THE CRUSTAL VELOCITY AND DENSITY STRUCTURE OF NORTHWEST EUROPE: A 3D MODEL AND ITS IMPLICATIONS FOR ISOSTASY

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Abstract

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A new seismic P-wave velocity model has been constructed for a region of northwest Europe encompassing Britain, Ireland and the surrounding marine sedimentary basins. This model has considerably higher resolution than previously published velocity models for the region and is unique in being quantitatively constrained. The velocity model provides a tool for future work in seismology, allowing crustal correction for teleseismic arrivals recorded in Britain and Ireland, refined local earthquake location and provides a starting model for local and regional seismic tomography. The velocity model has been developed in conjunction with a crustal density model. Together the two models provide information on the nature of the crust and have been interpreted in a geological context. The new models indicate that: the mean crustal velocity and density are related to the near-surface geology; the near-surface velocity is well correlated with the near-surface geology; there may be a regional trend from high velocity crust near the northwest continental margin to lower velocity crust under continental Europe; there may be a weak trend of increasing density with increasing crustal thickness; and that the crustal structure beneath the sedimentary basins varies from basin to basin. In addition to the geological interpretation, the density model has been used to investigate isostasy in the region using a newly developed method for estimating the elastic thickness of the lithosphere. This approach uses the relationship between the predicted height of topography for local isostatic equilibrium and the true topography to estimate the degree of compensation, which is in turn related to the elastic thickness. However, the results of this investigation are inconsistent across the different wavelengths tested and so the elastic thickness is resolved with less constraint than the range in published values for the region.
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Chapter 1

Introduction

Accurate models of the crust are extremely useful tools to many branches of Earth Science. Crustal velocity models in particular are used extensively within the field of seismology, where crustal corrections are made in tomographic studies, earthquake location and nuclear test monitoring. As a result many crustal velocity models have been constructed both for the whole globe and for restricted areas. The main objective of this study is to produce a new model of the crustal velocity structure of northwest Europe, owing to the relatively poor resolution of those that exist at present. The large quantity of wide-angle reflection and refraction seismic data that is available allows for a significant improvement in model resolution and accuracy. Accompanying the velocity model an associated crustal density model is created, allowing refinement of the velocity model and providing further information on the crustal structure. The secondary objective of this project is to use the new density model to investigate isostasy in the region; although a large body of work has been produced relating to isostatic compensation, there are still significant disagreements regarding the strength of the lithosphere, i.e. the importance of flexure compared to local compensation. This work returns to this problem with the benefit of the new high resolution crustal model.

1.1 Northwest Europe

The region of northwest Europe investigated in this project is shown in Figure 1.1. As an introduction to the scientific and commercial interest in the area the main topographic
Figure 1.1: The topography and bathymetric features of northwest Europe referred to in this work. Key: red line = boundary of the final velocity model; RB = Rosemary Bank Seamount; AD = Anton Dohrn Seamount; MV = Midland Valley; SU = Southern Uplands. The image is plotted on a transverse Mercator projection centred on 3.4° W, 57.15° N. The plot ranges from 1300 km west of the projection centre to 1300 km east and from 1400 km south to 1400 km north. Bathymetric and topographic data from Smith and Sandwell (1997) are contoured at 1 km intervals. This figure and many others presented in this work have been produced using GMT (Wessel and Smith, 1998).

and geological features are discussed in this section.

1.1.1 Basement architecture

The dominant fabric in the basement is inherited from the Caledonian Orogeny, the compressional event associated with the closing of the Iapetus and Tornquist Oceans and the combining of Laurentia with the continental fragments of Baltica and Avalonia (430–390 Ma). The southern extremities of Britain and Ireland have undergone subsequent deformation during the Variscan Orogeny while much of Scandinavia has remained undeformed since the Pre-Cambrian; however, the basement of the majority of Britain, Ireland and the areas occupied by the present day sedimentary basins preserve the tectonic fabric generated in the Caledonian Orogen (Fig. 1.2). The broad chronology and structure
of the Caledonian Orogeny is well understood and has been described in numerous publications (e.g. Woodcock and Strachan, 2000). The main phase of the orogeny occurred during the Silurian with oblique convergence of Avalonia and Baltica with Laurentia as the Iapetus Ocean and Tornquist Sea closed (Fig. 1.3). The deformation associated with the convergence continued into the Devonian with major sinistral strike-slip faulting along the Highland Boundary and Great Glen Faults in Scotland and Ireland, \( \sim 425-420 \) Ma (Woodcock and Strachan, 2000). The collision of two separate continental fragments with Laurentia results in a three branched orogenic zone: the Appalachian Belt and North Atlantic Caledonian Belt, generated by the closure of the Iapetus Ocean, passes through western North America, Ireland, Britain and Greenland/Norway (Fig. 1.2) and the Tornquist belt, formed by the closure of the Tornquist Sea, runs through Denmark and Eastern Europe. The Appalachian/North Atlantic arms produced a regional northeast-southwest structural grain through Ireland, Britain, the northern North Sea and Norway (Fig. 1.2).

Further collisional events associated with regrouping of continental fragments rifted from Gondwana caused deformation in the south of Britain, Ireland and throughout mainland Europe (Fig. 1.2) and resulted in the formation of the supercontinent Pangaea. The
Figure 1.3: Global palaeocontinental reconstruction for the closure of the Iapetus and Rheic Oceans (redrawn from Woodcock and Strachan, 2000).
Armorican fragment collided with Avalonia resulting in the the Acadian Orogen, \(~400\) Ma (Woodcock and Strachan, 2000), with the join between the two continental fragments marked by the Rheinohercynian Suture (Fig. 1.2). Iberia collided shortly after (~390–370 Ma) and the collisional events culminated with the impact of the main Gondwanan continent between \(~370\) and \(~290\) Ma (Woodcock and Strachan, 2000) forming Pangaea (Fig. 1.3). These individual collisional events are grouped together as the Variscan Orogeny. The Variscan Belt of southern and western Europe (Fig. 1.2) forms part of a broad orogenic zone extending from the Gulf of Mexico, through north Africa, southern Europe and into eastern Europe.

1.1.2 Sedimentary cover

The North Sea

Early rift basins were formed in the North Sea region during the Triassic, with the main phase of rifting occurring in the Late Jurassic to Early Cretaceous (Bartholomew and Peters, 1993) creating the Central Graben to the south and the Viking Graben to the north, with the Moray Firth rift branching out to the west (Fig. 1.4). The Central Graben cuts across the Iapetus Suture at a high angle and all three rifts show little influence of the regional northeast-southwest Caledonian trend. The late Cretaceous saw regional post-rift thermal subsidence, which out-stripped deposition, creating the North Sea topographic depression. The Central Graben now contains approximately 6 km of sediment, with around 4 km in the Viking Graben and 3 km in the Moray Firth (National Geophysical Data Center, 2004). The sediment on the thermally subsided flanks of the rifts reaches a maximum thickness of approximately 3–4 km around the Central Graben (England, 2000; National Geophysical Data Center, 2004). The basin contained significant hydrocarbon reserves; although following several decades of exploitation, production is now in decline.

The Porcupine Basin

To the west of Ireland the Porcupine Bank separates the Rockall Trough and the Porcupine Basin, located beneath the Porcupine Sea Bight (Fig. 1.1) and represents a continuation of the Irish Mainland Shelf (Hauser et al., 1995). The Porcupine Basin was generated via
1. Introduction

Faroe Islands

Deep Cretaceous basins

Late Cretaceous-Paleocene deformation

Cretaceous platforms - shallow basins

Tertiary domes

High/ridge

(a) Faroe-Shetland Basin (after Jolley et al., 2005)

(b) Voring Basin (after Eldholm et al., 2002)

(c) North Sea (Woodcock and Strachan, 2000) and Rockall and Porcupine Basins (Naylor et al, 1999)

Figure 1.4: Major structural elements of the sedimentary basins described in the text.
1. Introduction

episodic rifting and thermal subsidence through the Permo-Triassic and further post-rift thermal subsidence (Shannon et al., 1994). The basin's north-south orientation is notably different to the other basins to the north and west of the UK (Figs 1.1 and 1.4), with the main bounding faults oriented approximately orthogonal to the Mesozoic extension direction, rather than following the Caledonian trend (Shannon, 1991). Extension increases to the south, with $\beta$ factors up to 6 inferred for the southern part of the basin (Hauser et al., 1995) and there is a thick sedimentary succession, up to $\sim$9 km in the deepest areas (National Geophysical Data Center, 2004). The basin has been subjected to far greater investigation by the petroleum industry than the more westerly Rockall and Hatton Basins (29 exploration and appraisal wells have been drilled in the Porcupine Basin, whilst none have been drilled in the Rockall Basin), resulting in more detailed documentation on the sedimentary succession (e.g. Croker and Shannon, 1987; Moore, 1992; Moore and Shannon, 1995; Butterworth et al., 1999; McDonnell and Shannon, 2001).

Hatton and Rockall

The Hatton Bank marks the edge of the European continental lithosphere bounded to the northwest by the Hatton Continental Margin, which formed in the latest Mesozoic-Cenozoic time, prior to the onset of sea floor spreading in the early Eocene (Magnetic Anomalies 24–25, 56–59 Ma) (Laughton, 1971; Srivastava, 1978). Southeast of the Hatton Bank are the Hatton Basin, the Rockall Bank and Rockall Trough (Fig. 1.1). The region formed as a linked system of basins and continental blocks in the Late Palaeozoic and Early Mesozoic during the break-up of Pangaea (O'Reilly et al., 1998), with the basin bounding faults generally following the northeast-southwest Caledonian trend (Fig. 1.4). The Hatton and Rockall bathymetric lows have been sediment starved basins since at least the Late Mesozoic resulting in a much thinner sequence than is preserved in the other shallow-sea basins of western Europe (O'Reilly et al., 1998). The Hatton Basin contains $\sim$1.5 km of sediment and the Rockall Trough approximately 4 km (National Geophysical Data Center, 2004). Seismic images of the crust and granulite samples from the Rockall Bank confirm that the crust beneath the sediment is of continental affinity throughout the region (Miller et al., 1973; Morgan et al., 1989; Hauser et al., 1995; O'Reilly et al., 1998; Vogt et al., 1998). Seismic wide-angle reflection models show that the extension in the
region, as measured by the crustal $\beta$ factor, varies from $\sim$5 on the continental slope to 1.3 on Hatton Bank to 2 in the Hatton Basin and up to 6 in the Rockall Trough (Morgan et al., 1989; O'Reilly et al., 1995). The bathymetry of the northern Rockall Trough is pockmarked with seamounts (Fig. 1.1) generated by Tertiary volcanic activity, possibly associated with the Iceland Hotspot (White and McKenzie, 1989). At the southern end of the Trough wide-angle seismic data suggest that the transition from continental to oceanic crust occurs at the Charlie Gibbs Fracture Zone; a complex transform zone developed by at least 82 Ma (Shannon et al., 1994; Hauser et al., 1995).

Along the continental margin the crust has been thickened by the addition of large quantities of igneous underplating (Roberts et al., 1988; Fowler et al., 1989; Barton and White, 1997; Vogt et al., 1998). However, underplating appears to be limited to the continental margin and is not seen on seismic profiles in the Rockall Trough (except at the far northern end, where there may be underplate beneath northern Lewis (AMP line E, AMP Exclusive Report 01/3/5, 2001)).

**Faroe-Shetland Trough**

The opening of the Faroe-Shetland Trough has involved at least 6 phases of extension from the Devono-Carboniferous to the Paleocene (Kimbell et al., 2004). The first seaway connecting the Arctic sea to the north and the Tethys Ocean to the southwest may have opened by the early Jurassic (Ziegler, 1988). However, the main rifting event, associated with the opening of the North Atlantic, occurred in the mid Cretaceous, creating a series of basins in the region of the present Faroe-Shetland Trough (Fig. 1.4) (Hughes et al., 1998; Hitchen et al., 2003; Kimbell et al., 2004; Jolley et al., 2005). Continued subsidence and sea level rise through the Cretaceous submerged all, or almost all, of the highs by Maastrichtian times (74-65 Ma) (Hitchen et al., 2003). The northeast-southwest orientation of the Trough indicates that it too is influenced by the Caledonian structural grain. The Trough now contains over 8 km of sediment in the deepest regions (National Geophysical Data Center, 2004). No evidence has been found for lower crustal underplate beneath the Trough, except under the northwestern (seaward) side of the Faroe Islands along the Faroe-Iceland Ridge (Richardson et al., 1998).
1. Introduction

Norwegian Continental Margin

The development of the Norwegian Margin can be split into 3 major extensional tectonic events: 1) Carboniferous-Permian; 2) late Middle Jurassic to Early Cretaceous; and 3) late Cretaceous to early Cenozoic (Brekke et al., 1999; Kimbell et al., 2004). The Carboniferous saw the development of the Norwegian-Greenland rift system, with extension spreading down the Norwegian coast from the north (Ziegler, 1988). The Voring and Møre basins (Fig. 1.4) were formed mainly during the second of these episodes (Eldholm et al., 2002; Kimbell et al., 2004). During the third phase the axis of maximum extension shifted to the west; this final stage of extension lasted 18-20 Ma until continental break-up occurred in the late Paleocene/early Eocene (Mjelde et al., 2003; Fernández et al., 2004).

The Voring Plateau is a roughly semi-circular shallow bathymetric feature on the Norwegian Continental Margin (Fig. 1.1). Dense coverage of wide-angle seismic models has revealed that the feature is partially continental crust and partially oceanic (Mjelde et al., 1998; Berndt et al., 2001; Mjelde et al., 2001; Raum et al., 2002). The ocean continent transition appears to be a gradational boundary 30-50 km wide located just to the seaward side of the Voring Escarpment (Mjelde et al., 2001). Thick bodies of igneous underplate are present along most, if not all of the Norwegian Margin (Mjelde et al., 1998; Berndt et al., 2001; Mjelde et al., 2001; Raum et al., 2002; Raum, 2003).

1.1.3 The Iceland Plume

The northwestern region of the area shown in Figure 1.1 has been greatly affected by the Iceland mantle plume. The magmatic underplate, surficial lavas and seaward-dipping reflector sequences along the northwest continental margin are generally interpreted as evidence for continental break-up in the presence of the plume. The wide geographical spread of contemporaneous high-temperature volcanic rocks along the margin suggests that the continental break-up may coincide with the impact of the plume head on the base of the lithosphere (e.g. White, 1997; Saunders et al., 1997). The ongoing plume activity throughout the last 60 Ma is recorded by a number of geological features, including the increased thickness and unfractured nature oceanic crust that formed through this time (White, 1997), with much of the oceanic crust northeast of the Hatton Bank ~10 km thick.
or more (e.g. Smallwood et al., 1999; Weir et al., 2001) and Iceland itself reaching \(\sim 40\) km thickness (e.g. Darbyshire et al., 2000; Kaban et al., 2002; Foulger et al., 2003). The present day thermal anomaly is imaged by global and regional seismic tomography (e.g. Bijwaard and Spakman, 1999; Ritsema and Van Heijst, 2000; Foulger et al., 2001; Pilidou et al., 2004) and represents a strong seismic velocity anomaly down to \(\sim 400\) km with some ongoing debate as to whether the anomaly is seen in the lower mantle (e.g. Bijwaard and Spakman, 1999; Foulger et al., 2001).

The mantle plume had both permanent and transitory effects on the region. The volcanic material added to the continental crust, the thickened oceanic crust and the depletion of the mantle as a result of this melt extraction are permanent effects recorded in the density of the mantle and the thick ocean crust formed throughout the last \(\sim 60\) Ma. The transient effects, such as increased heat flow and dynamic topography, are seen associated with the current location of the Iceland Plume, but the geographical and temporal extent of such features in the past can only be postulated using the permanent effects and present day analogues. Therefore, the thermal anomaly associated with the plume and other transient effects are associated with uncertainty when considering the geological past.

**Faroe-Iceland Ridge**

Other than Iceland itself, the most notable effect of the Iceland Plume is the Faroe-Iceland Ridge (Fig. 1.1). The ridge was generated as a region of thickened oceanic crust as the North Atlantic opening propagated northwards (with Greenland separating from the Faroes Block \(\sim 54\) Ma). The crust beneath the ridge is approximately 30 km thick (along the bathymetric crest) and is aseismic. The shorter Greenland-Iceland Ridge to the north-west is likely to be a continuation of the same system across the mid-Atlantic ridge, the asymmetry being the result of westwards migration of the spreading centre and a change in spreading direction at \(\sim 44\) Ma (Bott and Gunnarsson, 1980). The transition from the continental Faroe Block to oceanic crust occurs as a gradual transition 40 to 100 km to the northwest of the Faroe Islands (Richardson et al., 1998).
1.1.4 Summary

Northwest Europe underwent significant widespread extension and rifting during the Mesozoic and subsequent thermal subsidence. As a result the present day continental crust, investigated in this project, is largely below sea level. Several of the sedimentary basins that formed during the Mesozoic rifting have developed considerable hydrocarbon reserves and have been extensively investigated using reflection seismology and exploration wells. As a result the structure and evolution of the North Sea Basin, the Faroe-Shetland Basin, the Porcupine Basin and the Norwegian margin are known in great detail. The sections above provide a brief overview and introduction to these basins. The deeper water Rockall and Hatton Basins have received far less attention from industry and as a result less detail is known about these regions. Beneath these sedimentary basins, the basement of the region preserves a record of the Silurian–Devonian aged Caledonian Orogeny and the Carboniferous–Permian aged Variscan Orogeny.

1.2 Crustal models

From a global perspective, the area of extended continental crust investigated in this project is among the most densely covered regions for seismic refraction/wide-angle reflection surveys. Such surveys provide the best constraints available on both the thickness of the crust and the velocity structure within the crust. In this work the refraction/wide-angle reflection models are compiled and used to produce a velocity model for the continental crust of the region shown in Figure 1.1.

Deep seismic refraction profiles have been acquired around the world since the 1950s, providing snapshots of the crustal structure beneath the survey areas. As the quantity of data has grown numerous authors have brought together the individual surveys into global or regional compilations and used the data to map out the thickness and structure of the crust (Mooney et al. (2002) list 38 compilations of crustal structure data published between Closs and Behnke (1961) and Mooney et al. (1998)). With ever increasing quantities and quality of individual surveys, there has been a continual improvement in the detail of these models. However, the global coverage of seismic data is still quite sparse and data are very irregularly distributed, limiting the global models to coarse grid dimensions (e.g. 2°
for CRUST2.0 (Bassin et al., 2000)). Regional studies in densely surveyed areas allow significant refinement of the models and thus much greater detail may be included (a regional model for the Barents Sea, currently under construction, will have a geographical resolution of 50 km (Bungum et al., 2005)).

The large regions of the globe with very poor or no data coverage force global models to rely on assumptions about crustal structure related to crustal type and age; the assumption that a velocity structure and crustal thickness of, for example, Archean crust is similar in all parts of the globe allows a type section to be created and applied to regions of Archean crust where no data are available (e.g. Hahn et al., 1984; Mooney et al., 1998; Bassin et al., 2000). With the high density of data in this region of northwest Europe such assumptions are not required. Simple interpolation between data is sufficient to complete the model, allowing a greater level of objectivity in the finished product.

In this section the previously published seismic crustal models are reviewed. Global velocity and crustal thickness models are discussed, followed by regional models that encompass all or some of the area covered in this work.

1.2.1 Global models

The first global map of crustal thickness was produced by Soller et al. (1982). This map is interpolated from 2508 refraction and surface wave data points and, although the coverage in Europe was good with respect to much of the globe, many features such as the Moho deflections associated with the North Sea Central Graben and the Rockall Bank are not resolved in this work (Fig. 1.5).

By the 1990s the amount of wide-angle reflection and refraction data had increased considerably (in Eurasia, Australasia and North America) allowing more detailed maps of variation in crustal thickness and seismic velocity (e.g. Fig. 1.7). Within the growing group of global models the CRUST models became (and still are) the dominant reference models for crustal structure following the publication of CRUST5.1 (Mooney et al., 1998). CRUST5.1 is a 5° by 5° model based on seismic refraction data published between 1948 and 1995. CRUST2.0 (Bassin et al., 2000) is a refined version, with a 2° by 2° grid and greater sensitivity to crustal type (Fig. 1.6).

The modelling approach used for the CRUST models was to compile the available
1. Introduction

Figure 1.5: Soller et al. (1982) map of crustal thickness for the North Atlantic and northwest Europe.

Figure 1.6: CRUST2.0 map of crustal thickness (Bassin et al., 2000).
wide-angle reflection and refraction data and use these to construct representative one

dimensional models for 14 different types of geologically and tectonically similar crust

(Archean, Early/Mid Proterozoic, Late Proterozoic, Phanerozoic, platform, extended crust,
orogen, forearc, continental arc, submarine plateau, shelf, continental margin, melt affected
ocean and standard ocean). Each of these 1D crustal models consists of 8 layers: ice, water,
soft sediments, hard sediments, upper crust, middle crust, lower crust and uppermost
mantle. Each of these layers has an assigned thickness, P-wave velocity, S-wave velocity
and density. The model is then defined by referencing the appropriate crustal code for
each cell.

This approach produces a model that can be defined by just two ascii files; one a matrix
of the longitude and latitude, giving the crustal code, and a second file giving the crustal
structure for each crustal type. This is a very elegant method as it allows a relatively
detailed crustal model (CRUST2.0 has 16200 tiles, each assigned one of 360 1D crustal
models) to be defined in two small files, creating a model that is easily transported, stored
and used. The disadvantage of the method is that any subtle variations in layer thickness
or velocity between cells can only be accommodated by construction of a new 1D model.
Thus any increase in the resolution of the model requires defining a large number of new
crustal types; the refinement from CRUST5.1 to CRUST2.0 increased the number of crustal
reference types from 139 to 360. Therefore, although this is a very tidy way of producing
a global model, the method would become cumbersome if applied to regions where there
is data available to constrain a significant proportion of the cells; well constrained regions
will have variations in layer thickness or velocity between most cells, thus requiring nearly
as many reference models as cells.

1.2.2 Regional models

Over the last ~20 years numerous maps of crustal thickness or Moho depth have been
produced for specific regions of northwest Europe (e.g. Meissner, 1986; Meissner et al., 1987;
Ziegler, 1990; Kinck et al., 1993; Chadwick and Pharaoh, 1998; England, 2000; Dèzes and
Ziegler, 2001; Yegorova and Starostenko, 2002; Clegg and England, 2003). Unlike the global
models, many of these maps integrate crustal-scale normal incidence data along with the
refraction and wide-angle reflection models. Europe has a very extensive network of crustal
Figure 1.7: Kinck et al. (1993) map of Moho depth for Fennoscandia.

normal incidence profiles, particularly in the offshore areas, collected by national programs such as BIRPS (Klemperer and Hobbs, 1991), DEKORP (Meissner and Bortfeld, 1990) and ECORS (e.g. Cazes and Torreilles, 1988); however, these profiles only poorly resolve the velocity structure and are therefore difficult to depth-convert. Consequently they are most useful when coincident with wide-angle reflection profiles (as the latter constrain the velocity structure). Generally a new map is produced every few years, reflecting the rapid acquisition of new data and refining the details of the Moho depth maps. The most recent maps for the area are:

**Kinck et al. (1993) Fennoscandian Moho depth map.** Kinck et al. compiled a Moho depth map for Fennoscandia from seismic reflection and refraction/wide-angle reflection profiles (Fig. 1.7). Reflection data are the primary data source for the Skagerrak, with the refraction data used to calibrate the results. For the rest of the
Figure 1.8: Maps of depth to the Moho produced from BIRPS data.

map refraction/wide-angle reflection data provide the depth control.

**British and North Sea Moho depths.** Chadwick and Pharaoh (1998) produced a map of the crustal thickness of Britain and the North Sea from BIRPS data, with additional constraint from the LISPB refraction profile (Barton, 1992) to compensate for the lack of deep reflection data onshore. The reflection two-way travel times were depth converted using velocity information from well data, BGS regional guides, the BIRPS Atlas (Klemperer and Hobbs, 1991) and several additional sources. In total up to 5 velocity layers were used for the sediments and 2 layers for the crystalline crust.

The BIRPS data were modelled again in 2000 to produce maps of two-way travel time and depth to basement and Moho for the North Sea and the waters around Ireland (England, 2000). There are two differences in method between this later model and the Chadwick and Pharaoh (1998) map: 1) the contouring method; and 2) the depth conversion procedure. The depth conversion for the England (2000) maps consists
of a simple two layer model of sediments (velocity 2.4 km/s) and basement (velocity 6.0 km/s) and the data are contoured digitally rather than manually. This process involves gridding the data onto a regular 15 minute grid and interpolation using GMT software (Wessel and Smith, 1991). The digital contouring produces more objective maps; with the contouring dependent solely on the data and independent of the author's knowledge of surface geology and topography. The two maps are shown together in Figure 1.8.

Dézes and Ziegler (2001) European Moho map. Dézes and Ziegler compiled the most up to date regional Moho depth maps, including Chadwick and Pharaoh (1998) and Kinck et al. (1993), to produce a regional map for Europe (Fig. 1.9).

The only regional model of crustal velocity specific to the area investigated here is the work of Clegg and England (2003). These authors mapped the depth to the Moho and the crustal velocity structure for a region encompassing Ireland, the UK and the North Sea using 19 wide-angle reflection and refraction seismic models. The method employed to interpret the crustal thickness and velocity structure is a 4 step process as follows:
1. Introduction

Figure 1.10: Maps of the crustal thickness and average crustal velocity from Clegg and England (2003).

- Construction of 1D velocity-depth profiles at 5 km intervals along each model. Velocity discontinuities are recorded at their precise depth;

- Regridding the velocity-depth profiles at 2 km vertical spacing using GMT (Wessel and Smith, 1998) to correct for the irregular depth spacing of velocity discontinuities;

- Contouring each 5 km by 2 km gridded profile using the Smith and Wessel (1990) minimum curvature algorithm, effectively reproducing a digital version of the published profile;

- Extracting a contour of a given value from each of the 19 models and contouring again in the third dimension, thus constructing a map.

They publish maps of the depth to the Moho together with a series of maps on the region’s velocity structure; namely the average crustal velocity, the depth to the 6 and 7 km/s isovelocity contours, and the thickness of crust with a P-wave velocity in excess of 7 km/s. The maps of depth to Moho and average crustal velocity are shown in Figure 1.10.
1. Introduction

As well as the development in detail of the Moho maps and crustal models over the last ~20 years, there has also been an evolution in the motivation behind the production of the models. The earliest models and Moho maps were generally produced simply to investigate crustal thickness, velocity structure and the relationship between deeper structure and the near-surface geology and topography (e.g. Soller et al., 1982; Meissner, 1986; Meissner et al., 1987). Increasingly the models have been published not only with interpretation of crustal structure, but are also used as tools for use in various fields within Earth Science. For example, the publication of CRUST5.1 (Mooney et al., 1998) included a study of surface wave data with the model used for crustal corrections. Another example is the BIRPS data based model of England (2000) which was used to investigate the role of basement geometry on basin development. This trend is not surprising as the early models produced the first look at crustal structure in any detail and as such produced some unexpected results (such as the lack of deep crustal structure or significant Moho root associated with the Caledonian and Variscan Orogenies (Meissner, 1986; Meissner et al., 1987)). The later models refined the detail of the structures, but were unlikely to show anything entirely new; therefore, their interpretation as free-standing information was likely to be limited, but their improvement in terms of tools for other fields was very significant. However this trend was only generally followed; exceptions are the Hahn et al. (1984) global crustal model which was produced solely to model the Earth's magnetic field and Clegg and England (2003) whose model was used only to investigate the crustal structure.

1.2.3 Summary

The progression seen from the early Moho depths maps (e.g. Soller et al., 1982) to the most recent crustal models (e.g. Mooney et al., 1998; Bassin et al., 2000; Clegg and England, 2003) is mainly the result of increasing quantity and quality of available data. The models are generally based largely, if not entirely, on refraction/wide-angle reflection seismic data. In the ~20 years that separate the earliest and most recent models a large quantity of seismic data has been acquired, especially in northwest Europe (the database used to construct the crustal model presented in this work contains 108 2D crustal models published between 1969 and 2005). Additionally the huge advances in data storage capabilities and computing power over the same period have allowed significant increases in both the density
of data acquired for each survey and refinement of the modelling process of this data. The one-dimensional gradient and intercept analysis and time-term modelling of the 1970s has given way to two- or three-dimensional ray tracing and tomographic methods. As a result not only are there more data to constrain the global and regional crustal models, the data are also more detailed and more accurate.

1.3 Isostasy

1.3.1 Introduction

As stated at the beginning of this chapter, the primary aim of this work is to construct crustal velocity and density models for northwestern Europe and the secondary aim is to use the density model to investigate isostatic compensation in the region. To address the second of these objectives the effective elastic thickness of the lithosphere is investigated. To introduce this work, the following section reviews the theories of isostasy and the specific work that has been undertaken for northwestern Europe.

Isostasy is the concept that the outer layers of the Earth respond to changes in load (e.g. topography) in such a way as to tend toward an equilibrium state where the loads are supported by the subsurface density distribution and rigidity of the lithosphere. The first ideas of isostasy were the those of Airy and Pratt in the 19th century (developed into full models by Hayford and Heiskanen in the early 20th century (Watts, 2001)). Both models of isostasy are based on the assumption that hydrostatic equilibrium prevails, with no lateral pressure variation beneath a constant depth of compensation. The models differ in their assumptions about how hydrostatic equilibrium is obtained. The Airy-Heiskanen model of isostasy assumes that the crust and mantle are both laterally constant density but that the crust varies in thickness. Topographic highs are therefore produced by regions of thicker crust, with mass of the mountains offset by a root projecting into the denser mantle. Therefore, topography can be maintained when the mass deficiency of the root balances the mass excess of the topography:

\[ r = \frac{h \rho_c}{(\rho_m - \rho_c)} \]

\[ a = \frac{W_d(\rho_c - \rho_w)}{(\rho_m - \rho_c)} \]  

(1.1)
where $\rho_c, \rho_m$ and $\rho_w$ are the densities of the crust, mantle and water and $r, h, a$ and $W_d$ are the heights of the root, topography, anti-root and water depth respectively (Fig. 1.11).

The Pratt-Hayford model assumes that topography is supported by lateral variation in the density above the compensation surface. The model was based on Pratt's belief in the contraction theory where high topography forms in warm (thermally expanded) regions and is therefore underlain by less dense material and low topography forms in cooler areas where contraction has occurred:

$$\rho_I = \rho_{sc} \frac{D_c - h}{D_c}, \quad \rho_o = \frac{(\rho_{sc}D_c - \rho_wW_d)}{(D_c - W_d)}$$ (1.2)

where $\rho_I, \rho_{sc}, \rho_w$ and $\rho_o$ are the densities of the crust inland, at the coast, water and the crust offshore respectively and $D_c, h$ and $W_d$ are the depth to compensation, height of topography and water depth (Fig. 1.11).

Both these models assume that the density variations supporting the topography are located directly beneath the topographic features. This requires a weak crust and mantle, whereas observations of broad deflections associated with topographic loads such as ocean islands suggest that the lithosphere has finite strength. Therefore, a significant change of approach was taken to reconcile a strong crust with isostasy and models moved away from local isostasy to a regional approach involving flexure of the lithosphere. This development represented a significant shift in the conceptual basis of isostasy. Airy and Pratt's views of flotation and contraction generating topography evolved into the concept of the lithosphere
as a rigid layer upon which topography and geological features (such as volcanos and river deltas) acted as loads causing flexure (Watts, 2001).

The flexural models are generally based on elastic plate theory, i.e. the lithosphere deforms as a rigid elastic plate above a weak inviscid fluid. Using this model the flexural rigidity of the plate \((D)\), a measure of the resistance to flexure, is given by

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

(1.3)

where \(E\) is the Young’s modulus of the plate, \(T_e\) is its elastic thickness and \(v\) its Poisson’s ratio. The general equation for the deflection of an elastic beam in an inviscid substratum is

\[
D \frac{d^4y}{dx^4} + (\rho_m - \rho_{infill})yg = 0
\]

(1.4)

where \(\rho_m\) is the density of the substratum, \(\rho_{infill}\) is the density of the material infilling the deflection and \(y\) is the deflection of the plate (Turcotte and Schubert, 2002). Specific equations for flexure can be generated for different load geometries and relationships between the density of the plate, loads, material infilling the flexure and the substratum.

This approach to isostasy assumes that geological features (e.g. volcanos and sedimentary basins) and variations in topography represent loads applied to the lithosphere. Although mostly marine, the region of northwest Europe investigated here shows significant undulations in bathymetry and topography, from over 3 km of water at the southern end of the Rockall trough to \(~1\) km of topography in the Scottish Highlands (Fig. 1.1). Along with this topography the sedimentary basins and areas of magmatic underplating represent additional loads. This results in an area of great heterogeneity in load and lithospheric structure.

### 1.3.2 Isostasy in northwest Europe

Any investigation into isostasy and the flexural strength of the lithosphere requires data on the load and flexural response of the lithosphere and a theoretical model relating load to flexural response and strength of the lithosphere. The most common approach for acquiring data on load and response is through the analysis of the relationship between gravity and topography.
1. Introduction

The relationship between gravity and topography

The flexural model of isostasy predicts that small loads will be supported by the strength of the lithosphere and are therefore associated with free air gravity anomalies, but (for topographic loads) little or no Bouguer anomaly. On the other hand large features will cause the crust to flex and will be supported by displacement of higher density mantle and hence will produce Bouguer gravity anomalies, but little or no free air anomaly. As the degree of compensation is reflected in the gravity anomaly the relationship between gravity and topography can be used to investigate the wavelengths over which isostatic compensation applies and thus determine the flexural rigidity of the lithosphere.

Since the mid-1980s the coherence between the topography and Bouguer gravity field has been the most widely used approach to estimating the elastic thickness in continental settings. The coherence is a measure of the correlation between the two fields in the wavenumber domain. The general form of the coherence is given by

\[
\gamma^2(k) = \frac{C_c(k)C_{c*}(k)}{E_{\Delta g}(k)E_t(k)}
\]

where \(C_c(k), E_{\Delta g}(k)\) and \(E_t(k)\) are the cross spectrum of gravity and topography, the power spectrum of gravity and the power spectrum of topography respectively. The * denotes the complex conjugate. The reason for the popularity of the coherence method (as opposed to the admittance method discussed later) is the relative insensitivity of coherence to buried loading. In flexural analysis, topography and sedimentary basins are modelled as loads applied to the top of the lithosphere, i.e. surface loads; however processes such as magmatic underplating can apply loads deeper in the lithosphere, i.e. buried loads. The ratio of surface to buried loading is often unknown and therefore significant uncertainty is present in elastic thickness estimates. However, the coherence between the Bouguer gravity anomaly and the topography is relatively insensitive to the loading ratio (Forsyth, 1985). For short-wavelength loads there will be little flexure of the lithosphere. In the case of surface loading the Bouguer anomaly will be small but there will be topography. Conversely, for buried loading the lack of flexure will result in a Bouguer anomaly, but little topography; either way the coherence between the topography and the gravity anomaly is low. For long-wavelength loads there will be significant flexure; therefore, with either
surface or buried loading, there will be both a Bouguer anomaly and topographic expression and so the coherence will be high (Watts, 2001).

In northwest Europe the coherence method has been used by Poudjom Djomani et al. (1999); Daly et al. (2004); Pérez-Gussinyé et al. (2004) and Pérez-Gussinyé and Watts (2005) to calculate the elastic thickness of specific regions. Poudjom Djomani et al. (1999) calculate the variation in elastic thickness across Fennoscandia, modelling the observed coherences with buried loads assumed to be applied to the base of the crust. Their elastic thickness estimates range from ~8 km along the Norwegian coast to a maximum of ~70 km in the Baltic shield (Fig. 1.12). Daly et al. (2004) estimate the elastic thickness and amount of buried loading in the Irish Atlantic margin; they find the elastic thickness varies from ~6 km on the eastern margin of the Rockall Trough to ~18 km on the Edoras Bank (Fig. 1.12). The elastic thickness of Fennoscandia was also investigated by Pérez-Gussinyé et al. (2004) and expanded to cover all of Europe by Pérez-Gussinyé and Watts (2005). This most recent work shows highly variable elastic thickness estimates from values less than 10 km for much of the marine area to values greater than 60 km for England, the North Sea and eastern Europe.

Elastic thickness estimates based on the Bouguer coherence have been questioned by McKenzie and Fairhead (1997) who suggest that the effect of erosion on topography will tend to remove topography and reduce the coherence of the Bouguer anomaly and topography and therefore such calculations are closer to maximum bounds for the elastic thickness than to accurate estimates. They advocate calculating the elastic thickness from the free air gravity anomaly, using only the portion of the anomaly that correlates with the topography. This removes any part of the spectrum that has poor coherence due to erosion, since where there is little effect from erosion partially compensated topography will be correlated with the free air gravity anomaly. Whereas, if erosion has reduced the coherence between the topography and gravity the correlation between the two will be poor. This approach has been used by Tiley et al. (2003) to estimate the elastic thickness of Britain and Ireland; they estimate a $T_e$ of 5 ±2 km. However, the method has not been widely accepted, Simons et al. (2000) and Banks et al. (2001) argue that this approach neglects internal loading, while Swain and Kirby (2003) point out that large free air anomalies that are not associated with topography must indicate that the lithosphere is supporting
Figure 1.12: The elastic thickness of Fennoscandia (Poudjom Djomani et al., 1999), the Irish Atlantic margin (Daly et al., 2004) and Europe (Pérez-Gussinyé and Watts, 2005) obtained from coherence analysis.
subsurface loading, and is therefore strong. McKenzie (2003) attempted to address these objections, providing further explanation of the inclusion of loads in the theoretical admittance curves. However, the controversy has not died away. Most recently Pérez-Gussinyé et al. (2004) and Pérez-Gussinyé and Watts (2005) argue that McKenzie’s method is inaccurate as analytical predictions of admittance are compared to observations acquired from finite windows. They make the case that the calculated admittance is dependent on the size of the window and therefore must be compared to a theoretical relationship that is based on finite data. By reformulation of the predicted admittance to account for this, the estimated elastic thicknesses calculated using admittance are increased and the discrepancy with the coherence based estimates is greatly reduced (Pérez-Gussinyé and Watts, 2005).

Not all investigations of the elastic thickness of the lithosphere involve the relationship between gravity and topography. Alternative approaches, which avoid the controversy surrounding admittance and coherence, include forward and inverse modelling of sedimentary basin evolution and estimation of degree of compensation directly from the density structure of the lithosphere.

Modelling basin evolution

The combination of wells and reflection seismic data provides great detail on the subsidence histories and geometries of the sedimentary basins in the marine parts of the study region. Through backstripping (the stepwise removal of sedimentary sequences, with correction for subsidence caused by sediment compaction, sea-level change and sediment and water loading) the tectonic subsidence/uplift history of sedimentary basins can be retrieved. This data on the basin evolution can be used to place constraints on a conceptual model for basin formation, allowing investigation of parameters such as the amount of extension and elastic thickness of the lithosphere.

The earliest application of this approach in the region covered in this project was by Barton and Wood (1984) in the North Sea. Using well data derived subsidence curves and crustal structure information from seismic modelling, they model the evolution of the North Sea’s post-Cretaceous sedimentary load, thermal load (estimated from post rift subsidence) and topographic load (from palaeobathymetry). They determine that in order to fit the subsidence curves and present day gravity anomaly with a uniform, pure shear
stretching model the elastic thickness is required to be less than 5 km for the central North Sea.

More recently models of crustal extension and basin formation have evolved to account for the change from brittle to ductile deformation in the mid-crust and to allow for accommodation of strain through either pure shear or simple shear (e.g. the flexural-cantilever model where the faulted brittle upper crust behaves as interacting flexural cantilevers over a ductile, pure sheared, lower crust and upper mantle (Marsden et al., 1990; Kusznir et al., 1991)). The flexural-cantilever model has been applied to the Viking Graben in the northern North Sea by Marsden et al. (1990). Using fault locations and slips determined from seismic data, they conclude that the subsidence and basin structure are best modelled with an elastic thickness of 6 km for the Triassic rifting, reducing to 3 km for the Jurassic rifting phase.

In the Danish Central Graben Korstgård and Lerche (1992) forward modelled the basin structure imaged on 5 deep seismic profiles by regarding the base of the Triassic sediments as the top of a preflexed elastic cantilever, with the overlying sediment representing additional load. By forward modelling the effect of the load on the cantilever they determined the flexural parameters (rigidity, bending moment and dip) that best reproduce the observed structure. They concluded that the subsidence generated by the Triassic to present day sediment load is best matched by a preflexed plate with an initial flexural rigidity of $10^{20.240.2}$ Nm. Using Equation 1.3, a Young's modulus of 70 GPa and Poisson's ratio of 0.25, this flexural rigidity is equivalent to a $T_e$ of $\sim 3$ km.

On the volcanic Hatton margin, Watts and Fairhead (1997) investigated the elastic thickness and the relative importance of surface sediment loading to buried loading from the magmatic underplate. Using backstripping to isolate initial load on the continental margin caused by the sediments and the underplate they calculate the individual components of the gravity anomaly caused by the two loads and the crustal structure. The flexure caused by these loads modifies the predicted gravity anomaly; therefore, by calculating the gravity anomaly resulting from flexure controlled by different elastic thickness and comparing this to the observed anomaly the elastic thickness was retrieved. Watts and Fairhead conclude that the underplating signal dominates the gravity anomaly and that the elastic thickness is $< 5$ km.
Recently onshore Britain has received some attention with investigation of the evolution of the South Wales coal field (Burgess and Gayer, 2000) and the Quaternary uplift of south central England (Watts et al., 2000, 2005). Burgess and Gayer (2000) use forward modelling of a flexural foreland basin in front of the propagating Variscan load to fit backstripped subsidence curves. Their modelling suggests that the elastic thickness of southwest Britain during the Carboniferous was \( \sim 40 \) km. Watts et al. (2000, 2005) model the Quaternary uplift of southern England using a flexural unloading model; they suggest that the topographic highs of the Cotswold Hills and Forest of Dean were uplifted in response to excavation of the Vales of Evesham and Gloucester, concluding that the present day topography is best matched by unloading of an elastic plate 5–10 km thick.

Direct estimates from crustal structure

The response of the lithosphere to applied loads can be viewed as a continuum: very small loads are supported entirely by the strength of the lithosphere, with no discernable flexure; intermediate loads are partially supported by the lateral strength of the lithosphere and partially by displacement of dense mantle beneath the load as the lithosphere flexes; and very large loads are supported hydrostatically by displacement of the mantle beneath the load. To what extend a load is supported hydrostatically is therefore a function of the wavelength of the load and the strength of the lithosphere. Turcotte and Schubert (2002) define a parameter \( C \), the degree of compensation, which is the ratio of the observed deflection of the lithosphere \( (w_0) \) to the deflection that would be caused by hydrostatic equilibrium \( (w_{0\infty}) \):

\[
C = \frac{w_0}{w_{0\infty}} \quad (1.6)
\]

The degree of compensation depends on the wavelength of the load \( (\lambda) \), the density structure of the crust \( (\rho_c) \) and mantle \( (\rho_m) \) and the flexural rigidity \( (D) \) of the lithosphere. In the case of periodic surface loading the degree of compensation is related to these parameters by:

\[
C = \frac{(\rho_m - \rho_c)}{\rho_m - \rho_c + \frac{D}{g} \left( \frac{2\pi}{\lambda} \right)^4} \quad (1.7)
\]

where the flexural rigidity is related to the elastic thickness by Equation 1.3 (Turcotte and Schubert, 2002). Therefore, if the wavelength of the load is known and the degree of
compensation can be calculated from a crustal model the elastic thickness can be extracted without further modelling. This approach has been taken, with relation to regions of northwest Europe, on two occasions; for the sedimentary basin in the English Channel (Warner, 1987) and for Scotland and northern England (Barton, 1992).

Warner (1987) estimated the degree of compensation for the English Channel using the two way travel time on BIRPS deep seismic profiles. Assuming compensation at the Moho and allowing for uncertainties in the velocities and densities of the sediments and crust, he showed that the near constant travel time to the Moho indicates that the compensation of the ~100 km wavelength basin is between 50 and 100%, which using Equation 1.7, requires an elastic thickness of less than 5 km.

Barton (1992) estimated the degree of compensation of features along the LISPB profile by comparing observed topography to the hydrostatic topography predicted by the gravity model of the profile. Although the uncertainties are large, the partial compensation of the Southern Uplands and almost complete compensation of the Grampian Highlands requires an elastic thickness of less than 5 km.

1.3.3 Summary

The syn- and post-rift sedimentary basins and the continental margin of northwest Europe provide the main loads used for studying isostatic compensation in the region. Most of the recent work on the flexural strength of Europe has utilised coherence or admittance analysis of the topography and gravity data, although other approaches involving modelling of topographic/basin evolution have also been applied. These investigations (summarised in Table 1.1) generally agree that the continental margins and the highly extended crust to the west of Ireland have a low elastic thickness (<20 km); however there is considerable discrepancy with regards to the strength of Britain and the North Sea. The main disagreement involves the best approach to interpreting the relationship between topography and gravity, particularly with regards to estimating the size and importance of buried loads. As this work involves the construction of a new highly detailed density model it provides the opportunity to investigate the flexural strength of the continental lithosphere with greater knowledge of the mass distribution within the crust.
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<td>~40 km</td>
</tr>
<tr>
<td>Watts et al. (2000, 2005)</td>
<td>Modelling topography with flexural unloading model</td>
<td>south central England</td>
<td>5–10 km</td>
</tr>
<tr>
<td>Poudjom Djomani et al. (1999)</td>
<td>Bouguer coherence</td>
<td>Fennoscandia</td>
<td>mapped across the region; minimum = 8 km along the Norwegian coast, maximum = 70 km in the Baltic shield</td>
</tr>
<tr>
<td>Tiley et al. (2003)</td>
<td>Free air admittance</td>
<td>British Isles</td>
<td>5 ± 2 km</td>
</tr>
<tr>
<td>Daly et al. (2004)</td>
<td>Bouguer coherence</td>
<td>Irish Atlantic margin</td>
<td>mapped across the region; minimum = 6 km in the eastern Rockall Trough, maximum = ~18 km on Edoras Bank</td>
</tr>
<tr>
<td>Pérez-Gussinyé et al. (2004)</td>
<td>Bouguer coherence &amp; Free air admittance</td>
<td>Fennoscandia</td>
<td>mapped across the region; minimum = 20–40 km in Norway, maximum = 70–100 km in Karelia province (southern Finland) mapped across the region with a range from &lt; 10 km near continental margins to &gt; 60 km for southern Britain, North Sea and Eastern Europe.</td>
</tr>
<tr>
<td>Pérez-Gussinyé and Watts (2005)</td>
<td>Bouguer coherence &amp; Free air admittance</td>
<td>All of Europe</td>
<td></td>
</tr>
</tbody>
</table>

Table 1.1: Summary of the previous studies into the elastic thickness of northwest Europe.
1.4 Summary and Thesis Review

This first chapter has primarily functioned as a review of previous work related to the two objectives of this study: modelling the crustal velocity and density structure of northwest Europe and interpreting this model in terms of isostasy in the region. Following this introduction the next chapter is a review of the present modelling methods for wide-angle reflection seismic data; included as the crustal velocity and density models are constructed from wide-angle reflection and refraction seismic models. The chapter includes a brief description of the historic modelling methods, but is mainly a detailed review of the current techniques and includes the processing of an example dataset. The third chapter details the construction of the velocity model, then presents and interprets the completed velocity model. Chapter 4 covers the conversion of the velocity model to a density model and presents the final crustal model. In Chapter 5 the velocity and density model are discussed in further detail, with particular reference to the geological implications of the model, and in Chapter 6 the density model is used to investigate the flexural strength of the region. The final chapter summarises the preceding chapters and highlights the main results.
Chapter 2

Modelling Wide-Angle Reflection Data

The crustal model presented in Chapters 3 and 4 is based largely on two-dimensional models of the crust generated from wide-angle data. Therefore an understanding of the modelling procedure and accuracy of the resulting crustal sections is needed to appreciate the strengths and limitations of the input data for the regional crustal model. In this chapter the historical and current techniques used to determine crustal structure from wide-angle reflection seismic data are reviewed in detail. To assist in demonstrating the current modelling procedure an example dataset is worked up, the data is MONALISA Profile 3 from the southern North Sea.

2.1 Historical Techniques

The oldest and most basic form of analysis for refraction/wide-angle reflection data is the gradient-intercept method. This approach was commonly used in the 1950s (e.g. Hill and King, 1952; Day et al., 1955) and continued to be used occasionally (normally for small surveys only sampling the near surface and shallow crust) through into the early 1980s (e.g. Sellevoll and Warrick, 1971; Whitmarsh et al., 1974; Jones et al., 1984). This analysis simply involves fitting a series of best-fit linear equations to a plot of arrival time versus offset between shot and receiver. The general equation for arrival time ($t_n$) for the wave
refracted along the top surface of the \( n^{th} \) horizontal layer is:

\[
t_n = \frac{x}{v_n} + \sum_{i=1}^{n-1} \frac{2z_i \cos \theta_{in}}{v_i}
\]  

(2.1)

where \( x \) is the source-receiver offset, \( v_n \) is the velocity of the \( n^{th} \) layer, \( z \) is the layer thickness and

\[
\theta_{in} = \sin^{-1}(v_i/v_n)
\]  

(2.2)

thus by starting at the shallowest layer and working downwards, the velocity and thickness of the layers can be estimated from the gradients and intercepts of the best-fit lines. Alternative equations can be generated for dipping or irregular (non-planar) interfaces. The most general of these forms the basis of time-term analysis.

Time-term analysis, developed by Willmore and Bancroft (1960), is based on the following general equation for the travel time \( t_{ij} \) between any source-receiver pair:

\[
t_{ij} = \frac{\Delta_{ij}}{v} + a_i + b_i
\]  

(2.3)

where \( \Delta_{ij} \) is the source-receiver offset; \( a_i \) is the time-term associated with the shot and \( b_i \) is the time term associated with the receiver. The first term in Equation 2.3 is the time taken for the energy to travel along the refracting interface between the source and receiver and the time-terms are the additional time taken for the seismic wave to travel between the refracting interface and the surface.

For \( n \) sources and \( m \) receivers there is a maximum of \( nm \) equations of type 2.3 (if the arrival is seen on every source-receiver pair). If the refracting interface is approximately level then the time-term for a given receiver will be similar for all sources recorded at that station, likewise the time-term at a given source will be similar for any of the receiver stations. If this applies and the velocity is laterally constant then \( v, a_i \) and \( b_i \) are constants and so there are only \( n + m + 1 \) unknowns. The data can then be regrouped into equations.
relating to a single source \((n_j)\) or receiver \((m_i)\), such as

\[
n_j(b_j) = \sum_{i=n}^{i=1} (t_{ij} - \Delta_{ij}/v) - \sum_{j=1}^{j=m}(a_i) = \sum_{j=1}^{j=m}(b_j) (2.4)
\]

\[
m_i(a_i) = \sum_{j=m}^{j=1} (t_{ij} - \Delta_{ij}/v) - \sum_{j=1}^{j=m}(b_j) (2.5)
\]

These can then be used to construct a series of \(mn\) simultaneous linear expressions that can be solved to obtain the time-terms and velocity.

The conversion of the time terms to a crustal model is straightforward as time-terms can be defined as:

\[
a_h = \int_0^h \frac{(v_h^2 - v_z^2)^{1/2}}{v_h v_z} dz (2.6)
\]

where \(a_h\) is the time-term for depth \(h\); \(v_h\) is the velocity immediately below the horizon at depth \(h\); and \(v_z\) is the velocity at depth \(z\) \((h > z)\) (Berry and West, 1966). Therefore once the time-terms are known, velocity-depth profiles can be constructed under the sources and receivers.

An additional approach occasionally used to estimate the crustal structure from reflection data is the \(T^2 - X^2\) method (e.g. Bamford et al., 1976; Brooks et al., 1984). This method uses the following relationship between two-way travel time and source receiver offset for reflections from horizontal constant velocity layers:

\[
T_x^2 = T^2_0 + \frac{X^2}{v^2} (2.7)
\]

where \(T_x^2\) is the two-way travel time at offset \(X\), \(T^2_0\) is the two-way travel time at zero offset and \(v\) is the velocity between the surface and the reflector. By plotting a best-fit line through the Moho (or mid-crustal) reflections an estimate of the depth to the reflector and average crustal velocity above the reflector is obtained.

These early modelling approaches suffer from the need to use very simple structure to approximate the subsurface, generally simple planar boundaries separating constant velocity layers. The presence of complex structure on velocity discontinuities or lateral variation in velocity cannot be incorporated into the models. In addition much of the potential information from the data is not used, these methods use only the arrival time.
of the refracted (time-term) or reflected ($T^2-X^2$) energy, but do not incorporate both sets of arrivals into a single model nor do they use the amplitude of the arrival.

In the early 1980s these techniques were replaced by modelling methods based on ray tracing (e.g. Červený and Pšenčík, 1981; Chapman and Drummond, 1982; Červený and Pšenčík, 1983). By forward modelling the path of the wavefront through the ground complex structure and lateral variations in velocity could be included in the model. Additionally, the amplitude of the arrival could be predicted and compared to the observed arrival providing more data with which to constrain the models. This development allowed the production of significantly more realistic models with 2D velocity inhomogeneity. However, in the few instances where data originally analysed using the older approaches described above have been remodelled with modern ray tracing techniques, the main features of the crust (such as basement and Moho depth or presence of high velocity lower crust) change very little. It is generally only the more detailed velocity structure of the crust that is refined by the use of ray tracing (e.g. LISP (Bamford et al., 1978; Barton, 1992) and ICSSP (Jacob et al., 1985; Al Kindi, 2002)).

With rapidly increasing computing power the progression in modelling techniques continued with the development of inversion routines for the ray tracing codes (e.g. Zelt and Smith, 1992) and new crustal tomographic modelling methods in the 1990s (e.g. Hole et al., 1992; Zelt and Barton, 1998). In the following sections the current approach to modelling wide-angle seismic reflection data is described in detail and the advantages over older techniques discussed.

### 2.2 Current Modelling Techniques

The normal approach for modelling wide-angle data is to produce a minimum structure model as such a model should represent the most simple solution. The simplest solution that fits all available data is preferred to any more complicated solutions, as additional complexity represents unconstrained structure, which by definition cannot be justified. The primary measures of the structural complexity are the spatial variation in the velocity and the velocity gradients.

At present the most popular method for modelling data involves ray tracing using
RAYINVR (Zelt and Smith, 1992). However many other ray tracing and tomographic codes exist and other commonly used codes include those of Hole et al. (1992) and Zelt and Barton (1998). The popularity of RAYINVR is possibly due to the following features which distinguish it from other methods (Zelt, 1999):

- The spacing and number of model nodes specifying velocities and interface depths can be uniform or irregular,
- Any or all of the model parameters may be selected for inversion at each iteration,
- The velocities are ‘tied’ to layer interfaces,
- Simultaneous inversion of all arrival types is possible,
- Vertical velocity gradients and layer thicknesses may be maintained during the inversion,
- ‘Floating’ reflectors may be included that are not tied to the velocity field.

These features allow additional information such as geological constraints or prior information from gravity and amplitude modelling to be included. RAYINVR produces a coarse, irregular gridded, near minimum parameter model. Such a model is constructed so that the removal of any one of the nodes produces a reduction in the fit between the observed and calculated travel times. The construction of such a model generally supports the aim of producing a minimum structure model (as minimising the number of parameters generally minimizes the variation in velocity). However, in a few cases departing from a minimum parameter model allows simplification of the velocity structure, without reducing the fit to the data. In these cases the non-minimum parameter model is considered to be closer to minimum structure and so preferred.

A detailed modelling strategy is outlined in Figure 2.1, this strategy was used to model the MONALISA Profile 3 data. The progression from receiver gathers to velocity model can be broken into four stages (as shown by the colours in Figure 2.1): 1) Pre-modelling; 2) construction of a starting model; 3) model development; and 4) model investigation and assessment.

The MONALISA project was designed to investigate the location of the suture between Avalonia and Baltica (the Thor Suture (Berthelsen, 1998)) in the North Sea (MONA
Figure 2.1: Flowchart showing the typical development of a P-wave velocity model, as used for the MONALISA data.
2. Modelling Wide-Angle Reflection Data

LISA working group, 1997a,b). A combination of normal incidence and wide-angle seismic reflection data were collected over two years on four profiles fanning across the central North Sea (Fig. 2.2). The profiles cross the dominant regional structures; the Mid-North Sea - Ringkøbing-Fyn High (MNSH-RFH) and the Central and Horn Grabens. Wide-angle data were collected on Profiles 1-3. The Profile 1 and 2 data were modelled by Abramovitz and Thybo (1998, 2000) and the Profile 3 data by Nielsen et al. (2000). Profile 3 was chosen as the example dataset as it represents a good ‘average’ for the input data for the velocity model of northwest Europe. The data were acquired in 1995 with fairly typical shot and receiver spacings for the mid-90’s (nominal shot spacing ~100 m and 8 receivers distributed along a 381 km profile).

2.2.1 Pre-modelling

Any modelling approach based on travel time data, be it ray tracing or tomography, will start with the following three stages: 1) phase identification; 2) analysis of travel time uncertainty; and 3) checking travel time picks for reciprocity. For the MONALISA data the phase identification was conducted in ZPLOT (Zelt and Smith, 1992) with different phases identified by changes in the apparent velocity of the arrivals. Visual inspection of the receiver gathers shows that there is considerable structure affecting the arrival times,

Figure 2.2: Location of the MONALISA profiles. The cross-hatching marks the approximate area of the MNSH-RFH, HG indicates the Horn Graben.
complicating the identification of different crustal phases. This is not unusual in wide-angle modelling and so it is standard practice to reassess the assignment of picks to crustal phases on occasions during the modelling. For the MONALISA data the reassessment of the picks during travel time modelling included forcing the arrivals into both deeper and shallower layers within the model; this checked that the assigned layer produced the fastest arrival, reducing the subjectivity associated with assigning picks to certain layers (Zelt et al., 2003).

The estimates of travel time uncertainty were derived from a qualitative measure of the signal strength to noise ratio, as cycle skipping is the largest likely source of error. Due to the increasing strength of the received signal through the first few oscillations the first break can be lost when the signal to noise ratio deteriorates, therefore the apparent first break is possibly one of the later cycles. The assigned uncertainties for each phase in the MONALISA data are given in Table 2.1.

The picked phases must be checked for travel time reciprocity. As the number of truly reciprocal pairs is often very low, geometries giving a close approximation to reciprocity are also checked. Two types of near reciprocal geometries, as defined by Zelt (1999), have

<table>
<thead>
<tr>
<th>Phase</th>
<th>Assigned uncertainty (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>C1</td>
<td>±0.050 &lt; 1 km offset</td>
</tr>
<tr>
<td></td>
<td>±0.025 ≥ 1 km offset</td>
</tr>
<tr>
<td>C1C</td>
<td>±0.075</td>
</tr>
<tr>
<td>C2</td>
<td>±0.025 &lt; 7 km offset</td>
</tr>
<tr>
<td></td>
<td>±0.050 ≥ 7 km offset</td>
</tr>
<tr>
<td>C2C</td>
<td>±0.050</td>
</tr>
<tr>
<td>C3</td>
<td>±0.100</td>
</tr>
<tr>
<td>C3C</td>
<td>±0.200</td>
</tr>
<tr>
<td>Pb</td>
<td>±0.100</td>
</tr>
<tr>
<td>PiP</td>
<td>±0.200</td>
</tr>
<tr>
<td>Pg</td>
<td>±0.100 &lt; 80 km offset</td>
</tr>
<tr>
<td></td>
<td>±0.150 ≥ 80 km offset</td>
</tr>
<tr>
<td>PmP</td>
<td>±0.250</td>
</tr>
<tr>
<td>Pn</td>
<td>±0.200</td>
</tr>
<tr>
<td>PnP</td>
<td>±0.250</td>
</tr>
</tbody>
</table>

Table 2.1: C1 and C2 are the refracted arrivals from the two post-rift sedimentary layers; C1C and C2C are the reflections from the base of the two layers. C3 is the refracted arrival from the syn-rift sediments and C3C is the reflection from the base of the rift. Pb are the refactorions from within the basement/top layer of the crust, with PiP reflections from the base of this layer within the crust. Pg and Pn are the crustal and mantle refractions respectively and PmP the Moho reflections. PnP are the sub-Moho refactorions.
been used for the MONALISA data. Firstly, interpolation was used to approximate the travel time where no shot of the correct offset was available (Fig. 2.3). Secondly, where the maximum offset picked from one OBH falls just short of another OBH then that reduced offset value was used for both OBHs to test reciprocity of the travel times (Fig. 2.3).

2.2.2 Construction of the starting model

The second stage for travel time modelling of wide-angle reflection data is the construction of a starting model. The choice of precise modelling method/package will affect the form of the starting model, but all ray tracing and tomographic methods require an initial model.

Velocity models produced from wide-angle seismic travel time data are non-unique as velocity and depth are not independent controls on the travel time, but trade off against one another (i.e. the same travel time can be obtained through two layers where one is thin and slow and the other layer thicker but faster). With complicated models, such as those generated from crustal scale surveys, it is impossible to fully investigate all the possible variations in parameter values; therefore the selection of an appropriate, geologically reasonable starting model is extremely important as it focuses investigation of model space into a geologically plausible region.

Starting models are therefore often generated using reliable external information to add constraint to the model. The potential sources of external data are vast, however a few are particularly common, these are:

Figure 2.3: Near reciprocal geometries used to check the consistency of the travel time picks (Zelt, 1999).
2. Modelling Wide-Angle Reflection Data

- Proximal normal incidence data (e.g. Horsefield et al., 1994; Mjelde et al., 1997; Barton and White, 1997; McCaughey et al., 2000; Raum et al., 2002),

- Well or borehole data (e.g. Horsefield et al., 1994; Barton and White, 1997; Raum et al., 2002),

- Proximal wide-angle/refraction surveys or expanding spread profiles (ESP) or previous models of the dataset (e.g. Morgan et al., 1989; Barton, 1992; Horsefield et al., 1994; McCaughey et al., 2000; Raum et al., 2002).

Simple 1D modelling of the main wide-angle dataset is normally used to produce an approximate starting velocity structure. In regions of relatively low structural complexity the single 1D velocity profile that is the best-fit to the complete dataset may be all that is used (e.g. Grandjean et al., 2001). Alternatively collating individual 1D models produced for each shot (or receiver in the case of marine data) to form a rough 2D section may be used (e.g. Roberts et al., 1988; Lowe and Jacob, 1989; Barton and White, 1997; Masson et al., 1998).

The starting model for the RAYINVR travel time modelling of the MONALISA data was produced from a combination of tomographic modelling of the first arrivals in FAST (Zelt and Barton, 1998) and preliminary modelling using ray tracing in RAYINVR. FAST is a 3D or 2D code for velocity modelling of wide-angle reflection/refraction seismic data; the code is described in Appendix A.1. The starting model used in FAST was a one dimensional velocity profile designed to fit the average arrival times for all of the OBHs. The free parameters controlling the modelling (the cell size for the forward and inverse steps, the relative weighting of time misfit to model smoothness, and the weighting of vertical to horizontal model smoothness) were optimised using checkerboard tests based on the starting model (Fig. 2.4). The checkerboard tests were conducted on a range of anomaly dimensions to ensure that the parameters allowed anomalies of different wavelengths to be recovered. The dimensions of the individual anomalies ranged from 25 km to 100 km horizontal extent and 5 km to 20 km vertical scale. The amplitude of the anomalies is tapered from a maximum of 0.5 km/s in the centre to 0 at the edges. The FAST velocity model produced using the optimised parameters is shown in Figure 2.5.

To construct a starting model for the RAYINVR inversion, the results of the FAST model
Figure 2.4: An example of the checkerboard tests used to optimise the inversion parameters used in FAST. In this example the optimum model cell size was determined to be 1 km as this produced the best retrieval of anomalies throughout the crust. The 0.5 km cells produce significant smearing along the ray paths even at very shallow levels within the crust. The 2.0 km cells smear the checks, causing poor resolution of the anomalies.
Figure 2.5: The P-wave velocity model produced in FAST and the number of rays sampling each grid cell.
were combined with preliminary forward modelling conducted in RAYINVR. The layers from the simple forward model were superimposed on the FAST model. This has the effect of visually highlighting the features in the FAST model that are particularly robust and visually suppressing features that are less robust (Fig. 2.6(a)). This combined image was then used to qualitatively determine appropriate layers and velocities for the starting model (Fig. 2.6(b)).

2.2.3 Model development

Travel time modelling

The third stage is the development of the model through the chosen modelling method. In the case of the MONALISA data this involves forward and inverse travel time modelling in RAYINVR and the associated damped least squares inversion routine (Zelt and Smith, 1992). RAYINVR is the most popular modelling method applied to the data used to construct the regional 3D model presented in Chapter 3, it has been used for 45% of the 2D profiles in the model. The RAYINVR code is described in Appendix A.2.

The model was improved by minimising the difference between the calculated and observed travel times using a top-down approach, as is standard in 2D modelling. A top-down approach involves refining the upper layers then fixing them before improving the lower layers. As is normally the case with wide-angle reflection surveys, the inversion was not constrained well enough to apply directly to a simple starting model. Instead forward modelling was used to define the general features of the model and then, once a model had been produced with a reasonable fit to the data, the inversion routine was used to optimise the features. The degree to which the model matches the observed data was tested through three statistics; 1) the number of observed arrival locations met by the synthetic rays traced through the model; 2) the travel time residual; and 3) the Chi-squared value. The Chi-squared value (Equation 2.8) is used to determine when the travel time residual has been reduced to the minimum required by the data:

\[
\chi^2 = \frac{1}{n-1} \sum_{i=1}^{n} \left( \frac{t(i)_{\text{calc}} - t(i)_{\text{obs}}}{u(i)} \right)^2
\]

(2.8)

where \( t(i)_{\text{calc}} \) is the modelled travel time for the \( i^{th} \) data point, \( u(i) \) is its uncertainty,
2. Modelling Wide-Angle Reflection Data

Figure 2.6: The starting model for the RAYINVR inversion of the MONALISA data (b) was constructed by combining the FAST model with preliminary forward modelling in RAYINVR (a).

(a) The colour wash represents the velocity model developed in FAST, the dashed black lines are the layer boundaries from preliminary RAYINVR modelling.

(b) Starting model used for the model development in RAYINVR.
Table 2.2: Model statistics for the degree of fit between the final P-wave velocity model and the recorded data. \( n \) is the number of arrival picks met by rays propagated through the model and the percentage of the picked data this represents, \( T_{\text{rms}} \) is the average root-mean-square travel time residual. \( \chi^2 \) is a measure of the level to which the model fits the data, see text for details. Statistics are given for the final model of the MONALISA Profile 3 data, after inversion.

\( t(i)_{\text{obs}} \) is the recorded travel time and \( n \) is the total number of data points (RAYINVR code: Zelt and Smith, 1992). A \( \chi^2 \) value of one indicates that the travel time residual equals the uncertainty, and as such the model fits the data. A \( \chi^2 \) value of greater than one indicates that the travel time residual is greater than the pick uncertainty. Therefore, the model does not fit the data adequately. A value of less than one occurs when the travel time misfit is less than the uncertainty, indicating that the model 'over fits' the data. In this case the calculated arrival times fit the picks to greater precision than the uncertainty in the arrival times. Therefore, variations in arrival time that fall within pick uncertainty are being reproduced by the calculated arrivals, so the model is fitting variations that may well be noise. The travel time residual is recorded as the averaged root-mean-square travel time misfit of all the rays traced. Reducing the misfit is the primary objective of the modelling. The modelling statistics, relating the observed travel times to the MONALISA model, are given in Table 2.2. As a visual display of the similarity between the modelled and calculated travel times the receiver gathers for OBHs 32, 36 and 38 (receivers in the
Figure 2.7: Trace normalised amplitude receiver gather for OBH32. Plotted with a 6 km/s reduction velocity and with the picked arrivals and calculated travel times for the final model overlaid.

eastern, central and western portions of the data) are shown with the calculated travel times overlaid (Figs. 2.7–2.9). Similar, but larger sized plots for the all 8 OBHs are given in Appendix B. The velocity structure is only constrained in regions sampled by the recorded arrivals and is only well constrained in the regions sampled by crossing raypaths; however, it is not clear from the receiver gathers which regions of the model are sampled and which regions are not. Therefore, it is standard practice to either provide a ray path diagram to illustrate the parts that are constrained by ray coverage, or to mask out the regions of the model that are not sampled. A raypath diagram for the MONALISA data is shown in Figure 2.10. The final model is shown in Figure 2.11.

**Additional modelling**

Once the travel time modelling has produced a good fit to the data it is common to use additional information from seismic arrival amplitude and gravity modelling to refine the model (e.g. Barton and Wood, 1984; EUGENO-S working group, 1988; Morgan et al., 1989; Richardson, 1997; Raum et al., 2002). The amplitude and gravity data provide an emphasis on different elements of the crustal properties to the travel time data. The amplitude of the seismic arrivals has a far greater sensitivity (than the travel time data) to the velocity gradients in the crust. The gravity data can be used as supporting evidence for
Figure 2.8: Trace normalised amplitude receiver gather for OBH36. Plotted with a 6 km/s reduction velocity and with the picked arrivals and calculated travel times for the final model overlaid.

Figure 2.9: Trace normalised amplitude receiver gather for OBH38. Plotted with a 6 km/s reduction velocity and with the picked arrivals and calculated travel times for the final model overlaid.
features in the velocity model by converting the model to a density model using an assumed relationship for the two properties. The non-uniqueness of gravity models and the relatively weakly constrained velocity-density relationship (see Chapter 4 for discussion) limits the usefulness of the gravity data in investigating the crustal velocity model. However, the long-wavelength regional features tend to be less flexible in their structure and so the gravity modelling can be a useful constraint on the coarse features of the model, such as Moho structure.

Amplitude modelling is generally used to provide additional constraints on the velocity gradients. However, the recorded amplitudes are also sensitive to attenuation, geometrical spreading and local, small-scale structure through seismic wave focusing and scattering by relief on velocity discontinuities. The geometrical spreading and attenuation can be estimated during the modelling. However, as crustal velocity models do not contain small-scale structure the modelled amplitudes do not fully reproduce the scattering and focusing effects of such structure. To minimise the effects of the sub-model-scale structure the modelling of the MONALISA data has been restricted to qualitative comparison of the observed and modelled amplitudes using synthetic seismograms. To generate the synthetic seismograms the mean amplitude in a 250 ms window following the calculated arrival time is determined for each phase and then convolved with the apparent source function (Zelt and Forsyth, 1994). The amplitudes were calculated by forward modelling with TRAMP.
Figure 2.11: Velocity structure of the final model. Solid lines indicate layer boundaries associated with first order velocity discontinuities and dashed lines second order discontinuities.
(Zelt and Ellis, 1988; Zelt and Forsyth, 1994) using the same model parameterisation as the travel time models. A description of the TRAMP code is provided in Appendix A.3. The observed and synthetic receiver gathers are shown for OBH33 as this receiver recorded arrivals of typical amplitudes for each of the modelled layers (Fig. 2.12). The modelled amplitudes reproduce the general trends in amplitude variation with offset and relative amplitudes between the different phases, providing support for the high velocity gradients in the near surface and the lower gradients deeper in the crust.

The gravity data were forward modelled in the 2½-dimensional gravity and magnetic modelling package GRAVMAG (Pedley et al., 1993). An explanation of the code is provided in Appendix A.4. The standard approach when using gravity data to assist in constraining velocity models is to convert the velocities to density using an empirically derived relationship. The density model for the MONALISA data was created by breaking the velocity model into polygons along isovelocity contours, with each polygon representing a 0.25 km/s range in velocity. The polygons were produced from closely spaced velocity contours rather than layer boundaries so that the lateral velocity gradients were reproduced in the density model. To remove any edge effects from the ends of the modelled bodies, the profile was extended 1000 km in both directions and the polygons were extended 1000 km out of the plane of the model. The mean velocity of each polygon was then used to set the density. Three velocity-density functions were used. The first represents the best-fit relationship through the most appropriate well and laboratory data; the Hughes et al. (1998) sedimentary data from the Faroe-Shetland Channel for velocities up to ~4 km/s and the Christensen and Mooney (1995) data for continental crust at higher velocities. This relationship is given by:

\[ \rho = -0.1264 + 1.8731v_p - 0.5036v_p^2 + 0.0583v_p^3 - 0.0022v_p^4 \]  (2.9)

where \( \rho \) is the density (g/cm³) and \( v_p \) is the P-wave velocity (km/s) (Fig. 2.13). Figure 2.14 shows the density model. The other two relationships represent the approximate envelope of experimental data given by Ludwig et al. (1970), widened a little at the high velocity-density values to incorporate the Christensen and Mooney (1995) data. These two relationships produce the minimum possible density structure and the maximum possible
(a) True amplitude receiver gather for OBH33, with every 10th shot plotted

(b) Calculated synthetic amplitude section for the final velocity model

Figure 2.12:
2. Modelling Wide-Angle Reflection Data

Figure 2.13: The velocity density function used to convert the MONALISA velocity model to densities and the published relationships from Hughes et al. (1998) for the Faroe-Shetland Channel, Christensen and Mooney (1995) for 10 km depth and 50 km depth and the envelope of empirical data given by Barton (1986).

Figure 2.14: The density model constructed from the final P-wave velocity model using the preferred velocity-density relationship.
Two sets of Free Air gravity data are available for the MONALISA profile; Sandwell and Smith (1997) satellite data and British Geological Survey ship-borne data compilation (Banka et al., 2002). The two profiles are similar, the satellite data generally has slightly larger amplitude anomalies than the ship data, but the difference is always less than 10 mGal and generally less than 5 mGal. Both profiles show a gentle gradient along their length, with the anomaly increasing from approximately -20 mGal in the west to +20 mGal in the east. Superimposed on this gradient are short-wavelength anomalies up to approximately 25 mGal in size. The gravity anomalies predicted by the travel time derived crustal model are compared to this observed data in Figure 2.15.

The first observation is that the minimum and maximum of possible velocity-density relationships derived from experimental data generate a range in calculated gravity anomaly far greater than the observed anomaly. This is generally the case for this approach to gravity modelling (Barton, 1986). It implies that the uncertainties in the conversion from velocity to density are too large to be able to gain much information on the crustal structure through this approach. However this is not necessarily the case. If the results from a constant velocity-density relationship are compared to the observed anomaly then some observation about the crustal structure can be made.

Using the best-fit relationship for the MONALISA data it can be seen that the calculated gravity anomaly has a significantly greater gradient (from negative values in the west
to positive values in the east) than the observed data. This misfit is created by the lateral velocity gradient in the lower crust. This lateral velocity gradient is a robust feature, required by the travel time data. No model could be generated that produced an acceptable fit to the travel times without the lower crustal velocity gradient. Therefore, the misfit in the gravity data can only be reduced by employing a laterally varying velocity-density relationship. This is the solution employed by Nielsen et al. (2000) (Fig. 2.16) for this data and also for the wide-angle profile across the Central North Sea (Barton and Wood, 1984).

Such a change in the velocity-density relationship suggests that the composition of the crust is different on the eastern side of the Central Graben to the western side. The Thor and Iapetus sutures may be approximately coincident with the Central Graben in these two regions; therefore the change in velocity-density relationship across the Central Graben may be related to the difference between the crust of the Baltic Shield and the Caledonian crust of Avalonia/Laurentia (Abramovitz and Thybo, 1998, 2000; Nielsen et al., 2000).

**Summary of model development**

The final model for the MONALISA Profile 3 data is shown in Figure 2.11 and the statistics for the travel time modelling summarised in Table 2.2. The modelling of these data revealed several problems that are not uncommon for wide-angle datasets.

For the near surface layers the wide spacing of the receivers (in the case of marine surveys, or shots for land data) results in unreversed ray coverage for diving waves turning in the top few kilometers (3–4 km for MONALISA Profile 3). As a result the sedimentary
velocity structure is very poorly constrained. Therefore, if the model is based solely on the wide-angle data (as the MONALISA model is) then only a very simple structure can be justified in these near surface layers. In recent years larger numbers of OBS have been used in marine survey greatly reducing this problem (e.g. the RAPIDS3 profiles acquired in the southern Rockall Trough in 1999 have an OBS spacing of between 3.67 and 4.75 km and provide excellent constraint on the sediments (Mackenzie et al., 2002)). With older data, coincident normal incident data (if available) is often used to help constrain the shallow layers; either by a-priori constraint (e.g. Horsefield et al., 1994; Berndt et al., 2001) or by including the reflection data in the modelling/inversion process (e.g. Mjelde et al., 1997; McCaughey et al., 2000).

The significant acoustic impedance at the base of the crust normally results in strong PmP reflections. These reflections provide a significant amount of data and therefore generally allow the Moho to be modelled with reasonable constraint. However, as the uppermost mantle generally only has a low velocity gradient the Pn diving rays normally have a low amplitude. As a result it is common to lose the Pn arrival in the background noise; often Pn is only recorded on a small number of the OBHs (or shots for land data). The MONALISA data is no exception; the arrivals turning within the mantle (Pn) are few in number and so can only very poorly constrain the mantle velocity structure. However, the mantle immediately below the Moho has been modelled with a simple velocity gradient defined by just two nodes. Modelling suggests that cycle skipping is a particular problem for these arrivals as it is not possible to match the Pn picks without introducing a large low velocity zone beneath the Moho and then a very high vertical velocity gradient within the mantle. Such geologically unlikely features have been avoided by matching the overall shape of the Pn phase, but allowing the ray traced arrivals to occur earlier than the picks from OBH32, resulting in a poor match to the data but producing geologically plausible velocities.

2.2.4 Model investigation and assessment

The greatest advantage to using inversion in modern methods is not in refining the velocity models. Generally the data distribution is such that the inversion is only stable once a fairly accurate forward model has been produced. Instead the principle advantages to
using inversion lie in assessing the resolution of the model features. Inversion allows the simple and rapid calculation of several indirect assessment methods (i.e. approaches that evaluate the model reliability without producing alternative models (Zelt, 1999)). The use of an inversion routine also greatly reduces the time taken to derive good accurate models, allowing easier direct model assessment (i.e. approaches that evaluate the model through the generation of alternative models (Zelt, 1999)).

Not all indirect assessment techniques require inversion of the model. Techniques generally presented in publications of wide-angle models and that do not require the model to have been inverted are:

- **Modelling statistics**: These are normally presented in tabular form and contain the number of picks, the $\chi^2$ and rms travel time residual for each phase (e.g. Table 2.2).

- **Plots of calculated travel times**: It is standard to produce plots either comparing the calculated travel times to the picked times or overlaying the calculated travel times on the receiver (or shot) gathers (e.g. Figures 2.7–2.9). Such plots aid assessment of the level of fit between the data and the model.

- **Ray coverage**: Usually presented as a ray path diagram (e.g. Figure 2.10). Knowledge of the sampled regions of the velocity model assists in qualitatively assessing the model constraint.

- **Plots showing the results of amplitude modelling**: Most commonly these are shown as synthetic receiver (or shot) gathers (e.g. Figure 2.12). Zelt (1999) advocates showing plots of the observed and calculated amplitude distance curves (e.g. Zelt and Forsyth, 1994).

- **Plots showing the results of gravity modelling**: Normally presented as two plots; one showing the density model (e.g. Fig. 2.14) and one comparing the calculated and observed gravity anomalies (e.g. Fig. 2.15).

The most common indirect assessment that does require the model to be generated by an inversion routine is parameter resolution. The parameter resolution is estimated from the diagonal values of the resolution matrix during the inversion of the model. The values
in the resolution matrix range from zero to one for each parameter. The resolution matrix \( R \) is given by:

\[
R = \left( A^T C_t^{-1} A + D C_m^{-1} \right)^{-1} A^T C_t^{-1} A
\]  

(2.10)

Where \( A \) is the partial derivatives matrix, \( C_t \) is the estimated data covariance matrix (which is a vector of the square of the travel time uncertainties associated with each pick), \( C_m \) is the model covariance matrix (a vector of the parameter uncertainties squared) and \( D \) is the damping factor (Zelt and Smith, 1992).

The resolution of a parameter can range from 0 to 1 and a node with a resolution of 0.5 or greater is considered well resolved and reliable (Zelt and Smith, 1992). The resolution of the nodes in the MONALISA model are shown in Figure 2.17. However, this is effectively a measure of the number of rays that sample each node, and as such is more a measure of precision than of accuracy, i.e. a high value indicates that the minimum in the objective function is well constrained for that parameter, but provides no information on the shape of the objective function or the possible presence of other minima in the function.

Other less common indirect assessment methods are single parameter resolution tests and calculation of standard errors associated with nodes. Single parameter resolution tests estimate the spatial resolution of the model with respect to a specific node (Zelt and Smith, 1992; Zelt, 1999). To perform such a test a node is chosen and the value perturbed so as to produce a travel time anomaly that is significant compared with the pick uncertainties. Rays are then traced through the perturbed model and the calculated travel times saved. The perturbed model is then replaced with the original model and the travel time picks are replaced with the saved calculated travel times. The model is then inverted using all the same parameters that were involved in optimising the perturbed node during the production of the final model. The spatial resolution of the model around the perturbed node is indicated by the number of nodes adjacent to the perturbed node that change during the inversion. If the model is well resolved then the inversion should only alter the perturbed node, if the model is poorly resolved then the inversion will smear the perturbation into the surrounding nodes.

The standard error is the square root of the diagonal elements of the covariance matrix, \( C \), which can be calculated retrospectively from the resolution matrix and data covariance.
Figure 2.17: Resolution of model nodes determined during inversion for the final model
matrix by:

\[ C = (I - R)C_m \quad \text{(2.11)} \]

This calculation gives an indication of the uncertainty on the node; however the value is the lower bound of the possible errors (Zelt and Smith, 1992) and is taken as the relative uncertainty of model nodes as there are many sources of error not taken into account by this calculation, such as (Zelt and Smith, 1992):

- trade-off between model parameters
- phase mis-identification
- inappropriate model parameterisation (unrealistic nature of minimum parameter model)
- modelling the 3D Earth in 2D

The standard errors were calculated for the MONALISA model and converted into more realistic uncertainties by scaling the errors with results of single parameter perturbation tests. The perturbation tests involved iteratively adjusting the velocity of a node, then ray tracing through the model and calculating the \( \chi^2 \) and number of data points traced. The perturbed models were compared to the original model using an F-test (on the \( \chi^2 \) values) to determine the possible change in the velocity of the node before the model degradation became statistically significant. The results of the single parameter perturbation tests were then used to scale the standard errors, producing a more realistic estimate of the velocity uncertainty (Fig. 2.18). The uncertainties are greatest in the regions with the poorest ray coverage (see Fig. 2.10). The uncertainties appear to be less than \( \pm 0.1 \text{ km/s} \) for a significant proportion of the crust, however this is a misleading result generated by the single parameter resolution tests. The values of the two nodes on the 10 km deep boundary at 150 and 381 km model distance, are strongly controlled by the values of the velocity nodes at the top of the layer above (see Fig. 2.17 for the node distribution). So, although the nodes can only be altered a tiny amount in single parameter tests, in fact the uncertainty in this region cannot be less than that of the nodes in the crust above. Therefore the uncertainty values of less than \( \pm 0.2 \) are unrealistic (as they do not include the effect of the strong coupling to the nodes above them). A more realistic uncertainty is
obtained if the region of less than ±0.2 km/s is instead taken to be ±0.2 km/s (i.e. similar to the nodes to which they are strongly coupled).

The single parameter perturbation tests used to scale the standard errors represent an example of direct model assessment. A direct assessment technique that is commonly used is calculation of the maximum perturbation that is possible for the crustal thickness (Moho depth). The depth to the Moho in the MONALISA model was assessed through a layer thickness perturbation test; shifting the Moho vertically and re-inverting the lower crustal velocity nodes, then measuring the misfit of PmP. The results show that the crustal thickness is fairly well constrained. The Moho depth can only be altered by ±1 km or -0.7 km before the fit to the data is significantly reduced. The test also indicates that the lower crustal velocity structure is fairly insensitive to the thickness of the crust with variation remaining within the ±0.2 km/s uncertainty (Fig. 2.19).

Additional model assessment comes from comparing the wide-angle model with other datasets sampling the same region. If it is available and has not been used either to construct the starting model or in the inversion, then coincident near normal incidence
Figure 2.19: Constraint on Moho depth. The top graph shows the $\chi^2$ value obtained from an optimised model with the Moho fixed to the value on the x-axis. The best fit model is obtained with the Moho at 23 km depth (under the Central Graben). At 22.3 km and 24 km the reduction in fit to the data become statistically significant. The lower figure shows the velocity obtained for each of the 3 lower crustal velocity nodes for the optimised model with various Moho depths. This shows that a deeper Moho requires a faster crust; but that the effect on the velocity is relatively small.
reflection data provide a useful check on the model features. A comparison of the wide-angle model velocity discontinuities from the MONALISA data with the reflections in the normal incidence data is shown in Figure 2.20. Although the wide-angle model will not resolve all the details of the structure shown in the reflection profile, the two methods may be expected to agree on the coarser structure. However, it has been noted that often the depth to Moho on wide-angle models does not tie in with the depth from normal incidence reflection data (Jones et al., 1996). The two most likely causes for a mismatch between the wide-angle Moho depth and the normal incidence Moho depth are (Jones et al., 1996); 1) anisotropy (as wide-angle reflections are significantly closer to horizontal than normal incidence reflections, velocity anisotropy could cause significant differences in the estimated Moho depth) or 2) normal incidence and wide-angle reflections originating from separate events near the base of the crust.

If the modelled data intersects with other wide-angle surveys then the resulting models should agree. In the case of the MONALISA Profile 3 data, the survey intersects Profiles 1 and 2 from the same project (Fig. 2.2). The other MONALISA wide-angle models, Profiles 1 and 2 and the Nielsen et al. (2000) model of Profile 3 can therefore be used to assess the final model presented here. The Nielsen et al. model (Fig. 2.16) was produced using joint inversion of the travel time and gravity data. Their model and the model presented here are in general in good agreement. However, the Nielsen et al. (2000) model is notably more complex; the authors of this earlier model assign smaller uncertainties to the picks, allowing them to retrieve greater detail from the data. The model is compared to the Profile 1 and 2 models (Abramovitz and Thybo, 1998, 2000) by plotting 1D velocity-depth profiles at the intersection of the lines 322 km and 283 km respectively (Fig. 2.21). For the majority of the crust the 1D profiles from MONALISA 1 and 2 fall within the uncertainty on the velocity structure in MONALISA 3; the two zones for which this does not apply are the basement structure and the Moho. The mismatch at the basement level suggests that this feature is not well constrained, an observation that is supported by the travel time modelling, where it was found that considerable alteration could be made to the basement layer without significantly altering the travel time residuals. The model presented here is the simplest model that provided a good fit to the travel time and amplitude data. The misfit at the Moho, while greater than the uncertainty on the MONALISA 3 model, may be
Figure 2.20: Normal incidence reflection profile along MONALISA 3 and the wide-angle model layers converted to two-way-travel time, given by the black lines.
Figure 2.21: Velocity-depth profiles through MONALISA profiles 1, 2 and 3 at their intersections.
within error if there are similar uncertainties associated with the Moho in the MONALISA 1 and 2 models.

2.3 Summary

Models of refraction/wide-angle reflection seismic data have progressed from simple velocity models of horizontal, constant velocity layers to sophisticated 2D or 3D inhomogeneous models based on ray tracing and tomography.

As an example of the modern modelling techniques a velocity model has been constructed from wide-angle data that crosses the southern North Sea. The model was produced primarily through travel time modelling using RAYINVR, with additional investigation and support for the model gained through amplitude and gravity analysis and comparison with the coincident normal incident reflection section. This modelling procedure represents a standard, current approach to modelling such data.

The example model indicates that the crustal structure resolved by modern methods is reasonably robust; the model agrees with other wide-angle models, data from normal incidence reflection and gravity modelling. Generally the uncertainty in the crustal velocity structure (in regions constrained by ray coverage) is less than 0.5 km/s. The regions of wide-angle models that are likely to be more poorly constrained are the top few kilometers beneath the surface (where there will be a shortage of crossing ray paths from turning rays) and the base of the crust (the deepest turning rays rarely sample the lower crust). The uncertainty on the crustal thickness/Moho depth is likely to be in the order of 1–2 km.
Chapter 3

Velocity Model

The model described in this chapter has a finite-element structure, with cells representing the P-wave velocity between the land surface (or sea bed offshore) and the base of the crust. The base of the model is assumed to represent the change from crustal to mantle lithologies. However, as there is no method of remotely identifying the petrological base of the crust the seismic Moho is used as a proxy. The Moho depths and velocity structure of the model were interpolated from wide-angle seismic data. For this work the Moho is defined using the surface interpreted, by the authors of the original wide-angle model, as the base of the crust rather than a specific velocity. This definition is particularly pertinent for several of the wide-angle models where the authors have interpreted velocities below 7.8 km/s as mantle material (e.g. RAPIDS2: Hauser et al., 1995) or velocities greater than 7.8 km/s as crustal material (e.g. CSSP: Al Kindi, 2002).

3.1 Seismic Data

The wide-angle profiles used to construct the velocity model were taken from a variety of published and unpublished sources, maps showing the distribution of the velocity and Moho data are shown in Figure 3.1 and a full catalogue of the models used is given in Appendix C. One-dimensional velocity–depth profiles and Moho depths were digitised at 5 km intervals along the length of the seismic models. The 5 km spacing retains all the significant features in the models as the horizontal resolution/parameterisation of the wide-angle models is rarely greater than this. Velocity values were recorded at the exact
3. Velocity Model

3.1 Data uncertainties

As well as recording the given velocity and Moho depth data an uncertainty was assigned to each parameter. Where possible the uncertainty was taken from the published source of the model; however, it is relatively unusual for comprehensive uncertainty information to be given.

Where published uncertainty information was not available a semi-quantitative estimate of the value was made. A series of points concerning the data and modelling method were qualitatively considered, then a numerical uncertainty assigned to the velocity and Moho values. These points were as follows:

1. **Modelling method and use of amplitude data.** In a qualitative sense models constructed using ray tracing methods (e.g. RAYINVR) or methods based on travel-time tomography (e.g. FAST) were considered "good". Modelling using time-term methods, $T^2 - X^2$ and other miscellaneous methods were considered "poor". The simple fitting of constant velocity layers on the basis of the gradient and intercept time of arrivals on record sections was considered "very poor" (irrespective of whether this was done by eye or using least squares optimization). This qualitative assessment included consideration of whether the modelling was one- or two-dimensional, with 1D methods considered significantly poorer than 2D methods. The use of amplitude

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(a) Location of crustal velocity data  
(b) Location of Moho depth data

Figure 3.1: Distribution of wide-angle seismic reflection data.

depth of nodes or boundaries in the published profiles.
information was considered to significantly reduce the velocity uncertainty in the lower crust.

2. **Data coverage.** The density of data coverage (i.e. spacing of receivers and shots) and whether the survey had reversed coverage were considered almost as important to the uncertainty as the modelling method. The coverage was generally assessed through published ray path diagrams. Data coverage was considered particularly important for assessing Moho depth uncertainty, as it is not uncommon for the Moho reflections (PmP) to be concentrated into isolated groups, with large sections of the model unsampled, especially at the ends of the profiles. The quality of the data in terms of signal to noise ratio was also examined, although not considered as significant.

3. **Inversion of the model.** Whether the modelling approach used inversion methods or only forward modelling was taken into account, but considered significantly less important than the previous points. Inversion of the model is not considered particularly important as in the vast majority of cases it is only used at the later stages of modelling to refine the forward model and produce quantitative estimates of uncertainty.

4. **Gravity data.** The use of gravity data was considered to improve the uncertainty on the depth to the Moho, particularly in regions poorly constrained by PmP reflections, but to be very much secondary to the modelling method used and data coverage.

5. **Normal incidence reflection data.** Coincident normal incidence reflection surveys have been used in a number of ways: 1) To construct a starting model; 2) to constrain the sediment geometries and velocity structure; 3) used directly as additional data in the modelling and 4) as an independent source of data to assess the final model. When used in either the first or the second approach, the normal incidence data is considered to help produce a more accurate final model, but to have little effect on the constraint/uncertainty of that model. In the third method the data is considered during the assessment of data coverage (point 2) and therefore not included here. If the fourth approach is used, the data has no effect on the velocity constraint.
3. Velocity Model

Uncertainties have generally been assigned as percentages of the velocity, but representative uncertainties for the top of the upper crust and the base of the lower crust were also recorded, to help maintain consistency between profiles. Moho depth uncertainty was recorded as a value in kilometers.

The two most common modelling methods are ray tracing using **RAYINVr** and time-term analysis. Typical errors assigned to a **RAYINVr** model with good data coverage (e.g. receivers every 30–60 km and dense coverage of airgun shots) and consideration of amplitude data would be ±3% (approximately equivalent to ±0.2 km/s) for the upper crust and ±5% (~ ±0.35 km/s) at the base of the crust (see Section 2.2.4). The models based on time-term analysis are generally older than the ray traced models and so have poorer data coverage and amplitude data is not used. As a result such models are typically assigned errors of ±7% (approximately equivalent to ±0.3 km/s) for the upper crust increasing to ±10% (~ ±0.6 km/s), at the base of the crust. A full catalogue of the assigned uncertainties is given in Appendix D.

The depths to mid-crustal interfaces were not assigned uncertainties as this information is largely redundant. In the case of first order discontinuities, the uncertainty on mid-crustal interfaces is related to the velocity step across the interface. Interfaces associated with large velocity discontinuities are generally well constrained. Only those interfaces associated with small velocity steps (or highly uncertain velocities) are poorly constrained. Therefore, where interfaces are poorly constrained the velocity step across the interface is generally much smaller than the uncertainties in the velocity values either side. As a result the discontinuity is largely masked by the velocity uncertainties and the uncertainty on its depth becomes fairly irrelevant. Figure 3.2 shows an example of this; the two lowermost sedimentary layers have well constrained velocities and the step in velocity between them is well outside of the velocity uncertainty, consequently the depth to the interface between layers is very well constrained. Conversely, the low velocity zone in the middle of the sediments is poorly constrained, with a large range in possible velocity values, which significantly overlap with the sediments above. Therefore, although the depth of the top of the low velocity zone is poorly constrained the large errors associated with the velocities will overshadow the depth uncertainties and ignoring this uncertainty is unlikely to have a profound effect on the final model.
Figure 3.2: Probability density function of velocity with depth and time for ESP 89-1W in the Rockall Trough (Pearse, 2002)

The geographical location of the data is also associated with uncertainty. In general the position of the published velocity models was measured from a location map given in the original publication. Therefore, the accuracy of the data location depends on the detail and scale of this map, the projection used (locations can be measured with far greater accuracy from a Mercator projection than other geographical projections with curved lines of longitude or latitude) and the proximity of the profile to distinctive features (data located close to coastlines or distinctive bathymetric features are more easily located than those in relatively featureless regions such as the North Sea). When digitising the data an estimate of the location error was made by assuming that the data could be located to within 1 mm on the map and this distance converted to a distance in minutes. The mean location error is ~2 minutes and the maximum is for the Rockall Bank Profile (Morgan et al., 1989), which has a large-scale map with little detail and therefore the error in location is ~10 minutes.
3. Velocity Model

3.1.2 Error checking

The majority of the data in the wide-angle database were digitised by hand; entering the velocities and depths into a spreadsheet manually. Therefore, the likelihood of typographical mistakes in input data is high. To combat this a careful error checking procedure was used to remove such mistakes.

The first stage of the error checking was to plot graphs of the depth and velocity of the data points during the digitising, these graphs immediately indicate any gross mistakes, such as a dropped decimal point.

The second stage was to run all the data points through a FORTRAN code to extract any points that failed the following tests:

1. The data were all checked to ensure that no points fell outside the region of possible values:
   - The depths of the velocity data were checked to ensure that all were between 2 km above sea level and 50 km below sea level.
   - The velocities were checked to ensure that all values fell within the ranges of 1.5–8.8 km/s.

2. Each point was then compared to the adjacent points on the same input profile. If no adjacent node could be found at a similar depth with a similar velocity (to within 0.2 km/s of the node in question), or if the step in uncertainties between the adjacent nodes was greater than 0.1 km/s, then the location of the node was printed out and manually checked against the original models. Any areas of large velocity gradients or models with layers that pinch-out laterally failed this test and were checked against the published velocity models. If the lateral gradients/steps were a feature of the published model then the data point was accepted.

The digitised points form 1D velocity–depth profiles located at 5 km intervals along the published models. Linear interpolation was used to recreate these 1D profiles at 100 m depth resolution and a further visual inspection of the data was then undertaken. At least one 1D velocity–depth profile was printed out for each of the input models and compared to the original data. This provided both a check that the digitised model reproduced the
Figure 3.3: Example of the increase of velocity uncertainty needed at some intersections between profiles. For the COOLE–ICSSP/CSSP intersection the velocity uncertainties were increased by 0.05 km/s between 2.3 and 3.5 km depth. For the AMP-L–RAPIDS1 intersection the velocity uncertainties were increased by 0.3 km/s between 2.8 and 5.8 km depth.

The final stage in error checking was to compare the 1D velocity-depth profiles at locations where the published models intersect (or are within close proximity to one another). The database was searched for 1D profiles that are located within 3 minutes of one another. The 1D profiles were then compared to ensure that the velocities (and Moho depths) matched to within the assigned uncertainties. Any parts of profiles that did not match were reported (to allow manual checks against the published model) and the uncertainties increased to the point where the profiles were compatible. At intersections where an increase in the uncertainties was required the vast majority of the discrepancies were due to small differences in depths of mid-crustal interfaces or parameterisation of the model (Fig. 3.3). However, this error check did show that the NASP-BC line of Bott and Smith (1984) (which had originally been included in the database) is consistently different from more recent and better resolved models in the Faroe-Shetland Channel (CDP88,
3. Velocity Model

<table>
<thead>
<tr>
<th></th>
<th>Baltica Vp (km/s)</th>
<th>Avalonia Vp (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tertiary</td>
<td>1.8–1.9</td>
<td>1.8–1.9</td>
</tr>
<tr>
<td>Mesozoic</td>
<td>2.0–3.6</td>
<td>2.0–3.6</td>
</tr>
<tr>
<td>Palaeozoic</td>
<td>3.9–5.5</td>
<td>4.3–5.2</td>
</tr>
<tr>
<td>Upper crust</td>
<td>6.0–6.25</td>
<td>5.8–5.9</td>
</tr>
</tbody>
</table>

Table 3.1: Seismic velocity structure of the sedimentary rocks and upper crust of Baltica and Avalonia in the southern North Sea. Compiled from interpretation of commercial reflection profiles, the MONALISA reflection profiles, well logs, borehole information and seismic refraction data (Abramovitz and Thybo, 1998, 2000).

AMP-A, FLARE and AMP-D). As the region is well sampled, including the NASP-BC was considered to add nothing to the database and the data were removed.

3.2 Other Data

The top surface of the model is defined by the topography and bathymetry. The data used in the model were extracted from the Smith and Sandwell (1997) bathymetry and GTOPO30 topography.

The velocity structure of the model is, for the most part, defined by the velocities retrieved from wide-angle reflection seismic data. However, the majority of the surveys used to acquire wide-angle data have, historically, had either a shot (for marine data) or receiver (for land data) spacing in the 10's of kilometers. The coarse shot/receiver spacing results in poor resolution of the top few kilometers of the velocity model, as there are no crossing ray paths. This poor resolution presents a potential problem for much of northwest Europe as there are many regions where the top few kilometers contain significant velocity and density variations, associated with changes in physical properties of the sediments, or between the sedimentary rocks and the basement. For example, in the North Sea the velocity discontinuity between the pre-rift (Palaeozoic) sedimentary rocks and the basement is often in excess of 0.5 km/s (e.g. Table 3.1). The wide-angle seismic models often only reproduce the coarse, large-scale, features of sediments and base-sediment interface. However, the contrasting physical properties create significant steps in acoustic impedance. Thus, normal incidence reflection seismology is capable of producing detailed images of the sedimentary cover and there is abundant normal incidence data in the region. Therefore, rather than estimating the sediment structure from the wide-angle seismic data alone,
external sources of data and compilations using normal incidence data were used.

Adding this base-sediment interface creates a model with two separate volumes: The near surface sedimentary layer, built using data from external sources; and the main crystalline crustal layer built using wide-angle seismic data.

3.2.1 Base-sediment interface

The sources compiled to produce the surface were:

- **National Geophysical Data Center map.** The dataset that covers the greatest part of the model is the NGDC map of “Sediment Thickness in the World’s Oceans and Marginal Seas” (National Geophysical Data Center, 2004). This map was not used in the North Sea as the interface it follows is the base of the Mesozoic syn-rift sediments, failing to include the significant thickness of pre-rift Permian sediments in the region.

- **A compilation of BIRPS reflection data and wide-angle seismic data.** Depth information for the North Sea was taken from a compilation containing depth converted picks of the base-Permian from BIRPS reflection data and other wide-angle reflection/refraction seismic models (England, 2000, and Fig. 3.4). The data were interpolated using a tensioned minimum curvature algorithm (Smith and Wessel, 1990).

- **Voring Plateau wide-angle data.** Sediment thickness data for the Norwegian margin, in the region of the Voring Plateau, has been compiled from wide-angle seismic data (Mjelde et al., 1997, 1998; Berndt et al., 2001; Mjelde et al., 2001; Raum et al., 2002; Raum, 2003). These data were also interpolated using the Smith and Wessel (1990) minimum curvature algorithm.

- **BGS’ Variscan unconformity data.** For onshore Britain the Variscan unconformity is used to define the base of the sediments. A digital version of the Variscan unconformity was provided by the BGS; this data is also published in the Regional Guides (Chatin, 1961; Hains, 1969; George, 1970; Earp, 1971; Greig, 1971; Edmonds, 1975; Kent, 1980; Melville, 1982; Green, 1992; Sumbler, 1996; Aitkenhead, 2002).
• **Global 1° data.** In regions not covered by the datasets described above, the base of the sediments was taken from a 1° resolution, global map of sediment thickness (Laske and Masters, 1997). The only exceptions are Scandinavia, Scotland, Ireland and Brittany where the sediment thicknesses were set to zero as the 1° resolution results in artificial sediment thickness in regions of short-wavelength topographic change.

The completed surface and the coverage of each dataset are shown in Figure 3.5.

### 3.2.2 Sediment velocity data

A 1D profile of increasing velocity with depth was used to define the velocity structure of the sedimentary layer. This velocity–depth function was derived using the interval velocities from a number of normal incidence seismic reflection profiles. Velocities were taken from the selection of reflection lines given in Table 3.2. These data were chosen to cover each of the major sedimentary basins in the region. The velocity data were converted to a velocity–depth profile by calculating a power-law regression curve through the median value of velocity data (binned into 0.5 km depth intervals). An estimate of the uncertainties associated with the velocity of the sediments was acquired by fitting similar regression curves through the 5th and 95th percentiles of the binned data (Fig. 3.6). The
Figure 3.5: The surface defining the base-sediment interface and map indicating the areas of the surface covered by each dataset, see text for details.

<table>
<thead>
<tr>
<th>Profile</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>AMP line L</td>
<td>Rockall Trough</td>
</tr>
<tr>
<td>AMP line N</td>
<td>Porcupine Basin</td>
</tr>
<tr>
<td>FAST</td>
<td>Faroe-Shetland Trough</td>
</tr>
<tr>
<td>AMP line B</td>
<td>Faroe-Shetland Trough</td>
</tr>
<tr>
<td>AMP line C (the northern end)</td>
<td>northern North Sea/ Faroe-Shetland Trough</td>
</tr>
<tr>
<td>MONALISA 1</td>
<td>southern North Sea</td>
</tr>
<tr>
<td>MONALISA 3</td>
<td>southern North Sea</td>
</tr>
<tr>
<td>SWAT-4</td>
<td>Celtic Sea</td>
</tr>
<tr>
<td>SWAT-5</td>
<td>Celtic Sea</td>
</tr>
</tbody>
</table>

Table 3.2: seismic surveys used to construct the sediment velocity-density function.
3. Velocity Model

Figure 3.6: Interval velocities from 9 reflection profiles around the UK and velocity–depth relationships calculated from this data.

Equations that define the minimum, best-fit and maximum velocity values are:

\[
\begin{align*}
5^{th} \text{ percentile} & \quad v = 2.1648z^{0.2929} \\
\text{median} & \quad v = 2.9091z^{0.2255} \\
95^{th} \text{ percentile} & \quad v = 4.8018z^{0.0584}
\end{align*}
\]

where \( v \) is velocity (km/s) and \( z \) is the depth below the surface/sea bed (km).

3.3 Model Construction

The model is defined using cartesian coordinates, with distances measured in kilometers, based on a Transverse Mercator projection centered at 3.4° west, 57.15° north.

The Moho surface and velocity structure of the crystalline crust were constructed by interpolation of the digitised wide-angle seismic data. The interpolation method used was ordinary kriging (in 2D for the Moho data and 3D for the velocity data), using the Deutsch and Journal (1998) code KT3D.

3.3.1 Kriging

Kriging is an interpolation technique that utilizes knowledge of the spatial continuity of a variable to estimate its value away from data points. Any variable that is continuous from one place to another (e.g. the Moho) must be spatially correlated over short distances
(although it may not be over longer distances). Kriging assumes that the spatial correlation is known (in the form of the semivariogram or covariance) and uses this to weight the importance of local data points to estimate the value of the variable away from the known data. It is this use of the statistical model, based on the data, to produce the weights for the interpolation that makes kriging a superior technique compared to traditional methods, such as inverse distance interpolation, which use an arbitrary weighting function.

Ordinary kriging estimates the variable using:

$$\hat{Z}(x_0) = m \left( 1 - \sum_{i=1}^{k} \lambda_i \right) + \sum_{i=1}^{k} \lambda_i Z(x_i)$$  \hspace{1cm} (3.4)$$

where $\hat{Z}(x_0)$ is the estimate at location $x_0$; $m$ is the mean of variable $Z(x)$; $\lambda_i$ is the kriging weight associated with data point $x_i$, one of the $k$ nearest points used in the estimation; and $Z(x_i)$ the value of the variable at measured location $x_i$ (Davis, 2002). By forcing the kriging weights to sum to unity the first term is removed and the kriging estimate becomes independent of the mean. The kriging weights are forced to equal one by adding a Lagrange multiplier ($\mu$) to the weights.

The weights are assigned to minimise the variance of the errors, $\hat{Z}(x_0) - Z(x_0)$. This is done using matrices of the weights ($A$), the covariance of the $k$ data points ($W$) and the covariance of the data points and the estimate location ($B$) (Davis, 2002):

$$\Lambda = W^{-1}B$$  \hspace{1cm} (3.5)$$

$$\Lambda = \begin{bmatrix} \lambda_1 \\ \lambda_2 \\ \vdots \\ \lambda_i \\ \mu \end{bmatrix}$$  \hspace{1cm} (3.6)$$
The covariances are provided by a variogram model designed to fit the data covariances. The variogram is a function recording the variation in spatial continuity with lag and is normally presented graphically. For the variable being estimated there will be a single "true" variogram, but as the variable is only known at the sample/data locations this true variogram is not known; therefore, a variogram model is generated and used to retrieve the covariances. Once the kriging weights have been calculated the estimate is found by vector multiplication of the weights by a vector of the data ($\mathbf{Y}$):

$$\hat{Z}(x_0) = \mathbf{Y}^T \mathbf{\Lambda}$$

where,

$$\mathbf{Y} = \begin{bmatrix} Z(x_1) \\ Z(x_2) \\ \vdots \\ Z(x_k) \\ 0 \end{bmatrix}$$

The estimation variance (which is used later to calculate the total uncertainty in the data) is given by:

$$\sigma^2(x_0) = \mathbf{B}^T \mathbf{\Lambda}$$
As the kriging weights are constructed using the variogram model of the data covariance, the derivation of this model is of primary importance to the accuracy of the interpolated surface.

3.3.2 Variogram modelling

The model variograms used for the estimation of the Moho and velocity data were constructed with consideration of experimental variograms produced from the input data. Experimental variograms are estimations of the true variogram produced by measuring the spatial continuity/variation of the sampled data. This approach assumes that the sampled data will represent a thorough enough subset of the variable that the experimental variogram will reproduce the same features (e.g. range, sill, nugget; see Figure 3.7) as the unrealizable “true” variogram. In many situations, including the data presented here, the samples are not randomly distributed, but situated in regions where some feature of interest is expected. Therefore, it is good practice to treat the experimental variograms with caution and produce a variogram model that is as basic as possible, and supported by all available a priori knowledge of the physical properties of the variable (Isaaks and Srivastava, 1989). The variogram is often directionally dependent and so is generally produced for a range of azimuths.
Experimental variograms were constructed (using the Deutsch and Journal (1998) code GAMV) at a range of orientations, with one hundred 10 km lags (with 5 km lag tolerance). The directional tolerance for the variograms was 22.5°, out to a maximum band width of 50 km (Fig. 3.8). A large number of experimental variograms were produced, using a range of directions and each of the 4 standard measures of spatial variability of a single variable. These measures of variability are:

1. **Semivariogram.** This is half the average squared difference between the value at two points \((x_i \text{ and } y_i)\) separated by lag \(h\), defined by:

\[
\gamma(h) = \frac{1}{2N(h)} \sum_{i=1}^{N(h)} (x_i - y_i)^2
\]

where \(N(h)\) is the number of pairs (Gooverts, 1997; Deutsch and Journal, 1998).

2. **Covariance.** This is a measure of the similarity of the data about their means,
defined by:

\[ C(h) = \frac{1}{N(h)} \sum_{i=1}^{N(h)} (x_i - m_{-h})(y_i - m_{+h}) \] (3.13)

where \( m_{-h} \) is the mean value of the \( x_i \) values and \( m_{+h} \) is the mean value of the \( y_i \) values. The semivariogram can be converted to the covariance using \( C(h) = c - \gamma(h) \) where \( c \) is a constant (Davis, 2002). The covariance is used in the calculation of the kriging estimates, in preference to the semivariance, as this guarantees no zeros in the matrix (Goovaerts, 1997; Deutsch and Journal, 1998).

3. Correlogram. This is a standardized version of the covariance, defined by:

\[ \rho(h) = \frac{C(h)}{\sigma_{-h}\sigma_{+h}} \] (3.14)

where \( \sigma_{-h} \) and \( \sigma_{+h} \) are the standard deviations of the \( x_i \) and \( y_i \) data (Goovaerts, 1997; Deutsch and Journal, 1998):

\[ \sigma_{-h}^2 = \frac{1}{N(h)} \sum_{i=1}^{N(h)} x_i^2 - m_{-h}^2 \quad \sigma_{+h}^2 = \frac{1}{N(h)} \sum_{i=1}^{N(h)} y_i^2 - m_{+h}^2 \] (3.15)

4. Semimadogram. This measure of variability is similar to the semivariogram, but takes the absolute (rather than squared) difference between point \( x_i \) and \( y_i \):

\[ \gamma_M(h) = \frac{1}{2N(h)} \sum_{i=1}^{N(h)} |x_i - y_i| \] (3.16)

Semimadograms are good indicators of large scale structure such as range and anisotropy, but do not indicate the appropriate nugget effect for kriging (Deutsch and Journal, 1998).

A selection of the experimental variograms are shown in Figure 3.9. The semimadograms are shown for 4 directions, the correlogram, covariance and semivariogram are only shown for the omni-directional experimental variogram.
Figure 3.9: Moho experimental variograms. For the semimadograms black indicates a sample direction of 000°, red indicates a direction of 045°, green indicates a direction of 090° and blue indicates 135°. The other variograms are omni-directional. The red curve on the semivariogram is the model variogram used to interpolate the Moho data.
Table 3.3: Ranges of directions and lags used to calculate the velocity data experimental variograms.

<table>
<thead>
<tr>
<th>Dip</th>
<th>Azimuths</th>
<th>Lag (km)</th>
<th>Lag tolerance (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°</td>
<td>000°-315° at 45° intervals</td>
<td>20</td>
<td>9.0</td>
</tr>
<tr>
<td>30°</td>
<td>000°-315° at 45° intervals</td>
<td>10</td>
<td>4.5</td>
</tr>
<tr>
<td>60°</td>
<td>000°-315° at 45° intervals</td>
<td>2</td>
<td>0.90</td>
</tr>
<tr>
<td>90°</td>
<td>-</td>
<td>1</td>
<td>0.45</td>
</tr>
</tbody>
</table>

Velocity experimental variograms

The use of three-dimensional kriging for the velocity data requires a three-dimensional variogram model. Therefore, experimental variograms were constructed using a number of lag directions and dips, as this would show any 3D anisotropy. The geometrical proportions of the input data are very dissimilar (with the horizontal extent of the data greater than 2000 km in both the x and y direction, but only ~50 km of data in the vertical direction) and so the lag distance is required to change with dip in order to retrieve sufficient data to constrain the variograms at all lags. A summary of the lag directions and dips used to produce the experimental variograms is given in Table 3.3. All variograms were calculated with a 12° angular tolerance and a maximum horizontal bandwidth of 10 km and vertical bandwidth of 1.5 km (see Fig. 3.8 for an illustration of these parameters). The experimental and model semivariograms are shown in Figure 3.10.

Variogram models

In fitting a model structure to the experimental variograms the following points were considered:

1. **Anisotropy.** For both data sets the experimental variograms show evidence for anisotropy. In the Moho data (Fig. 3.9) there is zonal anisotropy (i.e. direction dependent sill, but constant range) with a minimum sill in a NE or NNE direction. Such a trend is likely to be inherited from the relatively high sampling of the continental margin between Hatton Bank and the Lofoten Islands. Along this margin the topography and Moho depths show the rapid change in the NW direction associated with the transition from continent to ocean, but far greater continuity in the NE direction, parallel to the margin. It is also possible that the NE-NNE structural trend
Figure 3.10: Experimental and model semivariograms for the crystalline crust velocity data. The title of each plot gives the dip of the variogram and the key gives the azimuth. Note that the horizontal scales are different for the four plots.
generated during the Caledonian Orogeny has some residual signature affecting the Moho data away from the continental margin. However, as the anisotropy is probably restricted to the northwest continental margin, the risk of over-interpreting and introducing erroneous structure into the interpolated Moho surface was considered too high to include anisotropy in the model.

The velocity data (Fig. 3.10) show unquestionable dip-dependent geometric anisotropy (i.e. dip-dependent range) and zonal anisotropy (i.e. dip-dependent sill), with the horizontal variograms exhibiting greater ranges and lower sills than the dipping variograms. This is consistent with what is known of the crustal velocity structure from two-dimensional models. A vertical profile through the crust may well show increasing velocity from ~4 km/s to ~7 km/s over a few 10s of kilometers depth, whereas the horizontal variation may well be less than 1 km/s along a 2D profile several hundred kilometers in length. Therefore, the horizontal variation is expected to be both smaller in magnitude and spatially less rapidly changing than the vertical variation. There is no clear evidence for azimuth-dependent anisotropy. Therefore, the model variogram was constructed to reproduce the dip-dependent zonal and geometric anisotropy, but to be azimuthally isotropic.

2. Sill & Range. For the Moho data the experimental variograms show a well developed sill with a range of approximately 800 km.

For the velocity data the horizontal variograms show a well developed sill, with a value of ~0.5 km²/s⁴, with a range of approximately 150 km. The sill for the dipping variograms is not well developed, but is far higher than 0.5 km²/s⁴. The curvature of the dipping variograms is consistent with a model with a vertical range of 35 km and sill of 1.2 km²/s⁴.

3. Near-origin behaviour. Variables that are highly continuous over short distances, such as depth or layer thickness data, usually exhibit parabolic behaviour near the origin of the variogram. As adjacent points on a continuous surface will be at almost identical height the variability at short lags is very small. Such surfaces are often modelled with a Gaussian function (Fig. 3.11) to reproduce this continuity. However, there is no evidence for this behaviour in the Moho data, which are approximately
linear at the origin. It is highly likely that the experimental variograms are affected by the mismatches in the Moho depth at profile intersections, increasing the variability between closely spaced data (this is certainly the cause for the reduction in continuity at the shortest lag). Even if such effects are concealing what would otherwise be parabolic behaviour, with no data to constrain a parabolic curve it is unreasonable to try and fit a Gaussian model to the data. Instead a model with linear behaviour near the origin was considered more appropriate.

For all the velocity variograms the short lags show near linear behaviour. This is consistent with the observation that the velocity structure can contain discontinuities and rapid changes within the crust.

4. Structure at intermediate lags. A transitional model (i.e. including a sill) with near linear behaviour near the origin is required to fit points 2 and 3 above. The two most common transitional models are the spherical model and the exponential model (Fig. 3.11). The models are broadly similar, differing only in the rate of change. As the Moho experimental semivariogram and correlogram (Fig. 3.9) show reasonably linear behaviour at the intermediate lags the spherical model was preferred to the exponential model.

For the velocities the dipping variograms show reasonably smooth variation at intermediate lags. However, the horizontal variograms have a sharp change in gradient of the variogram at a lag of approximately 50 km. This sharp change is reproduced by adding a second horizontal structure with a short range to cause the initial rapid increase, but retaining the overall the horizontal range of 150 km.
Table 3.4: Structures of the variogram model for the velocity data.

<table>
<thead>
<tr>
<th>Structure</th>
<th>Range</th>
<th>Sill</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spherical</td>
<td>50 km horizontal range, 35 km vertical range</td>
<td>0.2 km²/s⁴</td>
</tr>
<tr>
<td>Spherical</td>
<td>150 km horizontal range, 35 km vertical range</td>
<td>0.3 km²/s⁴</td>
</tr>
<tr>
<td>Spherical</td>
<td>infinite horizontal range, 35 km vertical range</td>
<td>0.7 km²/s⁴</td>
</tr>
</tbody>
</table>

The final variogram structure chosen to model the Moho data was a single, isotropic spherical structure (Fig. 3.9). The spherical structure is defined by the following equation:

\[
\gamma(h) = c \cdot \text{Sph} \left( \frac{h}{a} \right) = \begin{cases} 
  c \cdot [1.5 \frac{h}{a} - 0.5 \left( \frac{h}{a} \right)^3] & \text{for } h \leq a, \\
  c & \text{for } h > a
\end{cases}
\]  

(3.17)

where \( \gamma(h) \) is the semivariance at distance \( h \), \( a \) is the range, \( c \) is the sill (Deutsch and Journal, 1998). The range of the model variogram is 800 km and the sill is 49 km². The velocity variogram model consists of three structures, see Table 3.4.

### 3.3.3 Model element size and search parameters

To decide the areal dimensions of the model elements the Moho data where interpolated onto a range of grids. Investigation of the model dimension was based on the Moho data, rather than the velocity data, in order to save computational time. Given the similarity between the Moho and velocity distributions, and that velocity data were interpolated with very little weight allocated to data at a different depths, using 2D data instead of 3D data has very little impact on the evaluation of model parameters. In order to assess the model quality the kriging uncertainty was recorded for the results of the interpolation onto each of a range of grid sizes. The preferred grid has dimensions that minimise the rms uncertainty (recorded as variance). The results of interpolation onto grids with cell sizes between 10 and 80 km, show that the kriging uncertainty is relatively insensitive to the model dimensions, but that 40 km cells minimise the variance (Fig. 3.12).

The vertical element size was set to reflect the balance between the desire to have fine spacing, to reproduce the vertical variation seen in the input data, and the need to have coarser spacing to allow for the poor depth resolution of velocity structure in the lower crust. The Moho uncertainties (Appendix D.1), which have an average of \( \sim \pm 2 \) km, give an indication of the depth resolution in the lower crust. However, given that the resolution
in the upper crust is significantly greater (and that lower crustal velocity gradients are generally small), a vertical element dimension of 1 km was considered the most appropriate for the crustal layer as a whole.

The search parameters used in the final model required a minimum of 25 and maximum of 64 data points, with a maximum number of data points per octant of 8. This set-up produces an estimation that, with the optimal distribution of data, is based on 8 data points in each octant, while the poorest acceptable data distribution has 8 data points in 3 octants and 1 data point in a fourth octant. Given that the data are strongly clustered, with samples every 5 km along the seismic profiles and that the mean distance between each profile and its nearest neighbour is 80 km, the restrictions on distribution of data within the octant search are needed to guarantee a reasonable geographical spread in the data used in the kriging. With a 5 km sampling interval on profiles the 8 data points within each octant are likely to come from a single profile; this will result in slight smoothing of the interpolation, predominantly over 40 km, the cell size of the final model. The maximum search radius was set to 800 km allowing Moho depths to be estimated at all constrained locations. The limits of the Moho estimation define the spatial extent of the model. The same search range was used for the velocity data, ensuring that all cells within the model would be assigned a Moho depth and velocity.

### 3.3.4 Combining the data

In order to build the complete crustal model the kriged velocity data was combined with the sediment velocity estimates. The interpolated Moho (Fig. 3.13) and filtered versions of the topography and base-sediment surfaces were used to define the top and base of the crystalline crust and sediments. The interface between the sediment and crystalline
layers in general falls within one of the model cells (rather than falling exactly at the cell boundary). Therefore, to reproduce the velocity structure as accurately as possible the cell containing the base-sediment interface is assigned a weighted average of the sediment and crystalline crust velocities. The weighting is controlled by the fraction of the cell containing sediment in relation to that containing crystalline crust.

3.3.5 Model uncertainties

An estimation of the total uncertainty associated with the interpolated data was achieved by combining the input data uncertainties with the kriging uncertainties. In order to do this the distribution function of the input data uncertainties had to be estimated.

When modelling wide-angle data an estimation of the parameter uncertainty may be obtained by varying the value of the parameter until there is a significant reduction in fit between the model and the observed data. The reduction in fit is usually measured through $\chi^2$, possibly in association with the travel-time residual and the ray coverage. An example of the effect on $\chi^2$ of changing the Moho depth is shown in Figure 2.19 and for perturbed velocity in Figure 3.14. Both figures show that the rate of change of $\chi^2$ increases as the
Figure 3.14: The effect on $\chi^2$ of perturbing a velocity node. The lines represent tests performed separately on 3 different velocity nodes. The data are based on synthetic travel-times calculated through a simple velocity model. The travel-times fit perfectly for the starting model, hence the initial $\chi^2$ of 0, and are assigned arbitrary small uncertainties and thus produce very high $\chi^2$ for small perturbations.

Perturbation increases. The parameter uncertainty is generally taken as the maximum perturbation that does not result in a $\chi^2$ value indicating significant degradation in the model. However, as modelling inherently contains some simplifying assumptions and as models are often non-unique, there is not zero probability that the true value lies outside this limit; therefore, it is more reasonable to take this assessment of uncertainty as the limit of likely values, rather than the absolute limit. The probability that the true value lies outside the assigned uncertainty, and the distribution of this probability, are impossible to directly assess. However, it seems reasonable to assume that the distribution tails off, with decreasing likelihood at increasing distance from the modelled value. So, given the shape of $\chi^2$ within the likely limits of uncertainty and the assumption of a tailing-off probability outside these limits, it seems reasonable to assume an approximately Gaussian distribution for the input data uncertainties, with the given uncertainty representing 2 standard deviations from the modelled value (i.e. there is $\sim$95% probability that the value falls within the given uncertainty).

This assumed Gaussian distribution allows combination of the kriging uncertainty and input data uncertainty using the Law of Propagation of Errors. In the case of independent variables such as these, this law simplifies to:

$$V(y) = \sum_{i=1}^{n} \sigma_i^2 \left( \frac{\delta y}{\delta x_i} \right)^2$$  \hspace{1cm} (3.18)

where $y$ is a function of variables $x_1$ to $x_n$, $V(y)$ is the variance of $y$, $\sigma_i^2$ is the variance of $x_i$
and $\frac{\partial y}{\partial x_i}$ is the partial derivative of function $y$ with respect to $x_i$. For the combination of the kriging and input uncertainties this becomes:

$$
\sigma_{mod} = \sqrt{\sigma_{input}^2 + \sigma_{krig}^2}
$$

where $\sigma_{mod}$ is the standard deviation in the model parameter (assumed to be half the total uncertainty). $\sigma_{input}^2$ and $\sigma_{krig}^2$ are the variances in the input depth/velocity data and kriging respectively.

The input variances were propagated through the model using kriging, with the same parameters as for the data. The kriging variance was taken directly from the output of KT3D for the velocity and Moho data.

Through the continued use of the assumption that the uncertainty has a Gaussian distribution, the Moho depth data was converted into a three-dimensional probability model of crustal thickness (using the interpolated Moho as the mean together with the calculated standard deviation of the uncertainty). Slices through this probability model of crustal thickness are shown in Figure 3.15. The probability model shows that the crustal thickness is well constrained near the input data. For example, comparing the slices through the model just to the north of Scotland and through southern Germany (Fig. 3.15), the likely range of crustal thickness is much smaller and the chance of the mean value being accurate is much greater for the region north of Scotland. The model also shows that major features such as the thinning under the Rockall and Faroe-Shetland Troughs are well within the resolving power of the data and without question required by the model. An alternative view of the Moho uncertainty is a map of the absolute value of uncertainty (Fig. 3.16).

The uncertainties in the velocity values are shown alongside the final velocity model, see Section 3.5.

### 3.4 Model Verification

To test the complete crustal model two verification procedures were used. The first test was to convert the velocities to densities and compare the gravity anomaly predicted by
Figure 3.15: Slices (outlined in yellow) through the 3D probability model of crustal thickness. The red line marks the extent of the model (vertical scale 15 times horizontal scale).
Figure 3.16: Magnitude of uncertainty in depth for the interpolated Moho surface.

the model's structure to the observed Free Air anomaly. The resulting density model is not only a verification tool for the velocity model, it is a useful tool in its own right. Therefore, the construction and features of this model are discussed in detail in the following chapter. However, two results of this modelling have been filtered back into the velocity model and therefore affect the final model presented in the next section. One of the results of the density model was to alter the Moho structure; therefore, this new Moho must be prematurely presented here (Fig. 3.17) before it is incorporated into the final model. The other change is to the model boundaries. The main area of interest in the model is Britain, Ireland and the surrounding marine sedimentary basins. The coverage of the dataset was extended into continental Europe in order to avoid interpolation errors at the edge of the model. However, the data in Scandinavia is not a comprehensive compilation of the available seismic models and the coverage in mainland Europe, from France to Germany, is very sparse. This results in poor constraint in these two regions, far weaker than in the main areas of interest. Therefore, the model boundaries have been adjusted to remove onshore Norway and Sweden and continental Europe between France and Germany. The Norwegian margin has been retained in the model as it has excellent data coverage and
the tectonic setting is an extension of the Hatton-Faroes margin.

The second verification method was to extract slices through the model coincident to the input wide-angle seismic models to ensure that the final model is faithful to the data. A selection of these cross-sections are shown in Figure 3.18. The fit between the wide-angle models and the final 3D model is generally very good; there is some loss of detail due to the 40 km element size and there are small regions of mismatch between the base of the sediments in the regional model and the base in the wide-angle models. However, all the significant features of the wide-angle models are shown.

3.5 Final Model

The final velocity model is shown in Figures 3.19 and 3.20.

For the top few kilometers of the model there is very strong lateral variation in velocity, with the sedimentary basins clearly marked out by large areas of relatively low velocity and the crystalline basement showing higher and more variable velocities (Fig. 3.19). The use of a single 1D velocity-depth profile to populate the cells in the sediments results in
Figure 3.18: Sections through the final velocity model along wide-angle seismic profiles. Black dots indicate digitised points in on the seismic model; the solid black line shows the Moho in the seismic model.
Figure 3.19: Depth slices through the final crustal velocity model. Pale grey = outside the model area, dark gray = either below the Moho or above the topography/seabed. Velocity contours are at 1 km/s intervals up to 5 km/s, and at 0.25 km/s intervals for higher velocities. Uncertainty contours are at 0.5 km/s intervals.
Figure 3.20: Depth slices through the final crustal velocity model. Pale grey = outside the model area, dark gray = either below the Moho or above the topography/seabed. Velocity contours are at 1 km/s intervals up to 5 km/s, and at 0.25 km/s intervals for higher velocities. Uncertainty contours are at 0.5 km/s intervals.
relatively constant velocity in the basins at any given depth. With the crystalline basement assigned velocities from the input seismic models, there is markedly greater variation in the velocities.

The uncertainty in the model is also very different between the sedimentary and crystalline regions. The cell for 0–1 km below the topography/sea bed in the sedimentary basins has a uniform uncertainty of 0.42 km/s while the velocity uncertainties in the crystalline crust are generally higher and much more varied. The maximum uncertainty in the top few kilometers occurs in central England (in the Pennines) in the model layer for 0–1 km above sea level, here velocity has been extrapolated vertically (very limited continuity) and from some distance, resulting in an uncertainty of 1.72 km/s. Below sea level the uncertainties in the crystalline crust are slightly lower as less horizontal interpolation is required: no cell has a velocity uncertainty in excess of 1.72 km/s until a depth of 17–18 km below sea level (the northern North Sea uncertainty reaches 1.73 km/s, see Fig. 3.20).

The velocity structure of the crystalline region of the near surface is well correlated with the surface geology. At a depth of 0–1 km below sea level the Carboniferous units in the Midland Valley and across Ireland are lower velocity than other areas of Britain (Fig. 3.19). The Welsh basement, the granites in SW England and northern Scottish Proterozoic regions show relatively high velocity and the Archean units of the Outer Isles have the fastest values, up to $6.58 \pm 0.70$ km/s. The correlation to surface geology reduces with increasing depth; however, some features are fairly persistent, such as the elevated velocities in SW England and Wales, which are still seen at 3–4 km below sea level, and the high-velocity Outer Isles crust, which is present (although at a reduced contrast against its surroundings) down to 6–7 km below sea level (Fig. 3.19).

The vertical velocity gradient in the crystalline crust is significantly smaller than in the sediments, as can be seen by comparing the velocity change from 1–2 km to 3–4 km below sea level in the sediments and basement (Fig. 3.19): a cell in the basement of central Ireland increases in velocity from $5.72 \pm 0.50$ to $5.94 \pm 0.52$ km/s over this depth range whereas as a cell in the central North Sea sedimentary basin increases from $2.17 \pm 0.42$ to $3.52 \pm 0.58$.

The uncertainty distribution rapidly becomes less segregated between sediments and crystalline crust with depth. By 3–4 km below sea level the better constrained regions of
crystalline crust, such as southern Ireland and proximal to the W-reflector profile (Morgan et al., 2000) off the north coast of Scotland, have similar constraint to the sedimentary basins (Fig. 3.19). The uncertainty distribution at this depth (3-4 km) is beginning to be dominated by the location of seismic data (Fig. 3.1(a)), with the most poorly constrained regions in SE England/southern North Sea and off the coast of Brittany. The southern North Sea/SE England region is surrounded by seismic profiles (LISPB (Barton, 1992), EUGEMI (Aichroth et al., 1992) and the French profile of Sapin and Prodehl (1973), see Figure C.1) but data are too widely spaced to provide constraint. Off the coast of Brittany the situation is even worse, the azimuthal coverage of the data is restricted by the proximity to the continental margin and the southern regions of the model are far more sparsely covered than the northern regions (Fig. 3.1(a) or C.1).

By 6-7 km below sea level the model is into the crystalline crust or near the base of the sediments for most of the area (Fig. 3.19). There is less lateral variation in velocity; the standard deviation in velocity is 1.30 km/s for the model layer at 3-4 km below sea level and 0.65 km/s for the layer at 6-7 km below sea level. However, lateral variation in velocity is still dominated by the sediment distribution with lower velocities in, or under, the thick sedimentary basins than in the regions where sediment cover is thin or absent.

Uncertainties in the model generally increase with increasing depth, the mean uncertainties at depths of 0–1, 3–4 and 6–7 km are 0.70, 1.04 and 1.20 km/s respectively. However, as the standard deviation in velocity at 6–7 km depth is 0.65 km/s, approximately 95% of velocity variation at this depth is within a 1.3 km/s velocity range. The bottom right image in Figure 3.19 shows the uncertainties at this depth with the 1 km/s contour marked (separating yellow from orange regions). From this figure it is clear that the constrained regions are limited to narrow strips in the model, which are coincident with the seismic profiles (Fig. 3.1(a)). Away from the seismic profiles the crustal model represents a best estimate of the velocity, rather than a constrained value. This does not mean that the velocity model is incorrect, simply that it is unconstrained.

As mentioned above, by 6–7 km much of the model is in the crystalline crust and the vertical variation in velocity is far smaller than the vertical variation in the sediments. Therefore, variation with depth for the model as a whole becomes less rapid after ~7 km;
for example, the change in mean velocity between 3-4 km and 6-7 km is 0.78 km/s (5.07-5.84 km/s) whereas the change in mean velocity from 6-7 km to 9-10 km is 0.32 km/s (5.84-6.16 km/s). Therefore the intervals between the depth slices shown in Figure 3.20 are greater than those in Figure 3.19.

At all depths below ~6-7 km the uncertainty distribution is a record of the seismic data coverage (Figs 3.20 and 3.1(a)) with uncertainties greater than 1 km/s in all locations except cells encompassing input data. Second order variation in the uncertainty can be seen where the model is built from well- or poorly-constrained profiles. For example in the northern North Sea cells near the north-south oriented profile of Christie (1982) are more poorly constrained than cells near the adjacent east-west oriented W-reflector profile (Morgan et al., 2000) (Fig. C.1).

The reduced lateral velocity variation with depth (e.g. the standard deviation in velocity at 17-18 km below sea level is 0.25 km/s compared to 0.65 km/s at 6-7 km below sea level) and the increased uncertainty combine to cause the details in the variation in velocity seen at depth (e.g. Fig. 3.20) to be unconstrained. However, longer wavelength trends gain reinforcement by being repeated on multiple seismic profiles. For example at any given depth between 8-9 km and ~20 km below sea level there is a general trend of higher velocities in the NW of the model than in the SE (e.g. slices at -11.5 and -17.5 km in Figure 3.20). As this trend covers all input data it is a robust feature of the model, even through the amplitude of the variation is generally within the velocity uncertainty (e.g. at 17-18 km the long-wavelength variation is approximately 1-1.5 km/s while the mean uncertainty is 1.38 km/s).

Another more local feature of the lower crust, which is not constrained to within the uncertainties of the model but gains support by repeated occurrence on the input data, is the elevated velocity on the NE side of the Central Graben (e.g. depth slices at -11.5 and -17.5 km in Figure 3.20). The transition from relatively low to relatively high velocities is seen on the 3 MONALISA profiles (Abramovitz and Thybo, 1998, 2000, Chapter 2 of this work and Fig. 3.18) and the Barton and Wood (1984) profile. No other profiles cross the North Sea and hence this appears to be a highly likely feature. However, the northern North Sea is the most poorly constrained region of the model, with uncertainties up to 1.76 km/s at 26-27 km depth, and so details of the spatial distribution of this anomaly
Figure 3.21: Mean crustal velocity. The left-hand image is the mean velocity including the sediments, the right-hand image shows the mean velocity of the crust below the sediments.

cannot be constrained using the model.

Elevated velocities are also seen under Ireland and particularly the Irish Sea (Fig. 3.20). The high velocities at the base of the Irish crust gain reinforcement as Ireland is very densely sampled and modelled with a high-velocity Moho transition zone (e.g. ICSSP in Fig. 3.18) on all profiles except the LEGS models in SE Ireland (Lowe and Jacob, 1989; Masson et al., 1998; Landes et al., 2000; Al Kindi, 2002; Hodgson, 2002). The Irish Sea is less well constrained by data coverage, but the high velocities receive support as they have been inherited from the CSSP input model (Al Kindi et al., 2003). Again, as with the North Sea velocity variation, the model uncertainties which in the Irish Sea are generally between 1 and 1.5 km/s, make the specific shape of these anomalies unconstrained in the model.

An alternative view of the velocity structure is to map the mean crustal velocities of the model (Fig. 3.21). The map of average velocity calculated for the whole crust, i.e. including the sediments, shows dramatic lateral variation with a standard deviation in the mean velocity of 0.388 km/s. This variation is dominated by the spatial distribution and
3. Velocity Model

<table>
<thead>
<tr>
<th></th>
<th>median</th>
<th>mean</th>
<th>std dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>CRUST2.0 (inc. sediment)</td>
<td>6.243</td>
<td>6.198</td>
<td>0.205</td>
</tr>
<tr>
<td>CRUST2.0 (exc. sediment)</td>
<td>6.558</td>
<td>6.516</td>
<td>0.0975</td>
</tr>
<tr>
<td>The final model (inc. sediment)</td>
<td>6.188</td>
<td>6.127</td>
<td>0.388</td>
</tr>
<tr>
<td>The final model (exc. sediment)</td>
<td>6.439</td>
<td>6.443</td>
<td>0.149</td>
</tr>
</tbody>
</table>

Table 3.5: The median, mean and standard deviation in velocity values for CRUST2.0 (Bassin et al., 2000) and the new crustal model.

thickness of the sedimentary basins (Fig. 3.5) as is emphasised by the difference in the mean velocity calculated with and without sediments (Fig. 3.21). The minimum mean velocity for the whole crust is 3.91 km/s (on the Biscay Margin), whereas the minimum for the crystalline crust is 5.79 km/s (on the Møre Margin) and the standard deviation in mean velocity drops from 0.388 to 0.149 km/s with the removal of the sediments from the calculation. The distribution of the mean velocity of the crystalline crust shows some sign of variation related to sediment distribution, with the crust below the North Sea Basin, the Faroe-Shetland Basin and the Celtic Sea Basin of relatively low velocity. However, the crust below the Rockall Basin has areas of both high and low velocity and below the Væring Basin is generally of high velocity. The possibility of a link between crustal velocity and sediment thickness is discussed in Chapter 5.

3.5.1 Comparison with CRUST2.0

The most widely used crustal model is the CRUST2.0 global 2° model of Bassin et al. (2000); therefore the new model is compared to this existing model (Fig. 3.22). The most striking difference between the models is the significant refinement in cell size for the new model and the resultant far greater detail. This is most apparent for the marine sedimentary basins, for example, the central Rockall basin hardly shows at all on the CRUST2.0 model, but is well defined in the new model. The range and spread of velocities are similar (Table 3.5) but, comparing individual local features of the models, the distribution of velocity is quite different. To the west and north of Ireland the velocity structure of the crystalline crust in the two models is broadly similar. An exception to this is southwest Hatton Bank, which is assigned oceanic crust in the CRUST2.0 model and contains lower velocities. The change in velocity between the North Sea and Britain/Ireland is in the opposite sense in the two models, with CRUST2.0 predicting higher velocities in the marine
Figure 3.22: Comparison of CRUST2.0 (Bassin et al., 2000) and the new model.
areas than the onshore regions and the new model predicting higher velocities onshore. The distribution in the CRUST2.0 model is a result of the assigned crustal structure, with Britain and Ireland assigned generic "extended crust" and the marine regions mainly "continental shelf". The structure of extended crust is governed by the global distribution of this crustal type, which, due to geographic distribution, is weighted towards the structure of western America. This results in a low velocity structure that is not consistent with the seismic data. The new model is consistent with the available data. The North Sea and Celtic Basin are classified as continental shelf and therefore assigned relatively high velocities in the CRUST2.0 model, whereas the seismic data available in the two regions suggest variable, but generally lower velocities.

3.6 Summary

In this chapter a new high-resolution, quantitatively constrained, P-wave velocity model has been built to represent the crust of Northwest Europe.

The model was built from two layers (sediments and crust) and three surfaces (topography/bathymetry, base-sediment interface and Moho). The P-wave velocity and velocity uncertainty of the crystalline crust were defined by 3D kriging of data compiled from published wide-angle seismic velocity models. A total of over 31000 km of velocity models were compiled and 1D velocity-depth and velocity uncertainty-depth profiles extracted from these models at 5 km intervals along the profiles. The velocity structure of the sediments was taken from the interval velocities used to model 9 seismic reflection profiles sampling each of the main sedimentary basins around the UK and Ireland. The topography/bathymetry data, used to define the top surface of the model, were the Smith and Sandwell (1997) satellite altimetry data and the GTOPO30 topography. The boundary between the sediments and the crystalline crust was defined using a compilation of datasets giving sediment thickness information, with the most widespread coverage from the National Geophysical Data Center (2004) map and BGS data on the depth to the Variscan unconformity. The base of the model is defined by a surface originally constructed by 2D kriging of Moho depth data extracted from the same wide-angle model database as the crustal velocity data. The Moho depths in specific areas of the model, such as along the
Biscay continental margin, have been optimised during gravity modelling described in the next chapter. The final model was defined on a 40 by 40 by 1 km finite element grid; with the grid geographically referenced using a Transverse Mercator projection, with the origin at 3.4° west, 57.15° north.

The velocity structure of the final model shows strong lateral velocity variation within the top few kilometers, with the sedimentary basins showing considerably lower velocities than regions of near-surface basement. Velocity variation is recorded within the basement and, for the top few kilometers, appears to be well correlated with changes in surface geology. For the crystalline crust in general the uncertainty in velocity is strongly controlled by the data distribution, with the interpolated velocity structure only constrained for cells adjacent to input data. Away from the data the modelled velocity is a regional average and is a largely unconstrained estimate. Despite the high uncertainties in the crystalline crust some robust trends are seen: the vertical and lateral velocity gradients both decrease with depth; the velocities in the lower crust on the northeast side of the central graben are higher than those on the southwest; and there is a long-wavelength trend from higher velocities in the northwest of the model to lower velocities under continental Europe.
Chapter 4

Density Model

The crustal density model has been developed to fulfill a number of uses. Firstly it provides a method for the verification of the velocity model. The conversion of velocities to densities allows the regional variations in the model to be compared to the Free Air Gravity anomalies, providing independent information on the model features. In poorly resolved areas of the velocity model, gravity modelling can allow features of the model to be sharpened-up. However, only tentative conclusions can be drawn as there is significant flexibility in the conversion between velocity and density. The second use of the density model is as a source of information on the nature of the crust and lateral variation in its properties. Densities are as useful as velocities for defining the physical nature of the crust and combination of the two data sets provides the opportunity to compare heterogeneities in each property, allowing better definition and investigation of regional and local features (see Chapter 5). The third use of the density model is as a tool for the further investigation of crustal properties and evolution. As all geological processes are affected by gravitational forces to some degree, knowledge of the crustal density structure can lead to a greater understanding of crustal evolution. For example, in Chapter 6 the density model is used to investigate the flexural strength of the lithosphere and England et al. (submitted) use the model presented here to investigate the causes of the Cenozoic exhumation of the UK.

The gravity anomaly predicted by the density model was calculated using the 3D gravity-modelling package Gmod (Dabek and Williamson, 1999). This code requires the model to be defined as a series of layers defining properties (with either constant or laterally varying density) and continuous 2D grids defining the depth to the layer boundaries. The
conversion of the finite-element structure of the velocity model to a layered density model is outlined in Section 4.1. Gmod’s embedded optimisation routine was used to alter the model properties during the development phase (Sect. 4.2). Both the model optimisation and the field calculation are undertaken in the wavenumber domain (implementing the algorithms of Parker (1972)), requiring the model to be defined on a regular grid. As the crustal model has an irregular boundary peripheral padding was required, the development of the padding is described below.

4.1 Construction of the Starting Model

4.1.1 Main crustal model

For the central region, covered by the velocity model, densities were assigned through a numerical velocity–density function. The velocity model used to define this region was the original interpolated model, with base of the crust and the spatial extent of the model defined by the interpolated Moho (see Fig. 3.13 for model area).

Many laboratory-based studies and in-situ borehole measurements have been used to define empirical relationships between velocity and density in crustal rocks (e.g. Birch, 1961; Hamilton, 1978; Bartetzko et al., 2005). These measured velocity–density values show a large scatter, introducing a significant uncertainty into the relationship between velocity and density, as summarised by Barton (1986). Of the published compilation of velocity–density relationships Christensen and Mooney (1995) provide the most appropriate data for this work. Their results are specifically aimed at characterizing continental crust, with a large number of velocity and density values measured in a range of crustal rocks at high temperatures and pressures (equivalent to burial depths of up to 50 km). Regression curves are given for these data to produce the velocity–density function for continental crust at 10 km depth intervals. The linear functions from Christensen and Mooney (1995) were used to convert the crustal P-wave velocities ($V_p$) to densities ($\rho$). The function is of the form:

$$\rho = a + b V_p$$

with the value of $a$ and $b$ varying with depth (Table 4.1). The gradient of the velocity–
Table 4.1: Coefficients for the Christensen and Mooney (1995) velocity-density conversion.

\[
\rho = a + b V_p, \quad S(\rho, V_p)
\]

is the standard error in the density estimate.

density function increases with depth; therefore, the calculated densities in the model have greater sensitivity to velocity structure at increasing depth. The difference amounts to nearly a 20% increase in sensitivity between 10 and 50 km, with a 0.1 km/s change in velocity at 10 km generating a 0.036 g/cm³ step in density compared to a change of 0.043 g/cm³ for the same velocity change at 50 km depth.

As both the velocity model and the velocity–density function contain a considerable amount of uncertainty it would be an over-interpretation of the data to preserve all the small-scale variation in the best-fit velocity model. Therefore, in building the density model the structure was significantly simplified to produce a 4-layer model, with 2 sediment layers and 2 crustal layers. The two sediment layers each equate to half the total sediment thickness and the upper and lower crustal layer represent half the thickness of crust between the base of the sediments and the Moho. As the gravity field is relatively sensitive to lateral variations in density the 40 km lateral resolution of the velocity model was preserved. Therefore, to construct the density model the cells of the crystalline crust in the velocity models were converted to densities and then the column of cells at every location averaged to give the values of the two layers in the density model (Fig. 4.1).

As the velocity structure of the sediments is poorly constrained, the sediment density structure was calculated from the Sclater and Christie (1980) compaction curve for shale, in preference to a velocity–density function. Thus the density (\(\rho\)) is given by:

\[
\rho = 2.72 - 1.065e^{-0.51d}
\]  \hspace{1cm} (4.2)

where the density is given in g/cm³ and \(d\) is the depth below the sea bed in kilometers.

This density structure takes no account of basin inversion and variations in lithology and
compaction of the sediments. To account for such variation would require detailed investigation of individual basins, which is beyond the scope of this work and would be severely restricted by data coverage in the deep water basins. Kimbell et al. (2004) give the density–depth variation in Cretaceous aged sediments from the UK and Irish Atlantic margins (Fig. 4.2) and state that this trend of reducing variation with depth is generally seen in well data in the northeast Atlantic margins. Given this observation and the relatively low resolution of the model (compared to the size of the sedimentary basins) the use of a single compaction curve is not considered to cause significant degradation in the quality of the density model.

4.1.2 Continent padding

The sediments were modelled in the padded regions (Fig. 4.3) using the same approach as in the main model space, namely using the Sclater and Christie (1980) compaction curve and the base-sediment interface compiled when building the velocity model.

The Moho depth and density structure of the crystalline crust for the padded regions were taken from the CRUST2.0 model (Bassin et al., 2000) as this is the highest-quality pre-existing crustal model. CRUST2.0 is defined using 2 sediment and 3 crustal layers and
Figure 4.2: Relationship between burial depth and density for Cretaceous units in the northeast Atlantic (Kimbell et al., 2004). Different symbols indicate different wells in the Irish and UK sectors.

Figure 4.3: Regions of padding, stippling indicates areas of continental padding and zigzag fill indicates areas of oceanic padding.
4. Density Model

so the structure required simplification to be consistent with the main model region. For
the region of the CRUST2.0 model covering the area modelled here the mean density of
the upper, middle and lower crustal layers are 2.70, 2.88 and 3.03 g/cm$^3$ respectively and
the lower crust has the greatest variation in density. Therefore to simplify the CRUST2.0
model to 2 layers the upper of these layers was assumed to be 2.75 g/cm$^3$ and the lower layer
density set to the value required to give a mean density consistent with the mean density of
the CRUST2.0 model. If this required the lower crustal density to be 3.0 g/cm$^3$ or greater,
then the upper crust density was reset to 2.80 g/cm$^3$ and the lower crust recalculated.

The area of padded continental crust in the southern Rockall Bank/Edoras Bank region
could not be defined using the CRUST2.0 model as this model assigns oceanic crust to
the region. Therefore, the Moho and crustal properties were assigned by extrapolation
(using the tensioned minimum curvature algorithm of Smith and Wessel (1990)) from the
Hatton/Edoras Bank.

4.1.3 Ocean padding

The oceanic crust to the north of Edoras Bank formed over the last 60 Ma under the
influence of the Iceland plume. Therefore, it cannot be assumed that all oceanic regions
in the model have the thickness and density structure of standard ocean crust. To allow
for variation in ocean crust structure three categories were used to build the padding:
1) Anomalous crust adjacent to continental margins; 2) general anomalous crust; and 3)
"standard" oceanic crust. The distribution of these crustal types is shown in Figure 4.4.
The primary data used to assign regions of the model to one of these 3 categories was the
deviation of the observed bathymetry from that predicted by the plate cooling model of
Parsons and Sclater (1977), with standard crust assigned to regions where the bathymetry
is within 1.2 km of that predicted by the plate cooling model. Any region where the
bathymetry is more than 1.2 km shallower than the plate cooling model was considered
to be anomalous crust. The only exception to this is the the finger of standard crust near
Hatton Bank, which was assigned standard crust following White (1997). The anomalous
crust in close proximity to continental margins was separated from general anomalous crust
to allow the crustal structure to be defined in a manner consistent with the observation of
Fedorova et al. (2005), who found the relationship between Moho depth and topography
Figure 4.4: Regions of each category of ocean crust used to produce the padding for the crustal model. Pale blue indicates standard ocean crust; purple indicates anomalous crust associated with continental margins; and mid-blue represents general anomalous crust.

near the continental margins was significantly different to that of other regions of anomalous crust.

Standard crust was assigned a simple two-layer structure with 2.3 km of upper crust (below the sediment) of density 2.80 g/cm³ (layer 2) and 4.7 km of lower crust of density 2.95 g/cm³ (layer 3), consistent with the models of Müller and Smith (1993).

The structure of anomalous ocean crust was based on the findings of a range of seismic and gravity investigations. In particular there have been a number of separate studies into the relationship between topography and Moho depth in the anomalous regions of the North Atlantic, summarised in Table 4.2. Many of these studies use dry residual topography; where “dry” indicates that the bathymetry is converted to the equivalent surface height if the water was replaced by an equivalent mass of crustal material (Kaban et al. (2002) calculate an adjusted topography that also includes converting the density of ice sheets and the crust above sea-level to 2.67 g/cm³). “Residual” refers to the removal of the age-related decay in topography predicted by the plate cooling model. In order to build the oceanic padding the dry residual bathymetry was calculated using the Parsons and Sclater (1977) plate cooling model, with the mass of water converted to crustal material with density 2.8 g/cm³ (Fig. 4.5). The age of the ocean crust was taken from Müller et al.
### Table 4.2: Summary of the relationship between topography and Moho depth for the North Atlantic.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Area</th>
<th>Data</th>
<th>( dt/dM )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Menke (1999)</td>
<td>Onshore Iceland</td>
<td>( t_{raw} )</td>
<td>0.03554 ± 0.0064</td>
</tr>
<tr>
<td>Kaban et al. (2002)</td>
<td>Onshore and Offshore Iceland</td>
<td>( t_{dry+red} )</td>
<td>0.0578</td>
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<tr>
<td>Gudmundsson (2003)</td>
<td>Onshore proximal to the Mid Atlantic rift</td>
<td>( t_{dry} )</td>
<td>0.030 ± 0.005</td>
</tr>
<tr>
<td>Gudmundsson (2003)</td>
<td>Offshore Iceland proximal to the Mid Atlantic rift</td>
<td>( t_{dry} )</td>
<td>0.116 ± 0.012</td>
</tr>
<tr>
<td>Fedorova et al. (2005)</td>
<td>Onshore Iceland</td>
<td>( t_{raw} )</td>
<td>0.04038</td>
</tr>
<tr>
<td>Fedorova et al. (2005)</td>
<td>Faroe-Iceland Ridge</td>
<td>( t_{dry+red} )</td>
<td>0.0469</td>
</tr>
<tr>
<td>Fedorova et al. (2005)</td>
<td>Jan Mayen Block</td>
<td>( t_{raw} )</td>
<td>0.0396</td>
</tr>
<tr>
<td>Fedorova et al. (2005)</td>
<td>Continental Margin regions</td>
<td>( t_{dry+red} )</td>
<td>0.0715</td>
</tr>
<tr>
<td></td>
<td></td>
<td>( t_{raw} )</td>
<td>0.0788</td>
</tr>
<tr>
<td></td>
<td></td>
<td>( t_{dry+red} )</td>
<td>0.1923</td>
</tr>
<tr>
<td></td>
<td></td>
<td>( t_{dry+red} )</td>
<td>0.1962</td>
</tr>
</tbody>
</table>

\( dt/dM \) is the rate of change in topography with change in Moho depth. In the data column \( t_{raw} \) indicates that the true topographic heights where used, \( t_{dry+red} \) indicates that dry residual topographic heights were used, see text for details.

(1997).

There have also been numerous combined seismic and gravity investigations into the crustal structure of Iceland and the surrounding oceanic regions. The models resulting from these investigations are summarised in Figure 4.6. In order to combine these results and project them across the entire oceanic region of the model a simplified structure was designed based on the data presented in Figure 4.6 and Table 4.2:

- The crustal thickness is assumed to be 10 km where the dry residual topography is 1.2 km. This is broadly in keeping with the results of Smallwood and White (1998) and RISE (Weir et al., 2001) on the Reykjanes Ridge.

- For the general anomalous crust the relationship between change in excess topography and change in Moho depth is assumed to be \( dt/dM = 0.120 \), while the crust is less than 21 km thick (i.e. where residual topography is less than 2.52 km). This is consistent with the results of Gudmundsson (2003) for the Reykjanes Ridge.

At the continental margins the ratio between topography and crustal thickness is assumed to be 0.196, following Fedorova et al. (2005).

The crust is assumed to be 1/3\(^{rd}\) layer 2 and 2/3\(^{rd}\) layer 3, which allows smooth scaling from normal oceanic crust and is broadly consistent with the data in Figure 4.6.
Figure 4.5: Residual topography after conversion of bathymetry to ‘dry’ topography and removal of the age-controlled decay predicted by Parsons and Sclater (1977) plate cooling model.

Figure 4.6: Summaries of the density structure in the anomalously thick oceanic crust of the North Atlantic. Shading on the models is related to the density values; green regions have densities similar to standard oceanic layer 2; blue are similar to layer 3 densities; red indicate densities of ~3.09 g/cm³; purple are intermediate between layer 3 and 3.09 g/cm³; and turquoise are intermediate between layers 2 and 3. ICEMELT, SIST and B96 from Darbyshire et al. (2000), FIRE from Smallwood et al. (1999) and RISE from Weir et al. (2001).
If the general anomalous crust is greater than 21 km thick the relationship between topography and Moho depth is taken as $\frac{dt}{dM} = 0.052$. This rate of change is midway between the values for onshore Iceland from Kaban et al. (2002) and Fedorova et al. (2005) and combines with the points above to produce a Moho that is similar to the predictions of Darbyshire et al. (2000); Kaban et al. (2002); Foulger et al. (2003) and Fedorova et al. (2005) under most of Iceland.

The structure is assumed to be 7 km of layer 2, 14 km of layer 3 and with the remaining crustal thickness assigned a density of 3.09 g/cm$^3$. This structure allows: a smooth transition from the thinner crust; is consistent with predictions of upper crust thickness from Foulger et al. (2003); and coarsely follows the deeper structure modelled from the ICEMELT and FIRE projects (Fig. 4.6).

This structure also allows the general anomalous crust to be in isostatic equilibrium, without requiring significant flexure or short-wavelength subcrustal density variation. In the case of crust less than 21 km thick the model has a constant mean crustal density regardless of the change in crustal thickness; therefore, isostatic compensation at the Moho requires:

$$\frac{dt}{dM} = \frac{\rho_m - \rho_c}{\rho_c} \approx \frac{3.25 - 2.90}{2.90} \approx 0.121$$

For crust greater than 21 km thick upper and lower crust thickness are constant, only the thickness of high density layer varies, and so compensation at the Moho requires:

$$\frac{dt}{dM} = \frac{\rho_m - \rho_{3.09}}{\rho_{3.09}} \approx \frac{3.25 - 3.09}{3.09} \approx 0.052$$

A cross section through the model demonstrating the structure of the oceanic crust is shown in Figure 4.7, along with maps illustrating the starting model densities and depths to the base of the sediments and the Moho.
Figure 4.7: Maps and cross-section showing the structure and properties of the starting model. The cross section depths are in kilometers.
4. Density Model

<table>
<thead>
<tr>
<th>Lithosphere age (stability zone)</th>
<th>( \rho_{20} ) (g/cm(^3))</th>
<th>( a_0 \times 10^{-4} )</th>
<th>( a_1 \times 10^{-8} )</th>
<th>( a_2 )</th>
<th>( 1/\beta )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Phanerozoic (spinel)</td>
<td>3.36</td>
<td>0.27768</td>
<td>0.95451</td>
<td>-0.12404</td>
<td>128</td>
</tr>
<tr>
<td>Phanerozoic (garnet)</td>
<td>3.37</td>
<td>0.2697</td>
<td>1.0192</td>
<td>-0.1282</td>
<td>130</td>
</tr>
<tr>
<td>Proterozoic</td>
<td>3.34</td>
<td>0.27014</td>
<td>1.05945</td>
<td>-0.1243</td>
<td>130</td>
</tr>
</tbody>
</table>

Table 4.3: Parameters for calculating mantle density (Poudjom Djomani et al., 2001). Legend: \( \rho_{20} \), density at 20°C; \( a_0-a_2 \), constants for estimation of coefficient of thermal expansion; \( \beta \), compressibility.

4.1.4 Sub-crustal density variations

The region covered by the model incorporates: oceanic lithosphere, ranging in age from 0 to over 120 Ma; Proterozoic continental lithosphere of the Baltic Shield; and Phanerozoic continental lithosphere, that has experienced Mesozoic extension, Alpine compression and plume initiation in the Tertiary. Therefore, it is necessary to include sub-crustal density variations that accommodate the lithosphere age and temperature variations.

In the oceanic regions the thermal structure of the upper mantle was estimated using the Parsons and Sclater (1977) plate cooling model. An adiabatic mantle temperature structure was assumed below the base of the plate cooling model (i.e. below 125 km), with the adiabat defined by a potential temperature of 1300°C, and a temperature gradient of 0.5°C/km. This adiabat predicts a temperature of 1363°C at 125 km and the plate cooling model assumes 1350°C at 125 km; therefore the temperature step at the change from plate cooling model to adiabatic model is minimal. This thermal structure was used to adjust the density of peridotite at 20°C for the effects of compression and thermal expansion with depth. Following Poudjom Djomani et al. (2001) the density due to thermal expansion \( \rho_T \) was calculated using:

\[
\rho_T = \rho_{20}(1 - \alpha T) \tag{4.5}
\]

\[
\alpha = a_0 + a_1 T + a_2 T^{-2} \tag{4.6}
\]

where \( T \) is the temperature in °C, \( \rho_{20} \) is the density at 20 °C and \( \alpha \) is the thermal expansion coefficient, which is controlled by constants \( a_0-a_2 \) given in Table 4.3. This density was then corrected for compression using:

\[
\rho = \rho_T(1 + \beta P) \tag{4.7}
\]
where $\beta$ is the compressibility, given in Table 4.3 and $P$ is the pressure in GPa. The spinel peridotite-garnet peridotite transition was included at 55 km depth.

In the continental regions the temperature in the mantle was modelled using an estimate of lithosphere thickness and the following assumptions:

- Adiabatic sub-lithospheric temperature, with the same adiabat as for the oceanic regions.

- The temperature at the base lithosphere is equal to the mantle adiabat at this depth. This assumption neglects the complication in thermal structure at the base of the lithosphere caused by small-scale convection (e.g. Jaupart and Mareschal, 1999; McKenzie et al., 2005). However, given that the mantle density is modelled as a single layer extending from the Moho to 200 km depth, this simplification is insignificant.

- The heat flow at the Moho is 20 mW/m$^2$ (Cermak, 1993).

- There are no heat producing elements within the mantle.

In the Proterozoic Baltic Shield the lithosphere was taken to be 150 km thick (Artemieva and Mooney, 2001). The boundary between the Pre-Cambrian and Palaeozoic lithosphere was defined using seismic tomography, rather than surface geology. The tomography directly images the deep structure; therefore, it provides a more appropriate control on the mantle properties than extrapolating the near surface rock age below the Moho. The limit of the Baltic Shield was defined using the S20RTS tomographic model (Ritsema and Van Heijst, 2000). This global shear-wave velocity model shows a strong high-velocity lithospheric lid associated with the East European Craton and the Baltic Shield. For the Phanerozoic lithosphere under the rest of continental Europe a simple starting model of a 125 km thick lithosphere was used. This thickness was chosen to be consistent with the asymptotic thickness of oceanic lithosphere (in the Parsons and Sclater (1977) model). The Mesozoic extension generated an elevated geotherm that has decayed through time, regenerating the lithosphere. It is assumed that this re-growing lithosphere is the same as the developing oceanic lithosphere. Measurement of ocean bathymetry suggests that lithosphere growth is limited after $\sim$70-80 Ma. Therefore, it is assumed that the Mesozoic perturbations have decayed back to equilibrium, resulting in a 125 km thick lithosphere.
This model takes no account of the disruption to the mantle geotherms that will have resulted from the Alpine Orogeny (associated with sub-lithospheric remnants of subduction in the southern regions of the model and widespread regional compression) and the Tertiary initiation of the Iceland mantle plume (an event that would have generated a large, widespread thermal perturbation in lithosphere that was still recovering from the Mesozoic extension). However, the effects of these two events on the lithosphere are not well enough understood to be included. Instead, the simple starting model was produced and the required deviations from this model were calculated during the model development (Sect. 4.2).

The effects of compression and thermal expansion were calculated using Equations 4.5–4.7 and the constants in Table 4.3. Examples of geotherms and mantle density structure are shown in Figure 4.8.

The primary reason for including sub-crustal density variation is to correct the long-wavelength gravity anomalies generated by the variation in temperature and bulk chemistry across the ocean basin and within the continent. Therefore, the mantle was modelled as a single layer extending from the Moho to 200 km depth. Sensitivity tests on the model indicated that including a more complex mantle structure, with up to 4 sub-crustal layers, produced results that were generally within 10 mGal of the calculated anomaly from a model with a single mantle layer.

4.2 Model Development

The starting model produces a gravity anomaly that is significantly different from the observed Free Air anomaly; the root-mean-square (rms) residual is 63 mGals and is dominated by a high amplitude long-wavelength misfit (Figs 4.9 and 4.10). The observed gravity data are Sandwell and Smith (1997) global data (version 9.1), which uses satellite altimetry derived data in marine areas and the EGM96 geopotential model (Lemoine et al., 1998) onshore. Given the relatively simple structure used for sub-crustal densities and the failure to include the effects of the Iceland plume and the Alpine Orogeny in the starting model, the subjective decision was made that the long-wavelength, high amplitude misfit was more likely to be generated by inaccuracies in the mantle structure than the crustal model. The
Figure 4.8: Examples of geotherms and density structure of the mantle. The thin solid lines give the geotherms, the thick solid lines the density. The density structure is shown relative to the reference Earth models of PREM (Dziewonski and Anderson, 1981) and AK135 (Kennett et al., 1995).
crustal model could not be optimised without either filtering out the long-wavelength misfit or optimising the mantle structure as a preliminary stage to developing the crustal model. Therefore, the model was improved using a two-stage process, with the long-wavelength misfit removed prior to consideration of the small-scale local anomalies.

4.2.1 Long-wavelength adjustment

This optimisation was performed within Gmod and involved simply varying the density of the mantle layer, it did not include any recalculation of the thermal structure; however, in the following paragraphs the results are discussed in terms of implications for the thermal structure. The optimisation was restricted to features with a wavelength of $\sim 1000$ km or greater.

The results of this investigation show that, if all long-wavelength misfit is corrected in the mantle, the density is required to be lower under the North Atlantic and substantially higher under northern France and the Baltic Shield (Fig. 4.11(a) and (b)). Reducing the density under the North Atlantic seems a reasonable result given the (unaccounted for) thermal anomaly in the region. Under the Baltic Shield the mantle was originally predicted to be lower density than in the Phanerozoic regions due to age-related variation in the composition of newly formed sub-continental lithospheric mantle (Poudjom Djomani et al., 2001). However, the density structure of the crust along the eastern extremity of the model requires the density to be higher than the Phanerozoic mantle. As this area lies in the the model padding, rather than in the main model space, this difference is accepted, but not considered robust enough to be assigned any significance. The increased density in northern France would require substantial reduction in the temperature or change in the chemistry of the mantle. A strongly suppressed geotherm is unlikely as this is a region of high surface heatflow (Fig. 4.12) and has been associated with Cenozoic volcanism in the Massif Central and Rhine Graben regions. Additionally seismic tomography models do not suggest anomalously fast mantle (e.g. Bijwaard et al., 1998; Goes et al., 2000; Piromallo and Morelli, 2003; Pilidou et al., 2004) and a significant change in bulk chemistry in the region has not been previously postulated. Therefore, other features of the starting model that could be responsible for the residual anomaly in the region were considered. The southern North Sea and mainland Europe are not well constrained regions of the velocity model,
Figure 4.9: Observed (Smith and Sandwell, 1997), calculated and residual Free Air Gravity anomalies associated with the starting density model, with coastlines and bathymetric contours, at 1 km intervals, in black.
Figure 4.10: Power spectrum of the residual gravity anomaly of the starting model.

Figure 4.11: Density structure of the sub-crustal layer. (a) shows the starting model; (b) the optimised structure if all misfit with a wavelength $\geq 1000$ km is accounted for by changing the mantle; (c) similar to (b), but with the SE corner of the model corrected by altering the crustal densities.
as discussed in Section 3.5; therefore the crustal density structure was investigated as a possible cause for the misfit. The mean crustal density in the south east of the starting model is low compared to the density of other land regions (Fig. 4.13). The values of upper crustal and lower crustal densities required to remove the long-wavelength residual anomaly in mainland Europe, with half the change allocated to each layer, are shown in Figure 4.14. This density structure remains broadly consistent with the mean densities for the rest of the model (2.772 and 2.957 g/cm³ mean density for the upper and lower crust of the starting model) and therefore was considered far more likely than the alternative of increased density mantle. The density structure of the crust was changed to incorporate these values and the mantle density recalculated to remove the long-wavelengths elsewhere (Fig. 4.11(c)).

After the correction for the long-wavelength misfit the residual gravity anomaly was reduced to 33.67 mGal (Fig. 4.15).

4.2.2 Short-wavelength adjustment

The removal of the long-wavelength misfit generates a residual map with a series of positive and negative misfits (Fig. 4.15). In order to optimise the model each region of misfit was investigated and the required changes to the sediment thickness, upper and lower crust density structure and Moho depth explored. Through these investigations the best method to reduce the residual was assessed and the model altered. The resulting changes to the
Figure 4.13: Mean density of the crystalline crust in starting model.

Figure 4.14: Upper and lower crustal densities for mainland Europe, optimised on wavelength greater than ~1000 km.
The padded regions (Fig. 4.16(a)). For the continental regions of padding, built using CRUST2.0, the main cause of misfit is the lack of short-wavelength variation in density and Moho depth caused by the 2° resolution of the CRUST2.0 model. As much of the continental padding covers regions of substantial topography, it was considered extremely likely that the inability to produce short-wavelength Moho structure would generate misfit. Therefore, the regions built using the CRUST2.0 model were optimised by altering the Moho depth.

In the regions of oceanic padding the gravity residual was removed by altering the crustal thickness and restricting the crustal structure to that used to generate the original padding (i.e. for crust less than 21 km thick 1/3rd of the thickness is layer 2 and 2/3rd layer 3, for crust greater than 21 km thick, 7 km is layer 2, 14 km layer 3 and the remaining crust has a density of 3.09 g/cm³). In several regions this approach did not produce satisfactory results and additional modelling was required, described below.

Iceland (Fig. 4.16(b)). The Moho depth was restricted to 45 km under Iceland, to remain similar to the seismically derived Moho depth maps (Darbyshire et al.,
4. Density Model

Figure 4.16: Areas covered for the regional adjustments to the model described in the text.

Figure 4.17: Thickness of low density material needed to remove the residual anomaly associated with Iceland.

2000; Kaban et al., 2002; Foulger et al., 2003; Fedorova et al., 2005). This restriction leaves a substantial misfit in the calculated gravity anomaly. However, the initial model did not include the several kilometers of low density, near-surface material that is often present (see Fig. 4.6) and by adding a layer of density 2.60 g/cm\(^3\) the residual was removed (Fig. 4.17).

**Lofoten region** (Fig. 4.16(c)). The density structure of the CRUST2.0 model required an extremely deep Moho under the Lofoten Islands and onshore Norway. However, reducing the densities of the upper and lower crust to 2.75 and 2.95 g/cm\(^3\) respectively (similar to the values in the model just to the south) allowed the Moho to shallow to values more consistent with the adjacent areas of the model.
4. Density Model

Figure 4.18: Areas covered for the regional adjustments to the model described in the text.

Scandinavia (Fig. 4.16(d)) is a poorly constrained region of the model, with only limited input data and is included in the model primarily to reduce extrapolation errors in the North Sea and along the Norwegian Margin. The Moho depths in the starting model are, in general, shallower than those modelled by Kinck et al. (1993) in their more comprehensive compilation of Fennoscandian seismic data (Fig. 1.7). Therefore, in order to reduce the model misfit 50% of the residual was removed through altering the Moho depth, with the remaining 50% split evenly between the upper and lower crust densities.

The Porcupine region and the Biscay Margin (Fig. 4.18(a)) is a region of the model that suffers from unfortunate data distribution insomuch as the substantial bathymetric variation suggests that there may be significant Moho relief associated with the Porcupine Bank and Sea Bight and relatively sharp Moho relief on the Biscay Margin, but interpolation of Moho depths could not predict such variation from the data. Therefore, the Moho depths were optimised to remove the gravity misfit. The substantial change required has been accepted as the Moho contours of the final model echo the bathymetry to a high degree.

The Rockall Trough (Fig. 4.18(b)) is well constrained by seismic data. However, the densities in the lower crust are relatively high due to the Moho transition zone in the RAPIDS1 wide-angle model (O’Reilly et al., 1995). Additionally lower crustal velocities are generally less well constrained than upper crustal velocities. Therefore the majority of the misfit was accommodated in the lower crust, with a small alteration allowed in the upper crust density and the Moho structure. The final model
Figure 4.19: An example illustrating the unreasonable results of attempts to optimise the sediment thickness in deep basins. The small change in calculated gravity anomaly produced by the large change in sediment thickness is a result of the small contrast in density between the deep sediments and the upper crust.

contains some very strong lateral density changes, which may be partially due to inaccuracies in the sediment thicknesses being mapped into the crustal structure. However, attempts to optimise the base-sediment interface, allowing less crustal density variation, proved to be unstable and have unreasonable results (Fig. 4.19). The optimisation difficulties are generated by the very small density contrast at the base of such deep basins, which contribute very little to the total gravity anomaly, making the optimisation insensitive.

Ireland (Fig. 4.18(c)), like the Rockall Trough, has a tightly constrained Moho. Additionally several of the seismic models providing this constraint were produced with the assistance of gravity modelling (Lowe and Jacob, 1989; Masson et al., 1998; Al Kindi, 2002), allowing the model to be compared to the published 2D models (Fig. 4.20). This comparison shows that the 2D models use lower deep-crustal densities. These lower densities are more consistent with the surrounding regions. Therefore, the majority of the alteration to the starting model was assigned to the lower crust, but with a small amount of upper crustal density change also allowed.

The Shetland Platform (Fig. 4.18(d)) had the largest misfit inside the main model area, with the calculated anomaly more than 100 mGals too high under Shetland. The proximity of the seismic lines AMP-A and C and NASP-D mean that that
Figure 4.20: Published 2D models of the density structure of Ireland, (a) redrawn from Masson et al. (1998), (b) Al Kindi (2002) and (c) Lowe and Jacob (1989).
Figure 4.21: Areas covered for the regional adjustments to the model described in the text.

region is fairly well constrained and the misfit could only be removed by altering both crustal layer densities and the Moho depth (each accommodated $1/3^{rd}$ of the required change).

**Mainland Europe** (Fig. 4.21(a)). As this region had already been subjected to long-wavelength change in density structure, it was considered reasonable to further alter the densities to remove the short-wavelength misfit. For this optimisation $1/3^{rd}$ of the required change was assigned to the upper crust density and $2/3^{rd}$ to the lower crust. This distribution of change was chosen as it results in a relatively smooth variation in upper crust density and removes the very low density values seen in some regions of the lower crust in the original model.

**The remaining model space** (Fig. 4.21(b)) is all reasonably well sampled by wide-angle models; therefore, has fairly well constrained Moho depths. As a result the misfit was removed by assigning 50% of the required change to the upper crust density and 50% to the lower crust density.

Combining these regional changes produces a final model with areas of crustal density and Moho depth that differ from the starting model (due to the instabilities in optimisation, the sediment thickness is left unchanged). Maps showing the starting values, final values and the adjustment of each feature of the model are shown in Figures 4.22 and 4.23. After incorporating these changes the rms misfit between the observed and calculated gravity anomaly was reduced to 8.80 mGals.
Figure 4.22: Maps showing the change in density structure of the upper and lower crust.
Figure 4.23: Maps showing the change in Moho depth.
4. Density Model

Figure 4.24: The relationship between Moho depth/ocean crust thickness and residual "dry" topography for the final model. Red points indicate cells of ocean crust to the north of the Charlie Gibbs Fracture Zone and blue points cells to the south. Black lines indicate the $dt/dM$ values of 0.054 and 0.12 used to build the starting model while the dashed line is a relationship of 0.033.

4.3 Model Verification

To ensure that the final density model is consistent with all the available data on crustal structure a number of verification procedures were undertaken.

To investigate the ocean crust structure of the final model, the Moho depth and ocean crust thickness were plotted against the residual dry topography (Fig. 4.24). The graphs show considerable scatter and there are several features of note. Firstly, for the thickest crust (under Iceland and the Faroe-Iceland Ridge) there is only a very weakly defined $dt/dM$ relationship, but it appears to be smaller than the starting model. A $dt/dM$ of ~0.033 would fit this weak trend, and although it is a little on the low side of the observed values for Iceland (Table 4.2), given the poor correlation it is considered acceptable. Secondly, there is now a large scatter in the thickness of the ocean crust for the model to the south of Charlie Gibbs Fracture Zone, most notably the crust is effectively removed in some cells on the Biscay margin. This is a region where exhumation of the mantle has been postulated (see discussion in Section 4.4).

For the continental regions the model validation has involved: 1) Checking the velocity-density relationships of the final model against empirically derived relationships; 2) Comparing the final model to wide-angle seismic models; 3) Comparing the change in Moho
depth to the uncertainty; and 4) constructing alternative models to fit the gravity anomaly.

**Velocity-density relationship**

Given the difference in structure between the velocity and density models, with a number of velocities contributing to a single density value, the velocity-density relationship at any given location in the model can be constructed from two approaches. The individual cells of the velocity model can be compared to the density value at the equivalent position in the density model; or the density of a layer in the density model compared to the average velocity of cells that contribute to that layer. Velocity-density relationships, for the starting and final models, were constructed using both these approaches and are shown in Figure 4.25. The graph comparing density to average layer velocity for the starting model deviates from the Christensen and Mooney (1995) relationships at high velocities as the density of individual model cell was capped at 3.1 g/cm$^3$. This capping was applied in regions of magmatic underplating where very high velocities (up to 8.1 km/s) have been modelled. The capping assumes that the greatest density the crust can have is generated by a gabbroic composition with some garnet (a density of $\sim$3.0 g/cm$^3$ would represent pure gabbro (Christensen and Mooney, 1995)). The relatively thin crust of the region makes the widespread presence of garnet unlikely. The graph of cell velocity and density for the starting model shows more scatter, as cells with a range of velocities go into a single valued density layer, but the vast majority of the points fall comfortably within the Nafe-Drake limits. The two bulls-eyes on the graph show the tendency for densities to cluster near the typical values of 2.75 and 2.95 g/cm$^3$ for the upper and lower crust. In the final model the mean layer velocity-density graph shows more scatter, due to the adjustment to the layer densities, but almost all values plot comfortably within the Nafe-Drake range. The cell velocity-density graph for the final model shows more defined bulls-eyes than the starting model, as during model development there was a conscious preference for reducing the gravity misfit by reducing the variation in layer density.

**Comparison to wide-angle seismic models**

In order to compare the final density model to the wide-angle seismic models, sections have been extracted coincident with the seismic profiles. By overlaying the digitised points from
Figure 4.25: Velocity–density relationships for the starting and final models. On the four graphs the black lines indicate the upper and lower bounds of the Nafe-Drake velocity–density pairs (Barton, 1986), the blue lines the Christensen and Mooney (1995) velocity–density functions for 10 to 40 km (the gradient decreases with depth), see Table 4.1. The left-hand graphs show the individual cells of the velocity model compared to the density model. The right-hand graphs show the density compared to the average velocity of cells that contribute to the layer. The lowermost cartoon illustrates the relationship between the finite element velocity model and the layered density model.
the velocity models (layer boundaries in ray traced models and iso-velocity contours from
tomographic models) it is possible to compare the structure in the density model to that of
the original wide-angle seismic models. A selection of such comparisons (chosen to provide
coverage for all regions of the model) are shown in Figures 4.26–4.27. Generally the fit to
the seismic models is very good. A feature of note is the strong crustal density variations
invoked in the Rockall Trough in order to match the observed gravity anomaly. The
constraint on Moho depth from the seismic data was considered more reliable than the lower
crustal velocity or the velocity–density function. By restricting Moho change strong crustal
density variation was needed to remove the anomaly. There is also considerable difference
at the oceanic end of AMP profile E, here the difference arises where the profile moves
into oceanic crust. The crustal structure of the oceanic region was forced to meet that
described in Section 4.1.3, rather than following the wide-angle profile. In this particular
case this approach created significant change and removed the magmatic underplate.

Change in Moho depth

To compare the change in Moho depth to its uncertainty, the final Moho depth is compared
to the probability function of the Moho occurring at that depth. The construction of the
Moho depth probability function is described in Section 3.3.5. The final model Moho
is assessed in terms of the distance from the most probable depth, measured using the
standard deviation of the probability function (i.e. the likelihood of the Moho falling outside
of one standard deviation from the mean is ~31.73%; the likelihood of the Moho being
further than two standard deviations from the mean is ~4.55%; greater than three standard
deviations from the mean has only a ~0.27% probability). The results of this investigation
are shown in Figure 4.28. The considerable amount of purple and red on the plot show
that the final model Moho is, in places, at a depth that the interpolation suggests is very
unlikely. However, as discussed in Section 4.2.2 on model development, the distribution
of seismic data on the continental margin from the Porcupine Bank to the south of the
model is such that high relief on the Moho could not be predicted by the kriging. The
same situation applies under the Shetland Platform. Although this argument suggests
that the Moho uncertainties are untrustworthy, this is only the case where there is strong
Moho relief without adequate seismic sampling. For the majority of the model the Moho
Figure 4.26: Gravity profiles and sections through the density and original velocity model along wide-angle seismic profiles. Black dots indicate digitised points in the seismic model, the solid black line shows the Moho in the seismic model. For the gravity profiles the red line indicated the calculated anomaly and the black line the observed. Depths are in kilometers and gravity profiles in mGals.
Figure 4.27: Gravity profiles and sections through the density and original velocity model along wide-angle seismic profiles. Black dots indicate digitised points in the seismic model, the solid black line shows the Moho in the seismic model. For the gravity profiles the red line indicated the calculated anomaly and the black line the observed. Depths are in kilometers and gravity profiles in mGals.
Figure 4.28: Moho change with respect to uncertainty. On the left the change in Moho depth is plotted as a function of the standard deviation in the Moho probability function. The right-hand image shows the magnitude of the change in kilometres.

is relatively flat, or the sampling very dense, and the uncertainty values are likely to be reasonable.

**Alternative models**

The discussion of alternative models is restricted to models with laterally constant upper and lower continental crust densities. This simple crustal structure is commonly used in studies of flexural isostasy (e.g. Watts, 2001, and references therein) and regional gravity modelling, where information on crustal structure is either not available or coverage is not considered dense enough to provide good constraints (e.g. Kimbell et al., 2004). In terms of verifying the final density model, these alternative models provide information on the significance of crustal density variation on the calculated gravity anomaly and on the trade-off between crustal density and Moho depth encountered during model development (as both crustal density and Moho depth could be altered to remove misfit between the calculated and observed anomalies in the starting model, the changes in the density and crustal thickness to produce the final model are inter-related). The many possible models with laterally varying densities are not discussed, as the starting model presented above
Figure 4.29: The gravity anomaly predicted by the alternative models with constant density continental crust.

represents the best fit to the velocity model (using a geographically constant velocity-density function for the initial model) and the final model the minimal change to the starting model required to produce a good fit to the gravity data. Although the required change could be distributed differently between the layer densities and/or layer boundaries, the chosen distribution of change was justified in terms of the parameter uncertainties and external knowledge of the crustal structure. Alternative distributions would represent arbitrary choices. Therefore, such models were considered to provide no further information or constraint on the crustal structure.

Two groups of alternative models, with constant density continental crust, were produced. In all alternative models the lower crust density was set to 2.95 g/cm$^3$ and the upper crust to 2.8 g/cm$^3$ in the oceans and 2.75 g/cm$^3$ in the continental regions. The sediment density structure from the final model was preserved. The first group of alternative models uses the base-sediment interface and Moho depths from the final model. As the group differs from the final model only in the use of constant density crust, it shows the significance of the modelled density variations. The group consists of 2 models; (a) uses the simple mantle density structure from the starting model and (b) uses the mantle density structure of the final model. The calculated gravity anomaly for these two models is shown in Figure 4.29. It is apparent that applying a constant density crust introduces
significant misfit between the calculated and observed anomalies. The misfit is present at all wavelengths, but is dominated by longer wavelengths (Fig. 4.30) indicating that the crustal density variation is important at all wavelengths, but the differences between the density structure in different regions (e.g. the density structure of the Baltic Shield compared to the Phanerozoic crust) are more significant than local variations. The maps and power spectra of calculated gravity anomalies show that the sub-crustal density structure has a large effect on the calculated anomaly, greater than the effect of the crustal density variation (Figs 4.29 and 4.30). The choice of mantle density structure therefore has an important effect on the model of crustal structure.

The second group of models is the same as those described in the previous paragraph; however, for these models the Moho depth is also optimised so that the calculated anomaly matches the observed gravity data. Comparing these models with the final model provides some insight into the effect that lateral density variation has on the Moho structure. The Moho depths for these two models are shown in Figure 4.31. In the oceanic regions the alternative model 2a (with the starting model’s sub-crustal structure) shows the required Moho depth (i.e. crustal thickness) of the ocean crust if the mantle density follows that predicted by the Parsons and Sclater (1977) plate cooling model (Fig. 4.31 model 2a). However, it must be noted that in the alternative models the ocean crust structure in regions with crust greater than 21 km thick is different to that in the final model; in the alternative model the ocean crust has 1/3rd upper crust and 2/3rd lower crust (at density 2.95 g/cm³) regardless of the crust thickness and this difference will have an impact on
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Figure 4.31: Moho depths for the final model and alternative models with constant density continental crust.

the optimised Moho depths. South of the Charlie Gibbs Fracture Zone the Moho is fairly similar to the final model, but north of the Fracture Zone the crust is significantly thicker (approximately 14 km thick northwest of Hatton Bank compared to 8-10 km in the final model, Fig. 4.31(a) and (c)). This is consistent with the prior knowledge that the southern region formed before the initiation of the Iceland plume, whereas the crust formation in the northern region was affected by increased mantle temperatures. Therefore, the Parsons and Sclater (1977) model is expected to deviate from the true mantle structure north of the Charlie Gibbs Fracture Zone and the alternative model forces this deviation into the crust, generating erroneously large crustal thicknesses. The alternative model 2b (with the same mantle structure as the final model) has a very similar Moho for the thin ocean crust as the final model (Fig. 4.31(b)). This result is expected as the only difference in the ocean crust structure between the final and alternative model is the removal of the high density basal layer of the thick crust. This difference shows up markedly in the regions of thick crust, such as Iceland and the Faroes-Iceland Ridge, where the greater density contrast at the Moho allows a far thinner crust in alternative model 2b. The crustal structure of these regions is generally well constrained by seismic data and the final model is far more consistent with these data than the alternative model.
In the continental lithosphere the two alternative models (2a and 2b) have more short-wavelength structure on the Moho than in the final model (Fig. 4.31). However, within the continent most of this high-frequency structure is relatively low amplitude. The areas of high magnitude short-wavelength change are associated with the Rockall and Porcupine Basins and the continental margins. These areas are very similar between the final and alternative models and therefore the short-wavelength power of the Moho is similar for all models (Figs 4.31 and 4.32). The constant lateral density requires the Moho to reflect the combined effect of topography and sediment thickness variations. While this is perfectly reasonable in some places (e.g. the Porcupine Bank and Biscay margin) there are several places where the Moho is known not to echo topography and lateral changes in density are expected. For example, in Scotland the Moho is at its deepest under the Southern Uplands and Midland Valley, not under the topographically higher Grampian Highlands. Similarly, under Denmark there is very good constraint on the Moho structure which is not matched by the flat Moho predicted by the constant density crust.

At wavelengths greater than ~150 km both constant density models have less Moho depth variation than the final model (Fig. 4.32). This difference arises from the long-wavelength density variation in the final model in the continental regions (Fig 4.22); not including this structure produces a flatter Moho across much of mainland Europe (Fig 4.33). In western Europe the final model is poorly constrained and the flatter Moho and more constant density structure of the alternative models may well be more accurate. However, in Scandinavia the reduced Moho topography (Figs 4.31 and 4.33) is an artifact generated
Figure 4.33: Filtered Moho depths for the final model and alternative models with constant density continental crust. Filtering removes structure with less than 300 km wavelength.

by forcing densities in the Baltic Shield to values consistent with the Phanerozoic crust in western Europe.

4.4 Final Model

The final crustal model is shown, in Figures 4.34 and 4.35, with the regions of model padding partially masked. The model boundaries have been adjusted to remove onshore mainland Europe. This adjustment was made as these regions are not well constrained (thus, are more likely to contain errors than the parts of the model preserved) nor are these areas of primary interest in this study. They have been reclassified as padding so that they do not distract from the better resolved regions.

The model contains significant lateral density variation, both for the complete crustal model (with sediments included) and in the crust beneath the sediments (Fig. 4.35). For the complete crustal model the sediments have quite a strong controlling influence on the distribution of densities, with lower densities associated with the sedimentary basins (Fig. 4.35). However, unlike the velocity model, sediment thickness variation does not overpower variation in the crystalline crustal structure, which also contains significant lateral density variation (Fig. 4.34, upper and lower crust density maps). The standard deviation of the mean densities is 0.064 g/cm³ for the final model when including sediment
Figure 4.34: Maps and cross-section showing the structure and properties of the final model.
4. Density Model

Mean crustal density (inc. sediment) vs. Mean crustal density (excluding sediment)

Figure 4.35: Mean density of the final crustal model. The left-hand image includes sediments in the average, the right-hand image is for the crystalline crust only.

(0.061 g/cm³ for the whole model space, i.e. including regions of padding) and drops a little over 15% if sediments are removed from the calculation, to 0.054 g/cm³ (0.046 g/cm³ for the whole model space). The main reason sediments dominate the lateral velocity variation but do not dominate the density variation is that the difference in density between sediment and basement is small compared to the difference in velocity between sediment and basement. In addition to this, the velocity-density relationship becomes more sensitive with depth (Table 4.1) counteracting the reduced velocity variation in the lower crust.

For most of the region covered by the final model the Moho structure is largely unchanged from the interpolated Moho presented in the previous chapter (Fig. 4.23). For much of the padded region of the model the Moho has been significantly altered, but for the central model space only the north Bay of Biscay Margin, Porcupine Trough and Bank and the Shetland region have been significantly changed. In these areas the Moho has gained more structure and the modelled variation echos the bathymetric contours to a high degree (Fig. 4.34).

On a local-scale the most striking feature is the density structure of the Rockall Trough
region where strong lateral variation is modelled (Fig. 4.34, upper and lower crust densities). The region has densities up to 2.97 g/cm³ in the upper crust and 3.10 g/cm³ in the lower crust with standard deviations of 0.089 g/cm³ and 0.095 g/cm³ in the upper and lower crust respectively.

The crust beneath the other sedimentary basins shows different density structure for each of the basins. In the Faroe-Shetland basin the densities of the upper and lower crust are both low compared to the surrounding regions, whereas the Porcupine Basin and Western Approaches Trough show no significant density difference between the sedimentary basins and the surrounding basement highs (Fig. 4.34). In the North Sea, both the upper and lower crust show suppressed densities under the Central and Moray Firth Grabens (as low as 2.67 g/cm³ in the upper crust and 2.75 g/cm³ in the lower crust), but relatively high density in the Viking Graben and in the regions of the basin, flanking the grabens, effected by only the thermal subsidence (reaching 2.81 g/cm³ in the upper crust and 3.07 g/cm³ in the lower crust (see Fig. 4.34). Beneath the Møre and Voring basins the crustal density is highly variable (standard deviations of 0.058 g/cm³ in the upper crust, compared to 0.042 g/cm³ for the whole of the final model, and 0.085 g/cm³ in the lower crust, compared to 0.075 g/cm³ for the whole of the final model) and generally increases to the north (Fig. 4.34).

The upper crust in the onshore regions of Britain and Ireland is relatively constant density. However, there are low density regions associated with the granites of the Scottish Highlands, the Leinster Granite and both the onshore and offshore granites of SW England.

In the lower crust, another local feature of note is the low density lower crust to the west of the Outer Isles where densities as low as 2.74 g/cm³ are modelled in the lower crystalline layer (Fig. 4.34).

Local regions of relatively high density lower crust are seen in regions where magmatic underplating has been suggested, such as along the northwestern margin of the model and under the Irish Sea (Fig. 4.34). However, these anomalies are of a similar, or sometimes lower, magnitude then variations in regions where underplate has not been postulated, such as the northwestern side of the Central Graben in the North Sea and the Rockall Bank.

A local scale feature of very different character is the absence of crust along a section
4. Density Model

Figure 4.36: a) Map of crustal thickness in the Bay of Biscay region, with area covered in (b) marked by a bold box and the region of crust <1 km thick marked by a thick dashed line; b) Depth to the seismic reflector marking the top of an anomalous velocity body (from Thinon et al., 2003).

of the Biscay continental margin, mentioned in Section 4.3. Here the observed strong negative gravity anomaly is not compatible with the thick sediment and normal oceanic crust. Although this region is largely in the model padding and therefore only weakly constrained, it is notable that it coincides with a previously identified anomalous area. The NORGASIS wide-angle seismic profile (Thinon et al., 2003) imaged a body with anomalous seismic velocities (higher than standard ocean crust and lower than standard mantle) in this region. The top of this body coincides with a reflector seen in the normal incidence data that can be traced across a broad swath of the ocean continent transition (Fig 4.36).

Given the similarity of the seismic and magnetic structure of this region to the exhumed mantle on the West Iberian margin the body is interpreted to represent altered mantle (Thinon et al., 2003) and therefore the reflector defining the top of the body also marks the base of the crust. This regional reflector is show in Figure 4.36, the correlation between the shallowest regions of the reflector and the region of model with less than 1 km of crust is striking. Therefore, the total removal of oceanic crust may well be reasonably accurate (however the model would still contain inaccuracies in that any serpentinised mantle has not been accounted for).
4. Density Model

<table>
<thead>
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<th></th>
<th>median</th>
<th>mean</th>
<th>standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
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<td>2.832</td>
<td>0.0434</td>
</tr>
<tr>
<td>CRUST2.0 (exc. sediment)</td>
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<td>2.873</td>
<td>0.0290</td>
</tr>
<tr>
<td>The final model (inc. sediment)</td>
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<td>2.807</td>
<td>0.0636</td>
</tr>
<tr>
<td>The final model (exc. sediment)</td>
<td>2.848</td>
<td>2.843</td>
<td>0.0540</td>
</tr>
</tbody>
</table>

Table 4.4: Distribution of mean velocity values in CRUST2.0 (Bassin et al., 2000) and the new crustal model (not including padding).

4.4.1 Comparison with existing models

CRUST2.0

The most widely used crustal model is the CRUST2.0 global 2° model of Bassin et al. (2000). To assist in comparison between the new model and the CRUST2.0 model, the Moho depth and the average density structure of the two models are shown together in Figure 4.37. The Moho depths are generally very similar; the only significant regions of mismatch falling in the Rockall Trough, the Hatton Bank and the north Atlantic. In the Rockall Trough the mismatch appears to be generated by an anomalously deep (greater than 30 km) cell in the CRUST2.0 model. The difference in the Hatton Bank region is due to the CRUST2.0 model defining this as an oceanic area. In the North Atlantic the CRUST2.0 model has less Moho relief and is generally deeper than the new model. This area of the new model is built entirely from gravity modelling; therefore the Moho depths are uncertain.

As with the comparison of velocities in the previous chapter, the most striking difference between the density structure of the models is the significant refinement in cell size for the new model and the far greater detail that results. Summary statistics for the density structure of the two models show some differences, with the new model having a slightly lower density and being more variable (Table 4.4). Although they have been removed to the model padding, it is worth noting that the new model has lower density crust under Scandinavia and has higher density crust in northern France and the Netherlands than CRUST2.0. Although not important to the crustal model, these two differences will affect the longer-wavelength mantle structure under the continent.
Figure 4.37: Comparison of CRUST2.0 (Bassin et al., 2000) (lower 2 rows) and the new model (top row).
Regional Moho maps

The most recently published regional maps of crustal thickness and Moho depth reviewed in Chapter 1 are compared to the best-fit Moho depths of the new model. The crustal thickness/Moho depth maps are expected to show some deviation from the Moho used in the model as these maps are generally hand contoured (rather than computer interpolated) and often based on normal-incidence reflection data as well as wide-angle data. In the southern 2/3\textsuperscript{rd} of the model space the most up-to-date map is the Dèzes and Ziegler (2001) compilation, while for northern Fennoscandia the Kinck et al. (1993) Moho depth map is the most recent (Fig 4.38). Generally the Moho used in the crustal model is very similar to the published Moho maps, particularly to the west of Britain and under the central and southern North Sea. However there are some notable differences. Northern France is deeper in the new model than in the Dèzes and Ziegler (2001) map; however, this area has been reclassified as model padding and therefore does not affect the final crustal model. The northern North Sea is also significantly different, with the Dèzes and Ziegler (2001) map predicting significant Moho relief under the Shetland Platform and Viking Graben. This Moho structure is not seen in the model as there is no wide-angle seismic data across this region to image such relief. However, crustal scale normal incidence data in the region does suggest that there is significant Moho uplift under the Viking Graben. Therefore, this area of the model would benefit from further modelling using the normal incidence seismic data. The third area of difference is Scandinavia; here the model has shorter wavelength variation on the Moho (as a result of the gravity modelling) than is imaged by the smoothly contoured Kinck et al. (1993) map. At longer wavelengths, the Moho depths are generally similar, but with the model's Moho often a kilometer or two shallower. The exception is the deep area near the Lofoten Islands where the modelled Moho is approximately 10 km deeper. However, as with northern France, Scandinavia is padding in the density model and not included in the velocity model. In the offshore region, under the Vøring Plateau, the Mohos are almost identical.
Figure 4.38: Comparison of the new crustal model with existing regional Moho maps.

(a) The new crustal model

(b) Dézes and Ziegler (2001)

(c) Kinck et al. (1993)
4.5 Summary

The construction and development of a new crustal density model has been described in this chapter. The model has a simpler structure than the velocity model of the previous chapter, with four laterally varying layers; two defining the density and thickness of the sediments and two defining the density and thickness of the crystalline crust.

The starting model was built from the crustal velocity model, using the interpolated Moho and velocity structure to define the densities and model boundaries. This model was extended to form a regular rectangular grid by adding padding to the edges of the model. In the continental regions the padding was constructed from the CRUST2.0 global crustal model, while in the oceanic region a generalised structure of North Atlantic oceanic crust was developed and used to construct the padding. The model was also extended vertically to include the mantle lithosphere, with the thermal and density structure of the upper mantle built using the plate cooling model in the oceans and an assumed lithosphere thickness in the continental regions.

The starting model produced a free air gravity anomaly that was significantly different from the observed anomaly and was therefore developed to remove this discrepancy. This development was undertaken as a two stage process with the mantle densities altered to remove the high-amplitude long-wavelength anomaly prior to correcting the short-wavelength misfit. The shorter wavelength residuals were removed through a combination of changes to the upper and lower crustal densities and the Moho depths. Where available external information was used to assign variation between the densities and Moho structure. Elsewhere change was divided equally or assigned so as to reduce the lateral complexity of the model.

The final model has been used to refine the velocity model presented in the previous chapter, with the optimised Moho incorporated into the velocity model. The altered densities were not used to adjust the velocity structure as there is greater flexibility in the velocity–density relationship than the magnitude of the density change. However, the density structure of the model has been described and is discussed in a geological context in the next Chapter. The density model also provides information on the isostatic compensation mechanisms that support topography and bathymetry in the region and this is described
in Chapter 6.
Chapter 5

Geological Interpretation

The main interest in the final crustal velocity and density models is in the investigation and interpretation of regional similarities, differences and trends. As the velocity and density models are constructed from published 2D and local 3D models, with the separation between the input seismic models generally greater than the distance the velocity structure can be confidently extrapolated, there is only very limited opportunity for the models to provide new information on specific local features. However, by combining the local datasets, longer-wavelength features, general trends and differences between regions can provide new information on the nature of the crust in northwest Europe.

5.1 Relationship to Near-Surface Geology

The first stage in discussing the final models is to tie-in the surface geology to the deeper velocity and density structure. To investigate the relationship between the near-surface geology and the modelled structure a simplified geological map of the region has been constructed, with a representative age for the near-surface geology assigned to each grid of the model space (Fig. 5.1(a)). This map was constructed from the regional maps of Sigmond (2002); British Geological Survey (1996); Chantraine et al. (1996) and Naylor et al. (1999).

It was noted in Section 3.5 that the velocity structure of the near-surface crystalline crust is well correlated with the surface geology. With the Carboniferous units in the Midland Valley and across Ireland showing lower near-surface velocities than the areas of
(a) near-surface geological age defined on the scale of the crustal model.

Figure 5.1: Comparison of geological age with mean crustal properties. For the graphs the number of cells is given at the base of the column and the dotted line marks one standard deviation in the value.
Britain with near-surface metamorphic basement (Fig. 3.19). The Welsh basement, the granites in SW England and northern Scottish Proterozoic regions show relatively high velocity and the Archean units of the Outer Isles record the fastest values at sea level, up to 6.58 ± 0.70 km/s. In this section these observations are extended to include trends in the deeper structure/whole crust and in the density model.

To complement visual comparison of the model features and the geological map, plots recording the mean velocity and density for each near-surface age have been constructed (Fig. 5.1).

As a result of the construction of the starting density model from the velocity model the two models show similar relationships between the near-surface geology crustal structure. For both models the Tertiary and younger regions are generally low velocity and density, but are also the most diverse cells, containing some relatively high-valued regions. This is a record of the mixed nature of the crust in this group. In some regions, such as the North Sea and the Rockall Trough, the Tertiary material is the top unit of a thick, low velocity and low density sedimentary succession; however, in other regions, such as the Faroe Islands, the Tertiary units may be extrusive volcanic rocks topping a crust containing large volumes of high velocity and high density volcanic material. Regions where older units are seen in the near-surface have either not formed sedimentary basins or these basins have been partially/completely inverted. Hence the relatively weakly compacted low velocity and density sediments are not seen and the mean velocity and density of the crust tend to be slightly elevated relative to the Tertiary and younger regions.

Where Palaeozoic units dominate the near-surface, the crust is generally relatively high density and velocity, the exception being a few low velocity regions with Ordovician surface rocks, located along the Norwegian coast. The lower Palaeozoic units of Britain are exposed due to basin inversion and denudation. It has been suggested that this exhumation was caused, at least in part, by Tertiary magmatic underplating (Tiley et al., 2004; England et al., submitted). While the model cannot provide causative information, it does indicate a geographic correlation between high velocity, high density crust and exposed lower Palaeozoic strata. Although it must be noted that the older units are likely to have increased velocity and density irrespective of the presence or absence of magmatic underplate, as any older units of sedimentary origin are likely to have undergone substantial
5. Geological Interpretation

Burial, compaction and possibly metamorphism.

Areas of the model where the crust is predominantly of Pre-Cambrian age, particularly northern Scotland and Northern Ireland, appear to differ from the Palaeozoic regions, showing average velocity for the model but relatively low density (Fig. 5.1). This could indicate a different deep crustal composition for the oldest units; but, given the small number of cells and the limited resolution of the model, this difference cannot be stated with any confidence.

5.2 Magmatic Underplating

One of the interesting features of the crust in northwestern Europe is the presence and distribution of magmatic underplating. The seismic profiles across the northwestern continental margin generally image a high velocity (>7.2 km/s), lens shaped body just above the seismic Moho, believed to be magmatic material trapped at, or near, the base of the crust. An additional high velocity body, interpreted to be underplate, has been imaged under the Irish Sea and the Southern Uplands of Scotland (Al Kindi et al., 2003). The new crustal models provide the opportunity to map zones of high velocity and density throughout the region and therefore produce 3D estimates of the distribution and volume of magmatic underplate. However, an investigation on the Vøring margin into the relationship between the high velocity and density bodies and the surface rifts suggests that these anomalous bodies may be more complicated, including a significant quantity of high-pressure metamorphic basement as well as igneous material (Gernigon et al., 2003, 2004). Therefore, it cannot be assumed that all high velocities and densities indicate underplate. However, isolating such regions does provide the data upon which such discussions can be based.

As the velocity model has a finite element structure isolating high velocity regions is a straightforward procedure. The thickness of crust with a velocity greater than 7.2 km/s is shown in Figure 5.2. However, the simplified structure of the density model results in the smearing and averaging of any high density regions across the model layer. Therefore, in order to extract an estimate of the quantity of high density material in the crust this smearing must be counteracted. To resolve this problem the coarse assumptions were
made that standard lower crust has a maximum density of 3.0 g/cm³ and any high density material, e.g. underplate, has a density of 3.1 g/cm³. These assumptions allow a very approximate estimation of the thickness of high density material from areas where the model density is between 3 and 3.1 g/cm³ (Fig. 5.2).

The high velocity regions are quite limited and form isolated patches along the north-western continental margin, more so than the high density regions which are more continuous along the margin. This difference is an artifact of the method of approximating the high density regions as the gravity modelling generally resulted in lower densities along the margin than predicted by the velocities, rather than higher densities (Fig. 4.22). The only exception was along a relatively short section of continental margin to the southwest of the Faroe Islands, where the mean density was increased and therefore a greater thickness of high density material should be estimated. In the original 2D seismic models the high velocity bodies are common to almost all profiles that cross the Hatton and Voring continental margins and along the Faroe-Iceland Ridge (Morgan et al., 1989; Planke et al., 1991; Eldholm and Grue, 1994; Barton and White, 1997; Mjelde et al., 1997, 1998; AMP Exclusive Report 99/3/1, 1999; Mjelde et al., 2001; Raum et al., 2002; AMP Exclusive Report 02/3/1, 2002); therefore, the high velocity body might be expected to be continuous along the margin in the interpolated velocity model. That it isn’t can be explained by the
Figure 5.3: Thicknesses of crust with velocity of 7.0 km/s or greater.

use of automated interpolation to build the velocity model. The underplate bodies on the 2D seismic lines are usually quite limited in geographical extent and, to reduce contouring artifacts in the model as a whole, the interpolation is required to take data from at least 4 octants when estimating a cell property. These two factors combine to produce kriging estimates that often use a combination of data points within and outside the underplate; resulting in estimates of lower velocities than expected for underplate. The velocities in the model are elevated along the margin with respect to the adjacent model space, but not elevated to velocities high enough to predict underplate, as is seen by plotting the thickness of crust with velocity greater than or equal to 7.0 km/s (Fig. 5.3).

Other regions of the model also record high velocities and densities (Fig. 5.2). In the Irish Sea and across the Southern Uplands both values are high (although the high densities are more sparse than the high velocities), consistent with the previously published interpretation of underplate in the region (Al Kindi et al., 2003). Under Ireland the gravity modelling reduced the density; therefore the final model exhibits normal densities but high velocities, associated with the Moho transition zone. In the padded regions of the model high densities are seen under northern France and the Baltic Shield. However, as these regions have not experienced recent volcanism the high values are more likely to be
generated by a dense, high velocity mineral assemblage in a metamorphic lower crust than magmatic underplate. Other isolated regions of high density are seen in the eastern and northern North Sea, particularly to the east of the Central Graben (Figs 5.2 and 4.34). The Moho in the northern North Sea, under the Viking Graben, is likely to be deeper in the model than is truly the case (see Section 4.4.1) and this is probably in part responsible for the increased crustal densities. However, the change in velocity and density across the North Sea is a feature common to all the 2D seismic profiles that cross the Central Graben (Barton and Wood, 1984; Abramovitz and Thybo, 1998, 2000, Chapter 2, this work). This change may represent a difference between the lower crust of the Baltic Shield and Caledonian crust of Avalonia/Laurentia (Abramovitz and Thybo, 1998, 2000; Nielsen et al., 2000).

5.3 Trends in the Crustal Models

5.3.1 Velocity and density variation with crustal thickness

Density and seismic velocity at any spatial location in the crust are controlled by the composition, temperature and pressure at that position. Temperature and pressure both increase with increasing depth through the crust and have a profound effect on the velocity and density. Increases in pressure and temperature have an almost immediate effect on the microcrack structure and elastic moduli, which can be measured in laboratory conditions. It is well documented in rock physics literature that velocity and density increase rapidly for pressures equivalent to the top ~5–10 km of continental crust as a result of microcrack closure (e.g. King, 1966; Nur and Simmons, 1969; Christensen and Mooney, 1995; Khaksar et al., 1999). At greater depths density continues to increase, but at a slower rate, and velocity gently reduces with depth through the crust as the pressure and temperature cause changes in the elastic moduli (Fig. 5.4). In addition to the change in velocity and density caused by changes in the elastic moduli, pressure and temperature increase can cause a far more significant change in velocity and density through recrystallisation in the rock and changes in mineral assemblage (Fig. 5.5). Therefore, the change in mean velocity and density may show a coherent relationship with the crustal thickness, without requiring changes in bulk composition, as increasing pressure in thickened crust may result
Figure 5.4: Variation in density and velocity with depth. Empirical data with pressure and temperature estimates for a normal geotherm. Data from Christensen and Mooney (1995).

Figure 5.5: Modelled changes in stable mineral assemblage for a typical lower crustal mafic granulite and the effect on the density of the rock. Left image, density profile of the rock down to 2 GPa. Right image, mineral assemblage for equilibrium along a geotherm reaching 300°C at 2 GPa. The vertical axis is the same scale for both images. Redrawn from Jull and Kelemen (2001). Legend: ol, olivine; cpx, clinopyroxene; opx, orthopyroxene; plg, plagioclase; gt, garnet; ky, kyanite; qtz, quartz.
Figure 5.6: Change in mean velocity and mean density with thickness for the crystalline crust of the final models. The left-hand image shows the velocity-thickness relationship for the final model; the central image is the density-thickness relationship for the final model; and the right-hand image is the density-thickness relationship for the final model including the model padding. The bold line in the right-hand figure show a possible trend with gradient $0.0054 \text{ g cm}^{-3} \text{ km}^{-1}$.

in significant changes in mineral assemblage.

The velocity-crustal thickness relationship and the density-crustal thickness relationship for the final models are shown in Figure 5.6, the data are for the crystalline parts of the crust to remove the effects of large differences in composition and compaction between the sediments and the basement. The velocity-thickness data show no trend, this suggests that regional variations in bulk composition and/or variation in water content restrict any regionally coherent changes in mineral assemblage with increased temperature and pressure. The density-thickness data show no correlation for thicknesses less than $\sim$20 km; however at greater thicknesses there may be a weak trend of increasing density with increasing crustal thickness (Fig. 5.6). This possible trend is statistically very weak, with the correlation coefficient between density and crustal thickness (for crust greater than 20 km thick) only 0.314 for the final model and 0.490 for the final model and the continental padding.

The geographical distribution of the regions that create this weak trend is shown in Figure 5.7. The trend is formed by the low-lying land and shallow marine regions, areas that are generally unexceptional in other respects. The areas of particularly high or low density and the regions of magmatic underplate generally plot away from the trend.

If this trend is genuine, it is for a higher rate of change than for increasing density
due to simple compression with constant mineral assemblage, which produces an increase of ~0.001–0.0015 g cm\(^{-3}\) km\(^{-1}\) (Fig. 5.4); it must have some compositional/mineralogical element. Therefore, the trend suggests a partial coherence between composition and crustal thickness. Possible explanations for this would include:

**Mineral assemblage change with depth.** The lithology of the lower crust is undoubtedly very heterogeneous as is indicated by the wide variety in composition of lower crustal xenoliths from Scotland and Ireland, which range from felsic granulites to mafic garnet granulite to plagioclase and pyroxenite cumulates (e.g. Halliday et al., 1993; Downes, 1993; Anderson and Oliver, 1996; Upton et al., 1998; Downes et al., 2001; Upton et al., 2001). However, if pressure and temperature conditions produce a change in equilibrium mineral assemblage below ~20 km, which results in significant increase in density for a reasonably prevalent rock type, then this may produce a trend of increasing density strong enough to show despite the complexity of the lower crust. The prime candidate would be garnet growth; however, it must be noted that garnet granulites are extremely rare in the Scottish and Irish xenolith suites (Halliday et al., 1993; Downes, 1993).
Crustal thinning mechanisms that preferentially remove denser crust. If the region where the trend applies is envisaged as having once been relatively thick and higher density than at present, but which has since been thinned in such a way as to reduce the average density as well as the thickness, the trend of reducing density with thinning crust could be formed. Mechanisms such as lower crustal flow, depth dependent stretching and lower crust delamination could have this effect.

Lower crustal flow has been proposed as the explanation for the lack of a crustal root below the Caledonian and Variscan mountain belts in Scotland and continental Europe (Kusznir and Matthews, 1988). If crust that was previously thickened has been thinned by lower crustal flow, the removal of dense lower crustal material could result in a combined reduction in mean density and crustal thickness.

Depth dependent stretching has been observed on the continental margins, where greater extension is seen in the lower crust and whole lithosphere than in the upper crust (Davis and Kusznir, 2004); however there is no evidence in the continental rifts of greater extension of the lower crust than for the crust as a whole (Davis and Kusznir, 2004). Depth-dependent stretching has therefore only been observed in the regions of the model that do not show any relationship between density and crustal thickness (Fig. 5.7).

Lower crustal delamination has been suggested to account for the high Mg to Fe ratio (Mg #) of global continental crust (e.g. Kay and Kay, 1991; Jull and Kelemen, 2001). The removal of dense lower crust would have the effect of both thinning and reducing the mean density of the crust and could therefore produce the trend seen on Figure 5.6. Delamination (or lower crust convective instability) requires the lower crust to be denser than the mantle and the viscosity of the mantle (and crust for convection) to be low enough to allow flow. These conditions are not readily met in the lower continental crust. Modelling suggests that convective instabilities can only occur where there is a mafic lower crust and Moho temperature are in excess of 700°C (although high strain rates can make instabilities more likely); therefore, delamination probably only occurs under arcs, young volcanic margins, orogens and continents underlain by a mantle plume (Jull and Kelemen, 2001).
Figure 5.8: The effect of erosion to sea level on mean density and crustal thickness. The two outer columns represent crustal sections at some arbitrary earlier time and may or may not be in local isostatic equilibrium. The two inner columns represent the same two sections after erosion and attainment of local isostatic equilibrium, the pressure at the compensation depth is the same for the two columns but, as the mean density is different, the Moho depth/crustal thickness is different.

A trend in the model is formed by a geographically widespread region, not limited to the ancient Caledonide and Variscide mountains nor to the northwest volcanic margin and the regions most affected by the Iceland plume (Fig. 5.7). It is therefore unlikely that the trend was formed by lower crust delamination.

Crustal thickening mechanism that adds denser material. The trend of increasing mean density with increasing crustal thickness could be generated if variable amounts of high density material have been added to the crust in regions where the trend is seen. The prime candidate for such a process is magmatic underplating; however as previously noted, the areas associated with underplating do not generally fall on this trend. The density of such underplated regions is normally higher than predicted by the mean density-crustal thickness trend.

Erosion, combined with local isostasy, can result in the coherent relationship between density and crustal thickness. Much of the region that follows this trend is relatively close to sea level (Fig. 5.7) and therefore thickness variation can generally be viewed as variation in Moho depth. If the area is in hydrostatic equilibrium then the thickness of the crust at sea level is a measure of the density contrast between the mantle and crust (Fig. 5.8). The change in thickness of the crust at sea level, $\Delta T$, caused by a
change in mean crustal density between two crustal columns may be written as:

\[ \Delta T = T_1 \left( \frac{\rho_m - \rho_1}{\rho_m - \rho_2} - 1 \right) \]  

(5.1)

where \( \rho_1 \) and \( \rho_2 \) are the mean density of the two regions and \( T_1 \) is the thickness at sea level of the first column. This equation does not produce a linear relationship between crustal thickness and density, but can produce very steep gradients in the density-crustal thickness plot. For example with \( \rho_1 = 2.85 \text{ g/cm}^3 \), \( \rho_2 = 2.90 \text{ g/cm}^3 \), mantle density of 3.33 g/cm\(^3\) and \( T_1 = 31.5 \text{ km} \), the change in crustal thickness at sea level would be 3.662 km and the gradient 0.0137 g cm\(^{-3}\) km\(^{-1}\) (compared to the apparent trend of \( \sim 0.0054 \text{ g cm}^{-3}\text{ km}^{-1} \) in the density model).

If the trend in crustal thickness and mean density shown in Figure 5.6 is genuine, then the most likely causes are changes in mineral assemblage with increased pressure and temperature within the thickened crust and/or erosion.

5.3.2 Velocity and density variation with sediment thickness

It was noted in Sections 3.5 and 4.4 that the velocity and density structure of the crystalline crust appears to be lower under the North Sea rifts and Faroe-Shetland Basin than the basement highs, while the crust beneath the Rockall Basin is variably both high and low velocity and density with respect to its surroundings and beneath the Voring Basin is generally relatively high velocity. These observations are tested by plotting the sediment thickness against the mean velocity and density of the crystalline crust (Fig. 5.9). The first observation of these plots is that, with the possible exception of the velocity structure of the Rockall Basin, the model properties show no correlation with sediment thickness: the mean velocity and mean density beneath the deepest regions of the sedimentary basins are no different from those beneath the shallower regions on the basin margins. The Rockall Basin appears to be different, with an increasing velocity beneath the deeper parts of the Basin. This plot does not clearly show the velocity and the density in the adjacent basement highs, this is better judged using Figures 3.19–3.21 and 4.34–4.35. However, it can be seen on Figure 5.9 that the mean velocity of the crust beneath the North Sea Basin and the Faroe-Shetland Basin is lower than the mean velocity for the model in
Figure 5.9: Relationship between sediment thickness and mean velocity and density of the crystalline crust. Legend: cor, correlation coefficient; V.B., Vöring Basin; N.S.B., North Sea Basin; R.B., Rockall Basin; P.B., Porcupine Basin; F-S.B. Faroes-Shetland Basin. Grey dots indicate data in areas of the model not highlighted on the location map in the top right corner; red indicates the North Sea Basin; dark blue indicates the Rockall Basin; green the Porcupine Basin; pink the Faroe-Shetland Basin; and light blue the Vöring Basin.

general (6.443 km/s), supporting the earlier observation that the velocity beneath these two basins is lower than the velocity of the surrounding basement. What might cause this difference is not clear. The geographical location of the basins may have been controlled by rheological differences between the highs and the areas now containing sedimentary basins; however for this to show in the velocity structure of the crust would require the rheological differences to be related to variation in bulk compositions or lithology. Such differences are possible, but very difficult to test and would also be expected to show in the density structure, but differences between the basement highs and sedimentary basins are less developed in the density model. If the differences between the basement highs and the crust beneath the basins were due to differences in pressure and temperature beneath the basins a correlation between the crustal properties and the sediment thickness would be expected, but is not generally seen (Fig. 5.9).

The Vöring Basin has immensely scattered velocity and density values; almost the entire range of velocity and density seen in the model is present in this area. As a result there is no correlation between properties and sediment thickness. This scatter arises as the crust under the basin contains highly variable amounts of high velocity and density magmatic underplate and there is often only a thin layer of crust between the underplate and the sediments.
For the Rockall Basin the trend of increasing velocity with increasing sediment thickness is generated by the difference between the northern and central parts of the Rockall Trough (see Figs 3.21 and 4.35 for velocity and density structure and 4.34 for sediment distribution). The high velocities in the central Rockall Trough are generated by the combined effect of greater thinning of the upper crust than the lower crust and a Moho transition zone (O'Reilly et al., 1995). The lack of trend between model properties and sediment thickness beneath the other basins suggests that they have not formed with differential stretching between the upper and lower crust.

5.4 Effects of the Iceland Plume

The mantle density structure of the starting model was constructed to represent the temporal evolution of mantle composition from the Proterozoic to the Palaeozoic (with Scandinavia assigned reduced densities with respect to Palaeozoic continental Europe) and the predicted cooling of the oceans with age (Fig. 4.11(a)). However, in the final model the density of the mantle was further reduced under the North Atlantic and the northwestern continental margins (Fig. 4.11(c)). The likely cause of this anomaly is the Iceland Plume, with two effects of the plume potentially lowering the density. Firstly, in the areas of the mantle currently affected by the thermal anomaly of the plume, thermal expansion will cause a reduction in density. Secondly, the extraction of melt from the mantle lowers the Fe/Mg ratio in the residue and hence causes a permanent reduction in the mantle density (Oxburgh and Parmentier, 1977), this second effect may account for the reduced density under the continental margins.

5.5 Summary

The near-surface layers of the velocity model are well correlated with the near-surface geology. For the crust as a whole, differences in the mean density and velocity of the crust are seen between areas with different aged near-surface rocks. The crust beneath areas with outcrops of Tertiary or younger units is highly variable in both mean velocity and density, reflecting the presence of deep sedimentary basins and of crust containing Tertiary magmatic material. The lower Palaeozoic rocks of Britain generally crop out in inverted
sedimentary basins and are underlain by relatively high velocity and density crust, which may be related to Tertiary magmatic underplating.

The models do not add new information on the distribution of underplating in northwest Europe. The interpolation method used to build the velocity model rapidly reverts to mean values away from control data points and therefore tends to predict velocities that are lower than those characteristic of underplate, even in areas where underplating is likely to occur and input data has relatively good coverage such as the northwest continental margin.

There may be a weak trend linking increased density with increased crustal thickness for crust greater than ~20 km thick. If this trend is genuine, then the most likely causes are changes in mineral assemblage with increased pressure and temperature within the thickened crust and/or erosion of surface topography. The main sedimentary basins in the region show some features that differ from basin to basin. The Rockall Basin exhibits a trend, which is not seen in the other basins, of increasing velocity with increasing sediment thickness. The trend is formed as sediment thickness is greatest in the central part of the basin where a combination of a Moho transition zone and differential stretching, preferentially removing the upper crust, produce higher velocities. The crust beneath the Faroe-Shetland Basin is relatively low velocity and density compared to the surrounding basement highs; however, the cause of the low values is not clear and the properties show no correlation with increasing sediment thickness.
Chapter 6

Isostasy in Northwest Europe

The new high resolution model presented in this work provides the opportunity to investigate the support of topographic load in Britain and Ireland and bathymetric features of the surrounding marine basins. It is well established that long-wavelength topographic features are maintained through hydrostatic equilibrium and short-wavelength features are either partially or totally supported by the lateral strength of the lithosphere (e.g. Watts, 2001; Turcotte and Schubert, 2002). The rheology of the lithosphere controls the degree of lateral support afforded to topographic loads. As the amount of lateral support, generally envisaged as the flexure of an elastic plate, is dependent on the rheology the degree of flexure generated by a known topographic load can be used to investigate the lithospheric strength. The relationship between load and flexure can be quantified using the degree of compensation $C$, which is the ratio of observed deflection of the lithosphere to the maximum deflection, i.e. for hydrostatic equilibrium. For a periodic topographic load applied to constant density crust and mantle the degree of compensation is given by

$$C = \frac{(\rho_m - \rho_c)}{\rho_m - \rho_c + \frac{D}{g} \left(\frac{\lambda}{\lambda}\right)^4}$$

(6.1)

where $\rho_m$ and $\rho_c$ are the density of the crust and mantle, $D$ is the flexural rigidity of the lithosphere and $\lambda$ is the wavelength of the topographic load (Turcotte and Schubert, 2002). A degree of compensation of 1 indicates the crust is in local isostatic equilibrium and 0 indicates the topography is entirely supported by the lateral strength of the lithosphere.

There is considerable ongoing debate as to the flexural rigidity (commonly expressed as
6. Isostasy in Northwest Europe

Figure 6.1: Variation in the degree of compensation with wavelength of topographic load and elastic thickness of crust/lithosphere. Values calculated for a crustal density of 2.8 g/cm³, mantle density of 3.3 g/cm³, Young's Modulus of 70 GPa and Poisson's Ratio of 0.25.

In northwest Europe stated values of elastic thickness for the marine sedimentary basins are generally less than ~15 km (e.g. Daly et al., 2004), with the exception of the North Sea where values range from less than 5 km to more than 70 km (e.g. Barton and Wood, 1984; Pérez-Gussinyé and Watts, 2005). Elastic thicknesses determined for onshore Britain and Ireland vary hugely, from less than 5-10 km to well in excess of 70 km (e.g. Tiley et al., 2003; Pérez-Gussinyé and Watts, 2005), see Chapter 1 for further detail. Such a range of values has an enormous effect on the mechanism of isostatic compensation (i.e. hydrostatic or flexural support) for most wavelengths relevant to geological features (Fig. 6.1) and imply very different rheologies for the lithosphere.

In this chapter the new density model is used to estimate the degree of compensation for a range of wavelengths and the results are discussed with regards to the elastic thickness of the lithosphere and the isostatic compensation mechanism for the topography/bathymetry. The approach used here differs from the majority of recent work in that the elastic strength of the lithosphere is estimated from the relationship between the density structure of the
crust and topography, rather than the relationship between the gravity field and topography. Thus this work side-steps the controversy that has engulfed admittance and coherence methods.

6.1 Methodology

The elastic thickness has been estimated using a similar approach to that of Barton (1992). Barton used a density model of the LISPB seismic profile to calculate the topography that would be expected were the region in local isostatic equilibrium. By comparing the true topography to this predicted topography she estimated the degree of compensation of features of various wavelengths along the profile; concluding that the partial compensation of the Southern Uplands and almost complete compensation of the Grampian Highlands requires an elastic thickness of less than 5 km. Here this approach is extended to apply to the 3D density model and to investigate a range of wavelengths in all locations.

To estimate the hydrostatic topography (i.e. the topography expected were the lithosphere in local isostatic equilibrium) the pressure at the base of the density model (200 km below sea level) was calculated after reducing the top of the model to sea level. Where there is positive topography the mass of the land above sea level is not included in the calculation of pressure and where the surface is below sea level the bathymetry is filled with material of density 2.67 g/cm$^3$ (Fig. 6.2). The mean crustal density is used to calculate the pressure at the base of the crust. For regions of positive topography it is assumed that the mean density of the crust is not affected by excluding the material above sea level. The variation in pressure is then used to calculate the surface topography required to remove the pressure differences at the base of the model, assuming the topography has a density of 2.67 g/cm$^3$ and is overlain by air.

The assumption that all topography is overlain by air is clearly erroneous. However, it has the advantage that the calculated surface relief is solely dependent on the variation in pressure at the base of the model and not on the absolute value of pressure. If the presence of water is included in the calculation then the location of sea level relative to the variation in topography becomes relevant to the amplitude of the topography (for any given change in pressure at the base of the model, the required change in bathymetry overlain
by water is greater than the required change in topography overlain by air). If the position of sea level is known relative to the estimated dry hydrostatic relief, then the relief can be adjusted to give the true hydrostatic topography/bathymetry by increasing the bathymetry to remove the volume of crust equivalent to the mass of water added. Alternatively, to make the amplitude of dry hydrostatic relief equivalent to observed bathymetry, the observed values can be converted to dry topography by adding the mass of water to the crust. This second approach is preferred as the sea level is known relative to the observed topography/bathymetry, whereas it can only be estimated by use of an assumed datum for the calculated hydrostatic relief. The topography and bathymetry data used for the observed values are the Smith and Sandwell (1997) bathymetry, derived from satellite altimetry and the GTOPO30 onshore topography data. With bathymetry derived from gravity data there is the potential for considerable error in the values, particularly in shallow marine areas or in the presence of thick sediment accumulations. To minimise these errors Smith and Sandwell (1997) use ship-track depth soundings to constrain the bathymetry where such data are available and to construct the gravity–topography transfer function in the areas proximal to the data. The coverage of ship-track data for the modelled region is generally good, with only the North Sea (which has very little relief) suffering from a lack of data (Fig. 6.3) and so the bathymetry is reliable for the modelled area.
Although the amplitudes of the observed dry topography (i.e. the topography and bathymetry after removing the mass of water) and the dry hydrostatic topography (i.e. the calculated surface relief without the mass of water included) are comparable, acquiring the absolute value of the hydrostatic topography at any given location still requires equating it to a reference datum. Although the methodology used here to investigate isostasy has been developed specifically for this project, numerous previous studies of isostasy have required a means of relating predicted and observed topography. Various reference points have been used in these previous investigations including the free asthenospheric surface (Lachenbruch and Morgan, 1990; Zoback and Mooney, 2003), the density structure of old oceanic lithosphere (Kaban et al., 2003) and for the calculation of the expected topography on the LISPB profile Barton (1992) used an arbitrary point at the end of the profile. Here the mean value of the hydrostatic topography has been set to minimise the rms misfit with the observed dry topography for the geographical area of the final model, not including the model padding (Fig. 6.4). However, to remove the significance of the datum the two surfaces are only compared using spectral analysis of the relative amplitudes of the topography. However, it is apparent from Figure 6.4 that there is considerable mismatch on the short wavelengths whilst on a regional scale the observed and hydrostatic topography are very similar.
To calculate the degree of compensation the amplitude of the hydrostatic topography is compared to the amplitude of the observed dry topography. If the model is in local isostatic equilibrium then the two values will be the same. If there is an element of lateral support then the amplitude of the observed topography may be greater or less than the hydrostatic topography (Fig. 6.2). The compensation can therefore be calculated as the fraction of dry topography, $t_{obs}$, explained by the hydrostatic topography, $t_{hydro}$: (Fig. 6.5):

$$C = \begin{cases} 
\frac{t_{obs}}{t_{hydro}} & \text{for } t_{hydro} > t_{obs}, \\
\frac{t_{hydro}}{t_{obs}} & \text{for } t_{hydro} \leq t_{obs}
\end{cases}$$

(6.2)

By definition the flexure generated by a periodic load must be the same wavelength and phase as the load. Therefore, in estimating $C$ it is fundamental that the dry topography is
Figure 6.6: Illustration of the calculation of the degree of compensation. Example waves for the observed and hydrostatic topography are shown on the left of the figure. Through Fourier analysis these waves are separated into cosine (real) and sine (imaginary) components. The compensation of the cosine phase is then calculated from the ratio of the real components and the compensation of the sine phase from the relative amplitudes of the imaginary components. The overall degree of compensation, $C$, is then calculated by weighting the real and imaginary compensations by their relative amplitude and summing.

only compared to the amplitude of hydrostatic topography that is in phase with it. Any out of phase component must be the result of lateral support. In order to calculate the in phase components of the two surfaces the Fourier transform of each was calculated, using the 2D algorithm “fourn” (Press et al., 2003). The real component of the transforms is the amplitude of the cosine phase of the surface and the imaginary component the amplitude of the sine phase. Thus, by transforming both the observed dry topography and hydrostatic topography, both the real and imaginary components of the two transforms must be either in phase or $\pi$ out of phase, in which case the amplitude of one transform will be positive and the other negative (Fig. 6.6). The compensation of the cosine phase of the topography is calculated from the relative amplitudes of the real components and the compensation of the sine phase from the amplitudes of the imaginary components. If either the real or imaginary components of the two transforms are $\pi$ out of phase the compensation for
that component is 0. The overall degree of compensation is then calculated by multiplying the two individual components by their relative amplitude and summing. The relative amplitude is determined from the magnitude of the denominators in the calculation of the real and imaginary components of $C$:

\[
\begin{align*}
    re &= \begin{cases} 
        \Re\{|T_{\text{hydro}}|\} & \text{for } \Re\{|T_{\text{hydro}}|\} > \Re\{|T_{\text{obs}}|\}, \\
        \Re\{|T_{\text{obs}}|\} & \text{for } \Re\{|T_{\text{hydro}}|\} \leq \Re\{|T_{\text{obs}}|\}
    \end{cases} \\
    im &= \begin{cases} 
        \Im\{|T_{\text{hydro}}|\} & \text{for } \Im\{|T_{\text{hydro}}|\} > \Im\{|T_{\text{obs}}|\}, \\
        \Im\{|T_{\text{obs}}|\} & \text{for } \Im\{|T_{\text{hydro}}|\} \leq \Im\{|T_{\text{obs}}|\}
    \end{cases} \\
    W_{re} &= re/(re + im) \\
    W_{im} &= im/(re + im) \\
    C &= C_{re}W_{re} + C_{im}W_{im}
\end{align*}
\] (6.3)

where $|T_{\text{hydro}}|$ indicates the amplitude of the transformed hydrostatic topography, $|T_{\text{obs}}|$ indicates the amplitude of the transformed observed dry topography, $\Re\{}$ indicates the real component of the transform, $\Im\{}$ the imaginary component and $C_{re}$ and $C_{im}$ the compensation calculated for the cosine and sine phases respectively (Fig. 6.6).

The degree of compensation was calculated for topography with a wavelength of 160, 320, 640 and 1280 km. These wavelengths were chosen as they can be sampled regularly by the 40 km cell size of the model and furthermore these wavelengths are $2^n$ grid cells ($n = 4, 8, 16$ and 32 respectively). The hydrostatic and observed topography surfaces were filtered to the chosen wavelength (to remove any problems with spectral leakage for the transform) and broken into windows with dimensions equivalent to one wavelength. The Fourier transform of the window was then used to obtain a degree of compensation for that subsection of the model. Thus, by moving the location of the window variation in the degree of compensation and therefore the rheology of the lithosphere were mapped. To improve the robustness of the calculated $C$ values, section averaging was applied. Any given cell in the model can be included in a number of different windows, each of which will estimate a different value for the degree of compensation (Fig. 6.7); therefore, the model cells are assigned the mean degree of compensation of all possible windows containing
Figure 6.7: Example of the calculation of degree of compensation, $C$, for a specific location and wavelength. The two larger images show the observed and hydrostatic topography, filtered to the wavelength of interest (320 km in this example, with a cosine tapered filter cutting wavelengths above 440 km and below 200 km and passing only 320 km), with an individual cell singled out by the black square. The degree of compensation for the cell can be calculated from any window that includes the cell, e.g. the red and blue boxes. The four smaller maps show the windows marked by the blue and red boxes and below these images is the degree of compensation calculated from the window and the weighting assigned to this estimate.
that cell. This effectively works as a running mean, reducing the spatial resolution of the calculation, but increasing the robustness. A weighted average of $C$ values was also calculated with the compensation of each window weighted by the sum of the power of the two topographic surfaces defining $C$. The justification for this weighting is outlined in the discussion of errors below.

### 6.2 Errors in Calculating the Degree of Compensation

The three possible sources of error in calculating $C$ and estimating the elastic thickness from the compensation are: 1) Errors in the density model, leading to error in the estimated hydrostatic topography; 2) Spatial averaging caused by the need to calculate $C$ for an area covering one wavelength; and 3) error in the theoretical model used to convert $C$ into an estimate of elastic thickness.

#### 6.2.1 Errors in the density model

The accuracy of the density model is of fundamental importance as any errors will map directly into the calculation of the hydrostatic topography and hence introduce error into the calculated value of $C$. Unfortunately the uncertainty in the density model is not known and is difficult even to estimate. The original density model was constructed from a velocity model that is generally quite poorly constrained (Section 3.5) using a poorly constrained velocity–density conversion (Section 4.1.1). Propagation of errors would therefore suggest the density model is very weakly constrained. However, the requirement that the density model produces a gravity anomaly that matches the observed anomaly removes some of the uncertainty, although the model is still far from unique.

The significance of potential error in the density model can be investigated through the difference between the calculated hydrostatic topography and the observed topography (Fig. 6.8). The difference at the longest wavelength is a record of the DC offset between the surfaces and has no physical meaning as the zero contour of the hydrostatic topography was fixed to minimise the difference between the two surfaces for the main model area. For the other wavelengths the difference between the surfaces is fairly constant at $\sim 60–80$ m, even though the amplitude of topography increases with wavelength (Fig. 6.8). Although
Figure 6.8: Comparison of the observed dry topography and the calculated hydrostatic topography. The 3 maps show the topography surfaces and the difference (calculated by subtracting the hydrostatic topography from the observed topography) scaled in kilometers. The two spectra show the magnitude of the difference in topography and the amplitude of the observed topography, in meters.
this difference takes no account of the phase of the surfaces or variation across the model, it does indicate that the calculation of $C$ is based on very small differences between the two surfaces; a change in the predicted hydrostatic topography of only a few 10's of meters could vastly change the difference between the topographies and hence the value of $C$. Given a density contrast between the topography and air of 2.67 g/cm$^3$ a change of 70 m in the hydrostatic topography would require a change in the average density of 30 km thick crust of only 0.006 g/cm$^3$, a far smaller difference than the uncertainty in density. However, although the gravity anomaly is not used directly in the calculation of $C$, the density model is required to produce a gravity anomaly that is comparable to the observed anomaly and this does provide some restriction on the possible hydrostatic topography. For example the 0.006 g/cm$^3$ change in density for a cylinder with a radius of 40 km extending from sea level to 30 km would produce a ~5 mGal change in gravity anomaly (Turcotte and Schubert, 2002, Equation 5-112), a similar magnitude to the misfit of the final density model and the observed gravity field. Therefore, although the change of 0.006 g/cm$^3$ in mean density is within errors predicted by the velocity-density conversion, the magnitude of the gravity anomaly it generates causes it to be close to the maximum acceptable change. Even so, a change of ~70 m could theoretically result in the removal of all the difference between the hydrostatic topography and the observed topography, or a doubling of the difference, thus potentially creating highly erroneous $C$ values.

However, in calculating $C$, the phase of the topography is taken into account and the amplitude is measured over one wavelength for each window, with the final $C$ value taken from the average of several windows. If the errors in the density model are incoherent with the observed topography they will produce increased $C$ for some windows and reduced $C$ in others, resulting in highly variable $C$ values for overlapping windows not coherently changed $C$ values for the windows. Highly variable $C$ for overlapping windows will result in a highly uncertain calculation of the degree of compensation; in effect recording the errors in the density model. It is therefore important to know whether errors in the density model are coherent with the topography.

To investigate this the coherency between the starting density model and the topography and between the changes in density during the modelling and the topography were calculated. Although this is not a direct measure of the errors in the density, if both the
starting model and the changes in density are incoherent with the topography then there is no reason why errors in the model should be coherent. However the coherence is highly variable for all wavelengths, both for the starting model and for the changes applied to the density structure during development of the model (Fig. 6.9). It is therefore possible that there are some errors in the density model which are coherent with the topography and may produce a biased estimation of $C$.

To further investigate the potential for false measures of $C$, the degree of compensation calculated from an alternative density model has been compared to $C$ from the final density model. The hydrostatic topography and $C$ (for a single wavelength) have been calculated for one of the alternative density models discussed in Section 4.3. This model has constant crustal density, the final model’s sub-crustal density structure and Moho optimised to remove the misfit between the observed and calculated gravity anomaly (Fig. 6.10). Despite the very large change in the density model the difference in the calculated values of $C$ is relatively small. Within the area covered by the final model (i.e. not including the padded regions) the maximum difference is -0.23 (at the southwest end of the Rockall Trough) and the rms difference in only 0.067. The only area of widespread and relatively large difference is in the oceanic lithosphere of the north Atlantic, particularly associated with Iceland and the Faroe-Iceland Ridge, where the alternative density model is highly inconsistent with the seismic constraint and therefore not truly a viable alternative model (Section 4.3). This result suggests that the requirement that the density model fits the observed gravity anomaly has the result that errors in the density model do not have as significant an impact on $C$ as suggested by the initial discussion on errors based on the difference between the observed and hydrostatic topographies.

As $C$ is calculated for specific wavelengths any variation in the density uncertainty with wavelength must be considered, not just the absolute uncertainty in the model. As discussed in the previous chapters, the short wavelength structure in the velocity model (and therefore the starting density model) is generated by the interpolated structure from the wide-angle seismic profiles and this interpolation rapidly loses constraint away from the input data. Therefore, in the velocity model, the short-wavelength structure is generally poorly constrained. The long-wavelength structure is the result of variation between the seismic profiles and therefore the long-wavelength trends in the velocity model are
Figure 6.9: The coherence between the topography and the crustal density structure. The coherence is shown for 4 profiles running at different azimuths though the centre of the model. For each profile a quad of plots and a location map are shown: the top left plot shows the profiles in the spatial domain (the left-hand axis gives both density of the starting model and change in density, ticks are at 0.05 g/cm³); the top right plot shows the power spectra for the three profiles. The bottom left plot gives the coherence between the starting model and the topography and the bottom right the coherence between the change in density and the topography. The black line in the top left plot and the black data points in the top right plot show the true topography data; the blue line and blue data points show the starting density model; and the red line and red data points show the change in density.
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Figure 6.10: Comparison of the degree of compensation for topography with 320 km wavelength for two density models.
more robust. However, as discussed above much of the constraint on the density model comes from the gravity modelling rather than being inherited from the velocity model. For the starting density model the gravity residual was dominated by long-wavelength misfit (Fig. 4.10). The smaller misfit for shorter wavelengths suggests that the method of constructing the model (through a single velocity-density conversion) is more accurate at short-wavelengths. Therefore, although the velocity model is better constrained at long-wavelengths, the starting density model may be more accurate at short-wavelengths. As the optimisation of the density model was highly non-unique, if shorter wavelengths are more accurate for the starting model they will also be less uncertain for the final model. This suggests that, with respect to the uncertainty inherited from the density model, the estimates of $C$ based on the shorter wavelengths may be more accurate than those based on the longer wavelengths. Furthermore, the investigation into the coherency between topography and the density (Fig. 6.9) suggests that the shorter wavelength (< 200 km) contain slightly less coherent structure than the longer wavelengths. Therefore, although short wavelengths may still contain some errors that are coherent with topography, they appear to be a little less likely to produce false estimations of $C$.

For any errors present in the density model that are independent of the surface topography and bathymetry the uncertainty in $C$ will be independent of the amplitude of topography. As a result, errors present in the density model will have a more significant effect on the estimation of $C$ in low amplitude regions then in high amplitude regions (Fig. 6.11). To accommodate this amplitude dependent variation in uncertainty, a weighted degree of compensation was calculated from the windows contributing to a single value of $C$. The

Figure 6.11: Cartoon illustrating the reduced effect of errors in the density model in high amplitude regions. The black curves represent the hydrostatic topography in two regions, the grey area the uncertainty inherited from the density uncertainty.
weighting was assigned using the power of the two topographies of the windows contributing to a single value of $C$, with greater importance assigned to $C$ calculated from windows with high amplitude in both the observed and hydrostatic topographies. Therefore, the weighted $C$ is skewed towards the more robust measures of compensation.

6.2.2 Spatial averaging

The calculation of $C$ as a single value for an area covering one wavelength of topography results in averaging any variation in properties on a sub-wavelength scale. This may cause errors when using the longer wavelengths as the area being used to calculate $C$ is likely to straddle a variation in tectonic settings and is therefore likely to contain variable strength. This is an unavoidable problem that affects all attempts to estimate the elastic thickness using any spectral method.

6.2.3 Errors in the theoretical model

The advantage of using the approach developed here to estimate the elastic thickness is that the theoretical model is relatively simple and, by using pressure rather than gravity anomaly, avoids the complications associated with depth of loading that effect gravity based models. However, the theoretical model still assumes fixed values for the Young’s Modulus, Poisson’s ratio, crustal density and mantle density. These simplifications of the true situation introduce some errors into the conversion of $C$ to an estimate of the elastic thickness. Fortunately, even a fairly substantial change in one of the parameters produces only a relatively small change in the theoretical curve for the elastic thickness (Fig. 6.12). Therefore, although variation in the parameters controlling the elastic thickness prediction are not accounted for, the errors introduced will be insignificant compared to the uncertainty in estimating $C$ from the density model.

6.3 Results

The degree of compensation is shown for the final density model (including the model padding) for wavelengths of 160, 320, 640 and 1280 km in Figure 6.13. The maps show that $C$ increases with increasing wavelength of topography, from a mean value of 0.243 for
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Figure 6.12: The effect of changing the controlling parameters in the relationship between degree of compensation and elastic thickness. The solid line on the plots shows the relationship used for all later plots and dashed lines shows the effect of altering one of the parameters. In the top left the mean density of the crust is changed, in the top right the mean density of the mantle varied, the bottom left shows the effect of changing Young’s modulus and the bottom right Poisson’s ratio.

160 km wavelength topography to a mean of 0.769 for 1280 km wavelength topography. However, there is also considerable lateral variation within each wavelength, especially at the shorter wavelengths (Fig. 6.13). The standard deviation in $C$ is 0.142 for 160 km wavelength, with values within the main model areas ranging from 0.005 in the central North Sea to 0.695 on the Porcupine Margin. For $C$ calculated using a 1280 km wavelength the range in values within the main model area is reduced to 0.188 (from 0.668 in the southern North Sea to 0.856 on the Vøring Margin). The reduced variation at long-wavelengths is the inevitable effect of the increased spatial averaging generated by the bigger windows needed for the longer wavelengths.

The spatial distribution of $C$ varies for the different wavelengths, regions with lower $C$ than their surroundings in the calculation based on one wavelength do not necessarily have lower values than their surroundings for the other wavelengths (Fig. 6.13). For example, the northeast Rockall Trough has very low $C$ at 160 km wavelength (less than 0.2), but higher values than its surroundings for 320 km wavelength topography. Some features, such as the relatively low values in the central North Sea/Denmark/southern Scandinavian region, are
Figure 6.13: Maps of the degree of compensation for given wavelengths. The colourwash and black contours give $C$ and the white contours indicate the coastline and bathymetry at 1 km depth intervals.
more persistent; however even these are not reproduced on the 1280 km wavelength map. In general the values of $C$ calculated using 160 km and 320 km show better correlation than the other wavelengths. These differences are reduced when using the weighted values of $C$ (Fig. 6.14); however, the differences in spatial distribution are still considerable.

The values of $C$ for each wavelength can be converted into estimates of elastic thickness by re-arranging Equation 6.1 and substituting $D$ into Equation 1.3 (Fig. 6.15). The first observation of the $T_e$ estimates is that the values are different for each wavelength and generally increase with increasing wavelength: the mean $T_e$ (including the padded regions) is $\sim$13 km, $\sim$16 km, $\sim$32 km and $\sim$63 km for wavelengths of 160 km, 320 km, 640 km and 1280 km respectively. Some specific areas have similar estimates of elastic thickness for two of the wavelengths, e.g. Edoras Bank and the Trøndelags Platform (near the Norwegian coast) for 160 and 320 km wavelengths and southernmost Sweden for 640 and 1280 km wavelengths (Fig. 6.15). The south central North Sea is the most consistent area with $T_e$ estimates $\sim$30 km for topography of wavelengths 160 km, 320 km and 640 km. However, most areas show increasing values of $T_e$ at longer wavelengths. The elastic thickness values also inherit the inconsistent spatial distribution across wavelength seen in the $C$ values. However, this calculation of $T_e$ contains no data on uncertainty in $C$. A degree of information in the uncertainty can be introduced by recording the variability in $C$ for each window used to calculate $C$ for a specific location (Fig. 6.16). The error bars in Figure 6.16 represent the standard deviation in $C$ calculated for each of the possible windows that contain the model cell. Therefore, they are a representation of the local variability in the calculated $C$. The error bars do not explicitly contain information on the uncertainty in the density model and how this maps into the calculated values of $C$ for any one of the windows. However, the variation in $C$ for the windows is a combination of genuine lateral variation and variation generated by errors in the density model, as changes in the amplitude of topography between windows will change the importance of any errors enhancing the variation (Fig. 6.11). Therefore, the error bars do contain some implicit information on the errors in the density model; however, they represent minimum estimates of uncertainty. The results of these plots (Fig. 6.16) generally agree with the results of the $T_e$ maps (Fig. 6.15) in that the elastic thickness estimates increase with increasing wavelength and the standard deviation in $C$ from the individual wavelengths is insufficient
6. Isostasy in Northwest Europe

Figure 6.14: Maps of the degree of compensation, $C$, for given wavelengths. Values are weighted to give greater preference to calculated values of $C$ based on high amplitude topography. The colourwash and black contours give $C$ and the white contours indicate the coastline and bathymetry at 1 km depth intervals.
Figure 6.15: Maps of the elastic thickness calculated for a given wavelength. The elastic thickness was calculated from the weighted mean $C$, using $E = 70$ GPa, $v = 0.25$, $\rho_m = 3.30$ g/cm$^3$ and $\rho_c = 2.80$ g/cm$^3$. The black contours give $T_e$ at 5 km intervals and the white contours indicate the coastline and bathymetry at 1 km depth intervals.
Figure 6.16: Degree of compensation calculated for specific wavelengths for the Voring Margin, the Rockall Trough, Central England and the southern North Sea. The circles and error bars give the mean and standard deviation of $C$ calculated for every possible window sampling the cell. The star gives the weighted mean, with the weighting calculated from the power of the observed and hydrostatic topography. The curved lines give the theoretical relationship of $C$ with wavelength of topography for a range of elastic thicknesses and $E = 70$ GPa, $v = 0.25$, $\rho_m = 3.30$ g/cm$^3$ and $\rho_c = 2.80$ g/cm$^3$. 
to accommodate the disparity. The mismatch between 160 and 320 km is generally less than the error bars and for some locations a single value of $T_e$ can fit all wavelengths up to 640 km, e.g. the southern North Sea (Fig. 6.16); however, the 1280 km wavelength data is always inconsistent with the shorter wavelengths.

The optimum $T_e$ value was calculated by minimising the misfit between the calculated $C$, for each wavelength, and predicted $C$ for different elastic thicknesses:

$$H = \left[ \frac{1}{N} \sum_{n=1}^{N} \left( \frac{C_o - C_p}{\Delta C_o} \right) \right]^{1/2} \quad (6.4)$$

where $H$ is the misfit, $N$ is the number of wavelengths used, $C_o$ is the observed/calculated $C$, $C_p$ is the $C$ predicted for a given elastic thickness and $\Delta C_o$ is the standard deviation in $C_o$. This is an equivalent calculation of misfit to that used by McKenzie (2003) in optimising admittance estimates. As the longest wavelength persistently predicts stronger lithosphere than the three shorter wavelengths, the optimisation was performed twice, once using all four $C$ values and once using only the degree of compensation calculated for the three shorter wavelengths. Generally, the difference in optimum $T_e$ for the two calculations is very small (Fig. 6.17).

In keeping with the previous observations, this analysis shows that the value of optimum $T_e$ is very poorly constrained, with large changes in elastic thickness causing little change in misfit (Fig. 6.17). This is the case for optimisation using just the 3 shorter wavelengths or all 4 wavelengths; however, the optimisation is a little more sensitive to variation in elastic thickness when using only 3 wavelengths (Fig. 6.17).

The variation in optimum elastic thickness can be mapped across the area (Fig. 6.18) along with the values of $T_e$ at which the misfit increases to twice that of the optimum value. This is an arbitrary choice of cut-off intended only to illustrate the variation in constraint across the model. The maps of optimum $T_e$ are relatively similar for both calculations (all wavelengths or only the wavelengths < 640 km) but show very sudden steps from $T_e \approx 20$ km to $T_e \approx 40$ km around the Shetland Islands and further north and west (Fig. 6.18). These steps occur as a result of the limited number of wavelengths in the optimisation (the elastic thickness predicted by the 320 km wavelength topography is $\sim 20$ km and $T_e$ predicted by the 640 km wavelength topography is $\sim 40$ km. The optimal
Figure 6.17: Examples of the calculation of optimum $T_e$ for four locations. Legend for the plots of misfit: solid red line = misfit for wavelengths $< 640$ km; dashed red line = misfit for all wavelengths; solid, vertical black line = optimum $T_e$ for wavelengths $< 640$ km; dashed black line = optimum $T_e$ for all wavelengths. Legend for the plots of degree of compensation against wavelength: solid red line = optimum $T_e$ for wavelengths $< 640$ km; dashed red line = optimum $T_e$ for all wavelengths; other symbols the same as Figure 6.16.
Figure 6.18: Maps of optimum $T_e$. The two central maps show the optimal elastic thickness calculated using either all wavelengths (upper image) or only wavelengths 160–640 km (lower image), elastic thickness is contoured at 5 km intervals. The smaller maps show the value of $T_e$ for which the misfit is twice the misfit of the optimal $T_e$, for these maps the elastic thickness is contoured at the same intervals as the scale bar.
value jumps between the two depending on the exact value and standard deviation for 160 km (Fig. 6.17(a) and (b)). The range of $T_e$ that can be accommodated without the misfit exceeding twice the minimum is far greater than the variation in optimum $T_e$ across the model; the median values for these minimum and maximum $T_e$ estimates are 6 and 53 km (for optimisation using 3 wavelengths) compared to a standard deviation for the optimum $T_e$ of 11.6 km (Fig. 6.18). For much of the model the misfit for a $T_e$ of either 0 km or 160 km does not exceed twice the minimum misfit (shown by black and white areas on Figure 6.18), indicating that for these locations either the minimum or maximum elastic thickness is effectively unconstrained. This is the result of the inconsistent $T_e$ for the 4 wavelengths investigated, which produces a very poorly constrained optimum.

6.4 Discussion

With discrepancies in the estimates of $T_e$ for the different wavelengths it is impossible to draw any strong conclusions on the strength of the lithosphere. However, it is possible to discuss the potential causes for the discrepancies between the wavelengths and to compare the results to previously published work on the elastic thickness of the region.

The inconsistent values of $T_e$ for different wavelengths may be the result of errors in the density model or in the Fourier transform of the topography surfaces. The effect of errors in the density model were discussed above and it was noted that small errors in the density model have a potentially significant effect on the calculation of $C$.

The Fourier transform itself is also inexact as it requires the data to be periodic. The windowing of the data forces appropriate periodicity onto the data, but numerous studies have shown that the results of the transform are dependent on the method of windowing and the size of the window (e.g. Simons et al., 2000; Ojeda and Whitman, 2002). The approach taken here is equivalent to taking a hanning window of the data; which has been shown to be quite susceptible to corrupting the spectrum (Ojeda and Whitman, 2002). However, usually all frequencies in the transform are used, whereas in this work only the wavenumber that is naturally periodic within the window and is unaffected by the filtering is used; therefore the transform should be relatively robust. However, information on other wavelengths, that may help constrain $T_e$, is lost. Therefore, it may be a profitable
avenue of future work to investigate alternative approaches to windowing the data, such as multi-taper windowing or alternatives to Fourier analysis such as wavelet transforms.

The approach used to estimate the elastic thickness is implicitly based on the assumption that the lithosphere is in isostatic equilibrium. However, the region investigated here is still undergoing a degree of dynamic adjustment following the Late Pleistocene deglaciation of northern Europe. GPS surveys of Britain and Fennoscandia indicate that there is ongoing vertical movement (e.g. Milne et al., 2001; Teferle et al., 2002), with uplift rates of up to 1 cm/year in northern Sweden (McConnell, 1968; Milne et al., 2001). There may be as much as 175 m of incomplete topographic adjustment in Sweden and Finland (Fig. 6.19). There is no information available on the unrecovered adjustment in the rest of the area; however, as the original ice load in Scotland was smaller than the Scandinavian ice sheet and present day vertical adjustment in Britain and Ireland is far smaller than in Fennoscandia, the unrecovered glacial adjustment is likely to be far less. Spectral analysis of uplift data from Scandinavia indicates that shorter wavelength isostatic anomalies decay more rapidly than longer wavelength anomalies (McConnell, 1968; Wieczerkowski et al., 1999; Klemann and Wolf, 2005); therefore the unrecovered topography is dominated by long-wavelengths (Fig. 6.19). The unrecovered topography will not affect the calculation of the hydrostatic topography (as it is based on the mass between sea level and the base of the density model); however it does effect the observed topography. The difference between the observed dry topography and the hydrostatic topography is 10's to hundreds of meters, depending on the location and wavelength (Fig. 6.8). Therefore, in Scandinavia and for the long-wavelength calculations, $C'$ is derived from topographic differences of similar scale to
the potential error in topography generated by incomplete glacial isostatic adjustment. As
the incomplete glacial adjustment is likely to be more important at the long-wavelengths
than the short-wavelengths, the unrecovered topography may have a significant role in the
inconsistent calculation of $C$ for the different wavelengths, effecting the 1280 km calculation
far more than the 160 or 320 km calculations. Pérez-Gussinyé et al. (2004) investigated the
effect of unrecovered post-glacial topography on coherence estimates of the elastic thickness
of Fennoscandia and found that the potential unrecovered topography shown in Figure 6.19
did not effect their elastic thickness estimates. This suggests that coherence based analysis
of $T_e$ is less sensitive to post-glacial adjustment than the method developed here.

Although weakly defined, the optimised elastic thickness estimates can be compared
to the results of previous investigations into the flexural strength of the region. Single
values of $T_e$ for specific areas of the model have been published for the English Channel,
several locations in the North Sea, the Hatton margin, South Wales, Scotland and northern
England, south central England and the whole of the British Isles. These values of elastic
thickness are compared to the optimum value from the 160–640 km topography (and the
$T_e$ for twice the minimum misfit) in Table 6.4. The method used to produce these published
elastic thickness estimates vary, but can be broadly classified as three groups:

**Direct assessment of $C$.** Two of these published elastic thicknesses, Warner (1987) and
Barton (1992), are calculated from a similar approach to the one used here, with
the degree of compensation estimated directly from the crustal model and used to
calculate $T_e$. Warner (1987) uses a very simple model of the crust to estimate $C$ for
the sedimentary basin in the Channel. The model, derived from the BIRPS data,
consists of the sedimentary basin (assumed to be 2.4 g/cm$^3$ with a P-wave velocity
of 4 km/s and two-way travel time to the base of the sediments of ≤~4 s) and the
crust beneath the basin (density 2.75 g/cm$^3$, seismic velocity of 6 km/s and a two-
way-travel-time to the Moho, both beneath the basin and beneath the sediment free
flanks, of ~10 s). This structure suggests a degree of compensation for the 100 km
wavelength basin of ~0.80. The calculation is reasonably insensitive to changes in
the density and thickness of the sediment and crust, hence the published range in
possible $C$ is 0.5–1.0 (equivalent to a $T_e$ of 0–4.7 km). However, the wavelength of the
Table 6.1: Comparison between the optimum $T_e$ from this work and the previous studies that have produced single values of $T_e$. Values in brackets give the $T_e$ for twice the minimum misfit.

<table>
<thead>
<tr>
<th>Region</th>
<th>Reference</th>
<th>Elastic thickness</th>
<th>Elastic thickness from this work</th>
</tr>
</thead>
<tbody>
<tr>
<td>The Channel</td>
<td>Warner (1987)</td>
<td>$&lt; 5$ km</td>
<td>$\sim 25$ km ($\sim 7/\sim 80$)</td>
</tr>
<tr>
<td>Central North Sea</td>
<td>Barton and Wood (1984)</td>
<td>$&lt; 5$ km</td>
<td>$\sim 20-45$ km ($\sim 2-15/\sim 160$)</td>
</tr>
<tr>
<td>Viking Graben</td>
<td>Marsden et al. (1990)</td>
<td>6 km/3 km for Triassic/Jurassic</td>
<td>$\sim 15-30$ ($\sim 8/\sim 160$)</td>
</tr>
<tr>
<td>Danish Central Graben</td>
<td>Korstgård and Lerche (1992)</td>
<td>$\sim 3$ km</td>
<td>$\sim 50$ ($\sim 10/\sim 160$)</td>
</tr>
<tr>
<td>Hatton margin</td>
<td>Watts and Fairhead (1997)</td>
<td>$&lt; 5$ km</td>
<td>$\sim 20$ km ($\sim 5/\sim 40$)</td>
</tr>
<tr>
<td>Scotland and northern England</td>
<td>Barton (1992)</td>
<td>$&lt; 5$ km</td>
<td>$\sim 20$ km ($\sim 10/\sim 80$)</td>
</tr>
<tr>
<td>South Wales</td>
<td>Burgess and Gayer (2000)</td>
<td>$\sim 40$ km</td>
<td>$\sim 20$ km ($\sim 7/\sim 40$)</td>
</tr>
<tr>
<td>south central England</td>
<td>Watts et al. (2000, 2005)</td>
<td>5-10 km</td>
<td>$\sim 20$ km ($\sim 6/\sim 45$)</td>
</tr>
<tr>
<td>The British Isles</td>
<td>Tiley et al. (2003)</td>
<td>$5 \pm 2$ km</td>
<td>$\sim 15-20$ km ($\sim 0-5/\sim 40-80$)</td>
</tr>
</tbody>
</table>
basin is taken from the width on the 2D seismic profile, which crosses approximately perpendicular to the long axis of the elongate basin. Given the elongate nature of the basin (Fig. 3.5) estimating the effective wavelength of load on the lithosphere is not trivial; however, the ~50 km width of the basin (100 km wavelength) is the minimum. It is very possible the effective wavelength of the load is ≥300 km, which would make the degree of compensation of ~0.8 very similar to the estimate based on 320 km topography in this work (Fig. 6.13).

Barton (1992) used a method very similar to the one developed here; comparing the expected hydrostatic topography to the observed topography along the LISPB profile. She determined that the elastic thickness of north Britain is less than 5 km. The process of extracting numerical values for $C$ is not explained in the paper, only a qualitative description of the similarity between the predicted and observed topography. The two estimates of $C$ that lead to the conclusion of $T_e < 5$ km are the values of ~0.2–0.8 for the ~130 km wavelength Southern Uplands and ~0.8–1.0 for the ~220 km wavelength Grampian Highlands. In the absence of an explanation of the method of quantifying $C$ it is impossible to tell to what degree the phase and amplitude of the predicted and observed topography were taken into account. The work presented here shows that visual inspection of similarity and a quantitative analysis can produce different results. For example, a visual comparison of the hydrostatic topography and observed topography shown in Figure 6.4 suggests that the two surfaces are very similar for northern Britain, even on short-wavelengths, suggesting high $C$ values such as those estimated by Barton (1992); however, the $C$ calculated in this work, taking into account phase and amplitude of the signals, is < 0.4 for 160 km wavelength and 0.4–0.6 for 320 km wavelength (Fig. 6.13). Therefore, if Barton (1992) used only a visual comparison of the hydrostatic and observed topography the estimate of elastic thickness may be considerably lower than would be produced quantitatively.

**Forward modelling loading.** The majority of the $T_e$ estimates summarised in Table 6.4 were calculated using some form of forward modelling of the load and flexure. Barton and Wood (1984) and Marsden et al. (1990) modelled subsidence for areas of the
North Sea sedimentary basin using kinematic models of basin formation. The two significant drawbacks to this approach are the reliance on an appropriate kinematic model (which will at best be a simplification of the true process of basin formation) and that the retrieved elastic thickness will relate to the time of basin formation not the present day. Although Korstgård and Lerche (1992) used a different approach to modelling the basin formation - numerically simulating the flexure generated by adding the sedimentary load to a pre-flexed finite elastic plate - this approach will also only retrieve the elastic thickness at the time of loading, not the present day strength of the lithosphere.

For onshore Britain, Burgess and Gayer (2000) investigated the South Wales coal basin through forward modelling the formation of a flexural foreland basin and comparing predicted subsidence of the basin to backstripped subsidence curves. This model has a large number of controlling parameters: load-advance rate; initial and final load width; initial and final distances between the load and the basin; load density; density of displaced material; and elastic thickness of the lithosphere. Many of these parameters are correlated making the solution non-unique; for example, an increase in subsidence could be generated by increasing the load density or decreasing the elastic thickness. Burgess and Gayer (2000) provide one alternative model for a lower elastic thickness (25 km) and show that the weaker lithosphere delays the onset of subsidence in the more distal parts of the basin, reducing the fit between the forward modelled and backstripped subsidence curves. However, the authors do not provide a full analysis of the uncertainty in \( T_e \). They also do not indicate whether (in the single alternative model provided) the backstripped subsidence curves were also recalculated with the alternative \( T_e \), or whether the forward model with weak lithosphere was compared to subsidence curves calculated assuming strong lithosphere. As with the other studies of sedimentary basin formation discussed above, this work will retrieve the elastic thickness at the time of loading (the Carboniferous) not the present day strength.

In a different study of the strength of onshore Britain Watts et al. (2000) and Watts et al. (2005) forward modelled the Quarternary uplift of the hills in south central
England and eastern Wales in response to excavation of the Vales of Evesham and Gloucester. The later paper builds on the earlier by adding further discussion on the timing of the uplift and calculating the flexural response to the unloading in 2D, rather than the 1D approach used in the 2000 paper. Watts et al. (2005) show that the topography predicted by flexural unloading with a $Te$ of 25 km is largely indistinguishable from that predicted by a $Te$ of 5 km, at least for the region proximal to the Severn River Valley. Their $Te$ estimate of 5–10 km is based mainly on the balance between observed and calculated volumes of material excavated from the rivers.

Watts and Fairhead (1997) estimated the strength of the lithosphere on the Hatton continental margin by modelling the gravity anomaly produced by the sum of three components of the margin (the sediments, the crust and the underplate), with flexure caused by the sediment load and magmatic underplate load. The $Te$ controls the degree of flexure and therefore the size of the gravity anomaly resulting from the sediment and underplate. This work is hindered by being one-dimensional, effectively assuming laterally constant crustal structure, sediment thickness and density structure and underplate thickness and density structure, which may introduce uncertainty in the result. Additionally the model uses the same density (2.85 g/cm$^3$) for both the continental crust and the oceanic crust, which is likely to introduce error into the calculated gravity anomaly.

**Admittance analysis.** The last $Te$ estimate in Table 6.4, Tiley et al. (2003), was derived from analysis of the admittance function between free air gravity and topography. Pérez-Gussinyé et al. (2004) and Pérez-Gussinyé and Watts (2005) argue that this method is inaccurate as analytical predictions of admittance are compared to observations acquired from finite windows, they state that the theoretical prediction should be based on finite data. An additional major problem with the use of admittance, or coherence, to investigate the elastic thickness of the continental lithosphere is the complexity of continental crust. In oceans the crust is approximately constant density and thickness for large areas, allowing relatively simple conceptual models to be formed that relate topography to gravity in the presence of isolated deviations.
from the standard crust (e.g. ocean islands). However, in the continents the crust varies hugely in both thickness and density structure and so the conceptual models relating topography to gravity must be made far more complex. Despite the added complications of buried loads and the considerable recent debate on how best to estimate these loads, the conceptual models are still based on highly simplified views of the crust: 1–3 constant density layers with isolated loads located at the Moho, a mid-crustal interface or at the surface. Whereas the model presented in Chapter 4 shows that the continental crust contains significant lateral variation. The elastic thickness estimates retrieved from admittance or coherence methods can only be as good as the conceptual model relating topography and gravity. Furthermore, the theoretical model is assumed to be accurate when calculating the uncertainty in $T_e$ and therefore uncertainties in the theory inherited from the simplified crustal structure do not carry into the elastic thickness uncertainties. These problems effect all estimates of $T_e$ in continents derived from admittance or coherence methods.

Most of these calculations of a single elastic thickness value predict weak lithosphere, with $T_e$ generally less than 10 km, with the exception of the calculation for the South Wales coal basin of ~40 km (Burgess and Gayer, 2000) (Table 6.4). The work presented here has optimal elastic thicknesses of approximately 20 km or greater for the same locations. However, the range of possible elastic thickness generally allows for $T_e$ estimates in this work to approximately match the published values (Table 6.4, Fig. 6.18 for $T_e$ values in general and 6.17(c) for south central England) and the elastic thickness calculation based solely on the 160 km topography is generally comparable to the published estimates for Britain (Fig. 6.15); therefore no significance can be attributed to the difference between the published and optimal $T_e$ estimates in England and Scotland. However, there is a mismatch between these published elastic thickness estimates for the North Sea (<10 km) and the $T_e$ estimate for 160 km wavelength topography, which is generally greater than 15 km peaking at 40 km in the south central North Sea (Fig. 6.15). Even using only the results from the 160 km topography and taking account of the uncertainties in $C$ for this wavelength, the results suggest the elastic thickness is unlikely to be less than 5 km in the North Sea (Fig. 6.16).
In addition to these individual values of elastic thickness, several previous studies have mapped out the variation in $T_e$ across all or part of the region. To the west of Ireland Daly et al. (2004) map $T_e$ variations between 6 and 18 km (using coherence calculated from wavelet analysis), with the weak area centred on the south eastern Rockall Trough/Porcupine Bank and the stronger region near the Hatton continental margin. For the same area the optimised $T_e$ for this work is generally higher, ranging between ~10 and ~33 km, and has a somewhat different distribution (Fig. 6.20). There are a few locations where the elastic thickness estimates are similar (both works predict $T_e \approx 16$ km for the northern Rockall Trough, the northern Rockall and Hatton Banks and the area around the Edoras Bank) or only differ by a few kilometers (such as onshore Ireland where Daly et al. (2004) calculate $T_e \approx 12$ km and this work has a mean $T_e$ of 16 km). However, the calculated $T_e$ values are very different for the northwest and southwest corners of the area (where Daly et al. (2004) predict values of ~14-16 and ~12-14 km respectively and this work calculated $T_e$ values in excess of 25 km) and for the region around the Rockall Trough/Porcupine Bank (where Daly et al. (2004) have their lowest elastic thicknesses of 6–10 km and this work calculates relatively high values of ~16–20 km). However, as noted above the optimal elastic thickness estimates for this work are very poorly constrained and uncertainty can account for the majority of these discrepancies (Fig. 6.16, Rockall Trough plot, and Fig. 6.18). Therefore the $T_e$ values calculated by Daly et al. (2004) are encompassed by the range of values estimated in this work.

In Fennoscandia, Poudjom Djomani et al. (1999) mapped out elastic thicknesses using the coherence method and calculated values between 8 km and 70 km. They predict the strength of the lithosphere increases to the east, with the $T_e$ of Norway less than 20 km and Finland generally greater than 40 km (Fig. 1.12). However, as this area lies in the padding of the density model and the Moho depths are a few kilometers different from the more reliable Kinck et al. (1993) map (Section 4.4.1) the calculated hydrostatic topography is not considered reliable. Therefore, the elastic thickness estimates cannot be considered reliable. As a result there is very little significance in the observation that the elastic thickness estimates for Scandinavia (~15–45 km, Figure 6.18) are very similar to those predicted by Poudjom Djomani et al. (1999).
Figure 6.20: Comparison of the elastic thickness to the west of $T_e$ for all of Europe calculated by Pérez-Gussinyé and Watts (2005) and the work presented here. To assist in comparing the models the optimised $T_e$ for this work has been plotted using the same projection and similar colourwash to the previously published work. The white line on (d) shows the boundaries of the main density model.
The most recent work on the strength of the European lithosphere is that of Pérez-Gussinyé and Watts (2005). Using coherence they map out variation in elastic thickness across the continental lithosphere of Europe and find a strong Pre-Cambrian core to Avalonia and eastern Europe ($T_e > 70$ km) with lower elastic thicknesses associated with the Caledonides of Scotland and Norway (15–25 km) and the Variscan and Alpine orogenic belts (10–50 km) in mainland Europe (Fig. 6.20). Comparing this distribution to the results of the work presented here, there is good correlation in Scandinavia, to the west and northwest of Ireland, along the Norwegian continental margin and near the Alps (Fig. 6.20). However, as Scandinavia and the Alps are in the padded part of the density model there is little confidence in the calculated hydrostatic topography and therefore little significance in the correlation with the Pérez-Gussinyé and Watts (2005) map. The only notable difference between the two calculated distributions is that the work presented here does not predict a strong lithosphere for Avalonia (‘Av’ on Fig. 6.20). However, the uncertainty in the calculated elastic thickness of Britain and the North Sea does allow for values to be similar to those calculated by Pérez-Gussinyé and Watts (2005) (Fig. 6.18). But, as mentioned above, the uncertainty also allows the values calculated here to be consistent with the < 10 km predicted by other studies of Britain. Thus this work provides no further constraint on the previously published range of elastic thickness for northwest Europe.

6.5 Summary

A new approach to estimating the elastic thickness of the continental lithosphere has been developed. This method, an expansion of the work by Barton (1992), calculates the elastic thickness from the relationship between the true topography and the topography that would occur were the lithosphere in local isostatic equilibrium. By measuring elastic thickness directly from the density model, through the calculation of hydrostatic topography, this approach bypasses the current debate over coherence and admittance methods.

The results of elastic thickness calculated using this approach are inconclusive as there is a considerable discrepancy between the strength predicted by short-wavelength topography and the strength calculated using long-wavelength topography. The elastic thickness estimate from topography with 160 km wavelength is broadly consistent with the published
$T_e$'s of less than 10 km for Britain and and 6-18 km for the Irish Atlantic margin. However, the strength estimated using 640 km topography is more in keeping with the strong Avalonian and Scandinavian lithosphere calculated by Pérez-Gussinyé and Watts (2005), but predicts elastic thicknesses of 25-30 km for Britain and the Irish Atlantic. Thus the uncertainty in the optimum elastic thickness for this data is as great as the variation in published $T_e$ values.

The uncertainty in optimum value of elastic thickness is generated by a combination of: errors in the degree of compensation, inherited from the density model and generated through inaccuracy in the Fourier transform; spatial averaging of the lithospheric strength; and incomplete isostatic equilibrium from unrecovered glacial isostatic adjustment.
Chapter 7

Conclusions

7.1 Summary

This work has been directed towards two goals: Firstly, the construction of a new, high-resolution P-wave velocity model for northwest Europe; and secondly, investigation of the strength of the European lithosphere using the density model affiliated to the velocity model.

Prior to building the velocity model the historical and modern approaches to modelling wide-angle seismic reflection data were reviewed as an introduction to wide-angle data and the limitations of the models produced from such data. The review was required as the regional 3D velocity model was built primarily from such data.

The final velocity model was defined on a 40 km by 40 km by 1 km finite element grid; with the grid geographically referenced using a Transverse Mercator projection, with the origin at 3.4° west, 57.15° north. The model extends from the topography/bathymetry surface to the base of the crust, defined using the seismic Moho, and covers a geographical region encompassing Britain, Ireland and the surrounding marine sedimentary basins.

A crustal and upper mantle density model was constructed to provide verification of the velocity model and supplementary information on the crust itself. The density model is defined by five laterally varying layers; two sedimentary, two crustal and one mantle layer.

As well as interpreting the geological significance of the structures resolved in the velocity and density models, the density model was used to investigate isostasy and the
strength of the lithosphere in the region, using a new method to calculate the elastic thickness of the lithosphere.

7.2 Conclusions

The final crustal velocity model provides a useful tool for future work in seismology. The new high resolution model will allow better crustal correction in teleseismic earthquake studies in the region, better location of local earthquakes and provides a good starting model for local and regional tomographic studies.

The velocity and density models also provide information on the structure of the crust, which has been interpreted in relation to the near-surface lithologies and geological history of the region:

- The near-surface layers of the velocity model are well correlated with the near-surface geology, with regions with young sedimentary cover recording low velocity while regions of Palaeozoic basement have high velocity.

- The velocities in the lower crust on the northeast side of the North Sea's Central Graben are higher than those on the southwest, which may be related to the change from crust affected by the Caledonian Orogeny to the Baltic Shield.

- There is a long-wavelength trend from higher velocities in the northwest of the model to lower velocities under continental Europe.

- Differences in the mean density and velocity of the crust are seen between areas with different aged near-surface rocks, with elevated mean velocity and density beneath the inverted Palaeozoic basins and Pre-Cambrian basement with respect to the areas covered by Mesozoic and younger sediments.

- There may be a weak trend linking increased density with increased crustal thickness for crust greater than ~20 km thick. If this trend is genuine, then the most likely causes are changes in mineral assemblage with increased pressure and temperature within the thickened crust and erosion of surface topography.
• The main sedimentary basins in the region show some features that differ from basin to basin. The Rockall Basin exhibits a trend of increasing velocity with increasing sediment thickness, formed as sediment thickness is greatest in the central part of the basin where a combination of a Moho transition zone and preferential extension of the upper crust, result in elevated velocities. The crust beneath the Faroe-Shetland Basin is relatively low velocity and density compared to the surrounding basement highs; however, the cause of the low values is not clear and the properties show no correlation with changes in sediment thickness.

Despite the high resolution, the models do not add new information on the distribution of underplating in northwest Europe. The interpolation method used to build the velocity model rapidly reverts to mean values away from control data points and therefore tends to predict velocities that are lower than those characteristic of underplate, even in areas where underplating is likely to occur and input data has relatively good coverage such as the northwest continental margin.

The investigation into isostasy in the region was based on a new approach using the relationship between the true topography and the topography that would occur were the lithosphere in local isostatic equilibrium. However, the results of elastic thickness calculated using this approach are inconclusive as there is a considerable discrepancy between the strength predicted by short-wavelength topography and the strength calculated using long-wavelength topography. The elastic thickness estimate from topography with 160 km wavelength is broadly consistent with the published $T_e$'s of less than 10 km for Britain and and 6-18 km for the Irish Atlantic margin. However, the strength estimated using 640 km topography is more in keeping with the strong Avalonian and Scandinavian lithosphere calculated by Pérez-Gussinyé and Watts (2005), but predicts elastic thicknesses of 25-30 km for Britain and the Irish Atlantic. Thus the uncertainty in the optimum elastic thickness for this data is as great as the variation in published $T_e$ values.

### 7.3 Future Work

Areas of the work presented here that could be developed in the future are:

• The Moho structure could be further investigated and refined by adding normal
incidence reflection data. These data were not used in the work presented here as the depth conversion of travel times is highly sensitive to the velocity structure, which the normal incidence reflection data do not constrain well. However, now that the velocity model has been constructed this data could be used to depth convert the normal incidence data. This additional data could help resolve structure in the more uncertain areas of the model, such as the Viking Graben in the northern North Sea.

- The velocity model can be developed by adding further wide-angle seismic models as they are published.

- The velocity model could be developed through seismic tomography based on local earthquakes, this would provide an alternative data source to verify and refine the velocity structure. Work in this area is being undertaken, at the University of Leicester, for a region surrounding north Wales. Preliminary results indicate that the structure of this area of the final model is reasonably robust, but some refinement will be possible (Hardwick et al., 2005).

- The estimation of the elastic thickness of the lithosphere could be developed to use more wavelengths through the use of a robust windowing technique such as multitaper windowing or an alternative to Fourier analysis such as wavelet analysis. Investigating the elastic thickness using a greater range of wavelengths may improve the constraint on the optimum value, giving greater confidence in the results.
Appendix A

Wide-Angle Reflection Seismic Modelling Software

A.1 FAST

FAST (First Arrival Seismic Tomography) (Zelt and Barton, 1998) is a 3D or 2D code for velocity modelling the turning rays in wide-angle reflection/refraction seismic data. The code produces a continuous velocity model defined on a regular grid, the node spacing of which is defined by the user. The model is produced iteratively with each step consisting of a forward and inverse step. The forward step calculates ray paths using the Hole and Zelt (1995) modification of Vidale's 1990 finite differencing method of calculating the travel times on an expanding cube. Vidale calculates the travel times using three schemes based on the eikonal equation,

\[
\left( \frac{\delta t}{\delta x} \right)^2 + \left( \frac{\delta t}{\delta y} \right)^2 + \left( \frac{\delta t}{\delta z} \right)^2 = s^2(x, y, z) \tag{A.1}
\]

where \( t \) is the travel time; \( x, y \) and \( z \) are the Cartesian co-ordinates and \( s \) is the slowness. These three schemes use finite differencing to calculate the time to a grid node by using the known travel time to adjacent nodes; the three schemes differ only in the relative positions of the node of unknown time and the nodes with known travel times, as illustrated in figure A.1. The 3 methods are used to propagate the travel times through the model by calculating the times to the nodes on the faces, then edges, then corners of a cube which
is then steadily increased in size until the times for the whole model are calculated. As the travel times are calculated on an expanding cube this method fails when the velocity structure generates head waves that refract so as to travel back towards the source.

The Hole and Zelt modification introduces extra finite difference operators that allow wave propagation along the edges of the grid cells and introduces reverse propagation, by initiating expanding cubes on different walls of the model, to calculate travel times when refractions travel back towards the source (Hole and Zelt, 1995). These improvements allow the calculation of ray paths across high velocity gradients and velocity discontinuities.

The inverse step employed in FAST reduces an objective function that combines travel time residuals and model roughness,

$$\Phi(m) = \Delta t^T C_d^{-1} \Delta t + \lambda \left( m^T C_h^{-1} m + s_z m^T C_v^{-1} m \right)$$

where $\Phi(m)$ is the objective function, $m$ is the model adjustment vector, $\Delta t$ the data residual vector, $C_d$ is the data covariance matrix, $C_h$ and $C_v$ are the horizontal and vertical roughness matrices and $\lambda$ and $s_z$ are the weighting factors. $\lambda$ controls the relative importance of minimising roughness to minimising the travel time residual and $s_z$ the weighting of vertical smoothness compared with horizontal smoothness. The search method used to minimise the objective function is a least-squares variation of the conjugate gradient method (Zelt and Barton, 1998).

User input is required to provide a starting model and set the grid size for the forward
modelling stage. For the inverse stage 3 parameters are needed: 1) the cell size; 2) \( s_z \), the weighting of horizontal to vertical smoothness; and 3) the initial values for \( \lambda \), the parameter controlling the trade-off between model roughness to data misfit. However after supplying these parameters the model development is a fully automatic process requiring no user input to produce the final model.

A.2 RAYINVR

RAYINVR (Zelt and Smith, 1992) is a forward modelling ray tracing code with an associated damped least squares inversion routine. The velocity model is constructed as a series of layers, each of which is defined by one or more depth nodes (for the top of the layer), one or more velocity nodes along the top of the layer and one or more velocity nodes along the base of the layer (the base of the layer is defined by the top of the layer beneath). Nodes are connected by linear interpolation.

Ray tracing is performed by solving Červený’s two dimensional equations (Equations A.3 and A.4) along short steps which combine to make the total ray path. The relative step length is automatically scaled along the ray path so as to be shorter in regions of greater bending, absolute scaling of the step length is defined by the user (Zelt and Smith, 1992). Where ray paths intersect layer boundaries Snell’s law is applied.

\[
\frac{dz}{dx} = \cot \theta, \quad \frac{d\theta}{dx} = \frac{(v_z - v_x \cot \theta)}{v} \\
\frac{dx}{dz} = \tan \theta, \quad \frac{d\theta}{dz} = \frac{(v_x \tan \theta - v_z)}{v}
\] (A.3) (A.4)

where \( \theta \) is the angle between the ray and the \( z \)-axis; \( v \) is the velocity, \( v_z \) and \( v_x \) are the partial derivatives with respect to the \( z \) and \( x \)-axes (Červený et al., 1977). The initial conditions for the equation set \( x \), \( z \) and \( \theta \) to the source locations and the ray take-off angle. When the ray is near horizontal Equation A.3 is solved by integrating with respect to \( x \); when the ray is near vertical A.4 is solved with \( z \) as the integration variable. All or a subset of the rays are traced through the model and the travel time residual and Chi-squared values are then calculated for the ray groups used.

The RAYINVR inversion routine uses a damped least squares approach to minimize the
travel time residuals. A Taylor series expansion about a starting model, with the higher order terms neglected (Equation A.5), forms the basis of the inversion routine.

\[ \Delta t = A \Delta m \]  

(A.5)

where \( \Delta t \) is the travel time residual vector; \( A \) is the partial derivatives matrix, which has dimensions \( M \times N \) (where \( M \) is the number of data points and \( N \) the number of model parameters) and each element is the partial derivative of the travel time with respect to the model parameter, i.e. \( \delta t_i / \delta m_j \); \( \Delta m \) the model adjustment vector, is the difference between the final model and the starting model. Both \( \Delta t \) and \( A \) are calculated during the forward step of \textsc{rayinvr} (Zelt and Smith, 1992).

The model adjustment is determined using a damped least squares solution of Equation A.5:

\[ \Delta m = (A^T C_t^{-1} A + D C_m^{-1})^{-1} A^T C_t^{-1} \Delta t \]  

(A.6)

where \( C_t \) is the estimated data covariance matrix, which is a vector of the square of the travel time uncertainties associated with each pick; \( C_m \) is the model covariance matrix, which is a vector of the parameter uncertainties squared; \( D \) is the damping factor (Zelt and Smith, 1992).

### A.3 TRAMP

The model parameterisation and ray tracing in \textsc{tramp} (Zelt and Ellis, 1988; Zelt and Forsyth, 1994) are the same as in \textsc{rayinvr}, therefore amplitude modelling of the travel time derived velocity models is a simple procedure and the results of the two approaches are easily combined to iteratively improve the velocity model.

The amplitude modelling of diving and reflected rays in \textsc{tramp} is based on the zero-order asymptotic ray theory of Červený et al. (1977). The amplitude of a ray at its endpoint \( (A) \) is given by,

\[ A = \frac{A_0 q (-1)^e}{L} \]  

(A.7)

where \( A_0 \) is the initial ray amplitude, which is set to 1 (therefore all amplitude calculation in \textsc{tramp} are relative values only); \( q \) is the factor controlling the partitioning of energy at
the model boundaries; $L$ is the reduction in amplitude due to geometrical spreading and $(-1)^{e}$ is a correction for the change in direction of the positive SV oscillation. SV is the shear wave that oscillates in the plane perpendicular to the model interfaces; when the ray intersects with a velocity discontinuity the direction of the positive SV oscillation can flip, changing the phase of the arrival by 180°. $e$ is the number of interfaces at which this effect occurs. For MONALISA profile 3 only P wave data are used, therefore $e = 0$.

The factor $q$ is given by

\[
q = \left( \frac{v_0 \rho_0}{v_r \rho_r} \right)^{1/2} \prod_{i=1}^{n} \left( \frac{v_i \rho_i}{v_i' \rho_i'} \right)^{1/2} Z_i
\]  

where $n$ is the number of model boundaries encountered by the ray; $Z_i$ is the Zoeppritz coefficient at the $i$th boundary; $v_0$ and $\rho_0$ are the velocity and density at the source (their product is the impedance); $v_r \rho_r$ is the impedance at the receiver; $v_i \rho_i$ and $v_i' \rho_i'$ are the impedances at the points of incidence and emergence at the $i$th boundary, respectively. The density is estimated from the velocity using either $\rho = 0.252 + 0.3788 v_p$ or $\rho = 1.732 v_p^{1/4}$, the choice being made by the user.

The effects of attenuation can be estimated by assigning varying Q-attenuation values to the model. The assumption is made that the attenuation is independent of frequency, allowing a scale factor $(A_Q)$ to be calculated on the dominant frequency ($\omega$) and applied equally at all frequencies. This assumption is permissible if the energy is concentrated within a narrow band and the attenuation is not too large (Zelt and Ellis, 1988). This scale factor is calculated by,

\[
A_Q = \prod_{i=1}^{N} \exp \left[ \frac{-\omega l_i}{2 v_i Q_i} \right]
\]  

where $N$ is the number of line segments in the complete ray path, $l_i$ is the length of ray path segment $i$, $v_i$ is the average velocity for the ray segment and $Q_i$ is the $Q$-attenuation value for the ray segment.

### A.4 GRAVMAG

GRAVMAG is a 2.5 dimensional interactive gravity and magnetic modelling code built by the British Geological Survey (Pedley et al., 1993). The model is constructed as a series
of polygons in the $xz$-plane with each polygon extended along the $y$-axis to create a 3 dimensional body (Fig. A.2). The length of the body in the $y$-axis is the strike length of the body and is symmetrical about the $xz$-plane and therefore entered as the half strike length ('$Y$' in Fig. A.2). This ability to model bodies with a finite length in the third dimension makes the code more than two dimensional, hence the 2.5 D description. Each polygon and the model background have a density value and magnetic attributes assigned, creating the gravity and magnetic fields.

The vertical component of the gravity field generated by each body is calculated at the origin of the co-ordinate system by summing the contribution of the $N$ vertices of the polygon defining the body (Equation A.10) (Rasmussen and Pedersen, 1979); the origin of the co-ordinate system is shifted across the model to calculate the total field from the body. The total field generated by all the polygons is then simply found by summing the field generated by each body.

$$
\Delta g_z(r_0) = -2YG \rho \sum_{i} \hat{z} \cdot \hat{n}_i \log \frac{u_{i+1} + R_{i+1}}{u_i + R_i} + 2G \rho \sum_{i} \frac{x_i z_{i+1} - z_i x_{i+1}}{\Delta x_i^2 + \Delta z_i^2} \left( \Delta z_i \log \frac{r_{i+1}(Y + R_i)}{r_i(Y + R_{i+1})} + \Delta x_i \left( \tan^{-1} \frac{u_{i+1}Y}{w_i R_{i+1}} - \tan^{-1} \frac{u_i Y}{w_i R_i} \right) \right)
$$

(A.10)
where the location of the polygon vertices are defined in 2 co-ordinate systems; the $x, y, z$ system of the model space and a rotation of this system into a $u, v, w$ system such that $u$ is parallel to the side of the polygon connecting the current vertex to the next one, $v$ is the same as $y$, and $w$ is perpendicular to both $u$ and $v$ (Fig. A.2). The additional parameters in Equation A.10 are as follows: $Y$ is the half strike length; $\rho$ is the density contrast between the body and the model background; $\hat{n}$ is a unit vector normal to the polygon surface and outward-directed; $r_{i+1}$ is the distance from the origin to the next vertex (in the rotated co-ordinates) and $R_{i+1}$ is the equivalent point on the end face of the body i.e. 
$r_{i+1} = (u_{i+1}^2 + w_i^2)^{1/2}$, $R_{i+1} = (u_{i+1}^2 + w_i^2 + Y^2)^{1/2}$; $\Delta x_i$ and $\Delta z_i$ are the distance step in $x$ and $z$ between the current vertex and the next vertex (Rasmussen and Pedersen, 1979).
Appendix B

MONALISA Profile 3 Receiver
Gathers
Figure B.1: Trace normalised amplitude receiver gather for OBH31. Plotted with a 6 km/s reduction velocity and with the picked arrivals (coloured dots) and the calculated travel times (red lines) for the final model overlaid. See Table 2.2 for phase nomenclature.
Figure B.2: Trace normalised amplitude receiver gather for OBH32. Plotted with a 6 km/s reduction velocity and with the picked arrivals (coloured dots) and the calculated travel times (red lines) for the final model overlaid. See Table 2.2 for phase nomenclature.
Figure B.3: Trace normalised amplitude receiver gather for OBH33. Plotted with a 6 km/s reduction velocity and with the picked arrivals (coloured dots) and the calculated travel times (red lines) for the final model overlaid. See Table 2.2 for phase nomenclature.
Figure B.4: Trace normalised amplitude receiver gather for OBH34. Plotted with a 6 km/s reduction velocity and with the picked arrivals (coloured dots) and the calculated travel times (red lines) for the final model overlaid. See Table 2.2 for phase nomenclature.
Figure B.5: Trace normalised amplitude receiver gather for OBH35. Plotted with a 6 km/s reduction velocity and with the picked arrivals (coloured dots) and the calculated travel times (red lines) for the final model overlaid. See Table 2.2 for phase nomenclature.
Figure B.6: Trace normalised amplitude receiver gather for OBH36. Plotted with a 6 km/s reduction velocity and with the picked arrivals (coloured dots) and the calculated travel times (red lines) for the final model overlaid. See Table 2.2 for phase nomenclature.
Figure B.7: Trace normalised amplitude receiver gather for OBH37. Plotted with a 6 km/s reduction velocity and with the picked arrivals (coloured dots) and the calculated travel times (red lines) for the final model overlaid. See Table 2.2 for phase nomenclature.
Figure B.8: Trace normalised amplitude receiver gather for OBH38. Plotted with a 6 km/s reduction velocity and with the picked arrivals (coloured dots) and the calculated travel times (red lines) for the final model overlaid. See Table 2.2 for phase nomenclature.
Appendix C

Wide-Angle Seismic Model

Database

The following table catalogues the wide-angle reflection/refraction data used to construct the crustal velocity and density model. The first column of the table refers to the numbers on Figure C.1; the second column gives the profile name, if one is given in the publications; the third column contains an x if the profile contains velocity data; the fourth contains an x if the profile extends down to the Moho; the fifth column gives an approximate location for the data; and the sixth column gives the reference for the published models.

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### C. Wide-Angle Seismic Model Database

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### C. Wide-Angle Seismic Model Database

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Table C.1: Catalogue of wide-angle reflection/refraction models.

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Figure C.1: Location map for the wide-angle data used to construct the regional crustal model, numbers refer to the first column of Table C.1.
Appendix D

Database Uncertainties

D.1 Moho Uncertainties

The following table catalogues the highest and lowest uncertainties assigned to the Moho input data. The first column of the table refers to the numbers on Figure C.1; the second column gives the profile name, if one is given in the publications; the third column contains the lowest uncertainty assigned to the Moho depth; the fourth column the highest uncertainty assigned to the Moho; and the fifth column the reference for the published models. The uncertainties are in italics if they have been taken from the original publication and in standard font if no uncertainties were provided, or the provided uncertainties have not been used.

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D.2 Velocity Uncertainties

The following table catalogues the uncertainties assigned to the wide-angle reflection/refraction velocity data. The first column of the table refers to the numbers on Figure C.1; the
second column gives the profile name, if one is given in the publications; the third column
gives a representative percentage uncertainty for the upper crust; the fourth contains a
representative uncertainty for the lower crust; the fifth column gives a representative un­
certainty for the low velocity zones (if any exist in the model); and the sixth column gives
the reference for the published models. The uncertainties are in italics if they have been
taken from the original publication and in standard font if no uncertainties were provided,
or the provided uncertainties have not been used.

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### D. Database Uncertainties

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