ORIGIN OF RIFTED CRATONIC BASINS: TESTING THE SLOW STRETCHING MODEL

A Thesis

Presented to

The Faculty of the Department of Earth and Atmospheric Sciences

University of Houston

In Partial Fulfillment

of the Requirements for the Degree

Master of Science in Geology

By

Alli Oluwaseun Oyepeju

August 2013

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Finally I am eternally grateful to God for the gift of life and who has kept me thus far.

ABSTRACT

Cratonic basins are large depressions filled with sediments that are located on cratonic shields. They are huge repositories of hydrocarbons, fresh water aquifers, and other important resources, and therefore are of important economic importance. Cratonic basins are characterized by having a long history of subsidence, a saucer-oval shape and 2-9 km of sediment thickness. Despite this economic and geodynamic importance, the process (or processes) that form these basins are still debated. One widely accepted hypothesis suggests that cratonic basins are formed by slow rifting of the lithosphere followed by a long period of thermal subsidence.

In this work I investigated this hypothesis by using 2D geodynamic numerical models of the lithosphere. These models are developed to understand the thermal subsidence during and following slow rifting, the deformation of the crust and mantle lithosphere, and the lithospheric stress field. Several models were tested in which I varied the lithosphere thickness, strain rates, and stretching factors. The model allows the lithosphere to rift, thereby creating accommodation space, shallowing of the Moho, and allowing passive upwelling of mantle material. I found that minor rifting is followed by a phase of ~175 million years of cooling of the lithosphere, and thermal subsidence. So, according to the models, the slow rifting hypothesis for the origin of cratonic basins can account for the long subsidence phase experienced by these basins. For cratonic basins where there is no record of minor rifting, other processes such as downgoing mantle flow, dynamic topography, and phase changes drive the slow subsidence.

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CHAPTER 1: INTRODUCTION

1.1 OVERVIEW OF CRATONIC BASINS

Cratonic basins are large saucer-shaped sedimentary basins located on stable cratonic or Proterozoic lithosphere. They experienced a long phase of subsidence, sometimes interrupted by periods of regional uplift (Armitage and Allen, 2010). These basins are often referred to as intra-cratonic (Caldwell, 1986; Quinlan, 1987), interior continental sag (Leighton et al., 1990), interior cratonic (Klein and Hsui, 1987), and cratonic basins (Bally and Snelson 1980). In this thesis I refer to these basins as cratonic basins. The basins are structural and sometimes topographic depressions located some distance away from plate boundaries (divergent or convergent continental margins), and are not associated with plate boundaries. Most cratonic basins (Figure 1) are believed to be inactive (for example, the Michigan, Illinois, Williston, Western Siberian, Anadarko, and Paris basins), with the exception of the Hudson, Chad, and Congo basins (Downey and Gurnis, 2009).

Buiter et al. (2012) summarized recent studies on cratonic basins, and suggested that the formation mechanism of cratonic basins is still unknown. This is probably related to the fact that these basins seem not to be clearly related to tectonic processes or plate boundaries (Ahern and Dikeou, 1989, Ingersoll and Busby, 1995). Faults and other tectonic structures such as anticlines are usually present in cratonic basins, but they do not dominate the structure of the basin. Also, the region of largest deposition is always consistent for successive units accumulated over periods of more than a 100 m.y. as suggested by Nunn and Sleep (1984).

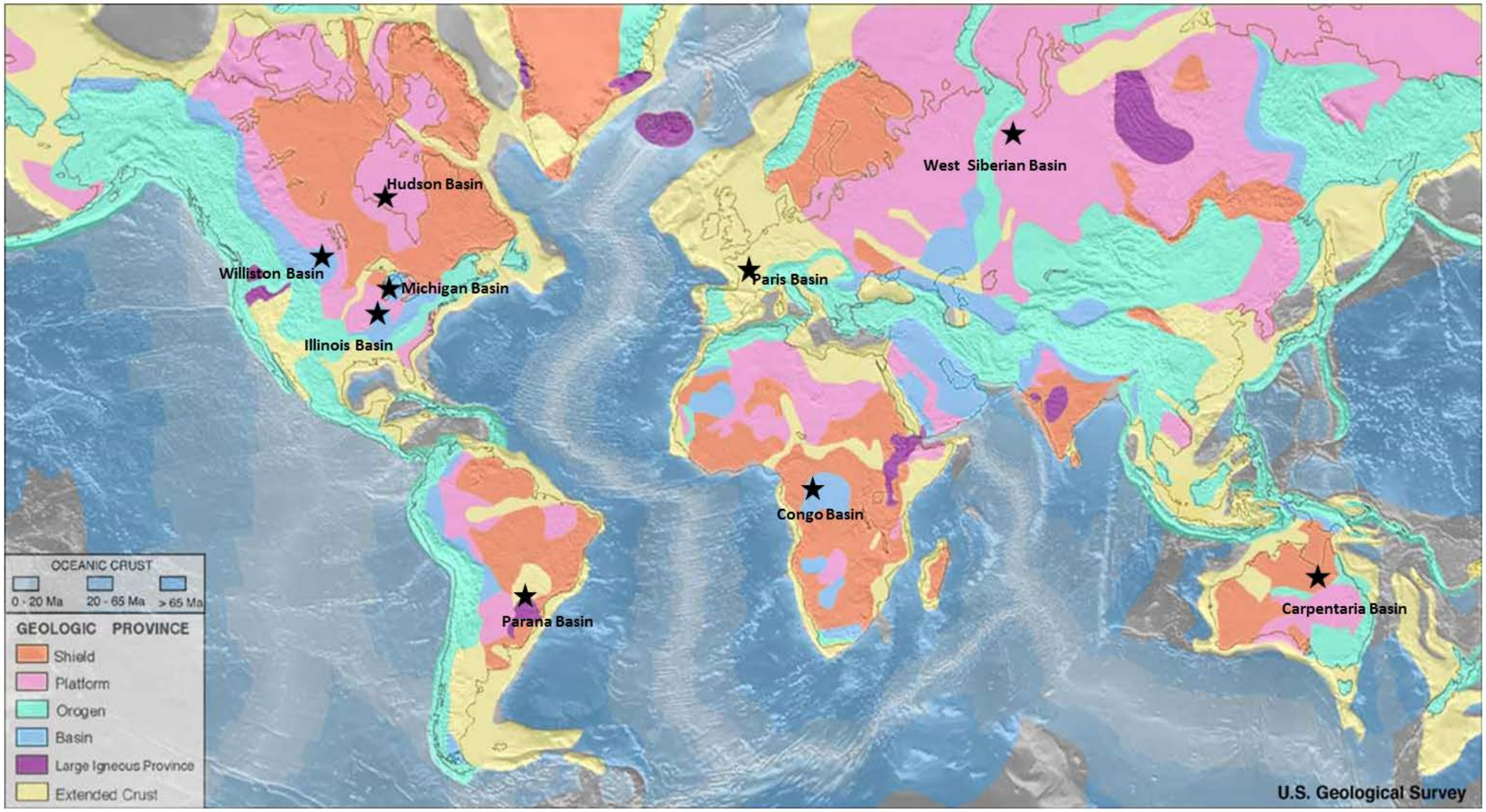


Figure 1. Map of the world showing the location of a selection of cratonic basins located on continental shields. Map with geologic provinces from U.S.G.S.1997.

All cratonic basins are different, but they do share some common attributes:

- Cratonic basins are characterized by a prolonged phase of subsidence in tectonic subsidence curves (Xie and Heller, 2009). The prolonged duration of subsidence is sometimes interrupted by unconformities marking uplift related to plate scale tectonics (Zalán et al., 1990).
- Cratonic basins show no sign of major rifting, although minor horst and graben structures are often present.
- Cratonic basins are structural saucer- shaped depressions. Sometimes they are regions of low surface topography (Heine et al., 2008).
- Some cratonic basin are associated with regional volcanism (intrusive and extrusive), such as the eruption of large flood basalts in the Parana Basin and West Siberian Basin (Saunders et al., 2005).
- Cratonic basins are easily distinguished from other basins based on their oval or circular shape with sizes varying from extensive depressions, e.g., the West Siberian Basin (3.5×10^6 km^2, Vysstoski et al., 2006), the Congo basin (1.2×10^6 km^2, Downey and Gurnis 2009), the Paraná basin (1.4×10^6 km^2, Zalán et al., 1990), to smaller ones such as the Paris Basin(110,000km^2 , Perrodon and Zabek.,1990).

The basins are characterized by shallow marine and terrestrial sedimentation with a simple layer –cake stratigraphy (Sloss and Speed, 1974), implying that sedimentation was at the same pace as tectonic subsidence throughout basin formation. Sediment accumulations are up to 3-10 km with the largest sediment thickness present at the center of the basin. Tectonic basins are situated inboard of margins, and sometimes are linked to ocean basins by failed rifts (for instance the Illinois Basin (Braile et al., 1986) and West Siberian Basin (Vyssotski et al., 2006).

Klein and Hsui (1987) recognized that cratonic basins are not initiated regularly during geological time. Some initiated coeval with the breakup of Precambrian supercontinents (Figure 7) with the African, and South and North American basins as examples. Bond and Kominz (1984) therefore proposed that their formation is related to supercontinent breakup.

Cratonic basins hold approximately 61% of the world oil reserves (Condie, 1997), yet their formation is poorly understood. The basins are loci of huge fresh water aquifers (Gossel et al., 2004), and they are important sediment sinks preserving and recording changes in climatic and tectonic processes occurring on the earth surface (Allen and Allen, 2005). In spite of their geodynamic and economic significance, there has not been a hypothesis that has been able to explain their mechanism of formation, evolution, and subsidence history clearly, without any ambiguity. Different hypotheses have been put forward by different authors (Chapter 3 is dedicated to the major hypotheses). The basin’s structure and evolution cannot unequivocally be classified under the major basin formation processes (such as flexural basins or rift basins). Allen and Allen (2005) observed that the majority of cratonic basins are formed by the isostatic response of alterations in the crustal thickness which may either be related to stretching or to thermal decay due to the cooling.

One of the hypotheses for cratonic basin formation suggests that they are formed as a result of slow extension of the lithosphere followed by a long period of thermal cooling and subsidence (Armitage and Allen, 2010). Armitage and Allen (2010) tested this mechanism using one-dimensional instantaneous models of extension and cooling. Since the lithosphere below rift basins also cools laterally, findings of one-dimensional modeling studies may not hold up in two-dimensions because the amount of cooling is underestimated. Further, cooling may be significant during the syn-rift phase. In this study, I used a 2-D numerical modeling approach to investigate this "slow extension" model for the formation of cratonic basins. I also tested several scenarios of varying thickness of the lithosphere and extension rates.

1.2 MOTIVATION

Since most cratonic basins show evidence of minor rift structures, the "slow extension" mechanism for their formation is tested here. The following approach is adopted:

- Two-dimensional geodynamic models are set up to simulate slow extension of the lithosphere followed by a phase of no extension and thermal cooling
- By varying strain rates, the position of cratonic basin in the rift-drift suite of basins formed by extension is investigated
- Surface subsidence during the cooling phase is tracked to explore the relation between cratonic subsidence and thermal cooling
- The thickness of the lithosphere is varied to study the effect on rifting, cooling, and subsidence

I hope that the results obtained in this study can be used to explain the role of stretching in the formation of rifted cratonic basins, and I aim to explore the limitation and merits of the hypothesis.

CHAPTER 2. CHARACTERISTICS OF CRATONIC BASINS

The purpose of this chapter is to discuss the structural characteristics of active and inactive cratonic basins. I use examples of several well-studied cratonic basins, including the Congo Basin, the Michigan Basin, Illinois Basin, Williston Basin, Paranã Basin, and West Siberian Basin.

2.1 CONGO BASIN

The Congo Basin in Africa (Figure 1, 2) is one of the few cratonic basins believed to be active today (Downey and Gurnis, 2009); Delvaux and Barth, 2010). The basin covers an area of 3.7 million km^2, and is located on the Congo craton and confined mainly today within the boundaries of the Democratic Republic of Congo. The basin is surrounded by numerous Archean and Paleoproterozoic blocks (Kabongo et al, 2011). It has a distinct circular depression (Buiter et al., 2012) with a negative Free-Air gravity anomaly. The basin has experienced a long period of subsidence (Kabongo et al., 2011, Daly et al., 1992). The basin's Neo-Proterozoic to Neogene sediment thickness is ~9 km (Daly et al., 1992, Downey and Gurnis, 2009) with five stratigraphic sequences (Figure 2). The thickness of the lithosphere, inferred from several tomographic studies, is believed to be ~200-250 km (Downey and Gurnis, 2009; Kabongo et al., 2011, Buiter et al., 2012). The lithosphere is believed to be cold and strong (Crosby et al; 2010). During the Late Palaeozoic the basin occupied a central position in the center of Gondwana. During the breakup of the Pangaea in the Mesozoic the basin remained in its present intracratonic position within Africa.

The origin of the Congo Basin is debated. Sachse et al. (2012) propose that the basin formed after a failed rift event. Downey and Gurnis (2009) believe that a density anomaly at 100 km below the basin had induced active subsidence in the Congo Basin. Hartley and Allen (1994) postulated that dynamic subsidence related to a downward flow in the mantle would best explain the continued subsidence of the Congo basin. Also, Kabongo et al. (2011) and Buiter et al. (2012) hypothesized a rift origin for the basin; they observed that the Congo Basin subsidence is a consequence of a Neo-Proterozoic failed rift, and verified this by several tomographic studies. Also, the magnitude and long term subsidence of the basin are consistent with the thermal time constant of 200-250 m.y., and backstripping curves and a narrow gravity anomaly found under the basin are further indications of crustal thinning associated with rifting.

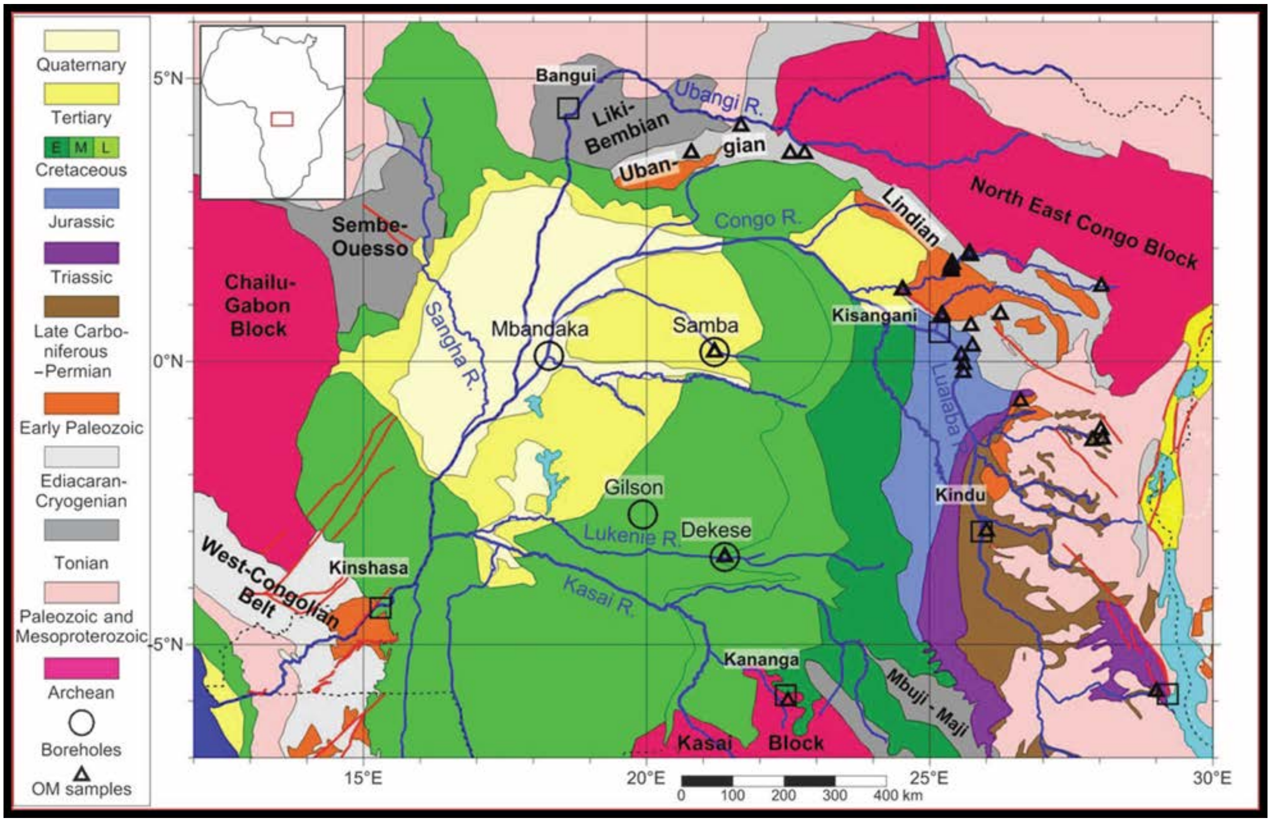


Figure 2. Simplified geologic map of the Congo Basin, from Sachse et al. (2012).

2.2 MICHIGAN BASIN

The Michigan Basin is located on the peninsula of Michigan (Ahern and Dikeou, 1989). It has an almost circular shape (Figure 3), covering an area of 207,000 km (Leighton et al., 1990). The sediment thickness of Cambrian to Jurassic sediments is about 4 km (Figure 3) (Haxby et al., 1976,Arhern and Dikeou, 1989). The basin has a typical sag or saucer-shape, and its sediment package is divided into six sequences (Fisher et al., 1988). The oldest strata in the basin are of Middle and Late Ordovician age (Haxby et al., 1976), while the youngest strata record the subsidence of Pennsylvanian age. The sediments are predominantly limestone, dolomite and shales, with very little sandstone or shale (Nunn et al., 1984).

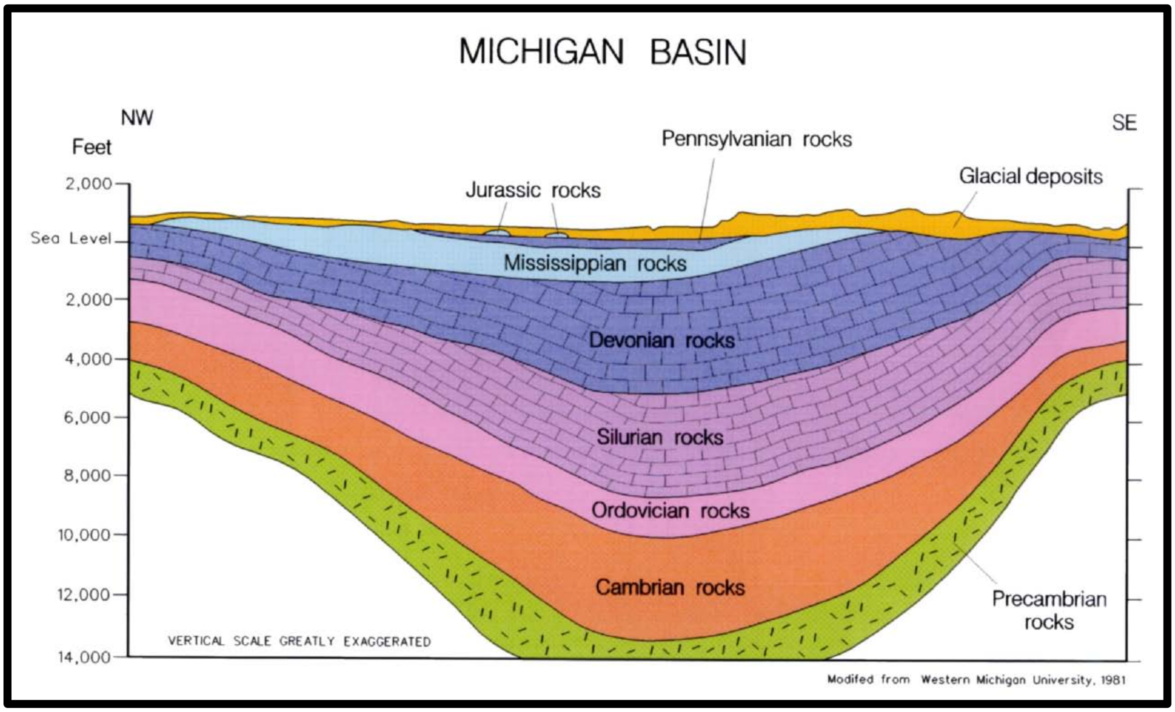


Figure 3. Cross-section of the Michigan basin modified from the Western Michigan University (1981).

The Michigan Basin is bounded to the north by the Canadian Shield, to the east and south-east by the Algonquin and Findlay Arches, to the south-west by the Kankakee Arch and to the north-west and west by the Wisconsin Arch and Wisconsin Dome (Ells, 1969). Nunn and Sleep (1984) observed the basin decreased in width with time.

It has often been suggested that the circular shape of the basin resulted from a flexural loading effect on the lithosphere (Haxby et al., 1976; Ahern and Dikeou, 1989). The origin of the load and subsidence stages of the basin led Haxby et al. (1976) to suggest that the load was created by a mantle diapiric intrusion into the lithosphere to Moho depths. Hereby the lower crust was heated by the ascending diapir which altered the gabbroic rocks of the crust to eclogite through thermal metamorphism. These authors suggested that the basin should show evidence of an uplift event during the early part of the basin's evolution, but there has not been any evidence for this. The authors further explained that if the diapir penetrates to its equilibrium position the mantle material cools to new state (high density) as compared to the lower density crust, causing subsidence of the earth surface. Haxby et al. (1976) believe the mantle diapir may have been a mantle plume. Other studies (Ahern and Dikeou, 1989) argued that a single thermal event occurred at 500 Ma, and that the rapid subsidence in the Cambrian was as a result of the alteration of greenshist to amphibolite in the lower and middle crust while the subsidence in the Ordovician was due to thermal decay. They further suggested that thermal contraction was the cause of the gradual thickening of the lithosphere over time, as evidenced by the changing shape of the basin. McKenzie (1978) proposed that a stretching model would explain the rapid Cambrian subsidence, while the slow subsidence in the Ordovician would have occurred as a result of the cooling and contraction of the lithosphere due to conductive heat loss. Sleep (1971) suggested that the long subsidence period of the basin (>75 m.y.) is constituent with the conductive decay of a thermal anomaly in the lower crust or upper mantle.

2.3 ILLINOIS BASIN

Also the Paleozoic Illinois Basin has a typical sag or saucer shape (Figure 4). The basin is an oval depression that encompasses about 155,000 km^2, accumulating close to 6000 m of Cambrian to Pennsylvanian sediments (Kolata and Nelson, 1990; Klein and Hsui, 1987). It is located in Illinois, Indiana and Kentucky (Figure 4). The basin is a broad structural depression (Figure 4), and it has an oval and elongated shape from north to south

The sediments were initially deposited in a southerly- facing open embayment which may have been formed by rifting (McGinnis, 1970; Heidaluf et al., 1986). Sedimentation trend was controlled by four distinct phases of tectonic subsidence (Heidaluf et al., 1986). The sediments were deposited during the Middle Cambrian rifting phase, Cambrian – Middle Ordovician , Late Ordovician through Early Mississippian, and during the final Middle Mississippian through Early Permian subsidence phase. Shutting of the embayment by the Pascola Arch uplift during the Pennsylvania and Cretaceous time (Marcher and Stearns ,1962) formed the present day oval shape of the basin.

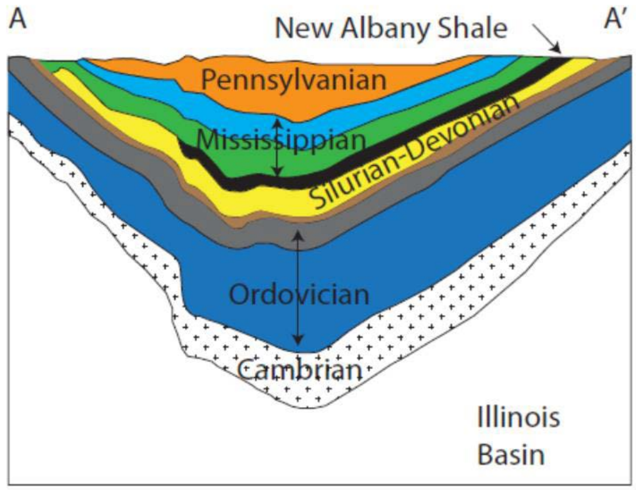


Figure 4. Cross section of the Illinois basin redrawn from Schlegal et al. (2011).

The New Madrid Seismic zone is located on the southwestern side of the basin. Braile et al. (1986) postulated that the New Madrid rift complex was formed during the late Precambrian, resulting in intrusions of mafic rocks which formed the mass excess along the axis of the rift. Cordell (1977) confirmed that there is a mass excess by finding a positive gravity anomaly. This positive anomaly seems to be related to a zone of high seismic wave velocities in the lower crust (Mooney et al., 1983). Braile and his co-workers suggested that during continental stability the excess masses were supported by the strength of the lithosphere, but an increase in geothermal gradient and or regional stress which reduced the viscosity in the lithosphere would have made the excess mass unstable driving subsidence. Their hypothesis is similar to that of DeRito et al. (1983).

Heidaluf et al. (1986) suggested that the basin subsidence is driven by an initial fault subsidence followed by thermal controlled subsidence in response to a mantle plume. The rising plume would have formed a three-arm rift, with one arm below the Illinois Basin.

Sleep et al., (1980) suggested otherwise, and believed that a single thermal event led to the subsidence that occurred during Late Cambrian – Early Mississippian. Also Kolata and Nelson (1990) and Burke and Dewey (1973) proposed that there exist a failed rift arm beneath the basin.

2.4 WILLISTON BASIN

The Williston Basin is a large circular shaped basin (Ahern and Mrkvicka, 1984), located within the United States-Canada border region and lies within the geographic boundaries of North and South Dakota, Montana, and Saskatchewan (Gerhard et al., 1982). The basin is a repository for sedimentary rocks ranging in age from Cambrian to Tertiary age. The sedimentary rocks are divided into six major unconformity bound sequences as proposed by Sloss (1963). The sediment thickness is largest with approximately 4900 m of sediments in North Dakota. There has been little tectonic deformation in the basin, but the basin was affected by the Laramide Orogeny which formed anticlines and hydrocarbon traps. Gerhard et al. (1982) discussed the structural features within the Williston Basin. The most important structures of the basins are the Poplar Dome, Cedar creek Anticline and the Nesson Anticline. Crustal studies of the basins reveal that the Williston basin has a very thick crust and high velocities in the lower crust and upper mantle (Hamdani et al., 1994). Kaminzki and Jaupart (2000) used a 1D thermal model for the Williston Basin, suggesting a lithosphere thickness around 270 km.

The origin of the Williston Basin is debated. Hamdani et al. (1994) suggested that an increase is lower crustal density following the re-crystallization of the mafic lower crust to eclogite caused the crust to subside. Fowler and Nisbet (1985) found the tectonic subsidence of the basin to be approximately steady and also they believed that this was a consequence of transformation of a mafic subcrustal body beneath the basin to a denser phase such as eclogite.

Bond and Kominz (1991) used tectonic subsidence curves for the Middle Paleozoic history of the Williston Basin and suggested a basin –forming mechanism related to large scale compressional tectonics. In their model, they incorporated deep mantle convection models, where down welling regions are under regional horizontal compressive stress. The authors suggested that pre-existing positive and negative lithospheric deflections will be reactivated and enlarged, arches will move upward and the basin undergoes subsidence. But the ability to attain or not to attain this “critical level” of stress before buckling of the lithosphere occurs is still contentious and will mostly depend on the assumed lithospheric rheology (Turcotte and Schubert 1982).

2.5 PARANÃ BASIN

The Paranã basin is located in South America and found within the boundaries of Brazil, Argentina, Paraguay and Uruguay (Zalan et al., 1990), Figure 1. It contains sedimentary rocks ranging in age from Ordovician to Cretaceous (~460-65 Ma) (Milani et al., 1998). It covers a land mass of over 1,500,000 km^2 and the depositional sequence is overlain by extensive lava flows (flood basalts) (Zalan et al., 1990). The flood basalts are associated with Tristan da Cunha mantle plume which was active around the same time as the opening of the southern Atlantic Ocean (Peate et al., 1990). The basin is bordered on the west by Asuncion arch (a flexural feature that may represent the peripheral bulge of the sub Andean foreland basin), bordered on the northeast by the Alto Parnaiba arch, and on the northwest it is bordered by the Pre-Cambrian Paraguai/Araguaia fold belt (Zalan et al., 1990). The basin contains about 7,000 m of sediments.

De Brito Neves et al (1984) suggested that the Cambro-Ordovician molasses deposits of the Brasiliano cycle are interpreted as remnants of crustal stretching. Zalan et al. (1990) observed that crustal stretching may have been the initial force behind subsidence. They observed a centrally located linear depocenter with two subordinate branches similar in manner to a rift triple junction.

The basin underwent three phases of subsidence as shown by subsidence curves (Oliveira, 1987). The subsidence is divided in three phases; Silurian-Devonian, Permian-Carboniferous, and Late Jurassic-Early Cretaceous (Oliviera, 1987; Zalan et al., 1990). The first phase of subsidence is believed to have been driven by regional thermal subsidence and sediment loading caused by increased temperatures of the Gondwana shield. Cooling may have induced regional subsidence (Zalan et al., 1990). The proceeding subsidence event may have been a combination of subtle stretching and rifting of the crust with flexure due to the load of the extensive continental glaciers which dominated the Perm-carboniferous times (Zalan et al., 1990). The late Jurassic and Early Cretaceous marked the last phase of subsidence of the basin, which was driven by the flood basalt load deposited on the basin during the opening of the south Atlantic basin.

2.6 WEST SIBERIAN BASIN

The 3.5 million km^2 depression often referred to as the West Siberian Basin (Figure 5) is amongst the largest cratonic basins in the world (Holt et al., 2012). Formation of the basin has been related to the Siberian Flood basalt province, which forms the largest Phanerozoic continental flood basalt province, erupting at the Permian-Triassic boundary (Vyssotski et al., 2006). The igneous activity was limited to the NNE sector of the basin. Saunders et al. (2005) suggested that there was a regional delay in sedimentation (60-90 m.y.) after initial rifting, which may have been as a result waning of the uplift generated by the thermal effect of a mantle plume (Campell and Griffiths, 1990).

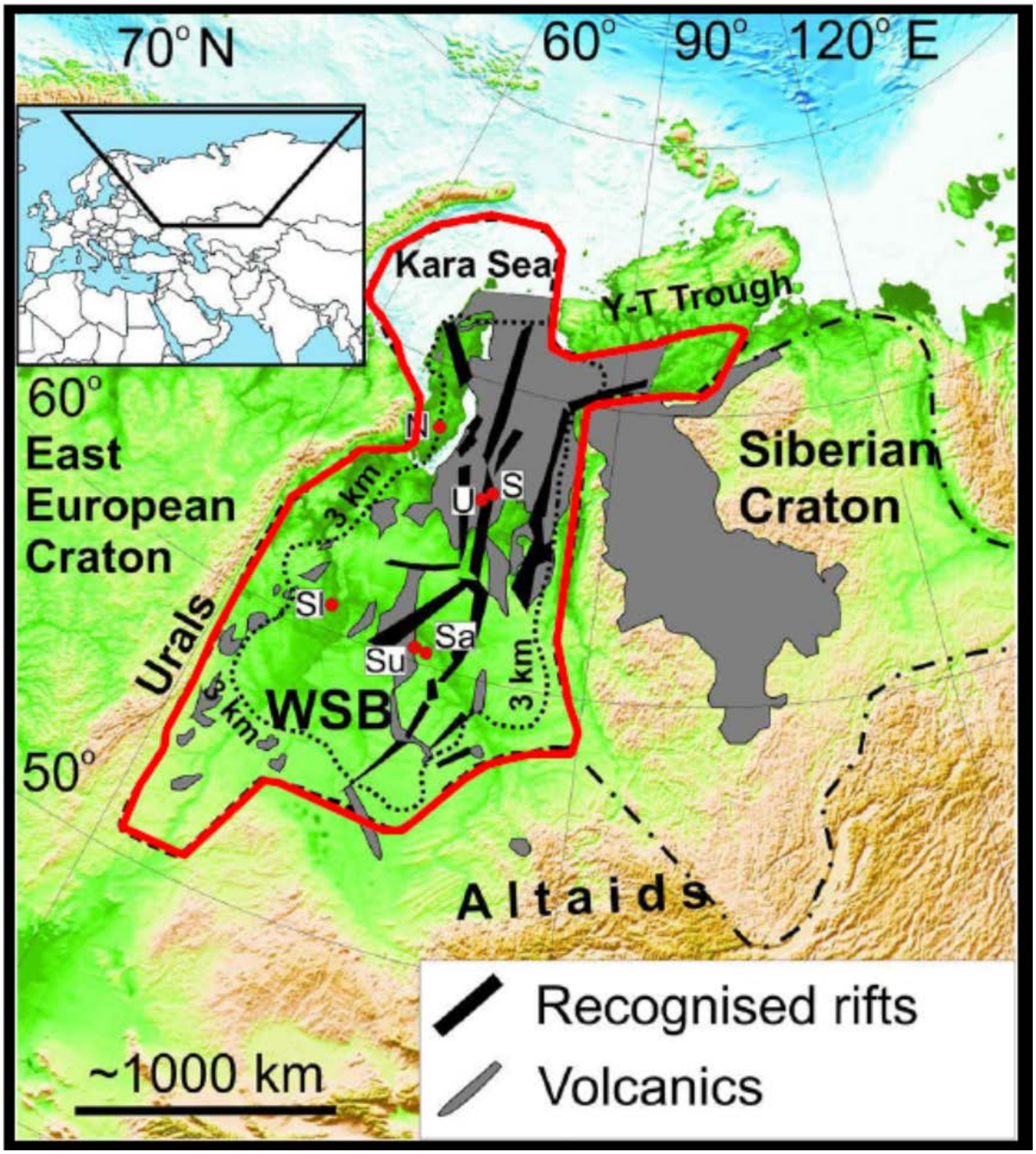


Figure 5. Location of the West Siberian Basin (red line), the 3 km sedimentary thickness contour (dotted line) and the neighboring Siberian Craton (dot dash line). Wells: N=Novoporto-130, S=SG-6, Sa=Samotlar-39, Sl=Salym-184, Su=Surgut-51, U=Urengoy-414, Y-T Trough=Yenisey-Khatanga Trough. Modified from Holt et al. (2012).

The basement beneath the basin is formed from segments of oceanic crust, Proterozoic micro-continents, and other terraces (Saunders et al., 2005). There was localized deposition of Paleozoic sediments over the more stable micro-continents (Peterson and Clarke, 1991). At the end of the Permian (250 Ma) the stress affecting the basin area changed from compressional to extensional, possibly with a component of right-lateral shear between the Siberian craton and Baltica (Allen et al ., 2006). This led to rifting within the basin which was coeval with eruption of the flood basalts. Vyssotski (2006) observed that the rifts are more pronounced and prevalent in the north but are subtle to almost uncertain in the south. He therefore divided the basin into a rifted zone and a passively subsiding zone using the 64° N line of latitude. Figure5 shows the locations of the rifts, and the relation between the basin and the flood basalt province. Rifting continued during the eruption of the flood basalts between 234 and 248 Ma (Saunders et al., 2005).

CHAPTER 3: HYPOTHESES FOR THE FORMATION OF CRATONIC BASINS

Numerous hypotheses have been postulated to explain the long phase of subsidence of cratonic basins. Since the subsidence and tectonic history of these basins differ widely in terms of duration, pattern, and formation period in geologic time (as discussed in Chapter 2), and the lack of modern active cratonic basins (with the exception of the Congo and Chad basins), there is active debate on their formation mechanism(s) (Hartley and Allen, 1994). A hypothesis that will explain the origin of these basins must be consistent and account for their long subsidence history. A major discussion concerns the continuity of the subsidence history. Fowler and Nisbet (1985), Hamdani et al. (1994), and Ahern and Mrkvicka (1984) suggested that subsidence is continuous throughout the basin's history. Other workers concluded that subsidence is episodic and not continuous. For example, DeRito et al. (1983) predicted that the subsidence was episodic and Sleep (1971) observed that major gaps in the cratonic basins record may have been caused by eustatic changes in the order of 100 m. Other studies have suggested that cratonic basins originated during the breakup of a late Precambrian continent (Klein and Hsui , 1987). In this chapter I will summarize the hypotheses for cratonic basin formation.

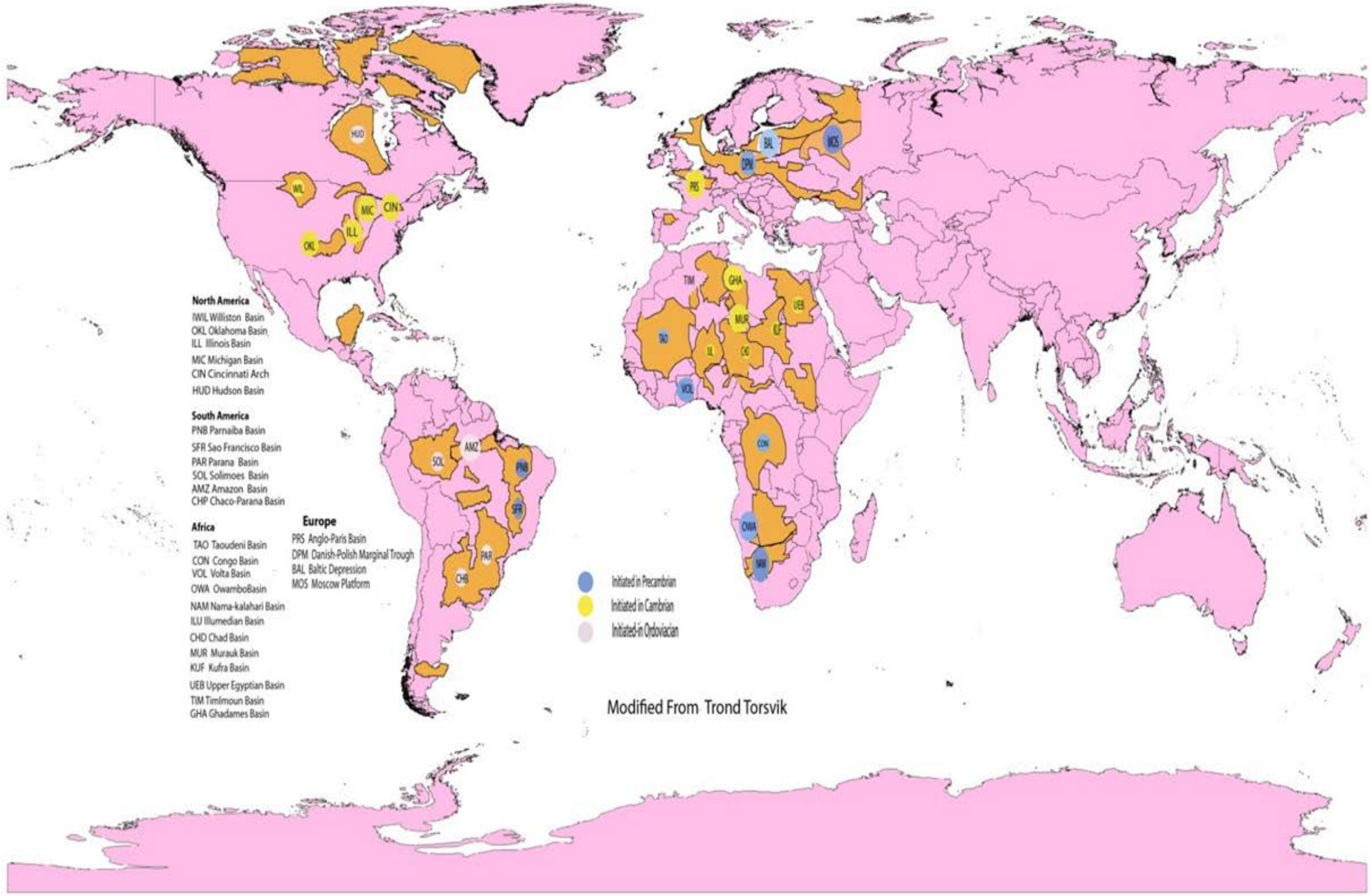


Figure 6.Cratonic basins with their different ages of formation which are coeval with the breakup and assembling of supercontinents. Redrawn from Busby and Perez (2012)

3.1. CRATONIC BASINS FORMED BY A LOAD IN THE LITHOSPHERE

DeRito et al. (1983) suggested that cratonic basins are the result of flexural deformation of the lithosphere. They noted that cratonic basins are usually located above ancient rift sites, characterized by positive linear Bouguer anomalies. These positive Bouguer anomalies indicate excess mass in the lithosphere, which may reflect magmatic intrusions during the rifting phase. Examples include the Chad basin (Burke, 1976), Illinois basin (McGinnis, 1970), and Michigan basin (Haxby et al., 1976). DeRito et al. (1983) inferred that the link between ancient rifts and cratonic basin may be related to structural and mass changes that were produced by rifting such as magmatic intrusions. They divided the evolution of rift basins into four stages: (a) the rift valley stage (e.g., East African rift system), (b) the youthful stage, (c) the mature stage, and (d) the tectonic subsidence stage. During stages a and b the crust and lithosphere are altered by mass and thermal changes (i.e. magmatic intrusions, heating of the lithosphere) which they believe to cause the mass excess in the crust as shown by a sharp positive Bouguer anomaly.

DeRito et al. (1983) modeled the subsidence of cratonic basins by a lithospheric flexure model with a nonlinear Maxwell viscoelastic rheology. They suggested that the mass excess (basaltic intrusion) may act as a load if it is not isostatically compensated. They suggested that isostatic compensation may take many millions of years, and as long as the basin is not isostatically compensated, flexural deformation will continue. When the load ceases to exist, flexural deformation will come to a halt. McGinnis (1970) proposed that when such mass exists, it results in a gravity anomaly.

Their models show that the lithosphere responds to uncompensated intrusions in three stages: an initial stage of instantaneous elastic subsidence, followed by a period of rapid subsidence, and finally a period of very slow subsidence. As expected, subsidence ceases after the mass excess is compensated. The basin may enter another phase of subsidence if reactivation occurs via an increase in geothermal gradient or compressional stress in the tectonic plate. The increase in the geothermal gradient may be as a result of an increase in heat flow from the upper mantle. They concluded that a continental or worldwide phenomenon in heat flow may have been responsible for the reactivation of simultaneous subsidence of cratonic basins.

DeRito et al.(1983) and Downey and Gurnis (2009) suggested in their study of the Congo Basin that the large gravity anomaly in this basin is related to a large positive upper mantle shear wave velocity anomaly. They did not consider pure flexural deformation, but suggested that a downward dynamic force on the lithosphere resulting from this high density object within the mantle lithosphere was responsible for the anomaly. The high-density body is located at 100 km depth within the mantle lithosphere, and has a density anomaly of 27-60 kgm^{-3}. Also Downey and Gurnis (2009) proposed that the anomalous masses became unstable/were formed during a global tectonic event, which is similar to Derito et al's(1983) suggestion that they arise from a global increase in geothermal gradient.

Hartley and Allen (1994) also explored the Congo Basin. They suggested that the Bouguer anomalies occurred because of the presence of a deep, cold region in the mantle lithosphere beneath the basin. They also proposed a dynamic downward –acting force on the base of the lithosphere, a characteristic of convective downwelling.

Heine et al. (2008) studied over 200 cratonic basins using an analytical flow model and global plate kinematics over the last 70 Ma. They tracked the subsidence of basins and observed that anomalous subsidence (subsidence induced by the mantle flow) occurred due to basins drifting away from regions of negative dynamic topography. They related these events to continental dispersal of the Pangaea supercontinent and proposed that deep process in the deep-earth may be responsible for subsidence of the basins.

3.2 SUBSIDENCE OF CRATONIC BASINS DRIVEN BY PHASE CHANGE AND THERMAL PROCESSES

Hamdani et al. (1994) investigated the subsidence of the Williston Basin and focused on evidence that the crust is very thick, with high seismic wave velocities in the lower crust and upper mantle, and with very little thinning of the crust. They suggested that this proves that the Williston Basin was not formed as a result of stretching and rifting, and that subsidence is not due to thermal contraction as a result of rifting, but rather as a result of mantle plume activity and phase changes in the lower crust (Sleep, 1971; Armitage and Allen, 2010). This in contradiction with the subsidence records of some other cratonic basins (the Michigan, Illinois and Hudson basins) which had experienced accelerated subsidence after they were initiated, pointing at rifting (Nunn and Sleep, 1984). Hamdani et al. (1994) concluded that the Williston Basin was formed as a result of subsidence in conjunction with a phase change in the lower crust that resulted from mantle plume activity followed by a cooling and thermal contraction phase. They believe that a change in heat flow at the LAB (lithosphere –asthenosphere boundary), which is assumed to be associated with the Late Proterozoic breakup of the supercontinent (Klein and Hsui, 1987), is responsible for the phase change. The change in heat flow at the LAB would result from an increase in mantle plume activities below the supercontinent. This led to the penetration of the asthenosphere and lithosphere by mantle plumes, weakening and heating the lithosphere beneath the basin to experience an increase in temperature compared to its flanks. After the heating phase, cooling would result, and thermal contraction and a phase change, resulting in a long period of subsidence. They concluded that their model could account for the multiple subsidence phases of cratonic basins. A change in the heat flow boundary conditions explains the long duration of subsidence, and the delay of phase change subsidence explains the acceleration of subsidence in the early stages of the basins evolution.

Xie and Heller (2009) studied subsidence curves of some cratonic basins (Figure 7) and noticed that they are exponential in shape similar to passive margins but generally lacking a rapid initial subsidence phase. They saw a consistency match when compared with seafloor subsidence, but with a longer decay constant. This led some authors to suggest that thermal decay was the process responsible for the subsidence of cratonic basins. Xie and Heller (2009) also observed that a thicker lithosphere (as found below cratonic basins) is normally associated with a longer decay constant to reach a thermal equilibrium, explaining the long subsidence histories. They concluded that various interacting tectonic mechanisms may be responsible, including multiple thermal permutations or changes in the rheology of the lithosphere.

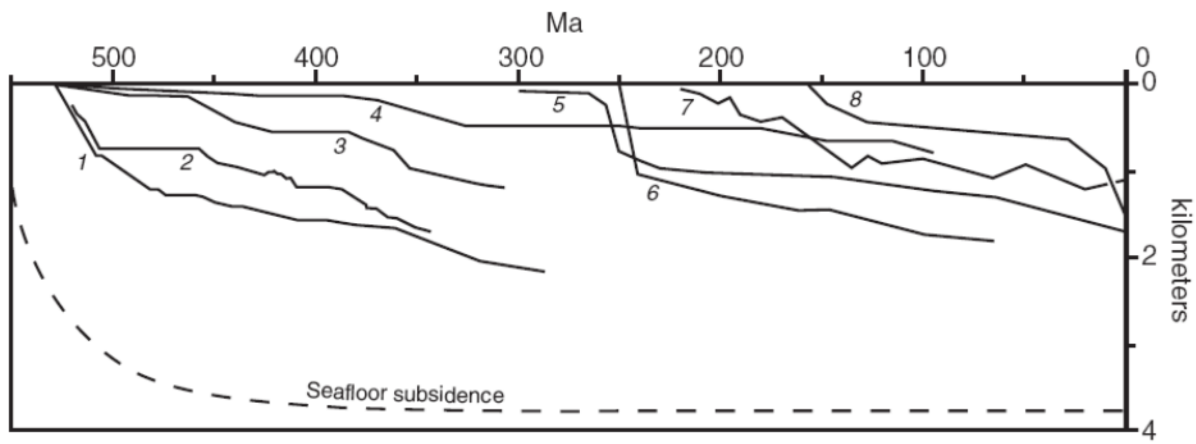


Figure 7. This figure shows the tectonic subsidence of cratonic basins. Taken from Xie and Heller (2009).

3.3 LOCALIZED EXTENSION RELATED TO MAGMATIC UPWELLING

In North America, subsidence of the Williston Basin started around 540-500 Ma (Fowler and Nisbet, 1985; Ahern and Ditmars, 1985). Around the same time (~520-460 Ma), the Michigan Basin started subsiding (Sleep and Sloss, 1978), and subsidence of the Illinois Basin started around 520-510 Ma (Heidlauf et al., 1986). Paleozoic basins in Africa, South America, and Europe started forming at around this same time (Bond and Kominz ,1984), Figure 6. According to Klein and Hsui (1987), these basins have more common characteristics: they have similar sediment accumulation histories (sequences, temporal trends in thickening) and the timing of major interregional unconformities corresponds. Klein and Hsui (1987) therefore propose a common formation mechanism for these basins (Figure 6).

They suggest that during supercontinent periods, the continent acts as a heat lens, trapping heat below the lithosphere. As a result partial melting of the upper mantle and lower crust may have been increased. As a result of the lower crustal melting, the remainder of the crust would be thinned (Klein and Hsui, 1987). These granitic intrusions regionally modified the rheology of the crust, weakening it and focusing rifting during the time of continental lithosphere extension when the supercontinent started to break up. The supercontinent-wide granitic intrusions and extension of the lithosphere would result in an almost contemporaneous onset of cratonic basin formation above the intrusions. The cratonic basins would thus form above the intrusions by thermal cooling after rifting ended. The timing of the onset of cratonic basin formation would correspond with the timing of the Pre-Cambrian supercontinental break up.

This mechanism is illustrated in Figure 8. Klein and Hsui (1987) suggested that the basin subsidence occurred in different stages (Figure 8 A,B,and C). Before continental breakup (stage A), during the supercontinent stage, there was limited heat loss and the continent acted as a heat lens. Partial melting of the asthenosphere, mantle lithosphere,and lower crust occurred, emplacing granites in the crust and thereby thinning the crust due to the melting. Positioning of anorogenic granites changed the physical properties of the crust (rheology, thermal structure) (Figure 8B), and weakening of the crust. During this phase the crust is under extension as the supercontinent starts to rupture. This was followed by the breakup of the continent which was accompanied by extension, and rifting was localized to zones where anorogenic granites were emplaced. These locations became the sites for rift - related basaltic intrusions (Figure 8C). In summary, they hypothesized that the location of cratonic basins was driven by global partial melting and intrusions of granites during the Precambrian supercontinent breakup. This may have been responsible for synchronicity of interregional unconformities, and the thermal decay that occurred after the emplacement of anorogenic granitic, and basaltic volcanics accounted for the thermal subsidence.

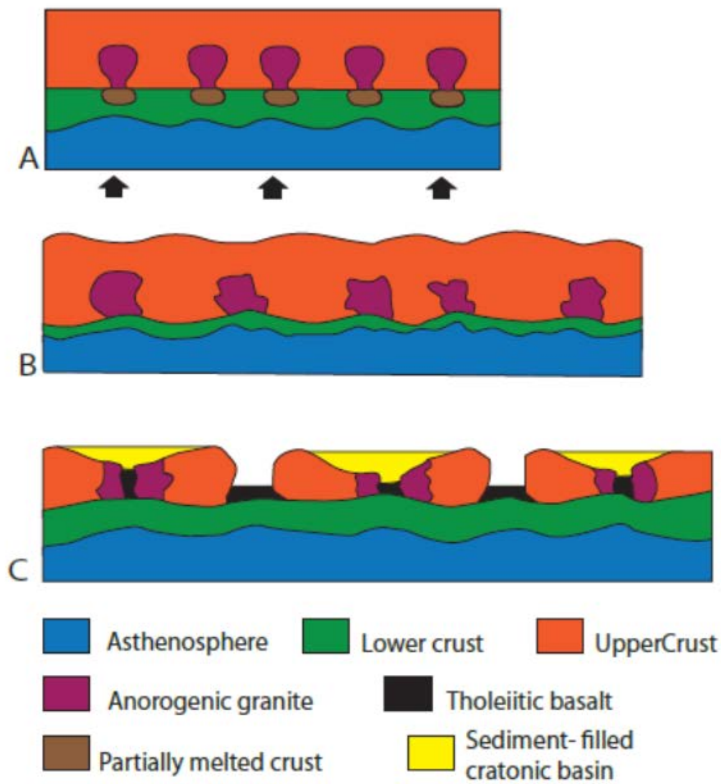


Figure 8.Three evolutionary phases of cratonic basin formation redrawn from Klein and Hsui (1987). A) emplacement of granites resulting from partial melting below the supercontinent; B) extension of the supercontinent results in rift localization and onset of subsidence above the intrusions; C) supercontinental breakup and contemporaneous formation of cratonic basins on several continents.

3.4 DOWNWELLING OF A MANTLE PLUME

Middleton (1989) proposed that a downgoing mantle plume may drive the formation of a cratonic basin. This hypothesis explains the initial sag in which the basin is formed, cooling without initial heating at the base of the lithosphere (thermal subsidence), and uplift at the end of the subsidence phase. An example of a cratonic basin formed by this process is the Canning Basin in Western Australia.

The descending plume develops within the mantle convection system (Figure 9), and the downwelling plume is associated with the formation of a depression at the surface. This initial depression can be of the order of 600 m, and is caused by dynamic topography. If filled with sediments, this initial depression may grow to a basin about 2.5 km deep. The basin continues to subside as long as the downwelling flow continues. Cooling above the downwelling may result in thermal subsidence following this initial phase, for a period of 50-100 m.y. Subsidence will come to a halt and uplift may occur when the convective downwelling changes and the downgoing plume is removed. Then, a phase of uplift and erosion may occur.

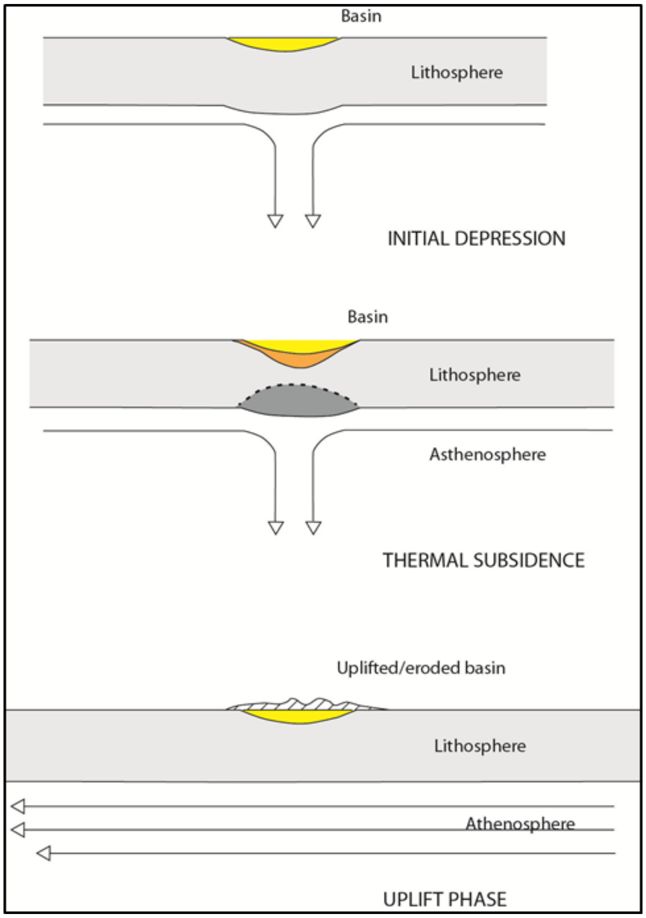


Figure 9.Descending plume model redrawn from Middleton (1989). An initial depression is formed above a downwelling. Thermal subsidence above the downwelling results in continuing subsidence. A change in the convective pattern is followed by uplift and erosion

3.5 LONG PHASE OF THERMAL SUBSIDENCE FOLLOWING SLOW EXTENSION OF THE LITHOPSHERE

Armitage and Allen (2010) investigated the long thermal subsidence phase following minor continental rifting. According to their hypothesis, cratonic basins are formed by slow cooling of the lithosphere following a long and slow rifting episode. They tested this hypothesis using a 1D forward modeling approach. They kinematically modeled lithosphere thinning using a Precambrian lithosphere of about 200 km in thickness. Stretching under slow extension and strain rates (10^{-16} s^{-1}) caused in their models permanent, long-lived thermal subsidence of cratonic basins.

Armitage and Allen (2010) suggested that at low strain rates the subsidence history departs from the uniform instantaneous stretching model (Figure 10) and that heating due to upward advection is countered by thermal diffusion. After stretching ends, the basin starts a slow thermal subsidence phase. Their models were one-dimensional and could not account for the heat loss in the horizontal directions. They may have overestimated the timing of cooling. They concluded that Cratonic basins belong to the extreme part of the rift-drift suite as suggested by Allen and Allen (2005), Figure 10.

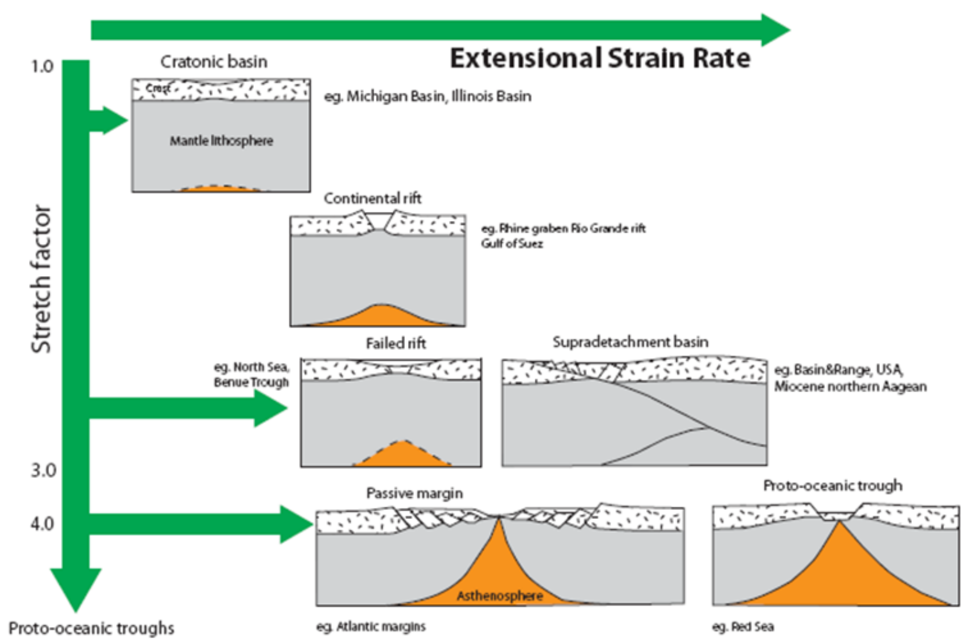


Figure 10. The rift-drift-suite of cratonic basins redrawn from Allen and Allen (2005).

CHAPTER 4: METHODOLOGY

The “slow extension” hypothesis for the formation of cratonic basins explains the long phase of subsidence by cooling and thermal contraction following a slow, long phase of rifting. To test this hypothesis, I used a two-dimensional numerical modeling approach to study the thermal evolution and deformation of the lithosphere under and after extension. Armitage and Allen (2010) used a one-dimensional modeling approach that did not account for lateral conductive heat loss. The two-dimensional finite element code was developed academically and has been used in previous studies (e.g., van Wijk et al., 2001; van Wijk and Cloetingh, 2002). The code is based on the Lagrangian formulation, which means that the finite element mesh deforms when the lithosphere deforms. This facilitated tracking of boundaries (such as the Moho) in time and space. A remeshing routine is included to facilitate large grid deformations.

The models describe visco-elastic deformation of the lithosphere, with a parameterization for brittle behavior. Approximately the upper half of the thermal lithosphere behaves elastically on geological time scales while in the lower half stresses are relaxed by viscous deformation. This behavior is described in the finite element model by a Maxwell body. Density in the models is temperature dependent, i.e. buoyancy forces are included. A parameterization for brittle behavior is included following the Mohr-Coulomb criteria, but there is no slip along the faults which infers that when the yield criterion is attained, stresses are adjusted. The heat flow equation is solved to obtain temperature. There is heat production in the crust.

The model domain is two –dimensional, and represents a vertical slice through the lithosphere (Figure 11). The model domain is 1000 km wide and 125 km deep, divided into 2560 triangular shaped elements. The temperature at the base of the model is 1333°C, and at the top 0°C. The heat flow through the sides of the domain is zero. The model runs start from thermal equilibrium of a 250 m.y. old lithosphere. Velocity boundary conditions are applied to the left and right sides of the model domain to extend the lithosphere. The velocity was varied for the different tests. In the models, the lithosphere is extended for a certain period of time (for example, 50 m.y.), after which extension stops and the lithosphere cools.

Rheological and thermal parameters that were used are shown in Table 1. The model domain is partitioned into an upper crust (granite), lower crust (dolerite) and mantle lithosphere (peridotite).

Parameter	**Value**
Density	2700(u.c.), 2800(l.c.), 3300 (m.l.) [kg/m3]
Crustal heat production	$1*10^{-6}$ [W/m3]
Specific heat	1050 [J/kg/K]
Conductivity	2.6(crust), 3.1(mantle) [W/m3]
Bulk modulus	$3.3*10^{10}$ (crust), $12.5*10^{10}$(mantle)[Pa]
Shear modulus	$2*10^{10}$ (crust), $6.3*10^{10}$ (mantle) [Pa]
Power law exponent n	3.3 (u.c.), 3.05 (l.c.), 3.0 (m.l.)
Activation energy Q	186.5 (u.c.), 276 (l.c.), 510 (m.l.) [kJ/mol]
Material Constant A	$3.16*10^{-26}$ (u.c.), $3.2*^{10-20}$(l.c.), $7.0*10^{-14}$ (m.l.) [Pa]
Friction angle	30°
Dilatation angle	0°
Cohesion factor	$20*10^{6}$ [Pa]
(Initial) thickness of lithosphere	125, 150, 200 [km]

Table 1. Rheological and thermal parameters used in this study. The parameters are from Ranalli (1995) and Turcotte and Schubert (2002). u.c. (upper crust), l.c. (lower crust), m.l. (mantle lithosphere).

In the extension models, a rift zone will form where the lithosphere is weakest. The lithosphere consists of a strong mantle layer and a weaker crustal layer. One mechanism to decrease the lithospheric strength locally in geodynamic models is by increasing the crustal thickness. In this way, strong mantle lithosphere material is replaced by weaker crust material. In my models the crust was thickened (and the lithosphere weakened) in a zone several hundreds of kilometers wide in the center of the model domain (see Figure 11). Upon extension, a rift zone forms in the models in the center of the model domain.

This initial setup was the same for all tests, except for a series of tests in which the thickness of the lithosphere was varied. In those tests, the model domain was different (up to 200 km deep), and the crust was thicker.

When the rift zone is formed, the lithosphere starts to thin, which is visible in the models by upwelling warm asthenosphere material. Also the crust thins, and a depression forms at the surface. I performed a series of tests in which the extension velocity, duration of rifting, and thickness of the lithosphere were varied. The results are presented in Chapter 5.

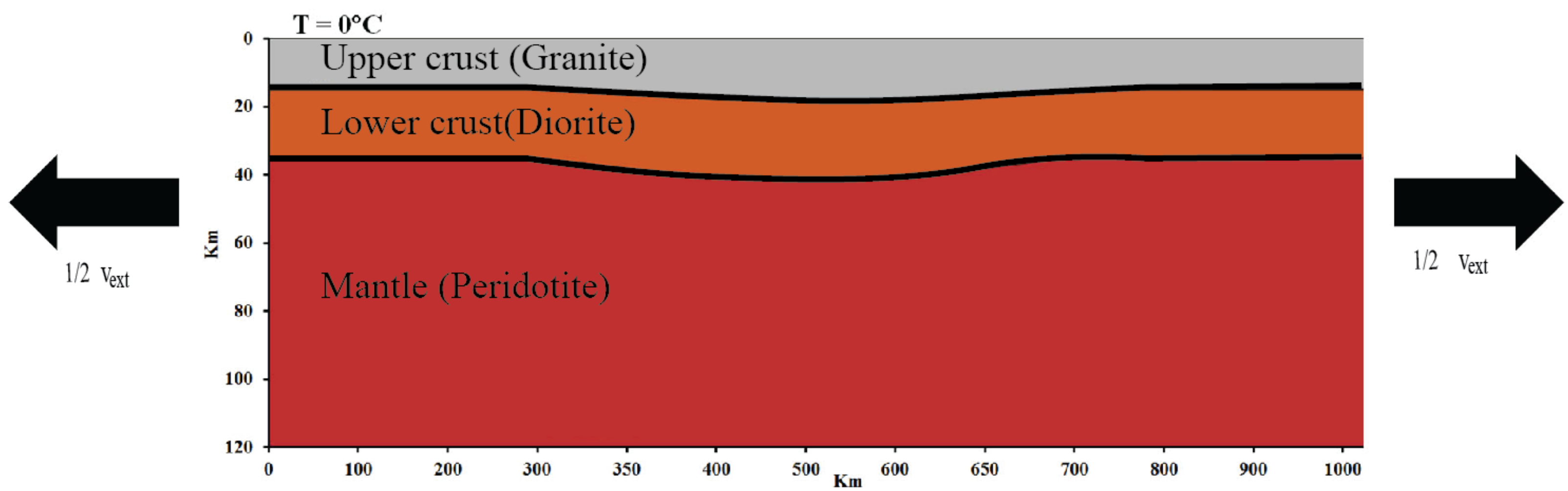


Figure 11. Initial model configuration used for the extension

CHAPTER 5: RESULTS

Results presented here illustrate the deformation of the lithosphere as a result of relatively slow extension of continental lithosphere. The extension rate varied in different tests from ~1 mm/yr to ~3 mm/yr. The initial thickness of the lithosphere was varied as well, between 125 km, 150 km, and 200 km. The tests conducted have a similar model setup that was discussed in Chapter 4. The results are shown at different times during and after rifting to observe the deformation of the lithosphere, its thermal evolution, and subsidence patterns.

Four models are presented. Since the left and right sides of the model domain are far away from the rift zone, it is assumed that horizontal size of the model domain does not have an influence on the results.

The model tests build on the previous investigations carried out by Armitage and Allen (2010). My investigation differs in two important aspects. First, the heat loss in both the horizontal and vertical directions has been accounted for in my two-dimensional models, while Armitage and Allen (2010) used one-dimensional models. Second, the lithosphere thickness was varied with different values primarily to see the response to the changes in stress and extension rate. Effects of sedimentation and erosion are not considered in the models used for this thesis, although the type of sediments deposited in the basin may influence the subsidence rate, as for example, in the case of anhydrites which make basins subside rapidly (personal communication with Kevin Burke, 2013).

In Model 1 the lithosphere thickness was 125 km and the extension rate 3 mm/year. In Model 2 the lithosphere thickness was 150 km and the extension rate 3 mm/yr, and in Model 3 the lithosphere thickness was 200 km and the extension rate 3 mm/yr. The fourth model has a lithosphere thickness of 125 km but an extension rate of 1mm/year. Within the context of lithosphere deformation and the thermal evolution, velocity and strength profiles are plotted, as well as thermal structures and the stress fields. The lithosphere was slowly stretched with a given constant velocity rate for 50 m.y. which resulted in a β-factor of 1.3-2.0, corresponding to the rift-drift suite of cratonic basins (Figure 3) as shown by Allen and Allen (2005). The initial thermal profiles of models 1,2, and 4 are shown in Figure 12.

It is assumed that a prerequisite for cratonic basins formed by slow stretching is a concentration of tensional stresses or a zone of weakness in the continental lithosphere (England, 1983) so that rifting will localize. In my models the crust was thickened in the center of the domain, which resulted in a weaker lithosphere in this location and the formation of a rift zone under extension. If extension is very slow (England, 1983), as is the case in model 4, it is observed that the lithosphere gains strength through the extension phase, and basin formation comes to a halt. This will be discussed later.

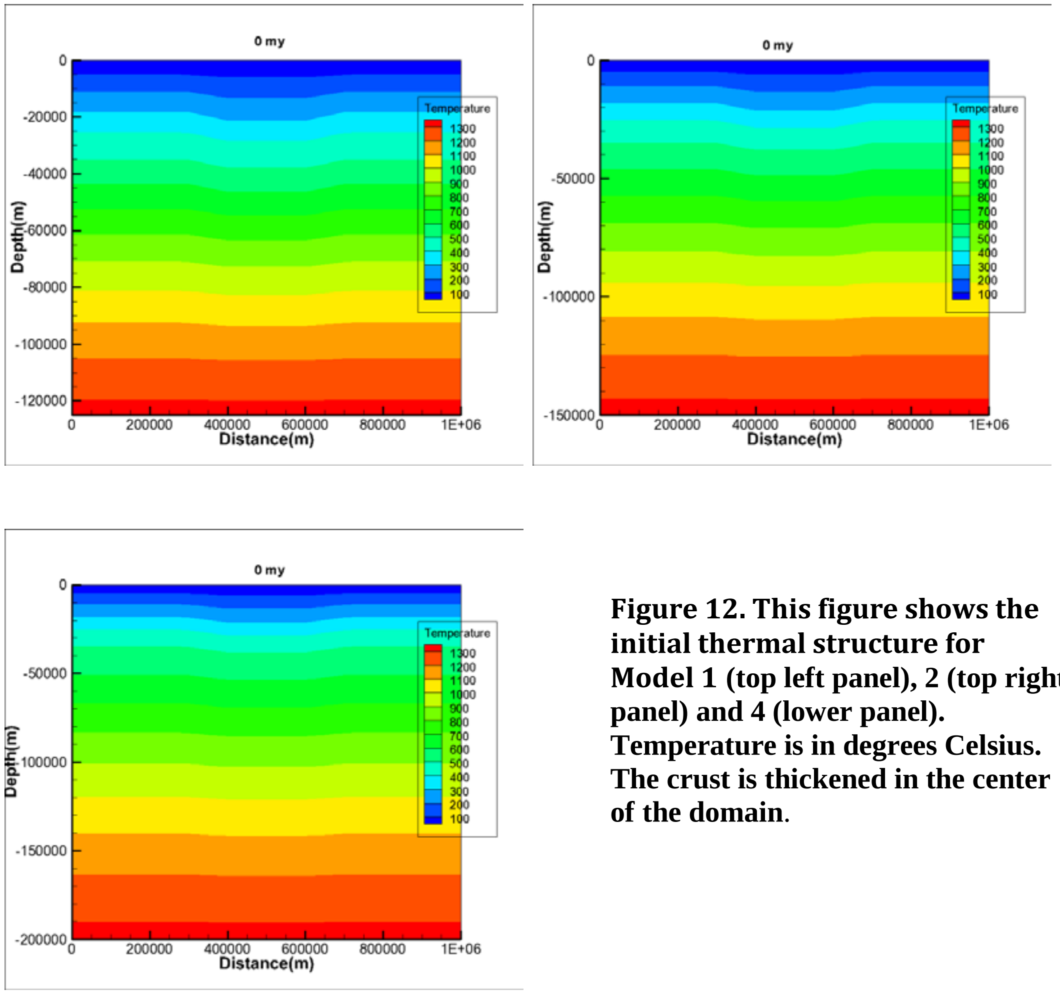
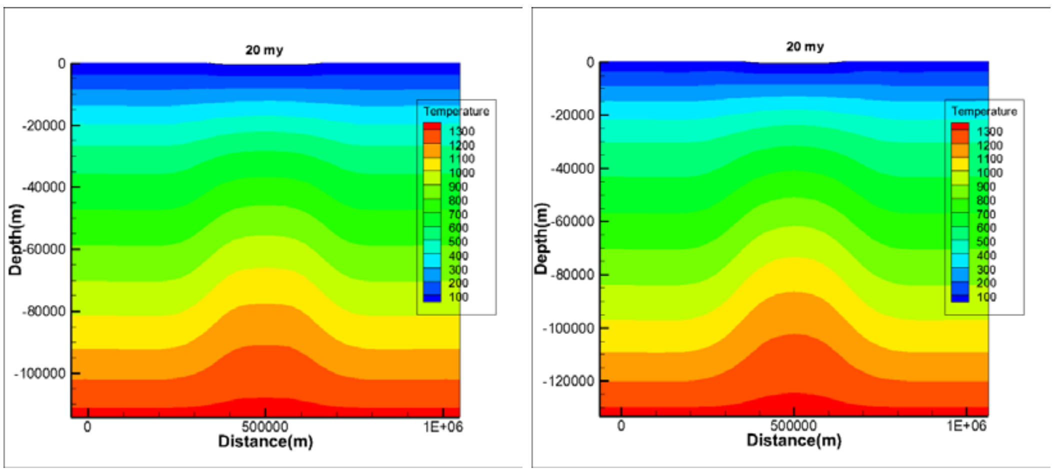


Figure 12. This figure shows the initial thermal structure for Model 1 (top left panel), 2 (top right panel) and 4 (lower panel). Temperature is in degrees Celsius. The crust is thickened in the center of the domain.

5.1. THERMAL EVOLUTION

Upon extension of the lithosphere, a rift is formed in the center of the model domain. Here, the lithosphere is thinned, and warm asthenosphere material wells up (Figure 13). Figure 13 shows the thermal structure after 20 m.y. of extension. A comparison of Figure 13 upper left panel with Figure 13 lower panel shows that a higher extension rate results in more asthenopshere upwelling below the rift.



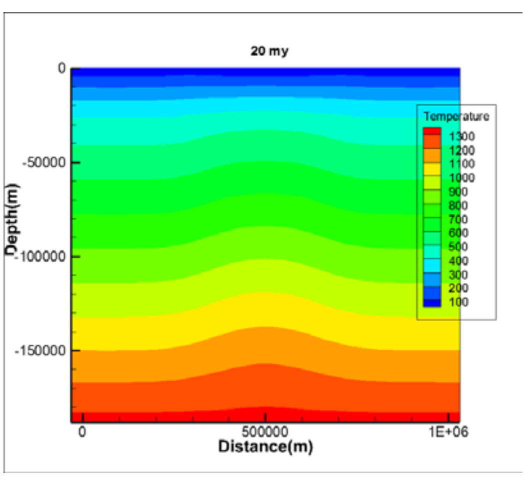
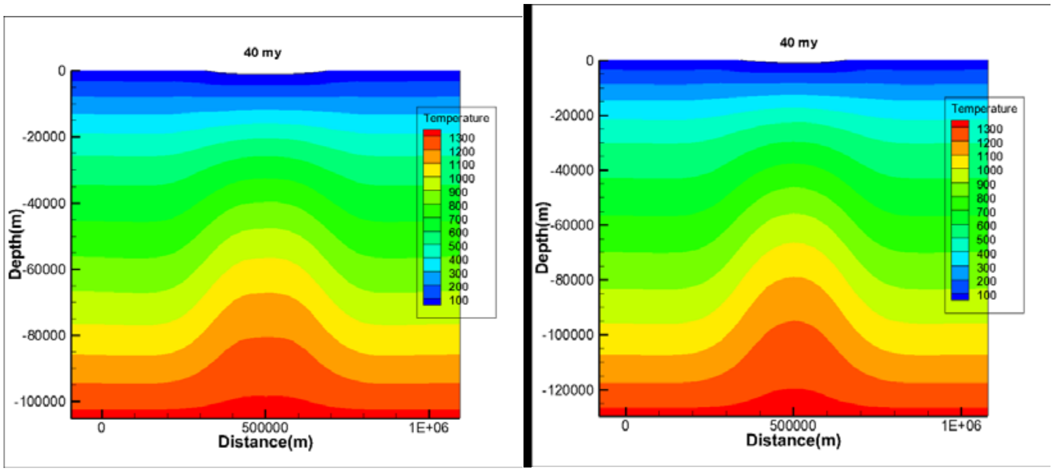


Figure 13. Passive asthenosphere upwelling after 20 m.y. of rifting below the rift is pronounced in Models 1 and 2 (upper panels left and right). In Model 4 (lower panel) the upwelling is very limited.



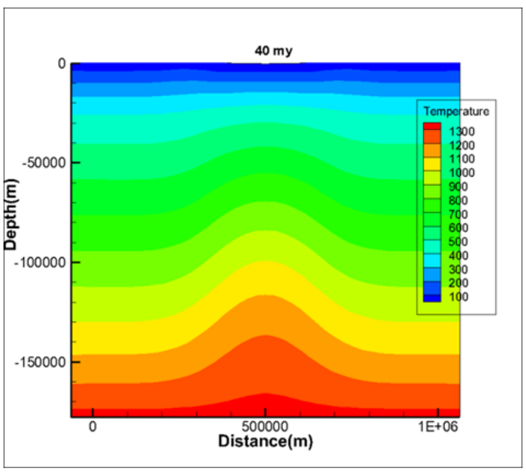
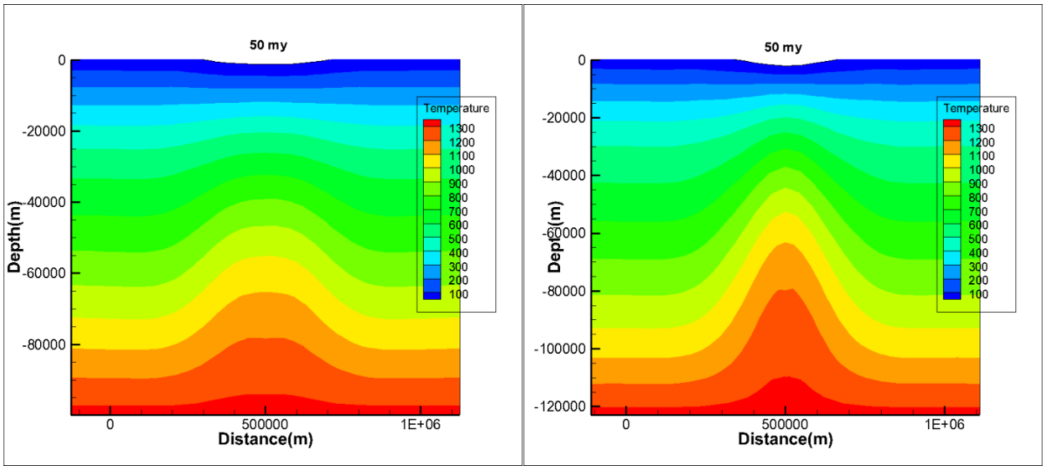


Figure 14. Passive asthenosphere upwelling after 40 m.y. of rifting below the rift is pronounced in Models 1 and 2 (upper panels left and right). In Model 4 (lower panel) the upwelling is less. Temperature in °C.

When extension of the lithosphere continues, the lithosphere progresses to thin and warm asthenosphere wells up. After 40 m.y. of extension the 1200°C isotherm is found at about 90 km depth below the rift in Models 1 and 2, and at about 140 km depth in Model 3 (Figure 13, 14, and 15).



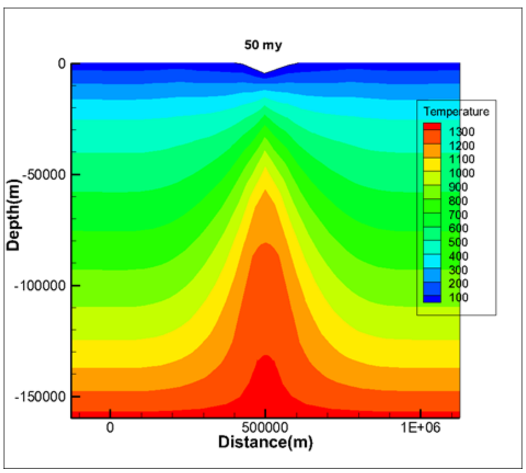


Figure 15. Passive asthenosphere upwelling after 50 m.y. of rifting below the rift is pronounced in all Models. Models 1 and 2 (upper panels left and right); Model 4 (lower panel). Temperature in °C.

Extension of the lithosphere stops after 50 m.y. of rifting. At this point in time (Figure 15), the mantle upwelling below the rift is at its maximum. After 50 m.y. of extension the boundary conditions of the model change and the extension velocity is set to zero. The model is then run for another 100-200 m.y. to study the thermal subsidence phase.

In Model 3 the thickness of the lithosphere was 200 km. Such a lithosphere thickness has been observed below the Congo (Buiter, 2012) and Western Siberian (Saunders et al, 2005) Basins. Here, thinning of crust became noticeable at ~20 m.y., and the rift was very localized after about 50 m.y. of extension. The shape of the basin was deep and narrow compared to that of a lithosphere thickness of 125 km.

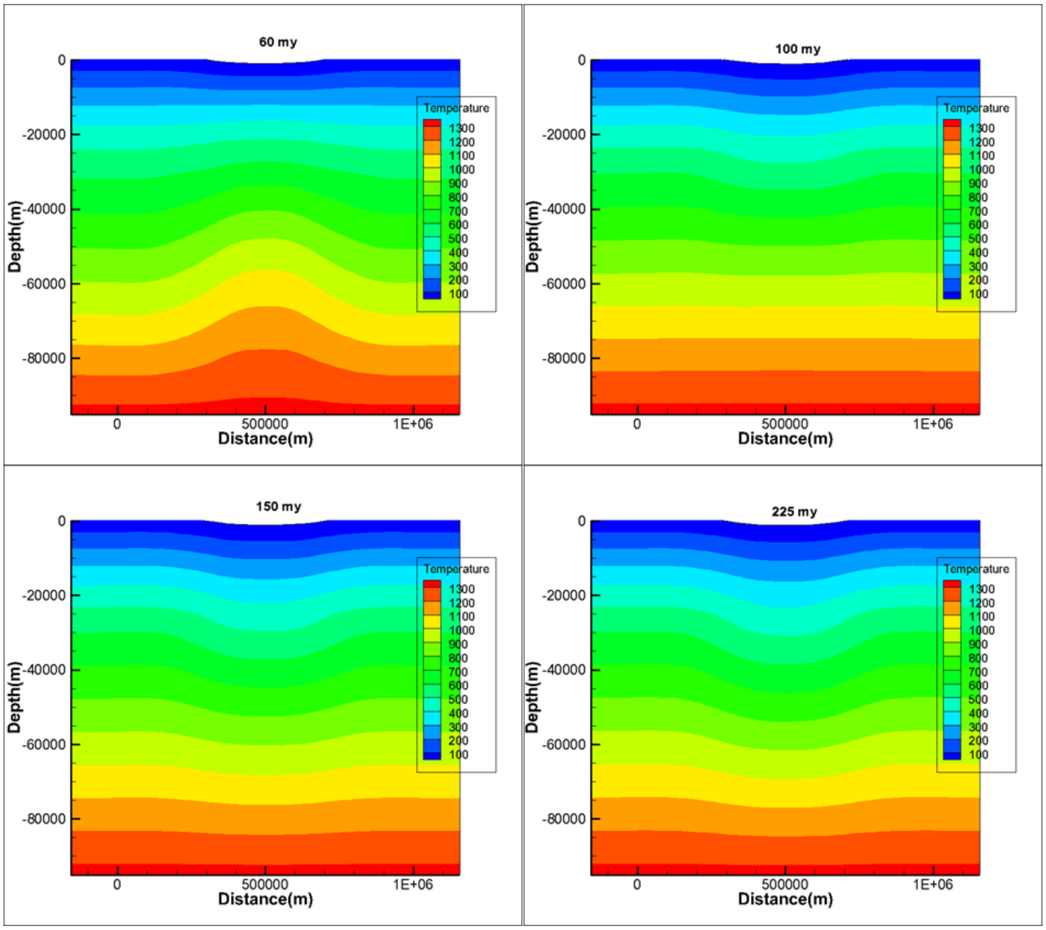


Figure 16. Model 1. The panels show the thermal evolution of the model after rifting Upper panel left: 60 m.y. (this is 10 m.y. after rifting stopped), upper panel right: 100 m.y., lower panel left: 150 m.y., lower panel right: 225 m.y. Temperature in °C.

During the first 20 m.y. after the onset of the stretching, rifting localized in the zone of weakness as observed previously by McKenzie (1978), Houseman and England (1986), and van Wijk and Cloetingh (2002). The upwelling was at its peak at about 40 m.y. in Model 1 with the shape of the upwelling becoming broader and waning at/after 50 m.y.

When extension of the lithosphere ends, rifting ends and the rifted lithosphere starts to cool (Figure 16). As a result, the density in the lithosphere increases, resulting in subsidence.

Figure 16 further shows that a long time after extension has ended, the rift basin shows up as a “cold” area in the lithosphere. This is because the heat-producing crust has been thinned, and therefore less heat is produced with as a result that the shallow lithosphere is colder. Because this is now the coldest area, this old rift zone will also be the strongest part of the lithosphere. This is discussed later.

In Model 4, with a lithosphere thickness of 125 km, there was no noticeable deformation through the slow stretching phase, as observed previously by van Wijk and Cloetingh (2002) and Houseman and England (1986). When rifting is this slow, syn-rift cooling is important and the lithosphere gains strength during rifting. This inhibits further development of the rift zone.

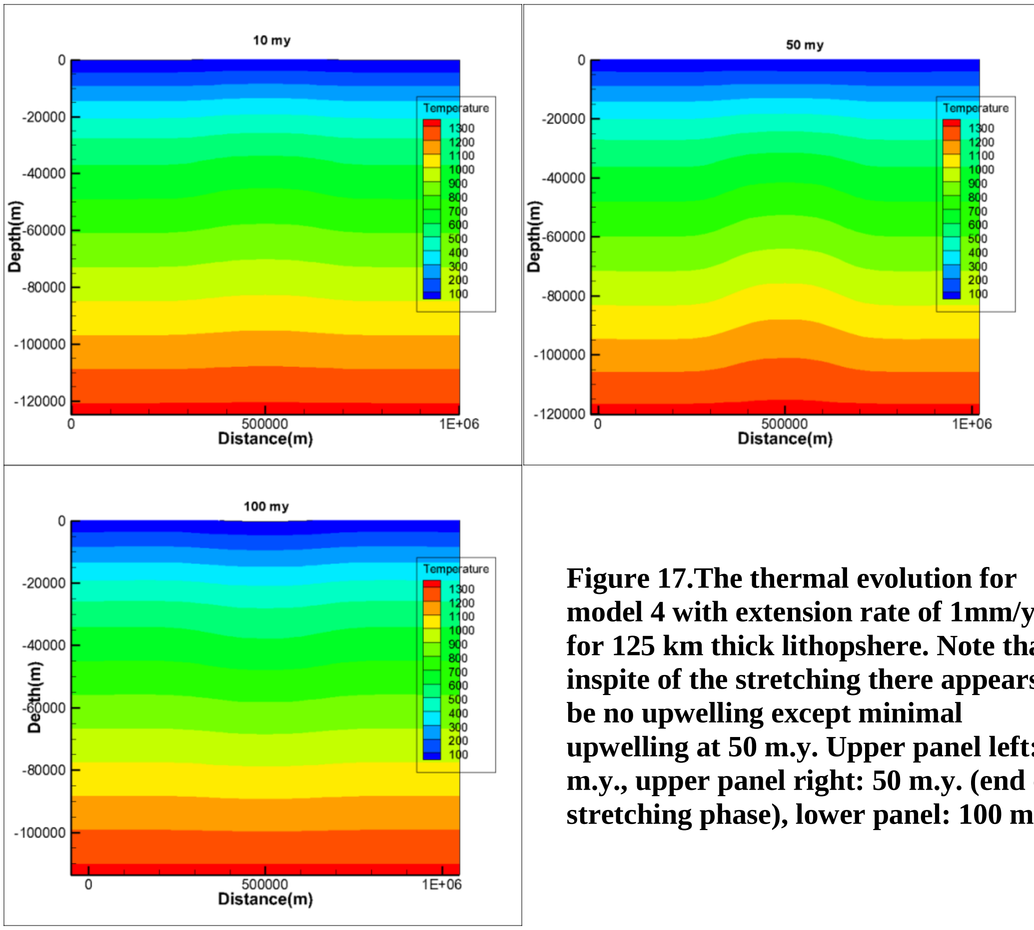


Figure 17.The thermal evolution for model 4 with extension rate of 1mm/year for 125 km thick lithopshere. Note that inspite of the stretching there appears to be no upwelling except minimal upwelling at 50 m.y. Upper panel left: 10 m.y., upper panel right: 50 m.y. (end of stretching phase), lower panel: 100 m.y.

Model 4 is similar to Model 1, except for the extension rate, which is now 1 mm/yr. This extension rate is so low that rifting does not occur (Figure 17).

5.2 CRUST AND LITHOSPHERE DEFORMATION

Upon extension the crust and mantle part of the lithosphere thin in the rift zone. Figure 18 shows that after 50 m.y. of extension, the stretching factor of the crust (ß) is 2.0 in Model 1. At this point in time the model stopped extending and the crustal thinning factor does not vary further. Also the mantle lithosphere thins (Figure 18). Mantle lithosphere thinning is not concentrated clearly in one location but is distributed in a wider zone with two maxima.

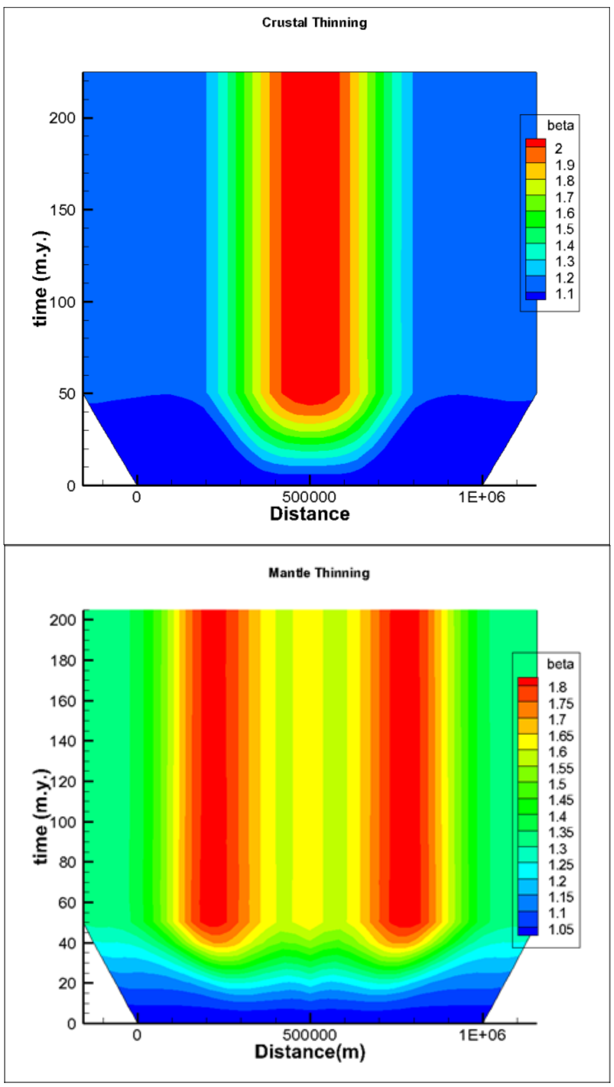


Figure 18. Evolution of crust and mantle thinning. The vertical axis represents the time in m.y., while the distance (m) or model width is depicted on the horizontal axis. The width of the model domain increases as extension progresses. Beta is the stretching factor (ratio between initial crust/mantle lithosphere thickness and present thickness).

The horizontal and vertical components of the velocity field are shown in Figures 19-22, for different time intervals. In the case of Model 1 at time interval of 20 m.y. (Figure 20), there is a zone of strong upwelling. Mantle material wells up below the rift zone. In the horizontal (vx) panels (Figure 19), I observed that below the lithosphere in the asthenosphere, mantle material moves towards the left on the right side of the center, and towards the right-on the left-hand side of the center, hence the movement of mantle material is towards the rift center in the asthenosphere. By the end of the extension phase (50 m.y.), the upwelling slows down and the velocity is one order of magnitude reduced. The vertical component of the velocity field (vy) at 70 m.y. (i.e. 20 m.y. after extension stopped) is almost reduced to zero, and the single upwelling zone is replaced by two slight downwelling zones, resulting from thermal cooling and isostatic readjustment. At 100 m.y. (50 m.y. after extension stopped) it is very obvious that the thermal decay had set with faint remains of downwelling zones.

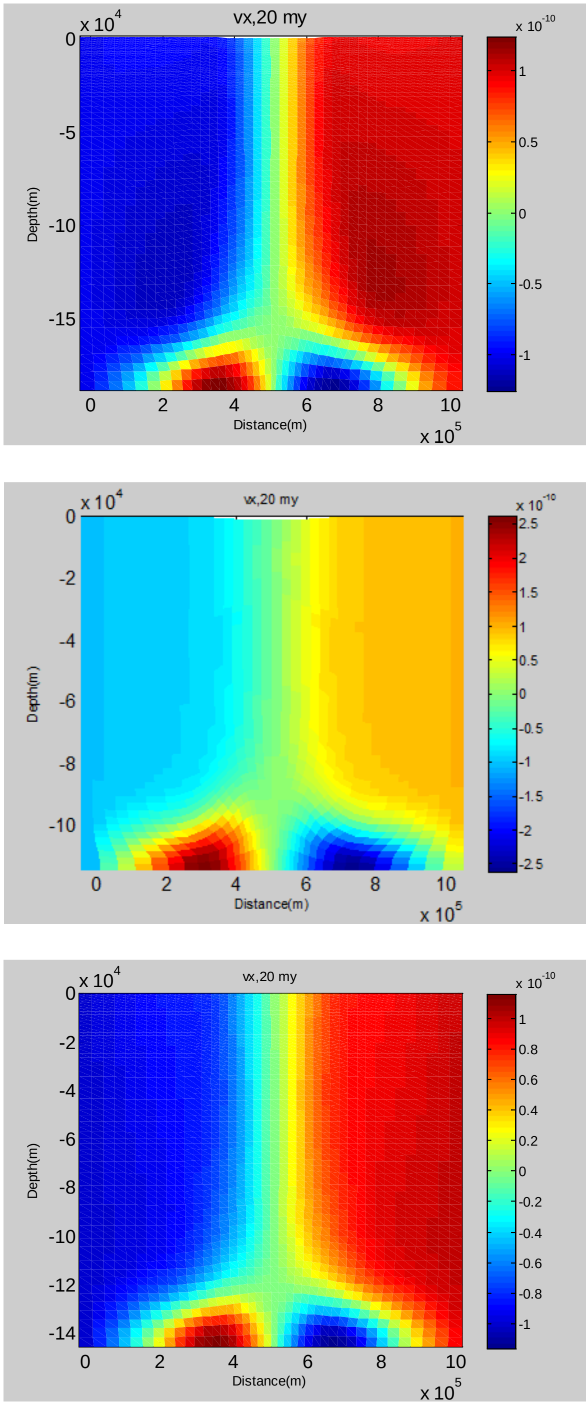


Figure 19. Horizontal component of velocity field after 20 m.y of extension. Left upper panel Model 1, right upper panel Model 2, lower panel Model 3. Velocity is in m/s. Velocities are positive (red) to the right, negative (blue) to the left.

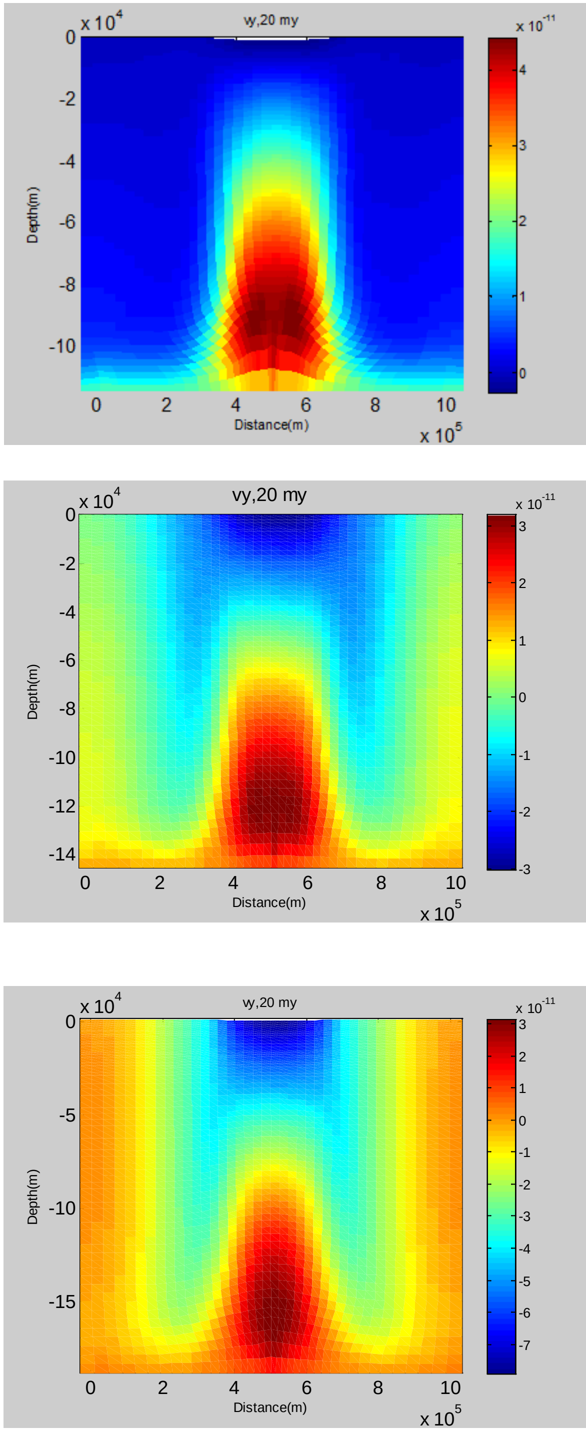


Figure 20. Vertical component of velocity field at 20 m.y for a lithosphere 125 km thick (Model 1, left upper panel), 150 km (Model 2, rigth upper panel) and 200 km (lower panel, Model 3). Velocities are positive upward (red). Velocities in m/s.

5.3 STRESSES AND STRENGTH OF THE LITHOSPHERE

The strength of the lithosphere is a function of composition, rheological and compositional layering, and temperature. Figure 21 shows the horizontal deviatoric stresses (σ_{xx}) for Model 1 at time 20 m.y. The lower crust is weaker than the upper crust, and the mantle just below the Moho is the strongest layer in the lithosphere. In the rift zone the lithosphere is weaker than in surrounding areas. After 150 m.y. (i.e. 100 m.y. after extension has ended) the lithosphere in the rift area is the strongest lithosphere. This is caused by the lower temperatures in the old rift.

Vertical transects through the horizontal deviatoric stress field (Figure 21) give the strength profiles (Figure 22) of the lithosphere. Strength of the lithosphere is approximated here by the depth integral of σ_{xx}. Here, the weak lower crust is clearly visible, as well as the strong mantle layer. Also, the lithosphere becomes weaker when extension proceeds.

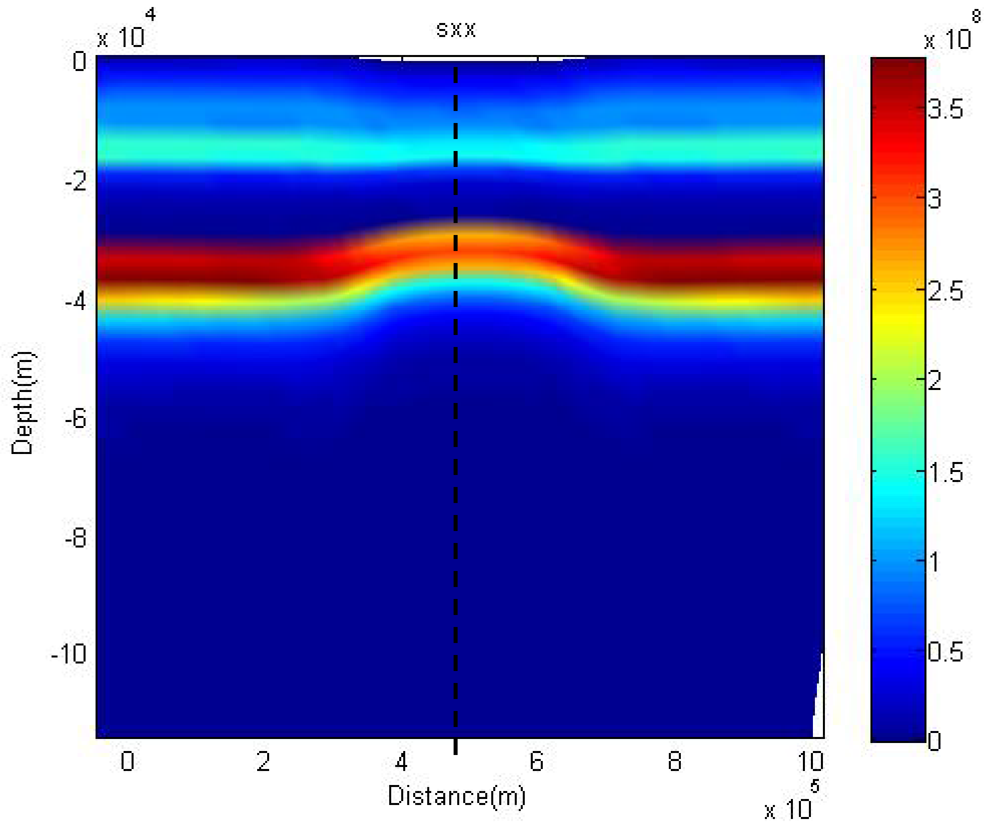
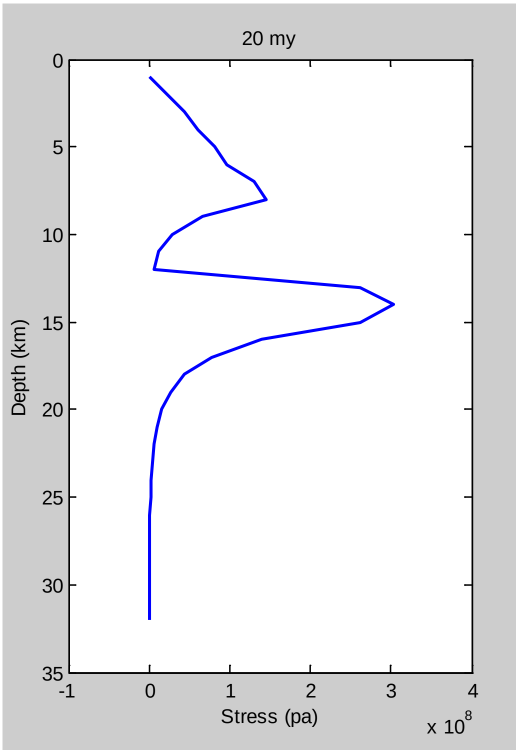
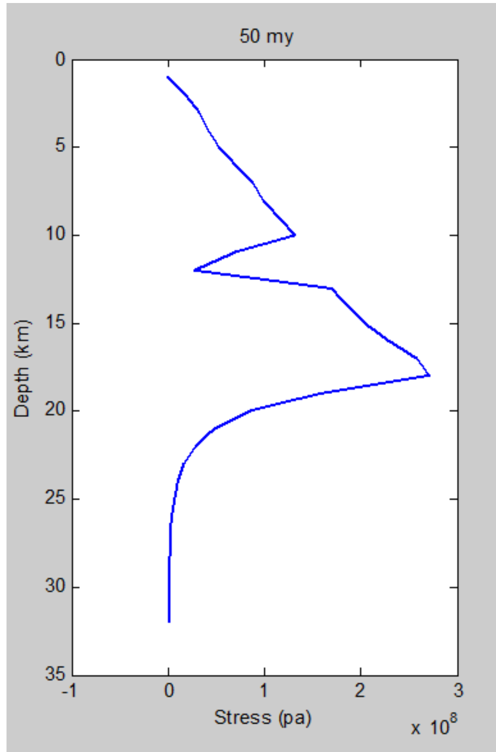


Figure 21. The horizontal deviatoric stress field (σxx) at 20 m.y . for Model 1. Units in Pa. The dashed line indicates the location of the vertical strength profiles shown in Figure 20. At this point in time the lithosphere is weakest in the rift area





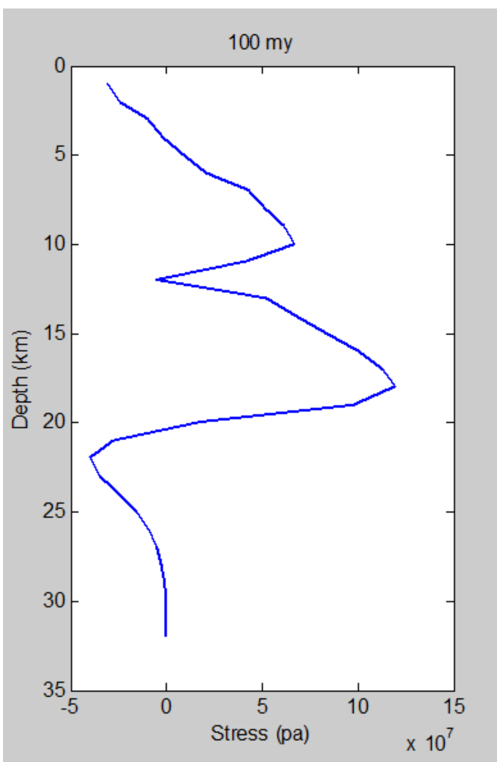


Figure 22: Strength profiles locations shown in Figure 19). These profiles show how the strength of lithosphere changes with depth and time. Units in Pa).

The strength profiles (Figure 22) also show that the weakest layer in the lithosphere is the lower crust, and the strongest layer the mantle part of the lithosphere just below the Moho. This remains the same during and after rifting. After extension ended (100 m.y. panel) the strength of the lithosphere is still low. Interestingly, the negative values of σ_{xx} in the 100 m.y. panel in the upper mantle indicate that this layer is under (very slight) compression at this point in time. This is attributed to ongoing thermal subsidence of the basin, which causes this layer to be under compression.

5.4 (THERMAL) SUBSIDENCE OF THE SURFACE

All models predict that during the post-extension thermal cooling phase, the surface continues to subside. This is illustrated in Figure 23. Figure 23 shows that upon rifting, a basin is formed in the center of the domain. When the extension phase comes to a halt (around 50 m.y.), cooling of the lithosphere continues and this results in thermal contraction and subsidence. The amount of thermal subsidence predicted by the models in the first 50 m.y. after rifting ended is about 200 m. This is tectonic subsidence, and if this accommodation space would have been filled with sediments, about 600 m of total subsidence would have occurred.

Since volume is conserved in the models, areas of subsidence (the rift) are compensated by areas of uplift. This characteristic of the finite element model inhibits us to study over how wide an area the thermal subsidence could have been expected.

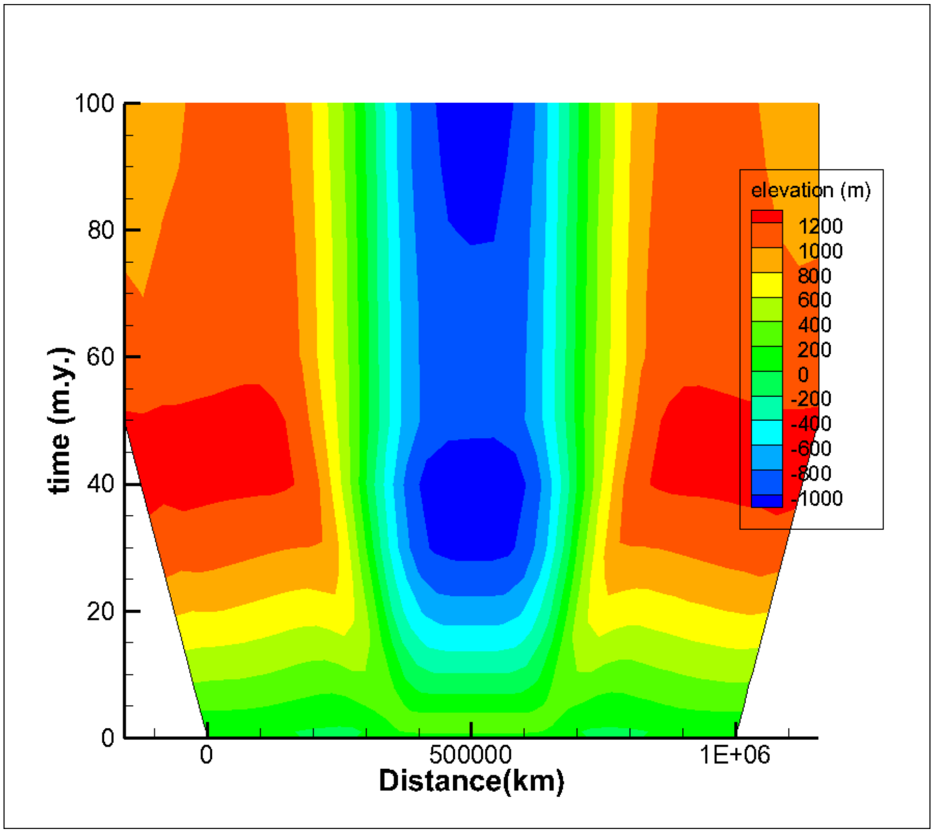


Figure 23. Surface elevation during and after rifting.

CHAPTER 6: DISCUSSION

Rifted cratonic basins are unique in the sense that they are underlain by a failed rift structure. These rift structures indicate minor rifting, and they are thought to have caused the onset of rifted cratonic basin formation. Armitage and Allen (2010) proposed that they formed from slow stretching. I investigated this hypothesis by using two-dimensional geodynamic models of lithosphere extension and thermal cooling. In this chapter I discuss the implications of my model results for the formation of rifted cratonic basins.

6.1 FORMATION OF RIFTED CRATONIC BASINS

I tested the effect of different extension velocities on rift evolution. Models 1 and 4 both had the same lithosphere thickness of 125 km but with different extension velocities; 3 mm/year and 1 mm/year respectively for a duration of 50 m.y. It was clearly observed that rifting localized very quickly at approximately 10 m.y in Model 1, while in Model 4 there was no rifting and no upwelling at this time. Extension was so slow in Model 4 that there was no heating of the lithosphere, and no weakening of the lithosphere. This prevented rifting from localizing. The Rio Grande rift in the southwestern U.S. is an example of a rift zone that currently opens with a rate of about 1 mm/yr (Ebinger et al., 2013); the western branch of the East African rift zone currently opens with a rate of about 3 mm/yr (Ebinger et al., 2013).

Also the lithosphere thickness was varied between the models. Models 1, 2, and 3 had a lithosphere thickness of 125 km,150 km, and 200 km respectively. These values were chosen because the lithosphere thickness observed by Kaminski and Jaupart (2000) beneath the Michigan Basin is 115 km and that of the Williston Basin was between 200 and 270 km, and 200 km for West Siberian Basin (Saunders et al., 2005) and Congo Basin (Kabongo et al., 2011). In Model 4 rifting started to localize only after about 40 m.y., while in Model 1 rifting localized already at 10 m.y. So, it took much longer to form the basin when the lithosphere thickness was larger. In that case, the upwelling below the rift was less. I also observed that the shape of the upwelling is different in Model 1 and 3, with Model 3 showing more of a broader and dome shape.

Model 3 shows the formation of a more narrow basins, which is in agreement with Buck (1991). Buck (1991) concluded that narrow rifts develop in thick cold lithosphere, such as the East African rift and the Caspian basin (Artyushkov, 2010).

Total subsidence of a basin consists of several components: tectonic subsidence, which is the result of a tectonic process, such as a rifting phase or loading of the lithosphere, and thermal subsidence, which results from thermal cooling and contraction. When this accommodation space is filled with sediments, the burial history of a basin is obtained. In my models, initial subsidence results from the rift phase, and this phase is followed by a phase of thermal subsidence. The initial subsidence during extension at 30 – 45 m.y. was about 800 meters in the models. This would correspond with about 2500 m of subsidence when sedimentation would have been included. The thermal subsidence was several hundred meters maximum. These values are in good agreement with tectonic subsidence curves of cratonic basins (Xie and Heller, 2009). Many rifted cratonic basins show a more discontinuous trend of subsidence, but this could result from multiple rift phases, or other tectonic processes such as a phase of compression during continent assembly periods.

My models suggest that a long, slow phase of minor rifting could form rifted cratonic basins. The prolonged subsidence phase could result from decay of the thermal anomaly that resulted from rifting.

6.2 FORMATION OF CRATONIC SAG BASINS

Some cratonic basins show no evidence for a rift structure. These include the Williston Basin in North America. These cratonic sag basins must have been formed by other process(es). Many of the hypotheses that were put forward to explain the formation of these basins are not supported by any direct evidence. DeRito et al. (1983) for example suggested that a body of high density material beneath the basins would drive subsidence until isostatic equilibrium would have been obtained. Downey and Gurnis (2009) postulated that the Congo Basin subsides as a result of a high density body. Free -Air anomalies in some cratonic sags do show anomalies that are in accord will a high density body in the lower crust or mantle lithosphere, but there is no further evidence that this mechanism is responsible for the formation of the cratonic sags. Heine et al. (2008) suggested that subsidence of the Congo Basin was due to dynamic topography as a result of downward mantle flow, and low seismic wave velocities in the upper mantle may support this hypothesis. However, Buiter et al. (2012) and Kabongo et al. (2011) stated that the Congo Basin was initiated as a result of a Neoproterozoic rift, with subsequent influence of mantle flow.

Middleton (1989) postulated that subsidence in the Canning Basin and Carnarvon Basin may have been formed as a result of a downwelling plume. There is however no evidence to support this mechanism. Klein and Hsui (1987) suggested that the coeval ages of cratonic basin found in North America, South America, and Africa suggests that they may have been formed as a result of the breaking of the Precambrian supercontinent. This would place these basins in the rift-drift suite as well.

A phase change from lower density gabbro to higher density eclogite was suggested to have been responsible for the formation of depressions on cratons by Fowler and Nisbet (1985) and Haxby et al. (1976). This mechanism was proposed as well for the origin of the Williston Basin. Fowler and Nisbet (1985) found that the subsidence record of the Williston basin is linear rather than exponential which supports this hypothesis and is not in agreement with subsidence driven by thermal contraction.

Since only a few cratonic basins (Congo Basin for example) are actively subsiding today, it is difficult to find evidence for support of one of the alternative mechanisms. The origin of these cratonic basins is therefore still debated.

CHAPTER 7: CONCLUSIONS

Each cratonic basin is unique. They are different in subsidence history and structure. Similarities include the long phase of slow subsidence, and their location away from active plate boundaries on cratonic lithosphere. Many processes have been proposed for their formation. One of the models for their origin describes cratonic basins as failed rift basins that underwent a long phase of thermal subsidence. These rifted cratonic basins were the topic of this study.

In this study, two- dimensional finite element models were used to study the rifted cratonic basin development. The models consist of visco-elastic lithosphere with a parameterization for brittle behavior, and parameters that were varied in the models included lithosphere thickness and extension velocities. The aim of this modeling study was to study the evolution of rifted cratonic basins from initiation to thermal subsidence. In the models the lithosphere was slowly extended for 50 million years, after which the lithosphere was allowed to cool. The study did not focus on the cause of extension.

The models predicted the prolonged subsidence history of cratonic basins following a slow rift phase. Models with a thick lithosphere did not focus rifting very effectively, but after a longer phase of extension a narrow rift basin developed. The extension velocity was also important. When the extension velocity was too low (1 mm/yr), the rift did not localize and no rift basin developed. The models predicted several hundred meters of thermal subsidence in the first 50 million years after rifting ended. This corresponds well with published tectonic subsidence curves of cratonic basins. The slow extension model seems thus to work well for the formation of rifted cratonic basins.

Arguments against this hypothesis state that the subsidence records of some rifted cratonic basins are linear rather than exponential, which would argue against a thermal cooling phase. Also, some tectonic subsidence curves are irregular and the basin might have undergone multiple phases of subsidence or uplift. This may reflect the long phase of basin formation which spans more than one super continent cycle. The stretching model may not work for basins not underlain by rifts.

A hypothesis that deals with the origin and evolution of cratonic basins such as the slow extension hypothesis should be able to account for the long period of subsidence and match the subsidence curves obtained from sedimentary successions. Based on the modeling results, I suggest that the slow stretching model can explain the origin of rifted cratonic basins. I also suggest that cratonic sags (such as the Carnarvon and Williston Basins) were formed by other processes.

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