

Gravity Anomalies, Flexure, and the Long-term Rigidity of the Continental Lithosphere

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The cause and distribution of spatial variations in the mechanical properties of the continental lithosphere are fundamental questions for modern geology. In this study variations in long-term lithospheric rigidity have been investigated. These investigations used profile- and grid-based flexural models of the lithosphere's response to geologically imposed topographic, or buried, loads. These models were constrained by topographic and gravity data allowing recovery of best fitting rigidity values.

In Oman a Cretaceous ophiolite acts as a significant load on the continental crust. Flexural models along profiles orthogonal to the ophiolite strike show that the observed gravity data can be best modelled by an elastic beam with standard thickness (T_e) of 30 km. Along strike there is shown to be significant variation in the foreland shape and the observed gravity signal. This, it is proposed, relates to the complex tectonic processes which occurred as the ophiolite was obducted.

The Himalayan foreland has been the focus of controversy over the recovered long-term rigidity of the continents, with recovered T_e values ranging from 40 to over 90 km. Both profile- and grid-based techniques show that T_e is high (>70 km) in the foreland region. Across the India-Eurasia collisional system as a whole T_e values are variable. Beneath the Tibetan plateau recovered values are generally low (<10 km), while the plateau margins are marked by regions of higher rigidity.

Recovered T_e values across the Arabia-Eurasia collisional system range from over 60 km in the foreland region to close to zero beneath the high Zagros mountains. In the eastern part of the foreland, flexural models match the gravity data; however, they disagree with sediment thickness data for the material infilling the foreland. This discrepancy is interpreted in terms of de-coupling of the flexural lithosphere from the shallower crustal levels, caused by the presence of significant salt deposits in this region.

Application of grid-based techniques to South America, North America and Europe recover a broad range of T_e values from ~ 0 to over 90 km. The low T_e values are explained in active orogenic belts in terms of current processes acting to weaken the lithosphere, and in the continental interiors as the relics of past orogenic events. High T_e values in the continental interiors correlate with ancient cratonic cores which have undergone little deformation since their formation in the Archean.

This study shows that T_e variations have a critical influence on the development of large compressional orogenic belts. In the Himalayan and Andean orogens there is a correlation between the over-thrusting of the orogenic belt and high T_e foreland regions. Where lower T_e regions are seen, less over thrusting is apparent, and in the case of the India-Eurasia collisional system out-flow of lower crustal material may be occurring.

Extended Abstract

The lithosphere forms the outermost layer of the earth, and comprises the crust and non-convecting part of the upper mantle. It is composed of a mosaic of tectonic plates, the relative motion of which is the underlying cause of all large orogenic belts seen both today and in the geological record. Variations in the mechanical properties of the lithosphere are an important factor in tectonic processes, and may also represent a fingerprint of past tectonic events.

The oceanic lithosphere is relatively young ($< \sim 250$ ma), while the continental lithosphere records a tectonic history, in places going back over two billion years. The long and complex history of the continents leads to lateral variation in lithospheric properties, the cause and spatial distribution of which are poorly understood.

Geophysical techniques provide a key tool for the investigation of large-scale variations in lithospheric properties. In the current study the rigidity of the continents is assessed. Rigidity can be assessed at a number of different time scales, depending on the technique applied. Seismic tomographic techniques, for example, assess the lithosphere's response to loading on a scale of seconds to minutes. Analysis of the distribution of earthquakes gives a view of the lithospheric properties on the time scale of the seismic cycle of years to thousands of years. On long (> 1 Ma) time scales rigidity can be assessed by analysis of the lithospheric response to geologically imposed loads, such as orogenic belts or ice caps. In this thesis it is the long term rigidity of the continents that is investigated. Recovery of the long term rigidity allows both an investigation of its impact on current tectonic processes, and the causes of the recovered variations.

The long term rigidity of the lithosphere is typically defined as the thickness (T_e) of a standard elastic plate that bends to the same extent as the observed lithosphere when subjected to an imposed load. To recover T_e variations in the continental lithosphere the isostatic compensation of imposed loads is considered. The extent of this compensation is reflected in the observed gravity anomalies that arise from deflection of the Moho and development of sediment-filled foreland basins. T_e can be recovered from this observed gravity data in two distinct ways: either by consideration of models of flexural surfaces due to known loads, such as surface topography, or by investigation of the spectral wavelengths across which the transition from Airy compensation to rigid support of loads occurs. In this study model-based techniques are preferred as they allow explicit consideration of the imposed loads.

The model-based techniques applied consider topographic and gravity data either along profiles or across grids. In the case of profile-based studies a range of flexural models was constructed, each with a different input T_e , using an assumed geological load (either topographic or buried). The model, and hence T_e , which produced the calculated gravity anomaly most closely matching the observed anomaly was then recovered. To recover the T_e from grids of data only the topographic load was considered. The flexural surfaces resulting from this load were modelled using a program that considered a spatially variable T_e . From these surfaces a flexurally derived gravity anomaly was calculated. By iteratively varying the T_e structure the pattern of variation that produced a calculated gravity anomaly most closely matching the observed anomaly was found. These techniques were tested using synthetic topography and gravity data derived from a known input T_e . Both profile- and grid-based techniques were shown to recover well the input T_e structure.

Compressional orogenic settings are the focus of this study, as they are typically associated with over-thrusting of material and the generation of significant surface topography. The over-thrust material acts as a significant, well defined load on the

continental lithosphere, allowing for the construction of well constrained flexural models. T_e variations are also recovered away from active orogenic regions as the topography, if of sufficiently long wavelength, must be isostatically compensated, and hence be associated with gravity anomalies originating from the Moho. In these regions, however, the distribution of the driving load is less clear.

The first region investigated was Oman. Here there is known to be a sliver of Cretaceous oceanic lithosphere, or ophiolite, thrust onto the continental margin. This ophiolite acts as a significant buried load, and hence, a distinct flexural signal would be expected. The results of flexural modelling on a series of profiles along strike of the orogen confirm a clear flexural signal in some parts of the foreland, which can be best matched with a T_e of ~ 30 km. In other parts of the foreland region, however, no clear flexural signal can be seen, despite the presence of very similar ophiolitic material. This variation is interpreted in terms of the tectonic processes which occurred as the ophiolite was obducted.

The next investigation focused on the India-Eurasia collisional system, from the Indian peninsula to north of the Tibetan plateau. This region was investigated because of the wide range of T_e values (from 40 km to over 90 km) recovered by previous authors in the Himalayan foreland. This variation leads to fundamentally different views of the mechanical behaviour of the lithosphere. The results of profile-based techniques confirm that T_e in the central foreland region to the south of the Himalaya is high (~ 70 km). The grid-based modelling technique was applied across the entire region. This technique similarly confirms the high T_e values in the foreland, which implies that a significant proportion of the lithospheric mantle must, in this regions, be rigid. The Tibetan plateau forms a low rigidity ($0 < T_e < 20$ km) region which is bounded by a series of more rigid blocks, such as the Himalayan foreland. Where the plateau is not bounded by rigid lithosphere there is a significant change in tectonic style, showing that T_e has a direct impact on how the orogen develops.

The Arabia-Eurasia collisional system presents a similar tectonic setting to the India-Eurasia collisional system, with continent-continent collision occurring today. The Zagros mountains represent the topographic expression of this orogenic system, while the Persian Gulf, in part, marks the foreland developed due to the orogenic loading. The T_e structure was investigated based on profile- and grid-based modelling techniques. These show that the best fitting foreland T_e values are high (>60 km), while recovered T_e values are much lower beneath the mountains. In the central and western parts of the foreland, the best-fitting flexural model, derived from comparisons of calculated and observed gravity data, matches the width and depth of the foreland as seen in isopach and stratigraphic data. Towards the east of the orogenic belt there is a significant discrepancy between the best fitting flexural model and the isopach data. It is proposed that this relates to the de-coupling of the deeper, flexing lithosphere from the shallower sedimentary sequence seen in the isopach data. This de-coupling is related to the presence of significant salt deposits in this region.

To further investigate the T_e variation across the continents three additional regional studies were conducted in South America, North America and Europe. These used extensive regional grids of topography and gravity data and the grid-based modelling technique. A clear distinction can be made between the low T_e values recovered in currently active orogenic belts and the more varied T_e values found in other continental regions. There is not likely to be single cause for the low values recovered in orogenic belts; however, the high heat flow and displacement of rigid mantle by less rigid continental crust are likely to be the most significant factors. Away from active orogenic belts the pattern of recovered T_e appears to correlate with older tectonic provinces. High T_e values are generally recovered in the ancient cratonic cores, while lower T_e values are typically associated with suture zones or traces of mantle plumes.

The T_e variations recovered by this study show that continental rigidity is highly

variable. For younger tectono-thermal lithospheric ages there appears to be a correlation between T_e and age, however, for older regions this correlation breaks down. It is therefore concluded that continental T_e is a balance between cooling with time, which acts to increase the rigidity, and tectonic processes that act to reduce rigidity, but may not re-set the tectono-thermal age. The highest T_e values recovered by this study of >70 km significantly exceed the local crustal thickness. This supports the view that in places the lithospheric mantle beneath the continents is, despite its aseismic nature, rigid over long time scales.

The tectonic impact of T_e variations is clearest in the largest orogenic belts. In the Andes there is a correlation between the orogen width and foreland T_e . This may relate to the shallower dip of the rigid foreland favouring overthrusting of the continental interior by the orogen. In the India-Eurasia collisional system margins of the Tibetan plateau bordered by high T_e regions show steep topographic fronts and overthrusting of the orogen, while the other regions show lower topographic gradients and may be the sites of lower crustal flow, away from the central orogenic belt.

As T_e variation appears to correlate with large-scale variations in tectonic processes it is proposed that the characteristics of continental orogenic belts are fundamentally controlled by the rigidity of the colliding plates. Three end-member situations are envisaged. Firstly where rigid and non-rigid plates collide. In this case the less rigid plate will preferentially deform and eventually be overthrust onto the rigid plate. The more rigid the overridden plate, the further the propagation of the orogen and the broader the region of deformation that will develop. Secondly, where two low rigidity plates collide over-thrusting of the orogenic belt onto the surrounding continental region will be retarded by the steep foreland and basement involved thrusting. Lastly where two rigid plates collide deformation would be limited at the suture, and stresses may be transferred to other less rigid regions.

Declaration

I hereby declare that the dissertation, submitted in partial fulfilment of the requirements for the degree of Doctorate of Philosophy and entitled Gravity Anomalies, Flexure, and the Long-term Rigidity of the Continental Lithosphere represents my own work and has not been previously submitted to this or any other institution for any degree, diploma or other qualification.

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Chapter 1

Introduction

The mountain ranges seen today are the most recent topographic expression of the slow but immensely powerful tectonic forces which have been acting to shape this planet over the last four billion years. How the crust and uppermost mantle, together known as the lithosphere, respond to these tectonic forces is a fundamental question which underpins modern geology. In this study the lithospheric response to long term geologically imposed loads is investigated. These investigations are based on topographic and gravity data sets which are used to construct and constrain models of the lithospheres response to loading. The results show that there is wide variation in the response of the lithosphere to loading, which in turn impacts on how tectonic processes develop. The wide variation in the response of the continental lithosphere has no single explanation, however, there does appear to be a correlation with geological structures suggesting that tectonic processes can permanently modify the lithospheric structure as they occur.

This chapter gives an overview of the lithosphere and some of the geophysical techniques used to investigate its structure. A brief history of the investigation of support of geologically imposed loads is given, from the earliest geodetic surveys in the eighteenth century to the modern tectonic era. The results of recent studies of continental rigidity, a measure of how loads are supported, are then discussed. Finally the structure of the remainder of this thesis is laid out.

1.1 Investigating the lithosphere

The lithosphere is the upper layer of the solid earth comprising the crust and the upper, non-convecting, part of the mantle. It is divided into a series of tectonic plates whose relative motion generates the orogenic belts seen today, and in the geological record.

Geophysical techniques provide a key tool for investigating the structure of the lithosphere due to its great thickness. These techniques include, for example, studies of long term ($> 10^6$ years) lithospheric rigidity based upon topographic and gravity data, studies of heat flow variation, earth-quake location studies, and seismic tomography. Each of these techniques investigates the character of the lithosphere in a different way and on a different time scale.

1.1.1 The elastic lithosphere

The elastic lithosphere is considered to be the part of the lithosphere which can rigidly support geologically imposed loads over time scales of $> 10^6$ years. Lateral variation in the elastic lithosphere can be discussed in terms of variations in recovered flexural rigidity. This describes how a theoretical elastic plate would respond to the same imposed forces. The forces considered in this study relate to vertical loads generated by surface topography or buried high density material.

Early studies of lithospheric rigidity such as Watts & Cochran [1974] returned values of flexural rigidity (D) in terms of N m. More recent studies of lithospheric rigidity, however, typically return their results as values of equivalent elastic thickness or T_e (e.g. Watts [1992]). T_e is the thickness of an elastic plate, which is related to D by Young's modulus and Poisson's ratio, and which flexes to the same extent as the observed lithosphere. To ensure consistency between studies, standard values of Young's modulus and Poisson's ratio must be assumed. T_e was thought to be more useful than D because it can be directly equated to lithospheric thickness. Returning rigidity values in terms of T_e worked well for the oceanic plates, where

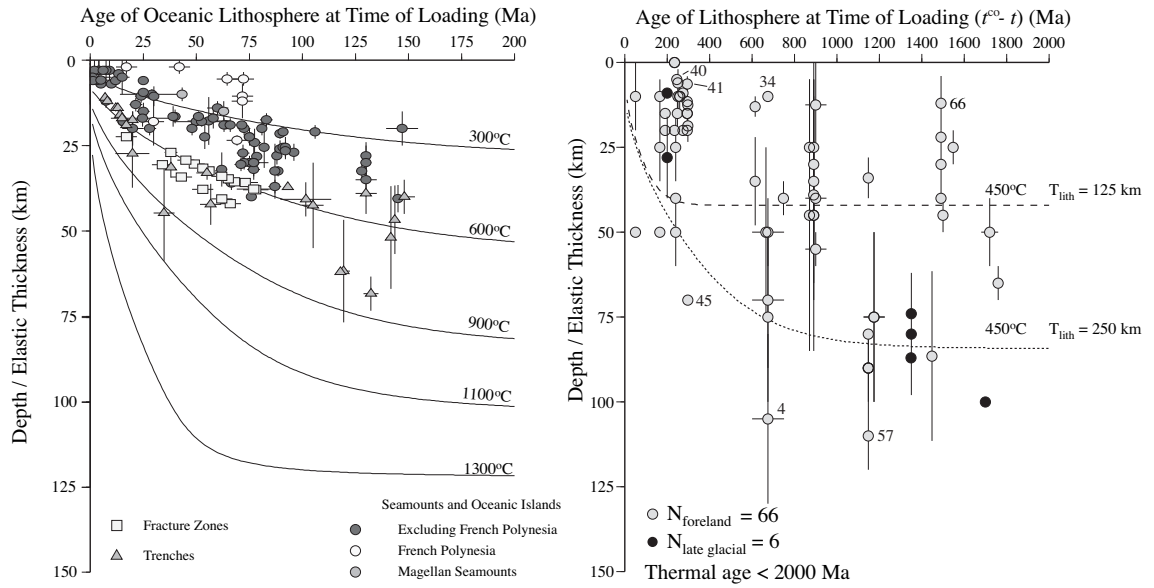


Figure 1.1: Compilation of oceanic and continental T_e values plotted against age at time of loading from Watts [2001]. a) Oceanic T_e values. b) Continental T_e values.

variations in T_e could be predicted based on the age of the sea floor and simple cooling plate models, as shown in Figure 1.1a from Watts [2001]. In continental regions it was also believed that T_e was a function of lithospheric age, at the time of loading, for example Karner & Watts [1983]. However, the correlation between T_e and age breaks down when significantly more estimates of continental rigidity are considered, for example Figure 1.1b from Watts [2001]. This figure shows that for young continental lithosphere T_e appears consistently weak, however, old lithosphere gives both high and low values with no clear pattern being discernible.

Yield-strength envelopes

The rigidity of the lithosphere can be described in relation to its chemical make up and the physical forces imposed upon it by consideration of a theoretical yield strength envelope, such as that shown in Figure 1.2. This provides a graphical representation of the amount of stress the lithosphere can support without yielding (Watts & Burov [2003]). Figure 1.2 shows the simple case of the oceanic crust. At shallow depths the amount of stress the lithosphere can support before yielding

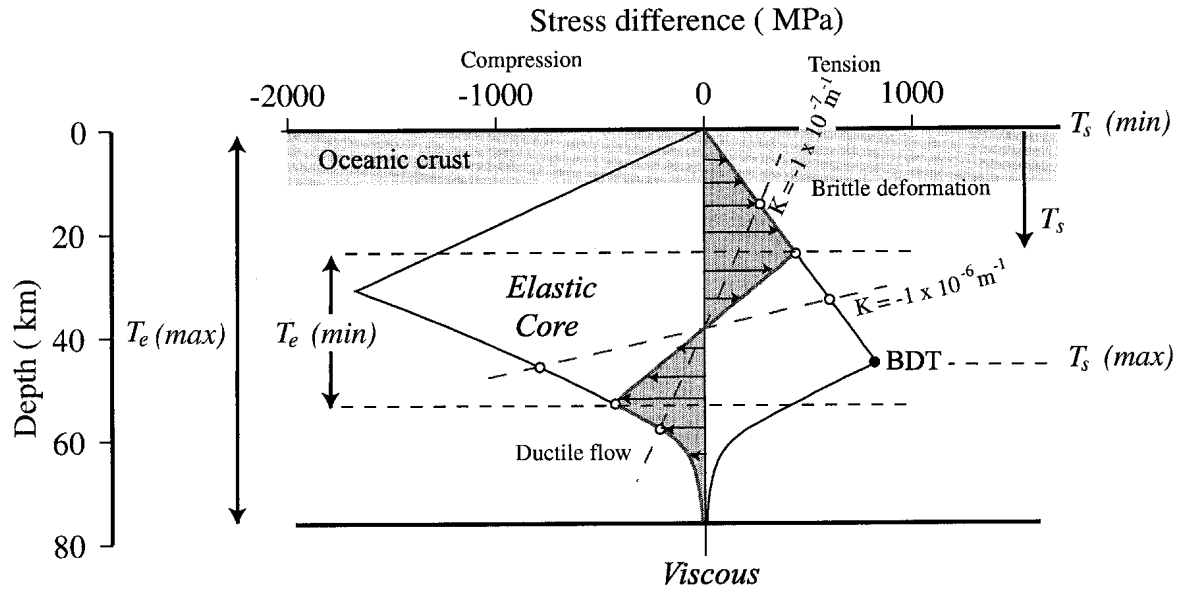


Figure 1.2: Yield strength envelope for 80 Ma oceanic lithosphere from Watts & Burov [2003]. BDT is the Brittle Ductile Transition. Thin solid line shows yield strength envelope. Thick diagonal line shows the stress difference generated for a curvature, K , of $5 \times 10^{-6} \text{ m}^{-1}$. For this curvature the load is supported partly by the unyielded elastic core and partly by the brittle and plastic strength of the lithosphere outside this core. Diagonal dashed lines are for K values of $1 \times 10^{-7} \text{ m}^{-1}$ and $1 \times 10^{-6} \text{ m}^{-1}$ which is suggested to be the approximate range of K for oceanic-trench flexural systems. T_s is the depth to which brittle failure, and hence seismicity, will occur, given the stresses generated by bending the beam. $T_s(\text{max})$ is the maximum depth to which seismogenic brittle failure can occur given additional imposed stresses.

increases linearly with depth according to Byerlee's law of brittle failure. As temperature increases with depth dislocation creep and other forms of plastic failure become more important and eventually supersede brittle failure as the dominant cause of yielding.

The rigidity of the lithosphere can be determined based upon yield strength envelopes as the integrated amount of stress that can be supported without yielding. Watts & Burov [2003] emphasised the point that this elastic thickness includes

the material which has not yielded, but would eventually yield by plastic failure. When curvature of the beam is small then the entire thickness of the beam will act elastically in response to an imposed load. However, as curvature of the beam increases the upper and lower surfaces are stressed beyond the limits of the yield strength envelope and will yield. This situation is shown in Figure 1.2 by the diagonal lines, these represent the stresses associated with beam curvature that develops due to flexure. In a high curvature state the integrated strength of the lithosphere includes both the unyielded elastic core, and the residual brittle and ductile strength of the lithosphere (Watts & Burov [2003]).

In the oceans the yield strength envelope has a relatively simple shape because the field of brittle failure rapidly exceeds the crustal thickness and the shape of the yield strength envelope is dominated by the behaviour of olivine with depth. The oceanic lithosphere therefore acts as a simple, single layered, beam where the thickness is controlled by how much the lithosphere has cooled, and hence its age at the time of loading. In continental regions the crust is much thicker and made of more silica rich material, this can yield due to plastic deformation at lower temperatures than mantle rocks at an equivalent depth. This leads to the more complex multi-layer yield strength envelope seen in Figure ???. Different crustal similar thicknesses, chemistry and thermal settings, stemming from the long and complex tectonic evolution that many continental regions have undergone cause significant differences in the shape of the continental yield strength envelope for regions of a similar age. As a result the recovered T_e in continental settings may not correlate directly with age.

1.1.2 The thermal lithosphere

The thermal lithosphere is defined as the thickness of crust and non-convecting mantle material overlying the convecting asthenospheric mantle. This thickness is derived from thermal models constrained by heat flow measurements. The re-

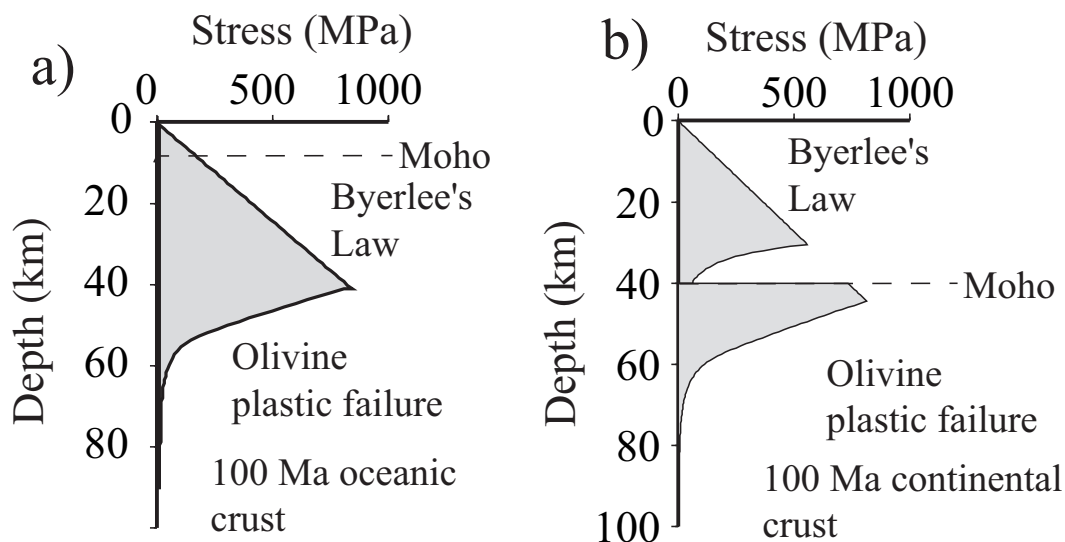


Figure 1.3: Yield strength envelopes for oceanic and continental regions after Watts [2001]. a) Oceanic yield strength envelope, simply controlled by the brittle - ductile transition point of Olivine. b) Continental yield strength envelope. Shallow part controlled by brittle ductile transition of silica rich crustal rocks, below the Moho controlled by the response of Olivine.

covered thickness represents a view of the lithospheric structure on time scales approximately equivalent to the age of the last tectonic event to occur in a region.

It has been shown in the oceans that heat flow and ocean depth are a function of age, due to cooling of the lithosphere as it moves away from the ridge where it was created. The variations in heat flow and ocean depth related to cooling allowed Parsons & Sclater [1977] to define a thermal lithospheric thickness based on a simple cooling plate model. In the continents the relationship between heat flow and age is not as distinct due to geothermal heat production within the crust locally modifying heat flow. Calculation of a thermal lithospheric thickness in the continents is more difficult than in the oceans for the same reason. Artemieva & Mooney [2001] solved the problem of recovering a thermally derived thickness for stable continental lithosphere across broad regions by considering the presence, or absence, of heat producing elements in the crust within average regional blocks. Having removed the effect of crustal heat production the thermal thickness of

the lithosphere was calculated from heat flow data assuming a conductive layer overlying the mantle with an adiabat of 1300°C. This led to the recovery of thermal lithospheric thicknesses, that is non-convecting thickness, of 200 to 350 km for Archean and early Proterozoic cratons. This thickness is much greater than the crustal thickness derived from seismic studies which is at most ~ 80 km (Nelson *et al.* [1996]). This implies that, as in the oceans, a significant proportion of the thermally defined lithosphere is made up of mantle material in addition to crustal material.

1.1.3 The seismogenic lithosphere

The seismogenic thickness (T_s), is the part of the lithosphere where earthquakes commonly occur and defines the depth to which brittle failure is occurring in a seismogenic fashion. Investigations of T_s assess the properties of the lithosphere on the time scale of the seismic cycle, which can be on the order of tens to thousands of years in length.

In the oceans values for T_s are approximately equivalent to T_e and increases with age in a similar manner from close to zero at the ridges to ~ 50 km in the oldest parts of the oceans, for example Watts & Burov [2003]. Within the continents the seismogenic thickness appears to be confined to the crust, Maggi *et al.* [2000a]. This observation led Jackson [2002b] and Jackson [2002a] to propose that continental T_e should always be less than the crustal thickness as the brittle region is the only part of the crust with any long term strength. Watts & Burov [2003] however argue that the seismogenic layer represents the region where the lithosphere is failing. In many regions they propose the upper mantle is too strong to fail, and is therefore not seismogenic because it is strong, rather than because it is weak. It is also argued by Watts & Burov [2003] that, as the T_e represents the integrated strength of the lithosphere, all the area within the yield strength envelope which has not yielded contributes to the observed rigidity, whether the lithosphere will eventually

fail brittly or plastically.

1.1.4 The seismic lithosphere

An additional view of the lithosphere comes from seismic tomographic studies. This type of study typically uses shock-waves from earthquakes on a time scale of seconds to minutes to build up a picture of variations in lithospheric structure.

These studies show that the mantle immediately below the crust has a significantly higher velocity than the mantle material at deeper levels. This mantle material appears to be differentiated from the lower velocity layer below, suggesting it forms part of the rigid lithosphere, rather than being part of the more mobile asthenosphere below. Beneath the oceans the seismic thickness of the lithosphere increases with age in a manner predictable by the cooling plate model (Nishimura & Forsyth [1989]). The cratonic cores of continents typically have a high seismic wave velocity to a greater depth than seen in the oceans. This suggests that the mantle beneath parts of the continents is cooler than mantle at an equivalent depth beneath the oceans. These high velocity lithospheric roots have been shown to extend several hundred km beneath many cratonic regions, for example Jordan [1978], Lerner-Lam & Jordan [1987], Fischer & van der Hilst [1999], James *et al.* [2001].

1.1.5 Focus on rigidity

The structure of the elastic lithosphere, as shown by variations in recovered T_e , is the focus of this study. It provides a view of the lithospheric structure on time scales similar to many geological processes. As T_e reflects the integrated view of the thermal and chemical structure of the lithosphere, variations in recovered T_e may result from variations in the processes that have formed the lithosphere, as well as tectonic processes that have since altered it. In addition T_e impacts directly on current tectonic processes, for example controlling the shape of foreland

basins (Watts [1992]). T_e variations therefore may be a powerful tool for explaining variations in contemporary tectonic processes.

1.2 Isostasy

Geologically, isostasy is the support of surface topography by buoyant forces from below. The theory of isostasy is important to any study of continental rigidity as it describes how the lithosphere responds to long term loads. An in-depth introduction to the historical background of the subject of flexural isostasy is given in Chapters 1 and 2 of Watts [2001]; a brief summary of the key events is given below.

1.2.1 Simple isostatic models

The theory of isostatic balance was developed in conjunction with the first regional geodetic surveys. One such survey was carried out by the French in an attempt to determine the meridian arc and hence the flattening of the earth. On this expedition Bouguer [1749] noted that the error due to the attraction of the mass of the Andes was much less than expected, given their size. Boscovich [1755] explained the smaller than expected error as a result of expansion, and hence the lower density, of material beneath the mountains. This expansion explained both the formation of the mountains and their relative lack of gravitational attraction.

A fuller consideration of the nature of the support of mountains was not made until the geodetic survey of India from 1840 to 1859 by the surveyor general of India, G. Everest. Everest used both geodetic and astronomical methods to measure positions of stations across India. In the Gangetic plain he observed a difference of $5''.236$ between the two methods. Pratt [1855] tried to explain this discrepancy as being due to the effect of the topography of the Himalaya and Tibet to the north, the gravitational attraction of which was deflecting the plumb-line used to

level the geodetic instruments. Pratt calculated that the discrepancy between the geodetic and astronomical position should have been $15''.885$. Everest's observed discrepancy, which was much smaller than that calculated by Pratt, was in line with the earlier observations of smaller than expected errors in the Andes by Bouguer [1749].

Airy [1855] showed that if the surface topography was isostatically balanced by low density material at depth, the conflict between Pratt's calculated error and Everest's observed error could be resolved. The low density material at depth, in place of higher density material, reduces the overall attraction of the mountain belt. Figure 1.4a illustrates the essential concept of Airy isostasy. Airy considered a crust 10 miles thick floating on a high density fluid 'lava' and that the depth of projection downward of the crust was proportional to the height of the surface topography, in the same manner as a raft of floating logs. Airy [1855] suggested that broad regions of elevated topography were bounded by 'vertical fissures' allowing them to maintain buoyant equilibrium. For local regions of elevated topography, such as individual mountains in Scotland, it was suggested that a compensating root was not needed.

Pratt [1858] agreed that low density material at depth could explain the discrepancy in the deflection of the pendulum, however, he disagreed with the method of compensation proposed by Airy [1855] for three reasons. Firstly the proposed crustal thickness of Airy [1855] was thinner than generally accepted. Secondly solidification from a fluid typically involves a density increase, and hence crust would not be expected to float. Lastly Pratt [1858] believed topography was created by thermal contraction and could not see how this process could create the different crustal thicknesses needed for Airy compensation to work. In place of the Airy model of support for topography Pratt [1858] suggested that lateral variations in density of the crust could account for the differential elevations of the mountains and the apparent mass deficiency at depth. This model is shown in Figure 1.4b.

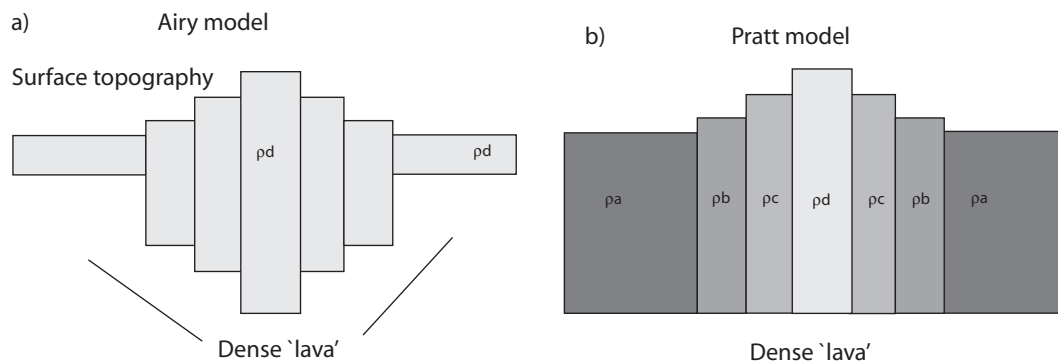


Figure 1.4: Models for isostatic compensation of topography. a) Airy model, where crustal density is constant, but depth of root varies. b) Pratt model where depth to compensating surface is constant while lateral density variations generate topography, i.e. $\rho_a > \rho_b > \rho_c > \rho_d$.

Today the Airy model of isostatic balance is widely accepted as an explanation for the compensation of large scale continental topographic features such as mountain ranges. The ultimate success of the Airy model comes from seismic studies, such as that of Nelson *et al.* [1996], which show that major mountain ranges are underlain by a significant crustal roots, displacing the higher density mantle as the Airy model suggested. The Pratt model, although not generally applicable to features on the continent, does explain the broad scale variations in elevation of the oceanic crust. Hot, lower density, mantle material is known to underlie the ridges, which are relatively shallow, while denser, cooler, mantle underlies the broad, deeper, abyssal plains (Parsons & Sclater [1977]).

1.2.2 Isostasy and rigidity

The early models for isostatic balance of topography generally worked well for correcting geodetic surveys; however, geological situations proved more complex. An early example of this complexity was demonstrated by Gilbert [1890]. Gilbert measured the elevations of paleo shore-lines around Lake Bonneville, a Pleistocene lake in the U.S. which is today almost completely drained. Gilbert [1890] showed

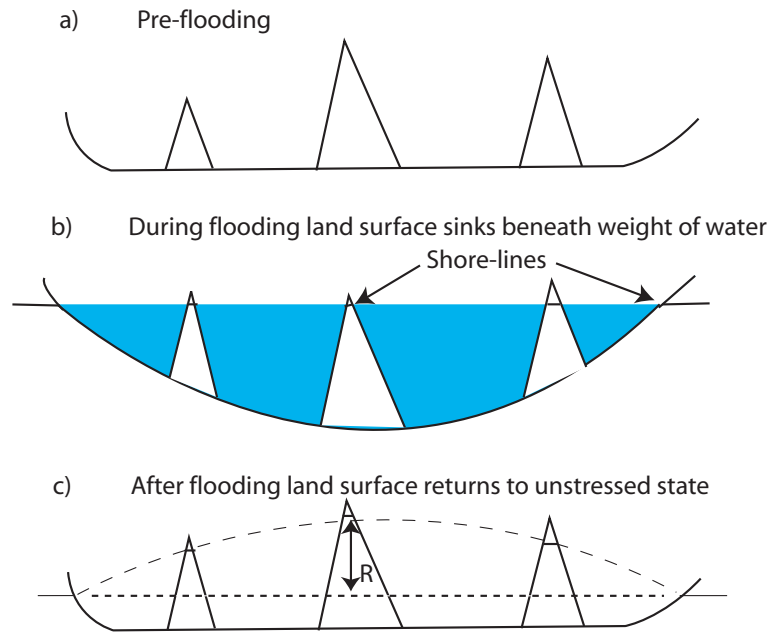


Figure 1.5: Model of response of surface to the loading and unloading of Lake Bonneville. a) State prior to loading. b) Loading by water mass and development of shorelines. c) Response to unloading of water mass. Islands return to their former elevations leading to doming of the paleo-shorelines. R = maximum rebound.

that the paleo shore-lines around islands in the centre of the lake were now higher than those of the same age on the flanks. This, it was suggested, related to the isostatic compensation of the weight of the water which had pushed the centre of the lake down more than the margins. When the water drained out the crust rebounded, raising the shorelines in the centre of the lake more than those at the margins, as shown in Figure 1.5.

An initial model based on an Airy response to the water loading showed that the expected maximum rebound (R) was around 90 m. The observed rebound was only on the order of 39 m. This finding suggested that the crust was not responding to loading of the water by buoyant compensation, or by total rigid support, the two options suggested by Airy [1855]. Instead Gilbert [1890] showed that the crust could be responding as an elastic beam where the load is, in-part, supported by the rigidity of the beam.

Barrell [1914] made the first extensive study of continental rigidity based on gravity anomalies, the results of which were published in a series of papers in the *Journal of Geology* volume 22. Barrell noted the existence of systematic residual anomalies after correcting for local compensation of topography using the Pratt model. These indicated that the crust was not in a state of local isostatic balance. In place of local isostatic balance Barrell suggested that the crust could support loads regionally and that the residual anomalies represented stress differences supported by the rigidity of the crust. This is in line with the earlier findings of Putnam [1895]. Putnam highlighted the fact that the residual anomalies calculated using Pratt isostasy assumed local compensation of topography down to an implausibly small scale of 2 m. When the average topography over a significantly larger region was considered, the isostatic discrepancy was much reduced.

The theory of regional isostatic support of topography was progressed by authors such as Vening-Meinesz [1948] and Gunn [1949]. Instead of looking at how large an area needed to be considered for ‘regional’ Pratt isostasy to work, they considered the Airy model. In this case they viewed the surface topography as a load which had to be compensated at depth by a low density root. As the surface topography was seen as a load Vening-Meinesz [1948] considered the possibility that the crust could support long term loads, in part, by its own internal rigidity as suggested previously by Barrell. Vening-Meinesz [1948] modelled the crust as an elastic plate overlying an inviscid substratum (fluid) and used this approximation to significantly reduce the residual anomaly after isostatic correction of gravity anomalies across the Hawaiian ridge. Gunn [1949] independently developed elastic-isostatic models which allowed the relative rigidity of different regions to be assessed based upon the width of residual anomalies remaining after local isostatic correction.

Early workers, such as those discussed above, were working before the development of plate tectonic theory and the idea that the earth surface was made up of a

series of rigid plates in relative motion had not been conceived. However, since the development of the theory of plate tectonics in the 1960s the idea that the earth's surface is made up of a series of rigid lithospheric plates moving over a relatively weak substratum has become widely accepted. The structure of these lithospheric plates has since been studied by many authors, using a variety of techniques discussed above. In the current study it is the variation in long term plate rigidity which is assessed. However, the recovered rigidity variations are often compared with the results of other techniques to build up a more complete picture of the lithospheric structure.

1.3 Current views of T_e variation

1.3.1 The range of recovered T_e values

Recent regional studies of T_e based upon loading models include, for example, those of Stewart & Watts [1997], Braitenberg *et al.* [2002], Wyer [2003] and Wyer & Watts [2006]. Stewart & Watts [1997] recovered T_e variations along the foreland regions associated with the Andean, Alpine, Carpathian and Apennine mountain belts. A wide range of T_e values from 5 to over 80 km were recovered. Braitenberg *et al.* [2002] recovered T_e values between ~ 2 and ~ 20 km based upon both topographic loading and the loading caused by sedimentary basins associated with the Alpine orogenic belt. Wyer [2003] and Wyer & Watts [2006] investigated the T_e variation along the passive margin on the east coast of the U.S. based on sedimentary loading. Here variations in recovered T_e ranged from 5 to 30 km along strike of the margin.

Recent regional studies based upon transfer function (spectral) techniques include those of Pérez-Gussinyé *et al.* [2004] and Pérez-Gussinyé & Watts [2005] across Europe and the study of Flück *et al.* [2003] in Canada. These authors recovered significant T_e variation from less than 5 km to over 60 km. McKenzie & Fairhead [1997] and McKenzie [2003] had previously argued that high T_e val-

ues, over ~ 40 km, recovered by both modelling and transfer function techniques were incompatible with physical models of the lithosphere. This was based on the assumption of, for example, Maggi *et al.* [2000a], discussed previously, that only the brittle, and therefore seismogenic part of the lithosphere could support long term loads. McKenzie [2003] applied transfer function techniques and recovered T_e values consistently below ~ 40 km for a range of continental regions. However, Pérez-Gussinyé *et al.* [2004] showed that by reformulating the technique of McKenzie [2003] high values of T_e could be recovered, in accord with the results of previous transfer function and modelling studies.

1.3.2 Causes of T_e variation

The maximum values recovered by Wyer [2003] and Wyer & Watts [2006] are significantly lower than the maximum values recovered by Stewart & Watts [1997]. It has been proposed that the lower values found at rifted margins such as the eastern U.S. and offshore West Africa (Wyer [2003] and Watts & Stewart [1998]) are related to the rifting processes which have occurred in these regions. Watts & Stewart [1998] suggest that after rifting the crust and mantle across the continental margin may remain decoupled. This would mean that even if the mantle regained significant strength it would not be reflected in the recovered T_e . In addition the removal of Archean lithospheric mantle and its replacement with younger, potentially more hydrous mantle material may leave rifted margins permanently weakened (Pérez-Gussinyé & Watts [2005]). Lin & Watts [2002] demonstrated that where rifted margins were later incorporated into compressional orogenic systems, such as Taiwan, the low rigidity of the margin was inherited and reflected in the shape of the developing foreland basin. This inheritance of low rigidity due to previous continental rifting may be a contributing factor, in addition to the more complex continental yield strength envelope, to the broad spread of recovered continental T_e values.

Compressive orogenic belts are often typified by low T_e values in their cores. This was demonstrated for example by Karner & Watts [1983] and Stewart & Watts [1997] who showed that the best flexural model in the orogenic regions considered was one where the elastic plate was completely broken beneath the core of the orogenic belt. Braitenberg *et al.* [2002] also recovered low T_e values within the Alpine orogenic belt, with no prior assumption about the presence or absence of a broken plate. There are likely to be multiple causes for the weakness of the lithosphere seen within orogenic belts. In compressional settings the two sides of the orogen were never originally connected so it is unlikely they will act as a single rigid elastic plate. Additionally crustal thickening in orogenic belts displaces mantle material from around 40 km, where it may provide a significant contribution to the lithospheric rigidity, to up to 80 km. The material displacing the mantle is much more silicic crustal material which, it is known from yield strength envelope studies, yields more easily at these depths and associated temperatures. The Wilson cycle of continental rifting and suturing over time also tends to lead to the development of orogenic belts in regions such as continental margins where T_e is typically low (Watts & Burov [2003]).

High T_e values in the range of 60 to 80 km have been recovered by previous authors such as Stewart & Watts [1997] and Pérez-Gussinyé & Watts [2005]. These high values are typically seen in the cores of old cratonic regions such as the Central Brazilian shield in South America and the Baltic shield in Europe as recovered by Stewart & Watts [1997] and Pérez-Gussinyé & Watts [2005] respectively. These high values of T_e significantly exceed the crustal thickness in these regions, suggesting that a large proportion of the continental rigidity in these areas is derived from high rigidity lithospheric mantle. The very rigid continental blocks seen in Europe generally exceed 1.6 Ga in age. The high values of T_e recovered in these old cratonic regions and in other cratonic regions around the globe is proposed by Pérez-Gussinyé & Watts [2005] to relate in-part to the long time they have had to

cool. In addition it is suggested by Pérez-Gussinyé & Watts [2005] that a transition from the formation of rigid and stable lithospheric mantle, to the formation of less stable lithospheric mantle occurred, associated with cooling and de-watering of the athenosphere over time. During the Archean higher mantle temperatures and fluid content would have led to greater proportion of partial melting within the upper mantle. If this melt was extracted the residue would be more depleted than in cooler, more modern settings. This highly depleted residue it is proposed forms the thick stable roots beneath many of the cratonic cores where high T_e values are recovered today (Doin *et al.* [1997] and Lenardic *et al.* [2003]).

1.4 This study

The aim of this study is to assess variations in the rigidity of the continental lithosphere, expressed in terms of T_e . A broad range of continental regions are considered based on the T_e which gives the best fit between calculated and observed gravity anomalies. Calculated anomalies are derived from flexural models of isostatic compensation of geological loads. The loads considered are either surface loads as derived from observed topography, or buried loads as inferred from observed gravity anomalies. Recovered T_e variation is discussed in terms of both the past tectonic history of a region, and its impact on present day tectonic processes.

1.4.1 Chapter 2, Methods

Chapter 2 presents a discussion of the model-based methods for recovering T_e and T_e variation as applied by previous authors. The specific model-based methods applied in this study are outlined and the results of synthetic checks of these techniques presented. Finally, transfer function (spectral) methods are introduced. These have been widely used by previous authors for recovery of variable T_e structures.

1.4.2 Chapter 3, Oman: flexure and buried loading

Chapter 3 introduces the geology of the Oman region where there is known to be a Cretaceous ophiolite loading the continental lithosphere. New gravity data collected as part of this study is presented in addition to existing data derived from published sources. A flexural model across the foreland where the new data was collected is first considered. This model represents the deflection generated by the ophiolite. Associated calculated gravity anomalies are presented. The concept of buried loading is introduced, as flexure driven solely by loading of the surface topography is shown to be insufficient in explaining the observed Bouguer anomaly. The flexural rigidity which gives the best fit between the observed and calculated gravity anomaly is then recovered. A series of modelled profiles is presented along strike of the foreland basin and variations in recovered T_e are discussed.

1.4.3 Chapter 4, T_e and the India-Eurasia collisional system

Chapter 4 investigates the flexural rigidity across the India-Eurasia collisional system. Initially a series of profiles are considered along the length of the Gangetic foreland to the south of the Himalayan mountains. The results from a 2D model-based technique for inverting T_e from gravity and topography data across a broad region are then presented. These results are discussed in terms of the tectonic history of the region and the impact of T_e variations on the modern tectonic regime.

1.4.4 Chapter 5, The Arabia-Eurasia collision: T_e in the Persian Gulf

Chapter 5 studies T_e variation along the length of the Persian Gulf, which in part comprises the foreland basin to the south of the Zagros mountains. A simple profile method is applied to recover the foreland T_e . T_e variation across strike of the mountain belt and foreland basin system is also investigated based upon profiles of

data. A series of these profiles is considered along strike of the foreland to provide an assessment of T_e variation. A full consideration of the spatial variation of T_e in the region is then presented based upon a 2D modelling inversion technique. The results of the modelling studies are assessed in relation to the current tectonic and geological setting of the region. Discrepancies between the flexural model, which provided the best fitting gravity anomaly, and the observed foreland basin shape are noted towards the eastern end of the foreland. These are explained by the presence of significant salt deposits in the region.

1.4.5 Chapter 6, T_e variations across South America, North America and Europe

Chapter 6 presents the results of the application of the 2D modelling inversion technique to three different continental areas: South America, North America, and Europe. This chapter presents maps of T_e variation which confirm that there is significant lateral variation in T_e within the continental lithosphere. In South America there is seen to be a correlation between T_e and the physiographic properties of the Andean orogen such as width and height. In Europe and North America agreement is shown between results of transfer function (spectral) techniques applied by previous authors, and the results of the 2D technique applied in this study. The correlation between T_e variation and seismicity, geology and geophysical properties of the lithosphere, such as seismic velocity and heat flow, is also discussed.

1.4.6 Chapter 7, Discussion

Chapter 7 presents a discussion of the implications of the recovered T_e variations. The variation of T_e with age is considered, as are the implications of the broad range of recovered T_e values on continental rheology. Finally the impact of T_e variations on continental tectonics is discussed.

1.4.7 Chapter 8, Conclusion and further work

Chapter 8 presents the main conclusions of the thesis and suggestions for future work.

Chapter 2

Methods

2.1 Introduction

Methods for recovering the rigidity of the lithosphere (T_e) fall into two broad groups; those based on comparison of observed data with models of flexural surfaces and associated gravity anomalies; and spectral techniques based on analysing the relationship between topography and gravity data as a function of wavelength.

Modelling techniques calculate flexural surfaces using mathematical solutions for the bending of a plate or beam. Simple elastic, or visco-elastic models can be used for calculating these surfaces. Watts & Zhong [2000] showed that in the case of the oceanic lithosphere, on time scales greater than 1-2 Ma, there is no significant difference between reasonable multi-layer visco-elastic and elastic models. Simple elastic models are used in this study of the continental lithosphere as the loads considered are generally >1 Ma. If the continental lithosphere deformed viscously over long ($>\sim 100$ Ma) time scales, old loads would always be associated with low recovered rigidities. Studies such as Karner & Watts [1983] showed that continental rigidity in regions such as the Appalachians appears to have remained high for ~ 300 Ma, supporting the use of an elastic model.

Model flexural surfaces can either be compared directly to geological surfaces (e.g. Jordan [1981]), or used to calculate a theoretical gravity anomaly which can be compared with observed data (e.g. Karner & Watts [1983] and Lyon-Caen &

Molnar [1983]). The flexural rigidity, or equivalent elastic thickness (T_e), that gives the best fit between calculated and observed data is assumed to represent the integrated long-term strength of the lithosphere (Watts & Burov [2003]). A key region where this type of analysis is carried out is in foreland basins associated with over-riding orogenic belts. In these regions distinct orogenic loads can be defined and the foreland represents the flexural depression created by regional compensation of these loads.

In spectral techniques the relationship between gravity and topography as a function of wavelength is computed and compared to transfer functions derived from isostatic models (e.g. Dorman & Lewis [1970]; Forsyth [1985]). These transfer functions show at which wavelengths the transition from full rigid support of loads to full Airy compensation occurs, and are a function of T_e . These techniques are based on topographic loads, buried loads or some combination of these. These techniques are often applied with no prior assumption about a specific geological load i.e. away from major modern orogenic belts. In addition due to the long wavelengths considered they tend to be applied across broad regions.

In this study modelling techniques are used to recover variations in T_e from observed gravity and topographic data. The techniques are applied to both 1D profile data and to 2D gridded data. Modelling methods were preferred as, in certain settings, they allowed consideration of the geological process occurring to create the flexure, such as emplacement of buried loads. Where no consideration of process is made, modelling methods assuming solely surface loading can provide an independent check on the results of spectral techniques.

In the following sections an introduction is made to both the profile and the grid based models of flexural surfaces used in this study. Methods for calculating gravity anomalies based on these models are also presented. An assessment is then made of techniques for recovering T_e , based on the flexural and gravity modelling methods discussed. This assessment has, in part, been presented by Jordan & Watts [2005].

Finally a brief introduction is made to spectral techniques. These were not directly applied in this study, however, the results of the modelling techniques are compared to the results of spectral techniques applied by other authors.

2.2 1D flexural models

2.2.1 Flexure on profiles

Many early studies of T_e in collisional systems were based on profiles and vertical line loads and/or moments applied to the end of a semi-infinite (i.e. broken) plate as in Karner & Watts [1983] and Lyon-Caen & Molnar [1983]. These methods were based upon analytical expressions for the deflection of thin elastic plates, with a constant T_e , overlying an inviscid fluid (e.g. Hetényi [1979]).

In the method of Hetényi [1979] the flexure, w , of an elastic plate in response to a load function $q(x)$, in the absence of horizontal forces, is described by:

$$D \frac{d^4 w}{dx^4} + gw(\rho_m - \rho_i) = q(x) \quad (2.1)$$

where g is the acceleration due to gravity, ρ_m is the density of the material underlying the flexed plate, ρ_i is the density of material infilling the flexural depression, and D is the flexural rigidity defined as:

$$D = \frac{ET_e^3}{12(1 - v^2)} \quad (2.2)$$

where E is Young's modulus, v is Poisson's ratio, and T_e is the elastic thickness of the plate.

More recent studies use distributed rather than line loads and either a finite difference (e.g. Stewart & Watts [1997]) or a finite element model (e.g. Cattin *et al.* [2001]) to simulate the bending of an elastic beam with variable T_e along its length. The profile based method used in this study follows the approach of Stewart & Watts [1997], in which the 1-dimensional finite difference method of

Bodine [1980] is used to calculate the flexure, and reducing T_e to zero simulates a plate break.

Bodine [1980] solved Equation (2.1) for the case where D is a function of x using finite difference approximations and a series of fundamental assumptions laid out below.

1) Linear elasticity i.e. that the stresses that develop due to flexing of the plate are linearly proportional to the strain.

2) The ‘thin plate approximation’, this requires that vertical stresses within the plate are negligible. This is valid in many geological settings where deflections are small, relative to the total lithospheric thickness. Where the thin plate approximation breaks down it has been shown by Comer [1983] and Wolf [1984] that the maximum discrepancy between the thin plate and the more complete thick plate solution is of the order of 5-10%. The greatest difference between the thick and thin plate models occurred beneath the load, however, the models are approximately equivalent beyond the load. Both Comer [1983] and Wolf [1984] concluded that simple elastic plate models were adequate for describing most geological situations.

3) The assumption of small deflections relative to the plate thickness, which means strains are infinitesimal and no stretching occurs along the neutral surface.

4) Cylindrical bending, meaning that any distortion is limited to the extreme edges of the plate. This means that flexure of a beam of unit width can represent the deflection of a plate, far from its edges, as long as the extra support due to it being part of a larger plate are considered. Timoshenko & Woinowsky-Krieger [1959] showed that when a plate that is three times longer than it is wide, and loaded at its edges, the resulting flexure will be cylindrical.

5) The plate overlies an inviscid substratum which subjects the plate to a buoyant restoring force proportional to the deflection, and the density contrast between the infill and the underlying substratum. It has been shown by Watts [2001] that the viscoelastic response of the oceanic lithosphere has a typical relaxation time,

τ , of 0.5 to 1.0 Ma. This is dependent on the viscosity of the mantle which controls how quickly it can flow to one side to allow compensation. Watts [2001] also provides estimates of τ in continental regions, based on isostatic rebound after ice loading, of 1.7 to 8.0 ka. Therefore, for ages greater than ~ 1 Ma, the assumption that the lithosphere can be modelled as a simple elastic plate overlying an inviscid substratum is valid, as any viscous support will have fully relaxed.

Bodine [1980] constrained his flexural models using a series of boundary conditions, allowing the solution of Equation (2.1) for loading due to a line load on a plate with varying rigidity. For a 1D profile, deflection beneath a spatially distributed load, $q(x)$, can be solved as a summation of the deflection due to a series of line loads calculated by subdivision of the distributed load. The boundary conditions used by Bodine [1980] are that the gradient of the flexural surface at the origin is zero and as x approached infinity the deflection and gradient go to zero.

van Wees & Cloetingh [1994] suggested that the original formulation of Bodine [1980] may not be correct due to his omission of certain cross-terms from the general flexure equation. To test this, the flexure computed using the Bodine method was compared with the analytical solutions of Hetényi [1979]. A 5 km high, 200 km wide rectangular load was considered, with one edge on the end of a broken elastic plate, $T_e = 90$ km and other parameters as defined in Table 2.1. In Stewart & Watts [1997] application of the Bodine method the plate break was simulated by a T_e that increased from 0 km to 90 km over a 100 km interval. The finite difference method gave a value of 20.28 km for the deflection at the plate break, while the analytical solution showed the maximum flexure to be 19.67 km. The two methods are, therefore, in close agreement.

Parameter	Symbol	Value
Poisson ratio	E	0.25
Young's modulus	v	10^{11} Pa
Substrate density	ρ_m	3330 kg m^{-3}
Infill density	ρ_i	2650 kg m^{-3}
Load density	ρ_l	2650 kg m^{-3}

Table 2.1: Parameters assumed in the flexure and gravity modelling.

2.3 2D flexural models

Spatial variations in the T_e structure may be inferred from a series of individual profiles. However, to further assess these variations a method is needed to compute the flexure due to a distributed load that is emplaced on a plate with a spatially variable T_e . The model used was based on the technique for modelling flexure of a plate with a spatially varying T_e proposed for example by Stewart & Watts [1997]. This method is based on a finite difference solution of the flexural equations for a 2D plate rather than a 1D profile. The finite difference method has previously been used by van Wees & Cloetingh [1994] and is as described in the appendix of Stewart & Watts [1997] and by Wyer [2003].

The equation for flexure (w) of a loaded thin elastic plate in the $x - y$ plane, overlying an inviscid substratum, with no horizontal forces applied to it is given by (e.g. Timoshenko & Woinowsky-Krieger [1959]):

$$\begin{aligned}
& D \nabla \nabla w + 2 \frac{\partial D}{\partial x} \frac{\partial}{\partial x} \nabla w + 2 \frac{\partial D}{\partial y} \frac{\partial}{\partial y} \nabla w + \nabla D \nabla w \\
& -(1 - v) \left[\frac{\partial^2 D}{\partial x^2} \frac{\partial^2 w}{\partial y^2} - 2 \frac{\partial^2 D}{\partial x \partial y} \frac{\partial^2 w}{\partial x \partial y} + \frac{\partial^2 D}{\partial y^2} \frac{\partial^2 w}{\partial x^2} \right] + \nabla \rho_1 g w = \nabla \rho_2 g h \quad (2.3)
\end{aligned}$$

Where ρ_1 = density contrast between infill and substrate, ρ_2 = density contrast

between load and the material it displaces, h is the load height, and

$$\nabla = \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} \quad (2.4)$$

In the method of Wyer [2003] the rigidity derivatives of Equation (2.3) are recovered by differentiation of the input variable T_e structure in the Fourier domain. Evaluation of the derivatives in Equation (2.3) can be approximated by substitution of centered finite difference terms as described by Gahli & Neville [1989].

The computer code for the flexural model used was initially developed by Bodine [1980] and Stewart & Watts [1997]. Wyer [2003] further improved the program by changing the way the matrixes involved in the calculation were handled, this considerably improved the speed of the program and made possible the analysis of larger grids. The boundary conditions imposed by Wyer [2003] are those of Airy compensation at all margins of the model, this was achieved by tapering the input T_e structure to zero at the margins. These boundary conditions were chosen as they remove the possibility of topography at the margins of the model which is not isostatically compensated, which would occur if flexure was constrained to zero at any margins of the model, as in Stewart & Watts [1997]. The problem of the margins of the model had previously been avoided by padding the data with zero values, however this generated very large grids, across which it takes a significant amount of extra time to calculate the flexural surface. Wyer [2003] has shown that the technique for modelling 2D flexure due to loading yields a flexural surface that agrees to within 1% of the analytical solution for a simple disc-shape load on an elastic plate with uniform T_e (Brotchie & Silvester [1969]).

2.4 Gravity anomalies on profiles

Loading and the resultant flexure generates deflection of horizontal surfaces within the lithosphere. This leads to the development of lateral density gradients, which generate gravity anomalies. In this study the anomalies that arise from these

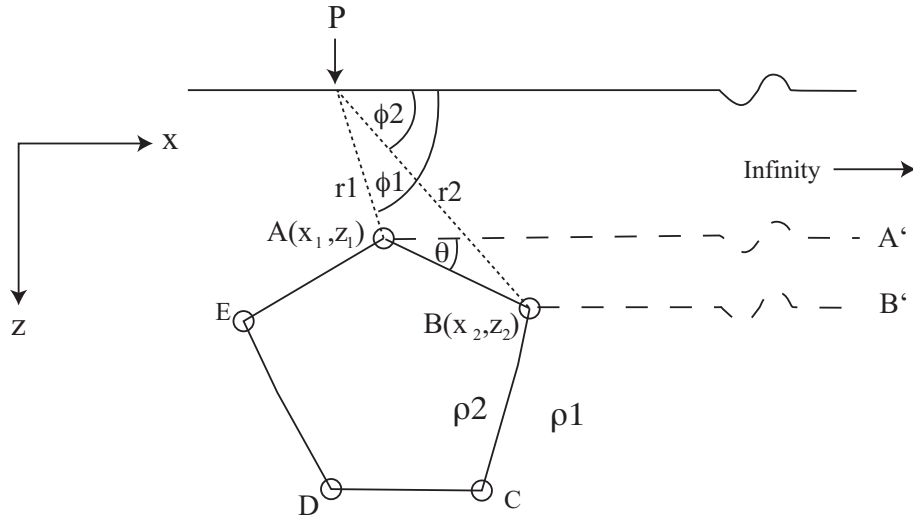


Figure 2.1: Geometry of body made up of a series of infinite horizontal layers. The gravity effect at P is the result of the sum of the effects of interfaces A-B, B-C, C-D, D-E and E-A.

deflected surfaces was calculated using the program **gbott**. **Gbott** is a 2D line integral method based on Talwani *et al.* [1959].

Any body, or undulating surface, can be broken down into a series of semi-infinite layers, each with a sloping end. **Gbott** considers the sum contribution of the gravity effect of the sloping ends of all these layers. An example of a closed body is shown in Figure 2.1, one layer within this body would be body A-A'-B'-B. As the layer has infinite extent in one direction the gravity anomaly is related only to the angle and height of the end interface AB. The gravity anomaly at a point P in Figure 2.1 resulting from the interface AB is given by Talwani *et al.* [1959] as:

$$\Delta g = 2G\Delta\rho(-[x_1\sin\theta + z_1\cos\theta][\sin\theta\log\frac{r_2}{r_1} + \cos\theta(\phi_2 - \phi_1)] + z_2\phi_2 - z_1\phi_1) \quad (2.5)$$

where G is the universal gravitation constant, $\Delta\rho$ is the contrast between ρ_1 and ρ_2 , and other parameters are as defined in Figure 2.1. The calculated anomaly at point P is then the sum of the effect of all the end interfaces on that point.

2.5 Gravity anomalies across grids

The calculation of the gravity anomaly derived from the flexure associated with variable rigidity grids, uses the method of Parker [1972] for rapid calculation of potential field anomalies resulting from 3D surfaces. Parker [1972] showed that the gravity anomaly, Δg , at a surface, $z = z_0$, for a layer that has an upper surface at $z = z_1$ and a lower surface at $z = z_2$ is:

$$\mathcal{F}[\Delta g] = -2\pi G e^{-|k|z_0} \sum_{n=1}^{\infty} \frac{(|k|)^{n-1}}{n!} \mathcal{F}[\Delta \rho(z_1^n - z_2^n)] \quad (2.6)$$

where $\mathcal{F}[\]$ represents the Fourier transform of the variable in brackets, G is the universal gravitation constant, k is the wave number and $\Delta \rho$ is the density contrast across the surface. Equation 2.6 can be solved by Taylor's series expansion. Only terms up to $n=4$ of the expansion are considered in the calculation of the gravity anomaly, as terms above this have been shown to have a negligible impact on the modelled gravity anomaly (Parker [1972]).

2.6 Recovery of 1D T_e

2.6.1 Assessment of the 1D, profile based, finite difference technique

In the 1D method of Stewart & Watts [1997] modelled flexural profiles across foreland regions are calculated using the method of Bodine [1980] for a range of T_e values and plate break positions. **Gbott** is then used to calculate the gravity anomalies resulting from two interfaces, one interface is considered to be at the foreland basin/crust boundary and the second is considered to be at the crust/mantle boundary. These calculated gravity anomalies are then combined and compared to observed Bouguer gravity anomaly data. The best fit is then determined as the

value of T_e that minimises the difference (i.e. minimum Root Mean Square (RMS)) between the calculated and observed anomaly.

To verify the recovery of T_e using the method outlined above, a synthetic gravity anomaly was produced, based on the flexural surface derived from the analytical solution for a rectangular load given by Hetényi [1979]. A five km high and 200 km wide surface load, as shown in Figure 2.2a, was considered. This synthetic model represented the gravity anomaly derived from the two interfaces discussed previously.

For this test of the profile method of Stewart & Watts [1997], gravity anomalies were calculated for a range of load widths from 50 to 350 km and for T_e values ranging from 0 to 200 km. The results of the comparison between the synthetic anomaly and the 1D profile method are shown in Figure 2.2.

The T_e found to best fit the analytical solution by the 1D finite difference model was 93 km, only 3.3 % higher than that input in the synthetic. The plate break position which best fit the synthetic model was at 200 km from the load front, the same as that used in the synthetic model. This shows that the profile method of Stewart & Watts [1997] recovered well the input T_e value and plate break position, confirming the robustness of this method.

2.6.2 Alternative techniques

An alternative technique for recovering T_e along profiles, that has previously been applied in foreland settings, is to solely consider the shape of the observed gravity anomaly beyond the load edge (e.g. McKenzie & Fairhead [1997] and McKenzie [2003]). In this technique the shape of the observed Free air anomaly is matched by a theoretical curve derived from the flexural equation;

$$w = e^{\lambda x}(A \cos \lambda x + B \sin \lambda x) \quad (2.7)$$

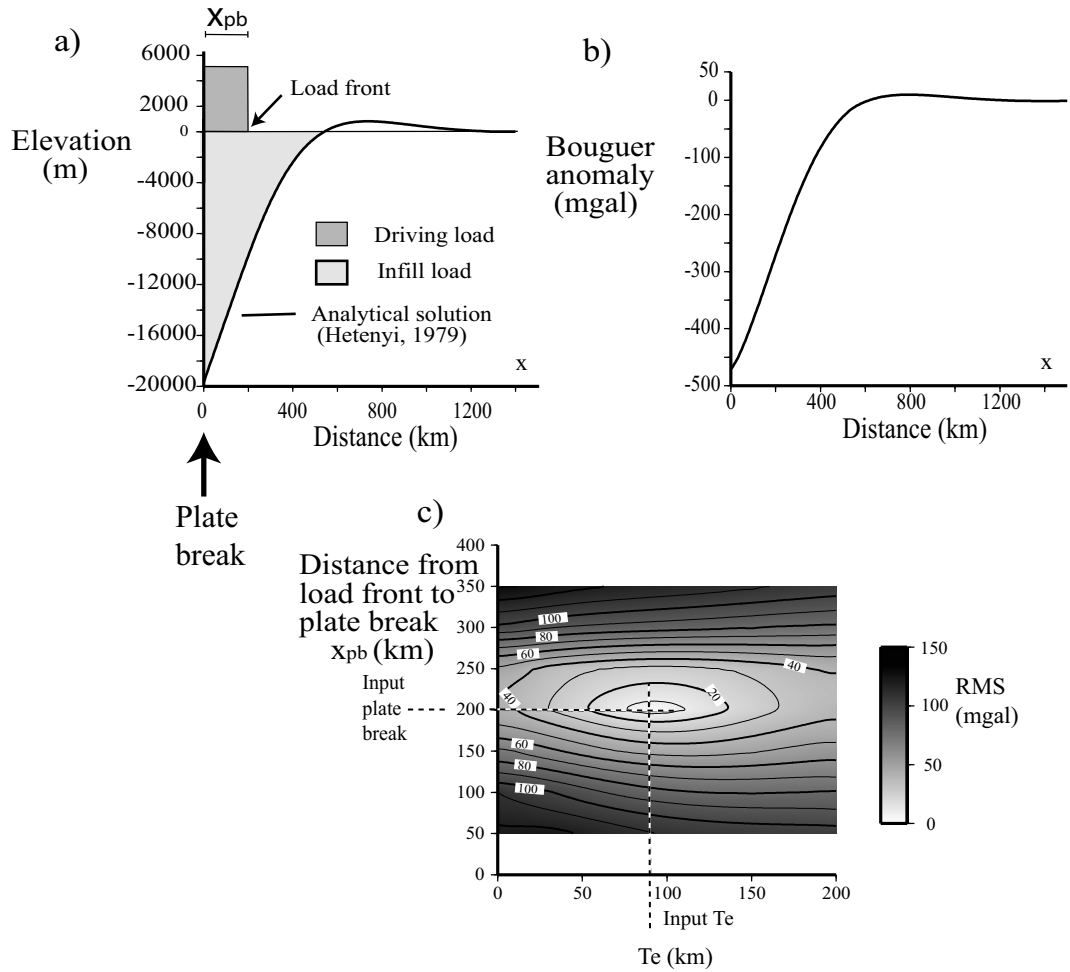


Figure 2.2: Comparison of the analytical solution of Hetényi [1979] for a discontinuous (i.e. broken) plate and the finite difference method. a) Analytical solution based on an input $T_e = 90$ km and distance from plate break to load front (x_{pb}) = 200 km. The driving load is 5 km high and 200 km wide. The infill load is the material that infills the flexure. b) Calculated Bouguer gravity anomaly assuming two uniform density contrast interfaces, one at the top of the flexed crust and the other at the base. The position of load front was fixed to be the same as the Synthetic model while the position of the plate break was varied. c) Plot of the RMS difference between the calculated Bouguer gravity anomaly, beyond the load front, based on the analytical solution of Hétenyi and a ‘synthetic’ anomaly calculated using a finite difference model with a range of T_e values and plate break positions.

where A and B are constants and,

$$\lambda = \left[\frac{(\rho_m - \rho_i)g}{4D} \right]^{1/4} \quad (2.8)$$

In this technique there is no consideration of load size or plate break position, the best fit between observed and calculated curves is simply recovered by varying A , B and D (and hence T_e). Both the shape fitting and finite difference methods are based on the observed gravity anomaly and should yield a similar recovered T_e .

Figure 2.3a shows the range of calculated Bouguer gravity anomalies for a fixed load width and plate break, based on the finite difference method, assuming two interfaces, between foreland basin sediments and crust, and crust and mantle. Also shown is the RMS difference between these models and the synthetic anomaly derived from Hetényi [1979] for this load and interface combination. This is equivalent to a section through Figure 2.2c for a plate break position of 200 km. Figure 2.3b shows the curves for various T_e values, based on the method of McKenzie & Fairhead [1997], which best fit the synthetic gravity anomaly derived from *only* the Moho interface and the flexural model of Hetényi [1979]. Initially only the Moho interface was considered as the method of McKenzie & Fairhead [1997] assumes the observed anomaly stems from a single interface.

The inset in Figure 2.3b highlights the recovery of the correct T_e value by the method of McKenzie & Fairhead [1997], although the RMS minimum is less well defined than for the finite difference case. Figure 2.3b shows the best fitting flexural shapes for a range of T_e values based on the method of McKenzie & Fairhead [1997]. As no driving load is considered, the flexural curves are not as well constrained as in the finite difference technique. The best fitting low T_e curves flex up at the load edge, while the flexural bulge of the high T_e curves are closer to the load edge than could be explained by the input driving load. Including load parameters, as the finite difference technique does, effectively limits the parameter space assessed and hence leads to a better defined minimum. The method of McKenzie & Fairhead [1997] was applied to a multi-interface synthetic model equating to the free air

anomaly, assuming 3 interfaces; the Moho, the base of the foreland basin and the flexural bulge. The results of this test are shown in Figure 2.3c. In this case a poorly defined RMS is recovered, the minimum of which does not match the T_e assumed in the synthetic model. The results of these synthetic tests suggest that a finite difference method which takes into account the contribution of the Bouguer anomaly of both the top and base of the flexed crust, gives more realistic solutions for the flexed surfaces and a better defined RMS minimum than the method of McKenzie & Fairhead [1997].

The 1D profile techniques of previous authors such as Karner & Watts [1983] and Stewart & Watts [1997] are therefore followed for recovering T_e from profiles. In particular it is assumed that the topography defines the surface load and that the gravity effect of the flexure arises from two interfaces, not one. The parameters used throughout this study are as defined in Table 2.1. The values of Young's modulus and Poisson's ratio are widely used in continental flexural studies. These values may vary but to allow comparison with the results of other studies these values were considered to be fixed. The densities of load and infill vary. However, they have a relatively small impact on the recovered T_e , as shown by Lyon-Caen & Molnar [1985]. Further discussion is given to these parameters in later sections.

2.7 Recovery of a 2D spatially varying T_e

In this study maps of spatially varying T_e were recovered based on the techniques for modelling flexure and the derived gravity anomalies across grids outlined above. The technique applied is similar to the 1D finite difference technique for recovering T_e along a profile, as the calculated, flexurally derived, gravity anomaly is compared, in an RMS sense, to observed data. However, the fit between calculated and observed anomalies is calculated in overlapping windows and local 'best fitting' T_e values are recovered.

The process for recovering the spatially varying T_e across a region has two

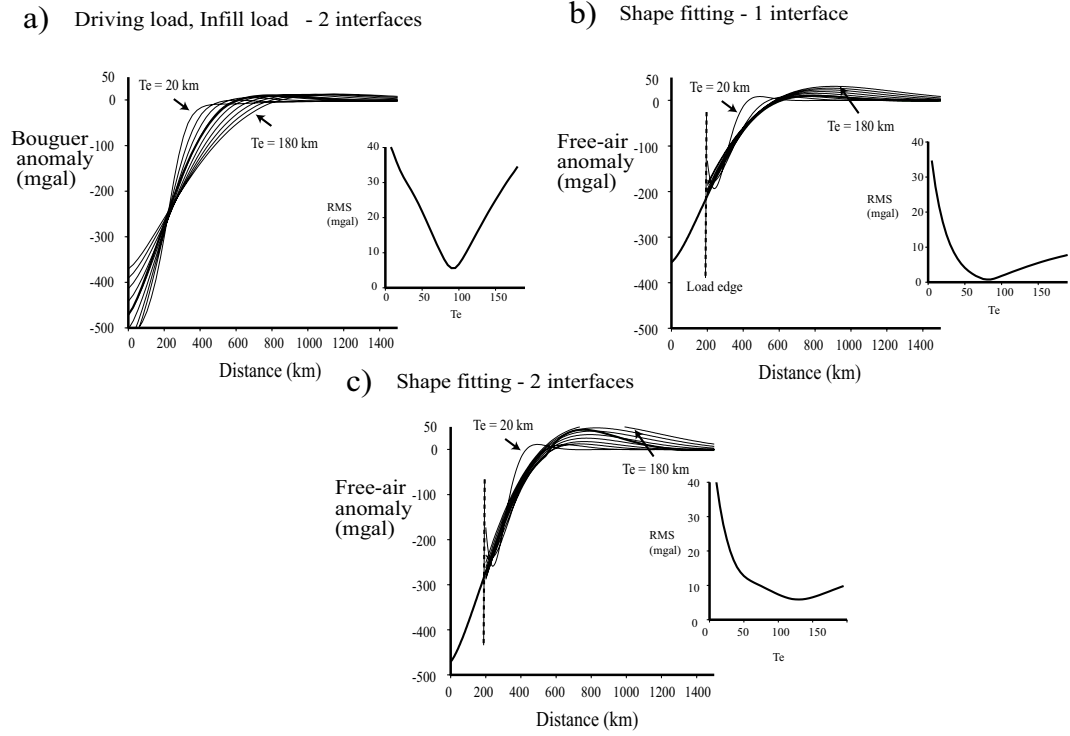


Figure 2.3: Comparison of the finite difference method and the shape fitting technique of McKenzie & Fairhead [1997] when recovering an input T_e from a synthetic gravity anomaly defined by the analytical solution of Hetényi [1979]. a) Calculated Bouguer gravity anomalies, based on the finite difference method (thin lines). Thick line shows a ‘synthetic’ anomaly based on $T_e = 90$ km, $x_{pb} = 200$ km and two interfaces. Inset shows RMS difference between Bouguer anomaly associated with the finite difference model and the synthetic anomaly. b) Calculated gravity anomalies based on the shape-fitting technique, beyond the load edge, of McKenzie & Fairhead [1997] (thin lines) assuming a single uniform density contrast interface at the Moho. The thick line shows the synthetic for the Moho interface only. c) Calculated gravity anomalies, beyond the load edge, based on the shape-fitting technique of McKenzie & Fairhead [1997] (thin lines) compared to a multiple interface (free air) synthetic.

parts. The first follows Braitenberg *et al.* [2002] and Wyer [2003], and is based on a comparison of observed Bouguer gravity anomaly data and calculated gravity anomalies. For each of a series of plates with laterally homogeneous T_e values from 0 to 150 km, in 10 km steps, the flexure below the input topographic load and its associated Bouguer gravity anomaly is calculated. The calculated anomaly is based on two assumed interfaces, one at the infill/crust boundary and a second at the crust/mantle boundary. Each calculated anomaly is then compared to the observed Bouguer anomaly based upon a ‘5 by 5’ point window and the RMS difference calculated. The RMS difference for each window is assigned to the window centre and the window is shifted by one point. This process of windowed comparison generates a RMS difference map for each constant T_e structure. The T_e is selected which gives the lowest RMS difference at each point in the grid. Finally, the resulting T_e structure is smoothed using a Gaussian filter to produce an ‘initial variable T_e structure’.

The necessity of filtering the T_e structure is shown in Figure 2.4, and stems from the fact that the derivatives of the rigidity structure are calculated using a FFT routine during the flexural calculation. This returns results which are infinite if the input T_e gradients are too high. Figure 2.4 shows that when the T_e structure is unfiltered the steep gradient prevents a stable flexural surface being calculated. As the amount of filtering applied to the T_e structure increases this instability decreases but it does not disappear fully until a filter of 300 km width is applied.

Various filter types were examined (see Figure 2.5). A band pass filter was rejected as it created a distinct ringing in the filter output when a very sharp edge was encountered. This can be seen to perturb the T_e structure at long distances from the edge in question, as shown in Figure 2.5. Also negative values are introduced into the T_e structure which are not physically possible. Box car, Gaussian and cosine filters were also assessed. The Box car filter appears to give a relatively sharp transition from constant high, to decreasing T_e , while the Gaussian

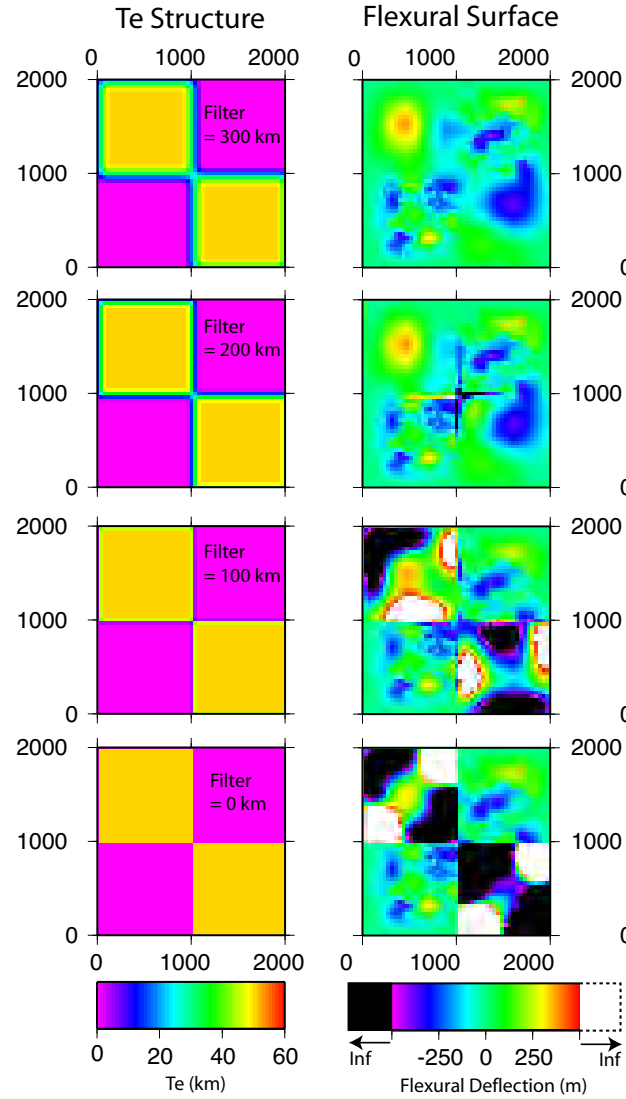


Figure 2.4: The result of loading a check-board T_e structure having applied various levels of Gaussian smoothing (taper) to the input T_e structure. The input T_e structure had values of 0 or 50 km for T_e and a grid point spacing of 40km. The input topography was of the order of -630 to 660 m. The output flexure with various filter widths (W) was as follows: $W=0$ km -3406 to 14202 m, $W=100$ km -1697 to 1220 m, $W=200$ km -1575 to 797 m, $W=300$ km -401 to 372 m.

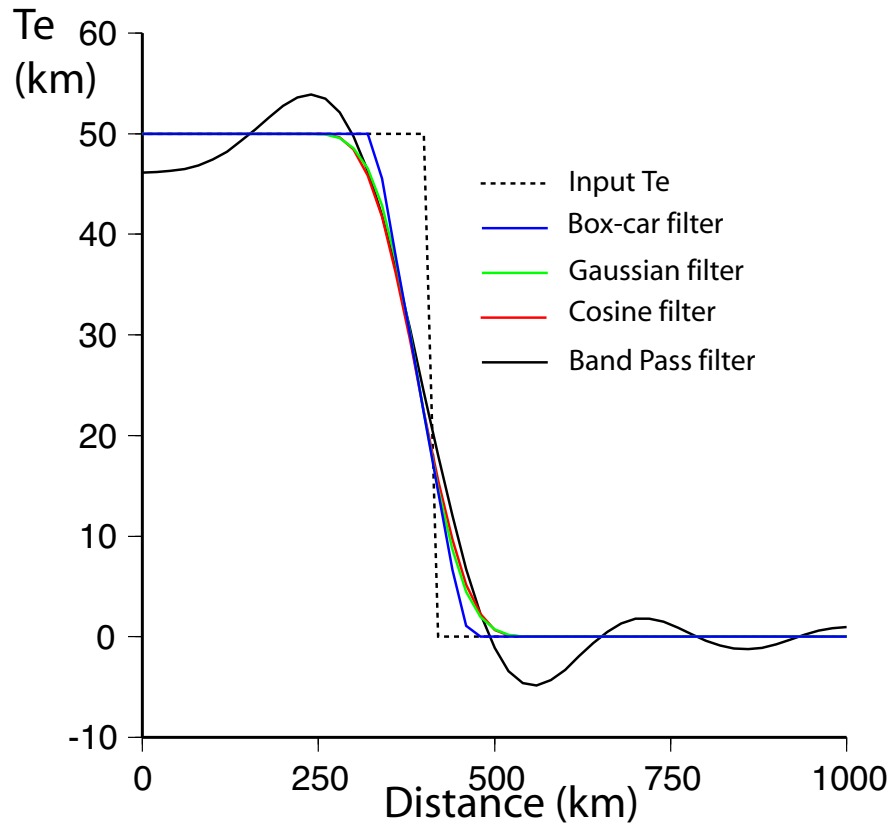


Figure 2.5: Cross sections showing a comparison of original and filtered T_e structures. Space domain filters are; Red Cosine (300 km), Green Gaussian (300 km), Blue Box Car (150 km). The band pass filter (black solid line) was designed as a low pass filter, passing wave lengths above 300 km, cosine tapering wavelengths between 300 and 200 km and cutting wavelengths below 200 km.

and cosine filters give a smoother transition. As both the cosine and Gaussian filters appear to give an equally smooth transition between high and low T_e , the Gaussian filter was arbitrarily chosen.

The second stage in recovering the regional T_e structure is an iterative one which refines the ‘initial variable T_e structure’. This technique follows that used by Wyer [2003] and Wyer & Watts [2006] for recovering the T_e structure along the eastern continental margin of the United States. These workers iteratively compared observed free air gravity anomalies to calculated anomalies predicted from a process orientated model of rifting and sedimentation on the margin.

In the current study the flexure and calculated Bouguer gravity anomalies associated with the ‘initial variable T_e structure’ are calculated, based on loading with the observed surface topography. Also calculated are flexural surfaces and gravity anomalies associated with shifting the entire ‘initial variable T_e structure’ by -4, -2, +2 and +4 km. At points where T_e was shifted to below 0 km the value at that point was reset to 0 km. RMS difference maps between the five calculated and the observed Bouguer anomaly were then produced using the same windowing technique as in the first section.

The shift in T_e at each point which gives the lowest RMS difference is applied to the ‘initial variable T_e structure’. This new variable T_e structure is then smoothed with a Gaussian filter. This structure is passed back to the start and the process repeated until the model is deemed to be stable. Finally, an assessment of the stability of the model is made based on how much T_e has changed over the final five iterations.

2.7.1 Tests of the 2D iterative technique

Several assessments of the reliability of the technique were carried out based on synthetic data sets generated using a known T_e structure.

Synthetic tests

The reliability of the recovery of a variable T_e structure by the 2D iterative method was tested by first constructing a synthetic gravity anomaly and topographic data set based on a known T_e structure. The 2D iterative technique was then applied to the synthetic data and the input T_e recovered.

These tests of the 2D iterative technique were presented in the paper of Jordan & Watts [2005]. The synthetic data is generated using the technique of Pérez-Gussinyé *et al.* [2004]. The modelled flexure is due to a surface fractal load (L_s) (Peitgen & Saupe [1988]; Macario *et al.* [1995]) superimposed on a 2-dimensional

elastic plate with a spatially varying rigidity (Pérez-Gussinyé *et al.* [2004]; Wyer [2003]). The resulting flexure (F_s) was calculated using the program of Wyer [2003], with an infill density of zero. The effect of buried loading was also included in some of the synthetic models using the technique of Pérez-Gussinyé *et al.* [2004]. These buried loads (L_b) were assumed to be emplaced at the Moho, to be fractal, and statistically un-correlated with the surface topography. The size of the buried load was scaled so that in the case where the ratio of surface to buried loading (f) is equal to one, $L_bmax \times (\rho_b - \rho_m) = L_smax \times \rho_s$, where L_bmax and L_smax are the magnitudes and ρ_b and ρ_s are the densities of the buried and surface loads respectively. The flexure due to the buried loads (F_b), applied to the variable T_e structure, was then calculated using the same method as for the surface loads. As the infill density assumed was zero the loads 'fall into' the flexural depressions they create and the final topography (T) is therefore calculated as:

$$T = L_s + F_s + F_b \quad (2.9)$$

and the final Moho M was calculated as:

$$M = L_b + F_b + F_s \quad (2.10)$$

The Bouguer gravity anomaly was calculated, assuming a single interface at the crust mantle boundary M (i.e. load density = crust density).

Figure 2.6 shows how the iterative method recovers T_e for a range of noise levels in the synthetic data and ratio, f , of surface to sub-surface loading (Forsyth [1985]). The figure shows that in the absence of noise (i.e 0%) and sub-surface loading (i.e. $f = 0$) the shape of the input T_e structure is recovered well. The maximum value of T_e recovered was 58.2 km compared to 50.0 km that was input. This overestimate is probably due to the fact that the modelled topography includes flexural bulges. The bulges are treated as loads by the iterative technique and therefore should be removed from the input topography. Unfortunately, this is difficult to do without

a priori knowledge of the T_e structure.

The recovery of T_e is not as good when either the noise levels or f are increased. The sensitivity to noise was evaluated by adding 5% and 10% random Gaussian noise to the synthetic Bouguer gravity data for $f = 0$. Although the shape of the input T_e structure is retained, the absolute values of the recovered T_e increase, particularly over the highest T_e areas. This increase, however, is relatively small (maximum T_e with 10% noise is 64 km compared to 50 km). The sensitivity to f was considered for values of f of 0.25 and 0.75. Again, the shape of the input T_e structure is retained. However, as the amount of sub-surface loading increases ($f = 1$ corresponds to an equal amount of surface and sub-surface loading), the maximum values recovered decrease. This is because at high T_e sub-surface loads have small topographic expressions relative to the mantle topography, which generates the Bouguer gravity anomaly. When the T_e is low, however, (as it is towards the edges of the model) then the recovery is good.

Stability and sensitivity

The stability and sensitivity of the results of the 2D iterative technique were assessed as discussed below. The stability of the final model is assessed based on the standard deviation of the recovered T_e at each grid point over the final five iterations. This test was carried out on the recovered synthetic T_e structure assuming no additional noise and a f ratio of zero. The results of this assessment are shown in Figure 2.7a. The variation seen at a single stable and a less stable point are shown in Figure 2.7b and c respectively.

The sensitivity of the model is a measure of how much the RMS changes at each point with variation of T_e . This test is based on the 5 RMS maps produced during the final iteration of the model. The calculated anomalies are based on shifts applied to the T_e structure of -4, -2, 0, 2 and 4 km. When the model is sensitive to variation in T_e a broad range of RMS values are seen at a point, and

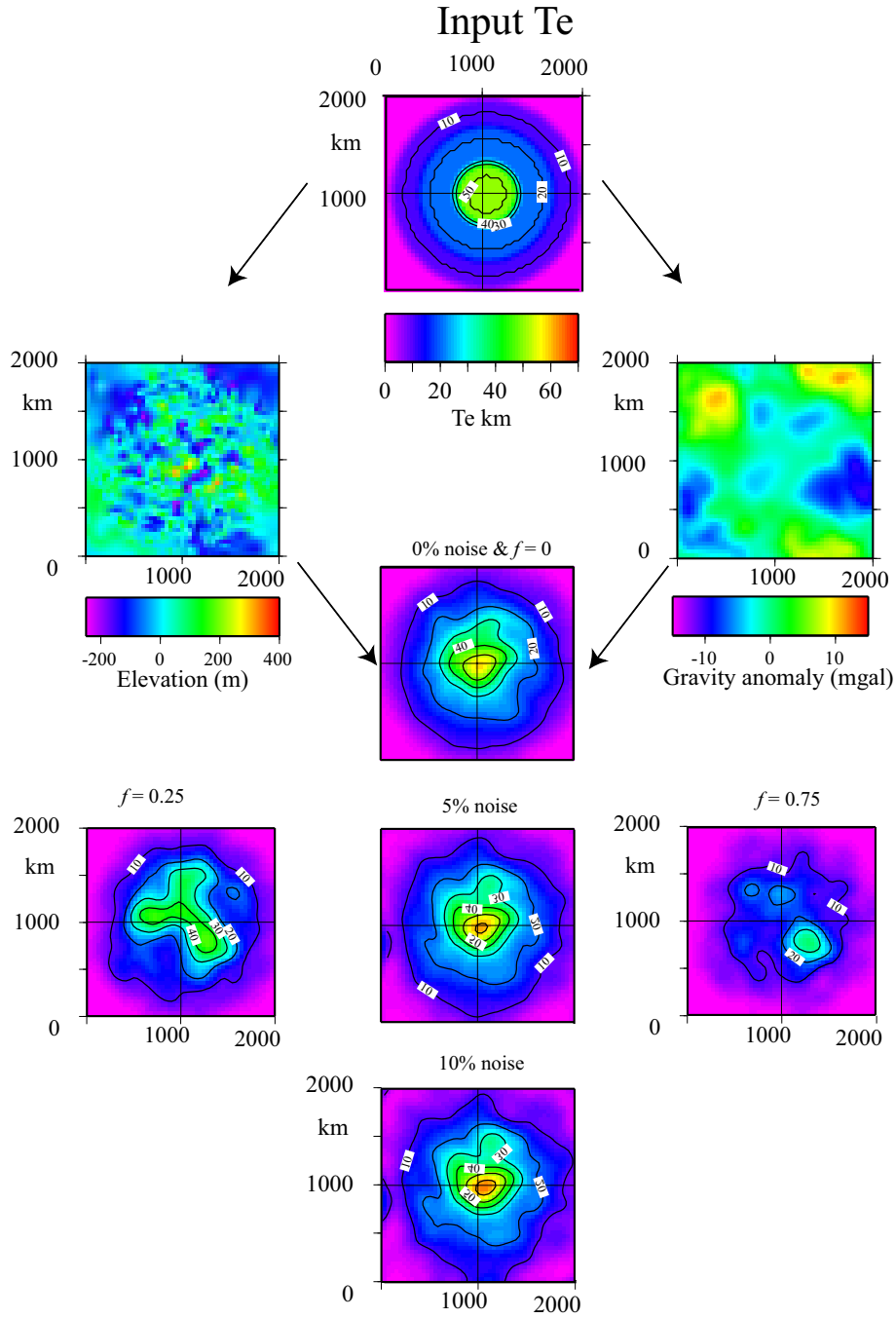


Figure 2.6: Synthetic tests of the 2-dimensional iterative method. The upper 3 panels show the synthetic topography and gravity generated by loading an elastic plate with the T_e structure shown. The central 3 panels show the recovered T_e structure with surface loading only (i.e. $f = 0$) and increasing amounts of noise. The lower left and right panels show the recovered T_e structures with $f = 0.25$ and $f = 0.75$, no noise, and with sub-surface loading occurring at the base of the crust. The recovered T_e structures were all smoothed using a 400 km Gaussian filter.

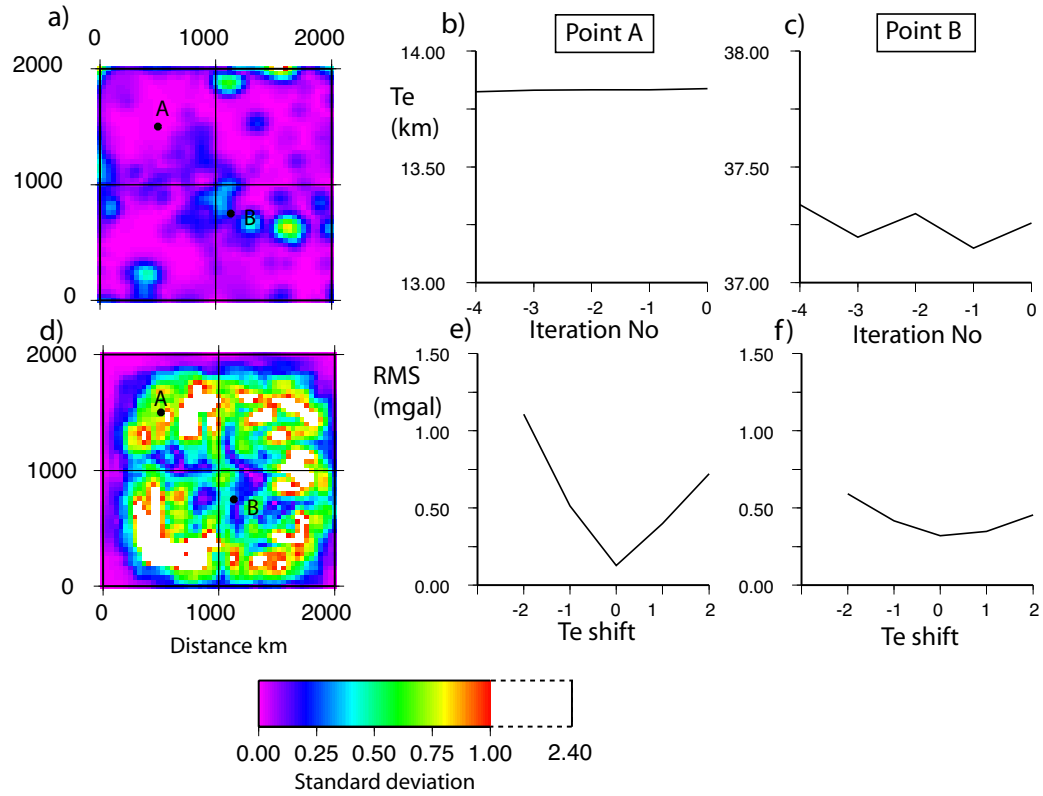


Figure 2.7: Assessment of stability and sensitivity of final iteration of the 2D iterative technique. a) Stability of the final model represented by the standard deviation of T_e at each point over the final 5 iterations of the model. b) Example of T_e variation at a stable point such as point A over the final 5 iterations of the model. c) Example of T_e variation at a less stable point such as B. d) Sensitivity of model to T_e variation as defined by the standard deviation of the RMS error in mgal between observed and calculated anomalies induced by a ± 4 km T_e variation. e) Variation in RMS between observed and calculated anomaly at high sensitivity point A. f) RMS variation at low sensitivity point B.

the associated standard deviation of RMS will be high. Where the model is less sensitive to T_e variation, low standard deviations of the RMS will be found. The results of this are shown in Figure 2.7d. The RMS variation at two single points is shown in Figure 2.7e and f, representing high and low sensitivity respectively.

These tests with a synthetic gravity and topography data set show that the 2D iterative method recovers an input T_e structure well. The best recovery is for low noise levels and small amounts of sub-surface loading. This is true even in the

case of subdued topography, the maximum used in the synthetic tests being in the range ± 400 m.

Incorporation of mechanical discontinuities, such as breaks, into a 2D elastic plate model, such as that of Wyer [2003], is not possible, as the FFT routines involved require a smoothly varying T_e structure. Wessel [1993] and Wessel [1996] have included the presence of plate breaks within grid based models of flexure, however, these methods require the assumption of the position of the plate break that, in many geological settings, is not reliably known. In addition, it has been shown in tests with synthetic data, using 1-dimensional applications of the finite difference method to calculate flexure, that a T_e which decreases to 0 km over a short distance is, in fact, a good proxy for a plate break (see Section 2.6.1).

Figure 2.8 shows a comparison between an analytical solution for flexure due to a rectangular driving load and the 2D flexural model of Wyer [2003] for a bar shaped load with the same cross section. The analytical model is based on the 1D analytical solution of Hetényi [1979]. The results of the 1D solution were extrapolated in the y direction to generate a gridded flexural surface. The bulge positions match well between the two models and the point where maximum flexure occurs is displaced by ~ 40 km. This agreement suggests that T_e decreasing to zero is a valid approximation for a broken plate in the 2D case. The maximum difference between the two models is 2173 m which is $\sim 11\%$ of the total deflection at that point. The discrepancy is larger than for the 1D finite difference comparison due to the shallower gradient over which T_e falls from high to low. The shallower gradient of T_e variation is unavoidable in the 2D flexural program as an infinite result was given if the gradient of the T_e structure was higher.

In the 2D case T_e falling to zero along strike of the entire length of a mountain belt would indicate the presence of a broken plate, as the surface loads on either side of the low T_e zone would not effect the flexural response on the opposite side. The filter imposed to prevent steep gradients in the T_e structure would act to reduce

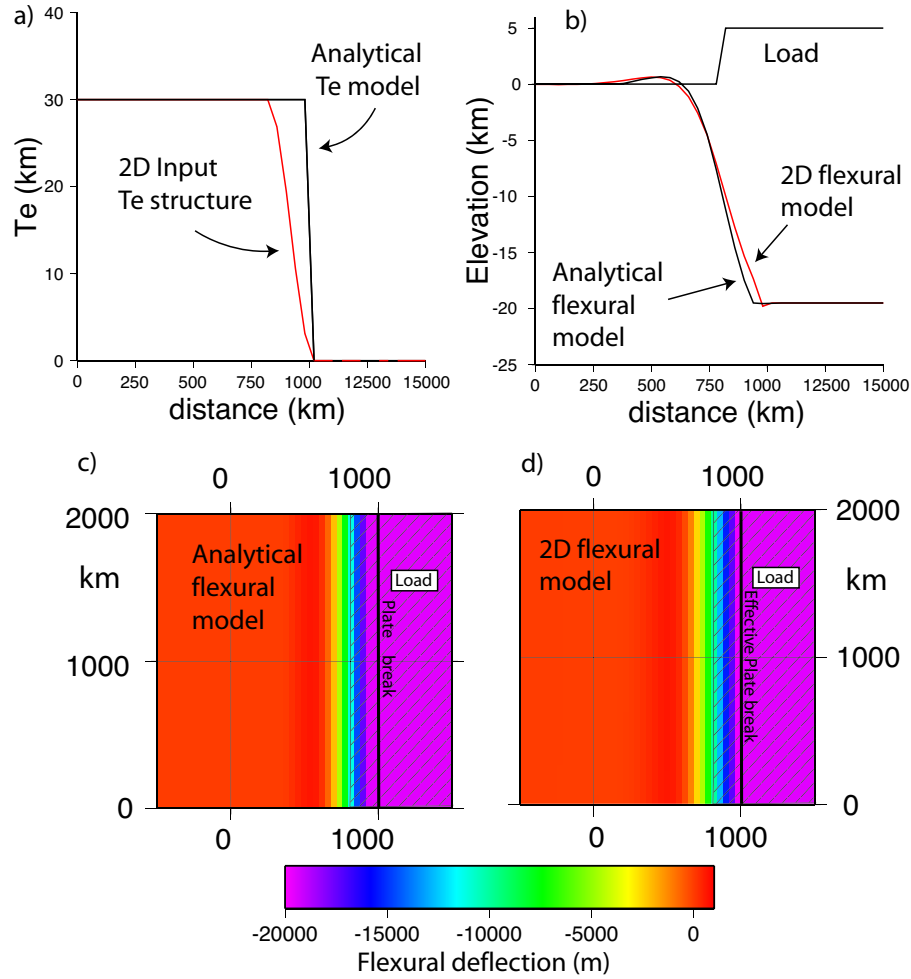


Figure 2.8: Comparison between an analytical flexural model for loading of a broken plate and the 2D flexural model where T_e falls to zero. Analytical model is based on the 1D solution of Hetényi [1979]. 2D flexural model based on the computer program of Wyer [2003] with the input T_e structure shown in a. a) Input T_e structures, black = analytical model, red = 2D model. b) Sections through resulting flexural surfaces. c) Map of flexural surface based on extrapolation of a 1D analytical model in the y direction. d) Flexural surface derived from 2D model.

the width of any low T_e zone and hence some coupling across an actually broken plate boundary may occur.

2.8 Spectral techniques

These techniques rely on analysis of the relationship between gravity and topography as a function of wavelength. This is done either by recovering the transfer function (admittance) between the observed topography and free air or Bouguer gravity anomalies, or the coherence between observed topography and the observed Bouguer gravity anomaly. The admittance, or coherence, of topography and gravity data has generally been assessed based on analysis the frequency spectra derived from Fourier transforms of the observed data. More recently wavelet transforms have been used to derive the power-spectra (e.g. Kirby & Swain [2004]). Fourier transforms consider the transform of all wavelengths either assuming infinite periodicity or all wavelengths within a certain window size. Wavelet transforms effectively consider a smaller window for short wavelengths and larger windows for long wavelengths. Kirby & Swain [2004] suggest that this improves the resolving power relative to Fourier transform techniques. In the formulation of Kirby & Swain [2004] ‘equivalent Fourier wave numbers’ are calculated using wavelet transforms, thereby making this transform technique applicable to Fourier transform methods of recovering admittance and coherence estimates of T_e previously developed (e.g. Forsyth [1985]).

In oceanic regions, transfer function techniques based on free air anomaly and bathymetric data have been very effective in recovering T_e , e.g. McKenzie & C.Bowin [1976], Watts [1978]. In continental regions, which are the focus of the current study, the Bouguer gravity anomaly has commonly analysed, and it is this application which is predominately discussed below.

2.8.1 Bouguer admittance

For a periodic topography, short wavelength features can be rigidly supported by an elastic plate with no deflection of the plate. For very long wavelength features the response of the same elastic plate will be equivalent to that predicted by the Airy model of isostasy. Topography between these two end-member situations will be flexurally supported.

Forsyth [1985] states that the gravity anomaly resulting from support of various wavelengths of topography by an elastic plate can be described by the isostatic response function, or admittance, $Q(k)$. This can be calculated from Dorman & Lewis [1970]

$$B(k) = Q(k).H(k) \quad (2.11)$$

where $B(k)$ and $H(k)$ represents the fourier transforms of the space domain Bouguer gravity and topography respectively, and k is the wave number $= 2\pi/\lambda$. To avoid bias due to noise the average admittance ($\hat{Q}(k)$) within discrete wave number bands is estimated from:

$$\hat{Q}(k) = \frac{\langle B \cdot H^* \rangle}{\langle H \cdot H^* \rangle} \quad (2.12)$$

where $\langle \rangle$ indicate averaging over discrete wave number bands and $*$ indicates the complex conjugate.

The Bouguer admittance for surface loading of a thin elastic plate is given by Banks *et al.* [1977] as:

$$Q(k) = -2\pi G \rho_c e^{-kt_c} / \zeta \quad (2.13)$$

where t_c is the mean depth to the Moho and ζ is given by

$$\zeta = \left[1 + \frac{Dk^4}{(\rho_m - \rho_c)g} \right] \quad (2.14)$$

where ρ_m is the mantle density and ρ_c is the crustal density.

If the admittance can be calculated for observed data based on Equation (2.11) it can be compared to theoretical curves resulting from Equation (2.14) such as the lower dashed lines shown in Figure 2.9a and the best T_e value recovered. The admittance technique has been applied by previous authors such as Dorman & Lewis [1970] and Banks *et al.* [1977].

2.8.2 Bouguer coherence

Forsyth [1985] considered, in addition to the effect of surface loading, the effect of buried loading on the observed transfer functions between topography and Bouguer gravity data. Forsyth [1985] suggested that it was reasonable to construct transfer functions based on loading of the lithosphere from beneath, representing, for example, magmatic underplating at the Moho. As is shown in Figure 2.9a the admittance response due to buried loading is very different to that for surface loading. Forsyth [1985] argued that this sensitivity to buried loading reduced the accuracy with which T_e could be recovered if buried loads were present. This was shown to be particularly important when terranes with a range of rigidities were analysed together.

To improve the recovery of T_e Forsyth [1985] instead considered the coherence between the Bouguer gravity anomaly and the topography. Consider an infinitely rigid beam with two equivalent loads imposed on it some distance apart, one from beneath (L_b) and one on the surface (L_s). The gravity anomaly derived from L_b at the Moho will have no correlation with the surface topography L_s . If the beam was considered to have a rigidity of zero the surface load L_s would be the same as the compensation of the buried load, F_b and vice versa. The gravity anomaly associated with the compensation of the surface and buried loads would exactly correlate with the surface topography. Between these two end members the coherence changes in a predictable manner, determined by the T_e and the ratio f

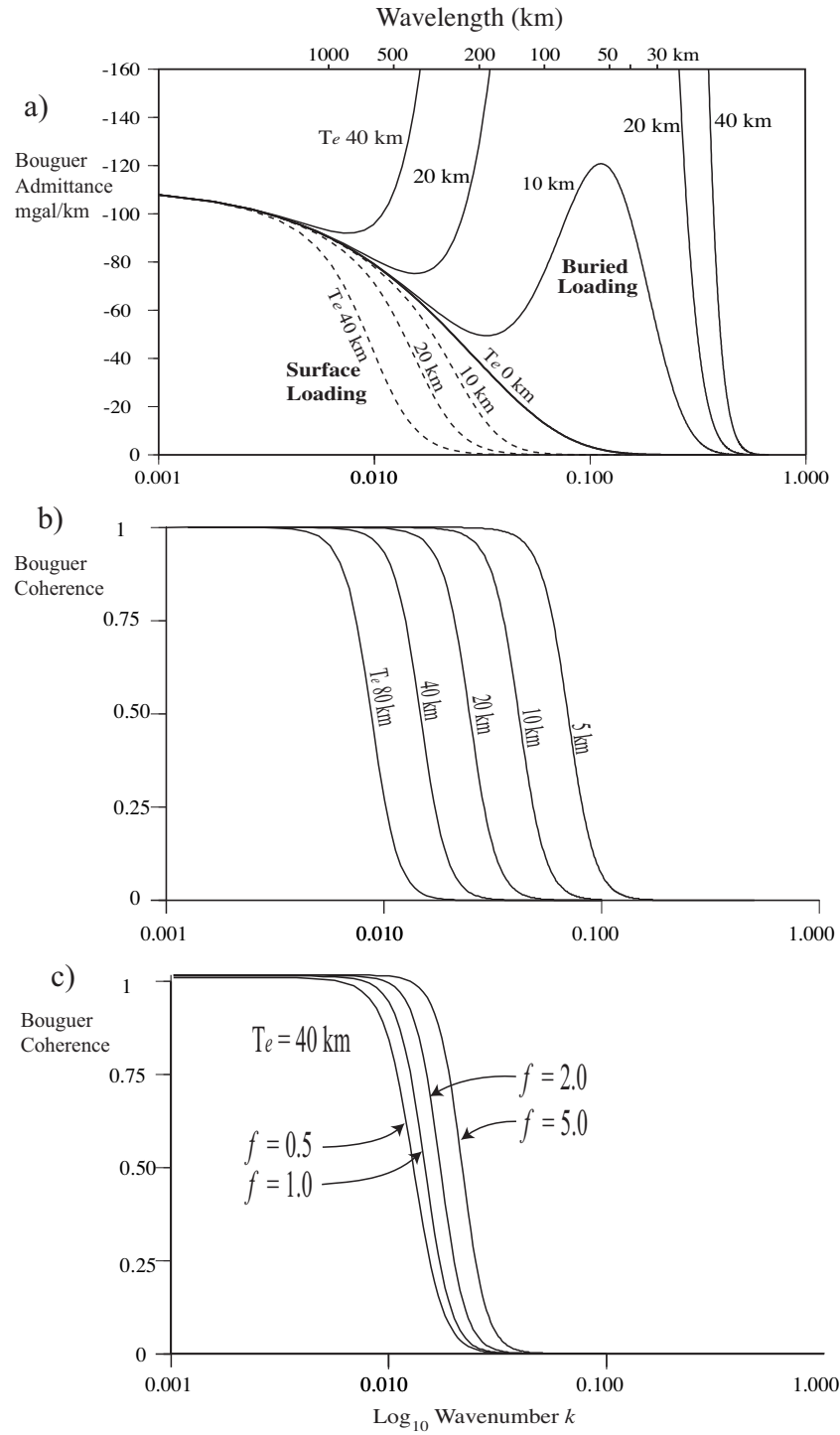


Figure 2.9: Response functions for theoretical admittance and coherence between gravity and topography data after Watts [2001]. a) Bouguer admittance functions for T_e values of 0, 10, 20 and 40 km. Dashed lines show response for surface loading only while solid lines show the predicted admittance curves given sub surface loading. b) Theoretical coherence between gravity and topography for T_e values of 5, 10, 20, 40 and 80 km and a ratio of buried to surface loading (f) of 1. Coherence is defined as the cross-spectra between topography and gravity anomaly divided by the spectra of the topography multiplied by the spectra of the gravity anomaly. c) Variation in coherence due to variation in the proportion of buried loading f .

of surface to buried loading. The variation of the coherence response function in response to T_e is shown in Figure 2.9b. The f ratio affects the coherence response function, however, Forsyth [1985] showed it has less impact than T_e . Figure 2.9c shows the variation in coherence response function due to variation in f . In the method of Forsyth [1985] a best fitting T_e is first recovered, the f ratio is then adjusted to fit the data and the T_e is reassessed using the new f ratio. This process is repeated until a stable result is achieved.

The Bouguer coherence technique have been used by several authors to recover continental T_e estimates in the presence of buried loading, for example, in western Canada by Flück *et al.* [2003] and in Europe by Pérez-Gussinyé *et al.* [2004].

Recently McKenzie [2003] argued that the Bouguer coherence technique only provides an upper bound for continental T_e . It was argued that in regions of subdued topography, with considerable sub-surface variation in density, the Bouguer gravity anomalies will tend to remain uncorrelated with topography until longer wavelengths than expected given the T_e , leading to artificially high recovered T_e values. Instead the technique of recovering continental T_e based upon free air admittance was preferred as McKenzie [2003] argued that correlated buried loads which alter the shape of the admittance function were not a significant factor. Using free air admittance techniques McKenzie [2003] recovered consistently low values of T_e compared to the coherence technique.

Pérez-Gussinyé *et al.* [2004] showed that when the models of coherence and admittance were formulated in the same way they recovered the same results, and both high and low T_e values were recovered. Pérez-Gussinyé *et al.* [2004] assessed theoretical admittance curves for various T_e values based on observed topography, within the same window, with the same tapering parameters, as was used for the recovery of the observed admittance. McKenzie [2003] had compared observed admittance curves to purely theoretical admittance curves based on infinitely large windows. Pérez-Gussinyé *et al.* [2004] showed that use of purely theoretical admit-

tance curves gave a systematic bias towards the recovery of lower T_e values unless windows of $\gg 2000$ km were analysed. If such a large region was considered the presence of both high and low T_e regions would mean the average T_e finally recovered for such a window would never be the maximum.

2.8.3 Spatially varying T_e based on spectral techniques

Spectral techniques can recover spatial variations in T_e by analysis of multiple overlapping windows of data. These techniques have been applied by authors such as Flück *et al.* [2003] and Pérez-Gussinyé *et al.* [2004]. However, the Fourier transforms utilised for calculating the spectra in many applications of these transfer function techniques, rely on windows of fixed size. This has been shown for example by Pérez-Gussinyé & Watts [2005] to introduce an intrinsic trade off between high spatial resolution, which would require small window sizes, and recovery of high T_e values which requires large window sizes. Pérez-Gussinyé & Watts [2005] solved this problem in-part by recovering the optimal window size which gave the most stable estimate of T_e , and then combining the results from different windows.

Kirby & Swain [2004] used wavelet transform techniques to estimate the gravity and topography spectra at a localised point, without consideration of specific window sizes. They showed that this allows a better resolution of the T_e structure. Daly *et al.* [2004] have applied the wavelet transform technique to the continental margin west of Ireland and recovered a T_e structure which is in good agreement with that recovered by standard Fourier methods.

2.9 Summary

There are several techniques for recovering T_e and T_e variations. In this study techniques based on models of flexure driven by known, or calculated, geological loads are applied. Along 1D profiles these allow consideration of the geological

process involved in the generation and compensation of observed features which is not available with the use of spectral methods.

For recovery of 2D maps of variation in T_e modelling methods are also applied. These consider a smaller window size (~ 200 km) than the spectral techniques ($> \sim 800$ km), allowing for theoretically better resolution of the T_e structure. For this technique of T_e recovery all flexure is assumed to be driven by surface loading. Where there is negligible topography but a significant proportion of buried loading synthetic tests show that T_e tends to be under estimated. Where there is significant topographic loading, and buried loading, such as in Taiwan (Lin & Watts [2002]) or the Appalachians (Karner & Watts [1983]), the results of the 2D technique applied in this study must be treated with caution as not all the loads present have been considered.

Oman: flexure and buried loading

3.1 Introduction

Oman lies on the north-eastern margin of the Arabian Plate and is the site of one of the best preserved ophiolite complexes in the world (Searle *et al.* [1983]). The ophiolite represents up to 14 km of oceanic crust and upper mantle (Shelton [1984], Searle *et al.* [1983]) that has been obducted onto the continental margin, and as such, it provides an excellent opportunity to study one of the classic occurrences of lithospheric flexure; flexure beneath an overriding thrust and fold load (e.g. Jordan [1981], Karner & Watts [1983], Snyder & Barazangi [1986]). Previous workers, discussed below, have recovered values of flexural rigidity for the foreland basin that developed by flexure in front of the North Oman Ophiolite, but they did not consider along strike variations in rigidity (Patton & O'Connor [1988], Shelton & Egan [1991], Ravaut *et al.* [1997]). Along-strike variation in flexural rigidity is potentially a key factor in the development of foreland basins as it controls their width and depth for a given load size (e.g. Stewart & Watts [1997], Lin & Watts [2002]).

Gravity data provides an important constraint on the extent of foreland flexure, and hence rigidity, because the sediment-filled foreland basin and deflected Moho give rise to distinctive gravity anomalies, which can be modelled. In this study a discussion of existing gravity anomaly data is given, then new data collected during

the 2003/2004 field season, in conjunction with U.A.E. University at Al Ayn, is presented. Application of modelling techniques to profiles across the foreland, associated with the Oman ophiolite, show that simple surface loading is not sufficient to explain the observed gravity anomalies. Models of flexure driven by sub-surface or buried loading are therefore introduced. The rigidity along a series of profiles is then recovered. Towards the south-east end of the Oman mountains the flexural signal observed in the gravity data is less distinct. This observation is discussed in relation to the observed geology and the obduction process.

The main tectonic elements of the northern Oman margin today are shown in Figure 3.1. This illustrates the location of the ophiolitic material within the mountains bounding the northern edge of the continent and other geological features of the area.

3.1.1 Geological history

The Arabian Plate finally amalgamated in the late Pre-Cambrian, as part of the Pan-African Orogeny. At this time a series of island arc terranes and pre-existing continental fragments up to 2000 Ma were sutured together, to form the Pangaeen super-continent (Husseini [1988], Gass *et al.* [1990]).

From the Pre-Cambrian to the Devonian the entire region underwent considerable extensional deformation, which led to the formation of extensive rift-type basins (Husseini [1988], Sharland *et al.* [2001]). Throughout the Late Devonian to Mid Permian period the Arabian plate was situated in a tectonically stable setting within the Gondwanan super-continent (Sharland *et al.* [2001]).

During the Carboniferous to Permian break-up of the Gondwanan super-continent a passive continental margin developed along the northern margin of Arabia, as the North West Iran, Alborz and Lut blocks rifted away and new oceanic crust was generated at the resulting mid ocean ridge (Stampfli [2000], Stampfli *et al.* [2001], Robertson & Searle [1990], Sharland *et al.* [2001]). In Oman little is known of

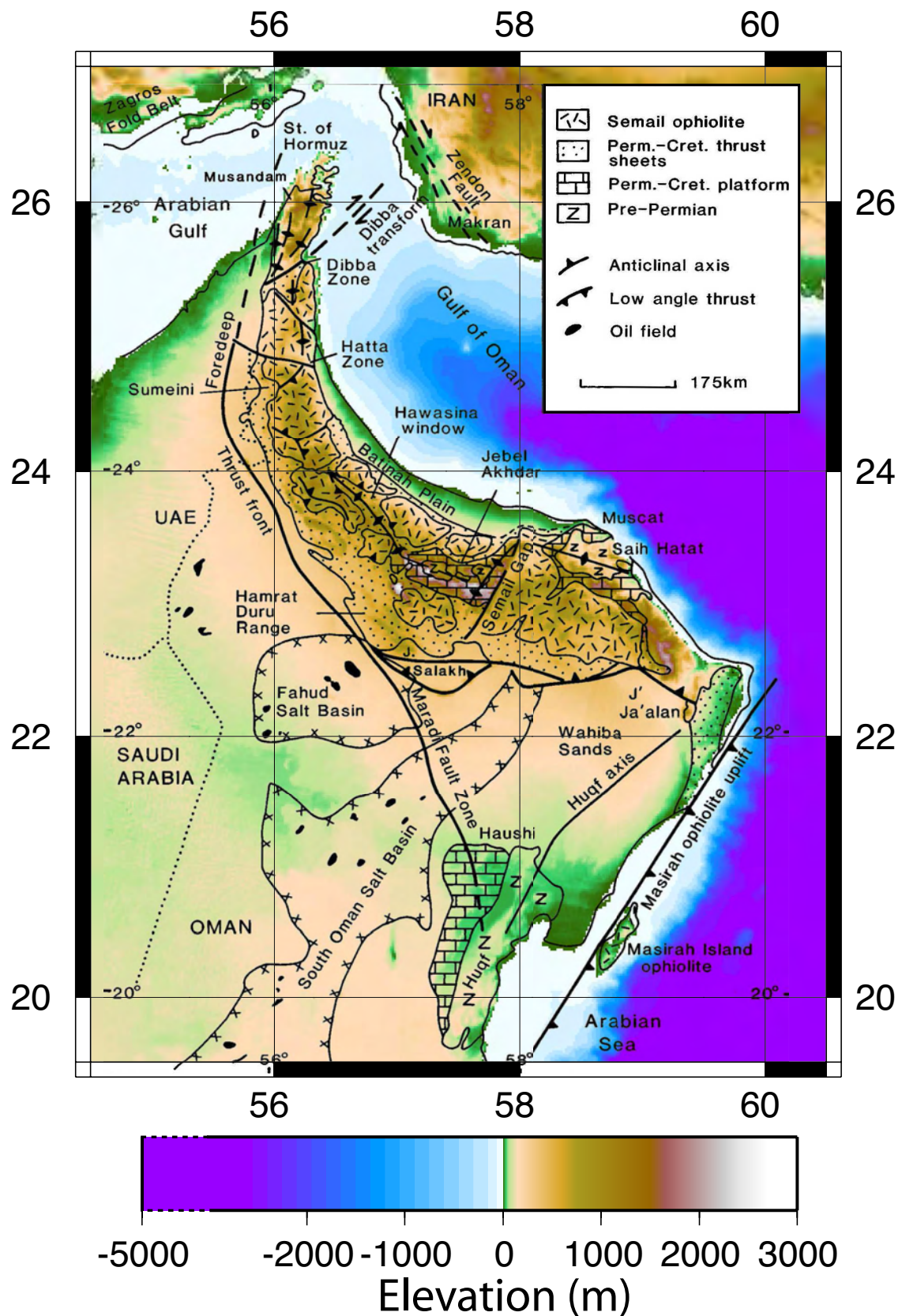


Figure 3.1: Outline geological map after Robertson & Searle [1990], showing major tectonic elements and topography and bathymetry from the GEBCO [2003] 1-minute global grid.

the syn-rift geology as much of it has been obscured by emplacement of the North Oman Ophiolite. However, studies of the highly metamorphosed terranes exposed in windows through the ophiolite by authors such as Warren *et al.* [2005] show a sequence of rocks which may represent the Permian syn-rift material in this area. These range from quartz rich schists, derived from more siliclastic material, mafic volcanically derived material, and more carbonate rich schists, which may be related to a carbonate passive margin sequence. The age of these rocks is not known but the presence of crinoid fossils shows that they are younger than the Cambrian. From the Late Permian to Late Cretaceous the passive-margin developed as a site of extensive carbonate deposition, as the margin subsided slowly due to post-rift thermal subsidence (Terkin [1999], Sharland *et al.* [2001], Stampfli *et al.* [1991]).

In Oman, this quiescent period was interrupted in the Cenomanian and Turoonian by uplift of the passive-margin and the development of a regionally extensive unconformity, the Wasi-Aruma break. This was followed by collapse of the margin and rapid subsidence (e.g. Glennie *et al.* [1973], Robertson & Searle [1990], Patton & O'Connor [1988]) due to the emplacement of the obducting Semail, or North Oman, ophiolite. The uplift, and subsequent subsidence, derive from the south-westward migration of the flexural bulge and foreland basin associated with the ophiolite (Robertson & Searle [1990], Patton & O'Connor [1988]).

It is generally accepted that the North Oman Ophiolite was emplaced onto a passive continental margin above an ocean-ward dipping subduction zone. Evidence for this comes from structural studies of high pressure assemblages beneath the ophiolite (Searle *et al.* [2004], Warren *et al.* [2005]). These show that very high pressure metamorphic material, formed by subduction of continental material, has been exhumed from the north west, indicating that the subduction zone originally dipped in this direction, away from the continental interior. In addition geochemical studies of the ophiolite itself suggest it developed in an oceanic back-arc setting (Alabaster *et al.* [1982]). Convergence was eventually halted when

buoyant continental material entered the subduction zone.

As the Northern Oman continental margin was overridden by the North Oman ophiolite, it flexed down beneath the propagating load, generating a flexural foreland basin which will be investigated in this study.

A cartoon of the Carboniferous to Recent evolution of the northern Oman continental margin is shown in Figure 3.2.

3.1.2 Previous estimates of the flexural rigidity of Oman

Previous studies of the flexural rigidity of the foreland basin have been based on both stratigraphic and gravity data. Patton & O'Connor [1988] recovered the shape of the flexural bulge associated with the initial emplacement of the North Oman Ophiolite onto the continental margin in the U.A.E. from the amount of material eroded from the regionally extensive and homogeneous Mauddud and Shilaif/Mishrif formations. They then fitted the shape of the recovered bulge with a broken plate flexural model. Values for flexural rigidity (D) of 3.87×10^{22} or 1.18×10^{22} N m were recovered based on an infill of water or sediment respectively. This equates to a T_e of $\sim 11 - 16$ km based on values of 10^{11} for Young's modulus and 0.25 for Poisson's ratio. Patton & O'Connor [1988] suggest that these relatively low values may relate to the rifting processes that created the passive-margin over which the ophiolite was later obducted. However, isopach data presented by these authors for the later flexural basin, associated with the final emplacement of the North Oman Ophiolite, shows a significantly broader basin. This suggests the T_e of the continent at the point of final loading was higher.

Shelton & Egan [1991] recovered a T_e value of around 10 km, based on an observed width of the final foreland basin of around 50 km and a depth of 4 km. These authors used a load thickness of 23 km, which was thought to be the maximum stratigraphic thickness of the ophiolite, and assumed the lithosphere to be represented by a continuous elastic plate.

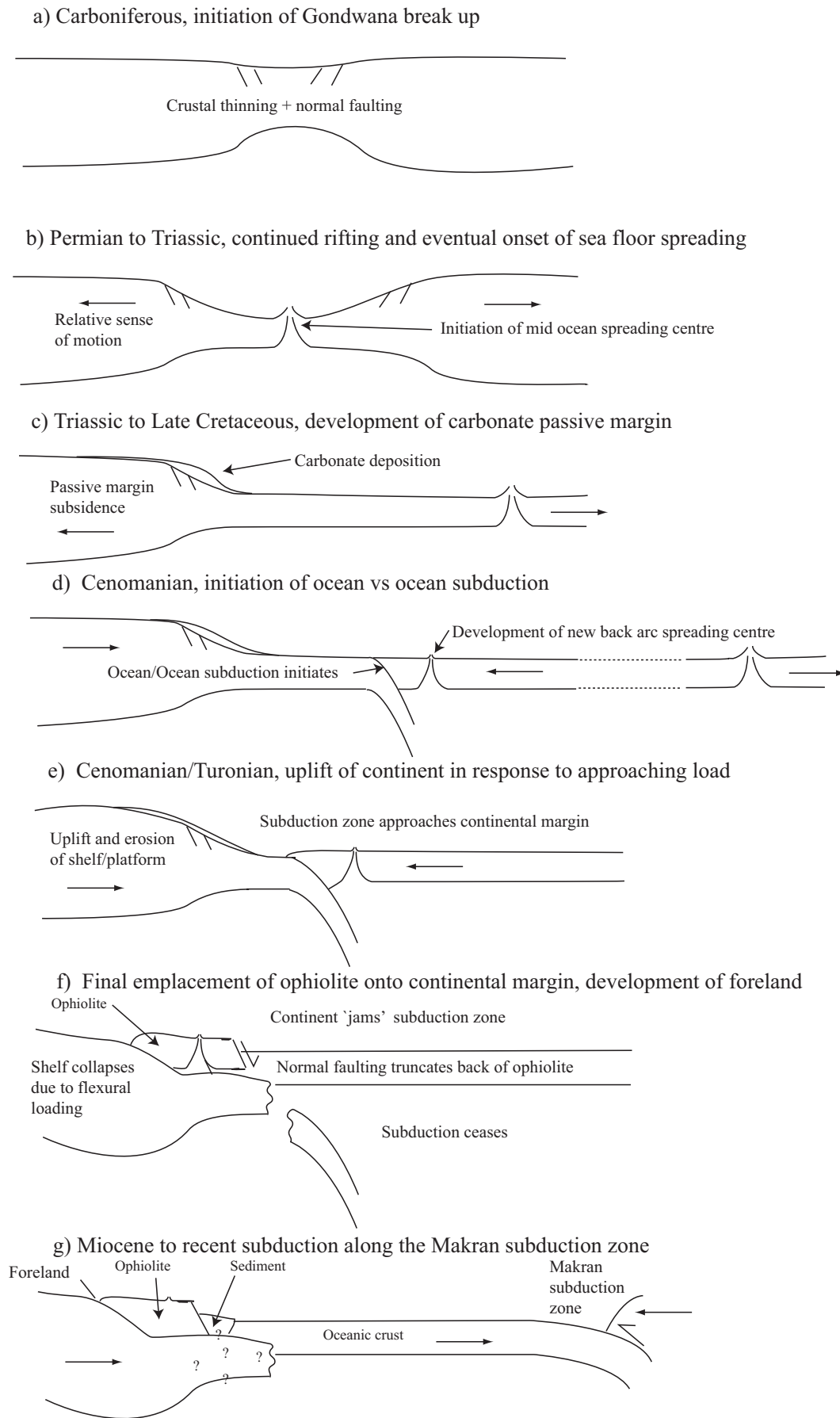


Figure 3.2: Geological cartoon of Carboniferous to Recent evolution of the Northern Oman continental margin after Terkin [1999], Sharland *et al.* [2001], Stampfli *et al.* [1991], Robertson & Searle [1990], Searle *et al.* [2004], Warren *et al.* [2005].

Ravaut *et al.* [1997] also assessed the flexural rigidity of the region based on a comparison of flexural surfaces, derived from a 2D finite element flexural model, and surfaces derived from isopach data which were thought to represent observed flexural surfaces. The flexural model was for a plate, broken along one edge, that allowed for a variable T_e structure, however, only regions of constant T_e were considered. The observed positive Bouguer gravity anomaly was used to constrain the ophiolite load size. The preferred value of D proposed by Ravaut *et al.* [1997] is 5.1×10^{23} N m for the foreland and 10^{22} N m beneath the load, equating to a T_e of ~ 38.5 km and ~ 10.4 km respectively. Although the study of Ravaut *et al.* [1997] considers a 2D model of flexure with two T_e values there is no consideration of variations along strike of the foreland.

The studies above are based predominantly on methods that compare the base of the foreland basin sequence, as defined by isopach data, to predictions of flexure based on simple elastic plate models. An alternative approach, which has been applied successfully in other parts of the world, is to analyse the observed gravity anomalies along profiles. This approach has typically been carried out based on forward models of the observed gravity anomalies, derived from models incorporating the effect of surface or buried loads bending an elastic plate (e.g. Karner & Watts [1983]). An advantage techniques based on gravity data present is the abundance of gravity data, relative to isopach and seismic data. This allows for an analysis of multiple profiles along strike of an orogen. The analysis of multiple profiles enables an assessment of the lateral variation of T_e and its impact on the regional geology.

3.2 Data

3.2.1 Existing gravity data

A large amount of both industry and academic gravity data exists for the Oman part of the ophiolite and foreland system. This has been collated by Ravaut & Warsi [1997a] and is available from the Ministry of Petroleum and Minerals, Sultanate of Oman in the form of a 5 mgal contoured, colour shaded Bouguer gravity anomaly paper map at a scale of 1:1,000,000. The Bouguer gravity anomaly data has been reduced using a density of 2670 kg m^{-3} and has been terrane corrected out to a radius of 167 km (i.e. Hayford zone 0).

This data-set was manually digitised from the northern tip of Oman to 21°S . This extended over 100 km beyond the southern edge of the pronounced negative Bouguer gravity anomaly associated with the foreland system (Ravaut *et al.* [1997]). This data was interpolated into a grid format using the **GMT** functions **blockmean** and **surface** with a tension factor of 0.25. A tension factor of 0.25 is that suggested by Smith & Wessel [1990] for gridding potential field data, to suppress the generation of spurious maxima and minima which may occur if a simple minimum curvature technique is used. The resulting map is shown in Figure 3.3.

3.2.2 Data collected as part of this project

The data-set compiled by Ravaut & Warsi [1997a] contains no data from outside Oman. It was therefore decided to carry out a field-based gravity survey in the neighbouring region of U.A.E. This survey was carried out in conjunction with a team from Al Ayn university, lead by Professor Ahmed Mahmoudi, which provided much of the logistical support that made the survey possible. Additional support was given by Dr Mohammed Ali from the Petroleum Institute, Abu Dhabi. The rationale for the survey design was to sample across the strike of the foreland basin, in the region flanking the ophiolite. The location of the gravity stations was, in

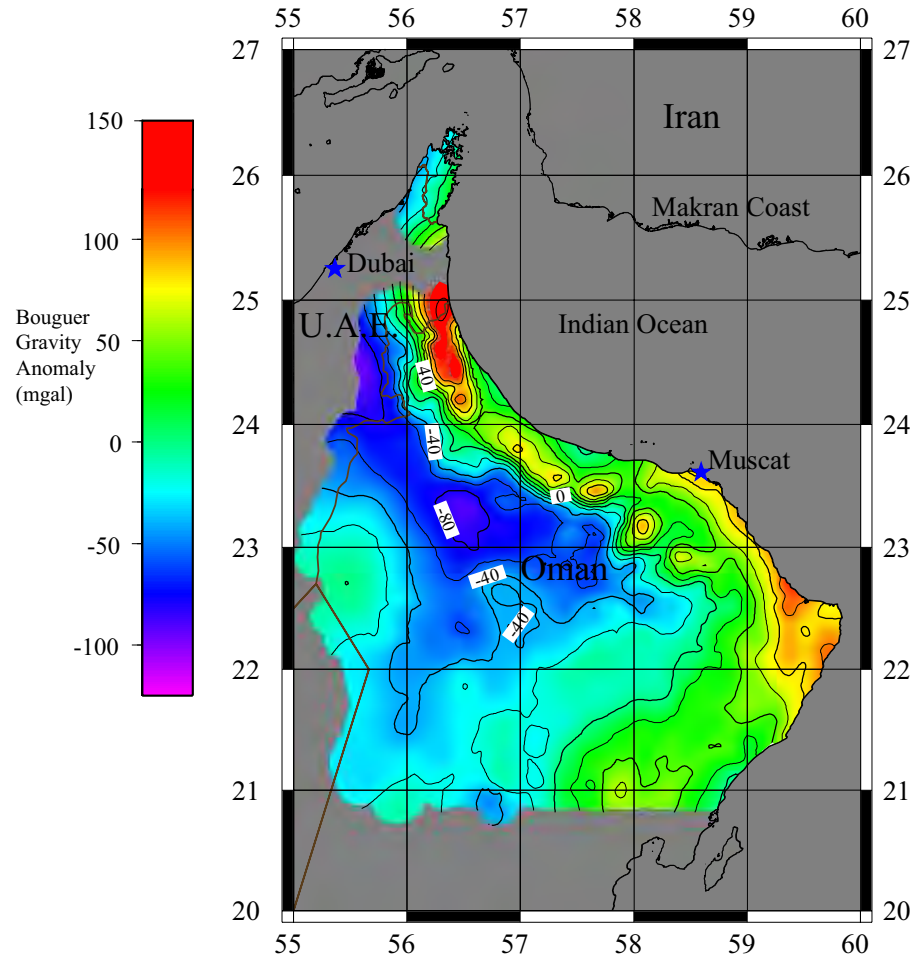


Figure 3.3: Bouguer gravity anomaly data from Ravaut & Warsi [1997a]. Contour interval of 20 mgal. The total data range is from -98 to 150 mgal.

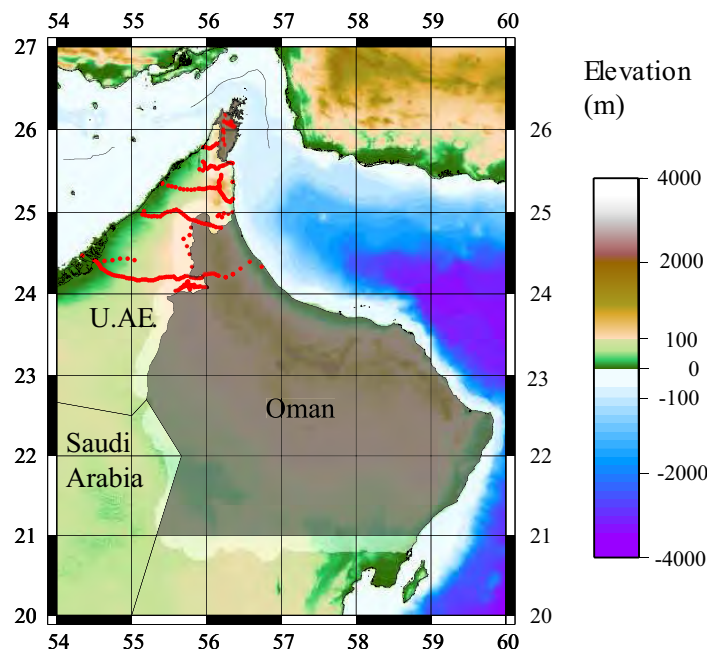


Figure 3.4: Location of gravity data collected during the 2003/2004 field season. Region shaded in grey marks the extent of the data from Ravaut & Warsi [1997a].

part, constrained by road access, although the road network provided several major east-west highways along which a survey could be conducted.

3.2.3 Data collection and processing

Gravity data was collected using a Lacoste and Romberg model G land gravity meter (model No 645) hired from Skiner Geophysics. Gravity values were tied to the IGSN71 global gravity network via an absolute base station close to Seib International (SI) airport Oman. See Appendix A for details of the base station.

During the survey gravity values were tied to the same point at the start and end of each day to constrain the meter drift. However, the drift rate used for final reduction of the data was taken from the occupations of the SI base station at the start and end of the survey. The drift rate recovered was ~ 0.0168 mgal per day. This instrument drift was removed from the data assuming linear drift at this rate.

Location and initial elevation data was collected using a Garmin GPS with

inbuilt barometric altimeter. The manufacturer stated horizontal accuracy was to ± 4 m and the stated vertical accuracy was ± 5 m which equates to an error of ± 1.5 mgal. The altimeter was calibrated by re-setting it to +2 m at the high tide line. When working away from the coast elevation was set to spot height values taken from high quality 1:25,000 maps of the area.

Comparison of the GPS/barometric elevations with data from the SRTM-90m (Rodriguez *et al.* [2005]) topographic data-set show there to be significant discrepancies between the two: this can be seen in Figure 3.5. The mean discrepancy is 1.72 m and the standard deviation of the discrepancy is 8.34 m. Some of these discrepancies may lie in the fact that the SRTM data-set loses accuracy in regions with steep topography (Rodriguez *et al.* [2005]). Differences may also stem from the fact that the SRTM data is a 90 by 90 m grid while the GPS/barometric elevations are point measurements. In areas of steep topography this can lead to large errors. Along the main foreland basin profile, highlighted in Figure 3.5, the discrepancy is less but still significant (>5 m). The error distribution appears to be non-random, with some parts of the profile consistently appearing higher than the SRTM data and others lower, suggesting that changes in barometric pressure through the day and errors in setting the meter pressure may be systematically biasing the GPS/barometric elevations. It was therefore decided for consistency to use the elevations from the SRTM-90 m data set for the correction of the gravity anomaly. The performance of the SRTM data-set for the Africa region (including Arabia) in comparison with over 300,000 kinematic GPS ground control positions gives a mean difference of 1.3 m, a standard deviation of 3.8 m and a 90 % absolute error of 6.0 m (Rodriguez *et al.* [2005]). This would give a theoretical mean error of 0.39 mgal and a 90 % absolute error due to elevation of 1.8 mgal. In regions of steep topography this error could be significantly higher as the SRTM value only represents the average elevation within a given 90 by 90 meter area.

Observation points were located, where possible, with no significant ($< \sim \pm 2$ m)

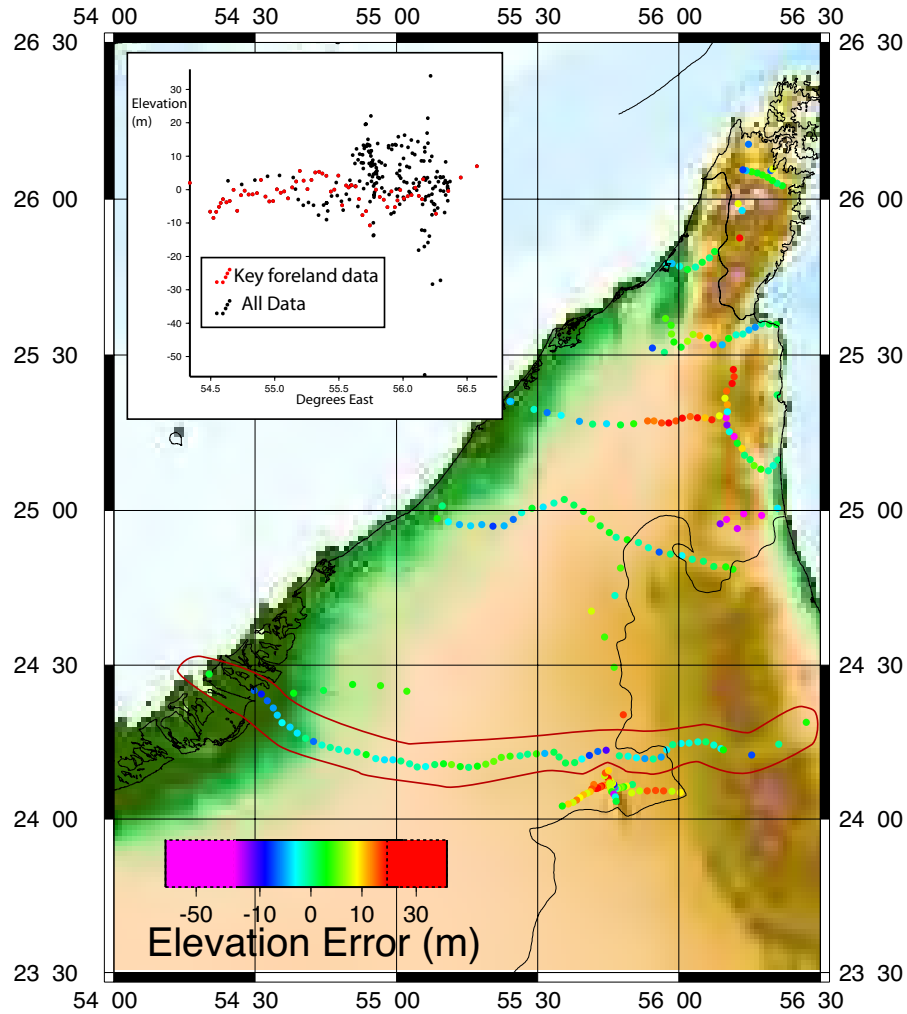


Figure 3.5: Elevation error map. Inset shows vertical difference between recorded GPS/Barometric values and SRTM-90 data. Dots on map show locations of data points, points within red line form the main foreland profile.

topographic relief within 100 m to reduce the effect of near field terrane corrections discussed below. This was easily accomplished in the foreland region, but in the mountains was not always achievable.

Data reduction

Data was processed using the standard formulas outlined below. The aim of the corrections applied to gravity data is to remove all non-geological factors from the observed signal. The observed gravity signal (g_{obs}) is corrected for: the change in shape of the earth from pole to equator; the centripetal force due to the earth's rotation; elevation above or below the geoid (mean sea level); the rock mass between the observation point and the geoid; and the gravity effect of the surrounding terrain.

Firstly, the theoretical gravity anomaly at each point was calculated based on Clairaut's formula. This compensates for the flattening of the Earth and the centripetal forces associated with its rotation.

Clairaut's formula, Equation (3.1), relates gravity at sea level to latitude on a rotating reference ellipsoid.

$$g_{\Phi} = g_0(1 + k_1 \sin^2 \Phi - k_2 \sin^2 2\Phi) \quad (3.1)$$

Where g_{Φ} is the gravity at latitude Φ , g_0 is gravity at the equator (978031.85) and k_1 and k_2 are constants which depend on the ellipsoid considered. The values used for k_1 and k_2 were 0.005278895 and 0.0000059 respectively.

The next correction is the free air correction (FAC). This corrects for the height of the station above or below the geoid. Together with the latitude correction this produces the Free Air Anomaly (FAA), where h is the height of the observation point above the geoid.

$$FAA = g_{obs} - g_{\Phi} - 0.0386h \quad (3.2)$$

The Bouguer correction accounts for the mass of material between the observed point and the geoid. In the case of the simple Bouguer correction, the correction

is based on the assumption of the presence of an infinite slab of material with a specific density beneath the observation point. The equation for the simple Bouguer correction is:

$$B_c = 2\pi G\rho h \quad (3.3)$$

Where ρ is the density of the material between the observation point and the geoid, and G is the Gravitational constant. In this correction a standard value for the correcting density (ρ) of 2670 kg m^{-3} was used, to be consistent with the data of Ravaut & Warsi [1997a].

The final correction is for the terrain around the observation point. This accounts for the gravity effect of the terrain, above and below the observation point, which is not included in the simple Bouguer slab correction. Details of how the terrain correction was applied are outlined in Appendix A. The terrain correction was carried out using the SRTM-90 data-set for the reference topography and considered all topography within a 1-degree box centred on the observation point. The topography and its terrane effect on an example point shown in Figure 3.6a and b. The total correction applied to all data points is shown in Figure 3.6 c and ranged from 0.024 to 16.96 mgal. Higher terrane correction values are associated with regions of steep topography, however, along the main foreland profile the corrections were always less than 1 mgal, and often less than 0.1 mgal.

3.2.4 U.A.E. profiles

Figure 3.7 shows a compilation of all data collected across the U.A.E. part of the foreland of the North Oman Ophiolite. The distinct positive anomaly associated with the exposed ophiolite can be seen in all cases. To the west of this positive anomaly there appears to be a significant negative Bouguer gravity anomaly. This pattern of gravity anomalies is suggestive of a flexural signal, as proposed by, for example, Karner & Watts [1983] and Stewart & Watts [1997]. However, the shape of the negative gravity anomaly associated with the foreland basin can only be

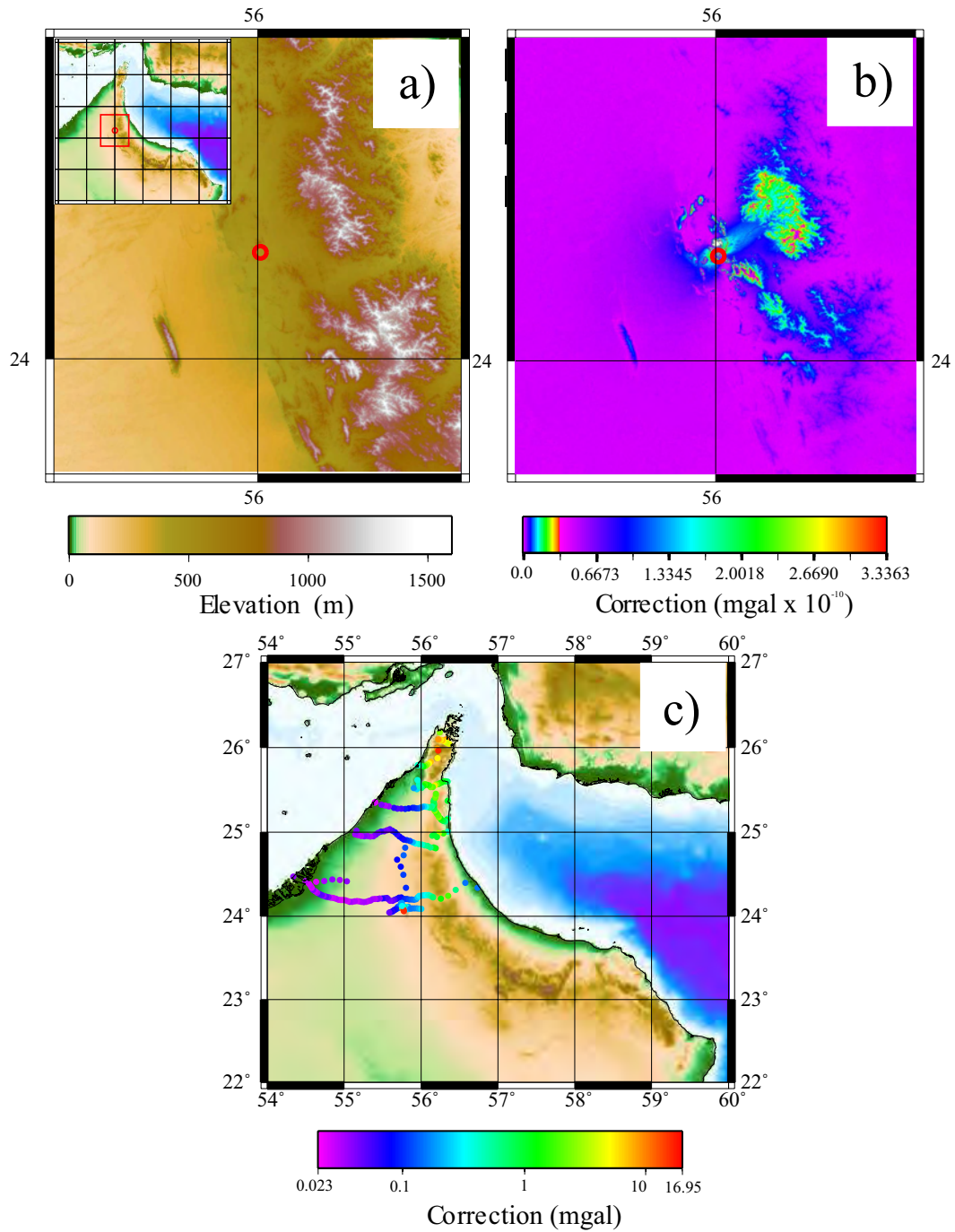


Figure 3.6: Terrane corrections. a) Topography around example point considered in the terrane correction, inset shows regional setting. b) Gravity effect of each topographic column around example point. c) Total terrane corrections applied at each observation point.

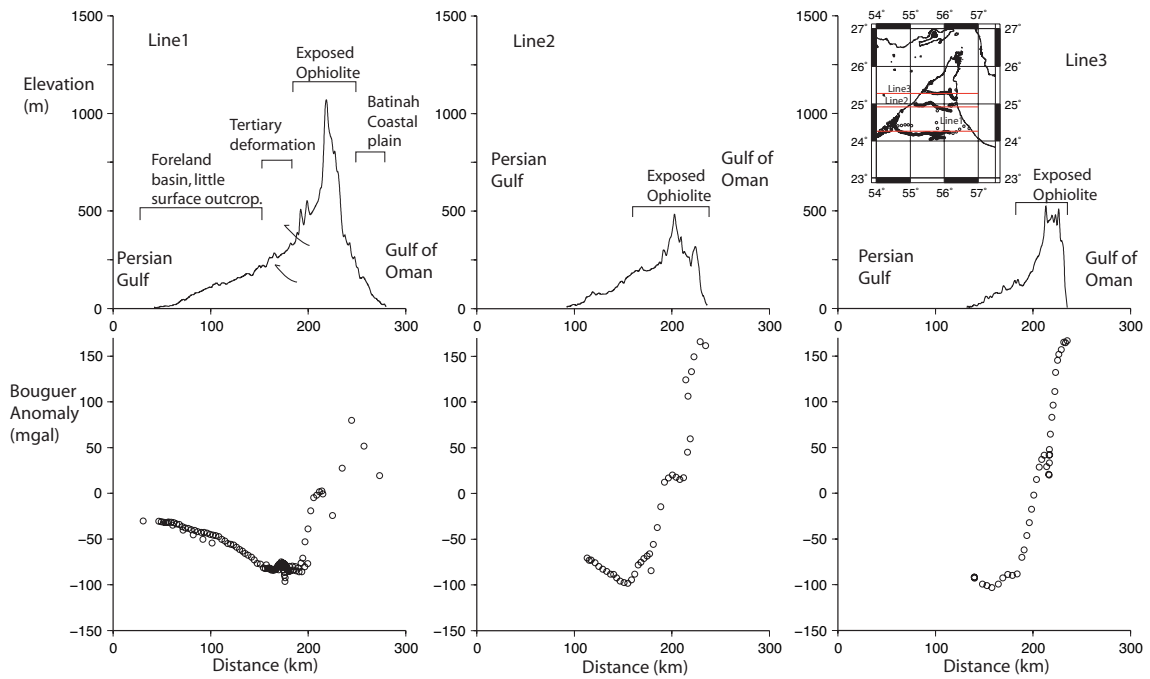


Figure 3.7: All data collected across the U.A.E part of the foreland of the North Oman Ophiolite projected onto 3 E-W profiles.

clearly defined in the Southern-most profile. This restricts any interpretation of foreland rigidity along the northern-most profiles.

3.2.5 Regional corrections

The obduction of the North Oman Ophiolite onto the continental margin of Arabia causes flexure which should be manifest in the gravity field. Figure 3.8 shows that the Oman ophiolite is generally associated with Bouguer high up to ~ 150 mgal, which is flanked to the west by a low of down to ~ -90 mgal. This high-low ‘couple’ is typical of thrust loads and their associated foreland basins.

There may exist, however, significant signals in the Bouguer gravity data that are not related to the process of orogenic loading. These may be both long or short wavelength.

To separate the shorter wavelength effects of crustal scale processes associated with the obduction of the ophiolite, from the effect of density variations deep in

the mantle, a long wavelength gravity anomaly was subtracted from the data-set. The long wavelength field was derived from perturbations of satellite orbits described by the model OSU91a of Rapp & Pavlis [1990]. The satellite derived field, complete to degree and order 40, was removed from the observed data, which equates to removing wavelengths $>\sim 1000$ km. These wavelengths were considered as they reduced the Bouguer gravity anomaly to approximately ~ 0 mgal along the U.A.E. coast. A Bouguer anomaly of zero would be expected in this region, in the absence of geological processes other than the obduction of the ophiolite, as it is known to lie at the edge of the flexural foreland from isopach data e.g. Patton & O'Connor [1988] and Ravaut *et al.* [1997]. The long wavelength correction is shown in Figure 3.8b. On the profile across the U.A.E. the field, complete to degree and order 40, of OSU91a was up to 3 mgal different from the more modern EGM-96 field (Lemoine *et al.* [1998]) at the same degree and order. However, the different long-wavelength anomalies make no significant change to the shape of the corrected foreland anomaly.

Figure 3.8a shows that there are significant positive Bouguer anomalies along the coast close to Muscat, and to the east, that do not appear to correlate with significant exposures of ophiolitic material. These high Bouguer values could be misinterpreted as being due to a large ophiolite body with no surface exposure. However, the continental shelf is very narrow in this region and the positive anomaly is therefore largely accounted for by the shallowing of the mantle across the ocean-continent transition. The anomaly due to the shallowing of the mantle can be calculated assuming the known water depth is balanced by an ‘anti root’ of high density mantle material, in the same way that onshore surface topography is balanced by a root of low density crustal material in the Airy model. Only the effect of the offshore ‘anti root’ was calculated as onshore the Moho position and associated anomaly is related to the obduction process. Water depth offshore from Oman was taken from the GEBCO [2003] 1-minute grid and a continental crustal density of

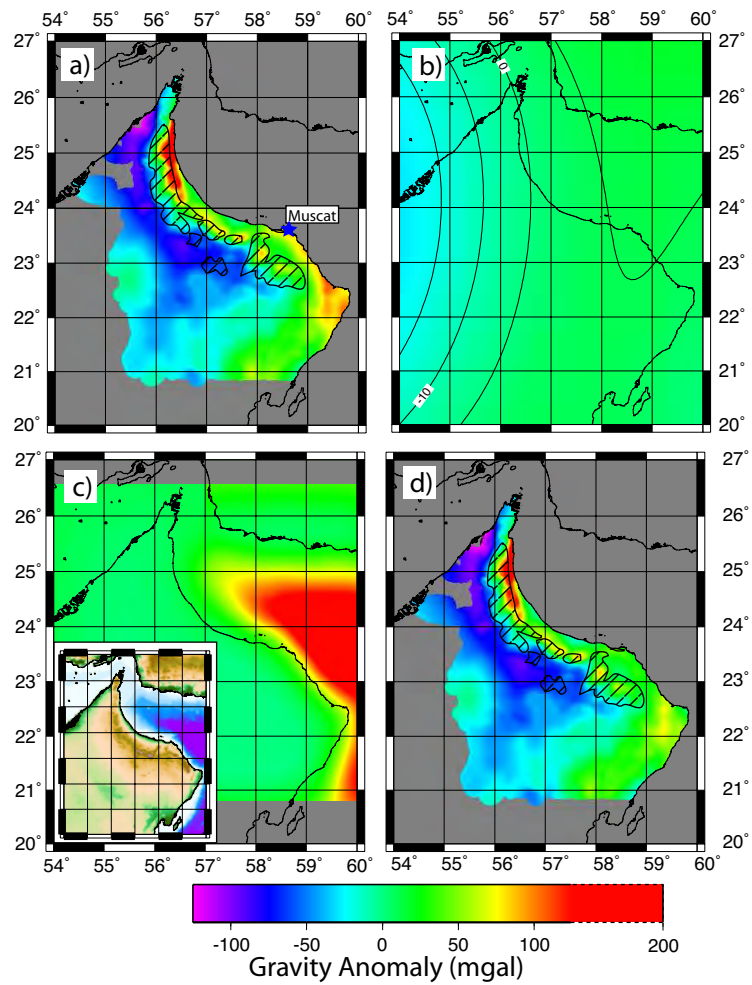


Figure 3.8: Regional corrections applied to Bouguer gravity data in the Oman and U.A.E. region. a) Uncorrected Bouguer gravity anomaly. Hashed areas are exposed ophiolite. b) Degree and order 40 satellite derived gravity anomaly, contoured at 10 mgal intervals. c) Correction for shallowing mantle due to isostatic compensation of the bathymetry across the continental margin. Inset map shows regional topography and bathymetry, only the bathymetry was considered for this correction. d) Bouguer anomaly corrected for shallowing Moho offshore and Degree and order 40 satellite derived gravity field.

2800 kg m^{-3} was assumed.

3.3 Modelling on profiles

The technique for recovering T_e based on modelling the flexure along profiles driven by surface loading, as proposed by previous authors such as Karner & Watts [1983] and Stewart & Watts [1997] discussed in Section 2.2, was initially considered. The technique was first applied to the Bouguer gravity anomaly data collected along the southern most profile across the foreland in the U.A.E. during 2003/2004 shown in Figure 3.9. The model of flexure driven by surface loading proved to be incomplete as significant buried loads are present in this region. Models based upon buried loading such as those used by Lin & Watts [2002] were therefore applied to this profile and provided a good fit to the observed data in this region. The technique for recovering foreland T_e based on buried loading was then applied to a series of profiles along strike of the ophiolite to the south-west.

3.3.1 Surface loading only

The methods of modelling flexure and gravity anomalies derived from only surface loading (proposed by Karner & Watts [1983] and Stewart & Watts [1997]) were initially applied, using a typical surface load (ρ_l) and infill (ρ_i) density of 2650 kg m^{-3} . A series of flexural surfaces were first calculated, assuming the application of the surface load to the end of a series of broken plates with a range of constant T_e values away from the plate break. The gravitational effect for each of these flexural surfaces was then calculated assuming two interfaces: one sediment/crust ($2650 \text{ kg m}^{-3}/2800 \text{ kg m}^{-3}$) and one crust/mantle ($2800 \text{ kg m}^{-3}/3330 \text{ kg m}^{-3}$). The results of this model are shown in Figure 3.10.

Lyon-Caen & Molnar [1985] showed that varying load and infill densities had little effect on the recovered T_e . To assess this, flexural profiles for a range of load and infill densities about the typical values assumed by Karner & Watts [1983] and Stewart & Watts [1997] were produced. These flexural profiles and the gravity anomalies derived from them are shown in Figure 3.11.

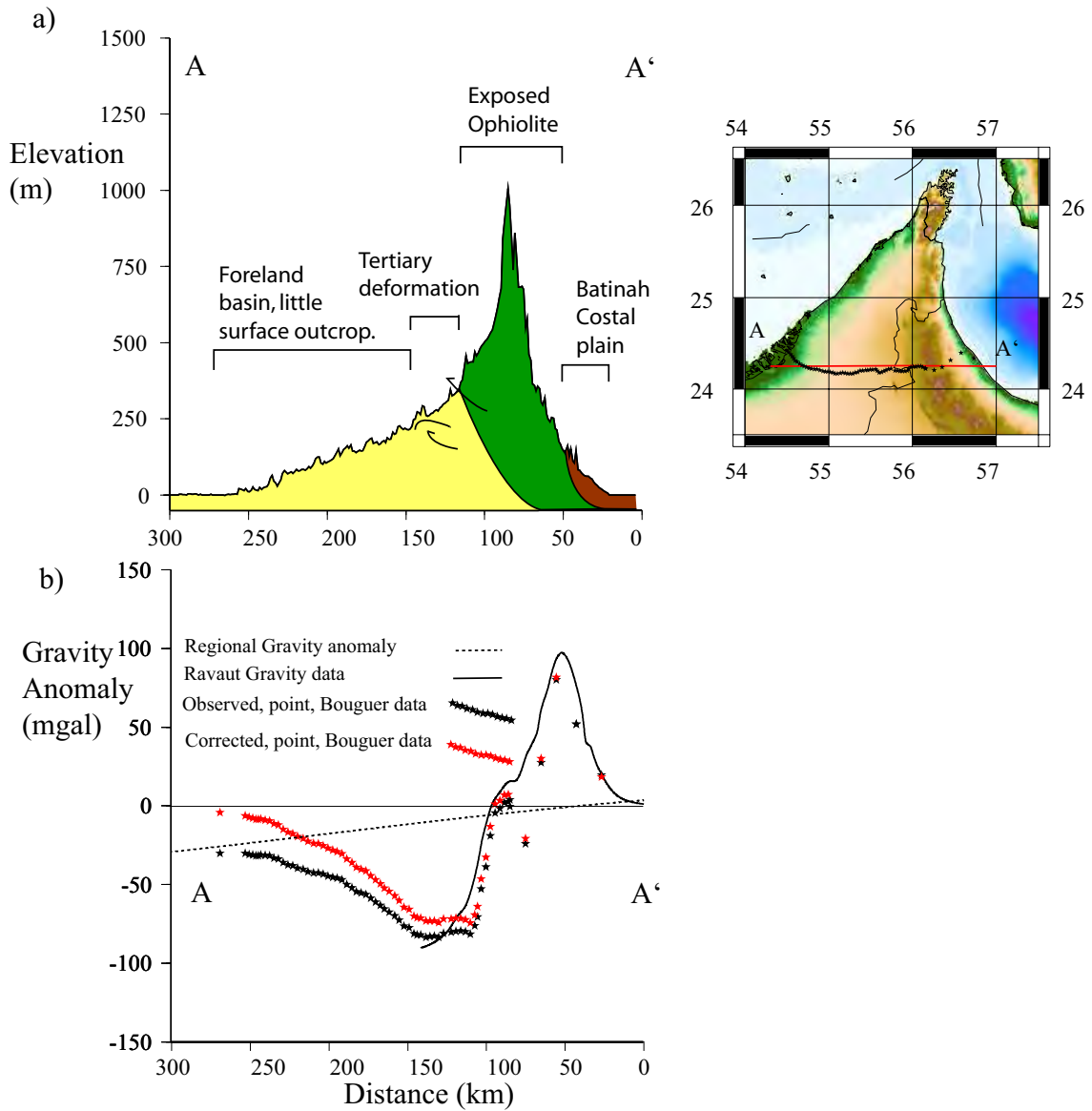


Figure 3.9: Bouguer gravity anomaly and topographic data, projected onto the line A-A' (profile 1). a) Surface topography along modelled profile. The approximate geological structure across the region is also shown. Inset map shows the location of the data points as stars and the E/W line that the data was projected onto in red. b) Bouguer gravity anomaly data from Ravaut as a black line. Bouguer gravity anomaly data from the 2003 field season is shown as black stars. Bouguer gravity data corrected for the regional gravity field, shown as red stars. Dashed line shows sum of degree and order 40 satellite derived gravity anomaly and anomaly due to shallowing of mantle offshore.

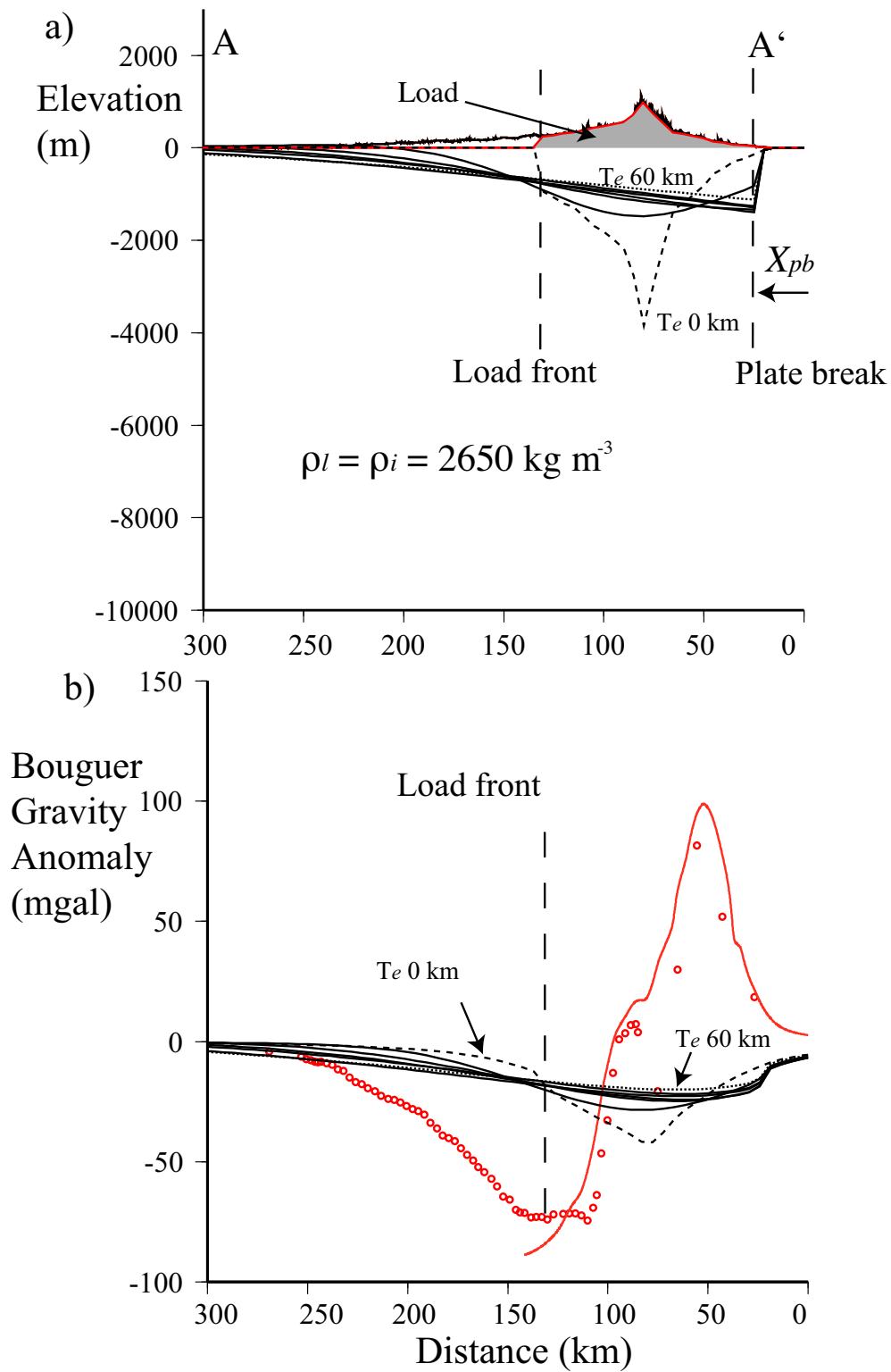


Figure 3.10: Results of modelling flexure based on loading solely due to observed surface topography. ρ_l and $\rho_i = 2650 \text{ kg m}^{-3}$. a) surface load (material above 0 m elevation) shaded grey, black lines show modelled flexure based on T_e ranging from 0 to 60 km. Vertical dashed line shows plate break and load front defined by the most negative value of the observed Bouguer anomaly. b) Calculated and observed Bouguer anomalies. Thin lines are models and red circles and thick red line are Bouguer data corrected for the long wavelength satellite derived anomaly.

Figure 3.11 shows the flexure and gravity anomaly differs little with varying load density by $\pm 100 \text{ kg m}^{-3}$. The most significant discrepancies occur beneath the load rather than in the foreland region where the flexural profile is assessed. The variations in flexure due to varying infill density are larger than those seen when varying load density. However, the calculated gravity anomalies show a much more subdued pattern. In-fact the model with denser infill and a greater magnitude of flexure is associated with a less negative calculated gravity anomaly. This is because the effect of the increase in the size of the flexural basin and additional deflection of the Moho is offset by the decrease in the density contrast between foreland fill and crust. These results confirm the suggestion of Lyon-Caen & Molnar [1985] that variations in either load or infill density, about typical values, will have little impact on the recovered T_e .

Figure 3.10 clearly shows that surface loading, as formulated, cannot fully account for the magnitude of the observed Bouguer gravity anomaly. In addition Figure 3.11 shows that limited variation of load and infill densities will have little impact on the recovered T_e value. However, if the surface load comprises ophiolitic material, a significantly higher driving load density (e.g. 3100 kg m^{-3}) is implied. The assumption that much of the surface topography within the load region is ophiolitic is reasonable given the mapped outcrop of the ophiolite shown in Figure 3.1. Figure 3.12 shows the results of modelling the flexure, assuming the load to be ophiolite with a density of 3100 kg m^{-3} and the infill to be sediment with a density of 2650 kg m^{-3} . The flexure is slightly greater than for the lower density load, however, the magnitude of the observed gravity anomaly is still significantly underestimated.

The fact that an ophiolite load, where the load size is derived from the surface topography, with an infilling load of sediment, does not generate enough flexure to match observed gravity anomaly should be expected. The ophiolite is considered to be over 10 km thick in places (Robertson & Searle [1990], Shelton [1990]) and

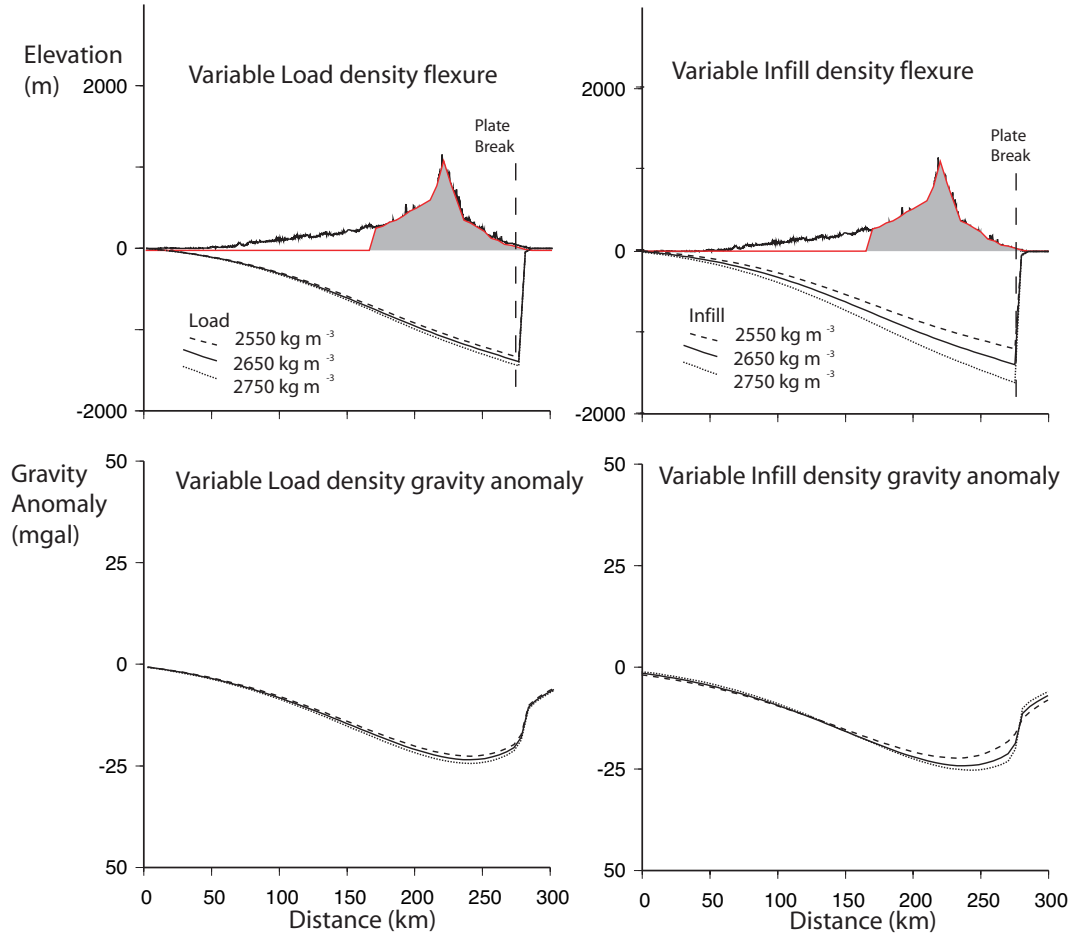


Figure 3.11: Effect of varying load and infill densities between 2550 kg m⁻³ and 2750 kg m⁻³ given a constant T_e of 30 km and other parameters as defined in Chapter 2. The applied load is the region shaded grey. Gravity anomalies based on two interfaces: one at the crust/mantle boundary and one at the foreland basin/crust boundary.

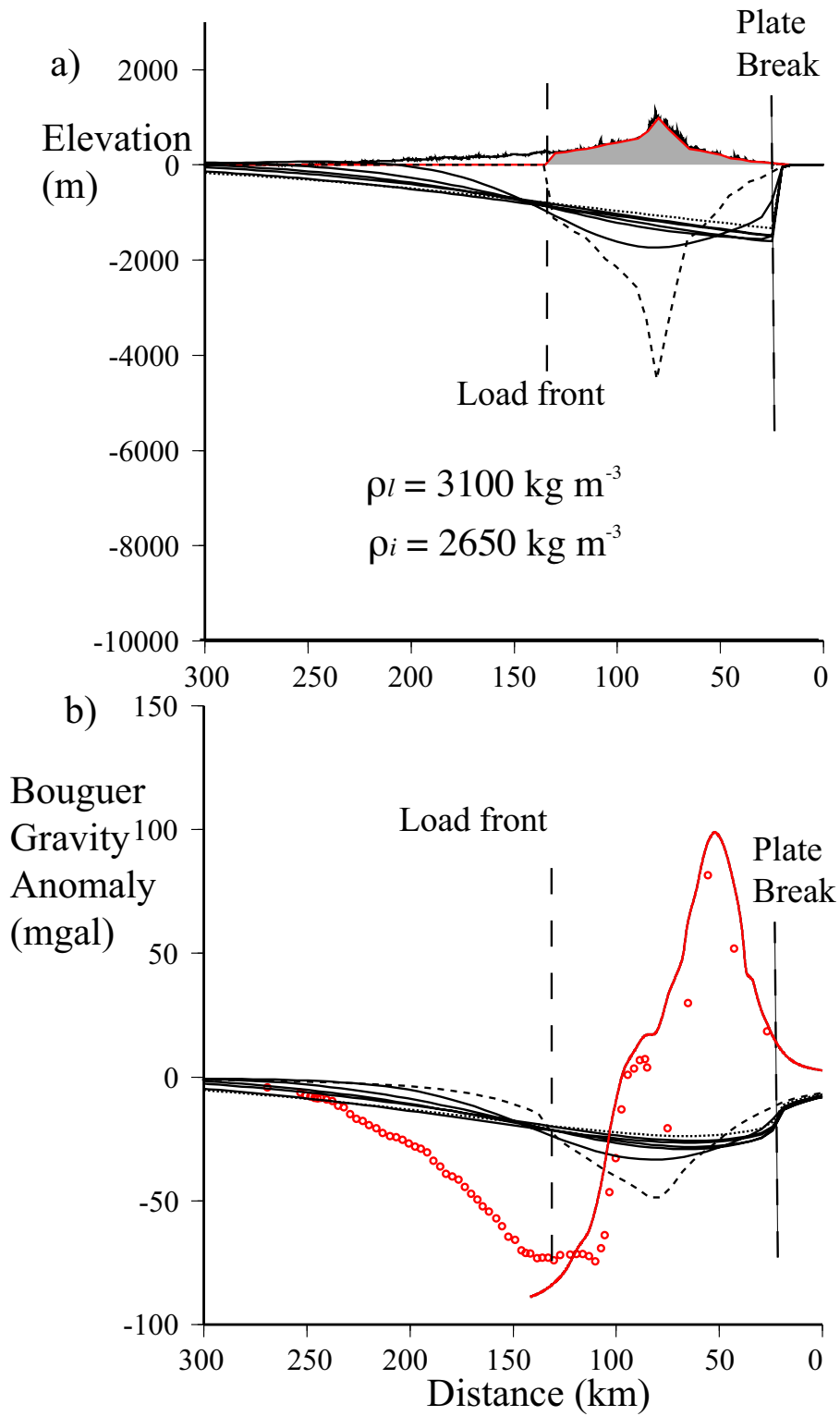


Figure 3.12: Results of modelling flexure derived from the observed surface topography made up of ophiolite with a density of 3100 kg m^{-3} , displacing air, as the driving load, and sediment acting as the infilling load. a) Modelled flexure for T_e values of 0 (long dashed line) to 60 km (fine dashed line). b) Calculated anomalies (thin black lines) and observed Bouguer anomalies (red circles and thick red line). Long dashed line $T_e = 0 \text{ km}$ and fine dashed line $T_e = 60 \text{ km}$.

given that there is only 1 or 2 km of exposed topography the load size is clearly under estimated. The model was therefore re-run assuming both the surface and infill loads to have a density of 3100 kg m^3 . This takes into account the fact that a substantial amount of ophiolitic material lies beneath the surface load. The results of this model are shown in Figure 3.13.

When both load and infill are assumed to be ophiolite the extent of flexure is much greater, as seen in Figure 3.13a. However, when an infill of ophiolite was considered for the calculation of the gravity anomaly, almost no flexurally generated anomaly was seen (Figure 3.13b). The small modelled anomaly arises from the presence of ophiolitic material filling the entire flexural depression. The density contrast in the foreland region is therefore positive, instead of negative as it is when the foreland is considered to be filled with sediment. The modelled positive anomaly in the foreland counteracts the additional negative anomaly derived from the flexed Moho, the sum of these two effects gives rise to the small total model anomaly.

This set of models highlights the fact that a model of flexure driven by surface loading with a constant infill density cannot account for both the magnitude of the flexure and the observed gravity anomaly. A key issue is that the infill density below the driving load may differ from that beyond the edge of the driving load, within the foreland.

3.3.2 Buried loading

To solve the problem of spatially varying infill densities we turn to techniques for assessing the flexure due to buried loading. Buried loads are mass anomalies, within the crust, that have little or no surface expression. Such loads have been recognised in many, but not all of the world's mountain belts and compressive margins (e.g. Karner & Watts [1983], Stewart & Watts [1997], Lin & Watts [2002]).

Recovering the buried load size allows the consideration of the entire ophiolite as

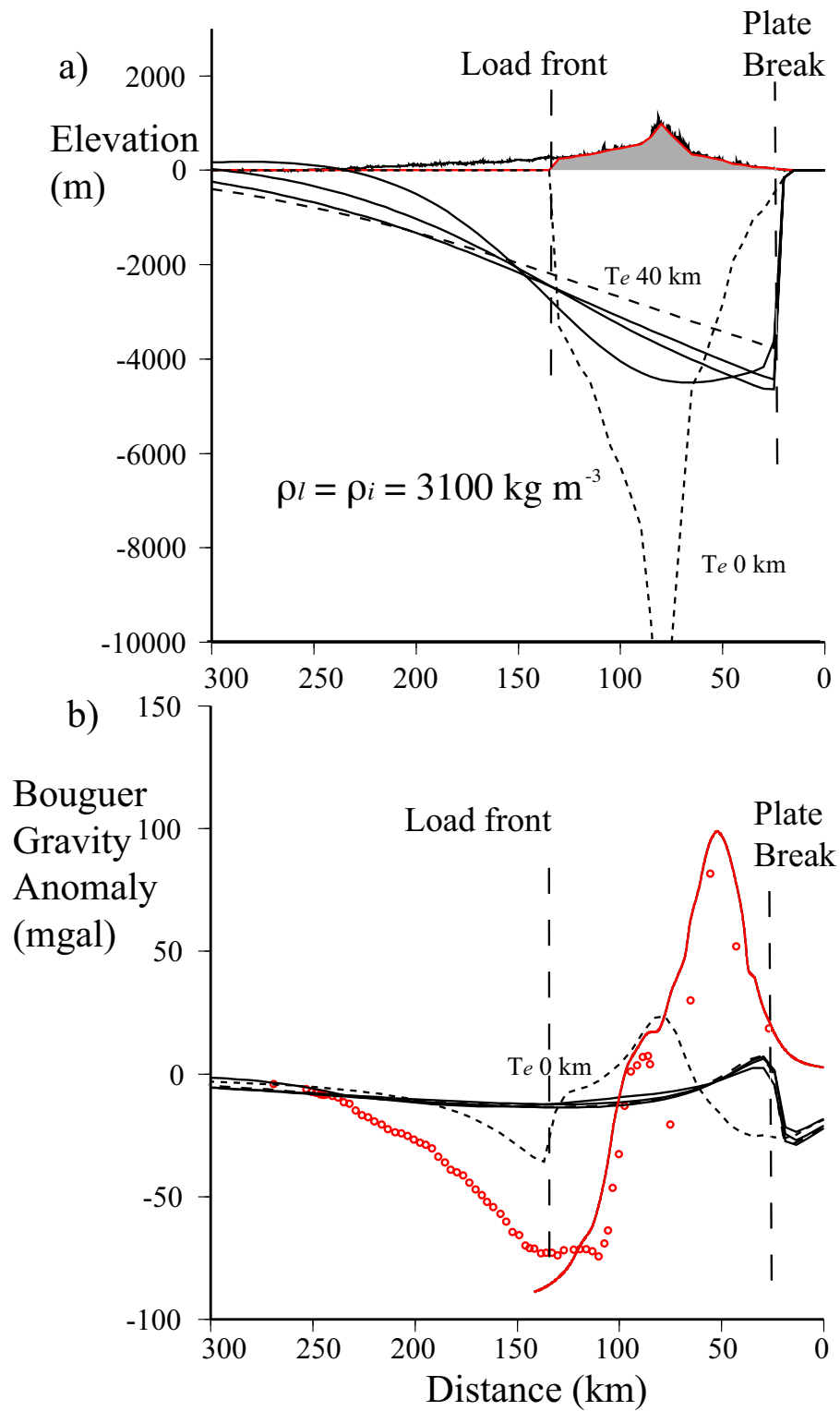


Figure 3.13: The results of modelling flexure due to a driving surface load of ophiolite and an infilling load of ophiolite for T_e values of 0 to 40 km. a) Modelled flexure. b) Modelled and observed Bouguer gravity anomalies, lines as used in Figure 3.12.

a single driving load. Infill densities for the flexural models can then be determined based on a process orientated model of the evolution of the ophiolite obduction. This effectively separates the flexure due to the ophiolite from the material infilling the foreland. The magnitude of the flexure, due to the large ophiolite, and the gravity anomaly due to the low density foreland material can therefore both be calculated, giving a more accurate model of the foreland region, and allowing the recovery of the best fitting T_e value.

One way to assess the size and contribution of such buried loads is to compare the gravity anomaly assuming Airy isostatic compensation of the surface topography with the observed Bouguer gravity anomaly. Positive parts of the Bouguer gravity anomaly typically equate to buried, high density, loads within the crust. Simply assuming all the positive part of the Bouguer anomaly is related to buried loads would underestimate the load size as downward deflection of the Moho would create a negative component to the observed anomaly. Assuming the positive residual, when the observed anomaly is compared with an Airy isostatic model, equates with the anomaly due to the buried load, is a better approximation to the gravity effect derived from the buried load, as part of the Moho deflection has been taken into consideration. When the gravity contribution of the buried load has been isolated the load size can be calculated assuming a specific density contrast between the buried load and the surrounding material.

A set of comparisons, from orogenic belts across the globe, between the model gravity anomaly derived from Airy compensation of local topography and observed Bouguer anomaly is shown in Figure 3.14. In cases where flexure is driven predominantly by surface loading, such as the Bolivian Andes or the Eastern Alps (Stewart & Watts [1997]), the main discrepancies between the observed and modelled gravity anomaly lies on the foreland basin side of the thrust front. In cases with little or no buried loading the highest topography is coincident with both the most negative observed Bouguer anomaly and the most negative modelled Airy anomaly.

Where significant buried loads are present, such as in the Appalachians, Western Alps and Taiwan (Karner & Watts [1983], Stewart & Watts [1997], Lin & Watts [2002]), the main departure from the Airy model is often beneath the highest topography. The departure comprises a high and a flanking low, this gravity anomaly pattern has been interpreted as a buried load and its associated flexure.

Figure 3.14 clearly shows that northern Oman, which has been plotted at a similar scale as the other orogenic belts, is associated with significant buried loads. The modelling applied in this region therefore follows the approach of Karner & Watts [1983] for the Appalachians and Lin & Watts [2002] for Taiwan.

First the Airy isostatic compensation of surface topography was calculated assuming a surface load of 2650 kg m^{-3} , a root (crust) with density of 2800 kg m^{-3} and a mantle density of 3330 kg m^{-3} .

Second, the Airy gravity anomaly was calculated assuming 2 interfaces for load/crust and crust/mantle, with the density values as above. The positive discrepancy between the observed and modelled anomaly was assumed to reflect the buried load, while the negative discrepancy related to flexure caused by the buried loading.

Lastly, Green's equivalent layer theorem was used to calculate the buried load thickness (t) assuming the buried load was an ophiolite body, with a density of ρ_{oph} , surrounded by sediment with a density of 2650 kg m^{-3} . The equation for this is:

$$t = \frac{\delta g}{2\pi G \Delta \rho_b} \quad (3.4)$$

Where δg is the positive residual gravity anomaly (observed Bouguer anomaly - model Airy gravity anomaly), G is the universal gravitation constant and $\Delta \rho_b$ is the density contrast between the buried body and the surrounding material. An example of the results of this process is shown in Figure 3.15.

The final step in the recovery of the buried load size is similar to the technique

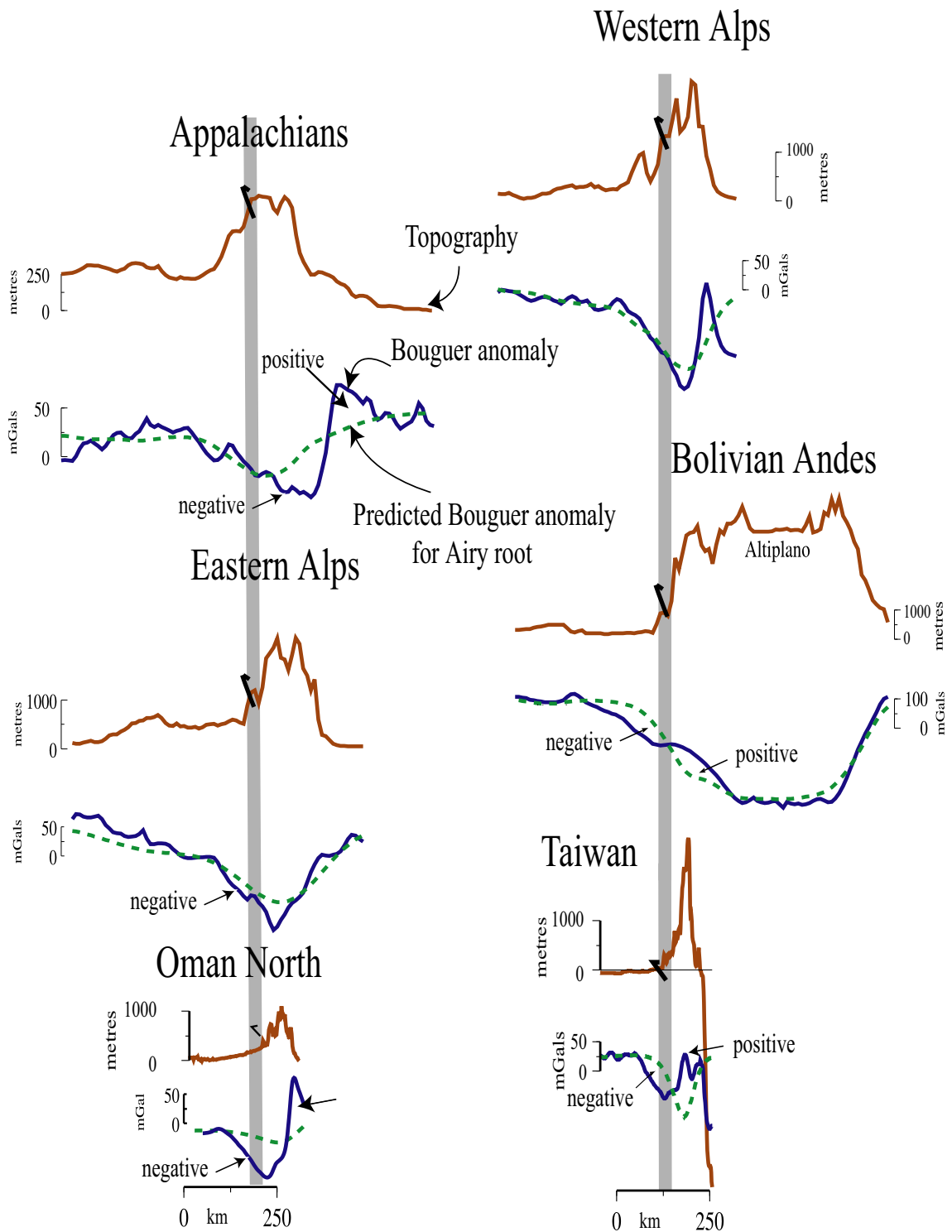


Figure 3.14: Comparison of the Bouguer gravity anomaly to the calculated gravity effect of compensation of surface topography based on an Airy model. After Stewart & Watts [1997] with the addition of data for Taiwan from Lin & Watts [2002] and models for Oman from this study. Grey line marks thrust front (thrusting propagating from right to left).

used by Ravaut *et al.* [1997], although in that study the anomaly due to the ophiolite was recovered by first removing the effects of low density sediment, followed by subtraction of a low order polynomial function representing the regional field.

To calculate the model flexure and associated gravity anomaly the process associated with the obduction of the ophiolite was considered. A simplified sketch of this process is shown in Figure 3.16. This allowed consideration of the contribution of each step to the final flexure and hence to the observed gravity anomaly.

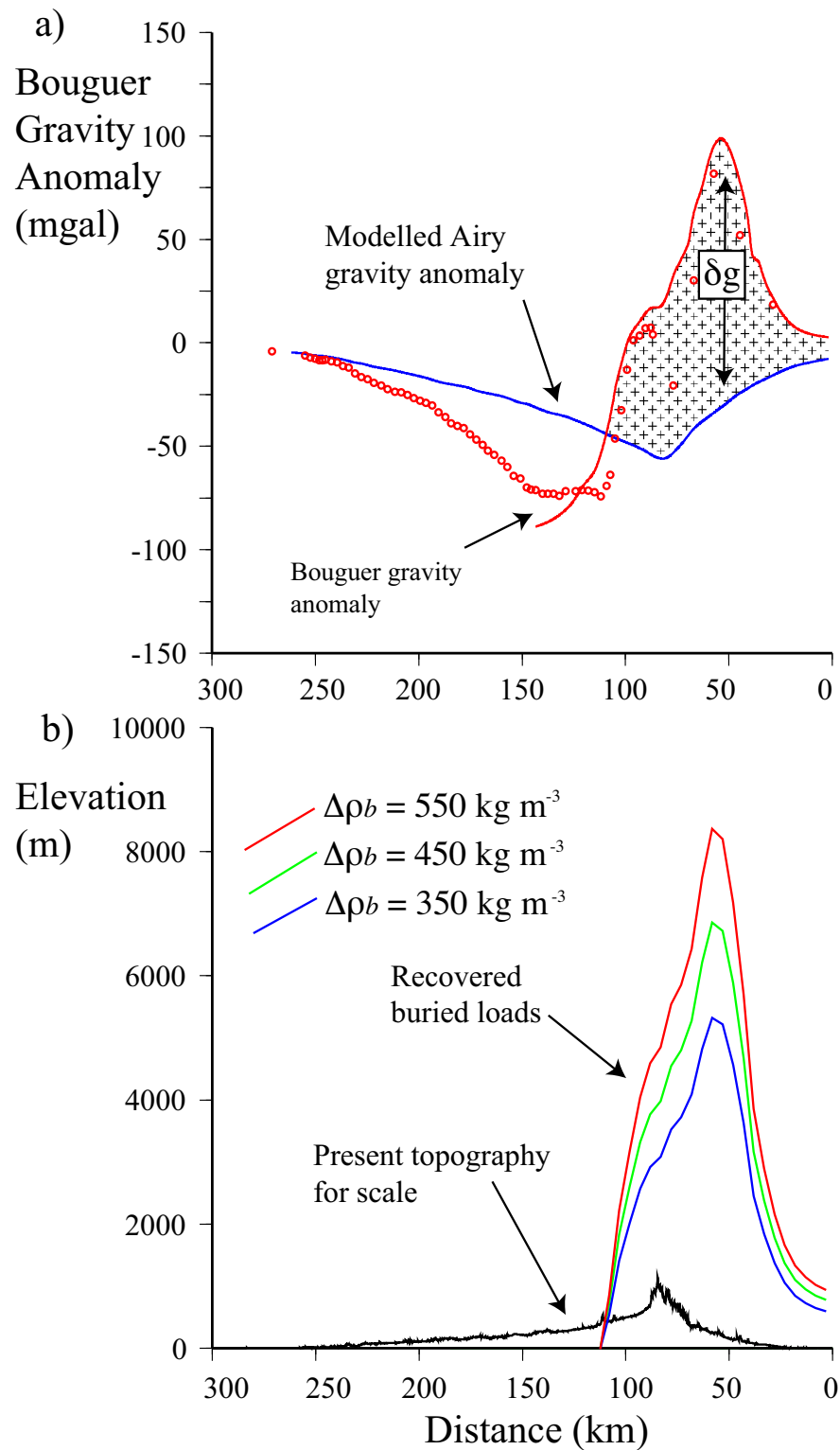


Figure 3.15: Recovery of buried load size using the difference between observed Bouguer gravity anomaly and modelled Airy gravity anomaly. a) Observed Bouguer gravity anomaly - red dots observed points and solid red line from Ravaut & Warsi [1997a]. Blue line modelled gravity anomaly, assuming Airy compensation of surface topography. Hashed area, proportion of the observed Bouguer anomaly attributed to the excess mass of a buried load. b) Observed surface topography and estimated size of the buried load, assuming a range of density contrasts.

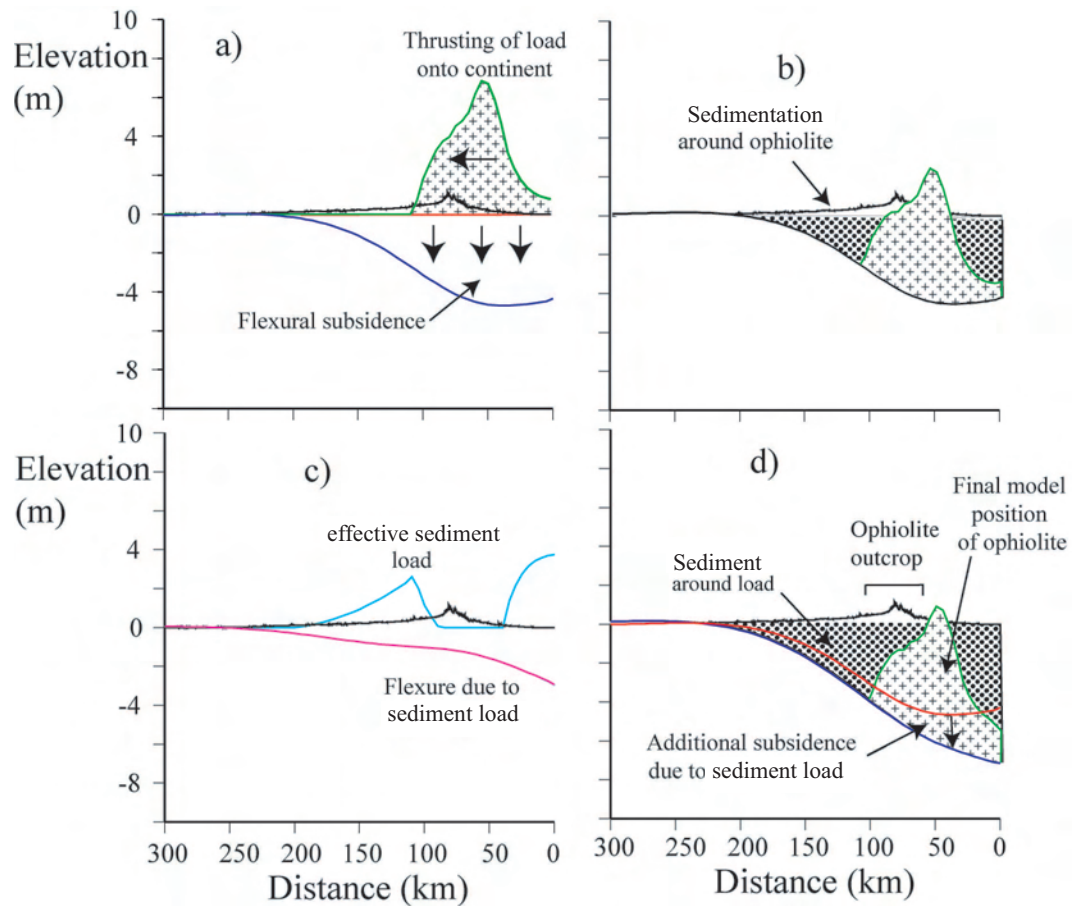


Figure 3.16: Highly simplified model of the obduction process and subsequent flexural relaxation. a) Ophiolite load (green body) is thrust onto continent, this causes flexure (blue line). b) Ophiolite 'falls into' flexural depression which fills with sediment (dashed region). c) Sediment represents a load (pale blue line) which causes further flexure (purple line). d) Sum flexure for both ophiolite and sediment loads leads to final blue line and a final position of ophiolite shown in green.

3.3.3 Parameters

There are several key parameters, aside from T_e , which must be put in to the flexural and gravitational calculations, and were, in part discussed in section 2.2. However, in the case of buried loading on a passive-margin additional parameters can be varied. A summary of these is shown in Table 3.1. Three types of parameters were considered: firstly, constants which are typical values in flexural/geophysical studies such as that of Karner & Watts [1983] (e.g. crust and mantle densities); secondly, model parameters which depend on how the obduction process is viewed (e.g. infill or displacement densities); and finally, variable parameters which, like T_e , can be assessed within a reasonable range to recover a best fitting and hence most likely value.

Parameter	Value	Type
Crustal thickness	21 to 35 km	Variable
Crustal density	2800 kg m ⁻³	Constant
Mantle density	3330 kg m ⁻³	Constant
Infill density	1010 to 3100 kg m ⁻³	Model
Load density	2650 to 3250 kg m ⁻³	Variable
Displace density	0 or 1010 kg m ⁻³	Model
Plate break	0 to 50 km	Variable

Table 3.1: Parameters assumed in the flexure and gravity modelling. Constants are fixed values which within their potential range have little effect on the recovered T_e value (Lyon-Caen & Molnar [1983]). Model parameters are values which are varied between certain fixed values depending on the geological model of the obduction process. Variable parameters are those where a best fitting value is determined by searching parameter space.

Before a search of parameter space can be performed to recover best fitting values for the ‘variable’ parameters outlined in Table 3.1, the ‘constant’ and ‘model’

parameters must be defined for each step of the loading process. The ‘model’ values used at each step of the model can be assessed in two ways. Firstly by consideration of the model for the obduction process (Figure 3.16) and secondly by comparison of the model results with the observed data.

Model load

To calculate the flexure due to the buried load, the entire load recovered by Equation (3.4) was considered to be the driving load. This was because it is not known which part of the load should be considered as the infilling load and which part should be considered as the driving load. This assumption can be made if the infilling material is considered to be water as the ophiolite can ‘fall into’ the depression it creates. The consideration that the infilling material was water is valid for the process of ophiolite emplacement, as the foreland is known to have subsided and drowned coincident with the final emplacement of the ophiolite (e.g. Patton & O’Connor [1988]).

Load, displacement and infill densities

For the case of surface loading the load density applied has little impact on the resulting flexure as has been shown in Figure 3.11 and previously by Lyon-Caen & Molnar [1983]. However, in the case studied here the size of the buried load, and hence the resulting flexure, depends critically on the density considered for the ophiolite and the contrast with the surrounding material. The ophiolite density has been estimated by previous authors, based on both displacement experiments and consideration of the ophiolite mineralogy (Shelton [1984] p121, Shelton [1990], Ravaut *et al.* [1997]) to be in the order of 2800 to 3100 kg m⁻³ depending on the extent of serpentinisation and the proportion of mantle vs crustal material present in the ophiolite. Given this large range it was decided to vary the ophiolite density to assess which value resulted in the best fitting flexural profile. The reference

density of the material surrounding the ophiolite was kept constant at 2650 kg m^{-3} . This is in-line with values suggested by Shelton [1984]. This value is also very close to the standard value of 2670 kg m^{-3} used for the Bouguer correction. If the material between the observation point and the geoid is ophiolite, and hence has a higher density, part of the positive Bouguer anomaly will stem from this dense ophiolitic material, rather than dense material below the geoid. By considering the buried ophiolite to be surrounded by material with approximately the same density as used in the Bouguer correction, the thickness of the ophiolite above the geoid will be included in the total load thickness calculated.

The material that the ophiolite body displaces during emplacement makes a significant difference to the modelled flexure and hence the resulting calculated gravity anomalies, as can be seen in Figure 3.17. The displacement density is significant because the driving load in effect weighs less when its mass is in part buoyantly supported when surrounded water, compared to when it is surrounded by air. The figure suggests that the model considering ophiolite displacing air gives a better fit to the observed Bouguer anomaly, however, further consideration is made of this parameter based on a discussion of the obduction process and geological observations.

Initially a displacement density of water was considered which implies that the ophiolite was emplaced in a submarine setting. A displacement density of water was considered as the ophiolite load is oceanic crust which must have been thrust up into the water column. Also the foreland is known to have been submerged due to flexural down-warping when the ophiolite was finally emplaced (Patton & O'Connor [1988]). However, with a displacement medium of water, as models of increasing T_e are considered, the flexural depression below the ophiolite became significantly smaller than the load height. This means that a large proportion of the model load is above the top of the flexed plate, assumed to approximately coincide with sea level. This is inconsistent with the fact that of the $\sim 10 \text{ km}$

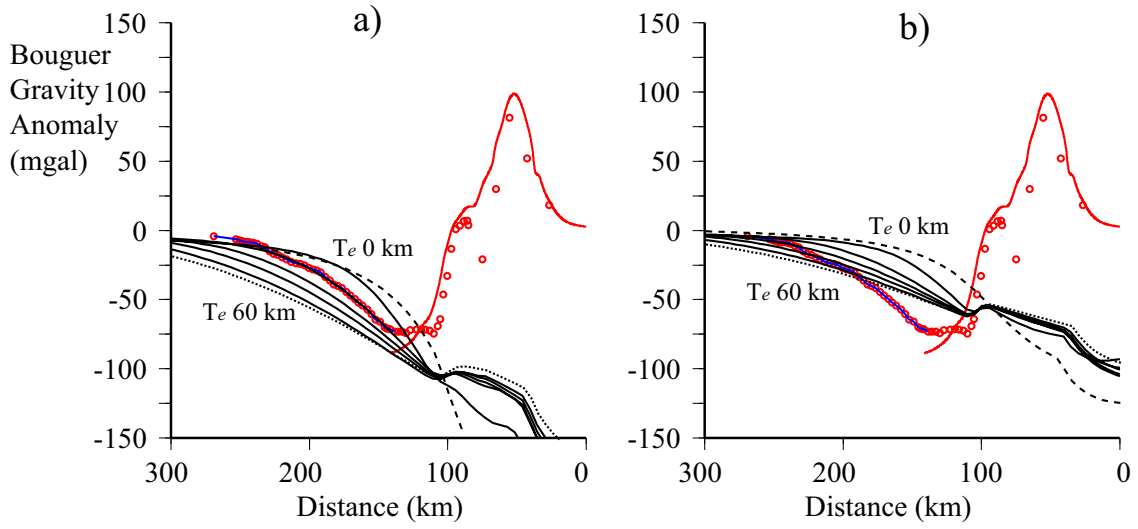


Figure 3.17: Comparison of modelled Bouguer gravity anomalies based on ophiolite loading with two alternative displacement densities for T_e values from 0 to 60 km. Red lines and points are Bouguer gravity data corrected for long wavelength effects. a) Initial ophiolite load ($\rho_{oph}=3200 \text{ kg m}^{-3}$) displacing air. b) Initial ophiolite load displacing water with a density of 1010 kg m^{-3} .

of ophiolite, calculated to be present from the observed gravity anomaly, $<2 \text{ km}$ is today expressed as surface topography. An additional line of evidence comes from Patton & O'Connor [1988] who show that some of the material infilling the foreland basin contains clastic sediments made up of ophiolitic fragments derived from the thrust load. This reinforces the view that the thrust load was exposed and displacing air rather than water soon after emplacement.

The displacement density for the sediment filling in the flexural depression around the ophiolite load in the second flexural step was considered to be water. This is reasonable as the later deposited foreland fill is predominantly marine shales, marls and limestones, intercalated with more clastic rich material (Boote *et al.* [1990], Patton & O'Connor [1988]).

Infill density values were chosen according to the model used. It is important that the correct value (e.g. water or sediment) is used, depending on the model of emplacement. An infill density of water is used when modelling the flexure due

to the ophiolite load. This allows the consideration of the deflection caused by the force due to the entire ophiolite without considering ophiolitic material in the foreland region. An infill density of sediment was considered when modelling the flexure due to filling of the foreland basin with sediment.

Crustal thickness, crustal density and mantle density

Continental crustal thickness was initially taken to be a constant 35 km, which is the global average value for crustal thickness taken from Fowler [1990] p351. This value is less than the 38 to 40 km recovered from seismic refraction studied over the central Arabian shield (Mooney *et al.* [1985]). However, varying the background crustal thickness from 35 to 40 km can be shown to have little effect on the modelled gravity anomaly and hence on the recovered T_e and load density. Crustal thickness did have to be varied beneath the load for reasons discussed later in section 3.3.4.

Crustal density is assumed to be a standard 2800 kg m^{-3} in all calculations. This density does not affect the amount of flexure but does affect the modelled gravity anomaly due to deflection of the Moho.

Mantle density was kept constant at 3330 kg m^{-3} . Previous authors such as Lyon-Caen & Molnar [1985] have shown that this parameter does not have a significant effect on the recovered value of T_e .

Plate break

In orogenic settings the loaded plate is almost invariably modelled as broken (e.g. Stewart & Watts [1997]). In the case of the Oman Ophiolite a broken plate is justified for two reasons. Firstly, it is bounded to the north by a large scale normal fault (e.g. Searle *et al.* [1983] and Searle *et al.* [2004]). This fault represents a large-scale break in the crust, and hence a significant part of the elastic lithosphere, across which only limited amounts of stress could be supported (Bird [1995]). Secondly, gravity and seismic studies (Shelton [1984], Shelton [1990] and Ravaut *et al.* [1997])

indicate that the oceanic crust making up the ophiolite is not continuous with the oceanic crust off shore, again suggesting a broken plate. However, the exact position of the plate break with respect to the ophiolite load is not known. The position of the plate break was therefore varied to allow the recovery of a best fitting break position.

3.3.4 Results

Initially the gravity effects of two interfaces were modelled: one at the base of the foreland basin (sediment vs crust) and the other at the base of the crust (crust vs mantle). Both interfaces were considered to continue 200 km beyond the origin, at their origin depth, to prevent edge effects entering into the calculated anomalies. The calculated gravity anomaly was compared to the observed Bouguer gravity anomaly across the foreland region.

The best fitting parameter set (T_e , $\rho_{ophiolite}$ and plate break) is defined as that which yields the smallest RMS difference between observed and calculated gravity anomalies. RMS difference was calculated as shown in equation 3.5.

$$RMS = \sqrt{\sum_{i=1}^n \frac{(x_i - \bar{x})^2}{n}} \quad (3.5)$$

Where x_i is the difference between model and observed at each point and $\bar{x}$ is the mean difference between model and observed. The results of the models varying T_e , load density and plate break position are shown in Figure 3.18.

Figure 3.18 shows that good fits to the data over the foreland region are achieved with T_e values of around 30 km and ophiolite densities of 3150 to 3250 kg m⁻³, depending on the position of the plate break. Plate break positions closer to the foreland can be seen to need higher load densities to achieve the same fit between calculated and Bouguer gravity anomalies. The figure shows that the modelled anomalies do not fit the observed Bouguer anomaly beneath the ophiolite; therefore the anomaly generated by the ophiolite within the flexural depression and

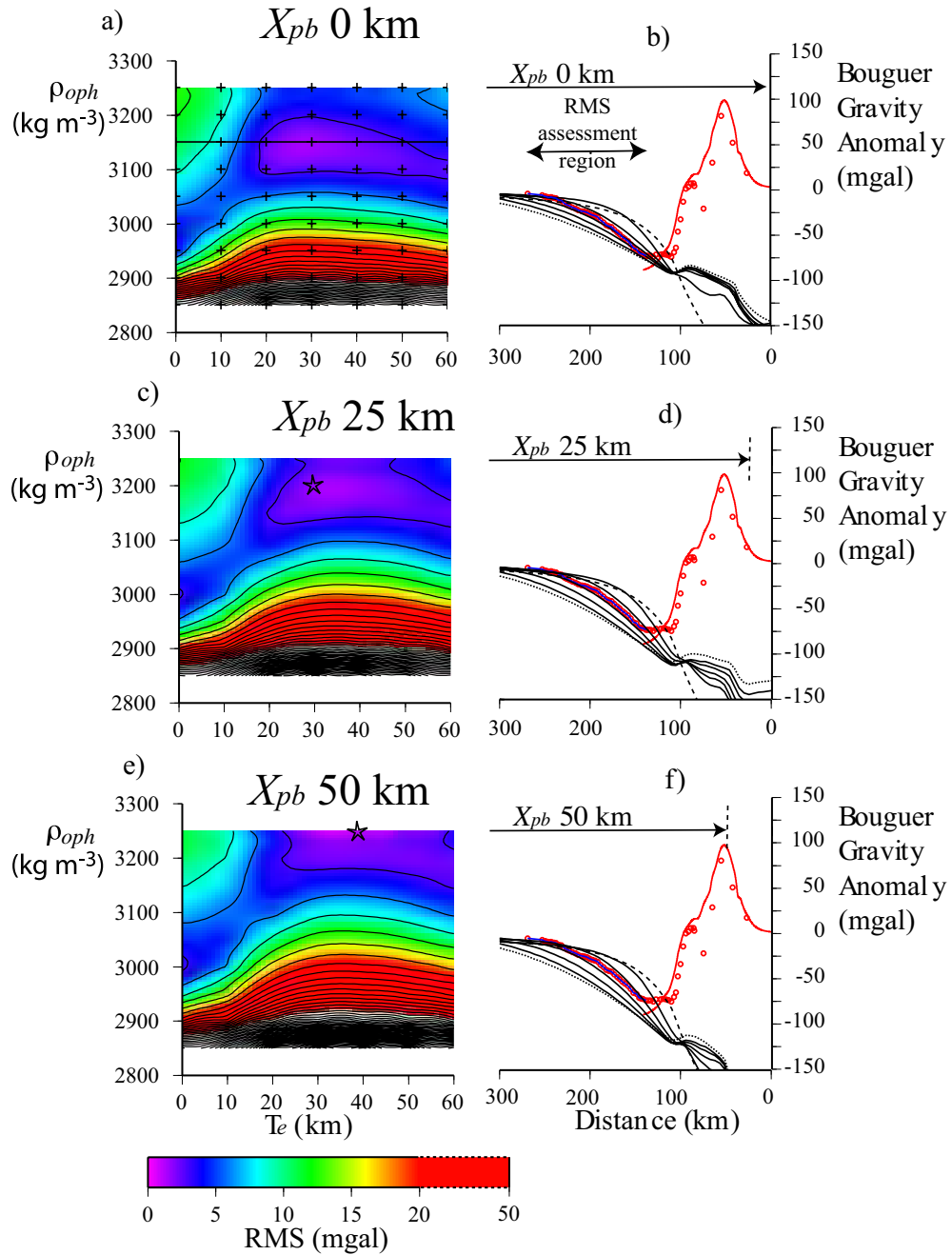


Figure 3.18: Results of models of gravity anomalies driven by ophiolite loading with various input parameters. a) RMS difference between calculated and Bouguer gravity anomaly data over the foreland region, contoured at 3 mgal intervals. Crosses mark parameter pairs assessed in all subsequent models. The T_e of the modelled plate falls to 0 km at the plate break (X_{pb}) and is constant and high beyond that into the foreland. b) Thin black lines, calculated anomalies based on an $\rho_{oph} = 3150 \text{ kg m}^{-3}$ which gave the best-fit to the observed data, for $X_{pb} = 0$. Red stars and thick red line, Bouguer anomaly data corrected for long wavelength anomaly. c) Same as a) but with a plate break 25 km from the origin. d) Same as b) but with $X_{pb} = 25 \text{ km}$. e) Same as a) but with $X_{pb} = 50 \text{ km}$ from the origin. f) Same as b) with ρ_{oph} of 3200 kg m^{-3} and $X_{pb} = 50 \text{ km}$ from the origin.

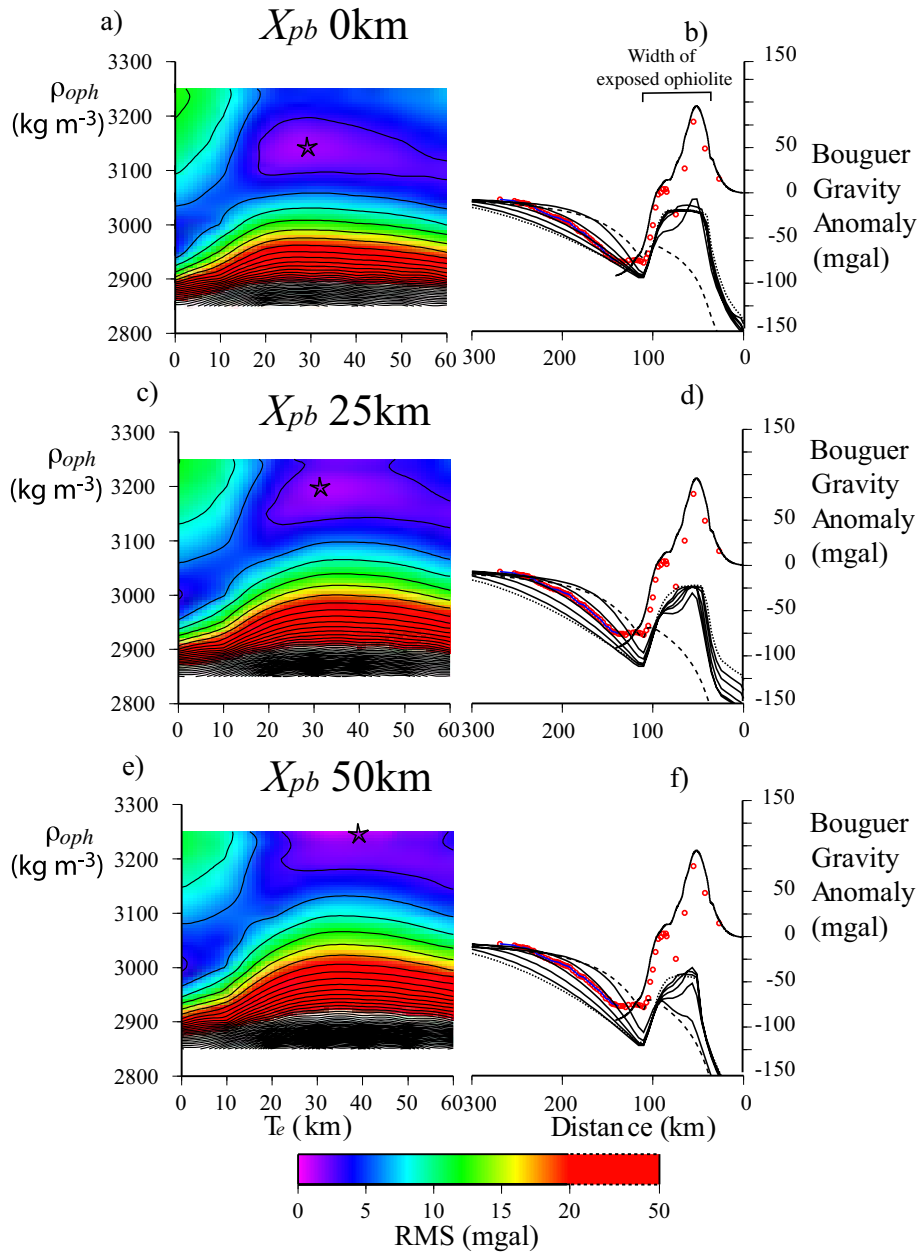


Figure 3.19: Modelled Bouguer gravity, including effect of ophiolite body. a)-f) have the same structure as in Figure 3.18 with the anomaly due to the ophiolite body added.

surrounded by sediment was calculated and added to the modelled anomalies.

The addition of the anomaly due to the ophiolite body, shown in Figure 3.19, can be seen to reduce the discrepancy between the model and the data over the ophiolite. However, the magnitude of the observed Bouguer anomaly positive is not recovered. This is because the flexure due to the load deflects the Moho to such

an extent that the magnitude of the negative anomaly cannot be compensated for by the ophiolite body.

Crustal thinning

An explanation for the discrepancy between the observed and calculated positive parts of the gravity anomaly, would be that the crust flexed by the ophiolite was thinned. This would be expected given that the ophiolite was obducted over a passive-margin sequence (Searle *et al.* [1983]) where the continental crust would normally be thinned (e.g. Stampfli *et al.* [1991], Wooler *et al.* [1992], Whitmarsh & Wallace [2001], Corfield *et al.* [2005]). The amount of stretching at a margin can be defined by the stretching factor β which is the ratio of the thickness of un-stretched to stretched crust in that region. A range of stretching factors (β) for rifted margins have been recovered. These include β values of 1.13-1.14 for the margin which existed to the north of India before the collision with Eurasia Corfield *et al.* [2005]. β values of 1.25 to 1.3 from Wooler *et al.* [1992] for peri-tethian margins in Europe. Davis & Kusznir [2002] recover stretching factors ranging from ~ 0.8 to ~ 3.4 for a range of modern margins across the globe and Whitmarsh & Wallace [2001] imply β values of over 10 for the modern Iberian margin. Davis & Kusznir [2002] also present widths for the regions of continental thinning ranging from ~ 75 to ~ 275 km across a range of modern passive-margins.

Figure 3.20b shows the residual anomaly for the modelled flexure given a driving ophiolite load with density of 3150 kg m^{-3} , a plate break at 25 km and a T_e of 30 km. This was the configuration in Figure 3.19 which gave the best-fit over the foreland. This residual can be accounted for by assuming the first 50 km of the over-ridden crust had thinned by 13 km.

Figure 3.20b and d shows that when only the first 50 km of the model crust is thinned a significant positive anomaly is seen up to 200 km westward over the foreland region. The model was therefore re-run to include this crustal thinning

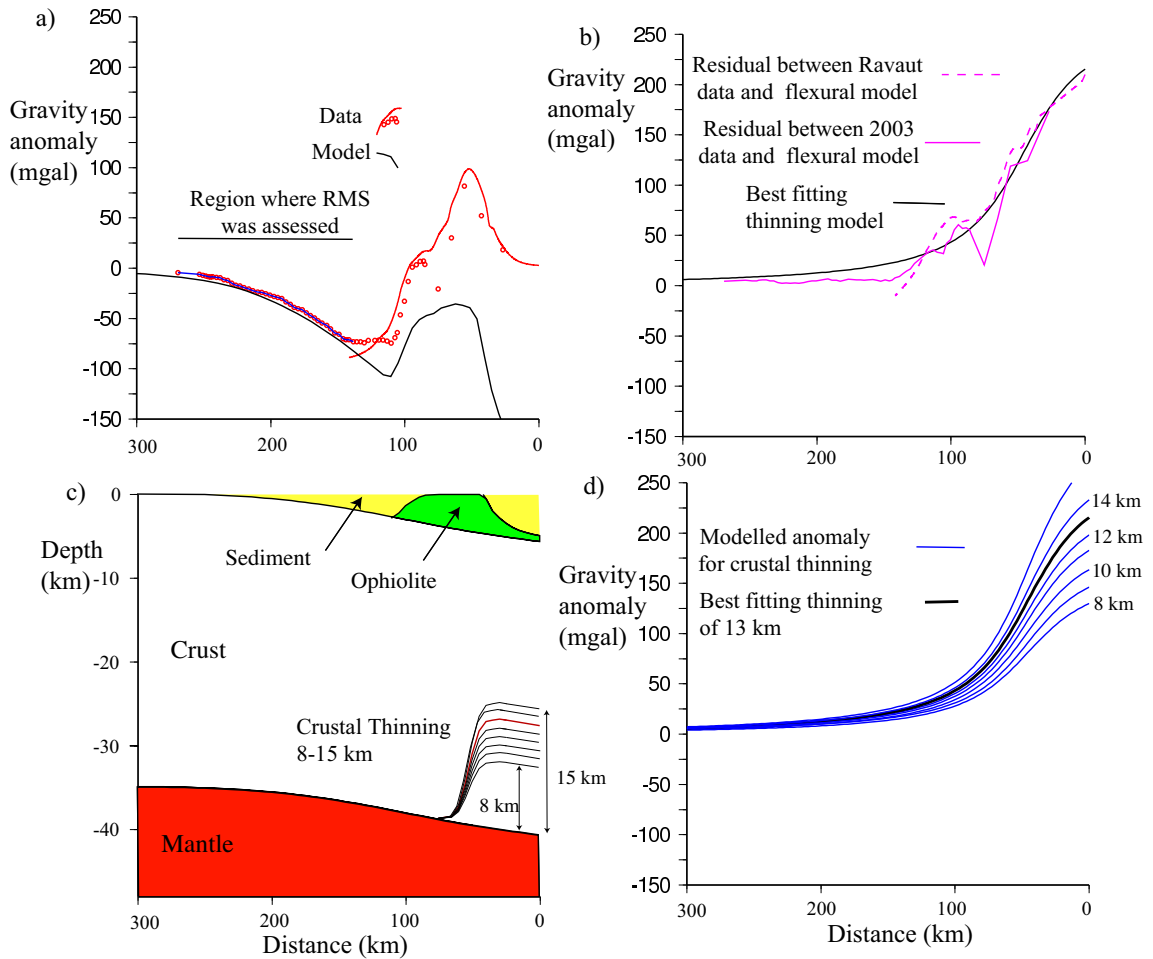


Figure 3.20: Difference between modelled and observed Bouguer gravity anomalies. a) Model and observed gravity anomaly given a driving ophiolite load with density of 3150 kg m^{-3} , a plate break at 25 km and a T_e of 30 km. Parameters based on the best fitting values from Figure 3.19. b) Residual between model and observed anomaly shown in purple. Best model shown in black. c) Crustal structure used to produce model anomaly shown in a, and region of thinned crust used to create models in d. Red line shows amount of crustal thinning which best fits residual in b. d) Anomaly due to thinning of the continental crust for thinning amounts of 8 to 15 km. Black line is best fitting model

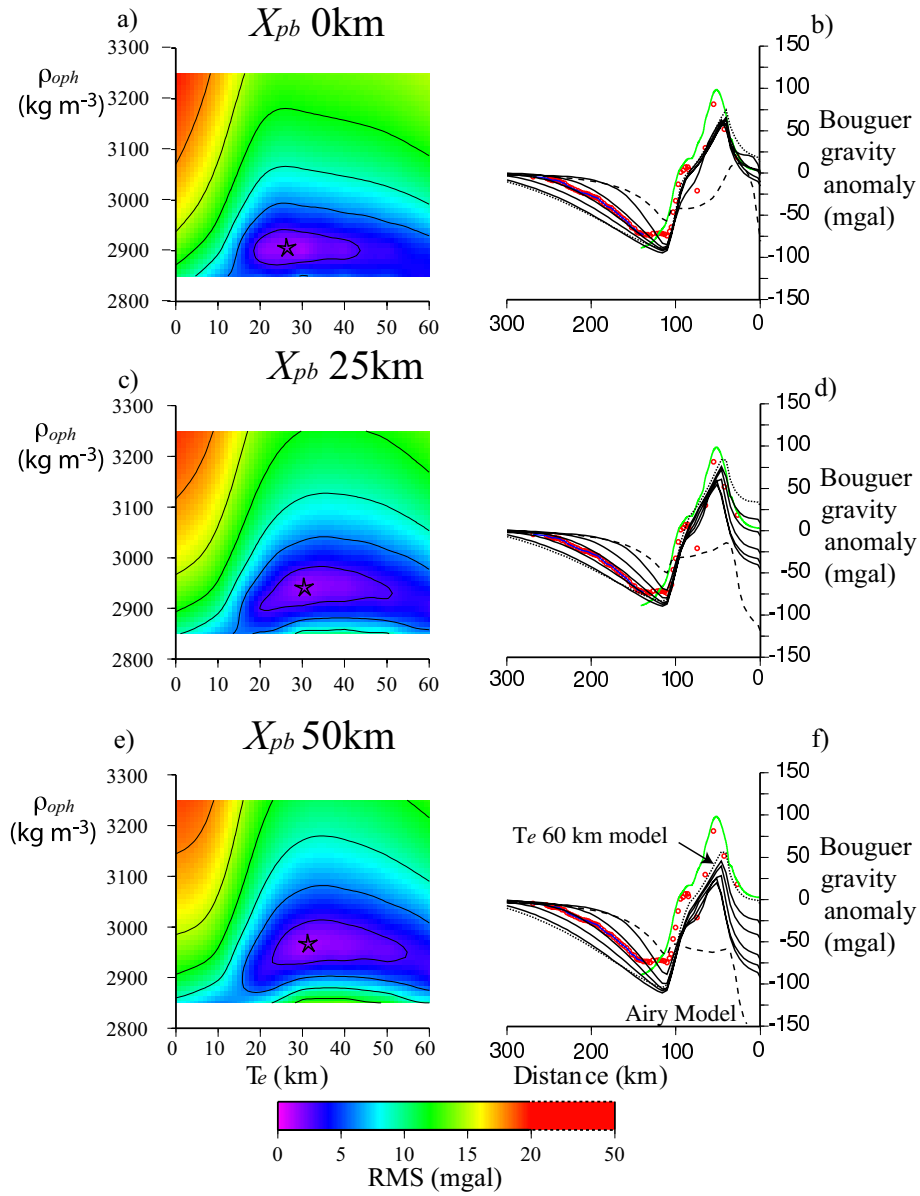


Figure 3.21: Results for models including 13 km of crustal thinning for the first 50 km of the model below the ophiolite. Best fitting values are $T_e = 30$ km, $\rho_{oph} = 2900$ kg m $^{-3}$ and plate break position at the origin. a-f) show the same parameters as in earlier figures.

factor. The model was re-run for all values of T_e , load density and plate break position that had been considered previously, assuming the first 50 km of the crust below the ophiolite had thinned by 13 km. The results are shown in Figure 3.21.

The best fitting load density is now different, as the anomaly resulting from

crustal thinning is affecting the modelled anomaly in the foreland region where it is compared to the observed Bouguer anomaly. The gravity anomaly over the load is also not completely explained so a reassessment was made of the residual anomaly based on the best parameters derived from Figure 3.21. This reassessment shown in Figure 3.22 shows that thinning of 14 km is more appropriate.

Figure 3.23 shows the RMS for various T_e values and plate breaks. The best-fit T_e is approximately 30 km and best load density is close to 2900 kg m^{-3} . Plate break position does not dramatically affect this outcome. Figure 3.24 shows the gravity model which best fits the foreland data (by RMS) and the load region by eye. Figure 3.25 shows the physical model which generated the best fitting gravity model.

3.3.5 Profile 1 summary

The gravity data collected in the 2003/2004 field season shows, in relation to the observed topography, a distinct flexural shape typical of a foreland basin created predominantly by buried loading (Karner & Watts [1983] and Lin & Watts [2002]).

The best fitting value of T_e recovered for this profile is ~ 30 km. Plate break position has a minor impact on the recovered value of T_e in this region with best fitting values of 30 to 40 km being recovered for all plate break positions. This differs from the results seen in the synthetic tests for the surface loading method shown earlier where plate break position was easily recovered. The difference stems from the variable load density. The increase in load density compensates for the decrease in load width and acts to maintain effectively the same load on the plate.

The crustal thinning needed to fit the positive part of the Bouguer gravity anomaly does not appear to alter significantly the recovered T_e values. This is partly due to the RMS error not being calculated over this region. However, the change in crustal thickness does have some impact on the foreland which is compensated for by the variation in load density.

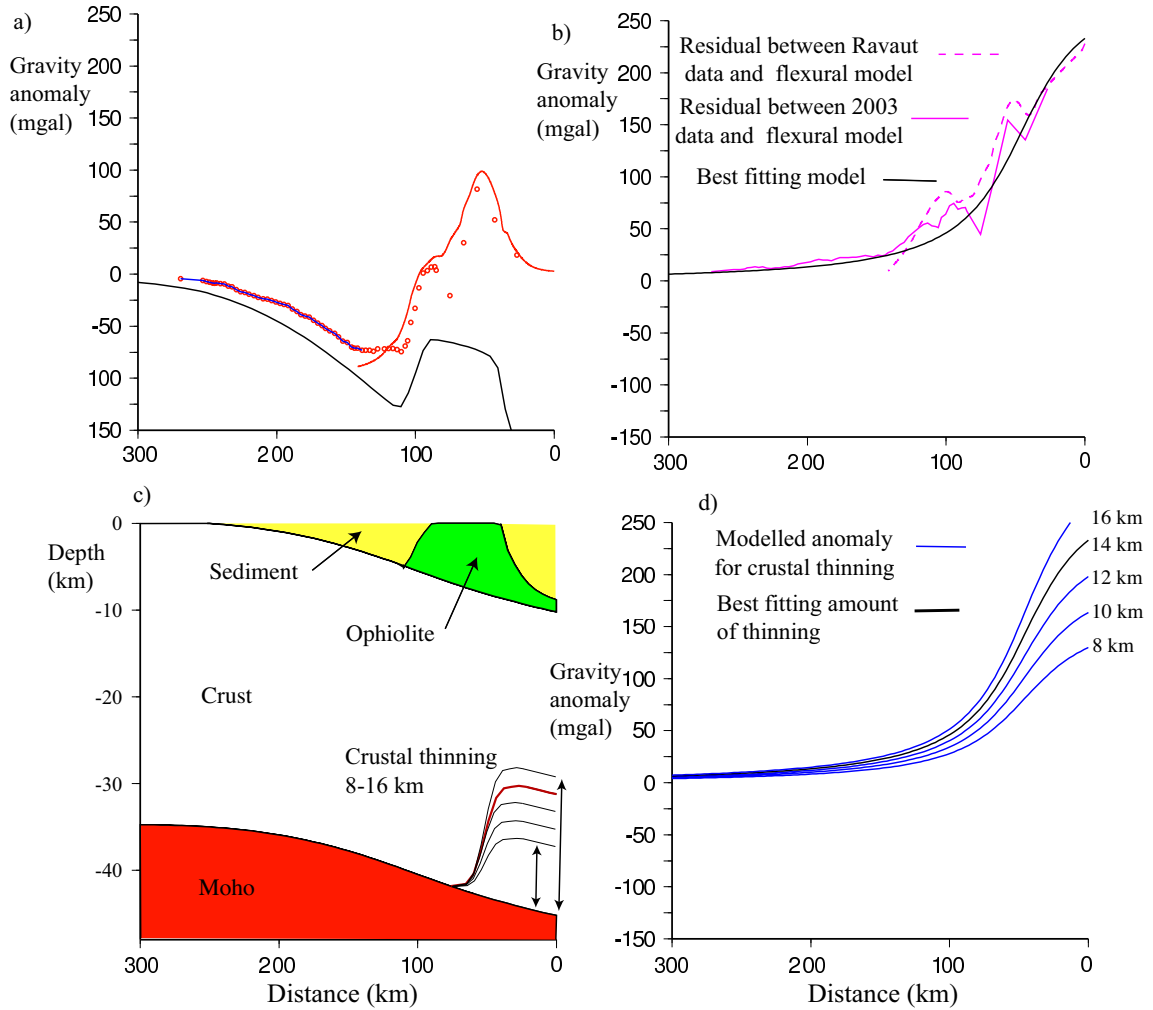


Figure 3.22: Reassessment of residual anomalies based on new best fitting values for load density and plate break determined from Figure 3.21. a) Calculated and Bouguer gravity anomalies calculated *without* crustal thinning using the new best-fit values for plate break and load density seen in Figure 3.21. b) Residual between calculated and observed Bouguer gravity anomalies in purple. Black line equates to best fitting amount of thinning of 14 km. c) Crustal structure generated by best flexural model and various amounts of crustal thinning that generated anomalies shown in d). d) Modelled anomalies due to crustal thinning of 8, 10, 12, 14 and 16 km.

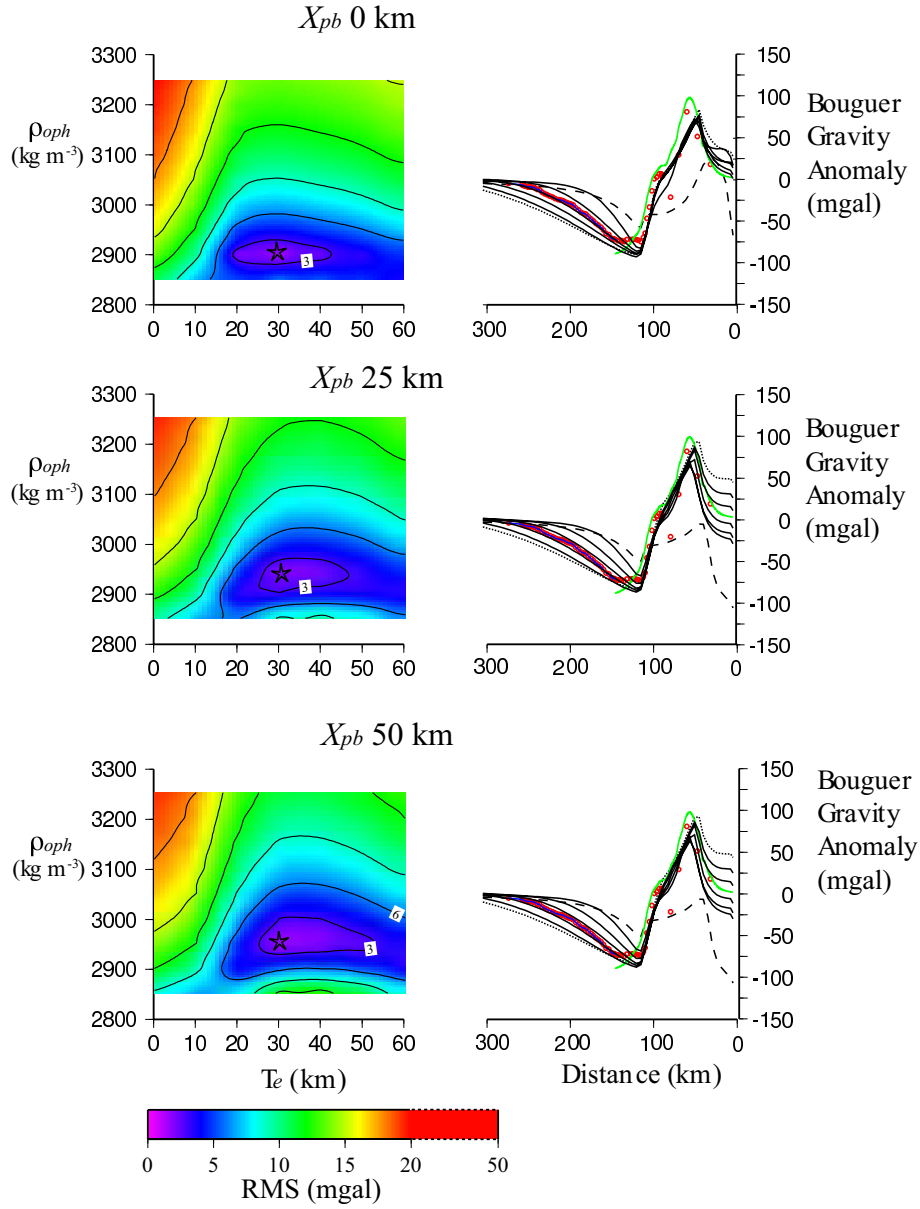


Figure 3.23: Results for modelling Bouguer gravity anomalies with 14 km of crustal thinning imposed on the first 50 km of the profile. a)-f) show the same as previous figures.

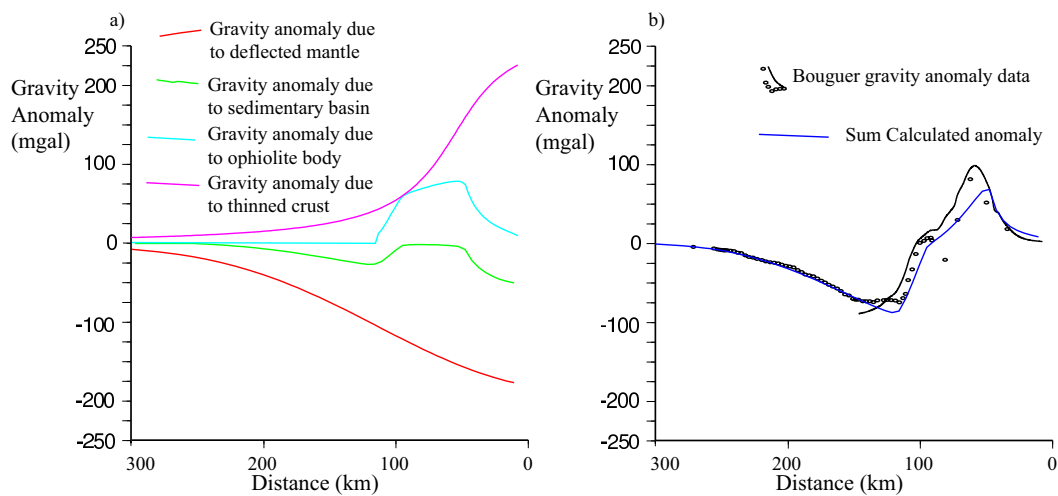


Figure 3.24: Modelled gravity anomaly which gives the best RMS fits the foreland gravity data and visual fit to the load part of the gravity data. $T_e=30$ km, ophiolite density = 2900 kg m^{-3} , a plate break at the origin and 14 km of crustal thinning. a) Component anomalies. b) Sum calculated anomaly in blue and observed Bouguer anomaly data in black.

The best fitting value of ophiolite load density for the final model, including crustal thinning, ranges from ~ 2900 to $\sim 2950 \text{ kg m}^{-3}$, depending on the position of the plate break. These values are in agreement with those recovered by Shelton [1984] p121, Shelton [1990], Ravaut *et al.* [1997]). These values are dependent on the amount of crustal thinning used. When no thinning was included in the model, as shown in Figure 3.19, best fitting values for the ophiolite density were closer to 3150 kg m^{-3} . However, these high values are at the absolute upper limit suggested by previous authors. This indicates that the model including crustal thinning is a better geological representation.

The values of T_e recovered by this study agree with those recovered by Ravaut *et al.* [1997] but are almost a factor of two higher than those recovered by Patton & O'Connor [1988] and Shelton & Egan [1991]. The values recovered by Patton & O'Connor [1988] are for the flexural bulge associated with initial ophiolite emplacement, which lies beneath the final foreland basin. They therefore reflect the T_e of a portion of the continental crust which lay closer to the ocean-continent

transition than that finally loaded by the ophiolite. Crust closer to the ocean-continent transition is likely to have been more effected by rifting processes which lead to weakening of the lithosphere (Watts [1992]). It is not immediately clear to which along strike section of the foreland the values of T_e recovered by Shelton & Egan [1991] refer, and they may therefore represent the T_e of a different part of the foreland.

The modelled anomaly which best fits the observed Bouguer anomaly data uses an ophiolite load thickness of ~ 12 km which agrees with values for the ophiolite thickness suggested by previous authors (Robertson & Searle [1990] Shelton & Egan [1991], Ravaut *et al.* [1997]). Shelton & Egan [1991] assumed a load height of 23 km based on a model of the obduction process, however, gravity data suggests that this is not present today.

When the load, recovered from the gravity data, is placed in the final flexural depression it can be seen that there is a residual topography with a peak height of ~ 4 km (Figure 3.25). This is in excess of what is observed today but not by a large amount. The discrepancy can be largely accounted for by the broad scale of the model not incorporating local density variations. These could mean that in places higher density material may be present, a smaller volume of which would give the same driving load. Also, erosional redistribution of dense ophiolitic material by mass-wasting of the ophiolite and deposition of eroded material into the foreland would be expected to change the surface expression of the load without significantly altering the amount of flexural subsidence.

Patton & O'Connor [1988] present isopach data for the sediments deposited in the foreland after the final emplacement of the ophiolite at ~ 90 Ma, until 65 Ma. This data suggests that the final position of the flexural bulge is over 200 km from the load front. This differs from the position of the flexural bulge which is ~ 150 km from the load front, seen in the model with a T_e of 30 km which best fitted the observed Bouguer anomaly. Digitised points from their data-set are shown in

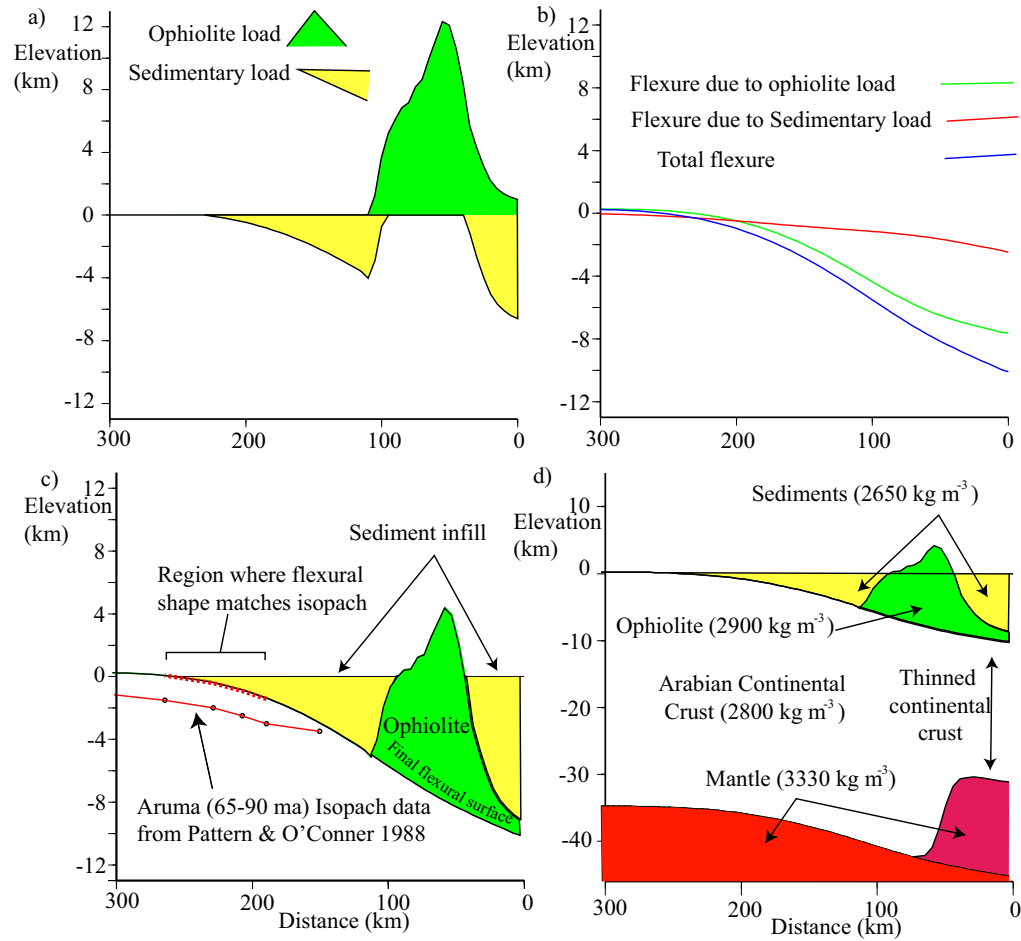


Figure 3.25: Results for load sizes and modelled flexure based on best fitting values of T_e (30 km), ρ_{oph} (2900 kg m⁻³) and plate break at the origin. a) Ophiolite load derived from Equation 3.4 and sediment load derived from filling in flexural depression around ophiolite. b) Flexure due to ophiolite, sediment and the sum of both. c) Final model configuration for the foreland area. Also shown is the isopach data from Patton & O'Connor [1988]. d) Final total crustal structure.

Figure 3.25 which can be seen to lie below the model flexed surface. However, since the isopach data spans a 25 Ma interval significant sedimentation may have occurred, in addition to that filling the flexural depression, in effect ‘blanketing’ the flexural surface. If the data from Patton & O’Connor [1988] is shifted upward by ~ 1.5 km a section of the data fits the shape of the flexural curve close to the bulge region found in this study. It is therefore proposed that the isopach data from Patton & O’Connor [1988] supports the result that the best fitting T_e value is ~ 30 km, once the effect of blanketing sediments have been removed.

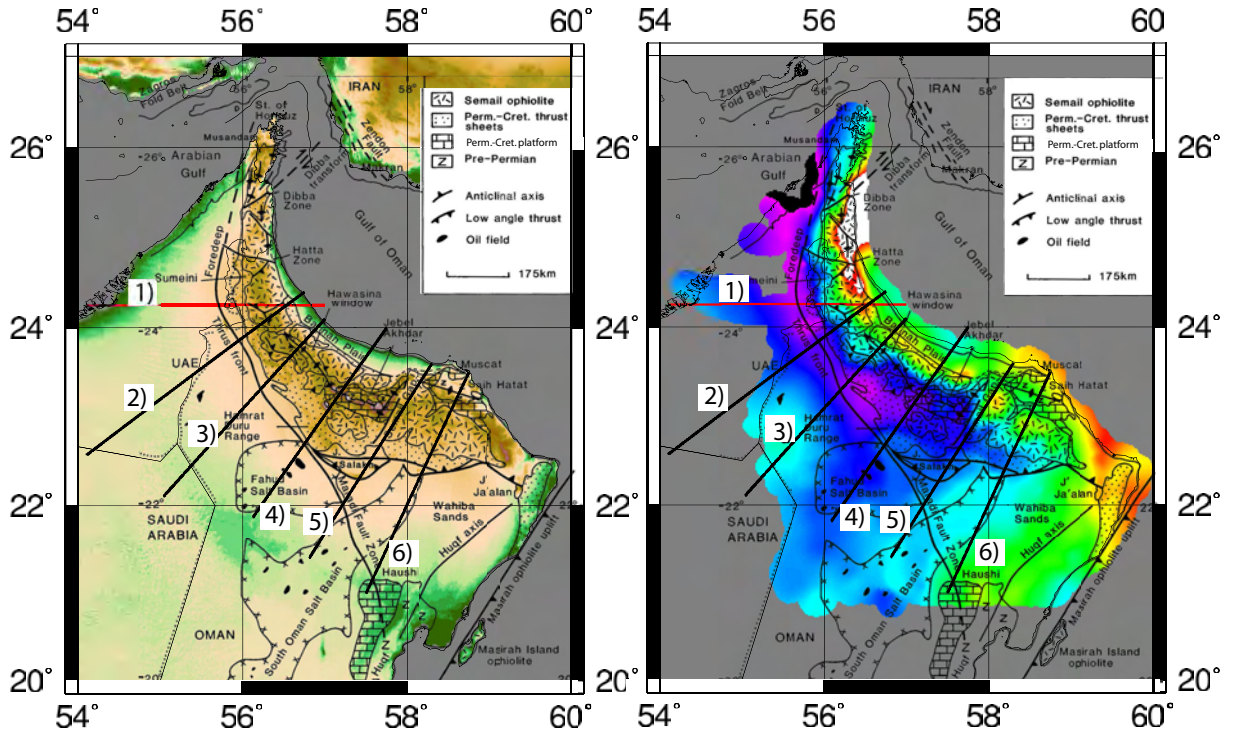


Figure 3.26: Location map of Oman profiles, and geological structures from Robertson & Searle [1990]. a) Topography. b) Bouguer gravity data. Red line shows profile A-A', black lines show profiles further south considered in this section.

3.4 Other profiles

Having recovered T_e using flexural techniques that incorporated buried loading across the U.A.E. part of the foreland, further profiles were assessed along strike of the foreland to the south east. These were designed to examine the spatial variation in flexural rigidity within the Oman Ophiolite foreland system. Profile locations are shown in Figure 3.26.

Using the same parameters as in the previous section the flexural part of the observed gravity data was again fitted, i.e. the negative Bouguer anomaly across the foreland. For these profiles data solely from Ravaut & Warsi [1997a] was used.

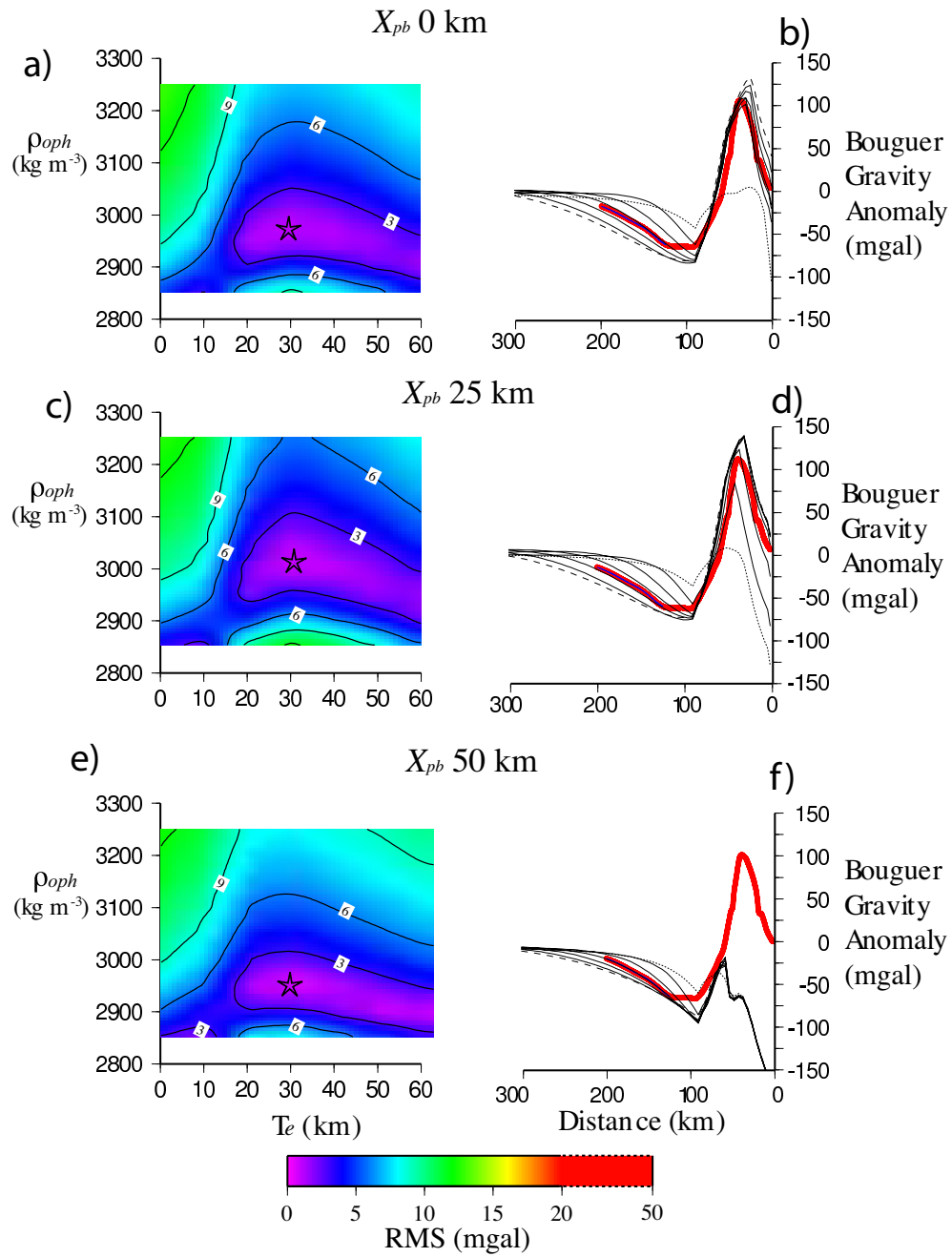


Figure 3.27: Model of Profile 2. Panels as in Figure 3.21.

3.4.1 Profile 2

Figure 3.27 shows the results of a range of flexural models applied along profile 2. A T_e of close to 30 km is recovered for this profile based on load densities of 2950 to 3000 kg m⁻³. However, the RMS minima is less well defined than in the previous profile (Figure 3.23). This is due to Bouguer gravity anomaly data not extending across the full width of the foreland region. The best position for the plate break is not clearly defined by the different RMS minima. However, with the plate break 50 km from the origin (i.e. the case with the narrowest load) the modelled anomaly does not fit the observed data beneath the load.

This result suggests that over the northern part of the foreland, as sampled by profile 1 in the U.A.E. and profile 2 in Oman, there is little change in rigidity. T_e values of 30 km being recovered in both cases.

3.4.2 Profile 3

The same technique and parameters for recovering T_e in the presence of buried loads were applied to data from profile 3. This profile lies approximately 25 km to the south of profile 2. The results of these models are shown in Figure 3.28.

In contrast to the earlier profiles, Figure 3.28 clearly shows that there is no set of parameters, within the range searched, that give a good fit between the calculated gravity anomaly and observed Bouguer anomaly. The best RMS fits are for ophiolite densities which are low compared to the 2 profiles to the north and the values presented by previous authors (Shelton [1990], Ravaut *et al.* [1997]). When modelled profiles for the best fitting load densities are visually assessed (as in Figure 3.28b, d and f) the observed Bouguer gravity anomaly can clearly be seen to be poorly matched.

To ensure that the data on a single profile gave a good representation of the local field all data within $\pm \sim 27$ km of the line was considered. The mean recovered was almost indistinguishable from the single profile so the discrepancy is not related to

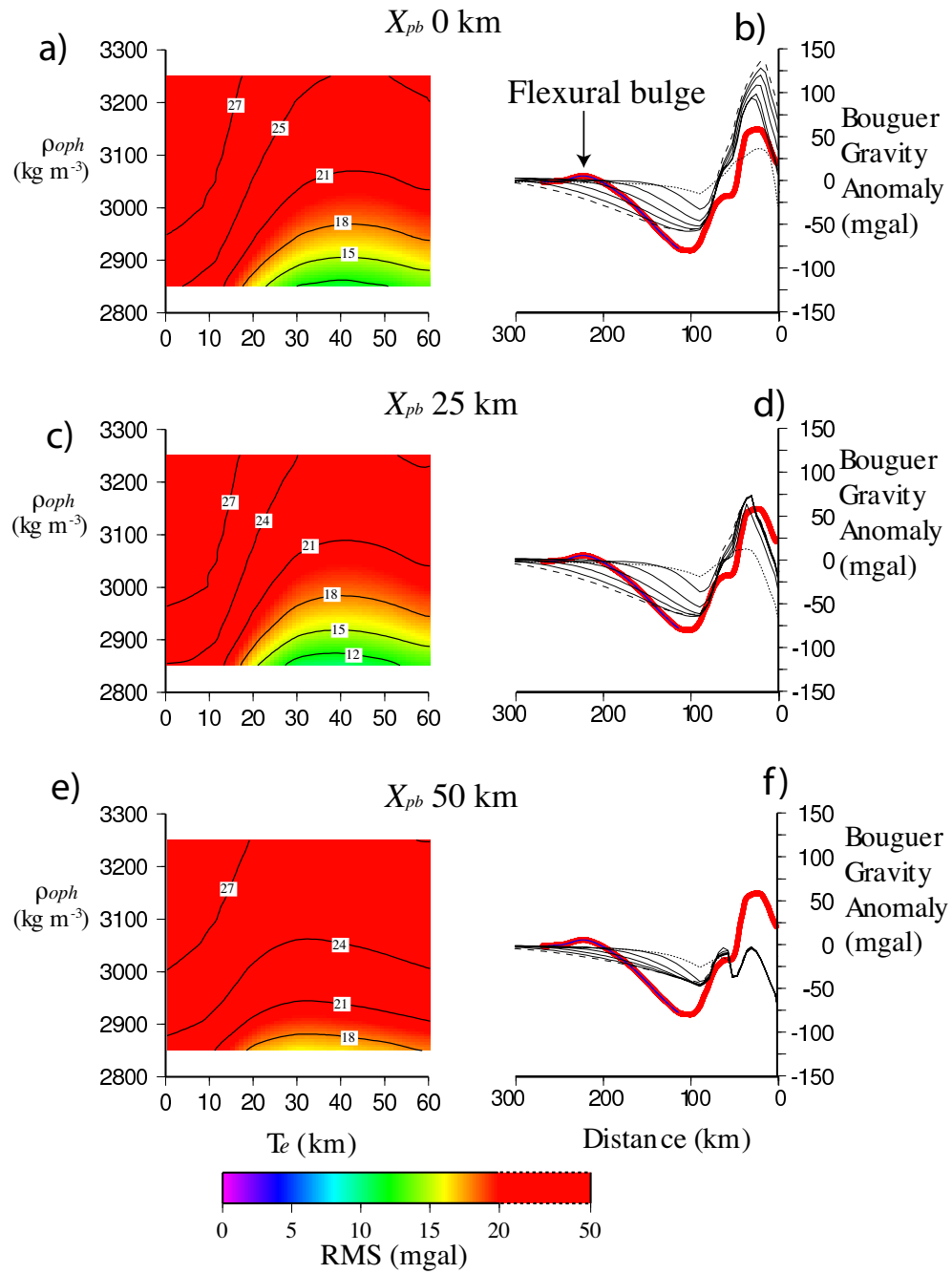


Figure 3.28: Models along profile 3. Panels as in Figure 3.21.

a localised sampling artefact.

For all positions of the plate break considered, Figure 3.28b, d and f shows that the first node, where the flexure, and hence the associated modelled gravity anomaly, becomes positive instead of negative, is well beyond the position seen in the observed Bouguer gravity data. It also appears that the gradient of the observed Bouguer gravity anomaly is much greater than can be explained by an ophiolite acting as the driving load on the end of a simple broken elastic plate.

Two breaks?

The RMS plots shown in Figure 3.28 indicate that loading with a buried ophiolite load, for the full range of densities considered, cannot explain the observed Bouguer gravity anomaly and hence the associated flexure. However, there is a distinctly linear negative gravity anomaly in front of the positive anomaly associated with the ophiolite, which is highly suggestive of a flexural origin for this basin. A solution to this problem may be that the continental crust in this region is not a simple discontinuous or broken plate that can be modelled with a single T_e along its length; rather, it is acting like an elastic beam with a more complex structure where T_e varies along the length of the beam.

If the plate was broken in two locations, beneath the load and close to the bulge crest, then the extent of the flexure would be much larger over the bulge region and the steepness of the foreland flexure would be greater for the same load. A second break beneath the bulge crest, as seen from gravity data, is proposed as a second break or weak zone in the plate will act to ‘trap’ and accentuate the flexural bulge. All deformation will then be concentrated on the load side of this weak zone.

This situation can be modelled in a similar manner to a simple broken plate, by reducing T_e to 0 in the region of the plate break. Figure 3.29 shows this case. The input T_e was reduced to 0 km between 220 and 260 km along the profile as well as at the plate break beneath the load considered previously.

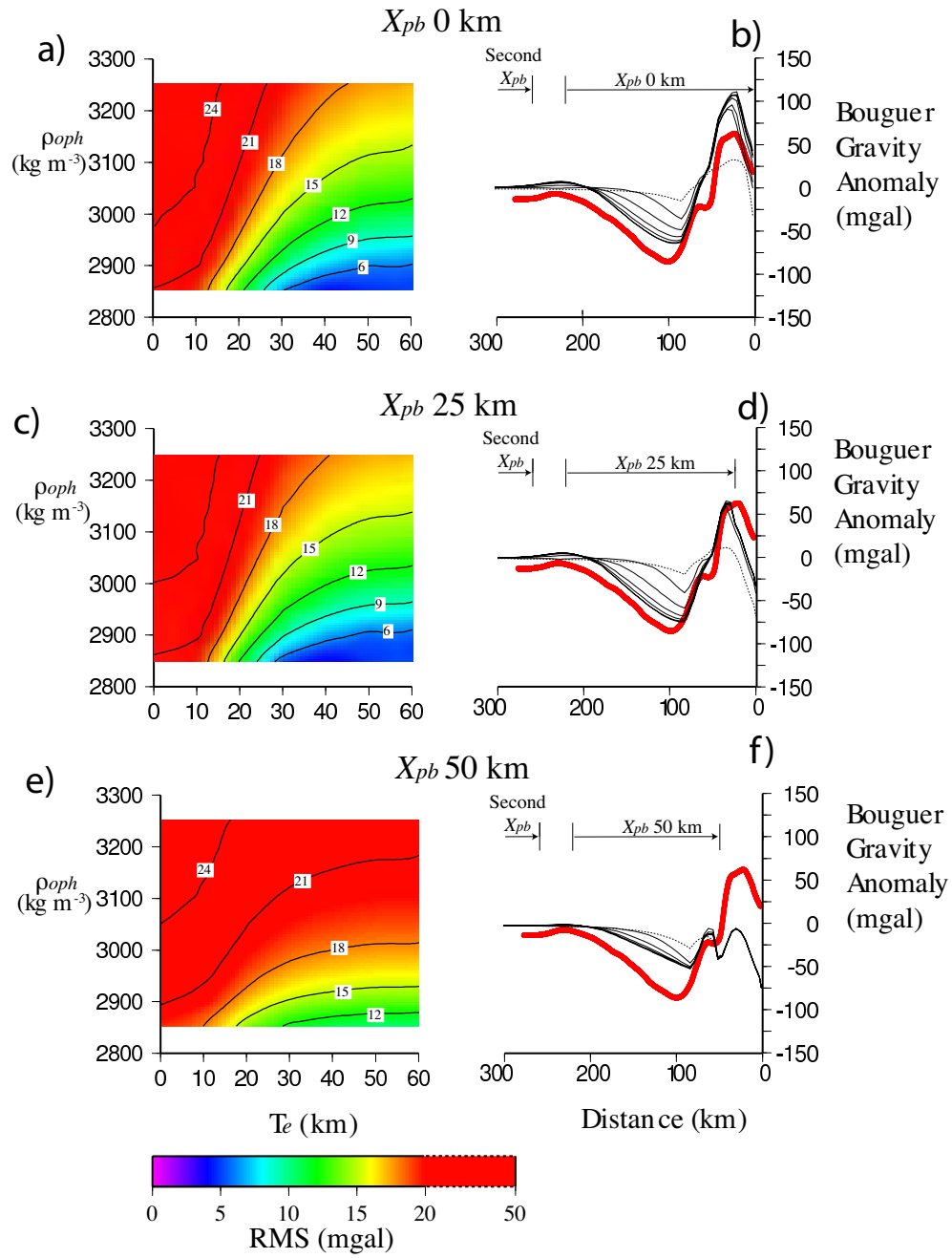


Figure 3.29: Models of flexure along profile 3 assuming two breaks in the T_e structure. a-f are as shown in Figure 3.28.

Figure 3.29a, c and e show significantly lower RMS values than Figure 3.28, although low values for ophiolite density are still preferred.

Figure 3.29b, d and f show the modelled gravity anomalies based on the best fitting load densities shown in a, c and e. There is a better fit to both the amplitude and the position of the Bouguer gravity high proposed to be associated with the flexural bulge. This suggests that the ‘two breaks’ hypothesis is consistent with the observed Bouguer gravity anomaly.

It is also apparent from Figure 3.29 that there are no distinct minima derived for the best fitting T_e value. Instead, from the profiles, it appears that once a critical T_e is reached the plate simply pivots about the second break. This means that there is no upper bound of T_e for this profile. However, T_e must exceed ~ 30 km in the foreland. Otherwise, the flexural effects would not extend far enough to match the observed anomaly.

Why the plate possesses a secondary break in the bulge region is not immediately clear. One explanation would be that the stresses induced in the plate by loading caused it to fail. The flexural bending stresses in a semi-infinite beam (equivalent to a broken plate) that has been loaded by a line end load can be shown to peak half way between the forward edge of the load and the first node (Watts [2001] P108). Figure 3.28 shows the first node for a plate with a single break and a T_e of 30 km lies ~ 150 km from the load front. A T_e of 30 km was assessed to match the best fitting models to the north. The second break, in the bulge region, needs to be at ~ 100 km not 75 km from the load front suggesting that bending stresses in a plate with a T_e of 30 km did not cause it to fail.

Patton & O’Connor [1988] and Warburton *et al.* [1990] argued that for at least parts of the Oman foreland the flexural bulge did not migrate. This can be accounted for by a pre-existing weak zone in the plate. This zone would ‘trap’ the bulge as it migrated ahead of the foreland basin. Given this line of evidence it is proposed that a pre-existing fault or weak zone acted as a secondary break in the

crust, preventing bulge migration and focussed the observed flexure into a narrow region.

3.4.3 Profiles 4, 5 and 6

Figure 3.30 shows a compilation of the profiles across the foreland of the Oman Ophiolite, from the north to the south-east. Topographic and Bouguer gravity anomaly data is shown along with the calculated anomaly based on Airy compensation of the observed topography.

Profiles 1-3 show distinct positive departures from the Airy model close to the mountains, associated with the ophiolite load, and negative departures from the Airy model over the foreland region associated with flexure. This is in line with the models discussed in the previous sections.

However, profiles 4-6 show very different patterns when compared to the Airy isostatic model. Profiles 4 and 5 show a low over the foreland region compared to the anomaly predicted from an Airy isostatic model. However, these low values, compared to the Airy model, continue to the south and do not appear to significantly converge on the Airy model as was seen on profiles 1, 2 and 3. This suggests that a non-flexural process may be biasing the anomaly downward, as where flexure is the only process active in a region, the Bouguer gravity anomaly should converge on zero towards the margin of the foreland (Lin & Watts [2002]). Alternatively very high T_e values in the foreland region could produce this type of anomaly. However, calculated anomalies along profiles 1 and 2 with relatively high T_e values of 60 km showed a steeper gradient than is seen in this region.

3.4.4 Airy isostatic anomaly

An additional way to look at gravity anomaly data is to assess the Airy isostatic anomaly in map view. This allows spatial variations in the extent of flexural compensation to be assessed. The Airy isostatic anomaly is calculated by subtracting

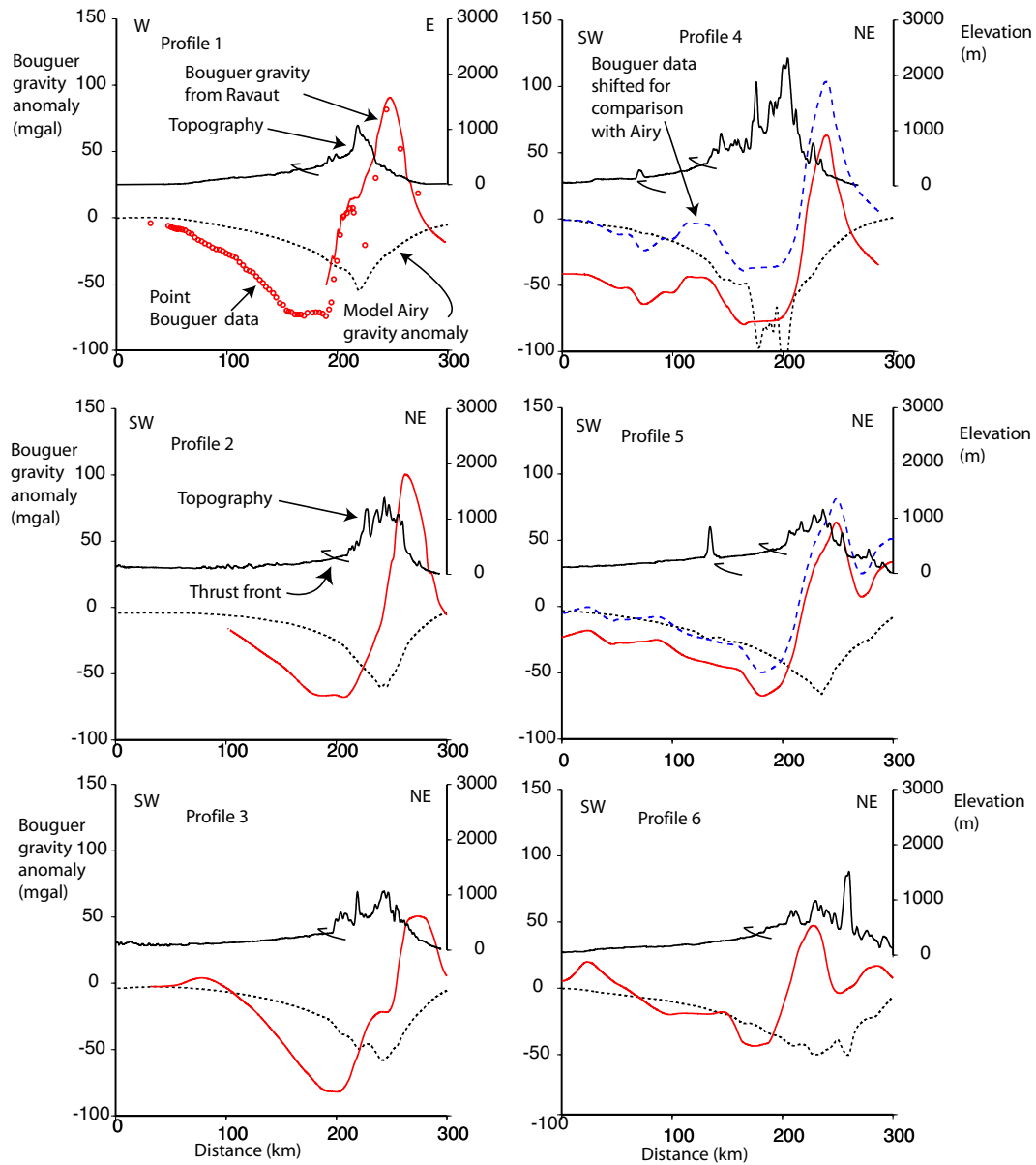


Figure 3.30: Compilation of profiles. Red lines and circles show Bouguer gravity anomaly data corrected for the long wavelength regional field and proximity to the shelf break. Fine dashed black line, model based on Airy compensation of surface topography. Red solid line, Bouguer gravity anomaly. Long blue dashed line, DC shifted Bouguer gravity data to allow comparison with shape of calculated anomaly based on Airy compensation.

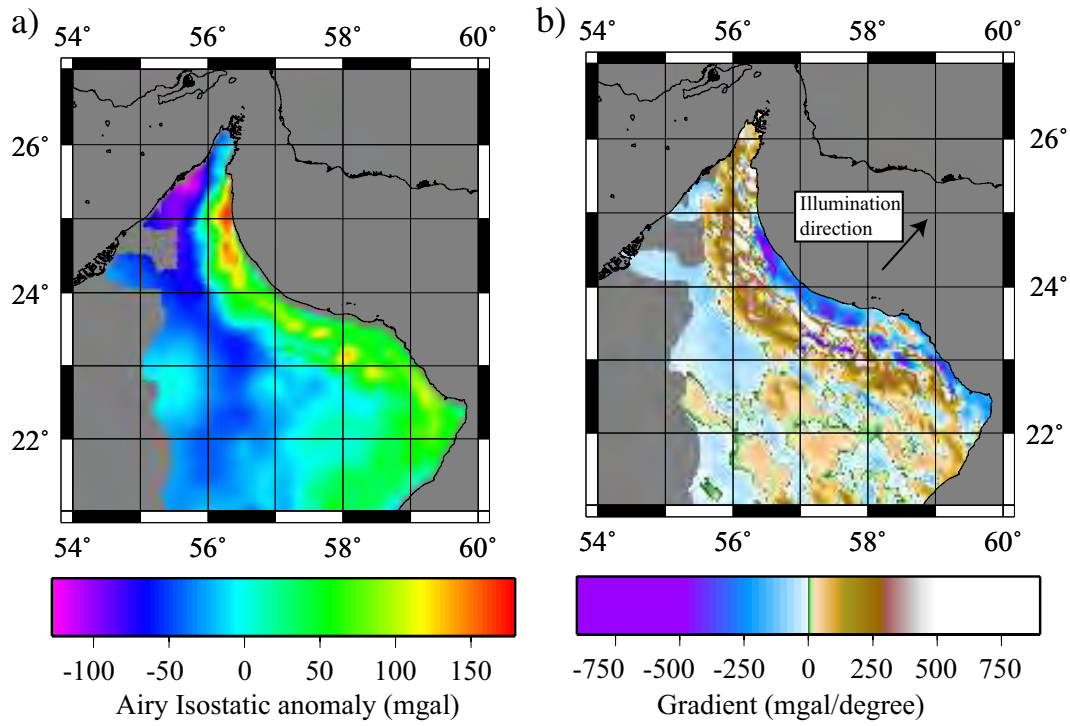


Figure 3.31: Airy isostatic anomaly maps across Oman. a) Airy isostatic anomaly. b) Horizontal gradient of the Airy isostatic anomaly in a north-east direction.

the calculated gravity effect based on Airy compensation of the surface topography, from the observed Bouguer gravity anomaly. The results of this calculation are shown in Figure 3.31.

Figure 3.31a shows the Airy isostatic signal to be strongly negative across the foreland basin in the north. However, the size of the anomaly decreases abruptly south of 23°S and east of 56.5°E. This change is highlighted in Figure 3.31b which shows the gradient of the Airy isostatic anomaly. The gradient was calculated in a direction that is approximately orthogonal to the strike of the ophiolite. Blue values in the foreland region in the north west show that the isostatic anomaly increases towards the mountain front in this region, as would be expected for a flexural foreland basin. The isostatic anomaly associated with the foreland ceases to simply increase towards the mountain front in the south east part of the foreland region. Instead the anomaly appears to undulate, dipping towards and away from

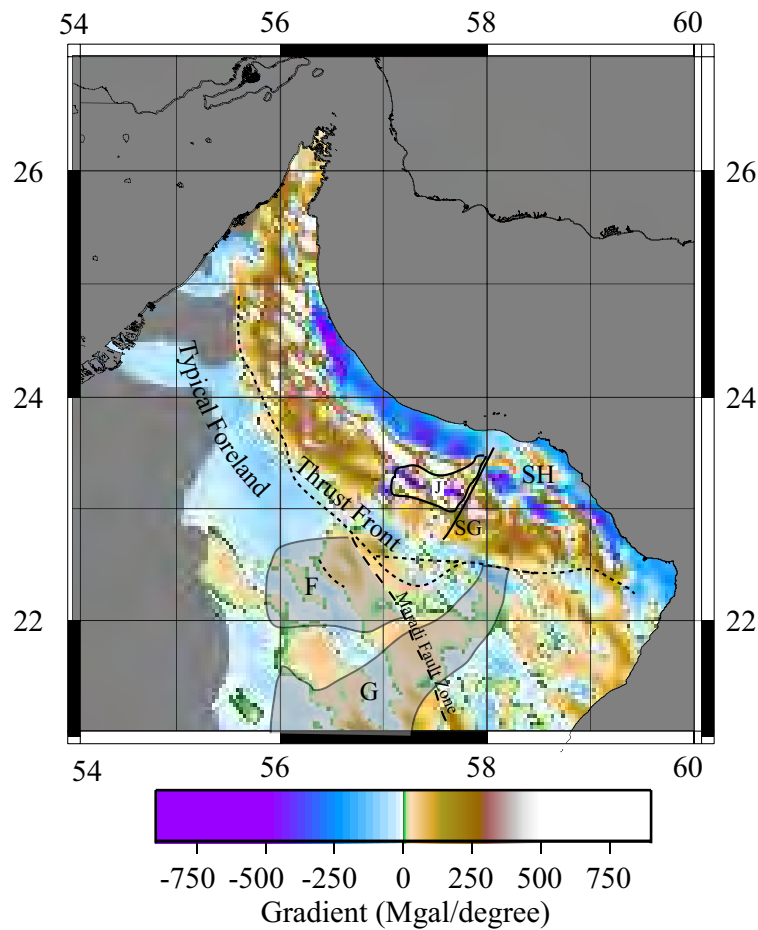


Figure 3.32: Gradient of isostatic anomaly and geological data from Robertson and Searle. G=Ghaba or South Oman salt basin, F=Fahud salt basin, J=Jebel Akhdar, SG=Semail Gap, SH=Sai Hattat.

the mountain front by around the same amount. This strongly suggests that flexure driven by loading to the north east is not an important consideration in this region.

Figure 3.32 shows the change in gradient of the isostatic anomaly from strongly towards the north-east to undulating, to be correlated with the northern margin of the Fahud salt basin. This correlation between a change in the gravity anomaly across the foreland and the edge of the Fahud salt basin strongly suggests that a change in the geology of the foreland has occurred which is affecting the flexural response of the foreland.

Figure 3.33 shows the depth to the top of the Natih unit (~ 89 Ma) taken from Terkin [1999] and unpublished data kindly provided by Hunt Oil Oman. The Natih

underlies the foreland sequence and hence provides a guide to the foreland depth and width. In the north-west of the area a relatively deep, broad, foreland can be seen. However, in the region of the Fahud salt basin the Natih is much shallower and there is no obvious foreland basin. There is also significant evidence of faulting/deformation in this area. Filbrandt *et al.* [2006] has shown that this deformation is related to mid/late Cretaceous normal faulting and oblique deformation. This is seen in the Fahud basin as normal faulting and towards the south east, along the Maradi fault zone, as trans-tensional deformation. These structures have been reactivated in the Tertiary to generate compressional fold structures. Filbrandt *et al.* [2006] suggests that there are predominantly trans-tensional faulting patterns orientated NW/SE seen in 3D seismic data across both the Fahud and Gahaba salt basins at the time of ophiolite obduction. These, they suggested, relate to the Campanian tectonic event associated with the emplacement of the Maserah Ophiolite (See Maserah Island in Figure 3.26) over 300 km to the south-east.

The cause of the change in foreland shape and observed gravity anomaly pattern along the mountain front, could relate to a number of factors. An initial impression would suggest that T_e in the foreland region had fallen dramatically and all topography is now being compensated by Airy isostasy, however, the low topographic gradient and low elevations in the foreland region make this a difficult model to prove as very high rigidities would give a similar result. Airy support of the topography also fails to explain the presence of large positive Bouguer gravity anomalies representing loads within the crust. If these loads were supported by displacement of the Moho the positive anomaly from the ophiolite would be largely balanced by a negative anomaly of approximately the same magnitude as seen in Figure 3.19. However, crustal thinning could counter this as suggested for profiles 1 to 3.

A second interpretation would be that the crust is not weak but instead extremely rigid in this region, despite the presence of the Maradi fault zone. The

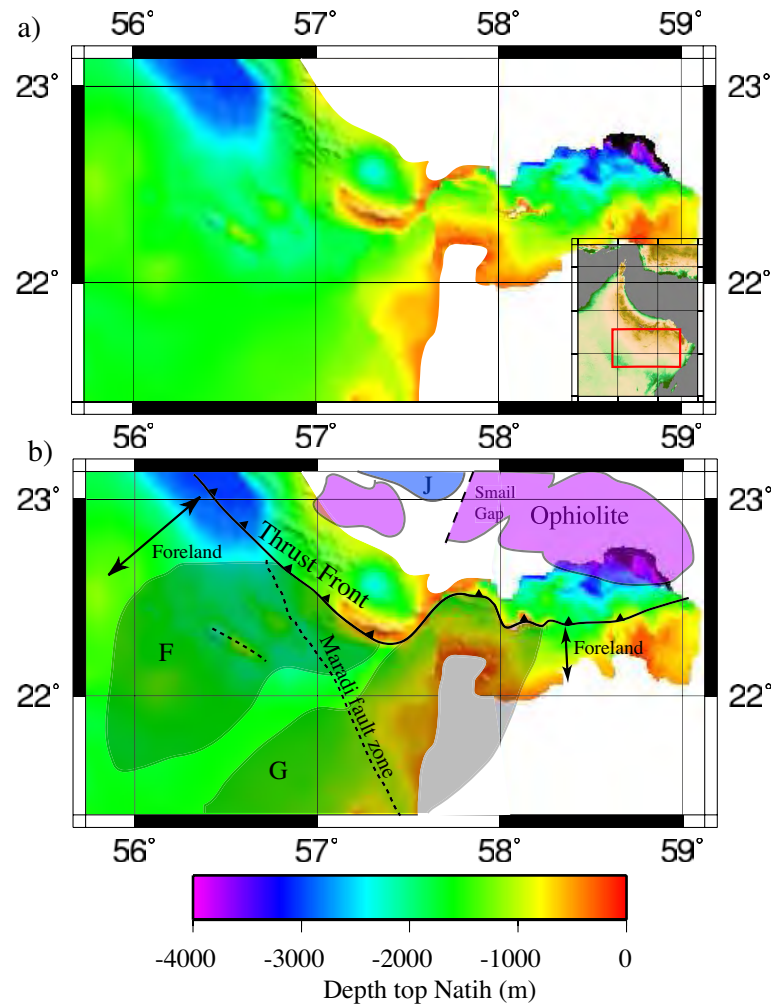


Figure 3.33: Depth to the top of the Natih formation from Terkin [1999] and Hunt Oil Oman. a) Isopach data, b) Structural elements overlaid on isopach data showing broader foreland in the north-west and narrower foreland in the south-east.

ophiolite load would then simply not be compensated by a mantle root. Models of the profiles to the north, such as that shown in Figure 3.23, show that for continental crust with a relatively high T_e of 60 km a significant flexural signal would still exist. Profile 6 is close to the south-eastern end of the ophiolite so a 3D model may show less deflection in the foreland region given high T_e plates, however, profiles 4 and 5 are close to the centre of the ophiolite load so the profile approximation is valid.

Given the proposed load thickness of approximately 10 km of ophiolitic material, it is highly unlikely that no flexural signal would be generated. Also, Figure 3.1 and Figure 3.32 show these areas to over-lie salt formations which are thought to have been deposited in a rifted environment and hence on stretched continental crust (Husseini [1988]). Areas of stretched continental crust are typically associated with low rigidities (Watts [1992]).

When the geological structures are plotted on the map of the gradient of the Airy isostatic anomaly as in Figure 3.32 an additional interpretation is possible. In the region towards the north-west end of the mountain range where the ‘typical foreland’ signal is seen, the Airy isostatic anomaly over the ophiolite has a single peak. Where the ‘abnormal’ signal is seen, towards the south-east, the anomaly over the load is split into two peaks. This can be seen in profiles 5 and 6 of Figure 3.30 and by the inversion of the Airy isostatic gradient in the region of Jebel Akhdar in Figure 3.32. This splitting of the ophiolite is significant in that it suggests re-activation since the initial emplacement of the ophiolite.

In the Jebel Akhdar and Saih Hatat regions the basement sequences, over which the ophiolite was emplaced, are exposed (Coffield [1990], Cawood *et al.* [1990]). The Jebel Akhdar window is interpreted to be due to out of sequence thrusting within the basement, after initial emplacement of the ophiolite. It has been suggested by several authors that the dramatic thickening and uplift of the crust in these areas would have caused gravitationally driven normal faulting on their flanks and reac-

tivation of thrusting in the foreland in a process of culmination collapse (Coffield [1990], Cawood *et al.* [1990]).

As this culmination collapse propagated into the existing foreland basement involvement in thrusting may have occurred. This would act to obscure the flexural signal of the foreland basin. If the ophiolite was effectively regionally isostatically compensated, by flexure of the lithosphere and generation of a foreland, movement into and deformation of the foreland would not create great amounts of additional subsidence in the foreland region. However, as Figure 3.24 shows, a significant part of the flexural signal is derived from deflection of the mantle. This would not be expected to be reduced by thin-skinned thrusting into the foreland so this reactivation can not account for the entire discrepancy.

Searle *et al.* [2004] has suggested that reactivation of faulting at the base of the ophiolite during the culmination must be relatively minor. They suggest instead that the uplift relates to predominantly vertical movement of crustal material up and through the ophiolite. In this interpretation the driving mechanism for uplift could have been ‘excessive’ subduction towards the north-east of buoyant continental material. This would have been dragged down the subduction zone by a dense down-going slab (Searle *et al.* [2004], Warren *et al.* [2005]). When this slab broke off, the down-going buoyant material would return up the subduction zone and cause uplift of the ophiolite from beneath. This has been inferred as the most likely explanation for the high pressure eclogitic rocks found on the coast in the Saih Hatat window area (Searle *et al.* [2004], Warren *et al.* [2005]).

If obduction in this region was driven predominantly by the pull of a dense down-going slab, the significant north-east/south-west strike slip activity seen in the foreland by Filbrandt *et al.* [2006] can be explained as a response to this. If the plate was in tension from the down-going slab then far field forces from the docking of the Masirah Ophiolite along the eastern coast of Oman could be felt in place of compressional ones from the North Oman Ophiolite. This strike slip activity

would significantly alter the way the foreland developed. It would not be until the slab finally broke away, towards the end of the emplacement episode (Warren *et al.* [2005]) that the continental margin would be truly flexurally loaded by the full weight of the North Oman Ophiolite. If significant amounts of buoyant material were at this time returning up the subduction zone the ophiolite load would be supported by this, so no significant foreland basin would develop.

Isopach data shown in Figure 3.33 shows that a foreland did develop in front of parts of the ophiolite in the far south-east. However, compared to the basin further to the north it is narrow, only ~ 50 km wide. This narrow basin correlates with the region of the Airy isostatic anomaly which can be seen to be dipping steeply towards the mountain front at the north-east end of the Gahaba salt basin in Figure 3.32. Figure 3.33 shows that the region to the far south-east has undergone some amount of regional uplift relative to the region to the north-west as the Natih formation is exposed at the surface. Filbrandt *et al.* [2006] suggests there has been between 500 and 1000 m of uplift in this area based on seismic velocity anomalies and Searle *et al.* [2004] proposes over 2000 m of uplift in the mountains. This uplift could have lead to erosive removal of much of the foreland. However, seismic reflection data provided by Hunt Oil Oman shown in Figure 3.34 highlights the fact that the foreland material remains approximately horizontal, despite the uplift. This suggests that any uplift must have been regionally uniform so the foreland shape should be preserved. The seismic data also shows that sediments belonging to the continental passive-margin sequence below the foreland are approximately horizontal south of the exposure point of the base of the foreland, suggesting the foreland width shown by the isopach map to be approximately correct.

The seismic data shown in Figure 3.34 is approximately coincident with Profile 6 and shows the dip of the base of the foreland to be constant in this region. This suggests hinge-like behaviour as was seen in profile 3 rather than a simple flexural basin. The slight negative Bouguer anomaly compared with the Airy model close to

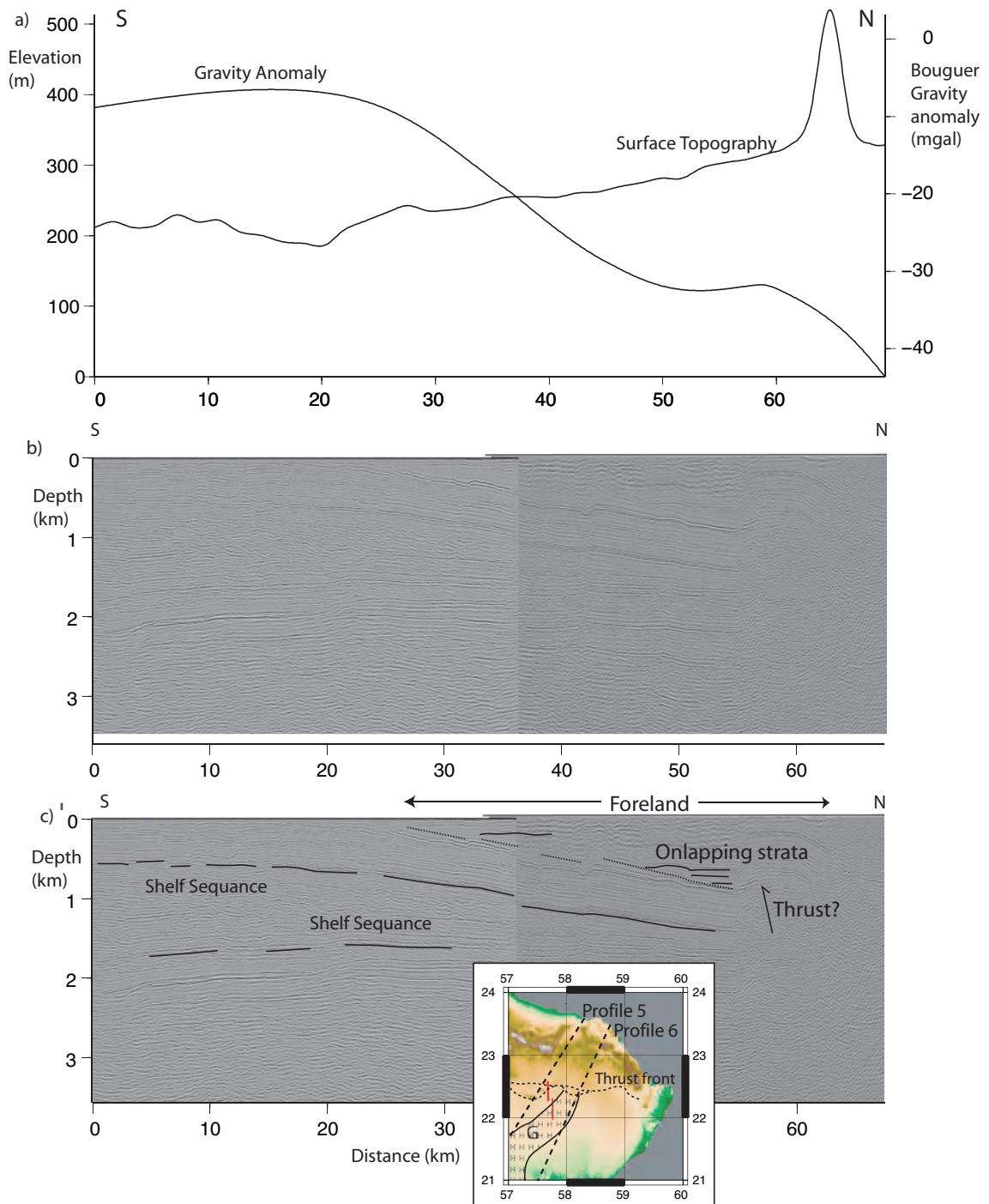


Figure 3.34: Compilation of two seismic lines provided by Hunt Oil Oman. a) Surface topography and Bouguer gravity anomaly data. b) Un-interpreted seismic line. c) Interpreted seismic data, black lines highlight traceable surfaces. Inset shows location of seismic profiles in red and modelled gravity profiles in black, also shown are the location of the Fahud and Ghaba (G) salt basins.

the mountains seen in Profile 6 is the gravity anomaly associated with this flexure.

3.5 Oman summary

Flexural modelling of the north-eastern continental margin of Oman and the U.A.E. has been carried out assuming deflection of an elastic plate, broken at one end, driven by loading of the existing passive-margin by the North Oman Ophiolite. These flexural models allowed the calculation of gravity anomalies based on two interfaces, one between the sediments infilling the flexural foreland and the crust and a second interface at the crust mantle interface. These models explain well the width and magnitude of the observed Bouguer gravity anomaly towards the northern end of the ophiolite. The crustal rigidity as represented by T_e , which provides the best fitting model to the Bouguer anomaly data, is of the order of 30 km over the north-western part of the foreland of the Oman Ophiolite (Profiles 1 and 2).

To match the positive part of the Bouguer gravity anomaly beneath the ophiolite 14 km of crustal thinning was needed which, for an initial crustal thickness of 35 km equates to a β factor of ~ 1.6 which is similar to other values of 1.13 to 1.3 recovered for the Tethyan margin by Corfield *et al.* [2005] and Wooler *et al.* [1992].

Towards the central part of the ophiolite (Profile 3) the observed Bouguer gravity anomaly can only be modelled as flexural if the continental crust has a second weak zone close to the observed southern edge of the foreland Bouguer gravity anomaly. The most probable cause of this weak zone is a pre-existing fault or fracture system within the basement.

Within the Fahud Salt basin no clear flexurally derived gravity anomaly or flexural basin can be seen. This, it is proposed, relates to the presence of a dense down-going slab off-shore of this area. This slab allowed trans-tensional deformation within the foreland region until it broke off. The buoyant return of material from the subduction zone then acted to support the load and inhibit flexural sub-

sidence.

In the Gahba salt basin region at the south-eastern end of the ophiolite there is a foreland basin but it is relatively narrow compared to the north-east and poorly defined in the gravity anomaly data. The dip of the foreland in this area from seismic data appears constant, which strongly suggests hinge-like rather than true flexural behaviour in this region.

Overall the rigidity structure of the Arabian plate loaded by the Oman ophiolite appears to vary from relatively simple and constant in the north, to more complex in the south east. This variation appears to correlate with a change in the basement of the foreland and the presence of culmination structures within the mountains.

T_e and the India-Eurasia collisional system

4.1 Introduction

The findings presented in this chapter has been previously presented in the paper of Jordan & Watts [2005]. Flexural rigidity (T_e), as discussed in Chapter 1, provides a measure of the long-term strength of the lithosphere. Previous studies suggest continental T_e is in the range of 5 to 125 km (see summary in Watts [2001]), with the highest values being associated with cratonic shields and the lowest values with extensional rift-type basins. Recently, McKenzie & Fairhead [1997] have questioned the validity of continental T_e values of > 25 km, especially those based on the Bouguer coherence spectral technique (e.g. Bechtel *et al.* [1990]; Zuber *et al.* [1989]; Ussami *et al.* [1993]). The controversy over high values of T_e has focussed on the India - Eurasia collisional system in the region of the Himalayan foreland. Lyon-Caen & Molnar [1983] and Karner & Watts [1983], for example, used forward modelling techniques to show that the Bouguer gravity anomaly over the Ganges basin could be explained by the flexure of the Indian continental lithosphere. They modelled the flexure by applying forces that were representative of the load of the Himalaya and Tibetan plateau on the end of an elastic beam that overlies an inviscid fluid. By comparing the observed gravity data to model predictions

they showed that T_e of the continental lithosphere was in the range 80 - 110 km (Lyon-Caen & Molnar [1983]; Karner & Watts [1983]). McKenzie & Fairhead [1997], however, challenged these high values using spectral estimates of the free-air admittance and a non-spectral shape-fitting technique along profiles of free-air gravity anomaly data. They concluded that the T_e of Indian lithosphere was low (24 km by admittance and 42 km by shape-fitting) and significantly less than that obtained by previous workers.

The low values derived by McKenzie & Fairhead [1997] are of the order of the thickness of the seismogenic layer, T_s , which is the depth range over which earthquakes occur (Maggi *et al.* [2000a]). This led Jackson [2002b,a] to question earlier rheological models of a strong upper crust, weak lower crust and strong mantle proposed, for example, by Chen & Molnar [1983]. They proposed instead a single strong crustal layer that 'floated' on an underlying weak, non-seismogenic, mantle.

The inference that a lack of earthquakes implies a weak mantle has subsequently been questioned. Watts & Burov [2003], for example, suggest that tectonic stresses (including those associated with flexure) are not usually large enough to exceed the brittle strength of the sub-crustal mantle and generate earthquakes. Handy & Brun [2003] also dispute the link between the presence or absence of earthquakes and the long-term strength of the lithosphere. Based on structural and laboratory-based studies of the deep-structure of orogens and rifted margins, they argue that the mantle must be strong. These studies suggest therefore that rather than being equal to T_s , T_e may very well exceed it.

Since the study of McKenzie & Fairhead [1997], there have been a number of forward modelling (e.g. Cattin *et al.* [2001]) and spectral (e.g. Tiwari & Mishra [1999]; Stephen *et al.* [2003]; Rajesh & Mishra [2003]; Rajesh *et al.* [2003]; Rajesh & Mishra [2004]) studies of the India-Eurasia collisional system. These yield T_e values in the range 10-50 km. Forward modelling along profiles (Cattin *et al.* [2001])

yield a single estimate of T_e as do spectral methods based on profiles (Tiwari & Mishra [1999]) and windows of fixed sizes (Stephen *et al.* [2003]; Rajesh & Mishra [2003]; Rajesh *et al.* [2003]). Only Rajesh & Mishra [2004] directly considered the possibility of spatial variations in rigidity by calculating the spectral estimates in multiple, overlapping, windows. Rather than considering the actual values of T_e , however, Rajesh & Mishra [2004] expressed the variations in terms of spatial changes in the Bouguer coherence transitional wavelength.

Spectral studies of continent-wide gravity anomaly and topography data elsewhere (e.g. North America - Lowry & Smith [1994]; Wang & Mareschal [1999]; Flück *et al.* [2003], South America - Ussami *et al.* [1993], Africa - Hartley *et al.* [1996], Fennoscandia - Pérez-Gussinyé *et al.* [2004]) suggest that T_e of cratonic regions is high and $>\sim 60$ km. However, as Pérez-Gussinyé *et al.* [2004] have recently shown, there are difficulties with the spectral estimation of T_e . One problem is that the ability to recover high T_e is dependant on window size. If the region of high T_e is of small areal extent, relative to the window, a lower average T_e may be recovered. Smaller window sizes will improve the resolution, but may not be large enough to recover the maximum value of T_e .

An alternative approach, which avoids some of the problems with spectral methods, has been suggested by Braitenberg *et al.* [2002]. These workers use the surface topographic load to calculate the flexure of the top and bottom of the crust for a range of uniform T_e values. The T_e which gives the best fit between an observed Moho, derived from gravity and seismic data, and the flexed Moho within a window is then assigned to its centre. By moving the window step-wise across the Tibetan plateau, Braitenberg *et al.* [2002] recovered a T_e structure that varied from 8 to 110 km over horizontal distances of less than 250 km. They found that the lowest values of T_e occurred in the centre of the plateau and the highest over flanking regions.

In this study both profile- and grid-based flexural and gravity modelling tech-

niques were used to determine the T_e structure of the India-Eurasia collisional system. The study is carried out in 3 main steps. First, profile based forward modelling techniques, introduced in Section 2.6.1, were used to show that the centre of the Himalaya foreland has a high T_e (≈ 70 km) while regions to the east and west have lower values (≈ 30 -60 km). Then the 2D iterative technique, introduced in Section 2.7, which synthetic modelling shows is capable of recovering a high resolution T_e structure from gravity anomaly and topography data, is used to verify these T_e variations. Finally, we examine the implications of the T_e structure for the terrane structure of peninsula India and for tectonic processes in the Tibetan plateau and surrounding regions.

4.2 Geological setting

The India-Eurasia plate boundary is one of the best-known examples of continent-continent collision on Earth. The collisional process has created the Himalayan mountains with many peaks over 7000 m and the Tibetan plateau which, in places, is over 1000 km wide and 3000 m high. The collision culminated during the Eocene (~ 50 Ma) when the Indian plate underthrust the southern margin of Eurasia. Paleomagnetic evidence suggests that during the Cainozoic, India 'indented' Eurasia by as much as ~ 2000 km (Windley [1988]). In the process, the crust has thickened to ~ 70 km below the Tibetan plateau and deformation has been distributed across a 1500 km wide region to the north of the Indian plate.

4.3 Gravity anomaly and topography profiles

Initially a series of profiles through the Himalayan mountain front were considered for analysis using the 1D modelling techniques introduced in Section 2.2. In Chapter 3 it was found that a significant proportion of the driving load in the Oman Mountains was buried. When buried loads play a significant role in driv-

ing flexure, such as in the Appalachians, Western Alps and Taiwan (Karner & Watts [1983], Stewart & Watts [1997], Lin & Watts [2002]) and Oman, a clear positive negative couple is apparent in the Bouguer gravity anomaly, with the positive appearing beneath the topography where a negative would be expected from an Airy model. However, such a ‘couple’ is not observed in the India - Eurasia collisional system (Karner & Watts [1983]; Lyon-Caen & Molnar [1983]; Caporali [1995]; Cattin *et al.* [2001]). It is therefore concluded that sub-surface loads do not significantly contribute to the gravity field in this region. This conclusion allows the modelling of the observed gravity anomalies due to flexure caused by a ‘driving load’ consisting solely of the present day topography.

Figure 4.1 compares the observed Bouguer gravity anomaly along 3 profiles of the eastern, central and western Himalaya foreland to calculated profiles. The observed profiles were constructed from a grid of Bouguer gravity and topography data compiled by GETECH (UK) as part of their South-East Asia Gravity Project (SEAGP). Bouguer gravity data was provided as 2.5 x 2.5 min average values which we have gridded using **GMT** function **surface** with a tension, T , of 0.25 and a grid spacing of 5 minutes. The Bouguer anomaly data is terrain corrected out to 167 km and has had a long wavelength satellite-derived gravity field, OSU-91, complete to degree and order 20 removed from it (Rapp & Pavlis [1990]). This field was chosen because it is dominated by the Indian Ocean gravity ‘low’ which most workers attribute to deep processes in the sub-lithospheric mantle. The discrepancy between OSU-91 and the more recent EGM-95 field (Lemoine *et al.* [1998]) is less than ± 2 mgal across the entire region, therefore using the older model will have little or no impact on the result. The calculated profiles assume that surface topographic loads in the Himalayas have flexed one end of a broken elastic plate. By calculating the RMS difference between observed and calculated Bouguer anomalies (Figure 4.2), both a best fit position of the plate break with respect to the mountain front, and a best T_e value, is estimated for each of the

profiles. Figure 4.1a-c show the topography and gravity anomalies stacked with respect the best fit plate break. The central profile (Figure 4.1b) reveals a best fit T_e of 70 km. Figure 4.2 shows the technique also recovers the position of the plate break. Figure 4.2b highlights the fact that no value of $T_e < 70$ km can account for the central profile, irrespective the plate break location. The eastern and western profiles, however, show significantly lower values of 50 and 30 km respectively (Figure 4.1a and c).

The results in Figure 4.1 are in accord with previous studies. In particular, they agree with the high T_e values reported by Lyon-Caen & Molnar [1983] and Karner & Watts [1983] for the central Himalaya foreland. They also confirm the earlier suggestion of Lyon-Caen & Molnar [1985] that T_e decreases to the west and east. Although the values recovered by the current study disagree with those recovered by Caporali [1995] in the west, based on end-conditioning loads, they agree well with those of Cattin *et al.* [2001] in the east who used actual topographic loads.

4.4 2D T_e variation

The 2D iterative method introduced in Chapter 2.7 was implemented to further constrain the T_e structure of the India-Eurasia collisional system, both in the foreland and in the surrounding regions. Figure 4.3e shows the recovered T_e together with the observed (Figure 4.3a) and calculated (Figure 4.3b) Bouguer gravity anomaly. The figure shows that T_e varies widely across the region, with recovered values ranging from 125 km in the Himalayan foreland to nearly 0 km in the Tibetan plateau. There is an excellent agreement between the observed Bouguer anomaly and the calculated anomaly based on the T_e structure.

A useful way to evaluate the role of flexure is by consideration of the isostatic gravity anomaly. Firstly the Airy isostatic anomaly is considered, this is defined as the difference between the observed Bouguer gravity anomaly and the gravity effect of the Airy-type compensation. This scheme of compensation considers the

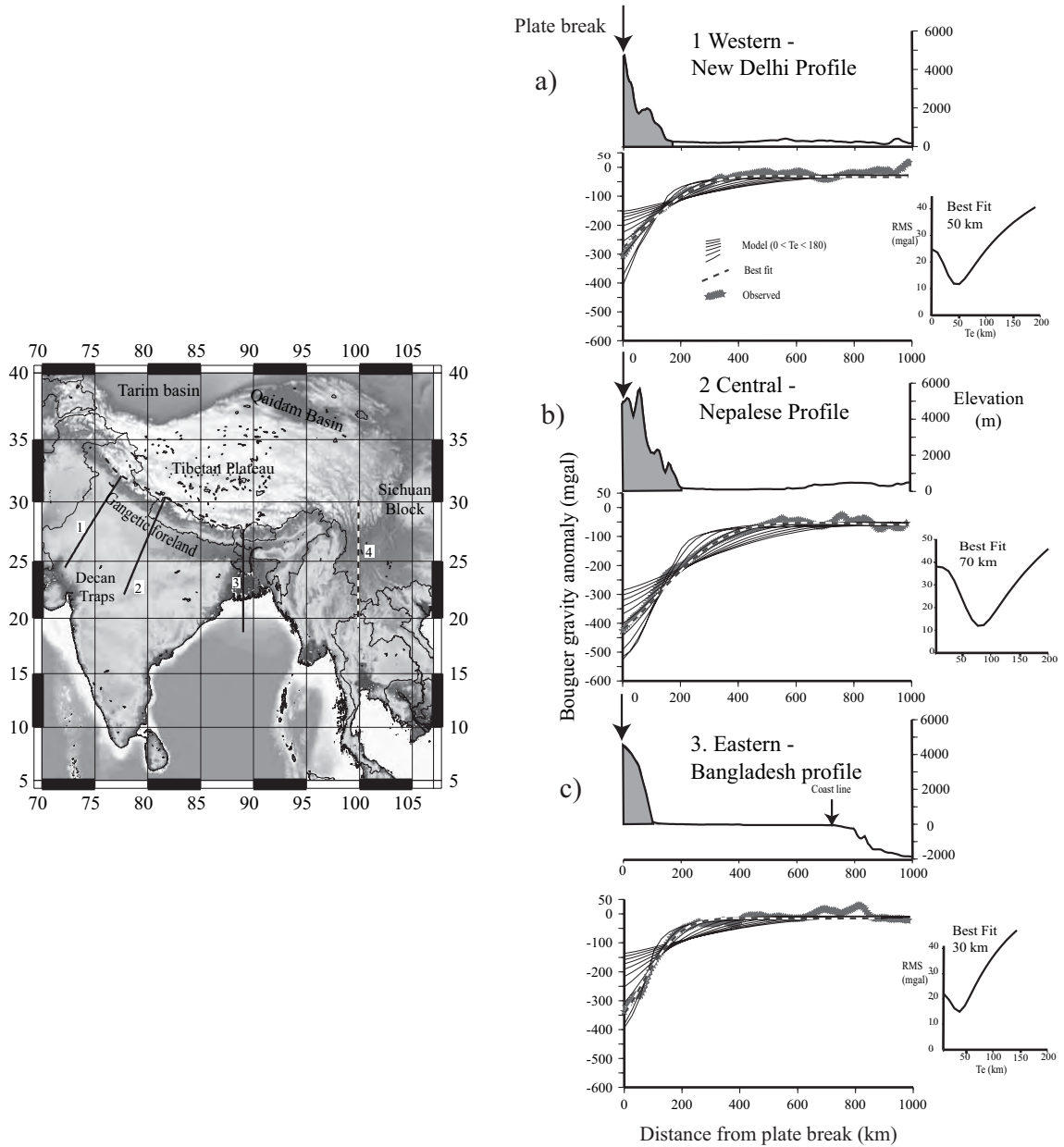


Figure 4.1: Location map and observed and calculated Bouguer gravity anomaly profiles of the Himalaya foreland region. The map is based on a GETECH '5 x 5' minute grid SEAGP [1991–1995]. Thick solid lines show the location of the eastern, central and western profiles. Thin dashed line shows the location of the additional profile in Figure 4.5. Thick dashed line shows the position of the 'best fit' plate break. a) Western profile. The upper profile shows topography. The lower profiles show the observed Bouguer anomaly (thick grey line) and calculated Bouguer anomaly profiles (thin solid line) for $0 < T_e < 180$. The 'best fit' profile is highlighted as a dashed line. The inset shows a plot of the RMS difference between observed and calculated gravity for different T_e . b) Central profile, profiles as in a). c) Eastern profile, profiles as in a).

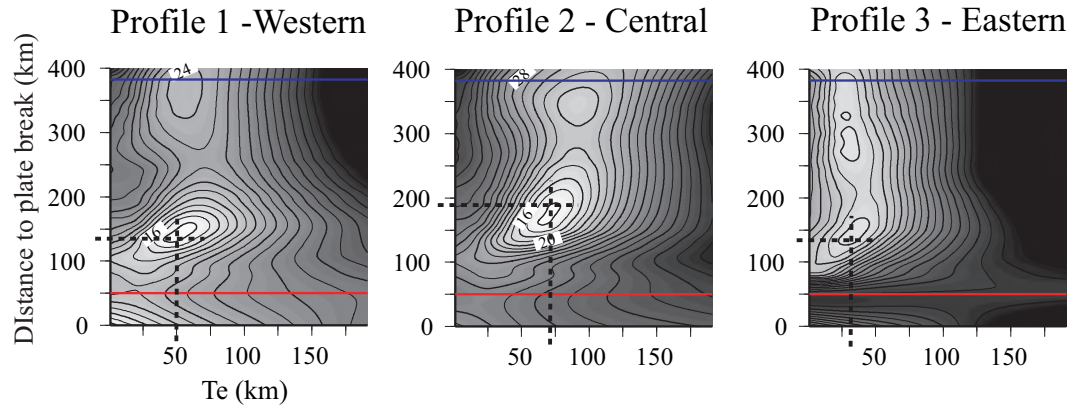


Figure 4.2: Plots of RMS error between observed and calculated Bouguer gravity anomalies for the western, central and eastern profiles (Figure 4.1). The thick dashed lines show the 'best fit' recovered values for the T_e and distance to the plate break from the mountain front. The western and central profiles show a well-defined minima and, hence, both the T_e and plate break are recovered along these profiles. The Eastern profile does not show a single well-defined minima, which we attribute to the fact that at low T_e the flexure is relatively insensitive to the width of the load.

topography as a load on the surface of an elastic plate with a T_e of 0 km. If all the topography in a region is locally compensated, then the Airy isostatic anomaly should be nearly zero. Alternatively, if the topography is flexurally compensated then a distinctive pattern of Airy isostatic anomalies should be seen.

Figure 4.3c shows that Airy isostatic anomalies are generally subdued, especially over the Indian peninsula, central Tibet, and Burma and south-west China. The most striking feature of the figure is the positive-negative 'couple' that correlates with the Himalayan mountain belt and its flanking foreland basin. The negative part of the 'couple' occurs over the foreland basin, suggesting that the Moho is deeper here than is predicted by Airy. The positive occurs over the Himalayan mountains, suggesting that the Moho is shallower here than is predicted by Airy. This 'couple' is attributed to flexure of the Indian lithosphere by the topographic loads of southern Tibet and the Himalaya. There is evidence from the amplitude of the 'couple' that the role of flexure decreases from the central part of the mountain

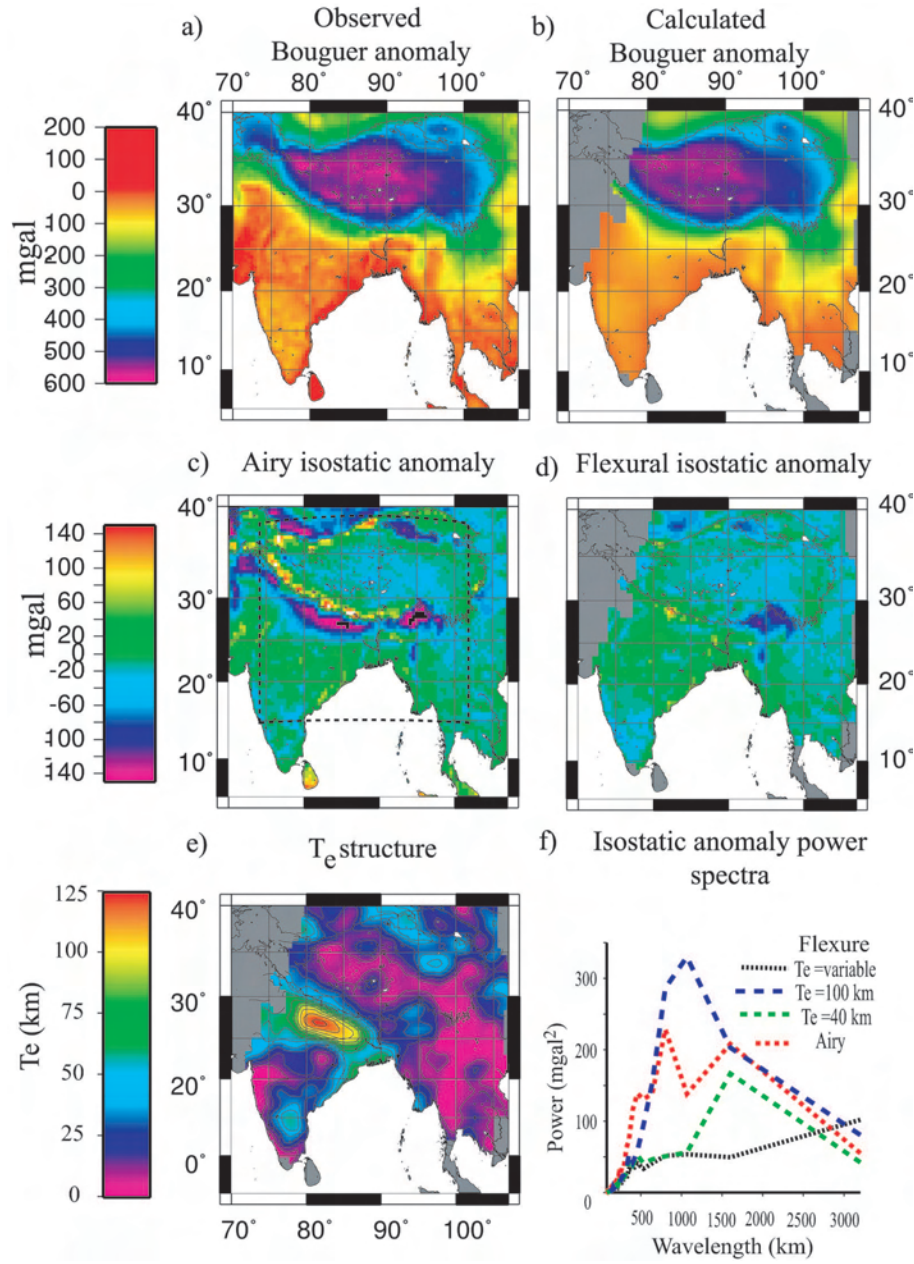


Figure 4.3: Bouguer anomalies, isostatic anomalies, and the recovered T_e structure in the India-Eurasia collisional system. a) Observed Bouguer gravity anomaly from the SEAGP [1991–1995] data base. b) Calculated Bouguer gravity anomaly (stabilised regions only) after 30 iterations of the model. c) Airy isostatic gravity anomaly, calculated by subtracting the gravity effect of the Airy (i.e. $T_e = 0$ km) compensation from the Bouguer anomaly. Dashed box outlines the analysis region for the power spectral plots in Figure 4.3f. d) Flexural isostatic gravity anomaly, calculated by subtracting the gravity effect of the flexural compensation from the Bouguer anomaly. e) Recovered T_e structure obtained after 30 iterations. Contour interval = 10 km. Output T_e structures were smoothed using a 350 km Gaussian filter. f) Power spectra of the Airy, $T_e = 40$ km, $T_e = 100$ km and variable T_e flexural isostatic anomalies. The isostatic anomaly based on the variable T_e spectra shows the least power. Parameters as in Chapter 3, Table 2.1.

front to the west and east. Northern and eastern Tibet also show a ‘couple’, although it is of smaller amplitude. Interestingly, the south-eastern corner of the plateau lacks a ‘couple’, suggesting this region is in Airy isostatic compensation.

Next the flexural isostatic anomaly was considered, this is defined as the difference between the observed Bouguer gravity anomaly and the gravity effect of the compensation based on the recovered T_e structure in Figure 4.3b. The flexural isostatic anomaly (Figure 4.3d) is generally of smaller amplitude than the Airy isostatic anomaly (Figure 4.3c), which is indicative of a regional rather than local-type compensation. The positive-negative ‘couple’ so visible in the Airy isostatic anomaly map, for example, is now absent.

The role of flexure is particularly well illustrated in power spectra plots of the different isostatic anomalies. Figure 4.3f shows, for example, power spectra plots for a central rectangular region of the study area (dashed line, Figure 4.3c). The plots show that the recovered T_e structure significantly reduces the power of the isostatic anomaly compared to Airy (i.e. $T_e = 0$ km) and uniform T_e values of 40 and 100 km. These considerations indicate that a spatially varying T_e describes well the state of isostasy in the region.

The main departures in the flexural isostatic anomaly in Figure 4.3d are limited to the Qaidam Basin, north of the Tibetan Plateau, and to the eastern syntaxis of the Himalayas. The Qaidam Basin is underlain by 7-10 km (Bally *et al.* [1990]) of alluvial plain sediments and it is likely that these sediments account for the discrepancy. The origin of the discrepancy at the eastern syntaxis, however, is not as clear. The region, which shows high uplift rates (3 to 5 mm/yr - Burg *et al.* [1997]), is associated with oblique convergence and strike-slip faults and we speculate that the discrepancy is in some way the consequence of this tectonic complexity.

The reliability of the T_e structure in Figure 4.3e can be assessed by consideration of its stability and sensitivity. The stability, which is defined as the change in T_e

structure over successive iterations, is shown in Figure 4.4a. The white areas in the figure show regions of convergence where little further change in T_e occurs with subsequent iteration. The black and white dashed line in Figure 4.4a shows the regions where the model appears to be stable. These include most of peninsula India, the Tibetan Plateau, and south-west China and Burma. The sensitivity, which is defined as the deviation in the Bouguer anomaly due to small shifts of the T_e structure, is shown in Figure 4.4b. The figure shows high sensitivity around the edges of the Tibetan Plateau, suggesting that T_e is well resolved in these regions. Although the sensitivity is low over peninsula India, it is non-zero and therefore some estimate of T_e can be recovered even in this topographically subdued region.

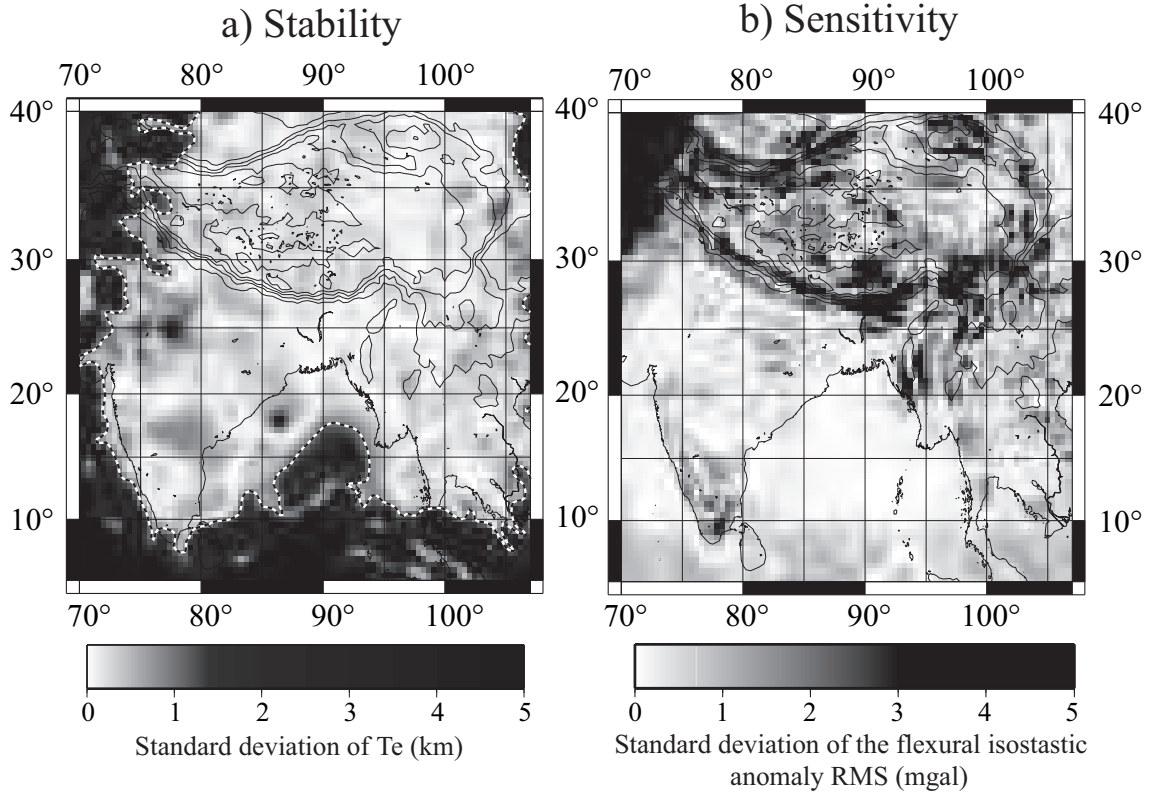


Figure 4.4: Reliability of the iterative 2-dimensional technique as expressed by stability and sensitivity. a) Stability as revealed by the standard deviation of the T_e over the final 5 iterations. The dashed line delimits the area of high stability. b) Sensitivity as revealed by the standard deviation of the RMS difference between observed and calculated gravity anomalies based on small shifts in the final T_e structure.

A final test of the iterative method is to visually compare observed and calculated gravity anomalies along selected profiles of the region. We selected 4 profiles, three of which coincide with the eastern, central and western profiles shown in Figure 4.1. The other profile (dashed line on the map in Figure 4.1) crosses the south-east margin of the Tibetan plateau.

Figure 4.5a-c show the Bouguer gravity anomaly based on the spatially varying T_e structure fits the observed data well along the western, central and eastern profiles. The calculated profiles explain the amplitude and wavelength of the observed profiles in the region immediately beneath the load and in flanking areas. On the central profile, no uniform T_e fully explains the observed Bouguer gravity anomaly. A uniform T_e of 100 km (which is similar to the value deduced in Figure 4.1b from 1-dimensional modelling) accounts for the data in the foreland region, but fails to explain the anomaly immediately beneath the load. A uniform T_e of 40 km accounts for the data beneath the load, but not in the foreland region. On the western and eastern profiles, however, a uniform T_e of 40 km appears to explain the gravity anomaly data both beneath the load and flanking regions. This is attributed to the fact that there is less variation in T_e along these profiles, as seen in the recovered T_e values which average ~ 40 km along much of their length.

In contrast to the western, central and eastern profiles, the south-east margin of Tibet profile shows a relatively constant, low, T_e structure Figure 4.5d. The Airy model (i.e. $T_e = 0$ km) and the variable T_e structure yield virtually identical calculated Bouguer anomalies. This could be a consequence of the long wavelength of the topography which makes a region appear weak, irrespective of its actual strength (e.g. Watts [2001]). However, a low value of T_e is preferred as high T_e values (40, 100 km) fail to adequately recover details of the observed profile, such as the small amplitude anomalies between 500 and 900 km.

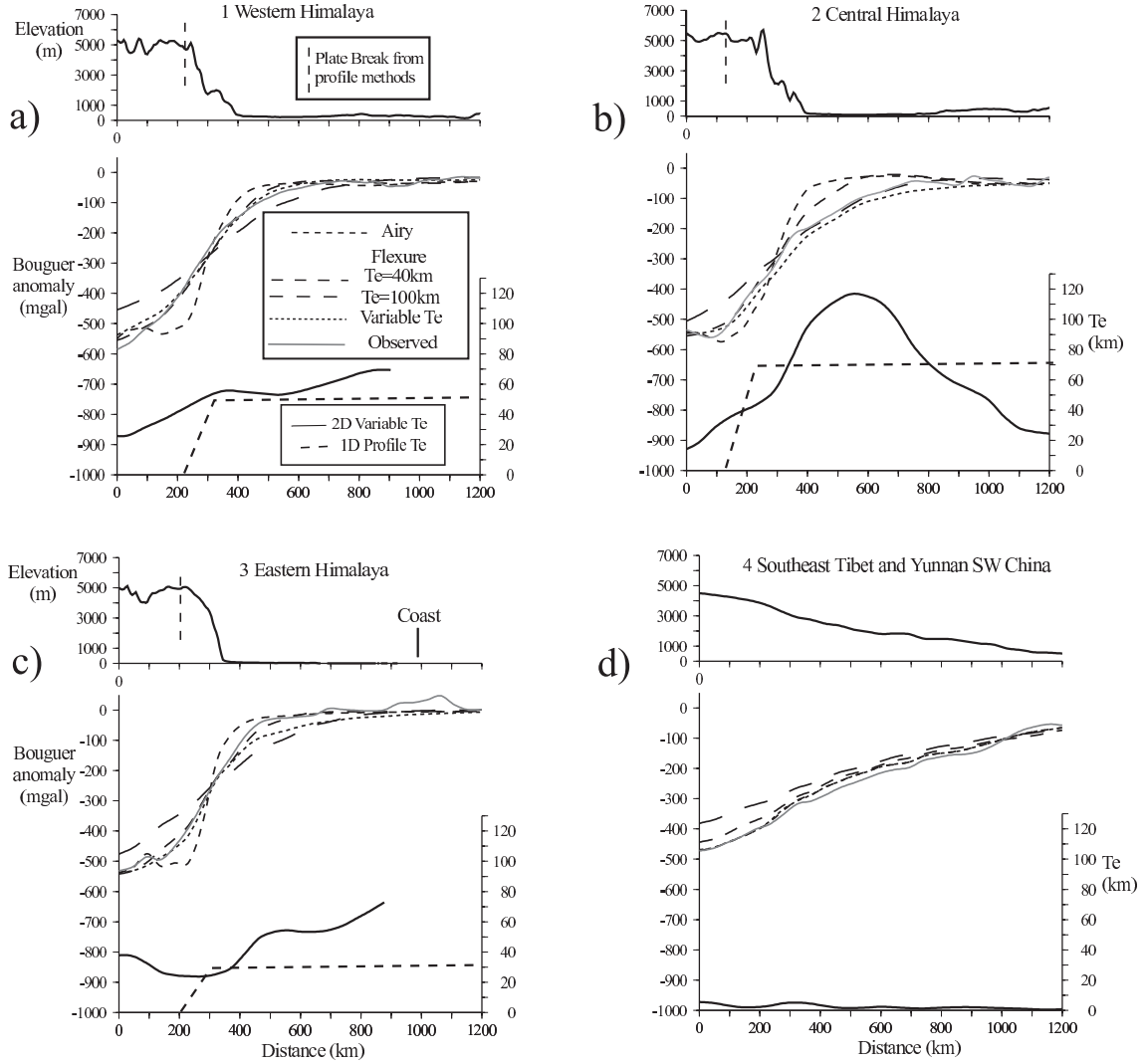


Figure 4.5: Comparison of observed and calculated gravity anomalies on profiles of the Himalaya foreland and southern margin of the Tibetan Plateau (see Figure 4.1 for location). The calculated anomalies are based on spatially varying T_e and uniform T_e structures of 0, 40 and 100 km. Also shown is the topography and recovered spatially varying T_e and the best fitting T_e structure from the 1D profile methods shown in Figure 4.1. a) Western profile of the Himalaya foreland. b) Central profile. c) Eastern profile. d) south-east margin of the Tibetan plateau and south-west China (Yunnan).

4.5 Discussion

4.5.1 Comparison with previous T_e estimates

There have been a number of previous estimates of T_e in the Himalayan foreland region. In the central region estimates from this study of 70 km based on forward modelling and 125 km based on the iterative method (Figure 4.1 and Figure 4.6) agree with previous results (Lyon-Caen & Molnar [1983]; Karner & Watts [1983]). The conclusions of McKenzie & Fairhead [1997] and Jackson *et al.* [2004] that the Northern Indian shield and Himalayan foreland is unusually strong are therefore confirmed. However, the low (24 to 42 km) T_e values that these authors recover are inconsistent with our results. As Figure 4.1 clearly shows, T_e values this low cannot account for the amplitude and wavelength of the observed Bouguer gravity anomaly in the central foreland region. The iterative method reveals that T_e decreases from the central foreland to the east and west. In the east, the values of $30 < T_e < 60$ km recovered by this study, agree with the result of Cattin *et al.* [2001] who recovered a T_e of 50 km in this area.

The most recent investigation of T_e across the central Himalayan foreland region is that of Hetényi *et al.* [2006]. These authors recover an EET (T_e calculated from a rheological model) of 80 km south of the foreland basin, in agreement with the results of this study, and supporting the suggestion that the mantle is strong. The model of Hetényi *et al.* [2006] shows T_e to fall dramatically to ~ 30 km at the southern edge of the foreland basin, in contrast both the profile and the 2D iterative techniques show T_e falling beneath the mountains. Hetényi *et al.* [2006] state that a small change in rheology could change the position where the EET falls to lower values, allowing rheological and purely elastic models to recover the same values.

Over the Tibetan plateau, a variable T_e in the range 5-35 km is recovered, with the lowest estimate corresponding to the central part of the plateau. Rajesh

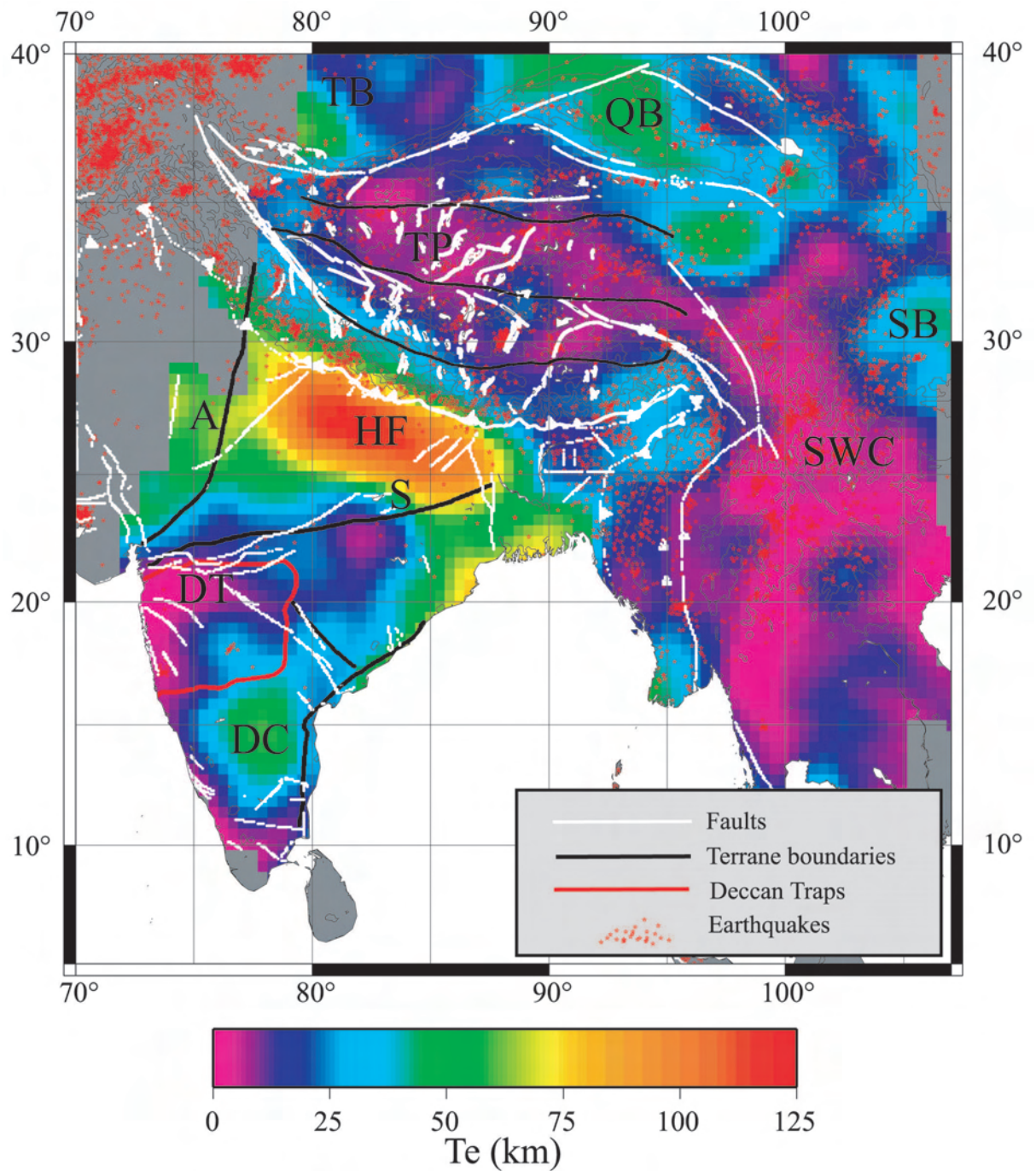


Figure 4.6: The T_e structure of the India-Eurasia collisional zone. Thin black lines show topographic contours at 500 m intervals. Thick black lines show the terrane boundaries from various authors (Yin [2000] Tibetan plateau and Sankran [1999]; Rajaram & Anand [2003]; Sarker *et al.* [2003] over Peninsula India). Thick white lines show faults from Yin [2000] and Searle (pers. com.) over the Tibetan plateau and Dasgupta *et al.* [2000] over peninsula India. Red stars show earthquakes (1979 to present), less than 50 km depth and all magnitudes USGS [2004]. HF = Himalayan foreland, DT = Deccan Traps, DC = Dharwar Craton, QB = Qaidam Basin, SB = Sichuan Block, TP = Central Tibetan plateau, SWC = Southwest China (Yunnan), A = Aravalli fold belt, S = Satpura Mobile Belt.

& Mishra [2003] recover a T_e of 50 km using spectral methods across the entire region, which may be the result of their averaging of the T_e structure of both the foreland and the plateau. Rajesh *et al.* [2003] used similar methods to recover a T_e of 35 km across the Himalayan foreland, 20 km in the central plateau region, and 25 km for the northern margin of the Tibetan plateau and the Tarim and Qaidam basins. The results in the central area are close to those recovered in this study, but the values recovered to the north and south, although higher than in the central region, are not as high as we derive. This is attributed to their use of relatively narrow analysis windows (500 km wide) which prevents the recovery of high T_e (Pérez-Gussinyé *et al.* [2004]; Daly *et al.* [2004]).

Braitenberg *et al.* [2003] recover a spatially varying T_e structure across the Tibetan Plateau which agrees well with our results, particularly over the Qaidam basin area where values of 50-60 km are recovered, while Braitenberg *et al.* [2003] recovers 60-80 km. Jiang *et al.* [2004] concluded that T_e along the northern margin of the plateau was in the region of 40-45 km assuming a broken plate along the Altyn Tagh and Kunlun faults. This in accord with the results of this study. Yin [2002] recover values from 60 to > 100 km over the central Tarim basin. This also agrees with values of ~ 50 km recovered by this study in the west of the basin. At the southern margin of the basin, however, these workers recover lower T_e values (24 km) which they suggest correlate with intensive and localised faulting. Finally, to the east of the plateau, Yong *et al.* [2003] have recovered T_e values of 43-54 km for a Triassic foreland basin, which are somewhat higher than those recovered in this study (20-45 km).

The iterative method reveals that the Indian peninsula, south of the Himalayan foreland, has a highly variable T_e structure. In the region of the Deccan Traps, T_e values of < 5 km on the coast, rising to approximately 30 km to the west across the shield are recovered. These values are significantly less than those estimated by Watts & Cox [1989] on the basis of the width of the lavas, but agree with the value

of 8 km recovered from spectral analysis along profiles (Tiwari & Mishra [1999]). Stephen *et al.* [2003] also used spectral methods to recover T_e values across the Southern Indian peninsula (Figure 4.6, areas DT and DC). They obtained relatively uniform T_e values of 11-15 km which concur with our results over the Deccan Traps, but disagrees over the Dharwar craton where values > 65 km have been recovered. This may be because of a ‘cap’ on the recovered T_e due to the window size used (Pérez-Gussinyé *et al.* [2004]).

Rajesh & Mishra [2004] determined the transitional Bouguer coherence wavelength in overlapping windows across peninsula India. The pattern of rigidity variation implied is very similar to the T_e structure recovered in this paper. However, the T_e values recovered (18-26 km in Northern India and 12-16 km in Southern India) are significantly lower than our values. This discrepancy is attributed to the fact that the T_e values recovered by these authors are lower bounds, as indeed their RMS plots of the difference between observed and calculated Bouguer coherence suggest (Rajesh & Mishra [2004]).

4.5.2 T_e and terrane structure

Figure 4.6 compares the T_e structure of the India-Eurasia collisional system to the main terrane boundaries, faults, and earthquakes.

Over peninsula India, there is a general correlation between T_e , terrane boundaries, faulting, and earthquakes. This is most clearly seen to the south of the Himalayan foreland where the region of high T_e is truncated by a fault which correlates with a suture zone (the Satpura Mobile Belt, Sankran [1999]). This and other terrane boundaries separate Archean-Proterozoic ‘blocks’ which appear to be more rigid and less prone to faulting than the intervening sutures. The terranes sutured together between 1600 and 500 Ma (Sankran [1999]). This suggests that suture zones are weak and have remained so over long periods of time. Evidence that old suture zones are areas of low rigidity also comes from spectral studies of

other continents. Simons *et al.* [2003] showed, for example, that in the Australian continent suture zones are weak. This is attributed to the fact that old suture zones tend to be the sites of intra-cratonic fault localisation (Sutton & Watson [1986]).

The terranes of the Tibetan plateau show a less obvious correlation with the T_e structure. There is a suggestion, however, that the southern most suture (the Zangpo suture) marks the limit of the high (>40 km) T_e zone which extends north from the Himalayan foreland. This is in accord with the results of the INDEPTH project (Nelson *et al.* [1996]) which show that north of the Zangpo suture the middle crust is partially molten. Seismic reflection profile data shows highly reflective ‘bright spots’ which are interpreted as fluid (probably magma) (Brown *et al.* [1996]) at depths of 15-18 km beneath the central plateau. The presence of magma suggests that the crust below these depths is close to melting and, hence, these regions would be expected to be weak. This is likely to be particularly true for the central and northern parts of the Tibetan plateau that are not underlain by Indian continental shield material.

The northern and eastern edges of the Tibetan plateau are bordered by the Tarim, Qaidam and Sichuan basins. Although these regions are not delimited by terrane boundaries in Figure 4.6, they are believed to be underlain by fragments of old cratonic ‘blocks’. The Qaidam basin, for example, is underlain by Palaeozoic crust (Zhu & Helmberger [1998]) and the Sichuan basin by Proterozoic basement (Chen *et al.* [1994]). These ‘blocks’ correlate with lower levels of seismicity and apparently are of higher T_e than surrounding regions.

Seismic attenuation (Q) variations, such as those recovered by Jemberie & Mitchell [2004], provide an additional indication of variations in the crustal (0 to 30 km) properties in this region. For depths less than 10 km there is little apparent correlation between the observed attenuation pattern and T_e . Between 10 and 30 km, however, there is a clear correlation. High Q values beneath the plateau and to the south east are regions of low T_e . To the south and east of the

plateau where high T_e values are recovered Q values are low, suggesting these parts of the crust are more rigid on both short seismic and longer flexural time scales.

Seismic tomographic studies such as that of Li *et al.* [2006] show there to be a correlation with T_e and lithospheric variations at sub-crustal levels. Fast regions thought to be associated with the cold Asian lithosphere overridden by the Tibetan orogen extend, at 140 km depth, to approximately the line of the Zangbo suture beneath the Tibetan plateau. This is in-line with the extent of the high T_e region recovered in this study. There also appears to be correlation between the other high T_e regions bordering the plateau and high modelled P-wave velocities. In the south east corner of the Tibetan plateau there appears to be a region of very low mantle P-wave velocity suggesting hotter mantle in this region. This correlates well with the low T_e region seen beneath this area.

4.5.3 T_e and crust and mantle strength

The T_e values recovered in this study are, in places, high and clearly exceed the seismogenic layer thickness, T_s as defined by Maggi *et al.* [2000a]. This is best seen in the central Himalayan foreland where T_e is ~ 100 km and T_s is ~ 40 -45 km (Figure 4.6). The T_e values greatly exceed the crustal thickness (40-44 km) (Murty *et al.* [2004]) which implies that the sub-crustal mantle must, in places, be rigid enough to significantly contribute to the long-term integrated strength of the lithosphere.

The fact that the sub-crust mantle can be involved in the support of long-term loads, is not surprising given that in oceanic areas, away from a mid-ocean ridge, T_e greatly exceeds the crustal thickness. As shown by Watts [2001] oceanic T_e values lie within the depth to the 300-600C° isotherms as defined by a cooling plate model. Therefore, a significant portion of the sub-oceanic mantle is strong - down to depths of ~ 50 km.

Seismic tomographic data shows that the mantle below parts of the continents

has a significantly higher shear wave velocity at 120 km depth than has the oceans (Ritsema *et al.* [2004]). If the mantle is cooler below the continents than the oceans, as these data suggest, then it is likely that it is rigid to a greater depth. Therefore continents, where they have cold mantle roots, would be expected to be stronger than the oceans.

Further lines of evidence that the continental mantle has some rigidity over long periods of time comes from direct samples and thermal modelling. McKenzie [1989], for example, showed that the geotherm of the Archean Kaapvaal craton in South Africa could be modelled as a plate with a mechanical boundary layer thickness of 165 km and a interior potential temperature of 1280 C°. This thickness compares to ~ 100 km for the equivalent layer in the oceans. Extrapolating the geotherm shown in McKenzie [1989] gives a depth to the 600 C° isotherm of 75-80 km. At these temperatures, by analogy with the oceans, the mantle, if it is dry, will be rigid enough to provide a significant contribution to the observed T_e . McKenzie [1989] also presents data from sheared peridotite nodules which originate from depths greater than about 160 km, geochemical studies of which show them to be part of the convecting mantle. Nodules shallower than 160 km are not sheared, suggesting that this part of the mantle is behaving rigidly when stressed by convection.

The great depth to the convecting mantle in continental regions is in accord with the results of Simons *et al.* [2003] based on spectral and tomographic studies in the Australian shield. They showed that a correlation exists between the isostatic (as defined by the directional Bouguer coherence) and seismic anisotropy down to a depth of almost 200 km and therefore that the lithosphere is apparently rigid to these depths. At greater depths, the correlation breaks down and while the mantle may have a preferred orientation on short seismic time-scales it appears weak on the long time-scales associated with flexural isostasy (Simons *et al.* [2003]).

If the Kaapvaal and Australian cratons are typical of Archean shield areas, then other shields, such as the Himalayan foreland, should also be rigid we believe to

great (150-200 km) depths. The high T_e recovered in parts of the Indian peninsula is therefore a manifestation of this rigidity. The actual thickness of the rigid layer can not be determined from isostatic studies because as a number of authors have shown e.g. Burov & Diament [1995], T_e is not the actual depth to which materials behave elastically, but reflects the integrated strength of the lithosphere.

4.5.4 T_e and ductile crustal deformation

The elevated regions of the south-eastern margin of Tibet and Yunnan are associated with low values of T_e , suggesting that they may be locally compensated (SWC - Figure 4.6). The region therefore differs from other parts of the plateau margin (e.g. the Himalayan foreland) which appear to be regionally compensated by a strong under-thrust lithospheric plate. If the south-eastern margin of the plateau, east of the eastern syntaxis, is not associated with under thrusting of a strong plate, then the question arises as to why it is elevated? One possibility is some form of lower crustal flow (Royden *et al.* [1997]; Clark & Royden [2000]; Wang [2001]). According to this model, low viscosity lower crustal material flows out from beneath the Tibetan plateau and ‘inflates’ south-west China.

Previous authors (e.g. England & Houseman [1985]) have argued that the Tibetan plateau can be modelled as a thin viscous sheet which is deformed by a rigid ‘indenter’ to the south. This model was able to incorporate higher viscosity (more rigid) blocks. These deformed much less than the surrounding material and generated regions of low relief which were suggested to be analogous to the Tarim basin to the north of the Himalayas.

Royden *et al.* [1997] pointed out that a viscous sheet, however, is unrealistic since such models do not incorporate a vertical stratified rheology. They therefore proposed instead a model in which the upper crust and mantle are effectively decoupled by a low-viscosity lower crust during compressional deformation.

Clark & Royden [2000] show that there are two ‘end-members’ for the topo-

graphic gradient of the Tibetan plateau margin; steep (e.g. southern Himalaya, Tarim and Sichuan basins) and shallow (e.g. Yunnan, south-western China). They model these end member cases in terms of a channelised viscous flow in the lower crust. In the steep case, the channel contains high viscosity material and a steep topographic gradient develops. In the shallow case, a low viscosity channel inflates the crust, giving rise to a shallow gradient. They showed that the Tibetan plateau was surrounded by a number of rigid ‘blocks’, which included the Tarim, Qaidam, and Sichuan basins and the Indian shield. The lower crust, they argued, would preferentially flow around these ‘blocks’ and in the process generate the shallow topographic gradients seen at the present day in south-west China.

The high T_e areas mapped in this study supports the view that deformation in the Tibetan plateau and surroundings has involved ‘blocks’ that have remained rigid for long periods of time. However, the mechanism of crustal thickening across the region is more complex, we believe, than has been previously suggested. We know that in the regions of steep topographic gradients, surface loads are compensated, at least in part, regionally not locally as has been assumed by previous authors (England & Houseman [1985]; Royden *et al.* [1997]; Clark & Royden [2000]). Moreover, it is hard to reconcile the underthrusting of the rigid Indian lithosphere beneath the Tibetan plateau (Nelson *et al.* [1996]) with viscous injection and flow of the lower most crust. Searle *et al.* [2003], for example, showed that in the Mt Everest region mid-crustal material has plastically extruded to the surface from depths of 12-18 km and laterally from approximately 200 km to the north. This suggests that the topography of the high Himalaya is generated, at least in part, by viscous flow of mid-crustal material that is over-riding the strong Indian plate, rather than by inflation of the lower crust (Clark & Royden [2000]).

However, the arguments of Clark & Royden [2000] for lower crustal flow in the shallow topographic gradient areas of south-west China (Yunnan) are preferred. Wang *et al.* [2001] has shown, from a GPS-derived velocity field, that central Tibet

is, on a large scale, deforming continuously and that material is being extruded from the Tibetan plateau to the south east. This is believed to represent movement of weak surface crustal material deforming and moving to the south east with the lower crust beneath it. The pattern of deformation is consistent with the results of this study, which show a broad region of low T_e beneath the Tibetan plateau which appears to run between high T_e regions and into south-west China (Yunnan). This region of low T_e may represent the isostatic signal of the weak, flowing, lower crust.

4.5.5 Extruded volume

If T_e variations can be used to define the area of extruded material then a minimum volume for this extruded material can be defined based on the observed topography. This estimate of the extruded volume can, however, only ever be a minimum as erosive processes are always working to reduce the topography created.

Figure 4.7 shows that there is a large region with a recovered T_e below 5 km to the SE of the high Tibetan plateau. This region correlates with the low topographic gradient region discussed in the previous section. If this topography is generated by crustal flow the extruded volume can be calculated by adding the surface volume in this area and the volume of the balancing Airy isostatic root.

The volume of material above sea level within the low T_e region is $\sim 1.023 \times 10^6$ km³. This needs an Airy root of $\sim 5.437 \times 10^6$ km³ to balance it, assuming crustal and mantle densities of 2800 kgm³ and 3330 kgm³ respectively. This gives a total volume of extruded material of $\sim 6.465 \times 10^6$ km³.

Given that this extruded material is driven out by indentation of the Indian continent into Eurasia, a minimum time for the observed topography to develop can be calculated. This calculation is based on the assumption that the total volume of India that is overridden by the Himalayas is laterally extruded. A length of the Himalayan mountain front of ~ 2400 km and a crustal thickness of 40 km was used to calculate the 'unit volume' of 1 km of overridden Indian Crust. To generate the

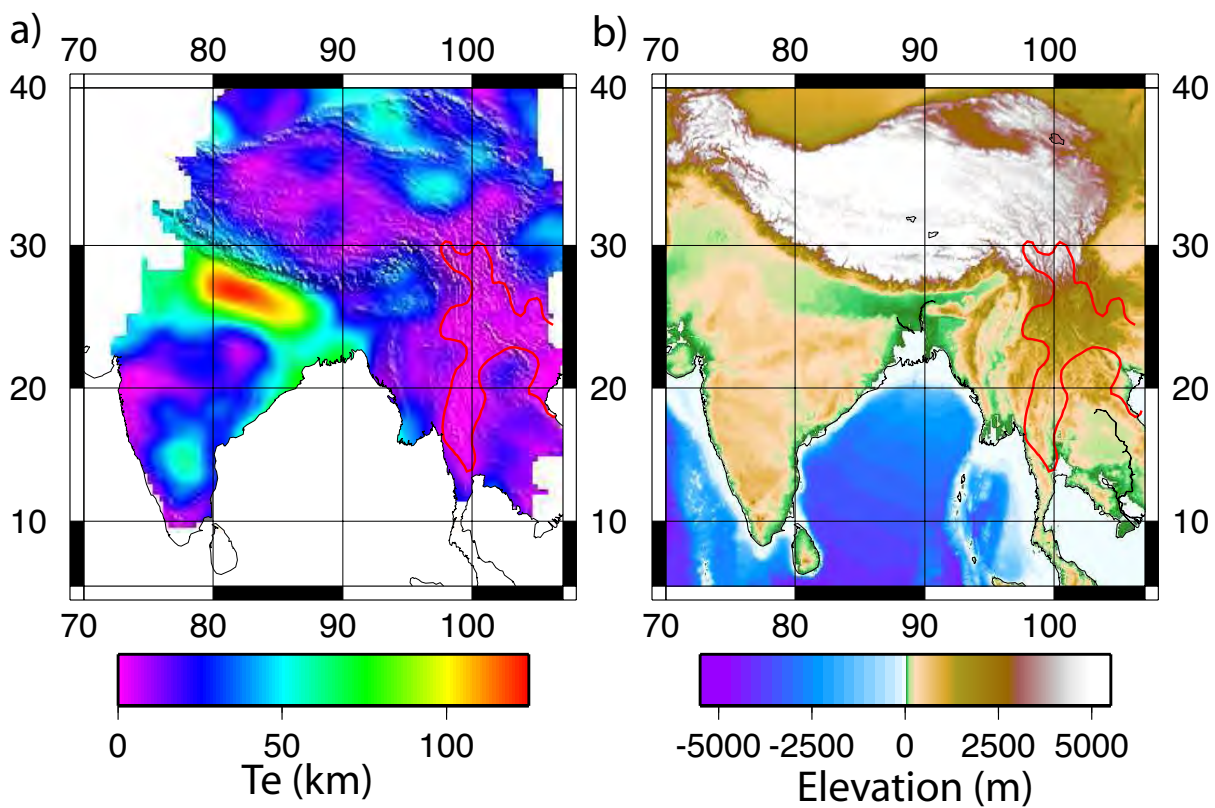


Figure 4.7: T_e and topography maps highlighting exceptionally weak region. a) T_e map with topographic gradient superimposed, largest area with T_e below 5 km highlighted in red. b) topographic map showing relationship between topography and region with $T_e < 5$ km.

total volume of extruded material a total of 67 km of Indian crust must have been pushed beneath the Himalayas.

GPS data suggests that although India is moving north at approximately 5 cm per year, the rate at which it is being overridden is closer to 1.6 cm per year (Wang *et al.* [2001]). At this rate the observed volume would take a minimum of ~ 4.2 ma to generate.

Clark *et al.* [2004] show that there has been approximately 2000 m of uplift with little horizontal deformation in this area, which they associate with lower crustal flow. The timing of this uplift is not well constrained; however, a speculative range of uplift rates of 0.07-0.2 mm/yr was proposed. A maximum possible uplift rate can be given for this area based on the minimum time needed to generate the topography by lateral extrusion. This calculation based on 2000 m of uplift in 4.2 ma gives a maximum uplift rate of 0.48 mm/yr.

The T_e structure of the region shows that there is a weak zone running around the eastern syntaxis of the Himalayas. This is proposed to be the site of lower crustal flow as material escapes from beneath the elevated Tibetan plateau. Consideration of the isostatically balanced volume of this region suggests that the topography could be created in a geologically plausible time frame, by lower crustal flow.

Further work would include constraining erosive sediment budgets to assess the ‘true’, uneroded volume of this region. Also other possible regions of lower crustal flow, such as to the north east of the Sichuan Block, need to be considered along with the growth of the Tibetan plateau with time. This would allow a full balance of the volumes of over thrust and extruded material, and a more accurate analysis of the timing of extrusion, to be made.

4.6 Summary

1. The iterative method is a non-spectral technique that successfully recovers a spatially varying T_e structure *directly* from observed topography and Bouguer gravity anomaly data.

2. Application of the method to the India-Eurasian collisional system reveals a complex T_e structure with both high and low rigidity areas.

3. The Himalaya foreland is a high T_e region ($40 < T_e < 100$ km) which forms a rigid ‘block’ with well-defined edges, as shown by localisation of faulting and deformation along its northern margin. Other rigid ‘blocks’ have been identified in the region of the Qaidam and Sichuan basins.

4. The Tibetan plateau is a low T_e region ($0 < T_e < 20$ km) which forms a weak zone that separates these rigid blocks. The weak region extends into south-west China (Yunnan) where T_e is also low.

6. Within cratonic regions there is evidence (e.g. Indian peninsula) that the rigid blocks are separated by weaker areas which correlate with ancient suture zones.

7. Tectonic styles in the India-Eurasian collisional system involve both rigid and non-rigid ‘blocks’. Where high rigidity ‘blocks’ are present the style is dominated by underthrusting of the more rigid ‘block’. Where the collisional zone is not constrained by rigid ‘blocks’ the style appears to be dominated by lower crustal flow and continuous deformation.

The Arabia-Eurasia collision: T_e in the Persian Gulf

5.1 Introduction

The northern part of the Arabian shield, and the Persian Gulf in particular, represents the largest oil producing province in the world. Much of the source rock and many of the traps for oil lie within the Tertiary flexural foreland of the Zagros Mountains which was formed during the final closure of the Tethys Ocean.

The formation of the Tertiary foreland was investigated based on 1D flexural models with constant T_e as discussed in Chapter 2 and applied to the foreland associated with the north Oman ophiolite in Chapter 3. In addition a technique for recovering T_e variations along a profile was used. The 2D iterative technique discussed in Chapter 2 and applied to the India-Eurasia collisional system in Chapter 4 was also applied in this region.

That the Tertiary foreland is of flexural origin has been previously suggested by Snyder & Barazangi [1986], based on modelling along a single profile across the mountains and foreland. However, the current study is a systematic attempt to recover the spatial variation in rigidity across the region.

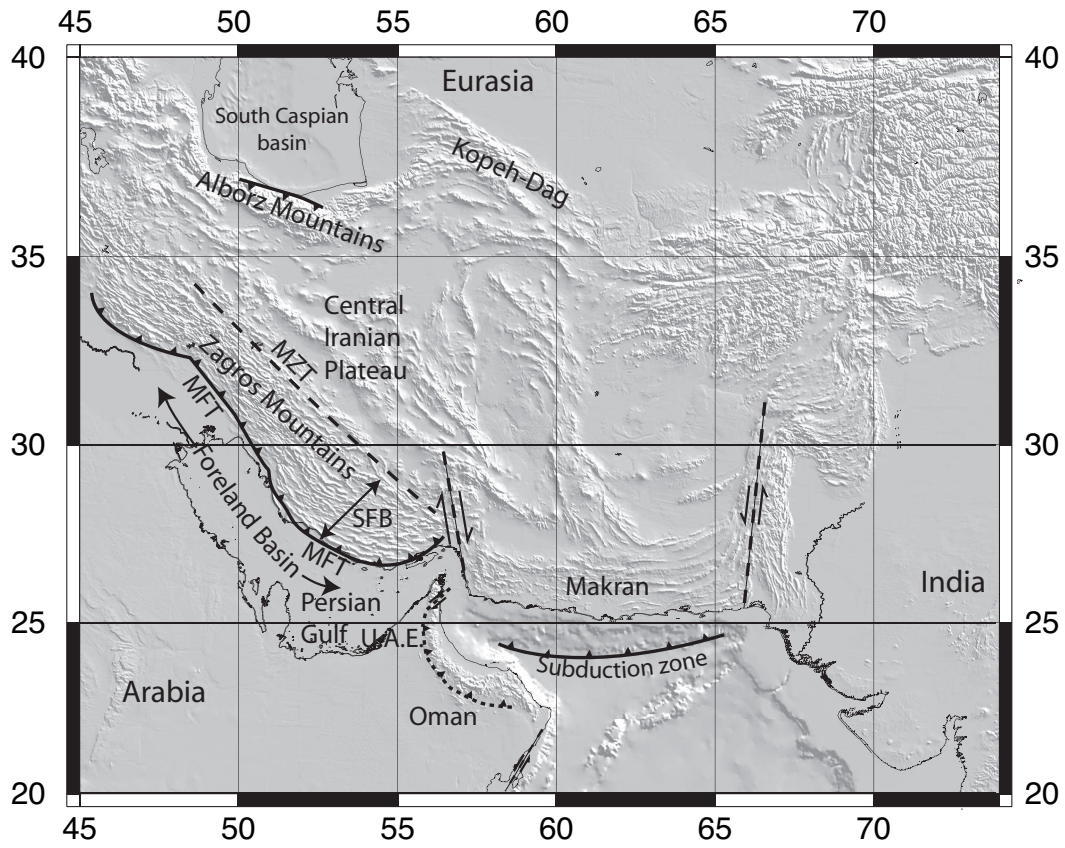


Figure 5.1: Shaded relief map of the Arabia-Eurasia collisional system, based upon data from the GEBCO [2003] one minute global elevation grid, illuminated at 45° . Main Frontal Thrust (MFT), represents approximate present day deformation front. Main Zagros Thrust (MZT) represents approximate position of Arabia Eurasia suture zone. Simply Folded Belt (SFB) represents the region of post Miocene deformation.

5.1.1 Geological History

The early history of the Arabian shield was discussed in Chapter 3. The Zagros region differs from Oman in that the transition from ‘oceanic vs continental’ to full ‘continental vs continental’ collision has been made.

The initial closure of the Tethyan ocean in Iran is marked by a zone of Cretaceous ophiolites. These outcrop from Turkey south-east along the Main Zagros Thrust (MZT) which marks the proposed suture zone between Arabia and Eurasia (Agard *et al.* [2005]). These ophiolites are ~ 90 Ma (Robertson [2000]) and are

approximately concordant in age with the Oman ophiolite discussed in Chapter 3.

Evidence for initial continent-continent collision, presented by Agard *et al.* [2005], first appears around 25 to 23 Ma from deformational structures located south of the MZT, shown in Figure 5.1. The progress of this collision is marked, around 20 Ma, by the onset of deposition, and subsequent deformation of terrestrially derived sediments (Koop & Stonely [1982], Hessami *et al.* [2001]).

McQuarrie *et al.* [2003] recover a minimum age of final continent-continent collision of 10 Ma, based on reconstruction of the plate margin and plate kinematic considerations. The most significant amount of shortening, within the Zagros mountains, of 57-140 km, has occurred since 5 Ma (McQuarrie [2003], Talebian & Jackson [2002]). This has also been noted by Allen *et al.* [2004] who showed that at the deformation rates determined from geological markers and modern slip rates, all the shortening observed would be completed in 3-7 Ma. Allen *et al.* [2004] proposed that before ~ 7 Ma the majority of deformation was taken up within the Central Iranian Plateau. When the Central Iranian Plateau achieved a critical elevation it became easier for deformation to propagate out into less elevated areas, which today form the Zagros Simply Folded Belt (SFB). This is a similar phenomenon to that suggested for the formation of the Himalaya by, for example, England & Houseman [1988]. An alternative cause for the change in shortening rates, presented by Molinaro *et al.* [2005], could be delamination of dense lithospheric mantle, leading to uplift of the central plateau and a change in the taper of the thrust wedge.

The modern tectonic setting of the Zagros mountains is shown in Figure 5.1. Today the convergence rate between Arabia and stable Eurasia is 2-2.5 cm/year from GPS measurements (Nilforoushan *et al.* [2003]) and 3-3.5 cm/year from global plate motion models (DeMets *et al.* [1994]). The Zagros mountains accommodate 15-30% of the shortening while the remainder is accommodated by convergence across the Alborz mountains in the north.

5.1.2 Foreland stratigraphy

The stratigraphy of the foreland region has been described by Alavi [2004] based upon a compilation of well and outcrop data in addition to both published and unpublished stratigraphic and sedimentological information. The succession of Pre-Cambrian to Late Cretaceous continental passive-margin and shelf sediments are overlain by a series of Late Cretaceous to recent marine and continental deposits. These deposits lie within elongate north west to south east asymmetric basins, related to development of the Zagros orogenic complex to the north east, where much of their sediment is sourced from. Today many of these basins have been incorporated within the developing Zagros mountain belt. Alavi [2004] considers the Late Cretaceous to recent sequence to consist of three separate sub-sequences.

The first sub-sequence from ~ 90 to ~ 68 Ma is related to emplacement of Neo-Tethyan ophiolites along the passive-margin from the north east. The basin fill, now seen within the imbricate thrust zone south of the MZT, consists of a variety of clastic sediments including the Amaran Formation which contains material derived from the ophiolites. The south west margin of this basin lay close to the modern deformation front of the Zagros origin and was marked by a transition from clastic to more carbonate dominated sedimentation.

The second sub-sequence ranges in age from ~ 66 to ~ 35 Ma. This sub-sequence is thought to be the response to slower convergence and loading of the continental margin which lead to the development of the Pabdeh foreland sequence. Erosional unloading during this time is thought to also be a factor in altering the position of the centre of deposition within the foreland basin from close to the mountains to a more distal setting. Much of this sub-sequence is now incorporated within the Zagros fold and thrust belt.

The final sub-sequence from ~ 35 Ma to recent results from progressive advance of the thrust load onto the Arabian shield. The lower formations, such as the regionally extensive Asmari carbonates, are thought to represent an initial flooding

event associated with loading to the north east. Overlying the Asmari Formation the Gashsaran, Agha Jari and Bakhtiari Formations all show a continued progradation of predominantly silicalastic strata which are coarser grained in the north east and finer grained towards the south west.

5.1.3 Geological summary

The shortening and thickening of the upper continental crust generates a load which must be, in some way, isostatically compensated. Snyder & Barazangi [1986] showed that the observed Bouguer gravity anomaly could be matched by a model where the topographic load of the Zagros Mountains created a flexural foreland basin and deflected the Moho south of the mountain front. It is this flexural foreland that was investigated in the current study. The flexural foreland relates to the material deposited during the final, ~ 35 Ma to recent, sequence described by Alavi [2004]. Earlier sequences, and the north eastern part of the final sequence, have now been incorporated into the fold and thrust belt so represent part of the thrust load.

5.2 Data

The topographic data used in this study was from the GEBCO [2003] 1-minute global topography and bathymetry grid, shown in Figure 5.2. This data set incorporates data from the Global Land One-km Base Elevation (GLOBE [1999]) project for the onshore topography. This data was re-sampled to a 10 km point spacing for the models applied along profiles.

The point Bouguer gravity data was provided by the Bureau Gravimetrique International (B.G.I. - see <http://bgi.cnes.fr> for more information). It represents the simple Bouguer corrected gravity anomaly based on a mean correction density of 2670 kg m^{-3} . The data, which is shown in Figure 5.3, was interpolated to create

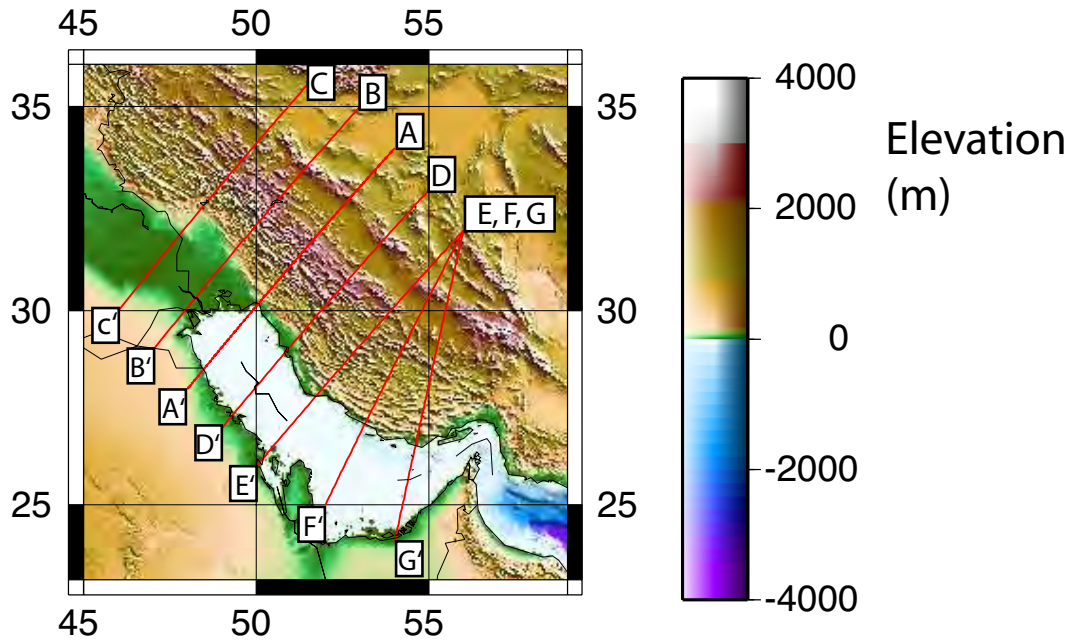


Figure 5.2: Location of profiles A to G analysed in this study. Map shows topographic data from the GEBCO 1-minute grid, illuminated from 45°

a grid using the **surface** function of the **GMT** software package. A tension factor of 0.25 was used as suggested by Smith & Wessel [1990] for the interpolation of potential field data. Gridding the data allowed for easy sampling for comparison to models along profiles and the application of the 2D iterative technique discussed in Chapter 2.

The long wavelength, satellite derived, gravity field was removed from the Bouguer gravity data before comparison with the calculated anomalies. The long wavelength anomaly chosen was complete to degree and order 20 which includes all wavelengths $> \sim 2000$ km. This field was removed in order to eliminate any contributions from sub-crustal sources, such as those associated with dense down going slabs. The satellite orbits were from the OSU91a model produced by Rapp & Pavlis [1990]. At degree and order 20 there is only $\pm \sim 2$ mgal difference across the entire region between the OSU91a and the more recent EGM-96 field (Lemoine *et al.* [1998]). Such small and very long wavelength discrepancies will have little effect on the T_e recovered by either the profile or the 2D iterative technique. The

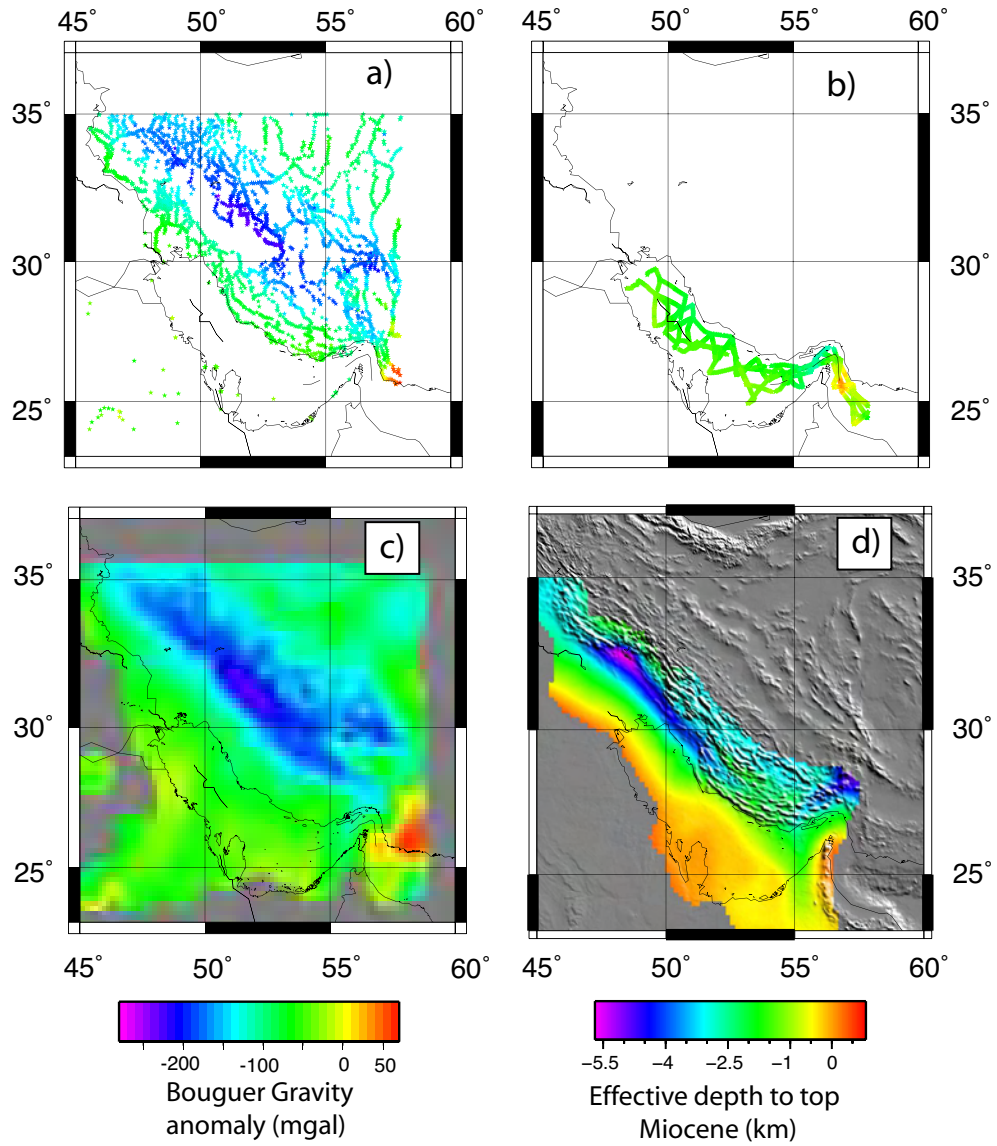


Figure 5.3: Data sets analysed in this study. a) Onshore point Bouguer gravity anomaly data. b) Offshore point Bouguer gravity anomaly data. c) Gridded gravity data. d) Depth to the top Asmari Formation from isopach data of Koop & Stonely [1982] for thickness of post Asmari (Miocene to recent) sediments. Overlain with shaded relief map illuminated at 45°.

results recovered based on Bouguer data corrected with the older field therefore remain valid.

Isopach data for the thickness of the Miocene to recent sediments (post Asmari Formation) was taken from Koop & Stonely [1982]. This data was digitised from a 500 m contour map and then converted to a grid using the **GMT** function **surface** with a minimum curvature solution (tension factor = 0). The stratigraphy of the region presented by Alavi [2004] suggests that the Miocene to recent sequence represents the foreland basin fill and is associated with progressive thrusting within the Zagros orogen to the north east of the foreland. Koop & Stonely [1982] state that within the fold belt the data are approximate as folding, uplift and erosion were occurring coevally with deposition. These processes mean that the thickness of material now seen within the mountains may not represent the thickness of material originally deposited.

5.3 Modelling along profiles

A series of profiles across the mountain range and foreland basin were considered. Profile locations are shown in Figure 5.2. These profiles allowed an analysis of variations in T_e along strike of the mountain front.

5.3.1 Profile A

Initially a profile through the region where the negative gravity anomaly below the mountains was most pronounced was considered. The profile was constructed by sampling topographic and Bouguer gravity grids along a line at right angles to the mountain front. The profile location and sampled data are shown in Figure 5.4.

The flexural response of the lithosphere to surface loading was modelled using a finite difference technique to describe the response of a series of broken elastic plates, with variable positions for the plate break (X_{pb}) and T_e . The gravity

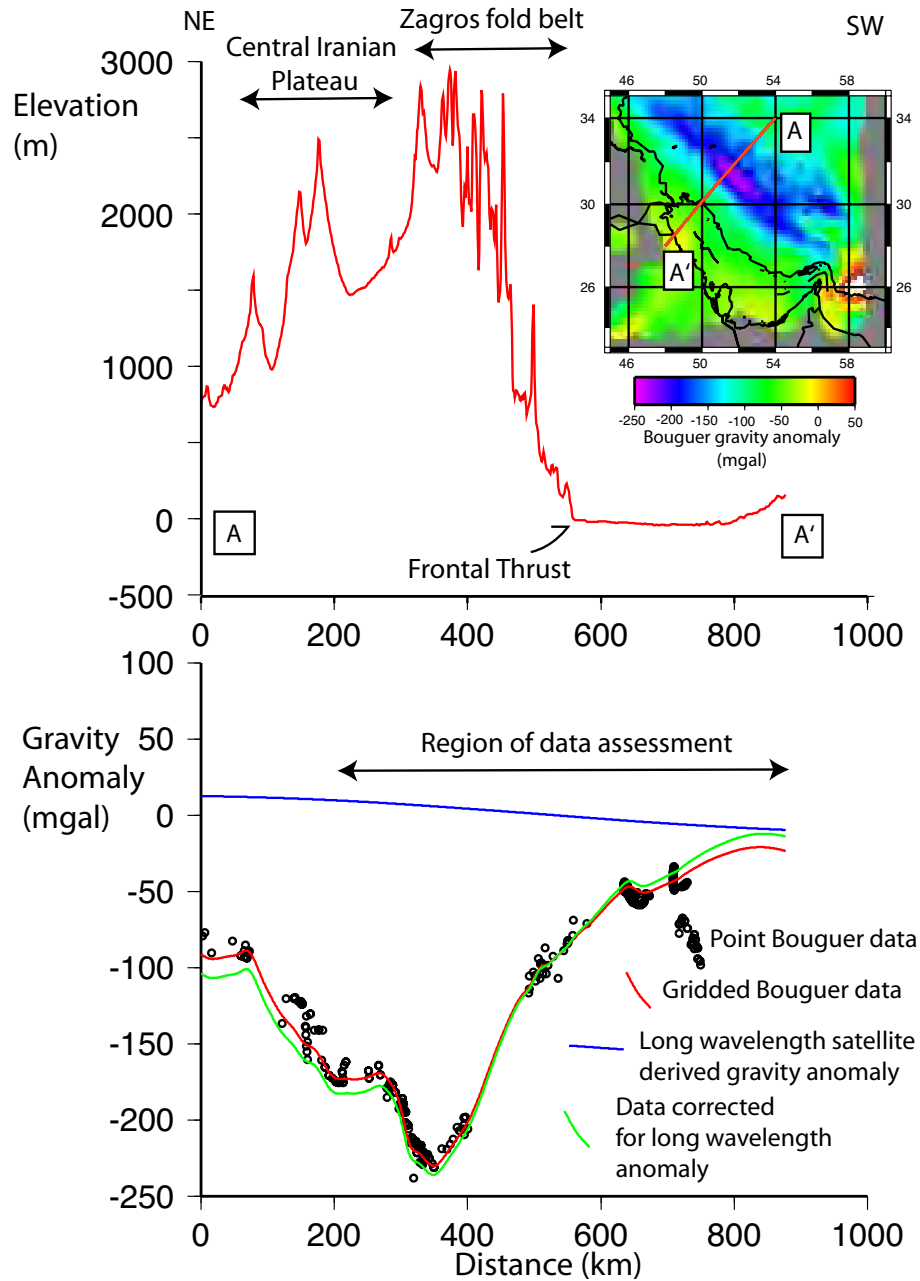


Figure 5.4: Topography and Bouguer gravity anomaly along profile A-A'. a) Topography data, inset map shows location of profile A-A' and observed Bouguer anomaly data. b) Bouguer gravity data. Black circles show point Bouguer gravity data within $\pm \sim 27$ km of the profile, gridded Bouguer gravity anomaly (red line), long wavelength satellite derived anomaly (blue line), and Bouguer anomaly corrected for long wavelength anomaly (green line).

anomaly was calculated assuming two interfaces, one at the Moho and one at the infill/basement interface. This follows the technique introduced in Chapter 2 and successfully applied to the Himalaya in Chapter 4. The surface topography was considered to be the only driving load as there are no distinct positive Bouguer gravity anomalies present beneath the mountains. Such anomalies would suggest the presence of a buried load on the scale seen, for example, in Oman (Chapter 3) or Taiwan (Lin & Watts [2002]). The topographic load front was defined by the northern coast of the Persian Gulf. For additional profiles, where the foreland is above sea level, the load front was defined by the first clear topographic discontinuity.

The parameters used in the model were as follows and are typical of values used in the previous chapters and shown in Table 2.1. Densities used were, $\rho_{\text{load}} = 2650 \text{ kg m}^{-3}$, $\rho_{\text{infill}} = 2650 \text{ kg m}^{-3}$, $\rho_{\text{crust}} = 2800 \text{ kg m}^{-3}$ and $\rho_{\text{mantle}} = 3330 \text{ kg m}^{-3}$. Initial crustal thickness was assumed to be 35 km. This is at the lower end of the range of values proposed by Hatzfeld *et al.* [2003] of 36 to 40 km from seismic refraction data.

5.3.2 Profile A-A' results

Broken plate gravity model

Figure 5.5 shows the results for the first set of models. Figure 5.5a shows a best-fitting X_{pb} is recovered at ~ 400 km, based upon an RMS comparison between observed and calculated anomalies along the ~ 600 km long section of the profile indicated in Figure 5.4b. X_{pb} values assessed ranged from the origin, i.e. the entire width of the mountains was considered to be load, to 600 km, where only the load of the frontal folds was applied to the elastic plate. To the north of X_{pb} the gravity anomaly was calculated based upon a model of Airy compensation of the surface topography. X_{pb} of 400 km therefore equates to a load width of 150 km, where load width is the distance from the load front in the south to the plate break in the

north. X_{pb} locations were assessed every 50 km from the profile origin to the load front. A best-fitting T_e of ~ 60 km is recovered for the foreland region, based on a X_{pb} at 400 km. T_e was assessed across the range of 0 to 100 km in 10 km intervals.

An additional test was carried out to assess the impact of the initial crustal thickness on the recovered plate break and T_e . The calculated gravity anomaly was re-calculated, for the same range of T_e and X_{pb} values assumed for Figure 5.5a, with a crustal thickness of 40 km in place of 35 km. The result of the RMS comparison between observed Bouguer and calculated anomalies derived from models with the thicker crust is shown in Figure 5.5b. This second test recovered best-fitting X_{pb} and T_e values the same as for the 35 km crustal thickness case, and a similar pattern of RMS variation. This indicates that the results of this technique, in this region, are insensitive to the input value of standard crustal thickness within the range proposed by Hatzfeld *et al.* [2003].

The range of calculated gravity anomalies based on a flexural model with X_{pb} 400 km from the origin and a crustal thickness of 35 km is shown in Figure 5.5c. There is, visually, a good fit between the shape of the blue, best-fitting calculated anomaly, based on a T_e of 60 km, and the green observed Bouguer gravity data, corrected for long wavelength effects. A DC shift of ~ 13 mgal would be needed to make the model values fit absolutely but this may relate to regional structures rather than flexural processes.

Figure 5.5d shows the effect of varying X_{pb} on the modelled gravity anomaly. Flexure, and hence the associated negative gravity anomaly beneath the load, initially increase as the plate changes from approximately continuous to broken. This can be seen best around 200 km from the origin. However, as the X_{pb} value passes 450 km the extent of flexure in the foreland decreases as the majority of the load is Airy compensated so in effect less material is acting to bend the plate.

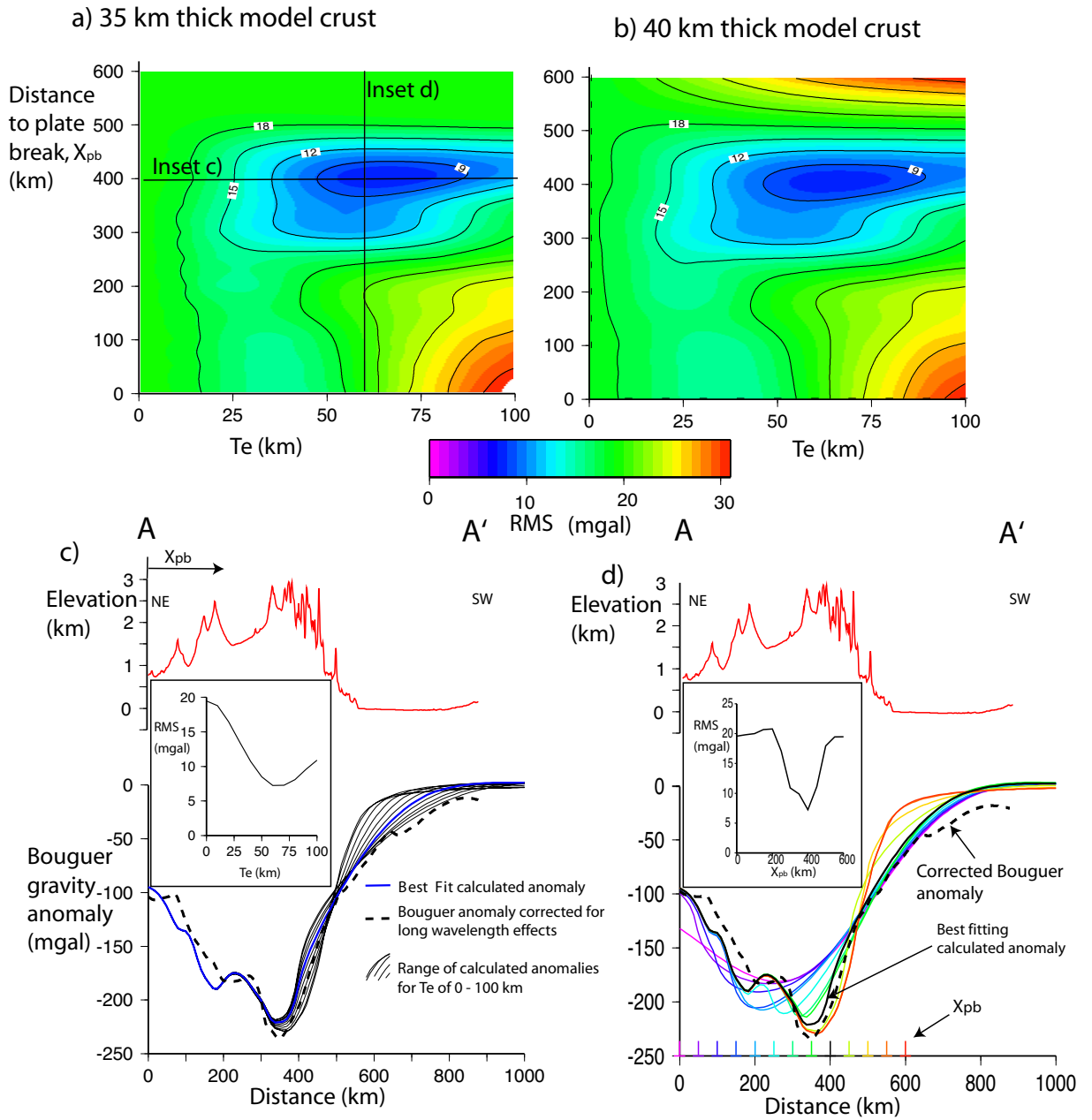


Figure 5.5: Calculated anomalies and RMS difference between calculated and observed Bouguer gravity anomaly across the foreland region of profile A. a) RMS difference between calculated and observed Bouguer gravity anomalies, based on various plate break positions (X_{pb}) and T_e values with a model crustal thickness of 35 km. Black lines mark the best-fitting profiles shown on the inset graphs in c) and d). b) As in a) but with a 40 km model crust. c) Calculated gravity anomalies for a range of T_e values with X_{pb} 400 km from the origin. d) Calculated gravity anomalies with a fixed T_e of 60 km and variable X_{pb} . X_{pb} locations are colour coded. Best fitting model and plate break position in black.

Isopach model

A similar technique for recovering T_e and plate break position was applied to the isopach data digitised from Koop & Stonely [1982]. This data was for the thickness of the Early Miocene (~ 20.3 Ma Hessami *et al.* [2001]) to recent sediments. This isopach relates to the depth of the top of the carbonate Asmari Formation. This isopach was assessed as it is thought to have formed coevally with significant mountain building within the Zagros and development of the current foreland basin (Alavi [2004]). Only data to the south west of the deepest point in the foreland was considered (green line in Figure 5.6c). The apparent shallowing of the Asmari Formation seen towards the north east cannot relate to a flexurally derived signal of foreland basin shape. Instead this shallowing relates to the significant folding, thrusting and erosion seen within the orogen which has perturbed the original thickness of the sediments. The isopach data towards the south west represents the relatively undeformed foreland sequence. Blind thrusting within the foreland sequence may, in places, increase the formation thickness by repeating units, however, this is unlikely to happen across a broad width of the foreland without topographic expression.

The RMS difference between modelled and observed flexural surfaces shown in Figure 5.6a indicates that two possible positions of the plate break provide a good fit to the data. Either a plate break in the far north, at the profile origin, with a T_e of ~ 40 km or a plate break at 350 km with a T_e of 50 to 60 km. The latter position of the plate break is preferred as it agrees more closely with the plate break position recovered from gravity data. The best fitting flexural model, with a T_e of 60 km, is highlighted in red in Figure 5.6b. This model shows a good match to the shape of the southern section of the isopach data but further to the north, closer to the mountain front, the isopach data appears much steeper than the modelled flexure.

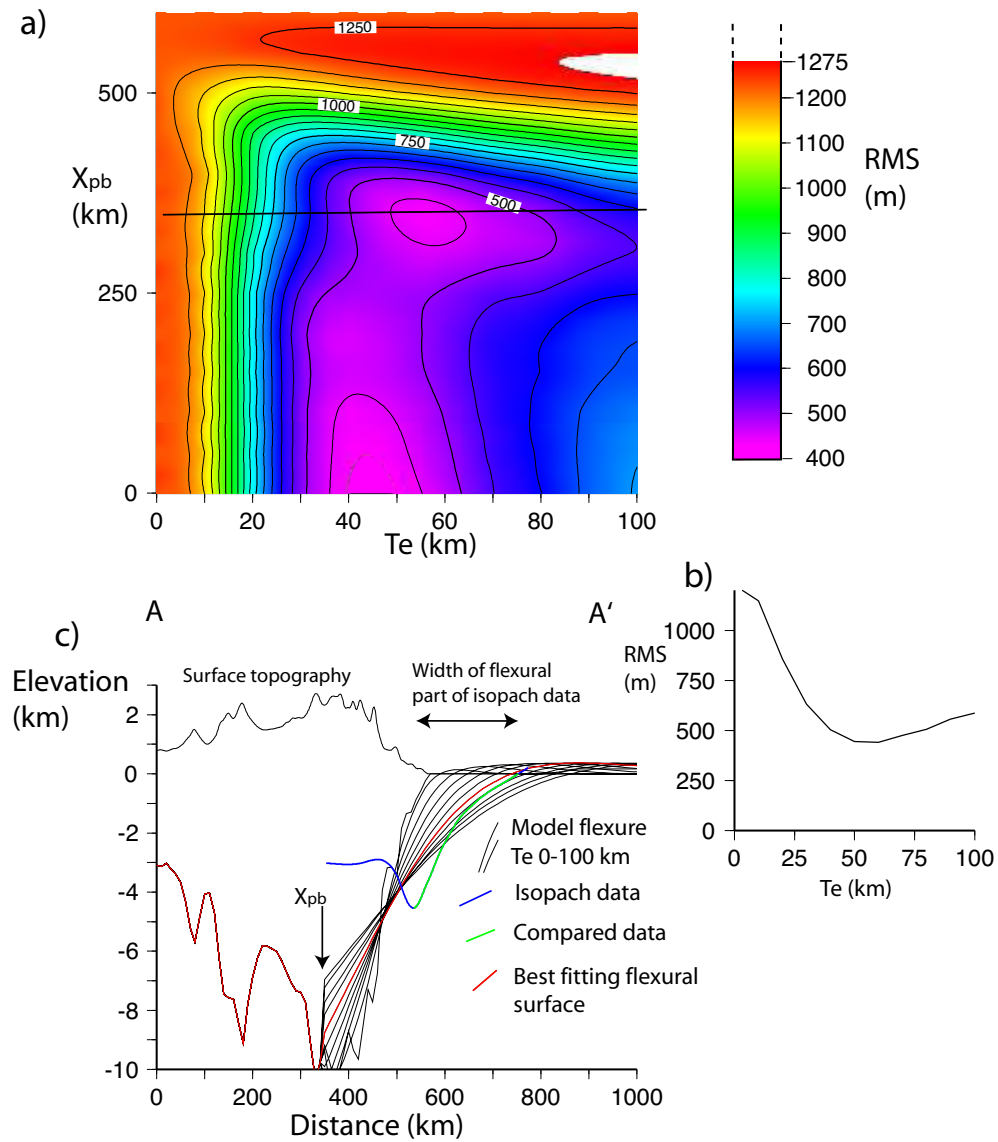


Figure 5.6: Results of modelling the flexed surface of the foreland basin along profile A-A'. a) Map of RMS difference between modelled flexure and foreland shape defined by thickness of Miocene to recent isopach. Model based on varying positions of plate break from 0 to 600 km from the origin and T_e values from 0 to 100 km. b) RMS profile for a plate break 350 km from the origin showing the best-fitting T_e for this plate break to be between 50 and 60 km. c) Modelled flexure for a plate break of 350 km and isopach data from Koop & Stonely [1982]. Blue line, all isopach data. Green line, isopach data over the foreland only. Red line, best-fitting modelled flexural profile.

5.3.3 Profile-based window technique

To further constrain the T_e structure for this profile a windowing technique was applied, capable of recovering T_e variations along the length of the profile. This technique followed the method set out for the first part of the 2D iterative technique presented in Chapter 2 but in this case applied to a profile of topographic and gravity data rather than a grid. The technique relies on a constant T_e structure giving a locally correct flexural shape, and hence gravity anomaly, for each segment of the profile assessed. Where the plate is broken there will be a transition from a calculated anomaly with high constant T_e best fitting the observed Bouguer anomaly, to a calculated anomaly based on a very low constant T_e , or Airy compensation, best-fitting the observed Bouguer anomaly. This technique is similar to that applied by Stewart & Watts [1997] to Bouguer anomaly data across the Sub-Andean fold and thrust belt. However, Stewart & Watts [1997] imposed a plate break on all the models rather than allowing a plate break to ‘appear’ as a region of low T_e .

For the profile-based window technique a series of elastic plates were modelled, with constant T_e values along their length. Each plate was loaded with the observed topography, above zero metres elevation, the expected flexure calculated, and the resulting gravity anomaly determined. Only topography to the north of the thrust front was applied as a load. This prevents the flexural bulge being treated as a ‘load’ when it is a derivative of the flexural process.

The RMS error between calculated gravity anomalies, based upon loading constant T_e profiles, and Bouguer gravity data within a series of 200 km long overlapping windows was considered. The Bouguer data for this comparison was sampled from the gridded Bouguer anomaly data to give an even point spacing along the profile. The data was also corrected for the long wavelength satellite derived anomaly discussed previously. The analysis was carried out from the origin of the profile in the north to the southern edge of the foreland, with the centre of each window

being moved 50 km each time. The first analysis window was centred on a point 100 km from the profile origin and the last was centred on a point 750 km from the origin. The best-fitting T_e for each window, as defined by the lowest RMS, was presumed to be the T_e for that section. This allowed the recovery of a T_e structure which varied along the length of the profile. Where resolution was low in the RMS data, or 2 minima appeared, such as at 650 km from the origin in Figure 5.7a, the value which best matched the pattern of variability along the profile was chosen.

The results of the windowed comparison are shown in Figure 5.7a. The results for 3 selected windows at 350, 450 and 550 km from the origin are shown in Figure 5.7b, showing the recovery of the RMS minima for each window. The combined windowed results indicate that the best-fitting T_e value is generally low beneath the mountains while high values are preferred over the foreland region. The RMS map also highlights the fact that T_e falls to around 0 km at ~ 400 km from the origin. This is the same position as the recovered plate break in the previous section, where a plate break was explicitly included in the model.

The calculated gravity anomaly resulting from loading the final variable T_e structure with the observed topography is shown in Figure 5.7c. Although the shape (curvature and amplitude) of the calculated anomaly matches the observed Bouguer anomaly well by eye, across the foreland, a DC shift of approximately -13 mgal is needed to match the observed values. This was also noted in the plate break model previously. This shift is not needed to improve the quality of the fit in the northern part of the profile. The discrepancy over the foreland could be explained by initially thicker Arabian crust compared to that in the northern Iranian region. If the southern half of the region initially had thicker crust the Moho would be relatively deeper leading to a negative shift in the gravity anomaly, relative to that seen in the north. Alternatively, the presence of a dense sub-crustal body, beneath the northern section of the profile, could account for the discrepancy.

The flexural curve generated by topographic loading of the variable T_e struc-

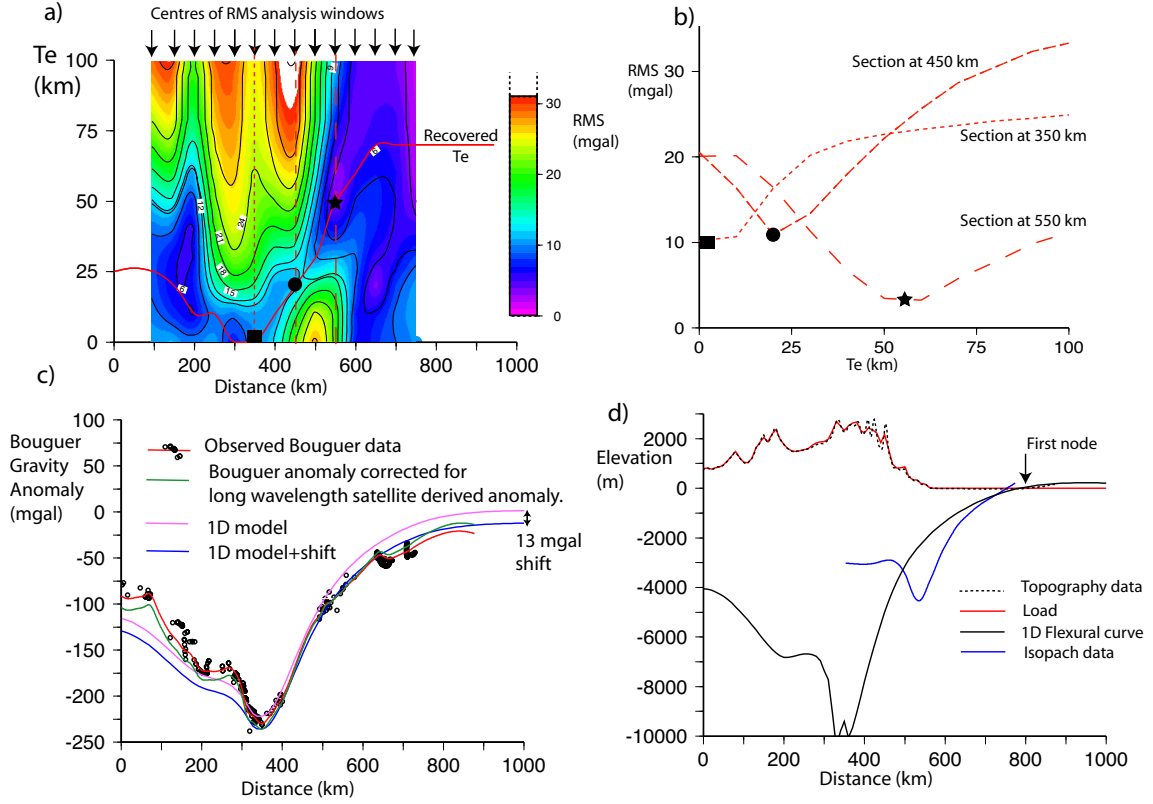


Figure 5.7: Results of analysis of a series of 200 km long segments of observed data along profile A. a) RMS values based on comparison of observed Bouguer gravity data and modelled gravity data resulting from loading of an elastic plate with constant rigidity. Red line shows chosen 'best fit' T_e structure. b) Results of RMS comparison between selected windows of calculated and observed Bouguer anomaly. Window centres were at 350, 450 and 550 km from the origin. c) Calculated and observed Bouguer gravity anomalies. Point data is shown to allow an assessment of data coverage. Models are the result of loading the 'best fit' T_e structure shown in a). d) Flexural curve generated by loading 'best fit' T_e structure and digitised isopach data. Dashed topographic line = high resolution data, red line = model load.

ture is shown in Figure 5.7d. The position of the first node, i.e. where the flexure becomes positive, correlates well with the present southern coastline of the Persian Gulf. This observation supports the T_e structure recovered by comparison of modelled and observed gravity anomalies, as it suggests the modelled width of the flexural basin is correct. However, when the Miocene to Recent isopach from Koop & Stonely [1982], which is thought to represent the main foreland sequence, is compared to the modelled flexure, the magnitude of the maximum flexure is clearly underestimated in the foreland region by up to ~ 2000 m close to the mountain front. Beneath the mountains the model flexural surface is deeper than suggested by the isopach data, however the isopach thickness in this region is not thought to be accurate due to tectonic reworking (Koop & Stonely [1982]).

Does the isopach represent a flexural surface?

To assess whether the base of the Miocene to recent isopach represents the flexural surface in the foreland region the gravity anomaly that would be produced if it did represent the flexural surface was calculated. This model assumed that the isopach thickness in the foreland represents the flexural basin shape and the Moho shape. This is correct in cases of thin plate elastic flexure as these two surfaces remain parallel. The isopach data within the mountains was not included in this calculation as in this region it cannot represent the flexural surface due to loading as it gets shallower towards the maximum load. The density contrasts assumed across the two interfaces (basin/crust and crust/mantle) were the same as those for the calculation of the gravity anomaly based on a flexural model. Figure 5.8a shows the shallow crustal bodies assumed, the Moho is considered to be parallel to the isopach defined basin but 35 km lower. For comparison Figure 5.8b shows the flexural model derived from loading the best-fitting T_e structure recovered in the previous section.

Figure 5.8c shows that the gradient of the calculated anomaly, resulting from

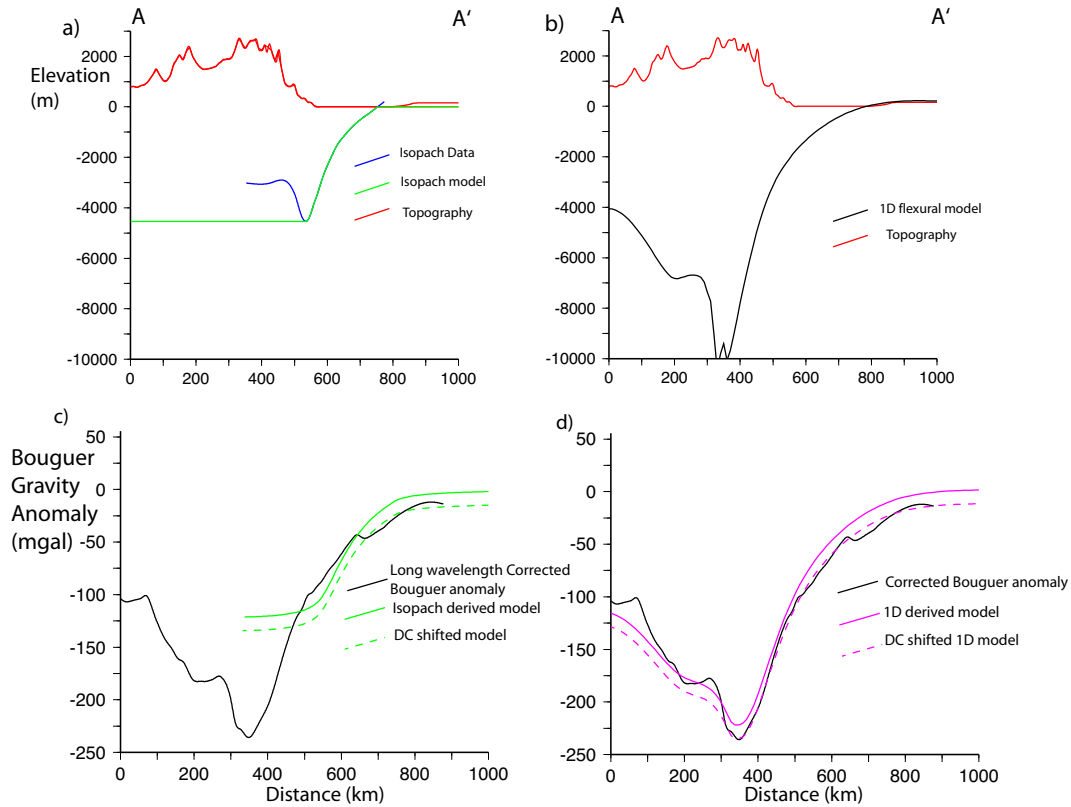


Figure 5.8: Calculated gravity anomalies, along profile A, based on either assuming the isopach shape, or flexural shape derived from loading, equates to the flexural surface. a) Model shallow crustal structure for the isopach case, the mantle parallels this. b) Flexural surface derived from topographic loading the ‘best fit’ T_e structure. c) Modelled gravity anomaly assuming isopach surface. d) Modelled gravity anomaly assuming flexural surface.

the isopach defined flexural surface, is steeper than the observed anomaly, this is most evident close to the mountain front. A closer match to the gradient of the observed Bouguer anomaly would be achieved if the density contrast at the two interfaces considered was significantly lower. However, the sediment/crust contrast of 150 kg m^{-3} is already relatively low and a crust/mantle contrast of closer to 330 kg m^{-3} in place of the more normally assumed 530 kg m^{-3} would be needed. Figure 5.8d shows how the calculated gravity anomaly derived from the modelled flexural surface fits the observed data.

The calculated gravity anomaly, assuming that the isopach represents a flexural surface, does not match the observed Bouguer anomaly. It is therefore concluded that the isopach data from Koop & Stonely [1982] does not represent a flexural surface generated by present day surface loading. This discrepancy is discussed further in the next section.

5.3.4 Profile B-B'

The same profile based technique of comparing calculated and observed anomalies within overlapping 200 km windows was applied to profile B in Figure 5.2 which lies $\sim 150 \text{ km}$ to the north-west of profile A.

For profile B the best-fitting T_e recovered over the foreland region appears to be $\sim 30 \text{ km}$ from the RMS difference between calculated and observed Bouguer anomalies (Figure 5.9a). However, this ‘best’ T_e structure produces a large flexural bulge with a first node approximately 100 km closer to the mountain front than the observed rise in the topography to the south of the foreland (Figure 5.9c). When a higher T_e value of 70 km was assumed in the foreland region, a better fit between the calculated flexural surface and the topography and the width of the foreland isopach data was observed. This is in agreement with the values of T_e suggested for the foreland in profile A. In addition the resolution of the ‘best’ T_e value in the foreland region is not good.

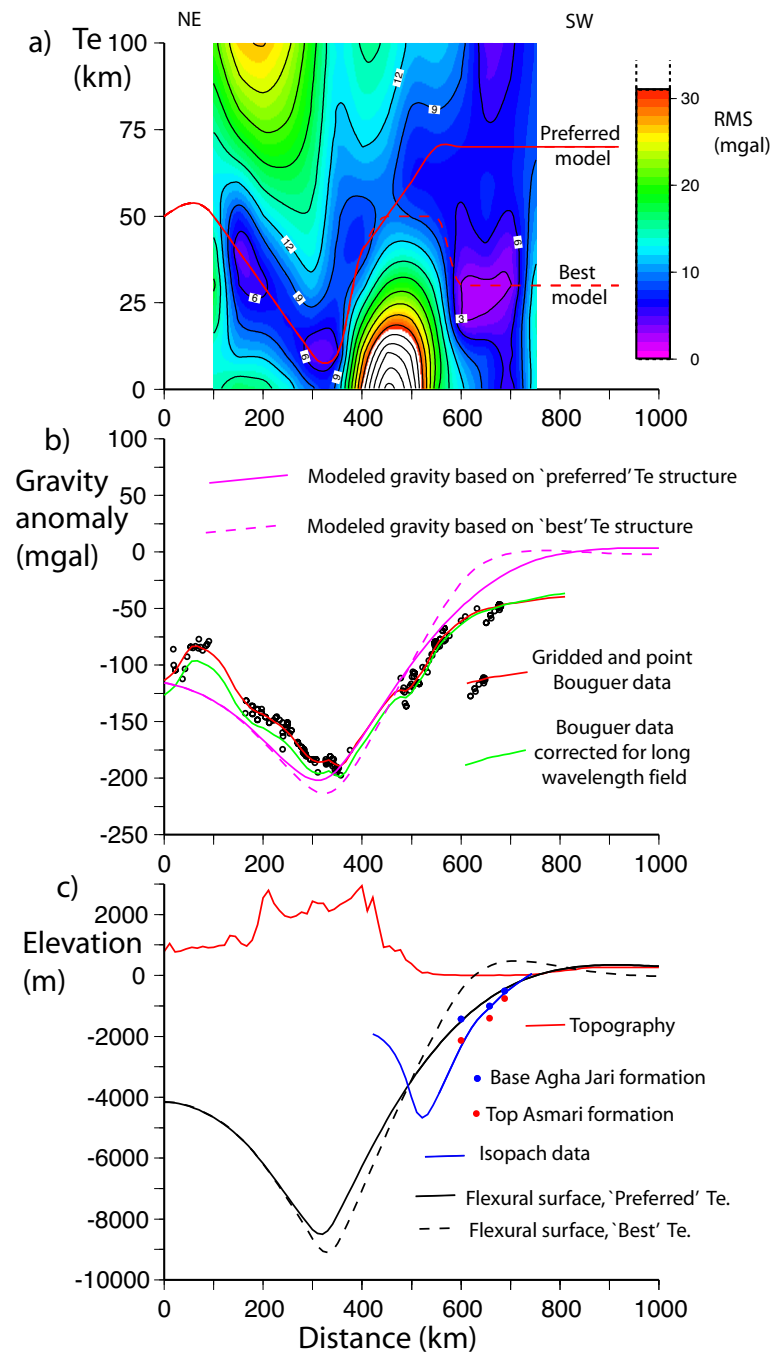


Figure 5.9: Profile B. a) Recovered T_e structure, shown are two possible best-fitting structures. b) Observed and modelled gravity anomalies. c) Modelled flexure, isopach and stratigraphic data.

Figure 5.9b shows the modelled and observed gravity data. The visual fit between the model and observed data is not as clear as on profile A, however, the lack of gravity data in the foreland makes a comparison difficult.

Figure 5.9c shows the modelled flexure, derived from loading the ‘preferred’ T_e structure. Also plotted is the digitised isopach data from Koop & Stonely [1982] and additional stratigraphic horizons taken from Alavi [2004]. The data for the depth to the top of the Asmari Formation from Alavi [2004] match the data from Koop & Stonely [1982] suggesting the older isopach data is accurate as a measure of the depth to this surface. However, Alavi [2004] shows that there are several younger unconformities across the region which could relate to flexurally driven subsidence. In particular the base of the Agha Jari formation. This unconformity is ~ 11 Ma and marks the base of a package of sediments which thicken towards the mountain front. When plotted it can be seen that the depth and pattern of thickening seen in sediments younger than the base of the Agha Jari Formation more closely match the ‘preferred’ flexural surface shown in Figure 5.9c.

Hessami *et al.* [2001] show, from an analysis of progressive deformation and unconformity patterns, how the deformation front migrated through time. It can be seen from Hessami *et al.* [2001] that in the early Miocene, when the Asmari Formation was truncated, the deformation front was around 100 km north of its present position. If the T_e structure had the same pattern as recovered today in the Miocene, with ~ 150 km of ‘load’ being applied to the end of a broken plate, the thickness of the top Asmari to base Agha Jari Formation, in the Persian Gulf, cannot represent infill of a flexural foreland. This is because the present Persian Gulf region would have been too far from the driving load.

It is therefore proposed that the observed gravity anomaly is predominantly generated by flexural loading of the Arabian shield after 11 Ma, corresponding to the deposition of the Agha Jari and later formations in the foreland basin. The isopach data from Koop & Stonely [1982] continues to be shown on later profiles as it includes data for the younger foreland sediments, and gives an impression of the overall change in foreland shape along strike.

5.3.5 Profiles C and D

Profile C lies in the far north-west of the study area, profile location is shown in Figure 5.2 and in the inset of Figure 5.10. The recovered T_e structure along this profile is similar to that recovered along profiles A and B. The match between the calculated anomaly based upon the recovered T_e structure and the observed Bouguer anomaly is not as clear as in the previous profiles. This may in part relate to basement involved thrusting bringing denser rocks closer to the surface in this region altering the observed gravity anomaly. Blind thrusting within the foreland is implied for example by Berberan [1995] who shows the presence of seismogenic faults within the foreland, which have little or no associated topographic relief.

There is no observed data over the foreland region for this profile and this prevents an accurate assessment of the foreland rigidity. However, using a T_e of 70 km in the foreland region gives a good fit between the observed rise in topography and the modelled first node shown in Figure 5.10c.

Profile D lies ~ 150 km to the south-east of profile A, profile location is shown in Figure 5.2 and in the inset of Figure 5.10. Profile D exhibits a similar recovered T_e structure to profile A. However, the modelled gravity anomaly shown in Figure 5.10e does not match well the data over the foreland with the calculated anomaly being consistently more positive than the observed. The discrepancy may be caused by the presence of lower density sediment close to the surface, in the Persian Gulf region, than assumed in the calculation of the gravity anomaly from the flexural model.

5.3.6 Profiles E, F and G

Profiles E, F and G, lie further to the south-east than the four profiles discussed so far, profile locations are shown on the inset map in Figure 5.11 and in Figure 5.2. These three profiles appear to have a different character to those discussed so far. Applying the same 1D windowing technique used on profiles A-D recovers the T_e

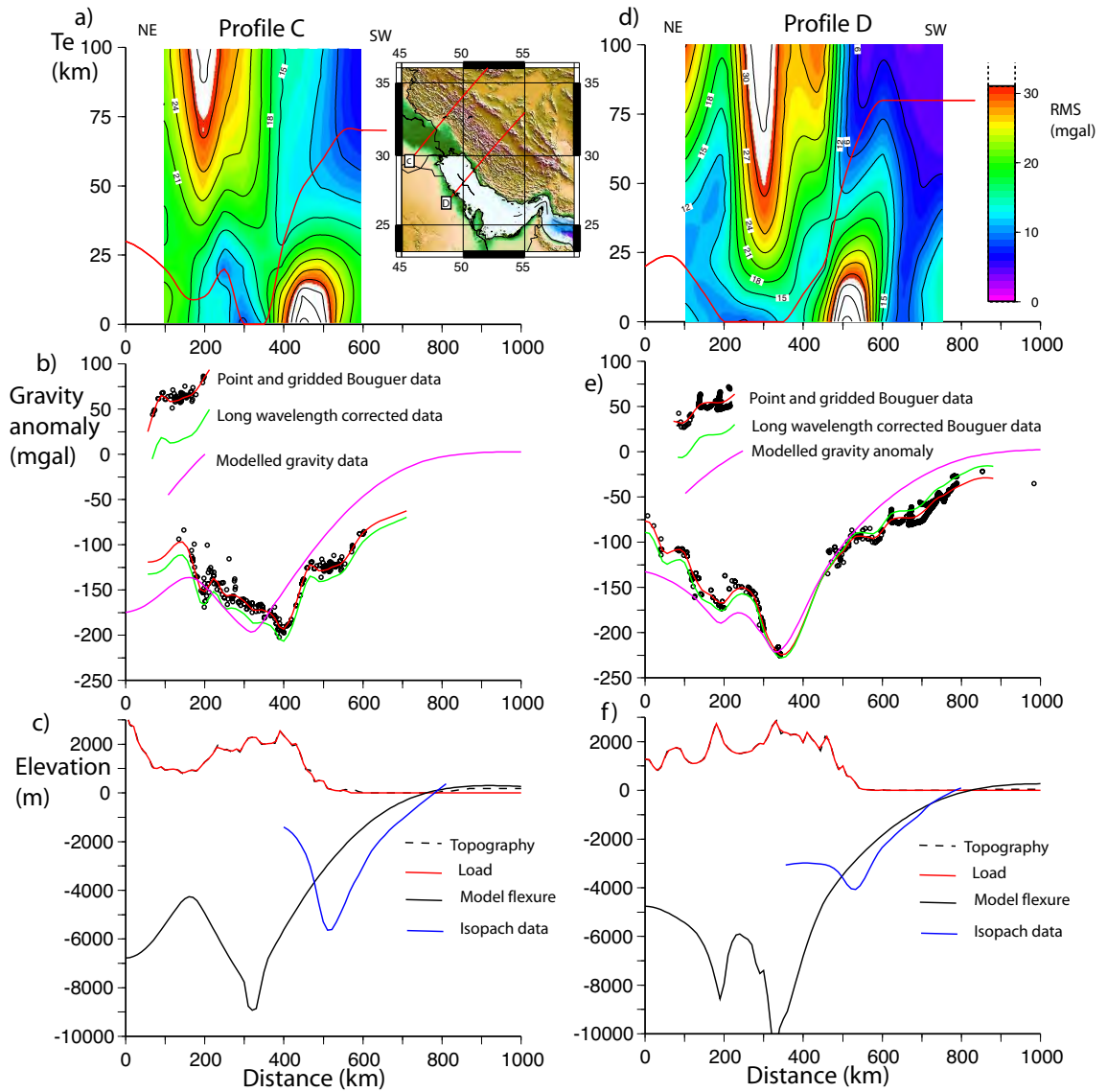


Figure 5.10: Profiles C and D. a) RMS map and recovered T_e structure for Profile C. b) Observed and modelled gravity anomalies, based on recovered T_e structure, for Profile C. c) Modelled flexure and isopach data for Profile C. d) RMS map and recovered T_e structure for Profile D. e) Observed and modelled gravity anomalies, based on recovered T_e structure, for Profile D. f) Modelled flexure and isopach data for Profile D.

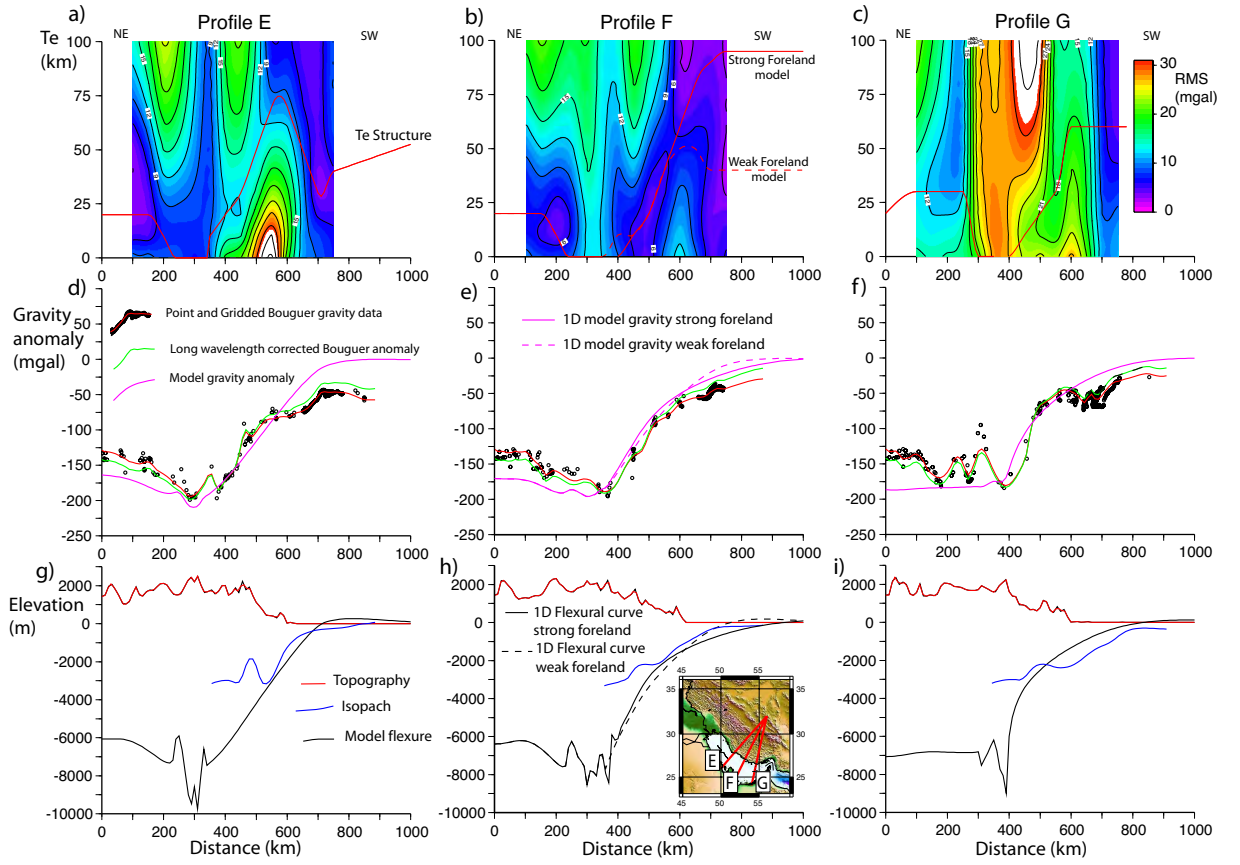


Figure 5.11: Profiles E, F and G. Top three figures show recovered T_e structure. Middle three figures show modelled and observed gravity. Lower three figures show modelled flexure and isopach data.

structures, shown in Figure 5.11a, b and c. However, calculated gravity anomalies, derived from these T_e structures, do not appear to give a good fit to the observed gravity data shown in Figure 5.11d, e and f. This result is irrespective of whether a DC shift is applied to the data or not.

Profiles E and G show a pronounced steepening in the observed Bouguer gravity anomaly beneath the mountains at around 500 km from the start of the profiles. Profile F shows a more gradual slope to the Bouguer anomaly. However, the RMS plot does not show a clear resolution of the T_e which suggests that lower or higher T_e values could fit equally well in this region.

The isopach data, shown in Figure 5.11g, h and i, also appear different to earlier profiles. Profiles A-D show a broad foreland approximately 200 km wide. The depth

of the foreland in lines A-D, from the isopach data, is between 4 and 6 km and consistently exceeds the modelled flexure. The isopach data below profiles E-G is shallower than in profiles A-D, with maximum thicknesses of ~ 3 km occurring. The foreland basin defined by the isopach data is therefore shallower than the modelled flexure in this region, and along profiles E and F the foreland appears to be less than 100 km wide. This difference in foreland shape along strike of the mountain front can be clearly seen in map view in Figure 5.3d. This shows that towards the south-east of the Zagros mountains the deformation front is overlying a Miocene surface < 1000 m deep, rather than over 4000 m deep as seen in the north-west.

On profile F an alternative very strong foreland was considered with a T_e of over 90 km. This gave a relatively good fit to the low angle slope of the gravity anomaly in the foreland region. However, the extent of the foreland flexure exceeds that seen in the isopach data which is regarded as the absolute maximum depth of the modern foreland. This discrepancy is discussed further in Section 5.5.2.

5.4 Grid based modelling

To further assess the spatial variability of the T_e structure across the Zagros collisional system the 2D iterative technique for recovering spatially varying T_e across broad regions, described in Chapter 2 and applied to the India-Eurasia collisional system in Chapter 4, was implemented. As the region considered here is smaller than in Chapter 4, a finer grid with a point spacing of 20 km was used. This allowed the preservation of resolution across smaller structures. The Gaussian filter width was also reduced from that used in the Indian case study to 200 km. For filter window sizes below this the finite difference technique for calculating flexure began to fail due to steep gradients in the T_e structure.

Input data

A 2D grid of Bouguer gravity anomaly data was constructed based on the point Bouguer data provided by the Bureau Gravimetrique International (B.G.I.) discussed previously in Section 5.2. In addition point data from the field-work in Oman (presented in Chapter 3) was included in the construction of the grid. Regions where data gaps persisted were supplemented using data from the ‘marine gravity from satellite altimetry data set’ by David T. Sandwell and Walter H.F. Smith available from Scripps Institution of Oceanography, described by Sandwell & Smith [1997]. Onshore this data is based on the EGM-96 model of the gravity field complete to degree and order 360 (Lemoine *et al.* [1998]). As this data is free air data the simple Bouguer correction was made using the GEBCO [2003] 1-minute data-set which incorporates GLOBE [1999] onshore topography data. This topographic data has a shorter wavelength than the EGM96 data and hence may introduce artefacts into the input grid. The final Bouguer gravity grid was created using the **GMT** functions **blockmean** and **surface** with a grid spacing of 5 minutes and a tension factor of 0.25. This final grid was, in addition, corrected for the same long wavelength, degree and order 20, satellite derived gravity anomaly as the data used in the profile study presented in Section 5.2.

The topographic data was taken from the GEBCO [2003] 1-minute grid. The data was clipped so only topography above zero metres remained. This allowed the consideration of a simple model, based purely on surface topographic loading.

All data was projected into a Cartesian coordinate system and re-sampled onto a 20 km grid. The edges of the data were smoothed using a cosine taper and the data padded with zero values up to 500 km from the load edge. This was carried out to prevent edge effects when the gravity anomaly was calculated.

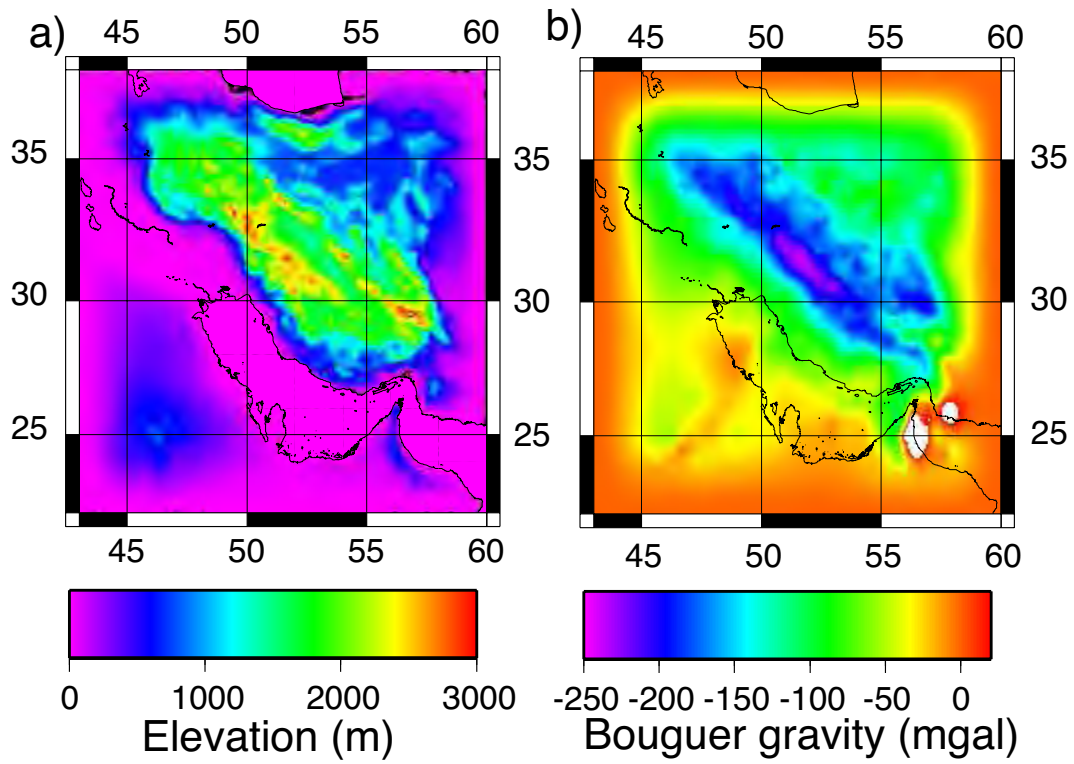


Figure 5.12: Tapered topography and gravity data sets. a) Elevation data showing tapered edges and clipping to 0 in offshore areas. b) Bouguer gravity anomaly data.

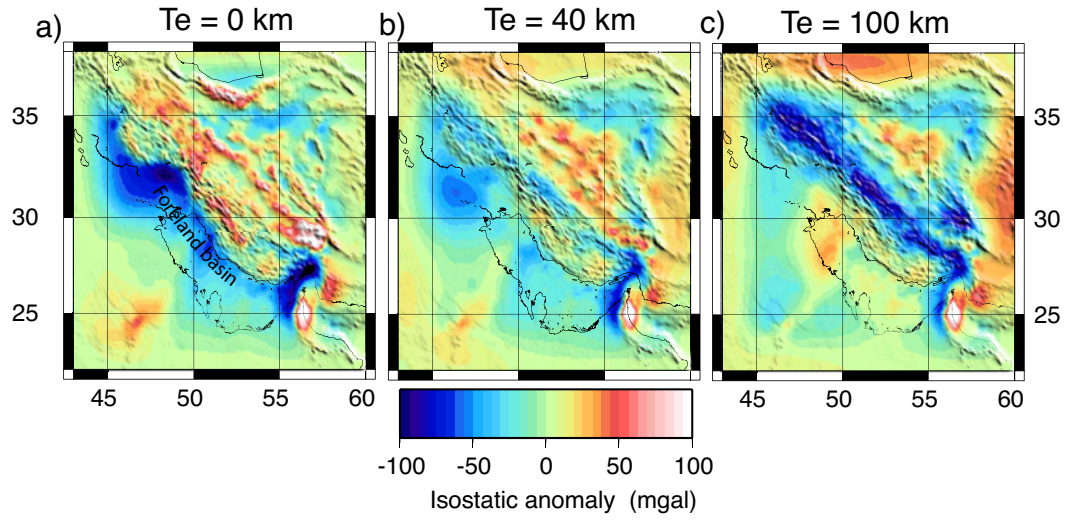


Figure 5.13: Isostatic anomalies based on comparison of calculated and gridded Bouguer anomaly data shown in Figure 5.12b. Topographic gradient data is overlain for reference. a) Airy Isostatic anomaly. b) $T_e = 40$ km, isostatic anomaly. c) $T_e = 100$ km, isostatic anomaly.

5.4.1 2D Results

Isostatic comparison

An initial assessment of the type of compensation in the region was made by comparing observed Bouguer gravity anomalies to calculated anomalies based on models of isostatic compensation of the surface topography, above 0 m elevation, with constant T_e structures. The isostatic anomalies were calculated for a T_e of 0 km (local Airy compensation), 40 km and 100 km and are shown in Figure 5.13.

The Airy ($T_e = 0$ km) case shows a prominent negative isostatic anomaly along the length of the foreland shown in Figure 5.13a. The maximum anomaly of ~ -100 mgal is seen in the north west part of the foreland and its magnitude decreases along the length of the basin to ~ -35 mgal at the south-east end of the foreland. The negative anomaly is highly asymmetric across strike with values becoming more negative towards the Zagros mountains which lie to the north east of the foreland. This is typical of a flexurally derived foreland basin as the largest de-

viation between the Airy and regional isostatic compensation models will be seen close to the margins of the load.

At a constant value of T_e of 40 km the negative isostatic anomaly over the foreland region persists, indicating that the isostatic support of the topography in this region is not matched by this T_e . Below the centre of the Zagros mountains a linear negative appears of up to ~ -45 mgal. This indicates that not enough deflection of the plate occurs beneath the mountains to generate a calculated anomaly that will match the observed Bouguer anomaly when T_e is this high.

When a constant T_e structure with a value of 100 km is considered, the negative isostatic anomaly in the foreland region, seen in the earlier models, is largely eliminated. This suggests that this high value more closely approximates the actual T_e in the foreland area. However, beneath the central part of the Zagros mountains a strong negative isostatic anomaly is present of up to ~ -90 mgal indicating that, for this model of isostatic support, the central part of the Zagros mountains is not accurately recreated.

Iterative results

The iterative part of the 2D iterative technique was run through 40 cycles to produce the best-fitting gravity anomaly shown in Figure 5.14a. The final modelled gravity anomaly was compared to the observed Bouguer gravity data and the regional isostatic anomaly calculated and is shown in Figure 5.14b.

Figure 5.14b shows that the iterative technique has reduced the overall isostatic anomaly relative to the anomalies produced for the constant T_e structures. This is particularly clear when the foreland and central mountains are considered, as no constant T_e structure could match both these regions. However, no clear reduction in the isostatic anomaly is observed in the Oman region and to the north. An additional view of the reduction in the isostatic anomaly can be derived from the power-spectra of the isostatic anomalies shown in Figure 5.14c. Power-spectra

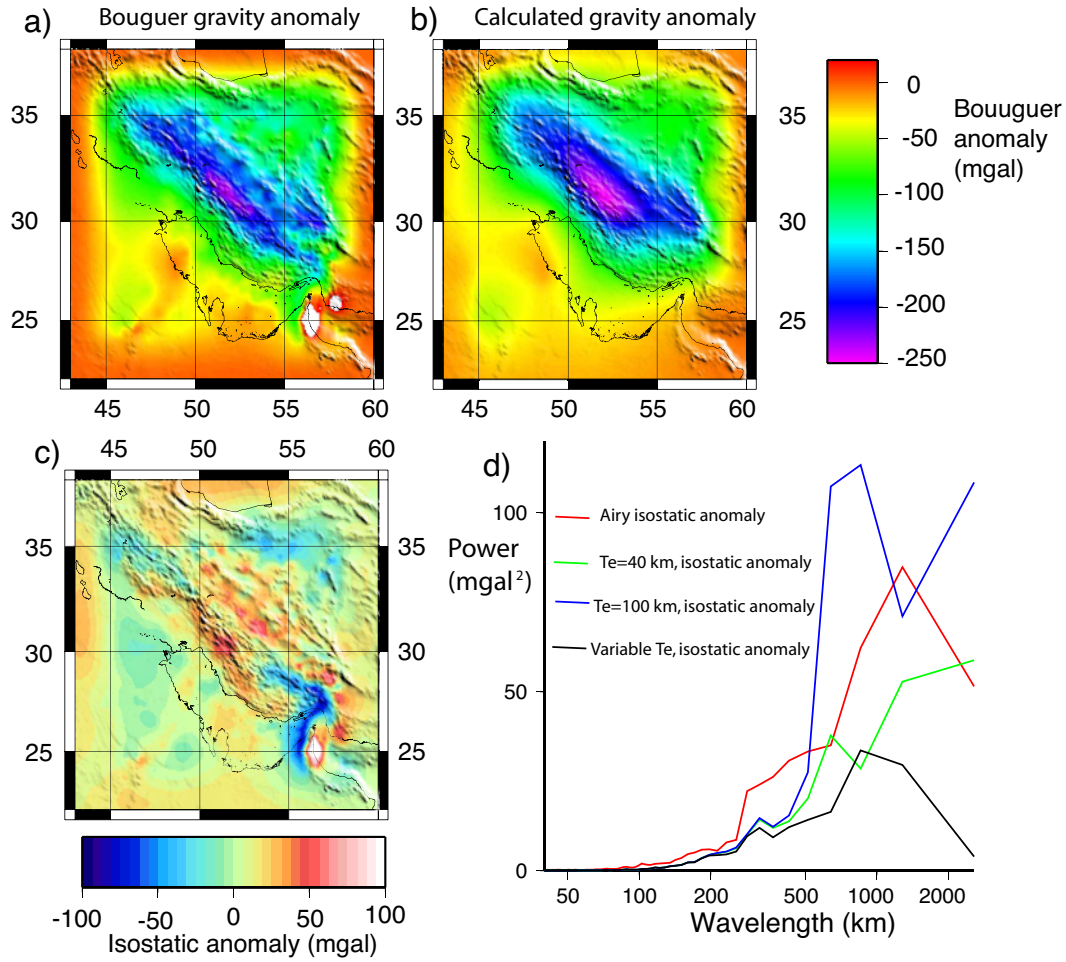


Figure 5.14: Modelled gravity anomaly and isostatic gravity anomaly after 40 iterations of the 2D iterative technique. a) Final modelled gravity anomaly based on loading a 2D variable T_e structure. b) Isostatic anomaly between observed and modelled gravity anomaly. c) Spectra of constant rigidity isostatic anomalies and regionally variable isostatic anomaly.

of the isostatic anomalies are assessed as they give a simple way to compare the reduction in the residual isostatic anomaly derived from the different compensation schemes, i.e. Local (Airy) compensation, regionally constant T_e compensation, or regionally variable T_e compensation. The spectra show that the regionally variable T_e structure gives the lowest amplitude power-spectra. This confirms that for most wavelengths the regionally variable T_e structure gives the best-fit to the data. The cause of the discrepancy close to 1000 km wavelengths where a constant T_e of 40 km gives the lowest power is not clear.

5.4.2 Statistics

Sensitivity and stability tests were carried out on the results of the 2D iterative model, as in Chapter 2. Figure 5.15a shows the sensitivity of the model to variations in T_e . High sensitivity equates to a deep, well defined, RMS minima while low sensitivity means that variation of T_e has little effect on the RMS between modelled and observed data. Sensitivity is non-zero over much of the model, making interpretation of the results valid. Clearly sensitivity is high beneath the mountains where there is a significant driving load.

Figure 5.15b shows the stability of the model based on how much the T_e structure changed over the final 5 iterations. Over much of the central region beneath the mountains the model appears to have stabilised. However, over the south-eastern edge of the region this is not the case.

T_e structure

The T_e structure recovered after 40 iterations of the 2D iterative technique shown in Figure 5.16a is highly variable across the region. However, many parts of the model are unconstrained by land gravity data and also, in some regions, the model has not stabilised. Regions of the T_e structure where there was no land data coverage, or the standard deviation of T_e with time was greater than one, suggesting poor

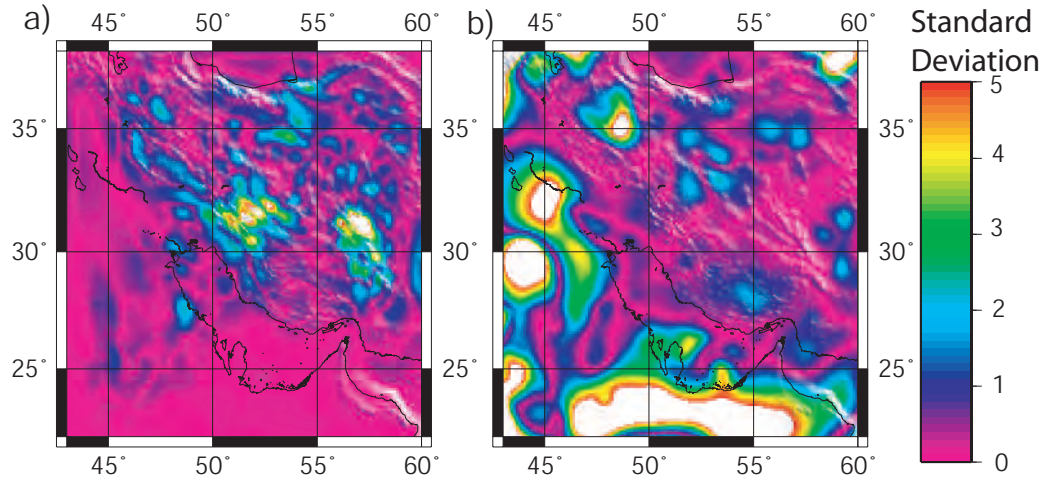


Figure 5.15: Statistics for the 2D iterative technique. a) Sensitivity to T_e variations (standard deviation of RMS with T_e variation of ± 4 km). b) Stability of model after 40 iterations (standard deviation of T_e over last 5 iterations).

stability, were masked out as shown in Figure 5.16b.

The T_e structure shown in Figure 5.16b, recovered using the 2D iterative technique, shows large variations in T_e across the region. Beneath much of the Zagros mountains and in the north east of the study area T_e appears low (~ 0 -15 km). North of the Zagros mountains there is a region of intermediate T_e with values rising up to ~ 40 km. To the south of the Zagros mountains, in the central foreland region, T_e is higher ranging from 75 to over 100 km north of Qatar. Towards the north-west end of the foreland, close to the Dezful embayment, T_e falls to values of closer to 30 km.

Within the Zagros mountain chain, towards the south-east end, there is a region of pronounced high T_e , with values of ~ 75 km recovered. This area correlates with the region with a pronounced anomaly in the regional isostatic anomaly map (see Figure 5.14). It also correlates with the strong positive Airy isostatic anomaly shown in Figure 5.13a. The Airy isostatic anomaly suggests a dense buried load may be present in this area. Molinaro *et al.* [2005] show, based on object orientated models of the free air gravity anomaly, that this region has anomalous dense

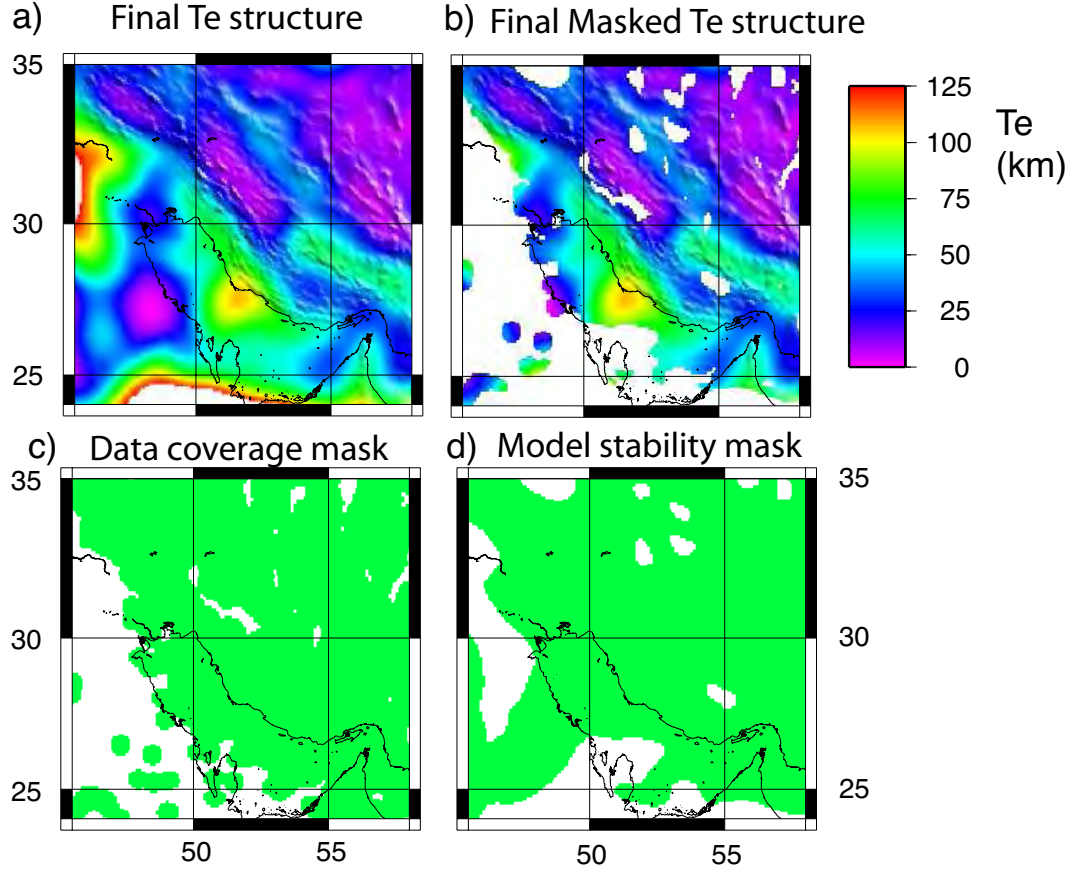


Figure 5.16: Recovered T_e structure and masks applied to it. a) T_e structure after 40 iterations of the 2D model. b) Final T_e structure. c) Data coverage, used to constrain reliable T_e estimates. Green areas have data, white areas have no data. d) Stability mask used to further constrain the T_e estimates. Green regions show stable areas.

upper crust and volcanic material present at shallow levels. These features would essentially act as buried loads and prevent the 2D iterative model attaining an accurate result. The results for the Oman peninsula region, where a buried load is known to be present, are not considered reliable for the same reason. In Chapter 2 synthetic tests with uncorrected buried loading showed anomalous low T_e values to be recovered. In this region the buried, positive load appears to be strongly correlated with elevated topography. In this case the model will return excessively high T_e values because they locally reduce the amount of isostatic compensation of the topography, and hence a less negative anomaly will be calculated which is closer to the observed trend.

5.4.3 Comparison of profile and grid based recovery of T_e

Figure 5.17 shows the comparison of the 1D profile and 2D techniques along profiles A and F.

Along profile A the 2D iterative technique recovers a similar T_e structure to the profile based segmented technique. The 2D technique recovers its lowest values around 350 km from the origin, close to the position for the plate break recovered by the ‘broken plate’ model applied previously. Associated with the similar T_e structures recovered for profile A, similar calculated gravity anomaly are recovered. Over the foreland region the 2D calculated gravity anomaly is shifted down by around 25 mgal. This shift is related to the presence of the flexural bulge acting as a load in the 2D model but not in the profile model. The maximum T_e recovered by the 2D technique occurs closer to the mountain front and is higher relative to the profile T_e . However, the 2D technique recovers a low T_e on the south west section of the profile. In this region there is little data so this result may not be reliable. To compensate for this low T_e in the bulge region the 2D model produces a higher T_e over the foreland region to match the observed gravity anomaly.

Figure 5.17 also shows the comparison of the profile and 2D techniques along

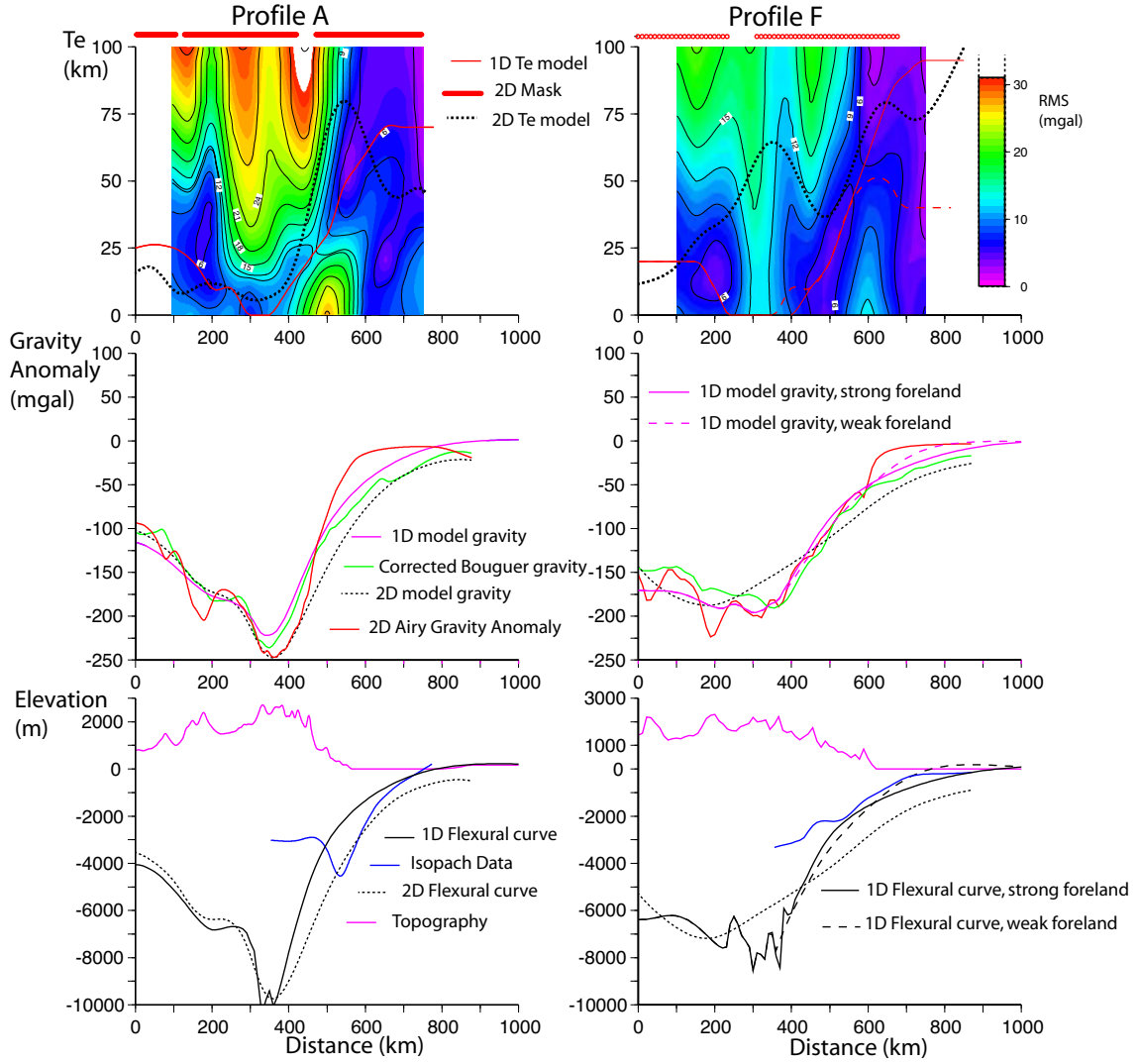


Figure 5.17: A comparison of the T_e structure, gravity anomalies and flexure derived from the 2D regional model and the 1D models along profiles A and F.

profile F. The T_e structure recovered by the profile and 2D techniques clearly differ in this area, particularly beneath the highest part of the mountains. The high values of T_e recovered beneath the high mountains by the 2D iterative technique are anomalous because of the presence of dense buried loads as discussed in the previous section.

In the foreland region to the south of the mountains, along profile F, the 2D method matches the shape of the Bouguer gravity data well, however beneath the mountains the fit is poorer. Beneath the frontal part of the mountain belt (350 to 600 km along the profile) the profile technique appears to fit the data well, however, a 2D Airy model also appears to fit the data. This, it is suggested, relates to the long wavelength of the observed topography. The anomaly generated by flexure beneath very long wavelength topography will tend to generate an anomaly close to that predicted by Airy isostasy even if T_e is high. This effect was discussed in Chapter 4 in relation to the south-east margin of the Tibetan plateau where low (<5 km) T_e values had been recovered in a region of long wavelength topography. However, in south-east Tibet the model sensitivity was high and short wavelength features could be seen in both the calculated and observed gravity anomalies which suggested a low T_e model was indeed correct. In profile F there do not appear to be any matching short wavelength features between the Airy model and observed data and sensitivity is very low. T_e therefore remains unresolved.

From 0 to 300 km along profile F neither the profile or 2D model fits the observed Bouguer anomaly well, although the profile model achieves the closest match to the shape. This poor fit relates to the presence of buried loads that generate a gravity anomaly that a simple surface loading flexural model can not fit.

5.5 Discussion considering geological context

The profile and 2D techniques for recovering T_e variation in this region show T_e to vary from generally high beneath the foreland to low beneath the mountains.

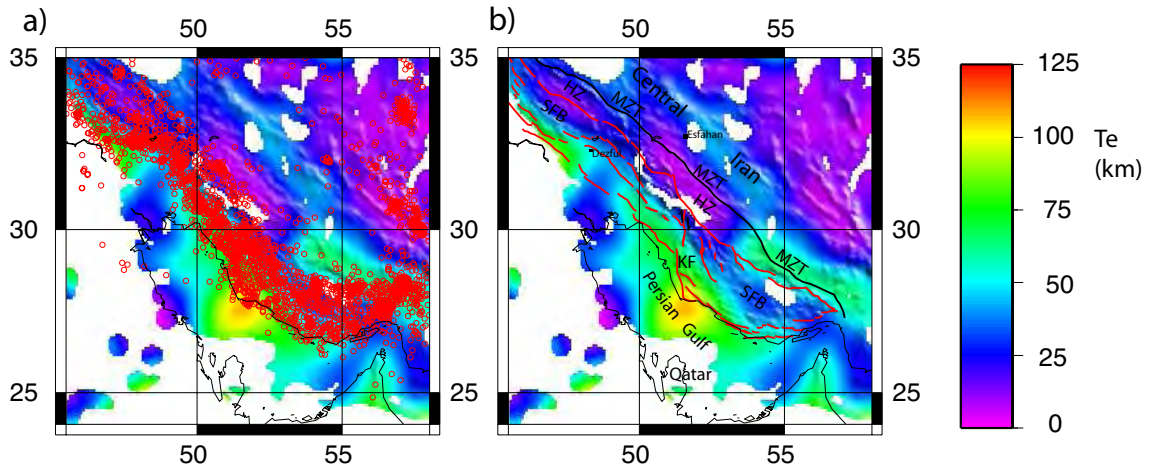


Figure 5.18: Geological and earthquake data for the Iran and Persian Gulf region plotted upon the recovered T_e structure. a) T_e map and earthquake location data for events over magnitude 4 from the USGS [2004] database. b) Main geological elements across the Zagros region from Talebian & Jackson [2004]. SFB, Simply Folded Belt; HZ, High Zagros; MZT, Main Zagros Thrust; KF, Kazerun Fault.

There appears to be considerable along strike variability in the T_e structure and in the reliability of its recovery. How this correlates with along strike variations in gravity, foreland basin shape, topography and geology will be discussed below.

5.5.1 The central Zagros mountains

Central Iran, north of the Main Zagros Thrust (MZT), appears rigid, relative to the region to the north east. This region correlates with the relatively aseismic region of the central Iran block of Jackson *et al.* [2002]. This block is Cambro-Ordovician in age and correlates with the Sanandaj-Sirjam belt which was marginal to the Palaeo-Tethys (Husseini [1992]). Further to the north east seismicity increases and T_e decreases.

Towards the mountain front, south of the High Zagros (HZ) region, the T_e increases towards its maximum close to the southern edge of the simply folded belt (SFB). This rise in T_e correlates with the zone of intense earthquake activity. Typically significant earthquake activity is associated with low T_e values, for example

beneath the Tibetan plateau as discussed in Chapter 4. However, south of the HZ most earthquakes are less than 20 km deep with the majority being less than ~ 10 km deep (e.g. Maggi *et al.* [2000b] and Tatar *et al.* [2004]). The earthquake distribution and high T_e (~ 60 km) suggests that the overlying material within the SFB is failing and deforming while the under-thrusting plate remains rigid and relatively undeformed. The region of high T_e beneath the mountains is therefore likely to represent the distance the rigid Arabian shield has penetrated beneath the mountains.

Between the HZ and the frontal thrust in the region of, $\sim 50^\circ\text{E}$ and 53°E , there appears to be a consistent positive isostatic residual of up to ~ 50 mgal between the final gravity anomaly calculated from the iterative model and the observed Bouguer anomaly. This may relate to dense material within the crust which is not included in the isostatic model. East of $\sim 53^\circ$ there is a clear positive residual anomaly of over 50 mgal and an associated negative which correlates with the MZT. This region also exhibits high recovered T_e values, however, these results are not thought to be reliable due to the presence of potentially significant buried loading in this area as proposed by Molinaro *et al.* [2005] and discussed previously.

5.5.2 Variations in foreland T_e and structural style

The T_e recovered by the 2D iterative technique in the foreland region varies along strike from ~ 40 km, close to Dezful in the north east, to over 100 km north of Qatar. T_e values reduce to ~ 75 km further to the south-east.

Lower values of T_e would be expected in the Dezful region as the isopach data suggests the foreland is at its deepest here. Given the loading mountains are approximately uniform in height the simplest explanation for this increase in depth would be a decrease in strength. A decrease in T_e in the foreland region is not clear from the results along profile C which runs through this region. In the profile-based model a high T_e of ~ 70 km is preferred in the foreland. This high value is needed

to fit the position of the first node seen in the topography. An explanation for the discrepancy between the profile and 2D results is that the profile model does not take into account loads which are off the profile. This is significant in the Dezful region as the mountain front, and hence the assumed load front, is not linear. Out of plane loads, east and west of the profile and further south into the foreland, would cause the flexural first node to appear further south for a lower T_e . Blind thrusts within the foreland may be acting as additional loads in this region, generating a deeper foreland with a more negative associated Bouguer gravity anomaly. The presence of such faults has been shown by, for example, Talebian & Jackson [2004]. However, the Airy isostatic anomaly in the Dezfu region does not feature a large positive anomaly as would be expected if buried loading was a very significant feature.

The structural style in the Dezful region is different to other parts of the foreland system. The front of the mountains is not coincident with the frontal thrust, which lies further south within the foreland. The SFB, separating the mountain front and the MZT, is at its narrowest in this region. These features may be a consequence of the low T_e in the region altering the foreland shape. The steeper flexural foreland, caused by loading a lower T_e plate, would provide more resistance to the overthrusting orogen. This would slow forward migration of the topographic front and hence a narrower zone between the initial suture zone (MZT) and the topographic front to the south. This is similar to the explanation proposed by Stewart & Watts [1997] for the variation of structural styles in the Andean foreland where more thin skinned folding and thrusting was seen associated with the more rigid foreland than in parts of the orogenic belt with weaker forelands. The position of the thrust fault within the foreland may be related to lateral propagation of the thrust from regions to the north east and south west where the frontal thrust and mountain front are coincident.

The presence of higher rigidities north of Qatar may represent the rigidity of the

Precambrian basement associated with the Qatar Arch (Edgell [1996]). This region represents a ridge of basement material that appears to have remained unaffected by the Precambrian rifting which led to the formation of the salt basins seen each side. In this region the mountain front appears approximately coincident with the thrust front suggesting the mountain belt is more easily over-riding the Arabian shield.

The T_e in the foreland region east of Qatar is some-what ambiguous in that it is hard to distinguish between a high and a low T_e beneath the SFB. In this region, along Profile F, a high T_e leads to a calculated anomaly which appears to be a good approximation to the observed Bouguer anomaly. However, the flexural model needed to generate the calculated anomaly has a greater magnitude of flexure in the foreland region than suggested by the isopach data, which marks the absolute maximum deflection. A resolution to this discrepancy comes when the process of mountain building in this region is considered. It has long been thought that the Hormuz salt has a controlling role in the deformational style of the Zagros fold and thrust belt (e.g. Bahroudi & Koyi [2003]). The models of Bahroudi & Koyi [2003] show that the presence of a viscous decollement layer, such as a salt deposit, causes crustal shortening to be dominated by folding. When there is no ductile layer, and the decollement surface is frictional, crustal shortening is accommodated by thrusting and thrust stacking. A sketch of these two models is shown in Figure 5.19.

Isostatic models state that any significant surface topography must be either locally or regionally isostatically supported by downward deflection of the Moho. However, most models of deformation and building of topography above a ductile detachment simply have material moving over a horizontal rigid surface. These, therefore, do not consider the compensation of the surface topography. Where topography is built by folding, isostasy would suggest that the frontal folds should subside as the topography behind is isostatically compensated. This is not seen in the Persian gulf region, instead rising folds typically mark the start of the defor-

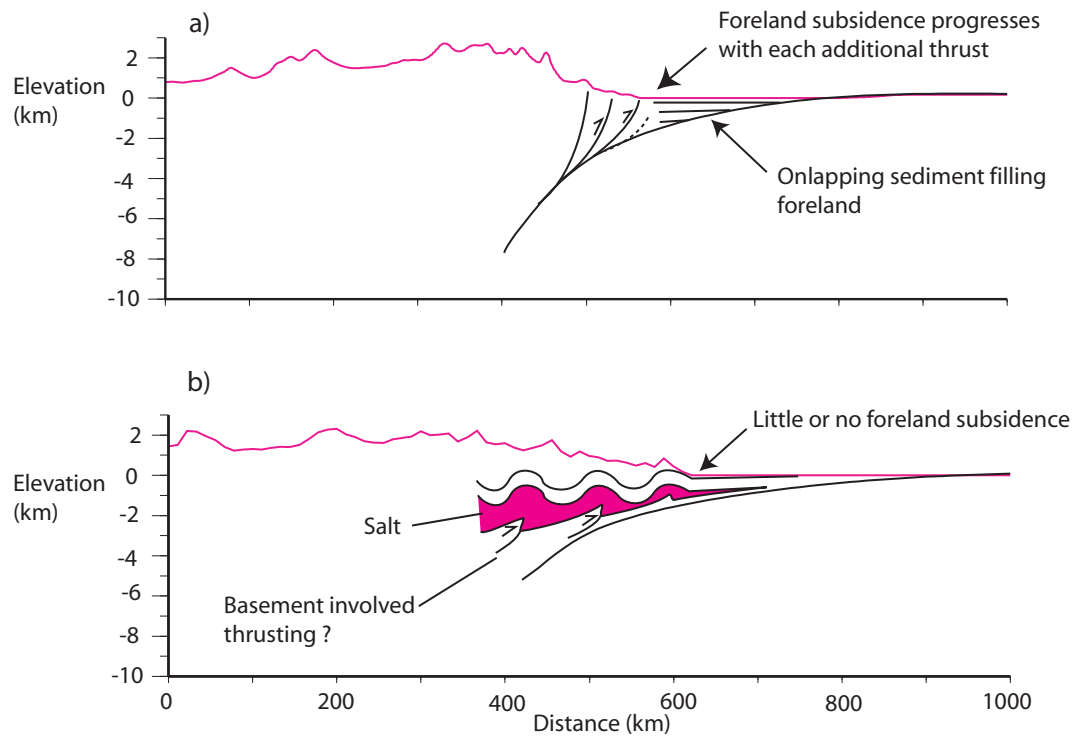


Figure 5.19: Models of crustal thickening. a) Thickening by stacking of thrust sheets. b) Thickening by folding above a viscous decollement. Surface topography and resulting flexural shape are taken from profiles A and F respectively.

mation. This can be explained if the basement in the foreland is also de-coupled from the surface stratigraphy by salt. This is shown in Figure 5.19b. As the plate flexes to compensate for the topographic load, salt moves from beneath the mountains and fills the space created within the foreland, preventing extensive foreland subsidence.

The result of de-coupling flexure from foreland generation can explain the discrepancy between the observed gravity anomaly, which can be modelled flexurally, and the lack of deflection of the foreland seen in the isopach data. In this region the modelled foreland basin only contributes ~ 8 mgal to the overall negative anomaly at the mountain front, while Moho deflection contributes ~ 35 mgal to the negative anomaly. Therefore, absence of a foreland does not significantly alter the longer wavelength anomaly derived predominantly from deflection of the Moho.

The best 1D model for the isostatic support of the surface topography along profile F is the one with a high T_e of ~ 90 km in the foreland region. This agrees with the results of the 2D iterative technique in this region which recovers T_e values of ~ 75 km.

5.6 Oman, Zagros comparison, T_e and yielding

The two studies of the Arabian shield, based on the Zagros foreland, presented in the current chapter, and the Oman foreland in Chapter 3 recover very different T_e values. In the Zagros foreland T_e values of over 60 km were recovered while in Oman the highest value recovered was 30 km. Lower values of T_e in Oman have previously been attributed to rifting processes (Patton & O'Connor [1988]), however, in northern Oman and the U.A.E. the two forelands are very close and almost overlap. An alternative explanation for the low T_e recovered in Oman may relate to the curvature of the lithosphere. It is known that the curvature of the lithosphere can induce failure, and a reduction in the recovered T_e relative to that predicted from the assumed thermal and lithological properties of the lithosphere

(Watts & Burov [2003]).

Figure 5.20a and b shows two flexural profiles, based on the broken plate model: profile A-A' from Chapter 3 across the U.A.E. part of the Oman ophiolite foreland, and profile A-A' from the current chapter across the Zagros foreland. Also shown is the curvature of the flexural surface. The foreland in Oman has experienced flexural bending to a greater extent, and hence the curvature seen in the foreland is much larger.

Figure 5.21 shows the reduction of T_e with increasing curvature, based on a modelled continental type yield strength envelope, from Watts & Burov [2003]. When the recovered T_e values are plotted against their respective curvatures they can be seen to lie approximately on the predicted curve for reduction in T_e . This suggests that, at least in the north, the Oman ophiolite may have crossed the weak passive-margin and been obducted onto strong continental crust. The low recovered T_e therefore relates to the large buried load present in the case of the ~ 10 km thick Oman ophiolite relative to the smaller load of the 2-3 km high Zagros mountains.

5.7 Summary remarks

There are significant T_e variations both across and along strike of the Zagros mountain range and foreland system.

Typically T_e in the foreland region is high (50-90 km), though the 2D iterative technique suggests this falls to closer to 30 km in the Dezfu embayment. In regions with higher T_e the thrust front and the topographic front are closely associated and the fold and thrust belt is relatively wide. Where T_e is low, as in the Dezfu embayment, the frontal thrust appears to have propagated into the foreland without significant forward migration of the topographic front. The associated fold and thrust belt to the north east of this region is narrower than elsewhere along the foreland.

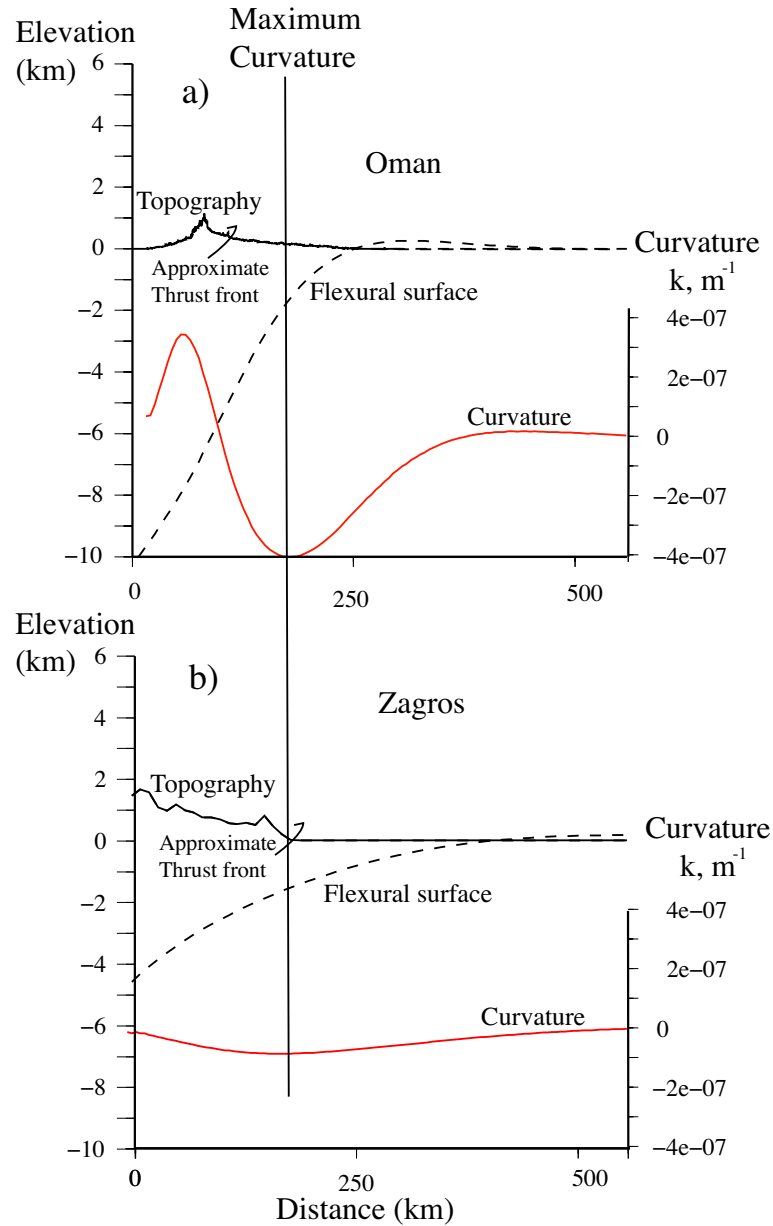


Figure 5.20: Curvature variation across the Oman and Zagros forelands. a) Profile across the U.A.E. section of the foreland of the Oman ophiolite. Solid black line observed topography. Dashed black line flexural curve based on best-fitting broken plate model. Red line curvature of the flexural curve. b) Profile across the centre of the Zagros foreland. Lines as in a).

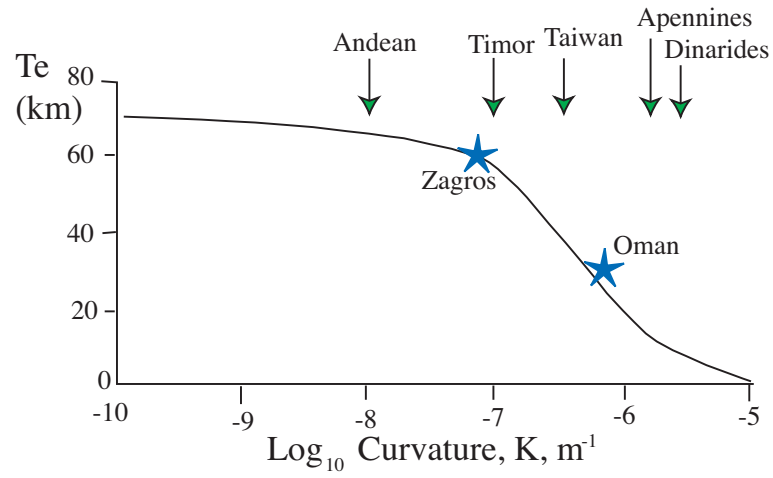


Figure 5.21: Variation of T_e with curvature from Watts & Burov [2003] based on a model continental lithosphere with a thermal age of 400 Ma and a thickness of 50 km. Blue stars mark the data from the U.A.E and Zagros mountain ranges. Arrows mark the range of curvature values recovered for other orogenic regions; the sub-Andean (Stewart & Watts [1997]), West Taiwan (Lin & Watts [2002]), Dinarides Waschbusch & Royden [1992], Apennines Kruse & Royden [1994] and the Australian plate in the Timor region (Lorenzo *et al.* [1998]).

Recovered T_e values fall to approximately zero within 150 to 200 km north of the mountain front. This is seen in both the profile- and grid-based techniques. North-west of 30°N this correlates with the start of the High Zagros zone. The profile technique applied also recovers low T_e values for the region south-east of 30°N but the presence of buried loads makes interpretation difficult.

The foreland shape derived from isopach data does not show the same flexural form as that indicated by the Bouguer gravity anomaly along the entire length of the foreland. This, it is proposed, is due to de-coupling of upper stratigraphic layers from the flexing basement below, along the Precambrian Hormuz salt.

The discrepancy between the T_e recovered in the Oman and Zagros forelands relates to the disparity in load size between the two regions. In Oman, although the topographic expression of the orogen is more minor, the actual load applied to the continental margin is significantly larger when the buried ophiolitic load is

considered. The extra load in Oman caused yielding of the elastic lithosphere and hence recovery of lower T_e values.

T_e variations across South America, North America and Europe

6.1 Overview

Many gravity data sets are now available for continental regions. Publicly available data sets include the Decade for North American Geology project (DNAG) compilation released by NOAA [1990] and the EGM-96 satellite derived potential field model (Lemoine *et al.* [1998]). Commercial compilations also exist, such as those produced by GETECH. These include the South East Asia Gravity Project (SEAGP [1991–1995]), the South American Gravity Project (SAGP, Green & Fairhead [1991]) and the West East European Gravity Project (WEEGP, Green & Fairhead [1994]). This gravity data, in conjunction with topographic data, and the 2D iterative technique discussed in Section 2.7, allows the recovery of the rigidity structure across broad continental areas. These T_e maps provide a new insight into the lithospheric structure in these regions.

In this chapter the T_e structure of a range of continental areas is examined. The recovered T_e maps are compared, wherever possible, to the results of other authors and generally show good agreement. The new maps produced here show

correlation between low T_e and geological features such as suture zones and active orogenic belts. There is also an apparent correlation between T_e in the foreland and the structural style within some orogenic belts.

6.2 South America

Due to the presence of an extensive active orogenic belt and good gravity data coverage, outside the Amazon basin, South America presents an ideal region in which to investigate the spatial variation of T_e and its impact on tectonic processes. The continent is bordered in the west by the Andean mountain chain; which is over 3000 km long, 500 km wide and 4 km high. Today the Andes are associated with ocean-continent collision as the oceanic Nazca and Antarctic plates are subducted eastwards beneath the western margin of the South American plate. On the eastern margin of the Andes there has, in places, been significant tectonic shortening within the sub-Andean fold and thrust belt (e.g. Lamb *et al.* [1997]) as the orogenic belt has been over-thrust towards the stable continental interior. Thus the Andes represent a significant topographic load upon the continental crust; arguably second to the Himalayas in size.

A link between T_e variations and the tectonic evolution of the region has previously been suggested (e.g. Watts *et al.* [1995], Stewart & Watts [1997] and Whitman *et al.* [1996]). To further investigate this link the 2D iterative technique was applied to recover a map of spatial variations in the T_e structure. The recovered T_e structure was compared to the observed geological and geomorphological features of the region. This comparison confirmed that T_e variations correlate with both modern variations in tectonic processes and the pattern of past orogenic events.

6.2.1 Data

The topographic data used in this study was derived from the GEBCO [2003] 1-minute (~ 2 km) grid which incorporates onshore data from the GLOBE [1999] data set. The region assessed is shown in Figure 6.1a. The GEBCO data was re-sampled onto a 40 km grid to allow rapid calculation of the flexural surfaces. Data below sea level was set to zero, as topography above sea level was assumed to be the only load. This assumption ignores the change in isostatic balance, and hence Moho depth, across the ocean-continent transition. However, models of this effect, such as that calculated for the Oman region (Figure 3.8), where the continental shelf is very narrow, indicate that there is no significant impact on the observed anomaly > 100 km from the coast. Where the continental shelf is wide there is little impact on the observed on-shore gravity field.

Bouguer gravity anomaly data for the South American region was from the SAGP database described by Green & Fairhead [1991]. The terrain corrected, point Bouguer data has been re-sampled onto a 5-minute grid using the **GMT** functions **blockmean** and **surface** with a tension factor of 0.25. The tension factor was that suggested by Smith & Wessel [1990] for gridding potential field data. The gridded Bouguer gravity anomaly and locations of the binned data are shown in Figure 6.1b and c respectively. Data coverage is generally good across much of the region. However, in the Amazon basin the coverage is sparse and the recovered T_e in this area is therefore less well constrained. For the 2D iterative technique the data was projected onto a Cartesian grid with a 40 km point spacing using a Mercator projection. The long wavelength gravity anomaly for wavelengths up to degree and order 20 ($> \sim 2000$ km), shown in Figure 6.1d, was also subtracted from the observed anomaly. These long wavelength anomalies, shown in Figure 6.1d are associated with deep mantle processes and therefore do not represent a flexural signal. Zhong & Davies [1999], for example, suggest that geoid anomalies up to degree and order 10 can be explained in terms of dense slabs and variations in

mantle viscosity. In addition, Watts [2001] showed that even if T_e was ~ 90 km across the entire region topography with wavelengths greater than degree and order 20 would be Airy isostatically compensated and hence the free air anomaly from this topography would be zero. The gravity anomalies of this order shown in Figure 6.1d must therefore be due to deeper processes than isostatic compensation of surface topography. The long wavelength anomaly was derived from an analysis of satellite orbits by Rapp & Pavlis [1990]. At these wavelengths the more recent EGM-96 model (Lemoine *et al.* [1998]) is less than 2.1 mgal different, use of the older model will therefore have little impact on the recovered T_e pattern.

Parameters used in this analysis were the same as those used in Chapter 3 for the recovery of T_e in Oman and in Chapter 4 over the Eurasian collisional system. Values used were; Poisson ratio 0.25; Young's modulus 10^{11} Pa; crustal density 2800 kg m^{-3} ; mantle density 3330 kg m^{-3} ; infill density 2650 kg m^{-3} ; load density 2650 kg m^{-3} . The only parameter varied between the Eurasian study and this one was the zero elevation crustal thickness (T_c). T_c was initially set to be 35 km for South America but the effects of varying it are discussed below (Section 6.2.3).

6.2.2 Results

Constant T_e

Initially, the type of compensation in South America was investigated by analysis of isostatic anomalies that were calculated assuming topographic loads were emplaced on the surface of an elastic plate with a uniform T_e structure.

Figure 6.2a shows that an Airy model gives a relatively subdued isostatic anomaly over much of the region. However, a distinctive isostatic positive-negative couple is associated with the eastern Andean mountain front of a similar form to that associated with the Himalayas (see Section 4.4, Figure 4.3). The positive anomaly is associated with the eastern margin of the mountains and the negative residual with the foreland region to the east. The couple suggests the eastern

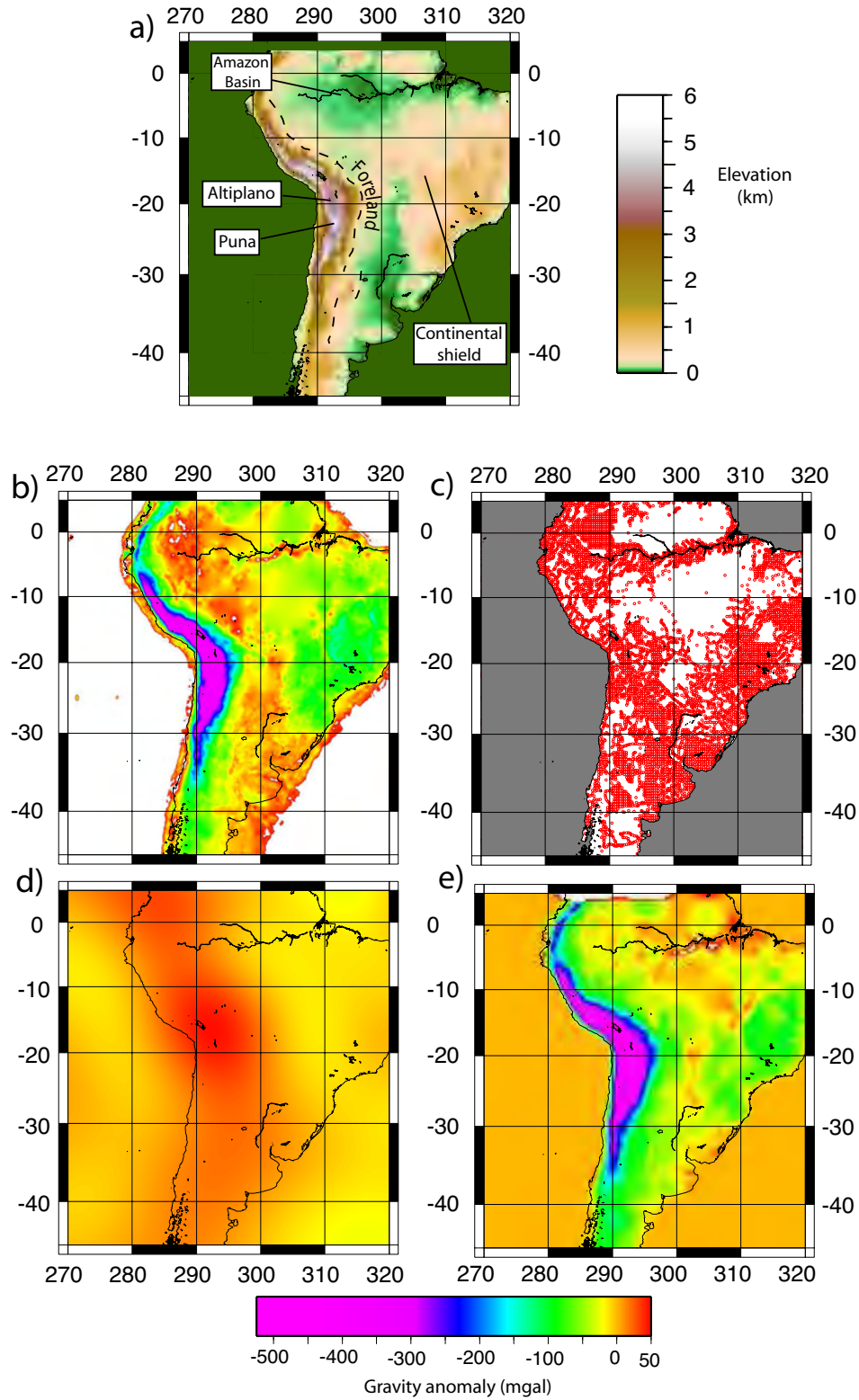


Figure 6.1: Data used in 2D iterative recovery of T_e in South America. a) Input topography data. Dashed lines show approximate position of the Frontal thrust of the Andean orogenic belt. b) Gridded SEAGP Bouguer gravity anomaly. c) data distribution in 2.5x2.5' mean bins. d) Satellite derived gravity anomaly based on coefficients of Rapp & Pavlis [1990] to degree and order 20. e) Bouguer gravity anomaly after removal of the long wavelength satellite derived gravity anomaly shown in d.

Cordillera of the Andes and the flanking Chaco Plain is regionally, rather than locally, compensated. Figure 6.2b shows that when a uniform plate with T_e of 40 km is loaded the high part of the high/low couple around 20° S, close to the mountain front, is reduced. This suggests that this T_e value provides a relatively good model of the T_e in this area as the model gives a good match to the observed data. However, the observed anomaly in the foreland region to the east of the mountains is not adequately matched by this constant T_e . In addition the gravity field beneath the Altiplano is not modelled accurately as significant isostatic anomalies remain in these areas. A constant T_e of 100 km accounts for the negative isostatic values in the foreland region and, in fact, makes them slightly positive as shown in Figure 6.2c. These positive values in the foreland suggest that this T_e value is too high, as the calculated negative anomaly is broader than the observed. Beneath the Andean mountains the isostatic anomaly is lower than -125 mgal for a model T_e of 100 km. This shows that the model does not match the observed gravity data well in this area. This is because the model Moho is too shallow in this region when high T_e values are assumed beneath the mountains.

T_e map

The recovered T_e structure, following 30 iterations of the 2D iterative technique, is shown in Figure 6.3. The figure shows a highly variable T_e structure with values ranging from 120 km, in parts of the foreland basin region, to less than 10 km below the Andes. At ~20°S, within the continental shield, east of the high T_e values in the foreland, a secondary region of low T_e is seen. Figure 6.3b displays the point locations for the data from which the T_e structure has been derived, good coverage is seen in the central region suggesting results in this region are robust. However, within the Amazon basin the data coverage is sparse, T_e values from this region are therefore less reliable.

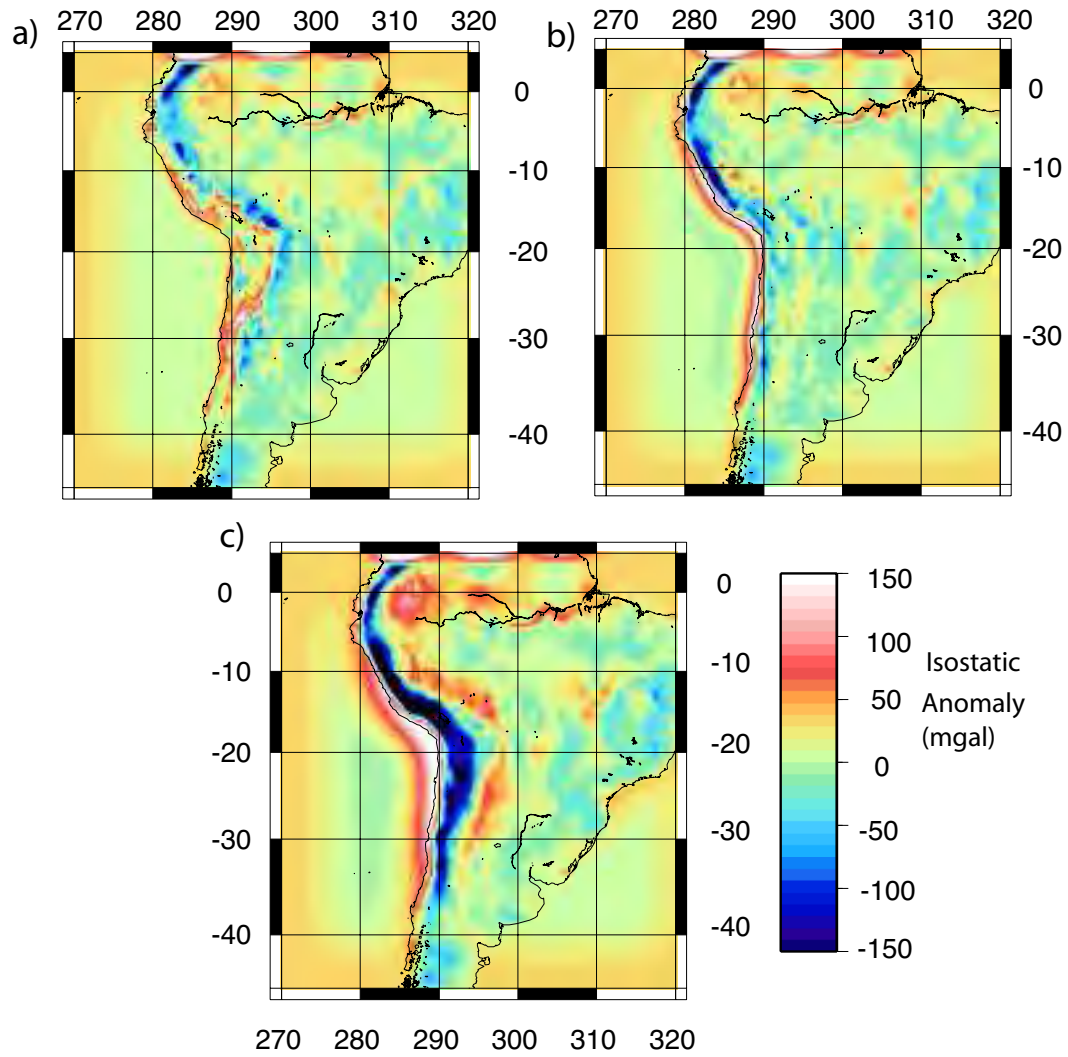


Figure 6.2: Isostatic anomalies based on the assumption of a constant T_e . a) Airy isostatic anomaly ($T_e = 0$ km). b) Isostatic anomaly for a T_e of 40 km. c) Isostatic anomaly for a T_e of 100 km.

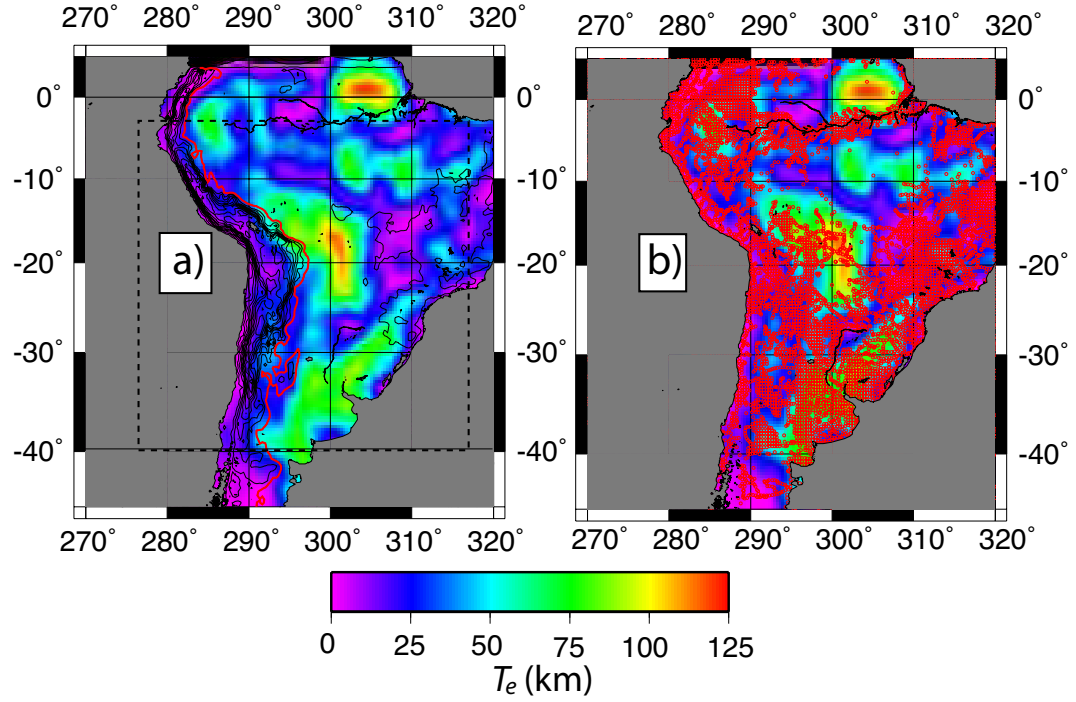


Figure 6.3: Recovered T_e structure and data coverage. a) T_e of South America after 30 iterations of the 2D iterative technique. Also shown is the regional topography, contoured at 500 m intervals. The approximate position of the main frontal thrust is highlighted in red, based on the 500 m contour. Dashed box marks boundary of tapered data. b) Locations of binned raw gravity data.

Variable T_e

The observed Bouguer anomaly and the calculated anomaly derived from the iterative technique shown in Figure 6.4a and b are very similar to each other, however, they are dominated by the negative values beneath the Andes which would be seen irrespective of the input T_e . To assess the quality of fit between the two anomalies the isostatic residual was calculated (Figure 6.4d). This isostatic residual shows a significant reduction in amplitude across the Andean region relative to the constant T_e isostatics shown in Figure 6.2. In the shield areas to the east of the mountains there is a slight reduction in the amplitude of the isostatic anomaly compared to the constant T_e isostatics.

The spectra of the isostatic anomalies, shown in Figure 6.4c, confirm that the variable T_e model significantly reduces the spectral power of the residual isostatic anomaly at all wavelengths compared to any uniform T_e structure. This suggests that the final variable T_e structure is the best model for the regional isostatic compensation of the present day topography.

6.2.3 Crustal thickness

The standard Moho depth is one of the parameters input into the 2D iterative technique. To verify this value, in the absence of seismic data, spectral techniques such as those of Karner & Watts [1983] may be used. These recover the mean depth to source for the interfaces which create the observed anomaly.

The gravity anomaly ($g(x)$) due to an interface at a mean depth d is given, in the frequency domain by:

$$G(k) = 2\pi\gamma\Delta\rho e^{-kd}H(k) \quad (6.1)$$

Where k is the wave number, γ is the gravitational constant, $\Delta\rho$ is the density contrast across the interface and $H(k)$ is the Fourier transform of the topography ($h(x)$). The average power spectra $|G(k)|^2$ given a random distribution of sources

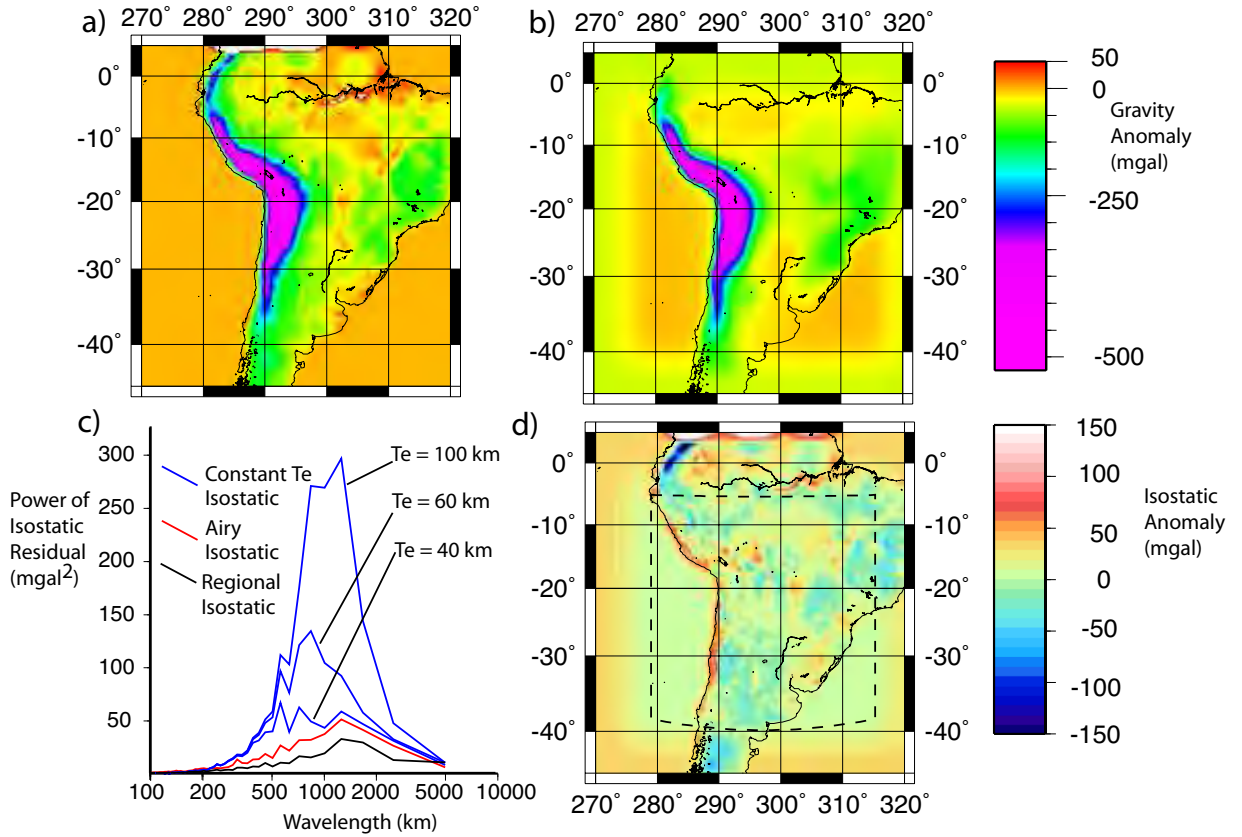


Figure 6.4: Results of iterative model a) Input Bouguer gravity anomaly data. b) Modelled Bouguer gravity anomaly based on variable T_e structure. c) Spectra of isostatic anomalies based constant on T_e values of 0 km (red line), 40, 60 and 100 km (blue lines). Additionally the spectra of the isostatic anomaly derived from a regionally variable T_e structure is shown in black. d) Isostatic anomaly based on difference between modelled and observed Bouguer gravity anomaly. Dashed box, outline of area assessed for isostatic spectra.

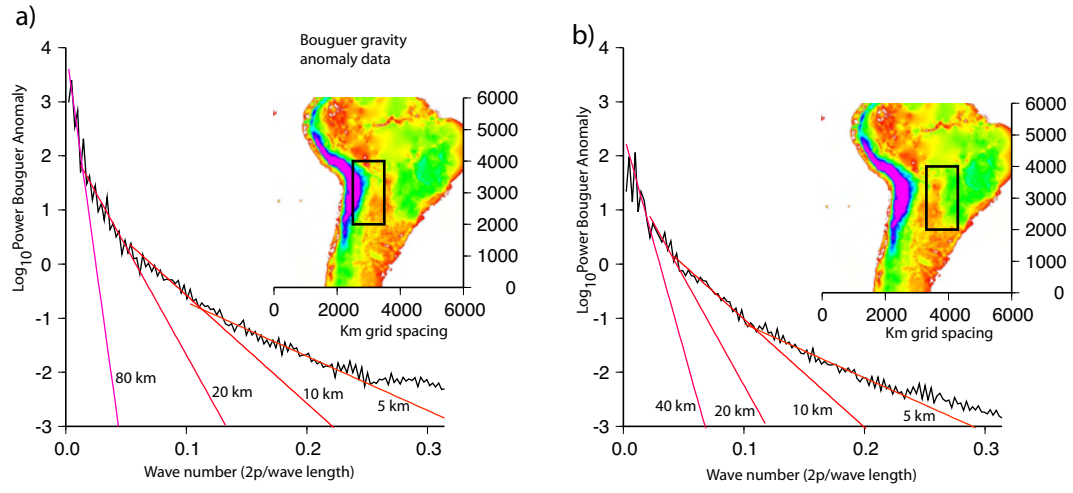


Figure 6.5: Spectra over sub-regions of the observed Bouguer gravity anomaly. a) Spectra over the Andean mountains and foreland. b) Spectra over the cratonic shield region.

spread over a depth range of $d \pm \Delta d$ is given by,

$$\langle |G(k)|^2 \rangle = 4\pi^2 \gamma^2 \langle \Delta \rho^2 \rangle \langle e^{-2k\bar{d}} \rangle \langle |H(k)|^2 \rangle \quad (6.2)$$

Where $\bar{d}$ is the mean compensation depth. The logarithm of the exponential component can be approximated as a straight line with a slope of $-2k\bar{d}$. This represents the slope of the log power spectra of a single interface. The power spectra of the gravity anomaly resulting from a series of discrete interfaces will have a series of linear segments with slopes given by the mean depths.

Two areas, shown in Figure 6.5, were chosen for analysis. One included both the foreland and the Andean mountains and the second was over the continental shield. Figure 6.5a displays the spectra over the foreland and Andean region. The segments of the spectra are most adequately matched by depth values of 5, 10, 20 and 80 km. The deepest of the recovered sources is likely to reflect the major density interface at the Moho. This recovered depth for the Moho agrees with that suggested by Zandt *et al.* [1994] for the depth of the Moho beneath the Andes from seismic studies.

Figure 6.5b shows the spectra for the central continental region. A number of linear segments are apparent; the gradients of which correlate with mean depths of 5, 10, 20 and 40 km respectively. The spectra over the central region suggests that a standard crustal thickness of 40 km would have been preferable to the value of 35 km, used in the iterative technique.

To assess the effect of altering the standard crustal thickness on the 2D iterative technique the model was re-calculated with a crustal thickness of 40 km. Figure 6.6a and b indicate that the recovered patterns of T_e variation for 40 km and 35 km crustal thickness are very similar. The absolute value of T_e recovered generally varies by less than $\pm \sim 5$ km, although in some areas this increases to a difference of $\pm \sim 10$ km. The most significant difference between the T_e structure recovered for a 40 km and a 35 km crustal thickness occurs in the foreland region at 20° S. Here a stronger foreland and a weaker zone beneath the mountains is recovered in the case with thicker standard crust.

To further analyse the effect of crustal thickness on the recovered T_e pattern the 2D iterative technique was re-applied with a standard crustal thickness of 30 km. Figure 6.6d shows that the recovered pattern of T_e variation determined for 30 km and 35 km crustal thickness are very similar. The difference between the recovered T_e and that recovered for the 35 km thick crust was $\pm \sim 10$ km. However, the sense of the difference was inverted from the thicker crustal case, with a weaker foreland and higher values recovered beneath the mountains.

These tests show that crustal thickness variations do not have a significant impact on the recovered pattern of T_e values. The 2D iterative technique is therefore valid even in regions of poorly known initial crustal thickness, as the pattern of recovered T_e is relatively insensitive to this model parameter.

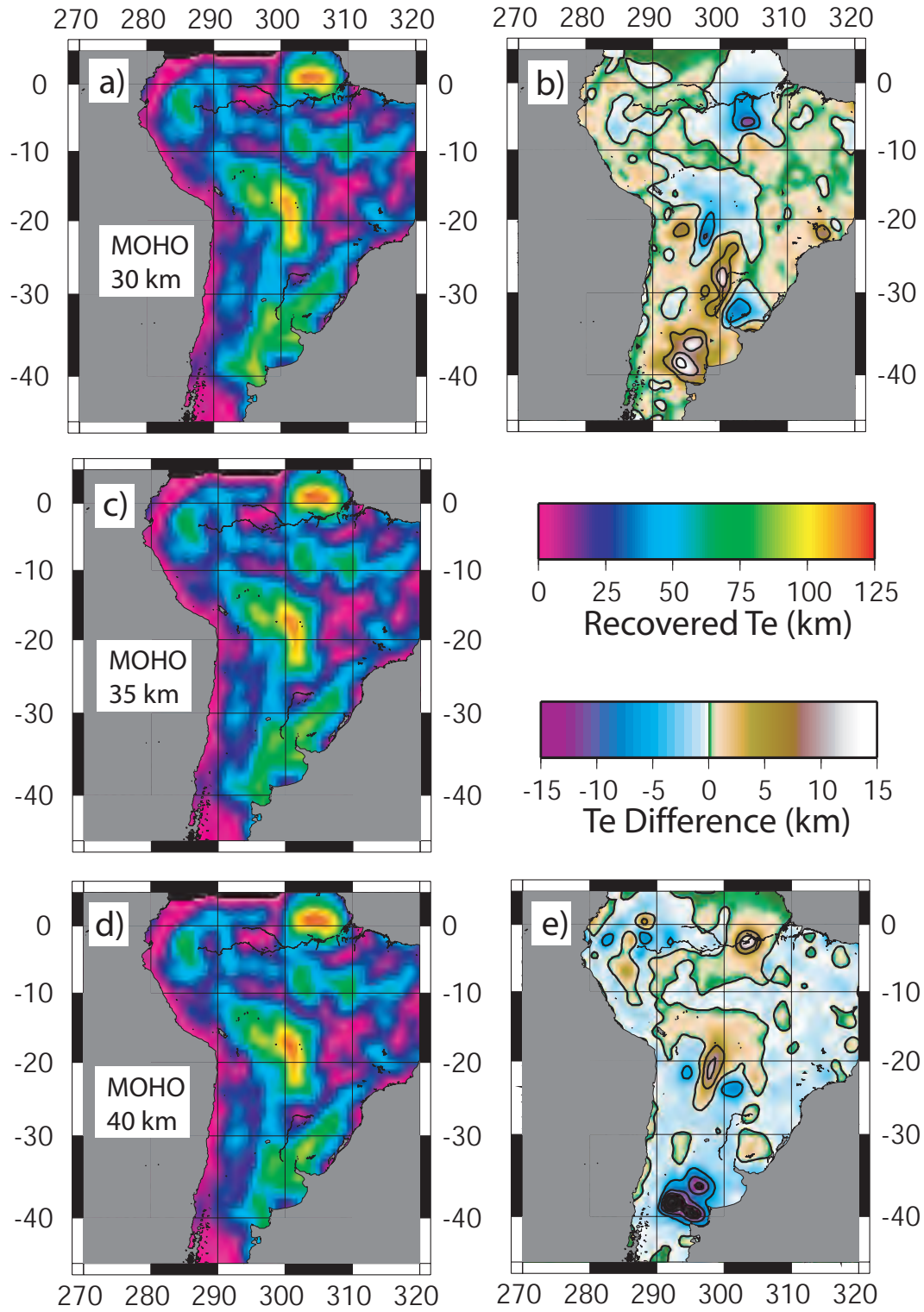


Figure 6.6: Effect of varying T_e on the recovered T_e structure. a) T_e structure recovered by 2D iterative technique based on a thinner standard crustal thickness of 30 km, b) Difference between T_e recovered for 30 and 35 km crustal thickness. c) T_e recovered based on a standard crustal thickness of 35 km. d) T_e recovered based on a standard crustal thickness of 40 km. e) Difference in recovered T_e structure for 40 and 35 km crustal thickness.

6.2.4 Previous work

A range of T_e values have been recovered along the length of the Andean foreland by previous authors. Lyon-Caen *et al.* [1985] recovered values of 25 to 55 km, based on gravity data, along a profile at 20°S. Whitman [1994] recovered values between 6 and 12 km, based primarily on Moho geometry derived from seismic studies, at approximately 24°S. Watts *et al.* [1995] recovered variations in T_e across the foreland region based on 2D constant rigidity flexural models, values of 25 km north of 15°S and south of 22°S were recovered. Between these regions Watts *et al.* [1995] recovered a higher T_e of 50 km.

A regional study of the T_e variation in the South American plate was previously carried out by Stewart & Watts [1997]. This was based on forward modelling of the observed Bouguer gravity anomalies along 31 profiles across the Andean mountain belt, from approximately 10°N to 35°S. In a similar manner to the 2D iterative technique used in this study, and the technique of Braitenberg *et al.* [2002], a series of plates with constant T_e were loaded with the observed topography (and in some cases an additional shear force) and the best fitting T_e for each section of the profile was recovered. In the study of Stewart & Watts [1997] the elastic plates were broken beneath the mountains. The plate break was considered to occur where a simple Airy model and the observed Bouguer anomaly coincided. To the west, beyond the plate break, an Airy model was assumed, as shown in Figure 6.7b.

Figure 6.7 shows that the profile method of Stewart & Watts [1997] recovers a similar transition from low (zero) T_e beneath the orogen, to high T_e in the foreland region as the 2D iterative method. This change in T_e is compatible with the plate being broken beneath the mountains as discussed in Chapter 2. The along strike variation of foreland T_e recovered by the 2D iterative method is also reflected in the variation in T_e recovered by Stewart & Watts [1997]. Values decrease from ~75 km in the central region to 50 km in the north and south in both studies.

Although the pattern of T_e variation matches well between the 2D iterative

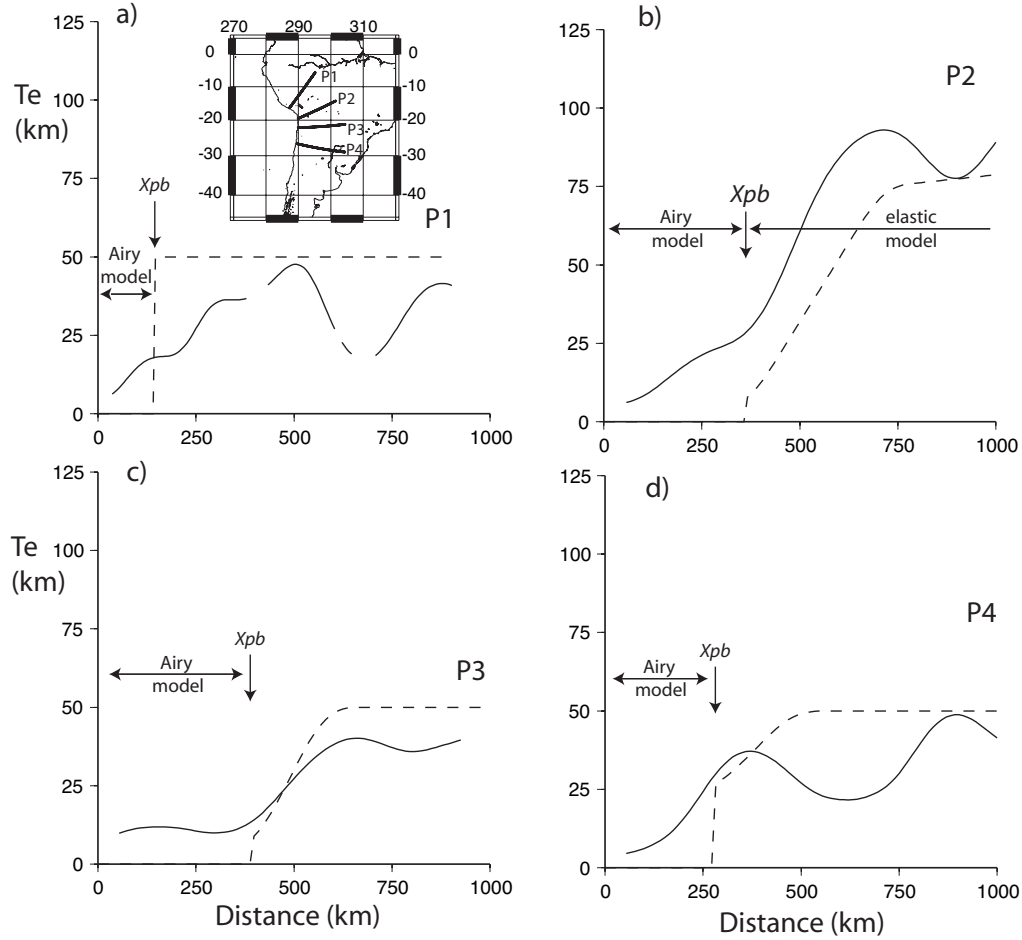


Figure 6.7: Comparison between results of this study (solid black line) and four profiles from the study of Stewart & Watts [1997] (dashed black line). X_{pb} marks the plate break positions assumed by Stewart & Watts [1997]. Gaps in profile P1 relate to regions with poor data coverage.

technique and the results of Stewart & Watts [1997] the absolute value of T_e recovered differs. To the north and south of the central region the 2D iterative technique recovers T_e values ~ 25 km lower than the profile methods. This discrepancy may be due to inclusion of an additional shearing force by Stewart & Watts [1997] on these profiles and is discussed below.

Tassara [2005] carried out a study based on profiles across the entire Andean mountain chain. A horizontal stress component, resulting from external compressive forces, was included in this study, in addition to the forces related to surface loading. Tassara [2005] iteratively varied T_e along the profile, and the amount of horizontal stress applied at the edges, until a best fit between observed and calculated Bouguer gravity anomalies was achieved.

The results of Tassara [2005] are consistent with those of the 2D iterative technique, i.e. they show low T_e beneath the Andean mountain range and higher values, in the order of 70 km in the foreland region between 15°S and 20°S . To the south the foreland rigidity falls to between 47 and 57 km. The horizontal stress term used by Tassara [2005] is $\sim 2 \times 10^{12}$ N/m but is only needed north of 17°S and south of 30°S . This corresponds to the regions where Stewart & Watts [1997] achieved a better fit between calculated and observed Bouguer anomaly by imposing additional bending moments on the end of the beam. The fact that additional stresses do not need to be applied in the central part of the mountain range suggests that omitting this term from the 2D iterative model will not change the pattern of recovered T_e in this region. Towards the north and south the 2D iterative technique recovers lower T_e values than the profile studies of Stewart & Watts [1997] and Tassara [2005]. This is because less force is applied to the plate in the 2D iterative technique, hence to achieve the same curvature, and associated calculated gravity anomaly, a lower T_e is required.

Cardozo & Jordan [2001] conducted a detailed study of a relatively small region between 28.5°S and 31.5°S based on reconstruction of isopach data and topographic

loads for the 9 Ma Bermejo basin. Using a finite element grid method they analysed the T_e variation of the region. The best fit T_e structure was low (0 km) beneath the high mountains, rising to 30 km below the frontal thrust. Beyond the thrust there is a narrow zone (50 to 100 km wide) of low T_e which increased to 30 km to the east. The resolution of the study of Cardozo & Jordan [2001] is much finer than that assessed using the 2D iterative technique, so the results could be expected to differ in detail. In this region the 2D iterative technique simply recovered a linear trend from approximately 5 km beneath the orogen to 50 km towards the mountain front.

The most recent study of T_e in South America is that of Tassara *et al.* [2006]. This was based on a analysis of the spectral coherence between satellite derived Bouguer gravity data and topography, using wavelet transforms. This recovered a T_e structure similar to that presented in this chapter, with low (<15 km) values recovered beneath the Andean mountain belt and high (~100 km) values recovered beneath the Amazonian craton. However, the model of Tassara *et al.* [2006] shows a smoother transition between high and low T_e across the mountain front. This is most likely to relate to the longer wavelength (>125 km) gravity anomaly considered by these authors.

6.2.5 T_e and the terrane structure of South America

T_e and the orogenic belt

There is a clear correlation between low T_e and the elevated topography of the Andean mountain belt seen in Figure 6.8. These low T_e values could be predicted in such a region for several reasons. Firstly the thickened crust beneath the orogen would displace mantle material from ~ 40 km to over 70 km. Mantle material between 40 and 80 km deep may retain significant rigidity; continental crust, however, yields easily at these depths, if it is silica rich, due to the relatively high temperatures (e.g. Cloetingh & Burov [1996]). Secondly potential ‘delamination’ of old

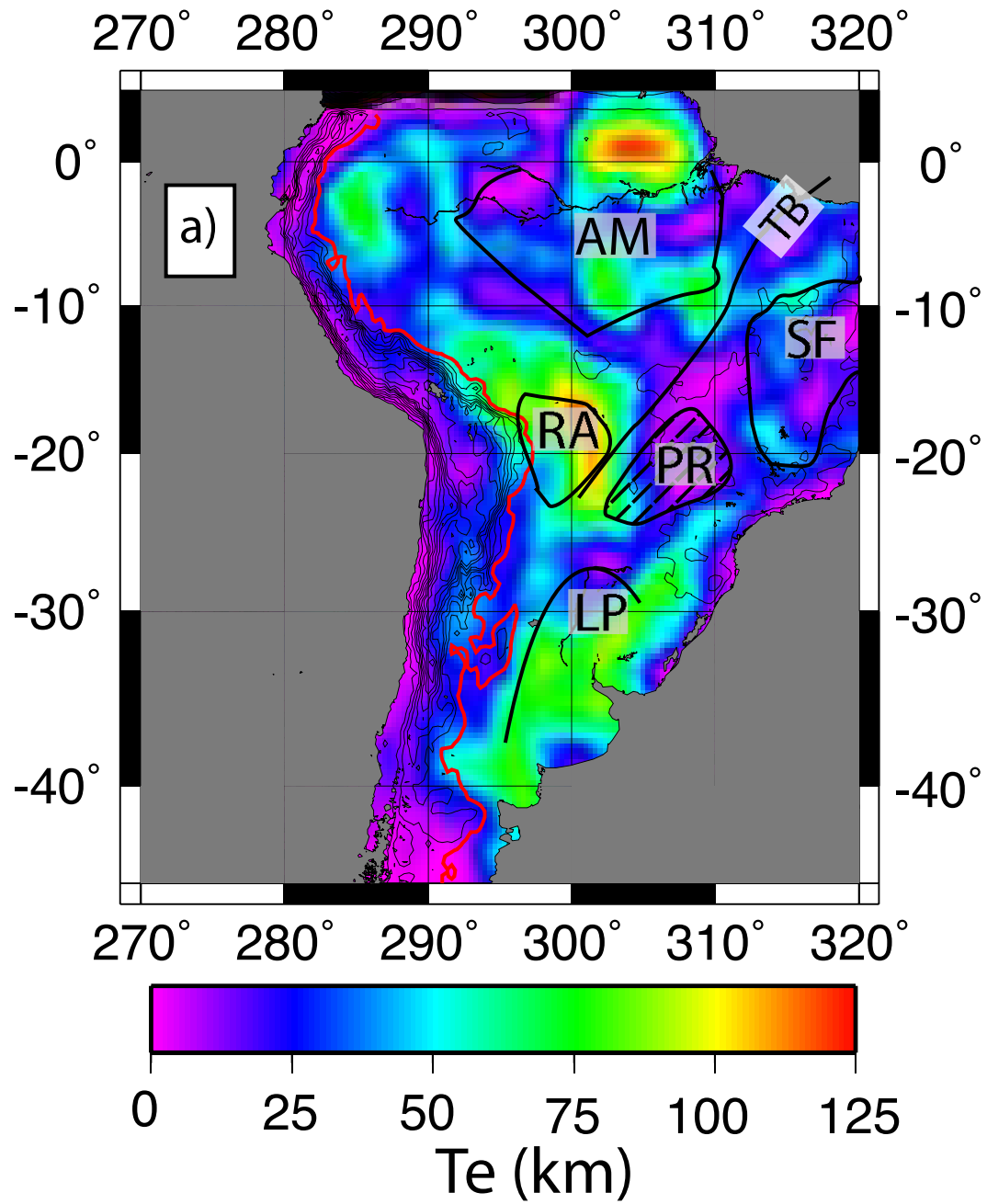


Figure 6.8: Map of recovered T_e and geology after Cordani *et al.* [2003]. AM, Amazonia; RA, Rio Apta; LA, Luis Alves; SF, Sao Francisco; PR, Paraná; TB, Transbrasiliano lineament.

lithospheric mantle that is colder than the surrounding material, as suggested for example by Hoke *et al.* [1994] and Whitman *et al.* [1996]. This would bring hot, weak mantle to the base of the crust. Seismic evidence for such delamination has been reported by Whitman *et al.* [1996]. Other evidence for mantle delamination, presented by Hoke *et al.* [1994], comes from the distribution of mantle derived He isotopic ratios. Hoke *et al.* [1994] showed that a mantle derived signature can be found across the entire width of the Altiplano to the Eastern Cordillera. This mantle derived signal is typically only found in regions with hot mantle close to the base of the crust such as at mid oceanic ridges or continental rifts. Lastly, it is known that dehydration of the down-going oceanic crust releases large amounts of water (e.g. Pawley & Holloway [1993]). This causes mantle melting and advection of hot magma into the overlying continental crust, which weakens it further.

T_e across the continental shield

The regional geology is shown superimposed over the recovered T_e structure in Figure 6.8. There is a correlation between the high recovered T_e region between 12°S and 25°S in the foreland and the Rio Apta (RA) craton. This craton is thought to be an ancient micro-continent formed during the Palaeo-Proterozoic (1.8-2.5 Ga) (Cordani *et al.* [2003]). However, the region of high T_e exceeds the area of the cratonic core proposed by Cordani *et al.* [2003]. This suggests that the cratonic core may be larger at deeper lithospheric levels than proposed by Cordani *et al.* [2003] from geological observations. The Luis Alves (LA) cratonic block, towards the south east of the area investigated, also appears as a region of relatively high, ~ 75 km, T_e .

The Parana flood basalt region appears to correlate with a region of low T_e , in a similar manner to the low T_e values found in the Deccan province of India, discussed in Section 4.5.1. The peak eruption of the Parana flood basalts occurred between 134 and 130 Ma (Kirstein *et al.* [2001]). It has been shown by O'Connor

& Duncan [1990] that this volcanism was associated with the Walvis ridge - Rio Grande hot spot system. The effects of a hotspot on the lithospheric mantle are likely to have been dramatic thinning and replacement of old, potentially cold, mantle with younger, hotter, mantle material. In addition Pérez-Gussinyé & Watts [2005], for example, have suggested that lithospheric mantle younger than ~ 2.5 Ga may be relatively more hydrated due to less partial melting, which removes volatile compounds, related to falling mantle temperatures with time. Younger, hotter, and potentially more hydrous mantle related to the hotspot would be weaker than the surrounding regions and could explain the lower recovered T_e values in this region.

Correlation of T_e with mountain height and structural style

It has been suggested by previous authors (e.g. Watts *et al.* [1995], Stewart & Watts [1997] and Whitman *et al.* [1996]) that there is a correlation between structural style and foreland T_e . They suggest that high rigidity in the foreland basin leads to low-angle, thin-skinned, thrusting of the orogenic belt over the continental interior. Conversely a weak foreland would lead to thick skinned deformation with extensive basement involvement preventing extensive over-thrusting by the orogenic belt. If there is significant thin skinned thrusting along the eastern margin of the Andes frictional coupling across the subduction zone would become relatively more important, and act to push the orogen onto the continent. If there is significant thick skinned deformation in the sub Andes, forward propagation of the orogen is retarded, and a larger proportion of the total convergence is accommodated by subduction of the oceanic crust. A sketch of this scenario is shown in Figure 6.9.

In the case where the continent is over-thrust the orogenic belt will potentially become more developed (Himalayan in nature) with growth by crustal thickening, facilitated by under thrusting of continental material. Delamination of old, cold mantle material may also occur leading to extensive crustal melting. If the conti-

ment is not over-thrust more oceanic material will be subducted, leading to crustal thickening predominately by addition of volcanic material.

To assess the correlation between foreland T_e and the orogen structure T_e at the frontal thrust and the average topographic parameters (width and height) of the orogen were assessed. The frontal thrust was taken to be approximately equivalent to the 500 m contour on the east of the mountains and the orogen edge was defined by the same contour in the west. The average height on an east-west line enclosed by the 500m contour was calculated, as was the width of the mountain range in km. Figure 6.10 shows the results of this analysis.

Figure 6.10c shows there is generally a positive correlation between T_e and the average elevation of the mountain belt, with a correlation coefficient of 0.12 recovered for this data. Figure 6.10d however, shows that this correlation breaks down at 24°S where the average elevation remains close to 3000 m while T_e falls to ~25 km. Whitman *et al.* [1996] discusses the elevation of the topography in this region and suggests from seismic refraction data that hotter mantle in this region may be supporting the topography, rather than under thrust continental lithosphere. Width shows a better correlation with T_e , with a correlation coefficient of 0.25 recovered for the data in Figure 6.10e. Width also visually matches the rise and fall in foreland T_e more closely in Figure 6.10f. The observed correlation supports the suggestion of Stewart & Watts [1997] that the amount of continental deformation is related to T_e of the foreland region.

Kley & Eisbacher [1999] compared the topographic and geological properties of the Central Andes, the Himalayan and the Alpine orogenic belts. They highlighted the fact that the central Andes around ~20°S appears to be closer to a Himalayan style continent-continent collision than a Cordilleran (ocean-continent) collisional system, i.e. crustal thickening is predominately due to under thrusting of continental crust rather than emplacement of mantle derived magmatic material. This study suggests that the key to transforming from one style to another may lie in

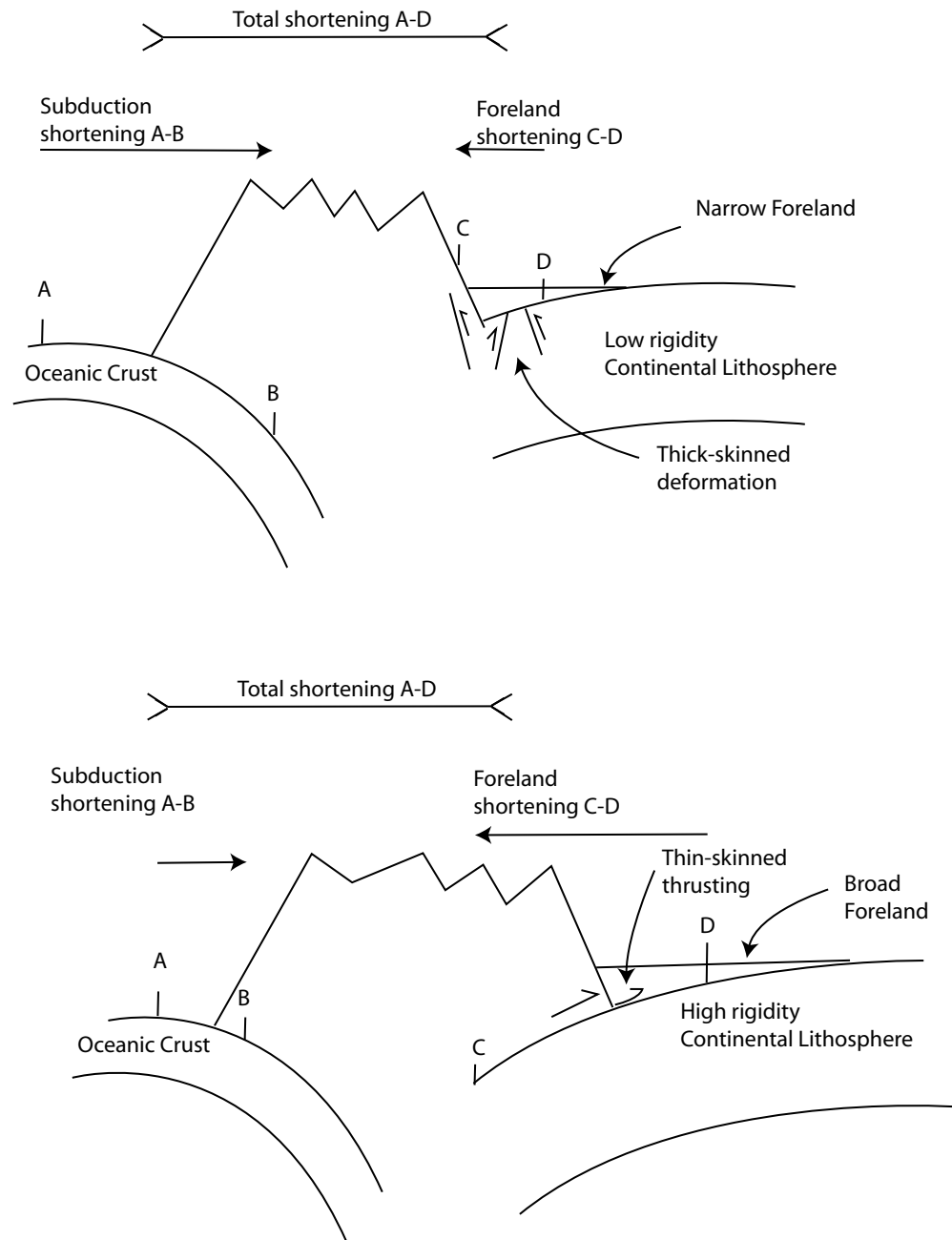


Figure 6.9: Cartoon showing different proportion of ocean, orogen convergence depending on thick skinned vs thin skinned deformation in the orogen, continent collisional zone. Markers A, B, C and D show how much relative convergence occurs in each area per unit time. In 1 time step marker A moves to the position of B and D to the position of C.

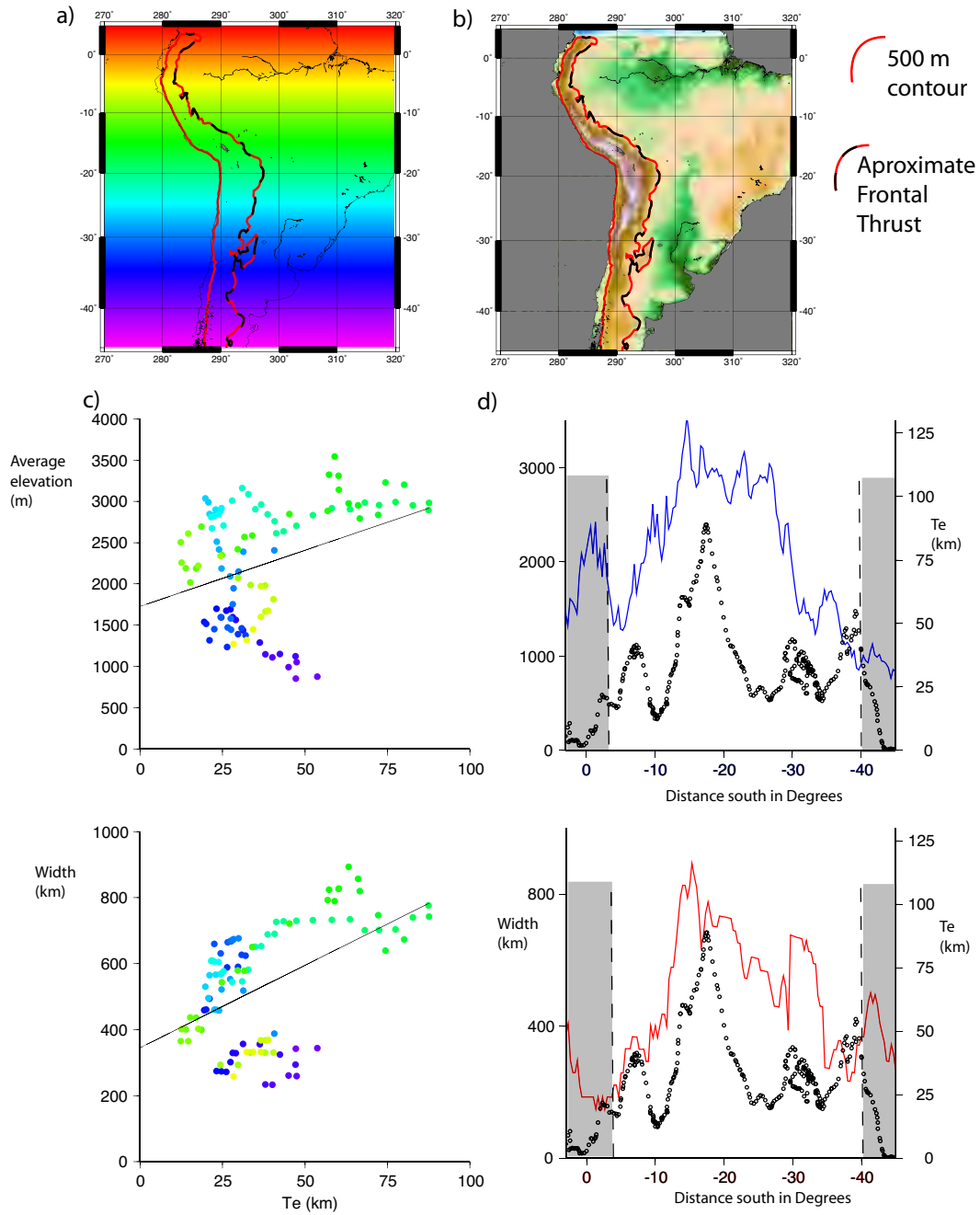


Figure 6.10: Correlation of orogen elevation and width with T_e . a) Latitudinal colour code for data points in c and e. b) Definition of the orogenic belt. c) Average orogen elevation against T_e . Black line shows best fitting trend with a slope of 13.61 m/km. d) Variation of average elevation across the E-W width of the orogen (blue line) and T_e (dotted black line) with latitude. Grey regions show margins of T_e model, not included in analysis. e) E-W orogen width against T_e . Black line shows best fitting trend with a slope of 5 km/km. f) Orogen width and T_e with latitude.

the rigidity of the foreland on the continental side of the orogen. Kley & Eisbacher [1999] suggested that the difference between an Alpine type continent-continent collision and Himalayan or central Andean style one partly lies in the nature of the foreland. The Central Andean and Himalayan forelands being very rigid and the Alpine foreland being relatively weak. The nature of the crust itself may also be a factor. The European crust involved in the Alpine collision was thinned by rifting and the development of the Tethys therefore there was a smaller volume of crustal material available to build the orogen.

6.2.6 Summary

- 1 The 2D iterative technique has been applied to the South American continental area and a regionally variable T_e structure has been recovered.
- 2 The 2D iterative technique recovers a similar pattern of T_e variation, irrespective of the assumed initial crustal thickness.
- 3 The recovered T_e values agree with past estimates of T_e for the foreland region of the Andean mountains.
- 4 T_e variations correlate with the terrain structure observed in the South American continent. Low values are observed beneath the Andean orogenic belt, while distinct regions of high T_e correlate with older cratonic regions.
- 5 T_e variations along the length of the mountain chain appear to correlate with its width, suggesting a wider orogen results from a stronger foreland.

6.3 North America

Large geophysical data sets are available across the North American continent. This data, in addition to the well studied structural history of the region, make North America an ideal region for application of the 2D iterative technique. The 2D iterative technique was applied to two overlapping areas, one comprising Canada and the northern United States and one including the northern, central and southern United States. Two overlapping regions were assessed, as using a single map projection for conversion into Cartesian co-ordinates in both regions created an unacceptable amount of distortion of the input data due to the high latitudes involved.

6.3.1 Data

Data for the assessment of T_e was taken from the DNAG (Decade of North American Geology) database published by the National Oceanic and Atmospheric Administration NOAA [1990]. Data for the northern region is shown in Figure 6.11. Figure 6.11a and b show the topographic data and Bouguer gravity anomaly data respectively. The gravity coverage is generally good. However, in the north-west of the analysed region there are significant data gaps present (Figure 6.11d). The point data was interpolated onto a 5x5-minute grid using **GMT** function **surface** with a tension factor, T , of 0.25 (Smith & Wessel [1990]). For the Canadian region the data was projected into a Cartesian reference system using a Lambert conic projection and re-sampled onto a 25x25 km grid. The edges of the data set were tapered to prevent edge effects, related to the modelled gravity anomalies, affecting the recovered T_e structure. For the United States region the data was projected into a Cartesian reference system using a Mercator projection. The data was re-sampled onto a 40x40 km grid and the edges were again tapered, see Figure 6.12. A coarser grid was used in the southern region to allow faster calculation of the flexural surfaces.

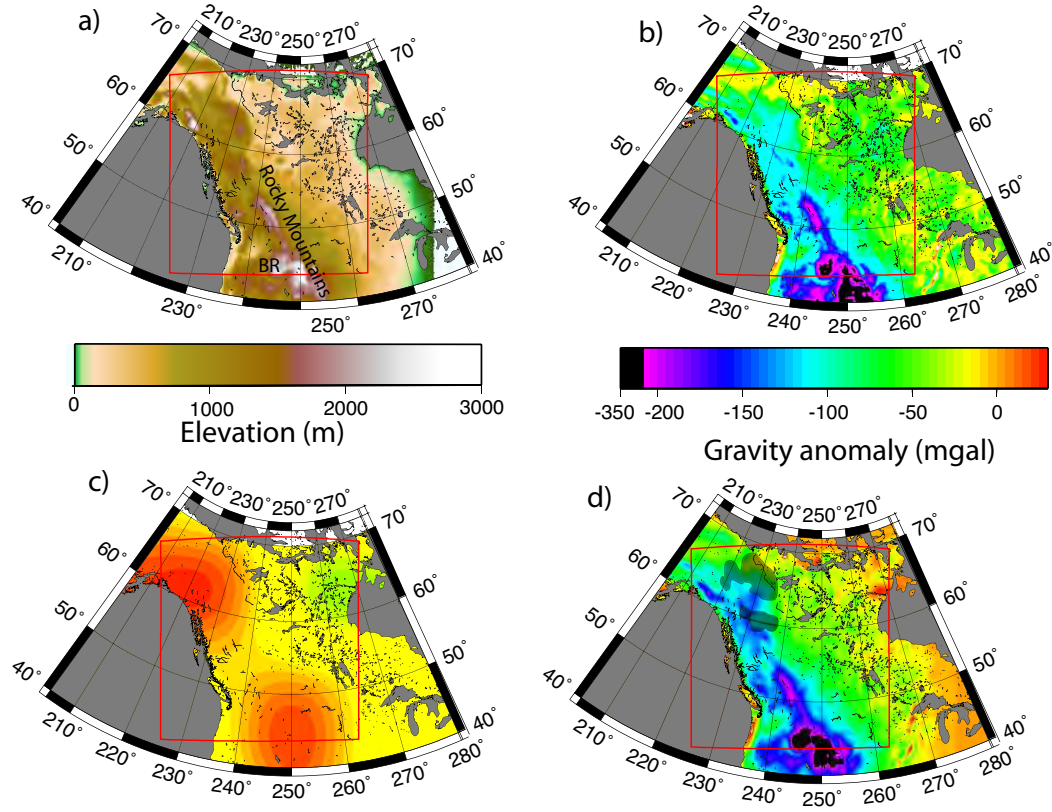


Figure 6.11: Input data for the northern U.S. and Canada. a) Resampled and cosine tapered topographic data. Box marks edge of untapered region. BR = Basin and Range province. b) Bouguer gravity anomaly data. c) Satellite derived gravity field complete to degree and order 20 Rapp & Pavlis [1990]. d) Tapered Bouguer gravity data with satellite derived field, shown in c) subtracted from it. Grey regions are areas with no data coverage.

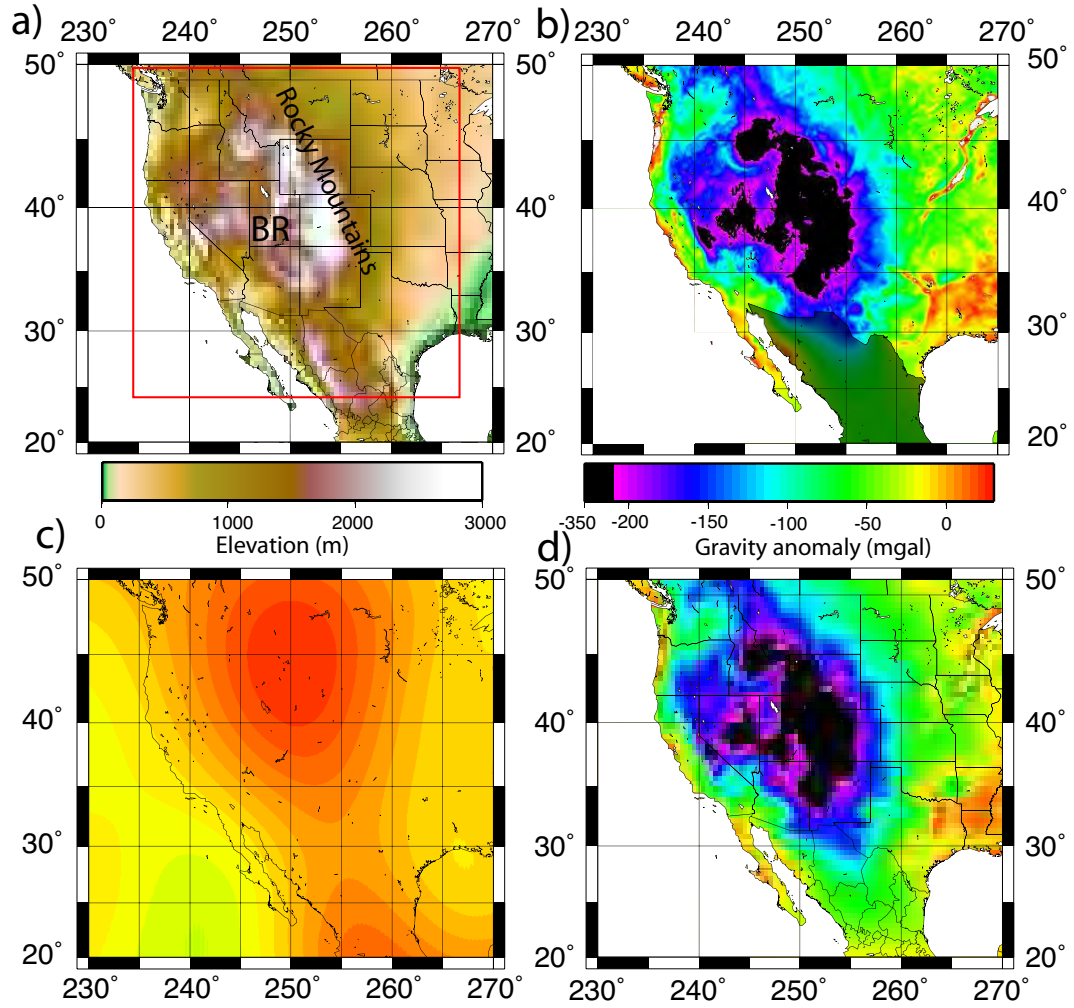


Figure 6.12: Input data for the central and southern U.S. region. a) Topography data. Red box marks limits of taper. BR = Basin and Range province. b) Bouguer gravity anomaly data. Grey regions are areas with no data coverage. c) Satellite derived free air gravity field, complete to degree and order 20. d) Bouguer gravity data with satellite derived field in c) subtracted from it.

6.3.2 Results

The results for the northern U.S. and Canadian region, and the central and southern U.S. region, are considered separately. The northern model was run for 20 iterations while the southern one was run for 60 iterations. Both models assumed standard crustal thickness of 35 km and other parameters as shown in Table 2.1 and used previously in the South American study. For both regions a series of isostatic anomalies were calculated. These were based on either Airy, or constant rigidity regional isostatic support of the observed topography. These allowed an additional assessment of the isostatic compensation to be made.

Isostatic anomalies across the northern U.S. and Canada

The isostatic anomalies resulting from models using constant T_e structures are shown in Figure 6.13. These show that Airy compensation, i.e. zero T_e , gives the best fit beneath the mountain range compared to higher T_e , regional compensation. Higher values of T_e generally appear to increase the magnitude of the isostatic anomaly; particularly in the west, beneath the mountains. In the north-west the Airy isostatic anomaly shows a definite high low couple, centred on 130°W which is significantly reduced when a higher T_e is used. However, this area correlates with an area of no data coverage so no interpretation can be made.

Figure 6.14 shows the observed and calculated gravity anomalies derived from the 2D iterative technique across the northern U.S. and Canada. Figure 6.14c shows the isostatic anomaly based on the regionally variable T_e model. This isostatic anomaly does not appear to give a significant improvement over the Airy isostatic anomaly shown in the previous figure. However, power spectra of the isostatic anomalies show that the regionally variable T_e structure gives the best reduction of power at most wavelengths. At wavelengths of over 1000 km the constant T_e structures appear to reduce the power to a greater extent. This discrepancy may relate to the inclusion of offshore regions, where the gravity model is unconstrained,

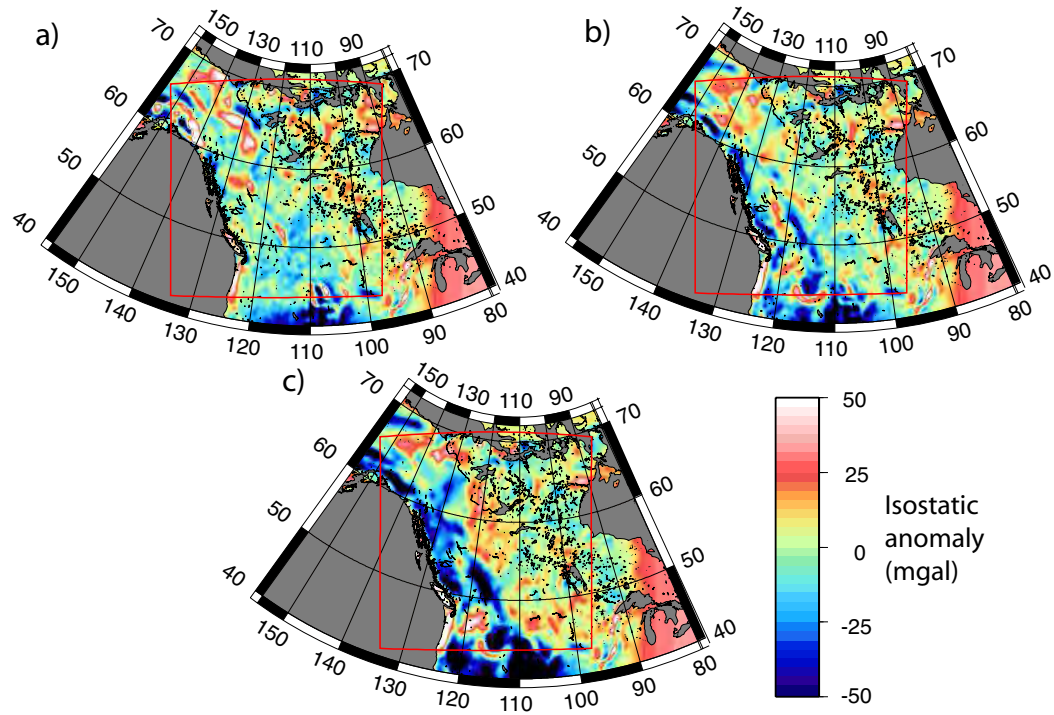


Figure 6.13: Isostatic anomalies resulting from models of compensation of surface topography by constant rigidity plates. a) Airy ($T_e=0$ km) Isostatic anomaly, b) Isostatic anomaly based on a constant T_e of 40 km, c) Isostatic anomaly based on a constant T_e of 100 km. Red box marks area inside which taper applied to input data is 0.

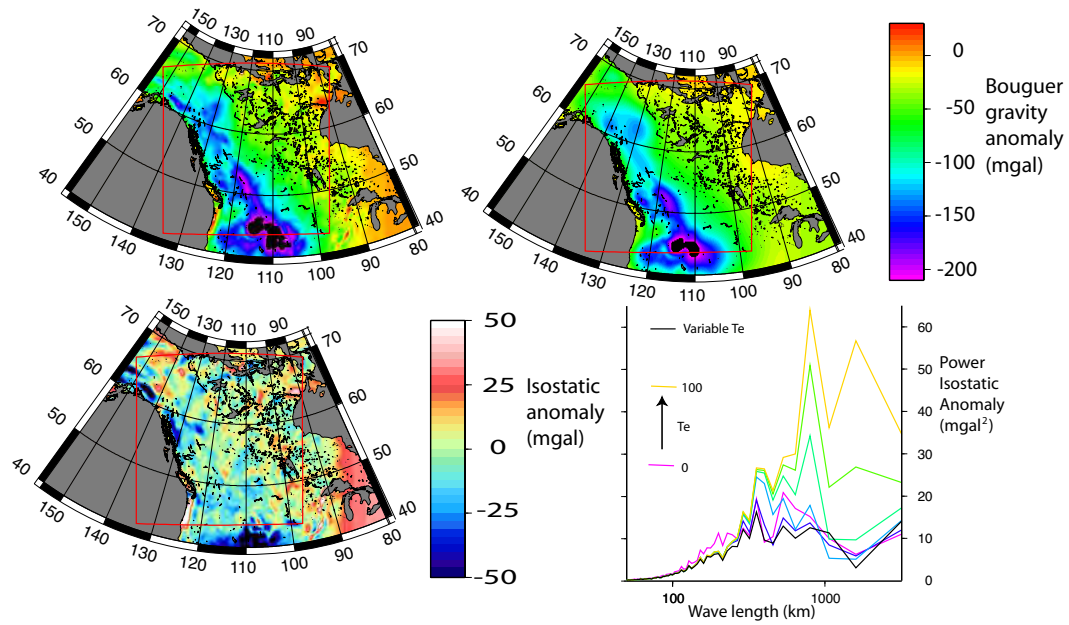


Figure 6.14: Comparison of gravity anomaly resulting from the 2D iterative technique and the observed Bouguer gravity anomaly. a) Observed Bouguer gravity anomaly data (tapered). b) Modelled Bouguer gravity anomaly based on a regionally variable T_e structure. c) Isostatic residual, derived from difference of a) and b). d) Power-spectra of isostatic anomalies. Spectra for constant T_e isostatic anomalies for T_e values of 0, 20, 40, 60, 80 and 100 km in colour. Black line, isostatic anomaly based on recovered variable T_e structure.

within the region where the power spectra were calculated.

T_e map of the northern U.S. and Canada

The T_e map shown in Figure 6.15 displays a wide variation in recovered T_e values ranging from <5 km to over 100 km across the region. The most obvious pattern seen in the T_e map is the pronounced low beneath the west coast of the United States. This low runs inland beneath the Basin and Range region and also below much of the Rocky mountain range. Beyond this low region T_e is much higher, with values of 60 km in the central United States and up to 100 km in north-east Canada.

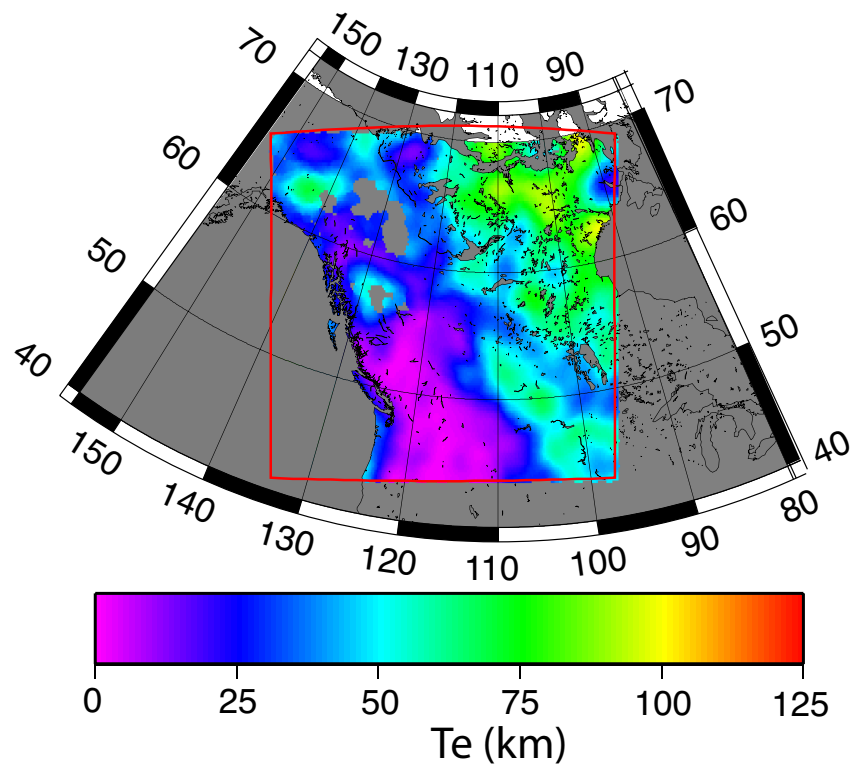


Figure 6.15: T_e map of northern North America and Canada. Regions with no data have been masked out.

Isostatic anomalies across the U.S.

The isostatic anomalies for local (Airy) and regional, (constant T_e), compensation of the observed topography are shown in Figure 6.16. The Airy isostatic anomaly (Figure 6.16a) shows a relatively subdued pattern across much of the region. However, on the eastern margin of the mountains, at around 40°N, there appears to be a high-low couple in the isostatic anomaly, associated with the front of the Rocky Mountains. This couple has same sense as in the Andes and Himalayas, with high values beneath the mountains and low values over the foreland. This suggests that regional compensation of the surface topography is an important factor in this region. As T_e values are increased this couple decreases but large isostatic anomalies develop beneath the mountains and towards the west coast (Figure 6.16b and c).

Figure 6.17 shows the final gravity anomalies calculated from the 2D iterative technique based on a regionally variable T_e . The isostatic anomaly shown in Figure 6.17c shows a slight improvement on the Airy case. The magnitude of the high low couple, close to 40°N, at the mountain front, is decreased but not removed. A spectral assessment of the various isostatic anomalies is shown in Figure 6.17d and demonstrates that the variable T_e structure gives the best reduction in power of the isostatic anomaly.

Recovered T_e map across the U.S.

As found in the northern area Figure 6.18 shows that low T_e values are recovered beneath the Rocky mountains and towards the West Coast across the Basin and Range province. Towards the central United States there is a broad region of high T_e with values ranging from 75 to over 100 km.

Combination of the Canadian and U.S. results

Figure 6.19a and b shows the superimposition of the two recovered T_e structures, and shows that generally the pattern of recovered anomalies is similar. In the

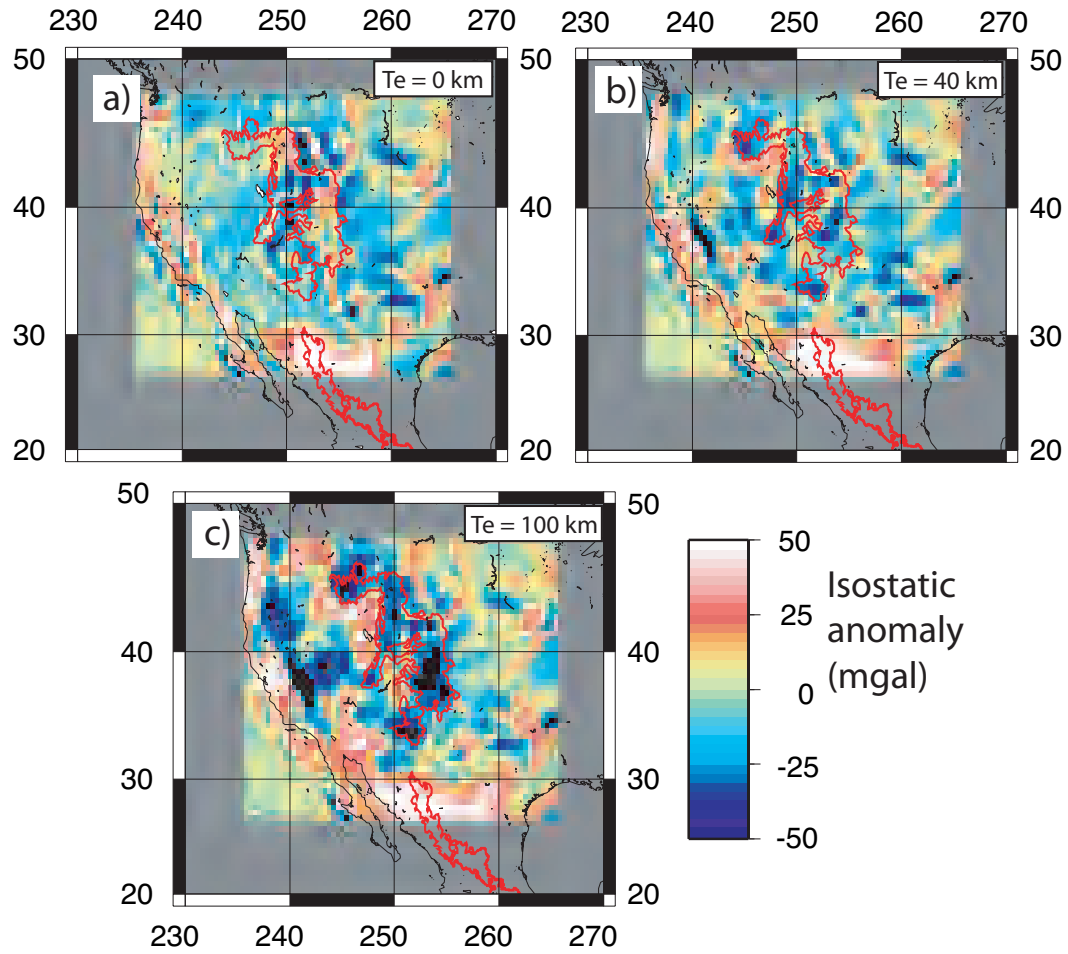


Figure 6.16: Constant T_e isostatic anomalies. Red contour = large areas above 2000m. a) Airy Isostatic anomaly. b) Isostatic anomaly, $T_e = 40$ km c) Isostatic anomaly, $T_e = 100$ km.

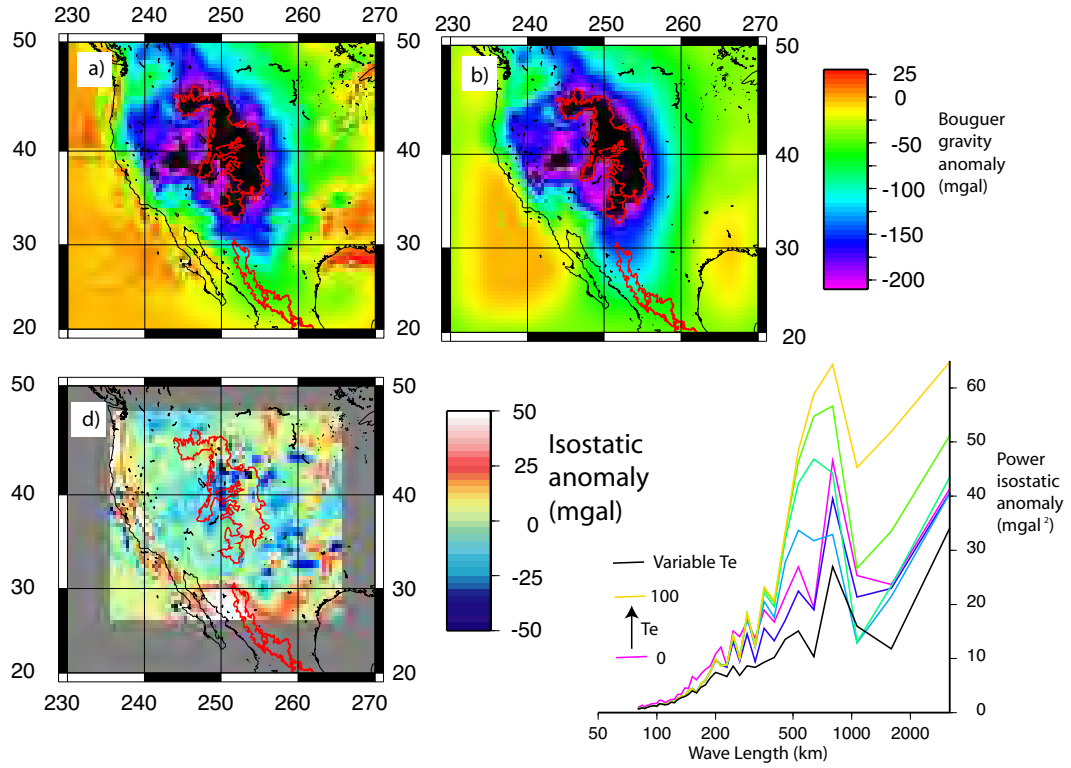


Figure 6.17: Comparison of input Bouguer gravity data and the result of the 2D iterative model. a) Input Bouguer gravity anomaly. b) Calculated Bouguer gravity anomaly based on final T_e structure. c) Isostatic anomaly derived from a and b). d) Spectra of isostatic anomalies. Constant T_e values of 0, 20, 40, 60, 80 and 100 km in color and variable T_e in black.

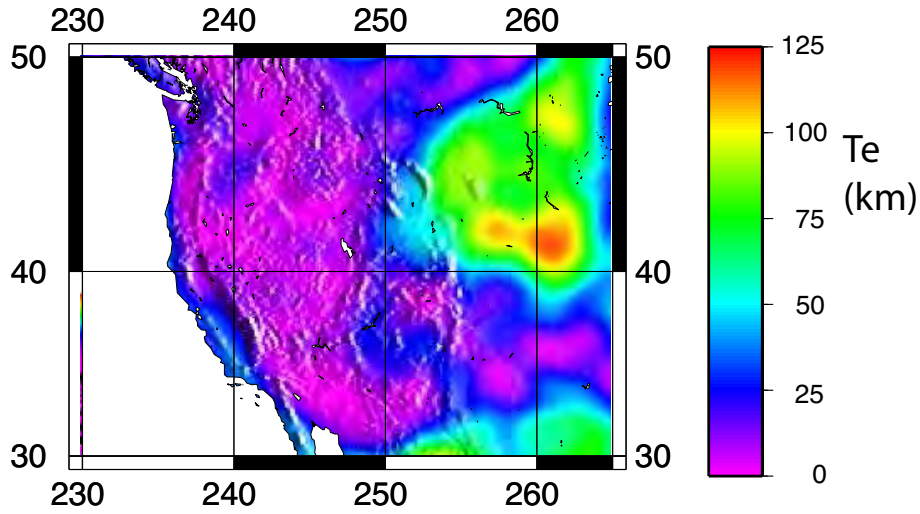


Figure 6.18: Recovered T_e structure across the U.S.

mountainous part of the overlapping region, along the west coast, the results are consistent, with low (<10 km) T_e values being recovered. Over the continental interior, although generally higher T_e values are recovered in both cases, significant discrepancies between the absolute recovered values are seen. Figure 6.19c shows that the difference in recovered T_e may be as large as ± 50 km between the two models. This highlights the need for caution when deriving absolute values from the 2D iterative technique in areas of limited topography. The largest discrepancies in T_e are also close to the margins of the models where T_e is least constrained. When the two models were stitched together the northern model was therefore used in the northern part of the overlapping region, (where the residual T_e was negative), and the southern model was used in the south, (where the residual T_e was positive).

Comparison with previous authors

The most comprehensive study of the regional variation in T_e across the Canadian region was carried out by Flück *et al.* [2003]. These authors used a spectral technique based on the Bouguer coherence method. This recovered a spatially varying T_e structure with maximum values of over 130 km in the north-east of the region and lower values of 10 to 20 km in the coastal mountains. A comparison of the

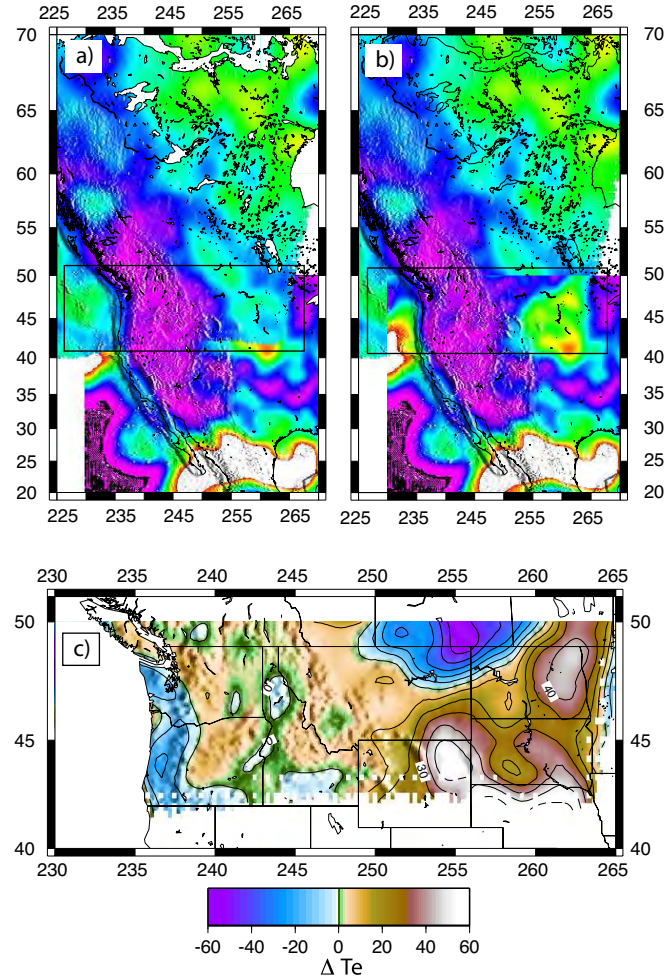


Figure 6.19: Superimposition of T_e structures of Canada and the U.S. a) Northern T_e structure on top, black box marks region of overlap b) Southern T_e structure on top, c) Discrepancy between two T_e structures (Southern - Northern).

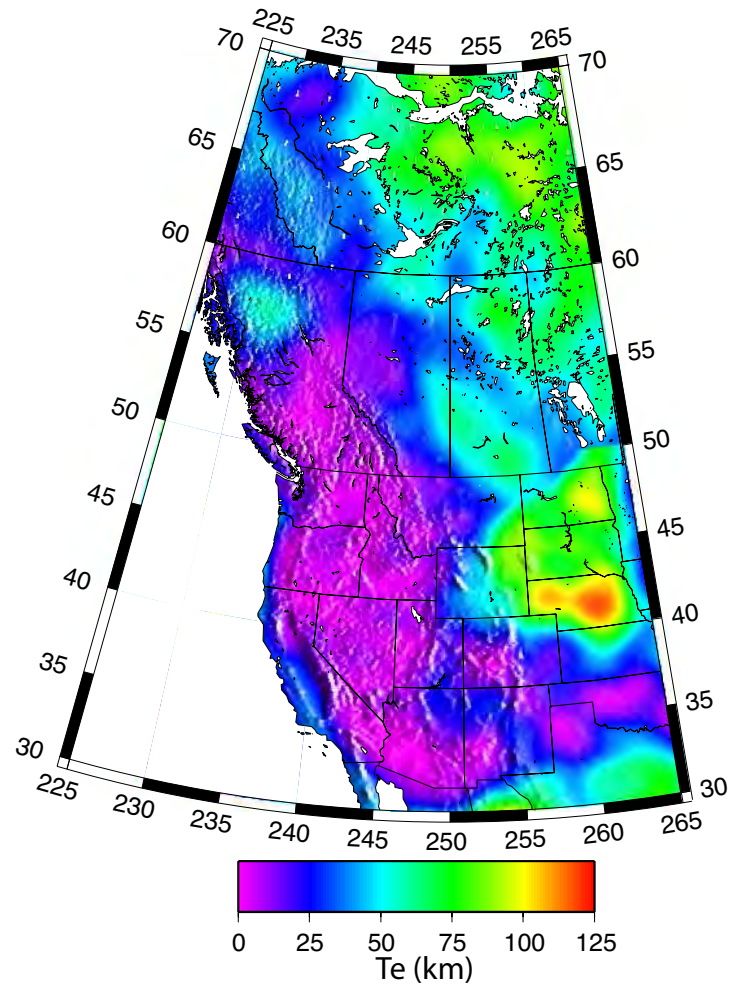


Figure 6.20: Final combined T_e structure for north America.

results of Flück *et al.* [2003] and the 2D iterative technique is shown in Figure 6.21. There is a marked similarity in location of the onset of high T_e in the north-east corner of the study area, with values recovered by both techniques in excess of 80 km in this area. Towards the west coast both techniques recover lower T_e values in the order of 10 to 20 km. These low values are similar to those recovered beneath the central Andes and Tibet, and appear to be typical for modern active orogenic belts. The region of high T_e that is recovered by the 2D iterative technique in the mountainous area, $\sim 57^\circ\text{N}$, doesn't agree with these typical values. This region of high T_e is believed to be an artefact of a lack of gravity data in this region, (Figure 6.22). The lack of data over a region of high topography, leads to the **GMT** program **surface** to underestimate the maximum negative gravity anomaly. The resulting anomaly is equivalent to the topography being supported by a higher T_e .

Other authors such as Wang & Mareschal [1999] and Audet & Mareschal [2004] also recover T_e variations across the Canadian shield using spectral techniques. These are in general agreement with both the results of the 2D iterative technique and the results of Flück *et al.* [2003]. High values are recovered over the Archean terranes in the East while low values are recovered over the mountains to the west. Intermediate values (~ 50 km) are recovered between the Superior ($\sim 50^\circ\text{N}$) and Slave ($\sim 65^\circ\text{N}$) provinces by Audet & Mareschal [2004].

Grotzinger & Royden [1990] have applied forward models to the northern part of the Canadian shield and recovered T_e values of $12 \text{ km} \pm 4$ for the Proterozoic Kilohigok basin within the Archaean Slave Craton. This basin is a 1.9 Ga flexural foreland and the T_e values were recovered by matching flexural models to the shape of the reconstructed foreland stratigraphy. Grotzinger & Royden [1990] suggested that although the low T_e value represented the T_e at the time of loading there has since been significant recovery in the strength of the lithospheric mantle in the area. This would explain the discrepancy between estimates of T_e based on modern loads, such as the 2D iterative technique and the spectral techniques of Flück *et al.*

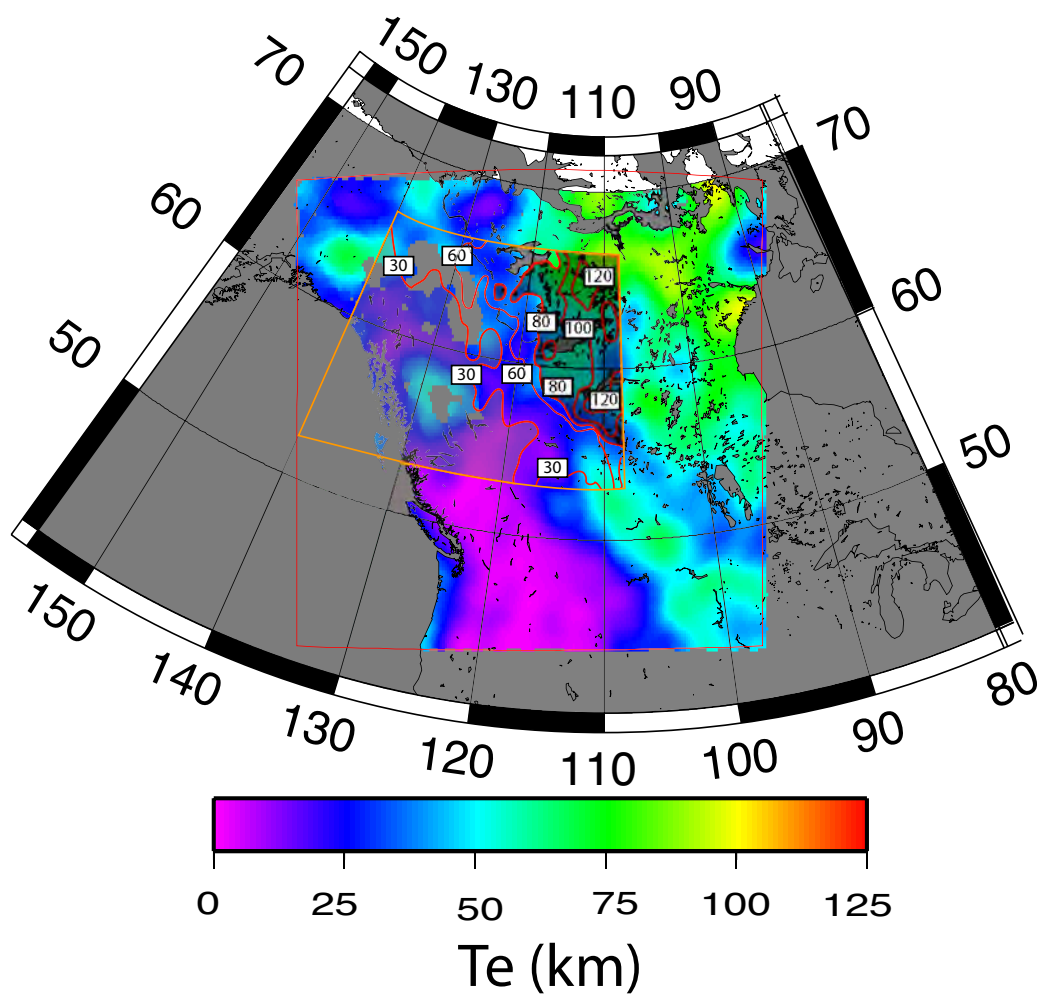


Figure 6.21: Comparison of T_e structure recovered by the 2D iterative technique and by Flück *et al.* [2003] using a Bouguer coherence, spectral technique. Orange box approximate region assessed by Flück *et al.* [2003], Red lines, contours of T_e from Flück *et al.* [2003]. Dark shaded region generally high T_e (>80 km). Light shaded region generally low T_e (<30 km).

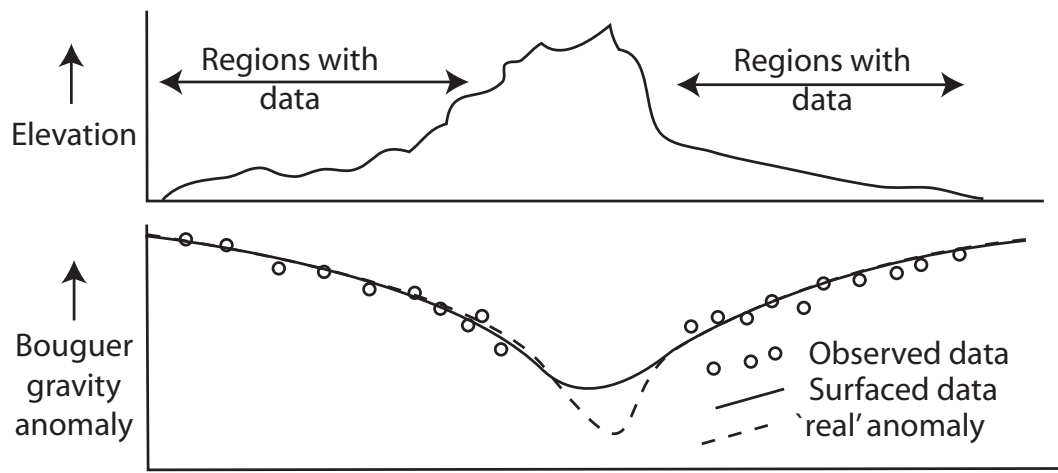


Figure 6.22: Effect of data gap beneath substantial topography. Maximum negative gravity anomaly is clearly underestimated.

[2003] and others, and those T_e values recovered based on Proterozoic loads which no-longer exist.

In the U.S. there have been several studies of T_e based both on spectral estimates and forward modelling. The classic forward modelling study was carried out by Jordan [1981] on the Cretaceous foreland of the Rocky mountains in Wyoming. This recovered a flexural rigidity of 10^{23} N m which equates to a T_e of ~ 22 km. This value is slightly lower than that recovered by the 2D iterative technique in the region. However, Jordan's model was based on a continuous plate rather than a broken or spatially varying one. A continuous plate requires a lower T_e to recover the same basin width due to a driving load. Thus the discrepancy between the two studies is likely to be related to the model used.

Work by Granger & Stock [2004] based on tilted cave carbonates recovered a T_e of 5 km in the Sierra Nevada region. These deposits were laid down horizontally during the last glacial maximum, since this time regional isostatic rebound, due to the removal of the ice load, has lead to regional tilting. These values approximate to those recovered by the 2D iterative technique, which are generally low in this region.

Karner & Watts [1983] applied a 1D flexural model to the Appalachians and

recovered T_e values in the order of 82-131 km. Stewart & Watts [1997] recovered T_e values ranging from 40 to 70 km in the same region with distinct variability along strike. The Appalachians lie to the east of the area considered by the present study. However, the studies of Karner & Watts [1983] and Stewart & Watts [1997], provide estimates of T_e values for the U.S. continental interior. The values reported by these authors are close to those recovered in the continental interior by the present study, supporting the results of the 2D iterative technique.

Bechtel *et al.* [1990] utilised a spectral technique to recover the spatially varying T_e across the North American continent. They recovered similar values to those from the 2D iterative technique. These ranged from ~ 4 km beneath the Basin and Range province to over 100 km across the continental core.

Barbeau [2003] recovers a palaeo T_e value for Utah. This was based on the shape of the Paradox basin which was assumed to be flexural. The recovered T_e value needed to fit the basin shape was 25 km. This is higher than the value recovered by the 2D iterative technique, which is around 10-15 km. That T_e has reduced with time is contrary to what is normally expected due to cooling of the lithosphere with time e.g. Grotzinger & Royden [1990]. This reduction in rigidity suggests the region analysed has become more involved with the orogeny over time. Increased heat flow associated with orogenic processes may have acted to weaken the crust and potentially rejuvenated the mantle, leading to the recovery of a lower T_e .

From this brief discussion of previous investigations of T_e across the north American continent the results of the 2D iterative technique can be seen to compare well with previous estimates of T_e . This is true of both spectral and non spectral techniques.

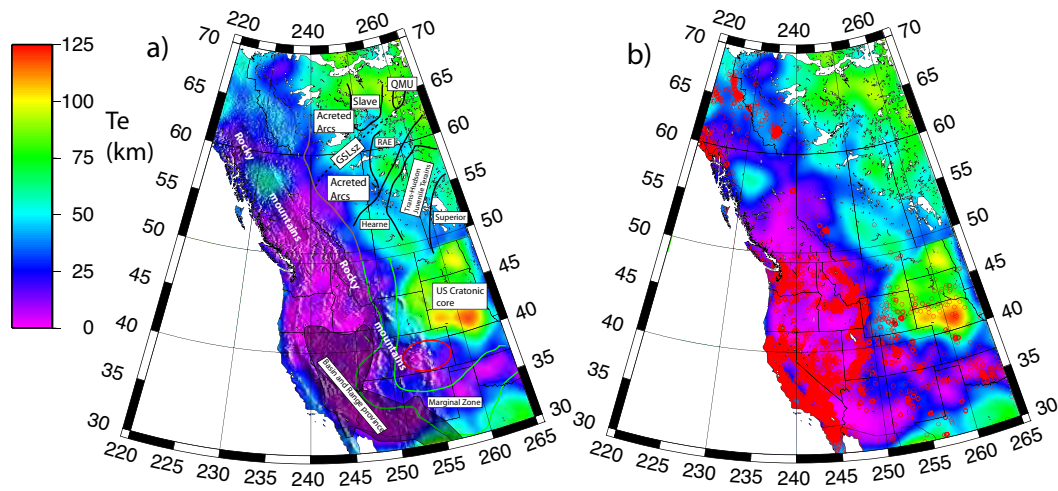


Figure 6.23: Geological data and T_e . a) Geological and physiographic provinces. Terrane boundaries and names for the northern part of the area comes from Hoffman [1987]. QMU = Queen Maud Uplift; GSLsz = Great Slave Lake suture zone. Brown line is the approximate border of the Cordilleran Origin. The boundary of the United States cratonic core, marked in green, comes from Maughan & Perry [1986]. Red oval marks a region of unusually thick crust from Tonnsen [1986]. b) Earthquake distribution from the USGS [2004] earthquake data base, 1973 to recent.

6.3.3 Geological and geophysical correlation

The correlation of recovered T_e with both geological observations, and geophysical measurements of the crust and lithosphere is discussed in this section. Figure 6.23a shows the distribution of geological terranes taken from the literature (Hoffman [1987], Maughan & Perry [1986] and Tonnsen [1986]). Figure 6.23b shows the distribution of all recorded earthquakes from the USGS [2004] earthquake database from 1973 to recent.

The high T_e values in the north-east of Figure 6.23a correlate with the Slave province which had amalgamated by 2.6 Ga (Cook *et al.* [1999]). Cook *et al.* [1999] show seismic reflection data from one of the Lithoprobe Slave-Northern Cordillera Evolution (SNORCLE) profiles. This, they propose, shows mantle reflectors from as deep as 70 km, suggesting the presence of significant thickness of stable Archean

lithospheric mantle beneath this region. The thick Archean mantle pinches out towards the west, beneath the accreted arcs of the Hotta and Fort Simpson Terranes. This boundary correlates with the transition from high (>50 km) T_e , in the north-east of the region, to lower (<50 km) in the south-west.

A key feature of the T_e structure recovered by both the 2D iterative technique and by Flück *et al.* [2003] is the change in strike of the edge of the high T_e zone that appears to be associated with the Great Slave Lake shear zone (GSLsz) in the north-east of the area. Hoffman [1987] suggested that this shear zone marked a transform boundary between an ‘indenting terrane’ in the form of the Slave province and an indented region in the form of the Queen Maud Uplift (QMU). These were seen as analogous to Peninsular India and the Tibetan plateau respectively. If the QMU was analogous to Tibet low T_e values might be expected (e.g. Jordan & Watts [2005]), however, the T_e across this region is consistently high ($>\sim 75$ km). This suggests that T_e has recovered significantly since 1.9 Ga.

In the central U.S. the high T_e region correlates with the area known to be the Precambrian core on the North American craton (Maughan & Perry [1986], Tonnsen [1986]). The margin of this cratonic core is marked in green on Figure 6.23a. The area highlighted by a red oval in Figure 6.23a and associated with generally low T_e values (<25 km) is thought to have been an ocean basin in the Archean (Tonnsen [1986]). Today it appears to be a region of anomalously thick crust with little or no topographic expression (Crosswhite & Humphreys [2003]). If this thick crust was acting as a significant buried load and the expected topography was suppressed by a high rigidity lid then very strong negative isostatic anomalies would be expected. This is not seen in Figure 6.16a, or in the study of Crosswhite & Humphreys [2003]. The interpretation given by Crosswhite & Humphreys [2003], is that the root must be compensated by dense material very close to it within the lithosphere. This compensating mass may be eclogitised oceanic crust preserved from the original suture zone. The recovered low T_e values in this region

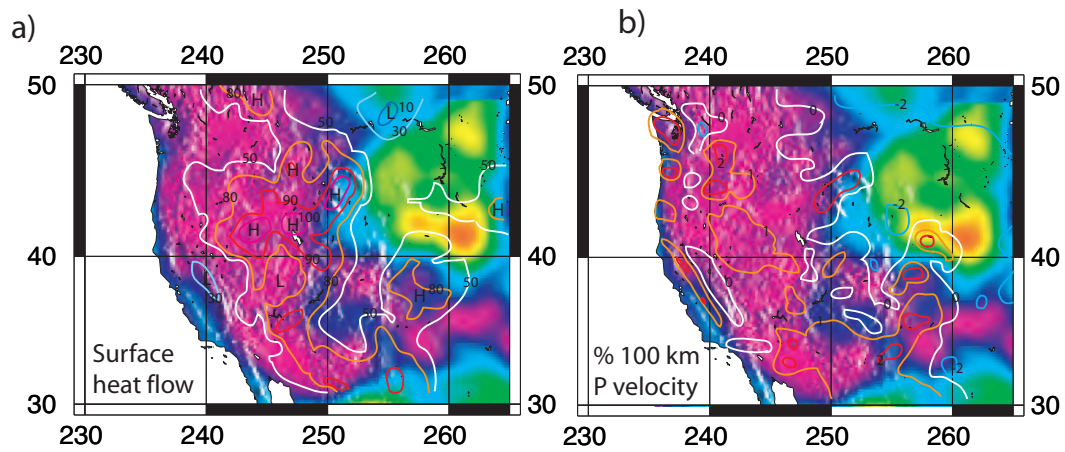


Figure 6.24: Correlation of recovered T_e and geophysical data. a) Correlation of surface heat flow and T_e . b) Correlation of variations in P wave velocity at 100 km depth with T_e .

may be due to the presence of a significant buried load being present which is not correlated with surface topography. However, lower T_e values than the adjacent Archean shield, which is thought to have a thick stable mantle root, would also be expected.

Figure 6.23b shows the distribution of earthquakes and T_e . The margin of the low T_e region in the western U.S. is rimmed by a zone of high earthquake activity. Towards the south-east there is a band of low T_e which correlates with a region of relatively high earthquake occurrence compared to the higher T_e region to the north. This runs along the margin of the Precambrian shield defined by Maughan & Perry [1986]. The correlation of low T_e with earthquake activity suggests that extensive fracturing of the crust weakens it. The localisation of earthquakes along the margin of low T_e region suggests that stresses may be localised along the margin of the high rigidity region.

Heat flow and P-wave velocity variations across the United States from Humphreys *et al.* [2003] are shown in Figure 6.24. There appears to be a clear inverse correlation between these two parameters. This would be expected as low seismic velocity is thought to reflect the underlying mantle temperature, although compositional

variation can also have an effect (e.g. Perry *et al.* [2006]). These variations in heat flow and P-wave velocity also appear to correlate with regional variations in T_e , high T_e being associated with fast P-wave velocities and low heat flow while low T_e regions are associated with slow P-wave velocity and high heat flow. This observation suggests that T_e variation is in part controlled by variations in mantle temperature, supporting the view of for example Watts & Burov [2003] that mantle rigidity is an important contributor to the overall recovered T_e .

6.3.4 Summary

- 1 The T_e variations recovered across the North American continent appear typical of the T_e structure recovered in other regions such as the India-Eurasia collisional system and South America.
- 2 The modern active orogenic belt along the western margin of the continent is marked by low T_e .
- 3 Towards the centre of the continent the T_e recovered is high reflecting the thicker older lithosphere.
- 4 Regions of low T_e appear to correlate with areas where other geophysical indicators, such as high heat flow and low P-wave velocity, indicate the mantle is hotter than normal.

6.4 Europe

Continental Europe has been the subject of many geological and geophysical investigations. The potential geological causes for recovered T_e variation are therefore well constrained. To investigate the link between geology and T_e variation the 2D iterative technique was applied across Europe, from the Mediterranean in the south to Scandinavia in the north and to the Black Sea in the east. The results of this were compared with geological and geophysical features such as suture zones, earthquake distribution, thermal structure and S-wave velocity. An additional reason to recover the T_e structure across Europe using the 2D iterative technique are the studies of Pérez-Gussinyé *et al.* [2004] and Pérez-Gussinyé & Watts [2005]. These recover maps of T_e variation across Europe based on spectral techniques. Comparison between these previous studies and the 2D iterative technique show good correlation. This helps to confirm that the 2D iterative technique is a robust method for recovering T_e .

6.4.1 Data

The topographic and gravity data used in this investigation are shown in Figure 6.25. The data is from the WEAGP database (Green & Fairhead [1994]). Onshore data coverage is generally good. However, coverage is poor for Serbia, Montenegro, and Turkey. The data was re-projected onto a Cartesian grid using a Lambert conic projection. This grid was then sampled up to 25x25 km grid spacing to allow the iterative program to run quickly while maintaining resolution. The edges of the topographic data were tapered using a cosine taper to prevent edge effects entering the flexural model. Topographic elevation offshore was set to zero.

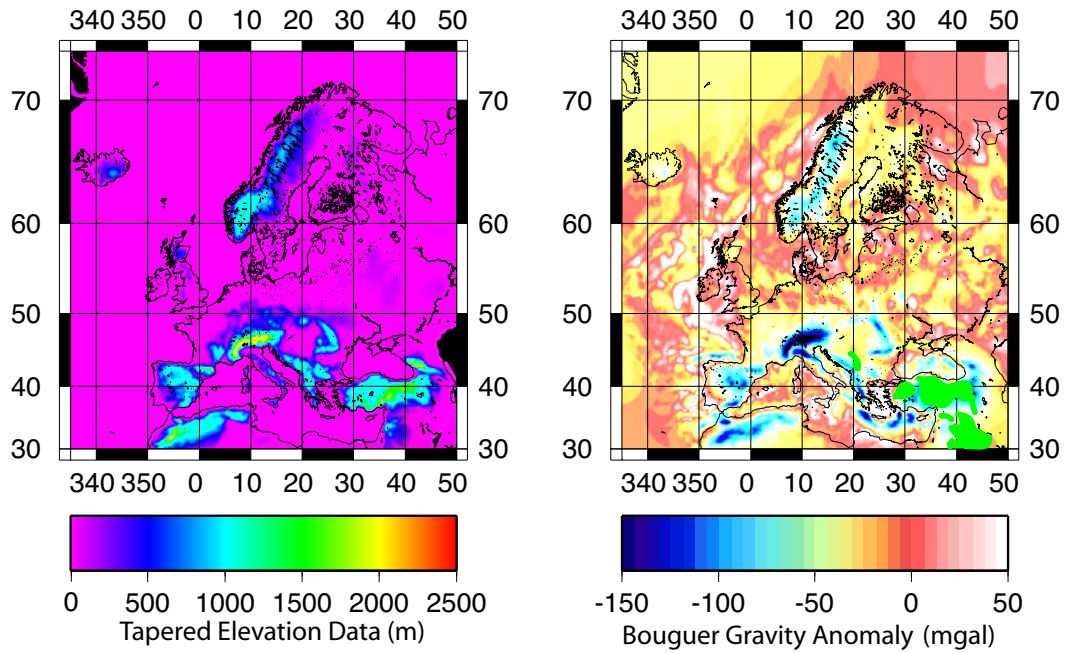


Figure 6.25: Data used in the 2D iterative technique. a) Elevation data, tapered at the edges. b) Bouguer gravity anomaly data. Green areas mark regions with no data coverage.

6.4.2 Results

Isostatic anomalies

Initially the flexural compensation of the topography was assessed from isostatic anomalies generated by considering Airy, or constant rigidity regional isostatic compensation of the observed topography. The results of this assessment are shown in Figure 6.26. The isostatic anomalies indicate that Airy compensation of the topography beneath the Alps gives the best fit to the data. However, in other regions no T_e gives a clear reduction in the observed isostatic anomaly.

Despite the lack of clear evidence of regional isostatic compensation of the surface topography, derived from constant T_e isostatic anomalies, the 2D iterative technique was applied to the data set. After 50 passes of the 2D iterative technique the isostatic anomaly was reassessed, based on a comparison of the observed Bouguer gravity anomaly and the calculated gravity anomaly. These are shown in

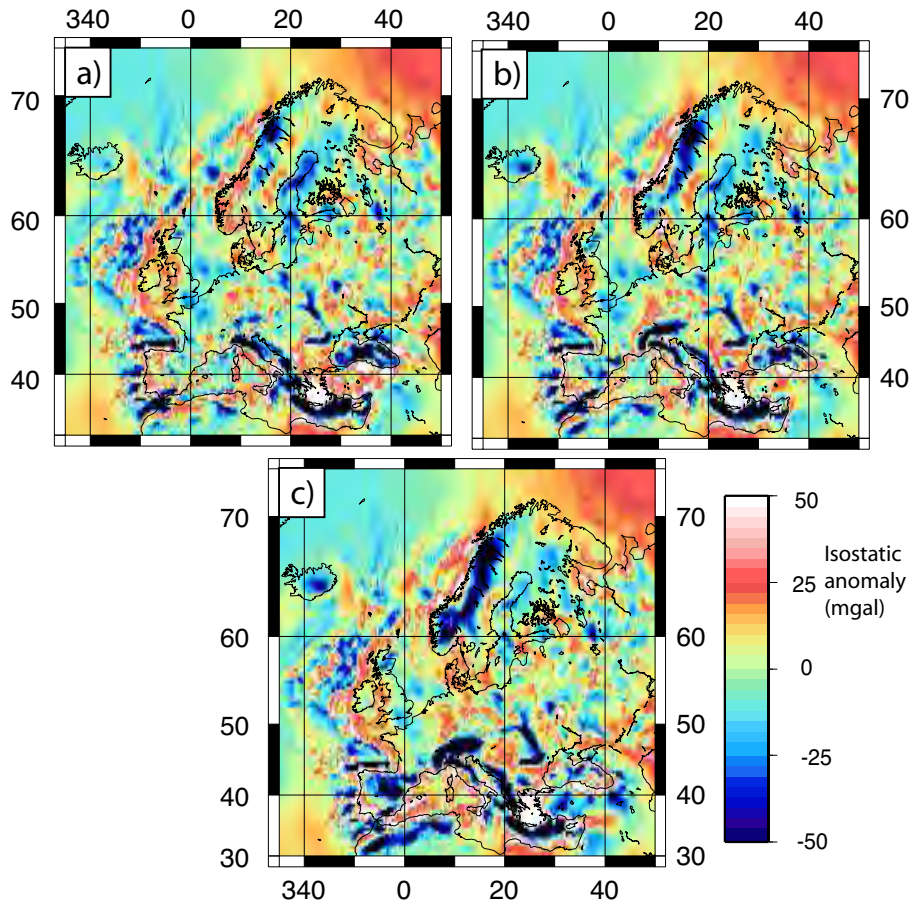


Figure 6.26: Modelled gravity anomalies based on local or regional compensation of the surface topography, and associated Isostatic gravity anomalies. Models are for constant T_e structures of a) 0 km (on-shore Airy compensation), b) 40 km and c) 100 km.

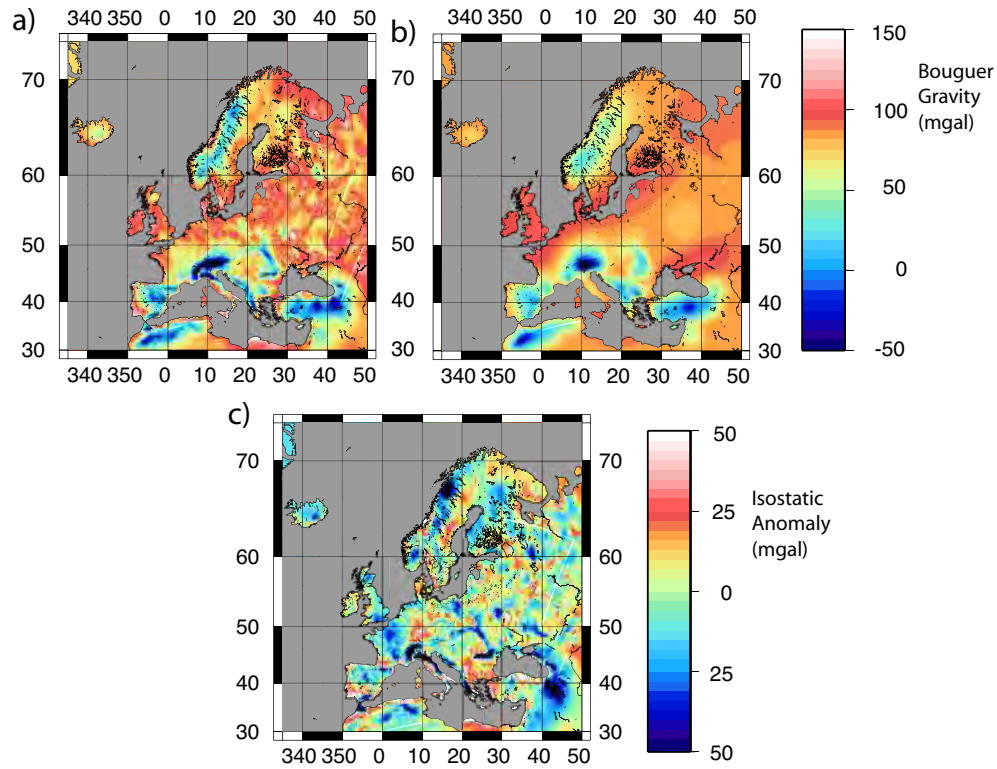


Figure 6.27: Input Bouguer gravity anomaly and calculated gravity and isostatic anomalies after 50 iterations of the 2D iterative technique. a) Input Bouguer gravity data. b) Calculated Bouguer gravity data based on regionally variable T_e structure. c) Regional isostatic anomaly.

Figure 6.27a and b. The regional isostatic anomaly shown in Figure 6.27c shows no clear visual improvement over the Airy or other constant T_e isostatic anomaly shown in Figure 6.26. However, Figure 6.28a shows that the power spectra of the isostatic anomaly derived from the regional T_e model, has less spectral power than any of the other constant T_e models. Hence, the regionally variable T_e model does provide the best fit to the observed Bouguer anomaly for a model based on the isostatic compensation of the present day surface topography.

The range of wavelengths over which the variable T_e structure gives the greatest reduction in power is for wavelengths between 300 km and 1600 km. This range is within the ‘diagnostic waveband of flexure’ shown in Figure 6.28b. This range of wavelengths was proposed by Watts [1982] and Watts [2001], as the range of

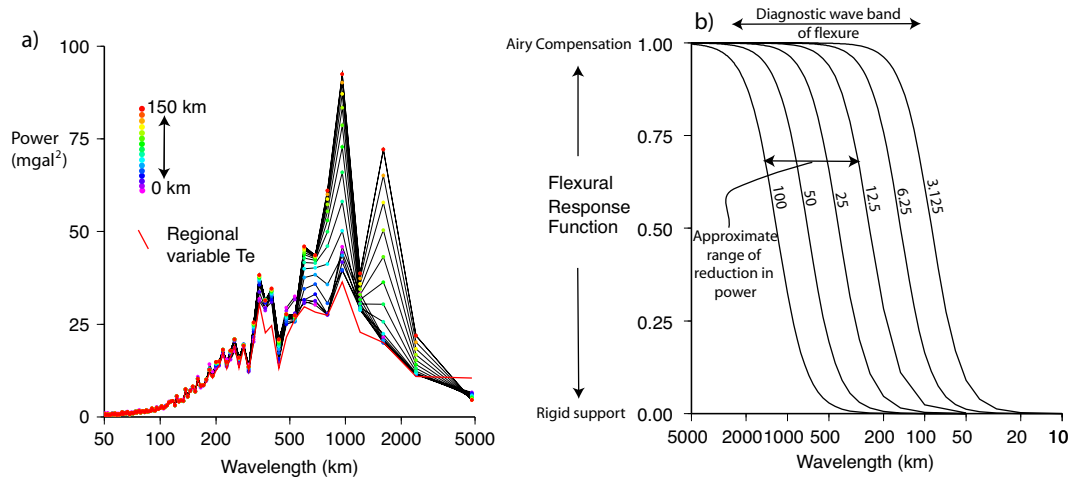


Figure 6.28: Spectral analysis of isostatic anomalies and flexural response function. a) Power spectra of isostatic anomalies based on loading of various T_e structures. b) Flexural response functions for various T_e values.

topographic wavelengths where a flexural response will typically be seen. For a given T_e , above a certain wavelength, topography will appear Airy compensated, while below a certain wavelength it will cause no flexure and be rigidly supported. Between these wavelengths, the topography will be flexurally supported. If a range of T_e values exists there will be a range of topographic wavelengths where flexure contributes significantly to the support of the topography. It is within this band of wavelengths where a flexural model, such as the final result from the 2D iterative technique, would be expected to give the largest reduction in spectral power. Figure 6.28a shows that there is little reduction in the power of the isostatic anomaly below 300 km for the variable T_e model. From Figure 6.28b this suggests that lower T_e values, below ~ 12.5 km, are having little effect on the regional gravity anomaly we are assessing.

T_e map

Figure 6.29a shows the best fitting T_e structure for Europe after 50 passes of the 2D iterative technique. The T_e structure recovered is highly variable; Low values, generally less than 25 km, are recovered in southern Europe. Conversely Northern

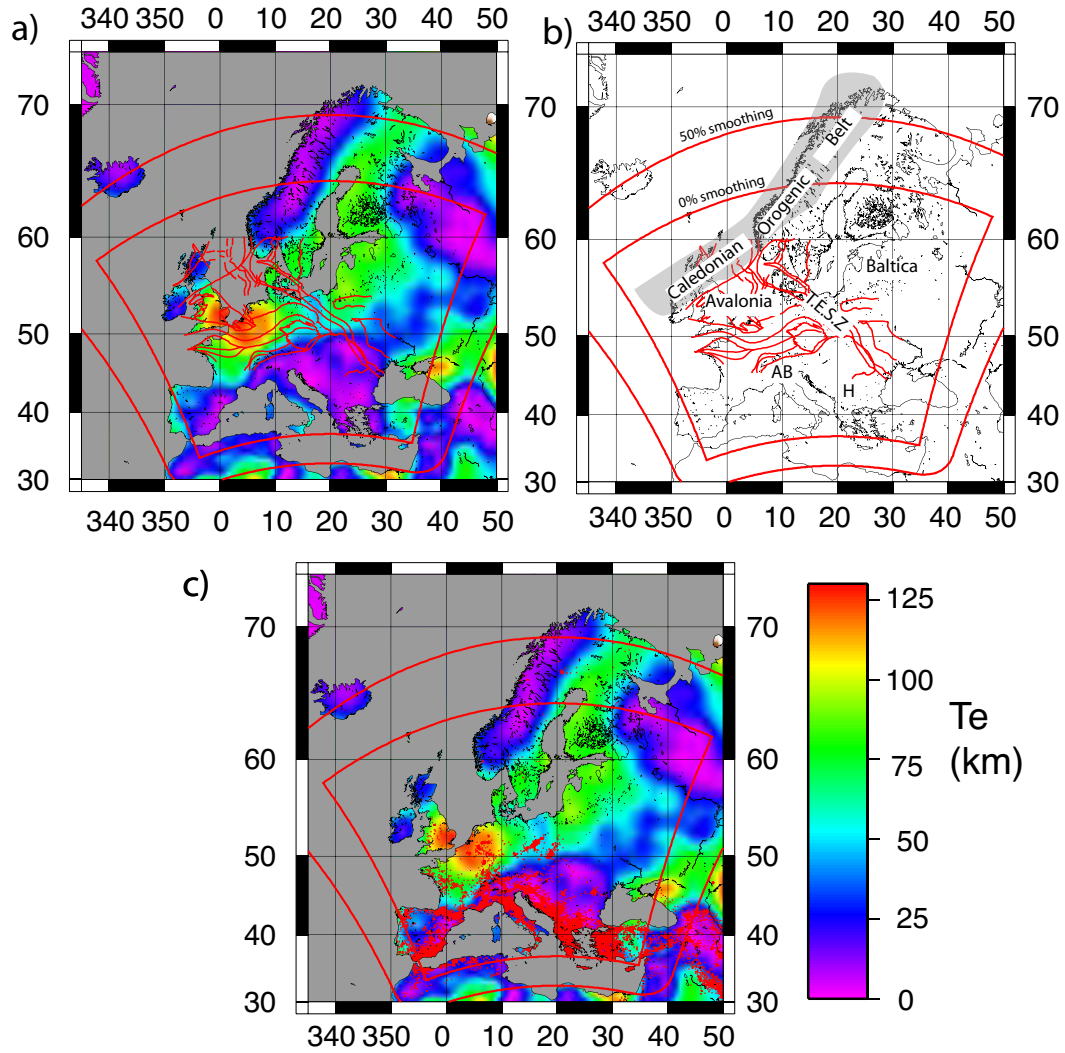


Figure 6.29: Recovered T_e map and geological structures a) T_e map of Europe after 50 iterations with geological province boundaries overlaid. b) Geological provinces (labelled) and suture zones (thin red lines) across Western Europe from, for example, Pharaoh [1999]. Red boxes mark limits of tapered region. AB, Alpine Belt; H, Hellenic region; T.E.S.Z, Trans European Suture Zone. c) T_e map of Europe and earthquake data from the USGS data base for 1980 to 2004, magnitudes >3 (USGS [2004]).

Europe is predominately more rigid (T_e values over 50 km).

It was this T_e structure which, when loaded with the observed topography, generated the final model Bouguer gravity anomaly. This was in turn used to generate the regional isostatic anomaly shown in Figure 6.27.

6.4.3 Previous work

Profile based models

The results of this study broadly agree with those of previous authors. Karner & Watts [1983] and Royden [1993] calculated the rigidity of the foreland basin in the Alpine region based on 1D profiles. Royden [1993] recovered values of T_e in the region of 40 and 20 km to the north and south of the Eastern Alps respectively and 50 and 15 km to the north and south of the Western Alps based on gravity and basement depth data. Karner & Watts [1983] recover values of 25 to 50 km along the length of the northern Alpine foreland based on Bouguer gravity anomaly data. The 2D iterative technique recovers T_e values of approximately 25 to 50 km beneath the frontal thrusts of the northern Alpine zone, with the higher values being found in the west.

Although the foreland values of T_e agree well between the 1D studies and the 2D iterative technique, Royden [1993] proposed the region of high T_e ran further beneath the mountain belt to where the plate was modelled as broken. The 2D iterative technique however, finds low T_e values beneath much of the mountain belt. This discrepancy may relate to the presence of buried loads suggested by both Karner & Watts [1983] and Royden [1993]. These extra loads would need a more rigid plate beneath the mountains to support them. In this study buried loading is not taken into account. However, given that all three studies agree relatively well with regard to the foreland T_e value, the 2D iterative technique seems robust in areas where there are significant surface loads.

2D models

Gaspar-Escribano *et al.* [2001] recovered spatial variations in T_e associated with the Ebro basin in the NE of Iberia. This study was largely based on loading a broken plate model with the topography present in the Pyrenees. However, buried loading beneath the mountain ranges was needed to fit the final flexure. Despite the presence of a broken plate, the T_e structure recovered by Gaspar-Escribano *et al.* [2001] decreases from 30 km to 5-20 km away from the mountains. The results of the 2D iterative technique also show T_e to decrease away from the mountains in this region. The 2D iterative technique does not appear to recover a plate break (region where T_e falls to zero) beneath the mountains. This may reflect the relatively coarse scale of the 2D iterative model.

Braitenberg *et al.* [2002] analyzed the T_e structure of the Eastern Alps using a similar technique to that used in this study. The window size used by Braitenberg *et al.* [2002] was 105 km and it was shifted by 15 km for each RMS analysis. This is close to the window size used in the 2D iterative technique of 125 km with a shift of 25 km each time. Figure 6.30 shows both studies recover a similar pattern of T_e variation. A broad low is seen to be present beneath the central part of the Eastern Alps, and a more rigid region to the north. However the T_e minima do not completely match between the two studies. The map recovered by the 2D iterative technique shows the minimum T_e values to lie directly below the highest topography while the minima in the study of Braitenberg *et al.* [2002] lies further to the north. This may reflect the incorporation of sedimentary loads in the study of Braitenberg *et al.* [2002].

The absolute values of T_e recovered by the 2D iterative technique also differ from those recovered by Braitenberg *et al.* [2002]. In the central region higher T_e values of 6 to 8 km are recovered compared to the lower values of 2 km recovered by Braitenberg *et al.* Over the foreland region the discrepancy is even more marked with the maximum value recovered by Braitenberg *et al.* being 19 km while the

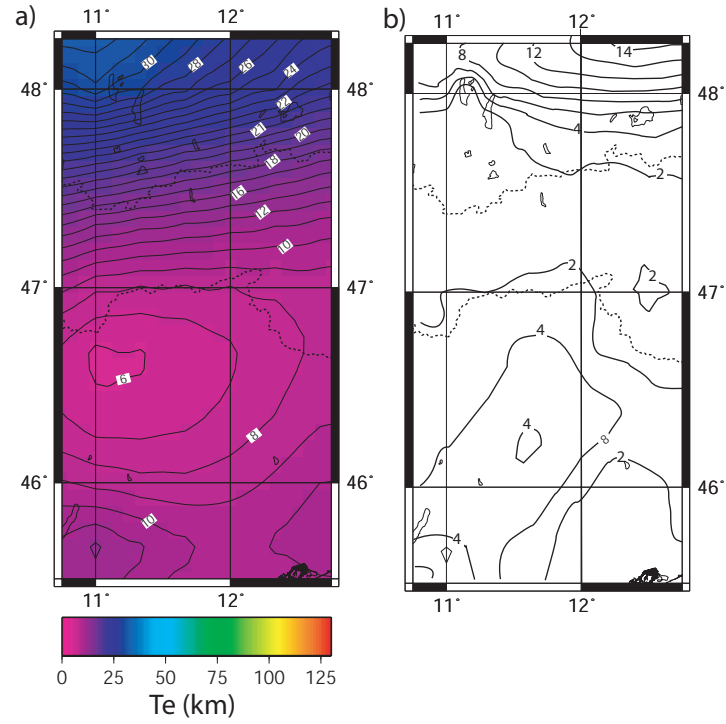


Figure 6.30: Comparison of results from the 2D iterative technique and results of Braitenberg *et al.* [2002]. a) Results from a subsection of this study, interpolated onto a 5 min grid. b) Results of the study by Braitenberg *et al.* [2002].

present study recovered T_e values of approximately 30 km. This reflects the action of the Gaussian filter which is applied to the T_e structure recovered after each iteration. This filter would have the effect of increasing T_e values close to the mountains if there was a region of higher T_e further out in the continental shield.

An alternative technique for calculating T_e , from deflection caused by erosional unloading rather than loading effects, has been applied by Watts *et al.* [2000] to the uplifted region surrounding the Severn Valley, England. The technique of Watts *et al.* [2000] was based on fitting the magnitude and extent of the surrounding escarpment. The T_e value which best fitted the observed data was 10 km. This contrasts strongly with the results for the Severn Valley region recovered by the 2D iterative technique, which are in the order of 100 km. The high values recovered by the 2D iterative technique may reflect the smoothing of high T_e values from the Avalonian continental block into regions which actually have a lower T_e . However, direct comparison between the two studies is difficult because the Severn Valley is a relatively small sample area and the driving ‘unload’ is only 25 to 50 km across while the present study involves the assessment of regional T_e variations on the order of 100s of km. In addition Figure 6.28b shows that a driving load of this size would only show a flexural response if T_e was less than ~ 10 km. Thus any technique applied to this specific region would not recover high T_e values even if they were present. It has been suggested by Westaway *et al.* [2002], that in the Severn Valley region the upper crust and mantle have been decoupled from each other by a low viscosity layer. This would mean that relatively small loads/unloads would only be affected by the rigidity of the top layer. The T_e recovered based on small local loads would therefore be small, while larger more regional loading would be affected by both layers and hence regional studies would recover a much higher T_e .

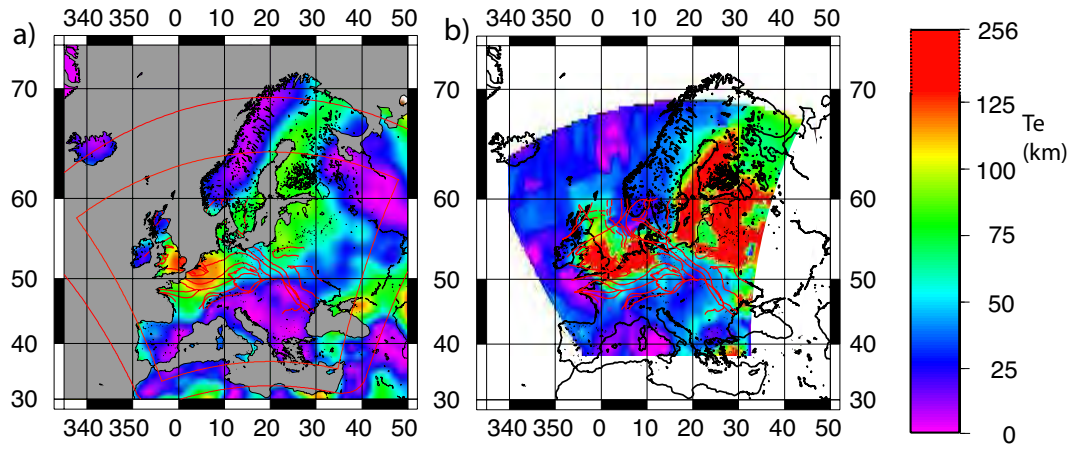


Figure 6.31: Comparison of results of Pérez-Gussinyé & Watts [2005] and this study. a) T_e structure recovered in this study. b) T_e structure recovered based on a spectral, free air admittance, technique from Pérez-Gussinyé & Watts [2005].

Spectral techniques

Pérez-Gussinyé *et al.* [2004] and Pérez-Gussinyé & Watts [2005] used alternative spectral techniques to assess the T_e structure of the European continent from Scandinavia in the north to the Mediterranean in the south. Figure 6.31 compares the T_e structure recovered by Pérez-Gussinyé & Watts [2005] to that recovered by the 2D iterative method in the present study. The results of the 2D iterative technique concur with the pattern of variation shown by Pérez-Gussinyé & Watts [2005] across central and northern Europe. The most significant discrepancy is in the north-east of the study area in Russia. This region has generally subdued topography (<250 m) and a large number of relatively short wavelength, positive Bouguer gravity anomalies are observed. This suggests that there are a significant number of buried loads not associated with surface topography. Pérez-Gussinyé *et al.* [2004] is based on the technique of Forsyth [1985] which takes into account buried loading. Buried loads which are not correlated with surface topography have, however, been shown in synthetic tests in Section 2.7.1 to lead to the recovery of anomalously low T_e values when the 2D iterative technique is applied.

High T_e values are recovered by Pérez-Gussinyé & Watts [2005] over much of

the Baltic shield region while lower values are recovered over the Alpine orogenic belt. The spectral techniques, especially free air admittance, recover a low T_e zone associated with the Trans-European Suture Zone (Pharaoh [1999]). This region of low T_e is also seen in the results of the 2D iterative technique.

The absolute values reported in Pérez-Gussinyé & Watts [2005] run up to 256 km for the very rigid regions. This is far in excess of the values recovered by the 2D iterative technique. However, the very high values were not considered by Pérez-Gussinyé & Watts [2005] to be a true reflection of the T_e of the region. Pérez-Gussinyé & Watts [2005] proposed a maximum T_e value of 60 km where values were known confidently, above this value T_e is known to be >60 km but its value is undetermined. It is difficult to confirm the absolute value of T_e in very rigid regions using spectral techniques because each analysis uses a fixed T_e across the entire analysis window. Pérez-Gussinyé *et al.* [2004] showed that when recovering the rigidity structure for an isolated area of high T_e there is an intrinsic trade off between spatial resolution of the T_e structure and the recovery of its absolute value. Small (800 km) windows were shown to give relatively good resolution but the recovered T_e values significantly overestimate the true T_e . Large (1200 km) windows gave a more stable recovery but the T_e structure tended to be smoothed across the area considered and the maximum recovered T_e was lower.

The 2D iterative technique used in this study suffers less from this problem as the T_e structure can vary within an analysis window. However, the 2D iterative technique does not account for the presence of buried loads. The use of both spectral and modelling techniques therefore gives the best recovery of T_e across a geologically complex region such as Europe.

6.4.4 Correlation of T_e variation and geological features

Low T_e and active tectonics

The low T_e regions across much of Southern Europe correlate with regions of active tectonic deformation. This is highlighted in 6.29c which shows that the low T_e regions in the compressional Alpine and Pyrenean mountain belts, and in the extensional regions of the Balkans and Greece, are all regions of relatively intense earthquake activity. This correlation suggests that brittle fracturing of the crust, as revealed by earthquakes, is reducing the integrated strength of the lithosphere as a whole.

Figure 6.32 shows that, as noted by Pérez-Gussinyé & Watts [2005], low T_e also correlates with geophysical indicators of anomalously high mantle temperatures. Such geophysical parameters include low upper mantle seismic velocity (e.g. Perry *et al.* [2006] and Boschi *et al.* [2004]) and high predicted mantle temperatures from heat flow data (Artemieva & Mooney [2001]). That mantle derived signatures show a correlation with T_e suggests that variations in the properties of the lithospheric mantle also play an important role in controlling the T_e structure across much of Europe.

Cratonic regions

There are two main cratonic regions in Europe, Baltica and Avalonia, and these appear to be regions with generally high recovered T_e . Baltica has existed as a distinct entity since the Proterozoic when it rifted away from the Rodinian supercontinent (Pharaoh [1999]). Baltica stretches from the North Sea in the west to the Urals in the east. Internally it has remained largely undeformed since its rifting from Rodinia. Around its margins it has been tectonically reworked by successive cycles of collision and rifting. As discussed previously the low T_e values recovered in the north east of the region, across Russia, are believed to be an artefact of buried loading. The results of spectral studies such as Pérez-Gussinyé & Watts

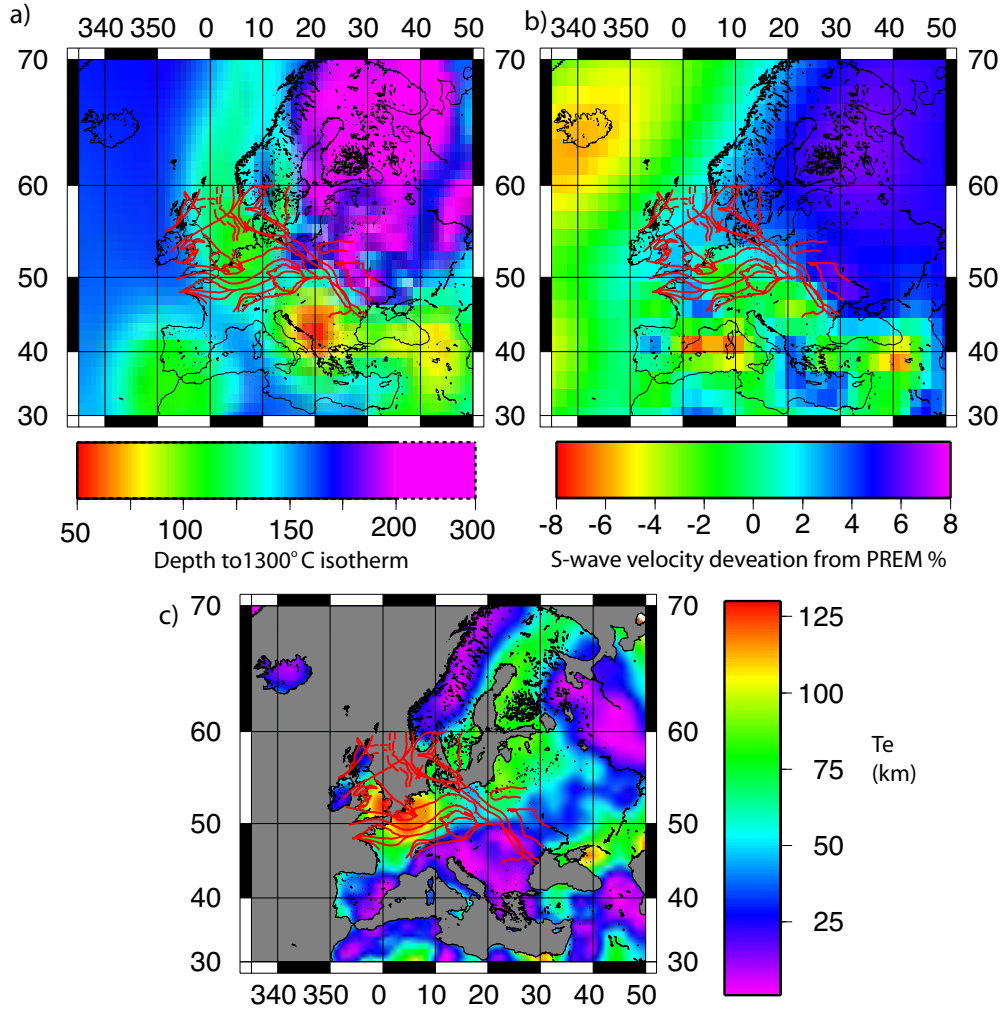


Figure 6.32: Comparison of geophysical observations, modified from Pérez-Gussinyé & Watts [2005], and the recovered T_e . a) Depth to 1300 °C Isotherm from Artemieva & Mooney [2001]. b) S-wave velocity at 200 km depth, departures from PREM described in Boschi *et al.* [2004]. c) T_e structure recovered by the 2D iterative technique.

[2005] which recover $T_e > 60$ km in this region are therefore preferred.

Avalonia is a much smaller microcontinental fragment that lies to the south of Baltica. It consists predominantly of accreted terrane fragments around the core of the ‘Midlands micro craton’. This cratonic core has remained largely undeformed while the terranes around its margins have been repeatedly tectonically reactivated (Pharaoh [1999]).

The central Iberian region also shows up as an area of relatively high T_e (~ 50 km). This area consists largely of Palaeozoic basement material and constitutes a fragment of the Gondwanan super-continent (Catalán *et al.* [2004]). The high T_e region in central Iberia is ringed with earthquakes. The margins of the cratonic fragment therefore appear free to deform while the core remains rigid. This may present an analogous situation to the lack of deformation within the Midlands micro craton within Avalonia.

The correlation of high T_e with old cratonic regions outlined above suggests that T_e is, in part, a function of age as suggested by for example Karner & Watts [1983] and Pérez-Gussinyé & Watts [2005]. The lack of earthquakes within these old cratonic regions suggests that deformation localises in weaker, lower T_e , areas at their margins. This explains why these regions appear to remain relatively unaffected by tectonic processes which preferentially deform the regions with lower T_e .

Sutures and T_e

In central and northern Europe the T_e is higher than in the south, and the earthquake activity is less intense. However, variations in T_e do appear to correlate with geological structures and hence show the imprint of previous geological activity. This is particularly true for the Trans-European Suture Zone (T.E.S.Z.) which marks the geological boundary between Baltica and Avalonia, Pharaoh [1999]. This suture also shows up clearly as a shallowing of the 1300° isotherm towards the south

and a decrease in S-wave velocity at 200 km depth towards the south (Figure 6.32). This suggests that the variations in T_e are reflecting fundamental variations in lithospheric structure across the suture.

Fennoscandia in the northern part of the study area marks the northern margin of the Baltica cratonic fragment. The northern edge of this region appears to have a very low rigidity, which correlates with the Scandinavian Caledonides. This region of low T_e , associated with the Caledonian suture, also marks the northern boundary of the Avalonian block.

The Caledonian suture occurred in the mid Ordovician when Laurentia collided with Baltica and Avalonia as the Iapetus ocean closed (Friedrich *et al.* [1999]). However, the Caledonian Orogeny (and orogenic belt) has a much longer history ranging from Cambrian to Early Devonian times with the successive collision of island arc terranes before the final closure (McKerrow *et al.* [2000]). The T.E.S.Z. is of a similar age, but appears to present a much narrower band of low T_e . This difference may relate to the different scale of the collision. The T.E.S.Z. results from the ‘docking’ of a relatively small microcontinental fragment and hence the associated orogeny would be expected to be more minor. Associated with the T.E.S.Z. there is evidence only of short lived magmatism associated with continental subduction (Verniers *et al.* [2002]). In contrast the Caledonian Orogeny was a more significant collision between two large continental land masses which led to a relatively large amount of magmatism (Watson [1984]). It has been suggested that much of the Caledonian magmatism in Scotland is mantle derived. Watson [1984] suggests that such mantle derived magmatism may result from hydration of mantle material due to subduction of hydrated oceanic crust or from the loss of an old, cold mantle root from below the orogen as collision progressed.

As discussed in Section 4.5.3 the high rigidities recovered of >40 km in some regions must relate to variations in the mantle. The correlation of low T_e values with other geophysical signs of hot mantle, discussed previously in this section,

also suggests significant mantle control on T_e . Rejuvenation of the mantle beneath the Caledonian Orogen, either by hydration or delamination, may therefore explain the different rigidity structures of the two sutures.

6.4.5 Summary

1 The maximum topographic loads are much less significant in Europe than in the India-Eurasia Collisional system or South America, however, the 2D iterative technique still produces a T_e map that shows relatively good correlation with regional geology.

2 The estimated T_e values recovered are, in most areas, consistent with the results of other authors, (based on both 1D and 2D modelling, and spectral techniques).

3 The recovered T_e variation is typical of that observed in continental regions, with values of up to 120 km over shield areas and low values down to < 5 km in both extensional and compressional active orogenic belts.

4 There is a clear correlation of low T_e values and other geophysical signatures indicating hot mantle, such as the thermal structure derived from heat flow data, or S-wave velocities. This confirms that the mantle plays a significant role in determining the recovered T_e across the region.

6.5 Evaluation

- 1) The 2D iterative technique works in a variety of continental areas, as long as the proportion of topographic loading significantly exceeds the amount of buried loading.
- 2) The resulting T_e structure generally agrees with the pattern and order of magnitude of the T_e recovered by spectral techniques.
- 3) There is a correlation between the recovered T_e and the geological structures in an area. Primarily old shield regions are rigid while active modern orogens are weak.
- 4) Suture zones between continental fragments are generally weak, as seen in Europe.
- 5) During the Archean continental crust could recover its strength (as in the Slave province in Canada). Since the Archean, if continental crust becomes weak it remains weak.
- 6) There is a correlation between T_e and width of orogeny in collisional systems. Rigid forelands appear to lead to thin skinned thrusting and wider, more Himalayan, mountain belts.
- 7) There is a correlation between geophysical indicators of a hot mantle, such as seismic wave velocity and heat flow data, and low T_e . Given the high T_e values of >80 km recovered in continental shield regions this suggests that the mantle is contributing significant strength in these regions, when it is cool.

Chapter 7

Discussion

This study has recovered spatial variations in T_e across broad regions of the continental lithosphere using both profile- and grid-based modelling techniques. The T_e values recovered range from less than 5 km to over 80 km over distances of ~ 400 km. These variations are shown to correlate with past and present variations in geological processes, as outlined in the previous chapters.

In this discussion the broader implications of the T_e variations recovered by this study are considered. These include the relationship between recovered T_e and lithospheric age; T_e variation and the implications for continental rheology; and the impact of variation in T_e on continental tectonics.

7.1 T_e and lithospheric age

The relationship between recovered continental T_e and the age of the lithosphere at the time of loading is complex. Studies such as Karner & Watts [1983] suggested that continental T_e was controlled by the depth of the 450°C isotherm, as it appeared to be in the oceans. T_e was therefore linked to lithospheric age at the time of loading due to cooling of the lithosphere with time. Watts [2001], however, showed that the broad range of continental T_e recovered by a larger number of studies could not be explained only in terms of the depth of a single isotherm. The more recent study of Pérez-Gussinyé & Watts [2005] across Europe showed a dis-

tinct decrease in recovered T_e between > 1250 Ma Archean and Early Proterozoic, and more recent terranes, proposed to be linked with a decrease in partial melting of the mantle with time. This change in partial melting lead to the formation of less depleted mantle lithosphere, which is weaker than that found in older regions.

A distinction must be made, when investigating the relationship of T_e and age, between absolute lithospheric age as seen today and the age of the lithosphere at the time of loading. If rigidity increases with age the same region could give different recovered T_e values depending on the age of the load considered. This effect was discussed in Section 6.3.2 in relation to northern Canada, where T_e values of 12 and >60 km have been recovered in the same region. The maps of T_e variation presented in previous chapters are based on the present-day topography and therefore represent a view of continental T_e due to isostatic compensation of modern loads.

Lithospheric ages considered in this chapter are from the global compilation made by Artemieva [2006]. These ages relate to the last major tectono-thermal event to affect a region, rather than the juvenile (oldest recovered) age in each area. Considering T_e variation relative to the last major tectono-thermal event is useful, as orogenic processes appear to re-set T_e to low values, as seen for example in the Himalayas, Andes and along the west coast of North America where consistently low ($< \sim 15$ km) are recovered. Assessing T_e against the tectono-thermal age therefore shows how T_e recovers after an orogenic event. A problem with considering this measure of lithospheric age, as noted by Pérez-Gussinyé & Watts [2005], is that the core of a cratonic region may not be affected by orogenic processes which occur at its margins.

The maps in Figure 7.1 show the recovered T_e and age variations from the four largest regions investigated. In general, continental lithosphere with an age of less than 50 Ma is associated with low T_e values. Older terranes tend to be associated with a broader range of T_e values ranging from less than 5 km to over 100 km.

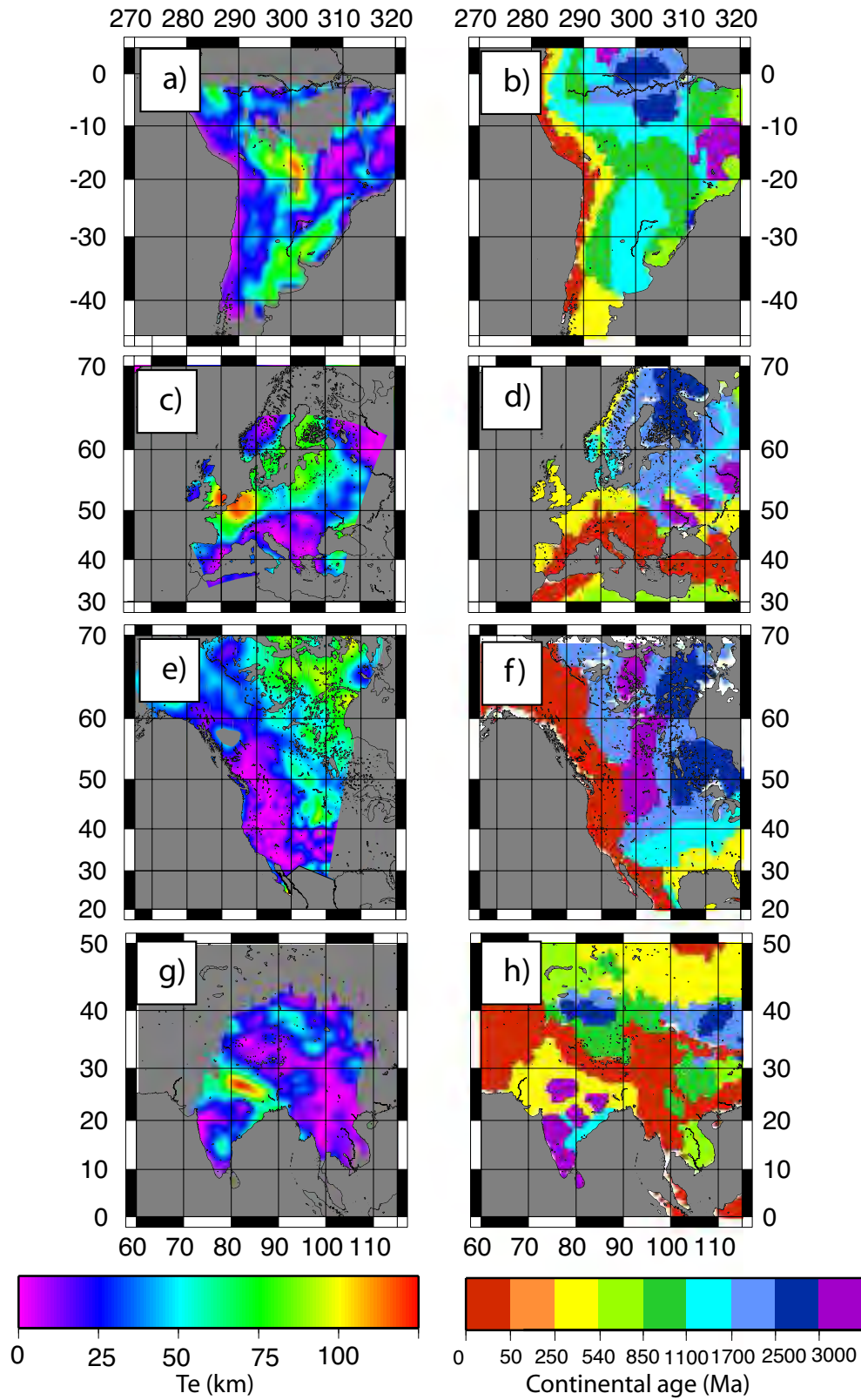


Figure 7.1: T_e and continental age maps from the four largest regions investigated in this study. T_e estimates are the results of the 2D iterative technique and age estimates are from Artemieva [2006]. a) T_e of South America. b) Age map of South America. c) T_e map of Europe. d) Age map of Europe. e) T_e of North America. f) Age map of North America. g) T_e map of India Eurasia collision zone. h) Age map of India Eurasia collision zone.

Plots of frequency distribution of recovered T_e for the different age bins are shown in Figure 7.2. These confirm the concentration of lower T_e values in very young (0-50 Ma) terranes suggested by the maps. For ages between 250-540 Ma a broad distribution of T_e values is seen in both Europe and India (Figure 7.2a and b). This may be a result of the lithospheric age model used. In some areas it assumes a lower age for the regional basement than is known from local studies; for example, much of north west Europe falls within the 250 to 540 Ma range when the Avalonian micro-continent, which forms the core of this region, is known to be older (e.g. Pharaoh [1999]). In Europe and India lithosphere with an age range of 540-850 Ma has a very high percentage of low T_e values; however, these regions represent a relatively small area and may not therefore be representative of the T_e distribution for this age range. As lithospheric age increases towards 1100 Ma the most commonly recovered T_e values increase from generally less than 20 km to the 25 to 50 km range in all continents studied. As lithospheric age increases further in some regions T_e appears to decrease towards an average value of close to 25 km. This pattern is most clearly seen in South America.

The global compilation of T_e and age data shown in Figure 7.2e confirms the pattern observed within the individual continental regions. The general increase in recovered T_e with age from 0 to 1000 Ma suggests that lithospheric age does play a role in controlling the recovered T_e . It also shows that the majority of the recovered T_e values fall within the range predicted by the 450°C isotherm, based on a thermal thickness of the lithosphere of 250 km. This supports the proposal of, for example, Karner & Watts [1983] that cooling of the continental lithosphere with time is an important factor in controlling the recovered T_e .

The apparent decrease in T_e for lithosphere older than 1000 Ma reflects processes which act to reduce the recovered rigidity of the lithosphere. There are four main causes of the recovery of low T_e in old continental regions.

Firstly, mantle plumes may weaken the continental lithosphere due to thermo-

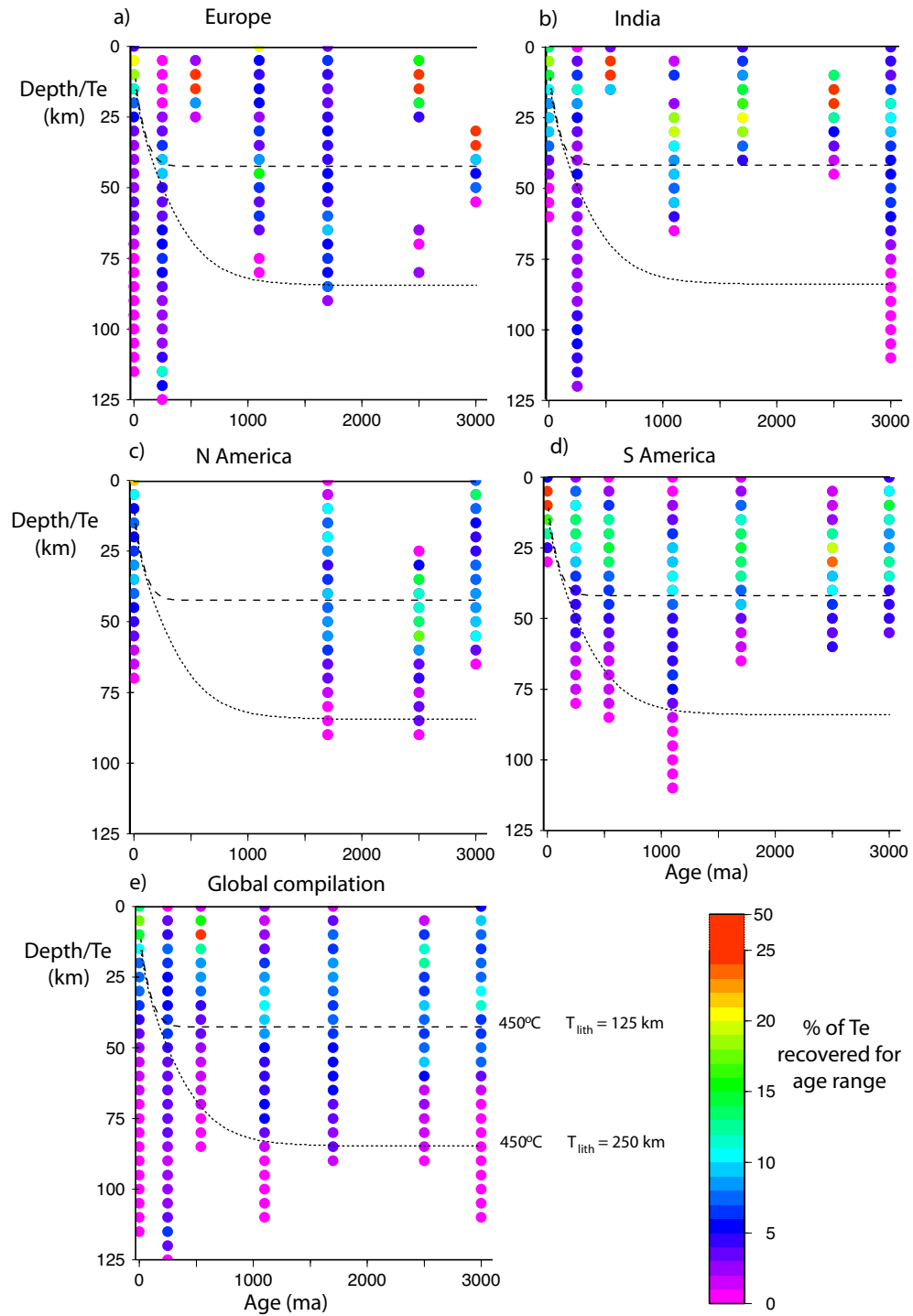


Figure 7.2: Distribution of recovered T_e with age. Each column represents a normalised histogram of T_e for that age range, based on a bin width of 5 km. Dashed lines show the increase in depth to the 450°C isotherm with time given a 125 or 250 km thick thermal lithosphere, from Watts [2001].

mechanical erosion of the mantle, as suggested by Burov & Guillou-Frottier [2005]. An example of this is the low T_e values seen along the west coast of India associated with the Deccan Traps. Here, despite the presence of continental material over 3000 Ma, low T_e values are recovered.

Secondly, orogenic processes may rejuvenate the lithospheric mantle; for example, re-hydration due to de-watering of subducted oceanic material, or convective removal of cold dense mantle material. This is the proposed cause of the low T_e values recovered in north west Scotland and discussed in Section 6.4.4.

Thirdly, yielding and de-coupling of the rigid layers within the continental lithosphere may reduce the recovered T_e value compared to that predicted solely from lithospheric age (e.g. Burov & Diament [1995], Cloetingh & Burov [1996]). Yielding occurs when curvature, associated with flexure, causes stresses that exceed the yield strength of the lithosphere. This failure decreases the thickness of the lithosphere which can behave elastically and hence T_e . De-coupling in the context of the continental lithosphere is associated with yielding of the lower crust. This allows the upper crust and mantle to behave as separate beams. Burov & Diament [1995] showed this may lead to a reduction in T_e of over 50%. They also suggested that this drop in T_e would not be seen in regions with strong lower crust which prevents decoupling. The effect of yielding and de-coupling was discussed in Chapter 5 to explain the different T_e values recovered for the Arabian shield associated with the Zagros foreland (~ 60 km) and the Oman ophiolite (~ 30 km). The 2D iterative technique recovers T_e values of over 70 km, associated with the Himalayan and Andean forelands, showing that for regions with high initial T_e , yielding is limited.

Lastly, synthetic tests show that the 2D iterative technique tends to underestimate the recovered T_e value in the presence of significant uncorrelated buried loading (Section 2.7.1). This is thought to account for the very low recovered T_e values seen in the north-east corner of Europe. Buried loads are often generated by the same orogenic processes that create the elevated topography seen in mod-

ern mountain belts. These mountain belts are subject to erosive processes which remove them over time leaving the buried loads, which, if T_e is high, remain uncorrelated with the surface topography. Uncorrelated buried loads are therefore more likely to occur in older lithospheric settings as erosive processes have had more time to modify the topography. T_e estimates, based on the 2D iterative technique, in older cratonic regions may therefore be biased towards lower recovered values.

7.1.1 T_e and age summary

The recovered T_e distribution is a result of the interplay of cooling over time, which acts to increase lithospheric rigidity, and processes which act to reduce the recovered T_e . For lithospheric ages of less than 1000 Ma, and particularly less than 250 Ma, when the lithosphere is cooling fastest, lithospheric age is likely to be the controlling factor. For older regions processes acting to reduce the recovered T_e become more important, leading to the wider scatter of recovered T_e values.

7.2 T_e and continental rheology

The spatial and temporal variations in T_e recovered by this study reflect variations in the rheological properties of the lithosphere. These variations suggest that unlike the oceans, the continental lithosphere does not have a simple rheological structure.

The rheological properties of the lithosphere describe where and how the lithosphere yields in response to imposed loads. They relate to the composition and thermal state of the lithosphere. An example of the impact of these properties is given by Watts [2001] and is shown by the yield strength envelopes (Y.S.E.) in Figure 7.3. These show how much stress can be supported by each part of the lithosphere before it yields. T_e is related to these Y.S.E. models and rheology because it is a measure of the integrated strength of the lithosphere, and hence is a measure of the amount of material present that is supporting any imposed stresses

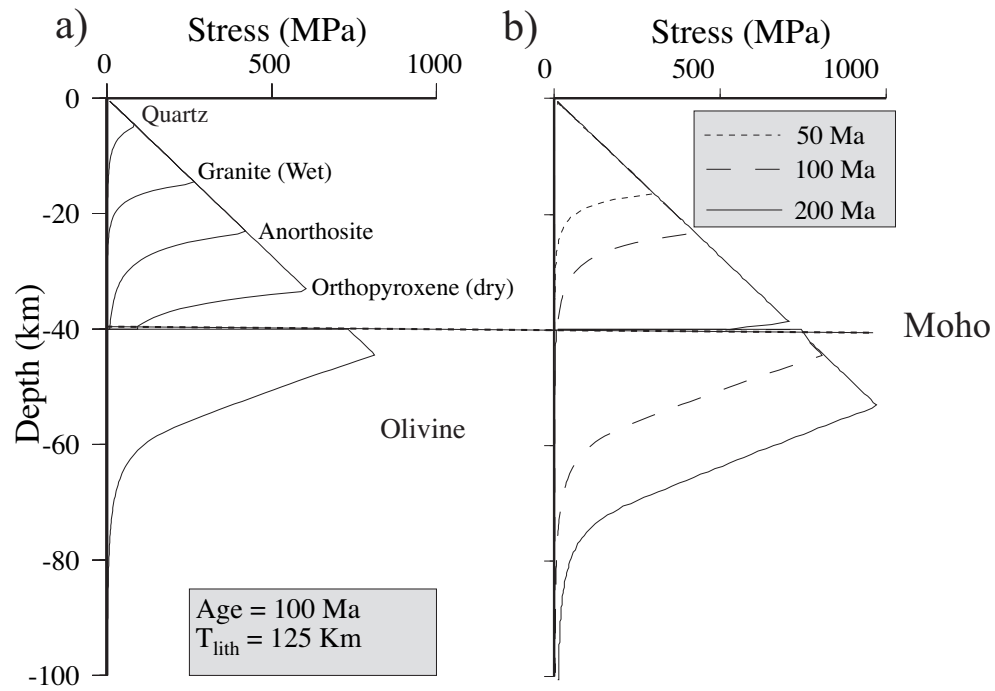


Figure 7.3: Impact of composition and temperature on continental Y.S.E. from Watts [2001], based on a strain rate of 10^{-14} s^{-1} and a cooling plate model with a thermal thickness of 125 km and a crustal thickness of 40 km. Larger areas enclosed by the Y.S.E. relate to higher T_e values. a) The impact of crustal composition on the Y.S.E. for 100 Ma lithosphere. b) The impact of cooling with time on the continental Y.S.E. assuming an anorthositic crust.

on geological time scales (e.g. Watts & Burov [2003]).

Rheological models are generally constrained by laboratory-based rock mechanical data, extrapolated onto geological time scales, combined with observed or predicted geothermal data. In contrast T_e values are recovered with few assumptions about the actual rock mechanical properties of the lithosphere, instead considering the lithosphere's response to real loads which are applied over geological time scales.

Theoretically rheological models could be used to predict the T_e recovered by simple elastic models (Burov *et al.* [1998]). The use of rheological models in this way faces two fundamental problems. Firstly, the extrapolation from laboratory-based experiments, which may last for days or months, onto geological time scales

may not be realistic (Burov [2003]). Secondly, there is no clear agreement on which set of rheological properties best describes the continental lithosphere. As shown by Figure 7.3 these have an important impact on the Y.S.E and hence on the overall recovered rigidity.

Current views of the rheology of the continental lithosphere fall into two broad groups: those where most strength resides in the crust (e.g. McKenzie & Fairhead [1997], Maggi *et al.* [2000a] and Jackson [2002a]); and those where most strength resides in the mantle (e.g. Cloetingh & Burov [1996], Lavier & Steckler [1997], Watts & Burov [2003] and Hetényi *et al.* [2006]). Evidence for a model dominated by a strong crust came initially from the recovered distribution of earthquakes, which occur predominantly within the crust (Maggi *et al.* [2000a]). The lack of mantle earthquakes was seen as a result of weakness and plastic yielding of a more hydrous continental lithospheric mantle, relative to oceanic regions (Jackson [2002a]). The same lack of mantle earthquakes has been viewed by other authors such as Watts & Burov [2003] as evidence that the mantle is so strong that it does not fail brittly given the expected flexural and tectonic stresses. Handy & Brun [2003] question whether any conclusions about the long term rigidity of the continents can be drawn from the distribution of earthquakes. Instead these authors argue that the mantle is strong from structural observations of rifted margins and orogens in addition to laboratory-based studies.

As there is no clear agreement on a model of continental rheology, recovered T_e values can themselves be used to constrain the properties of the lithosphere as suggested by, for example, Burov & Diament [1995] and Burov *et al.* [1998]. As T_e is the integrated strength of the entire lithosphere the recovered values may relate to either strength within the mantle, or crust, or a combination of the two. Studies of T_e have been used as evidence for both strong and weak mantle rheologies. McKenzie & Fairhead [1997] recovered T_e values always less than the local crustal thickness for a broad range of continental regions and supported the weak

mantle view. Other investigations such as those of Flück *et al.* [2003], Braitenberg *et al.* [2003] and Pérez-Gussinyé & Watts [2005] recovered T_e values that, in places, exceeded the local crustal thickness suggesting a strong mantle.

The results presented in previous chapters show that T_e values do often exceed local crustal thickness, indicating that the mantle is playing an important role in providing the strength of the continental lithosphere. This result is best constrained in northern India, where studies along individual profiles show that T_e must be close to 70 km to explain the observed gravity anomalies. Additionally the correlation seen between high T_e and high seismic wave velocity in the upper mantle of Europe, noted by Pérez-Gussinyé & Watts [2005], and also seen in North America in Chapter 6, suggests this part of the mantle is contributing to the high recovered T_e values.

Although important, it is unlikely that the mantle is the sole contributor to the long term strength of the lithosphere. Lavier & Steckler [1997] have shown that a rheology with a strong mantle and a strong plagioclase-dominated crust is needed to explain the high ($\sim 90 \pm 25$ km) T_e presented by previous authors for areas such as the Gangetic foreland. If the lower crust was dominated by quartz or granite it would be very weak, as shown in Figure 7.3. This would lead to de-coupling of the crust and mantle parts of the lithosphere and a dramatic reduction in recovered rigidity. Additionally the correlation between areas of intense seismic activity and lower recovered T_e values supports the view that the crust must contribute to the overall lithospheric rigidity. If it did not, seismicity and the associated fracturing of the crust would not be expected to impact on the recovered T_e .

Since recovered T_e values show that the lithospheric mantle can be strong, the question arises as to why low recovered T_e values occur in old continental regions. Burov & Diamant [1995] suggested there were three key controls on recovered T_e : temperature, decoupling and yielding. Cloetingh & Burov [1996] incorporated these factors into rheologically derived T_e estimates for Europe and predicted T_e

values of between 60 and 90 km, for all regions older than 800 Ma. As discussed in the previous section, the recovered range of T_e for old continental regions far exceeds this. It is therefore proposed that mantle plumes and orogenic activity fundamentally alter the rheology of the lithospheric mantle. This would lead to a more heterogeneous mantle than assumed by Cloetingh & Burov [1996], allowing the prediction of a broader range of T_e values from rheological models. This finding is in line with seismic tomographic studies such as that of Fischer & van der Hilst [1999] which describe significantly more heterogeneity with the ‘seismic keels’ beneath the continents than was previously supposed by older, lower resolution, models.

7.2.1 T_e and continental rheology summary

The results from this study confirm that the continental lithospheric mantle, in regions such as the Himalayan and Andean forelands, is strong over long time scales. The broad scatter of T_e values recovered in old continental regions, from ~ 5 to over 70 km, does not support a model with a single strong mantle rheology. Yielding and decoupling of the upper crust and mantle are likely to be important factors in increasing the range of recovered T_e . However, processes which fundamentally alter the properties of the lithospheric mantle must also play a part.

7.3 T_e and continental tectonics

Rheology, and hence T_e , is not only a significant factor for how the lithosphere responds to vertical loads; it also controls how it responds to tectonic forces. Variations in recovered T_e can therefore play an important role in interpreting and understanding tectonic processes within the continental lithosphere.

The theory of plate tectonics was developed to explain the relative motion of the continental land masses and the narrow zones of seismicity which appeared

to divide many oceanic basins. It also succeeded in explaining the behaviour of oceanic plates, which form along narrow mid oceanic ridges and are destroyed at subduction zones. In detail, however, continental deformation differs from the original model of internally rigid plates in relative motion, with deformation concentrated at their margins. Instead diffuse deformation is seen, with seismicity occurring across broad regions of the continents. Molnar [1998] suggested that this was because the continents are fundamentally weaker than the oceans due to the presence of the thick, silica rich, continental crust. Authors such as Steckler & ten Brink [1986] showed that the weakness of the continental lithosphere relative to the oceans was a key factor in controlling the distribution of rifting and faulting on a tectonic scale.

In contrast to the view that the continents are weak, this study recovers T_e values for the continents higher than those of the most rigid parts of the ocean, as do many other investigations (e.g. Stewart & Watts [1997], Tassara [2005], Braitenberg *et al.* [2003] and Pérez-Gussinyé & Watts [2005]). The impact of these rigid continental regions on tectonic processes has been discussed previously by, for example, Watts *et al.* [1995] and Stewart & Watts [1997]; however, the new maps of T_e presented in this study allow for further interpretation.

The expected effect of T_e variation on tectonic processes would be the partitioning of deformation. Regions with high T_e , and hence strong rheology, would deform less while low T_e regions would deform more. A simple examination of the topographic features associated with the India-Eurasia collisional system shows strong partitioning of deformation. Rigid regions, such as the Gangetic foreland, Qaidam Basin and Sichuan Block, have subdued topography compared to the lower T_e regions within the orogen, which generally have very extreme topography. Additional evidence for partitioning of deformation in the continents comes from the distribution of seismicity. Although seismicity is distributed over broad regions these regions have, in places, distinct margins. An example of this is the south edge

of the Tibetan plateau and Himalayan mountain chain: here there is a transition from highly seismic within the orogen to relatively aseismic beyond. This change is associated with a dramatic rise in T_e from <10 to over 70 km, showing that the high T_e region is not deforming.

T_e variations may also define how deformation occurs, as well as where. In South America Watts *et al.* [1995] and Stewart & Watts [1997] proposed that differences in T_e correlated with differences in the width of the orogenic belt. In Chapter 6 this correlation was confirmed. Watts *et al.* [1995] and Stewart & Watts [1997] suggested that this correlation occurs because high T_e in foreland regions preferentially leads to thin-skinned thrusting. This thrusting allows the orogen to more easily override the craton and hence leads to the development of a broader orogenic belt. Low T_e foreland regions tend to favour thick-skinned faulting which retards the progress of the orogen over the surrounding area.

The most extreme effect of very weak continental crust on tectonic processes can be seen in the India-Eurasia collisional system. Towards the south east margin of the plateau, where the topographic gradient is lowest, Clark & Royden [2000] propose that the lower crust is so weak that it is flowing out from the centre of the plateau. The low T_e values recovered in this region support the presence of a very weak rheology, and hence the idea that the lower crust may be flowing. The other margins of the plateau have steeper topographic gradients and appear to be bounded by high T_e regions. It is therefore proposed that the high T_e regions act to constrain the weak lower crust beneath the plateau, only allowing lower crustal flow where the lower crust was initially weak.

7.3.1 T_e and tectonics summary

It is concluded that the patterns seen in continental collisional systems relate to the interplay of weak and strong regions within the continental lithosphere. Three end-member collisional systems are envisaged; strong-weak convergence, weak-weak

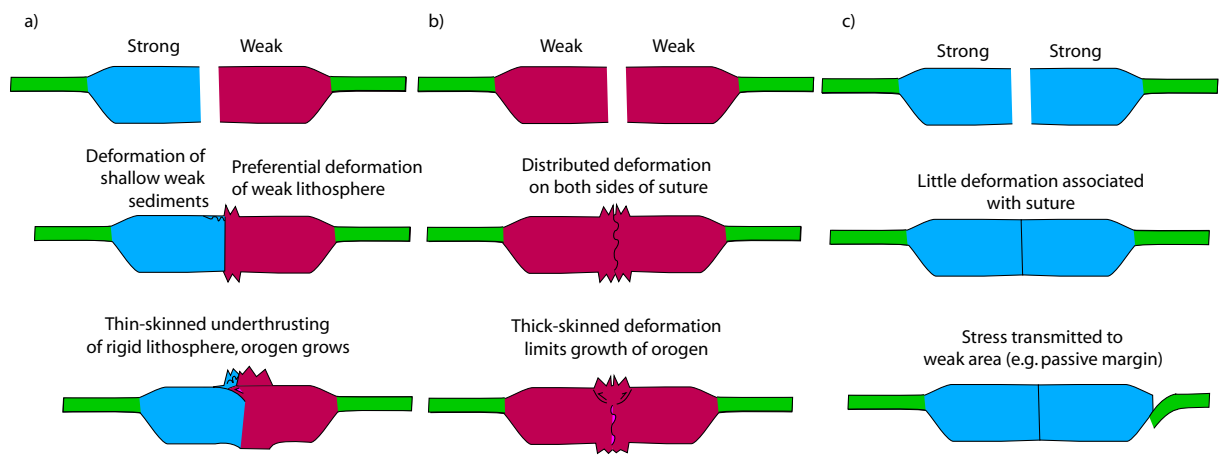


Figure 7.4: End-member collisional systems. a) Strong-weak, where the weaker lithosphere and shallow sediments preferentially deform leading to over-thrusting of the more rigid continent. b) Weak-weak, where deformation is equal each side of the suture, however, thick-skinned thrusting limits the orogen's development. c) Strong-Strong, where little deformation occurs along the suture and stress is transferred to a weaker region.

convergence and strong-strong convergence, as shown in Figure 7.4.

Where weak and strong continents collide, the weaker lithosphere will preferentially deform, generating elevated topography. The weak sediments overlying the rigid continental region may also be involved in this deformation. The elevated topography will tend to over-thrust the lower, undeformed, rigid part of the continent allowing further convergence. Over-thrusting generates a load on the crust and the development of a foreland basin, the shape of which is controlled by the local T_e . The shape of the foreland basin determines the ease with which convergence continues, with more rigid forelands allowing greater convergence. The best examples of this are the India-Eurasia collisional system or the centre of the Andean orogenic belt.

Where weak regions of the continents are forced together overall less deformation will occur, as thick-skinned faulting will tend to limit any over-thrusting. This may in part explain the smaller size of the Alpine orogenic belt, relative to

the Himalaya.

Where two rigid continental blocks collide internal deformation is likely to be limited, and stresses may be transferred to other parts of the continent where the rheology is weaker. There is no clear example today of this type of collision. However, the Trans European suture zone, which marks the join of the Baltic and Avalonian cratons, may mark such an event. It is not thought that a significant orogenic belt was generated here, although the strike slip nature of this collision may have also contributed to this.

Conclusion and further work

8.1 Introduction

The aim of this investigation was to study the variations in rigidity of the continental lithosphere, and to assess the causes and impact of these variations in a geological context. Model-based techniques were used, which explicitly calculated the deflection of an elastic beam or plate due to imposed vertical loads. The best fitting T_e structure was constrained by comparison of observed Bouguer gravity anomalies with calculated anomalies based on the modelled deflection. This study focused on the response of the lithosphere to loading by surface topography, the isostatic compensation of which is known to give rise to observable gravity anomalies. In addition, where it was possible to constrain their size, sub-surface loads were also considered.

8.2 Conclusions

The key conclusions of this study are laid out in the following section.

- Continental T_e is highly variable, with recovered values ranging from <5 km to over 70 km.

a) In Oman and the U.A.E., profile-based flexural models of a plate with a T_e of ~ 30 km, broken beneath the load, best matched the observed data. However, to the south east, along strike of the obducted ophiolite, flexural models incorporating

a decrease in T_e to 0 km in the bulge region were needed to match the observed gravity anomaly.

b) In the Himalayan foreland, T_e values of between 30 and 70 km were recovered along strike, based on simple flexural models along profiles, with the highest values being recovered in the centre of the foreland. This pattern of variation was confirmed by the grid-based 2D iterative technique. This recovered T_e values from 50 km in the east and west of the foreland to over 100 km in the central region. Beneath the Tibetan plateau T_e values of <5 km are recovered.

c) In the Zagros foreland system T_e values of between 0 and 90 km were recovered using both profile- and grid-based techniques. Beneath the centre of the orogenic belt T_e is low (<~15 km), while for the Arabian shield to the south T_e is high (>~60 km).

d) Investigations of North and South America and Europe using the 2D iterative technique also recover a broad range of T_e values from 0 to ~125 km.

- Present day orogenic belts are regions of low T_e (<20 km).

a) Profile based techniques including a plate break, such as applied to recover T_e in Oman and the Himalayan foreland, all recovered a best-fitting position for the plate break beneath the orogenic belt. This suggests that T_e reduces to approximately zero in these areas.

b) The 2D iterative technique recovered generally low T_e values in all the present day orogenic regions investigated. These included: compressional regions such as the Alps, Rocky mountains, and Zagros mountains; high plateaus, such as the Altiplano and Tibet; and extensional regions such as the basin and range province in the U.S. and Greece.

- Cratonic regions in the centre of continents generally have high recovered T_e values compared to orogenic belts.

a) The 2D iterative technique recovered T_e values of over 70 km in parts of all the continental areas investigated.

- T_e variations are correlated with other geophysical parameters which describe the lithosphere.
 - a) High T_e regions in North America and Europe are associated with high upper mantle (100 and 200 km depth) seismic velocity.
 - b) T_e appears to be inversely correlated with measures of mantle temperature such as heat flow or depth to a specific isotherm.
 - c) Low T_e is often associated with an increased occurrence of earthquakes. Lack of earthquakes in high T_e regions make comparison of T_e and T_s (the range of depths where earthquakes occur) difficult.
- For ages less than 1 Ga the T_e recovered in the continents generally increases with age. For lithosphere with tectonic ages older than 1 Ga a broad scatter of T_e values are recovered.
 - a) The initial increase in T_e with age suggests that cooling of the lithosphere with time is the dominant factor contributing to lithospheric strength.
 - b) The broad scatter of T_e values recovered in older regions is related to processes which modify the lithospheric mantle, such as proximity to an orogenic belt or a mantle plume event.
- T_e variations recovered by this study place a constraint on the rheological properties of the lithosphere.
 - a) High T_e values, which exceed the local crustal thickness, indicate that the lithospheric mantle in these regions must be rigid on geological time scales.
 - b) High T_e values in foreland regions suggest that the lower crust must also be strong, as yielding of a weak lower crust would reduce the recovered T_e in these regions.
 - c) The broad range of T_e values recovered in regions with old lithosphere show that the lithospheric mantle is not homogeneous and that tectonic processes act to modify it over time.
- T_e has a direct impact on the evolution of tectonic processes. T_e may therefore

be a controlling factor in the development of continental orogenic belts.

a) The Tibetan plateau is surrounded by a number of regions of high T_e . In these regions the topographic gradient is high, and deformation progresses by over-thrusting of the orogen. In the south east of the plateau T_e is low, as is the observed topographic gradient. In this region there is no over-thrusting and it has been previously proposed that the topography is generated by lower crustal flow from beneath the Tibetan plateau.

b) In South America, there is a correlation between the width of the Andean mountain belt and T_e in the foreland region. It is proposed that this correlation is associated with variation from thick-skinned to thin-skinned faulting. Thick-skinned faulting is due to a weak, high curvature, foreland which retards over-thrusting of the orogenic belt. Thin-skinned thrusting is due to a rigid, low angle foreland, which forms a ramp, promoting over-thrusting of the orogenic belt onto the stable continental interior.

c) Three end-member continental collisional situations are envisaged. Firstly, weak-strong convergent situations that are dominated by over-thrusting of the weaker region onto the more rigid region. This over-thrusting leads to the development of a broad, high orogenic belt such as the Andes or Himalayas. Secondly weak-weak convergent situations, which will be dominated by thick-skinned deformation and a lesser degree of overall convergence. An example of this setting may be the Alpine collisional belt in Europe. Lastly, where two rigid continental fragments collide, it is likely that deformation at the suture will be limited and stresses will be transmitted to weaker regions.

8.3 Future work

A key challenge for further work on continental T_e is the presence of buried loading. This is a particular problem when recovering T_e across broad regions, both when using the 2D iterative modelling technique applied in this study, and the spectral

techniques applied by other authors. The technique applied in Oman to recover buried load size from the observed gravity anomaly would not work across broad regions of the continents, as in regional studies the observed anomaly is assumed to represent the isostatic signal due to compensation of the load.

An alternative to deriving the buried load sizes from the gravity anomaly is to use seismic data to define the subsurface structure. This technique has been applied, for example, by Braitenberg *et al.* [2003], and successfully used to constrain T_e in the presence of sedimentary loads within the crust. Application of this technique to broad continental regions will require compilation of large continent-wide seismic data sets such as those derived from the European Geotraverse (e.g. Wever *et al.* [1990]). If the density structure of the continental crust was known, flexural models could be constructed accounting for both the isostatic compensation of the surface topography and any buried loads.

A second area of future work is to further assess the implications of the T_e structures recovered by this study. The results of this study show that variations in T_e can control the nature and extent of deformation associated with orogenic belts. However, work is needed to incorporate the variations in rheology implied by the recovered T_e structure into dynamic models of the lithosphere's response to both vertical and horizontal stresses.

Regions within continental areas which have low recovered T_e and do not appear to be currently deforming are of particular interest. These weak but non-deforming regions do not appear to be compatible with models of tectonic processes driven by forces at the plate margins, such as ridge push, slab drag and slab suction (e.g. Conrad & Lithgow-Bertelloni [2004]). Instead convective traction on the base of the continental lithosphere may play a role in moving the entire plate, carrying the low T_e regions with little or no internal deformation (e.g. Ziegler [1993]).

Thirdly, further work is needed where geological processes have altered or obscured the simple isostatic signal associated with regional compensation of loads.

Towards the south-eastern end of the North Oman ophiolite, the nature of the foreland and the associated T_e have not been fully explained due to the complex nature of the obduction process. Further work is needed in this region to quantitatively assess the impact of return flow of material from the subduction zone on the isostatic compensation of the ophiolite and the development of the associated foreland basin.

In the Arabian gulf the presence of significant salt deposits within parts of the foreland region seems to have altered the response to thrust loading. In the south eastern part of the foreland high T_e values (>80 km) were proposed from gravity data, but not supported by estimates of foreland size from isopach data. Seismic refraction studies in this region would help constrain the T_e of this part of the foreland by revealing the sub-salt crustal structure, which is invisible to traditional seismic reflection techniques.

The current study has considered the regional isostatic support of surface topography, and in some cases buried loads, to reveal the long term rigidity structure of the lithosphere. Other types of support of the long term continental topography have been proposed, such as dynamic topography related to subducting slabs (Burgess & Moresi [1999]). This effect is thought to be of generally long wavelength (>1000 km), however, it may cause surface deflections of up to 2000m. The long wavelength of the deflections associated with dynamic topography mean that the gravity signal associated with it may have been removed, as part of the satellite derived gravity anomaly, before recovery of T_e . The dynamic forces which drive surface deflection will be modulated by the rigidity of the lithosphere. Modelling of the dynamic forces and applying them to the base of the variable rigidity T_e structures recovered in this study may therefore allow more accurate prediction of the expected deflection associated with dynamic mantle forces.

Appendix A

Gravity Data

A.1 Base station information

The gravity data collected in Oman was tied into the IGN71 global reference frame by 4 measurements made at the absolute gravity base near Seeb International Airport (SIA) shown in Figure A.1. The original base established at Seeb International Airport by Shelton [1984] has now been destroyed, however, before its destruction it was linked to a base at Sultan Qaboos University (SQU). The SQU base has since been tied to a publicly accessible base close to Seeb International Airport by Ravaut & Warsi [1997b]. The current appearance of the base and its location are shown in Figure A.2. The gravity value at the base is given as 978921.955 mgal.

Station ID	reading 1	reading 2	reading3	reading average	Meter Factor	Conversion Level	Conversion Mgal	mGal meter	Absolute Gravity	Day	Month	Year	hour	min
Seib B1	2649.7	2649.69	2649.71	2649.7	1.01811	2600	2646.41	2697.01	978921.9550	21	1	4	16	30
Seib B2	2649.71	2649.7	2649.7	2649.7	1.01811	2600	2646.41	2697.01	978921.9550	23	1	4	10	0
Seib B3	2650.04	2649.99	2650	2650.01	1.01811	2600	2646.41	2697.33	978921.9550	10	2	4	2	17
Seib B4	2650.11	2650.07	2650.09	2650.09	1.01811	2600	2646.41	2697.41	978921.9550	17	2	4	7	40

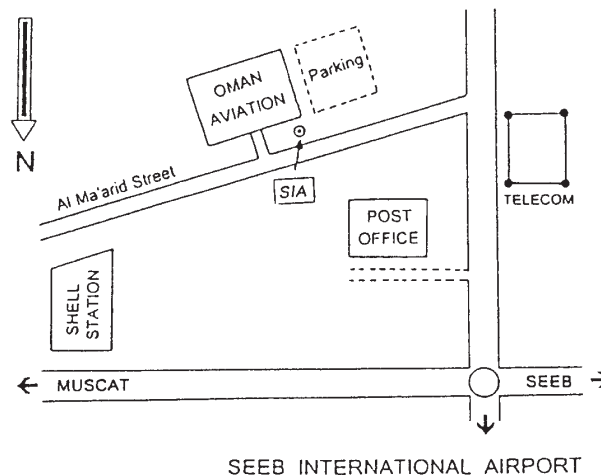
Figure A.1: Readings taken at the IGSN base close to Seeb International Airport



STATION : SIA

COUNTRY : OMAN

b)



The station is located at NSA GPS point 7911001 near Seeb International Airport, outside the Oman Aviation Services compound. There is a protected subsurface marker and a concrete pillar. The measurement was made at ground level. g (IGSN71) = 978921.955 mgals

Latitude : 23° 35.00' N

Longitude = 58° 17.68' E

Elevation = 22 m (approx.)

Figure A.2: IGSN gravity base station close to Seeb International Airport a) View of gravity base station, looking north east. Note the number on the pillar is 7911001. b) Location map of IGSN 71 base.

A.2 Terrain corrections

The terrain correction accounts for the gravity effect of the material either side of an observation point which is not included in the simple Bouguer slab correction. If the observation point was at the base of a hill as shown in Figure A.3 the centre of mass of the hill exerts a force (f_1) on the gravity meter at the observation point. This can be resolved into two components, a horizontal force (fh) and an upward vertical one (fv). If the hill was removed, to recreate the ideal infinite flat Bouguer slab, the upward (negative) force vf would be removed. Therefore to correct for this, the positive value vf must be added to the final anomaly. In the case of a valley beside the point, the extra downward force expected if the valley was filled in to recreate the infinite slab must be calculated and added to the final anomaly in the same way.

The terrain correction can be calculated by breaking down the topography into a number of discrete blocks. The gravity anomaly due to each block can then be calculated using equation A.1 from, for example, Lowrie [1997].

$$\Delta g_T = G\rho\phi\left\{\left(\sqrt{r_1^2 + h^2} - r_1\right) - \left(\sqrt{r_2^2 + h^2} - r_2\right)\right\} \quad (\text{A.1})$$

Where ρ is the density of the topography considered, G is the Gravitational constant, h is the height of the block relative to the observation point, and the other parameters are as shown in Figure A.3b.

The sum of the gravity effect of all blocks on the measurement point is the terrain correction. The terrain correction was carried out using the SRTM-90 data set for the reference topography. Equation A.1 is for a segment of a ring with an inner radius of r_1 and an outer radius of r_2 . To calculate the terrain effect around each observation point from the SRTM90 data set each grid point more than 90 m from the observation point was considered. First the distance from the observation point to the grid point (r) was calculated. Then r_1 and r_2 were set to $r-45$ m and $r+45$ m respectively and ϕ was calculated as $2(\tan^{-1}(45/r))$. This constructed a

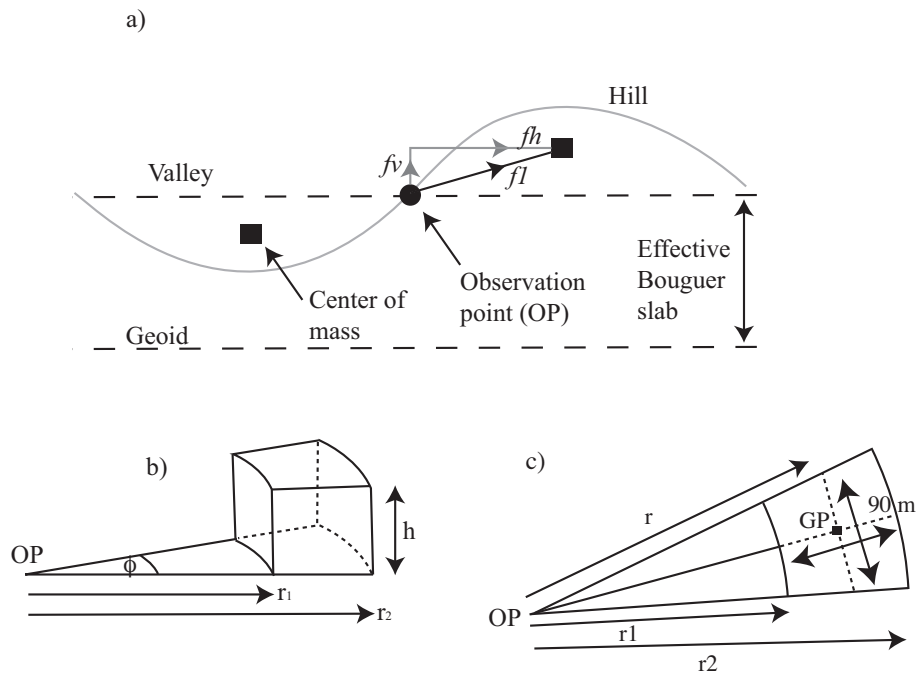


Figure A.3: Terrain correction model. a) The effect lateral topography on the observation point. b) The parameters defining each terrain component as shown in equation A.1. c) Plan view of terrain component as defined about a grid point (GP) from the SRTM 90 m data base.

segment centred on the SRTM grid point that was 90 by 90 m across as shown in Figure A.3c. The gravity effect of the segment of topography thus defined was calculated using equation A.1. Close to the observation point the approximation that the segment of topography has the same effect as a square pillar breaks down as the surface area, and hence volume, of the segment is less than the square pillar defined by the SRTM90 grid point. The correction for each segment was therefore divided by the area of the segment and multiplied by the area of the 90 by 90 m square pillar it represented.

The sum of the effect of all the terrain segments was the terrain correction applied. Typically the terrain correction is applied radially about the observation point, however, as can be seen in Figure 3.6c the magnitude of the far field component is negligibly small. Therefore the sum of all components within the 1 degree box was considered. The correction was calculated using the program shown below

in Figure A.4.

The effect of offshore bathymetry on the terrain corrections was not considered for two reasons. Firstly the lower implied density contrast offshore ($\sim 1660 \text{ kg m}^{-3}$ in-place of 2670 kg m^{-3}) means the correction will be just over half that for on-shore regions for the same topography. Secondly in the Arabian gulf close to the study area, the water depth is less than 50 m with little relief, which will also make the correction very small, particularly along the main foreland profile.

```

/* Standard terrain correction */
// compile using gcc -lm terrain.c -o terrain.exe
#include<stdio.h>
#include<math.h>
#include <stdlib.h>
main()
{
FILE *fp;
FILE *f;
char ref_name[20], dat_name[20];
int nx, ny, lmax, j;
static float NUMX[1500000], NUMY[1500000], DAT[1500000], ANS[1500000];
float xp, yp, hp, r1, r2, rr, theta, h, delg, corection, a, b, degree, pi;
float area, dif, outer, inner;
corection = 0;
pi = 3.142;
// Get data
// printf("hello start\n");
// printf("enter reference file name\n"); // Contains Lon, Lat, z data within region considered for terrain correction
scanf ( "%s", &ref_name);
scanf ( "%i", &nx); /*printf("enter number of x points in grid \n");*/
scanf ( "%i", &ny); /*printf("enter number of y points in grid \n");*/
scanf ( "%i", &lmax); /*printf("enter total number points in grid \n");*/
scanf ( "%f", &xp); /*printf("enter x point\n"); // Lon in deg of observayion point */
scanf ( "%f", &yp); /*printf("enter y point\n"); // Lat in deg of observayion point */
scanf ( "%f", &hp); /*printf("enter point height\n"); // Enter elevation of observation point*/
f=fopen( ref_name, "r"); /*Get topo data from xyz file*/
for(j=0; j<lmax; j++)
{
    fscanf(f, "%f %f %f", &NUMX[j], &NUMY[j], &DAT[j]); /* read gridded data*/
}
fclose(f); /* close file*/
// Loop through all points in grid considered
for(j=0; j<lmax; j++)
{
    // Get distance from observation point to grid point
    rr=111000*(sqrt(((NUMX[j]-xp)*(NUMX[j]-xp))+((NUMY[j]-yp)*(NUMY[j]-yp))));
    if ( rr>90 ) //Excludes grid points < 90 m from observed point//
    {
        r1=rr-45; /*Inner radius*/
        r2=rr+45; /*Outer radius*/
        theta=2*(atan(45/rr)); /*theta*/
        h=DAT[j]-hp; /*height diference */
        a = (sqrt((r1*r1)+(h*h)))-r1;
        b = (sqrt((r2*r2)+(h*h)))-r2;
        delg = 2650*theta*(a-b); /*grav correction due to single coloum */
    }
    if ( rr<=90 )
    { delg = 0; }
    if ( delg<=0 )
    { delg = 0; }
    // area effect, corrects for diferance between prismatic and square piller.
    degree = theta*(180/pi);
    inner = ((pi*((rr-45)*(rr-45)))/360)*degree;
    outer = ((pi*((rr+45)*(rr+45)))/360)*degree;
    area = outer-inner;
    dif = (90*90)-area;
    delg = (delg/area)*(90*90);
    if ( area<=0 )
    { delg = 0.000001; }
    corection = corection + delg; /*Sums corrections for all prisms*/
    printf ( "%f %f %f\n", NUMX[j], NUMY[j], delg ); /*Output correction due to every prism*/
}
printf ( "%f %f %f\n", xp, yp, ((6.67259e-11)*corection)/1e-5 ); /* prints final corection in mgal*/
}

```

Figure A.4: Terrain Correction program

A.3 Data tables

Point Name	Lat	Lon	R1	R2	R3	Conetion Factor	Conetion Level	Mgal	Meter gravity	Drift Correction	ABS Gravity Mgal	Theoretical Gravity	Elevation GRS80	Elevation SRTM90	FAA SRTM	BAA SRTM	Terrain Correction	BAA Full	Day	Month	Year	Hour	MinTime since 01/01/04	
degrees	degrees	meter unit	meter unit	meter unit	meter unit	= 1 mgal	Level	Level	mgal	mgal	mgal	mgal	mgal	m	m	mgal	mgal	mgal					hours	
MU 1.10 56.24678	26.175889	2728.17	2718.18	2728.17	2724.84	1.01818	2700	2748.22	2776.90	0.12993	979001.72	979040.14	8	14.23	-34.03	-35.61	1.92	-33.70	29	1	4	9	4	681.12
MU 1.10 56.32578	26.001944	2673.51	2673.50	2673.51	2673.51	1.01811	2600	2646.41	2721.25	0.12968	979046.07	979034.14	318	934.87	-12.18	-23.89	2.46	-21.43	29	1	4	5	4	681.75
MU 1.30 56.33988	26.006000	2796.27	2796.25	2796.26	2796.26	1.01818	2700	2748.22	2785.14	0.13003	979009.95	979034.43	14	5.71	-22.72	-23.35	5.26	-18.09	29	1	4	10	15	682.25
MU 1.50 56.34678	26.043883	2713.33	2713.33	2713.33	2713.34	1.01818	2700	2748.22	2781.79	0.13003	979086.61	979030.69	169	166.58	7.32	-11.18	3.15	-8.03	29	1	4	11	7	683.12
MU 1.60 56.34672	26.050778	2708.26	2708.26	2708.27	2708.26	1.01818	2700	2748.22	2756.64	0.13002	979881.45	979031.21	170	160.92	2.55	-16.27	2.67	-15.60	29	1	4	11	32	683.53
MU 1.70 56.31450	26.070028	2710.58	2710.55	2710.56	2710.56	1.01818	2700	2748.22	2758.97	0.13108	979031.84	979031.84	150	147.32	-2.59	-18.84	3.06	-13.89	29	1	4	11	41	683.78
MU 1.80 56.29616	26.077422	2707.65	2707.66	2707.65	2707.65	1.01818	2700	2748.22	2759.24	0.13117	979894.05	979033.11	133	129.36	-8.03	-22.60	2.93	-19.68	29	1	4	11	54	683.90
MU 1.90 56.27222	26.084111	2706.42	2706.40	2706.41	2706.41	1.01818	2700	2748.22	2756.01	0.13137	979890.56	979033.11	115	112.65	-12.37	-26.73	2.91	-23.82	29	1	4	12	2	684.03
MU 1.10 56.28278	26.088111	2707.36	2707.35	2707.37	2707.36	1.01818	2700	2748.22	2754.75	0.13146	979890.54	979033.58	100	96.33	-23.61	-34.30	3.49	-30.61	29	1	4	12	10	684.17
MU 1.11 56.28262	26.091659	2703.12	2703.10	2703.10	2703.11	1.01818	2700	2748.22	2751.98	0.13160	979876.19	979034.12	114	118.29	-22.04	-34.96	3.50	-31.45	29	1	4	12	30	684.30
MU 1.12 56.28556	26.093681	2694.88	2694.85	2694.84	2694.86	1.01811	2600	2646.41	2742.97	0.13175	979867.78	979034.28	132	141.57	-22.04	-38.53	3.99	-34.54	29	1	4	12	43	684.72
MU 1.13 56.21022	25.984556	2497.75	2497.75	2497.75	2497.75	1.01797	2400	2442.81	2542.32	0.13277	979876.13	979028.61	114	1135.40	90.90	-35.16	4.46	-30.74	29	1	4	14	10	686.17
MU 1.14 56.22861	25.984556	2403.61	2403.59	2403.62	2403.61	1.01797	2400	2442.81	2542.32	0.13306	979876.13	979028.61	1579	1589.55	134.62	-41.10	9.59	-31.51	29	1	4	14	35	686.58
MU 1.15 56.21697	25.875750	2593.00	2593.01	2593.01	2593.01	1.01804	2500	2544.61	2639.30	0.13362	979864.11	979018.76	707	672.96	53.03	-21.89	2.88	-18.81	29	1	4	15	23	687.38
MU 2.10 56.34630	25.371944	2881.13	2881.16	2881.14	2881.14	1.01825	2800	2850.04	2932.66	0.21298	979157.39	979863.25	4	3.06	175.08	174.74	0.56	175.31	3	2	4	8	45	800.75
MU 2.20 56.39067	25.162444	2856.36	2856.35	2856.37	2856.36	1.01825	2800	2850.04	2907.44	0.21348	979132.17	979868.64	6	5.90	165.35	164.70	0.16	164.86	3	2	4	9	28	801.47
MU 2.30 56.33864	25.145139	2852.20	2852.17	2852.19	2852.19	1.01825	2800	2850.04	2903.18	0.21362	979127.81	979867.44	7	8.61	163.14	162.18	0.32	162.50	3	2	4	9	40	801.67
MU 2.40 56.31678	25.127667	2847.19	2847.21	2847.20	2847.20	1.01825	2800	2850.04	2898.10	0.21377	979122.83	979866.22	32	32.48	166.64	163.03	0.24	163.27	3	2	4	9	53	801.88
MU 2.50 56.29000	25.133278	2833.01	2833.01	2833.00	2833.00	1.01825	2800	2850.04	2883.64	0.21389	979108.37	979866.61	70	66.24	162.20	154.85	0.32	155.16	3	2	4	10	4	802.07
MU 2.60 56.28606	25.144722	2822.79	2822.80	2822.82	2822.80	1.01825	2800	2850.04	2873.28	0.21399	979098.01	979867.41	101	95.64	160.12	149.50	0.43	149.93	3	2	4	10	9	802.20
MU 2.70 56.25219	25.163778	2810.81	2810.79	2810.80	2810.80	1.01825	2800	2850.04	2861.55	0.21413	979085.77	979868.73	132	131.10	157.49	142.94	0.46	143.40	3	2	4	10	24	802.40
MU 2.80 56.22458	25.200028	2767.43	2767.44	2767.43	2767.43	1.01818	2700	2748.22	2841.55	0.21422	979066.28	979869.73	163	162.69	160.12	128.88	0.58	130.26	3	2	4	10	32	802.53
MU 2.90 56.20363	25.200028	2745.06	2745.02	2745.03	2745.04	1.01818	2700	2748.22	2794.07	0.21454	979018.80	979871.25	239	236.78	119.40	108.09	1.25	109.34	3	2	4	10	42	802.70
MU 2.10 56.20736	25.228028	2725.60	2725.60	2725.60	2725.60	1.01818	2700	2748.22	2774.28	0.21468	979018.80	979871.25	283	283.14	119.40	108.09	1.25	109.34	3	2	4	10	52	802.87
MU 2.11 56.19519	25.234056	2701.28	2701.28	2701.29	2701.28	1.01818	2700	2748.22	2749.53	0.21474	979018.80	979871.25	314	314.13	96.80	61.70	0.98	61.17	3	2	4	11	1	803.02
MU 2.12 56.19333	25.234056	2675.76	2675.76	2675.76	2675.76	1.01818	2600	2646.41	2702.44	0.21483	979018.80	979871.25	364	378.84	111.46	80.19	1.42	94.57	3	2	4	11	16	803.27
MU 2.13 56.16870	25.275417	2655.00	2655.03	2655.02	2655.02	1.01811	2600	2646.41	2702.44	0.21483	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.14 56.16767	25.298167	2655.00	2655.03	2655.02	2655.02	1.01811	2600	2646.41	2702.44	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.15 56.17181	25.340750	2653.54	2653.56	2653.55	2653.55	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.16 56.17100	25.340750	2653.54	2653.56	2653.55	2653.55	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.17 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.18 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.19 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.20 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.21 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.22 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.23 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.24 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.25 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.26 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.27 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.28 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.29 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2	4	11	24	803.40
MU 2.30 56.16319	25.361681	2624.40	2624.45	2624.43	2624.43	1.01811	2600	2646.41	2692.95	0.21504	979018.80	979871.25	442	450.07	90.74	39.76	0.26	46.36	3	2				

Point Name	Lat	Lon	R1	R2	R3	R average	Conversion Factor	Level	Mgal	Meter	Drift	ASS Gravity	Theoretical Gravity	Elevation	SSRTM	BAA	Train	BAA	Day	Month	Year	Hour	Min Time since 01/01/04		
	degrees	degrees	meter unit	meter unit	meter unit	meter unit	1 mgal	Level	Level	mgal	mgal	mgal	mgal	mgal	m	SRTM	SRTM	mgal	mgal					hours	
MU 3.80	56.176417	25.553639	2678.64	2678.66	2678.63	2678.64	1.0181	2600	2646.41	2726.48	0.23003	978951.18	978966.00	136	133.15	-3.73	-18.52	1.60	-16.92	4	2	4	10	24	826.40
MU 3.90	56.153600	25.553000	2683.56	2683.55	2683.55	2683.55	1.0181	2600	2646.41	2700.93	0.23112	978952.39	978964.55	221	225.45	0.66	-24.37	0.91	-23.46	4	2	4	10	50	826.63
MU 3.11	56.102869	25.553444	2685.60	2685.60	2685.60	2685.60	1.0181	2600	2646.41	2693.62	0.23123	978952.00	978966.06	183	183.69	-0.12	-31.45	0.88	-30.77	4	2	4	10	50	826.67
MU 3.12	56.074056	25.552472	2682.15	2682.14	2682.14	2682.15	1.0181	2600	2646.41	2699.28	0.23153	978952.23	978966.62	163	163.44	-0.16	-46.34	0.41	-36.14	4	2	4	11	15	827.25
MU 3.13	56.074056	25.552472	2682.15	2682.14	2682.14	2682.15	1.0181	2600	2646.41	2699.28	0.23153	978952.23	978966.62	163	163.44	-0.16	-46.34	0.41	-36.14	4	2	4	11	32	827.53
MU 3.14	56.074056	25.552472	2682.15	2682.14	2682.14	2682.15	1.0181	2600	2646.41	2699.28	0.23153	978952.23	978966.62	163	163.44	-0.16	-46.34	0.41	-36.14	4	2	4	11	47	828.07
MU 3.15	56.030939	25.554906	2681.19	2681.19	2681.19	2681.19	1.0181	2600	2646.41	2699.28	0.23201	978952.25	978966.38	124	117.07	-35.96	-48.96	0.46	-48.48	4	2	4	12	17	828.07
MU 3.21	55.960806	25.509861	2690.33	2690.32	2690.32	2690.33	1.0181	2600	2646.41	2695.98	0.23223	978952.31	978966.44	107	104.65	-39.45	-51.05	0.17	-50.88	4	2	4	12	17	828.25
MU 3.21	55.960806	25.509861	2690.33	2690.32	2690.32	2690.33	1.0181	2600	2646.41	2695.98	0.23223	978952.31	978966.44	107	104.65	-39.45	-51.05	0.17	-50.88	4	2	4	12	17	828.25
MU 3.21	55.960806	25.509861	2690.33	2690.32	2690.32	2690.33	1.0181	2600	2646.41	2695.98	0.23223	978952.31	978966.44	107	104.65	-39.45	-51.05	0.17	-50.88	4	2	4	12	17	828.25
MU 3.16	55.960806	25.509861	2694.81	2694.84	2694.81	2694.82	1.0181	2600	2646.41	2699.12	0.23240	978952.76	978966.85	64	51.31	-83.21	-88.91	0.06	-88.65	4	2	4	17	11	833.18
MU 3.17	55.974611	25.568083	2694.25	2694.25	2694.25	2694.25	1.0181	2600	2646.41	2699.12	0.23240	978952.76	978966.85	64	51.31	-83.21	-88.91	0.06	-88.65	4	2	4	12	30	833.18
MU 3.17	55.974611	25.568083	2694.25	2694.25	2694.25	2694.25	1.0181	2600	2646.41	2699.12	0.23240	978952.76	978966.85	64	51.31	-83.21	-88.91	0.06	-88.65	4	2	4	12	30	833.18
MU 3.19	55.974611	25.568083	2694.25	2694.25	2694.25	2694.25	1.0181	2600	2646.41	2699.12	0.23240	978952.76	978966.85	64	51.31	-83.21	-88.91	0.06	-88.65	4	2	4	12	30	833.18
MU 3.19	55.974611	25.568083	2694.25	2694.25	2694.25	2694.25	1.0181	2600	2646.41	2699.12	0.23240	978952.76	978966.85	64	51.31	-83.21	-88.91	0.06	-88.65	4	2	4	12	30	833.18
MU 4.10	56.127194	25.832000	2667.06	2667.04	2667.05	2667.05	1.0181	2600	2646.41	2714.68	0.23408	978939.40	979015.65	186	188.88	-19.51	-39.93	2.84	-37.09	4	2	4	14	54	830.90
MU 4.20	56.107611	25.812417	2670.78	2670.79	2670.82	2670.80	1.0181	2600	2646.41	2716.51	0.23419	978943.22	979014.27	154	154.16	-23.47	-40.59	3.13	-37.46	4	2	4	15	3	831.05
MU 4.30	56.082222	25.797778	2668.52	2668.52	2668.54	2668.53	1.0181	2600	2646.41	2716.19	0.23433	978940.90	979013.23	126	124.11	-34.03	-47.81	1.81	-46.00	4	2	4	15	15	831.25
MU 4.40	56.058194	25.782167	2667.77	2667.76	2667.78	2667.77	1.0181	2600	2646.41	2716.42	0.23444	978940.13	979012.12	86	87.18	-46.09	-54.77	1.46	-53.32	4	2	4	15	25	831.42
MU 4.50	56.030939	25.775528	2667.25	2667.25	2667.24	2667.25	1.0181	2600	2646.41	2706.69	0.23468	978939.38	979011.65	52	49.92	-46.55	-72.39	0.86	-71.54	4	2	4	15	34	831.57
MU 4.60	56.002750	25.785111	2668.97	2668.98	2668.97	2668.97	1.0181	2600	2646.41	2696.27	0.23479	978939.38	979012.53	23	25.28	-83.55	-86.36	0.43	-85.93	4	2	4	15	45	831.75
MU 4.70	55.937417	25.793244	2651.09	2651.08	2651.10	2651.09	1.0181	2600	2646.41	2696.44	0.23492	978932.15	979012.65	-2	0.00	-89.71	-89.71	0.23	-88.46	4	2	4	16	6	832.10
MU 5.31	56.191687	24.810556	2677.48	2677.46	2677.48	2677.47	1.0181	2600	2646.41	2725.29	0.24885	978949.99	979944.29	267	263.42	86.99	-57.74	0.52	-56.27	5	2	4	12	0	853.00
MU 5.30	56.165444	24.817500	2682.10	2682.08	2682.11	2682.10	1.0181	2600	2646.41	2708.64	0.24905	978934.34	979944.77	274	270.42	73.02	-43.00	0.52	-43.52	5	2	4	12	17	853.28
MU 5.29	56.134278	24.819861	2686.64	2686.65	2686.65	2686.65	1.0181	2600	2646.41	2672.54	0.24914	978988.24	979944.53	315	313.94	50.19	15.53	0.42	-15.76	5	2	4	12	25	853.42
MU 5.28	56.088806	24.836222	2618.97	2618.95	2618.98	2618.97	1.0181	2600	2646.41	2665.73	0.24925	978988.43	979946.06	349	346.56	51.94	13.24	0.24	-13.61	5	2	4	12	34	853.57
MU 5.27	56.047861	24.842611	2616.87	2616.81	2616.91	2616.89	1.0181	2600	2646.41	2665.61	0.24947	978988.30	979946.50	375	373.61	57.72	16.01	0.17	-16.25	5	2	4	12	42	853.70
MU 5.26	56.010194	24.854194	2624.75	2624.74	2624.77	2624.75	1.0181	2600	2646.41	2671.63	0.24967	978989.18	979947.63	348	350.99	57.34	18.37	0.17	-18.54	5	2	4	13	2	853.83
MU 5.25	55.970083	24.865028	2627.57	2627.56	2627.57	2627.57	1.0181	2600	2646.41	2671.63	0.24967	978989.18	979947.63	348	350.99	57.34	18.37	0.17	-18.54	5	2	4	13	2	853.83
MU 5.24	55.929722	24.865533	2628.50	2628.48	2628.49	2628.49	1.0181	2600	2646.41	2671.63	0.24967	978989.18	979947.63	348	350.99	57.34	18.37	0.17	-18.54	5	2	4	13	10	853.17
MU 5.23	55.882111	24.880194	2610.95	2610.94	2610.92	2610.94	1.0181	2600	2646.41	2657.42	0.24967	978988.22	979949.06	285	286.43	43.34	10.49	0.12	-10.61	5	2	4	13	17	853.28
MU 5.22	55.863631	24.906222	2595.44	2595.44	2595.45	2595.44	1.01804	2500	2544.61	2641.78	0.24966	978988.22	979949.06	285	286.43	43.34	10.49	0.12	-10.61	5	2	4	13	25	853.42
MU 5.21	55.816833	24.905694	2583.19	2583.18	2583.21	2583.19	1.01804	2500	2544.61	2629.32	0.24966	978985.02	979950.85	205	202.62	-34.61	-57.00	0.07	-56.93	5	2	4	13	35	853.58
MU 5.20	55.778083	24.912750	2574.86	2574.85	2574.88	2574.86	1.01804	2500	2544.61	2629.32	0.25006	978985.02	979950.85	205	202.62	-34.61	-57.00	0.07	-56.93	5	2	4	13	41	853.65
MU 5.19	55.754556	24.926839	2575.50	2575.49	2575.52	2575.50	1.01804	2500	2544.61	2629.32	0.25006	978985.02	979950.85	205	202.62	-34.61	-57.00	0.07	-56.93	5	2	4	13	51	853.85
MU 5.18	55.727417	24.935000	2575.32	2575.30	2575.32	2575.32	1.01804	2500	2544.61	2629.32	0.25006	978985.02	979950.85	205	202.62	-34.61	-57.00	0.07	-56.93	5	2	4	13	51	853.85
MU 5.17	55.693639	24.947861	2574.17	2574.14	2574.16	2574.16	1.01804	2500	2544.61	2629.32	0.25006	978985.02	979950.85	205	202.62	-34.61	-57.00	0.07	-56.93	5	2	4	14	37	854.72
MU 5.16	55.663639	24.957083	2573.85	2573.82	2573.85	2573.85	1.01804	2500	2544.61	2629.32	0.25006	978985.02	979950.85	205	202.62	-34.61	-57.00	0.07	-56.93	5	2	4	14	37	854.72
MU 5.15	55.636833	25.003000	2570.23	2570.23	2570.24	2570.23	1.01804	2500	2544.61	2612.65	0.25114	978980.81	979958.12	150	146.02	-75.93	-86.87	0.05	-86.87	5	2	4	14	56	855.35
MU 5.14	55.594083	25.003000	2570.23	2570.23	2570.24	2570.23	1.01804	2500	2544.61	2612.65	0.25114	978980.81	979958.12	150	146.02	-75.93	-86.87	0.05	-86.87	5	2	4	14	56	855.35
MU 5.13	55.557667	25.004056	2566.80	2566.82	2566.83	2566.82	1.01804	2500	2544.61	2612.65	0.25114	978980.81	979958.12	150	146.02	-75.93	-86.87	0.05	-86.87	5	2	4	15	16	855.37
MU 5.12	55.520167	25.011194	2568.47	2568.46	2568.48	2568.47	1.01804	2500	2544.61	2612.65	0.25136	978983.02	979959.13	110	111.40	-87.32	-98.69	0.02	-98.66	5	2	4	15	25	855.58
MU 5.11	55.479833	25.007444	2573.58	2573.57	2573.60	2573.58	1.01804	2500	2544.61	2624.00	0.25160	978944.22	979959.14	100	100.50	-87.18	-98.68	0.02	-98.66	5	2	4	15	47	855.78
MU 5.10	55.448806	24.992972	2577.98	2577.98	2577.97	2577.97	1.01804	2500																	

Point Name	Lat	Lon	R1	R2	R3	R average	Convention	Factor	Level	Mgal	Meier	Drift	ASS Gravity	Theoretical	Elevation	FAA	BAA	Train	BAA	Day	Month	Year	Hour	Min	Time since	
	degrees	degrees	meter unit	meter unit	meter unit	meter unit	= 1 mgal		Level	Level	mgal	mgal	Correction	Mgal	mgal	GravBaro	SR7M	SR7M	Correction	Full				Time since		
																SR7M50								07/07/04		
HA 1.22	55.797694	24.103361	2480.52	2480.47	2480.51	2480.50	1.01797	2400	2442.81	2524.75	0.26938		978749.44	978866.12	320	314.47	-49.63	-84.55	0.36	-84.19	6	2	4	17	20	881.33
HA 1.21	55.795200	24.102139	2480.47	2480.51	2480.49	2480.48	1.01797	2400	2442.81	2524.75	0.26934		978748.45	978866.04	324	317.50	-48.63	-83.89	0.45	-83.44	6	2	4	17	14	881.13
HA 1.20	55.781778	24.100750	2479.91	2479.95	2479.93	2479.93	1.01797	2400	2442.81	2524.75	0.26934		978748.45	978866.04	330	324.99	-46.66	-83.33	0.62	-82.29	6	2	4	17	8	881.03
HA 1.19	55.781778	24.098361	2479.20	2477.15	2477.16	2477.16	1.01797	2400	2442.81	2521.36	0.26917		978748.03	978865.85	336	335.51	-46.28	-83.33	0.14	-82.70	6	2	4	17	2	881.03
HA 1.18	55.781778	24.098361	2479.20	2475.28	2475.27	2475.28	1.01797	2400	2442.81	2517.97	0.26899		978748.03	978865.85	346	351.60	-41.36	-81.56	0.15	-79.92	6	2	4	16	55	880.92
HA 1.17	55.771250	24.108361	2477.63	2477.67	2477.65	2477.65	1.01797	2400	2442.81	2523.86	0.26893		978748.53	978866.47	347	346.93	-43.19	-81.60	0.76	-80.84	6	2	4	16	41	880.68
HA 1.15	55.780238	24.113000	2481.96	2481.93	2481.95	2481.95	1.01797	2400	2442.81	2526.23	0.26866		978758.91	978867.77	328	372.93	-41.57	-79.10	0.43	-78.67	6	2	4	16	35	880.58
HA 1.14	55.784881	24.113000	2481.96	2481.93	2481.95	2481.95	1.01797	2400	2442.81	2523.45	0.26872		978758.91	978867.77	303	303.21	-47.31	-80.98	0.40	-80.58	6	2	4	16	29	880.38
HA 1.13	55.784881	24.120417	2480.49	2480.49	2480.51	2480.50	1.01797	2400	2442.81	2523.45	0.26872		978758.91	978867.77	303	295.61	-46.36	-79.20	0.29	-78.91	6	2	4	16	23	880.38
HA 1.12	55.784881	24.120417	2480.49	2480.49	2480.51	2480.50	1.01797	2400	2442.81	2523.45	0.26864		978758.91	978867.77	286	286.41	-47.12	-78.53	0.27	-78.43	6	2	4	16	16	880.27
HA 1.11	55.784881	24.120417	2480.49	2480.49	2480.51	2480.50	1.01797	2400	2442.81	2523.45	0.26864		978758.91	978867.77	286	286.41	-47.12	-78.53	0.24	-77.44	6	2	4	16	10	880.17
HA 1.10	55.784881	24.113000	2482.48	2482.51	2482.50	2482.50	1.01797	2400	2442.81	2526.97	0.26847		978761.65	978868.99	289	289.37	-45.43	-77.78	0.38	-77.09	6	2	4	16	5	880.03
HA 1.09	55.784881	24.113000	2483.04	2483.02	2483.06	2483.04	1.01797	2400	2442.81	2526.97	0.26847		978761.65	978868.99	289	289.37	-45.43	-77.78	0.34	-77.09	6	2	4	15	53	879.88
HA 1.08	55.784881	24.110000	2485.93	2485.98	2485.95	2485.95	1.01797	2400	2442.81	2524.65	0.26829		978765.14	978868.60	280	286.03	-48.47	-78.34	0.36	-77.06	6	2	4	15	46	879.77
HA 1.07	55.784881	24.110000	2485.93	2485.98	2485.95	2485.95	1.01797	2400	2442.81	2524.65	0.26821		978765.14	978868.60	276	282.75	-49.18	-78.34	0.29	-76.06	6	2	4	15	39	879.65
HA 1.06	55.784881	24.108917	2487.92	2487.95	2487.97	2487.92	1.01797	2400	2442.81	2524.35	0.26812		978767.22	978868.49	270	269.55	-49.49	-78.20	0.25	-77.94	6	2	4	15	32	879.53
HA 1.05	55.784881	24.108917	2487.92	2487.95	2487.97	2487.92	1.01797	2400	2442.81	2524.35	0.26803		978767.22	978868.49	265	248.58	-51.26	-78.95	0.21	-78.40	6	2	4	15	8	879.25
HA 1.04	55.784881	24.105028	2489.84	2489.86	2489.84	2489.84	1.01797	2400	2442.81	2524.81	0.26784		978768.12	978868.23	259	240.57	-51.35	-78.61	0.18	-78.40	6	2	4	15	8	879.25
HA 1.03	55.784881	24.103361	2489.84	2489.86	2489.84	2489.84	1.01797	2400	2442.81	2524.81	0.26784		978768.12	978868.23	254	240.57	-51.35	-78.61	0.18	-78.40	6	2	4	15	8	879.25
HA 1.02	55.784881	24.107122	2489.84	2489.86	2489.84	2489.84	1.01797	2400	2442.81	2524.81	0.26784		978768.12	978868.23	254	240.57	-51.35	-78.61	0.18	-78.40	6	2	4	15	8	879.25
HA 1.01	55.784881	24.102500	2489.84	2489.86	2489.84	2489.84	1.01797	2400	2442.81	2524.81	0.26784		978768.12	978868.23	254	240.57	-51.35	-78.61	0.18	-78.40	6	2	4	15	8	879.25
HA 2.01	55.694722	24.112444	2489.53	2489.54	2489.57	2489.55	1.01797	2400	2442.81	2524.17	0.26165		978768.04	978868.73	248	235.67	-55.19	-81.96	0.09	-81.27	8	2	4	15	26	927.43
HA 2.02	55.694722	24.112444	2489.53	2489.54	2489.57	2489.55	1.01797	2400	2442.81	2524.17	0.26165		978768.04	978868.73	245	231.73	-56.76	-82.49	0.09	-82.41	8	2	4	15	33	927.55
HA 2.03	55.694722	24.108914	2487.67	2487.72	2487.69	2487.69	1.01797	2400	2442.81	2524.17	0.26165		978768.04	978868.73	245	227.60	-58.54	-83.81	0.08	-83.73	8	2	4	15	40	927.67
HA 2.04	55.694722	24.108914	2487.67	2487.72	2487.69	2487.69	1.01797	2400	2442.81	2524.17	0.26165		978768.04	978868.73	233	224.99	-58.59	-83.87	0.06	-83.91	8	2	4	15	46	927.77
HA 2.05	55.662278	24.078684	2486.66	2486.64	2486.65	2486.66	1.01797	2400	2442.81	2524.12	0.30198		978768.06	978868.46	229	218.38	-61.19	-85.44	0.05	-85.39	8	2	4	15	54	927.90
HA 2.06	55.662278	24.078684	2486.66	2486.64	2486.65	2486.66	1.01797	2400	2442.81	2524.12	0.30206		978768.06	978868.46	229	218.38	-60.52	-84.73	0.03	-84.70	8	2	4	16	1	928.02
HA 2.07	55.630000	24.062528	2487.62	2487.58	2487.60	2487.60	1.01797	2400	2442.81	2524.12	0.30213		978768.06	978868.46	227	210.82	-61.50	-84.91	0.02	-84.88	8	2	4	16	7	928.15
HA 2.08	55.630000	24.054472	2487.62	2487.58	2487.60	2487.60	1.01797	2400	2442.81	2524.12	0.30223		978768.06	978868.46	223	217.19	-59.10	-83.21	0.03	-83.19	8	2	4	16	17	928.25
HA 2.09	55.630000	24.049056	2489.74	2489.70	2489.72	2489.72	1.01797	2400	2442.81	2524.12	0.30231		978768.06	978868.46	215	204.71	-60.32	-83.05	0.02	-83.03	8	2	4	16	22	928.30
HA 2.10	55.597083	24.042750	2502.57	2502.55	2502.53	2502.55	1.01804	2500	2544.61	2547.19	0.30240		978771.83	978882.04	209	206.20	-56.58	-79.47	0.01	-79.46	8	2	4	16	30	928.57
HA 7.40	55.742694	24.224750	2488.82	2488.83	2488.83	2488.81	1.01797	2400	2442.81	2524.38	0.28329		978768.04	978904.31	262	272.70	-52.12	-82.40	0.03	-82.36	7	2	4	13	12	901.20
HA 7.60	55.712639	24.220167	2488.52	2488.57	2488.54	2488.51	1.01797	2400	2442.81	2524.12	0.28342		978767.78	978904.00	253	259.45	-56.15	-84.96	0.03	-84.93	7	2	4	13	22	901.37
HA 7.50	55.684139	24.206633	2489.13	2489.10	2489.16	2489.13	1.01797	2400	2442.81	2524.12	0.28351		978768.04	978904.31	248	255.68	-55.78	-84.17	0.03	-84.15	7	2	4	13	32	901.53
HA 7.70	55.636311	24.185500	2500.58	2500.56	2500.57	2500.57	1.01804	2500	2544.61	2546.23	0.26361		978770.96	978901.29	229	241.84	-57.84	-84.89	0.02	-84.87	7	2	4	13	39	901.63
HA 7.80	55.636311	24.185500	2500.58	2500.56	2500.57	2500.57	1.01804	2500	2544.61	2546.23	0.26361		978770.96	978901.29	229	241.84	-57.84	-84.89	0.02	-84.86	7	2	4	13	50	901.85
HA 7.90	55.636311	24.185500	2500.58	2500.56	2500.57	2500.57	1.01804	2500	2544.61	2546.23	0.26361		978770.96	978901.29	229	241.84	-57.84	-84.89	0.02	-84.86	7	2	4	13	57	901.85
HA 7.90	55.636311	24.185500	2500.58	2500.56	2500.57	2500.57	1.01804	2500	2544.61	2546.23	0.26361		978770.96	978901.29	229	241.84	-57.84	-84.89	0.02	-84.86	7	2	4	13	57	901.85
HA 7.90	55.636311	24.185500	2500.58	2500.56	2500.57	2500.57	1.01804	2500	2544.61	2546.23	0.26361		978770.96	978901.29	229	241.84	-57.84	-84.89	0.02	-84.86	7	2	4	13	57	901.85
HA 7.90	55.636311	24.185500	2500.58	2500.56	2500.57	2500.57	1.01804	2500	2544.61	2546.23	0.26361		978770.96	978901.29	229	241.84	-57.84	-84.89	0.02	-84.86	7	2	4	13	57	901.85
HA 7.90	55.636311	24.185500	2500.58	2500.56	2500.57	2500.57	1.01804	2500	2544.61	2546.23	0.26361		978770.96	978901.29	229	241.84	-57.84	-84.89	0.02	-84.86	7	2	4	13	57	901.85
HA 7.90	55.636311	24.185500	2500.58	2500.56	2500.57	2500.57	1.01804	2500	2544.61	2546.23	0.26361		978770.96	978901.29	229	241.84	-57.84	-84.89	0.02	-84.86	7	2	4	13	57	901.85
HA 7.90	55.636311	24.185500	2500.58																							

Point	Name	Lat	Lon	R1	R2	R3	R average	Conversion Factor	Conversion Level	Mgal Level	Meter gravity	Drift mgal	ABS Gravity Mgal	Theoretical Gravity mgal	Elevation GPS/baro m	Elevation SRTM-90 m	FAA SRTM	BAA SRTM	Terrain Correction mgal	BAA Full mgal	Day	Month	Year	Hour	Min	Time since 01/01/04 hours
MU	7 42	54.623139	24.299806	2600.57	2600.53	2600.55	2600.55	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	19	22.83	-30.73	-33.27	0.01	-33.26	8	2	4	10	52	922.87
MU	7 43	54.596028	24.313288	2605.06	2605.02	2605.04	2605.04	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	3	5.86	-32.33	-32.97	0.01	-32.97	8	2	4	10	59	922.88
MU	7 44	54.579194	24.337444	2606.37	2606.34	2606.35	2606.35	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	1	4.99	-32.89	-33.44	0.01	-33.44	8	2	4	11	6	923.10
MU	7 45	54.564639	24.361472	2607.32	2607.30	2607.34	2607.34	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	3	8.14	-32.54	-33.44	0.01	-33.44	8	2	4	11	14	923.23
MU	7 46	54.543981	24.385000	2610.00	2610.00	2610.00	2610.00	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	-2	4.65	-32.53	-33.04	0.00	-33.04	8	2	4	11	22	923.50
MU	7 47	54.522278	24.405556	2611.42	2611.39	2611.40	2611.40	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	-2	6.47	-31.97	-32.69	0.00	-32.69	8	2	4	11	30	923.60
MU	7 48	54.506444	24.438889	2612.70	2612.67	2612.71	2612.69	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	2	7.62	-31.10	-31.93	0.00	-31.94	8	2	4	11	39	923.65
MU	7 49	54.537917	24.471444	2617.80	2617.84	2617.82	2617.82	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	1	0.00	-31.86	-31.86	0.01	-31.85	8	2	4	12	17	924.28
MU	8 01	54.63972	24.409639	2608.68	2608.63	2608.64	2608.65	1.01811	2600	2646.41	2646.97	0.29846	978871.62	978909.33	6	3.37	-35.96	-36.33	0.01	-36.33	8	2	4	12	52	924.87
MU	8 02	54.742472	24.417972	2598.84	2598.80	2598.82	2598.82	1.01804	2500	2544.61	2545.22	0.30001	978869.87	978917.42	32	30.45	-38.15	-41.54	0.01	-41.53	8	2	4	13	5	925.08
MU	8 03	54.845139	24.437139	2593.18	2593.14	2593.17	2593.16	1.01804	2500	2544.61	2545.22	0.30015	978884.11	978918.72	42	39.19	-42.53	-46.88	0.01	-46.86	8	2	4	13	17	925.28
MU	8 04	54.943889	24.433028	2583.06	2583.04	2583.07	2583.06	1.01804	2500	2544.61	2545.22	0.30025	978883.82	978918.44	69	65.03	-44.55	-51.77	0.01	-51.76	8	2	4	13	26	925.43
MU	8 05	55.037806	24.415611	2573.33	2573.33	2573.33	2573.33	1.01804	2500	2544.61	2545.22	0.30037	978883.91	978917.26	94	88.75	-45.66	-55.62	0.01	-55.61	8	2	4	13	36	925.60
MU	9 01	55.78667	24.205750	2493.07	2493.11	2493.08	2493.09	1.01797	2400	2442.81	2537.56	0.31367	978762.19	978903.03	299	299.01	-48.56	-81.76	0.05	-81.71	9	2	4	8	36	944.60
MU	9 02	55.819694	24.204806	2490.93	2490.96	2490.94	2490.94	1.01797	2400	2442.81	2535.38	0.31376	978760.02	978902.96	311	313.18	-46.30	-81.07	0.05	-81.02	9	2	4	8	44	944.73
MU	9 03	55.850083	24.199389	2487.30	2487.33	2487.32	2487.32	1.01797	2400	2442.81	2531.70	0.31383	978756.33	978902.60	328	331.03	-44.11	-80.87	0.05	-80.81	9	2	4	8	50	944.83
MU	9 04	55.879528	24.195667	2482.56	2482.59	2482.57	2482.57	1.01797	2400	2442.81	2526.86	0.31393	978751.49	978902.47	347	352.26	-42.27	-81.39	0.06	-81.33	9	2	4	8	58	944.97
MU	9 05	55.909333	24.192667	2477.22	2477.26	2477.24	2477.24	1.01797	2400	2442.81	2521.43	0.31402	978746.06	978902.35	370	370.54	-41.94	-83.08	0.07	-83.02	9	2	4	9	6	945.10
MU	9 06	55.937389	24.202694	2478.29	2478.30	2478.30	2478.30	1.01797	2400	2442.81	2522.52	0.31410	978747.15	978902.82	390	395.25	-33.70	-83.08	0.08	-83.02	9	2	4	9	13	945.22
MU	9 07	55.968778	24.225139	2481.89	2481.93	2481.90	2481.91	1.01797	2400	2442.81	2526.18	0.31422	978750.81	978904.34	409	412.25	-26.30	-72.08	0.09	-72.08	9	2	4	9	23	945.38
MU	9 08	55.978528	24.242833	2497.05	2497.01	2497.02	2497.03	1.01797	2400	2442.81	2541.57	0.31432	978766.20	978905.54	430	430.11	-6.60	-54.36	0.13	-54.22	9	2	4	9	32	945.53
MU	9 09	56.007417	24.246500	2504.96	2504.92	2504.94	2504.94	1.01804	2500	2544.61	2549.64	0.31441	978774.27	978905.78	458	460.67	10.65	-40.50	0.10	-40.41	9	2	4	9	39	945.65
MU	9 10	56.036583	24.247972	2520.45	2520.41	2520.43	2520.43	1.01804	2500	2544.61	2554.41	0.31450	978779.67	978906.12	480	481.98	32.89	-20.62	0.11	-20.51	9	2	4	9	47	945.78
MU	9 11	56.066722	24.251528	2529.90	2529.88	2529.89	2529.89	1.01804	2500	2544.61	2564.61	0.31467	978799.67	978906.12	506	507.62	55.20	-6.17	0.13	-6.04	9	2	4	9	55	945.92
MU	9 12	56.096083	24.251667	2527.16	2527.18	2527.15	2527.16	1.01804	2500	2544.61	2575.25	0.31467	978799.67	978906.12	536	535.45	50.89	-3.47	0.14	-3.33	9	2	4	10	2	946.03
MU	9 13	56.123361	24.241056	2523.29	2523.31	2523.27	2523.29	1.01804	2500	2544.61	2568.30	0.31474	978792.93	978906.42	569	570.90	63.69	0.30	0.17	0.44	9	2	4	10	8	946.13
MU	9 14	56.149250	24.229639	2516.53	2516.52	2516.50	2516.52	1.01804	2500	2544.61	2561.41	0.31483	978786.04	978904.64	602	604.89	68.06	0.90	0.22	-2.20	9	2	4	10	15	946.25
MU	9 15	56.158028	24.225278	2515.49	2515.51	2515.48	2515.49	1.01804	2500	2544.61	2555.06	0.31488	978784.99	978904.35	595	591.91	63.30	-2.42	0.22	-2.20	9	2	4	10	20	946.33
MU	9 16	56.167894	24.209833	2510.10	2510.50	2510.28	2510.22	1.01804	2500	2544.61	2555.06	0.31499	978779.67	978903.24	485	482.24	28.34	-26.32	0.64	-25.68	10	2	4	10	55	970.82
MU	9 17	56.176722	24.386278	2503.68	2503.65	2503.66	2503.66	1.01804	2500	2544.61	2555.06	0.31500	978786.04	978903.24	485	482.24	28.34	-26.32	0.64	-25.68	10	2	4	11	6	971.10
MU	9 18	56.185161	24.314611	2503.68	2503.68	2503.67	2503.67	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 19	56.193722	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
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MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.68	2503.68	2503.66	2503.66	1.01811	2600	2646.41	2723.78	0.33241	978988.57	978993.58	342	342.42	62.85	24.83	0.36	78.32	10	2	4	11	22	971.27
MU	9 20	56.201528	24.386278	2503.																						

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