Chapter 30

Cratonic basins

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ABSTRACT

Cratonic basins are sites of prolonged, broadly distributed but slow subsidence of the continental lithosphere, and are commonly filled with shallow water and terrestrial sedimentary rocks. They remain poorly understood geodynamically. A number of models have been proposed that fall into families involving cooling of stretched continental lithosphere, cooling related to mantle flow (dynamic topography), densification of the underlying lithosphere due to phase changes, the surface response to magmatism and/or plume activity, and long-wavelength buckling under in-plane stresses.

The timing of initiation and spatial distribution of cratonic basin formation are linked to geodynamic phases within the overall framework of plate amalgamation and supercontinental break-up and dispersal. Many cratonic basins initiated in the Neoproterozoic and Cambrian-Ordovician. Some suites of cratonic basins originated as broad ramp-like realms of subsidence tilting down to the adjacent passive margin, and were later "individualized" by secondary processes such as, for instance, reactivation of tectonic structures during intracontinental orogeny, and the emergence of intervening arches and domes.

Several different mechanisms may therefore control the geological evolution and subsidence history of cratonic basins during their long life-times. We propose that a model of low strain rate extension accompanied and followed by cooling of the underlying lithosphere satisfactorily explains the long-term subsidence history of a range of cratonic basins. However, the precise role played by dynamic topography transmitted from large-scale mantle flow in initiating or modifying the elevation history of continental interiors remains an intriguing focus for further research.

Keywords: continental lithosphere; stretching; strain rate; subsidence; stratigraphy

INTRODUCTION

"Intracratonic basins," "cratonic basins," "interior cratonic basins," and "intracontinental sags" are circular to oval-shaped crustal sags, located on stable, relatively thick continental lithosphere (Sloss and Speed 1974; Sloss 1988). We restrict the use of the term "cratonic basin" to those basins located some distance from stretched or convergent continental margins, distinct from rifts where a history of continental extension is unequivocal, but located on a variety of crustal substrates, irrespective of whether they are crystalline shields *sensu stricto*, accreted terranes, or ancient foldbelts and rift complexes. The important consideration is that the lithosphere behaves stably (Sloss 1988).

Cratonic basins are characterized by prolonged, predominantly shallow-water and terrestrial sedimentation and a gross layer-cake type of stratigraphy (Sloss and Speed 1974; Quinlan 1987; Sloss 1990; Leighton et al., 1991). Their subsidence history is prolonged, occasionally marked by an initial stage of relatively fast subsidence, followed by a period of decreasing subsidence rate (Nunn and Sleep 1984; Stel et al., 1993; Xie and Heller 2009) (Fig. 30.1), somewhat similar to that of ocean basins (Sleep 1971). Cratonic basins generally lack

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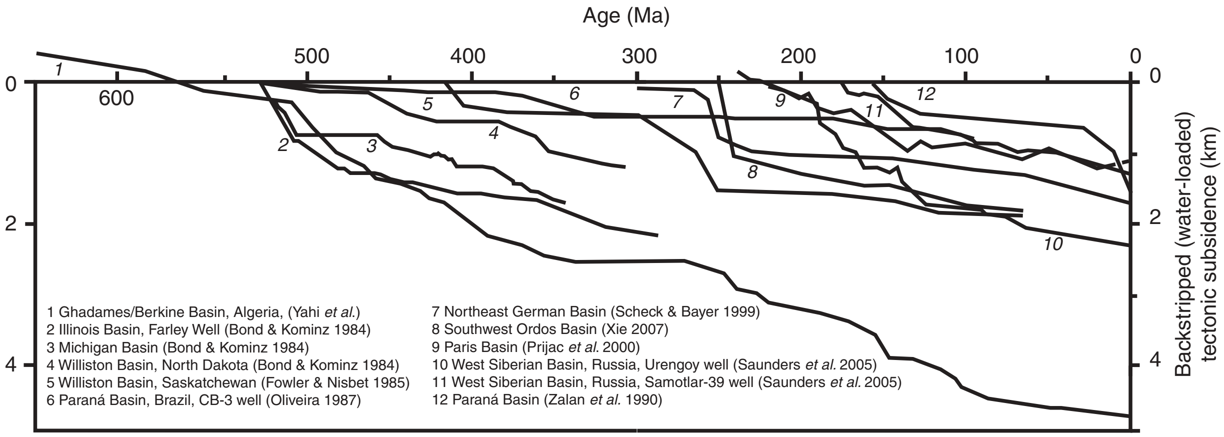

Fig. 30.1. Compilation of intracratonic basin subsidence curves. *Source*: Extended From Xie and Heller (2009).

well-developed initial rift phases, marked by arrays of extensional faults and associated graben and half-graben, though this may be due in part to the poor seismic imaging of the base of cratonic basins preserved on land. Cratonic basins are located on continental lithosphere, away from the plate margin, but in some cases connected by a rift or failed rift zone to the ocean, as in the Neoproterozoic Centralian Superbasin of Australia (Walter et al., 1995; Lindsay 2002), the Lower Paleozoic Illinois and Oklahoma basins of USA (Braile et al., 1986; Kolata and Nelson 1990), and the Mesozoic phase of the Chad Basin of north-central Africa (Burke 1976). This geometry suggests that many cratonic basins lie at the tips of failed rifts extending into the continental plate at a high angle from the extensional plate margin, which may be the site of former triple junctions (Burke and Dewey, 1973).

Although cratonic basins have their own individuality, which is to be expected in such long-lived basins, there are a number of prominent features common to the majority of examples. At first order, these include the following:

(1) The surface area enclosed by the zero isopach of the basin-fill is commonly circular or elliptical, and large, with surface areas ranging from the relatively small Anglo-Paris Basin (10^5 km^2), through the large Hudson Bay (1.2×10^6 km^2) and Paraná basins (1.4×10^6 km^2) to the giant Centralian Superbasin (2×10^6 km^2) and West Siberian Basin (3.5×10^6 km^2) (Leighton and Kolata, 1990, p. 730; Sanford, 1987; Walter et al., 1995, p. 173; Vyssotski et al., 2006).

(2) In cross-section, cratonic basins are simple saucers, lacking major syn-tectonic faults (late post-sedimentary faulting during transpression/transtension is more common), with sediment thicknesses typically less than ca. 5 km, and rarely $<$6–7 km (as in the West Siberian, Illinois, and Paraná basins). However, in some cases, the circular planform shape is a result of later compartmentalization of a previously more extensive platform or ramp, as in the cratonic basins of north Africa, such as Al Khufra, Murzuk, and Ghadames (Selley, 1972, 1997; Boote et al., 1998).

(3) The duration of subsidence is very long, measured in hundreds of millions of years (e.g., Aleinikov et al., 1980), and backstripped tectonic subsidence histories are commonly sublinear to gently negative exponential (Xie and Heller, 2009) (Fig. 30.1). These long lifetimes of subsidence commonly comprise several basin phases separated by unconformities, giving superimposed megasequences reflecting changing patterns of the governing plate-scale tectonics. Taking the depositional megasequences most safely attributable to cratonic basin subsidence, and neglecting long periods of non-deposition at megasequence boundaries, the cratonic basins of North America accumulated sediment at rates of 20 to 30 m Myr^{-1} (Sloss 1988), which is extremely slow compared to rifts, failed rifts, young passive margins, foreland basins, and strike-slip basins (Allen and Allen 2005), but relatively fast compared to the adjacent platforms. Laterally equivalent platformal areas, such as the

Transcontinental Arch of USA, accumulated ca. 1 km if sediment between Cambrian and Permian, at a rate of 3–4 m Myr^{-1} (Sloss 1988).

(4) Stratigraphy is predominantly terrestrial to shallow-water, indicating that sedimentation kept pace with tectonic subsidence throughout. Cratonic basin megasequences commonly start as a broad regional tilting of the continent, as in the latest Proterozoic-Early Ordovician "Sauk" sequence of east-central North America (Sloss, 1963, 1988) (discussed below) and the Early Paleozoic (pre-Silurian) of north Africa (Selley, 1997). However, facies belts in bull's-eye and teardrop patterns (exemplified by the Silurian carbonates and evaporites of the Michigan Basin; Nurmi and Friedman, 1977) indicate that the circular outline of some cratonic basins is a primary, syndepositional feature and is not a result simply of post-sedimentary tectonic deformation dissecting a previously more extensive depocenter. Cratonic basins in low paleolatitudes are commonly dominated by chemical sediments, showing that particulate sediment supply was modest and that high topographic relief was absent around basin margins at these times.

(5) Cratonic basins are commonly regularly spaced with their centers located about 10^3 km apart. In North America, they line up a certain distance (hundreds of km) from the Early Paleozoic edge of the North American plate, as, from south to north, the Illinois, Michigan and Hudson Bay basins (Leighton et al., 1991).

(6) Some cratonic basins are associated with widespread magmatism, such as the eruption of large volumes of basalts, as in the West Siberian Basin and the Paraná Basin (Thompson and Gibson, 1991; Saunders et al., 2007). However, the precise causal links between basaltic volcanism and basin development are unclear.

Cratonic basins are very long-lived (Fig. 30.1), but it is important to recognize that the basin-fill is commonly composed of a number of different megasequences (or *sequences* of Sloss, 1963), some of which may be associated with entirely different mechanisms of formation, such as strike-slip deformation, flexure and unequivocal stretching. Consequently, it is important, wherever possible, to extract the cratonic basin megasequence from the polyhistory basin-fill (Kingston et al., 1983) for analysis. In other cases, basins have remained as cratonic basins throughout their history but have existed long enough to have been strongly affected by several tectonic mechanisms of subsidence and uplift. Consequently, there may be a primary mechanism for basin formation, and different secondary mechanisms for later modification.

A large number of mechanisms have been invoked to explain cratonic basins (see Hartley and Allen, 1994, table 1; and review by Klein, 1995). Models, which are partly overlapping, include the following:

- Thermal contraction following heating (Haxby et al., 1976; Sleep and Sloss, 1980; Kaminski and Jaupart, 2000)
- Localized extension related to magmatic upwelling that may be associated with plume activity (Klemme, 1980; Keen, 1987; Klein and Hsui, 1987; Ziegler and van Hoorn, 1989; Ziegler, 1990; Neumann et al., 1992; Zhao et al., 1994)
- Deep crustal phase changes (De Rito et al., 1983; Fowler and Nesbit, 1985; Helwig, 1985; Artyushkov and Bear, 1990; Artyushkov, 1992; Artyushkov et al., 2008)
- Reactivation of pre-existing sags under in-plane stress or flexural loading (Quinlan and Beaumont, 1984; Quinlan, 1987; Beaumont et al., 1987)
- Emplacement of basaltic underplates during anorgenic magmatism, and slow thermal contraction in a non-extensional setting (Stel et al., 1993)
- Subsidence due to negative "dynamic topography" – the topography resulting from the transmission to the Earth's surface of stresses caused by large-scale mantle flow – over a downwelling or convective instability in the mantle (Liu, 1979; Middleton, 1989; Hartley and Allen, 1994; Heine et al., 2008; Farringdon et al., 2010) or related to the subduction of cold oceanic slabs (Mitrovica et al., 1989; Burgess et al., 1997)
- Subaerial erosion over a thermal uplift followed by sediment loading (Hsu, 1965; Le Pichon et al., 1973; Sleep and Snell, 1976; Sahagian, 1993).

An association with continental stretching followed by flexural compensation of the thermal contraction load is explicit or implicit in many models (e.g., Sleep and Snell, 1976; Kaminski and Jaupart, 2000) and classification schemes (e.g., Klemme, 1980; Kingston et al., 1983) of cratonic basins. In the following section we briefly review the geological context for cratonic basin formation in order to support this view of the fundamental role played by the cooling of thermal anomalies related to mechanical stretching *sensu stricto* or related to plate configurations in which regional extension accompanies cratonic basin initiation.

THE TIMING OF INITIATION AND GEOLOGICAL CONTEXT OF CRATONIC BASINS

It is impossible to understand cratonic basins (and sedimentary basins in general) without some reference to their previous geological history. Most fundamentally, the timing of initiation of cratonic basins needs to be placed in the context of relative plate motion and the continent's state of stress. As previously recognized (Klein and Hsui, 1987) the initiation of cratonic basins is not uniformly or randomly distributed in geological time: instead, they initiate at times of break-up of supercontinents (Rodinia, Gondwana, and Pangaea over the last billion years) (Fig. 30.2), particularly at 550–500 Ma. The basins of Africa, North America and South America (Fig. 30.3) illustrate this well (Table 30.1)

In South America, a number of basins can be confidently attributed to rifting in the Jurassic, whereas more typical cratonic basins initiated predominantly in the Precambrian or Early Paleozoic. A major phase of cratonic basin formation started in the Ordovician, *ca.* 450–500 Ma, coinciding with the development of passive margins along the trailing continental edges caused by the separation of South America-Laurentia from the Gondwanan superassembly. Note that the timing of initiation in the Ordovician lags the timing (Cambrian) in the conjugate plate of Africa (see below).

In North America, basin initiation in the Cambrian coincided with the break-up of the North American plate and the formation of passive margins along its perimeter. The timing of initiation is identical to that of many cratonic basins in Africa.

In Africa, there are many basins attributed to rifting or eventual passive margin development, mostly initiated in the Jurassic-Cretaceous. This widespread suite of rift basins is related to the plate-wide tension during the formation of the Atlantic Ocean. A further set of rift-related basins initiated in the late Precambrian-Cambrian. More typical cratonic basins have ages of initiation traced principally to the Cambrian, coeval with the fragmentation of the Gondwanan-Laurentian

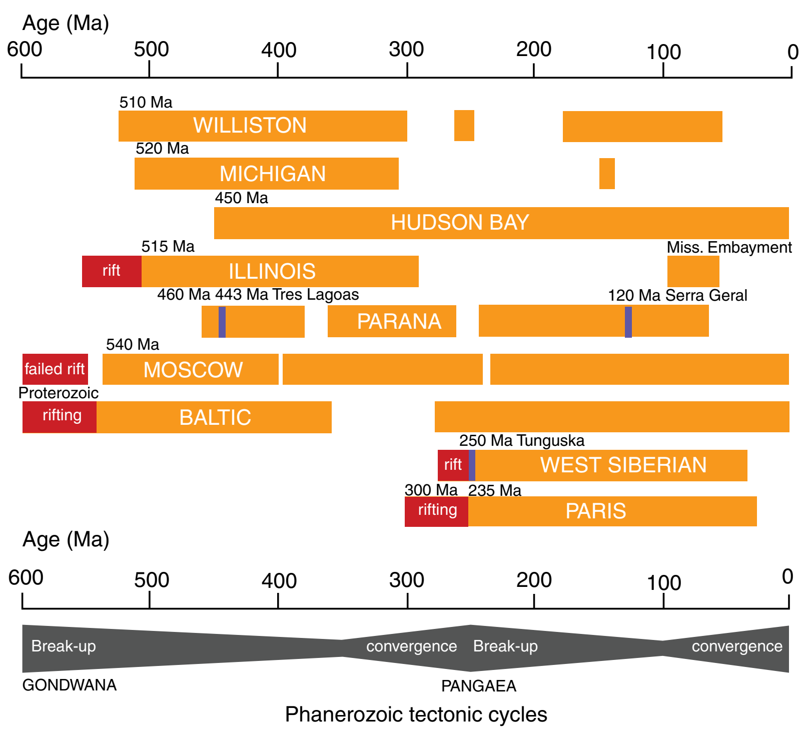

Fig. 30.2. Selected cratonic basins, showing timing of basin-fill megasequences in relation to the two great tectonic cycles of the Phanerozoic. Some cratonic basin megasequences are preceded by rifting, and some have important magmatic episodes (blue bands).

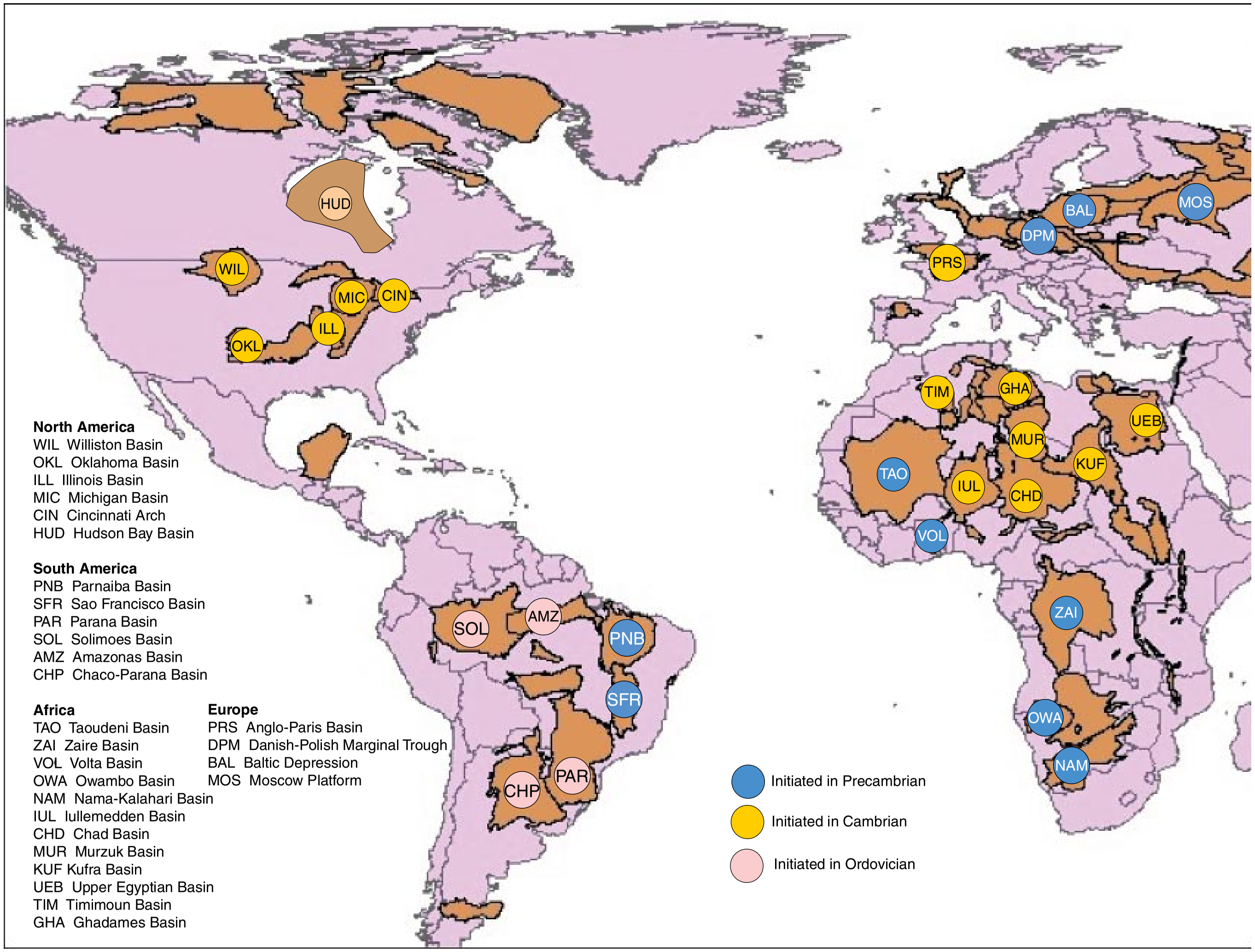

Fig. 30.3. Distribution of intracontinental basins on the continents surrounding the Atlantic Ocean, with typical cratonic basins highlighted. Other basins shown are commonly unequivocally associated with extensional tectonics. Selected basins are colour-coded according to the timing of initiation. *Source*: Base map and basin outlines provided by Trond Torsvik.

Table 30.1. Selection of basins on the North American, South American and African plates with their ages of initiation

Basin type	Basin name (location)	Age of initiation
South America		
Linked to rifting	Reconcavo (Brazil), San Jorge (Argentina-Chile-South Atlantic), Parecis-Alto Xingu (Brazil), Alto Tapajos (Brazil)	Jurassic
Other cratonic basins	Solimoes (Upper Amazon, Brazil), Amazonas (Brazil), Paraná (Brazil, Paraguay, Argentina)	Ordovician
	Parnaiba (Brazil)	Late Precambrian
North America		
Cratonic basins	Williston Basin (Canada, USA), Oklahoma Basin (USA), Illinois Basin (USA), Michigan Basin (USA)	Cambrian
	Hudson Bay Basin (Canada)	Ordovician
Africa		
Linked to rifting	Melrhir Trough (Algeria, Tunisia), Qattara Ridge (Egypt), Cyrenaican Platform (Libya)	Cambrian
	Iullemedden Basin (Niger, Mali, Nigeria, Algeria, Benin)	Cambrian
	Bida Trough (Nigeria), Gongola-Maiduguri Trough (Nigeria), Doba Trough (Chad, Cameroon), Bongor Trough (Chad), Doseo Trough (Chad, Central African Republic), Muglad Basin (Sudan), Melut (Sudan, Ethiopia)	Cretaceous
	East African Rift System western branch, Salamat Basin (Central African Republic, Chad)	Jurassic
Other cratonic basins	Tilrhemt Uplift (Algeria), Bechar Basin (Algeria-Morocco), Mouydir Basin (Algeria), Ahnet Basin (Algeria), Timimoun Basin (Algeria), Murzuq Basin (Libya, Niger, Chad, Algeria), Al Kufra Basin(Chad, Sudan, Libya, Egypt, Niger), Upper Egyptian Basin (Egypt, Sudan Libya), Chad Basin (Niger, Chad, Nigeria, Algeria, Cameroon)	Cambrian
	Taoudeni (Mali, Mauritania), Volta (Ghana, Togo), Zaire (Congo), Owambo (Namibia, Angola), Nama-Kalahari (Botswana, Namibia, Zimbabwe)	Precambrian

superassembly, and, importantly, at the same time as a suite of rifts (Guiraud et al., 2005). During the Cambro-Ordovician, the north African area was one extensive ramp, which was subdivided in the Silurian, and basins became further individualized in phases of contractional tectonics in the Late Carboniferous-Permian and Late Cretaceous-Early Tertiary (Boote et al., 1998). The cratonic basins therefore originated essentially under the same stress regime as the rifts. A subordinate number initiated in the Precambrian.

In Australia, a similar long-term evolution can be recognized. A large ($2 \times 10^6\,km^2$ surface area) cratonic basin known as the Centralian Superbasin formed at the time of break-up of Rodinia (ca. 800 Ma; Korsch and Lindsay, 1989; Walter et al., 1995), but became compartmentalized by phases of intracontinental orogeny in the Petermann (570–530 Ma) and Alice Springs (ca. 400–300 Ma) events (Hand and Sandiford, 1999) to give the individualized Officer, Amadeus and Georgina basins. The Centralian Superbasin was connected to the ocean via the extensional corridor of the Adelaide Geosyncline oriented at a high angle to the plate margin.

The large-scale geodynamic context is therefore strongly suggestive of cratonic basins forming initially at times of plate-wide extensional stress, with or without the involvement of mantle plumes. This is supported by the details of the geological history of selected basins.

The West Siberian Basin

With a surface area of 3.5 million km^2, the post-Permian West Siberian Basin is the largest cratonic basin known. It overlies deeply eroded (during the Permo-Triassic), folded and metamorphosed orogenic belts and platformal blocks stitched together during the Late Paleozoic into the Pangaean superassembly (Ziegler 1989; Sengör and Natal'in 1996).

The Moho depth decreases from 50 km beneath areas fringing the basin (Urals, Altay-Sayan foldbelt and East Siberian Platform) to 38 km beneath the center of the basin. There is more uncertainty in the regional mapping of the base of the lithosphere (Morozova et al., 1999; Artemieva and Mooney 2001), but a depth of 200 km under the basin, decreasing toward the Altay is probable. Extensional rifts are found particularly in the northeast part of the basin and in the Kara Sea, but extend southwards in two arms – the Urengoy and Khudosey Rifts (Nikishin et al., 2002; Saunders et al., 2005) (Fig. 30.4).

The basin-fill is underlain by the 250 Ma Tunguska flood basalts, which cap Permian continental deposits basin-wide, and overlie a rift sequence in the northeast. The initial stretching therefore predates the magmatism, which in turn predates the main period of cratonic basin subsidence. Rifting and the eruption of large volumes of basalts have been attributed to the emplacement of a Permian plume (Nikishin et al., 2002). The basalts spread much wider than the extent of grabens, so basalt occurrence is not closely or directly related to graben formation (Vyssotski et al., 2006). Pre-volcanism rifting is controversial, but is well documented in the Pur-Taz region and Kara Sea: N-S rifts are imaged on seismic reflection profiles, and are presumably filled with Permian sedimentary rocks. But there is no good evidence of pre-Permian rifts south of Pur-Taz (lat of 64°N). The post-basalt Triassic-Middle Jurassic is termed "pre-basin evolution," comprising continental siliciclastics and volcanics deposited in low-relief sag basins, with rifts restricted to the north. From Callovian times (Mid-Jurassic, beginning 159 Ma) basin-wide transgression took place, with sediments fed from a major supply in the southeast, marking the start of the West Siberian Basin as a distinct sag-type depocenter. During the Kimmeridgian-Barremian water depths reached 1200 m in the basin center, showing significant underfilling of the basin, with sediment supply from the Siberian Platform (none from the Urals). In the Turonian marine transgression reopened the West Siberian Basin to the Arctic Ocean, making use of the corridor running above previously stretched lithosphere. The basin filled again by the Oligocene and has remained subaerial since that time. Jurassic-Tertiary subsidence is thought to be due to thermal re-equilibration of the lithosphere-aesthenosphere system following plume activity (Vyssotski et al., 2006).

The Paraná Basin

The Paraná Basin is thought to be a classical or "typical" intracratonic basin (Zalan et al., 1990) (Fig. 30.5). Mesozoic sedimentation along the

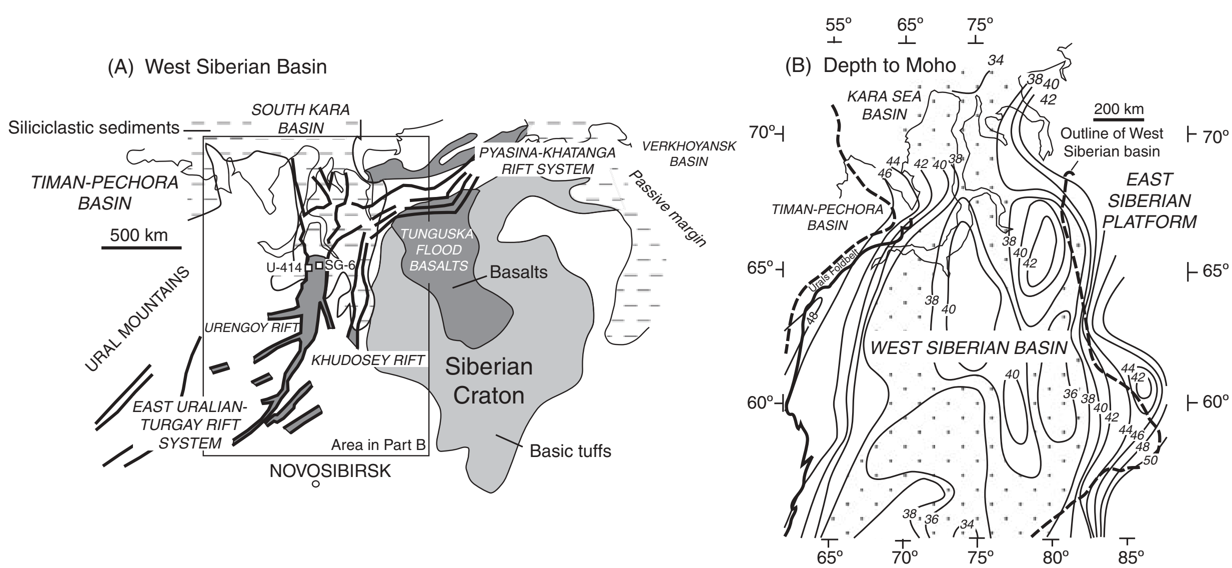

Fig. 30.4. The West Siberian Basin. (A) Location of basin in relation to the extensional plate margin and the Siberian craton. Present day extent of Tunguska flood basalts and basic tuffs shown, with remnants preserved in graben underlying the West Siberian Basin. Boreholes U-414 and SG-6 used in subsidence analysis are also shown in the northern part of the Urengoy Rift. (B) Depth to Moho (km) showing crust stretched to 34–36 km depth beneath the main rifts in the substrate of the basin, compared to >40 km on the basin flanks. *Source*: Modified from Vyssotski et al. (2006).

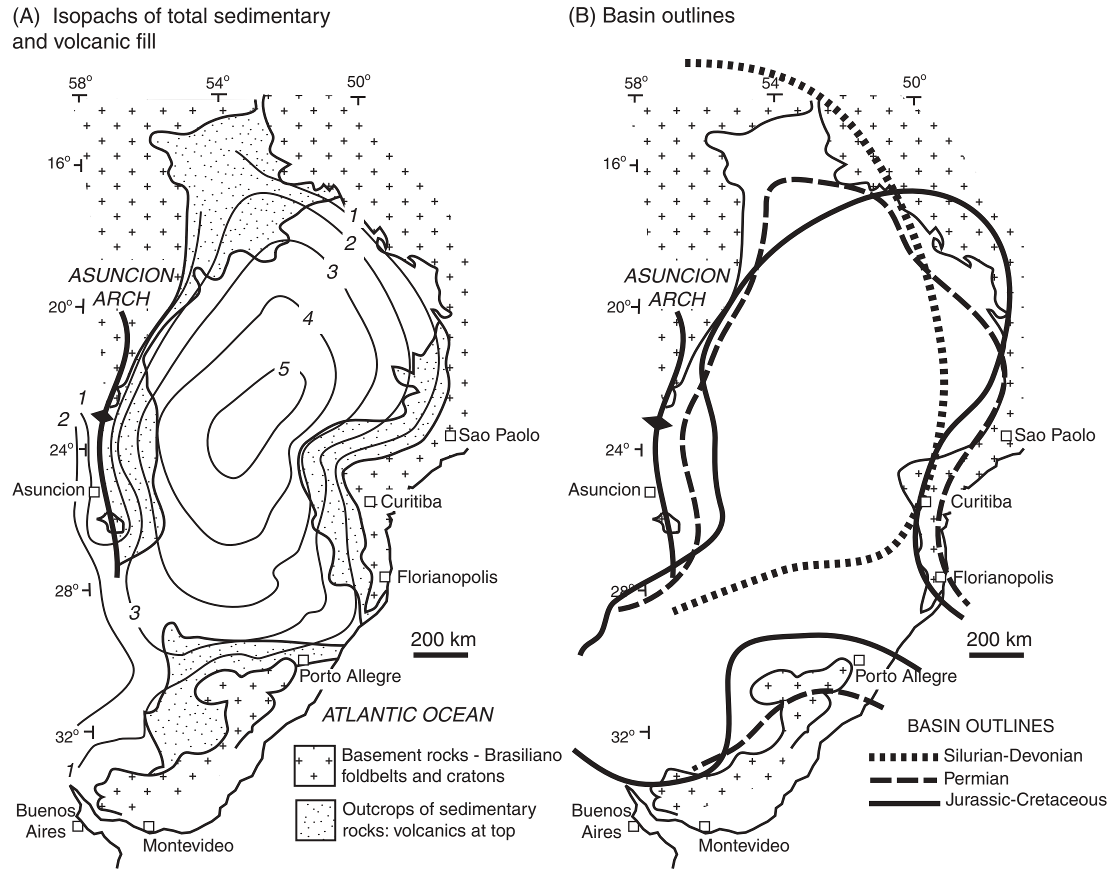

Fig. 30.5. The Paraná Basin. (A) Isopachs of the total sedimentary and volcanic fill, reaching >5 km in the basin center, with a concentric pattern. (B) The Basin outlines at three different stages. The Early Paleozoic basin thickened toward the Asuncion Arch in the west, whereas the Permian and Mesozoic depocenters are concentric with an open corridor to the Atlantic. *Source*: Modified from Zalan et al. (1990).

Brazilian continental margin started with the break-up of Pangaea in the Late Jurassic-Early Cretaceous. The rifting of the southernmost part of the South American plate was heralded by the extrusion of flood basalts in the Paraná, Campos and Santos Basins (Cainelli and Mohriak, 1999), originating presumably from the Tristan plume (Peate, 1997). As in the West Siberian Basin, the flood basalts (Serra Geral volcanics, 137–127 Ma, Turner et al., 1994) are much more extensive than the Paraná Basin itself. They occur at the initiation of subsidence of a megasequence that is clearly related to continental stretching at the margin of the South American plate. In the coastal basins, the syn-rift phase is up to Albian, followed by a drift phase from Albian to Quaternary. This rift-drift phase is therefore only the latest megasequence of the Paraná Basin. The important question is "When did the Paraná Basin become a cratonic basin?"

In the future region of western Gondwana, a suite of sedimentary rocks (Quartzite-Pelite-Carbonate suite of Brito Neves, 2002) formed on the continental margins of the cratonic elements produced by the break-up of Rodinia, including the Paraná fragment. The ocean basins were closed during Neoproterozoic deformation known as the Brasiliano orogenic events, to produce the supercontinent of Gondwana. These late orogenic to postorgenic basins are termed the "transition stage" (Almeida et al., 2000), occupying a position between Brasiliano orogenesis and the subsequent stage of extensive stable platforms in the "cratonic stage" of the Ordovician (Soares et al., 1978). Western Gondwana was covered by vast amounts of sedimentary cover in the Paleozoic, with successions preserved in large basins extending from the Solimoes (upper Amazonia) in the north to the Paraná in the south. Brito Neves et al. (1984) suggested these basins, and their equivalents in

Africa, formed by thermal contraction of the continental lithosphere following the Brasiliano-Pan-African tectonic cycle. In basins such as the Chaco-Paraná and Solimoes basins, the cratonic basin megasequence began in the Early and Middle Ordovician, underwent contractional deformation during the final accretion of Pangaea (ca. 230 Ma; Veevers, 1989) and extensional deformation during its break-up (225 to 100 Ma). The eruption of the Serra Geral flood basalts in the Paraná basin therefore postdates the Ordovician to Early Triassic age of the cratonic basin megasequences in South America, whereas the Tres Lagoas basalts of the neo-Ordovician (443±10Ma) suggests a rift under the basin or at least the passage of melts through a fractured cratonic basement (Julia et al., 2008).

The North American cratonic basins

The advantage offered by the North American basins is that a suite of examples subsided and evolved in a coordinated fashion across the craton. Inspection of maps of basin development on the North American craton (Fig. 30.6) leaves no doubt that cratonic basin mechanics are closely related to plate boundary processes. The Phanerozoic history of plate tectonics in the North American region, from the break-up of Rodinia (ca. 800–700 Ma) to the reassembly of the Pangaean supercontinent (ca. 300 Ma) and its subsequent dispersal (Dalziel, 1991; Hoffman, 1991; Rogers, 1996) provides an essential backdrop for the evolution of its cratonic basins. Here we concentrate on the Paleozoic tectonic cycle.

During the latest Proterozoic to Early Ordovician (>543 Ma to 485 Ma), corresponding to the Sauk sequence of Sloss (1963), the North American craton was extensively inundated, leaving a NE-SW oriented Transcontinental Arch marked by condensed deposition ($<5\,\mathrm{m\,Myr^{-1}}$) or non-deposition (Fig. 30.6A). Although there were elongate, linear regions of accelerated subsidence striking inland at high angles to the plate margin, such as the Central Montana Trough connecting the Williston Basin to the western plate margin, and both the Reelfoot Rift connecting the Michigan and Illinois Basins, and the Southern Oklahoma rift connecting the proto-Anadarko Basin to the Ouachita margin in the southeast, the widespread distribution of sediment accumulation suggests a ramp-like tilting of the craton toward its margins. Such long-wavelength tilting took place at a time of extensional break-up as Laurentia was set adrift from its Gondwanan mooring (Sloss 1988; Bally 1989), leading to the expansion of the Iapetus Ocean. The majority of cratonic basins in North America therefore initiated at a time of extension on the adjacent continental margins, but their characteristically circular planform shape was defined later under a different stress regime.

The history of Phanerozoic subduction can be divided into two main stages: the subduction of Iapetus oceanic crust along the eastern margin during the Paleozoic, and the subduction of Pacific and Farallon oceanic crust along the western margin from the late Paleozoic to the Cenozoic. Iapetus subduction began at ca. 480 Ma and continued to ca. 420 Ma (van der Pluijm et al., 1990; McKerrow et al., 1991), within the Tippecanoe sequence of Sloss (1963) (Fig. 30.6b). Westward subduction of an Iapetus slab is supported by eastward tilting in Middle Ordovician time of sedimentary rocks in the Michigan Basin (Coakley and Gurnis 1995) and was used in the numerical simulation of cratonic sequences by Burgess et al. (1997). The Paleozoic history of the Cordilleran margin of Laurentia is well known but complex (Dickinson 2009), starting as a passive continental margin following rifting of Rodinia, but was strongly modified by Mediterranean-style subduction zone roll-back and extensional back-arc formation during the Antler and Sonoma orogenies. The onset of subduction of Pacific ocean floor is variously dated at Precambrian (Scotese and Golonka 1992) and late Paleozoic (Burchfiel and Davis 1972; Miller et al., 1992). Burgess et al. (1997, Fig. 4, 1519) provided a model with subduction starting at 500 Ma (Cambrian-Ordovician boundary), far from the present western edge of the North American craton.

Cratonic depocenters became individualized during the time span from Ordovician to Mississippian (Tippecanoe and Kaskaskia sequences, Figure 30.6b–d) when the plate margins were undergoing subduction of oceanic crust or continent-continent collision. Subduction of the Iapetus ocean slab in the southeast changed the far-field distribution of stress in the overlying plate, causing widespread subaerial emergence and the formation of an incised unconformity surface between the Sauk and Tippecanoe sequences (ca. 480 Ma). This phase of uplift has been interpreted as the early stages of orogeny (Taconic) in northeastern USA (Hatcher and Viele, 1982), but Burgess et al. (1997) viewed the elevation changes of the

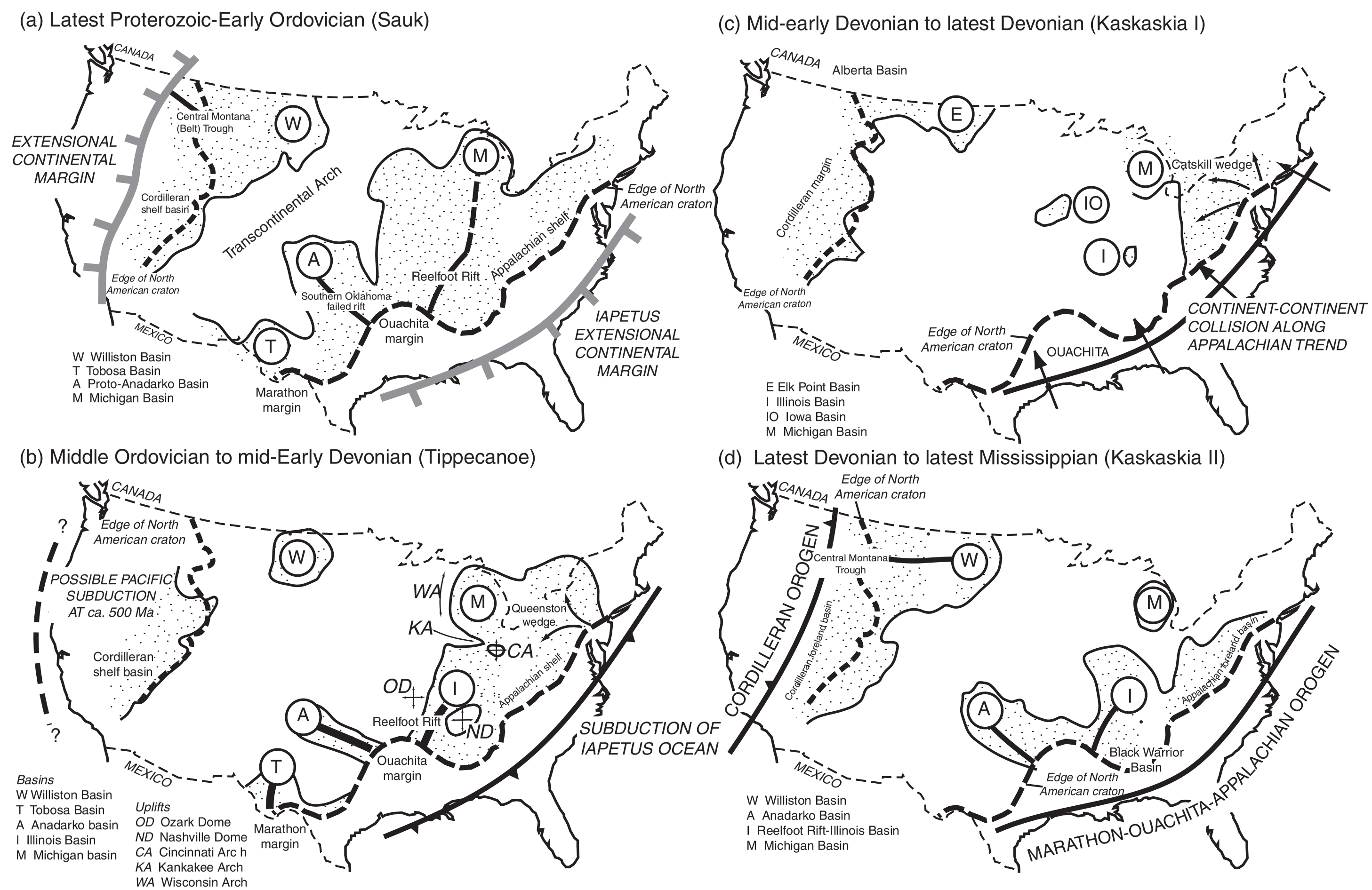

Fig. 30.6. The Paleozoic history of cratonic basins in the USA, modified from Sloss (1988). (a) Latest Proterozoic to Early Ordovician, corresponding to the Sauk sequence. Several cratonic basins were initiated at the tips of rifts striking at a high angle to the extensional plate margin, within a region of very broad tilting of the continent toward both margins. (b) Middle Ordovician to Mid-Early Devonian corresponding to the Tippecanoe Sequence. Cratonic basins become individualized by the growth of intervening arches and domes while subduction of oceanic crust takes place along the eastern plate margin. (c) Mid-Early Devonian to Latest Devonian, corresponding to Kaskaskia I Sequence. Continent-continent collision caused regional uplift of the eastern part of North America and the shedding of major coarse clastic wedges onto the northeastern part of the plate. (d) Latest Devonian to Latest Mississippian, corresponding to Kaskaskia II Sequence. Cratonic basins in the central eastern USA are connected to the Marathon-Ouachita-Appalachian orogen and foreland basin system, while the Williston Basin is connected to the Cordilleran foreland basin system in the northwest.

craton as due to dynamic topography driven by the onset of subduction. Subsequently, the eastern margin of North America was subjected to pulses of ocean-continent collision and accretion along the Appalachian trend, leading to compressive stresses in the craton. By the Late Silurian, continent-continent suturing took place between the Baltica and Laurentian plates. At this stage, the Michigan Basin was individualized as a circular shaped depocenter and its history was yoked to that of the Appalachian orogen and its foreland basin system (Coakley and Gurnis, 1995). In the Early Devonian, the Transcontinental Arch, Ozark Dome, Cincinnati Arch and Nashville Dome were uplifted as the Caledonian orogenic trend became transcurrent and convergence took place between Gondwana and the North American plate. A proto-Tethyan ocean plate subducted under its southern margin at the end of the Devonian to Early Carboniferous. Widespread uplift and erosion allowed the sub-Kaskaskia II unconformity to be cut (368 Ma). At this time, foredeeps developed along the western margin of the North American craton (Bally, 1989), which caused the Williston Basin to be disconnected from the Alberta foreland basin system. Subsequently, by latest Mississippian, major continent-continent collision between Gondwana and the North American plate led to widespread uplift and erosion of the sub-Absaroka unconformity and the reactivation of old lineaments. The Absaroka I sequence (beginning 330 Ma) marks a change of style of sedimentation on the North American craton, due to a new cycle of extension, subduction, convergence and collision concentrated in the south and west of the craton.

The North American cratonic basins show unequivocally that basin subsidence began as a ramp-like tilting of the continent toward extensional, trailing-edge plate margins, with local rapidly subsiding troughs extending across weakly subsiding continental platforms into the cratonic interior at high angles to the plate margin. In eastern and southeastern North America, the onset of subduction of oceanic crust caused a change of state of stress, during which time regionally correlated unconformities were formed and basin stratigraphy tilted. Basins such as the Michigan and Illinois basins became "individualized" as roughly circular depocenters fringed by uplifting arches, domes and bulges. Consequently, the later evolution of cratonic basins is moderated by the various effects of dynamic topography set up by subduction of cold oceanic slabs or convective instabilities, in-plane stress transmitted from a convergent plate boundary, flexure caused by tectonic loading in convergent orogens, and the spreading of siliciclastics from foreland basin systems into the plate interior.

In the paragraphs that follow, we investigate the effects of stretching at low strain rate a relatively thick continental lithosphere as the dominant mechanism for cratonic basin inception and long-term evolution. Other effects, such as those suggested in the paragraph above, are treated as secondary.

MODELING CRATONIC BASIN EVOLUTION

Simple instantaneous uniform extension leads to fault-related subsidence as the lithosphere thins, and further post-rift subsidence as the thinned lithosphere cools back to its initial state (McKenzie, 1978). Cratonic basins may have formed due to extension by very small stretch factors, for example 1.05 to 1.2 in the Hudson Bay (Hanne et al., 2004) and <1.6 in the West Siberian Basin (Saunders et al., 2005). Assuming a normal initial lithospheric thickness of 125 km, instantaneous extension to these stretch factors produces subsidence that is too rapid and does not last long enough to reproduce typical cratonic basin subsidence history (Fig. 30.7). However, simple instantaneous extension of a 200 km-thick lithosphere produces subsidence that continues for more than 250 Myr due to the effect on the lithospheric time constant $\tau = z_0{}^2/\pi^2\kappa$ (Fig. 30.7), where z_0 is the initial lithospheric thickness and κ is the diffusivity.

The concept of instantaneous extension is useful for simple modelling of sedimentary basins that form due to extension at strain rates greater than $10^{-15}\,s^{-1}$ (ca. 30 nannostrain yr^{-1}) (Fig. 30.8). On this basis the assumption of instantaneous extension is justified (Jarvis and McKenzie, 1980). We have modelled the uniform extension in 1D, assuming a constant strain rate, to understand where this assumption of instantaneous extension breaks down and what the effect of low strain rates would be on subsidence (Armitage and Allen 2010). Temperature is calculated down a 1D depth profile. Extension in the horizontal plane leads to a proportional thinning in the vertical. This assumes that the upper lithosphere deforms plastically and conserves volume. The thermal

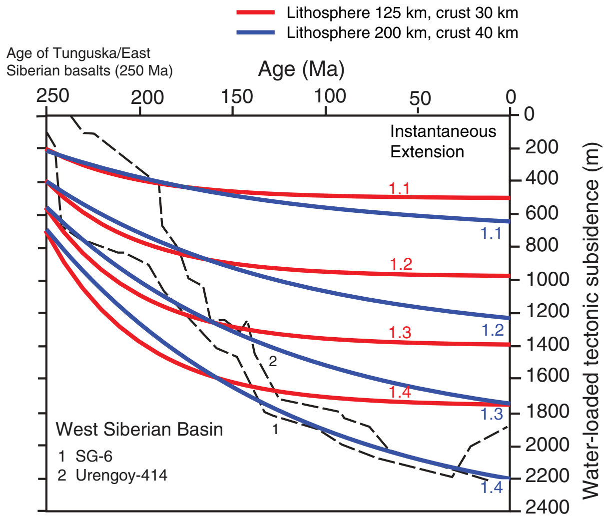

Fig. 30.7. Predicted tectonic water-loaded subsidence curves for instantaneous extension of the lithosphere for initial lithosphere thickness of 125 and 200 km at stretch factors between 1.1 and 1.4. The predicted subsidence is plotted together with the backstripped subsidence of boreholes SG-6 and Urengoy-414 (located on Figure 30.3) in the West Siberian Basin (Saunders et al., 2005). The model subsidence assumes that the temperature at the base of the lithosphere is 1330°C and that density varies with temperature within the lithosphere and is constant within the crust (McKenzie, 1978).

structure of the lithosphere is therefore subject to heat conduction, upward advection due to the lateral extension and internal heating due to radioactive decay. The thermal structure of the lithosphere, the geotherm, is then solved through time as the lithosphere extends and cools. The composition of the upwelled aesthenosphere is assumed to be identical to that of the lithospheric mantle it replaces so that there is no additional downward-acting load due to a change in chemical composition and therefore density (cf. Kaminski and Jaupart, 2000).

Subsidence is calculated assuming that the lithosphere is isostatically compensated locally (Airy) and that the initial configuration is at sea level. We assume that density remains constant within the crust at 2900 kg m^{-3}, and varies linearly due to thermal expansion within the mantle lithosphere. The density at the base of the lithosphere is assumed to be 3400 kg m^{-3}, and the temperature is assumed to be 1330°C. Within our model, the subsidence is relatively insensitive to the aesthenospheric temperature.

For very low rates of extension (strain rates of 10^{-16} s^{-1}) the heat conduction and upward advection of material become comparable. While the crust is mechanically stretched, warmer lithosphere advects upwards. If this upward advection is rapid, the surface is uplifted due to the less dense warm mantle beneath the crust. When the strain rates are sufficiently low, however, the upwelling mantle cools as it rises, causing the geotherm to increase less than in the situation where stretching is instantaneous or at a faster strain rate (Fig. 30.9). In this case the lithosphere does not become as buoyant so as to counter the subsidence due to the stretched crust. The result is syn-stretching subsidence that is more prolonged and greater when the lithosphere is stretched to the same stretch factor at lower strain rates (Fig. 30.8). Subsequent

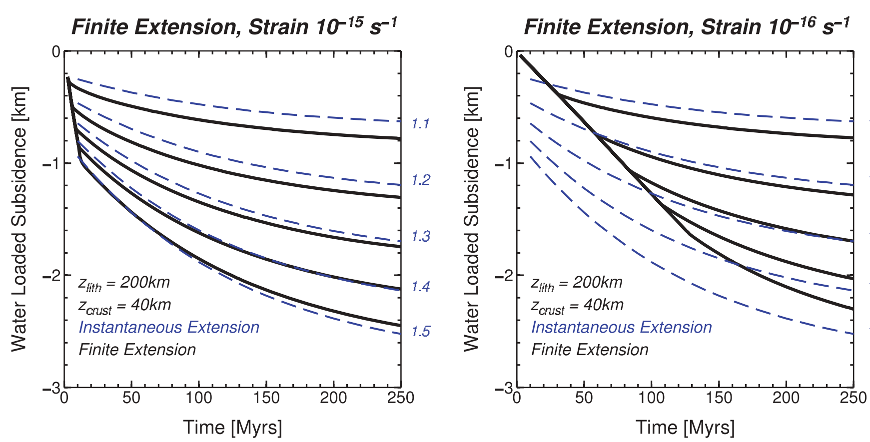

Fig. 30.8. Comparison of water loaded subsidence for instantaneous extension with finite extension at strain rates of 10^{-15} and 10^{-16} s^{-1}. The base of the lithosphere is assumed to be at 200 km depth with a temperature of 1330 °C. The crust is assumed to be 40 km thick.

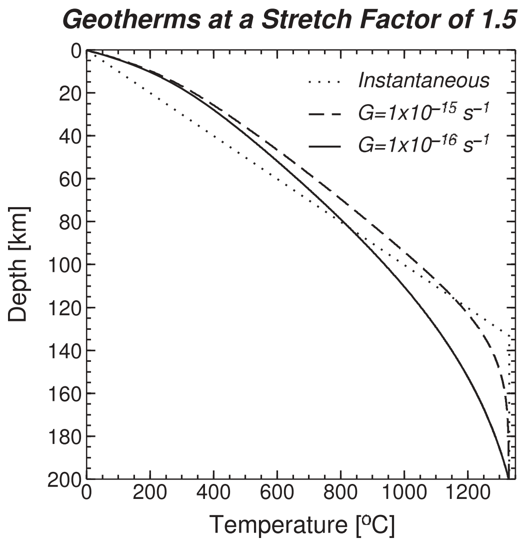

Fig. 30.9. Geotherms for extension at a stretch factor of 1.5. Instantaneous assumes no internal heating due to radioactive decay. The models of extension at strain rates, *G*, of 10^{-15} and $10^{-16}\,s^{-1}$ assume internal heating. The temperature at the base of the lithosphere is assumed to be 1330 °C, and the initial lithospheric thickness is 200 km.

post-stretching subsidence due to the thermal relaxation of the lithosphere is not as great compared to when the lithosphere is extended at faster strain rates. The subsidence profile therefore has a more constant slope, with a slight elbow when the stretching ceases and the lithosphere thermally relaxes (Fig. 30.9).

DISCUSSION

There are a number of uncertainties that make the understanding of cratonic basins problematical. One is the thickness of the thermal lithosphere at the time of basin formation, another is the impact of densification due to mineralogical changes on cratonic basin subsidence, and another is the state of stress of the continental plate at the time of basin initiation. We review these topics briefly.

We are able to explain the subsidence history of cratonic basins primarily by stretching of a thick continental lithosphere at low strain rates. But evaluating the physical state of the continental lithosphere at the time of formation of cratonic basins and during their long evolution is difficult. Lithospheric thickness, based on **S** wave velocities, shows strong variations from plate to plate and within large plates (Goes et al., 2000; Artemieva and Mooney 2001; Goes and van der Lee 2002; McKenzie et al., 2005), with large gradients across major tectonic lineaments (Pérez-Gussinyé and Watts 2005). Lithospheric thickness is also suggested to be affected by the long-term secular cooling of the mantle (Pollack, 1997; Herzberg, 2004). In the case of the West Siberian Basin, the Russian platform has a present-day lithospheric thickness of 160–210 km, with thinner regions (125 km) beneath the Central Russian Rift system and thicker regions under the Siberian shield (>350 km) (Artemieva and Mooney 2001; Artemieva, 2003). In contrast, thin lithosphere but normal crust might be expected in regions of long-term accretion of arcs, subduction zones and backarc lithosphere (Murphy and Nance 1991). Constraining the initial thickness of the lithosphere at the time of the initiation of cratonic basins is therefore difficult.

Subsidence results from changes in the density structure of the lithosphere during extension and once extension ends. It is normally assumed that density is a function of temperature. For rift margins, where melting is involved, there is an additional buoyancy due to melt retained within the mantle (Scott, 1992). Solid-state phase changes also alter the density structure of the lithosphere. For simple models of lithosphere extension, the phase change from garnet- to plagioclase-lherzolite may reduce the lithosphere density by up to 100 kg m^{-3} (Podladchikov et al., 1994, Kaus et al., 2005). The bulk composition of continental lithosphere is characterised by the average composition of peridotite xenoliths. For post Archaean lithosphere, we assume that the bulk composition of the lithosphere is that of an average spinel peridotite (McDonough, 1990). Using this bulk composition we calculated the density of the lithosphere using the thermodynamic code *Perple_X* (Connolly, 1990, 2005; Connolly and Kerrick, 2002). For continental lithosphere there is a slight alteration in the density due to the phase change from garnet- to plagioclase-lherzolite. However, the lithosphere would have to be extended by a large amount for this phase change to have an effect on subsidence, contrary to the evidence afforded by cratonic basins. Consequently, we do not expect any mineralogical changes that would significantly alter the density structure of the lithosphere during their formation.

More complex phase changes within the lithosphere and crust have been put forward to explain subsidence within sedimentary basins. For example the gabbro-eclogite phase change might potentially cause subsidence (Joyner, 1967; Haxby et al., 1976; Stel et al., 1993), as this causes a density increase of around 500 kg m^{-3}. This phase change has been used to explain the formation of the Williston and Michigan Basins in North America (Haxby et al., 1976, Baird et al., 1995) and the late-stage evolution of the Baltic Basin (Stel et al., 1993). However, to generate this phase change a source of heating such as a large upwelling of warmer material to the base of the crust is required (Baird et al., 1995), for which no consistent evidence is available.

We propose that although there are secondary mechanisms for subsidence during the long time evolution of cratonic basins, the primary mechanism is stretching of the continental lithosphere at a low strain rate followed by cooling. The origin of the far-field extensional stresses causing these low strain rates of the continental lithosphere is debatable, but the regional state of stress of a continent can be linked to its position relative to the geoid and to the impact of plate boundary forces (Coblentz and Sandiford, 1994; Coblentz et al., 1998; Hillis and Reynolds, 2000). Surrounded by mid-ocean ridges, like Africa today and Australia prior to 50 Ma, the continent should experience weak extensional deviatoric stresses.

The break-up of supercontinents and their subsequent dispersal are likely driven by the migration of continents from geoid highs to geoid lows (Gurnis, 1988). Numerical modelling has found that a stationary continental plate insulates the mantle below, generating an upwelling of mantle material as the warm mantle creeps laterally beneath the plate (Gurnis, 1988). Such an insulating effect has been inferred from the stability of continental cores (McKenzie and Priestley, 2008). Alternatively, plumes impinging on the base of the supercontinental lithosphere would also cause topographic doming (Zhong et al., 1996). Both processes would place the overlying continental lithosphere in a state of horizontal extensional stress. Tensional stresses would be relieved as the continent migrated to a region of geoid low and the state of stress switched to compressional (Gurnis, 1988).

The extensional stress driving the very low strain rates necessary to explain cratonic basins may therefore be part of the cycle of supercontinental break-up. Using an initial lithospheric thickness of 200 km and an extensional strain rate of 10^{-16} s^{-1}, the subsidence history of two boreholes in the West Siberian Basin can be matched with a stretch factor of 1.5 (Fig. 30.10). Further complexity is evident in the subsidence histories of the Michigan and Illinois Basins

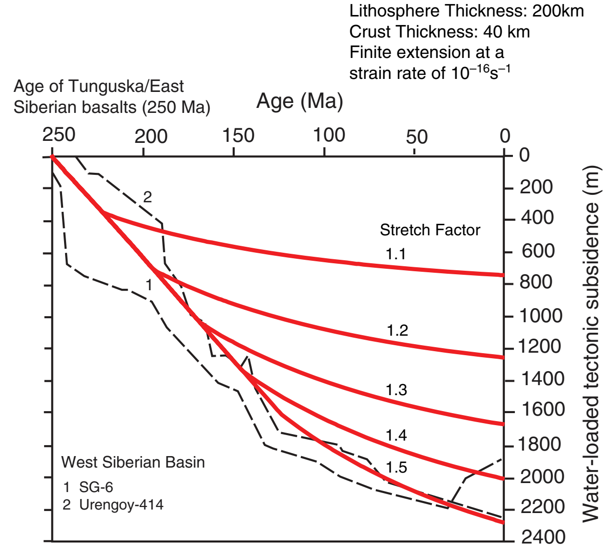

Fig. 30.10. Water loaded tectonic subsidence of the West Siberian Basin (boreholes SG-6 and U-414 located in Figure 30.3, after Saunders et al., 2005) compared with extension of 200 km-thick lithosphere with 40 km-thick crust at a low strain rate of 10^{-16} s^{-1} until stretch factors of 1.1 to 1.5 are reached. Different stretch factors are achieved by different durations of stretching at a constant strain rate.

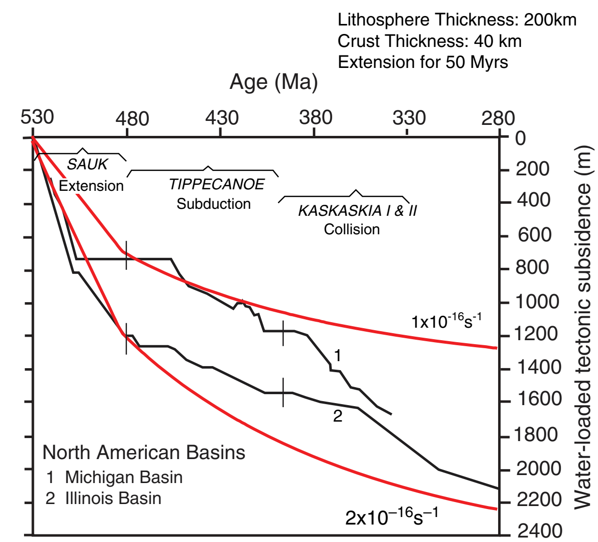

Fig. 30.11. Water loaded tectonic subsidence for boreholes representative of the Michigan and Illinois Basins (from Xie and Heller, 2009) with an envelope of model subsidence for strain rates of 10^{-16} s^{-1} and 2×10^{-16} s^{-1} imposed for 50 Myr, followed by cooling. For the lower strain rate after 50 Myr the stretch factor is 1.17. For the higher strain rate after 50 Myr the stretch factor is 1.37. Different segments of the subsidence curve correspond to the Sauk, Tippecanoe and Kaskaskia sequences. Secondary subsidence mechanisms superimposed on cooling affect the Tippecanoe and Kaskaskia sequences.

(Fig. 30.11). Extension is modelled from 530–480 Ma (Sauk sequence) at strain rates of 10^{-16} and $2 \times 10^{-16}\,s^{-1}$, followed by cooling. The onset of subduction of the Iapetus ocean (Tippecanoe), and the formation of a continental collision zone (Kaskaskia) (Fig. 30.6) can be recognized as departures from the background curve for thermal subsidence. During these basin-modifying stages, we envisage that the observed subsidence history is an amalgam of background subsidence due to cooling, dynamic topography due to the subduction of a cold oceanic slab, flexural subsidence related to orogenic loading, and the transmission of in-plane stresses from the eastern convergent plate boundary.

CONCLUSIONS

Cratonic basins are large saucer-like sedimentary basins with a roughly circular planform outline located on relatively stable continental lithosphere well inboard of the continental margin. They subside over very long periods of time but at a low rate. Their longevity means that the basin fill is commonly polyhistory (Kingston et al., 1983), with a number of megasequences representing distinct phases of basin evolution. Consequently, their subsidence histories are an amalgam of different forcing mechanisms rather than one.

Typical cratonic basins appear to initiate at times of rifting and drifting on the adjacent plate margin, under a state of extensional horizontal stress, by means of a regional tilting and flooding of the continental surface. Basin initiation is related to the break-up of supercontinents and the drifting away of continental fragments, in the Neoproterozoic, Early Paleozoic and Early Mesozoic. Later important modifications, including the individualization of circular depocenters, may be connected with the onset of subduction, the development of mantle flow instabilities and, later, continental collision. Consequently, the extensional primary mechanism for subsidence may be followed by the varied influences of dynamic topography related to the subduction of a cold oceanic slab or the formation of instabilities at steps in the base of the lithosphere, flexural subsidence by tectonic loading, and in-plane compression from the convergent plate margin.

The subsidence histories of cratonic basins cannot be explained by a model of instantaneous stretching of a normal thickness lithosphere. One way in which typical subsidence histories of cratonic basins can be explained is by a model of protracted stretching at low strain rate of an initially thick continental lithosphere.

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