Lithospheric cooling and thickening as a basin forming mechanism

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ABSTRACT

Widely accepted basin forming mechanisms are limited to flexure of the lithosphere, and lithospheric stretching followed by cooling and thermal subsidence. Neither of these mechanisms works for a group of large basins, sometimes known as "intracontinental sags". In this paper we investigate cooling and thickening of initially thin lithosphere as a basin forming mechanism, by a combination of forward modelling and a backstripping study of two Palaeozoic North African basins: Ghadames and Al Kufrah. These are two of a family of basins, once unified, which lie over the largely accretionary crust of North Africa and Arabia. Such accretionary crust tends to be juvenile, consisting of amalgamated island arcs, accretionary prisms and melanges, and typically has near-normal crustal thicknesses but initially thin mantle lithosphere. Post-accretion subsidence is modelled using a plate cooling model similar to cooling models for oceanic lithosphere. The crustal composition and thickness used in the models are varied around average values of accretionary crust to represent likely heterogeneity. The model allows the lithosphere to thicken as it cools and calculates the resulting isostatic subsidence.

Water-loaded tectonic subsidence curves from these forward models are compared to tectonic subsidence curves produced from backstripped wells from Al Kufrah and Ghadames basins. A good match between the subsidence curves for the forward model and backstripping is produced when the best estimates for the crustal structure, composition and the present day thickness of the lithosphere for North Africa are used as inputs for the forward model. The model produces sediment loaded basins of 2-7 km thickness for the various crustal assemblies over ~250 Myrs. This shows that lithospheric cooling provides a viable method for producing large basins with prolonged subsidence, without the need for initial extension, provided the condition of initially thin mantle lithosphere is met.

Keywords: basin, lithosphere, subsidence, North Africa

1. Introduction

Conventional basin formation mechanisms can be divided into two categories, lithospheric stretching followed by thermal subsidence proportional to the extension (rift basins) and flexure caused by tectonic loading (foreland basins) (Allen and Allen, 2005). However, there is scant evidence for either of these mechanisms forming a group of basins normally classified as "intracontinental sags" – a term that describes their geometry rather than the process of formation. Examples include the Williston and Michigan basins of North America (Klein, 1995), the Palaeozoic basins of North African and Arabia (Boote et al., 1998; Konert et al., 2001); and the Mesozoic Scythian and Turan platforms (Natal'in and Şengör, 2005). The basins are large features, commonly over 1000 km in length, with remarkably uniform, prolonged, gentle subsidence across them, lasting over 200 Myrs. They generally have a polyphase history with a main subsidence phase either preceded or followed by other periods of subsidence and uplift which modify the basin. Armitage and Allen, (2010) recently proposed that these basins are formed by stretching under low strain rates. They argued this based upon the modelling of rifting under low strain rates and the observation that the initiation of subsidence in many intracontinental basins coincides with supercontinent breakup and therefore a broad extensional regime. However, in many basins the evidence for rifting is poor and a number of other mechanisms have been proposed. Subsidence due to cooling of thermal anomalies in the lithosphere has been proposed for the North American intracontinental basins (Kaminski and Jaupart, 2000). The main evidence for this is matching modelling results with the shape and thickness of the present day sedimentary cover. A density change in the crust due to phase changes such as a basalt underplate changing to eclogite has also been suggested based on high velocities in the lower crust interpreted from seismic refraction data (Artyushukov, 2005). However, beneath the Barents/Kara sea region a high density area in the lower crust, suggested by modelling gravity data, has been deemed too local to cause subsidence across the basin. Instead Ritzmann and Faleide, (2009) have suggested that a deeper high velocity zone visible in seismic tomography is evidence of a thick cratonic lithosphere, which causes the subsidence. Heine et al. (2008) noted that many intracratonic basins overlie areas of mantle which have been down welling over the last 100-150 Myrs in their coupled plate and mantle flow model. They proposed that dynamic topography could form these basins. Other subsidence mechanisms and variations on those above have been suggested and are debated in more detail by Armitage and

Allen (2010) and Klein (1995). It is likely that one mechanism does not explain the formation of every intercontinental basin and in some cases the basin may be formed by a combination of mechanisms.

Here we show that cooling and thickening of initially thin mantle lithosphere, beneath crust of normal thickness (~30 km) is a viable mechanism for producing basin-scale subsidence. Such initial conditions are typical of accretionary crust, a term used to summarise the vast orogenic collages of largely juvenile crust and mantle lithosphere, formed by the collision of non-cratonic terranes: island arcs, accretionary prisms, ophiolites and isolated microcontinents (Murphy and Nance, 1991; Şengör et al., 1993). This would neatly explain the formation of many of the intracratonic basins on juvenile continental crust such as the Pan African mobile belt or the Scythian and Turan platform. However, we show that where the lithosphere is thinned by a thermal anomaly it is also possible to form broad, slowly subsiding basins.

This subsidence mechanism is discussed in greater detail in section 2 below, followed by a case study of two of the North African Palaeozoic basins. Thermal subsidence has been suggested as the cause of intracratonic basins before (Guiraud et al., 2005; Kaminski and Jaupart, 2000; Kominz, 1995). Our contribution is to model the subsidence, compare it to subsidence in two case studies and to discuss why the lithosphere is plausibly thin in the first place. The subsidence history of the basins is analysed using backstripping. This analysis is compared to results from a numerical forward model of thermal subsidence acting on accretionary crust, designed to test if it is a mechanism capable of producing the observed subsidence.

2. Geological background and hypothesis

Seismic refraction studies show that present day island arcs can have a crustal thickness of 25-35 km, similar to normal continental crust (Holbrook et al., 1999; Takahashi et al., 2007). However, seismic tomography shows slow velocities in the mantle wedge below island arcs which are interpreted as evidence for the presence of melts and thin (~20 km) mantle lithosphere because it is weakened by the addition of fluids from the subducting slab and then eroded by the corner flow in the mantle wedge (Gorbatov et al., 1999; Zhao et al., 1994). This is supported by numerical models of subduction (Arcay et al., 2006; Stern, 2002; van Keken, 2003), by geochemical evidence from the southward initiation of the Philippine subduction zone (Macpherson, 2008) and from the Cascades (Elkins Tanton et al., 2009). These studies suggest an average overall lithospheric thickness of about 50 km beneath island arcs (Fig. 1a).

Accretionary prisms may be 30 km thick, largely composed of off-scraped and imbricated fragments of oceanic crust and its sedimentary cover. Whilst subduction is active such prisms are underlain by the oceanic plate. When subduction has recycled the oceanic plate in to the mantle, the base of the prism may be in contact with the asthenosphere, particularly if ocean closure resulted in the collision of two such prisms, initially on opposite sides of the ocean, rather than collision of the prism with a continental margin. The Cenozoic East Anatolian Accretionary Complex may be an example of such a lithospheric structure, where tomographic studies suggest a thin or even absent mantle lithosphere (Zor et al., 2003).

A notable feature of accretionary orogenic belts is that they lack evidence for substantial crustal thickening (and presumably lithosphere thickening): there is rarely evidence for pre-collision passive continental margins, Alpine-type nappes, or overfilled foreland basins ("molasse") (Şengör and Okurogullari, 1991). This means that putative lithospheric delamination following an orogeny of this type, hypothesised by Ashwal and Burke, (1989), is not our preferred mechanism for thinning the lithosphere. However, it would produce similar starting conditions to those in our model.

As accretionary crust is assembled through subduction and collision, the thin mantle lithosphere of the original terranes is inherited by the final collage (Fig. 1b). We hypothesise that once accretion is completed, and subduction has ceased beneath an area, the underlying asthenosphere will cool, thickening the mantle lithosphere. This cooling will cause prolonged subsidence, forming basins (Fig. 1c).

Figure 1 should be placed here

Our model is similar to the thermal subsidence phase of McKenzie style rifting (McKenzie, 1978) or the subsidence of the ocean floor away from a mid ocean ridge, except that the crust involved is continental, albeit juvenile, and has not been thinned in any way. In this paper we proceed to show how this mechanism could produce the basins in North Africa. However, there are many other intracratonic basins on accreted crust where this mechanism could apply. Table 1 provides a sample of some of the basins we aware of, but is by no means an exhaustive list. Allen and Armitage, (in press) note a clustering in time of the initiation of intracratonic basins which they link to the breakup of supercontinents. In Table 1 we show the start of subsidence follows closely the end of accretion and the clustering may be related to the end of periods of accretion of crust. The basins are long lived features and so many have later phases of subsidence which potentially have other causes.

Table 1 should be placed here.

3. Tectonic subsidence history of backstripped North African basins

In order to test whether the proposed mechanism of subsidence provides a good explanation for anomalous basins developed over accretionary crust, subsidence histories for the Ghadames and Al Kufrah basins were investigated using backstripping. These are Palaeozoic basins situated on the North African crust, which was accreted in the Pan African orogeny during Neoproterozoic times (Caby, 2003; Stern, 1994). Most of the basement to North Africa and Arabia is juvenile, generated and assembled during the Pan African orogeny. The western margin is the West African Craton. The southern limit is the Congo Craton. The eastern and northern limits are not so well defined because of later rifting and collision with Eurasia. For example, similar basement underlies much of the territory of Iran, but with a more complicated Mesozoic and Cenozoic magmatic and tectonic history. The outcrop or sub-Mesozoic subcrop of the Palaeozoic strata of North Africa and Arabia is shown in Fig. 2.

Figure 2 should be placed here

The proportion of cratonic nuclei within this vast, 10,000,000 km^2 orogen is debated, but plausibly is small. There are indications of pre Late Proterozoic crust within North Africa, based on Meso- or Palaeo-Proterozoic isotopic model ages and detritial zircons in younger metamorphic terranes (Black et al., 1993; Sultan et al., 1990). But the Late Proterozoic tectonic overprint is severe, suggesting that extensive magmatic and metamorphic re-working and additions took place during the Pan African orogeny, and no single, regionally extensive block survived the Late Proterozoic orogeny with its original structure and boundaries preserved. A significant re-worked region has been called the Saharan metacraton (Abdelsalam et al., 2002), but this is not recognised by all workers e.g. (Bumby and Guiraud, 2005). Whatever the nature and origins of the Saharan metacraton, it appears to have had little impact on the overlying

Phanerozoic basins (Fig. 2). These are continuous across the metacraton margins, without discernible change in sedimentary thickness or composition.

The end of the Pan African orogeny was diachronous. Local timings for the last deformation vary from Late Precambrian to the Early Cambrian e.g. (Paquette et al., 1998). Present-day lithosphere thickness across North Africa is on the order of ~100 km (Pasyanos and Nyblade, 2007; Priestley and McKenzie, 2006); except for the West African Craton where it reaches >200 km. Crustal thicknesses are not well-constrained, but are variously estimated at 30 to 40 km thick from gravity and seismic interpretations (Seber et al., 2001), or 25 to 35 km from the inversion of surface waves (Pasyanos and Nyblade, 2007).

The Ghadames Basin is well studied because large hydrocarbon accumulations have been discovered within it (Echikh, 1998). Several wells have penetrated the crystalline basement and there are large amounts of seismic data for the basin. This means the geometry of the basin is well understood. This allows the most complete stratigraphic sections to be identified, which is helpful when backstripping. The Al Kufrah Basin is less well studied and has only two published wells which reached the crystalline basement (Grignani et al., 1991). To a first order, it has a similar Palaeozoic stratigraphy to the Ghadames Basin. Enough data are available for Al Kufrah to make it viable for modelling.

3.1 Methodology

Backstripping is a well known technique for calculating the tectonic subsidence of a particular horizon in a basin over time. We use the basement/cover boundary, which reveals the overall subsidence (Allen and Allen, 2005; Sclater and Christie, 1980). The technique first decompacts each sedimentary layer through time. This gives the total subsidence curve for the basement which is assumed to be composed of the tectonic subsidence, the isostatic effect (weight) of the sediments and the changes in water depth compared to the present day sea level. The effects of the sediments and sea level changes are removed so that the remaining subsidence is due purely to the tectonic driving force. A global eustatic sea-level curve of (Haq and Schutter, 2008) is used to remove the effects of sea-level changes.

The backstripping method assumes that the compaction of the sediments is purely mechanical (due to the weight of the sediments above) and ignores chemical processes, such as cementation, which are very difficult to take into account because they depend on a complex series of factors such as fluid flow, composition, temperature and pressure. Backstripping assumes that the sediments are laid down in successive layers throughout time and does not take into account periods of uplift and erosion or non deposition. These are difficult to include in backstripping because generally the amount of eroded material is poorly constrained. If the present day burial depth of the basement is the deepest it has been, then any erosion will make no difference because the sediments are at their peak pressure. Otherwise the compaction of the sediments will be underestimated. There are no porosity depth relationships available for the sediments in North Africa and we use standard relationships from published work (Sclater and Christie, 1980). We assume that the compaction curves, which are based on sediments from the North Sea, are applicable to the sediments in North Africa. This is justifiable because the North African sediments are entirely siliciclastic and are similar to those in the North Sea.

This method was applied to a composite well from the Ghadames Basin. A composite well was used because it gives the most complete section possible from the basin therefore showing the maximum subsidence and limiting the effect of erosion. The composite well was created using four unpublished wells from the transect shown on Fig. 2, to identify the most complete sections and to ensure that the differences in the thickness of the layers were due to variations in erosion rather than deposition rates. This minimises the errors related to eroded sections of the stratigraphy. It also produces a subsidence curve which emphasises the subsidence phases rather than any uplift. This allows us to see clearly the main subsidence phase associated with basin formation, rather than later phases of uplift or subsidence. This approach was not possible in the Al Kufrah Basin because of the scarcity of available well data, and instead the backstripping methodology was applied directly to the two available wells.

3.2 General Stratigraphy

Detailed descriptions of the stratigraphy in the Ghadames, Al Kufrah and other North African basins can be found in papers such as Bellini and Massa (1980), Echikh (1998), Fekirine and Abdallah (1998), Grignani et al. (1991) and Lüning et al. (1999). Therefore we present a brief summary of the evolution of the Palaeozoic basins on the accretionary crust of North Africa collated from Boote et al. (1998), Bumby and Guiraud (2005), Craig et al. (in press), Guiraud and Bosworth (1999) and Guiraud et al. (2005) alongside more detailed observations from the basins themselves. The basins are filled with a largely siliciclastic succession with some evaporites and carbonates towards the end of the Palaeozoic. The whole of North Africa subsided as a large platform from roughly the start of the Palaeozoic, depositing a wedge of sediments that thinned to the south (Selley, 1997). Only localised Late Proterozoic/Early Palaeozoic rifting is known (see Fig. 6 of Guiraud et al. (2005), and as several of the main basins are mature in terms of hydrocarbon exploration it is unlikely that major rifts have been missed. Seismic lines through the Kufrah Basin show Late Proterozoic/Cambrian rifts (Lüning et al., 1999), however these are only seen on seismic lines in the south of the basin (Ghanoush and Abubaker, 2007) and do not explain subsidence across the whole basin or in neighbouring basins. Nor do the dimensions of the subsiding area (>>1000 km) fit a flexural, foreland basin mechanism. In any case, there is no record of an appropriate orogeny lasting through the Palaeozoic along the Gondwanan continental margin (Stampfli and Borel, 2002). The initial sediments are largely fluvial sandstones and conglomerates in the Cambrian, changing to marine sandstones in the early Ordovician. The Cambrian age for the earlier sediments is inferred because there are very few trace fossils to date the sediments until the marine incursions (Grignani et al., 1991). In the mid to late Ordovician there was major glaciation creating a regional erosional surface (Le Heron et al., 2010). The amount of erosion varies greatly reflecting the geometry of the ice sheets. Locally in Al Kufrah the erosion cuts down to the Cambrian, but in other areas a complete section to the Mid Silurian is preserved. In these areas the erosion is minimal and the glaciation reflects a period of non deposition. The erosion becomes less severe to the north into the Murzuq and Ghadames basins (Bellini and Massa, 1980; Fekirine and Abdallah, 1998).

Coarse to medium grained sandstones and tillites related to the retreat of the glaciers were deposited at the end of the Ordovician. This was followed by a major marine transgression leading to the deposition of shales throughout the Early Silurian, but water depths decreased with time and the shales grade up to shoreface sandstones. The shales at the base of the Silurian are one of the main source rocks and record the maximum water depth experienced by the basins during the Palaeozoic. Turner (1978) interpreted it as a shallow shelf environment with water depths of about 200 m. Separation of individual basins began during the Silurian caused by a period of compression most likely related to the Acadian collision (Guiraud and Bosworth, 1999). This resulted in an unconformity across the basins of North Africa, but within basins it is localised over arches, such as uplift of the arch between the Ghadames and Murzuq basins (Boote et al., 1998). During the Devonian the basins were filled with shallow marine sandstones and siltstones, and fluvial sandstones, corresponding to fluctuations in sea-level. The transgressions and regressions, as well as prograding systems, can be correlated between basins (Bellini and Massa, 1980). The southern basins (i.e. Al Kufrah) show a greater continental influence compared with the basins further to the north (i.e. Ghadames) (Boote et al., 1998).

Deposition at the start of the Carboniferous was similar to the Devonian, with variations in siliciclastic sediments controlled by sea-level changes, but with some carbonates deposited locally in the Ghadames Basin (Fekirine and Abdallah, 1998). The Palaeozoic sediments were terminated by the Late Carboniferous-Permian Hercynian orogeny, which was an important period of compression, causing uplift along the flanks of many of the basins and forming much of the Palaeozoic subcrop pattern seen in Fig 2. In some areas, such as beneath the Sirte Basin, it eroded down to the Cambrian sediments (Abadi et al., 2008). Mesozoic and Cenozoic tectonics related to the opening and closing of Tethys to the north of North Africa, and the opening of the Atlantic Ocean, modified some of the basin geometries and subsidence patterns. The Sirte Basin began forming as a rift basin during the late Cretaceous (100-70 Myrs), and continued to the Early Eocene. Extension was due to far-field stresses related to changes in the spreading rate of the Central and South Atlantic (Guiraud et al., 2005; Janssen et al., 1995). However, this has recently been challenged by Capitanio et al. (2009) who suggest this does not fit the timing of extension, and that a better explanation is that the African plate was placed under tensile stress by avalanching of its subducted northern edge through the 660 km discontinuity, beneath the

Hellenic arc. This basin borders the Murzuq and Al Kufrah basins, but was formed later, and so is not analogous to the basins specifically considered in this study. There is little evidence of similar extension in the surrounding basins at this period as shown by the thickness of the Cretaceous deposits seen in the wells from Fig. 4. Similar Palaeozoic successions are seen in the Reggane, Ahnet and Tindouf basins to the north of the West African Craton (Boote et al., 1998). The major event affecting the Paleozoic basins is the Alpine unconformity caused by uplift related to collision of Africa and Europe at the end of the Eocene. See Brunet and Cloetingh (2003) and papers therein for descriptions of the Carboniferous-Recent evolution of basins within northern North Africa and Arabia. These events uplifted the basin margins, and in some areas obscured the earlier subsidence by removing significant thicknesses of sediments. Much of the erosion is focused on the flanks of the basins, although some does occur within the basins, as evidenced by missing Carboniferous section in the well A1 from Al Kufrah (Fig 3). We have sought to minimise the effects of erosion as outlined in our methodology. Details of the stratigraphy for the three wells which were backstripped are shown in Fig. 3.

Figure 3 should be placed here

3.3 Results

Fig. 4 shows the tectonic subsidence profiles derived from the backstripping. The profiles start at the beginning of the Cambrian, but the precise time of the onset of subsidence is poorly constrained because the sediments are fluvial sandstones, with no fossils. The subsidence rates are fairly rapid to begin with and slow gradually through time with little or no tectonic subsidence during the Mesozoic and Cenozoic (fastest rates are 22.2 m/Myr in Ghadames and

10.1 m/Myr in Al Kufrah). The subsidence profiles are steadily decaying curves without clearly separate rift and thermal subsidence phases, i.e. they resemble the thermal subsidence that typically follows rifting, without a rift phase being apparent. The subsidence curves show very little tectonic subsidence affected the basins after 250 Myrs, showing that later periods of subsidence, such as that in the Sirte basin, only slightly modified the Ghadames and Al Kufrah basins. The tectonic subsidence is not smooth; the curves show the periods of erosion and uplift evident in the sedimentary record, e.g. Late Palaeozoic, Hercynian, uplift.

Figure 4 should be placed here

Both basins show very similar tectonic subsidence patterns, both in the timing and the amount of the tectonic subsidence. This is evidence that one causal mechanism generated the Palaeozoic subsidence across North Africa. The total tectonic subsidence in the Ghadames composite well is ~2230 m. This is similar to the ~1740 m of subsidence calculated for the B1NC43 well in the Al Kufrah Basin. As Fig. 4 shows, well A1NC43 appears to have had less subsidence (~1260 m). However much of the Carboniferous stratigraphy has been removed by the Hercynian deformation, so this lower total is plausibly an effect of erosion rather than differential subsidence.

The uncertainties involved come from the input parameters and the assumptions. These include the thickness and depths of the sedimentary layers from errors in picking horizons from well logs or due to erosion. Estimates of these are not provided by Grignani et al. (1991). Assuming that compaction is entirely mechanical also introduces uncertainties. The errors from using the general compaction curves were estimated by carrying out the backstripping with the entire sedimentary column being made of either shale or sandstones which form the two end members of the compaction curves (Sclater and Christie, 1980). Depositional water depth can contribute to large uncertainties because it is difficult to tell the depth of sediments deposited in deep water. As shown in section 3.2, in the basins in North Africa this is not such an issue because all sediments were deposited in shallow waters less than 200 m (Grignani et al., 1991). The uncertainties in water depth estimates for shallow water sediments are much smaller (10s of meters). There are also errors related to corrections for the eustatic sea-level. The largest sea-level fluctuations do not exceed 200 m, so the uncertainties in water depths and eustatic sea-level will have little effect on the overall shape of the curves. The errors from the backstripping calculations are shown as the dashed lines in Fig. 4. The errors make a negligible difference to both the shape of the curves and the overall amount of subsidence.

4. Forward Modelling

Numerical forward modelling of basement subsidence was used to test whether lithospheric cooling and growth is a realistic mechanism for the subsidence of accretionary crust. This is compared to the backstripped results from basins in North Africa to determine if it fits the observed subsidence. The modelling was carried out using Matlab and the code was tested against analytical solutions.

4.1 Methodology of the forward modelling

The numerical subsidence model is based on and tested against the plate models for sea floor spreading (Parsons and Sclater, 1977; Stein and Stein, 1992). It is modified to include layered continental crust with radioactive heat production. It solves the vertical conductive heat flow through a one-dimensional column of the lithosphere and upper mantle.

$$\rho C_p \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}(k\frac{\partial T}{\partial z}) + A \qquad (1)$$

Equation 1 is solved using a finite difference technique with a grid resolution of 1 km. *T* is the temperature of the rock at a particular point in the grid at depth *z*. *A* is the contribution of radioactive heat production and *t* is the time over which the temperature is changing. Time-stepping is performed with an Euler forward time-integration scheme. The material properties of the rock are the thermal conductivity (*k*), the specific heat capacity (C_p) and the density (ρ). The density of the rock is dependant on the temperature and is calculated using equation 2.

$$\rho = \rho_0(1-\alpha(T-T_0)) \qquad (2)$$

In equation 2 the reference density (ρ_0) and the coefficient of thermal expansion (α) are dependant on the rock type (Turcotte and Schubert, 2002). The values used for all the parameters in equations 1 and 2 are shown in Fig. 5.

Figure 5 should be placed here

The model is set up with a two layer crust composed of a felsic upper crust and granulitic lower crust which is underlain by mantle lithosphere. The base of the lithosphere is purely thermal rather than compositional and describes the temperature below which the mantle rock does not deform significantly over geological timescales. The 1200 °C isotherm is used (Turcotte and Schubert, 2002). The model starts with a 20 km thick mantle lithosphere, which matches the starting conditions described in section 2. The temperature at the surface of the model is set as 0°C. The temperature at the base of the model is calculated using a potential temperature at the surface using an adiabatic gradient of 0.3°C per km. Constant temperature boundary conditions are used at the top and bottom of the model. The initial temperature conditions follow the mantle adiabat to up to a transition point 20 km below the crust above which they follow a linear gradient to the surface. The depth and therefore temperature at this transition point is dependent on the thickness of the crust and the potential temperature for the surface. The initial conditions for the model are shown in Fig. 5. The temperature profile is used to calculate the density at grid point in the column from the reference density using equation 2. The elevation is calibrated from the density profile of the column using a column of hot mid ocean ridge material with a 7 km thick basaltic-gabbroic crust at 3 km below sea-level as reference. Standard Pratt isostacy (Allen and Allen, 2005) is used. When the topography of the model drops below sea level the basin is filled with water. The thermal boundary condition is applied to the basement floor because water in the basin would have an almost uniform temperature due to convection. However, the water is included in the isostacy so the model effectively produces water loaded tectonic subsidence.

One inherent weakness of the plate model is that it only calculates the heat transfer by conduction (Parsons and Sclater, 1977). At the base of the plate heat transfer changes from conduction to convection (Huang et al., 2003; Richter and Parsons, 1975; van Hunen et al., 2005). The plate model does not explicitly describe the physics of this transition, but provides a very good fit to the bathymetry and heatflow data (Huang and Zhong, 2005) and included references. It does not help explain the physical process that causes the thermal boundary layer to become stable at this depth. We ignore the temperature dependence of k, C_p and α. This assumption slightly overestimates temperatures in the top half of the model and underestimates temperatures in the bottom half of the model (McKenzie et al., 2005). This means the depth to the base of the lithosphere is an upper limit.

4.2 Results: Effects of varying parameters and the best fit model for the backstripping

The model is set up so that the each of the parameters from equation 1 can be changed and the structure of the model, shown in Fig. 5, can be altered. Changing these will alter the subsidence produced. It is important to understand how these different parameters affect the subsidence before trying to find the best fitting model because there may be a trade off between different parameters.

The model is most sensitive to changes in the thickness of the crust and the plate thickness because they change the isostacy by varying the amount of low density crust and high density lithosphere. However, changing the plate thickness affects the timing of the subsidence whereas changing the crustal thickness only affects the total subsidence so their effects can be distinguished. There is not a direct trade-off between the two parameters. The model is less sensitive to variations in the potential temperature and radioactive heat production respectively (Fig. 6). In each case only the parameter under investigation is changed and the estimates for North African crust shown in Fig. 5 are used as the default the parameters. The amount of variation in the input parameters is based on the uncertainties and variation across North Africa from the literature. The present day thickness of the crust of North Africa was calculated to be between 30 and 40 km thick from low resolution gravity and seismic interpretations and gravity modelling in the 3-D crustal model of Seber et al. (2001) over the area of interest. This is thicker than the crustal thickness of 25-35 km estimated from the inversion of surface waves given by Pasyanos and Nyblade (2007). Due to the uncertainty in crustal thickness our model calculations were run for a range of crustal assemblages between 20 and 40 km thick. The results are shown in Fig. 6a. With a crustal thickness of 40 km, ~ 1.2 km of tectonic subsidence takes place, however the crust is too buoyant to drop below sea-level, so no basin is formed. When the crust is reduced to 36 km thick the total subsidence increases to ~1.9 km and 1 km of water loaded tectonic subsidence is recorded. A 20 km thick crust starts at 1.4 km below sea-level and experiences ~2.7 km of tectonic subsidence as the lithosphere cools and thickens. There have been no deep crustal seismic lines of North Africa published so the proportion of upper crust and lower crust is unknown. Therefore the model was run with a 30 km thick crust where the thickness of the upper crust was varied between 10 and 20 km. This alters the total subsidence by about 500 m. A 1 km decrease in the total crustal thickness causes ~150 m of additional subsidence, whereas changing 1 km of upper crust to lower crust only increases the total subsidence by ~ 50 m. The main reason both these parameters affect the overall subsidence is that they change the isostacy by varying the amount of low density crust. In all these scenarios the timing of the subsidence does not change, most of the subsidence occurs in the first 100 Myrs and the subsidence after 200 Myrs is negligible.

Figure 6 should be placed roughly here in the text

We also change the amount of radiogenic heat production. Estimations of heat production in the continental crust vary between 1.31 mWm^{-3} and 0.8 mWm^{-3} for the upper crust and 1.0 mWm^{-3} and 0.6 mWm^{-3} for the lower crust. The values reflect calculations of bulk crustal heat production from the literature (Christensen and Mooney, 1995; Gao et al., 1998; Shaw et al., 1986; Wedepohl, 1995). Changing the heat production for the upper and lower crust between these extremes causes a small variation of 100 m in the tectonic subsidence. The heat production of the bulk continental crust will be lower than that of accretionary crust and heat production would have been higher during the Palaeozoic, so the model was also run with a higher heat production of 2.0 mWm^{-3} (upper crust) and 1.2 mWm^{-3} (lower crust). This does reduce the overall amount of tectonic subsidence from ~ 1800 to 1550m, but does not alter the timing. Fig.

6b shows that varying the heat production in the crust has a small effect compared to varying the crustal thickness.

The other main parameter which has a big effect on the subsidence is the thickness of the model, referred to as the plate thickness (Fig. 6c). A plate thickness of 95 km, which is the preferred for the oceanic lithospheric cooling model of Stein and Stein (1992) results in 200 m of subsidence, of which 80% has occurred within ~70 Myrs, with a final lithospheric thickness (i.e. depth to the 1200 °C isotherm) of 81.1 km. At the other extreme when a plate thickness of 200 km is used, similar to an Archean craton (Priestley and McKenzie, 2006), ~3050 m of subsidence is produced, and it takes longer for the subsidence to tail off. 80% of the subsidence is completed within 200 Myrs. The lithosphere thickness after 500 Myrs is 166.5 km. As the plate thickness is increased the final lithospheric thickness also increases. A thicker, dense lithosphere results in more subsidence, occurring over a longer time period.

The temperature at the base of the plate is controlled by both the potential temperature at the surface (the temperature at which the mantle adiabat dissects the surface) and the thickness of the plate (Fig. 5). Increasing the potential temperature by 1°C increases the temperature at the base of the plate by 1°C whereas increasing the thickness of the model causes the temperature at the base of the plate to increase along the mantle adiabat i.e. 3°C for every 10 km. The potential temperature was varied from 1280 °C, which was suggested by McKenzie et al. (2005) to 1410 °C suggested by Parsons and Sclater (1977). As Fig. 6d shows this variation of 100 °C in the potential temperature produces less than 500 m difference in the overall subsidence.

Van Wees et al. (2009) report a trade off between parameters, such as the final lithospheric thickness (plate thickness) and crustal thickness, when investigating the uncertainties in surface heatflow using modelling. However, in this study we find that varying the plate thickness to fit the time span of the subsidence and then varying the crustal thickness to fit the amount of subsidence allows a unique solution to be obtained for these parameters. However, there is a direct trade off between the crustal thickness, heat production and potential temperature because they all only affect the total subsidence. The model is not sensitive to reasonable variations in the heat production or potential temperature, but is sensitive to variations in crustal thickness suggested for North Africa. Therefore the trade off between these parameters is not important when trying to fit the modelled subsidence to the observed subsidence.

The sensitivity of the model to k, C_p and α was also investigated, but because much less is known about the amount of variation that is found in the crust they are not discussed here and the standard values from the literature (Fig. 5) are used. The forward model produces significant subsidence for a large variety of crustal assemblages and for wide variation in the parameters. It supports the hypothesis that lithospheric cooling and thickening caused the subsidence seen across North Africa.

5. Discussion

5.1 Comparison of tectonic subsidence from backstripping and forward modelling.

The results of the forward modelling show a good fit to the backstripping from both basins as demonstrated by Fig. 7. Four forward model runs are shown on the diagram. One of which uses the best estimates for the crustal assemblage, thickness and plate thickness from the literature, one which is the best fit to the backstripping and then two runs which show the maximum and minimum subsidence for reasonable variations of the properties of the North African crust.

The forward model curves are smooth unlike the backstripped subsidence curves. This is because North African crust has experienced periods of uplift, and the forward model doesn't include these complexities. However, the forward model fits the broad shape of the backstripped subsidence curve very well. Both have a concave up shape, and the timescale and magnitude of subsidence are very similar. When the crustal thickness, final lithosphere thickness and heat production used are taken from the best estimates of the North African crust in the literature (Fig. 5), and therefore are independent of the backstripping, the forward model curve only underestimates the subsidence slightly, by <500 m in 350 Myrs. The discrepancy at 500 Myrs is ~700 m, because of the Mesozoic and Cenozoic subsidence of the Ghadames Basin. This is a separate tectonic event, related to Tethyan opening and evolution (Guiraud and Bosworth, 1999). The best fit model has been fitted iteratively by eye. The input parameters for the best fit model are the same as those from the literature except that the plate is 155 km thick, which results in a final lithospheric thickness of ~124 km. The crust beneath the basins cannot be assumed to be homogenous, in fact because it is accreted from numerous crustal fragments it is likely that it varies across North Africa. The maximum and minimum subsidence based on the variation found in the literature, discussed in section 4.2, is shown in Fig. 7. The upper limit has a 35 km thick crust with 30 km of upper crust and 5 km of lower crust, a heat production of 1.6 mWm^{-3} in the upper crust and 1.0 mWm^{-3} in the lower crust, a potential temperature of 1280 °C and plate thickness of 120 km. The lower limit has 25 km thick crust, where the top 5 km is upper crust and the rest is lower crust, a heat production of 1.0 mWm^{-3} in the upper crust and 0.6 mWm^{-3} in the lower crust, a potential temperature of 1380 °C and a plate thickness of 180 km.

Figure 7 should be placed here

There are a number of assumptions in both the modelling and backstripping. Both are oversimplified versions of the real situations and processes. The problem has been simplified to be one dimensional, whereas the basins are three dimensional features. We assume that the basins are large enough features that it is possible to remove effects of marginal uplift from the sedimentary record by choosing wells away from the basin flanks, or constructing composite wells. This allows the basins to be treated as one dimensional features. Another assumption is that North African crust resembles the starting conditions outlined in the hypothesis. This is difficult to test as there is no newly assembled accretionary orogenic belt on the North African scale that can be used as an analogue. However, the building blocks of accretionary crust are available to study. It seems reasonable that North African crust formed through subduction would share the thin lithosphere of its component island arcs and accretionary wedges. When North African crust is modelled using these starting conditions it produces a good fit to the subsidence. As mentioned in the hypothesis as long as the crust is of near normal thickness and has a thin mantle lithosphere these results still apply. The two North African basins have been chosen as case studies, but we propose that the lithosphere cooling and thickening hypothesis is applicable to all the Palaeozoic basins on the accretionary crust of North Africa and Arabia, and to other basins on accretionary crust around the world, such as the Mesozoic cover of the Scythian and Turan platforms of SW and Central Asia (Natal'in and Şengör, 2005), the Mesozoic cover of the Tasmanides of Eastern Australia (Gallagher et al., 1994) and the late Palaeozoic platform cover over the Altaid orogenic collage in Central Asia (Cook et al., 1994).

The good fit between modelled and backstripped subsidence curves do not prove that lithosphere thickening and cooling is the cause of the subsidence. Other authors have suggested that the Palaeozoic basins in North Africa formed by rifting (Lüning et al., 1999), orogenic collapse (Ashwal and Burke, 1989) and dynamic topography (Heine et al., 2008). Lüning et al. (1999), compared the Al Kufrah Basin to pull apart basins in Oman and Saudi Arabia, and suggested that the Najd fault system may extend north into the Al Kufrah Basin and be reactivated as a rift during the Precambrian. Seismic lines of the south of the Al Kufrah Basin do indeed have features which have been interpreted as rifts. However, they are only found in a small area of the basin and the synrift subsidence accounts for a sixth of the overall subsidence (Ghanoush and Abubaker, 2007). The Ghadames Basin has very similar stratigraphy to Al Kufrah, but shows no evidence of rifting. Rifting is not widespread or large enough to cause the subsidence across all the Palaeozoic basins in North Africa. This also means that orogenic collapse is an unlikely subsidence mechanism, given that this is a special case of rifting on thickened continental crust. Armitage and Allen (2010) suggest that these basins can still form due to extension, but when stretching is occurring at low strain rates the strain might not localise along faults. It is hard to distinguish between their model and the model proposed in this paper on the grounds of rifting. Subsidence due to extension at low strain rates produces a period of constant subsidence during stretching followed by decreasing subsidence during the thermal sag phase. Whereas cooling of a thin lithosphere produces subsidence which decreases smoothly throughout time. The latter model fits the backstripping results (Fig. 4) which do not have a kink, but are be closer to a smooth curve.

There are problems with using dynamic topography to explain the basins: it has a transient effect and will rebound once the down welling has ceased leading to uplift and erosion. Nor is it possible at present to model whether the basins in North Africa would have been affected by mantle down-welling during the Palaeozoic. Thermal subsidence on the other hand explains the lateral extent of the subsidence as the lithosphere would be thin across most, if not all, of the accretionary crust. Our results show that it is long lasting, fits the magnitude of the observed subsidence and is able to cause subsidence over a wide range of terranes.

Thermal subsidence has a number of important implications. It suggests that one of the fundamental properties of accretionary crust is that, in the absence of other competing tectonic forces, it will subside after its formation. This will alter the overall composition of the crust, adding several kilometres of sediments to crust made up largely of island arc and oceanic material. It also predicts that the lithosphere beneath accretionary crust is not depleted by the melting events which produced the crust, but is compositionally no different to upper mantle. This will make the lithosphere more susceptible to thermal erosion as it does not have compositional buoyancy and also affect the character of melts originating in it or passing through it. Ashwal and Burke (1989) also noted that this fits the geophysical measurements of the thickness and seismic properties of the lithosphere and with the geochemistry of Cenozoic volcanism in North Africa.

6. Conclusions

We have shown that results from backstripping wells in two North African basins and numerical modelling, are consistent with a basin forming mechanism of lithospheric cooling and thickening, underneath relatively juvenile, accretionary crust. The backstripping revealed that the Ghadames and Al Kufrah basins have experienced their highest rates of subsidence, (22.2 m/Myr and 10.1 m/Myr, respectively) at the start of the Palaeozoic, declining to almost zero at the end of the Palaeozoic. There is almost no tectonic subsidence during the Mesozoic and Cenozoic. The forward modelling shows that thermal subsidence of accretionary crust provides a viable explanation for the formation of the Palaeozoic basins of North Africa, fitting the available data better than previously suggested mechanisms.

We suggest that thermal subsidence following accretion would be expected in other areas of accretionary crust, and could have influenced the rifted West Siberian Basin. It also predicts that accretionary crust is underlain by fertile mantle lithosphere which may explain why Cenozoic volcanism in North Africa and Arabia seems to originate from a fertile source (Ashwal and Burke, 1989).

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Figure Captions

Fig. 1. Formation of accretionary crust and subsequent lithospheric thickening and subsidence. a) Assembly of accretionary crust from island arcs and other crustal fragments through subduction. b) Newly formed accretionary crust with a thinned lithosphere. c) Lithospheric thickening due to cooling causing subsidence.

Fig. 2. Present outcrop and sub-Mesozoic subcrop of Palaeozoic sediments across North Africa and Arabia (Boote et al., 1998; Konert et al., 2001). Four wells used to construct the Ghadames composite well are located along the line of the transect shown.

Fig. 3. Stratigraphy of the Ghadames and Al Kufrah basins used for backstripping analysis. The Ghadames composite well is constructed from four wells located along the transect shown in Fig. 2, to give the most complete sedimentary record. The locations of the wells from Al Kufrah are also marked on Fig 2.

Fig. 4. Tectonic subsidence curves for the Ghadames and Al Kufrah basins. The solid lines are the tectonic subsidence and the dotted lines show the subsidence if the stratigraphy is entirely shale or sandstone. This envelope is the maximum possible range of variation in the subsidence possible from varying the proportions of the lithologies.

Fig. 5. The right-hand side shows structure and initial temperature conditions for the numerical forward model. The solid line is the preferred temperature profile while the dashed line indicates the potential temperature. The left hand side shows the best estimates for the material properties for the different layers of crust and mantle from the literature. (Allen and Allen, 2005; Shaw et al., 1986; Turcotte and Schubert, 2002)

Fig. 6. The sensitivity of the forward model to variations **(a)** in the crustal thickness and proportion of lower crust (LC) and upper crust (UC), **(b)** the heat production, **(c)** the plate thickness and **(d)** the potential surface temperature of the mantle adiabat.

Fig. 7. Comparison of the subsidence from backstripping the composite well from the Ghadames Basin with a range of forward model runs. The hypothesised subsidence mechanism fits the observed subsidence well although there is a wide range of subsidence over the possible range of variations in the North African crust as demonstrated by the upper and lower limits.

Table Captions

Table 1 Compilation of intracratonic basins situated on accretionary crust from around the world. The table shows their locations and the temporal link between the end of the accretion event forming the underlying crust and the beginning of subsidence across the basin as a whole.

***Research Highlights**

- Thermal subsidence is viable for forming basins on accretionary crust.
- Accretionary crust starts with a thin lithosphere due to formation by subduction.
- Backstripping curves from North African basins suggest thermal subsidence.
- Forward modelling of thermal subsidence is a good match for the backstripping.
- This may also form basins in South America, Central Asia and Eastern Australia.

Lithospheric cooling and thickening as a basin forming mechanism

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ABSTRACT

Widely accepted basin forming mechanisms are limited to flexure of the lithosphere, and lithospheric stretching followed by cooling and thermal subsidence. Neither of these mechanisms works for a group of large basins, sometimes known as "intracontinental sags". In this paper we investigate cooling and thickening of initially thin lithosphere as a basin forming mechanism, by a combination of forward modelling and a backstripping study of two Palaeozoic North African basins: Ghadames and Al Kufrah. These are two of a family of basins, once unified, which lie over the largely accretionary crust of North Africa and Arabia. Such accretionary crust tends to be juvenile, consisting of amalgamated island arcs, accretionary prisms and melanges, and typically has near-normal crustal thicknesses but initially thin mantle lithosphere. Post-accretion subsidence is modelled using a plate cooling model similar to cooling models for oceanic lithosphere. The crustal composition and thickness used in the models are varied around average values of accretionary crust to represent likely heterogeneity. The model allows the lithosphere to thicken as it cools and calculates the resulting isostatic subsidence.

Water-loaded tectonic subsidence curves from these forward models are compared to tectonic subsidence curves produced from backstripped wells from Al Kufrah and Ghadames basins. A good match between the subsidence curves for the forward model and backstripping is produced when the best estimates for the crustal structure, composition and the present day thickness of the lithosphere for North Africa are used as inputs for the forward model. The model produces sediment loaded basins of 2-7 km thickness for the various crustal assemblies over ~250 Myrs. This shows that lithospheric cooling provides a viable method for producing large basins with prolonged subsidence, without the need for initial extension, provided the condition of initially thin mantle lithosphere is met.

Keywords: basin, lithosphere, subsidence, North Africa

1. Introduction

Conventional basin formation mechanisms can be divided into two categories, lithospheric stretching followed by thermal subsidence proportional to the extension (rift basins) and flexure caused by tectonic loading (foreland basins) (Allen and Allen, 2005). However, there is scant evidence for either of these mechanisms forming a group of basins normally classified as "intracontinental sags" – a term that describes their geometry rather than the process of formation. Examples include the Williston and Michigan basins of North America (Klein, 1995), the Palaeozoic basins of North African and Arabia (Boote et al., 1998; Konert et al., 2001); and the Mesozoic Scythian and Turan platforms (Natal'in and Şengör, 2005). The basins are large features, commonly over 1000 km in length, with remarkably uniform, prolonged, gentle subsidence across them, lasting over 200 Myrs. They generally have a polyphase history with a main subsidence phase either preceded or followed by other periods of subsidence and uplift which modify the basin. Armitage and Allen, (2010) recently proposed that these basins are formed by stretching under low strain rates. They argued this based upon the modelling of rifting under low strain rates and the observation that the initiation of subsidence in many intracontinental basins coincides with supercontinent breakup and therefore a broad extensional regime. However, in many basins the evidence for rifting is poor and a number of other mechanisms have been proposed. Subsidence due to cooling of thermal anomalies in the lithosphere has been proposed for the North American intracontinental basins (Kaminski and Jaupart, 2000). The main evidence for this is matching modelling results with the shape and thickness of the present day sedimentary cover. A density change in the crust due to phase changes such as a basalt underplate changing to eclogite has also been suggested based on high velocities in the lower crust interpreted from seismic refraction data (Artyushukov, 2005). However, beneath the Barents/Kara sea region a high density area in the lower crust, suggested by modelling gravity data, has been deemed too local to cause subsidence across the basin. Instead Ritzmann and Faleide, (2009) have suggested that a deeper high velocity zone visible in seismic tomography is evidence of a thick cratonic lithosphere, which causes the subsidence. Heine et al. (2008) noted that many intracratonic basins overlie areas of mantle which have been down welling over the last 100-150 Myrs in their coupled plate and mantle flow model. They proposed that dynamic topography could form these basins. Other subsidence mechanisms and variations on those above have been suggested and are debated in more detail by Armitage and

Allen (2010) and Klein (1995). It is likely that one mechanism does not explain the formation of every intercontinental basin and in some cases the basin may be formed by a combination of mechanisms.

Here we show that cooling and thickening of initially thin mantle lithosphere, beneath crust of normal thickness (~30 km) is a viable mechanism for producing basin-scale subsidence. Such initial conditions are typical of accretionary crust, a term used to summarise the vast orogenic collages of largely juvenile crust and mantle lithosphere, formed by the collision of non-cratonic terranes: island arcs, accretionary prisms, ophiolites and isolated microcontinents (Murphy and Nance, 1991; Şengör et al., 1993). This would neatly explain the formation of many of the intracratonic basins on juvenile continental crust such as the Pan African mobile belt or the Scythian and Turan platform. However, we show that where the lithosphere is thinned by a thermal anomaly it is also possible to form broad, slowly subsiding basins.

This subsidence mechanism is discussed in greater detail in section 2 below, followed by a case study of two of the North African Palaeozoic basins. Thermal subsidence has been suggested as the cause of intracratonic basins before (Guiraud et al., 2005; Kaminski and Jaupart, 2000; Kominz, 1995). Our contribution is to model the subsidence, compare it to subsidence in two case studies and to discuss why the lithosphere is plausibly thin in the first place. The subsidence history of the basins is analysed using backstripping. This analysis is compared to results from a numerical forward model of thermal subsidence acting on accretionary crust, designed to test if it is a mechanism capable of producing the observed subsidence.

2. Geological background and hypothesis

Seismic refraction studies show that present day island arcs can have a crustal thickness of 25-35 km, similar to normal continental crust (Holbrook et al., 1999; Takahashi et al., 2007). However, seismic tomography shows slow velocities in the mantle wedge below island arcs which are interpreted as evidence for the presence of melts and thin (~20 km) mantle lithosphere because it is weakened by the addition of fluids from the subducting slab and then eroded by the corner flow in the mantle wedge (Gorbatov et al., 1999; Zhao et al., 1994). This is supported by numerical models of subduction (Arcay et al., 2006; Stern, 2002; van Keken, 2003), by geochemical evidence from the southward initiation of the Philippine subduction zone (Macpherson, 2008) and from the Cascades (Elkins Tanton et al., 2009). These studies suggest an average overall lithospheric thickness of about 50 km beneath island arcs (Fig. 1a).

Accretionary prisms may be 30 km thick, largely composed of off-scraped and imbricated fragments of oceanic crust and its sedimentary cover. Whilst subduction is active such prisms are underlain by the oceanic plate. When subduction has recycled the oceanic plate in to the mantle, the base of the prism may be in contact with the asthenosphere, particularly if ocean closure resulted in the collision of two such prisms, initially on opposite sides of the ocean, rather than collision of the prism with a continental margin. The Cenozoic East Anatolian Accretionary Complex may be an example of such a lithospheric structure, where tomographic studies suggest a thin or even absent mantle lithosphere (Zor et al., 2003).

A notable feature of accretionary orogenic belts is that they lack evidence for substantial crustal thickening (and presumably lithosphere thickening): there is rarely evidence for pre-collision passive continental margins, Alpine-type nappes, or overfilled foreland basins (“molasse”) (Şengör and Okurogullari, 1991). This means that putative lithospheric delamination following an orogeny of this type, hypothesised by Ashwal and Burke, (1989), is not our preferred mechanism for thinning the lithosphere. However, it would produce similar starting conditions to those in our model.

As accretionary crust is assembled through subduction and collision, the thin mantle lithosphere of the original terranes is inherited by the final collage (Fig. 1b). We hypothesise that once accretion is completed, and subduction has ceased beneath an area, the underlying asthenosphere will cool, thickening the mantle lithosphere. This cooling will cause prolonged subsidence, forming basins (Fig. 1c).

Figure 1 should be placed here

Our model is similar to the thermal subsidence phase of McKenzie style rifting (McKenzie, 1978) or the subsidence of the ocean floor away from a mid ocean ridge, except that the crust involved is continental, albeit juvenile, and has not been thinned in any way. In this paper we proceed to show how this mechanism could produce the basins in North Africa. However, there are many other intracratonic basins on accreted crust where this mechanism could apply. Table 1 provides a sample of some of the basins we aware of, but is by no means an exhaustive list. Allen and Armitage, (in press) note a clustering in time of the initiation of intracratonic basins which they link to the breakup of supercontinents. In Table 1 we show the start of subsidence follows closely the end of accretion and the clustering may be related to the end of periods of accretion of crust. The basins are long lived features and so many have later phases of subsidence which potentially have other causes.

Table 1 should be placed here.

3. Tectonic subsidence history of backstripped North African basins

In order to test whether the proposed mechanism of subsidence provides a good explanation for anomalous basins developed over accretionary crust, subsidence histories for the Ghadames and Al Kufrah basins were investigated using backstripping. These are Palaeozoic basins situated on the North African crust, which was accreted in the Pan African orogeny during Neoproterozoic times (Caby, 2003; Stern, 1994). Most of the basement to North Africa and Arabia is juvenile, generated and assembled during the Pan African orogeny. The western margin is the West African Craton. The southern limit is the Congo Craton. The eastern and northern limits are not so well defined because of later rifting and collision with Eurasia. For example, similar basement underlies much of the territory of Iran, but with a more complicated Mesozoic and Cenozoic magmatic and tectonic history. The outcrop or sub-Mesozoic subcrop of the Palaeozoic strata of North Africa and Arabia is shown in Fig. 2.

Figure 2 should be placed here

The proportion of cratonic nuclei within this vast, 10,000,000 km^2 orogen is debated, but plausibly is small. There are indications of pre Late Proterozoic crust within North Africa, based on Meso- or Palaeo-Proterozoic isotopic model ages and detritial zircons in younger metamorphic terranes (Black et al., 1993; Sultan et al., 1990). But the Late Proterozoic tectonic overprint is severe, suggesting that extensive magmatic and metamorphic re-working and additions took place during the Pan African orogeny, and no single, regionally extensive block survived the Late Proterozoic orogeny with its original structure and boundaries preserved. A significant re-worked region has been called the Saharan metacraton (Abdelsalam et al., 2002), but this is not recognised by all workers e.g. (Bumby and Guiraud, 2005). Whatever the nature and origins of the Saharan metacraton, it appears to have had little impact on the overlying

Phanerozoic basins (Fig. 2). These are continuous across the metacraton margins, without discernible change in sedimentary thickness or composition.

The end of the Pan African orogeny was diachronous. Local timings for the last deformation vary from Late Precambrian to the Early Cambrian e.g. (Paquette et al., 1998). Present-day lithosphere thickness across North Africa is on the order of ~100 km (Pasyanos and Nyblade, 2007; Priestley and McKenzie, 2006); except for the West African Craton where it reaches >200 km. Crustal thicknesses are not well-constrained, but are variously estimated at 30 to 40 km thick from gravity and seismic interpretations (Seber et al., 2001), or 25 to 35 km from the inversion of surface waves (Pasyanos and Nyblade, 2007).

The Ghadames Basin is well studied because large hydrocarbon accumulations have been discovered within it (Echikh, 1998). Several wells have penetrated the crystalline basement and there are large amounts of seismic data for the basin. This means the geometry of the basin is well understood. This allows the most complete stratigraphic sections to be identified, which is helpful when backstripping. The Al Kufrah Basin is less well studied and has only two published wells which reached the crystalline basement (Grignani et al., 1991). To a first order, it has a similar Palaeozoic stratigraphy to the Ghadames Basin. Enough data are available for Al Kufrah to make it viable for modelling.

3.1 Methodology

Backstripping is a well known technique for calculating the tectonic subsidence of a particular horizon in a basin over time. We use the basement/cover boundary, which reveals the overall subsidence (Allen and Allen, 2005; Sclater and Christie, 1980). The technique first decompacts each sedimentary layer through time. This gives the total subsidence curve for the basement which is assumed to be composed of the tectonic subsidence, the isostatic effect (weight) of the sediments and the changes in water depth compared to the present day sea level. The effects of the sediments and sea level changes are removed so that the remaining subsidence is due purely to the tectonic driving force. A global eustatic sea-level curve of (Haq and Schutter, 2008) is used to remove the effects of sea-level changes.

The backstripping method assumes that the compaction of the sediments is purely mechanical (due to the weight of the sediments above) and ignores chemical processes, such as cementation, which are very difficult to take into account because they depend on a complex series of factors such as fluid flow, composition, temperature and pressure. Backstripping assumes that the sediments are laid down in successive layers throughout time and does not take into account periods of uplift and erosion or non deposition. These are difficult to include in backstripping because generally the amount of eroded material is poorly constrained. If the present day burial depth of the basement is the deepest it has been, then any erosion will make no difference because the sediments are at their peak pressure. Otherwise the compaction of the sediments will be underestimated. There are no porosity depth relationships available for the sediments in North Africa and we use standard relationships from published work (Sclater and Christie, 1980). We assume that the compaction curves, which are based on sediments from the North Sea, are applicable to the sediments in North Africa. This is justifiable because the North African sediments are entirely siliciclastic and are similar to those in the North Sea.

This method was applied to a composite well from the Ghadames Basin. A composite well was used because it gives the most complete section possible from the basin therefore showing the maximum subsidence and limiting the effect of erosion. The composite well was created using four unpublished wells from the transect shown on Fig. 2, to identify the most complete sections and to ensure that the differences in the thickness of the layers were due to variations in erosion rather than deposition rates. This minimises the errors related to eroded sections of the stratigraphy. It also produces a subsidence curve which emphasises the subsidence phases rather than any uplift. This allows us to see clearly the main subsidence phase associated with basin formation, rather than later phases of uplift or subsidence. This approach was not possible in the Al Kufrah Basin because of the scarcity of available well data, and instead the backstripping methodology was applied directly to the two available wells.

3.2 General Stratigraphy

Detailed descriptions of the stratigraphy in the Ghadames, Al Kufrah and other North African basins can be found in papers such as Bellini and Massa (1980), Echikh (1998), Fekirine and Abdallah (1998), Grignani et al. (1991) and Lüning et al. (1999). Therefore we present a brief summary of the evolution of the Palaeozoic basins on the accretionary crust of North Africa collated from Boote et al. (1998), Bumby and Guiraud (2005), Craig et al. (in press), Guiraud and Bosworth (1999) and Guiraud et al. (2005) alongside more detailed observations from the basins themselves. The basins are filled with a largely siliciclastic succession with some evaporites and carbonates towards the end of the Palaeozoic. The whole of North Africa subsided as a large platform from roughly the start of the Palaeozoic, depositing a wedge of sediments that thinned to the south (Selley, 1997). Only localised Late Proterozoic/Early Palaeozoic rifting is known (see Fig. 6 of Guiraud et al. (2005), and as several of the main basins are mature in terms of hydrocarbon exploration it is unlikely that major rifts have been missed. Seismic lines through the Kufrah Basin show Late Proterozoic/Cambrian rifts (Lüning et al., 1999), however these are only seen on seismic lines in the south of the basin (Ghanoush and Abubaker, 2007) and do not explain subsidence across the whole basin or in neighbouring basins. Nor do the dimensions of the subsiding area (>>1000 km) fit a flexural, foreland basin mechanism. In any case, there is no record of an appropriate orogeny lasting through the Palaeozoic along the Gondwanan continental margin (Stampfli and Borel, 2002). The initial sediments are largely fluvial sandstones and conglomerates in the Cambrian, changing to marine sandstones in the early Ordovician. The Cambrian age for the earlier sediments is inferred because there are very few trace fossils to date the sediments until the marine incursions (Grignani et al., 1991). In the mid to late Ordovician there was major glaciation creating a regional erosional surface (Le Heron et al., 2010). The amount of erosion varies greatly reflecting the geometry of the ice sheets. Locally in Al Kufrah the erosion cuts down to the Cambrian, but in other areas a complete section to the Mid Silurian is preserved. In these areas the erosion is minimal and the glaciation reflects a period of non deposition. The erosion becomes less severe to the north into the Murzuq and Ghadames basins (Bellini and Massa, 1980; Fekirine and Abdallah, 1998).

Coarse to medium grained sandstones and tillites related to the retreat of the glaciers were deposited at the end of the Ordovician. This was followed by a major marine transgression leading to the deposition of shales throughout the Early Silurian, but water depths decreased with time and the shales grade up to shoreface sandstones. The shales at the base of the Silurian are one of the main source rocks and record the maximum water depth experienced by the basins during the Palaeozoic. Turner (1978) interpreted it as a shallow shelf environment with water depths of about 200 m. Separation of individual basins began during the Silurian caused by a period of compression most likely related to the Acadian collision (Guiraud and Bosworth, 1999). This resulted in an unconformity across the basins of North Africa, but within basins it is localised over arches, such as uplift of the arch between the Ghadames and Murzuq basins (Boote et al., 1998). During the Devonian the basins were filled with shallow marine sandstones and siltstones, and fluvial sandstones, corresponding to fluctuations in sea-level. The transgressions and regressions, as well as prograding systems, can be correlated between basins (Bellini and Massa, 1980). The southern basins (i.e. Al Kufrah) show a greater continental influence compared with the basins further to the north (i.e. Ghadames) (Boote et al., 1998).

Deposition at the start of the Carboniferous was similar to the Devonian, with variations in siliciclastic sediments controlled by sea-level changes, but with some carbonates deposited locally in the Ghadames Basin (Fekirine and Abdallah, 1998). The Palaeozoic sediments were terminated by the Late Carboniferous-Permian Hercynian orogeny, which was an important period of compression, causing uplift along the flanks of many of the basins and forming much of the Palaeozoic subcrop pattern seen in Fig 2. In some areas, such as beneath the Sirte Basin, it eroded down to the Cambrian sediments (Abadi et al., 2008). Mesozoic and Cenozoic tectonics related to the opening and closing of Tethys to the north of North Africa, and the opening of the Atlantic Ocean, modified some of the basin geometries and subsidence patterns. The Sirte Basin began forming as a rift basin during the late Cretaceous (100-70 Myrs), and continued to the Early Eocene. Extension was due to far-field stresses related to changes in the spreading rate of the Central and South Atlantic (Guiraud et al., 2005; Janssen et al., 1995). However, this has recently been challenged by Capitanio et al. (2009) who suggest this does not fit the timing of extension, and that a better explanation is that the African plate was placed under tensile stress by avalanching of its subducted northern edge through the 660 km discontinuity, beneath the

Hellenic arc. This basin borders the Murzuq and Al Kufrah basins, but was formed later, and so is not analogous to the basins specifically considered in this study. There is little evidence of similar extension in the surrounding basins at this period as shown by the thickness of the Cretaceous deposits seen in the wells from Fig. 4. Similar Palaeozoic successions are seen in the Reggane, Ahnet and Tindouf basins to the north of the West African Craton (Boote et al., 1998). The major event affecting the Paleozoic basins is the Alpine unconformity caused by uplift related to collision of Africa and Europe at the end of the Eocene. See Brunet and Cloetingh (2003) and papers therein for descriptions of the Carboniferous-Recent evolution of basins within northern North Africa and Arabia. These events uplifted the basin margins, and in some areas obscured the earlier subsidence by removing significant thicknesses of sediments. Much of the erosion is focused on the flanks of the basins, although some does occur within the basins, as evidenced by missing Carboniferous section in the well A1 from Al Kufrah (Fig 3). We have sought to minimise the effects of erosion as outlined in our methodology. Details of the stratigraphy for the three wells which were backstripped are shown in Fig. 3.

Figure 3 should be placed here

3.3 Results

Fig. 4 shows the tectonic subsidence profiles derived from the backstripping. The profiles start at the beginning of the Cambrian, but the precise time of the onset of subsidence is poorly constrained because the sediments are fluvial sandstones, with no fossils. The subsidence rates are fairly rapid to begin with and slow gradually through time with little or no tectonic subsidence during the Mesozoic and Cenozoic (fastest rates are 22.2 m/Myr in Ghadames and

10.1 m/Myr in Al Kufrah). The subsidence profiles are steadily decaying curves without clearly separate rift and thermal subsidence phases, i.e. they resemble the thermal subsidence that typically follows rifting, without a rift phase being apparent. The subsidence curves show very little tectonic subsidence affected the basins after 250 Myrs, showing that later periods of subsidence, such as that in the Sirte basin, only slightly modified the Ghadames and Al Kufrah basins. The tectonic subsidence is not smooth; the curves show the periods of erosion and uplift evident in the sedimentary record, e.g. Late Palaeozoic, Hercynian, uplift.

Figure 4 should be placed here

Both basins show very similar tectonic subsidence patterns, both in the timing and the amount of the tectonic subsidence. This is evidence that one causal mechanism generated the Palaeozoic subsidence across North Africa. The total tectonic subsidence in the Ghadames composite well is ~2230 m. This is similar to the ~1740 m of subsidence calculated for the B1NC43 well in the Al Kufrah Basin. As Fig. 4 shows, well A1NC43 appears to have had less subsidence (~1260 m). However much of the Carboniferous stratigraphy has been removed by the Hercynian deformation, so this lower total is plausibly an effect of erosion rather than differential subsidence.

The uncertainties involved come from the input parameters and the assumptions. These include the thickness and depths of the sedimentary layers from errors in picking horizons from well logs or due to erosion. Estimates of these are not provided by Grignani et al. (1991). Assuming that compaction is entirely mechanical also introduces uncertainties. The errors from using the general compaction curves were estimated by carrying out the backstripping with the entire sedimentary column being made of either shale or sandstones which form the two end members of the compaction curves (Sclater and Christie, 1980). Depositional water depth can contribute to large uncertainties because it is difficult to tell the depth of sediments deposited in deep water. As shown in section 3.2, in the basins in North Africa this is not such an issue because all sediments were deposited in shallow waters less than 200 m (Grignani et al., 1991). The uncertainties in water depth estimates for shallow water sediments are much smaller (10s of meters). There are also errors related to corrections for the eustatic sea-level. The largest sea-level fluctuations do not exceed 200 m, so the uncertainties in water depths and eustatic sea-level will have little effect on the overall shape of the curves. The errors from the backstripping calculations are shown as the dashed lines in Fig. 4. The errors make a negligible difference to both the shape of the curves and the overall amount of subsidence.

4. Forward Modelling

Numerical forward modelling of basement subsidence was used to test whether lithospheric cooling and growth is a realistic mechanism for the subsidence of accretionary crust. This is compared to the backstripped results from basins in North Africa to determine if it fits the observed subsidence. The modelling was carried out using Matlab and the code was tested against analytical solutions.

4.1 Methodology of the forward modelling

The numerical subsidence model is based on and tested against the plate models for sea floor spreading (Parsons and Sclater, 1977; Stein and Stein, 1992). It is modified to include layered continental crust with radioactive heat production. It solves the vertical conductive heat flow through a one-dimensional column of the lithosphere and upper mantle.

$$\rho C_p \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}(k\frac{\partial T}{\partial z}) + A \qquad (1)$$

Equation 1 is solved using a finite difference technique with a grid resolution of 1 km. *T* is the temperature of the rock at a particular point in the grid at depth *z*. *A* is the contribution of radioactive heat production and *t* is the time over which the temperature is changing. Time-stepping is performed with an Euler forward time-integration scheme. The material properties of the rock are the thermal conductivity (*k)*, the specific heat capacity (C_p) and the density (ρ). The density of the rock is dependant on the temperature and is calculated using equation 2.

$$\rho = \rho_0(1 - \alpha(T - T_0)) \qquad (2)$$

In equation 2 the reference density (ρ_0) and the coefficient of thermal expansion (α) are dependant on the rock type (Turcotte and Schubert, 2002). The values used for all the parameters in equations 1 and 2 are shown in Fig. 5.

Figure 5 should be placed here

The model is set up with a two layer crust composed of a felsic upper crust and granulitic lower crust which is underlain by mantle lithosphere. The base of the lithosphere is purely thermal rather than compositional and describes the temperature below which the mantle rock does not deform significantly over geological timescales. The 1200 °C isotherm is used (Turcotte and Schubert, 2002). The model starts with a 20 km thick mantle lithosphere, which matches the starting conditions described in section 2. The temperature at the surface of the model is set as 0°C. The temperature at the base of the model is calculated using a potential temperature at the surface using an adiabatic gradient of 0.3°C per km. Constant temperature boundary conditions are used at the top and bottom of the model. The initial temperature conditions follow the mantle adiabat to up to a transition point 20 km below the crust above which they follow a linear gradient to the surface. The depth and therefore temperature at this transition point is dependent on the thickness of the crust and the potential temperature for the surface. The initial conditions for the model are shown in Fig. 5. The temperature profile is used to calculate the density at grid point in the column from the reference density using equation 2. The elevation is calibrated from the density profile of the column using a column of hot mid ocean ridge material with a 7 km thick basaltic-gabbroic crust at 3 km below sea-level as reference. Standard Pratt isostacy (Allen and Allen, 2005) is used. When the topography of the model drops below sea level the basin is filled with water. The thermal boundary condition is applied to the basement floor because water in the basin would have an almost uniform temperature due to convection. However, the water is included in the isostacy so the model effectively produces water loaded tectonic subsidence.

One inherent weakness of the plate model is that it only calculates the heat transfer by conduction (Parsons and Sclater, 1977). At the base of the plate heat transfer changes from conduction to convection (Huang et al., 2003; Richter and Parsons, 1975; van Hunen et al., 2005). The plate model does not explicitly describe the physics of this transition, but provides a very good fit to the bathymetry and heatflow data (Huang and Zhong, 2005) and included references. It does not help explain the physical process that causes the thermal boundary layer to become stable at this depth. We ignore the temperature dependence of k, C_p and α. This assumption slightly overestimates temperatures in the top half of the model and underestimates temperatures in the bottom half of the model (McKenzie et al., 2005). This means the depth to the base of the lithosphere is an upper limit.

4.2 Results: Effects of varying parameters and the best fit model for the backstripping

The model is set up so that the each of the parameters from equation 1 can be changed and the structure of the model, shown in Fig. 5**,** can be altered. Changing these will alter the subsidence produced. It is important to understand how these different parameters affect the subsidence before trying to find the best fitting model because there may be a trade off between different parameters.

The model is most sensitive to changes in the thickness of the crust and the plate thickness because they change the isostacy by varying the amount of low density crust and high density lithosphere. However, changing the plate thickness affects the timing of the subsidence whereas changing the crustal thickness only affects the total subsidence so their effects can be distinguished. There is not a direct trade-off between the two parameters. The model is less sensitive to variations in the potential temperature and radioactive heat production respectively (Fig. 6). In each case only the parameter under investigation is changed and the estimates for North African crust shown in Fig. 5 are used as the default the parameters. The amount of variation in the input parameters is based on the uncertainties and variation across North Africa from the literature. The present day thickness of the crust of North Africa was calculated to be between 30 and 40 km thick from low resolution gravity and seismic interpretations and gravity modelling in the 3-D crustal model of Seber et al. (2001) over the area of interest. This is thicker than the crustal thickness of 25-35 km estimated from the inversion of surface waves given by Pasyanos and Nyblade (2007). Due to the uncertainty in crustal thickness our model calculations were run for a range of crustal assemblages between 20 and 40 km thick. The results are shown in Fig. 6a. With a crustal thickness of 40 km, ~ 1.2 km of tectonic subsidence takes place, however the crust is too buoyant to drop below sea-level, so no basin is formed. When the crust is reduced to 36 km thick the total subsidence increases to ~1.9 km and 1 km of water loaded tectonic subsidence is recorded. A 20 km thick crust starts at 1.4 km below sea-level and experiences ~2.7 km of tectonic subsidence as the lithosphere cools and thickens. There have been no deep crustal seismic lines of North Africa published so the proportion of upper crust and lower crust is unknown. Therefore the model was run with a 30 km thick crust where the thickness of the upper crust was varied between 10 and 20 km. This alters the total subsidence by about 500 m. A 1 km decrease in the total crustal thickness causes ~150 m of additional subsidence, whereas changing 1 km of upper crust to lower crust only increases the total subsidence by ~ 50 m. The main reason both these parameters affect the overall subsidence is that they change the isostacy by varying the amount of low density crust. In all these scenarios the timing of the subsidence does not change, most of the subsidence occurs in the first 100 Myrs and the subsidence after 200 Myrs is negligible.

Figure 6 should be placed roughly here in the text

We also change the amount of radiogenic heat production. Estimations of heat production in the continental crust vary between 1.31 mWm^{-3} and 0.8 mWm^{-3} for the upper crust and 1.0 mWm^{-3} and 0.6 mWm^{-3} for the lower crust. The values reflect calculations of bulk crustal heat production from the literature (Christensen and Mooney, 1995; Gao et al., 1998; Shaw et al., 1986; Wedepohl, 1995). Changing the heat production for the upper and lower crust between these extremes causes a small variation of 100 m in the tectonic subsidence. The heat production of the bulk continental crust will be lower than that of accretionary crust and heat production would have been higher during the Palaeozoic, so the model was also run with a higher heat production of 2.0 mWm^{-3} (upper crust) and 1.2 mWm^{-3} (lower crust). This does reduce the overall amount of tectonic subsidence from ~ 1800 to 1550m, but does not alter the timing. Fig.

6b shows that varying the heat production in the crust has a small effect compared to varying the crustal thickness.

The other main parameter which has a big effect on the subsidence is the thickness of the model, referred to as the plate thickness (Fig. 6c). A plate thickness of 95 km, which is the preferred for the oceanic lithospheric cooling model of Stein and Stein (1992) results in 200 m of subsidence, of which 80% has occurred within ~70 Myrs, with a final lithospheric thickness (i.e. depth to the 1200 °C isotherm) of 81.1 km. At the other extreme when a plate thickness of 200 km is used, similar to an Archean craton (Priestley and McKenzie, 2006), ~3050 m of subsidence is produced, and it takes longer for the subsidence to tail off. 80% of the subsidence is completed within 200 Myrs. The lithosphere thickness after 500 Myrs is 166.5 km. As the plate thickness is increased the final lithospheric thickness also increases. A thicker, dense lithosphere results in more subsidence, occurring over a longer time period.

The temperature at the base of the plate is controlled by both the potential temperature at the surface (the temperature at which the mantle adiabat dissects the surface) and the thickness of the plate (Fig. 5). Increasing the potential temperature by 1°C increases the temperature at the base of the plate by 1°C whereas increasing the thickness of the model causes the temperature at the base of the plate to increase along the mantle adiabat i.e. 3°C for every 10 km. The potential temperature was varied from 1280 °C, which was suggested by McKenzie et al. (2005) to 1410 °C suggested by Parsons and Sclater (1977). As Fig. 6d shows this variation of 100 °C in the potential temperature produces less than 500 m difference in the overall subsidence.

Van Wees et al. (2009) report a trade off between parameters, such as the final lithospheric thickness (plate thickness) and crustal thickness, when investigating the uncertainties in surface heatflow using modelling. However, in this study we find that varying the plate thickness to fit the time span of the subsidence and then varying the crustal thickness to fit the amount of subsidence allows a unique solution to be obtained for these parameters. However, there is a direct trade off between the crustal thickness, heat production and potential temperature because they all only affect the total subsidence. The model is not sensitive to reasonable variations in the heat production or potential temperature, but is sensitive to variations in crustal thickness suggested for North Africa. Therefore the trade off between these parameters is not important when trying to fit the modelled subsidence to the observed subsidence.

The sensitivity of the model to k, C_p and α was also investigated, but because much less is known about the amount of variation that is found in the crust they are not discussed here and the standard values from the literature (Fig. 5) are used. The forward model produces significant subsidence for a large variety of crustal assemblages and for wide variation in the parameters. It supports the hypothesis that lithospheric cooling and thickening caused the subsidence seen across North Africa.

5. Discussion

5.1 Comparison of tectonic subsidence from backstripping and forward modelling.

The results of the forward modelling show a good fit to the backstripping from both basins as demonstrated by Fig. 7. Four forward model runs are shown on the diagram. One of which uses the best estimates for the crustal assemblage, thickness and plate thickness from the literature, one which is the best fit to the backstripping and then two runs which show the maximum and minimum subsidence for reasonable variations of the properties of the North African crust.

The forward model curves are smooth unlike the backstripped subsidence curves. This is because North African crust has experienced periods of uplift, and the forward model doesn't include these complexities. However, the forward model fits the broad shape of the backstripped subsidence curve very well. Both have a concave up shape, and the timescale and magnitude of subsidence are very similar. When the crustal thickness, final lithosphere thickness and heat production used are taken from the best estimates of the North African crust in the literature (Fig. 5), and therefore are independent of the backstripping, the forward model curve only underestimates the subsidence slightly, by <500 m in 350 Myrs. The discrepancy at 500 Myrs is ~700 m, because of the Mesozoic and Cenozoic subsidence of the Ghadames Basin. This is a separate tectonic event, related to Tethyan opening and evolution (Guiraud and Bosworth, 1999). The best fit model has been fitted iteratively by eye. The input parameters for the best fit model are the same as those from the literature except that the plate is 155 km thick, which results in a final lithospheric thickness of ~124 km. The crust beneath the basins cannot be assumed to be homogenous, in fact because it is accreted from numerous crustal fragments it is likely that it varies across North Africa. The maximum and minimum subsidence based on the variation found in the literature, discussed in section 4.2, is shown in Fig. 7. The upper limit has a 35 km thick crust with 30 km of upper crust and 5 km of lower crust, a heat production of 1.6 mWm^{-3} in the upper crust and 1.0 mWm^{-3} in the lower crust, a potential temperature of 1280 °C and plate thickness of 120 km. The lower limit has 25 km thick crust, where the top 5 km is upper crust and the rest is lower crust, a heat production of 1.0 mWm^{-3} in the upper crust and 0.6 mWm^{-3} in the lower crust, a potential temperature of 1380 °C and a plate thickness of 180 km.

Figure 7 should be placed here

There are a number of assumptions in both the modelling and backstripping. Both are oversimplified versions of the real situations and processes. The problem has been simplified to be one dimensional, whereas the basins are three dimensional features. We assume that the basins are large enough features that it is possible to remove effects of marginal uplift from the sedimentary record by choosing wells away from the basin flanks, or constructing composite wells. This allows the basins to be treated as one dimensional features. Another assumption is that North African crust resembles the starting conditions outlined in the hypothesis. This is difficult to test as there is no newly assembled accretionary orogenic belt on the North African scale that can be used as an analogue. However, the building blocks of accretionary crust are available to study. It seems reasonable that North African crust formed through subduction would share the thin lithosphere of its component island arcs and accretionary wedges. When North African crust is modelled using these starting conditions it produces a good fit to the subsidence. As mentioned in the hypothesis as long as the crust is of near normal thickness and has a thin mantle lithosphere these results still apply. The two North African basins have been chosen as case studies, but we propose that the lithosphere cooling and thickening hypothesis is applicable to all the Palaeozoic basins on the accretionary crust of North Africa and Arabia, and to other basins on accretionary crust around the world, such as the Mesozoic cover of the Scythian and Turan platforms of SW and Central Asia (Natal'in and Şengör, 2005), the Mesozoic cover of the Tasmanides of Eastern Australia (Gallagher et al., 1994) and the late Palaeozoic platform cover over the Altaid orogenic collage in Central Asia (Cook et al., 1994).

The good fit between modelled and backstripped subsidence curves do not prove that lithosphere thickening and cooling is the cause of the subsidence. Other authors have suggested that the Palaeozoic basins in North Africa formed by rifting (Lüning et al., 1999), orogenic collapse (Ashwal and Burke, 1989) and dynamic topography (Heine et al., 2008). Lüning et al. (1999), compared the Al Kufrah Basin to pull apart basins in Oman and Saudi Arabia, and suggested that the Najd fault system may extend north into the Al Kufrah Basin and be reactivated as a rift during the Precambrian. Seismic lines of the south of the Al Kufrah Basin do indeed have features which have been interpreted as rifts. However, they are only found in a small area of the basin and the synrift subsidence accounts for a sixth of the overall subsidence (Ghanoush and Abubaker, 2007). The Ghadames Basin has very similar stratigraphy to Al Kufrah, but shows no evidence of rifting. Rifting is not widespread or large enough to cause the subsidence across all the Palaeozoic basins in North Africa. This also means that orogenic collapse is an unlikely subsidence mechanism, given that this is a special case of rifting on thickened continental crust. Armitage and Allen (2010) suggest that these basins can still form due to extension, but when stretching is occurring at low strain rates the strain might not localise along faults. It is hard to distinguish between their model and the model proposed in this paper on the grounds of rifting. Subsidence due to extension at low strain rates produces a period of constant subsidence during stretching followed by decreasing subsidence during the thermal sag phase. Whereas cooling of a thin lithosphere produces subsidence which decreases smoothly throughout time. The latter model fits the backstripping results (Fig. 4) which do not have a kink, but are be closer to a smooth curve.

There are problems with using dynamic topography to explain the basins: it has a transient effect and will rebound once the down welling has ceased leading to uplift and erosion. Nor is it possible at present to model whether the basins in North Africa would have been affected by mantle down-welling during the Palaeozoic. Thermal subsidence on the other hand explains the lateral extent of the subsidence as the lithosphere would be thin across most, if not all, of the accretionary crust. Our results show that it is long lasting, fits the magnitude of the observed subsidence and is able to cause subsidence over a wide range of terranes.

Thermal subsidence has a number of important implications. It suggests that one of the fundamental properties of accretionary crust is that, in the absence of other competing tectonic forces, it will subside after its formation. This will alter the overall composition of the crust, adding several kilometres of sediments to crust made up largely of island arc and oceanic material. It also predicts that the lithosphere beneath accretionary crust is not depleted by the melting events which produced the crust, but is compositionally no different to upper mantle. This will make the lithosphere more susceptible to thermal erosion as it does not have compositional buoyancy and also affect the character of melts originating in it or passing through it. Ashwal and Burke (1989) also noted that this fits the geophysical measurements of the thickness and seismic properties of the lithosphere and with the geochemistry of Cenozoic volcanism in North Africa.

6. Conclusions

We have shown that results from backstripping wells in two North African basins and numerical modelling, are consistent with a basin forming mechanism of lithospheric cooling and thickening, underneath relatively juvenile, accretionary crust. The backstripping revealed that the Ghadames and Al Kufrah basins have experienced their highest rates of subsidence, (22.2 m/Myr and 10.1 m/Myr, respectively) at the start of the Palaeozoic, declining to almost zero at the end of the Palaeozoic. There is almost no tectonic subsidence during the Mesozoic and Cenozoic. The forward modelling shows that thermal subsidence of accretionary crust provides a viable explanation for the formation of the Palaeozoic basins of North Africa, fitting the available data better than previously suggested mechanisms.

We suggest that thermal subsidence following accretion would be expected in other areas of accretionary crust, and could have influenced the rifted West Siberian Basin. It also predicts that accretionary crust is underlain by fertile mantle lithosphere which may explain why Cenozoic volcanism in North Africa and Arabia seems to originate from a fertile source (Ashwal and Burke, 1989).

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Figure Captions

Fig. 1. Formation of accretionary crust and subsequent lithospheric thickening and subsidence. a) Assembly of accretionary crust from island arcs and other crustal fragments through subduction. b) Newly formed accretionary crust with a thinned lithosphere. c) Lithospheric thickening due to cooling causing subsidence.

Fig. 2. Present outcrop and sub-Mesozoic subcrop of Palaeozoic sediments across North Africa and Arabia (Boote et al., 1998; Konert et al., 2001). Four wells used to construct the Ghadames composite well are located along the line of the transect shown.

Fig. 3. Stratigraphy of the Ghadames and Al Kufrah basins used for backstripping analysis. The Ghadames composite well is constructed from four wells located along the transect shown in Fig. 2, to give the most complete sedimentary record. The locations of the wells from Al Kufrah are also marked on Fig 2.

Fig. 4. Tectonic subsidence curves for the Ghadames and Al Kufrah basins. The solid lines are the tectonic subsidence and the dotted lines show the subsidence if the stratigraphy is entirely shale or sandstone. This envelope is the maximum possible range of variation in the subsidence possible from varying the proportions of the lithologies.

Fig. 5. The right-hand side shows structure and initial temperature conditions for the numerical forward model. The solid line is the preferred temperature profile while the dashed line indicates the potential temperature. The left hand side shows the best estimates for the material properties for the different layers of crust and mantle from the literature. (Allen and Allen, 2005; Shaw et al., 1986; Turcotte and Schubert, 2002)

Fig. 6. The sensitivity of the forward model to variations **(a)** in the crustal thickness and proportion of lower crust (LC) and upper crust (UC), **(b)** the heat production, **(c)** the plate thickness and **(d)** the potential surface temperature of the mantle adiabat.

Fig. 7. Comparison of the subsidence from backstripping the composite well from the Ghadames Basin with a range of forward model runs. The hypothesised subsidence mechanism fits the observed subsidence well although there is a wide range of subsidence over the possible range of variations in the North African crust as demonstrated by the upper and lower limits.

Table Captions

Table 1 Compilation of intracratonic basins situated on accretionary crust from around the world. The table shows their locations and the temporal link between the end of the accretion event forming the underlying crust and the beginning of subsidence across the basin as a whole.

Table 1
Click here to download Table: Table 1.doc

Basin Name	End of basement accretion	Beginning of platformal subsidence
South America	~510 Ma (Cordani and Teixeira, 2007)	
Parnaíba Basin		Silurian (Oliveira and Mohriak, 2003) ≤ 444 Ma
Paraná Basin		Ordovician (Brito Neves, 2002; Zalán et al., 1990) ≤ 488 Ma
Choco-Paraná Basin		Ordovician (Brito Neves, 2002) ≤ 488 Ma
North Africa	~550 Ma (Caby, 2003)	
Mouydir Basin		Cambrian (Boote et al., 1998) ≤ 542 Ma
Ahnet Basin		Cambrian (Boote et al., 1998) ≤ 542 Ma
Reggane Basin		Cambrian (Boote et al., 1998) ≤ 542 Ma
Ghadames Basin		Cambrian (Boote et al., 1998) ≤ 542 Ma
Al Kufrah Basin		Cambrian (Boote et al., 1998) ≤ 542 Ma
South Africa	~550 Ma (Bumby and Guiraud, 2005)	
Cape Basin		Cambrian (Tankard et al., 2009) ≤ 542 Ma
Arabia	~550 Ma (Stern, 1994)	
Arabian Platform		Cambrian (Konert et al., 2001) ≤ 542 Ma
Central Asia	~210 Ma (Natal'in and Şengör, 2005)	
Scythian Platform		Jurrassic (Natal'in and Şengör, 2005) ≤ 200 Ma
Turan Platform		Jurrassic (Natal'in and Şengör, 2005) ≤ 200 Ma
Eastern Australia	~250 Ma (Glen, 2005)	
Eromanga Basin		Jurassic (Gallagher et al., 1994) ≤ 200 Ma
Surat Basin		Jurassic (Gallagher et al., 1994) ≤ 200 Ma

Figure 1

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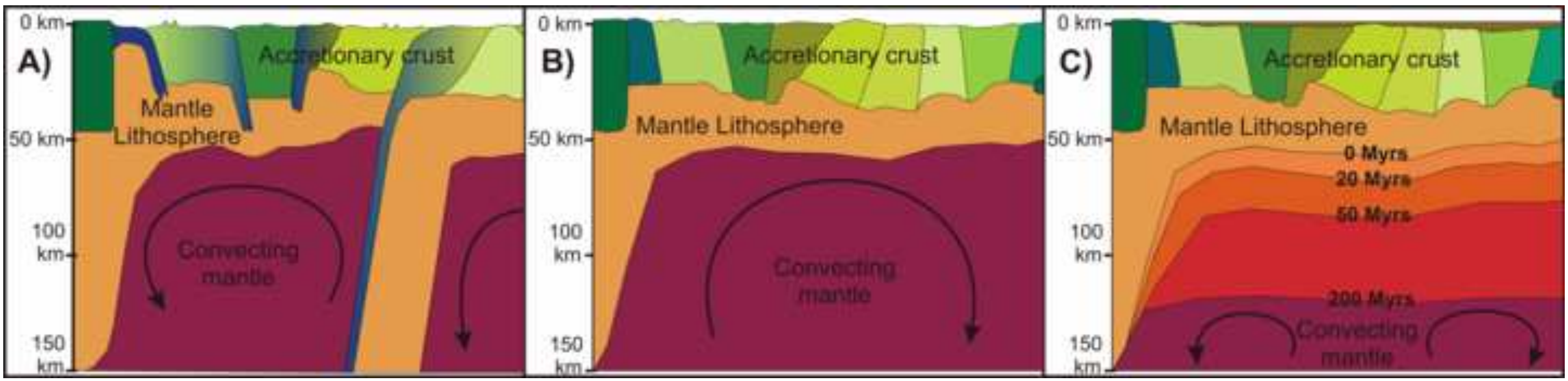

Figure 2
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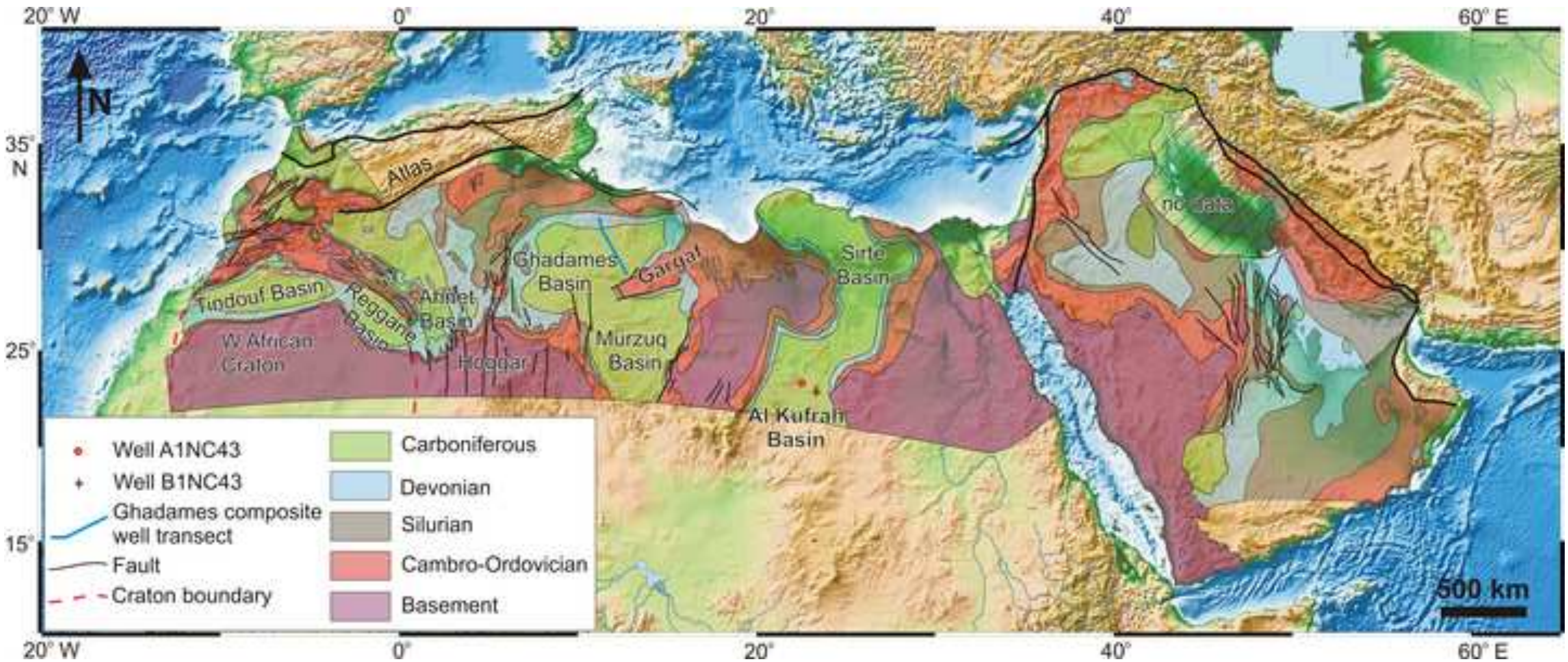

Figure 3

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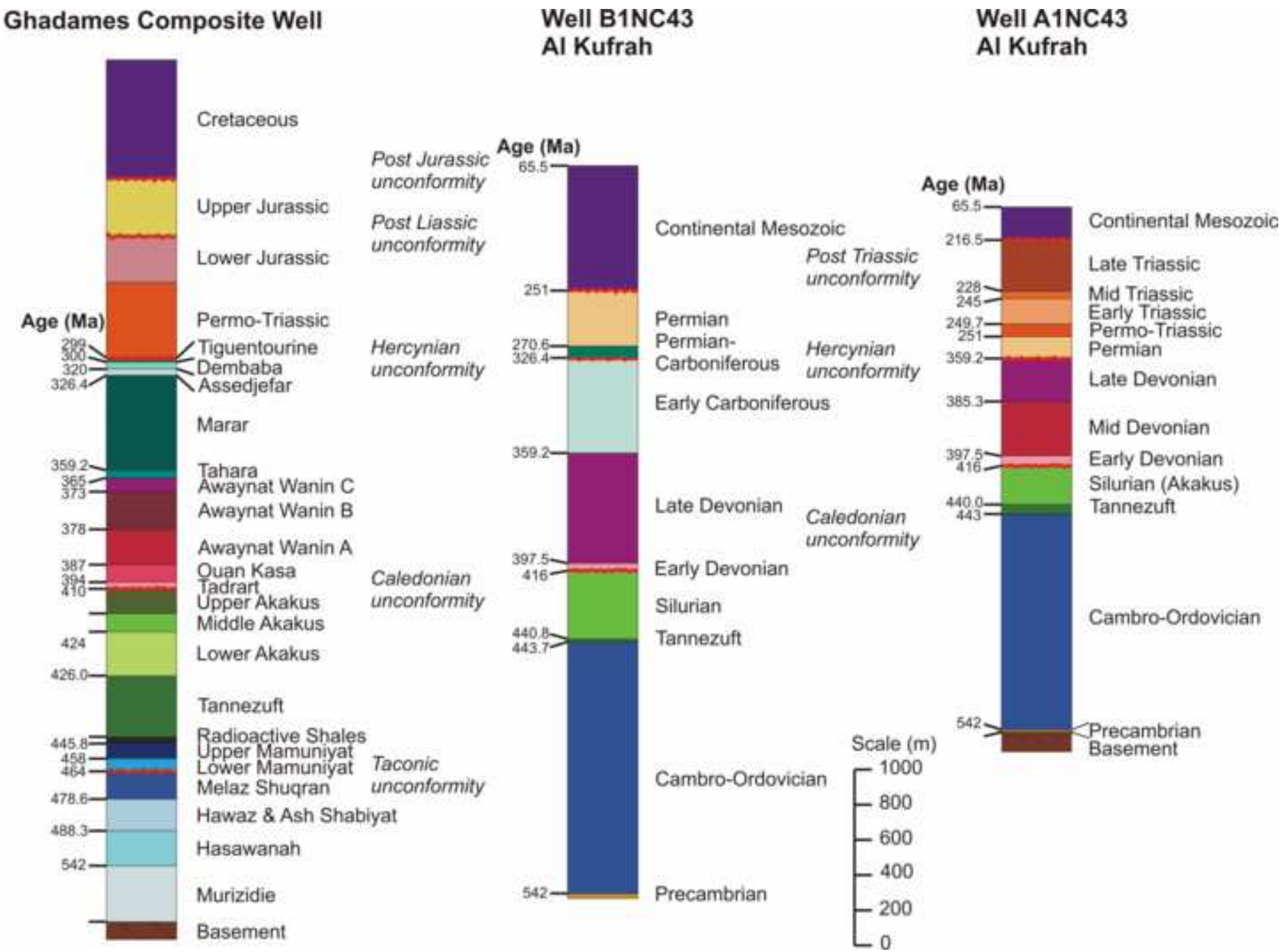
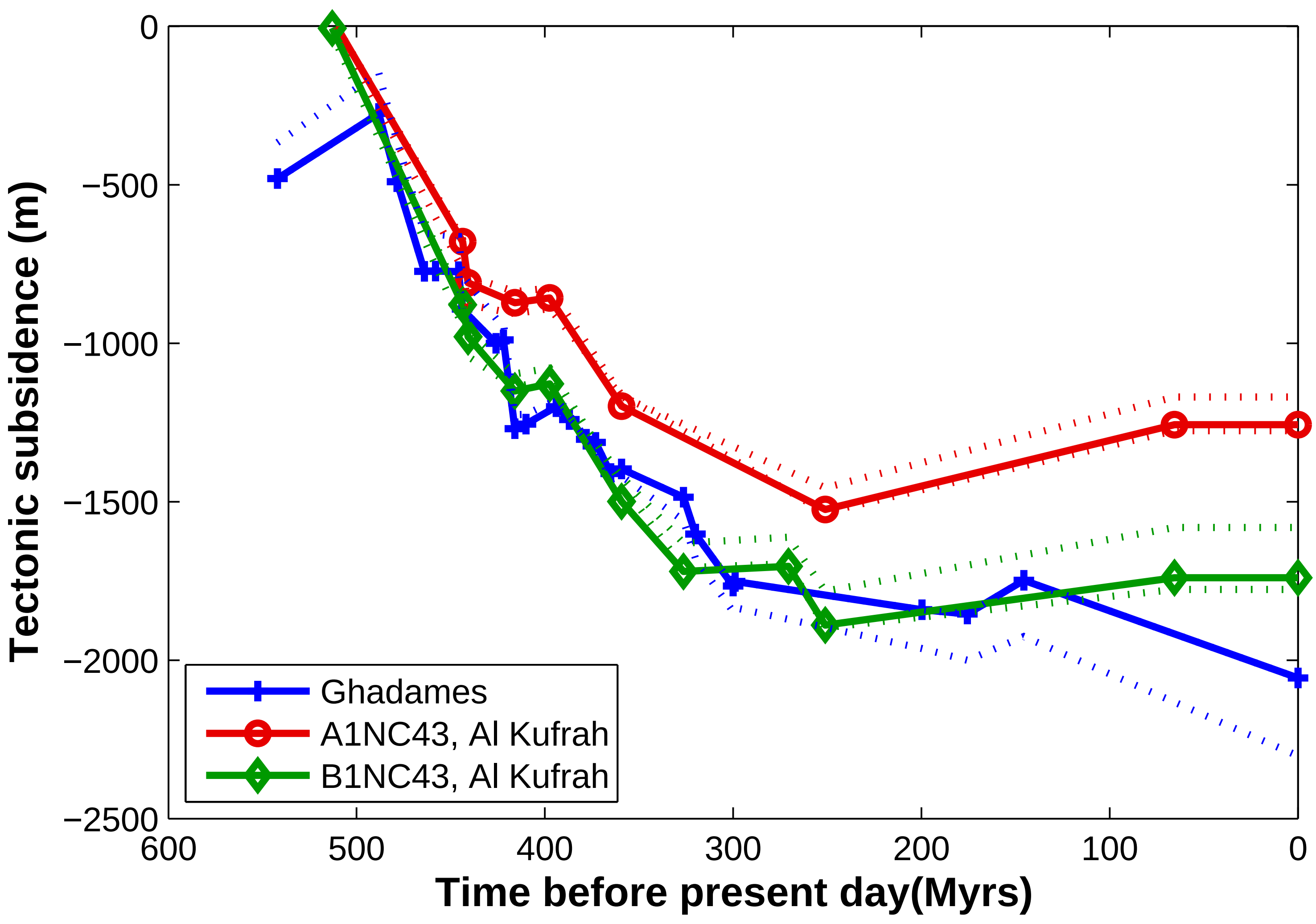
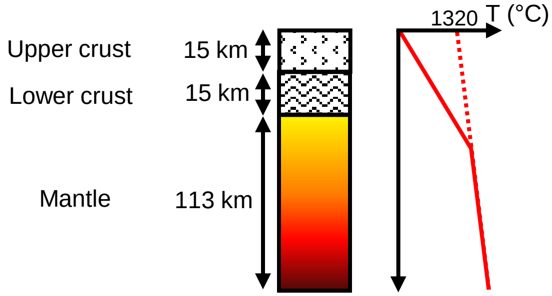


Figure 4



0
−500
−1000
−1500
−2000
−2500
600
500
400
300
200
100
0
Tectonic subsidence (m)
Time before present day(Myrs)
Ghadames
A1NC43, Al Kufrah
B1NC43, Al Kufrah

Figure 5



ρ_0 (kgm^{-3})	A (mWm^{-3})	α (K^{-1})	k ($Wm^{-1}K^{-1}$)	Cp ($Jkg^{-1}K^{-1}$)
2750	1.31	2.4×10^{-5}	3.1	790
2900	0.8	1.6×10^{-5}	2.1	790
3300	0.006	3.3×10^{-5}	3.3	790

Figure 6 Black and White version

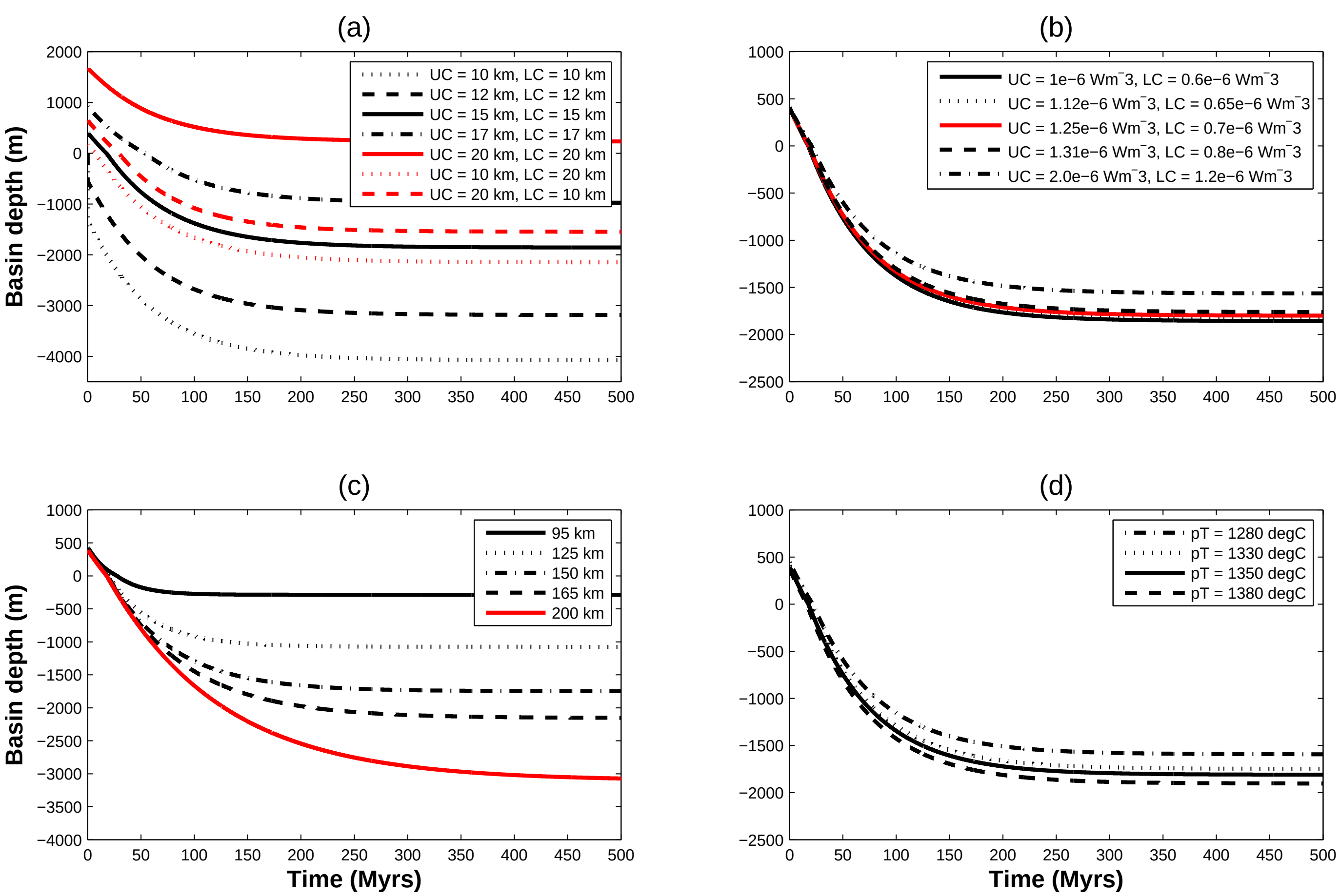


Figure 6 Colour version

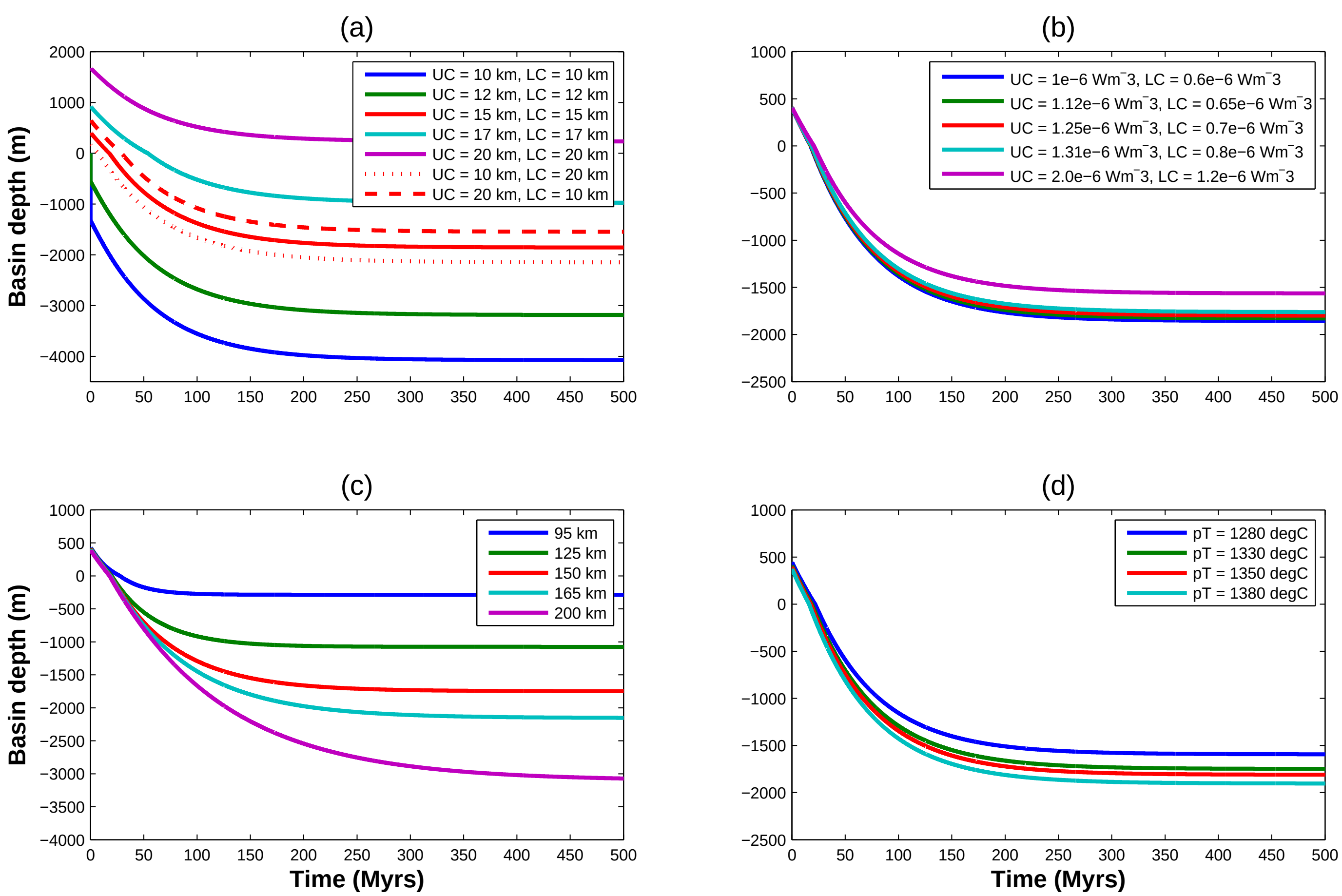
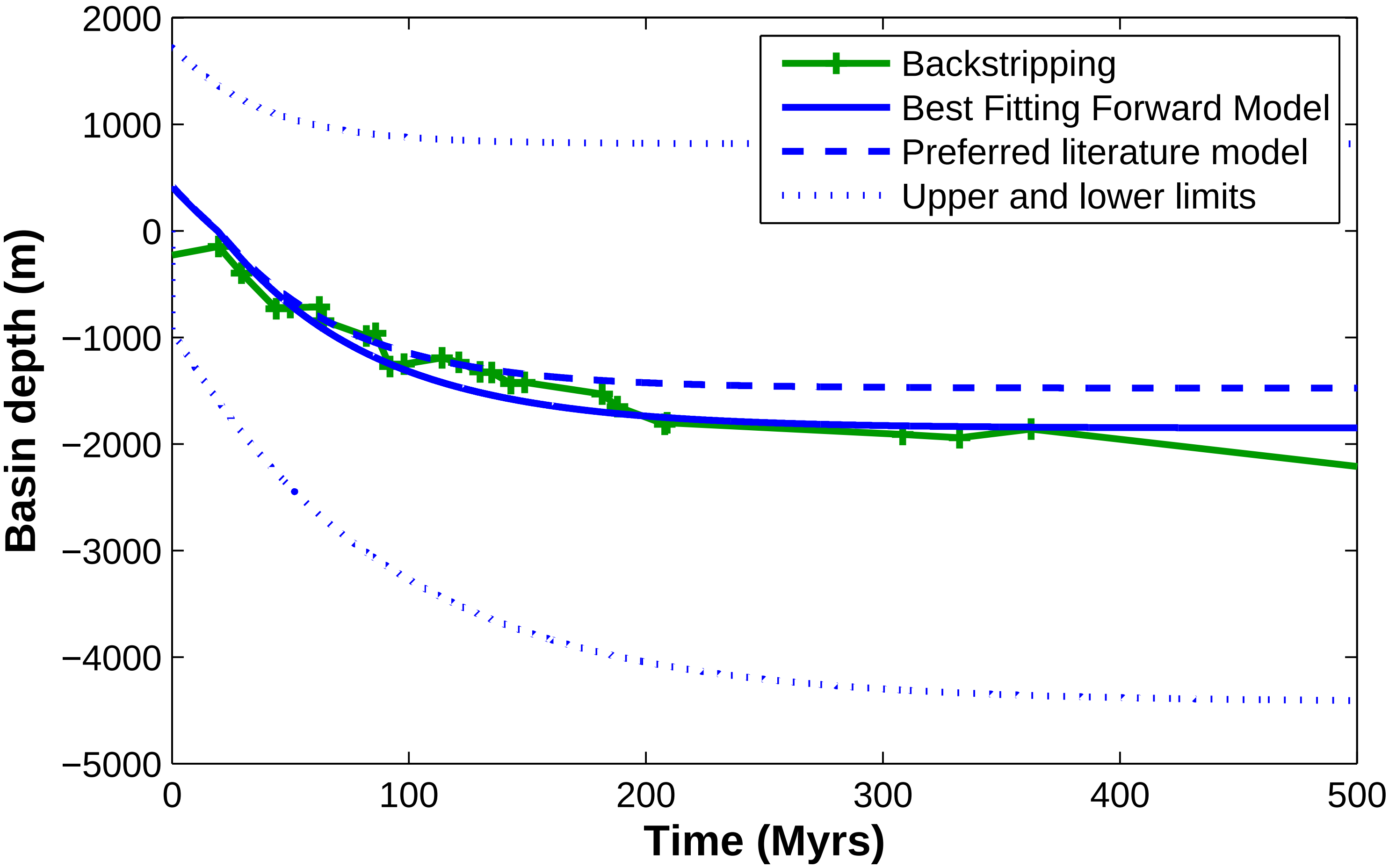


Figure 7



Backstripping
Best Fitting Forward Model
Preferred literature model
Upper and lower limits
Basin depth (m)
Time (Myrs)
2000
1000
0
−1000
−2000
−3000
−4000
−5000
0
100
200
300
400
500