

Cratonic basins and the Wilson Cycle: a perspective from the Parnaíba Basin, Brazil.

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Abstract

Cratonic basins may be considered an inherent component of the Wilson cycle, initiating after continental collision, craton and supercontinent development; and before rifting and continental break-up. They do not result directly from the horizontal plate motions characteristic of the Wilson Cycle, but from localised and long-lived, vertical subsidence. Covering over 10% of the world's continental crust, the majority of preserved cratonic basins initiated in the Early Paleozoic, after the formation of Gondwana and Laurentia. Recent investigation of the Parnaíba cratonic basin of Brazil reveals six features that characterize it, and potentially, cratonic basins generally: formation on thickened lithosphere (>150km); a pronounced, basal unconformity; a sub-circular outline and large area of the order of 0.5×10^5 to 2×10^6 kms²; long-lived (100-300 Ma), exponential tectonic subsidence of shallow marine and terrestrial sediments; no extensional strain features such as rift systems, crustal thinning or Moho elevation; a central, deep sourced, relative gravity anomaly high. These characteristics indicate basin initiation and development, simply by a thermally and mechanically driven, vertical subsidence of the lithosphere. The relaxation of a post-orogenic thermal and mechanical perturbation, after the continental collision phase of the Wilson cycle, appears the most likely driving mechanism. The source of that perturbation remains unclear.

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32 The Wilson Cycle describes the evolution of an ocean basin (Wilson 1966),
33 through continental rifting, ocean opening, oceanic subduction, arc and
34 continental collision and craton formation, all driven by the horizontal
35 movement of lithospheric plates. Each of these tectonic phases has been
36 characterized geologically through the recognition of ophiolite belts, collisional
37 orogens, rift systems and deformed passive margins (Dewey & Bird 1970). These
38 tectonic associations have enabled the interpretation of much of the worlds
39 Phanerozoic and Proterozoic geology in terms of plate tectonics (Burke & Dewey
40 1973, Dewey & Burke 1973, Dewey 1982, Sengor 1984). The Wilson Cycle built
41 on Wegener's (1929) contention that today's continents were derived from a
42 single ancestral continent, Pangaea. The Atlantic closure proposed by Wilson
43 (1966), a significant event in the formation of Pangaea, opened the possibility of
44 earlier supercontinents by repetition of the Wilson Cycle through time. The
45 notion of a supercontinent cycle developed from the periodicity in Precambrian
46 isotope ages (Gastil 1960). Four such supercontinents are now considered to
47 have existed Pangaea, Rodinia, Nuna and Kenorland (Evans et al. 2016).

48

49 Cratonic basins are significant features of continental geology, yet historically
50 they are not included in the Wilson Cycle. They cover over 10% of continental
51 crust, and are arguably developed throughout much of geologic time. Their
52 absence in the Wilson Cycle is in part a result of their enigmatic geodynamic
53 origin, and the fact that they occur dominantly within continents, rather than on
54 continental margins where the Wilson Cycle has left its greatest footprint. On the
55 basis of recent, detailed, characterization of the Parnaíba basin of Brazil (Daly et
56 al 2014, Tozer et al. 2016), we propose that the formation of cratonic basins be
57 considered an inherent component of the Wilson cycle. In contrast to most other
58 Wilson Cycle features, the formation of cratonic basins appears to be entirely a
59 process of vertical tectonics. No significant local extension is apparent
60 representing a preceding phase of lithospheric stretching or rift basin formation.
61 Equally orogenic loading and associated flexure is not a relevant formation
62 process either.

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Notwithstanding the uncertainty in their tectonic subsidence mechanism, cratonic basins appear to initiate during a particular part of the Wilson Cycle, after continent-continent collision and the formation of a large continental landmass, and well before continental break-up. Both Gondwana and Laurentia are particularly long-lived large landmasses (>400 Ma) and both have several examples of basins that fit this mode. In this paper we use recent insights from the Parnaíba cratonic basin of Brazil to outline some of the significant features of cratonic basins. We then discuss these in the context of basin formation, supercontinent formation and the connection of these large continental basins with the Wilson Cycle.

Characteristics of cratonic basins

Cratonic basins have long been recognized as a distinct basin type due to their within-continent location in cratonic and intra-cratonic regions of thick lithosphere (Figure 1). Characteristically they occur as singular, sub-circular shaped basins with long lived subsidence histories (Sloss 1963), and an absence of significant, horizontal strain features such as rifts (Sleep 1971, Fowler and Nisbet 1985, Hanne et al 2004, Allen and Allen 2013). Their singularity and sub-circular geometry is in marked contrast, to the linearity and thousand kilometer length of other common basin systems such as rifts, strike slip, passive margin and foreland basins (Roberts & Bally 2012).

Cratonic basins often comprise more than one basin forming stratigraphic unit, with the cratonic sequences being only a part of the tectono-stratigraphic history of the basin. The unconformity bounded, “cratonic megasequence” that is identified with the cratonic phase is specifically what we are dealing with in this paper.

Recent geophysical characterization of the Parnaíba cratonic basin of Brazil (Daly et al 2014, Tozer et al 2016) has emphasised six distinct features of cratonic basins and their megasequences that appear to apply globally. We outline these characteristics briefly below.

97

98 1. **Formation upon thick lithosphere (>150km).** Figure 1 shows the location
99 of 25 cratonic basins globally and their underlying lithospheric thickness as
100 determined by Rayleigh wave tomography (Priestley and McKenzie 2013). It
101 also shows the approximate age of their initiation.. The large majority of the
102 basins rest on seismic lithosphere between 150 and 250 km thick. The
103 singularity of the basins is occasionally disrupted by long-lived arches that
104 segment once contiguous basins. This is so in North Africa where the smaller
105 and adjacent Murzuq and Kufra cratonic basins appear to have been
106 contiguous and rest on thinner lithosphere. This thinner lithosphere may be a
107 feature of original accretion in a forearc setting (Holt et al 2014), or may also
108 be a result of extensive, post cratonic basin tectonics and crust and
109 lithospheric thinning due to Mesozoic basin formation (Selley 1997).

110

111 2. **A sub-circular outline of the order of 0.5×10^5 to 2×10^6 sq kms.** Figure 1
112 and figure 2 show the large scale and characteristic equant or sub-circular
113 shape of the cratonic basins as they are preserved today. Although believed
114 to be generally a close approximation, today's basin outlines are rarely the
115 original shape due to post cratonic basin deformation and erosion. This
116 element of post-depositional tectonics and erosion has resulted in significant
117 variety on the general sub-circular shape. This is particularly true where
118 apparently very large basinal areas have become separated, as appears to be
119 the case between the Amazon and Parnaíba cratonic basins in the late
120 Permian (Milani & Zalan, 1990) and the N African cratonic basins in the
121 Mesozoic (Craig et al. 2009)

122

123 3. **An underlying, profound, regional unconformity.** It appears that all
124 cratonic basin megasequences are underlain by a profound, sub-planar
125 erosional unconformity of basin wide extent. This is clearly seen on regional
126 seismic data in the Parnaíba basin of Brazil (Figure 3) where the cratonic
127 megasequence overlies a large planar unconformity of truncated Lower
128 Paleozoic and Neoproterozoic metasediments and crystalline rocks (Daly et
129 al 2014). The unconformity marks the planation of an earlier orogenic phase

and/or cratonic crystalline surface. In the case of the Brazilian and sub-Saharan African cratonic basins, the basement comprises Neoproterozoic to Lower Paleozoic (Brasiliano and Pan African) orogens (Brito Neves et al 2014).

There are several implications of this pronounced basal unconformity. Firstly, that there was a period of regional elevation and planation to sea level prior to the onset of the subsidence forming the cratonic basin. Sea level being registered by the marine sediments of the cratonic megasequence. Secondly, the erosion associated with that planation may have removed a significant volume of material and the resulting crustal thinning is potentially a component in the initiation of the basin (McKenzie & Priestly 2016). Finally, that the time encompassed in the formation of such a large regional feature is likely significant, and separates the subsidence driving mechanism from most preceding tectonics.

4. **Long-lived, quasi-exponential, subsidence profiles of 200-300 Ma.** The Parnaíba cratonic megasequence lasts from the Upper Ordovician to the end of the Triassic, a period of about 240 Ma. The sediments are generally shallow marine, becoming terrestrial towards the end of the basins life. Several hiatus characterise the stratigraphic sequence and enhance a very slow average rate of sediment accumulation of less than 15 m/Ma for the life of the basin. Tozer et al (2016) have shown that the backstripped tectonic subsidence in the Parnaíba basin is generally exponential and is fast initially and slow later on. This form of subsidence profile, similar across a number of cratonic basins, is also similar in form to the subsidence of rift-type basins (Allen & Armitage 2012). Also characteristic of the subsidence is an off-lapping geometry of the stratigraphic sequences within the cratonic basin megasequence.

5. **No evidence of lithospheric stretching preceding subsidence.** It has been proposed that cratonic basin formation is a result of protracted, low strain rate extension and consequent thermal relaxation (Allen and Armitage 2012).

Allen & Allen (2016) developed this idea and concluded that cratonic basins represent a low strain end member of conventional extensional rift basins. However, the crucial supportive evidence of the strain expected from this model appears to be missing. The Parnaíba basin is well imaged by regional deep reflection seismic (Daly et al. 2014) and a wealth of industry 2D and 3D reflection seismic data (De Castro et al. 2016, Porto et al. 2016). The seismic reflection data do not show an extensive rift system that records the strain preceding thermal relaxation of the Parnaíba basin. They do show localized, deformed, remnant basins interpreted as Cambrian in age, between large areas of Proterozoic crystalline basement (Porto et al. 2016). In addition, the current depth to the Moho shows no indication of elevation above the regional level of the Moho to the east and west, and therefore no thinning of the crust, with respect to the basin margins. Figure 3 shows the Parnaíba Moho slightly deeper than the flanks of the basin, an observation supported regionally by Luz et al (2016) from receiver function studies. Rather than subsiding due to crustal and lithospheric stretching, it appears the lithosphere has subsided due to a different mechanism of a purely vertical nature.

A similar situation is repeated in many, perhaps all, cratonic basins where authors have commented on the missing rift basins assumed to be required for their formation (Sleep et al. 1980, Allen and Allen 2012, Holt et al. 2014, McKenzie and Priestley 2016). However, a Precambrian rift interpretation of seismic reflection data has been used as evidence of extension preceding the formation of the Congo cratonic basin, (Crosby et al. 2010). The same seismic features have also been interpreted as indicating a compressional, pre-Ordovician, Pan African structuring of the Congo basin prior to cratonic basin formation (Daly et al. 1992, Linol et al. 2015). In conclusion, the missing rift system and undisturbed Moho appear to be a general characteristic of cratonic basins as they are increasingly well imaged, and lays open the issue of what has caused the thermal perturbation that has resulted in the exponential subsidence patterns widely observed.

6. **A mid-basin, deeply sourced, positive gravity anomaly.** The approximate centre of the Parnaíba basin is characterized by a localised gravity anomaly “high” which is superimposed on a regional gravity anomaly “low”. This high coincides with an area of mid-crustal reflectivity imaged by the regional seismic data (Daly et al. 2014). The anomaly has been modeled as an area of increased lower crustal density of $+185 \text{ kg m}^{-3}$ (Tozer et al. 2016). This is interpreted as evidence for a lower crustal area of magmatic intrusion. Similar dense bodies have been interpreted in the Amazon (Nunn & Aires 1988), Congo (Downey & Gurnis 2009) and Michigan (Nunn & Sleep 1984) basins.

These characteristic geological features of cratonic basins are summarized in the sketch of figure 4. Although based on the Parnaíba basin, these features appear common, and perhaps characteristic, of all cratonic basin and their megasequences.

Implications for the formation mechanism of cratonic basins

A large number of driving mechanisms have been proposed for the subsidence of cratonic basins (Hartley & Allen 1994). These can broadly be categorized as involving a thermal or density perturbation in the crust and lithosphere with subsequent relaxation and subsidence; or a stress related phenomena driven by surface stresses (Farrington et al. 2010 & Cloeting et al, 1988). The density perturbation may be developed from lithospheric stretching and thermal relaxation (Armitage & Allen 2010) or a deep thermal source (Kaminski & Jaubert 2000). A recent proposal by McKenzie & Priestley (2016) has suggested the creation of a thermal perturbation as a result of orogenic crustal thickening, re-equilibration of geotherms and then rapid erosion generating a thermal perturbation that then cools and drives the basin subsidence.

In Parnaíba, it appears the subsidence history has not been greatly affected by post depositional events and the subsidence curves resemble the post rift phase of extensional basin subsidence driven by post-rift thermal contraction

(McKenzie 1978, Cochran 1981), albeit, with a longer thermal time constant due to the thick lithosphere. Given this similarity, and the fact that cratonic basins appear to form on thick (>150 km) lithosphere, Allen and Armitage, (2010) and Crosby et al., (2010) have suggested cratonic basin subsidence is driven by thermal contraction following extension. However, mapping of the Parnaíba stratigraphic isopachs through time, at the scale of a stratigraphic group shows a consistent sub-circular form, and not the strong linearity characteristic of rift systems (Figure 2). These maps also demonstrate the off-lapping character of the sedimentary fill as the basin area shrinks with time. Such off-lap is evident in many cratonic basins, for example the Congo (Kadima et al. 2011), and Michigan basins (Nunn and Sleep, 1984).

These geometrical considerations apart, the extension initiated thermal subsidence model is also inconsistent with geophysical data that images a lithosphere that shows no evidence of thinning. A crustal thickness of 40 km, has been derived from the Parnaíba 2D deep seismic reflection velocity model (Daly et al. 2014), and a more accurate 42 ± 1 km, interpreted from wide angle reflection data (Tozer et al. 2016). These estimates are generally thicker than crust to the east and west of the basin (Luz et al 2015). They are also much greater (± 5 km) than that consistent with backstripping of Parnaíba subsidence. Equally importantly, the absence of widespread, extensional strain features such as rift basins in the basement of the Parnaíba basin is apparent from both academic and industry seismic reflection data (De Oliveira & Mohriak 2003, Daly et al. 2014, de Castro et al. 2016, Porto et al. 2016).

In the absence of lithospheric extension, an alternative basin driving mechanism is required that creates a thermal perturbation and its subsequent relaxation. This is a detailed and complex issue that is the subject of other papers in preparation.

Supercontinents, semi-supercontinents and cratonic basins in the Wilson Cycle

As mentioned in the introduction, the Wilson Cycle is an integral part of the formation and breakup of supercontinents through time. Continents largely resist the process of subduction due to their relative buoyancy. Hence the periodic aggregation of continents is an inevitable consequence of plate tectonics. Meert (2012) proposed the definition of a supercontinent as a single contiguous continent comprising more than 75% of extant continental landmass. Evans et al (2016) outlined four such supercontinents through time, Pangaea, Rodinia, Nuna and Kenorland, and defined the passage between them as a supercontinent cycle with a characteristic periodicity of 500-700 Ma. However, the definitions proposed excluded large landmasses such as Gondwana, Laurentia and Eurasia, all of which have been and are, crucial influences in our understanding of Earth's history. Evans et al (2016) recognized this shortcoming, but also recognized the power of the simplicity of a small number of supercontinental periods. They proposed an informal subset they labeled "semi-supercontinents", to include large and long lived landmasses that were the subsets of true supercontinents, such as Gondwana and Laurentia.

Figure 1 shows our best estimate of the age of initiation of the cratonic basins outlined, based on the oldest known sediments of the associated cratonic megasequence. It highlights pronounced Early Paleozoic cratonic basin initiation in Gondwana and Laurentia. In Eurasia the extensive impact of Cenozoic continental collision makes the identification of earlier cratonic megasequences uncertain. The map also shows preserved remnants of Precambrian cratonic basins in Australia, Africa and Canada.

The major Gondwana cratonic basins of Africa and South America initiated after the Late Neoproterozoic/Cambrian Pan African and Brasiliano orogenesis in the Cambro-Ordovician. The initiation of the Parnaíba, Amazon, Parana basins (figure 1) are well constrained as initiating in the Latest Ordovician on the basis of the oldest sediments above the regional unconformity (Vaz et al. 2007, Milani & Zalan 1999).

The Congo, Taoudeni, Kufra and Murzuq cratonic basins (figure 1) are less precisely constrained as being initiated during the Cambro-Ordovician. Seismic reflection data, constrained by wells, show that the Congo and Parnaíba cratonic sections unconformably overlie folded and thrust stratigraphic sections deformed during the Pan African/Brasiliano orogenic events (Daly et al. 1992, Kadima et al. 2011, Daly et al. 2014). A similar geometrical relationship is described by Craig et al. (2009) throughout the cratonic basins of North Africa, where a flat lying “cratonic” Cambro-Ordovician section unconformably overlies a deformed, then peneplained Neoproterozoic folded section.

Gondwana crustal development through the Early Paleozoic is therefore characterized by the formation of several, large, sub-circular basins of slow, long-lived subsidence, developed on lithosphere deformed during the Brasiliano/Pan African orogeny. The same Lower Paleozoic period exhibits no major rifting and breakup events in central Gondwana. In Laurentia Sleep (2009) and Armitage and Allen (2010) point out similar fundamental characteristics for the Williston, Illinois and Michigan basins. Most, if not all, of these large sub-circular areas of prolonged subsidence show no related extensional events immediately prior to the cratonic basin phase of subsidence.

The pronounced grouping of ages in Gondwana occurs after the stabilization of the collisional tectonics that created Gondwana through the Late Neoproterozoic-Ordovician (650-500 Ma) Brasiliano/Pan African orogenic period. The stabilized Gondwana then survived largely intact for the next 250 Ma before extensive break-up began in the Mesozoic. The observation of widespread cratonic basin formation in Gondwana, soon after Brasiliano/Pan African orogenesis and stabilization, places the formation of these basins firmly in the cratonic, post-collision and pre-break-up phase of the Wilson Cycle. This period is very long lived in Gondwana and Laurentia (>200Ma), but is much shorter in Pangaea where final collision and break-up followed relatively quickly (<50 Ma).

Two geodynamic implications of these observations may be significant in the origin of cratonic basins. Firstly, crustal thickening associated with the pre-

ceding collisional orogeny and continental accretion, will have resulted in a major thermal perturbation in the thickened crust and lithosphere. McKenzie & Priestley (2016) have modeled the rapid erosion of equilibrated thickened crust to create a thermal anomaly that cools and results in the subsidence we see as a cratonic basin.

Secondly, the mantle's convective response to the assembly of a large continental landmass like Gondwana over one or more subducted slabs, is likely to result in a period of thermal insulation (Philips & Coltice 2010) and a reorganization of the mantle convective system (Zhong et al 2007). The thermal consequences of this insulation and reorganization may have resulted in an early period of widespread plume activity with associated magmatic material that has intruded into the upper mantle or lower crust and created both the thermal and mechanical perturbation recorded in the Parnaíba and other Gondwana cratonic basins. Clearly, this connection to the accretion of a large continental landmass is unproven and difficult to test. However, the regional geodynamic context, the characteristics of basin subsidence outlined above and a significant gravity anomaly are all tangible data that support this interpretation.

These general observations of Gondwanan and Laurentian Paleozoic basins indicate that the initiation of cratonic basins preferentially occurs in the period post-continental collision and semi-supercontinent formation, and prior to the rifting and break-up processes of continental fragmentation (Figure 5). In Gondwana this period, from the Ordovician to the Triassic, lasts approximately 200 Ma, before significant rifting and the commencement of the break-up of Gondwana. During that long, static period, Gondwana and Laurentia developed most of the worlds preserved cratonic basins.

Precambrian Cratonic basins

In addition to the well preserved Phanerozoic cratonic basins, several Proterozoic cratonic sedimentary basins occur in the geological record (Allen and Allen, 2013), although preservation is usually poor and the available

subsurface data sparse. However, basement terranes have also been argued to provide evidence for cratonic basin formation processes, unrecognised due to thick sediment overburden in Phanerozoic basins (e.g. Stel et al. 1993).

The Paleoproterozoic, Mporokoso Basin of eastern Zambia (Figure 1) is a potential example of a remnant cratonic basin (Unrug, 1982). This triangular shaped cratonic megasequence is up to 5 km thick, covers over 60,000 sq km and consists of interbedded fluvial, aeolian and lacustrine quartz and arkosic arenites and variegated siltstones. The edges of the basin are defined by later zones of deformation and Cenozoic rifting, and the largely flat lying Paleoproterozoic rocks sit unconformably on a Paleoproterozoic basement of schists, gneisses, calc-alkaline granites and andesitic volcanics known as the Bangweulu craton (Andersen & Unrug 1984). The Mporokoso basin has no apparent rift precursor indicative of crustal stretching and, size apart, does have the other characteristics discussed above, including the apparently off-lapping stratigraphy diagnostic of a viscoelastic weakening lithosphere (e.g. Watts et al. 2013). Temporally the development of the Mporokoso basin post-dates the formation of the Paleoproterozoic Kenorland supercontinent (Evans et al. 2016).

In Australia, the Officer, Amadeus and Ngalia basins of central Australia (Figure 1) contain sediments of Neoproterozoic age (ca. 820-780 Ma) (e.g. Lindsay and Leven, 1996; Haddard et al., 2001). During this initial subsidence, these basins appear to have formed a single basin called the Centralian basin (Walter et al., 1992) of comparable size to that of a typical Phanerozoic cratonic basin. However, these basins have been later modified by several orogenic episodes resulting in inversion, compartmentalisation and overprinting by foreland style subsidence (e.g. Haddard et al., 2001). Hence, the initial Neoproterozoic subsidence is not well understood, but appears to form late in the accretion of the Rodinia supercontinent and to have many of the characteristics of well-preserved Phanerozoic cratonic basins.

Finally, Stel et al., (1993) inferred that Proterozoic sediments found in Southern Finland are remnants of a former cratonic basin. Based on observations of the

surrounding exposed basement geology, namely mafic dyke swarms and silicic dykes and A-type granite batholiths, these authors also developed a model for cratonic basin formation. They suggest that passive upwelling of asthenospheric material leads to the melting of the lower lithosphere and emplacement of basaltic melt at lower crustal levels with associated mafic dyke swarms. This in turn generates silicic melts that migrate to upper crustal levels, thereby forming the observed A-type granites and silicic dykes and leaving in the lower crust, mafic intrusive material. Upon cooling and crystallization this becomes a buried load thereby driving plate flexure and basin formation as seen in the cratonic megasequences described above.

Discussion

Sloss (1963) developed the early perspectives on cratonic basins in North America's Michigan basin. He pointed out the characteristic bowl shape, shallow water to terrestrial sedimentation and long-lived character. He perceived them as vertically subsiding areas in central continental locations. Sleep (1971) and Sleep et al. (1980) developed largely thermally driven solutions for the formation of the Michigan basin. Burke and Dewey (1973) suggested that cratonic basins formed at the tips of rifts propagating into continental crust from a divergent margin triple junction. However, the relevant cratonic basin is often older than the propagating rift, perhaps suggesting a different relationship. Armitage & Allen (2010) argued that cratonic basins are initiated as the low strain rate end-member of extensional type basins. They suggested that a low strain rate, extended over a large time period, could explain the subsidence in the Williston basin, although its tectonic setting suggests a significant foreland type contribution. They also acknowledged the absence of physical evidence for the proposed rifting and lithospheric thinning. Far field stress (Cloetingh 1988) and an upwelling plume (Middleton 1989) have also been proposed as viable basin forming mechanisms. A dense load hypothesis has also been widely argued, generated either by metamorphic phase changes (Haxby et al 1976) or igneous intrusion and phase changes in the lower crust or lithospheric mantle

(Fowler & Nisbit 1985; Sousa, 1997). More recently McKenzie (2016) has proposed an orogenic thickening, heating and unroofing model.

The well constrained geophysical and geological data set of the Parnaíba basin makes some of these proposed subsidence mechanisms inapplicable or not required to explain the evolution of this basin. In particular, there is no evidence of extensional strain in the crust or mantle (Daly et al., 2014, Porto 2016) and the gravity data is inconsistent with an eclogite phase change (Tozer et al. 2016). However, the initiation of basin subsidence due to a visco-elastic weakening response to a buried load, formed by a cooling igneous intrusion in the lower crust beneath the centre of the basin, does fit the data constraints. The orogenic unroofing model of McKenzie and Priestley (2016) may also be valid. However, it will require detailed P/T paths from metamorphic data on the Parnaíba basement to test its applicability.

In the context of the Wilson cycle (Figure 5), our data and observations indicate that cratonic basin subsidence is a predominantly vertically driven tectonic process. Subsidence occurred on thickened lithosphere and developed above a major, post-orogenic unconformity. The subsidence continued for ca. 250 Ma with a shallow, concave up, exponential form of decay.

The age of initiation groups the Gondwana cratonic basins of Africa and Brazil as post the Pan African/Brasiliano accretion of Western Gondwana (post 500 Ma). The basins formed during a relatively stagnant period of the Gondwana supercontinent, 200 Ma before it began to break up in the Permian. The Michigan, Illinois and Williston basins formed slightly earlier in a similar period of Laurentian continental stability.

The mantle processes active during periods of long lived super or semi-supercontinental stagnation and the convective response have been modeled by Zhong et al. (2007). They imply a period of increasingly active mantle plume generation as the large continental lid prevents the cooling of particular parts of the asthenosphere. This period of heightened plume activity may be the tectonic

link to the origin of cratonic basins and their place in the Wilson cycle. The post-orogenic, post-craton formation, supercontinent or semi-supercontinent phase of crustal evolution favours a period of heightened mantle plume activity. At least some of this activity may penetrate the lithosphere and result in the formation of magmatic intrusions in the mantle and lower crust. The cooling of this intrusion and thermal perturbation creates a dense load that initiates a large, subsiding, sub-circular, cratonic basin.

Conclusions

The geophysical characterisation of the Parnaíba basin of Brazil has improved the constraints on the formation of cratonic basins. These constraints are consistent with much of the evidence-based literature on cratonic basins and can be used to position these enigmatic basins within the Wilson Cycle. Specifically we conclude that:

- The Parnaíba cratonic basin initiated in the Late Ordovician, about 50 Ma after the accretion of Gondwana and the end of the Brasiliano and Pan African orogenic period (500 Ma). Gondwana remained intact for over 200 Ma, and did not commence extensive continental rifting and break-up until the Mesozoic. This post-orogenic period, of a long-lived, stable, semi-supercontinent, saw the proliferation of cratonic basins during the early years of Gondwana and also Laurentia.
- The Parnaíba cratonic basin megasequence sits upon a major, post-orogenic, erosional unconformity of regional proportions. The basin developed over a long time frame (>200 Ma), and is characterized by very slow (15 m/Ma average), exponential subsidence. Sedimentation was dominantly shallow marine becoming increasingly fluvial with time.
- There is an absence of evidence, in such a well imaged basin, of widespread crustal thinning through extension and rifting, or a regionally elevated Moho. This negative evidence, together with the sub-circular isopachs of the

stratigraphic Groups within the cratonic megasequence, lead us to believe that a process of continental extensional tectonics did not initiate the Parnaíba basin.

- A relative gravity anomaly high over the centre of the Parnaíba basin, coincident with an area of mid-crustal seismic reflectivity, suggests the presence of an anomalously dense lower crust that may be associated with the driving mechanism for the basin.

- A dense lower crust or upper mantle may have formed through magmatic or metamorphic processes. Its presence may have driven subsidence through viscoelastic relaxation and thermal contraction following plume activity beneath the large continental landmass.

- The prolonged existence of a stagnant to slow moving and long lived landmass of the scale of Gondwana or Laurentia, with a large area of contained, sub-continental asthenosphere, appear to be related to the preferential formation of cratonic basins.

- Cratonic basins would therefore occur cyclically within the Wilson Cycle; post major orogenic periods and the formation of large landmasses that are long lived (>50 Ma); and before significant continental rifting and break-up. Whilst the driving mechanism of these basins is vertically, not horizontally driven, their context and generation appears deeply connected to the activity and sequence of the Wilson Cycle.

Acknowledgements

The authors would like to thank BP do Brasil and the Parnaíba Basin Analysis Project (PBAP) partners for their support in this work. The seismic data supporting this paper are owned by Global Geophysical Services Incorporated (GGSI), 13927 South Gessner Road, Missouri City, TX 77489, USA. The data are proprietary to GGSI and are commercially available from them. The authors

524 would like to thank Global Geophysical Services Inc., for their permission to
525 publish the profile discussed in this paper.
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Figures:

Figure 1.

A world map showing lithospheric thickness after Priestley and McKenzie (2013), together with the location of selected cratonic basins (basin outlines after Torsvik, personal communication). The basins are coloured in line with the approximate age of their initiation. Note the proliferation of basins forming soon after the Cambrian stabilization of Gondwana and Laurentia..

Figure 2.

Isopach evolution of the Paleozoic lithostratigraphy of the Parnaíba cratonic basin, shown at the lithostratigraphic “Group” level (as defined by Vaz et al. 2007). The three Groups make up the cratonic basin megasequence.

Inset map shows the geographic location of the Parnaíba basin in South America and the position of the deep crustal seismic profile of figure 3.

Figure 3.

Parnaíba regional deep seismic reflection profile: (A) uninterpreted PSDM line; (B) line drawing interpretation; (C) geoseismic interpretation overlaid on line drawing (modified from Daly et al. 2014). The images show the scale and tectonic setting of the Parnaíba basin and its basement structure.

Key: colours from east to west: Copper, Borborema Province basement; Orange and Yellow, Parnaíba cratonic megasequences; Grey, Campo Maior pull-apart basin; Pink, Parnaíba basement; Green, zone of mid-crustal reflectivity; Mauve, amphibolite facies metasediments; Brown, greenschist facies metasediments; Purple, Amazonian craton.

Figure 4.

A geological and geophysical summary diagram of a ‘typical’ cratonic basin, based on the Parnaíba basin, with key points numbered.

Key: 1. Lithosphere Asthenosphere boundary >150 km; 2. The Moho beneath the cratonic basin shows a similar or greater crustal thickness to the flanks of the

basin, indicating no significant crustal thinning during basin formation. 3. Pre-existing continental crust largely hidden beneath the basin; 4. Pre-Silurian, deformed, remnant basins and small, late to post-tectonic Alkali granites possibly related to the emplacement of lower crustal igneous rocks prior to basin formation; 5. Major crustal scale structure not reactivated during the deposition of the cratonic megasequence, consistent with a purely vertical subsidence model. 6. Regional, planar unconformity that underlies the entire basin; 7. Off-lapping stratigraphic sequences of the cratonic megasequence; 8. Isolated, small graben features preserving cratonic megasequence sediments beyond the present outlines of the cratonic basin, indicating a much large footprint for the original basin; 9. A markedly erosional edge to the basin indicating a much larger footprint of the original basin.

Figure 5.

Sketch diagram showing cratonic basins in the Wilson cycle sequence.

1. Continental rifting; 2. Continental break-up, sea floor spreading and the onset of oceanic subduction as an ocean basin is created and begins to be subducted; 3. Continental collision and lithospheric thickening builds a mountain range and the subsequent onset of tectonic collapse, extrusion and erosion; 4. Stabilized post orogenic crust, peneplained and restored to sea level with saucer shaped subsidence defining the cratonic basin.

The cratonic basin is marked in yellow and develops between the continent continent collision event, crustal thickening, tectonic collapse, erosion and stabilization; and the onset of the next period of rifting and break-up (back to sketch 1). In the case of Gondwana that time period, from collisional stabilization to extensive break-up, was over 200 Ma.

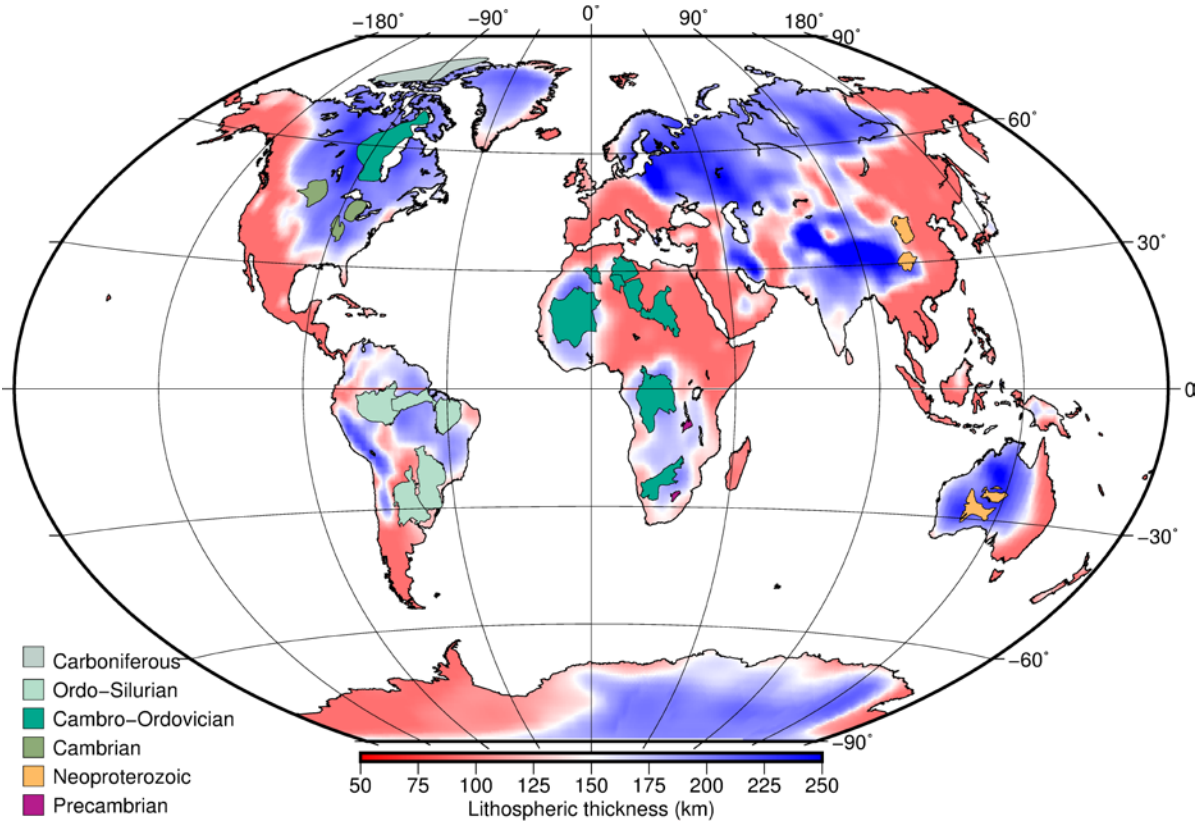
Key: 1. Lithospheric mantle; 2. Oceanic crust; 3. Continent I; 4. Continent II; 5. Continent III; 6. Syn-tectonic sediments; 7. Passive margin sedimentary wedge; 8. Cratonic basins.

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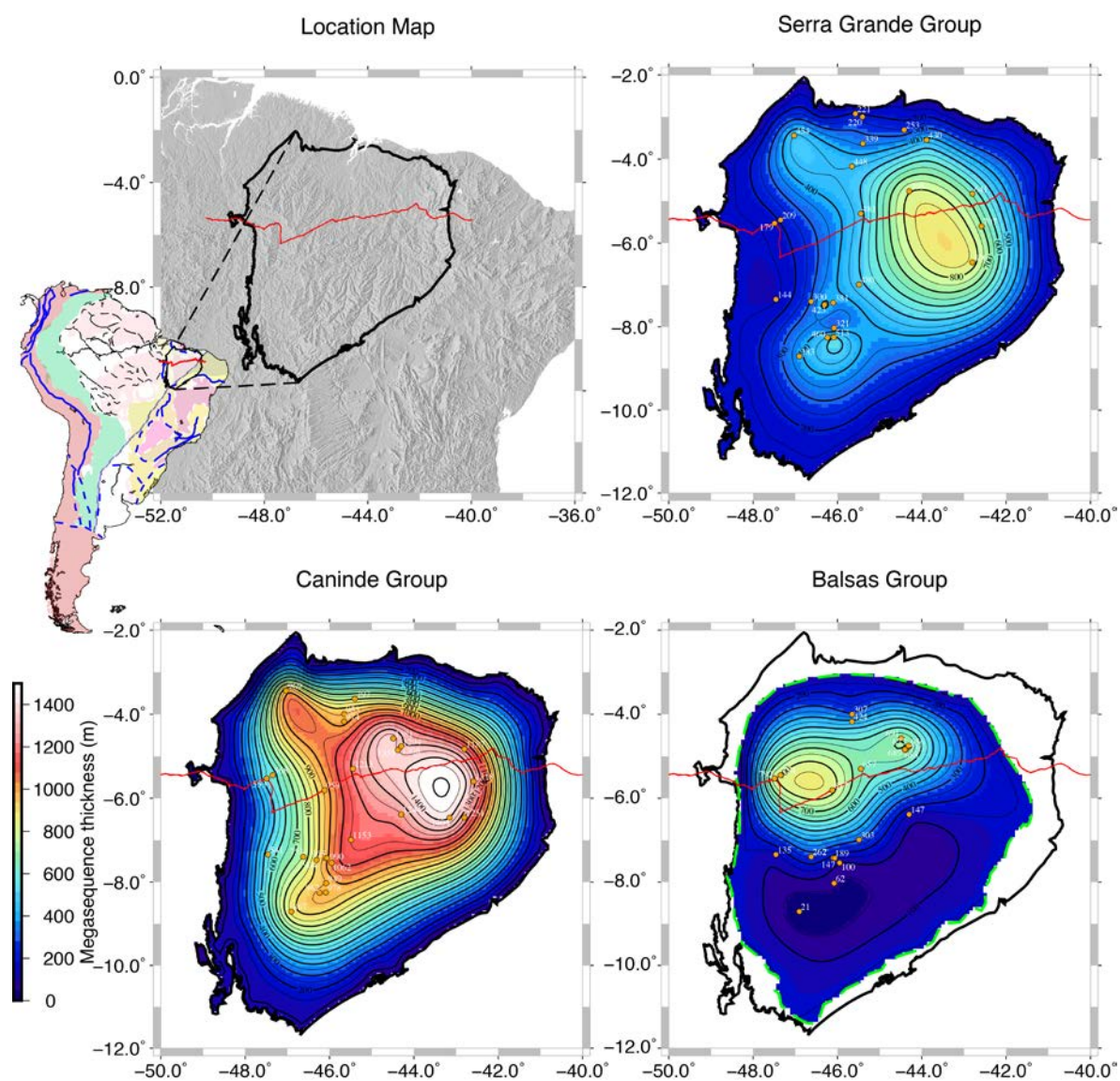
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Figures

Figure 1



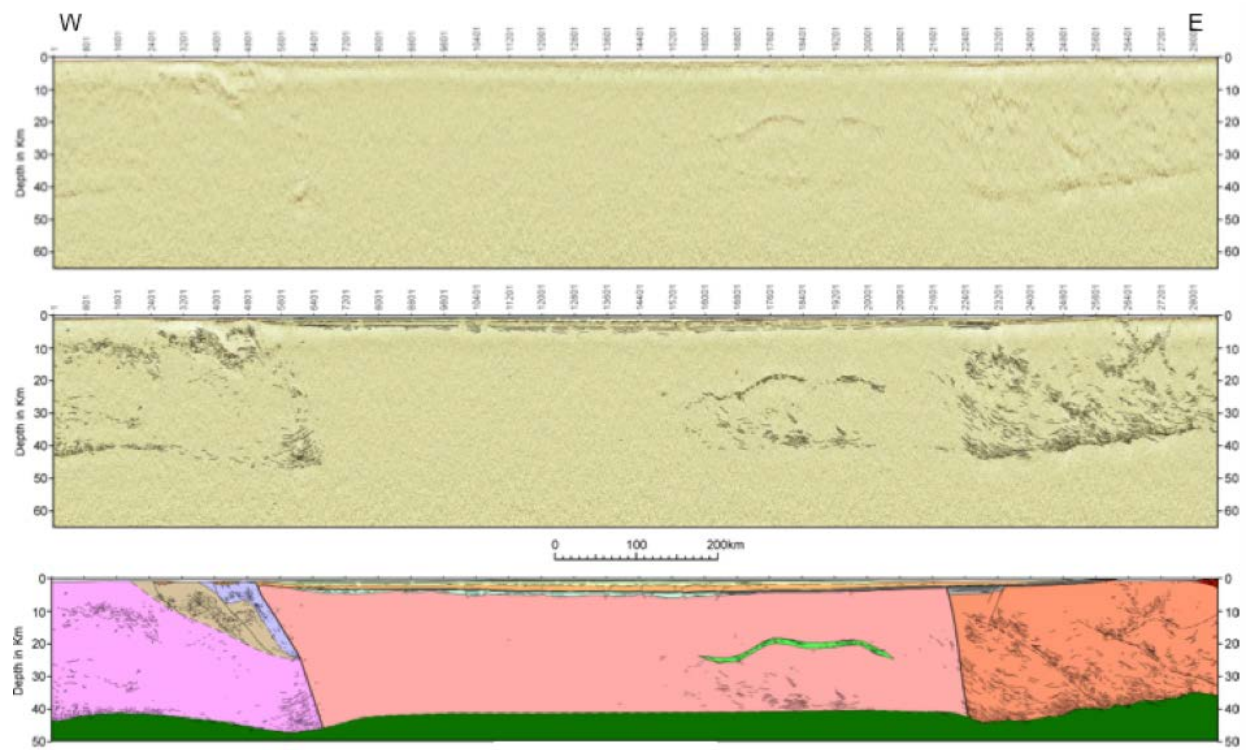
827 Figure 2



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830 Figure 3

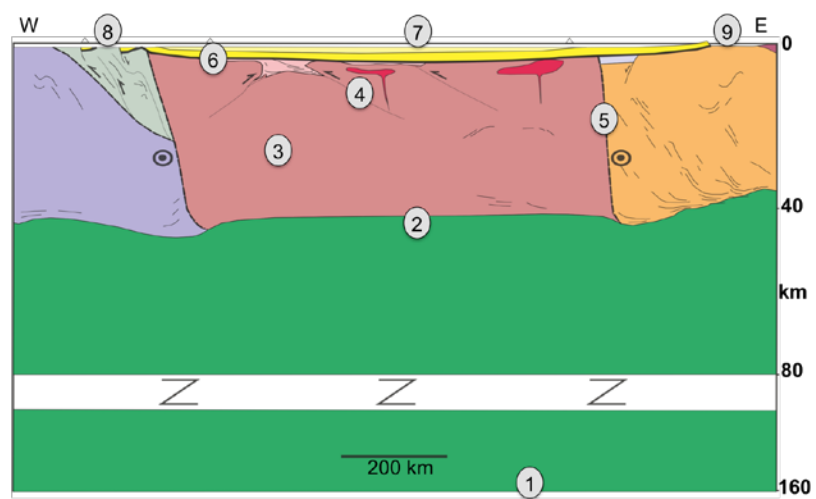


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834 Figure 4



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