Marine Geophysical Study of the Eurasian-African Plate Boundary in the Vicinity of Gorringe Bank

by

Nathan Hayward

Thesis submitted for the Degree of Doctor of Philosophy to the University of Oxford

Exeter College & The Department of Earth Sciences

Trinity Term September 1996
For my family and friends....
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Abstract: The Gorringe Bank region is located at the eastern end of the
Azores-Gibraltar plate boundary, which plate kinematic studies show to progressively change from extension at the Azores, through pure right lateral strike slip at the Gloria fault to compression at Gibraltar. The region is dominated by high relief (4-5 km), highly deformed, uncompensated, ENE-WSW trending seamounts and intervening abyssal plains with basin sediment thicknesses in excess of 4 km and minimal surface deformation.

Gorringe Bank, which was formed by overthrusting of the African plate upon the Eurasian plate at about 10 Ma along the plate boundary, is supported in part by flexure of the Eurasian plate, as indicated by pre-loading sediments and basement to the north which are tilted towards Gorringe Bank. Broken plate models show the Eurasian plate to have an elastic thickness of about 35 km which is in agreement with that expected for the crustal age (130-135 Ma) at the time of loading.

Coral Patch Ridge was formed by a combination of thrust faulting and whole crustal buckling resulting from the past 20 Ma compression and was partially uplifted before deposition of an olistostrome in the Middle Miocene.

Recent compressional deformation is distributed over a wide region, as indicated by the dispersed shallow seismicity and has a trend which rotates from approximately N45°E to N70°E from west to east across the region, near perpendicular to westward verging plate motion vectors. The majority of extensional and strike-slip deformation is explained by a regional strike-slip strain ellipse model, including an antithetic NNE-SSW strike slip fault between Gettysburg and Hirondelle seamounts which marks the boundary between the Eastern and Western Horseshoe Basins.

Isostatic models for the Madeira-Tore Rise, which initially formed at the Mid Atlantic Ridge, give an elastic thickness of approximately 15 km indicating that significant material was added to the Rise as it moved away from the Ridge.
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Extended Abstract: The structure of the Gorringe Bank region, located at the eastern end of the Azores-Gibraltar plate boundary, is a consequence of complex interactions between the Eurasian, Iberian and African plates during North Atlantic Ocean evolution. During early rifting in the Jurassic, uniformly magnetised oceanic crust produced the magnetic smooth zone observed east of 12°W in the Gorringe Bank region. This smooth zone precludes recognition of the ocean-continent transition zone and prevents accurate crustal dating prior to the first clearly identified magnetic chron M-25 west of the Madeira-Tore Rise.

From about M-4 to M-0 (126-119 Ma) an increased magma flux, moving south along the Mid Atlantic Ridge, possibly sourced by a mantle plume, resulted in formation of the Madeira-Tore Rise which bounds the Gorringe Bank region to the west. Isostatic gravity models reveal that the elastic thickness of oceanic crust in this region is about 15 km, translating to a crustal age of approximately 20 Ma at the time of Madeira-Tore Rise formation and indicating the addition of material during movement away from the Mid Atlantic Ridge. Miocene volcanic rocks from Ampere and Josephine seamounts and 60 Ma volcanic material recovered from Gorringe Bank indicate intermittent addition of material to seamounts of region, including the Madeira-Tore Rise.

Differential motion between the Eurasian, Iberian and African plates, during Eastern Atlantic evolution, created lines of weakness which subsequently influenced the creation of structures under a changing tectonic regime. Plate boundaries existed between the Iberian and African plates from the late Jurassic (M-25, 156 Ma) until approximately M-34 (84 Ma) when left lateral motion between the African and Eurasian plates ceased. Right lateral motion between Iberia and Africa commenced at about chron 19 (44 Ma) and had stabilised on the Azores-Gibraltar plate boundary between Eurasia and Africa by chron 6 (20 Ma). Plate kinematic studies show the current plate boundary to change progressively from a small component of extension on the Terceira Ridge, through pure right lateral strike slip at the Gloria Fault, to compression with a small component of right lateral strike slip in the Gorringe Bank region.
The Gorringe Bank region is characterised by ENE-WSW trending, 4-5 km seamounts and intervening abyssal plains. The seamounts, including Gorringe Bank, Coral Patch seamount and Coral Patch Ridge, which isostatic gravity models show to be uncompensated, were formed due to compressive motion of the past approximately 20 Ma which exploited crustal weaknesses and previous structures.

Gorringe Bank, from which upper and lower crustal rocks have been recovered, was formed by northward overthrusting of the African plate upon the Eurasian plate in the Middle Miocene at approximately 10 Ma, causing flexure of the Eurasian plate in the Tagus abyssal plain. Flexure resulted in the dip of the block faulted basement and pre-loading sediments towards Gorringe Bank with possible under-thrusting of pre-loading sediments trapped between basement block faults and an associated linear free-air gravity anomaly low parallel to the northwestern flank. The post loading or infill sediments which have an age of about Middle Miocene to Pleistocene show fanning reflections in the Tagus abyssal plain. This indicates that subsidence and flexure of Eurasian plate have not ceased, with a planar seafloor resulting from a sediment supply greater than the subsidence.

Broken plate flexure gravity models reveal that Gorringe Bank is supported in part by flexure of the Eurasian plate which has an elastic thickness of approximately 35 km. This is in agreement with that expected for the age of the crust at the time of loading (130-135 Ma). The whole of Gorringe Bank cannot be supported by the Eurasian plate, with models showing a well constrained fit to NNW half of the Bank supported by the Eurasian plate, the SSE must be supported by a regional stress field due to regional compression.

Major deformation moved south of the Horseshoe abyssal plain, concentrated on the northern flank of Coral Patch seamount and the southern edge of the Eastern Horseshoe abyssal plain where current convergence is preferentially accommodated. Compression resulted in large scale thrust faulting and folding of the sediments and basement rocks on Coral Patch Ridge and the formation of large folded ridges on Coral Patch seamount.

Gravity inverse and forward models show an elevated Moho beneath Coral Patch Ridge, as tentatively interpreted from seismic reflection profiles, which indicates a combination of thrusting and crustal buckling with intense sediment deformation formed as a result of the past 20 Ma compression, but partially uplifted before deposition of the olistostrome in the Middle Miocene. Combined thrusting and crustal buckling of a similar nature is observed in the central Indian Ocean in the region of the Afranasy-Nikitin seamounts, where deformation was focussed by pre-existing crustal folds caused by fold trough filling sediments and the load of the...
seamounts. Gorringe Bank and the great sediments thickness in the Horseshoe Basin may have similarly focussed deformation associated with Coral Patch Ridge.

Recent deformation in the Gorringe Bank region is widespread as implied by the shallow and apparently diffuse seismicity. The seismic trend can however be tied to the seaward extension of continental crustal fractures and as with the seismicity, much of the surface deformation observed primarily on GLORIA and seismic reflection profiles, can be explained on the basis of a regional model.

Although observed across the whole Gorringe region, recent deformation is concentrated on bathymetric highs, with minimal surface deformation in the abyssal plains. Folds and faults exhibit clockwise rotation from a trend of about N45°E at 15°W in the region of Josephine seamount to N70°E at 10°W in the region of Coral Patch Ridge. This rotation is in accordance with the variation of slip vector azimuths verging to an E-W trend towards the west.

Alignment of the direction of maximum principal compressive stress in a dextral strike-slip ellipse model near parallel to the plate motion vectors shows that the majority of these recent compressional features can be explained in terms of compression in a regional dextral strike slip zone. Extensional features observed on Josephine seamount are near perpendicular to the direction of minimum principal compressive stress $\sigma_3$ and the strike-slip fault between Gettysburg and Hirondelle seamounts is an antithetic $R_2$ Reidel shear which agrees with motion interpreted from a focal mechanism.

The dispersed shallow seismicity and focal mechanisms from the Gorringe Bank region also tie with the ellipse model, with the majority of strike-slip focal mechanisms aligned near to the $R_1$ Reidel shear and the strike of the thrust faulting mechanism of perpendicular to $\sigma_1$. However, a few focal mechanisms in the Eastern Horseshoe abyssal plain show that faulting is complex and variable in this area.

Sediments in the Gorringe Bank region show great thickness variation with thinner drapes on seamounts and substantial thicknesses in the Tagus, Seine and Western and Eastern Horseshoe Basins. These basins are divided by bathymetric highs except for the Eastern and Western Horseshoe Basins, which although currently represented by a single basin, were once separated by a basement shallowing and deformation associated with NNE-SSW faulting between Gettysburg and Hirondelle seamounts.

Sediments of the Tagus Basin have a thickness of up to 2.9 km (basement depth 8.6 km) which fill a linear trough across the base of the northern flank of Gorringe Bank. Thickness decrease is rapid towards Gorringe Bank and gradual towards the north due to flexure of the Eurasian plate under partial loading of Gorringe Bank.
To the west, upper units are draped over the flanks of Hirondelle seamount, but deeper units are partially confined by boundary faults.

The Eastern Horseshoe Basin has a primary depocentre to the east with up to 5 km of sediment (basement depth 10 km) which thins gradually towards the west and east. To the north sediments continue onto the southern flank of Gorringe Bank and to the south are offset by and show substantial thinning beyond the ENE-WSW striking thrust faults of Coral Patch Ridge. However, active basin boundary faulting truncates sediments to the southwest.

The Western Horseshoe Basin is comprised of two near circular sub basins forming a "dumbbell" shaped basin. The eastern sub basin has up to 4 km of sediment (basement depth 8.9 km) and is stretched slightly along a NNE-SSW trend related to faulting between Hirondelle and Gettysburg seamounts. The second sub basin has up to 2.8 km (basement depth 7.5 km) of sediment. The sediments thin onto faulted basement to the east, south and west. However, to the north and northeast they are constrained by classic steeply dipping basin bounding normal faults which have been active throughout basin formation.

Sediment thickness in the Seine Basin is poorly constrained, however, sediments rapidly attain a thickness of at least 2 km to the south of Ampere seamount and Coral Patch Ridge. Sediments drape and thin onto block faulted basement to the southwest of Coral Patch seamount and to ENE onto Coral Patch Ridge.

Basin isolation has resulted in diversity of reflection and sedimentary character and structure on seismic reflection profiles. Facies interpreted for individual basins have been tentatively correlated with the aid of the DSDP sites 120 and 135 to give four main sedimentary units and basement.

The youngest unit has an age from Pleistocene to Late Oligocene (Possibly Pleistocene to Middle Miocene in the Tagus Basin) and is characterised by medium-low amplitude, short wavelength, semi-continuous to continuous at depth parallel reflections. The unit forms a turbidite deposit, including a component of pelagic chalk ooze, with low angle basal onlap onto a regionally distinctive unconformity, often represented by a pair of continuous high amplitude reflections.

The second unit rests upon the major unconformity in the Tagus and Eastern Horseshoe Basins and is probably Middle Miocene in age. The unit is characterised by medium amplitude hummocky broken reflections with a high diffraction density. In the Tagus Basin the unit is restricted to the linear flexural moat to the northwest of Gorringe Bank, created by partial loading of Gorringe Bank upon the Eurasian plate and is interpreted as one or more debris flows derived from Gorringe Bank. In the Eastern Horseshoe Basin the unit represents an olistostrome derived from
the Straits of Gibraltar following formation of Gorringe Bank, as the unit is absent on its uplifted flanks.

The next unit has a probable age of Early Eocene to Early Aptian and is characterised by low amplitude, semi-continuous, parallel reflections which are slightly deformed and show low angle basal onlap onto fold highs.

The oldest unit of the Gorringe Bank region has a tentative age of Upper Jurassic to Early Aptian based on DSDP sites and estimates of crustal age, however, if the deepest deposits are remnants of prerift sediments they may predate the Jurassic. The unit fills basement troughs with onlap onto basement, faults and fold highs and may in part represent a synrift deposit with growth observed on a seismic reflection profile from the Western Horseshoe Basin.

The basement structure in the Gorringe Bank region is obscured by the great sediment thicknesses, but generally characterised by low-medium amplitude, broken hummocky reflections and a high diffraction density. Basement has influenced sedimentation, with decreasing effect over time, in most regions. Basement in the Western Horseshoe and Tagus basins shows a huge basement relief in excess of 1.5 seconds TWTT which has influenced sediment deposition. In contrast the Eastern Horseshoe Basin only shows a relief of about 1 second TWTT which appears to have had a lower effect on sedimentation than other basins. This may represent a different crustal nature as suggested by refraction models.
Acknowledgements

A time, now seemingly long ago, I first discovered geology shrouded in the background of a high school geography class. Were it not for David Lomax, I would never have chosen to further my study of the subject.

I was fortunate to have received much support at U.C.W. Aberystwyth and U.C. Santa Cruz. I am especially grateful to Robert Whittington, Bill Fitches, Alex Maltman and Eli Silver, whose influence inspired the continuation of my studies.

I wish to express my appreciation to Professor Tony Watts for the supervision he gave and to all at the Department of Earth Sciences, Oxford University, who assisted me throughout this project. My thanks go especially to Rupert Dalwood and Audrey Willet for their comradeship, Robin Owens for assistance with swath bathymetry, Martin Leese and Stephen Usher with computing, Tim Henstock with seismic processing, Geoff Peglar with seismicity and Marc Audet for mathematical guidance.

Tim LeBas, Doug Masson of the Institute of Oceanographic Sciences, I also thank for use of their GLORIA processing system.

This thesis was funded by the National Environmental Research Council whom I would like to thank for their support.

Special thanks go to my parents and friends who were always there.

Nathan H. September 1996
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### Glossary of terms and abbreviations

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<td>AF</td>
<td>African plate</td>
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<tr>
<td>CDP</td>
<td>Common depth point</td>
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<tr>
<td>COE</td>
<td>Cross-over error</td>
</tr>
<tr>
<td>DOBS</td>
<td>Digital ocean bottom seismometer</td>
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<tr>
<td>DSDP</td>
<td>Deep Sea Drilling Project</td>
</tr>
<tr>
<td>ECOE</td>
<td>External cross-over error</td>
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<tr>
<td>EU</td>
<td>Eurasian plate</td>
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<tr>
<td>FAA</td>
<td>Free-air gravity anomaly</td>
</tr>
<tr>
<td>GPS</td>
<td>Global positioning system</td>
</tr>
<tr>
<td>IB</td>
<td>Iberian plate</td>
</tr>
<tr>
<td>ICOE</td>
<td>Internal cross-over error</td>
</tr>
<tr>
<td>IGRF</td>
<td>International Geomagnetic Reference Field</td>
</tr>
<tr>
<td>Ma</td>
<td>Million years ago</td>
</tr>
<tr>
<td>NA</td>
<td>North American plate</td>
</tr>
<tr>
<td>OCT</td>
<td>Ocean continent transition</td>
</tr>
<tr>
<td>ODP</td>
<td>Ocean Drilling Program</td>
</tr>
<tr>
<td>PC</td>
<td>IBM compatible personal computer</td>
</tr>
<tr>
<td>SEG-Y</td>
<td>Trace sequential seismic recording standard (Barry <em>et al.</em>, 1975)</td>
</tr>
<tr>
<td>SCBA</td>
<td>Sediment corrected Bouguer gravity anomaly</td>
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<td>TWTT</td>
<td>Two-Way Travel Time</td>
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Chapter 1

Introduction

This thesis is concerned with the morphology, structure and formation of the Gorringe Bank region, located at the eastern end of the Azores-Gibraltar plate boundary in the Eastern Atlantic Ocean.

A new survey of the Gorringe Bank region and Alboran sea by the 64th cruise of the R.R.S. Charles Darwin during December, 1991 and January 1992 acquired in excess of nine thousand kilometres of marine geophysical data. The combination of these with previous geophysical data, which failed to produce a detailed image of the regional structure, provide a new base for interpretation of the tectonics of the region.

1.1 Aims of the study

There were two principal aims of this project. The first was to process and display all marine geophysical data acquired during R.R.S. Charles Darwin cruise 64 (CD64) in the Gorringe Bank region. This includes in excess of four thousand kilometres of seismic reflection, GLORIA mark II, 3.5 and 12 kHz bathymetry, gravity and magnetic data.

The second was to use these in collaboration with other available data, including underway geophysical data, published seismic reflection profiles, seismicity and sampling, to interpret the tectonic structure of the Gorringe Bank region. Focus is on investigation and modelling of the mechanism of formation seamounts including Gorringe Bank and interpretation of the tectonic and sedimentary structure of the region.
1.2 Data

Marine geophysical data for this study were acquired during R.R.S. Charles Darwin cruises CD64 and CD82 in 1991 and 1993, with additional seismic reflection data from the Iberian Atlantic Margin Project (Banda & Torne, 1995). During CD64 3.5 kHz and 12 kHz bathymetry, magnetic, gravity, seismic reflection and GLORIA Mark II sidescan sonar data were acquired for the Azores-Gibraltar plate boundary, Gorringe Bank region and Alboran Sea. Over four thousand kilometres of data collected to the west of Gibraltar (8°W), were used in this study, with data to the east of Gibraltar analysed in the study of Willet (1996). CD82 was primarily concerned with a geophysical survey of the Canary Islands (Dalwood, 1996). During transit to the Canaries a single line of gravity, magnetic, 3.5 kHz and 12 kHz bathymetry data and high resolution EM12 Simrad swath bathymetry were collected across the Gorringe Bank region. Additional gravity and bathymetry data were provided by the Generic Mapping Tools (G.M.T.) database (Wessel & Smith, 1991) which contains data for currently 2456 cruises worldwide from the years 1953 to 1993. Additional magnetic data was provided by the National Geophysical Data Center (N.G.D.C.) database. Data from previous studies is reviewed in chapter 2.

1.3 Thesis Outline

- **Chapter 1**: Introduction to the study, including the aims and data used.

- **Chapter 2**: An introduction to the morphology and structure of the Gorringe Bank region, including relevant previous research and data used in this study. A review of the seismicity, plate reconstructions for the Northern Atlantic Ocean and marine geophysical data is given.

- **Chapter 3**: is a description of the acquisition and processing the marine geophysical data from CD64 and 82 from the Gorringe Bank region, including gravity, magnetic, bathymetry, GLORIA Mark II and seismic reflection data.

- **Chapter 4**: Interpretation of the sediment and basin structure, primarily from seismic reflection data. Seismic facies are described and correlation made between basins and with Deep Sea Drilling Project data. Depth conversion of seismic reflection data reveal the sediment thickness variation and basement depth.

- **Chapter 5**: describes the interpretation of deformation from GLORIA Mark II and seismic reflection data. Structures are analysed with reference to the
tectonic framework, current plate kinematics and seismicity.

- Chapter 6: presents an interpretation of gravity and magnetic anomalies and models, including isostatic, inverse and forward gravity models of Coral Patch Ridge and flexural forward gravity models for Gorringe Bank.

- Chapter 7: Conclusions with a summary of the main results in relation to the tectonic evolution of the Gorringe Bank region from initial Atlantic Ocean rifting to the present day.
Chapter 2

Geological and Geophysical Setting

2.1 Morphology of the Gorringe Bank Region

The Gorringe Bank region is located in the Northeastern Atlantic Ocean off the coasts of Iberia and Africa (figure 2.1) and is dominated by a NNE-SSW rise to the west with two eastward stretching arms of ENE-WSW trending seamounts with intervening abyssal plains (figure 2.2).

At the eastern end of the northern seamount arm lies Gorringe Bank, the most studied feature of the Gorringe Bank region, which is a ridge comprising Ormonde and Gettysburg seamounts. A high resolution bathymetric image (figure 2.3) from EM12 Simrad (Beuzart et al., 1979) shows more accurately the surface structure of Gorringe Bank, including channel erosion features (LaGabrielle & Auzende, 1982).

Gettysburg seamount is slightly shallower than Ormonde, the whole of Gorringe Bank rising from abyssal depths of approximately 5000 metres to within 25 metres of sea-level. The Bank is 180 km long on a trend of approximately N50°E and 80 km wide at the base with steeper slopes on the northwestern flank than the southeastern flank.

The location of the current plate boundary separating the Eurasian and African plates is unknown within the Gorringe region, but to the west, stretching to the Azores, is represented by the Gloria Fault (figure 2.1). LePichon et al. (1970) proposed that the plate boundary in the Gorringe region was located at Gorringe Bank between 5-10 Ma and that the asymmetry of the bank resulted from formation by overthrusting of the African plate upon the Eurasian plate at that time.

This interpretation was advanced by Purdy (1974) who completed the most comprehensive study of the region to date and concluded that Gorringe Bank formed as a result of ”Flake Tectonics” (Oxburgh, 1972). Slivers of crust were
overthrust onto the Eurasian plate (figure 2.4), with the majority of the African plate forming the nascent subduction zone beneath the Horseshoe abyssal plain to the south (LePichon et al., 1970).

At the western end of the northern seamount arm is the small near circular Hirondelle seamount (figure 2.2), which rises with increasing gradient from abyssal depths to within 2400 metres of sea-level where it forms a sharp edged plateau.

The southern seamount arm is composed of the sparsely studied Ampere and Coral Patch seamounts. They form a pair of overlapping ridges with a trend of approximately N65°E, in contrast to Gorringe Bank which trends N50°E. Ampere seamount is located at the western end of the southern seamount arm with a width of approximately 80 km and length 120 km with a main peak and a few smaller lateral peaks (Litvin et al., 1982).

Detailed echo-sounding (Marova & Yevsyukov, 1988) revealed a main 59 metre eastern summit and a smaller 327 metre peak to the west. The eastern summit has the form of an inclined bowl which may represent a volcanic crater and the
upper slopes have a steep gradient which gradually decreases with increasing depth (Litvin et al., 1982). It has a Pliocene to Miocene age (Matveyenkov et al., 1994) with a volcanic origin and partial subaerial formation (Litvin et al., 1982; Marova & Yevsyukov, 1988). Coral Patch lies to the east of Ampere seamount with a width of approximately 80 km, a length of 150 km and a depth of within 1200 metres of sea level. To the east, Coral Patch seamount joins the Coral Patch Ridge which continues to the African margin.

The two seamount arms are bound to the west by the Madeira-Tore Rise, a NNE-SSW trending continuous ridge connecting Madeira Island (33°N, 17°W) and Tore seamount (39°N, 13°W). The Rise has variable dimensions of about 200 km width and rises 2000 to 3000 metres above the 5000 metre abyssal depths, due to the number of small seamounts on the crest and flanks. It was possibly formed following the southward movement of magma, vented from a mantle plume, along the ridge axis at a rate of 50 mm/yr from about M-4 to M-O (approximately 126-119 Ma) (Tucholke & Ludwig, 1982).
One of the major seamounts associated with the Madeira-Tore Rise, Josephine seamount, is a Late Miocene (12.5-7.7 Ma) volcano (Wendt et al., 1976) composed of homogeneous alkaline basalt, rising to within 200 metres of sea-level. The crest of the seamount has carbonate silt deposits in small basins formed by NW striking normal faults of Pliocene to Early Pleistocene age, interpreted from sampling and bathymetric surveying (Matveyenkov et al., 1994), created by tectonic movements along the Azores-Gibraltar plate boundary. These deposits were exposed to a shallow lagoon erosional environment during Pleistocene regressions (Matveyenkov et al., 1994).

The Iberian continental margin forms the eastern boundary of the seamount arms, where the Portuguese continental slope rises rapidly from abyssal depths, but to the south gradients towards the Straits of Gibraltar are lower. The whole margin and continental slope is cut by many submarine canyons, which are associated with
fault zones, including the Sao Vincent Canyon which provides the main supply route for terrigenous sediment from Iberia to the Horseshoe abyssal plain at the present time.

Enclosed by the seamount arms, the Madeira-Tore Rise and Iberian margin are three relatively small abyssal plains. They resemble the major abyssal plains of the Atlantic Ocean in that they are extremely flat with apparently undisturbed sediments, yet are of a far smaller scale. To the north is the near hexagonal Tagus abyssal plain which is the deepest of the three abyssal plains at approximately 5000 metres. It has a width of about 250 km and is bordered by gradual slopes except along the southern boundary, where Gorringe Bank and Hirondelle seamount rise abruptly. The Horseshoe abyssal plain is composed of the Western and Eastern Horseshoe abyssal plains which have a depth of 4800 metres, 200 metres shallower than the Tagus abyssal plain. The near circular Western Horseshoe abyssal plain is bound to the north and south by the steep slopes of Hirondelle and Ampere seamounts and shallower gradients towards the Madeira-Tore Rise. The Eastern and Western Horseshoe abyssal plains are separated by a narrowing at 12°30’W and elongated parallel to Gorringe Bank, Coral Patch seamount and Coral Patch
Ridge which bound it with steep gradients to the north and south. To the northeast the
gradients are lower onto the Iberian margin and to the east the are lower still
towards the Straits of Gibraltar. To the south, the Seine abyssal plain is the
shallowest abyssal plain in the region at approximately 4500 metres, the depth
having decreased overall from north to south across the three abyssal plains. The
northern boundary is quite a smooth transition with low gradient slopes onto the
base of Coral Patch and Ampere seamounts.

2.2 Sampling, Coring and Dredging

Sediment and basement rock sampling has been quite extensive, though localised
in the region, mainly concentrating on Gorringe Bank with additional sampling
of Ampere seamount and Coral Patch Ridge and minimal sampling of the abyssal
plains. Investigations into the sedimentary nature of the region began with deep sea
photography (LePichon et al., 1971) which showed oozes and silts in the Eastern
Horseshoe abyssal plain and piston core studies (Hoyt & Fox, 1977) which showed
that near surface turbidites within the Horseshoe abyssal plain could be correlated
over large distances, although showed no regular decrease of thickness or grainsize
with distance from the Iberian margin sources.

The most significant sampling in the Gorringe Bank region came from the Deep
Sea Drilling Project (DSDP) sites 120 and 135. DSDP site 135 (Hayes, 1971; Hayes
et al., 1972), located on Coral Patch Ridge at 35°20.80’N, 10°25.46’W, sampled
sediments (table 2.1) to a depth of 689 metres, but did not sample basement.
The cores consist of primarily Pleistocene to Late Oligocene pelagic deposits in the
upper section, probably deposited following formation of Gorringe Bank. These are
separated from Early Aptian to Early Eocene terrigenous sediments by an Early
Eocene to Late Oligocene unconformity.

DSDP site 120 (Ryan et al., 1973) is the only core to have sampled some of
the oldest sediments in the region which lie directly upon sampled basement rocks
(table 2.2). Cores were taken on the northern crest of the Gorringe Bank in the sad-
dle between Ormonde and Gettysburg seamounts at 36°42.12’N, 11°22.38’W. The
sedimentary deposits which were recovered to a depth of 246 metres have an age
from Barremian to Pleistocene with unconformities relating to a hiatus and change
in sedimentation between the Cretaceous and Miocene. The transition from deep
to shallow water fossils show that the upper pelagic sediments were deposited once
Gorringe Bank was near to sea-level and are probably similar to those observed in
DSDP 135. Since formation of Gorringe Bank pelagic deposition has prevailed on
## Table 2.1: DSDP Site 135 - Summary of lithologies, age and recovery depth.

<table>
<thead>
<tr>
<th>UNIT/DEPTH</th>
<th>DESCRIPTION</th>
<th>AGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Light grey nannoplankton chalk ooze. Pelagic with traces of quartz silt.</td>
<td>Pleistocene to Late Oligocene</td>
</tr>
<tr>
<td>- 0-325 m -</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Olive grey to brown mudstone, sand and brown clay. Terrigenous sediments with less than 25 % CaCO&lt;sub&gt;3&lt;/sub&gt;</td>
<td>Early Eocene</td>
</tr>
<tr>
<td>- 335-341 m -</td>
<td>Silty mud</td>
<td></td>
</tr>
<tr>
<td>- 341-350 m -</td>
<td>Quartz siltstone and mudstone</td>
<td>Cret. or Paleocene</td>
</tr>
<tr>
<td>- 431-435 m -</td>
<td>Silty mudstone, sand and clay</td>
<td>Early Maestrichtian</td>
</tr>
<tr>
<td>3</td>
<td>Black and green shale, siliceous mudstone, limestone and chert. Traces of CaCO&lt;sub&gt;3&lt;/sub&gt;. Rich in zeolite, with pyrite and 5-10 % terrigenous silt.</td>
<td>Cretaceous</td>
</tr>
<tr>
<td>- 563-570 m -</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Olive grey and black nannofossil laminated marl ooze and limestone. Some silty quartz layers and pyrite.</td>
<td>Early Aptian</td>
</tr>
<tr>
<td>- 684-689 m -</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

seamounts and abyssal highs, with turbidites and pelagic deposition on the abyssal plains. The lower terrigenous deposits at DSDP 120 give an indication of early facies of the region which lie upon the oceanic crustal basement, intercepted at between 251.7 and 253.4 metres. The cores were of course grained gabbroic breccia and metabasalts (Honnorez & Fox, 1973) with serpentinised olivine (antigorite) and chloritised pyroxene, a spilite of the ophiolite suite (Ryan et al., 1973). The original composition was of Ca-plagioclase, pyroxene and opaque minerals, but alteration under greenschist/amphibolite facies metamorphism produced amphiboles (actinolite and hornblende), chlorite and quartz, suggesting post formation heating perhaps associated with movement along the Azores-Gibraltar plate boundary.

DSDP 120 did not provide the first sampling of Gorringe Bank basement rocks which were initially recovered by the R/V Robert D. Conrad in 1965 with two piston cores (RC 9-206, RC 9-208) and a dredge (RC 9-6) (Gavasci et al., 1973). Samples comprised cataclastic metagabbros of Greenschist/Amphibolite facies meta-
Table 2.2: DSDP site 120 - Summary of lithologies, age and recovery depth.

morphism, an amphibole-biotite pyroxene and pyroxene bearing hornblendite of amphibolite facies, which are representative of lower crustal rocks (Gavasci et al., 1973). Samples of intermediate and mafic alkaline rocks also collected are thought to represent volcanism associated with a period of rapid subduction and uplift of Gorringe Bank (Gavasci et al., 1973). Further dredging by the R.R.S. Shackleton in (2/72) 1972 and (3/73) 1973 (Prichard & Cann, 1982) and a series of shallow cores drilled in Gorringe Bank by the R/V Bavenit in 1991 (Matveyenkov et al., 1994) recovered partially weathered serpentinites from Gettysburg seamount, with harzburgites in upper sections and lherzolites in the deeper section. Gabbros and amphibolitic dolerites from Ormonde seamount, with yellow ochreous shales on crest, probably formed from the decomposition of underlying volcanic rocks.

In 1977 Gorringe Bank was sampled at outcrop by the SP 3000 Cyana submersible of the Centre National pour l’Exploitation des Océans (CNEXO) (CYAGOR 1). From samples collected, Auzende et al. (1978) concluded that Gettysburg seamount is almost entirely composed of serpentinite, whilst Ormonde seamount is primarily gabbro. The ultramafic rocks from Gettysburg seamount exhibit 95-100% serpentinisation to antigorite and lizardite (Prichard, 1979), suggesting hydration at the time of formation or during heating associated with deformation. A second CYAGOR survey in 1981 (Auzende et al., 1982) with dives and dredges
provided a detailed analysis of the structure of Gorringe Bank. The southern flank of Ormonde has an upwardly varying sequence from gabbros near to the base with alkaline breccia at between 1500 and 1000 metres. Above, the gabbros are cut by north-south trending dolerite and variably trending alkaline dykes moving into a zone of course grained gabbros and finer grained dolerites which are of a similar composition to suspected pillow lavas at 500-400 m. The crest is a conglomerate of alkaline volcanic rocks, sourced by the alkaline dykes, in a limestone matrix. The northern flank of Gettysburg seamount is composed of tectonically layered serpentinites, gabbros and pyroxenes, with no alkaline volcanism. The majority of the serpentinites were derived from harzburgites formed under spinel-facies metamorphism and microstructural analysis shows a plastic flow structure which may be associated with ocean spreading. The gabbros are largely undeformed, but there are localised zones of shearing with high temperature metamorphism (Auzende et al., 1982; LaGabrielle & Auzende, 1982; Auzende et al., 1984).

The only other basement and sediment sampling in the Gorringe Bank region is from Ampere seamount where initially shallow cores recovered alkali basalts (Purdy, 1974). More comprehensive coring (Matveyenkov et al., 1994), dredging and submersible sampling (Litvin et al., 1982) revealed Ampere seamount to be composed of highly alkaline nepheline basaltoids, which are typical of the final stages of oceanic island volcanism (Matveyenkov et al., 1994). Outcropping ridges of dense reddish-ochre trachyte rocks on the upper slopes of the Western summit have rectangular jointing which gives the appearance of a stack of bricks (Litvin et al., 1982) and are separated by conglomerates formed under coastal island conditions (Marova & Yevsyukov, 1988), perhaps indicating fairly long periods between eruptions (Matveyenkov et al., 1994). Conglomerates (Litvin et al., 1982) and pillow basalts of underwater origin (Marova & Yevsyukov, 1988) have been found on the lower slopes of Ampere seamount.

2.2.1 Age and Formation of Gorringe Bank Basement Rocks

Samples collected by selected surveys have been investigated as to their metamorphic and formation history. According to Mevel (1988), two distinct phases of metamorphism are recognised in samples from Gorringe Bank. The first and major alteration is associated with seawater derived fluid reactions with the gabbros at the mid ocean ridge with shearing at high temperatures. The second is associated with water interactions, under amphibolite conditions, when the gabbros were tectonically emplaced near to the seabed and water penetrated the undeformed rocks and along shear zones associated with plate motions along the Azores-Gibraltar
boundary. Following cooling the gabbros were tectonically exposed at the seabed and underwent low temperature alteration with seawater.

Samples of gabbro from Gettysburg seamount from the cruises of the R/V Robert D. Conrad and the R.R.S. Shackleton 2/72 were dated with standard K-Ar methods and revealed three distinct groups. An age of $141 \pm 3$ Ma and $138 \pm 3$ Ma (Prichard & Mitchell, 1979) in (Feraud et al., 1986) which probably relates to the formation age of the rocks, $105 \pm 3$ Ma relating to a thermal event and $82 \pm 3$ Ma probably relating to shearing at the Azores-Gibraltar plate boundary (Prichard & Mitchell, 1979). Serri et al. (1988) concluded that Gorringe Bank has been tilted by $20^\circ$ to the east and is cut by NNE-SSW trending faults which are probably related left lateral motion.

Alkaline volcanic rocks, Ormonde gabbros and a dolerite from Gettysburg from CYAGOR and Gibraco 1972 were dated (Feraud et al., 1982, 1986) using a step-wise heating $^{40}$Ar/$^{39}$Ar method giving three distinct ages similar to the results of (Prichard & Mitchell, 1979). The formation age for the rocks was found to be approximately 143 Ma from Gettysburg dolerite and 140 Ma for Ormonde gab-bros. The second with an age of 110 Ma relating to a northeast tilt of Gorringe Bank and structures relating to the Madeira-Tore Rise. The youngest event at 75 Ma resulting from change to compressive movement along the plate boundary from strike-slip at this time was seen in the study of Feraud et al. (1982) but not of Feraud et al. (1986). The alkaline volcanism lasted for about 6 Ma from approximately 60 Ma and is probably linked to compressive events in the Upper Cretaceous (Feraud et al., 1982, 1986).

### 2.3 Gravity and Magnetics

Gorringe Bank is characterised by some of the worlds highest free-air gravity anomalies with amplitudes of up to approximately 480 mGal (+390 mGal peak) (Purdy, 1975). The Bank has a vertical gravity gradient of approximately 85 mGal/km and a steeper horizontal gravity gradient on the northern flank approaching 10 mGal/km (Purdy, 1975) than on the southern flank (figure 2.5) corresponding to the asymmetric bathymetry (figures 2.3 and 2.5). Gorringe Bank is flanked by relative gravity lows in the surrounding abyssal plains with a negative anomaly of -90 mGal in the Eastern Horseshoe abyssal plain and -60 mGal in the Tagus abyssal plain. The Western Horseshoe and Seine abyssal plains have local gravity lows of -76 mGal and -40 mGal respectively, contrasting positive gravity anomalies which mimic the bathymetry over seamounts and abyssal highs. Hirondelle
seamount has anomalies of greater than 100 mGal and Coral Patch and Ampere seamounts have anomalies of greater than 200 mGal and with lower horizontal gravity gradients than Gorringe Bank of about 4 mGal/km (Purdy, 1975).

Geoid anomalies (Souriau, 1984) derived from Seasat altimeter data give indications as to the deeper structure of the region than revealed by the gravity anomalies. The geoid anomalies which are longer wavelength than the corresponding gravity anomalies have heights of -3 metres over the Tagus abyssal plain and up to 8 metres over Gorringe Bank. The Geoid maximum over Gorringe Bank is shifted slightly to the south of the bathymetric peak which may represent a deep source, but is more likely due to the attractive affects of Coral Patch and Ampere seamounts to the south.

Previous gravity (e.g., LePichon et al., 1970; Purdy, 1974; Bergeron & Bonnin, 1991) and geoid (e.g., Souriau, 1984) models which have been constructed are presented and discussed in chapter 6, section 6.2.

Figure 2.5: Free-air and magnetic anomaly profile across Gorringe Bank, the Eastern Horseshoe abyssal plain and Coral Patch Ridge, from Shackleton 2-72 in Purdy (1975).

In contrast to the gravity anomalies, magnetic anomalies are generally of low
amplitude and comparatively smooth with a relief of generally less than 100 nT except for Gorringe Bank, Ampere seamount and to the northwest of Hirondelle seamount. Gorringe Bank has short wavelength local magnetic anomalies with wavelengths of about 1.0-1.5 km (Gorodnitskiy et al., 1988) and amplitudes of greater than 700 nT for Gettysburg seamount corresponding to horizontal gradients in excess of 600 nT/km. Magnetic anomalies over Ormonde seamount have wavelengths of about 2 km and amplitudes of up to about 1000 nT. Analysis of magnetic anomaly contour maps (Gorodnitskiy et al., 1988) reveals the predominance of two systems of magnetic lineations. The first has a trend of between 20-30°, similar to the orientation of faults observed to cut Gorringe Bank from bathymetry data. The second with an orientation of 110-130° correlates with steep vertical fault scarps on bathymetric slopes of a similar trend.

Ampere seamount has variable polarity anomalies with peak to peak amplitudes of greater than 1000 nT (Purdy, 1975), with the highest normal amplitudes corresponding to the eastern summit which is thought to be the paleovolcanic eruption centre (Marova & Yevsyukov, 1988) and high amplitude reversed anomalies associated with the western summit (Matveyenko et al., 1994).

To the northwest of Hirondelle seamount, magnetic anomalies with amplitudes of greater than 1000 nT and a trend of about N10°E are associated with seafloor spreading magnetic anomalies. The first identified chron M-0 (Klitgord & Schouten, 1986) (approx. 119 Ma) is located to the NW of Hirondelle (36°48'N, 13°33'W), but has not been identified in the Western Horseshoe abyssal plain. The J anomaly (Pitman & Talwani, 1972), which is associated with the formation of the Madeira-Tore Rise, is a linear zone of high amplitude magnetic anomalies of up to 1000 nT with an age of between M-0 (118 Ma) and M-1 (122 Ma) (Rabinowitz et al., 1979). It is located at approximately 13°18’W in the Tagus abyssal plain just to the east of M-0 marking the oldest recognised anomalies. Modelling of the magnetic anomalies (Pinheiro et al., 1992) in the Tagus abyssal plain revealed a clear N-S trending magnetic anomaly at 11°40’W which may represent chron M-11 (133 Ma). To the south of Coral Patch seamount the first identified seafloor spreading anomaly M-25 (Klitgord & Schouten, 1986) (approx. 156 Ma) is located at 34°36’N, 13°6’W. Chron M-0 is located at 34°42’N, 16°54’W and offset from M-0 northwest of Hirondelle seamount by about 3.3° to the west.

The region to the east of these earliest identified seafloor spreading anomalies is a magnetic smooth zone, as identified along the margins of NW Africa and Iberia (Vogt et al., 1970; Poehls et al., 1973), with low amplitude magnetic anomalies of about ±20-50 nT in contrast to about ±100-300 nT typical of the North Atlantic
CHAPTER 2. GEOLOGICAL AND GEOPHYSICAL SETTING

2.4 Seismic Reflection and Refraction Surveys

2.4.1 Seismic Reflection Data

One of the first reflection surveys in the region was by the 9\textsuperscript{th} cruise of the R/V Jean Charcot in 1970 (LePichon \textit{et al.}, 1970). The seismic profile across Gorringe Bank, the Eastern Horseshoe abyssal plain and Coral Patch Ridge was used to aid in the choice of location for DSDP sites 120 and 135 (Hayes \textit{et al.}, 1972). This profile collectively with data from seismic reflection surveys by Researcher 4-71, Shackleton 3-73 and Atlantis II were interpreted by Purdy (1974, 1975). A lack of velocity information, sparse data coverage and difficulty in reflection correlation across boundaries between sediment bodies, precluded a quantitative interpretation and defining of isopachs. The next survey in 1988 by the 14\textsuperscript{th} Cruise of the R/V Rift (Kogan, 1990) performed a grid survey across the Gorringe region and two multichannel seismic lines were collected in July 1992 by the R/V OGS-Explora across Gorringe Bank and Coral Patch seamount and intervening abyssal plains (Sartori \textit{et al.}, 1994).

Reflection data across Gorringe Bank clearly illustrate the asymmetry of the sedimentary cover with folding and slumping of fairly thick sediments on the southern flank and faulting and slumping of the thin sedimentary cover on the northern flank into the Tagus abyssal plain (Ryan \textit{et al.}, 1973). Based on this asymmetry, Purdy (1975) concluded that formation of Gorringe Bank could not have been by a simple vertical uplift. Further evidence was provided by the data of (Kogan, 1990; Sartori \textit{et al.}, 1994) from which it was concluded that Gorringe Bank overthrusted the Tagus abyssal plain along high angle sheet thrusts with total uplift of about 4-5 kilometres at post Early Eocene time, possibly corresponding to the unconformity observed in DSDP 135.

Data on the southern seamount arm reveal intense folding and south to north thrust faulting of sediments with a thickness of up to 2 seconds two way travel time (TWTT) (Sartori \textit{et al.}, 1994) on Coral Patch Ridge atop an uneven thrusted crystalline basement (Kogan, 1990), which contrasts the flat lying upper sediments of the Eastern Horseshoe abyssal plain to the north. The reflection Moho could not be identified in the Horseshoe abyssal plain from the data of Sartori \textit{et al.} (1994) and although poorly imaged, is interpreted to shallow beneath Coral Patch seamount, represented by the base of distinct seismic laminations, forming a discontinuous sub-horizontal reflection.
CHAPTER 2. GEOLOGICAL AND GEOPHYSICAL SETTING


Very thick sediments of up to 4 seconds TWTT in the Eastern Horseshoe abyssal plain dip slightly towards the centre of the basin in the near surface, but deeper sediments and the faulted basement are obscured by an olistostrome or disturbed zone which dips to the west from 0.5 seconds TWTT to 1 second TWTT (Purdy, 1975). Its surface is very irregular and it becomes less distinctive as it decreases in thickness towards the west. The origin of this zone may result from disturbance of the sediments before deposition of the undisturbed upper sediments or from a period of continual disturbance which decreases to the west with the upper unconsolidated sediments being undisturbed. Purdy (1975) favoured this second explanation with the zone being linked to a nascent subduction zone in the Eastern Horseshoe abyssal plain. However, the interpretations of Bonnin et al. (1975);
Auzende et al. (1981) conclude that current shortening in the Eastern Horseshoe abyssal plain is away from Gorringe Bank and proposed region of subduction and that the disturbed zone represents an olistostrome derived from the Straits of Gibraltar during the Middle to Late Miocene (15-5 Ma) rather than from Gorringe Bank.

Rona (1969); Pautot et al. (1970); Schneider & Johnson (1970) propose that salt deposits exist along the North Atlantic margin and Pautot et al. (1970); Purdy (1975) showed seismic evidence for salt diapirs in the Horseshoe abyssal plain and along the North African and Iberian margins. Diapirs in the Horseshoe abyssal plain, with diameters of about 12 km and near vertical walls in 4 km of sediment, sometimes pierce the seafloor. Salt responsible for these diapirs may be related to the initial rifting of the Atlantic, ceasing deposition when subsidence became too great (Schneider & Johnson, 1970). The salt would therefore be of Jurassic to Triassic age in the Gorringe region and would explain why no salt has been observed in cores which have sampled to an insufficient depth to sample sediments of this age.

The seismic structure of the Western Horseshoe abyssal plain differs from the Eastern Horseshoe abyssal plain in that the majority of the features exhibit little
Table 2.3: Seismic facies of the Tagus Basin from Mauffret et al. (1989).

<table>
<thead>
<tr>
<th>UNIT</th>
<th>DESCRIPTION</th>
<th>AGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1A - 1B</td>
<td>Turbidites with minor pelagics</td>
<td>Late Eocene to Present</td>
</tr>
<tr>
<td>2</td>
<td>Channelled Nanno chalks and red clays</td>
<td>Mid-Cenomanian to Mid-Eocene</td>
</tr>
<tr>
<td>3A</td>
<td>Black shales atop laminated shaley marls</td>
<td>Albian to Barremian</td>
</tr>
<tr>
<td>3B</td>
<td>Fill beneath a prominent reflector</td>
<td>Tithonian to Valanginian??</td>
</tr>
<tr>
<td>4</td>
<td>Thick chaotic facies of synrift deposits Possible evaporites</td>
<td>Oxfordian to E-Tithonian?? Triassic</td>
</tr>
</tbody>
</table>

recent activity. A 1-2 km basement relief is overlain by approximately 2 km of sediments which post date the basement features as no disturbance of the sediments abutting basement blocks is observed in the majority of the abyssal plain, but to the east some of the sediments appear to be affected by basement faulting (Purdy, 1974, 1975).

The Tagus abyssal plain as with the Western Horseshoe abyssal plain exhibits little evidence of recent activity with sedimentation not deposited under basement control. The large basement topography is overlain by 2-3 kilometres of flat lying sediments with deep sediments overlain by a possible olistostrome and debris flows derived from Gorringe Bank (Mauffret et al., 1989). These are topped by turbidites and Pliocene and Quaternary sediments in the upper section (Ryan et al., 1973) which terminate at the base of the Gorringe Bank in an orderly manner (Purdy, 1974, 1975). A comprehensive seismic reflection survey of the Tagus abyssal plain by Mauffret et al. (1989) on Lusitanie cruise 86 (figures 2.8 and 2.7) along with a single line (line 5) by Pinheiro et al. (1992) (figure 2.11) showed sediments of greater than 3 seconds TWTT which have been divided into units (table 2.3) based on seismic facies analysis and partial correlation with DSDP site 398 (Sibuet et al., 1979) and ODP site 103 (Boillot et al., 1987).

The Tagus abyssal plain is bound to the west by the Madeira-Tore Rise. Re-
reflection data is sparse, however data from the 6th cruise of the R/V Professor
Shtokman in 1981 (Yel’nikov et al., 1986) show uplifted blocks and thrust faulting,
with deformation prior to deposition of 1-2 kilometres of sediments and normal
or strike-slip block faulting to the north. To the west a single channel reflection line
acquired by RSS Discovery cruise 178 in 1988 (Peirce & Barton, 1991) reveals the
crest of Josephine seamount to be a sediment filled caldera rimmed by basement
highs. To the north of Josephine seamount lies a second smaller seamount ridge
with disturbed sediments tilted upwards towards the peak (Peirce & Barton, 1991).

The Iberian Atlantic Margin project (Banda & Torne, 1995) acquired a com-
prehensive near vertical incidence seismic reflection survey with a 5 km streamer
and a large (7524 cu.in.) airgun array source on-board the Seisquest and M/V
Geco-Sigma in 1993. The 3500 km of data acquired across the Iberian margin,
Gulf of Cadiz and several lines in the Gorringe Bank region when combined with
wide angle reflection and refraction data recorded by land stations and Digital
Ocean Bottom Seismometers (DOBS) is certain to provide new information as to
the structure of the region in the future.

2.4.2 Seismic Refraction Data

The results of the first refraction survey in the Gorringe Bank region were presented
by Purdy (1974, 1975). Four seismic refraction profiles (A-D) for the Tagus, East-
er Horseshoe, Western Horseshoe and Seine abyssal plains (figures 2.8 and 2.9)
were collected using the single ship shooting technique of Hill (1963), with three
Cambridge internally recording sono-radio buoys (Gray & Owen, 1969). Lines A, B
and C were collected on Researcher cruise 471 in 1971, whilst line D was collected
on Shackleton cruise 3073 in 1973. Recognition of wide angle reflections from the
upper layers was not possible and models were uncorrected for basement relief due
to the lack of reflection data control. Therefore these models do not represent ac-
curate crustal sections, but major discontinuities such as the Moho can be defined
with confidence (Purdy, 1975).

Line A in the Tagus abyssal plain shows 3 km of sediments overlying a 2.1
km thick, 4.9 kms$^{-1}$ Layer 2 with a 1 km thick, 6.99 kms$^{-1}$ layer 3. The Moho is
modelled at a depth of 11.1 km with a velocity of 7.6 kms$^{-1}$ which is lower than
typical oceanic crust. The reversed profile (Line A-R), could not be interpreted
as a conventional reversed profile and shows 2.9 km of sediments overlying a 1.9
km layer 2. Purdy (1974) assumed that line A represented the true structure of
the Tagus abyssal plain. However Pinheiro et al. (1992) say that the discrepancy
between profiles A and A/R, including the absence of a 6.5-7.2 kms$^{-1}$ layer 3 in

line A/R, represents a change in the nature of the crust along the profile.

Line B-B/R in the Eastern Horseshoe abyssal plain exhibits typical oceanic crustal characteristics, but an anomalous velocity structure which is probably associated with the uncorrected basement relief. Thickening of the upper layer from west to east is probably due to the source of terrigenous sediment supply from the Iberian margin. The 4.12 layer decreases in thickness towards the east and may represent a thin evaporite sequence in the depths of the basin (Rona, 1969; Pautot et al., 1970; Schneider & Johnson, 1970), for which a velocity inversion would be undetected by this experiment. Layer 3 dips to the west with a Moho depth of approximately 12 kilometres.

Line C in the Seine abyssal plain as with line B-B/R exhibits an oceanic crustal structure, but an anomalously low velocity upper mantle. The 4.11 km/s layer
may represent similar material to the 4.12 km s\(^{-1}\) of line B-B/R.

Line D in the Western Horseshoe abyssal plain gave the clearest results of the experiment, with oceanic crustal structure and velocities. Two kilometres of sediments, with a structure which differs from that in the Eastern Horseshoe abyssal plain, overlie a 2.8 km layer 2 with a velocity of 4.55 km s\(^{-1}\). Layer 3 has a thickness of 2.5 km and a velocity of 6.43 km s\(^{-1}\), to a Moho depth of 12 km.

The low velocities associated with the deepest layers observed on many of these profiles have also been found by refraction experiments for the eastern Atlantic (Ewing & Ewing, 1959), with velocities of 7.7-7.8 km s\(^{-1}\). If a nascent subduction zone exists in the Eastern Horseshoe abyssal plain, the upper mantle velocities in this region would be anomalous as areas of oceanic crustal consumption do not typically exhibit low velocities (Grow, 1973).

The results of a second survey presented by Bonnin (1978) were similar to those of Purdy (1974) in the use of sonobuoys, but the shorter lines only enabled investigation of the shallow crustal structure and on most lines only the sediment structure. Three lines 5M, 6M and 7M were collected on the 1972 cruise Gibrac of Ifremer and lines M-NOR and M-NOR/R were collected in 1979 by the cruise Noratlante (figures 2.8 and 2.10) (Bergeron & Bonnin, 1991). Lines M-NOR, MNOR/R and
4.16 2.34 1.5 2.0 1.5 1.5

1.8 3.1 4.76

6.8

Figure 2.10: Crustal sections from seismic refraction experiments of Bonnin (1978). Dots and grey shading represent sediments and crust respectively. Velocities in kms\(^{-1}\). See figure 2.8 for location.

6M only model to a depth of approximately 2 km into the sediment cover, so were unable to give information on the crustal structure. Line 5M in the Eastern Horseshoe abyssal plain shows a similar sediment structure to Line B/B-R of Purdy (1974) which lies in the same region, but the deeper structure is modelled with a lower velocity layer of 4.81 kms\(^{-1}\) at a depth of 10 km probably associated with crustal basement. Line 7M in the southwestern Tagus abyssal plain shows a roughly similar structure to line A of Purdy (1974) with sediments of velocity 1.8 to 3.1 kms\(^{-1}\) overlying a 4.76 kms\(^{-1}\) layer at a depth of about 8 km which probably represents crustal basement.

The next seven years saw significant advances in the methodology and technology employed to perform refraction surveys. Sonobuoys were superseded by Digital Ocean Bottom Seismometers (DOBS) (Kirk et al., 1982). These have a fixed position on the seafloor enabling the recovery of a higher signal to noise in the recorded signal and more accurate locations in comparison to floating sonobuoys which are liable to drift.

On cruise 161 of the R.R.S. Discovery in 1986, line 5, a reversed refraction profile
figures 2.8 and 2.11) in the Tagus abyssal plain, was acquired utilising two digital ocean bottom seismometers (DOBS) (Kirk et al., 1982) which were deployed at each end of the line. The interpretation of Pinheiro et al. (1992) (figure 2.11) is neither typical of oceanic or thinned continental crust. A very thin 2 km crust with a velocity of 4.4-6.3 kms\(^{-1}\) is underlain by material with velocities lower than expected for upper mantle, yet higher than expected for lower crust and may be related to serpentinised peridotite (Pinheiro et al., 1992). The low velocity zone in the centre of the profile has no evidence for a layer 3 and is thought to represent the ocean continent transition in this region (Pinheiro et al., 1992).

![Figure 2.11: Velocity-depth refraction model (Line 5) of Pinheiro et al. (1992) from the Tagus abyssal plain. Velocities in kms\(^{-1}\), assumed in brackets. Triangles mark DOBS locations. See figure 2.8 for location.](image)

A second survey using five digital ocean bottom seismometers was performed along a NW-SE refraction line (figures 2.8 and 2.12) providing information as to the crustal structure of the Madeira-Tore Rise, Josephine seamount and Western Horseshoe abyssal plain (Peirce & Barton, 1991). The sediment sequence was divided into 2 layers, based on seismic reflection data interpretations, with velocities of 1.8-3.7 kms\(^{-1}\). The upper crust was modelled with velocities of 5.5-5.6 kms\(^{-1}\), but the peak to the northwest which required lower velocities of 4.8-5.2 kms\(^{-1}\) may represent low density rock erupted in shallow water. The lower crust has a mean
velocity of 6.8 kms\(^{-1}\), but the central region shows high 7.4 kms\(^{-1}\) velocities which were required to model the observed Moho reflections. The crustal velocities and densities are neither typical of oceanic or continental crust, but are thought to represent oceanic crust non-tectonically thickened perpendicular to the east-west trend of the magnetic spreading anomalies. The depth of the Moho varies from 11 km north of the Madeira-Tore Rise to 17.5 km beneath Josephine seamount and 12 km beneath the Western Horseshoe abyssal plain. Models imply that Josephine seamount is locally compensated by a low density root and that Josephine seamount and Madeira-Tore Rise formed at or near to the Mid Atlantic Ridge at an age of between 119 Ma and 126 Ma (Tucholke & Ludwig, 1982).

Figure 2.12: Velocity-depth refraction model of Peirce & Barton (1991) from Josephine seamount. Dotted lines show velocity contours at 0.2 kms\(^{-1}\). Triangles mark DOBS locations. See figure 2.8 for location.

### 2.5 Seismicity of the Azores-Gibraltar Region

The Azores-Gibraltar region is characterised by large magnitude seismicity (figure 2.13) with six events of a magnitude greater than 7 and three with magnitude greater than 8 since 1931, including the Lisbon earthquake of 1755 which is possibly the world’s largest magnitude recorded earthquake producing the only known Atlantic earthquake tsunami (Muzcua et al., 1991).

A study and revision of earthquake data from 1912-1985 and historical seis-
micity for the 14-1900’s show a boundary which consists of a western extensional (Azores), a central strike slip (Gloria Fault) and an eastern compressional region (Gorringe Bank region) (Buforn et al., 1988).

Seismicity in the Azores region (figure 2.13), including the Azores Triple Junction and Terceira Ridge, is of generally moderate magnitude and shallow depth with the majority of epicentres located on the ridges. The seismicity of the Terceira Ridge is fairly complex, however normal faulting (principal axes of stress at an average of N25°E (Buforn et al., 1988)) and earthquake derived slip vectors (Udias & Buforn, 1991) indicate NE-SW extension.

Seismicity in the central section trends east-west with a linear form following the Gloria Fault. Focal mechanisms (principal axes of stress at approximately N35°W (Bufron et al., 1988)) and slip vectors parallel to the fracture zone (Udias & Bufron, 1991) are consistent with the right lateral strike slip motion (Udias et al., 1976; Bufron et al., 1988). The focal mechanism for the 1970 earthquake at 37°13.2′N, 14°55.8′W (Grimison & Chen, 1986) and a few of smaller magnitude give some indication that the Gloria Fault continues into the Madeira-Tore Rise, yet masked by the complex bathymetry. This may be the cause of the disturbed zone observed on the seismic data presented by Peirce & Barton (1991).

In the Gorringe Bank region, to the east of Josephine seamount, focal mechanisms show a domination of reverse faulting with a strike slip component (figures 2.13 and 2.14). Seismicity initially appears diffuse in a zone which increases in width towards the Iberian margin. However, the seismicity is not purely diffuse as the seismic trend bends to the ESE, rather than following the trend of an extrapolated Gloria Fault. With seismic strain distributed over a wide region (Vegas, 1991) a single plate boundary is unlikely (Grimison & Chen, 1986, 1988). P axes have a mean trend of N30°W (Bufron et al., 1988) and slip vectors, calculated from focal mechanisms, have a trend of about N35°W (Udias & Bufron, 1991). Many epicentres are related to the oceanic continuation of continental faults (figure 2.14) from the Iberian margin, including the Cadiz-Alicante, Alentejo-Plasencia (Messejana), Bajo-Tajo and Nazare Faults (Bufron et al., 1988; Moreira, 1985, 1991). Focal depths are deep for a region with no Beniof zone and increase in depth from 20°W to 2°W with a depth of up to 100 km at 10°W (Bufron et al., 1988), but most are less than 50 km corresponding to the 600°C isotherm (Chen & Molnar, 1983).

The presence of deep crustal faulting in this region is shown by the anomalously high heat flow to the west of the Madeira-Tore Rise, to the north of Gorringe Bank and in the Horseshoe abyssal plain. Heat flow in the Horseshoe abyssal plain has a mean of 70 mW/m² (Verzhbitsky & Zolotarev, 1989), which is far greater
Figure 2.13: Seismicity greater than Mb 4 in the Azores-Gibraltar region from the ISC. Focal mechanisms from the ISC, Grimison & Chen (1986, 1988) and Buorn et al. (1988), with depths in kilometres.
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Figure 2.14: Seismicity greater than Mb 4 in the Gorringe Bank region from the ISC. Focal mechanisms from Grimison & Chen (1986, 1988) and Buform (1988), with depths in kilometres. Solid white lines mark major mainland faults. White dash lines show possible continuation of mainland faults.

than the recorded 45 mW/m² in the central Atlantic abyssal plains adjacent to the Azores-Gibraltar fracture zone. Measurements of heavy helium (He4) seawater concentration, which is an indicator of deep crustal faulting, are 2.5 to 5 times greater than background.

One of the earliest studies of seismicity in the Gorringe region by Fukao (1973) re-analysed the 7.30 ± 0.28 Mb earthquake of the 28th February 1969 after differing results by McKenzie (1972) and Udias & Arroyo (1972). Based on the primary earthquake’s location (36°0.6’N, 10°34.2’W with a focal depth of 22 ± 6 km) and aftershocks which lie on a NE-SW plain corresponding to the first nodal plain of the focal mechanism, interpretation was of a thrust fault plain which dips at 52° to N35°W with a strike of N55°E through the upper oceanic lithosphere. This may represent a section of the present plate boundary (Fukao, 1973), but seems
unlikely to represent a shallow (Grimison & Chen, 1986) Beniof zone, based on seismic reflection evidence and the proximity to low density buoyant continental crust, which would counteract subduction (McKenzie, 1972).

### 2.6 Morphology and Evolution of the Eurasian-African Plate Boundary

Changes in the current morphology of the Azores-Gibraltar plate boundary, from the Azores Triple Junction to the Straits of Gibraltar, are due to the anticlockwise rotation of the African plate with respect to the Eurasian plate. Euler poles for the African plate with respect to the Eurasian plate have been calculated by various authors (table 2.4).

<table>
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<th>LATITUDE</th>
<th>LONGITUDE</th>
<th>REFERENCE</th>
</tr>
</thead>
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<td>9.3</td>
<td>-46.0</td>
<td>Le Pichon (1968)</td>
</tr>
<tr>
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<td>-28.2</td>
<td>McKenzie (1972)</td>
</tr>
<tr>
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<td>-23.5 ± 200</td>
<td>Chase (1978)</td>
</tr>
<tr>
<td>25.2 ± 500</td>
<td>-21.1 ± 100</td>
<td>RM2 (1978)</td>
</tr>
<tr>
<td>21.3 ± 70</td>
<td>-21.0 ± 60</td>
<td>Searle (1980)</td>
</tr>
<tr>
<td>21.0 ± 700</td>
<td>-20.6 ± 70</td>
<td>NUVEL 1 (1989)</td>
</tr>
<tr>
<td>21.0 ± 100</td>
<td>-21.0 ± 100</td>
<td>Wesatway (1989)</td>
</tr>
</tbody>
</table>


The most comprehensive model, NUVEL 1 (DeMets et al., 1990) utilised fault azimuths, focal mechanisms and slip vectors from the Gloria Fault and predicts a counter clockwise rotation of 0.12°/my, translating to a slip of about 4 ± 1 mm/yr along the Gloria Fault and extension on the Terceira Ridge of about 2 to 3 mm/yr. The rate of convergence at Gibraltar is about 4 mm/yr, which is lower than expected considering the large magnitude seismicity (Grimison & Chen, 1986), on a trend of N45°W, 10° counter clockwise of slip vectors based on focal mechanisms alone (Udias & Buform, 1991).

Extensional motion on the Terceira Ridge changes to right lateral strike slip forming the Gloria Fault at approximately 24°W. Bathymetry data show a 5-16
Figure 2.15: Tectonic map of the Azores-Gibraltar plate boundary. White lines mark current plate boundaries. NAM=North American plate. Black lines represent faulting and lineations: MAR=Mid Atlantic Ridge, TR=Terceira Ridge, EA=East Azores Fracture Zone, GF=Gloria, V=Vilarica, NZ=Nazare-Seia Lousa, TF=Lower Valley of the Tagus, AL=Alandroal, MJ=Messejana, LL=Loule, GQ=Gaudalquivir lineament, CA=Crevillente (Cadiz-Alicante) fault zone, SAF=Southern Atlas Fault zone. MA=Middle Atlas fault zone. Previous plate boundaries: B=Bay of Biscay, KT-ABR-NST=King’s Trough-Biscay Rise-North Spanish Trough. Arrows show plate motion relative to the Eurasian and African plates.
km wide V shaped valley and a ridge immediately to the south, associated with the Gloria Fault, which truncates and offsets the NNE-SSW trending bathymetry and magnetic spreading anomalies (Laughton et al., 1972; Whitmarsh & Laughton, 1974). Geological Long Range Inclined Asdic (GLORIA) data (Searle, 1979) show the Gloria Fault as straight narrow echos or a series of parallel to near parallel echos, possibly cut by weak WNW-NW normal faults near to 21°W. East of 20°W, the plate boundary becomes more complex with several ENE trending sonar echoes (Searle, 1979). In addition CD64 collected new GLORIA data along the GLORIA fault clarifying its location. At approximately 16°W the end of the Gloria Fault becomes masked by the Madeira-Tore Rise. In the Gorringe Bank region to the east of 16°W deformation is regional with localised deformation corresponding to the seamounts and no observed time progression of deformation across the region (Sartori et al., 1994). Focal mechanisms indicate thrust faulting with a right lateral strike slip component. To the east this region moves into a zone of continental collision close to the Straits of Gibraltar. There is some indication from analysis of magnetic spreading anomalies and seismicity that the pole of rotation is moving southwards, causing a reduction of extension and compression at the western and eastern ends of the Azores-Gibraltar plate boundary (Roest, 1987).

A single large magnitude seismic event (figure 2.13) with right lateral strike slip motion and associated smaller events to the south of the Gloria Fault at 36°N, 18°W may be related to a NE-SW trending fracture zone (figure 2.15) (Vegas, 1991). Intersection of the fracture zone with the NE-SW fault trends observed on the Moroccan margin near Agidir (Bufrom et al., 1988) may form the southern boundary of one or more counter clockwise rotating sub-plates (Vegas, 1991), bound to the north by the Gloria Fault and to the east by the Southern Atlas Fault (figure 2.15).

Evolution of the African-Eurasian plate boundary is coupled with the complex spreading history of the North American (NA), Eurasian (EU) and African (AF) plates and formation of the North Atlantic Ocean. Previous studies (e.g., Pitman & Talwani, 1972; Klitgord & Schouten, 1986) have primarily concentrated on spreading in the central Atlantic with respect to the NA and AF margins, as they show a more continuous and more easily decipherable magnetic anomaly record. Only a few studies have focused on the motion between the EU-AF plates (e.g., Srivastava et al., 1990; Roest & Srivastava, 1991).

Plate reconstructions for the North Atlantic from chron M-25 to chron 6 are shown in figure 2.16. Initiation of rifting between the NA-AF plates at about 175 Ma (Klitgord & Schouten, 1986) was prior to rifting of the NA-EU plates.
Figure 2.16: Plate reconstructions for the Northeast Atlantic from chron M-25 to 6 (Srivastava et al., 1990).
which commenced from south to north in the Jurassic at about 165 Ma (Sclater et al., 1977). During the Late Jurassic (M-25, 156 Ma) there was stretching between Iberia (IB) and the Grand Banks and continued spreading between NA-AF (figure 2.16a), with plate boundaries between IB-AF and IB-EU (Srivastava et al., 1990) and regional left lateral motion between EU-AF (Whitmarsh & Laughton, 1974). Mauffret et al. (1989); Malod & Mauffret (1990) propose the presence of an abandoned spreading ridge in the eastern Tagus abyssal plain with an age of oldest chron M-21 (150 Ma), abandoned by at latest chron 16 (142 Ma), but perhaps chron M-10 (130 Ma) when the ridge jumped to the west.

At chron M-10 (130 Ma) spreading was occurring between IB-NA (figure 2.16b) and the Flemish Cap separated from Galicia Bank (Srivastava et al., 1990). Stretching continued to the north with continued left lateral motion between EU-AF (Whitmarsh & Laughton, 1974) and boundaries between IB-EU and IB-AF.

Active seafloor spreading between New Foundland and IB (figure 2.16c) (Srivastava et al., 1990) commenced by M-O (119 Ma) resulting in possible right lateral motion between the EU-AF plates (Whitmarsh & Laughton, 1974). Stretching continued between NA-EU to the north with compression on the boundary between the EU-IB plates to the east. Paleomagnetic evidence constrain rotations of IB to be approximately 35° counter-clockwise with respect to EU (Galdeano et al., 1989), occurring between chron M-0 and 31, but that 30° of this rotation occurred between M-10 and M-0.

By chron 34 (84 Ma) (figure 2.16d) IB which was previously independent began to move as part of the African plate and the Bay of Biscay was mostly opened by EU-NA seafloor spreading by chron 33 (77 Ma) (figure 2.16e). Between chron 34 and 33 the relative spreading rates of the NA-EU and NA-AF plates were similar resulting in many left and right lateral movements between EU-AF, but little relative net motion (Whitmarsh & Laughton, 1974) along the E-W Bay of Biscay plate boundary (B) which was the main EU-AF plate boundary from chron 34 (84 Ma) to chron 17 (40 Ma) (Roest & Srivastava, 1991). From chron 34 (84 Ma) to chron 31 (69 Ma) (figure 2.16f) motion on this boundary was extensional (Roest & Srivastava, 1991), with continued extension with a strike slip component until chron 25 (59 Ma) (figure 2.16g). Almost entirely strike slip motion, but with gradually decreasing magnitude continued until chron 19 (44 Ma) and IB again began to move as an independent plate (Srivastava et al., 1990) with the formation of two new plate boundaries (figure 2.16i). To the north was the King’s Trough, Azores-Biscay Rise, North Spanish Trough boundary and to the south was a boundary with a similar location to the present Azores-Gibraltar fracture zone.
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Motion was compressional in King’s Trough and strike slip with some compression along the Azores-Biscay Rise and compressional in the North Spanish Trough and Pyrenees (Roest & Srivastava, 1991), with similar motion sense along the Azores-Gibraltar fracture zone.

By chron 6 (20 Ma) (figure 2.16j) motion along KT-ABR-NST became small and Iberia joined with Eurasia (Srivastava et al., 1990). Motion on the Azores-Gibraltar plate boundary has been fairly stable since (Argus et al., 1989), with extension on the Terceira Ridge, right lateral slip along the Gloria Fault and compression and crustal shortening of about 11 mm/yr for the past 10 Ma (Purdy, 1974, 1975) and currently about 4 mm/yr (DeMets et al., 1990) in the Gorringe Bank region.

2.7 The Ocean-Continent Transition

The Ocean-Continent Transition Zone (OCT) has not been identified in Gorringe Bank region. Differences in the nature of the crust between the Eastern and Western Horseshoe abyssal plains from refraction models (figure 2.9) (Purdy, 1975) were assumed to be due to tectonic alteration of the crustal structure in the Eastern Horseshoe abyssal plain in association with the shallow seismicity. The location of the OCT on either side of the Azores-Gibraltar plate boundary is likely to be different due to several phases of relative plate motion (figure 2.16). However, examination of regions to the north and south of the Gorringe region may give constraint to the location of the OCT in the Horseshoe abyssal plain.

A study in the Iberia abyssal plain (figure 2.17), to the north of the Tagus abyssal plain (Whitmarsh et al., 1990) places the OCT at 40-41°N, 12°10′ to 12°30′W with a trend of about N10°E. The identification of the OCT was based upon the crustal velocity structure from refraction models and a ridge and trough identified on seismic reflection data. Magnetic models indicate that continental crust must lie east of 40°35′N, 12°08′W, also providing constraint to the western extent of oceanic crust and location of the OCT.

In the Tagus abyssal plain the OCT was interpreted to be located at approximately 11°W (Mauffret et al., 1989) based on a low angle crustal reflection (reflection L) from interpretations of seismic reflection data, a linear magnetic ridge, gravity data and the refraction lines of Purdy (1974). However Pinheiro et al. (1992) interpret the OCT to be further west at about 11°40′W, based on interpretation of seismic refraction line 5 (figure 2.11) and magnetic models. These reveal a clear N-S trending magnetic anomaly related to chron M-11 (133 Ma) assuming
Figure 2.17: Location of the Ocean-Continent Transition zone (OCT) on the Iberian margin. Green lines (Mauffret et al., 1989), dots (Whitmarsh et al., 1990) and stripes (Pinheiro et al., 1992) show the interpretations as to the location of the OCT. Red boxes show the possible extent of thin crust underlain by serpentinitised peridotite (Whitmarsh et al., 1993). Black lines mark the location of magnetic seafloor spreading anomalies (Klitgord & Schouten, 1986) and black dash lines show the location of fracture zones (Klitgord & Schouten, 1986). Blue lines show the S magnetic anomalies (Rousert, 1982) and red lines indicate the location of the oldest seafloor spreading anomalies identified by modelling (Whitmarsh et al., 1993).
a constant spreading rate of 10 mm/yr. Further magnetic models (Whitmarsh et al., 1993) show that the anomalies to the west of 11°50’W can be interpreted as seafloor spreading anomalies, but to model those to the east would require a 150 km wide constant polarity zone. However, the magnetic intensity of this zone is an order of magnitude less. Therefore, crust to the east of 11°50’W is interpreted to be thinned continental crust with the OCT, which Whitmarsh et al. (1993) say is not associated with the L reflection hypothesis (Mauffret et al., 1989), located at between 12°W and 11°15’W (Whitmarsh et al., 1993).

To the south of the Horseshoe abyssal plain the location of the OCT is constrained to the west by the M-25 magnetic spreading anomaly at 12°30’W (Klitgord & Schouten, 1986) and may be represented by the S1 magnetic anomaly which is one of a group of anomalies on the African margin (figure 2.17). They are unlikely to be spreading anomalies which have similar amplitude reversed and normal periods, as they show lower negative than positive amplitudes. Seismic experiments have shown that the S2 to S4 anomalies to the east of S1 lie on continental lithosphere (Roeser, 1982), but the S1 anomaly located (Roest et al., 1992) at approximately 10°30’W, is oblique to the African margin and maybe related to initial rifting of AF-NA (Roeser, 1982).
Chapter 3

Data Acquisition and Processing

3.1 Charles Darwin Cruises 64 and 82

This chapter describes marine geophysical data acquired during R.R.S. Charles Darwin cruises CD64 and CD82 to the Azores-Gibraltar plate boundary region in 1991 and 1993. During CD64 over 9000 km of 3.5 kHz and 12 kHz bathymetry, magnetic, gravity, seismic reflection and GLORIA Mark II sidescan sonar data were acquired. Only data from, west of Gibraltar (-8 W), were used in this study (figure 3.1), with data to the east of Gibraltar analysed in the study of Willet (1996). CD82 was primarily concerned with a geophysical survey of the Canary Islands (Dalwood, 1996). During transit to the Canaries a single line of gravity, magnetic and bathymetry data, both high resolution EM12 Simrad swath bathymetry and 3.5 kHz and 12 kHz were collected across Gorringe Bank region.

3.2 Navigation

Navigation during cruises CD64 and 82 was based on the Global Positioning System (G.P.S.) (Dixon, 1991), which has now become the standard source of accurate position fixing for marine geophysical surveys, Transit satellites and a Trimble 4000AX receiver. G.P.S. position fixing relies on the measurement of signal travel times from a minimum of two satellites from which the distance to each is derived and the position calculated from these distances. Positioning from transit satellites is calculated from measurement of the doppler shift in the transmitted signal, caused by the relative motion of the satellite to the receiver. The Trimble 4000AX G.P.S. receiver currently achieves an accuracy of between 30 and 40 metres (about 0.015’ to 0.02’ latitude and longitude) (pers. comm. R. Lloyd 1993). To account for these inaccuracies the position data are averaged using a moving window with length approximately 5 to 10 minutes. The length of the window is dependent on
3.3 Bathymetry

Bathymetry data were acquired on CD64 and CD82 by 3.5 kHz and 12 kHz echo sounders. The 3.5 kHz fishes were towed behind the ship and the 12 kHz echo sounders were located on the hull. Transducers emit high frequency acoustic beams in the form of a cone having a circular footprint with a diameter of greater than 1 km in 3 km water depth. These frequencies are highly reflective with little sub-seafloor penetration making them ideal for depth sounding. The 3.5 kHz signal, however,
shows some sub-seafloor penetration of up to approximately 70 metres, dependent upon seafloor material properties, and can provide shallow structural information. Reflected acoustic energy is recorded by a receiver and the depth calculated from the time lapse between the transmission and reception of this energy (Johnson & Helferty, 1990) and an assumed velocity of sound in seawater. The reflections from the 3.5 kHz sounder were recorded on electro sensitive paper. The reflections from the 12 kHz sounders were recorded on paper and the calculated depths were merged with the smoothed navigation and digitally recorded.

![Histograms showing cross-over errors and statistics](image)

**Figure 3.2:** Bathymetry cross-over errors and statistics for G.M.T. data from the Gorringe Bank region.

The Generic Mapping Tools (G.M.T.) software package (Wessel & Smith, 1991), was used to "de-spike" the data and check for "rogue points". The de-spiked data were combined with bathymetry data from cruises in the G.M.T. database (Wessel & Smith, 1991), to provide high density data coverage over the region. The G.M.T. database contains the gravity, magnetic and bathymetry data for currently 2456 cruises from the years 1953 to 1993. Cross-over error (COE) analysis (Wessel & Watts, 1988) of the data in the Gorringe Bank region show a mean internal
cross-over error (ICOE) of -0.87 metres and a standard deviation of 133.32 metres. External cross-over errors (ECOE) have a mean of 1.61 metres and a standard deviation of 122.95 metres (figure 3.2). These errors are primarily due to the high bathymetric gradients of the region combined with navigation system inaccuracies. Additional errors may be the result of incorrect digitising of paper records on old surveys which dominate the database and incorrect estimations of seawater velocity (Smith, 1993). The combined data were gridded over a five minute grid, surfaced using a minimum curvature gridding algorithm (Smith & Wessel, 1990) and displayed (figure 3.3).

3.4 Magnetics

Total magnetic field intensity during CD64 and CD82 was measured by a Varion proton precision magnetometer which has a sensitivity of between 0.1 nT and 0.01 nT and is not sensitive to orientation (Dobrin & Savit, 1988). The magnetometer
was housed in a fish towed at a distance of between 200 and 300 metres to reduce interference from ship produced magnetic fields, from the port side of the ship on CD64 and from the starboard side on CD82. The data were logged, merged with the smoothed navigation and manually de-spiked.

Figure 3.4: Comparison of mean magnetic anomaly versus year for G.M.T. and N.G.D.C. data within quiet zone regions. Jurassic quiet zone (32° to 34° N, 13° to 11° W) and Cretaceous quiet zone (30° to 32° N, 24° to 22° W). Square brackets show number of N.G.D.C. cruises and round brackets show number of G.M.T. cruises.

Both the dipole and non-dipole components of the Earth’s magnetic field change with time. These changes comprise magnetic storms, diurnal variations and longer period secular variations. Magnetic storms have rapid magnetic oscillations of up to 1000 nT, whilst diurnal variations commonly have a period of one day with changes on the order of 25 nT. Secular variations, however, have a period on
the order of years and may have variable rates of change up to approximately \( \pm 200 \text{ nT/yr} \), but in general more typically \( \pm 30 \text{ nT/yr} \) depending on location and year (Merrill & McElhinny, 1983). Before constructing maps of magnetic data, it is important to calibrate all total field data which has been collected at different times and under different magnetic fields. This is achieved by reducing the data to a calculated model of the field for the period of data collection. One such model is the International Geomagnetic Reference Field (I.G.R.F.) (Barraclough, 1988), which is a model of the geomagnetic field based on spherical harmonic expansions of the Earth’s dipole field, incorporating a secular variation model for the prediction of future fields. This model of the dipole field only explains approximately 90\% of the total earth field, so the reduced anomalies will only approximate the true field.

The CD64 and CD82 magnetic data were combined with magnetic data from the National Geophysical Data Center (N.G.D.C.) database rather than the G.M.T. database. The total field N.G.D.C. data were all calibrated with the 5th generation of the 1987 I.G.R.F. (for the epoch 1940 - 1995). The magnetic anomaly data in the G.M.T. database have been reduced to I.G.R.F.’s for different epochs, resulting
in inconsistencies. This variation can be seen by plotting mean magnetic anomaly versus year for data within a Jurassic quiet zone (32° to 34°N, 13° to 11°W) and Cretaceous quiet zone (30° to 32°N, 24° to 22°W) region (figure 3.4). A large discrepancy would not be expected within a quiet zone and the N.G.D.C. data show a linear trend when compared with the G.M.T. data.

Figure 3.6: Magnetic anomalies of the Gorringe Bank region. Light lines represent the location of ship collected magnetic data which has been reduced to I.G.R.F. 87. Contour interval 100 nT.

Cross-over error analysis (Wessel & Watts, 1988), shows the mean COE’s for CD64 and CD82 with respect to the N.G.D.C. cruises to be -30.9 nT and -40.1 nT respectively. An explanation for this discrepancy is that these two cruises are the youngest in the database and that the I.G.R.F. for 1987 does not correctly model the field for 1991 and 1993. A constant shift of 30.9 nT and 40.1 nT was therefore applied to the respective cruises reducing the mean COE’s to 1.2 nT and 2.3 nT.

COE analysis of all magnetic data from the Gorringe Bank region (figure 3.5) shows mean and standard deviation of 6.35 and 24.28 nT for ICOE’s and a mean
and standard deviation of 1.03 and 32.45 nT for ECOE’s. The low magnitude of the errors is due in part to the low magnetic amplitudes within the magnetic smooth zone, but some error may be a result of inaccuracies in historical navigation systems on older surveys.

The combined magnetic data were gridded over a five minute grid, surfaced and displayed (figure 3.6) with the G.M.T. software package (Wessel & Smith, 1991) using the methods applied to the bathymetry.

### 3.5 Gravity

#### 3.5.1 Acquisition

The gravitational field was measured by Lacoste-Romberg model S40 and S84 gravimeters on CD64 and by an S84 gravimeter on CD82. Both the S40 and S84 are highly damped spring-type gravimeters and were mounted on gyro stabilised platforms (Valliant & LaCoste, 1967), designed to maintain the horizontal in a variety of sea states. During CD64 the S40 was located in the centre of the ship, the S84 in the upper photography laboratory off axis of the ship’s centre of gravity. The S84 was located in the centre of the ship on CD82.

![Figure 3.7: Comparison of S40 and S84 free-air anomalies from CD64.](image)

A Worden Portable gravimeter was used to take base station readings. The CD64 data were tied to the International Gravity Standardisation Network 1971, with land stations (adjusted to ship), at Lisbon, Portugal (980089.59 mGal) and Ponta Delgado, Azores (980116.50 mGal). The CD82 data were tied to base sta-
tions (adjusted to ship) at Barry, South Wales (981191.00 mGal) and Las Palmas, Gran Canaria (979371.77 mGal). The drift (assuming linear drift) was calculated to be 0.036 mGal/day (S40) and -0.038 mGal/day (S84) for CD64 and 0.035 mGal/day for CD82.

3.5.2 Initial Processing

The total field gravity data were corrected for instrument drift and the free-air anomaly was obtained by correcting the data for the Eötvös and latitude effects.

The Eötvös effect arises due to centrifugal accelerations which act on a moving vessel on a rotating Earth. The correction is positive when moving east, as when the outward acting centrifugal accelerations are increased the inward acting accelerations due to gravity are decreased. It is calculated using the equation (3.1) of Worzel (1959), where \( \Delta g = \text{Eötvös effect in mGal} \) and \( V = \text{velocity in knots} \), \( \phi = \text{heading in degrees} \) and \( \alpha = \text{latitude in degrees} \), derived from the smoothed G.P.S. positions.

\[
\Delta g = 7.487V \sin \phi \cos \alpha + 0.00415V^2
\] (3.1)

The latitude correction accounts for variations of gravitational acceleration with latitude due to the axial rotation of the non-spheroidal Earth, with greater acceleration at the poles. The latitude correction is calculated from the the smoothed G.P.S. positions by the International Gravity Formula (equation 3.2, where \( g(\alpha) = \text{latitude correction} \) and \( \alpha = \text{latitude, both in degrees} \) (Levallois, 1971) which is a model of the variation of gravity on an rotating ellipsoidal Earth with a flattening of 1/298.25.

\[
g(\alpha) = 978031.8(1 + 0.00518591581 \sin^2 \alpha - 5.7703876 \sin^2 2\alpha + \ldots)
\] (3.2)

Errors in the measurement of gravity at sea may result from various sources:

- Cross-coupling errors arise from horizontal accelerations, beam motions and the phase relations between them, due to movements of the ship not compensated for by the gyro stabilised platform. Residual effects such as translational horizontal movements will not be removed even when the platform is horizontal. They are the largest source of error and may be up to \( \pm 25 \text{ mGal} \), depending on the sea state and swell direction. The cross-coupling effect is removed from measurements of accelerations by accelerometers mounted orthogonally on the platform.
Non-linear drift between base stations is due to the gravimeter spring not showing purely elastic behaviour throughout the cruise, due to movement and changes in temperature.

Accurate calculation of the Eötvös and latitude corrections is dependent on precise knowledge of position, from which velocity and heading are calculated in the case of the Eötvös correction. A ship heading east to west at a latitude of 36°N and with a speed of 6 knots would experience an Eötvös effect of 36.49 mGal. An error in the heading of 1° would result in an error of 0.0055 mGal and an error in ship speed of 0.1 knots would result in an Eötvös correction which is in error by 0.9 mGal.

![Graph showing cross-over errors and statistics for S40 and S84 free-air anomalies from CD64.](image)

Figure 3.8: Internal cross-over errors and statistics for the S40 and S84 free-air anomalies from CD64.

### 3.5.3 Comparison of S40 and S84

The free-air anomaly data from the S40 and S84 gravimeters during CD64 show slight differences, so it is necessary to select either the S40 or S84 data for further
processing and interpretation. The differences are up to about ± 10 mGal, but generally less than ± 1 mGal (figure 3.7), with less oscillation of the S40 data over short wavelengths. The S40 and S84 would experience different movement characteristics due to their different locations. The S84 was mounted further from the ship's centre of gravity, so would experience a greater affect from the roll and yaw, especially during higher sea states. Comparison of the offset of the free-air anomaly from the two gravimeters with variations in bathymetry, ship heading and velocity found no direct connection. It is likely that the variation between the S40 and S84 was a result of the combination of these and other unknown factors.

Analysis of the internal cross-over errors gave a mean of -0.25 and -0.83 mGal and standard deviation of 4.2 and 4.61 mGal for S40 and S84 respectively (figure 3.8). Although histograms (figure 3.8) of the ICOE's show a narrower peak for S84 than S40, there are a number of points in the S84 data which show large COE’s. The S40 data were chosen for further processing due to their better statistics.
3.5.4 Further Processing

The CD64 (S40) and CD82 (S84) gravity data were combined with other data from the G.M.T. database (Wessel & Smith, 1991). Analysis of the gravity COE’s for the Gorringe Bank region (figure 3.9) gave ICOE’s with a mean of -1.72 mGal and standard deviation of 8.76 mGal. ECOE’s have a mean and standard deviation of 7.88 mGal and 75.46 mGal respectively. In addition to gravity errors previously reviewed, the COE’s may be in part a result of the combination of navigation system inaccuracies and the large gravity gradients of the Gorringe region. A cross-over correction, which removes the DC-shift and drift not previously corrected, by the fitting of regression lines to each cruise’s COE points in time and reducing them in a least squares sense (Wessel & Watts, 1988), was calculated and applied. DC-shift errors occur when the gravity data collected on a cruise is not accurately tied to the land stations, whilst drift errors result from inaccurate calculation of the drift, which may be non linear.
Data coverage is good (figure 3.10) except for a small region to the west of Ampere seamount (34°40’N, 14°30’W) and to the east of the Tagus abyssal plain (37°30’N, 14°W). Prior to contouring and display (figure 3.10) the data were gridded over a five minute grid and surfaced using a minimum curvature gridding algorithm (Smith & Wessel, 1990), similar to the bathymetry data.

3.6 GLORIA Acquisition and Processing

3.6.1 Acquisition

The Geological Long Range Asdic (GLORIA) was developed at the Institute of Oceanographic Sciences (IOS), Deacon Laboratories during the 1970’s. It is a side-scan sonar system that provides information on the horizontal disposition of the seafloor. The vehicle is deployed from the rear of the ship from an integral launching gantry and nose towed (figure 3.11) on a 300-400 metre heavy cable at a depth of approximately 40 to 80 metres dependent on speed (Somers & Searle, 1984), but at about 50 metres at 8 knots (Laughton, 1981). The tow depth on CD64 was probably slightly greater than 50 metres based on the survey speed of less than 7 knots.

Transducers on the port and starboard sides of the vehicle emit short sound pulses in the form of a 2° wide fan shaped beam (figure 3.12) at an angle of between
90° and 10-15°, with frequencies of approximately 6.8 kHz and 6.3 kHz respectively to minimise cross-talk from either side (Johnson & Helferty, 1990; Searle et al., 1990). Energy from these beams which impinge on the seafloor is mostly backscattered with minimal direct reflection. The degree of backscatter (echo strength) is dependent upon the incidence angle of the beam to the seafloor, the roughness of the seafloor and the acoustic impedance of the seafloor materials with respect to seawater. For example, backscatter will be greater from a fault scarp of rough outcropping crystalline rock sloping towards the vehicle, than flat lying mud sediments. Shadows may be produced behind features such as fault scarps which obscure the beams from material in their lea (Searle, 1979; Laughton, 1981; Johnson & Helferty, 1990; Mitchell et al., 1993). This variation in backscatter intensity is represented on the GLORIA image by grey shaded pixels, higher backscatter producing lighter shades (figure 3.11).

Returning energy from the seabed is recorded by voltage fluctuations in the same transducers which emit the sound pulses. In the near range almost directly below the vehicle, especially in deep water, the image is distorted due to the high incidence angle of the sound rays (Laughton, 1981). The maximum range is restricted by the beam path through the water column which is refracted towards the sea surface at low incidence angles and thus the beams fail to intersect the seafloor. The width of the swath within this range can be adjusted by altering the emission rate of sound pulses. This "pulse repetition rate" may be 20, 30 or 40 seconds giving a respective swath width of 15, 22.5 or 30 km (LeBas, 1995) with decreasing resolution. A pulse repetition rate of 30 seconds was used on CD64, giving a pixel resolution of approximately 45 m (Chavez, 1986).
3.6.2 Geometric and Radiometric Distortions

GLORIA acquisition characteristics introduce geometric and radiometric distortions into the data. Geometric distortions are related to the geometry and spatial relations of the pixels, including water column offset, anamorphic and slant range distortions. Radiometric distortions are products of the sonar beam pattern, geometrical spreading and absorption in the water.

Water column offset is a result of the commencement of recording immediately following transmission of the sound pulses and the low sampling rate, giving a time delay for the first returning energy (first arrivals) (Johnson & Helferty, 1990). This is represented on the image (figure 3.13) by an offset of the first arrivals to either side of the vehicle path by an amount related to the velocity of sound in water and the seafloor depth directly below the vehicle. A narrow band of approximately 20 out of 1024 noisy pixels are a consequence of changing from equal sampling in time to equal sampling in distance through the slant range correction. Knowledge of the depth and the velocity of sound in seawater allows the correction of the water column offset (Chavez, 1986).

Anamorphic distortions result from the difficulty of maintaining a constant vehicle speed over the seafloor, especially in poor sea states. The across track pixel resolution will be fairly constant on a flat seafloor of constant depth, but ship speed changes will cause along track variation in pixel resolution (aspect ratio) (Somers & Searle, 1984). Correction is provided by adjustments to the data with respect to the speed of the GLORIA vehicle from the ship’s G.P.S. navigation (Chavez, 1986; Johnson & Helferty, 1990).

Slant range distortions result from differences in travel time along a slanted ray-path from the vehicle to the seafloor to the horizontal range, in an inhomogeneous water column (Somers & Searle, 1984). Horizontal range is calculated by applying a slant range correction (Johnson & Helferty, 1990). This is calculated from a knowledge of the seafloor depth beneath the ship from the 6 minute echo soundings, the assumption that seafloor depths are constant either side of the ship track and that the sound pulses are of a constant velocity (1500 ms$^{-1}$ (LeBas, 1995)) and follow straight ray paths (Somers & Searle, 1984).

Spreading results from amplitude loss, which varies as the inverse square of the distance, due to seawater absorption. This reduction in intensity is receiver corrected (Somers & Searle, 1984) by the application of a time varying gain to the recorded signal. The time varying gain is an amplifying filter which changes non-linearly with time and enhances the amplitudes of the signal at greater distances, producing a signal whose amplitude is linear with time (Johnson & Helferty, 1990).
In addition the shape of the emitted beam (figure 3.12) produces a stronger mid-range signal (figure 3.13). Application of a shading correction will correct for the variation of beam shape and any gain not yet corrected, by enhancing or reducing the brightness of pixels across track (Chavez, 1986) resulting in a more even shading across the swath, giving far range features greater definition against background amplitudes.

Speckle noise is represented by anomalous noisy pixels within the image which may be the result of distortions at the edges of the beam footprint or momentary interference or signal cancellation. They may be removed by a combination of a high cut filter to identify noisy pixels and a smoothing filter to remove them. Striped noise perpendicular to the scanning direction and may be removed by a combination of high and low cut filters.

3.6.3 Processing

Initial processing occurred on-board CD64 with real time monitoring of the raw data, to access quality and the possible need for adjustment of acquisition parameters (Somers & Searle, 1984). Data were also recorded onto magnetic tape in the form of ".PIC" files for further processing. In addition the data were geometrically corrected and 35 mm negatives produced, from which 6 hour "passes" were printed (figure 3.13) and mosaiced by hand to provide an initial spatial inspection of the data.

Further processing was carried out at the Institute of Oceanographic Sciences, Deacon Laboratories utilising the Mini Image Processing System (M.I.P.S.). Initially the region mapped by GLORIA data was divided with the aid of track charts into 11 rectangular blocks with dimensions no greater than 2.5 by 2.5 degrees. Block boundaries were positioned to not cut major features of interest and to provide efficient regional coverage. The raw data (.PIC) files and the navigation file were transferred from the magnetic tapes to disk for processing.

The first step was to create a shading correction file (shade3.dat). A small featureless section of data from the Horseshoe abyssal plain which most closely reflected the regional background character was selected and geometrically corrected. The mean amplitude for each track parallel pixel column in the selected section was calculated and used to create a set of coefficients stored in the shading correction file. During the next processing step each column in the raw data was multiplied by its respective coefficient, increasing or decreasing its intensity and therefore normalising intensities to a background and removing radiometric spreading and variable beam shape effects.
Each block was processed individually under the control of the program "PROMOS" which required the input of: the block coordinates; the desired projection (Mercator, 30° standard latitude); a line increment parameter (20 km)(the length of track used to assess whether a course change has occurred greater than the maximum course deviation parameter); the maximum course deviation parameter (the angle above which a line will be divided into two separate tracks); the ".PIC" files for the particular block; an index file containing the data times and number of data lines for each ".PIC" file and the shading correction and navigation files.

Upon completion several ".GOM", geographically orientated mosaic track section, files are created, which are geometrically and radiometrically corrected. There is overlap with sections from other ".GOM" files, but not internally. The sections contained in the ".GOM" files were displayed in red, green and blue simultaneously and an electronic stencilling program was used to select the desired mosaic sections for the final image, including the removal of the edges of the GLORIA swaths which contain noisy data. In an overlap stencilling was performed to preferentially select the section exhibiting the best geological information, sometimes at the expense of a smooth intensity across the join. The final image for the block was recorded onto magnetic tape and PROMOS repeated for all other blocks. Images for the blocks were stencilled together to produce a final image of the Gorringe Bank region for display (figure 3.14).
Figure 3.14: Mosaic of processed GLORIA images from the Gorringe Bank region from CD64.
Photographic negatives were produced from the processed data by Richard Carlton at the Open University, Milton Keynes. The images were contrast stretched interactively within the Terra-Mar IDIMS image processing software system hosted on a Sun-Microsystems Sun4/360. An Optronics International Inc. Colorwrite C-4300 filmwriter was utilised to make 10” negatives on Agfa Avipan 150 PE film, which were developed with Agfa chemicals. Photographic prints at a relevant scale (4”/°) were produced from the negatives for display and interpretation.

3.7 Seismic Acquisition

Four thousand kilometres of seismic reflection data were acquired along eight profiles with an ENE-WSW orientation and two profiles with a generally NW-SE orientation (figure 3.1) providing ties to the previous lines and an alternative aspect to various features. The same technical setup (figure 3.15), which is now reviewed, was used throughout the cruise.

3.7.1 Equipment Configuration and On-board Processing

An airgun array of four Bolt 1500C airguns with chamber size 1.64, 2.62, 4.92 and 7.64 litres was utilised to give a strong source with a low peak frequency and small "tail". These source characteristics provide power to image the basement below sometimes 4-5 km sedimentary sequences in a water depth of 5 km, with a minimisation of bubble pulses due to the combination of different sized guns. A theoretical amplitude spectrum (figure 3.16) (Giles & Johnson, 1973; Hardy, 1991) of the source shows a low peak frequency of approximately 10 Hz with the majority of the energy below a frequency of 60 Hz. The three smaller airguns were towed on a boom at a probable depth of approximately 7.5 metres, estimated from the tow cable length and ship speed, and the 7.64 litre airgun was towed separately (figure 3.15). The shot interval was 16 seconds, with a ship speed of about 6 knots providing a shot spacing of 50 metres.

A 200 metre long streamer with four 50 metre sections containing groups of transducers providing a total of four channels was towed from the stern at a distance of 160 metres (figure 3.15). A longer and higher channel streamer could not be used due to potential interference with the GLORIA survey. Initially the tow depth was estimated to be 33 metres based on calculation of the geometry in reverse from the seafloor multiple. The tow rope was shortened during acquisition in an attempt to raise the streamer to a depth of between 10 and 20m, but with limited success.

Synchronised airgun firing and recording was operated by the personal computer
Figure 3.15: Configuration of Geophysical Equipment on CD64.
based Seismic Acquisition System (SAQ) (Owen & Sinha, 1990). A shot interval of 16 seconds and six seconds of data, with a sample interval of 2 ms (corresponding to a nyquist frequency of 250 Hz (equation 3.3, where \(|f| = \text{Nyquist frequency in Hertz and } d = \text{the sampling interval in milliseconds})\)) were recorded onto 2400 foot half inch 1600 pbi magnetic spool tapes in SEG-Y format (Barry et al., 1975). SEG-Y is trace sequential (Demultiplexed) and for each tape consists of a tape header followed by individual traces with trace headers. An anti-alias filter (125 Hz-24Db) was applied to the data prior to recording as aliasing will occur if the data frequency is greater than the nyquist frequency. A few traces of data were not recorded whilst loading spool tapes. A deep water delay was added to reduce the quantity of water column information recorded in deep water regions and was adjusted according to water depth throughout acquisition.

\[|f| = \frac{1}{2d}\]  

(3.3)
3.8 Seismic Processing

Standard methods (figure 3.17) were applied to process the raw data (figure 3.18) (e.g., Yilmaz, 1987). The SIOSEIS software package (Scripps Institute of Oceanography) was used during stacking, deconvolution and filtering of the data. FK filtering and Migration were performed utilising the PROMAX software package which was acquired after initial processing, all supported by Sun Sparc workstations at Oxford university.

Division of the survey was based on the eight ENE-WSW lines and the line at the end of the cruise was divided into two based on a gap in the data on the southwestern flank of Gorringe Bank (figure 3.1).
Figure 3.18: Single trace monitor from line 3. Plot parameters: Clip = 1.3, scalar = 1000, bias = -10.
3.8.1 Gather and Common Depth Point Stacking

Field tapes were sequentially read and sorted from common shot point to common mid point gathers on the basis of the acquisition geometry. The first receiver group was centred 200 metres behind the shot point with three others following at intervals of 50 metres. Shot spacing was assigned to be 50 metres giving a 25 metre distance between common mid-points. A move out correction was not necessary, as the short streamer gave no observable trace offset. Traces within a 10 metre smear distance were stacked for each reflection point giving 2 fold data and resampled to 4 ms (Nyquist 125 Hz). Stacked data (figure 3.19) for each field tape were assembled into the ten previously selected lines with trace gaps between field tapes padded with dead traces to retain spatial definition along the lines. Gaps were of about 50 common mid points corresponding to 1.25 km and were not large enough to significantly interfere with reflection interpretation.

Deep water delays were assigned in the trace headers to be in integer seconds, but correction for these was complicated by true delays of fractional seconds. These are due to the non synchronous trigger and firing of the guns of approximately 0.05 seconds. The data were shifted to correct for this.

3.8.2 Frequency-Wavenumber (FK) Filtering

FK filtering operates in the frequency-wavenumber domain and is able to remove low velocity coherent noise.

FK analysis (figure 3.20), which plots frequency versus wavenumber for a section of data aided in the selection of a "pie slice" with a velocity of 2500. Test panels of an FK filter applied to the stacked data (figure 3.21) reveal that small variations in the velocity do not have a great effect on the data, but in general the signal to noise improvement is dramatic (figure 3.22).

3.8.3 Predictive Deconvolution

A seismic signal can expressed as a convolution of the source signal with the seismic acquisition equipment and the earth. The earth can produce some undesirable effects such as attenuation, reverberation and multiples. Deconvolution is a method by which repeated events such as bubble pulses may be removed from seismic data. Predictive deconvolution attempts to predict, given a time series, the next repeated sequence in the series. It is assumed that the noise to be removed follows a repeated pattern with a linear offset or lag. The magnitude of this lag can be derived from an auto-correlation which gives a quantitative estimate of the similarity between a
Figure 3.19: Stacked section from line 3. Plot parameters: Clip = 1.3, scalar = 1000, bias = -10.
time series and a shifted copy of itself. Application of an auto-correlation function to the data from an abyssal plain section of line 3 with flat lying reflections gave a value of approximately 70 ms for the correlation lag for the deconvolution function. This was based on the repetitive reverberation from the upper second of data below the seafloor, caused by bubble pulses.

A small section of line 3 was deconvolved with various predictive lags from 0.06 to 0.18 seconds (figure 3.23) and various deconvolution filter lengths of between 0.1 and 0.75 seconds (figure 3.24).

Predictive distance or lag variation (figure 3.23) has a significant effect on the results. Selection of a predictive distance close to 100 ms gives greater reduction of the seafloor bubble pulse. And an auto-correlation of the deconvolved section reveal the section with an applied lag of 100 ms has the least residual repetition.

Filter length variation (figure 3.24) is not such an important factor as predictive distance variation. A filter length of 500 ms gave the best result with the greatest bubble pulse reduction and the corresponding residual auto-correlation shows lower repeated events.

A prewhitening of 0.1 percent was applied to add a lag to the zero lag of the auto correlation. Variation of the percentage of prewhitening did not significantly
Figure 3.21: FK-Filter test panels showing the results of variation of the pie slice velocity. Plot parameters: Clip = 1.3, scalar = 1500, bias = -10.
Figure 3.22: FK-Filtered section from line 3. Pie slice velocity = 2500. Plot parameters: Clip = 1.3, scalar = 1000, bias = -10.
Figure 3.23: Deconvolution test panels. Upper figures show the results of variations of the predictive distance with a fixed filter length of 0.5 s. Lower figures show the results of an autocorrelation on the processed data. Upper plot parameters: Clip = 1.3, scalar = 1000, bias = -10. Lower plot parameters: Clip = 1.3, scalar = 1, bias = -2.
Figure 3.24: Deconvolution test panels. Upper figures show the results of variations of the filter length with a fixed predictive distance of 0.1 s. Lower figures show the results of an autocorrelation on the processed data. Upper plot parameters: Clip = 1.3, scalar = 1000, bias = -10. Lower plot parameters: Clip = 1.3, scalar = 1, bias = -2.
Figure 3.25: Deconvolved section from CD64 - line 3. Predictive distance = 0.1 s. Filter length = 0.5 s. Plot parameters: Clip = 1.3, scalar = 500, bias = -10.
affect the result (figure 3.25).

3.8.4 Post Stack Frequency Filtering

![Frequency spectrum](image)

Figure 3.26: Frequency spectrum of seismic data from line 3 of CD64.

Band Pass filters allow the selection of desired frequencies between certain frequency bands. Low frequencies have greater penetration and are more powerful for imaging of deep structure, but will give low resolution. Higher frequencies will give a high resolution for imaging shallow structure.

Filter selection was based on the frequency spectra of a typical line 3 section (figure 3.26); the model of the source frequency spectra (figure 3.16); the results of test panels for various pass bands (figure 3.27) and on the depth of the features of interest.

The majority of the signal is below 40 Hz, but frequencies up to 60 Hz were preserved to retain high frequency information for greater resolution of the structure in the upper section. The water column noise has a frequency of less than
Figure 3.27: Frequency filter test panels. Frequency pass bands in steps of 10 Hz. Plot parameters: Clip = 1.3, scalar = 1200000, bias = -10.
Figure 3.28: Frequency filtered section from CD64 - line 3. Pass band 20 - 60 Hz. Plot parameters: Clip = 1.3, scalar = 1500000, bias = -10.
Figure 3.29: Migrated section from CD64 - line 3 with top mute. Migration velocity 1500 ms$^{-1}$. Plot parameters: Clip = 1.3, scalar = 400000000, bias = -10.
20 Hz and is not related to the small peak in frequencies beyond 70 Hz. All data were processed using a zero phase bandpass filter of 20-60 Hz, a length of 75 points and a slope of 30 decibels per octave for general imaging of the shallow structure (figure 3.28).

### 3.8.5 Post Stack Time Migration

Migration converts time series seismic data to give true dip and position to features. A knowledge of the velocity of all points at depth is required, as each point is repositioned based on its apparent angle and the velocity of the material above. This is unrealistic, but in practice using a rough estimation of the velocity structure may more clearly reveal the true structure. The method assumes that diffractions come from point sources and so a correct model of the velocity structure will result in a transform which collapses the diffractions.

The absence of accurate velocity information makes it impossible to accurately migrate the data from CD64. Migration was utilised to more clearly image features which were obscured by diffraction hyperbolae by collapsing them using a Stolt F-K Migration (Stolt, 1978) at a seawater velocity of 1500 ms$^{-1}$ (figure 3.29). The Stolt F-K Migration is rapid and only allows the use of a single migration velocity, so is ideal for the application required in this study. It involves a Fourier transform from the T-X (time-space) domain to the F-K (frequency-wavenumber) domain with mapping using the single velocity. The migration involves an inverse Fourier transform to move from the F-K to T-X Domain. This scheme proved to be very effective despite the lack of velocity information.

### 3.8.6 Display

Display was performed using the SIOSEIS package combined with UNIRAS 6v2a. with plotting of complete lines on a Versatec 24” plotter. Variable amplitude plots, with traces scaled to absolute values to preserve the trace to trace amplitude relationships, were found to be the most effective for interpretation and for display following each processing step.
Chapter 4

Sediment and Basin Structure

4.1 Seismic Facies and Basin Structure

On the basis of their seismic character, three major depocentres have been identified in the Gorringe Bank region. These include the Tagus Basin, Horseshoe Basin and Seine Basin, corresponding to their respective abyssal plains. Their individual character and sedimentation is a result of their isolation by the many ridges and seamounts (figure 4.1). The Horseshoe Basin is further subdivided into the Western Horseshoe Basin and Eastern Horseshoe Basin on the basis of their differing seismic character and a bathymetric narrowing between the two sub-basins where sediments thin.

Each basin has been interpreted independently as the identification of reflections across basin boundaries is complicated by the diversity of reflection style, as commented on by Purdy (1975). Seismic facies or units are defined on the basis of changes in reflection style including continuity, wavelength and amplitude (Mitchum et al., 1977), the recognition of reflection terminations and regional fill style of the unit. The basement structure and basin boundary style is described and probable ages and correlations between units from various basins made.

4.1.1 The Tagus Basin

The Tagus Basin was only imaged by a single seismic reflection line from CD64 (figure 3.1), although examination of the seismic reflection data from Mauffret et al. (1989) aided interpretation. The sediments have been subdivided into six seismic facies on the basis of reflection character and terminations shown on profile A-B (figure 4.2).
Figure 4.1: Location of seismic reflection profiles from CD64 and IAM project (Banda & Torne, 1995) presented in chapters 4, 5 and 6.

**Unit T-I** is the shallowest unit and exhibits medium amplitude, medium to short wavelength continuous parallel "striped" reflections. The unit was described by Mauffret et al. (1989) as a sheet drape (Mitchum et al., 1977) and although the unit contains a significant proportion of pelagic material, based on sampling at DSDP site 135 (Hayes et al., 1972), the major components of the unit are turbidites. These were deposited unconformably, with many low angle reflection terminations towards the ENE and WSW, upon an uneven surface, giving a variable thickness with a mean of about 0.85 seconds TWTT. Deformation is minimal with slight anticlinal folding, of the parallel conformable reflections, related to the upper extent of basement faults which terminate in the sub-seafloor.

**Unit T-II** is highly disturbed with many diffractions and medium amplitude semi-continuous hummocky reflections which lie unconformably upon an uneven lower surface. The unit is interpreted as one or more debris flows which were derived from Gorringe Bank. A decrease in thickness on Profile A-B (figure 4.2) from
Tagus Abyssal Plain

Unit T-I
Unit T-II
Unit T-III
Unit T-IV
Unit T-V
Unit T-VI
Basement

AB1

A

WSW

EN

TWTT (s)

0.0
5.0
10.0

KM

F

IAM 4
Figure 4.3: Extent of the olistostrome unit EH-II and debris flow unit T-II. Constrained by seismic reflection data from CD64, Lusitanie cruise 86 (Mauffret et al., 1989), the Iberian Atlantic Margin project (Banda & Torne, 1995) and R/V OGS-Explora 92 (Sartori et al., 1994).

0.3 seconds TWTT at the ENE end of the line to less than 0.1 seconds TWTT towards the WSW, suggests thinning away from the Iberian margin. However, correlation with seismic reflection data of Mauffret et al. (1989); Sartori et al. (1994) (figure 2.7) shows that the unit’s thickness also decreases with increasing distance from Gorringe Bank and that it has an extent which is probably constrained to a linear belt parallel to the base of Gorringe Bank (figure 4.3). Deposition of unit T-II may have occurred above the base of unit T-I, with a very thin layer of unit T-I beneath, yet obscured by diffractions from T-II. Where T-II is absent on the western flank of the Tagus Basin, T-I unconformably overlies T-III.

**Unit T-III** has a surface which is marked by a pair of semi-continuous medium-high amplitude reflections, below which low amplitude, semi-continuous, parallel reflections dominate. The thickness of the unit is about 0.45 seconds TWTT along
the majority of the profile, with thinning towards the ENE along the line due to slight basement controlled slumping and possibly to differential erosion during the period of non-deposition analogous to the unconformity between T-I and T-III. At the base, reflections onlap at a low angle apparently towards the ESE onto the top of unit T-IV.

**Unit T-IV** is marked at the top by a change to high amplitude, continuous reflections which dominate for the upper approximately 0.2 seconds TWTT. Lower reflections show a decrease to medium amplitude with onlap onto the top of unit T-V and basement highs. The unit has a near constant thickness of about 0.4 seconds TWTT, but thins and shallows slightly towards basement highs.

**Unit T-V** has a somewhat similar character to unit T-III with low amplitude, semi-continuous, parallel reflections which show onlap onto the top of unit T-VI and basement highs. The unit shows a high thickness variation of between 0.45 and 0.15 seconds TWTT, with substantial thinning associated with faulting and basement highs.

**Unit T-VI** is the deepest unit identified in the Tagus Basin, with low amplitude, semi-broken reflections and a high thickness variation due to the large undulations of the block faulted basement. Although no growth associated with basement block faults is clearly observed, the reflection character and basement trough fill style suggest that the unit may represent a synrift deposit.

**Basement** The basement is characterised by low amplitude, broken reflections and a high density of diffractions marking the crests of basement highs which have a huge relief of up to 1.7 seconds TWTT. Basement features which have affected the deposition of the deeper sedimentary units are probably fault blocks, but the basement feature AB1 (Profile A-B, figure 4.2) is possibly of diapiric origin due to the folding of reflections above and to the sides of the steep sided feature.

**Boundary Style** The style of the basin boundaries are poorly constrained, but deep lying boundary faults which confine the lower T-III to T-VI units to the deeper regions of the Tagus Basin are responsible for basement uplift on the WSW edge of the basin. Units T-I and perhaps T-III form a drape over the flanks of the basin and over the northern flank of Hirondelle seamount indicating that the bathymetry pre-dates these units. A thin veneer of the deeper units may exist
on these basement highs, with the majority of deposition, controlled by basement block faulting, in the central Tagus Basin. The ENE edge of the basin has thick margin sediments of up to 3 seconds TWTT which were difficult to sub-divide due to the folding, faulting and many reflection terminations relating to a complex layered structure.

4.1.2 The Western Horseshoe Basin

Six seismic facies were also identified in the Western Horseshoe Basin, although probably not all directly related to the six units of the Tagus Basin. Three seismic lines (lines 2, 3 and 4; figure 3.1) providing ENE-WSW coverage and line 10 providing ties enable the seismic facies in this region to be interpreted with confidence, as shown on profile C-D (figure 4.4).

**Unit WH-I** is characterised by low amplitude semi-continuous reflections which parallel the very flat seafloor. With increasing depth, reflections become more continuous and of a slightly higher amplitude giving a striped appearance. The unit has a fairly constant thickness of about 0.7 seconds TWTT, but thins gradually towards the west to less than 0.5 seconds TWTT with onlap at the base. The unit shows slight deformation due to upward extensions of basement faults which have introduced minor displacements into shallow depths.

**Unit WH-II** has a surface which is marked by a distinctive pair of continuous reflections composed of a medium-high amplitude reflection and high amplitude long wavelength reflection immediately below. Overall the unit is of similar character to WH-I with flat lying reflections and thinning towards the west. However reflections are more continuous and of slightly higher amplitude. The maximum thickness of the unit is approximately 0.3 seconds TWTT, but variable due to the infilling of fold troughs formed in the lower units and substantial thinning to approximately 0.2 seconds TWTT associated with onlap onto axial fold highs. Unit WH-II is only slightly affected by the folding, the majority of the deformation occurring prior to deposition of WH-II and ceasing before the deposition of WH-I. The unit also shows greater offset by basement faulting which is observed to terminate in WH-I.

**Unit WH-III** has a surface which is marked by a pair of gently folded high amplitude, very continuous reflections, below which medium amplitude, semi-continuous reflections grow in continuity with depth. The almost constant thickness of 0.35 seconds TWTT shows near uniform, long wavelength low amplitude folding. The
unit’s base is marked by a pair of high amplitude very continuous reflections which show an uneven surface with low angle onlap onto fold limbs.

**Unit WH-IV** shows slightly more continuous reflections than WH-III and a variable thickness due to the control of basement block faulting, but generally of about 0.4 seconds TWTT. The unit lies unconformably upon WH-V with onlap onto the fault associated fold troughs and axial highs of the lower unit.

**Unit WH-V** has a surface which is characterised by medium amplitude continuous reflections, with decreasing amplitude at depth to a zone of low amplitude semi-continuous reflections which exhibit folding. The maximum thickness of the unit is approximately 0.4 seconds TWTT, but highly variable due to deposition under the control of basement block faulting onto which reflections onlap. Toplap against the base of WH-IV indicates that deformation and erosion occurred before the deposition of WH-IV.

**Unit WH-VI** has a surface which is defined in most regions by a series of broken high amplitude reflections which decrease in amplitude and continuity with depth. The unit’s thickness is highly variable due to the infilling of basement troughs, but estimated to approach 1 second TWTT. Reflection character and fill style suggest that the unit represents a synrift deposit, although no growth was observed against faults. The primary reflection may be of negative polarity compared to the water-sediment interface, perhaps suggesting a negative impedance contrast and anomalous material with a lower velocity such as low velocity clays or sands. Salt associated with initial rifting (Pautot et al., 1970) could perhaps be responsible, but the Western Horseshoe Basin is probably at too great a distance from the margin for rifting associated salt to be deposited.

**Basement** The basement surface is poorly imaged in deeper regions of the basin, but where recognised is characterised by medium amplitude broken hummocky reflections and a high density of diffractions. The massive basement relief is greater than that observed in the Tagus Basin, with basement block faults of up to 2 seconds TWTT, but generally on the scale of about 1 second TWTT.

Block faulting has influenced sediment deposition with onlap of WH-III, IV and V, sometimes with small drag folds, onto faults associated with some of the larger blocks. Although the degree of control has decreased over time, even WH-I has been thickened under the control of the largest basement block faults with faulting
from deep levels effecting the deposition of the shallow layers (figure 4.5). The faults do not break the surface, perhaps indicating that basement control has now ceased or is only a small factor in further development of the Western Horseshoe Basin. This is also illustrated by a small half graben, with basement relief on the order of 1 second TWTT, showing growth in units below WH-I west of the centre of the Western Horseshoe Basin (figure 4.6). Here where sediment thicknesses are less than in the central Western Horseshoe Basin, there are magnetic lows of up to -90 nT associated with the half grabens opposed to +15 nT over the shallower basement between.

**Boundary Style**  The boundary style of the Western Horseshoe Basin is variable. To the east the boundary between the Western and Eastern Horseshoe Basins is represented by sediment thinning upon an uplifted and highly faulted basement region, with all units from WH-I to at least WH-IV overlying basement uplifts. To the northeast the boundary style exhibits classic steeply dipping basin bounding normal faults (figure 4.5) which have been active throughout basin formation with a great influence on basin development. Their length is greater than 3 seconds TWTT with an apparent fault gradient decrease with depth which may be due to a velocity pull up in the seismic data. They constrain all units of the Western Horseshoe Basin, except for WH-I and WH-II which overly fault crests, with onlap and drag folding of reflections onto faults. There are negative magnetic anomalies of -150 to -270 nT on the downthrown side of faults and free-air gravity anomalies which mimic the bathymetry, which may indicate that the structures are uncompensated. The magnitude of the faulting decreases to the north along the edge of the basin and around the northern to western edges the boundary faults are absent. In these regions units WH-I to at least WH-V form a steadily thinning drape away from the central basin, with the deposition of units WH-II and below affected and isolated by the block faulted basement. The style of the southern boundary is poorly constrained, but appears from an ENE-WSW profile running parallel with the boundary to be block faulted on a large scale with high amplitude, long wavelength folding and faulting of the sediments.

### 4.1.3 The Eastern Horseshoe Basin

Five units have been identified in the Eastern Horseshoe Basin based on reflection character and terminations. Three seismic lines (lines 4, 5, and 8; figure 3.1) provide ENE-WSW coverage with ties from lines 9 and 10. The structure, seismic character and units of the Eastern Horseshoe Basin are shown on profile J-K (figure 4.7).
Eastern Horseshoe Abyssal Plain

Unit EH-IA
Unit EH-IB
Unit EH-II
Unit EH-III
Unit EH-IV
Unit EH-V

Basement

NNW  SSE
Unit EH-I has been divided into 2 sub units. The upper unit EH-IA is characterised by low amplitude, continuous parallel reflections and EH-IB by medium amplitude, highly continuous stripy reflections, both units showing onlap at low angles towards the edges of the basin. The thickness of EH-I is quite variable within the basin with a maximum thickness of about 0.9 seconds TWTT (Unit EH-IA approximately 0.55 seconds TWTT) and thinning towards the edges of the basin. The base of EH-I is marked by a distinctive pair of continuous reflections with a medium-high amplitude reflection and high amplitude long wavelength reflection immediately below, similar to the base of WH-I in the Western Horseshoe Basin.

Unit EH-II consists of a series of medium amplitude hummocky reflections perhaps with some chaotic layering which are masked by a high density of diffractions. The maximum thickness is about 1 second TWTT to the east, but thins towards the west and pinches out by the western edge of the basin (figure 4.3). The unit is broadly folded on ENE-WSW profiles, but shows more intense folding on N-S orientated profiles, indicating that deformation follows a roughly N-S trend, probably related to the direction of regional compression (e.g., Grimison & Chen, 1988; Argus et al., 1989; Udias & Buñorn, 1991). The unit represents the olistostrome unit interpreted by Bonnin et al. (1975); Auzende et al. (1981) to have been derived from the Straits of Gibraltar during the Middle to Late Miocene (15-5 Ma). A plan of the extent of EH-II (figure 4.3), constructed from available seismic reflection data (Sartori et al., 1994; Banda & Torne, 1995) reveals the rough shape of the Eastern Horseshoe abyssal plain at the time of EH-II deposition to be similar to the present shape. However, the unit appears to cut across the base of Gettysburg seamount and the western end of the unit appears to have a NNE-SSW trend similar to the strike-slip fault identified by Purdy (1974) between Gettysburg and Hirondelle seamounts. Where EH-II is absent, towards the west of the Eastern Horseshoe Basin, EH-I rests upon EH-III.

Unit EH-III shows low amplitude, semi-continuous reflections and a thickness of approximately 0.7 seconds TWTT which decreases towards basin edges. Reflections at the base of the unit onlap at low angles onto the top of EH-IV.

Unit EH-IV has a surface which is marked by a series of approximately five parallel high amplitude continuous reflections. Amplitude rapidly declines with depth to semi-continuous, low amplitude reflections which terminate with onlap and thinning, from about 0.6 seconds TWTT to 0.2 seconds TWTT, onto basement
highs.

**Unit EH-V** consists of medium amplitude broken hummocky reflections and a top which is not always clearly imaged. The unit has a thickness which is probably greater than 0.5 seconds TWTT, but the base of the unit is poorly imaged.

**Basement** The basement is poorly imaged due to the great sediment thickness, with the best imaging associated with basement highs which have a relief of about 1 second TWTT and show highly diffracted and broken hummocky reflections. Although deeper units EH-III, IV and V of the Eastern Horseshoe Basin have been folded and faulted, it appears that there has been little basement control in contrast to the Western Horseshoe Basin. This may represent a change of crustal nature between the Eastern and Western Horseshoe Basins as interpreted (Purdy, 1975) from refraction models. These showed different crustal layer thicknesses and velocities for the Eastern and Western Horseshoe Basins, assumed to be a result of tectonic alteration of the crustal structure in the Eastern Horseshoe Basin with respect to the shallow seismicity in this region (e.g., Grimison & Chen, 1988).

**Boundary Style** As observed in the Western Horseshoe Basin, the style of the basin boundary in the Eastern Horseshoe Basin exhibits great variety. The northern boundary is characterised by the continuation of EH-I to probably EH-V, with the exception of EH-II, onto the southern flank of Gorringe Bank, indicating deposited before uplift. At the eastern edge of the basin the units all thin towards the Iberian margin where there is an abrupt change in bathymetry from the Eastern Horseshoe abyssal plain to a comparatively steep continental slope. Faulting at the slope break appears to be normal, but diffractions prevent a detailed interpretation. The southern edge of the basin is bound by large scale ENE-WSW striking thrust faults. EH-I to at least EH-IV continue south onto Coral Patch Ridge, but are offset from the Eastern Horseshoe Basin by the faults. However, EH-II does not appear to transgress the faults, indicating that some uplift in the region of Coral Patch Ridge preceded deposition of this unit. The southwestern corner of the Eastern Horseshoe Basin shows basin boundary normal faulting similar to the Western Horseshoe Basin, yet of lower magnitude. All units are truncated by the faulting which continues to be active in development of this area of the basin. EH-I and a thin sequence of unclassified older sediments continue to the WSW of the boundary faults with thinning onto Ampere seamount. The western end of the basin moves into the deformed region separating the Eastern and Western Horseshoe Basins.
and appears to contain units from EH-I to perhaps EH-IV, with the exclusion of EH-II.

### 4.1.4 The Seine Basin

The structure and style of the Seine Basin is poorly constrained, with only a single seismic line (line 6; figure 3.1) imaging the basin and subdivision at depth complicated by the structural style, lack of data and great sediment thickness. Profile L-M (figure 4.8) shows the three seismic units which have been defined and the structure and style of the northern Seine Basin.

**Unit S-I** has been subdivided into two sub-units. The upper subunit S-IA is of similar character to EH-I of the Eastern Horseshoe Basin with semi-continuous low amplitude reflections which become medium amplitude continuous stripy reflections towards the base. The thickness of the S-IA decreases towards the ENE and WSW of line 6 (figure 3.1) from a maximum of about 0.9 seconds TWTT to a minimum of about 0.5 seconds TWTT. Subunit S-IB was only imaged in the deepest regions of the basin, near to the centre of line 6 and shows a continuation of medium amplitude stripy reflections. The subunit has a maximum thickness of about 0.5 seconds TWTT and thins towards the ENE and WSW with low angle basal onlap of S-IB (S-IA where S-IB is absent). The whole of unit S-I has been affected by basement associated faulting.

**Unit S-II** has a surface which is marked by two continuous high amplitude, long wavelength reflections. The unit is characterised by semi-continuous slightly hummocky, medium amplitude reflections and has a variable thickness due to folding and quite intense faulting with offsets of up to 0.3 seconds TWTT. The thickness of the unit in unfaulted regions is about 0.4 seconds TWTT, but thickening of up to 0.4 seconds TWTT is associated with faulted regions. The unit’s base is marked by a pair of high amplitude continuous reflections, which are also faulted.

**Unit S-III** encompasses all sediments beneath S-II and is characterised by medium-low amplitude, semi-continuous reflections which are disrupted by diffractions related to faulting and diapirism. The maximum thickness of the unit is approximately 1 second TWTT, yet highly variable due to the basement relief.

**Basement** is poorly imaged in regions of great sediment thickness, but where the sediment thins to the ENE and WSW of line 6, consists of medium amplitude,
semi-continuous hummocky reflections as observed in other basins of the Gorringe Bank region. Considerable sediment disruption in the northwestern Seine Basin is a result of diapirs which appear to originate from a shallow depth. They are unlikely to be related to salt deposited during initial rifting of the EU-NA and AF-NA plates (Pautot et al., 1970) as this would originate from great depths and would not be expected at this distance from the Iberian margin. They may represent shallow, low density mud layers, perhaps activated by compression associated deformation.

**Boundary Style** The boundary style of the Seine Basin is not clearly constrained by seismic reflection data. The single line which images this region shows the WSW edge of the basin to be normally faulted with small displacements. Units S-I to S-III form a drape which continues across the boundary with thinning onto the block faulted basement. The ENE edge of the basin is characterised by a fairly continuous drape of units S-I to S-III onto Coral Patch Ridge, with some deformation, but no boundary faulting.

### 4.2 Age, Correlation and Lithologies

The age and composition of sediments in the Gorringe Bank region are poorly constrained due to a lack of detailed sampling. Although DSDP sites 120 (table 4.2) and 135 (figures 4.10, 4.9 and table 4.1) provide an indication of age and lithology for sediments on bathymetric highs, they do not contain a complete stratigraphic record for the basins.

DSDP 135 (table 4.1) shows an upper unit of pelagic chalk ooze which was deposited above a major unconformity between DSDP 135 unit 1 (DS135-1) and unit 2 (DS135-2). This unconformity represents an interval of Early Eocene to Late Oligocene and a lithological change to mudstones and clays which has resulted in an acoustic impedance which is clearly represented on seismic reflection data. Profile N-O (figure 4.10) skirts the location of DSDP site 135 (figure 4.1) on Coral Patch Ridge and shows a clear change in reflection style with a pair of high amplitude continuous reflections marking the transition. The transitions between DS135-2, 3 and 4 are not easily correlated with reflections due to the similar nature of their composition producing lower impedance contrasts across boundaries. However, a major reflection correlates with the base of DS135-4, which was also observed on seismic reflection data from the 9th cruise of the R/V Jean Charcot (LePichon et al., 1970).

DSDP 120 (table 4.2) provides the oldest sediment samples for the region (Ryan
<table>
<thead>
<tr>
<th>UNIT/DEPTH</th>
<th>DESCRIPTION</th>
<th>AGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Light grey nannoplankton chalk ooze. Pelagic with traces of quartz silt.</td>
<td>Pleistocene to Late Oligocene</td>
</tr>
<tr>
<td>- 0-325 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Olive grey to brown mudstone, sand and brown clay. Terrigenous sediments with less than 25 % CaCO$_3$</td>
<td>Early Eocene</td>
</tr>
<tr>
<td>- 335-341 m</td>
<td>Silty mud</td>
<td></td>
</tr>
<tr>
<td>- 341-350 m</td>
<td>Quartz siltstone and mudstone</td>
<td>Cret. or Paleocene</td>
</tr>
<tr>
<td>- 431-435 m</td>
<td>Silty mudstone, sand and clay</td>
<td>Early Maestrichtian</td>
</tr>
<tr>
<td>3</td>
<td>Black and green shale, siliceous mudstone, limestone and chert. Traces of CaCO$_3$. Rich in zeolite, with pyrite and 5-10 % terriginous silt.</td>
<td>Cretaceous</td>
</tr>
<tr>
<td>- 563-570 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Olive grey and black nannofossil laminated marl ooze and limestone. Some silty quartz layers and pyrite.</td>
<td>Early Aptian</td>
</tr>
<tr>
<td>- 684-689 m</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 4.1: DSDP Site 135 - Summary of lithologies, age and recovery depth.

et al., 1973). Of the four units identified, units 1, 2 and 3 (DS120-1, DS120-2 and DS120-3) are probably similar to DS135-1, all consisting of almost entirely pelagic material which was deposited subsequent to uplift, with deceasing influence of abyssal plain associated terriginous deposition from the base of DS120-3. The oldest sediments cored (DS120-4) lay upon basement rocks and were of Barremian age. It is likely that similar sediments lie between the base of DSDP site 135 and basement observed on seismic reflection data.

The correlation, age and examples of seismic character of units from the Tagus, Western Horseshoe, Eastern Horseshoe and Seine Basins are shown in table 4.3. These have been established on the basis of identification of similar reflection character and correlation with units sampled at DSDP sites 120 and 135, including isotopic dates for the Gorringe Bank basement rocks (e.g., Feraud et al., 1986) and estimates of crustal age from magnetic spreading anomalies (e.g., Klitgord & Schouten, 1986)
Figure 4.9: Sediment accumulation at DSDP site 135 (Hayes et al., 1972). Red solid line shows the age depth relationship. Green dash line represents the Early Eocene to Late Oligocene unconformity. Blue lines show sediment accumulation in metres per million years.

The youngest units in the region WH-I, EH-IA/B and S-IA/B are likely to be of a similar age as they all lie above the distinctive Early Eocene to Late Oligocene unconformity, observed in DSDP 135 (Hayes et al., 1972). The base of unit T-I may be slightly younger, based on tentative correlations with sequences IA of Mauffret et al. (1989) and DSDP site 120 (Ryan et al., 1973). Therefore the age of WH-I, EH-IA/B and S-IA/B would be between Pleistocene and Late Oligocene and the age of T-I would be from Pleistocene to Middle Miocene. Compositionally they may be similar to DS135-1 and DS120-1,2 and 3, but with a higher terrigenous component, perhaps similar to terrigenous material observed in DS120-3, as DSDP sites 120 and 135 sampled bathymetric highs which experience a higher concentration of pelagic material than the abyssal plains.

Units T-II and EH-II represent debris flow and olistostrome deposits derived from Gorringe Bank and Straits of Gibraltar respectively. Although unsampled, EH-II was deposited at earliest between the Early Eocene and Late Oligocene or slightly later in the Miocene (Bonnin et al., 1975; Auzende et al., 1981), with the disrupted reflection character and opacity possibly concealing a thin veneer of post Late Oligocene sedimentation. T-II is assumed to be of Miocene age by Mauffret
Figure 4.10: Seismic reflection profile N-O from CD64 in the region of DSDP site 135.
Table 4.2: DSDP Site 120 - Summary of lithologies, age and recovery depth.

et al. (1989). Both T-II and EH-II were deposited following uplift of Gorringe Bank, providing an upper limit to the age of uplift from Late Oligocene to Miocene. The compositions of T-II and EH-II are unknown, but T-II is probably composed of a melange of material similar to that sampled at DSDP site 120 (table 4.2) on the crest of Gorringe Bank.

Units T-III, EH-III, S-II and possibly WH-II/III may have an age and composition which correlates with units DS135-2,3 and 4 from DSDP 135 (table 4.1) and unit DS120-4 from DSDP 120 (table 4.2). The top of unit DS135-2 and base of unit DS135-4 correlate with seismic reflections which mark the top and base of units EH-III and S-II. These can be traced on seismic data in the region of DSDP 135 (Profile N-O, figure 4.10) into the Eastern Horseshoe and Seine Basins. Units T-III, EH-III, S-II and WH-II (possibly WH-III) probably have an age between Early Eocene and Early Aptian as they lie beneath the Early Eocene to Late Oligocene unconformity.

The age, correlation and composition of units below those previously discussed is complicated by the lack of sampling, their isolation from other basins and the variable reflection character between basins. An upper age constraint is provided by the Barremian (Lower Cretaceous) sediments cored in DSDP site 120 and the Early Aptian sediments cored in DSDP site 135, which have been correlated with T-
<table>
<thead>
<tr>
<th>TAGUS BASIN</th>
<th>WESTERN HORSESHOE BASIN</th>
<th>EASTERN HORSESHOE BASIN</th>
<th>SEINE BASIN</th>
<th>SEISMIC</th>
<th>DESCRIPTION</th>
<th>AGE</th>
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<tr>
<td>T-VI</td>
<td>WH-VI</td>
<td>EH-V</td>
<td></td>
<td></td>
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</tbody>
</table>

Table 4.3: Possible unit correlation between basins of the Gorringe Bank region and probable age based on DSDP site 135 and plate reconstructions. Brackets represent debris flow and olistostrome units.
III, EH-III, S-II and WH-II (possibly WH-III). The oldest sediment age is limited by the crustal age of the Gorringe Bank region, which may be estimated from Atlantic Ocean plate reconstructions, magnetic spreading anomaly recognition and basement rock samples.

An assumption that oceanic crust underlies the basins of the Gorringe Bank region gives an oldest age similar to initiation of Atlantic rifting (175 Ma for NA-AF (Klitgord & Schouten, 1986) and 165 Ma for NA-EU (Sclater et al., 1977)) for the oldest crust in the eastern Gorringe region. Crustal age decreases to the west, with oceanic crust in the Tagus Basin at an age of approximately 118 Ma at 14°W based on the first distinguishable magnetic chron M-0 (Klitgord & Schouten, 1986). An abandoned spreading centre in the eastern Tagus Basin may have introduced crust of an anomalously young age from 130 Ma to 119 Ma (Mauffret et al., 1989; Malod & Mauffret, 1990). Crust in the Seine Basin has an age of about 156.5 Ma (M-25) at 12.5°W, 149.5 Ma (M-21) at 14°W and 141.5 Ma (M-16) at 16°W (Klitgord & Schouten, 1986).

Although magnetic chrons have not been identified in the Horseshoe abyssal plain, a NNE extrapolation of the M-25 spreading anomaly from the Seine Basin would place M-25 (156.5 Ma) at approximately 12.5°W in the Horseshoe abyssal plain, but this is tentative due to the undetermined slip between the Seine and Horseshoe abyssal plains associated with the Azores-Gibraltar plate boundary. Isotopic dates from samples of Gorringe Bank basement gabbros and dolerites gave formation ages of 141 ± 3 Ma and 138 ± 3 Ma (Prichard & Mitchell, 1979) in (Feraud et al., 1986) by standard K-Ar methods and ages of 143 Ma from Gettysburg dolerites and 140 Ma for Ormonde gabbros (Feraud et al., 1982, 1986) using a stepwise heating 40Ar/39Ar method.

Together the data suggest that the oldest crust in the region is likely to be upper Jurassic to Lower Cretaceous decreasing in age towards the west, with the oldest sediments deposited upon embryonic oceanic crust. However, the oldest deposits in the region may be the remnants of pre-rift sediments, in accordance with Mauffret et al. (1989) who said that the oldest sediments in the Tagus Basin may be as old as Tithonian to Oxfordian and may include a sequence of evaporites and pre-rift sediments of Triassic age, which would indicate continental crust. Pautot et al. (1970) presented seismic reflection evidence for salt diapirs in the Eastern Horseshoe abyssal plain and along the North African and Iberian margins, possibly related to initial rifting of the Atlantic Ocean giving an age of Jurassic to Triassic for salt in the Gorringe Bank region. It may be that areas of the Gorringe Bank region are underlain by continental crust with sediments of a greater age than the oldest
The location of the Ocean Continent Transition (OCT) has been interpreted (figure 2.17) to lie between 12°W and 11°55′W in the Tagus Basin (Whitmarsh et al., 1993). To the south of the Seine Basin the OCT may be located at 10.5°W (Roeser, 1982; Roest et al., 1992), based on the S1 anomaly which provides a tentative eastern limit as to the location of the OCT on the African margin. A westerly constraint is provided by the M-25 magnetic spreading anomaly at 12.5°W (Klitgord & Schouten, 1986).

The ages of T-VI, WH-VI, EH-V and S-III, the oldest sediments in the Gorringe region, are poorly constrained. Unit T-VI is likely to be older than 118 Ma based on the M-0 magnetic anomaly at 14°W (Klitgord & Schouten, 1986) and the interpretation of seismic reflection data by Mauffret et al. (1989). Unit WH-VI probably overlies Middle Jurassic crust with an age of 156.5 - 149.5 Ma (Bathonian to Callovian) based on the identification of M-25 to M-21. Extrapolation of the M-25 spreading anomaly would give a crustal age of 156.5 Ma at 12.5°/dgr W at the western edge of the Eastern Horseshoe Basin, implying older crust to the east. However, ages from samples of Gorringe Bank basement rocks have given ages in the region of 140 Ma. Therefore the basal age of EH-V is under question, but probably from Lower Cretaceous to Upper Jurassic. The age of S-III is constrained by the magnetic anomalies identified in the Seine Basin with an age of 156.5 Ma (M-25) at 12.5°W (Klitgord & Schouten, 1986) giving a maximum age of Late Jurassic for the base of this unit.

### 4.3 Sediment Thickness and Basement Depth

#### 4.3.1 Calculation

Sediment thickness variation in the Gorringe Bank region has been constrained by basement picks in TWTT from interpreted CD64 seismic reflection profiles and other available seismic reflection data (LePichon et al., 1970; Mauffret et al., 1989; Sartori et al., 1994; Banda & Torne, 1995). Acoustic basement was clearly imaged on all seismic profiles except for the deepest regions of the Eastern Horseshoe Basin and a small region to the west of Hirondelle seamount, where the great sediment thickness prevented seismic penetration. Additional constraint was provided by the basement intercept depth of 251.7 m in DSDP site 120, a minimum basement depth estimated from DSDP site 135 and regions of zero sediment thickness (basement outcrop) from the interpreted CD64 GLORIA data (figure 5.2).

Accurate basement depth and sediment thickness constraint, in TWTT, is pro-
vided for the Eastern and Western Horseshoe Basins, with less constraint in the Tagus and Seine Basins. Although units have been defined (figure 4.3), poor constraint of their regional extent and continuity preclude construction of three-dimensional isopach maps for the Gorringe Bank region and assignment of step increases in velocity associated with lithological boundaries. Therefore the total sediment package was converted from TWTT to depth using the quadratic velocity-depth function (equation (4.1), where \( V_b \) = bulk velocity in \( \text{ms}^{-1} \) and \( X \) = depth in metres), derived by the following method.

\[
V_b = 1518.5 + 1.0702X - 0.00010003X^2
\]  

(4.1)

Initially formulae linking velocity and porosity (Wyllie et al., 1956; Hamilton & Bachman, 1982) were used to convert the empirical porosity versus depth curves (figure 4.11) (Bond & Kominz, 1984) to velocity versus depth (figure 4.12). These curves (figure 4.11) were constructed using general concepts of sediment diagenesis during burial and porosity values as a function of depth from drilled modern sedimentary deposits (e.g., Maxwell, 1964; Chilingarian & Wolfe, 1975, 1976; Magara, 1980).

Figure 4.11: Empirical porosity - depth curves from Bond & Kominz (1984).
The first was the empirical formula of Wyllie et al. (1956) (equation (4.2), where $V_b =$ bulk velocity, $V_f =$ pore fluid velocity, $V_m =$ matrix velocity all in ms$^{-1}$ and $\phi =$ porosity from 0-1 where 1 is 100% porous).

\[
\frac{1}{V_b} = \frac{\phi}{V_f} + \frac{1-\phi}{V_m}
\] (4.2)

A knowledge of the pore fluid and grain velocity of the constituents of the sediment is required. The pore fluid velocity is assumed to be 1500 ms$^{-1}$ based on standard seawater measurements. Brine type fluids within the sediments may have slightly higher velocity, but such variation does not significantly affect results. Grain velocity is composition dependent and experiments (Asquith & Gibson, 1982) show high grain velocity variation with sandstone 5400-6000 ms$^{-1}$, limestone 6400-7000 ms$^{-1}$ and dolomite 7000-7900 ms$^{-1}$. Examination of velocity results from equation (4.2) within the porosity range relevant to that expected for sediments in the Gorringe region showed that the results are fairly insensitive to changes in the grain velocity between 5000 and 6000 ms$^{-1}$.

The second was the formula of Hamilton & Bachman (1982) (equation (4.3), where $V_p =$ velocity in ms$^{-1}$ and $\phi =$ porosity in %) which is a direct relationship between porosity and velocity, but only applies to a porosity range of 35 to 85 %.

\[
V_p = 2502.0 - 23.45\phi + 0.14\phi^2
\] (4.3)

The accuracy of equations 4.2 and 4.3 was investigated for shallow depths by comparison of velocities calculated from porosity data from DSDP sites 135, 136, 370 and 400a (Hayes et al., 1972; Lancelot & Seibold, 1977; Montadert & Roberts, 1979) from the Iberian margin with the measured measured sonic log velocities (figure 4.13).

Equation (4.2) over predicts the sonic velocity by approximately 500 ms$^{-1}$ based on a grain velocity of 5500 ms$^{-1}$. To achieve a closer fit requires an unrealistically low velocity for the sediments of the Gorringe region. However, equation (4.3) quite accurately predicts the sonic log velocity for the upper approximately 800 metres sampled by the DSDP sites, but only applies to a 35 to 85 % porosity range, which is rapidly lost and therefore provide no information for deep sediments.

The velocity versus depth curves converted from the porosity versus depth curves (Bond & Kominz, 1984) ($V_m = 5500$ ms$^{-1}$ and $V_f = 1500$ ms$^{-1}$ for equation (4.2) and 35-85 % for equation (4.3)) were plotted (figure 4.12) with sonic log velocities from DSDP site 135, which provide the only direct measurements of sediments in the Gorringe region and refraction model velocities (Purdy, 1975;
Figure 4.12: Velocity-depth function derived from seismic refraction models (Purdy, 1975; Whitmarsh et al., 1990; Pinheiro et al., 1992), sonic logs from DSDP site 135 (Hayes et al., 1972) and the velocity calculated from the porosity-depth curves of Bond & Kominz (1984) by equation (4.3) of Hamilton & Bachman (1982). The velocity calculated from the empirical porosity-depth curves of Bond & Kominz (1984) by equation (4.2) of Wyllie et al. (1956) with a grain velocity of 5500 ms\(^{-1}\) are shown for comparison.

Whitmarsh et al., 1990; Pinheiro et al., 1992).

The curves of Bond & Kominz (1984) converted by equation (4.3) give good constraint on the upper sediment velocities, with near surface velocities close to the expected value of approximately 1500 ms\(^{-1}\) corresponding to 100 % porosity. Most of the refraction and sonic velocity data lie just below the curves of Bond & Kominz (1984) converted by equation (4.2), but follow a roughly similar trend with an exponential increase of velocity with depth.

A quadratic polynomial (equation 4.1) was fitted to the seismic refraction velocities, sonic log velocities and velocities derived from the curves of Bond & Kominz (1984) converted by equation (4.3) for a porosity range 35-85 % . The curves of Bond & Kominz (1984) converted by equation (4.2) were not used in the calculation of the velocity-depth function due to the discrepancies with sonic log velocities.
Figure 4.13: Comparison of sonic velocity from DSDP 400a (Montadert & Roberts, 1979) with velocity derived from the porosity-depth curves of Bond & Kominz (1984) by equation (4.2) (Wyllie et al., 1956) and equation (4.3) (Hamilton & Bachman, 1982).

(figure 4.13) and refraction data (figure 4.12), but were used as a check to the modelled exponential trend of increasing velocity with depth. The trend of the final velocity-depth function (figure 4.12) shows greatest similarity with the lower curves of Bond & Kominz (1984). These correspond to sediment which experience initial rapid decrease in porosity due to the loss of pore fluid under compaction close to the seafloor and do not experience early cementation.

The velocity-depth function (equation 4.1) was integrated to give depth as a function of TWTT (equation (4.4), where $X = \text{depth in metres}$, $T = \text{one way travel time in seconds}$, $a = 0.00010003$, $b = 1.0702$, $c = 1518.5$ and $P = \sqrt{b^2 - 4ac}$).

$$X = \frac{b^2 - P^2 - b^2 e^{PT} + P^2 e^{PT}}{2abe^{PT} - 2aPe^{PT} - 2ab - 2aP} \quad (4.4)$$
4.3.2 Description and Interpretation

Great sediment thickness variation is observed in the Gorringe Bank region (figures 4.14 and 4.15), with thinner drapes on bathymetric highs and substantial thicknesses in the basins, which correspond to shallow and deep basement depths (figure 4.16).

The Eastern Horseshoe Basin has the greatest sediment thickness of up to 5 km, with a corresponding basement depth of 10 km. The basin's shape roughly follows the surface expression, but with the main depocentre to the east and a narrowing and gradual sediment thinning to a minimum of just greater than 1 km (basement depth 6.3 km) towards the western edge of the basin. The decrease in sediment thickness and basement depth is quite rapid towards the southern flank of
the basin, which follows a near linear ENE-WSW trend associated with the thrust front identified in this region. The nature of the northern flank of the basin is more complex with sediment thinning onto the southern flank of Gorringe Bank and fairly steep basement shallowing.

Western Horseshoe Basin sedimentation is divided into two main depocentres forming a "dumbbell" shaped basin with an ENE-WSW trend which differs from the surface expression. The largest sub-basin to the ESE has up to 4 km of sediment (basement depth 8.9 km). The sub-basin is near circular, yet stretched slightly along a NNE-SSW trend related to the equally trending strike slip fault identified between Gorringe Bank and Hirondelle seamount on seismic reflection data (Purdy, 1974), with associated seismicity, including a strike slip focal mechanism (Buform et al., 1988). The second near circular sub-basin to the WSW has a much lower total sediment thickness of up to 2.8 km (basement depth 7.5 km). In general the sediments in the Western Horseshoe Basin thin more rapidly towards the N to NE due to the basin bounding faults, with lower rates towards Ampere seamount and
Figure 4.16: Basement depth in the Gorringe Bank region derived from the bathymetry and total sediment thickness. Contour interval 0.5 km.

the Eastern Horseshoe Basin.

Sediment thickness variation in the Tagus Basin is poorly constrained, yet the thickest sediments of up to 2.9 km (basement depth 8.6 km) appear to form a linear band across the base of the northern flank of Gorringe Bank, with rapid thinning towards the Bank and a gradual thinning towards the north.

Gorringe Bank has outcrop of basement rocks and sparse sedimentation on the northern flank, as sampled (e.g., Auzende et al., 1982, 1984), with only approximately 69 m of post uplift pelagic sediments out of a total of 251.7 m sampled at DSDP site 120 (Ryan et al., 1973) in the saddle between Ormonde and Gettysburg seamounts. Sediment thickness is more substantial on the southern flank and increases gradually to the south atop a fairly steeply deepening basement.

West of Gorringe Bank and over Hirondelle seamount sediment thickness is approximately 1 km with a complex variation in contrast with the basement which shallows with a fairly regular near circular form associated with Hirondelle seamount. Further west on Josephine seamount sedimentation is very low with large basement
outcrops (basement depth approximately 300 m), due to its Late Miocene volcanic history (Wendt et al., 1976), with only a short period for pelagic sediment accumulation. Low sediment thicknesses are also observed on other seamounts associated with the Tore-Madeira Rise.

Ampere seamount also has a sparse sediment cover, with many basement outcrops with a minimum depth of approximately 800 m at the crest. However, Coral Patch seamount has a sediment cover of approximately 1 km and a basement depth of about 3 km. To the east, along Coral Patch Ridge the sediment thickness increases to about 1.5 km, with an increase of the basement depth to about 6 km. The sedimentary cover on Coral Patch Ridge is greater than the other recent bathymetric highs suggesting that it was uplifted at a later time. The variation of sediment thickness and basement depth are poorly constrained in the northern Seine Basin, but it seems likely that sediments quickly reach a thickness of at least 2 km to the south of Ampere seamount and Coral Patch Ridge.
Chapter 5

Deformation and Structural Analysis

5.1 Introduction

GLORIA data is used to interpret the tectonics of the Gorringe Bank region with the aid of other marine geophysical data, including 3.5 kHz and seismic reflection data. The distribution of deformation in the Gorringe region is described and related to the orientation of features with respect to plate motion vectors.

5.2 Description of Deformation in the Gorringe Bank Region

GLORIA data collected along eight ENE-WSW trending profiles with a swath width of about 20 kilometres provides almost complete coverage of the Gorringe Bank region (figure 5.1), with two additional NNW-SSE trending lines giving overlapping information. The GLORIA image is a black and white representation of the reflectivity of the seafloor and is affected by the incidence angle of the beams. Textural and tonal changes are associated with seafloor characteristics which may be related to geological features. The geological significance of an observed feature on the image is not always clear, however combination of the images with bathymetry, 3.5 kHz and seismic reflection data aids interpretation.

Gorringe Bank is the most distinctive feature on the GLORIA images of the Gorringe region (figure 5.1). The brightest zone of reflectivity, associated with the outcrop of crystalline basement rocks and the high incidence angle of the GLORIA beams, is located on the northern flank. Patches of brighter and duller echoes within this zone are related to the thin discontinuous sediment cover. The northern flank’s texture has the appearance of Bank parallel ridges and blocks with an
Figure 5.1: Mosaic of GLORIA images from the Gorringe Bank region from CD64. See figure 5.2 for interpretation.
Figure 5.2: Interpretation of GLORIA images. Red solid and red/black dash lines show folding and faulting constrained by seismic profiles. Red dotted lines show unconstrained folds or faults. Orange dash lines show major fault trends, with motion sense shown by black vectors. Blue, green and yellow lines show channels, abyssal plain edges and the continental slope base respectively. "v" shows basement outcrop and speckle shows debris flows. Seismic profiles located by thick red lines. Bathymetry contour interval 0.5 km.
Figure 5.3: Location of seismic reflection profiles from CD64 and IAM (Banda & Torne, 1995) presented in chapters 4, 5 and 6.

ENE-WSW to bathymetry parallel trend, probably representing small terraced cliffs of fractured basement rocks with scattered fallen blocks, scree and pelagic sedimentation, as observed during submersible investigations (Auzende et al., 1982; LaGabrielle & Auzende, 1982; Auzende et al., 1984). These features represent the source of debris flows which have been observed on seismic reflection data in the southern Tagus Basin (e.g., Mauffret et al., 1989) (figure 2.7). The crest of the Bank shows a distinct textureless shadow zone, which mimics the bathymetry at a depth of approximately 400 metres, due to the shallow depth and low incidence angle of the GLORIA beams. The southern flank of the Bank is less reflective due to the thicker more continuous sediment cover. A narrow bright reflection around the southern edge of Gettysburg and Ormonde seamounts, at a depth of approximately 1000 metres, is related to an increase in gradient and on the SSE flank of Ormonde, a change from mud deposits to the outcrop of alkaline breccia, also observed by submersible. Below this edge a series of lower intensity broken echoes parallel to the upper edge probably represent small vertical cliffs, which are
of gabbroic basement rocks with a sedimentary drape on the SSE flank of Ormonde seamount (Auzende et al., 1984).

The SSE flank of Gorringe Bank is cut by a system of up to twenty narrow channels (figures 5.1 and 5.2), which all appear to originate from a depth of approximately 1300 metres where they show high reflectivity. Downslope they become wider and less reflective, mostly coalescing at a depth of greater than 3000 metres. Examination of EM12 Simrad swath bathymetry data (Beuzart et al., 1979) from Gorringe Bank (figure 2.3) reveals the channels to show active erosion, as interpreted by LaGabrielle & Auzende (1982) and probably formed subsequent to Gorringe Bank uplift.

The northern flank of Gorringe Bank is cut by a major faulting related feature which transects the base with a trend of N48°E, in some areas marking the edge of the Tagus abyssal plain. The feature is observed as a small slightly uplifted faulted and deformed region to the north of Ormonde seamount on seismic reflection profiles from CD64 and by a trough to the northwest of Gettysburg seamount.
although the characteristics of the feature are masked by a high density of deformation related diffractions and a change from sedimentary to basement rocks.

Figure 5.5: GLORIA image of Hirondelle seamount showing the linear N48°E trending deformation features and the N26°E trending outcrop of basement rocks associated with the scarp of a major strike-slip fault.

To the west of Gorringe Bank a feature comprised of a group of narrow linear echoes with a trend of approximately N26°E and a total length of approximately 100 km cuts across the northern seamount arm between Gettysburg and Hirondelle seamounts, continuing into the Tagus Basin and probably some distance into the Western Horseshoe Basin (figures 5.5, 5.1 and 5.2). Together these features represent outcrops of basement rock associated with the scarp of a seismically active major strike-slip fault, with a single focal mechanism from the magnitude 4.8 (Ms) earthquake of the 29th June 1965 (Buforn et al., 1988) indicating left lateral strike-slip motion. The fault was initially observed on the seismic reflection data of Purdy (1974) and was imaged by three reflection profiles from CD64, most clearly represented by profile P-Q (figures 5.3 and 5.4) which shows a thin layer
of sediments on the crest of a fault scarp which has basement outcrop on a 750 m high surface. The deep extent of the fault plane and a large fault block to the ENE, with a height of up to 1.5 seconds TWTT, bounds a great thickness of up to 2 seconds TWTT which appear to have been uniformly folded to produce a single large scale anticlinal fold with a surface relief of approximately 350 m.

Further to the west, Hirondelle seamount is recognised on GLORIA images by a radial zone of fairly bright echoes whose brightness decreases away from the crest of the seamount, related to the shallow depth and continuous thick sedimentary cover (figure 5.5). The whole of this region is cut by many linear features with a variable yet close spacing of about 2.5 km and almost identical trends of about N48°E, which is 22° greater than the trend of the strike-slip fault to the east. The features have a surface expression which is somewhat similar to this fault with many linear basement outcrops represented by bright echoes, sometimes linked by the narrower slightly darker echoes which represent the surface expression of compressional faulting and folding of the sediments, as observed on seismic reflection data from CD64, with no basement exposure.

West of Hirondelle seamount, Josephine seamount exhibits very bright echoes with a "V" shaped shadow zone relating to the shallow elongated crest of the seamount, masking the structure of the upper flanks. The lower slopes show a few bright echoes related to the basement outcrop with the appearance of many scarps and scree related blocks. A few bright narrow features on the southeastern flank are related to channels similar to those observed on Ormonde seamount and the small unnamed seamount just to the south which shows very bright echoes in a radial pattern related to basement outcrop channels around the flanks.

Located to the south of Gorringe Bank, Hirondelle and Josephine seamounts and the Horseshoe abyssal plain are Ampere seamount, Coral Patch seamount and Coral Patch Ridge (figure 5.2). The perimeter of Ampere seamount has medium to dark echoes related to the sedimentary cover (figure 5.6). To the southwest, west and northwest many linear features mostly unassociated with basement outcrops, but otherwise similar to features observed in the region of Hirondelle seamount have a lower trend of between about N20-30°E. They are related to thrust faulting and folding of the basement and sedimentary cover, with surface uplifts of up to 600 m observed on seismic reflection data from line 2.

Central Ampere seamount exhibits bright echoes of a near circular form mimicking the bathymetry, with very bright near vehicle echoes corresponding to the high incidence angle of the GLORIA beams. The seamount has the appearance of a volcano with a near circular plan and radiating ridges and channels similar to
other volcanic seamounts of the region, including Josephine seamount and others associated with the Madeira-Tore Rise. Matveyenkov et al. (1994) concluded that Ampere seamount is of volcanic origin on the basis of recovered samples (Litvin et al., 1982) of highly alkaline nepheline basaltoids. A radial pattern of bright narrow features around the base from a depth of between 1000 and 2000 metres to a depth of about 3000 metres probably correspond to outcrop ridges related to volcanic material such as pillow basalts (Marova & Yevsyukov, 1988) which have been observed on the lower slopes and to channels utilising the volcanic structure. Some linear bathymetry parallel outcropping ridges and a blocky texture similar to that observed on the northern flank of Gorringe Bank probably represent the
observed and sampled layers of rectangularly jointed dense reddish-ochre trachyte rocks (Litvin et al., 1982) and conglomerates.

Figure 5.7: GLORIA image of Coral Patch seamount, including radial outcropping ridges and channels.

Coral Patch seamount has a character on GLORIA images different to other seamounts of the Gorringe Bank region. The regional echoes are of light to dark grey tones indicative of a more continuous thicker sedimentary cover, with bright echoes on the crest of the seamount unrelated to basement outcrop, but to the shallow depth and high incidence angle of the GLORIA beams (figure 5.7). The
Figure 5.8: Seismic reflection profile R-S from CD64 of deformation ridges to the north of Coral Patch seamount.
entire northern flank is covered by in excess of thirty near linear N45°E trending echoes with wide shadow zones and a few bright narrow echoes associated with channels which have exploited the linear features. Many of these features continue obliquely over the crest and onto the southwestern flank of the seamount, but their concentration is higher to the north where they continue to a depth of 4800 metres corresponding to the edge of the Horseshoe abyssal plain. These "ridges and valleys" represent short wavelength folds and associated faulting as observed on seismic reflection profile R-S (figure 5.8).

These "ridges and valleys" are almost certainly tectonically related to the deformation, poorly imaged by GLORIA, which continues to the northwest of Coral Patch seamount. Profile T-U (figure 5.9) shows the two major steep basin bounding normal faults with a seafloor relief of about 1 km and a corresponding large scale basement relief of up to 1.5 seconds TWTT. The units of the Western Horseshoe Basin continue towards the southeast beyond these boundary faults, with the stripy reflections on the crest of the faults probably similar to unit WH-II. It appears that the majority of the deposition of unit WH-III and older occurred under the control of boundary faulting with thick deposits in the Western Horseshoe Basin, slightly thinner deposits around the basin margin and substantial thinning towards the southeast associated with a bathymetric high. Deposition of units WH-II and WH-I was more even with significant deposition to the southeast, yet still slightly less than in the Western Horseshoe Basin. This thinning may be related to the bathymetric position of the region to the southeast of the boundary faults being above the major deposition level of the Western Horseshoe abyssal plain, with uplift and deformation occurring during WH-I time associated with recent compression, perhaps including inversion along the boundary normal faults. Bottom current erosion may have also played a part in thinning of WH-I on the bathymetric highs.

Moving towards the east, from about 11.5°W, the "ridge and valley" lineations rotate near parallel to the bathymetric trend of approximately N70°E, continuing to Coral Patch Ridge. Here the echoes are weaker, with a few large slightly curved features associated with folding and thrust faulting observed on seismic reflection data in the region of DSDP site 135. Seismic reflection profile V-W-X (figure 5.10) shows the thrust faulting and folding of sediments in this region, associated with a regionally high sediment corrected Bouguer anomaly (Chapter 6, section 6.6.3). Seismic reflection profile W-X shows a pair of steep thrust faults with a throw of up to about 500 metres of the sediments and basement. Units on the overthrust block thicken slightly towards the fault suggesting that they represent an inverted half
graben structure. Sediments in the hanging wall show large hanging wall anticlines into the fault and those on the crests of thrust blocks exhibit long wavelength folding and faulting which often breaks the seafloor, indicating recent tectonic motion. Profile V-W shows similar widespread deformation, yet with several thrust faults confined to a narrower zone at the edge of the Eastern Horseshoe abyssal plain. This change may be represented on the GLORIA image by a concentration of deformation to the north of a small bathymetric high at 35.4°N, 10°W, compared to the fairly wide featureless region to the west across Coral Patch Ridge.

The abyssal plains show a regionally similar texture and tonal echo character probably due to their minimal surface deformation, relief and minimal material composition variation, with no basement outcrop within their bounds. The track parallel ripple texture observed on many swaths is probably an artifact produced by the interference of reflections from within the upper unconsolidated sediments, into which the beams penetrate, with backscattered energy from the seafloor (Belderson et al., 1972; Huggett et al., 1992).

The cloudy textured features (figures 5.1 and 5.2) in the southern Tagus abyssal plain (37°N, 10.8°W to 12°W.) probably represent geologically recent debris flows or turbidites derived from the Iberian margin, with no features linking them to Goringe Bank. Their appearance is of what might be expected from fine silt and mud at the toe of a turbidite deposit. Channels which supply turbidite sedimentation to the Eastern Horseshoe abyssal plain, probably similar to those associated with the Tagus abyssal plain, are observed to the eastern end of the Eastern Horseshoe abyssal plain at approximately 36°N, 9°W. Fairly wide solid bright echoes related to tributaries on the middle to upper continental slope of Iberia are confined within deep canyons, including the Sao Vincente Canyon, observed on bathymetry data. On the middle to lower continental slope, where there is a fairly gradual break in the slope gradient, the tributaries coalesce forming two major channels with a width of up to about 25 km represented by a bright echoes with a whispy texture. These continue into the Eastern Horseshoe abyssal plain and onto Coral Patch Ridge where they become indistinct. Small channels are imaged at the base of the continental slope at 37°N, 9.8°W, where the slope break is represented by a bright narrow linear feature, with a slight shadow zone to its east. Other channels and canyons observed on bathymetric maps of the Iberian margin to the east of the Tagus abyssal plain associated with many of the major Iberian mainland faults such as the Alandroal and Lower Valley of the Tagus faults are responsible for turbidite deposition in the Tagus Basin.

The abyssal plains generally exhibit little surface deformation. No surface fold-
Figure 5.11: Seismic reflection profile Y-Z and 3.5 kHz data of a recent fault in the Eastern Horseshoe abyssal plain from CD64. Red line marks the top of the olistostrome unit EH-II (See chapter 4). Arrows mark the location of the fault scarp on CD64 line 8. Green dash lines mark the location of the 3.5 kHz data on the seismic reflection profile.
Figure 5.12: EM-12 Simrad swath bathymetry of a recent fault in the Eastern Horseshoe abyssal plain from CD82. Contour interval 5 metres. Arrow marks the location of the fault scarp with a trend shown by a white dash line. Black dash line marks the location of seismic reflection and 3.5 kHz data from CD64.
Figure 5.13: Seismicity greater than Mb 4 in the Eastern Horseshoe abyssal plain from the ISC. Focal mechanisms from Grimison & Chen (1986, 1988) and Bu forn (1988), with depths in kilometres. Solid and dash white lines show the tracks of CD64 and CD82 respectively. The yellow line shows the location of EM12 simrad swath bathymetry shown in figure (5.12), the thin black line shows the location of seismic reflection profile Y-Z (figure 5.11) and the red line shows the location of 3.5 kHz data shown in figure (5.11). Contour interval 500 m.
ing or faulting has been observed on seismic reflection data from the Tagus abyssal plain. The Seine abyssal plain shows some surface deformation, but it is mostly confined to the southern flank of Coral Patch seamount. The only surface breaking feature imaged in an otherwise featureless Horseshoe abyssal plain is located in the southern central Eastern Horseshoe abyssal plain (35°48’N, 10°35’W). The feature was initially imaged by seismic reflection data and 3.5 kHz data from CD64 (figure 5.11). Subsequently it was imaged by EM12 Simrad swath bathymetry data from CD82 (figure 5.12). The sonar and seismic data suggest that the feature represents a fault with a trend of approximately N40°E and a surface throw of approximately 30 metres with downthrow to the NNW. Unit EH-I (figure 5.11, above the red line) is quite thin in this region of the Eastern Horseshoe abyssal plain and the olistostrome unit EH-II prevents the accessment of the deeper fault structure including estimates of total throw.

The nature of the fault is unclear on seismic reflection and 3.5 kHz data. It is located to the south of a cluster of shallow seismicity in the Eastern Horseshoe abyssal plain (figure 5.13), associated with mainly thrust and strike-slip focal mechanisms (Grimison & Chen, 1986, 1988; Buforn et al., 1988). Two focal mechanisms show motion with an anomalous orientation to the regional right lateral motion if either nodal plain is taken to represent the fault plain. Although the style of faulting in the Eastern Horseshoe abyssal plain is complex, the feature probably represents a thrust fault, based on its orientation and seismic character.
5.3 Interpretation of the 3.5 kHz data

The 3.5 kHz data image the surface characteristics of the seafloor with some sub-bottom penetration and the character of the reflections and diffractions provides information as to the nature of the seafloor material. A map (figure 5.14) of regional reflection character variation based on the classification of Damuth & Hayes (1977) (table 5.1 and figure 5.15) leads to the identification of zoned characterisation.

<table>
<thead>
<tr>
<th>ECHO TYPE</th>
<th>FORM DESCRIPTION</th>
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</thead>
<tbody>
<tr>
<td>DISTINCT ECHOES</td>
<td>PROLONGED</td>
</tr>
<tr>
<td>IA</td>
<td>Sharp continuous-No sub bottom reflections</td>
</tr>
<tr>
<td>IB</td>
<td>Sharp continuous-Numerous sub bottom reflections</td>
</tr>
<tr>
<td>INDISTINCT ECHOES</td>
<td>PROLONGED</td>
</tr>
<tr>
<td>IIA</td>
<td>Semi prolonged-Intermittent sub bottom reflections</td>
</tr>
<tr>
<td>IIB</td>
<td>Very prolonged-No sub bottom reflections</td>
</tr>
<tr>
<td>INDISTINCT ECHOES</td>
<td>HYPERBOLIC</td>
</tr>
<tr>
<td>IIIA</td>
<td>Large irregular hyperbolae with varying vertex elevations</td>
</tr>
<tr>
<td>IIIIB</td>
<td>Regular single hyperbolae with varying vertices and conformable sub bottom reflections</td>
</tr>
<tr>
<td>IIIC</td>
<td>Regular overlapping hyperbolae with varying vertex elevations</td>
</tr>
<tr>
<td>IIID</td>
<td>Regular overlapping hyperbolae with vertices tangential to the seafloor</td>
</tr>
<tr>
<td>IIIE</td>
<td>IIID with intermittent zones of IB</td>
</tr>
<tr>
<td>IIIF</td>
<td>Irregular single hyperbolae with non conformable sub bottom reflections</td>
</tr>
</tbody>
</table>

Table 5.1: 3.5 kHz echo character after Damuth et al. (1977).

The Tagus, Horseshoe and Seine abyssal plains all exhibit mostly IB character (Sharp continuous and numerous sub bottom reflections), associated with the very flat undeformed seafloor and highly stratified turbidites, which have been sampled in these regions (Hoyt & Fox, 1977). Margins of the abyssal plains exhibit greater
Figure 5.14: Simplified regional echo character classification (Damuth et al., 1977) of 3.5 kHz data from CD64 in the Gorringe region. Mixed zone includes IB, IIA, IIIA, IIIB, IIIC, IIID and IIIF. See figure 5.15 for character examples.

character diversity. Ampere, Gettysburg, Ormonde, Josephine and to the north of Coral Patch seamount exhibit IIIA character (Large irregular hyperbola with varying vertex elevations) or IIIC character (Regular overlapping hyperbola with varying vertex elevations), with the many hyperbolae corresponding to basement outcrop and deformed seafloor sediment. Coral Patch seamount and ridge differed in that they were the only regions to show IIIB character (Regular single hyperbolae with varying vertices and conformable sub bottom reflections). Diffractions related to the recent deformation of the stratified sediments which give the sub-bottom reflections confirm that Coral Patch Ridge and seamount are the focus of recent deformation, whereas previous studies (e.g., Purdy, 1974) concluded that major recent deformation is located at Gorringe Bank. The Iberian margin shows mainly IIA character (Semi prolonged with intermittent sub bottom reflections) and localised IIB character (Very prolonged with no sub bottom reflections) related to the complex sedimentary deposits in this region. The western edge of
Figure 5.15: Examples of 3.5 kHz echo character from CD64 classified according to (Damuth et al., 1977). IB 15:00-15:30/18/92. IIA 9:42-10:12/13/92. IIB 2:48-3:18/19/92.
Figure 5.15: Continued examples of 3.5 kHz echo character from CD64 classified according to (Damuth et al., 1977). IIIA 23:48/17/92-0:18/18/92. IIIB 8:09-8:39/19/92.
Figure 5.15: Continued examples of 3.5 kHz echo character from CD64 classified according to (Damuth et al., 1977). IIIC 2:42-3:12/360/91. IIID 22:36-23:06/14/92.
the Western Horseshoe abyssal plain shows mixed character which includes IB, IIA, IIIA, IIIB, IIIC and IID (table 5.1), due to great variation in seafloor character with sediment, isolated basement outcrop and deformation. Small isolated regions of IIIF character (Irregular single hyperbolae with non conformable sub bottom reflections) corresponding to the deformation and faulting are located between Gettysburg and Hirondelle seamounts, but are not shown on figure 5.14 due to their minor extent.
5.4 Structural Analysis of Deformation

Structures identified on GLORIA images of the Gorringe Bank region (figure 5.1) have been interpreted (figure 5.2) with the aid of seismic reflection data and shown to represent recently active folding and faulting, which may yield important information as to the tectonic structure of the region.

The trend of deformation features interpreted (figure 5.2) from the GLORIA image (figure 5.1) were calculated by regression fits to digitised locations. Analysis of these orientations gave a mean strike of N56.63°E and a standard deviation of 24.40° (figure 5.16). The majority of the features have a trend close to the mean with most lying between N35-60°E. Only a few deviate significantly from the mean and none reach an orientation perpendicular to the mean trend. An assessment of the trends of features weighted according to their length, so that longer features carry a higher weight, shows a mean trend of N55.60°E, almost identical to that obtained from the unweighted analysis, as would be expected due to the majority of features being of similar orientation. Weighting is complicated by surface expressions not representing total feature length, with many being components of larger features.

The mean strike of the features is near perpendicular to motion vectors of the African plate with respect to the Eurasian plate (figure 5.17). These have been derived from earthquake slip vectors (Fukao, 1973; Grimson & Chen, 1986, 1988; Buform et al., 1988; Udias & Buform, 1991) and plate models such as NUVEL-1 (DeMets et al., 1990) which predicts N45°W convergence at Gibraltar and the three plate model of Argus et al. (1989), which utilised earthquake slip vectors, magnetic spreading anomalies and fracture zone azimuths from the Eurasian, African and North American plates. The features mostly represent compression (observed on GLORIA and seismic reflection data) with an expected strike perpendicular to the main compression direction.

Examination of N-S to NW-SE trending seismic reflection profiles 9 and 10 from CD64 in the Horseshoe Basin (figure 5.18), near parallel to the main compression direction, gives a minimum estimate of shortening (horizontal versus the along reflection distance) which has been accommodated by the sediments. The shortening was calculated along continuous segments of interpreted reflections between faulting associated breaks. These reflections were the base of EH-I and EH-III from the Eastern Horseshoe Basin and the base of WH-I, WH-III, WH-IV and WH-V from the Western Horseshoe Basin. This only provides an estimate due to shortening of the sediments by folding and takes no account of layer thickening and faulting, as
CHAPTER 5. DEFORMATION AND STRUCTURAL ANALYSIS

Figure 5.16: Rose diagram of the strike of deformation features in the Gorringe region from the interpretation of GLORIA images (Red lines on figure 5.2).

Figure 5.17: Plate motion vectors of Africa with respect to Eurasia, derived from earthquake slip vectors (Fukao, 1973; Grimison & Chen, 1986; Grimison & Chen, 1988; Buforn, et al., 1988; Udias & Buforn, 1991), the three plate model of Argus et al. (1989) and the NUVEL-1 plate model of DeMets et al. (1990).
the true fault dips are unknown.

Shortening in the Horseshoe Basin is small with most estimates lower than 1%. The highest estimates of shortening are located on the margins of the abyssal plains and over abyssal highs, with the base of EH-I and WH-I giving 0.11-0.17% for Coral Patch Ridge, 0.72% to the NE of Ampere seamount and 0.44% for the southern Hirondelle seamount, related to the concentration of deformation in these regions. The higher values of shortening estimated for these regions, especially the 2.29% from NE of Ampere seamount, are probably due in part to the amplification of inaccuracies in the calculation due to the short profile lengths and to the intense deformation associated with faulting.

Estimates for the Eastern Horseshoe Basin from the base of EH-I (figure 5.18a and table 5.2) are from about 0.09-0.20% with slightly greater shortening at the base of EH-III (figure 5.18b and table 5.3) of about 0.17-0.27 %. This suggests that unit EH-III was partially deformed prior to deposition of unit EH-I, as observed on seismic reflection profile J-K (figure 4.7).

An increased shortening with depth is also shown in the Western Horseshoe Basin, where better imaging enabled the calculation of shortening for deeper units than was possible for the Eastern Horseshoe Basin. The base of units WH-I and WH-III (figure 5.18 and tables 5.2 and 5.3) have a shortening of about 0.07% and 0.26% respectively, similar to the shallow units of the Eastern Horseshoe Basin, perhaps indicating a similar deformation history. The base of the deeper units of the Western Horseshoe Basin give 0.38% and 0.35% for the base of WH-IV and WH-V respectively (figure 5.18b), indicating greater shortening than and deformation before the deposition of the younger units WH-I and WH-III. The unexpected lower shortening estimated for the base of unit WH-V in comparison to the base of unit WH-IV is probably due calculation inaccuracies, which may include incorrect depth conversion of the deep units and the short profile lengths.

The NUVEL-1 plate model (DeMets et al., 1990) for the Eurasian-African plate boundary predicts counter-clockwise rotation of the African plate with respect to the Eurasian plate. This translates to pure right lateral strike-slip motion along the Gloria fault which changes progressively to N45°W trending compression at Gibraltar (figure 2.15). Focal mechanisms (Grimison & Chen, 1986, 1988; Bufern et al., 1988) suggest (figure 2.14) that a motion in the Gorringe Bank region retains a component of right lateral strike slip motion to the east of the Madeira-Tore Rise and last recognisable location of the Gloria fault on seismic reflection data (Peirce & Barton, 1991). Plate reconstructions show that right lateral strike-slip motion has been fairly stable on the Azores-Gibraltar plate boundary since chron 6 (20
Figure 5.18: Percentage sediment shortening in the Horseshoe abyssal plain, calculated from the interpretation of seismic reflection lines 9 and 10 from CD64 for (a) the base of units EH-I and WH-I (table 5.2) and (b) the base of units EH-III, WH-III, WH-IV and WH-V (table 5.3).
### Table 5.2: Shortening calculated for the base of units WH-I and EH-I from seismic reflection profiles 9 and 10 from CD64 in the Western and Eastern Horseshoe abyssal plains.

<table>
<thead>
<tr>
<th>CDP</th>
<th>LENGTH (m)</th>
<th>SHORTENING</th>
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<td>REFLECTION</td>
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<td>35693</td>
<td>35725</td>
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<td>47474 - 47710</td>
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Table 5.3: Shortening calculated from seismic reflection profiles 9 and 10 for the base of units EH-III, WH-III, WH-IV and WH-V of the Western and Eastern Horseshoe abyssal plains.

Comparison of the tectonic setting of the Gorringe Bank region with a strain ellipse model (Anderson, 1951) (figure 5.19) for the deformation expected in a regional strike-slip zone, shows that many of the features can be explained on the basis of the model. In theory, initially two Reidel Shears \((R_1 \text{ and } R_2)\) will develop at approximately 30° to direction of maximum principal compressive stress \(\sigma_1\). Movement will usually be concentrated on \(R_1\) which is synthetic to the direction of maximum resolved shear stress with minimal motion on the antithetic system, \(R_2\). Folding and thrust faulting will be orientated perpendicular to \(\sigma_1\) and normal faulting perpendicular to the minimum principal compressive stress \(\sigma_3\). In some cases synthetic P and antithetic X shears and sub systems associated with any of the other fault systems may develop.

Alignment of the direction of maximum principal compressive stress \((\sigma_1)\) in
Figure 5.19: Regional strain ellipse associated with a strike-slip fault system. The Reidel shear faults are synthetic $R_1$ and antithetic $R_2$ systems. In some systems synthetic P and antithetic X shears may also develop. Folds and thrust faults are developed at 90° to $\sigma_1$, whereas normal faults are developed 90° to $\sigma_3$.

The dextral strike-slip ellipse model (figure 5.19) near parallel to the plate motion vectors for the Gorringe region (figure 5.17) enables many features to be resolved. The majority of the compressive deformation features interpreted from the Gorringe Bank region (figure 5.2) have a strike near perpendicular to plate motion vectors and the direction of maximum compressive stress ($\sigma_1$). These include: deformation over and to the north and northeast of Coral Patch seamount; faulting and folding of sediments in the region of Hirondelle seamount and thrust faulting which bounds the Eastern Horseshoe Basin to the south.

Northwest trending normal faults, interpreted from bathymetric surveying on Josephine seamount (Matveyenkov et al., 1994), but not observed on GLORIA images, are near parallel to the plate motion slip vectors and perpendicular to the direction of minimum compressive stress ($\sigma_3$). The strike-slip fault between Gorringe Bank and Hirondelle seamount (figures 5.4, 5.13 and 5.2) probably relates to the $R_2$ Reidel shear with a similar motion sense as deduced from the focal mechanism of Bufor et al. (1988). The minimal lateral displacement expected along this fault is confirmed by GLORIA and seismic reflection data which show little fault expression in the Tagus and Western Horseshoe basins. The recent fault
Figure 5.20: Strike of compressional features and small outcrops (Red lines and "v" pattern on figure 5.2) versus latitude and longitude for the Gorringe Bank region.
in the Eastern Horseshoe abyssal plain (figures 5.11 and 5.12) does not fit the strain ellipse model if we assume that the feature represents an extensional fault as was previously concluded. The style of faulting as deduced from earthquake focal mechanisms (figure 5.13) is quite variable indicating that the Eastern Horseshoe Basin has a complex faulting pattern and therefore the fault may represent minimal subsidence in the Eastern Horseshoe Basin.

The nature of the NE-SW trending fault across the base of the Gorringe Bank interpreted from GLORIA images (figures 5.1 and 5.2) is unclear. Examination of seismic reflection profiles from CD64 which show minor compressional deformation and correlation with the strike-slip ellipse model (figure 5.19), suggest that the fault is a strike-slip fault with a compressional component.

Many focal mechanisms in the Gorringe region fit the ellipse model with the majority of strike-slip focal mechanisms (Grimison & Chen, 1986; Buforn et al., 1988; Grimison & Chen, 1988) approximately aligned to the $R_1$ Reidel shear and the strike of the thrust faulting mechanism of Grimison & Chen (1986) perpendicular to $\sigma_1$. The trend of magnetic anomalies from Gorringe Bank have two dominant orientations of N20-30°E and N110-130°E (Gorodnitskiy et al., 1988), which have been correlated with northeast trending faults and steep vertical fault scarps respectively (Gorodnitskiy et al., 1988). The N20-30°E orientated faults are near perpendicular to $\sigma_1$, possibly indicating that they may represent thrusts or strike-slip faults and the N110-130°E trending fault scarps probably represent NW-SE trending normal faults, similar to those observed on Josephine seamount (Matveyenkov et al., 1994).

A comparison of the strike of compressional features and small fault related outcrops with latitude and longitude (figure 5.20) shows that although there is some scatter, the strike of the features does not significantly vary with latitude, with perhaps a slight increase in strike towards the north. However, the strike of the features shows a clockwise rotation towards the east from about N45°E at 15°W in the region of Josephine seamount to N70°E at 10°W in the region of Coral Patch Ridge. This agrees with the change in the orientation of slip (figure 5.17) and plate motion azimuths, verging to an E-W trend towards the west. The strike of the recent compressional features is different to strike of seamounts suggesting a change in the compression direction with some utilisation of older deformational structures.
Chapter 6

Gravity, Magnetics and Modelling

6.1 Free-air Anomaly

The Free-air Anomaly (FAA) for the Gorringe Bank region (figure 6.1) closely mimics the bathymetry (figure 6.2) in most areas, with positive anomalies related to all bathymetric highs and negative anomalies related to the intervening abyssal plains.

The highest amplitude anomalies of up to approximately 400 mGal are observed over Gorringe Bank with anomalies of up to 200 mGal over the crest of Coral Patch and Ampere seamounts. Josephine seamount has anomalies of about 200 mGal in a regional anomaly for the Madeira-Tore Rise of about 80 mGal. The Horseshoe and Tagus abyssal plains show regional negative anomalies of less than -60 mGal and the Tagus abyssal plain has a linear gravity low of less than -60 mGal running parallel to the trend of the northern flank of Gorringe Bank.

Free-air anomaly and bathymetry (figure 6.3) profiles across the Gorringe Bank region with a NNW-SSE trend, perpendicular to the major bathymetric trend, were constructed from the gridded datasets used to construct the FAA and bathymetry maps (figures 6.2 and 6.1). Examination of these profiles provides a more direct comparison between the bathymetry and the FAA more clearly representing some of the subtle relationships. Line 1 (figure 6.3) shows that the FAA associated with the Western Horseshoe abyssal plain is lower than over the Seine abyssal plain and the double peaked FAA over Josephine seamount is of a low amplitude when considering the scale of the bathymetry. Profile 2 shows a high amplitude gravity peak associated with Coral Patch seamount similar to Ampere seamount on profile 1 and a gravity low associated with the transition from the Western to the Eastern Horseshoe abyssal plain. Hirondelle seamount, topographically much smaller than Josephine seamount (profile 1), shows a low amplitude FAA compared to the amplitude of the bathymetry. Profile 3 is interesting in that although the...
anomalies mimic the bathymetry, the amplitudes observed in the gravity are greater than observed for similar bathymetric scales on profiles to the west. This includes the relatively long wavelength (approximately 100 km), twin peaked FAA over the eastern end of Coral Patch seamount and the high amplitude FAA associated with the western end of Gorringe Bank. The FAA is also observed to slope towards the northern flank of Gorringe Bank, even though the bathymetry of the Tagus abyssal plain is at a constant depth, as also observed to a lesser extent on profile 4. These sections cut the linear FAA low (figure 6.1) associated with the base of the northern flank of Gorringe Bank. Coral Patch Ridge (Profile 4, figure 6.3) correlates with a long wavelength high amplitude FAA which is greatly accentuated in comparison to the bathymetric scale. Ormonde seamount also has a very high amplitude anomaly. A significant change in the FAA is observed on profile 5 with low amplitude anomalies associated with smaller bathymetric features. This may be due to the proximity of the margin with compensated features and no increase

Figure 6.1: Free-air Anomaly of Gorringe Bank region with interpretation lines (see figure 6.3). Red dash line marks the location of the linear free-air gravity low. Contour interval 30 mGal.
in the FAA as the bathymetry shallows toward the African margin. Overall there is a change in the style of the FAA from west to east with an increase in the amplitudes of the FAA compared to the bathymetric scale and then a decrease in the amplitudes of the FAA towards the Iberian and African margins.
Figure 6.3: Free-air anomaly and bathymetry profiles from NW to SE across the Gorringe Bank region (see figure 6.1). Abyssal plains: WHAP = Western Horseshoe; EHAP = Eastern Horseshoe; SAP = Seine; TAP = Tagus. Seamounts: JOSEPHINE = Josephine; AMP = Ampere; CP = Coral Patch; GB = Gorringe Bank; HIR = Hirondelle. Other: M-T RISE = Madeira-Tore Rise; CPR = Coral Patch Ridge; AFM = African margin.
6.2 Previous Gravity Models

The large amplitude free-air anomalies of the Gorringe Bank region have attracted a number of studies. Previous models have concentrated on the structure of Gorringe Bank, the Horseshoe abyssal plain, Coral Patch seamount and Coral Patch Ridge along mainly two dimensional profiles perpendicular to the main bathymetric trends. Some models were constructed parallel to the trend of Gorringe Bank and a single model has been constructed for Josephine seamount and the Madeira-Tore Rise.

Figure 6.4: Location of previous gravity model profiles. Models: Red lines = forward; Blue lines = inverse; Green lines = isostatic; Black lines = flexure. Dash lines show bathymetry with a 500 m contour interval. A-A’ by Le Pichon et al. (1970); B-B’ by Purdy (1974); C-C’ by Purdy et al. (1978); D-D’ by Souriau (1984); E-E’ by Chen & Grimison (1989); F-F’, G-G’ and H-H’ by Bergeron & Bonnin (1991) and I-I’ by Peirce & Barton (1991)

The first gravity model of the Gorringe Bank region was by LePichon et al.
Figure 6.5: Free-air anomaly forward model of Le Pichon et al. (1970) for profile A-A’. See figure 6.4 for location. D = Possible salt diapirism. Dash and solid lines represent calculated and observed interfaces respectively. Densities in g/cm$^3$. 
Figure 6.6: Sediment corrected Bouguer gravity forward model (Purdy, 1974), for profile B-B’. See figure 6.4 for location. Densities in g/cm$^3$ with a Bouguer density of 2.6 g/cm$^3$. Shaded bodies represent sediments, defined by seismic reflection profiles, whose effect has been removed from the Bouguer anomaly. Stars mark deep seismic refraction lines A/A-R and B/B-R (Purdy, 1974).

(1970) for Gorringe Bank, the Horseshoe abyssal plain and the eastern end of Coral Patch Ridge, along a two dimensional WNW-ESE trending profile (figure 6.5). The forward model, which provides a good fit to the observed FAA, was based on early observations of the structure of the Gorringe Bank including the interpretation of a nascent subduction zone and an opaque zone of deformed sediment in the Eastern Horseshoe abyssal plain. Also incorporated into the model were the formation of Gorringe Bank by the overthrust of oceanic crust and the resulting asymmetric sediment distribution. The model assumes that the crust is locally uncompensated and required a high density mantle below to model the observed FAA, although no other evidence exists for this high density zone.

The second model by Purdy (1974) was a two dimensional forward model (figure 6.6) NW-SE across Gorringe Bank, the Eastern Horseshoe abyssal plain and Coral Patch Ridge. The sediment corrected Bouguer anomaly (SCBA) was used
due to improved sediment constraint from seismic reflection and refraction data, with sediments in the Tagus Basin interpreted to terminate at the base of Gorringe Bank, yet still poorly constrained in the Horseshoe abyssal plain. The model differed from that of LePichon et al. (1970) in that the large positive FAA over Gorringe Bank was attributed to a shallow high density body. Densities were derived from the refraction model velocities from lines A, B-BR and C (Purdy, 1974) based on the method of Nafe & Drake (1963). These gave a low density mantle of 3.18 g/cm$^3$, but to provide the reasonable fit to the SCBA the model required significant crustal thinning between Gorringe Bank and Coral Patch Ridge and a high density body beneath Coral Patch Ridge.

The SCBA forward model of Purdy (1974) was updated by Purdy et al. (1978) in Bonnin (1978) to give a similar model along a NNW-SSE profile across Gettysburg seamount, the Eastern Horseshoe abyssal plain and Coral Patch seamount.
Figure 6.8: Geoid and free-air gravity isostatic model of Souriau (1984) for profile D-D', with Gorringe Bank in isostatic equilibrium. See figure 6.4 for location. Densities in g/cm³.
Figure 6.9: Simplification and recalculated of the forward model of Le Pichon et al. (1970) by Souriau (1984) along profile A-A’, using the geoid and free-air anomalies. See figure 6.4 for location. Densities in g/cm³.
Figure 6.10: Simplification and recalculated of the forward model of Purdy et al. (1978) by Souriau (1984) along profile B-B’, using the geoid and free-air anomalies. See figure 6.4 for location. Densities in g/cm$^3$. 
Improved sediment constraint, including the asymmetry of sediments on Gorringe Bank was provided by additional seismic reflection data. The high density body beneath Coral Patch Ridge was replaced by thin crust. The densities were again derived from refraction models and the structures assumed to be locally uncompensated and supported by a strong regional stress field.

Figure 6.11: Geoid anomaly forward model of Chen & Grimison (1989) along profile E-E’, using the concept of broken flexed crust beneath the load of Gorringe Bank. The solid curves below the bathymetry show calculated deflections of the Moho with an elastic plate thickness of about 50 km. It is assumed that the bathymetric highs and infill have a density of 2.8 g/cm$^3$. The crust is modelled with a thickness of 10 km over a 3.3 g/cm$^3$ mantle. See figure 6.4 for location.

The model of Souriau (1984) (figure 6.8) was centred on investigations as to the structure of Gorringe Bank. The Bank was modelled as a simple structure in Airy isostatic equilibrium (locally compensated) with a Bank density of 2.8 g/cm$^3$ assuming serpentinite supported by a low density deep root of 3.0 g/cm$^3$ to the maximum seismicity depth of about 60 km. The geoid anomaly derived from Seasat altimeter data was used to give constraint to modelling as it has a better
Figure 6.12: Three sections (Profiles F-F’, G-G’ and H-H’) through the three dimensional inverse gravity models of Bergeron & Bonnin (1991). Crustal density 2.7 g/cm³, sediment density 2.37 g/cm³, mantle density 3.15 g/cm³ and a mean body depth of 14 km. See figure 6.4 for location.
signal to noise ratio at long wavelengths than gravity and is hence more sensitive to deep crustal structure. The geoid maximum for Gorringe Bank is shifted slightly to the south possibly indicating a deep dense source, but more likely due to the geoid pull of Coral Patch and Ampere seamounts (Souriau, 1984). Although the model provides a reasonable fit to the crest of the Bank, the flanks are poorly modelled, probably due to the simplified sediment structure. Souriau (1984) say that it is difficult to explain how a deep root may have formed based on geological observations.

Souriau (1984) also investigated the geoid anomaly associated with the previous models of LePichon et al. (1970) (figure 6.9) and Purdy et al. (1978) (figure 6.10). The geoid anomalies were calculated for these models and the fit was found to be satisfactory indicating that the models were plausible based on the geoid anomalies. The geoid misfit in the Tagus abyssal plain in the reworked model of Purdy et al. (1978) may be due to the poor constraint of structure from the limited seismic data.

The model of Chen & Grimison (1989) along a ENE-WSW profile (figure 6.11) across Gorringe Bank and Western Horseshoe abyssal plain, again utilised geoid anomalies. However the model differed from those previously with crustal flexure incorporating two discontinuities associated broken flexed crust beneath the load of Gorringe Bank based on seismological data and a break and deflection beneath the Madeira-Tore Rise was required to fit the geoid anomalies. The fit of the model to the observed geoid is quite poor and little geological evidence for deep discontinuities of this form and orientation have been observed in the region.

The gravity model of Bergeron & Bonnin (1991) approached the structure of the whole Gorringe region using three dimensional inverse gravity models based on the method of Cordell & Henderson (1968), which produces a non-unique solution. Models were calculated from the smoothed reduced Bouguer anomaly using a crustal density of 2.7 g/cm$^3$, a sediment density of 2.37 g/cm$^3$, a mantle density of 3.15 g/cm$^3$ and a mean Moho depth of 14 km. The trend of the computed Moho surface beneath Gorringe Bank has a trend which is $15^\circ$ less than the bathymetric trend, also observed in the Bouguer anomaly. Three profiles through the model (figure 6.12) show computed surfaces which roughly parallel the topography. However, the surfaces rise dramatically and are slightly offset to the north beneath Gorringe Bank in accordance with thrusting and show a small upward deflection beneath Coral Patch seamount.
6.3 New Gravity Models

Preliminary to improved interpretation of the Gorringe Bank region, available bathymetry data were used to construct a three dimensional Bouguer anomaly. A useful next step is to apply a sediment correction as the sediment corrected Bouguer anomaly is an indicator of the compensation style, for example whether the crust thins or is thickened beneath topographic features. Simple isostatic models were constructed in an attempt to explain the Bouguer anomalies. The results from these models lead to the investigation of Coral Patch seamount, Coral Patch Ridge and Gorringe Bank with two dimensional inverse and forward models. These models were able to resolve the structure of Coral Patch seamount and Ridge and gave important results for further modelling of Gorringe Bank using continuous and broken plate flexure models.

6.3.1 Estimation of Sediment and Basement Density

A vital factor in accurate gravity modelling is the selection of the correct sediment and crustal density responsible for the observed gravity anomalies. To provide the best possible constraint to the density of material in the Gorringe Bank region all available sampling information and models from other regions were used to provide density constraint. The seawater density was taken to be 1.03 g/cm$^3$ based on standard seawater density calculations and the sediment densities were chosen to be 2.3-2.4 g/cm$^3$ depending on the burial depth and age.

<table>
<thead>
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<th>LITHOLOGY</th>
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<th>DENSITY (g/cm$^3$)</th>
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</thead>
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<td>1.175</td>
</tr>
<tr>
<td></td>
<td>200</td>
<td>1.35</td>
</tr>
<tr>
<td>Diatomaceous Ooze</td>
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</tr>
<tr>
<td></td>
<td>500</td>
<td>1.45</td>
</tr>
<tr>
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<tr>
<td></td>
<td>500</td>
<td>2.0</td>
</tr>
<tr>
<td>Terrigenous Sed</td>
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<td>1.53</td>
</tr>
<tr>
<td></td>
<td>500</td>
<td>2.09</td>
</tr>
<tr>
<td></td>
<td>1000</td>
<td>2.29</td>
</tr>
</tbody>
</table>

Table 6.1: Typical sediment density according to facies (Hamilton, 1976).
Laboratory measurements of sediment density provide an initial constraint. The grain densities of standard rock samples are 2.715 g/cm³ for calcite, greater than 2.65 g/cm³ for shale and 2.65 g/cm³ for quartz and clay (Dobrin & Savit, 1988). Typical whole rock densities are 2.0-2.65 g/cm³ for shales, 2.2-2.7 g/cm³ for sandstones and 2.25-2.8 g/cm³ for limestones (Garland, 1979).

Modelling of data from various DSDP sites from the Northwest Atlantic Ocean (LeDouran & Parsons, 1982) gave a mean sediment density of 1.59 g/cm³ at the seawater interface and 1.87 g/cm³ at a depth of 1000 metres. Hamilton (1976); Hamilton & Bachman (1982) calculated the in-situ density of samples from selected DSDP data covering various sediment facies (table 6.1) and environments (table 6.2).

Examination of sediment density from selected DSDP sites from the Eastern Atlantic margins are summarised in table 6.3. At DSDP sites where sediment sample density was not given, the Gamma Ray Attenuation Porosity Evaluator (GRAPE) density was used. The GRAPE measures the in-situ wet bulk rock density by the emission and reception of gamma rays through the core liner. The rays are absorbed or scattered as they travel through a rock and the degree of attenuation is related to the density and mineralogy of the rock (Boyce, 1976).

Although these studies provide useful information for the upper approximately 1000 metres for various sediment facies and environments, the maximum sediment thickness in the Gorringe region is about 5750 metres with a mean of about 1535
metres from sediment thickness compilations (section 4.3). The density for thicknesses greater than 1000 metres was estimated from the DSDP measurements and from the empirical porosity versus depth curves (figure 6.13) of Bond & Kominz (1984). These were constructed from various types of sedimentary deposit using general concepts of sediment compaction and diagenesis during burial and porosity values as a function of depth from drilled modern sedimentary deposits (e.g., Maxwell, 1964; Chilingarian & Wolfe, 1975, 1976; Magara, 1980).

\[
\rho_b = \phi \rho_w + (1 - \phi) \rho_g
\]  

(6.1)

This porosity-depth relationship was converted to density-depth (figure 6.14)

<table>
<thead>
<tr>
<th>SITE</th>
<th>DEPTH(m)</th>
<th>DENSITY(g/cm³)</th>
<th>REFERENCE</th>
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<tr>
<td>DSDP 135 (GRAPE)</td>
<td>0.75</td>
<td>1.57</td>
<td>Hayes et al. (1972)</td>
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<td></td>
<td>80.75</td>
<td>1.76</td>
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<td>175.25</td>
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<td>DSDP 136 (GRAPE)</td>
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<td>1.67</td>
<td>Hayes et al. (1972)</td>
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</tr>
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<td>Sibuet et al. (1979)</td>
</tr>
<tr>
<td></td>
<td>1000</td>
<td>1.85</td>
<td></td>
</tr>
<tr>
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<td>146.32</td>
<td>1.97</td>
<td>Lancelot et al. (1980)</td>
</tr>
<tr>
<td></td>
<td>602.13</td>
<td>2.06</td>
<td></td>
</tr>
<tr>
<td></td>
<td>891.23</td>
<td>2.05</td>
<td></td>
</tr>
<tr>
<td>DSDP 547 (SAMPLE)</td>
<td>53.8</td>
<td>1.80</td>
<td>Hinz et al. (1984)</td>
</tr>
<tr>
<td></td>
<td>400.3</td>
<td>2.18</td>
<td></td>
</tr>
<tr>
<td></td>
<td>907.4</td>
<td>2.59</td>
<td></td>
</tr>
<tr>
<td>ODP 637 (SAMPLE)</td>
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<td>1.77</td>
<td>Boillot et al. (1987)</td>
</tr>
<tr>
<td></td>
<td>230</td>
<td>2.45</td>
<td></td>
</tr>
<tr>
<td>ODP 900 (SAMPLE)</td>
<td>0.24</td>
<td>1.68</td>
<td>Sawyer et al. (1994)</td>
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<tr>
<td></td>
<td>796.35</td>
<td>2.86</td>
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</tbody>
</table>

Table 6.3: Summarised sediment sample (SAMPLE) and GRAPE densities from selected DSDP sites from the Eastern Atlantic margin.
CHAPTER 6. GRAVITY, MAGNETICS AND MODELLING

using the empirical formula (equation 6.1, where $\rho_b = \text{bulk density}$, $\rho_g = \text{grain density}$, $\rho_w = \text{water density}$ all in g/cm$^3$ and $\phi = \text{porosity}$ from 0 to 1, where 1 = 100% porosity) assuming the lower and upper ranges of grain density to be 2.65 and 2.7 g/cm$^3$ respectively. A mean depth of 1535.6 m for sediments in the region gives a density of between 2.25 g/cm$^3$ and 2.51 g/cm$^3$. Examination of only curves related to sediments sampled at DSDP site 120 (Ryan et al., 1973) and 135 (Hayes et al., 1972) give values of 2.25 g/cm$^3$ to 2.3 g/cm$^3$ at this mean depth.

The selection of crustal densities for gravity modelling was estimated from sampling, standard crustal densities and those used in previous gravity models. Samples of the crustal basement in the Gorringe Bank region are mostly of gabbro and serpentinised peridotite from Gorringe Bank (e.g., Ryan et al., 1973; Auzende et al., 1978). These have typical densities of 2.85-3.12 g/cm$^3$ and 3.15-3.28 g/cm$^3$ respectively, with standard oceanic crustal density at about 2.9 g/cm$^3$ (e.g., Dobrin & Savit, 1988; Allen & Allen, 1990). Previous gravity models have used the geological observations and standard values to constrain crustal density. LePichon et al. (1970) used a layer 2 and 3 density of 2.84 g/cm$^3$, whilst Purdy (1974) used a layer 2 density of 2.6 g/cm$^3$ for Gorringe Bank and Coral Patch Ridge and 2.72-2.84 g/cm$^3$ for the deeper crust (layer 3). A higher density of 2.8 g/cm$^3$ was
Figure 6.14: Density - depth calculated from the empirical porosity - depth curves (figure 6.13) (Bond & Kominz, 1984) using the formula of Wyllie et al. (1956) and a Grain Density of 2.65-2.7 g/cm$^3$.

used by Souriau (1984) in the simple model for the Gorringe Bank and a Bouguer density of 2.67 g/cm$^3$ was used in the gravity models of Bergeron & Bonnin (1991). Peirce & Barton (1991) used a oceanic crustal density of 2.8 g/cm$^3$ for the Western Horseshoe abyssal plain and slightly higher densities of 2.85-3.0 g/cm$^3$ for crust beneath Josephine seamount. These densities were derived from the refraction model velocities by the method of Nafe & Drake (1963) and are quite reliable based on the good fit obtained by the combination of seismic refraction and gravity modelling.

### 6.3.2 Crustal Thickness

In the construction of gravity models it is essential to know the crustal thickness and the depth of the Moho, at which the density increases significantly. Selection of crustal thickness for use in gravity models was derived from previously collected refraction data and previous gravity models. Refraction studies in the North Atlantic Ocean show a mean crustal thickness of about 6-8 km (White, 1992) with typically 30% layer 2 and 70% layer 3. Crustal thickness in the Gorringe Bank region is significantly lower and more variable based on the refraction data of Purdy (1974),
who presented crustal thickness for the Tagus, Western and Eastern Horseshoe and Seine abyssal plains (table 6.4).

Refraction experiments in the Tagus abyssal plain by Pinheiro et al. (1992) gave total crustal thicknesses of about 1.8 km. However the refraction experiments of Mauffret et al. (1989) gave layer 2 and 3 thicknesses of 1.6 km and 1.1 km respectively, a total crustal thickness 2.7 km. The refraction line of Peirce & Barton (1991) gave a Moho depth of about 12.5 km in the Western Horseshoe abyssal plain and refraction experiments in the Iberian abyssal plain, to the north of the Tagus abyssal plain, gave thicknesses of 5.7 and 6.5 km for oceanic crust and 3.5 and 3.6 km for continental crust (Whitmarsh et al., 1990).

### 6.4 The Bouguer Anomaly

A three dimensional Bouguer Anomaly was constructed for the Gorringe Bank region. The Bouguer anomaly is calculated from the subtraction of a Bouguer correction from the free-air anomaly. The Bouguer correction is the gravitational effect of the bathymetry, due to the mass deficiency of the water column with respect to the seafloor. The Bouguer correction is the anomaly which would be produced if the seawater was replaced with material of a density contrast of crust minus water (2.65 - 1.03 g/cm³) giving a greater Bouguer correction for increased seafloor depth. Bouguer correction calculation is dependent upon knowledge of the seafloor crustal density (Bouguer Density). It is likely that there are wide variations in seafloor density due to variations in seafloor material, but an average of 2.65 g/cm³ is a reasonable value based on estimates from sampling (Section 6.3.1). The Bouguer anomaly if correctly calculated will remove the gravitational affects of the

<table>
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<th>LAYER 3</th>
<th>MOHO</th>
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<td>1.0</td>
<td>11.2</td>
</tr>
<tr>
<td>A-R</td>
<td>1.9</td>
<td>4.5</td>
<td>14.5</td>
</tr>
<tr>
<td>B</td>
<td>-</td>
<td>1.3</td>
<td>12.3</td>
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<tr>
<td>B-BR</td>
<td>-</td>
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<td>12.0</td>
</tr>
<tr>
<td>C</td>
<td>1.0</td>
<td>2.1</td>
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</tr>
<tr>
<td>D</td>
<td>-</td>
<td>2.5</td>
<td>12.0</td>
</tr>
</tbody>
</table>

Table 6.4: Crustal layer thickness derived from seismic refraction models of Purdy (1974). See figure 2.8 for locations.
Figure 6.15: Bouguer anomalies for the Gorringe Bank region. (a) Bouguer density $= 2.45 \text{ g/cm}^3$. (b) Bouguer density $= 2.65 \text{ g/cm}^3$. Contour interval 30 mGal.
Figure 6.15: Continued Bouguer anomalies for the Gorringe Bank region. (c) Bouguer density = 2.8 g/cm³. (d) Bouguer density = 2.85 g/cm³. Contour interval 30 mGal.
bathymetry, revealing information about the sub seafloor density distribution and structure.

The Bouguer correction was calculated using the "G3DGRIDG" software which utilises the Fourier transform method of Parker (1972) to calculate the gravitational effect of the uneven seafloor layer with a uniform density contrast. A range of Bouguer densities were chosen to access the sensitivity of the Bouguer correction to density variation (figures 6.15a to d).

A Bouguer density of between 2.45 and 2.85 g/cm$^3$ does not drastically affect the resulting Bouguer anomaly (figures 6.15a to d). Two Bouguer densities were chosen for interpretation. A density of 2.65 g/cm$^3$ (figure 6.15 b) was chosen based on standard Bouguer densities and 2.8 g/cm$^3$ (figure 6.15 c) based on the sampled high density rocks from Gorringe Bank (e.g., Ryan et al., 1973; Auzende et al., 1978). The anomalies resulting from the use of a higher Bouguer density have lower amplitudes, between bathymetric highs and lows, due to the amplification of the Bouguer correction for deeper regions, yet remaining small for shallow regions.

The Bouguer anomaly of the Gorringe Bank region (figures 6.15a to d) is dominated by a north-south regional high which extends from the Tagus to the Seine abyssal plain. The high is flanked by Bouguer anomaly lows associated with the Iberian margin to east and the Madeira-Tore Rise to the west. Superimposed on these regional anomalies are a number of short wavelength features. Gorringe Bank is associated with a high amplitude Bouguer anomaly which trends N44°E. Interestingly, this differs from the bathymetry of the Bank by approximately 14°.

The Bouguer anomaly is fairly narrow over the crest of Gettysburg and Ormonde seamounts, but with a broader base towards the northeast. Hirondelle seamount shows a lower Bouguer anomaly, with a relative low over the Horseshoe abyssal plain which closely follows the bathymetric trend and increasing anomalies to the south. Coral Patch Ridge shows a very long wavelength high amplitude Bouguer anomaly which decreases towards the west, with lower amplitude highs associated with Coral Patch and Ampere seamounts.

The Sediment Corrected Bouguer Anomaly (SCBA) is calculated from the Bouguer anomaly by the subtraction of a sediment correction. The sediment correction attempts to compensate for the mass deficiency of sub seafloor sediments which are of lower density than that assumed for Bouguer correction calculation. Accurate calculation of the gravity effect of the three dimensional sub seafloor sediment bodies is dependent upon knowledge of their shape and density. The form and thickness of the sediments shown in figure (4.14) and a sediment density of 2.3 g/cm$^3$, estimated from the values presented in section 6.3.1, were used for
Figure 6.16: Sediment Corrected Bouguer Anomalies. SD = 2.3 g/cm³. (a)BD = 2.65 g/cm³ and (b)BD = 2.8 g/cm³. Contour interval 30 mGal.
sediment correction calculation utilising the Fourier transform method of Parker (1972). The sediment corrected Bouguer anomaly calculated for Bouguer densities of 2.65 g/cm$^3$ and 2.8 g/cm$^3$ (figures 6.16a and b).

The main effect of the sediment correction was to raise the general level of the Bouguer anomaly associated with the Tagus, Horseshoe and Seine abyssal plains (figures 6.16a and b), as they have greatest sediment thicknesses of the region. This reduces the anomalies related to the short wavelength features along the N-S trending regional high. The flanking lows are also reduced, although they are still very apparent. A large gravity high associated with Gorringe Bank forms a distinctive NE-SW linear anomaly with a trend similar to the uncorrected Bouguer anomaly with little change in the observed anomalies. The gravity high associated with Coral Patch Ridge and seamount show similar characteristics to the uncorrected Bouguer anomaly with a large gravity high beneath Coral Patch Ridge and an ENE-WSW trending linear medium high anomaly over Coral Patch seamount, which begins to disappear for a Bouguer density of 2.8 g/cm$^3$.

6.5 Isostatic Gravity Anomalies

To investigate the isostatic characteristics of the Gorringe Bank region, isostatic anomaly models were produced. Isostatic anomaly models are derived from the subtraction of an isostatic correction from the uncorrected Bouguer anomaly and can reveal if features within a region are isostatically compensated or supported by crustal flexure. The isostatic correction is a calculation of the gravitational effect of the lower layers of the crust assuming airy isostasy or flexure. Isostatic anomalies were calculated with the "G3DGRIDG" software (Smith 1986), which utilises the Fourier transform method of Parker (1972), assuming a continuous three dimensional plate with elastic thicknesses of 0 (Airy), 5, 15, 35 and 75 km.

The density of the upper crust was taken to be 2.65 g/cm$^3$ with a 2.3 km thick layer 2 of 2.85 g/cm$^3$, a 5.0 km thick layer 3 of 2.9 g/cm$^3$ and an upper mantle density of 3.33 g/cm$^3$ based on standard values. The isostatic anomaly for an elastic plate thickness ($T_e$) of 15 km was calculated with a variation of crustal density from 2.45 to 2.85 g/cm$^3$ and mantle density of 3.13 to 3.53 g/cm$^3$ to access the isostatic anomaly sensitivity to density variation (figure 6.17). Changes in crustal density along line 4 (figure 6.1 for location) have little effect over the abyssal plains, but shows significant variation over Gorringe Bank. Mantle density variation (figure 6.17) shows little affect as would be expected due to the attenuation of deep sourced gravity anomalies.
Figure 6.17: Effect of crust and mantle density variation on isostatic anomaly models along line 4. Figure 6.1 for location.
Figure 6.18: Isostatic Anomalies: (a) Airy (b) $T_e = 5$ km. Crust = 2.65 g/cm$^3$, Layer 2 = 2.3 km at 2.85 g/cm$^3$, Layer 3 = 5.0 km at 2.9 g/cm$^3$ and mantle density = 3.33 g/cm$^3$. Contour interval 30 mGal.
Figure 6.18: Continued Isostatic Anomalies: (c) $T_e = 15$ km (d) $T_e = 35$ km. Crust $= 2.65 \text{ g/cm}^3$, Layer 2 = 2.3 km at 2.85 g/cm$^3$, Layer 3 = 5.0 km at 2.9 g/cm$^3$ and mantle density = 3.33 g/cm$^3$. Contour interval 30 mGal.
Figure 6.18: Continued Isostatic Anomalies: \( e \) for \( T_e = 75 \) km. Crust = 2.65 g/cm\(^3\), Layer 2 = 2.3 km at 2.85 g/cm\(^3\), Layer 3 = 5.0 km at 2.9 g/cm\(^3\) and mantle density = 3.33 g/cm\(^3\). Contour interval 30 mGal.

Application of an isostatic correction to the uncorrected Bouguer anomaly removes the N-S trending regional high and flanking lows indicating that this regional anomaly may be due to a crustal root.

The Airy Isostatic anomaly (figure 6.18 a) is very similar to the FAA (figure 6.1) with high amplitude anomalies associated with the seamounts and Gorringe Bank, as the roots are too narrow to have much effect on the Bouguer anomaly. Surprisingly though neither does the Madeira-Tore Rise.

Increases in elastic thickness (figures 6.18b to 6.18e) result in a gradual reduction in the gravity anomaly amplitudes. An elastic thickness of 5 km shows a slight reduction in amplitude of features to the west, including the Madeira-Tore Rise, but large gravity highs associated with the seamounts and relative lows over the abyssal plains.

At a \( T_e \) of 15 and 35 km there is little or no correlation with the bathymetry. Gorringe Bank shows some amplitude reduction and a quite substantial gradient decrease. The Bank also shows a counter clockwise rotation of 10\(^\circ\) to a trend of
N46°E, in comparison to a trend of about N56°E in the FAA (figure 6.1) and Airy isostatic anomaly (figure 6.18 a), similar to the inverse gravity models of (Bergeron & Bonnin, 1991). The ridge connecting Gorringe Bank to the Madeira-Tore Rise is decreased and the Madeira-Tore Rise itself is now marked by negative anomalies. Coral Patch Ridge, Coral Patch seamount and Ampere seamount show a linear E-W high which does not parallel the bathymetry (figure 6.2) or FAA (figure 6.1). The NE-SW trending anomalies near to the margin are more prominent as the margin effect has been removed.

A Tc of 75 km (figure 6.18 e) still does not remove the gravity effects of Gorringe Bank, Coral Patch Ridge and Coral Patch seamount showing that they are not in isostatic equilibrium, as found by gravity models (e.g., Souriau, 1984) or completely supported by crustal flexure and must therefore be supported by some other mechanism (Purdy, 1974). The observed reduction in the anomalies may indicate a component of regional compensation or that the features are uncompensated.

### 6.6 Crustal Gravity Models

#### 6.6.1 Madeira-Tore Rise Isostatic Models

The Madeira-Tore Rise, which forms the western boundary of the Gorringe Bank region, is a NNE-SSW trending continuous ridge connecting Madeira Island (33°N, 17°W) and Tore seamount (39°N, 13°W). The Rise has variable dimensions of about 200 km width and rising 2000 to 3000 metres above the 5000 metre abyssal depths, due to the number of small seamounts on the crest and flanks. It is thought to have formed as a result of the southward movement of magma along the Mid-Atlantic ridge axis vented from a mantle plume, from about M-4 to M-O (approximately 126-119 Ma) (Tucholke & Ludwig, 1982). The J magnetic anomaly (Pitman & Talwani, 1972) which is related to the formation of the Madeira-Tore Rise is thought to have an age of between M-0 (118 Ma) and M-1 (122 Ma) (Rabinowitz et al., 1979).

A NW-SE refraction line survey (figure 2.12) provided a model of the crustal structure of the Madeira-Tore Rise and Josephine seamount (Peirce & Barton, 1991). The crustal velocities and densities are neither typical of oceanic or continental crust, but are thought to represent oceanic crust non-tectonically thickened perpendicular to the east-west trend of the magnetic spreading anomalies. The Moho depth varies from 11 km north of the Madeira-Tore Rise to 17.5 km beneath Josephine seamount and 12 km to the south beneath the Western Horseshoe abyssal plain.

A gravity forward model (figure 6.19) based on seismic refraction model (fig-
Figure 6.19: Free-air gravity forward model calculated from the seismic refraction model of Peirce & Barton (1991) along profile I-I'. Densities in g/cm$^3$. See figure 6.4 for location.

The Madeira-Tore Rise exhibits positive Airy isostatic anomalies (figure 6.18a), indicating less contribution from a root and that the crust may be flexed. For a $T_c$ increasing from 5 km (figures 6.18b to 6.18e) the isostatic anomalies show a rapid decrease to highly negative anomalies, whilst other features of the Gorringe region still show significant, yet reduced, positive isostatic anomalies. This is related to the formation mechanism of the Madeira-Tore Rise.
To investigate the compensation of the Madeira-Tore Rise, two dimensional calculated isostatic models were constructed for lines A and C (figures 6.21 and 6.20). These are a model of the gravity effects of the bathymetry combined with the gravity effect of the compensation calculated for various values of elastic thickness. Line A produces a best fit to the observed FAA with an elastic thickness of about 15-25 km. However, the model is unable to explain the regional observed FAA gradient which is even more apparent in the two dimensional model for Line C (figure 6.21). This slope is probably a result of the three dimensionality of the structures and the flexural effects of the large loads of the Gorringe Bank region to the east which are depressing the FAA over the abyssal plains.

To account for the three dimensional effects, three dimensional calculated isostatic models were constructed for the Madeira-Tore Rise. Lines A to D (figures 6.20, 6.22 and 6.23 through the model, over estimate the regional anomaly gradient except for line C which shows a similar gradient to the observed. A best fit to the
observed FAA is at an elastic thickness of about 15 km, based on the amplitude and wavelength of the anomalies.

A $T_e$ of about 15 km suggests that the Madeira-Tore Rise in this region formed on 20 Ma oceanic crust based on the $T_e$ versus age curves (figure 6.24) (Watts, 1994) and the 450°C isotherm. This does not agree with Madeira-Tore Rise formation at the Mid Atlantic Ridge (Tucholke & Ludwig, 1982), which corresponds (figure 6.24) to an elastic thickness of lower than 5 km from the models of Peirce & Barton (1991). The higher $T_e$ suggests that a large volume of volcanic material was added as the Madeira-Tore Rise moved away from the ridge axis. Miocene to Pliocene age volcanics on Ampere seamount (Matveyenkov et al., 1994) and 60 Ma volcanic material recovered (Feraud et al., 1982, 1986) from Gorringe Bank indicate that volcanism has been active, since initial Madeira-Tore Rise formation, in the Gorringe Bank region and provide evidence for the addition of material to the Madeira-Tore Rise.

### 6.6.2 Gorringe Bank, Coral Patch Ridge and Coral Patch Seamount Inverse Models

In an attempt to resolve the deep structure of features of the Gorringe Bank region, which isostatic anomalies have shown to probably partially compensated or uncompensated, two dimensional inverse gravity models (figures 6.25 and 6.26a, b and c) were constructed for three profiles across Gorringe Bank, Coral Patch Ridge and Coral Patch seamount.

The sediment corrected Bouguer anomaly was used as it is likely that the principal contributor to these anomalies (especially the long wavelength component) is the Moho. The inverse model approximates the source of the sediment corrected Bouguer anomaly to a single surface at the approximate depth of the Moho.

Inverse models were calculated with the GRAVOS software (Bott, 1967; Tanner, 1967; Laving, 1972) which calculates the shape of a two dimensional outward sloping body with a single density contrast by linear approximations to a non-linear problem. With an approximation to the three dimensionality of structures provided by the application of end corrections (Nettleton, 1940). The top of the body, the constant density contrast (0.68 g/cm³) and the number of blocks into which the body is segmented, have to be defined. The sediment corrected Bouguer anomaly, with an assumed regional field removed is used as we are trying to model the Moho structure and not the long wavelength anomalies of the regional field.

The anomaly is represented by a surface mass distribution approximated by a number of rectangular blocks of varying density in two dimensions. The block’s
densities are adjusted iteratively to give a series of blocks whose calculated gravity anomaly nearly satisfy, in the case of the least squares solution, the sediment corrected Bouguer anomaly. The blocks are then transformed to blocks of uniform density contrast (Tanner, 1967). The solution is unique, but is unrealistic for multi-layered geological structures such as oceanic crust where a single body with a constant density contrast is unlikely to be responsible for the observed anomalies, but may give an indication of the general structural trends.

Analysis of the stability of the method with varying input parameters showed that the model was not affected by the resolution of the input gravity observations, as long as there were a number of points for each block. A large number of blocks into which the body is divided gives unrealistic results, but is stable for between 10 and 20 blocks for these profiles. Selection of the depth to the top of body produces more spikey results with depth, but also affects the body depth beneath the abyssal plains to a level comparable to the Moho modelled from refraction data (table 6.4) (Purdy, 1974). An increased density contrast results in a flatter body.

**Coral Patch seamount**

The structure of the Moho beneath Coral Patch seamount (Profile AG-BG, figures 6.25 and 6.26a) roughly parallels the basement topography with a peak under the crest of the seamount and a flat surface to the NNW indicating the presence of high density material at a shallow depth. The main Moho peak is offset to the NNW of the bathymetric peak as reflected in the SCBA. The flat surface which is not observed on the bathymetry, but observed in the SCBA, occurs in the region of deformation observed on seismic reflection profiles (figures 5.8 and 5.9) and GLORIA data (figures 5.1 and 5.2) to the north of Coral Patch seamount.

**Gorringe Bank**

The modelled structure for Gorringe Bank (Profile EG-HG, figures 6.25 and 6.26b) shows a wide Moho elevation beneath the crest of the Bank which is offset slightly to the north. This is in agreement with seismic reflection data (e.g., LePichon *et al.*, 1970; Sartori *et al.*, 1994) and sampling (e.g., Auzende *et al.*, 1982) which show basement outcrop on the northern flank and a fairly thick sediment cover on the southern flank. A second smaller elevation to the south is required by the inverse model to fit the associated minor peak in the sediment corrected Bouguer anomaly. Examination of this anomaly (figure 6.16a) reveals that the small SCBA peak is only present at the ENE end of the southern flank of Gorringe Bank and is likely to be the result of a shallow body or over compensation of the sediments due to
an incorrect sediment density estimation or poor sediment thickness constraint in this region. Overall the Moho is roughly parallel to the basement topography with deepening beneath the eastern Horseshoe Basin, but a slight peak in the centre of the eastern Horseshoe Basin.

**Coral Patch Ridge**

The model for Coral Patch Ridge (Profile HG-IG, figures 6.25 and 6.26c) shows an upwarp in the Moho which generally follows the form of the SCBA and not that of the basement topography. The SCBA high is offset slightly to the north of the topographic high and Moho highs are often offset from basement highs indicating a complex crustal structure with variable crustal thickness and the presence of high density material at an anomalously shallow depth.

### 6.6.3 Gorringe Bank and Coral Patch Ridge Forward Models

Results from the inverse models was used to provide initial constraint to forward models for profile DG-IG from line 4 (figure 6.25), for further investigation of the structure of Gorringe Bank and Coral Patch Ridge. Two models were constructed to model the SCBA with layers relating to a lower crustal layer and Moho beneath an upper crust with a Bouguer density of 2.65 g/cm$^3$ (figure 6.27a) and 2.8 g/cm$^3$ (figure 6.27b), a sediment density of 2.3 g/cm$^3$ and a water density of 1.03 g/cm$^3$. Moho depth and crustal thickness constraint were provided by previous seismic refraction data (e.g., Purdy, 1974, 1975) (See section 6.3.2). The lower crustal layer density was assumed to be 2.9 g/cm$^3$ with a near constant thickness of 2 km and the density of the upper mantle to be 3.33 g/cm$^3$. Calculation of the gravitational effect of lower crustal layer and Moho was provided by the "GBOTT" software which utilises the method of Talwani *et al.* (1959) for the calculation of the gravity effect of a series of semi-infinite slabs.

Overall a good fit has been obtained between the observed and calculated anomalies.

**Tagus abyssal plain**

In the Tagus abyssal plain it was necessary to have long wavelength oscillations in lower crustal layer at about 10-12 km for a Bouguer density of 2.65 g/cm$^3$ and greater at 10.5-12.5 km for a Bouguer density of 2.8 g/cm$^3$ to fit the observed long wavelength oscillations in the SCBA, although the general anomaly trend could be
fitted with a flatter lower crustal layer. The long wavelength anomalies are possibly a result of poor sediment thickness and density constraint in SCBA calculation, probably due to the variable basement relief and sediment thickness observed on seismic reflection data from the Tagus abyssal plain (figure 4.2 and Mauffret et al. (1989)). Towards Gorringe Bank, the lower crustal layer dips at about 5° (2.65 g/cm³) to 7° (2.8 g/cm³) to the SSW which may represent an underthrust slab of crust related to the thrusting mechanism of formation as proposed by LePichon et al. (1970)

**Gorringe Bank**

A fit to the high amplitude, short wavelength SCBA related to Gorringe Bank was obtained by a Moho depth of 6 km (2.65 g/cm³) to 7 km (2.8 g/cm³) with a horizontal length of about 30 km and the absence of a lower crustal layer. It is difficult to relate the modelled structure geologically, but a shallow Moho depth would be expected for an overthrust. The small SCBA peak to the south could be fitted with a second smaller shallow Moho peak, as was suggested by the inverse model (figure 6.26b), but was assumed to be due to poor sediment thickness or density control in this region.

**Eastern Horseshoe abyssal plain**

The central eastern Horseshoe Basin required a slightly upwarped lower crust and Moho to fit the subtle gravity anomaly high, leaving a very thin upper crust. This upwarp was slightly greater for a Bouguer density of 2.8 g/cm³ with an increased dip of the lower crust and Moho toward Gorringe Bank. The high may be due to poor constraint of the sediment density or thickness. Below the southern eastern Horseshoe Basin, the lower crust and Moho dip slightly towards the south, rising at an angle of about 3° (2.65g/cm³) or 7° (2.8 g/cm³) at the southern edge of the eastern Horseshoe Basin beneath Coral Patch Ridge.

**Coral Patch Ridge**

A good fit was obtained beneath Coral Patch Ridge with an elevated lower crust and Moho associated with the SCBA, but offset from the bathymetric and basement topographic highs, as observed in the inverse model (figure 6.26c). Use of a higher Bouguer density of 2.8 g/cm³ results in higher amplitudes of the lower crust and Moho, yet with a similar structure.

This suggests that the structure beneath Coral Patch Ridge is related to whole crustal buckling in combination with thrust faulting of the brittle upper crust.
Thrust faulting is observed to cut the sediments and upper crust on seismic reflection profiles (e.g., LePichon et al., 1970; Sartori et al., 1994) and profile V-W-X (figure 5.10) from CD64 and possible shallowing Moho may have been imaged by reflection profiles of Sartori et al. (1994). Thrust faulting related features have also been traced on GLORIA images from the region (figures 5.1 and 5.2).

6.6.4 Comparison with the Indian and NW Atlantic Oceans

The central Indian and northwestern Atlantic Oceans are regions where structural and gravity characteristics similar to the Gorringe Bank region have been observed.

The Central Indian Ocean in the region of the Afranasy-Nikitin seamounts (78°E to 88°E, 10°S to 5°N), has many features which are similar to those observed in the Gorringe Bank region, but specifically associated with the region of Coral Patch Ridge.

The seismicity of the central Indian Ocean is widely distributed (Weissel et al., 1980) and dominated by shallow compressive earthquakes associated with a complex roughly north-south trending convergence (Bull, 1990) of about 10 mm/yr (Karner & Weissel, 1990) between the Indo-Australian and Eurasian plates, with focal depths concentrated at 27-39 km (Bull & Scrutton, 1990).

Associated with the compression which was initiated in the late Miocene (7-8Ma) (Karner et al., 1993), possibly related to the Himalayan orogeny (Weissel et al., 1980), are high angle reverse faults (Weissel et al., 1980) and large scale folding (Weissel et al., 1980). The faults parallel the magnetic spreading lineations indicating that they are reactivations of ridge faults (Karner et al., 1993). They have a spacing of about 5-20 km (Karner & Weissel, 1990), throws of up to 600 m (Bull & Scrutton, 1992) and dips of 35-45° at basement depths to 40-90° in the sediment cover. Some of the faults have hanging wall anticlines, sometimes paired with footwall synclines (Bull & Scrutton, 1992). Heat flow is high at 107±36mW/m² (Weissel et al., 1980), considering the crustal age of 60 Ma, due to deep faulting and hydrothermal upwelling (Karner et al., 1993).

The large scale east-west trending sediments and basement folds, which are associated with the faulting, have wavelengths of 100 to 300 km and amplitudes of up to 3 km (Weissel et al., 1980; McAdoo & Sandwell, 1985; Karner et al., 1993), with associated gravity anomalies of 15-80 mGal (McAdoo & Sandwell, 1985; Karner et al., 1993). Some of these folds are cut by the reverse faults and some normal faults (Bull & Scrutton, 1992).

Deformation in the central Indian Ocean has shortened the crust mainly by faulting (Bull & Scrutton, 1992), but also by whole crustal buckling. This buckling
has been aided by the sediments which were deposited before deformation and the infilling of fold troughs by mainly turbidites during deformation (Zuber, 1987; Karner et al., 1993). The most intense deformation is located around the Afranasy-Nikitin seamounts which pre-date the compression (Karner et al., 1993), indicating that the pre-existing deflections caused by the seamounts, focussed later deflections.

Compressional deformation in the Gorringe Bank region has only been active for a slightly longer period than in the central Indian Ocean, probably with a similar convergence rate of about 11 mm/yr during the past 10 Ma (Purdy, 1974, 1975) (Current rate of about 4 mm/yr (DeMets et al., 1990)), widespread seismicity of a depth generally less than the 50 km (Chen & Molnar, 1983) and an anomalously high heat flow. Coral Patch Ridge shows many similar features to the central Indian Ocean, including high angle reverse faulting with throws of up to 500 metres (figure 5.10), broad sediment folding and high gravity anomalies (figure 6.1). Thrust faults in the central Indian Ocean formed along lines of weakness associated with the former ocean spreading ridge normal faults (Karner et al., 1993). Thrust faults on Coral Patch Ridge may similarly represent reactivations of lines of weakness related to movements along the Eurasian-African plate boundary. Short wavelength folding on Coral Patch Ridge is related to the faulting, with hanging wall anticlines and footwall synclines. Regional long wavelength folding is associated with undulations in the basement, which are related to the thrust faults. The wavelength of warps in the central Indian Ocean is 100-300 km and is modelled to be about 200 km for regional buckling of Coral Patch Ridge (figures 6.27a and 6.27b), indicating that buckling is a possible explanation. The shallowing of the Moho in the central Indian Ocean has been proposed to be by a combination of thrust faulting and whole crustal warping (Bull & Scrutton, 1992). The similar structure of Coral Patch Ridge to the central Indian Ocean and the results of gravity forward models (figures 6.27a and b) lead to the conclusion that Coral Patch Ridge was formed by a combination of thrust faulting and crustal warping as a result of compression between the Eurasian and African plates. In the central Indian Ocean the deformation is concentrated around the pre-deformation Afranasy-Nikitin seamounts. Gorringe Bank and the great thickness of sediments in the Eastern Horseshoe Basin, may have acted in a similar way in the deformation of Coral Patch Ridge.

The northwestern Atlantic Ocean in the region of the Fifteen-Twenty fracture zone (46-60°W, 14-18°N), east of the Lesser Antilles volcanic arc, also exhibits features with some similar characteristics to those of the Gorringe Bank region. Transtension and crustal thinning in the Late Cretaceous was followed by Tertiary
transpression (Muller & Smith, 1993), due to a differential rotation pole between the North and South American plates, along what may have been the plate boundary between the North and South American plates at this time (Roest & Collette, 1986). This compression was responsible for the formation of the Barracuda and Tiburon Ridges and related to the Fifteen-Twenty fracture zone. The Ridges are associated with large Bouguer anomalies of up to approximately 135 mGal and sediments show mainly normal faulting with a few thrust faults (Peter & Westbrook, 1976; Mascle & Moore, 1990), in contrast to thrust faulting in the Gorringe Bank region. Crustal gravity models suggest that the Moho is elevated by 2-4 km over short wavelengths of about 70 km at the Barracuda and Tiburon Ridges, with thinned crust with a thickness of 1.5-2 km. Focal mechanisms from north of the Barracuda Ridge indicate compression, with a strike-slip component, which presently maintains the uncompensated features. The elevation of the Moho is thought to be due to crustal faulting (Muller & Smith, 1993). Models show that crustal buckling, as observed in the central Indian Ocean (Karner & Weissel, 1990), cannot explain the short wavelength features of the Barracuda and Tiburon Ridges, as curvatures would be required in excess of the elastic limit of the crust (McNutt, 1984).

6.7 Gorringe Bank Flexure Models

Although inverse and forward modelling is useful, in that it provides information on the mass distribution required to account for the observed gravity anomaly for Gorringe Bank, it provides little information on the formation process. LePichon et al. (1970) proposed that Gorringe Bank was formed by overthrusting of the African plate upon the Eurasian plate. If this is the case we would expect to see flexural features to north of the Gorringe Bank. Evidence for flexure is provided by a linear gravity low parallel to the northwestern flank of Gorringe Bank in the Tagus abyssal plain (figure 6.1) and seismic profile IAM line 4 (Banda & Torne, 1995) (figure 6.28) in the Tagus abyssal plain. This clearly shows an undeformed wedge of post loading sediments separated from flexed pre-loading sediments, which fill the faulted basement blocks, by a distinct unconformity. The age of this unconformity may be related to the Early Eocene to Late Oligocene unconformity observed in DSDP 135 (Hayes et al., 1972). However tentative correlations with sequence IA from Mauffret et al. (1989) and DSDP site 120 (Ryan et al., 1973) give an age of oldest Middle Miocene for the unconformity in the Tagus abyssal plain. The post loading sediments show fanning reflections which dip slightly towards Gorringe
Bank. These indicate that the majority of Eurasian plate flexure occurred at the time of the unconformity. However flexure and subsidence have not ceased, with the planar character of the Tagus abyssal plain (figure 6.2) resulting from a sediment supply greater than the subsidence. Debris flows which lie upon or marginally above the unconformity were derived from Gorringe Bank during or just after overthrusting (figures 6.28 and 4.3)

6.7.1 Flexure Calculation Methodology

The lithosphere can be approximated by a thin rigid plate floating on a weak underlying substratum (Barrell, 1914a,b). When a load is applied to the plate by the formation of a volcanic seamount, as in the Hawaiian seamount chain (e.g., Walcott, 1970a,b; Watts, 1978), the plate will bend in an attempt to compensate for the additional load. Seismic data show that deformation at Hawaii can be approximated by a thin elastic plate, where the amplitude of bending is related to the strength of the lithosphere (Walcott, 1970b).

The basic differential equation (equation 6.2) for the bending of a thin elastic plate with a constant elastic thickness is given by:

\[
D \frac{d^4 w(x)}{dx^4} + (\rho_m - \rho_{infill}) gw(x) = q(x)
\]  

(6.2)

where \(w(x)\) = vertical plate deflection as a function of horizontal distance, \(\rho_m\) = mantle density, \(\rho_{infill}\) = infill density, \(g\) = gravitational acceleration, \(q(x)\) = load function as a function of horizontal distance and \(D\) = the flexural rigidity which is given by:

\[
D = \frac{ET_e^3}{12(1-\sigma^2)}
\]  

(6.3)

where \(T_e\) = Elastic thickness of the plate, \(E\) = Young’s Modulus and \(\sigma\) = Poisson’s Ratio.

The equation for flexure when \(T_e\) varies with distance or in the case of a semi-infinite slab (broken plate) the \(T_e\) becomes zero at the break is given by:

\[
D(x) \frac{d^4 w(x)}{dx^4} + 2 \frac{dD(x)}{dx} \frac{d^3 w(x)}{dx^3} + \frac{d^2 D(x)}{dx^2} \frac{d^2 w(x)}{dx^2} + (\rho_m - \rho_{infill}(x)) gw(x) = q(x)
\]  

(6.4)

This was derived from Hetenyi (1946) and has been solved by finite difference methods (Bodine, 1981; Stewart & Watts, 1996).
6.7.2 Gorringe Bank Continuous Plate Flexure Model

To test the applicability of flexure models to the Gorringe Bank region, a simple model was assumed in which Gorringe Bank loaded a continuous elastic plate. The $\tau_e$ was assumed to be constant, but the infill density was varied to take account of the higher density infill material directly beneath Gorringe Bank.

A two dimensional continuous plate flexure model (figure 6.29) has been applied to line 4 (figure 6.25) across the Tagus abyssal plain, Gorringe Bank and Eastern Horseshoe abyssal plain, centred on the crest of Gorringe Bank (FG, figure 6.29). The model assumes that the whole Gorringe Bank load is supported by a crustal plate of constant lateral strength. The load base, which correlates with the top of the pre-loading sediments, if it is assumed that the seafloor was originally undeformed, was estimated from equation (6.5) of Stein & Stein (1992):

$$d(t) = 5651 - 2473e^{-0.0278t}$$  

(6.5)

where $d(t) =$depth (metres) at time $t$ (Ma). This was derived from a study of heat flow and bathymetry data from the North Pacific and Northwest Atlantic Oceans. An age of approximately 140 Ma for the formation of the Gorringe Bank basement rocks (Prichard & Mitchell, 1979; Feraud et al., 1982, 1986), gives a load base depth of about 5600 metres.

The pre-loading sediment thickness is lower in the Tagus than the Eastern Horseshoe abyssal plain, probably due to differences in the turbidite supply. Estimation of the mean pre-loading sediment thickness is complicated by uncertainties in seismic reflection correlation between the abyssal plains and the great variation in sediment thickness due to the basement relief and block faulting. The pre-loading sediment thickness was estimated to be about 2 km from available seismic reflection data from the Tagus and Eastern Horseshoe abyssal plains.

For the purposes of gravity calculation, crustal and mantle density were taken to be 2.8 g/cm$^3$ and 3.33 g/cm$^3$ respectively based on standard values (e.g., Dobrin & Savit, 1988; Allen & Allen, 1990), with a crustal thickness of 3.5 km based on seismic refraction data (Whitmarsh et al., 1990; Peirce & Barton, 1991). The pre-loading sediments were estimated to have a density of 2.4 g/cm$^3$ based on data presented in section 6.3.1. The post-loading (infill) sediment thickness was estimated to be 1.2 km in the Eastern Horseshoe abyssal plain and 1 km in the Tagus abyssal plain from the top of the pre-loading sediments defined by the distinct unconformity to the seafloor, with a density of 2.3 g/cm$^3$. And the infill density was assumed to be 2.3 g/cm$^3$ for the post loading sediments on the flanks of Gorringe Bank and 2.8
A continuous plate flexure model with an elastic plate thickness of 35 km (figure 6.29), which is similar to that expected for the crustal age (approximately 140 Ma) of the Gorringe region (Watts, 1994) assuming an age of approximately 125 Ma at the time of loading, provides a good overall fit to the observed FAA. In the Tagus abyssal plain the fit is excellent with the misfit to the NNW associated with the Madeira-Tore Rise which was not incorporated in the model. The fit in the Eastern Horseshoe abyssal plain is slightly poorer probably due to the assumption of a similar original sediment thickness to the Tagus abyssal plain when it is greater in the Eastern Horseshoe abyssal plain and also due to the effects of Coral Patch Ridge which were not incorporated into the model.

Although a reasonable fit to the observed FAA is obtained, there is a significant depth misfit of the depth converted basement and Middle Miocene (Mauffret et al., 1989) unconformity from seismic reconnection profile 4A-4B (figure 6.28) from IAM line 4 (Banda & Torne, 1995). However, the model quite accurately predicts the curvature. The use of a shallower load base would decrease the misfit in depth, but would decrease the curvature.

It is difficult to propose a mechanism by which Gorringe Bank loaded a continuous plate, when seismic reconnection (e.g., LePichon et al., 1970; Sartori et al., 1994) and sampling (e.g., Ryan et al., 1973; Auzende et al., 1978) indicate overthrusting (LePichon et al., 1970). The crust in the region is also highly faulted with associated high seismicity due to the proximity of the EU-AF plate boundary, so it is unlikely that the crust could behave as regionally continuous plate.

### 6.7.3 Gorringe Bank Broken Plate Flexure Models

The most common model in compressional settings is a broken plate model. For example, deep sea trench systems (Watts & Talwani, 1974) and foreland basins/thrust fold loading (Karner & Watts, 1983). In these models the plate break is located at the trench axis or inland of the thrust/fold loads.

On the basis of evidence for formation of Gorringe Bank by overthrusting and evidence for flexure in the Tagus abyssal plain, a two dimensional broken plate flexure model (figure 6.30) was applied to profile CG-FG-GG from line 4 (figure 6.25) across the Tagus abyssal plain, Gorringe Bank and Eastern Horseshoe abyssal plain. A fraction of Gorringe Bank (African plate) is modelled to be supported by the Eurasian plate, causing flexure in the Tagus abyssal plain. The position of the plate break divides material to the north of the break which loads the Eurasian plate and material to the south, assumed to be maintained by a regional stress
field associated with Eurasia-Africa convergence. The northern half model was modelled as a broken plate flexure model and once a good fit had been obtained to the observed FAA from the Tagus abyssal plain, the southern half model was forward modelled.

The same density and primary crustal structure were used as for the continuous plate model (figure 6.29), with crustal and mantle densities of 2.8 g/cm$^3$ and 3.33 g/cm$^3$ respectively, a crustal thickness of 3.5 km and a load base of 5.6 km, estimated from equation (6.5) (Stein & Stein, 1992). The pre-loading sediment thickness was estimated to be about 2 km for the Tagus and Eastern Horseshoe abyssal plains with a density of 2.4 g/cm$^3$. The post-loading (infill) sediment thickness was taken to be 1.2 km for the Eastern Horseshoe abyssal plain and 1 km for the Tagus abyssal plain with a density of 2.3 g/cm$^3$ and an infill density of 2.8 g/cm$^3$ beneath Gorringe Bank.

The model is sensitive to variation of the modelled Eurasian plate elastic thickness and to the position of the plate break. This plate break position defines the amount of load that has been added to the Eurasian plate by the convergence. The position of the plate break is a measure of the amount of shortening/overthrusting that has occurred, but does not define the true overthrust distance as the load is sheet thrusted and may have some lateral strength.

Several models were constructed to access the stability of models to variations in elastic thickness (figure 6.31) and break position (figure 6.32). Elastic thickness variation has a great affect on the calculated FAA. A low elastic thickness (e.g. 15 km) gives too short-wavelength and a too large amplitude deflections compared to the observed FAA, whereas a high elastic thickness (e.g. 55 km) gives too long-wavelength and too small amplitude deflections. The best fit is obtained with an elastic thickness of approximately 35 km since it explains both the wavelength and amplitudes of the observed. The position of the plate break has a small, yet significant, effect on the deflections and amplitudes in the calculated FAA for small variations of up to 20 km. The deflection caused by the whole of Gorringe Bank loading the Eurasian plate is too great to fit the observed FAA, indicating that Gorringe Bank must be partially supported by a regional stress field.

The southern half model was forward modelled using a similar density structure and crustal thickness as the northern half model. However the basement depth was constrained by seismic reflection data and the pre-loading sediment thickness was estimated to be a maximum of about 3.4 km in the eastern Horseshoe Basin with a thinner post-loading sedimentation of about 0.6 km. The olistostrome, although younger than the formation of Gorringe Bank, was assumed to have a higher density
than the recent sedimentation and was incorporated into the pre-loading sediment body. The Moho was modelled to keep a near constant crustal thickness beneath the Eastern Horseshoe abyssal plain as was shown in forward models along line 4 (figures 6.27a and b). Beneath Gorringe Bank the Moho was modelled to align with the base of the crust from the northern half model.

The models show that flexure can be applied to the Gorringe Bank region, explaining the wedge-shaped infill sediments of the southern Tagus Basin and the FAA low to the north. The best fitting model to the FAA, is a broken plate with an elastic thickness of 35 km and a plate break directly beneath the topographic peak. A $T_e$ of 35 km is consistent with a model in which Gorringe Bank loaded approximately 130-135 Ma oceanic lithosphere (e.g. figure 6.24).

The depth converted Middle Miocene (Mauffret et al., 1989) unconformity and basement from seismic reflection profile 4A-4B (figure 6.28) from IAM line 4 (Banda & Torne, 1995), show good correlation with the modelled depth and deflection (figure 6.30). The basement shows a greater misfit to the SSE, which may be due to the block structure or inaccurate depth conversion, in an area where higher velocity debris flow material from the erosion of the high density Gorringe Bank basement rocks is expected.

A plate break directly beneath the topographic peak suggests overthrusting of approximately 50 km, which assuming an age of about 20 Ma for the unconformity in the moat, suggests a mean convergence rate of 2.5 mm/yr, but does not take sheet thrusting into account which would imply a higher rate.
Figure 6.21: Calculated 2D free-air isostatic models from lines A and C across the Madeira-Tore Rise (figure 6.20). Solid lines = calculated isostatic anomalies for various $T_e$. Dash lines = observed FAA. WHAP = Western Horseshoe abyssal plain. TAP = Tagus abyssal plain.
Figure 6.22: Calculated 3D free-air isostatic models for lines A and B across the Madeira-Tore Rise (Figure 6.20). Solid lines = calculated isostatic anomalies for various $T_e$. Dash line = observed FAA. WHAP = Western Horseshoe abyssal plain. HIR = Hirondelle seamount.

Line A: WNW ESE

Line B: WNW ESE

- $T_e = 15$ km

Te = 15 km

Observed
Figure 6.23: Calculated 3D free-air isostatic models for lines C and D across the Madeira-Tore Rise (figure 6.20). Solid lines = calculated isostatic anomalies for various $T_e$. Dash line = observed FAA. TAP = Tagus abyssal plain.
Figure 6.24: Plot of $T_e$ against age for the oceanic lithosphere at the time of loading (Watts, 1994). The solid and dash lines show the 300, 450 and 600°C isotherms based on a cooling plate model.
Figure 6.25: Location of 2D gravity models for the Gorringe region. Circles mark the end points of model profiles. Red = Inverse models (figure 6.26); Green = Forward models (figure 6.27); Blue = Broken plate flexure model (figure 6.30); Black = Location of plate break for 0 km for broken plate flexure model and central point for the continuous plate flexure model (figure 6.29) whose endpoints exceed the bounds of the figure from the ends of line 4.
Figure 6.26: Inverse gravity models of the SCBA. (a) Profile AG-BG from line 2 over Coral Patch seamount. (b) Profile EG-HG from line 4 over Gorringe Bank. (c) Profile HG-IG from line 4 over Coral Patch Ridge. Density contrast = 0.68 g/cm$^3$. Lower figures show the crustal structure: Grey = mantle, dash line = basement, solid line = sea floor. Middle figures show the misfit between the modelled and observed SCBA. Upper figures show the observed and calculated SCBA. Blue dash line = assumed regional background field.
Figure 6.27: Forward gravity models for profile DG-IG (figure 6.25) from line 4 across Coral Patch Ridge, Gorringe Bank and Tagus and Eastern Horseshoe abyssal plains. Modelled against SCBA with sediment density = 2.3 g/cm$^3$ (a) Bouguer density = 2.65 g/cm$^3$ (b) Bouguer density = 2.8 g/cm$^3$. Inverted triangles mark the intersections of seismic reflection profiles from CD64 which provide sediment thickness constraint. Diamond ended line shows the projected location of seismic profile V-W (figure 5.10).
E-Eocene to L-Oligocene Unconformity

Tagus Abyssal Plain

Undeformed Infill (T-I)

Tilt Towards Bank Increasing With Depth

Folded and Faulted Sediments (T-III to T-VI)

Debris Flow (T-II)

Flexed Block Faulted Basement

Gorringe Bank Basement Outcrop

Baseline: 0 mGal

TWTT (s)

0.0 5.0 10.0

KM

4A 4B
Figure 6.29: Continuous plate flexure model for line 4 centred on Gorringe Bank (GB). Elastic thickness \( (T_e) = 35 \) km. M-T Rise = Madeira-Tore Rise, TAP = Tagus abyssal plains, EHAP = Eastern Horseshoe abyssal plain and CPR = Coral Patch Ridge. Densities in g/cm\(^3\). Upper figure shows the observed and calculated Free-air anomalies in mGal.
Figure 6.30: Broken plate model for Gorringe Bank for profile CG-FG-GG from line 4 (figure 6.25). Lower figure shows the structure with the break position, directly beneath the crest of Gorringe Bank, marked by a white dash line. Mantle shown in red, oceanic crust in blue, pre-loading sediments in green and post-loading sediments in yellow. Elastic thickness ($T_e$) = 35 km. Densities in g/cm$^3$. Upper figure shows the observed (from 5 minute gridded data), projected and calculated Free-air anomalies, where the projected values are true observed values from within 5km of the profile. EHAP = Eastern Horseshoe abyssal plain.
Figure 6.31: Effect of varying the elastic thickness ($T_e$) in the broken plate flexure model for Gorringe Bank along line 4 (figure 6.25 for location). Break position directly beneath the crest of Gorringe Bank.
Figure 6.32: Effect of varying the break position (or amount of overthrust) in the broken plate flexure model for Gorringe Bank along line 4 (figure 6.25 for location), with a fixed elastic thickness of 35 km. Break position distances are positive south from the crest of Gorringe Bank.
6.8 Magnetic Anomalies

Magnetic anomalies may provide information as to the age and crustal structure of the oceanic lithosphere. A magnetic anomaly map (figures 6.33 and 6.34) of the Gorringe Bank region shows a rough zone to the west and a subdued zone to the east.
Figure 6.33: Magnetic anomalies of the Gorringe Bank region. White lines show the location of interpretation profiles. Dash line shows the track of CD82. Red lines show magnetic forward model profile locations. Yellow dash lines show the J anomaly (Verhoef et al., 1986) and chron M-O (Klitgord & Schouten, 1986). Orange dash line mark the approximate rough and subdued magnetic zone boundary. MTR = Madeira-Tore Rise. GB = Gorringe Bank. HIR = Hirondelle, AMP = Ampere and CP = Coral Patch seamounts. Contour interval 100 nT.
Figure 6.34: Magnetic anomaly and bathymetry lines 1 and 2 from NW to SE across the Gorringe region. Dash line marks the division between the rough and subdued magnetic zones. Abyssal plains: WHAP=Western Horseshoe; SAP=Seine. Seamounts: JOSEPHINE=Josephine; AMP=Ampere; CP=Coral Patch; HIR=Hirondelle. Other: M-T RISE=Madeira-Tore Rise. See figure 6.33 for location.

The rough zone shows high amplitude anomalies in excess of 800 nT associated with volcanic rocks and oceanic crustal rocks of the Madeira-Tore Rise and Hirondelle seamount and the NNE-SSW trending chron M-0 (located in the region by Klitgord & Schouten (1986)) and J anomaly (located in the region by Verhoef et al. (1986)). The J anomaly (Pitman & Talwani, 1972) is related to formation of Madeira-Tore Rise and has an age of between 118 Ma (M-0) and 122 Ma (M-1) (Rabinowitz et al., 1979). Hirondelle seamount shows quite large dipolar anomalies.
Figure 6.34: Continued magnetic anomaly and bathymetry lines 3 to 5 from NW to SE across the Gorringe region. Dash line marks the division between the rough and subdued magnetic zones. Dotted line marks the J anomaly. Abyssal plains: EHAP=Eastern Horseshoe; TAP=Tagus. Seamounts: GB=Gorringe Bank. Other: M-T RISE=Madeira-Tore Rise; CPR=Coral Patch Ridge; AFM=African margin. See figure 6.33 for location.
of -250 to 350 nT, perhaps indicating a volcanic origin similar to the Madeira-Tore Rise and Josephine seamount. A magnetic high, in excess of 400 nT, to the east of Hirondelle seamount is related to the antithetic strike-slip fault observed on seismic reflection data (figure 5.4), Gloria data (figures 5.1 and 5.2) and seismicity in this region (figure 5.13).

The subdued zone to the east is associated with the Tagus, Horseshoe and Seine abyssal plains and subdued anomalies over Gorringe Bank and Coral Patch seamount. Here the anomalies decrease in amplitude to a background of about -150 to 150 nT, but contain a number of low amplitude and long wavelength anomalies. The origin of this zone is unclear, but may correspond to the Middle Cretaceous uniform polarity interval (Vogt et al., 1970; Whitmarsh & Laughton, 1974) or non-magnetic stretched continental crust.

Ampere seamount is anomalous in the subdued zone in that it shows a large dipolar anomaly with amplitudes of up to 400 nT to the north and -200 nT on the south with the highest normal amplitudes corresponding to the eastern summit which is thought to be the paleovolcanic eruption centre (Marova & Yevsyukov, 1988) and high amplitude reversed anomalies associated with the western summit (Matveyenkov et al., 1994).

Gorringe Bank (line 4, figure 6.34) shows magnetic anomalies with a weak NNE-SSW trend and a peak of about 200 nT centred under Ormonde seamount, which is large when compared with the regional field and with negative anomalies of up to -100 nT beneath Gettysburg seamount. Coral Patch seamount shows magnetic anomalies of about -75 nT to 75 nT with a very weak trend roughly parallel to the orientation of the seamount. These low amplitude anomalies may be the result of deformation, Moho elevation and possibly crustal heating.

The abyssal plains generally show low amplitude anomalies with low amplitude positive anomalies of about 175 nT over most regions of the Horseshoe abyssal plain. A magnetic anomaly low of about -175 nT between the Eastern and Western Horseshoe abyssal plain continues the trend of the fault associated high amplitude anomalies to the east of Hirondelle seamount.
6.9 Magnetic Models

Two profiles were forward modelled to investigate the magnetic material properties across a region which exhibits a highly variable magnetic character. The first model AM-BM (figure 6.35) along line 1 (figure 6.33) was chosen to investigate the structure of the western Gorringe Bank region which has high amplitude anomalies, including the large dipolar anomaly associated with Ampere seamount.

![Magnetic forward model AM-BM along line 1. Shaded region shows a susceptibility (dimensionless SI units) of greater than 0.15. Dash line marks the division between the rough and subdued magnetic zones. Seamounts: JOS=Josephine; AMP=Ampere. Abyssal plains: WHAP=Western Horseshoe; SAP=Seine. See figure 6.33 for location.](image)

The second model CM-DM-EM (figure 6.36) was constructed across Gorringe Bank and Coral Patch Ridge close to line 4 to investigate their magnetic structure and why they exhibit such low magnetisation despite their dimensions. The profile was taken directly from the track of CD82 to provide greater accuracy, than data derived from the combined gridded data, in a region where the magnetic anomaly...
lies are only slightly greater than one standard deviation in the crossover errors (figure 3.5) (Wessel & Watts, 1988).

The method of Talwani & Heirtzler (1964) was used to compute the magnetic anomalies associated with two dimensional irregular polygons. The magnetic anomalies are modelled to be induced and originate from the crustal basement, constrained by seismic reflection data. The location of magnetic anomalies directly above basement topographic features suggests that remanent magnetisation is only a small factor and was not incorporated into the model.

Figure 6.36: Magnetic forward model CM-DM-EM along CD82 magnetic profile. Shaded region shows a susceptibility (dimensionless SI units) of greater than 0.15. Seamounts: GB=Gorringe Bank; CPR=Coral Patch Ridge. Abyssal plains: TAP=Tagus; EHAP=Eastern Horseshoe and SAP=Seine. See figure 6.33 for location.

Profile AM-BM (figure 6.35) was initially modelled as a single body with the induced magnetic anomaly calculated for various constant susceptibilities (All susceptibilities in SI units (dimensionless)) to gain a range of values appropriate for the region. This shows that values are likely to about 0.12 for Josephine seamount.
and about 0.06 for Ampere seamount. The forward model was created by sub-
division the basement body into many sub bodies with varying susceptibility. A
small susceptibility change across a body boundary results in a large change in the
amplitude of the calculated anomaly, meaning that the values are well constrained,
but that exact fitting of calculated to the observed anomalies is complicated.

The resulting model (figure 6.35) shows a reasonable fit of the calculated anom-
aliies with the most highly magnetic rocks beneath Josephine seamount and extend-
ing under the western Horseshoe abyssal plain. Although the body and susceptibility
pattern is fairly complex the susceptibilities for Ampere seamount are about an or-
der of magnitude smaller than Josephine seamount and the Western Horseshoe
abyssal plain.

Susceptibility is not indicative of rock type, as it is mainly influenced by abun-
dance of accessory magnetic minerals (Clark & Emerson, 1991), but a value of 0.02
to 0.06 for Ampere seamount is within the range expected for alkaline basalts which
have been sampled (e.g., Purdy, 1974; Marova & Yevsyukov, 1988; Matveyenkov
et al., 1994). Josephine seamount gave values of about 0.16 to 0.18 which also
falls within the range of expected values for oceanic crustal rocks such as Basalt,
Dolerite and Peridotite (Clark & Emerson, 1991). The slight skew of the calculated
magnetic anomalies over Josephine seamount (figure 6.35) may indicate the pres-
ence of remanent magnetisation which is often higher in rapidly chilled fine grained
basaltic rocks (Clark & Emerson, 1991), as expected for the volcanic formation of
Josephine seamount.

Profile CM-DM-EM (figure 6.36) was modelled in a similar manner to profile
AM-BM. A good fit was obtained to most of the features with similar values of
susceptibility of about 0.01 to 0.16, but about 0.02 for the majority of the profile,
which agrees with calculated susceptibilities of 0.006 to 0.02 for basalts sampled
from Ormonde seamount and overlaps with values of 0.01 to 0.775 for serpentinites
from Gettysburg seamount (Gorodnitskiy et al., 1988). The generally similar
susceptibilities along most of the profile are significantly lower than those obtained
for profile AM-BM perhaps indicating a similar crustal nature along profile CM-
DM-EM, but one which is different to that beneath profile AM-BM. A susceptibility
of about 0.02 is not indicative of a particular rock type, but is within the range
expected for typical oceanic crustal material (Clark & Emerson, 1991).
Chapter 7

Conclusions

7.1 Tectonic Evolution

7.1.1 Evolution of the Azores-Gibraltar Region

The structure of the Gorringe Bank region is a consequence of complex interactions of the Eurasian, Iberian and African plates during evolution of the North Atlantic Ocean. Rifting between North America and Africa initiated at approximately 175 Ma (Klitgord & Schouten, 1986) and at approximately 165 Ma (Sclater et al., 1977) between North America and Eurasia. During early opening in the Jurassic, uniformly magnetised oceanic crust produced the magnetic smooth zone observed east of approximately 12°W in the Gorringe Bank region. This smooth zone precludes recognition of the ocean-continent transition zone and prevented accurate crustal dating prior to the first accurately identified magnetic chron M-25 to the west of the Madeira-Tore Rise.

Formation of the Madeira-Tore Rise

From about M-4 to M-0 (126-119 Ma) (Tucholke & Ludwig, 1982) an increased magma flux, possibly from a mantle plume, moved south along the Mid Atlantic Ridge resulting in formation of the Madeira-Tore Rise and J magnetic anomaly. Isostatic gravity models show that oceanic crust in this region has an elastic thickness of approximately 15 km which translates to an age of up to approximately 20 Ma for the age of the crust at the time of Madeira-Tore Rise formation, based on Watts (1994). Although the results of Peirce & Barton (1991) suggest that formation occurred at the Mid Atlantic Ridge, a T_e of 15 km indicates the addition of material as the Madeira-Tore Rise moved away from the Mid Atlantic Ridge. Miocene volcanism on Ampere seamount (Matveyenkov et al., 1994) and 60 Ma volcanic material recovered (Feraud et al., 1982, 1986) from Gorringe Bank indi-
cate that volcanism has been active intermittently through time, so it is likely that material may have been added to the Madeira-Tore Rise since initial formation.

**Evolution of the Azores-Gibraltar Plate Boundary**

As the Eastern Atlantic evolved, differential motion between the Eurasian, Iberian and African plates created many lines of weakness which subsequently influenced the creation of structures under a different tectonic regime. Plate boundaries existed between the Iberian and African plates from the Late Jurassic (M-25, 156 Ma) until approximately M-34 (84 Ma) when left lateral motion between the African and Eurasian plates ceased (Srivastava *et al.* , 1990). Right lateral motion between Iberia and Africa commenced at about chron 19 (44 Ma) and became the only plate boundary between Eurasia and Africa by chron 6 (20 Ma) (Srivastava *et al.* , 1990). Motion on the Azores-Gibraltar plate boundary has been fairly stable since (Argus *et al.* , 1989). Seismicity (Buforn *et al.* , 1988) indicates that there is active extension on the Terceira Ridge, right lateral slip along the Gloria Fault and compression and crustal shortening of about 4 mm/yr (DeMets *et al.* , 1990) in the Gorringe Bank region.

**7.1.2 Evolution of the Gorringe Bank Region**

Following rifting of the Eurasian and North American plates, sediments of initially Upper Jurassic age were deposited upon block faulted oceanic crustal basement. Deposition continued throughout the region (figure 7.1), with sediment supply primarily from the Iberian margin, which bounds the region to the east. A boundary to the west was created by the Madeira-Tore Rise, following its primary formation in the Lower Cretaceous (Tucholke & Ludwig, 1982). A substantial sediment thicknesses had accumulated in the region by the Early Eocene, as imaged by seismic reflection profiles in the Tagus and Horseshoe Basins (figures 4.2 and 4.4). Sediments of the Tagus and Eastern Horseshoe Basins were deposited upon a mostly inactive block faulted basement. However, sediments of the Western Horseshoe Basin were deposited under a greater influence of active block faulting throughout deposition (figure 4.6) and large scale basin bounding normal faults (figure 4.5) confined the sediments at the base of Hirondelle seamount. Plate motion during this time was left lateral between Iberia and Africa (figure 2.16) (Srivastava *et al.* , 1990). However, little evidence of deformation associated with this motion is observed, but minor folding and block faulting with related onlap is observed on seismic reflection profiles in the Tagus Basin (figures 4.2 and 6.28).
When compressive plate motion commenced at about 20 Ma (Srivastava et al., 1990), pre-existing crustal fractures were exploited in the formation of the Gorringe Bank region, including the uplift of Gorringe Bank and Coral Patch Ridge and the compressional deformation of the sediments.

All sedimentary units of the Western and Eastern Horseshoe Basins older than the Early Eocene to Late Oligocene unconformity (Hayes et al., 1972) show features of deformation related to the compression which initiated at about 20 Ma (Srivastava et al., 1990). The units below the olistostrome unit EH-II, which has an age of Middle Miocene (Auzende et al., 1981), in the Eastern Horseshoe Basin show folding which pre-dates olistostrome deposition. This deformation is associated with the uplift of the Gorringe Bank and Coral Patch Ridge which had relief prior to olistostrome deposition. Deformation continued with folding of the sediments of the Western and Eastern Horseshoe Basins, including the folding of the olistostrome unit EH-II. Sediments which post date the Early Eocene - Late Oligocene unconformity (Hayes et al., 1972) (Miocene in the Tagus Basin Mauffret et al. (1989)) show minimal deformation in the basins, indicating a change in the style of deformation following the uplift of Gorringe Bank, with recent deformation concentrated on bathymetric highs (figure 7.1) and primarily to the south of the Horseshoe Basin (section 7.1.3).

Free-air gravity anomalies for the Gorringe Bank region reveal an increase then a decrease in amplitude from west to east in comparison with the bathymetric scale, related to a change in compensation across the region.

Isostatic models show that bathymetric features of the Gorringe Bank region are uncompensated. The Madeira-Tore Rise and Iberian margin may be compensated with increased elastic thickness, however Gorringe Bank, Coral Patch seamount and Coral Patch Ridge can not be compensated even by an elastic thickness in excess of 50 km. These uncompensated features must be supported in part by a regional stress field.

**Formation of Gorringe Bank**

Gorringe Bank was formed as a result of northward sheet thrusting of the African plate upon the Eurasian plate (LePichon et al., 1970). Gabbros and serpentinites, which represent upper and lower crustal rocks, have been recovered from Gorringe Bank (e.g., Auzende et al., 1978, 1982) and sediments show an asymmetric distribution on seismic reflection profiles, with thick layered sediments on the southeastern flank and basement outcrop on the northwestern flank. GLORIA images of the northern flank show small terraced cliffs of fractured basement rocks with
CHAPTER 7. CONCLUSIONS

scattered fallen blocks, scree and minor pelagic sedimentation as observed during submersible investigations (Auzende et al., 1982; LaGabrielle & Auzende, 1982; Auzende et al., 1984).

Overthrusting partially loaded Gorringe Bank upon the Eurasian plate causing flexure, which is associated with a linear free-air gravity anomaly low, parallel to the northwest flank of Gorringe Bank in the Tagus abyssal plain. Interpretation of seismic reflection profile IAM line 4 (figure 6.28) shows the dip of the block faulted basement and pre-loading sediments of the Tagus Basin towards and possibly under thrusting Gorringe Bank. Post loading or infill sediments which have an age of about Middle Miocene to Pleistocene (Mauffret et al., 1989) show fanning reflections in the Tagus abyssal plain (figure 6.28). This indicates that subsidence and flexure of Eurasian plate have not ceased, with a planar seafloor resulting from a sediment supply greater than the subsidence.

Broken plate flexure gravity models of partial loading of Gorringe Bank upon the Eurasian plate, give a well constrained elastic thickness of 35 km for crust beneath the Tagus abyssal plain. This is expected for crust of an age of about 130-135 Ma at the time of loading (Watts, 1994), which agrees with the crustal age of 140 Ma (Prichard & Mitchell, 1979; Feraud et al., 1982, 1986) for Gorringe Bank and overthrusting in the Middle Miocene at approximately 10 Ma, associated with the age of the unconformity (Mauffret et al., 1989) between the pre and post loading sediments.

A best fit to the observed FAA and seismic reflection profile 4A-4B (figure 6.28) from IAM line 4 (Banda & Torne, 1995) is obtained for loading of the northern part of Gorringe Bank, with a plate break directly beneath the bathymetric peak. This distance is related to, but does not represent the true overthrust distance, due to possible sheet thrusting and gives a minimum estimate of shortening of about 50 km. The model is quite sensitive to overthrust distance variations of up to 20 km, which cause a greater misfit with the observed FAA. Models of complete support of Gorringe Bank by the Eurasian plate cannot fit the observed gravity anomalies and indicate that Gorringe Bank is partially supported by a regional stress field.

Major deformation associated with the formation of Gorringe Bank ceased, probably due to the regional fractured nature of the crust which was easier to deform elsewhere, rather than to continue subduction of the Eurasian plate beneath the African plate.
Formation of Coral Patch Seamount and Ridge

Major deformation moved to the south of the Horseshoe abyssal plain, concentrated on the northern flank of Coral Patch seamount and the southern edge of the Eastern Horseshoe abyssal plain. Compression resulted in large scale thrust faulting and folding of the sediments and basement rocks on Coral Patch Ridge and the formation of large folded ridges on Coral Patch seamount.

Gravity forward models of Coral Patch Ridge show an elevated Moho beneath bathymetric highs, as tentatively interpreted from seismic reflection profiles (Sartori et al., 1994), which indicates a combination of thrusting and crustal warping. Combined thrusting and crustal warping of a similar nature, is observed in the central Indian Ocean in the region of the Afranasy-Nikitin seamounts, where compression of similar magnitude to the Gorringe Bank region has prevailed since the Late Miocene (Karner et al., 1993). High angle reverse faulting and large scale sediment folding (Weissel et al., 1980), similar in style to that observed on seismic reflection profiles from Coral Patch Ridge, are associated with large crustal folds (Weissel et al., 1980; McAdoo & Sandwell, 1985; Karner et al., 1993) and gravity anomalies of 15-80 mGal (McAdoo & Sandwell, 1985; Karner et al., 1993). Crustal buckling in the Indian Ocean was focussed by pre-existing crustal folds caused by fold trough filling sediments (Zuber, 1987; Karner et al., 1993) and the load of the Afranasy-Nikitin seamounts (Karner et al., 1993). Gorringe Bank and the great thickness of sediments in the Horseshoe abyssal plain may have similarly focussed deformation associated with Coral Patch Ridge.

Magnetic Structure of the Gorringe Bank Region

The Gorringe Bank region has high amplitude NNE-SSW trending magnetic anomalies in excess of 800 nT associated with the J anomaly and chron M-0 in the region of the Madeira-Tore Rise. Ampere and Hirondelle seamounts shows large dipolar anomalies with amplitudes 400 nT to -200 nT and 350 to -250 nT respectively, perhaps indicating a similar volcanic origin. An ENE-WSE trending 300-450 nT anomaly to the southeast of Hirondelle seamount is related to the strike slip fault between the Gettysburg and Hirondelle seamounts.

East of the Madeira-Tore Rise, Ampere and Hirondelle seamounts, anomalies are low amplitude at about ± 150 nT due to the magnetic smooth zone. Gorringe Bank shows magnetic anomalies with a weak NNE-SSW trend and a peak of about 200 nT centred under Ormonde seamount and negative anomalies of up to -100 nT beneath Gettysburg seamount. Coral Patch seamount shows magnetic anomalies
of about ± 75 nT with a very weak roughly seamount parallel trend. The abyssal plains generally show low amplitude anomalies with about 175 nT over most of the Horseshoe abyssal plain.

**Magnetic Forward Models of the Gorringe Bank Region**

Magnetic forward model investigations of the magnetic structure of the western Gorringe region show a susceptibility of about 0.16 to 0.18 for Josephine seamount, similar to that expected for oceanic crustal rocks (Clark & Emerson, 1991). Ampere seamount shows susceptibilities of about 0.02 to 0.06, an order of magnitude smaller than Josephine seamount and the Western Horseshoe abyssal plain, which is within the range expected for the sampled alkaline basalts (e.g., Purdy, 1974; Marova & Yevsyukov, 1988; Matveyenkov et al., 1994).

The eastern Gorringe Bank region across the Tagus abyssal plain, Gorringe Bank and Coral Patch Ridge gave susceptibilities of about 0.02 for the majority of the profile, which agrees with calculated susceptibilities for sampled basalts and serpentinites (Gorodnitskiy et al., 1988) and is typical of oceanic crustal rocks (Clark & Emerson, 1991). The susceptibilities are significantly lower than those to the west perhaps indicating a different crustal nature.

### 7.1.3 Recent Deformation in the Gorringe Bank Region

Recent deformation in the Gorringe Bank region is widespread as implied by the shallow and apparently diffuse seismicity (e.g., Grimison & Chen, 1986, 1988), also observed in the central Indian Ocean (Weissel et al., 1980). The trend of the seismicity can however, be tied to the seaward extension of continental crustal fractures (Moreira, 1985; Buforn et al., 1988; Moreira, 1991) and as with the seismicity, much of the surface deformation observed primarily on GLORIA and seismic reflection profiles, can be explained on the basis of a regional model.

Recent deformation although observed across the whole Gorringe region is concentrated on bathymetric highs, with minimal surface deformation in the abyssal plains. Folds and faults exhibit clockwise rotation from a trend of about N45°E at 15°W in the region of Josephine seamount to N70°E at 10°W in the region of Coral Patch Ridge. This rotation is in accordance with the variation of slip vector azimuths (e.g., Fukao, 1973; Grimison & Chen, 1988; Argus et al., 1989; Udias & Buforn, 1991) verging to an E-W trend towards the west.

Alignment of the direction of maximum principal compressive stress in a dextral strike-slip ellipse model near parallel to the plate motion vectors, shows that the majority of these recent compressional features can be explained by compression
in a regional dextral strike-slip zone. Extensional features observed on Josephine seamount (Matveyenkov et al., 1994) are near perpendicular to the direction of minimum principal compressive stress $\sigma_3$ and the strike-slip fault between Gettysburg and Hirondelle seamounts is probably an antithetic $R_2$ Reidel shear, which agrees with motion interpreted from the focal mechanism of (Buforn et al., 1988).

Many other focal mechanisms in the Gorringe Bank region also tie with the ellipse model, with the majority of strike-slip focal mechanisms (Grimison & Chen, 1986; Buforn et al., 1988; Grimison & Chen, 1988) aligned near to the $R_1$ Reidel shear and the strike of the thrust faulting mechanism of Grimison & Chen (1986) perpendicular to $\sigma_1$. However, a few focal mechanisms in the Eastern Horseshoe abyssal plain show that faulting is complex and variable in this area.

Magnetic anomalies from Gorringe Bank have two dominant orientations of 20-30° and 110-130° (Gorodnitskiy et al., 1988), which have been correlated with northeast trending faults and steep vertical fault scarps respectively (Gorodnitskiy et al., 1988). The 20-30° orientated faults are near perpendicular to $\sigma_1$, possibly indicating that they may represent thrusts or strike-slip faults and the 110-130° trending fault scarps probably represent NW-SE trending normal faults, similar to those observed on Josephine seamount (Matveyenkov et al., 1994).

### 7.1.4 Sedimentary Structure of the Gorringe Bank Region

Sediments in the Gorringe Bank region show a great thickness variation, with thinner drapes on seamounts and substantial thicknesses in the Tagus, Western Horseshoe, Eastern Horseshoe and Seine Basins. These basins are divided by bathymetric highs except for the Eastern and Western Horseshoe Basins which, although currently represented by a single basin, were once separated by a basement shallowing and deformation associated with NNE-SSW faulting between Gettysburg and Hirondelle seamounts.

Sediments of the Tagus Basin have a thickness of up to 2.9 km (basement depth 8.6 km) which fill a linear trough across the base of the northern flank of Gorringe Bank. Thickness decrease is rapid towards Gorringe Bank and gradual towards the north due to flexure of the Eurasian plate under partial loading of Gorringe Bank. To the west, upper units are draped over the flanks of Hirondelle seamount, with deeper units partially confined by boundary faults.

The Eastern Horseshoe Basin has a primary depocentre to the east with up to 5 km of sediment (basement depth 10 km) which thins gradually towards the west and east. To the north sediments continue onto the southern flank of Gorringe Bank and to the south are offset by and show substantial thinning beyond the
ENE-WSW striking thrust faults of Coral Patch Ridge. However, The active basin boundary faulting truncates sediments to the southwest.

The Western Horseshoe Basin is comprised of two near circular sub basins forming a "dumbbell" shaped basin. The eastern sub basin has up to 4 km of sediment (basement depth 8.9 km) and is stretched slightly along a NNE-SSW trend, related to faulting between Hirondelle and Gettysburg seamounts. The second sub basin has up to 2.8 km (basement depth 7.5 km) of sediment. The sediments thin onto faulted basement to the east, south and west. However, to the north and northeast, they are constrained by classic steeply dipping basin bounding normal faults (figure 7.1) which have been active throughout basin formation.

Sediment thickness in the Seine Basin is poorly constrained, however, sediments rapidly attain a thickness of at least 2 km to the south of Ampere seamount and Coral Patch Ridge. Sediments drape and thin onto block faulted basement to the southwest of Coral Patch seamount and to ENE onto Coral Patch Ridge.

**Seismic Facies of the Gorringe Bank Region**

Basin isosalation has resulted in diversity of reflection and sedimentary character and structure on seismic reflection profiles. Facies interpreted for individual basins have been tentatively correlated with the aid of the DSDP sites 120 and 135 (Ryan *et al.*, 1973; Hayes *et al.*, 1972) to give four main units and basement.

1. The youngest unit has an age from Pleistocene to Late Oligocene (Possibly Pleistocene to Middle Miocene in the Tagus Basin (Mauffret *et al.*, 1989)) and is characterised by medium-low amplitude, short wavelength, semi-continuous to continuous at depth parallel reflections. The unit forms a turbidite deposit, including a component of pelagic chalk ooze, with low angle basal onlap onto a regionally distinctive unconformity, often represented by a pair of continuous high amplitude reflections.

2. The second unit rests upon the major unconformity in the Tagus and Eastern Horseshoe Basins and is probably Middle Miocene in age. The unit is characterised by medium amplitude hummocky broken reflections with a high diffraction density. In the Tagus Basin the unit is restricted to the linear flexural moat to the northwest of Gorringe Bank, created by partial loading of Gorringe Bank upon the Eurasian plate and is interpreted as one or more debris flows derived from Gorringe Bank. In the Eastern Horseshoe Basin the unit represents an olistostrome derived from the Straits of Gibraltar (Bonnin
et al., 1975; Auzende et al., 1981) following formation of Gorringe Bank, as
the unit is absent on its uplifted flanks.

3. The next unit has a probable age of Early Eocene to Early Aptian and is
characterised by low amplitude, semi-continuous, parallel reflections which
are slightly deformed and show low angle basal onlap onto fold highs.

4. The oldest unit of the Gorringe Bank region has a tentative age of Upper
Jurassic to Early Aptian based on DSDP sites and estimates of crustal age.
However, if the deepest deposits are remnants of prerift sediments they may
predate the Jurassic. The unit fills basement troughs with onlap onto base-
ment, faults and fold highs and may in part represent a synrift deposit, with
growth observed on a seismic reflection profile from the Western Horseshoe
Basin.

5. The basement structure in the Gorringe Bank region is obscured by the great
sediment thicknesses, but generally characterised by low-medium amplitude,
broken hummocky reflections and a high diffraction density. Basement has
influenced sedimentation, with decreasing effect over time, in most regions.
Basement in the Western Horseshoe and Tagus basins shows a huge basement
relief in excess of 1.5 seconds TWTT, which has influenced sediment depo-
sition. In contrast the Eastern Horseshoe Basin only shows a relief of about
1 second TWTT which appears to have had a lower effect on sedimentation
than other basins. This may represent a different crustal nature, as suggested
by refraction models (Purdy, 1975).

7.2 Primary Conclusions

- Gorringe Bank, which was formed by overthrusting of the African plate upon
  the Eurasian plate, is supported in part by flexure of the Eurasian plate
  in the Tagus abyssal plain. The Eurasian plate has an elastic thickness of
  approximately 35 km which is in agreement with that expected for the age
  of the crust at the time of loading at approximately 10 Ma. This has created
  a free-air gravity low to the NNW of the Bank in the Tagus abyssal plain
  associated with infill sedimentation which is tilted slightly towards the Bank,
  with increased dip with depth.

- Coral Patch Ridge was formed by a combination of thrust faulting and whole
  crustal bucking, resulting from compression during the past 20 Ma and was
partially uplifted before deposition of the olistostrome in the Middle Miocene.

- The NNE-SSW fault between Gettysburg and Hirondelle seamounts is an antithetic strike-slip fault with a trend similar to the western end of the olistostrome and the paleo western boundary of the Eastern Horseshoe Basin.

- Recent deformation in the Gorringe Bank region is distributed over a wide region with extensional and compressional faulting and folding observed associated with all bathymetric highs and can be tied with a regional strike-slip strain ellipse model.

- The recent deformation trend rotates from approximately N45°E to N70°E from west to east across the Gorringe Bank region, near perpendicular to plate motion vectors whose orientation also changes from west to east.

- The Madeira-Tore Rise formed at the Mid Atlantic Ridge, but a modelled elastic thickness of approximately 15 km indicates that significant material was added as the Rise moved away from the Mid Atlantic Ridge.

### 7.3 Future Work

Although a greater understanding of the structure and formation of the Gorringe Bank region has been achieved from the integration of marine geophysical data. The most significant information lacking in this study is detailed sampling of the age and physical properties of sediments from the deep sedimentary basins. Adequate seismic reflection data, including new data from the IAM project (Banda & Torne, 1995), exists to accurately map the facies of the region. However, without knowledge of age for the various basins which are highly variable in character regional correlation is difficult. It would be useful to obtain further deep cores from the region to help in regional facies correlation.

The nature of the crust in the Gorringe Bank region and location of the Ocean Continent Transition are still under question. Acquisition of a reversed seismic refraction experiment utilising the latest D.O.B.S. technology, would accurately reveal the crustal structure beneath the Horseshoe abyssal plain.
Figure 7.1: Summary tectonic map of the Gorringe Bank region. Yellow zones show regions of recent compressional deformation. Heavy black lines show fault locations, with possible extension shown by dash lines. Triangles represent the location of thrust fronts. Ticks mark the location of basin bounding normal faults. Large red vectors show plate motion according to NUVEL-1 (DeMets et al., 1990). Small red vectors show fault displacements. Faults: GF = Gloria, TF = Tagus, AL=Alandroal, MJ=Messejana, LL=Loule, GQ=Gaudalquivir lineament, CA=Crevillente (Cadiz-Alicante) fault zone. Bathymetry contour interval 500m. Seismicity greater than Mb 4 from the ISC shown by red circles Focal mechanisms from Grimison & Chen (1986, 1988) and Buforn (1988).
References


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