There has been much speculation as to whether the first order of deformation represents buckling, inverse boudinage or some form of block faulting. The high-angle reverse faults have been less well studied and are the subject of this paper.

The high-angle faults were first documented by Eittreim and Ewing and then further studied by Weissel et al. and Geller et al. They found that they offset both basement and overlying sediments, dipped at angles >65° in the sedimentary cover and had a strike of 100° E. Weissel et al. suggested that they arose from the reactivation of the pre-existing structural fabric in basement. Recently, results from ODP Leg 116 (ref. 9) have shown that this reactivation started 7 Myr BP and Curray and Munasinghe have shown that the onset of deformation can be correlated over a wide area. Here we discuss results from RRS Charles Darwin Cruise 28, during which the first multichannel seismic reflection profiles were collected over the intraplate deformation (Fig. 1). These results not only confirm these previous observations but add important new findings about the character, origin and mode of reactivation of the faults.

The Darwin seismic reflection profiles clearly show that the vast majority of the high-angle faults are reverse in nature and few are normal, a conclusion which could not be made from the profiles used by Weissel et al. The reverse faults can be divided into those downthrowing to the north, and those downthrowing to the south. In the survey area approximately equal numbers of faults downthrow in each direction. The new multichannel profiles image (Fig. 2) the faults in basement that dip to the north, their dip being ~35–40° at this depth. In some instances the faults are resolved down to 10-s two-way travel time, roughly the expected level of the oceanic Moho. The faults, which appear slightly listric on the time section, are approximately planar after depth conversion. There are additional north-dipping events at lower angles that do not intersect the basement/cover interface that could be other, little reacti-
the mean depth of fault slip, it seems that faulting extends from 2 to 10 km below the sea floor for these earthquakes—into the upper mantle. Further evidence for the whole crustal extent of spreading-centre-formed faults is found in the western North Atlantic where White et al., image eastward-dipping reflectors which they interpret as representing inward-dipping normal faults. It should be noted, however, that these studies of brittle failure have been carried out on crust formed at slow spreading rates; little analogous earthquake work has been carried out on fast-spreading ridges, like the East Pacific Rise, similar to the one that produced the central Indian Ocean basin lithosphere.

Nevertheless, it is likely that there is a decrease in the thickness of the brittle layer at fast-spreading ridges in view of the evidence from slow-spreading regions that, in general, brittle layer thickness decreases with increasing spreading rate. For large earthquakes the brittle layer may be as little as 3 km thick when spreading rates of 2.5 cm yr$^{-1}$ are reached. Thus it would seem that faulting at the fast-spreading ridge in the Indian Ocean may not have originally penetrated down to Moho and uppermost mantle depths. The deepest parts of the fault planes seen on the multichannel seismic profiles data would then have been formed during reactivation.

Despite the clear evidence for faulting there is a lack of recorded earthquakes associated with the crustal deformation in the central Indian Ocean basin. This could be due to the brittle strength of the oceanic crust and uppermost mantle being too low to sustain significant stresses resulting in only low-magnitude seismicity that is below the level of detection on the worldwide seismograph network. The focal depths of large-magnitude earthquakes recorded in this area are concentrated at depths between 27-39 km (ref. 19), extending down to the expected depth of the brittle/ductile transition. It is interesting that this may be compared with the pattern of continental seismicity in the seismogenic upper crust (see ref. 20, for example) with the largest earthquakes nucleating at or near the base of the brittle layer. By analogy, recent brittle deformation in the central Indian Ocean basin may nucleate at upper-mantle depths and propagate into and through the crust, in the upper part by the reactivation of pre-existing spreading-centre-formed faults.

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Reduced original record for the digitised profile shown in Figure 3.2. For digitising purposes the full-sized record was used to discriminate between reflectors. Annotated are the seafloor (R1), top of basement (B1), refracted arrivals (Re) and the direct wave (Dw).
Enclosure 3 - Example of 3.5kHz record

This 3.5 kHz shows an active reverse fault breaking through the sea-floor. The hanging wall of the fault rises 400 metres from the undisturbed sea-floor. The position of this fault is shown as fault S in Figure 4.8. While this thesis has not studied the 3.5 kHz records, it is planned to use them in future to study in detail the relationships between sedimentation and tectonics.