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Precambrian crustal evolution of Peninsular India: A 3.0 billion year odyssey

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ABSTRACT

The Precambrian geologic history of Peninsular India covers nearly 3.0 billion years of time. India is presently attached to the Eurasian continent although it remains (for now) a separate plate. It comprises several cratonic nuclei namely, Aravalli–Bundelkhand, Eastern Dharwar, Western Dharwar, Bastar and Singhbhum Cratons along with the Southern Granulite Province. Cratonization of India was polyphase, but a stable configuration between the major elements was largely complete by 2.5 Ga. Each of the major cratons was intruded by various age granitoids, mafic dykes and ultramafic bodies throughout the Proterozoic. The Vindhyan, Chhattisgarh, Cuddapah, Pranhita–Godavari, Indravati, Bhima–Kaladgi, Kurnool and Marwar basins are the major Meso to Neoproterozoic sedimentary repositories. In this paper we review the major tectonic and igneous events that led to the formation of Peninsular India and provide an up to date geochronologic summary of the Precambrian. India is thought to have played a role in a number of supercontinental cycles including (from oldest to youngest) Ur, Columbia, Rodinia, Gondwana and Pangea. This paper gives an overview of the deep history of Peninsular India as an introduction to this special TOIS volume.

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1. Introduction

The Indian subcontinent covers approximately 5,000,000 km^2. Although India is connected geographically to the Eurasian continent, the subcontinent and Himalayan sectors make up a distinct lithospheric plate. Our goal is to outline the basic Precambrian geology of Peninsular India. Such an effort would require a tome-length treatise (e.g. Naqvi and Rogers, 1987; Naqvi, 2005; Balasubramanyan, 2006; Ramakrishnan and Vaidyanadhan, 2008; Sharma, 2009) and we apologize at the outset for any over simplifications or omissions and hope that the extensive reference list provides the reader with further information. In this paper, we summarize the geological history of Peninsular India including the constituent cratonic nuclei, their bordering orogenic belts, intrusive and extrusive cover and the sedimentary basins. We cover ~3.0 billion years of geologic history and intend to give the reader an overview of this important continental block. The odyssey begins with a view of the cratonic nuclei that collectively form Peninsular India (Aravalli–Bundelkhand Cratons, Singhbhum and Bastar Cratons, the Western and Eastern Dharwar Cratons and the Southern Granulite Province; Fig. 1). The description of each craton includes a discussion of the progressive stabilization of the block, post-stabilization intrusive events including mafic dyke swarms and a description of Proterozoic sedimentary basins developed within the cratons. Mafic dyke swarms are indicators of crustal extension and can represent supercontinent assembly and/or dispersal, subduction, large igneous province emplacement, and crust/mantle interaction. We use 'stabilization' in the same manner as Rogers and Santosh (2003) and consider a craton stabilized when it is intruded by undeformed plutons, when whole rock isotopic systems become closed and platform sediments are deposited on the newly formed basement. We conclude with a summary of the drift history of India and its position in the supercontinents of Columbia, Rodinia and Gondwana.

Geochronologic studies on the Indian subcontinent have a long history, but many of the published ages are of the older whole rock Rb–Sr and K–Ar variety that are considered less reliable by modern standards. Wherever possible, we state more recent vintage U–Pb, Pb–Pb, Ar–Ar and Sm–Nd ages. We caution that wherever the older studies are cited, they reflect our attempt to provide some broad constraints.

2. Aravalli and Bundelkhand cratons

The Aravalli–Bundelkhand protocontinent (Fig. 2) occupies the north-central region of the Indian subcontinent. The Great Boundary Fault (GBF) divides the protocontinent into two blocks: the

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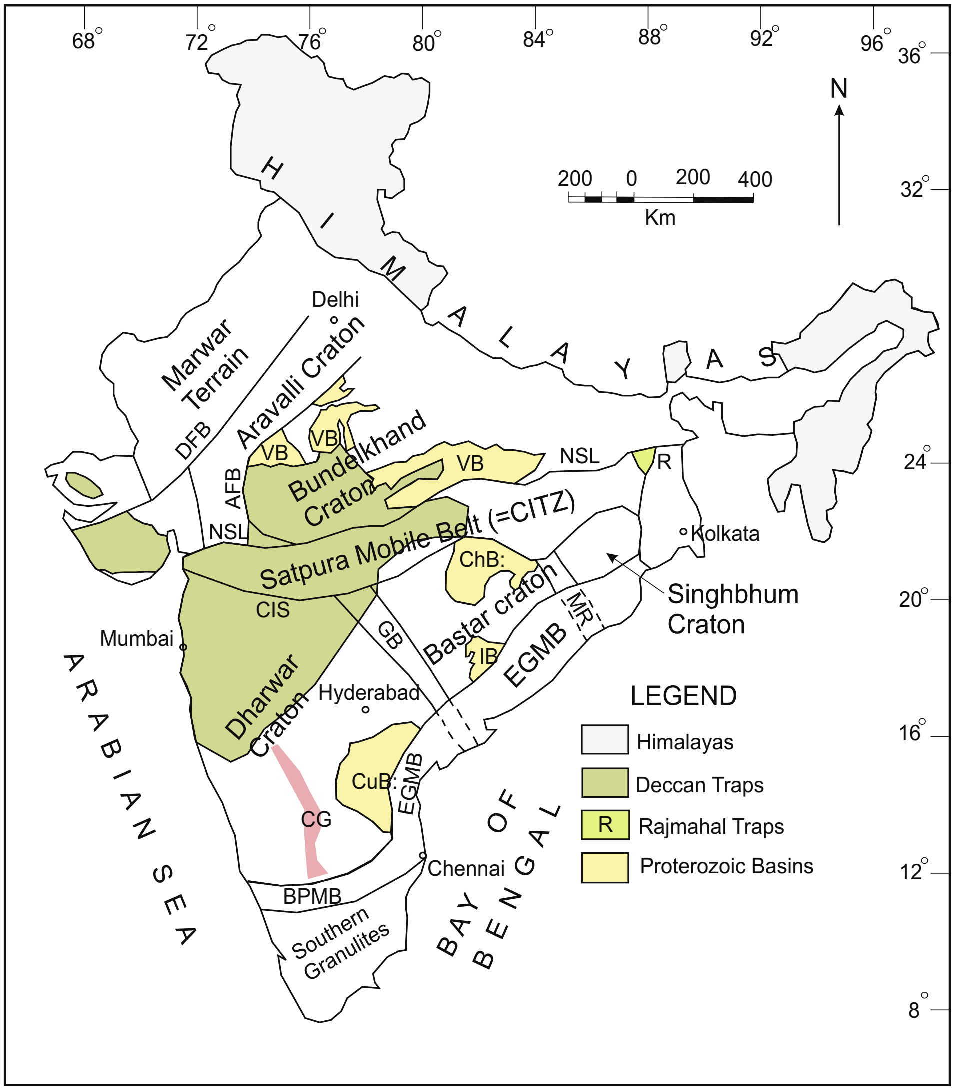

Fig. 1. Generalized tectonic map of Indian subcontinent: Precambrian Cratons, Mobile Belts and Lineaments. AFB = Aravalli Fold Belt, DFB = Delhi Fold Belt, EGMB = Eastern Ghats Mobile Belt, SMB = Satpura Mobile Belt, NSL = Narmada-Son Lineament, CIS = Central Indian Suture and BPMP = Bhavani–Palghat Mobile Belt. Proterozoic Rifts and Basins include IB = Indravati Basin, ChB = Chhattisgarh Basin, GB = Godavari Basin; MR = Mahandi Rift; CuB = Cuddapah Basin, VB = Vindhyan Basin. CG = Closepet Granite (modified from Rao and Reddy, 2002).

Aravalli cratonic block to the west of the GBF and the Bundelkhand–Gwalior block to the east. These cratons are bounded to the north-east by the Mesoproterozoic-aged Vindhyan Basin and Phanerozoic alluvium and to the south by the northern edge of the Deccan Traps volcanic rocks. The Bundelkhand and Aravalli Cratons are also separated from the Bastar and Singhbhum Cratons by the Narmada-Son lineament (Naqvi and Rogers, 1987: Goodwin, 1991).

Most of the Aravalli Craton is underlain by the 3.3–2.5 Ga Banded Gneissic Complex (Heron, 1953; hereafter BGC). The BGC is composed of migmatites, gneisses, meta-sedimentary rocks and minor amphibolite. The BGC in the Aravalli mountain region forms the basement for the supracrustal rocks of Aravalli and Delhi fold Belts. The ion microprobe $^{207}Pb/^{206}Pb$ zircon studies by Wiedenbeck et al. (1996), strongly suggest a $\sim$2.5 Ga stabilization age for the southern segment of the Aravalli Craton based on the uniformity of the Late Archaean and Early Proterozoic crystallization ages (see also Roy et al., 2005). The sediments of Aravalli Supergroup were unconformably deposited on this stabilized landmass. For a considerable time the poorly constrained ages of 2.5–1.9 Ga and 1.8–0.85 Ga, respectively, were the only available ages for constraining the Aravalli and Delhi Supergroups sedimentation (Gupta et al., 1980). Pb–Pb model ages for the Aravalli Supergroup range from 2075 to 2150 Ma (Deb, 1999; Deb and Thorpe, 2004). Deb (1999) reported a 2075–2150 Ma Pb–Pb age for galena, presumably syngenetic with the basal Aravalli volcanics. In the absence of any direct geochronologic evidence this age is taken to represent the initiation of Aravalli sedimentation. Further support is provided by the recent studies of Pandit et al. (2008) who describe the paleosol below the Aravalli Supergroup to have developed during the Great Oxidation Event (GOE). Intrusion of 1850 Ma Darwal Granite (whole rock Rb–Sr age – Choudhary et al., 1984) has generally been accepted as the closing age for deposition of the Aravalli Supergroup.

The ages of metamorphism in the Aravalli Craton are better constrained. Roy et al. (2005) argue that the main pulse of metamorphism took place between 1725 and 1621 Ma at the outset of the Delhi Orogenic Cycle. Buick et al. (2006) also obtained metamorphic ages of $\sim$1720 Ma for granulites of Sandmata Complex which was formerly thought to be part of the Archaean BGC. Kaur

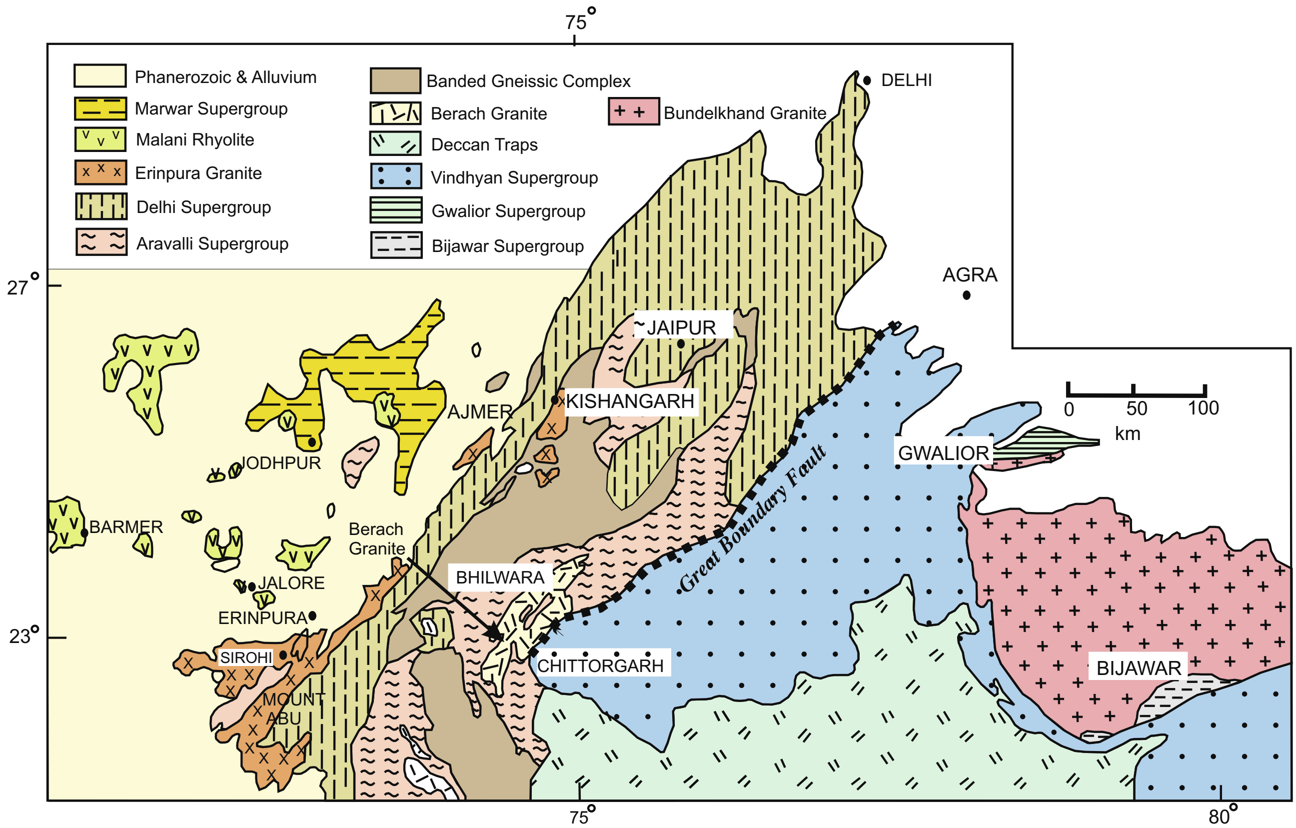

Fig. 2. Sketch map of the major units in the Aravalli–Bundelkhand protocontinent, NW India (after Naqvi and Rogers, 1987; Ramakrishnan and Vaidyanadhan, 2008).

et al. (2007) determined two distinct ages of granitoid intrusions in the northern sector of Delhi Fold Belt. The best constrained of these magmatic events took place between 1711 and 1660 Ma, that these authors attribute to an extensional setting. An A-type geochemistry and 1780–1710 Ma (zircon U–Pb–Th EPMA) age for various granitoids plutons in northern sector of Delhi Fold Belt was proposed by Biju Sekhar et al. (2003) implying a protracted period of extensional tectonics. These ages are supported by the ^{207}Pb/^{206}Pb and U/Pb isotopic data and appear to be coincident with the main phase of metamorphism documented by both Roy et al. (2005) and Buick et al. (2006). Kaur et al. (2007) have also quoted BRGM (French Bureau de Recherches Geologiques et Minieres) unpublished 1844 and 1832 Ma ages for granitoids and felsic volcanics. These results are further supported by the 1821 Ma single zircon Pb–Pb evaporation age of 'Andean type' Jasrapura granite (Kaur et al., 2009). Granitoid intrusions in the southern sector of Delhi Fold Belt are much younger and can be assigned to two distinct age groupings, the ~1 Ga calc-alkaline Sendra Granite (Pandit et al., 2003) and 860–800 Ma granitoids (Deb et al., 2001; Malone et al., 2008; van Lente et al., 2009), the latter popularly known as 'Erinpura Granite' (Heron, 1953). Significant tectono-metamorphic events in the Aravalli mountain region have been summarized in Deb et al. (2001).

Buick et al. (2006) also report a younger metamorphic (shear) episode that took place between ~950 and 940 Ma, based on their ^{235}U/^{206}Pb isotopic ratios on the meta-sedimentary rocks of the Mangalwar complex, and suggested that these ages may be part of a larger metamorphic and igneous event that occurred during this time interval. Support for this metamorphic event is also observed in detrital zircon spectra from the Sonia and Girbhakhar sandstones in the Marwar Supergroup of Rajasthan (Malone et al., 2008). This was followed by the onset of Malani felsic extrusive and intrusive igneous activity (800–750 Ma; Torsvik et al., 2001a,b; Gregory et al., 2009; van Lente et al., 2009). The Malani felsic province is overlain by the Neoproterozoic Marwar Supergroup.

The Bundelkhand Craton, to the east of the Aravalli-Delhi fold belt, is a relatively less studied region (Fig. 3). Sharma and Rahman (2000) divide the Bundelkhand Craton into three distinct litho-tectonic units: (1) Archaean enclaves constituted by highly deformed older gneisses–greenstone components – the Enclave suite; (2) Undeformed multiphase granitoid plutons and associated quartz reefs – the Granite Suite; and (3) Mafic dyke swarms and other intrusions – the Intrusive suite (Soni and Jain, 2001). The Archaean enclave suite is composed of intensely deformed basement rocks; predominantly, schists, gneisses, banded iron formations, mafic volcanic rocks and quartzites. These basement rocks are intruded by the Bundelkhand Igneous Complex that constitutes about 80% of the outcrop of the Bundelkhand Craton (Goodwin, 1991; Basu, 2007). Three generations of gneisses are thought to have formed at 3.2 Ga, 2.7 Ga and 2.5 Ga, respectively as indicated by ^{207}Pb/^{206}Pb isotopic data (Mondal et al., 1997). The latter 2.5 Ga age is also considered as the stabilization age of the Bundelkhand massif. The ages of the enclaves are not known, but there are a few ages on the granites that intrude them. The Bundelkhand granite is dated to 2492 ± 10 Ma (Mondal et al., 2002) and the Berach Granite to 2530 ± 3.6 Ma using U–Pb isotopic dating (R.D. Tucker, personal communication, see also Wiedenbeck et al., 1996). Numerous mafic dykes intrude the Bundelkhand Igneous Complex. Rao (2004) suggests that most of these mafic dykes were emplaced in two phases, one at 2.15 Ga and the second at 2.0 Ga, based on their ^{40}Ar/^{39}Ar isotopic analyses. An earlier study by Sarkar (1997) reported two distinct K–Ar age clusters (~1800 Ma and 1560 Ma) on mafic dykes intruding the Bundelkhand Province. A younger suite of mafic dykes may be related to the Late Cretaceous Deccan Traps igneous events. The paleomagnetic directions dem-

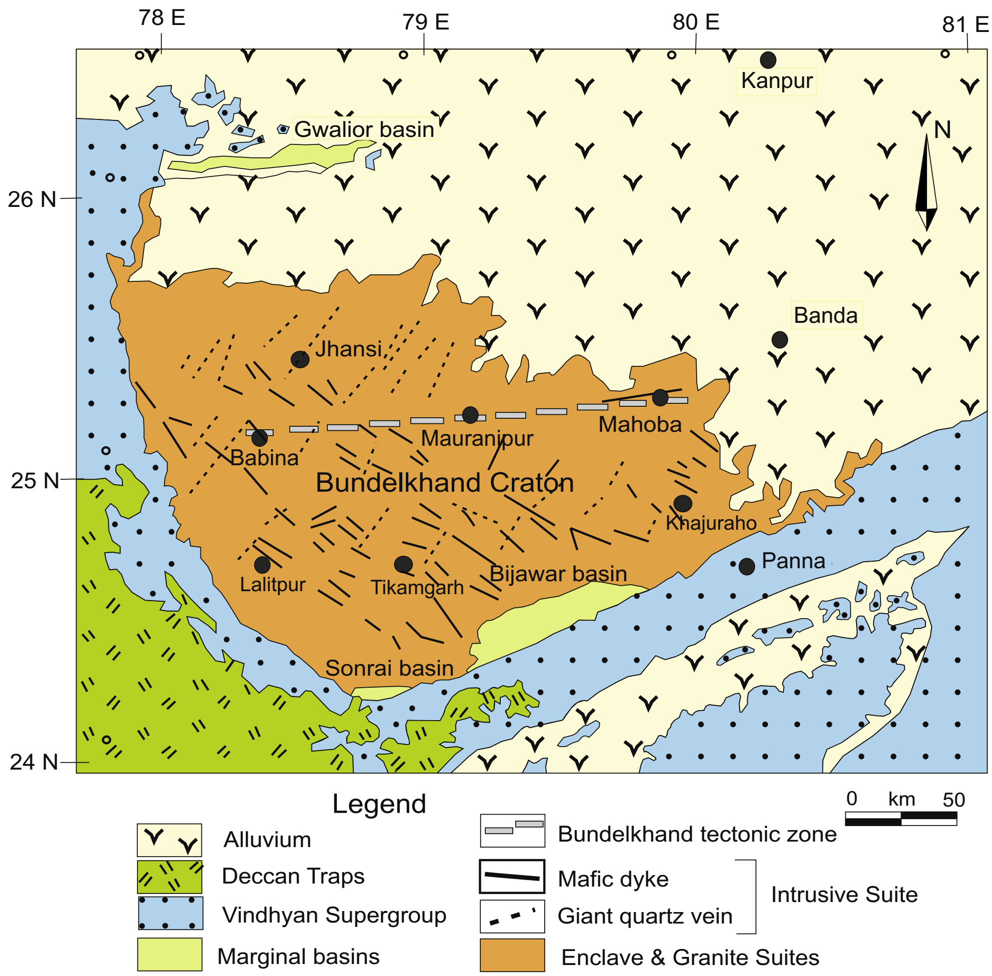

Fig. 3. Sketch map of the Bundelkhand Craton showing mafic dyke swarms including the Great Dyke of Mahoba (white-dashed line; after Malviya et al., 2006).

onstrate at least three generations of dykes in the BGM (Meert, unpublished data) that is consistent with the cross-cutting relationships and the available geochronological data. Ahmad et al. (1996) correlate the Bundelkhand mafic dykes to the Aravalli mafic dykes based on the similarity in their composition and Nd model ages, however, neither the ages nor the geochemistry are robust enough to confirm such a connection.

The old Hindoli Group (1854 ± 7 Ma; Deb et al., 2002) is interleaved with the basement complex. They are in turn overlain by the sedimentary sequences in the Vindhyan Basin (~1700–1000 Ma; Rasmussen et al., 2002; Ray et al., 2002, 2003; Sarangi et al., 2004; Malone et al., 2008). The Lower Vindhyan and lower part of Upper Vindhyan basinal sequences are intruded by the 1073 ± 13.7 Ma Majhgawan kimberlite (Gregory et al., 2006). The red bed and evaporitic sequences of Marwar basin, developed west to the Aravalli Range in northwestern Rajasthan, lie unconformably over the Malani Igneous Suite (MIS) and are assigned ages from Neoproterozoic to Early Cambrian (Pandit et al., 2001; Mazumdar and Bhattacharya, 2004; Kumar and Pandey, 2008). The Malani Igneous Suite unconformably overlies Paleo- to Mesoproterozoic metasediments, and basement granite gneisses and granodiorites (Pandit et al., 1999).

2.1. Other igneous events in Aravalli–Bundelkhand Cratons

Mafic intrusions thought to be associated with granitic intrusions date to 2828 ± 46 Ma. These are believed to be one of the earliest intrusive events into the 3.31 Ga tonalities in the Aravalli Craton (Gopalan et al., 1990).

There are other relatively minor dyke intrusions in the Aravalli Craton including the norite dykes of the Sandmata complex (~1720 Ma old; Sarkar et al., 1989); felsic dyke swarms of Sirohi and the albitite dykes in the NW part of the Aravalli Mountain (Ray, 1987, 1990; Fareeduddin and Bose, 1992). None of these minor dyke swarms have good age control, but all are believed to be Neoproterozoic in age on the basis of geological relationships.

The MIS to the west of Aravalli Mountains forms the largest felsic magmatic province of India (Fig. 2). It is characterized by voluminous magmatism that began with minor basaltic and predominant felsic volcanics followed by granitic emplacement while third and final phase of magmatic cycle is marked by predominately felsic and minor mafic dyke swarms. The MIS is often explained as 'anorogenic magmatism' related either to crustal melting during extension or to an active hot spot (Eby and Kochhar, 1990; Bhushan, 2000; Sharma, 2004). Alternatively MIS magmatism can be interpreted in the context of an Andean-type active margin (Torsvik et al., 2001a,b; Ashwal et al., 2002; Gregory et al., 2009). On the basis of geochemical similarities and anisotropy of magnetic susceptibility, Pandit and de Wall (in preparation) have described Mt. Abu batholith (previously considered as post-orogenic relative to the Delhi deformation) as the easternmost part of the Malani Igneous Province.

Previously reported ages (whole rock Rb–Sr) for the Malani magmatism range from 680 to 780 Ma spanning a period of

100 Ma. Dhar et al. (1996) and Rathore et al. (1999) reported whole rock Rb–Sr isochron ages for felsic volcanic rocks and granite plutons, emplaced during the first two stages of activity in the MIS, ranging from 779 ± 10 to 681 ± 20 Ma. Torsvik et al. (2001a) cited precise U–Pb ages of 771 ± 2 and 751 ± 3 Ma for rhyolite magmatism in the MIS although analytical data for those ages were never presented. Gregory et al. (2009) provide a robust U–Pb concordia age of 771 ± 5 Ma (MSWD = 1.5) for felsic volcanism in the Malani sequence. Additional constraints on igneous activity come from a variety of sources. van Lente et al. (2009) provide 767 ± 2.9 Ma, 765.9 ± 1.6 Ma and 761 ± 16 Ma U–Pb ages for the Sindreth felsic volcanics. The Sindreth felsic rocks in SW Rajasthan were thought to be the youngest part of Delhi Supergroup (Gupta et al., 1997; Chore and Mohanty, 1998). These new ages; however, seem to cement their relationship to the Malani magmatism. van Lente et al. (2009) also report ages of 800 ± 2 Ma and 873 ± 3 Ma for tonalitic basement rocks in the region. Pradhan et al. (2010) report an age of 827 ± 8.8 Ma for the Harsani granodiorite that also lies beneath the MIS. These data suggest that Malani magmatism was preceded by a protracted interval of granitic intrusion ranging from ~860 to 820 Ma (Erinpura Granite event; see also detrital zircon spectra in the overlying Marwar Supergroup by Malone et al. (2008) and Just et al. (in press)).

The Bundelkhand Craton in the Central Indian shield is also characterized by various Proterozoic extrusive and intrusive events. The NE–SW trending quartz reefs are the most spectacular feature in the Bundelkhand granitic massif (Basu, 1986). The majority of these quartz reefs are concentrated in the area bounded by Jhansi on the NW, Mahoba to the NE, Khajuraho on the SE and Tikamgarh to the SW (Fig. 3). These giant quartz reefs and veins along the brittle-ductile shear zones and fault planes mark extensive hydrothermal fluid activity following the crystallization of the granite plutons. The quartz reefs and associated hydrothermal activity are argued to have taken place in three phases based on the K–Ar geochronology: (1) 1480 ± 35 to 1660 ± 40 Ma, (2) 1790 ± 40 to 1850 ± 35 Ma and (3) 1930 ± 40 to 2010 ± 80 Ma (Pati et al., 1997). The broad age ranges reported here testify to the need for more robust dating of these intrusive events.

2.2. Sedimentary basins

The metasediments that make up the Aravalli and Delhi Supergroups represent the oldest sedimentary basins in the Aravalli–Bundelkhand Cratons. The younger intracratonic basins of Mesoproterozoic–Paleozoic age outcrop to the west of Aravalli Fold Belt and south and west of Bundelkhand Province: (1) The Marwar Supergroup in the western part of the Aravalli mountain range in Rajasthan and (2) the Vindhyan Basin located in central Peninsular India (Fig. 2).

2.3. Aravalli/Delhi sedimentary sequences

Both the Delhi and Aravalli meta-sedimentary sequences overlie the Banded Gneissic Complex and both show evidence of polyphase deformation and metamorphism. The Aravalli Supergroup has been subdivided into Lower, Middle and Upper Aravalli Groups (Roy et al., 2005) or Delwara (lower), Debari (middle) and Jharol (upper) Groups (Sinha Roy et al., 1998). The former two units represent shelf sedimentation while the latter one represents deep sea facies sedimentation. The Debari Group is older and is composed of a basal conglomerate overlain by metavolanics, quartzites, phyllites, carbonates. The younger Jharol Group is a turbidite sequence comprising phyllites, schists and a basal quartzite (Chakrabarti et al., 2004). A north–south trending ultramafic body divides the Aravalli Supergroup into eastern (shallow sea) and western (deep sea) segments.

The Delhi Supergroup has been subdivided into two main geographic terrains, referred to as the North Delhi Fold Belt and South Delhi Fold Belt (Sinha Roy et al., 1998). A three fold classification (Raialo, Alwar and Ajabgarh groups) has been retained for the northern sector while a number of Groups and smaller units, depending on location and author, have been proposed for the southern sector (Gupta et al., 1997). For simplicity, we use the Gupta et al. (1980, 1997) divisions. The lowermost Raialo Group is comprised of conglomerate, basic volcanics, marble and quartzite. Overlying the Railo Group is the Gogunda Group (=Alwar Group) and it is composed of quartzite, schists and metabasic rocks. The Gogunda Group is disconformably overlain by the Kumbhalgarh Group (=Ajabgarh Group). The main rock types of the Kumbhalgarh Group are carbonaceous shales, marbles, phyllites and relatively minor quartzites. In the sedimentary sequence, the Kumbhalgarh Group is overlain by the Sirohi and Punagarh (=Sindreth) Groups. The Sirohi Group is an association of marbles, pelitic schists and quartzites and the Punagarh Group is composed of arenaceous sediments, bimodal volcanics, shales, phyllites and schists.

2.4. Vindhyan basin

The Vindhyan Basin is one of several "Purana" (ancient) sedimentary basins of the Indian subcontinent. It is a sickle-shape basin that outcrops between the Archaean Aravalli–Bundelkhand province to the north and west and the Cretaceous Deccan Traps to the south; the Great Boundary Fault marks the western limit of this basin (Fig. 2). The Vindhyan Basin is composed of several smaller sub-basins, the largest of these are referred to as the Rajasthan sector and the Son Valley sector (Fig. 4).

The trondhjemitic gneisses of the Bundelkhand Igneous complex act as a basement ridge between the Rajasthan and the Son valley sectors (Prasad and Rao, 2006). In the western sector, the Vindhyan Basin is thought to have been deposited as an infill of the failed rifts on the Aravalli Craton (Mondal et al., 2002). Rifting thinned part of the crust along a series of east to west trending faults in a dextral transtensional setting (Bose et al., 2001). The observed volcaniclastic units, faults and paleoseismic sedimentary deformation in the lower part of the Vindhyan section support the rift origin of the basin but is also the source of some controversy (Bose et al., 2001). In a recent contribution, Raza et al. (2009) looked at the geochemistry of the basal volcanic sequence (Khairmalia and Jungel) and linked the formation of the Vindhyan Basin beginning at ~1800 Ma to collisional events in the Aravalli-Delhi Fold belt and the Central Indian Suture (CIS).

Stratigraphically, Vindhyan Basin can be divided into two sequences: The Lower Vindhyan Sequence formed by Semri Series (Satola, Sand, Lasrawan and Khorip Groups) and the Upper Vindhyan Sequence subdivided into the Kaimur, Rewa and Bhander Groups, respectively (Chaudhari et al., 1999; Prasad, 1984).

Age control on Vindhyan sedimentation is still the subject of considerable controversy as are the ages of the other Purana basins (Gregory et al., 2006; Patranabis-Deb et al., 2007; Azmi et al., 2008; Basu et al., 2008; Malone et al., 2008). In general, the age of sedimentation for the Lower Vindhyan is far better constrained than that of the Upper Vindhyan. The lower Vindhyan units are collectively designated as Semri Group. The Semri Group is made up of five alternating formations of shale and carbonates with areas of sandstones and volcaniclastic units. The Semri sediments unconformably overlie basement rock of either the 1854 ± 7 Ma Hindoli Group (Deb et al., 2002) or the 2492 ± 10 Ma Bundelkhand granites (Mondal et al., 2002) or terminal Archaean Berach Granite (~2.5 Ga). Ages from the Semri Series include a Pb–Pb isochron from the lower Kajrahat Limestone of 1721 ± 90 Ma (Sarangi et al., 2004); U–Pb zircon ages from the Porcellanites and Rampur

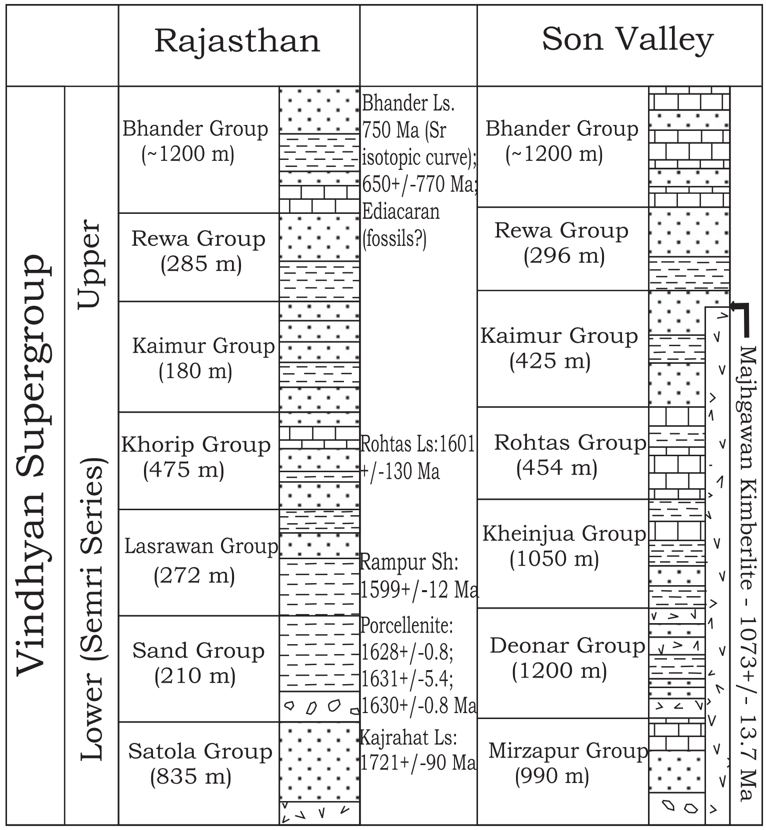

Fig. 4. Generalized stratigraphic columns of the Vindhyan Supergroup in the Rajasthan sector (left column) and the Son valley sector (right column) along with published age constraints (Malone et al., 2008). The stratigraphic representations here are designed to show correlations and age relationships rather than absolute thickness or stratigraphic continuity.

shale ranging from 1630.7 ± 4 to 1599 ± 8 Ma. The Rohtas limestone in the upper part of the Lower Vindhyan has Pb–Pb ages of 1599 and 1601 Ma (Rasmussen et al., 2002; Ray et al., 2003; Sarangi et al., 2004). Most authors (see Azmi et al., 2008 for an alternative view) agree with a Mesoproterozoic age for Lower Vindhyan sedimentation (~1750–1500 Ma).

The Semri Series is separated from the Upper Vindhyan by a basin wide unconformity between the Rohtas limestone and the overlaying Kaimur Group. The Kaimur rocks are intruded by the 1073 ± 13.7 Ma Majhgawan kimberlite (Gregory et al., 2006), that cross-cuts both the Semri Series and Kaimur Groups and is currently exposed in the Kaimur Group (Baghain sandstone). Up-section is the Rewa Group, a series of shale and sandstone formations that, in areas, contain kimberlite derived diamondiferous conglomerates (Rau and Soni, 2003). A thin shale unit marks the transition into the Bhander Group. The Bhander Group contains the only major carbonate unit in the upper Vindhyan system, a unit containing stromatolites, ooids, and micritic layers known as the Bhander or Lakheri limestone (Bose et al., 2001). The overlying lower Bhander sandstone marks a transition into shallower marine, sometimes fluvial, sandstone typical of the Bhander Group (Bose et al., 2001). The Sirbu shale overlies the lower Bhander sandstone, and is in turn overlain by the upper Bhander sandstone.

Age control on the Upper Vindhyan sequences is more problematic. The best age estimates come from the Majhgawan kimberlite, that intrudes the Lower Vindhyan and into the Baghain sandstone (Kaimur Group – Upper Vindhyan) near Panna. Possible Ediacaran fossils have been described in the Lakheri and Sirbu formations of the Bhander Group and could indicate an age <635 Ma for the Bhander (De, 2003, 2006). The Ediacaran disks reported by De (2003, 2006) have been challenged both in terms of their biologic nature (MacGabhann, 2007) as well as their age (Gibsher et al., submitted for publication). However, a Mesoproterozoic age for these discoidal organisms would lead to questions regarding the depth of metazoan evolution. In an attempt to further constrain the age of the uppermost Vindhyan, Malone et al. (2008) conducted a study of detrital zircon populations from the Bhander and Rewa Groups in the Rajasthan sector along with samples from the Lower Marwar Supergroup in Rajasthan. In that study, Malone et al. (2008) note that the youngest population of zircons from the Upper Bhander is older than 1000 Ma. That observation, coupled with the similarity in paleomagnetic directions from the Upper Vindhyan sedimentary sequence and Majhgawan kimberlite led Malone et al. (2008) to conclude that Upper Vindhyan sedimentation was completed by ~1000 Ma. This rather surprising conclusion is consistent with recent data from another of the Purana basins to the south (Patranabis-Deb et al., 2007).

2.5. Marwar basin

This Neoproterozoic to Cambrian age asymmetric intracratonic sedimentary basin lies unconformably over the Malani Igneous Suite (MIS) and the rocks of Delhi Supergroup. The Marwar Supergroup is found in western Rajasthan and the Marwar basin trends NNE–SSW with a slight westerly tilt. Lithostratigraphically, the Marwar Supergroup consists of Lower Jodhpur Group, Middle Bilara and Hanseran evaporite Group and the Upper Nagaur Group. The tectonic and thermal events in NW Rajasthan suggest that the Marwar Supergroup was developed by the reactivation of the NNE–SSW trending lineaments of Archaean and Proterozoic age.

The sedimentary sequences of the Marwar Supergroup interpreted as "Trans-Aravalli Vindhyans" (Heron, 1932) were correlated to the Upper Vindhyan Sequence and the Salt Range of Pakistan bracketing their age between Late Neoproterozoic – Cambrian (Khan, 1973; Pareek, 1981; Raghav et al., 2005). The Marwar Supergroup is a predominately deltaic to shallow marine facies sequence composed of evaporites, carbonates and sandstones. The total thickness of the Marwar Supergroup reaches a maximum of ~2 km (Pandit et al., 2001). At the base of the Marwar Supergroup is the Pokaran boulder bed containing cobbles of Malani and older igneous rocks (Chakrabarti et al., 2004; Ramakrishnan and Vaidyanadhan, 2008). The nature of the Pokaran boulder bed is the subject of some debate. Although the recent age estimates for the Lower Marwar place it in the proper time frame for either the Gaskiers (ca. 580 Ma) or Marinoan (ca. 635 Ma) glaciations, there is no evidence for a glacial origin (e.g. dropstones, striated clasts).

The middle section of the Marwar Basin contains evaporite (observed in boreholes only) and carbonate facies that are capped by sandstone. The age of the Marwar is considered to be of Ediacaran–Cambrian age (~635–515 Ma; Naqvi and Rogers, 1987; Pandit et al., 2001) but there are no robust radiometric data. Recent discoveries of medusoid fossils and traces in the lower section (Raghav et al., 2005) along with trilobite traces in the uppermost section (Kumar and Pandey, 2008) are consistent with an Ediacaran–Cambrian age for the Marwar Supergroup.

Recently, the $^{238}U/^{206}Pb$ isotopic analyses on the detrital zircon grains extracted from the Sonia and Girbakhar sandstones (Jodhpur Group) of the Marwar Supergroup by Malone et al. (2008) yielded a major age peak for the Marwar Supergroup centered between 800 and 900 Ma. These ages have been interpreted to reflect sedimentary input from the igneous rocks in the South China Craton, juvenile crust formed in the Arabian–Nubian Shield or igneous rocks emplaced along the western margin of the Delhi – Aravalli Fold Belt (Deb et al., 2001; Jiang et al., 2003; Xiao et al., 2007; van Lente et al., 2009; Pradhan et al., 2010).

3. Singhbhum Craton

The Singhbhum Craton (also called the Singhbhum-Orissa Craton; Fig. 5) lies along the eastern coast of India and borders the Mahanadi graben to the west, the Narmada-Son lineament, and the Indo-Gangetic plain. It is bordered to the north by the Tamar-Poropahar Shear Zone and the Chhotanagpur Granite–Gneiss Terrain (CGGT). The CGGT is thought to be an extension of the Central Indian Tectonic Zone (CITZ). The craton itself can be further subdivided into several different assemblages including the Older Metamorphic Group, the Singhbhum granite and the Iron Ore Group (IOG).

3.1. Chhotanagpur Terrain

The Proterozoic Chhotanagpur Granite–Gneiss Terrain (CGGT) occurs in the northern part of the Singhbhum mobile belt and consists of granitic gneisses, quartzo-feldspathoids, and intermittent mafic intrusives, all of which display varying degrees of metamorphism and tectonic deformation (Mahmoud et al., 2008). Age constraints for the 200 km wide by 500 km long terrain range from 1500 to 800 Ma on the basis of K–Ar dating, but these ages most certainly reflect disturbance of the K–Ar system (Naqvi and Rogers, 1987).

The CGGT hosts a wide variety of magmatic rocks, both mantle and crustal derived, that took part in the 2.1 billion year long cratonization process. The magmatism of the CGGT was largely controlled by intra-continental rift zones leading to extensional tectonic activity and magmatic intrusions (Ghose and Chatterjee, 2008).

The features displaying high-grade metamorphism characteristically contain intrusive pegmatites that display both concordant and discordant foliation. The pegmatites are thought to have intruded over a long time span and display indications of the different deformation episodes thought to have occurred in the area (Mahmoud et al., 2008). Two of the most prominent deformation features are the large geo-anticline of metamorphic schists and the shear zone that deforms the geo-anticline along its overfolded southern limb (Dunn and Dey, 1942).

3.2. Older Metamorphic Group–Singhbhum Nucleus

The Singhbhum nucleus is composed mainly of Archaean granitoid batholiths, including the Singhbhum Granite Complex. Enclaves within the batholiths include supracrustal rock complexes, the oldest of these is known as the Older Metamorphic Group (OMG). The OMG consists mainly of remnants in the form of micaceous schists, quartzites, calc-silicates, and para- and ortho-amphibolites (Naqvi and Rogers, 1987). Tonalite–Trondhjemite Gneisses (TTGs) are found along the contacts with some of the amphibolites (Fig. 5). Unfortunately, reliable age dates within the Singhbhum Craton are sparse. U–Pb zircon dating in the OMG supracrustal suite yields ages of 3.5, 3.4, and 3.2 Ga (Mondal et al., 2007). Gneisses in the OMG enclaves yield ages of 3.8 Ga with Sm–Nd dating and 3.2 Ga with Rb–Sr techniques (Naqvi and Rogers, 1987; Saha, 1994). The most comprehensive attempt to date the OMG is described by Misra et al. (1999). Detrital zircons from the OMG yielded age ranges between 3.5 and 3.6 Ga and other zircons have peaks at 3.4 and 3.2 Ga. Misra et al. (1999) conclude that the younger ages represent two distinct metamorphic events in the OMG. Basu et al. (1996) describe a Pb-loss event at 3352 ± 26 Ma in the OMG most likely related to the intrusion of the Singhbhum Granite (see below). The scarcity and nature of the OMG rocks make their relationship to surrounding rocks difficult to determine and numerous widely varied hypotheses exist that we consider too poorly constrained to merit further discussion.

3.3. Singhbhum Granite

The OMG is intruded by the approximately 10,000 km^2 Singhbhum Granite complex (Fig. 5). The complex includes 12 domal magmatic bodies that are independent of one another. Whether these bodies were emplaced in a single magmatic event or several remains a subject of debate; however older data suggest polyphase emplacement (Naqvi and Rogers, 1987).

The Singhbhum Granite (SG) complex includes two different types of granite. One set of granites displays HREE depletion and is dated at 3300 ± 7 Ma using U–Pb dating on zircons (Mondal et al., 2007). Other varieties of granites produce a fractionated LREE pattern and flat HREE and are dated at ~3.1 Ga using whole rock Pb–Pb techniques (Mondal et al., 2007). Misra et al. (1999) report an age of 3328 ± 7 Ma for the Singhbhum 'phase II" granites and ages of 3080 ± 8 Ma and 3092 ± 5 Ma for the Mayurbhanj granite. Reddy et al. (2008) report SHRIMP U–Pb and Pb–Pb ages from the Singhbhum Granite with a discordant upper intercept age of 3302 ± 14 Ma and a more robust $^{207}Pb–^{206}Pb$ age for the most concordant zircons of 3288^{+8}_{-2} Ma. The Sushina nepheline syenite body yielded the youngest (SHRIMP) ages from the SG complex of 922.4 ± 10.4 Ma (Reddy et al., 2008). Acharyya et al. (2008) reported U–Pb ages from an 'earliest' phase of Singhbhum Granite intrusion. Their two concordant ages of 3527 ± 17 Ma and 3448 ± 19 Ma would represent the earliest phases of granitic intrusion coeval with the formation of the Iron Ore Group (described below). These ages are consistent with the early Pb–Pb age for the early phase reported by Moorbath et al. (1986) of 3250 ± 5 Ma. The geochronological data from the Singhbhum Granites therefore favours at least a 4-stage emplacement. The oldest intrusions of granites at ~3.45–3.5 Ga were followed by a secondary emplace-

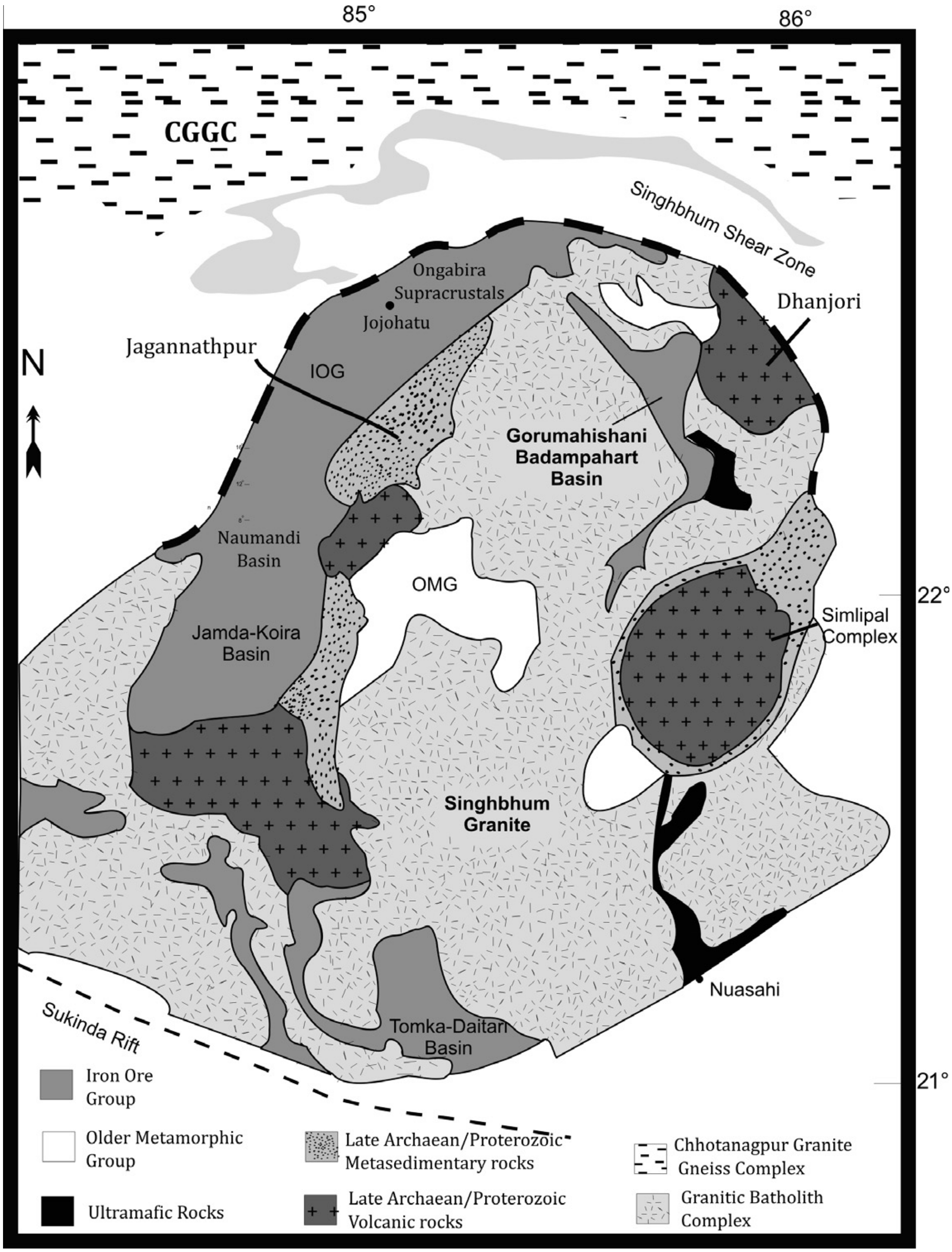

Fig. 5. Sketch map of the Singhbhum Craton (NE India) showing major elements comprising the craton and the surrounding region. IOG = Iron Ore Group; OMG = Older Metamorphic Group (after Naqvi and Rogers, 1987; Mondal et al., 2007; Ramakrishnan and Vaidyanadhan, 2008).

ment at 3.3 Ga, tertiary intrusions at 3.1 Ga and the youngest intrusive event at ~0.9 Ga. In addition, Misra (2006) discusses emplacement of Soda Granites within the existing shear zones of the Singhbhum Craton at 2.22 Ga, with subsequent shearing at 1.67–1.63 Ga.

3.4. Iron Ore Group

Cratonization of the Singhbhum Craton includes the formation of a greenstone–gneiss terrane known as the Iron Ore Group (IOG) (Eriksson et al., 2006; Mondal et al., 2007). The entire IOG occurs as a supracrustal suite composed of three fold belts: the Jamda-Koira, the Gorumahishani–Badampahar, and the Tomka-Daitari (Mondal et al., 2007). It is divided into an Older and a Younger section, with similar compositions but differing ages.

The Older IOG is comprised of clastic sedimentary rocks formed in a shallow marine setting along with syndepositional volcanic rocks that together suggest large scale rifting (Eriksson et al., 2006). The Older IOG formed prior to the intrusion of the Singhbhum Granite and was thought to have an age range between 3.3

and 3.1 Ga, based solely on associations to nearby rocks and available ages for related rocks (Eriksson et al., 2006; Mondal et al., 2007). A recent geochronologic study of the IOG (Mukhopadhyay et al., 2008) yielded an age of 3506.8 ± 2.3 Ma for a dacitic lava and confirms that the IOG formed just prior to, or slightly before the earliest Singhbhum Granites. Detrital zircons found in the TTG suite (see above) may have been derived from the IOG.

The Younger Iron Ore Group formed after the Singhbhum Granite cratonization event and has a suggested depositional age >2.55 and <3.0 Ga. It is comprised of shallow or shelfal marine greenstone deposits with banded iron formation (Eriksson et al., 2006).

3.5. Simlipal Basin

The relatively undeformed Simlipal Basin is a volcano-sedimentary basin composed mainly of tuffs, lavas, quartzites, arkoses and shales and is intruded by the fractionally crystallized Amjori Sill and various smaller intrusive bodies (Fig. 5). The volcanic complex displays evidence of caldera collapse features as units show dips toward the center of the basin. Early attempts at dating the complex yielded Rb–Sr ages of 2085 Ma (Naqvi and Rogers, 1987). This age appears to be too young as the Simlipal basinal rocks (including the Simlipal Complex) are intruded by the ~3.1 Ga Mayurbhanj granite.

3.6. Newer dolerites

Dyke swarms found within the Singhbhum granitic pluton vary in rock type from mafic to intermediate. The largest suite of dykes is the so-called "Newer Dolerites" (Bose, 2008). Nearly every group of rocks in the Singhbhum nucleus are intruded by the widespread Newer Dolerites. The Newer Dolerites are subdivided into at least two distinct generations, related by cross-cutting relationships and distinct geochemical signatures. Emplacement ages are poorly constrained, ranging from 1600 to 950 Ma, based mainly on K–Ar dating (Naqvi and Rogers, 1987; Srivastava et al., 2000; Bose, 2008). Bose (2008; and sources therein) suggests three distinct pulses of magmatism, based mainly on available K–Ar data, at 2100 ± 100 Ma, 1500 ± 100 Ma, and 1100 ± 200 Ma. These ages should be viewed with caution until more robust U–Pb ages become available.

The dykes vary from a few meters to 700 m in thickness and can extend for several kilometers. They predominantly strike NNE–SSW or NNW–SSE. The dolerites exhibit a variety of textures including fine, medium and coarse grained specimens, though the majority is of medium grain size. The main minerals seen in the slightly metamorphosed dolerites are augite and labradorite (Verma and Prasad, 1974).

3.7. Proterozoic sedimentation

3.7.1. Paleo-Mesoproterozoic history

Proterozoic sedimentation and volcanism found in the Singhbhum Craton is divided into several formations (Fig. 6) including (from oldest to youngest) the Dhanjori Basin the Singhbhum Group (composed of the older Chaibasa and younger Dhalbhum Formations), the Dalma Formation, and the Chandil Formation (Mazumder, 2005; Eriksson et al., 2006). The oldest two formations (Dhanjori and Chaibasa) represent periods of transgression and regression onto the continent, while the younger supracrustal units (Dhalbhum, Dalma, and Chandil Formations) reflect a possible mantle plume event at ca. 1.6 Ga (Eriksson et al., 2006). Mazumder (2005) suggested that the cooling of the massive Singhbhum Granite Complex induced isostatic readjustment within the craton and led to a new tectonic regime of tensional stresses and deep-seated fractures that in turn influenced the deposition of the Proterozoic formations.

The Dhanjori Basin formed unconformably over the Archaean basement and is the oldest sedimentary basin in the Singhbhum Craton. It is a terrestrial, primarily fluvial sequence composed of clastic sedimentary rocks overlain by mafic–ultramafic volcanic and volcaniclastic rocks (Mazumder, 2005; Eriksson et al., 2006; Bhattacharya and Mahapatra, 2007). Both massive and schistose amygdaloidal basalts are present in the basin. These low-Al tholeiites have been folded into a first order syncline (Naqvi and Rogers, 1987). Based on field and preliminary geochronologic results, Acharyya et al. (2008) propose a Late-Archaean to Early Paleoproterozoic age for the Dhanjori Basin (~2.5 Ga). In contrast, Mishra and Johnson (2005) report a much older 2858 ± 17 Ma Pb–Pb age for the volcanic rocks of the Dhanjori Formation making it considerably older than the estimates of Acharyya et al. (2008).

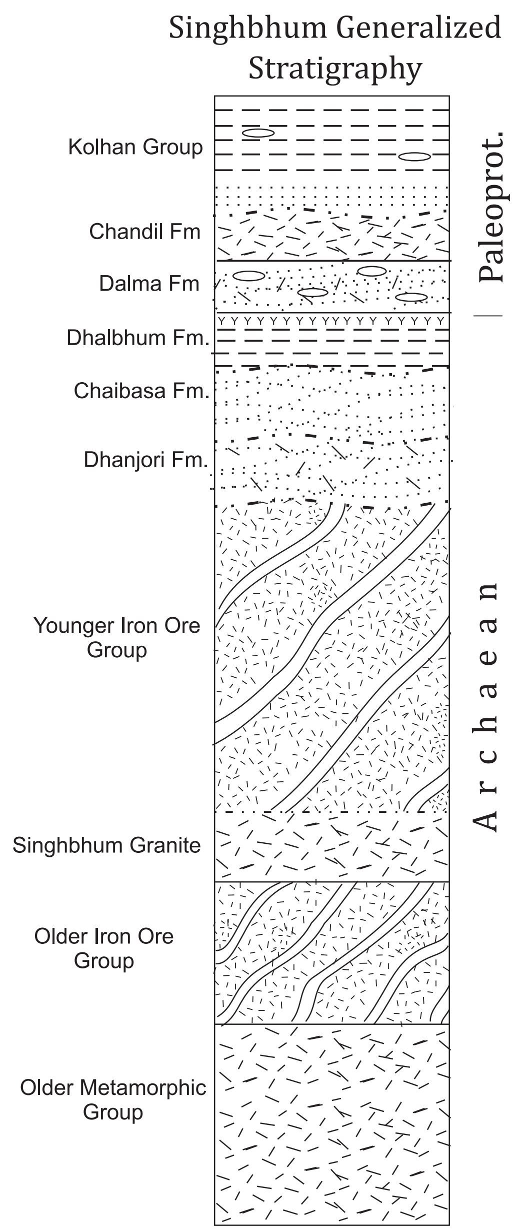

Fig. 6. Generalized stratigraphic column for the Singhbhum Craton after Ramakrishnan and Vaidyanadhan (2008). The stratigraphic representations here are designed to show correlations and relative age relationships rather than absolute thickness or stratigraphic continuity.

Stratigraphically above (?) the Dhanjori Formation is the Chaibasa Formation, the lower sequence of the Singhbhum Group, marked by a transgressive lag representing a major transgression on the Singhbhum Craton (Mazumder, 2005). The formation is comprised of tidal sandstones, a heterolithic facies generated by deposition during stormy and fair weather, and an offshore shale facies (Eriksson et al., 2006).

Unconformably overlying the Chaibasa Formation is the Dhalbhum Formation. It is composed of a layer of phyllites, shales, and quartzites overlain by volcanic tuff. Ultramafic–mafic intrusions and thin basaltic to komatiitic lava flows occur through the section (Eriksson et al., 2006). The depositional environment for the formation is thought to be largely fluvial/aeolian and thus representative of a dominantly terrestrial environment with a sequence boundary at the Chaibasa–Dhalbhum contact (Mazumder, 2005).

The Dalma Formation conformably overlies the Dhalbhum Formation and consists of ultramafic–mafic volcanic rocks. This thick sequence contains numerous lenses of high Mg agglomerates (Eriksson et al., 2006). A period of concurrent sedimentation and volcanism is inferred for the Dalma Formation (Mazumder, 2005). The formation shows evidence of a major folding event and displays widespread metamorphism (Dunn and Dey, 1942). The depositional history of the Dalma Formation is widely debated with hypotheses ranging from island arc to back-arc basinal settings or alternatively from continental volcanism to intra-continental rifting. A plume related origin also attracts many supporters, although no single theory has become generally accepted (Eriksson et al., 2006 and references therein).

The Chandil Formation exists as a belt of meta-sedimentary rock and volcanic rocks that separate the Dalma sequence to the south from the Chhotanagpur Granite Gneiss Complex to the north. Original depositional settings are interpreted as a combination of fluvial–aeolian and shallow marine sedimentation (Eriksson et al., 2006). The Chandil tuffs are dated to 1484 ± 44 Ma (Rb–Sr; Sengupta et al., 2000); however, this age is challenged and generally regarded as a metamorphic resetting rather than an eruptive age (Mazumder, 2005; Eriksson et al., 2006). Whole rock Rb–Sr dating on a cluster of granites intruding the Chandil yielded an age of 1638 ± 38 Ma are thought to provide a poorly constrained younger limit for the Chandil Formation (Mazumder, 2005).

3.8. The Kolhan Group

In the southern part of the Singhbhum Craton, there is a minor supracrustal suite known as the Kolhan Group. The age of the Kolhan Group is unconstrained, but Mukhopadhyay et al. (2006) argued that sedimentation likely began at about 1.1 Ga. The Kolhan Group formed in an intracratonic basin with a westward slope and was subsequently deformed into a synclinal structure. Elongate domes and basins and dome-in-dome structures dominate the eastern part of the basin (Naqvi and Rogers, 1987). The Kolhan Group is subdivided into three different formations; the Mungra sandstone (25 m thick), the Jinkphani limestone (80 m thick) and the Jetia shale (1000 m). As a whole, the Kolhan Group is a transgressive feature that is interpreted as having formed in a rift setting that is perhaps related to the fragmentation of the Rodinia supercontinent (Bandopadhyay and Sengupta, 2004; Mukhopadhyay et al., 2006).

4. Bastar Craton

The Bastar Craton (Fig. 7; also known as the Bhandara or Central Indian Craton) is bordered by the Pranhita–Godavari rift (to the south), the Mahandi Rift (in the northeast), the Satpura Mobile Belt (in the north), the Eastern Ghats Mobile Belt (to the east) and Deccan Traps cover (to the west). The craton consists mainly of granites and granitic gneisses and contains three major features: (1) a trio of supracrustal sequences, (2) mafic dyke swarms, and (3) the Satpura orogenic belt (Naqvi and Rogers, 1987; Srivastava et al., 2004).

The "Gneissic Complex" is dominated by TTG assemblages dated to between 2500 and 2600 Ma that are interpreted to reflect a major interval of crustal accretion (Ramakrishnan and Vaidyanadhan, 2008). A tonalite sample yielded a U–Pb upper intercept age of 3561 ± 11 Ma (Ghosh, 2004) thought to reflect the oldest age of the gneissic protolith. Sarkar et al. (1993) report an age of 3509 +14/−7 Ma for another gneissic protolith within the complex.

4.1. Supracrustal sequences

The Bastar Craton contains at least three major supracrustal sequences of rocks (Fig. 8), the Dongargarh, the Sakoli, and the Sausar suites. Of these three suites, only the Dongargarh Supergroup has been dated. Further subdivisions and supracrustal sequences are described in the Bastar Craton, but a lack of geochronological data makes it extremely difficult to discern if they are truly distinct units or merely poorly correlated across the craton. Thus we adopt a more simplistic subdivision for the purposes of this review.

4.2. Dongargarh supergroup

The Dongargarh Supergroup extends from the Chhattisgarh Basin in the east to the Sakoli in the west and is composed of three smaller groups of rocks, the Amagaon, Nandgaon, and Khairagarh groups. The Amagaon granites and gneisses are presumed to have formed during the Amagaon Orogeny at ca. 2.3 Ga and are sometimes excluded from the Dongargarh Supergroup. The group consists mainly of gneisses with secondary schists and quartzites (Naqvi and Rogers, 1987).

The Nandgaon Group contains two volcanic suites (the Bijli and Pitepani suites) that are dominated by rhyolites with secondary dacites, andesites, and basalts that show signs of fractionation (Neogi et al., 1996). The Bijli rhyolite was dated using Rb–Sr techniques at 2180 ± 25 Ma and 2503 ± 35 Ma (Sarkar et al., 1981; Krishnamurthy et al., 1988) and has localized inclusions of Amagaon granite (Naqvi and Rogers, 1987). The Dongargarh volcanic rocks have Rb–Sr ages of 2465 ± 22 Ma and 2270 ± 90 Ma (Sarkar et al., 1981; Krishnamurthy et al., 1988). Chakraborty and Sensarma (2008) largely dismiss the inconsistency within the Rb–Sr data and argue, on the basis of correlation with well-dated units in the Singhbhum Craton, that the Nandgoan Group was developed ~2.5 Ga. We view all these age estimates as tentative until more precise U–Pb ages are acquired.

The Khairagarh Group unconformably overlies the Nandgaon and consists of shales, sandstones, and igneous rocks. The basal formations are broken into the Basal Shale, the Bortalao formation and an inter-trappean shale, all conformably overlying each other (Naqvi and Rogers, 1987). It also contains two overlying volcanic suites, the Sitagota and Mangikhuta suites that are dominated by more primitive, unfractionated tholeiitic basalts (Neogi et al., 1996). The volcanic suites are divided by the Karutola sandstone (Naqvi and Rogers, 1987). The four volcanic suites within the Dongargarh Group erupted periodically between ca. 2462 and 1367 Ma (Neogi et al., 1996), but these ages are poorly constrained.

4.3. Sakoli Group

The Sakoli Group consists mainly of low grade metamorphic rocks of undetermined age in a large synclinorium. The Sakoli Group is a significant volcano-sedimentary deposit comprised of (youngest to oldest) slates and phyllites, bimodal volcanic suite

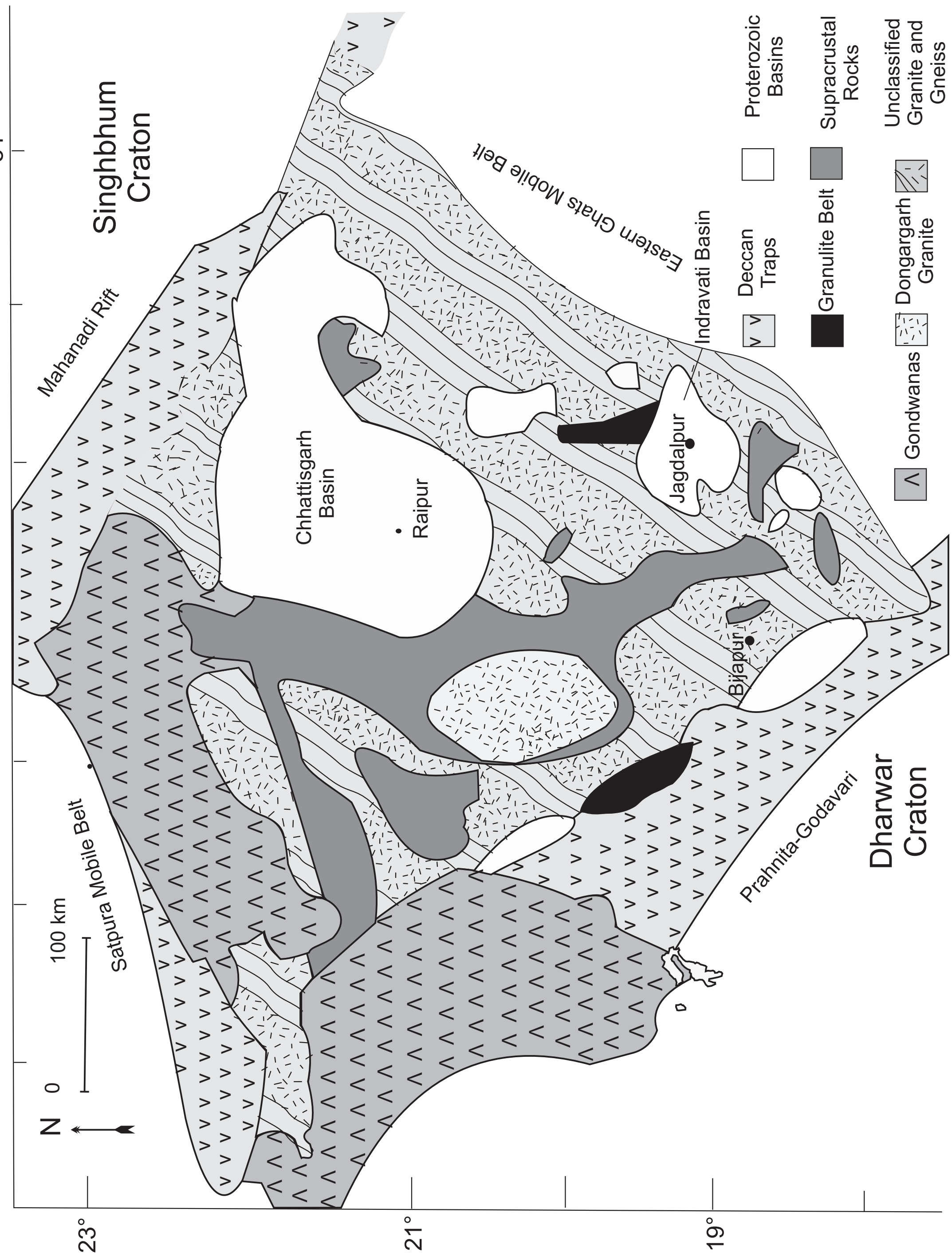

Fig. 7. Sketch map of the Bastar Craton after Naqvi and Rogers (1987) and Ramakrishnan and Vaidyanadhan (2008).

and schists, metabasalts and cherts and conglomerates and Banded Iron Formations (BIF's; Bandyopadhyay et al., 1990). Two stages of deformation are thought to have occurred, creating a sequence of overfolded bedding and a period of progressive metamorphism followed by retrogression (Naqvi and Rogers, 1987). Unconformably overlying the Sakoli Group are the sediments of Permo-Triassic Gondwana Supergroup and the Late Cretaceous Deccan basalts.

The age of the Sakoli Group is not known. Rb–Sr ages on metavolcanics and tuffs yield ages of 1295 ± 40 and 922 ± 33 Ma but the significance of these ages is difficult to interpret (Bandyopadhyay et al., 1990).

4.4. Sausar Group

The Sausar Group of metamorphosed sediments and manganese-bearing ores were once thought to be the oldest formations in Central India. The Sausar polymetamorphic belt that contains the sediments is part of the larger Central Indian Tectonic Zone (CITZ) and is approximately 300 km in length and 70 km in width (Naqvi and Rogers, 1987; Figs. 1 and 8). Detailed geochronologic studies are lacking within this belt; however, Roy et al. (2006) on the basis of Rb–Sr and Sm–Nd geochronological studies, argue that the main phase of metamorphism (amphibolite-grade) took place

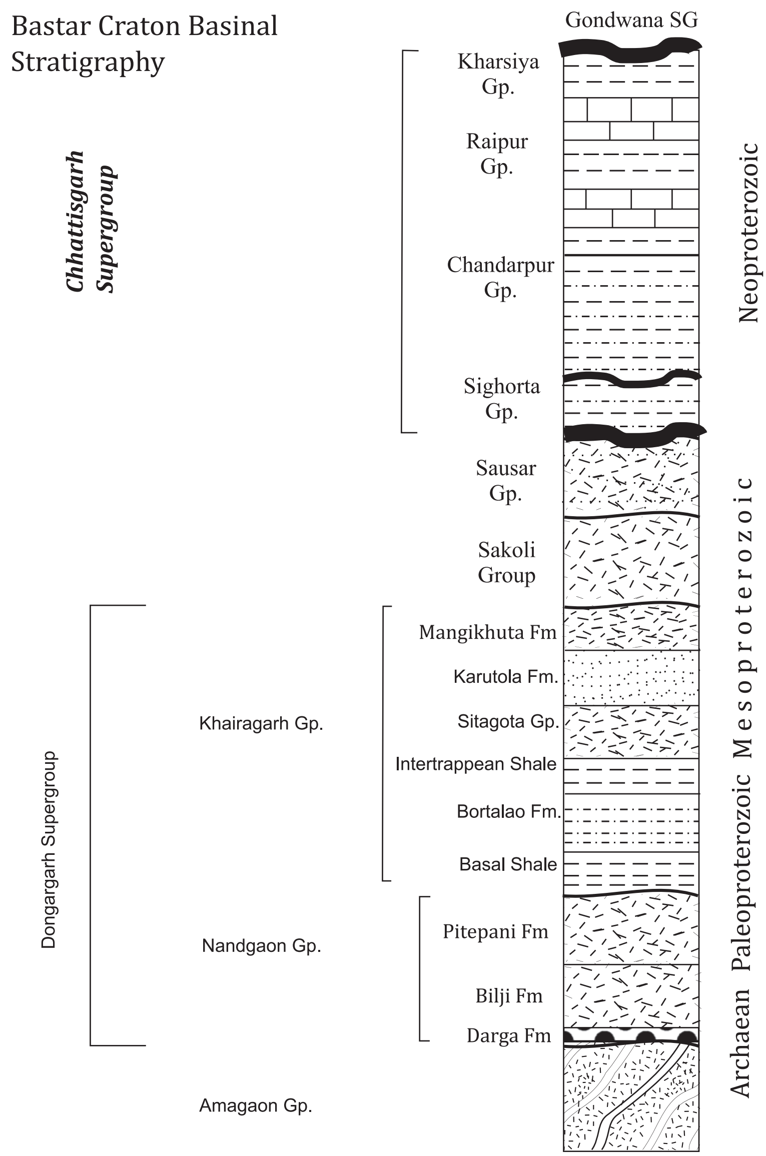

Fig. 8. Generalized stratigraphic column for the Bastar Craton after Ramakrishnan and Vaidyanadhan (2008). The stratigraphic representations here are designed to show correlations and age relationships rather than absolute thickness or stratigraphic continuity.

between 800 and 900 Ma. They also noted that the Sausar Belt was bounded on the north and south by granulite belts of different ages. The southern granulite belt hosts a charnockite that yielded a Sm–Nd isochron age of 2672 ± 54 Ma. A mafic granulite within the southern belt yielded Sm–Nd age of 1403 ± 99 Ma. The northern granulite yielded Sm–Nd age of 1112 ± 77 Ma. The granulites in the north and south also yield Rb–Sr isochron ages in the range of 800–900 Ma.

Roy et al. (2006) developed a tectonic model for the region whereby the Dharwar and Bastar Cratons were juxtaposed with the Bundelkhand Craton during the younger Sausar orogenic cycle as part of the larger assembly of Rodinia.

In contrast, Stein et al. (2004) argue that the juxtaposition between the northern and southern Indian cratonic nuclei along the Central Indian Tectonic Zone (CITZ) took place during the earliest Paleoproterozoic based on Re–Os ages from within the Sausar Belt (Malanjkhand granitoid batholith). They report a Re–Os age of 2490 ± 2 Ma for the granitoid that is nearly identical to U–Pb zircon ages of 2478 ± 9 Ma and 2477 ± 10 Ma for the same unit (Panigrahi et al., 2002). Cu–Mo–Ag mineralization ages associated with the intrusions ranged from 2446 to 2475 Ma (Stein et al., 2004). Stein et al. (2004) note that the region underwent significant ∼1100–1000 Ma reworking, but the main assembly of cratons occurred during the latest Archaean to earliest Paleoproterozoic (∼2.5 Ga) along the Sausar Belt (e.g. CITZ, see also Fig. 1).

4.5. Mafic Dyke Swarms

The Bastar Craton is intruded by numerous mafic dyke swarms, spanning an area of at least 17,000 km^2, that cross cut the various granitoids and supracrustal rocks of the region (French et al., 2008). The swarms are given regional names, but many may belong to the same intrusive episode. These include the Gidam-Tongpal swarm, the Bhanupratappur-Keskal swarm, the Narainpur–Kondagaon swarm and the Bijapur-Sukma swarm (Ramachandra et al., 1995). A majority of the dykes in the southern Bastar Craton trend NW–SE, paralleling the Godavari rift and these dykes are thought to have exploited preexisting faults. The northern dykes are oblique to the Mahanadi rift in a NNW–SSE direction (French et al., 2008).

Geochronologic constraints on many of the swarms are poor although recent work suggests a major episode of igneous activity and dyke intrusion around 1.9 Ga (French et al., 2008). The Paleoproterozoic dyke swarms are dated using U–Pb baddeleyite/zircon techniques at 1891.1 ± 0.1 Ma and 1883 ± 1.4 Ma and include boninite–norite and sub-alkaline mafic dykes, most of which display some degree of metamorphism (Srivastava et al., 2004; French et al., 2008; Srivastava and Gautam, 2008). Srivastava et al. (2004) and French et al. (2008) interpret the Precambrian dyke swarms as remnants of a large igneous province. French et al. (2008) noted that this activity is coeval with mafic magmatism in both the Superior Craton of North America and along the northern margin of the Kaapvaal Craton although they did not link the regions together paleogeographically and instead argued for a mantle upwelling on a global scale. In contrast, Srivastava and Singh (2003) linked the dykes to Laurentia and Antarctica in a "Columbia-type" paleogeography.

The younger dyke swarms represent the youngest igneous events in the Bastar Craton and mainly include metagabbros and metadolerites (Subba Rao et al., 2008). Hussain et al. (2008) postulated that these dykes were derived from subduction constituents that were altered in the mantle lithosphere. A subduction related genesis is consistent with the increased incompatible lithophile element concentration, seen in the geochemical analysis (Subba Rao et al., 2008), but this does not preclude different genetic models for the younger dykes. Age control on the dykes is lacking though they are younger than 1.9 Ga based on cross-cutting relationships with the older dykes.

4.6. Sedimentary basins

The Bastar Craton contains two major Proterozoic basins, the Chhattisgarh Basin, the Indravati Basin and six minor basins.

4.7. Chhattisgarh Basin

The 36,000 km^2 Chhattisgarh Basin is comprised of a ~1500 m thick sedimentary sequence (the Chhattisgarh Supergroup) of conglomerates, orthoquartzites, sandstones, shales, limestones, cherts, and dolomites (Fig. 7; Naqvi and Rogers, 1987). The sedimentary sequence has been divided into a basal Chandarpur Series and an upper Raipur Series (Fig. 8; Naqvi and Rogers, 1987; Patranabis-Deb et al., 2007). The Chandarpur Series consists of a shale-dominated sequence containing conglomerate and coarse arkosic sandstone formed as coalescing fan–fan delta deposits, storm-dominated shelf deposits, and high-energy shoreface deposits. The Raipur Series, however, underwent outer shelf, slope and basin deposition and consists of a limestone-shale-dominated sequence (Chaudhuri et al., 2002).

In the eastern part of the basin lies the "Purana" succession. The Purana contains a proximal conglomerate-shale-sandstone assemblage and a distal limestone-shale assemblage. The conglomerate-shale-sandstone assemblage unconformably overlies the basement and is thought to correspond to the Chandarpur Series. The limestone-shale assemblage, on the other hand, is thought to correspond to the Raipur Series (Deb, 2004).

Recently, the timing of deposition of the Chhattisgarh Supergroup is the topic of debate. A current 'consensus' places the dates of deposition in the Chhattisgarh between the Neoproterozoic to as young as 500 Ma (Naqvi, 2005). However, rhyolitic tuffs near the top of the Chhattisgarh sequence (the Sukhda and Sapos Tuffs) yielded ages of 1011 ± 19 Ma and 990 ± 23 Ma (Sukhda tuff) and 1020 ± 15 Ma (Sapos tuff) using U–Pb SHRIMP techniques on magmatic zircons (Patranabis-Deb et al., 2007). This led the authors of that paper to conclude that the Purana basins may up to 500 Ma older than the 'consensus' agreement. This conclusion is also supported by recent geochronologic studies from the lower part of the Chhattisgarh Basin by Das et al. (2009). In particular, zircon ages from the Khariar tuff show a concentration of ages around 1455 Ma.

4.8. Indravati Basin

The 9000 km^2 Indravati Basin (Fig. 7) consists of unmetamorphosed, unfossiliferous, largely undeformed shales, dolomites, sandstones, quartz arenites, limestones, and conglomerates. The sediments are thought to have a shallow marine or lagoonal depositional environment (Maheshwari et al., 2005). The basin is lithologically similar to the Chhattisgarh and it is postulated that at one point the two were connected and later eroded into discrete basins (Naqvi and Rogers, 1987). The sandstone member is correlated with the sandstone of the Chopardih Formation of the Chhattisgarh Basin, that has been dated using K–Ar methods at 700–750 Ma. Given the recent dating of the tuffaceous layers in the Chhattisgarh Basin, these K–Ar ages should be viewed with skepticism.

5. Eastern dharwar craton

The Dharwar Craton is split into Eastern and Western Cratons, with major differences in lithology and ages of rock units. The western boundary of the Eastern Dharwar Craton (EDC) is poorly defined and is constrained to a 200 km wide lithologic transitional zone from the Peninsular Gneisses of the Western Dharwar Craton to the Closepet Granite (Fig. 9). The Closepet Granite is a good approximation of the western boundary and is used as such in this paper (Ramakrishnan and Vaidyanadhan, 2008). The EDC is bounded to the north by the Deccan Traps and the Bastar Craton, to the east by the Eastern Ghats Mobile Belt, and to the south by the Southern Granulite Terrane (Balakrishnan et al., 1999). The craton is composed of the Dharwar Batholith (dominantly granitic), greenstone belts, intrusive volcanics, and middle Proterozoic to more recent sedimentary basins (Fig. 9; Naqvi and Rogers, 1987; Ramakrishnan and Vaidyanadhan, 2008).

5.1. Greenstone Belts

Greenstone and schist belts of the EDC are concentrated in the western half of the craton and are stretched into linear arrays. The belts continue to the east where they are covered by the Proterozoic Cuddapah Basin. The general trend of the belts is N–S and related belts are classified into supergroups (Ramakrishnan and Vaidyanadhan, 2008). Metamorphism of the belts is generally limited to greenschist to amphibolite facies with lower grades occurring in the larger belts and in the interior of smaller ones (Chadwick et al., 2000). Balakrishnan (1990) used whole rock Pb/Pb dating to constrain the age of the Kolar Schist Belt between 2900 and 2600 Ma. Nutman et al. (1996) and Nutman and Ehlers (1998) used SHRIMP U–Pb zircon methods to obtain ages of ~2725–2550 Ma in the Kolar Belt. Age trends in the belts generally infer a younging trend from west to east. These schist belts are all, to some degree, intruded by syn- and post-tectonic felsic rocks (Chadwick et al., 2000). Some of the more important greenstone-schist belts and supergroups that will be discussed briefly below include: the Sandur schist belt, the Ramagiri–(Penakacherla–Sirigeri)–Hungund superbelt (RPSH), the Kolar–Kadiri–Jonnagiri–Hutti superbelt (KKJH), and the Veligallu–Raichur–Gadwal superbelt (VRG).

The Sandur schist belt is characterized by dominant green schist facies metamorphism with higher, amphibolite grade rocks occur-

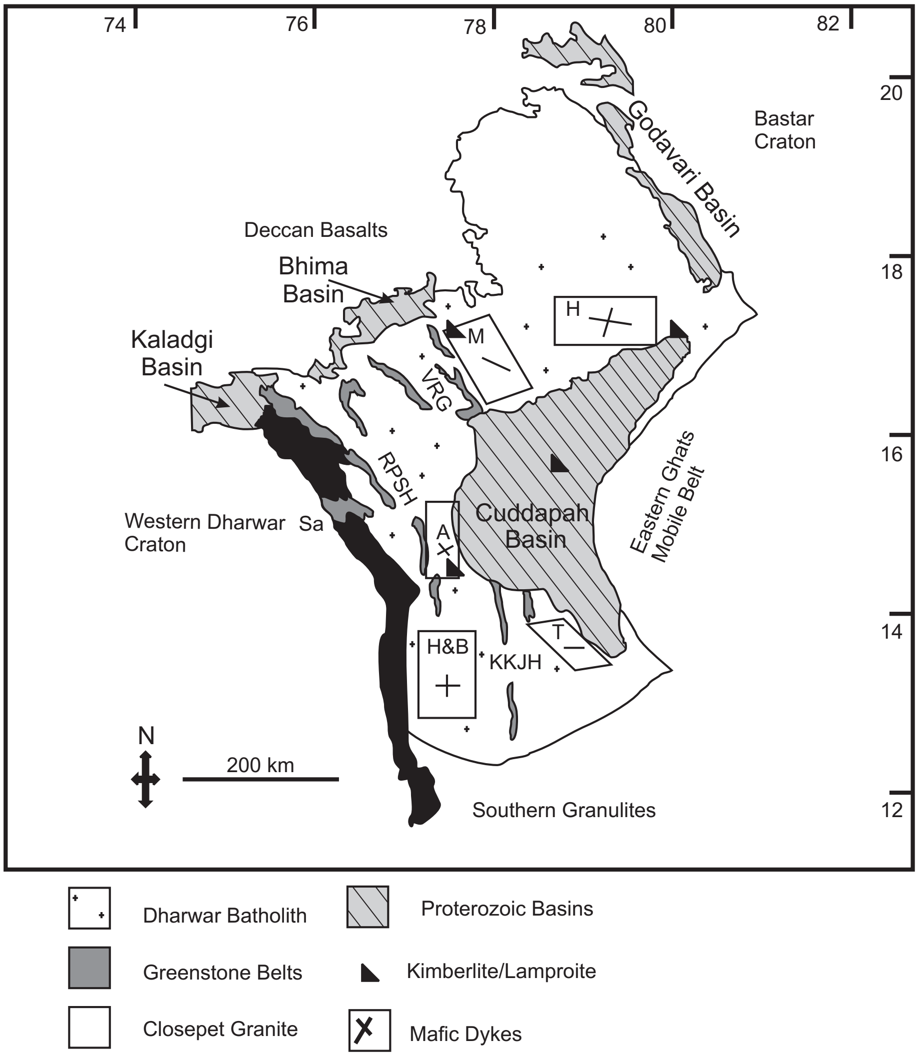

Fig. 9. Sketch map of the Eastern Dharwar Craton. Archaean assemblages associated with cratonization are shown here. Abbreviations for schist belts are as follows: Sa = Sandur, KKJH = Kolar–Kadiri = Jonnagiri–Hutti, RPSH = Ramagiri–(Penakacherla–Sirigeri)–Hungund, and VPG = Veligallu–Raichur–Gadwal superbelt. Dyke intrusions into the EDC are HandB = Harohalli and Bangalore swarm; A = Anantapur swarm; M = Mahbubnagar swarm; H = Hyderabad swarm. (Modified from Naqvi and Rogers, 1987).

ring at the margins (Naqvi and Rogers, 1987). It is located at the northern end of the Closepet Granite and differs from most of the belts in that it is not a thin N–S trending belt (Fig. 9). Granites in the center of the belt were dated using SHRIMP U–Pb at 2600–2500 Ma (Ramakrishnan and Vaidyanadhan, 2008). Rhyolites from the Sandur greenstone belt yield a SHRIMP zircon U–Pb age of 2658 ± 14 Ma (Nutman et al., 1996) and Naqvi et al. (2002) report Sm–Nd ages of 2706 ± 84 Ma for basalts and komatiites.

The RPSH consists of two discontinuous schist belts, the Ramagiri–Penakacherla–Sirigeri and the Hungund belts. Green schist facies metamorphism is most common in these belts although higher-grade metamorphic rocks occur locally. The RPSH belts are intruded by a series of granites and gneisses that provide minimum age constraints for the metamorphic protoliths of >~2500 Ma. Basalts from the Ramagiri greenstone belt are dated to 2746 ± 64 Ma (Pb–Pb; Zachariah et al., 1995). This result is consistent with ages of rhyolitic, basaltic and komatiitic lavas from the nearby Sandur greenstone.

The KKJH superbelt is located in the southern portion of the EDC and is a discontinuous band of linear belts (Fig. 9). The southern portion of the superbelt grades into a characteristic charnockitic terrain, while the north end (Kadiri belt) disappears beneath the Cuddapah Basin. The Kolar region contains mostly amphibolite grade metamorphic rocks. As with the other greenstone belts in the region, the KKJH are intruded by various felsic dykes that provide minimum age constraints. Pb–Pb isochron data provide an upper estimate at 2700 Ma for the protolith. This age is consistent with SHRIMP U–Pb zircon analysis of granites and gneisses located in the belt (Ramakrishnan and Vaidyanadhan, 2008). A second SHRIMP U–Pb zircon age of ~2550 Ma was found in various intrusions within the KKJH which provides a younger limit for the superbelt (Rogers et al., 2007).

The VRG Supergroup is located to the south of, beneath, and north of the Cuddapah Basin (Fig. 9). The group is split to the south of the basin and emerges to the north as a single unit before diverging again. The southern portion is divided by granite and is composed of metabasalts (amphibolites). The northern portion contains pillowed metabasalts and boninites that are typically formed during the early stages of subduction.

In summary, age constraints on the greenstone belts in the Eastern Dharwar Craton are known from only a few locations and all appear to be Neo-Archaean in age as compared to those in the Western Dharwar Craton described below. The relationships and stratigraphy of the gneissic rocks in the region are difficult to discern mainly due to the dismembered nature of the outcrop and the limited geochronology.

5.2. Dharwar Batholith

The Dharwar Batholith is a term first used by Chadwick et al. (2000) to describe a series of parallel plutonic belts (Fig. 10a and b). Previous works consistently used the term 'Peninsular Gneisses' to describe the majority of the EDC; however, it is compositionally different than the WDC gneisses, more granitic than gneissic, and hence the new terminology is more appropriate (Ramakrishnan and Vaidyanadhan, 2008). Age constraints from the WDC Peninsular gneisses suggest an early Archaean age, whereas the granitic gneisses of the Dharwar Batholith are of Late Archaean age. The plutonic belts are approximately 15–25 km wide, hundreds of km long and separated by greenstone belts (described above). They trend NW to SE except for in the south where the trends become predominately north–south. The belts are mostly mixtures of juvenile multipulse granites and diorites, and are wedge-shaped with steep granitic dyke intrusions (Chadwick et al., 2000; Ramakrishnan and Vaidyanadhan, 2008). Geochronologic information for this unit comes from SHRIMP U–Pb zircon measurements that constrain the emplacement of the Dharwar Batholith to 2700–2500 Ma (Friend and Nutman, 1991; Krogstad et al., 1995; Nutman et al., 1996; Nutman and Ehlers, 1998). Ages for granitic units appear to decrease from west to east; however, gneissic protolith ages of >2900 Ma are inferred from inherited zircons within younger dykes near Harohalli intruding the gneissic rocks (Pradhan et al., 2008).

5.3. Closepet Granite

The Closepet Granite is located on the western margin of the EDC and is a linear feature trending ∼N–S. The granite is 400 km long and approximately 20–30 km wide with shear zones on both sides. Recent studies suggest that the similar convexity of adjacent schist belts and granitic plutons may indicate that the Closepet Granite is a 'stitching pluton' formed during the suturing of the Eastern and Western Dharwar Cratons (Fig. 9; Ramakrishnan and Vaidyanadhan, 2008). The exposed rock is divided into northern and southern components by a part of the Sandur Schist Belt; however, both sections appear to be lithologically similar at the outcrop level (Naqvi and Rogers, 1987). The Closepet Granite is dated to 2513 ± 5 Ma (Friend and Nutman, 1991) and appears to be part of a widespread Neo-Archaean phase of plutonism (Mojzsis et al., 2003) in both the Eastern and Western Dharwar Cratons that we consider to mark the stabilization age for the WDC and EDC.

5.4. Cratonization history

Balakrishnan et al. (1999), Manikyamba et al. (2005), and others propose that the eastern portion of the Dharwar Craton formed as a result of island arcs accreting to an older (>3500 Ma), solid western craton through transpression. The linear schist belts represent back-arc basin environments that were metamorphosed during the accretion. The Eastern Ghats Mobile Belt is thought to mark the closure point during amalgamation of the EDC. Chadwick et al. (1996, 1999, 2000) suggest a similar idea; however, their model involves the 'Dharwar Batholith'. This idea suggests that already formed island arcs and granitic plutons constituted a single landmass. The Dharwar Batholith then obliquely converged with the WDC causing sinistral transpressive shear systems along the margins. Greenstone belts developed as a result of intra-arc basins associated with the batholith. The proposed timing of this collision history is between 2750 and 2510 Ma. These models suggest that the Closepet Granite was accreted onto the WDC, in contrast with the arguments presented below.

Jayananda et al. (2000), Chardon et al. (2002), and others propose that the most likely mechanism of formation of the EDC was through vertical tectonics. The plume model suggests a large mantle plume situated just beneath the EDC/WDC boundary in an enriched mantle. Further east, the plume began the melting of a colder and more depleted mantle. Induced melting from the plume is suggested to emplace juvenile magmas around 2500 Ma in the EDC. In this model, the greenstone belts result from inverse diapirism and resulting metamorphism. In this case, the Closepet Granite is simply a batholith rather than an accreted island arc or a stitching pluton between the EDC/WDC.

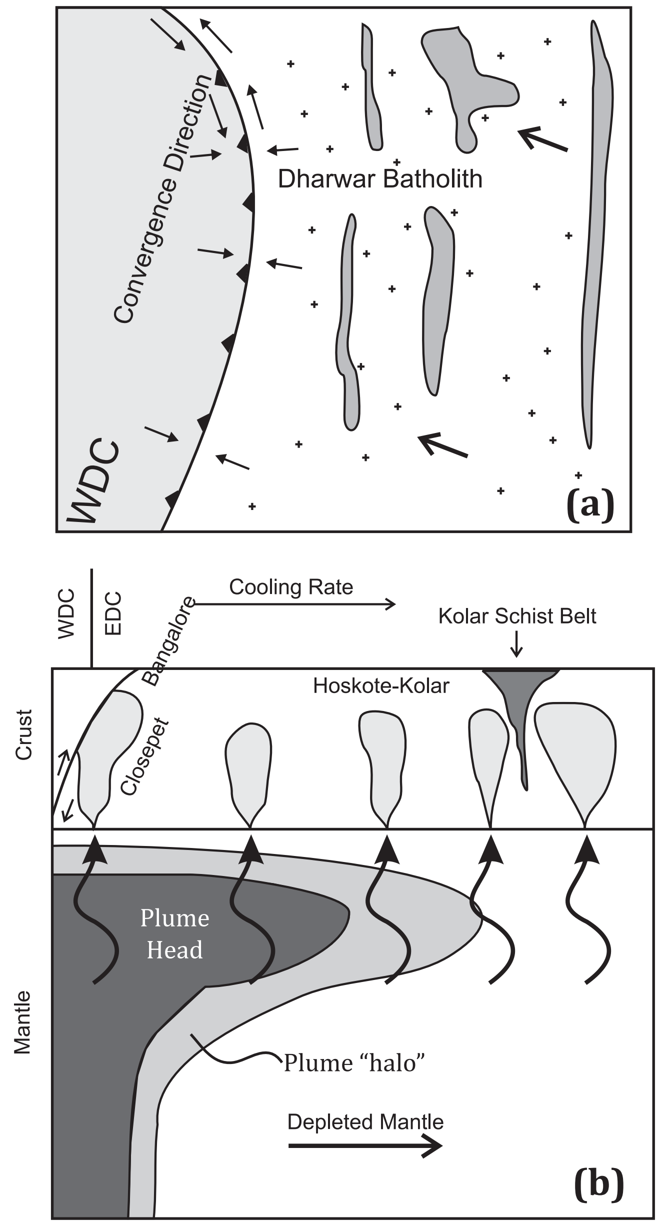

Fig. 10. (a) Model of cratonization of the Dharwar batholith. The WDC is believed to be the overriding plate and foreland contribution. The Dharwar batholith represents a series of juvenile granites sutured at the schist belts. (b) Superplume model for creating the granitic gneisses of the EDC. The mantle transitions from an enriched area near the plume to a depleted zone beneath the Kolar schist belt. Partial metling of the lower lithosphere creates plutons that form the mainland of the EDC (Modified from Jayananda et al., 2000).

5.5. Post-cratonization intrusive events

The majority of intrusive events of the Eastern Dharwar Craton (EDC) are represented by mafic dykes, kimberlites and lamproites.

Many of the clusters occur around the Cuddapah Basin and have three main trends: NW–SE, E–W, and NE–SW. These trends are associated with various paleostress orientations during the Proterozoic to Late Cretaceous (Srivastava and Shah, 2008). Most of the dykes disappear beneath the Cuddapah Basin, indicating that intrusion of the host granitic gneiss took place before the basin developed. These dykes all formed after the migmatitic activity of the host granitoids and are virtually free of any effects of metamorphism and deformation (Chakrabarti et al., 2004). Five major dyke clusters of the EDC, described below, include: (1) Hyderabad, (2) Mahbubnagar, (3) Harohalli/Bangalore, (4) Anantapur and (5) Tirupati (Fig. 9).

The Hyderabad cluster is located to the north of the Cuddapah Basin (Fig. 9). Widely spaced NNE–SSW to N–S trending dykes traverse ENE–WSW and WNW–ESE oriented dykes. The majority of the dykes present are doleritic in composition (Murthy, 1995). Whole rock K–Ar ages of local dykes indicate emplacement between 1471 ± 54 Ma and 1335 ± 49 Ma (Mallikarjuna et al., 1995), but these well may reflect a younger isotopic disturbance as at least some of the dykes in the Hyderabad cluster may be related to either the 1.9 Ga swarm in the Bastar Craton (French et al., 2008) or the ~2.2 Ga swarm near Mahbubnagar (French et al., 2004).

Located to the NW of the Cuddapah Basin (Fig. 9), the Mahbubnagar dyke swarm intrudes local granitic gneisses with Rb–Sr ages of 2.5–2.4 Ga and 2.2–2.1 Ga. The mafic dykes are predominantly gabbroic; however, dolerite and metapyroxenite are also present. They are oriented NW–SE and can be up to 50 km long and average 5–30 m wide. Chilled margins are common with coarse aphyric or plagioclase-rich interiors. Pooled regression results from Sm–Nd analysis gives an emplacement age of 2173 ± 64 Ma (Pandey et al., 1997). These results are duplicated by French et al. (2004), who obtained ages of ~2180 Ma using U–Pb techniques on nearby dykes. In light of the Sm–Nd and U–Pb ages for the dykes, it appears that the 2.2–2.1 Ga Rb–Sr ages cited above for the gneisses in the region may reflect disturbance due to dyke intrusion.

The Harohalli/Bangalore swarm is located between the southwestern portion of the Cuddapah Basin and the southeastern limb of the Closepet Granite (Fig. 9). The dyke cluster is split into an older group made up of dolerites, trending E–W (Bangalore dyke swarm), and a younger group of alkaline dykes that trend approximately N–S (Harohalli alkaline dykes; Pradhan et al., 2008). The Bangalore dyke swarm provided two robust U–Pb ages of 2365.5 ± 1.1 Ma and 2370 ± 1 Ma (French et al., 2004; Halls et al., 2007). Initial Rb–Sr whole rock measurements of the Harohalli alkaline dykes constrained ages to 850–800 Ma (Ikramuddin and Stueber, 1976; Anil-Kumar et al., 1989). However, recent U–Pb ages of 1192 ± 10 Ma produced by Pradhan et al. (2008) on the alkaline dykes challenge these earlier estimates.

Just west of the Cuddapah Basin is the Anantapur dyke swarm and south of the basin is the Tirupati swarm (Fig. 9). These two clusters are less studied than other areas; however, some poorly constrained ages are available. The NE–SW and ENE–WSW oriented dykes of the Anantapur swarm are dated using K–Ar measurements and are poorly constrained between 1900 and 1700 Ma and 1500 and 1350 Ma, respectively (Murthy et al., 1987; Mallikarjuna et al., 1995). More recently, Pradhan et al. (2010) dated the NE–SW trending 'Great Dyke of Bukkapatnam' of the Anantapur swarm to 1027 ± 13 Ma using U–Pb methods suggesting either a third phase of intrusion or more probably excess argon not discerned in the K–Ar data.

Dykes in the Tirupati swarm show two trends, the dominant trend is E–W and there are subordinate NW–SE trending dykes. There are K–Ar and Ar–Ar age determinations on dykes in the Tirupati swarm. The E–W trending dykes have K–Ar ages of 1073 and 1349 Ma and one Ar–Ar total fusion age of 1333 ± 4 Ma. NW–SE trending dykes have K–Ar ages of 935 and 1280 Ma (Mallikarjuna et al., 1995). Although there is a bit of agreement between two of the E–W dyke ages at ~1340 Ma, there was no clear plateau in the argon spectra making it likely that the reported K–Ar ages are integrating multiple episodes of disturbance.

Kimberlites and lamproites are found in relative abundance in four areas within the EDC. Concentrations can be found distributed around the Cuddapah Basin (Kumar et al., 2007; Fig. 9). They are characteristically potassic volcanic rocks that sometimes bear diamonds. The main areas of kimberlite–lamproite intrusions are known as the Wajrakarur, Narayanpet, Krishna and Nallamalai fields. Each of these fields contains multiple pipes. There are excellent age constraints on many of these fields. The Wajrakarur field is probably the best dated of the four. Rb–Sr ages on the Wajrakarur field form a tight cluster between 1091 and 1102 Ma and a recent U–Pb age on perovskite is 1124 +5/−3 Ma (Kumar et al., 2007). A newly discovered cluster at Sidanpalli (north of Wajrakarur) yielded an Rb–Sr whole-rock mineral isochron age of 1093 ± 4 Ma (Kumar et al., 2007). Miller and Hargraves (1994) report a U–Pb perovskite age of 1079 Ma for the Mulgiripalli pipe, but analytical details were not provided. Rb–Sr ages on kimberlites from the Kotakonda and Mudalbid kimberlite intrusions yielded ages of 1084 ± 14 and 1098 ± 12 Ma (Kumar et al., 2007). It should be noted that there are $^{40}Ar/^{39}Ar$ ages from the Kotakonda kimberlite that are much older. Chalapathi-Rao et al. (1999) obtained plateau ages of 1401 ± 5 Ma for a phlogopite separate from Kotakonda and 1417 ± 8 Ma from a lamproite at Chelima. The discrepancy in the Rb–Sr and $^{40}Ar/^{39}Ar$ ages from Kotakonda were recently addressed by Gopalan and Kumar (2008) who applied K–Ca dating to samples from the Kotakonda swarm and obtained ages of 1068 ± 19 Ma. Gopalan and Kumar (2008) argue that the $^{40}Ar/^{39}Ar$ results of Chalapathi-Rao et al. (1999) are affected by excess argon and the Kotakonda field is ~1100 Ma. It is unclear if the lamproite in Chelima represents an older suite of lamproitic intrusion. It is possible that the kimberlitic intrusions into the Dharwar Craton all occurred within a relatively narrow time frame from ~1050–1100 Ma. It should be noted that many other kimberlites around the globe were emplaced during this same interval of time including elsewhere in India (Majhgawan, Madhya Pradesh for example).

5.6. Proterozoic sedimentation

5.6.1. Cuddapah Basin

The Cuddapah Basin, located in the eastern portion of the EDC, is one of the most well studied basins in India (Fig. 9). It covers an area of approximately 44,500 km^2 and the convex western margin spans nearly 440 km. The eastern margin of the basin is represented by a thrust fault while all other boundaries are part of the Epi-Archaean Unconformity (non-conformity associated with undisturbed contact to older Archaean rocks). The sediments and minor volcanics of the basin are estimated to be approximately 12 km thick and made up of two distinct stratigraphic groups (Fig. 11). The Cuddapah Supergroup is the older unit and is present throughout the basin. The Kurnool Group was deposited unconformably over the Cuddapah rocks and is concentrated in the western portion of the basin. The basin is surrounded by granitic gneisses, dykes, and sills, all of which terminate at the basin boundary and appear to have formed before deposition. The youngest igneous activity in the basin is the kimberlite and lamproite field located near the basin center (Fig. 9 Chakrabarti et al., 2006).

Two competing hypotheses for the initiation of basinal subsidence and deposition were forwarded. Chatterjee and Bhattacharji (2001) propose that the basin was formed due to a mantle induced thermal trigger. Evidence for this comes from the presence of a large subsurface mafic body in the southwestern portion of the basin that provided episodic magmatism to form the abundant dykes and lava flows in and around the basin. These mantle flows may have been a result of collisional tectonics involving the Eastern Ghats Mobile Belt. A second hypothesis suggests that deep basin margin faults played a major role in controlling the evolution of the basin (Chaudhuri et al., 2002). Evidence for these marginal faults comes from seismic studies and Bouguer anomaly interpretations.

Lower limits for the onset of basin formation (assuming a thermal origin) can be inferred by ages of a mafic dyke on the southwest border of the craton and the Pulivendla Sill along the western margin of the basin. Chatterjee and Bhattacharji (2001) report a ^{40}Ar–^{39}40 age of 1879 ± 5 Ma for the mafic dyke that is coeval with the 1882 Ma U–Pb age on the Pulivendla Sill by French et al. (2008). Unpublished paleomagnetic data from the ~1.9 Ga Bastar dykes are identical to the Cuddapah traps volcanic (Clark, 1982). Overwhelming evidence suggests a thermal pulse of ~1.9 Ga for the initiation of basin formation in the Cuddapah Basin.

Cuddapah Basin sedimentation was discontinuous and numerous unconformities exist within the Cuddapah Supergroup. A major unconformity separates the Cuddapah Supergroup from the overlying Kurnool Group. Age constraints on the Kurnool Supergroup are lacking, but Goutham et al. (2006) correlate the Kurnool Group sediments with those in the Upper Vindhyan and assign all to the Neoproterozoic; however, such a correlation is based more on tradition rather than on strong correlative evidence and radiometric dating.

5.7. Pranhita–Godavari Basin

The Pranhita–Godavari (P–G) Basin is made up of two NW–SE trending sub-parallel basins sandwiched between the Dharwar and Bastar Cratons (Fig. 9). It is one of several Purana Basins formed (at least partially) on the Dharwar Craton. The Cuddapah (see above) lies to the south of the P–G Basin and the Bhima (discussed below) Basin lies to the southwest. The Paleozoic–Mesozoic aged Gondwana sediments lay between the eastern and western portions of the P–G Basin (Chaudhuri, 2003; Ramakrishnan and Vaidyanadhan, 2008). The sedimentary sequence within the basin consists of a series of unconformity-bounded packages reaching an aggregate thickness of ~6000 m. The rocks are mildly deformed and weakly metamorphosed. Age constraints are lacking within the basin although Chaudhuri (2003) gives a range between 1330 and 790 Ma for the sequence.

There are numerous stratigraphic interpretations (and names) for the P–G sequence, but we present the version favoured by Chaudhuri (2003). According to his classification, the basinal sediments are collectively referred to as the Godavari Supergroup and contain three unconformity-bounded groups (from oldest to youngest) known as the Pakhal Group, the Albaka Group and the Sullavai Group.

In the southwestern basin, the basal Pakhal Group is composed of two subunits called the Mallampalli and Mulug Subgroups. The Mallampalli Subgroup is predominately limestone and quartz arenite whereas the Mulug subgroup contains a basal conglomerate followed by a carbonate-rich shelfal sequence. Unconformably overlying the Pakhal Group is the Albaka Group composed of mature sandstones and shales. The uppermost Sullavai Group is floored by a conglomerate and sandstones of a primarily aeolian nature (Chaudhuri, 2003).

The northeastern basin contains only the Albaka and Sullavai Group sediments although a few authors have noted small outcrops of the Pakhal Group. The Proterozoic sedimentary sequence is unconformably overlain the Paleozoic–Mesozoic aged Gondwana Supergroup.

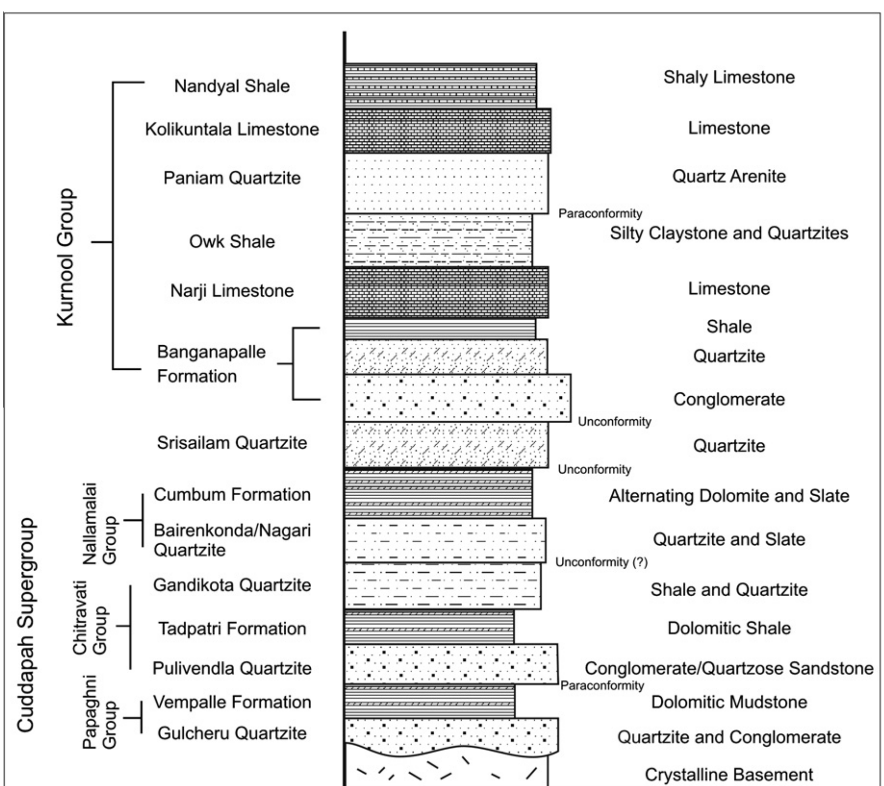

Fig. 11. Generalized stratigraphic succession of lithologies in the Cuddapah Basin (eastern India) after Naqvi and Rogers (1987) and Ramakrishnan and Vaidyanadhan (2008). The stratigraphic representations here are designed to show correlations and age relationships rather than absolute thickness or stratigraphic continuity.

5.8. Bhima Basin

The Bhima Basin is located between the northern margin of the EDC and the Deccan Trap flows (Fig. 9). The basin is much smaller than the Cuddapah and covers 5200 km^2 with the longest portion having an axis of 160 km (NE–SW). The southern portion of the basin is bounded by an unconformity with the underlying granitic gneisses while the E–W and NW–SE borders are bounded by faults. The full extent of the basin is unknown due to the Deccan Trap covering the basin in the north. The Bhima Group is predominantly composed of limestones; however, sandstone and conglomerate sediments exist between the basement and the upper sequence limestones. The oldest age for the formation of the Bhima Basin is constrained by the underlying granitic gneisses to $\sim$2500 Ma (Sastry et al., 1999). It is currently under debate as to whether the basin formed during the Meso- or Neoproterozoic (Patranabis-Deb et al., 2007; Malone et al., 2008).

6. Western Dharwar Craton

The Western Dharwar Craton (WDC) is located in southwest India (Fig. 12). It is bounded to the east by the Eastern Dharwar Craton, to the west by the Arabian Sea, and to the south by a transition into the so-called "Southern Granulite Terrane." The remaining boundary to the north is buried under younger sediments and the Cretaceous Deccan Traps. The division between the Western and Eastern Dharwar Cratons is based on the nature and abundance of greenstones, as well as the age of surrounding basement and degree of regional metamorphism (Rollinson et al., 1981).

The Archaean Tonalitic–Trondhjemitic–Granodioritic (TTG) Gneisses are found throughout the Western Dharwar Craton, dated at 3.3 to 3.4 Ga via whole rock Rb–Sr and Pb–Pb methods (Pitchamuthu and Srinivasan, 1984; Bhaskar Rao et al., 1991; Naha et al., 1991). U–Pb zircon ages ranging from 3.5 to 3.6 Ga have also been published. Three generations of volcanic-sedimentary greenstone granite sequences are present in the WDC: the 3.1–3.3 Ga Sargur Group, the 2.6–2.9 Ga Dharwar Supergroup (Radhakrishna and Vaidyanadhan, 1997) and 2.5–2.6 Ga calc-alkaline to high potassic granitoids, the largest of which is the Closepet Granite (Jayananda et al., 2008). The Dharwar supracrustal rocks uncomformably overlie widespread gneiss-migmatite of the Peninsular Gneissic Complex (3.0–3.3 Ga) that encloses the Sargur schist belts (Naqvi and Rogers, 1987).

The WDC shows an increase in regional metamorphic grade from greenschist to amphibolite facies in the north and granulite facies in the south. The metamorphic grade increase corresponds to a paleopressure increase from 3 to 4 kbar in the amphibolite facies to as much as 9–10 kbar (35 km paleodepth) in the highest-grade granulite-transition zone along the southern margin of the craton (Mojzsis et al., 2003). A nearly continuous cross section of Late Archaean crust that has been tectonically upturned and channeled by erosion is exposed in the WDC.

6.1. The Sargur Group

The Sargur Group greenstone belts display well-preserved volcano-sedimentary sequences. Generally these comprise of ultramafic to mafic volcanic rocks (komatiitic to tholeiitic sources) that shows an up-section transition to felsic volcanic rocks, often interpreted to be related to a calc-alkaline source (e.g. Naqvi, 1981; Srikantia and Bose, 1985; Charan et al., 1988; Srikantia and Venkataramana, 1989; Srikantia and Rao, 1990; Venkatadasu

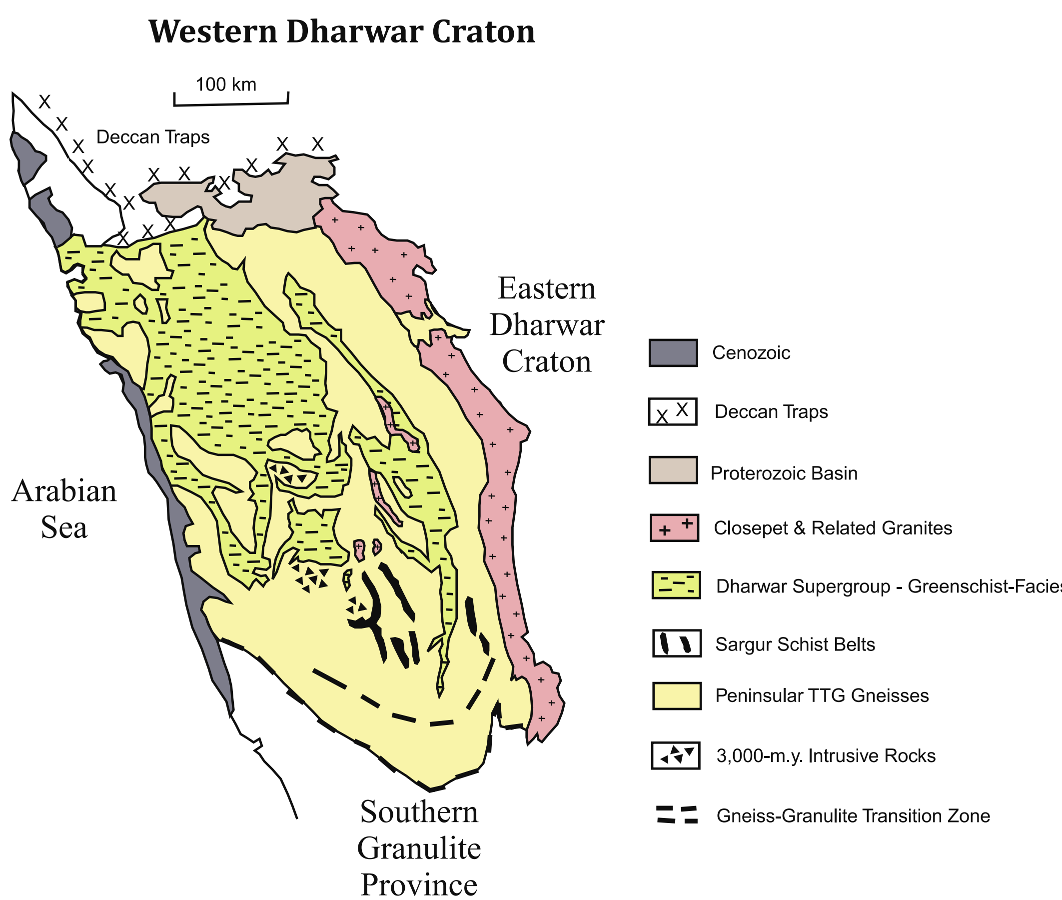

Fig. 12. Sketch map of the Western Dharwar Craton showing major lithologic boundaries after Naqvi and Rogers (1987) and Ramakrishnan and Vaidyanadhan (2008).

et al., 1991; Devapriyan et al., 1994; Subba Rao and Naqvi, 1999; Paranthaman, 2005). These include the Ghattihosahalli, the J.C. Pura, the Bansandra area, the Kalyadi area, and the Nuggihalli belt (Jayananda et al., 2008).

The Sargur Group developed from several distinct geodynamic processes across a span of millions of years. Detrital zircons from the schist yield a Pb–Pb evaporation age of ~3.3 Ga. SHRIMP U–Pb analysis yielded ages between 3.1 and 3.3 Ga, with some analyses yielding a 3.6 Ga age inherited from the protolith. Sm–Nd model ages of ~3.1 Ga were calculated from the ultra-mafic units. Rb–Sr dating on anorthosite fell into this range as well, resulting in a 3.1 Ga age for the unit. When taken together, this geochronologic dataset may constrain the age of the Sargur Group to ~3.1 Ga. The older ages present in these analyses are likely inherited from the basement material, and may represent the older limit of the group. The Sargur unit appears to have formed in a subduction setting, likely derived from the melting of oceanic slab materials (Martin, 1986). Komatiites found in the Sargur Group are interpreted by Jayananda et al. (2008) to be related to plume events, and may have originally been elements of oceanic plateaus. These accreted oceanic plateaus then served as a base for further subduction related processes, represented by the series of mafic to felsic volcanic units emplaced over and intruded the ultramafic plateau sequences (Jayananda et al., 2008).

Two commonly suggested single stage methods of generating the observed rocks in the WDC are suggested: (1) massive partial melting related to a plume event and (2) magmatism related to subduction processes. The ultramafic rocks in the greenstone belts show evidence of high (1600–1700 °C) eruption temperatures as well as elevated Mg, depletion in Al and low concentrations of incompatible elements. Most komatiite samples analyzed also show elevated Ni and Cr values, either absent or positive Nb anomalies, and negative Hf anomalies (Jayananda et al., 2008). Many authors (e.g. Campbell et al., 1989; Ohtani et al., 1989; Griffith and Campbell, 1992; Arndt, 1994, 2003; Arndt et al., 1997; Chavagnac, 2004) suggest that these characteristics are most easily explained by decompression melting of a mantle plume head at depth. Problems with a plume model arise, however, when the chemistry of associated felsic volcanics and TTG basement rock is considered (Jayananda et al., 2008). The subduction generation hypothesis offers a contrasting view, based on the assumption that lateral accretion of crust was a major factor in the formation of the cratonic continental nuclei (Martin and Moyen, 2002; Smithies et al., 2003). Taylor and McLennan (1985) proposed that continental crust, on average, is andesitic in composition. Following this logic, many workers suggested that subduction zones represent major sites of crust formation. Drury (1983) suggested that the mafic volcanic rocks of the WDC were formed through arc-related processes. The geochemical data, however, do not support this idea. The ultramafic volcanics display Al-depletion, absent or positive Nb anomalies, and Nb–U, Nb–Th, Nb–La, Th–U ratios that are not consistent with an arc setting (Jayananda et al., 2008). Also, even assuming the elevated Archaean geothermal gradient, it is difficult to explain the high eruption temperatures indicated by the ultramafic volcanic units in such an environment. Given the failure of one mode of generation to explain all of these features, a more complicated model must be considered.

The extensive ultramafic to basaltic flows present in the WDC may represent accreted oceanic plateau crust; indeed, the high degree of melting needed to form such plateaus reveal the enhanced thermal potential of a plume source. Isotopic and geochemical data of the volcanics are inconsistent with the assimilation of older, felsic crust into the komatiitic magmas erupting to form the ocean plateau sequences (Jayananda et al., 2008). The absence of such contamination by continental material suggests that the plume must have melted beneath and erupted onto oceanic lithosphere. Ubiquitous pillow lava textures are evidence of the oceanic character of the crust seen in the ultra-mafic units, diagnostic of submarine flows. Peucat et al. (1995) and Jayananda et al. (2008) conclude that ~50 Ma period separates the formation of the oceanic plateau ultra-mafic sequences and the formation of the subduction related felsic to mafic volcanic sequences above them on the basis of isotopic work. This thickened plateau crust, as well as the restite in the lithospheric mantle root beneath the plateau is unsubductable and therefore likely to accrete against any continental margin encountered (Cloos, 1993; Abbot and Mooney, 1995). The accretion of such a plateau along a continent would jam the existing subduction zone, forcing a "jump" in subduction from the old continental margin to the outboard plateau margin. This would subsequently result in the formation of a new igneous arc on the plateau, represented in the WDC as the felsic to mafic volcanic sequence. The deeper level intrusions emplaced beneath the remnant oceanic plateau are interpreted to serve as the protolith for the TTG's of the WDC basement (Jayananda et al., 2008).

The komatiitic–tholeiitic volcanism observed in the WDC is part of a larger scale process that led to the growth of the proto-craton. The 3.35 Ga volcanism appears to have been pene-contemporaneous with the formation of the TTG basement, and provided hosting for the intrusion of the TTG protoliths. The melting events that lead to the ultra-mafic volcanism occurred over a range of depths and co-existed with mantle peridotite; however, evidence for the presence of garnet in the residue is unclear (Jayananda et al., 2008). Trace element and Nd isotope data rule out the assimilation of continental materials into the magma. Instead, the komatiite magmas show the characteristic geochemical evidence of a depleted mantle source (Boyet and Carlson, 2005). Mantle depletion at 3.35 Ga is potentially significant. It would suggest that the earlier extraction of enriched materials from the mantle had depleted the upper mantle prior to 3.35 Ga. A mantle plume is a prime candidate for the eruption of such a massive volume of komatiitic magma; however, the later felsic to mafic igneous activity bears the signature of subduction processes (Jayananda et al., 2008).

6.2. The Dharwar Supergroup

The Dharwar Supergroup is exposed in two large schist belts that have been divided into two sub-sections, the Bababudan Group and the Chitradurga Group. The Bababudan Group is spread over a 300 km long and 100–150 km wide area, and is made up of the Babadudan schist belt, Western Ghats Belt, and the Shimoga schist belt. The Bababudan schist belt covers an area of approximately 2500 km^2. The base of this unit is represented by the Kartikere conglomerate that discontinuously extends along the southern margin of the belt for ~40 km. This unit grades into a quartzite. The detrital zircon population from the quartzite suggests that the sediments were mainly derived from the Chikmagalur granodiorite. The overlying formations typically consist of metabasalts with intercalated meta-sedimentary units, with occasional gabbroic sills, minor BIF, and phyllites. These are thought to represent a variety of terrestrial environments, ranging from braided fluvial systems to sub-aerial lava flows. The Western Ghats Belt is a large schist belt about 2200 km^2 in extent, and about 150 km by 15 km in dimension. The stratigraphy closely resembles the Babaudan belt; however, a major group of basalts, felsic volcanics, and pyroclastic units is also seen in the upper levels. The Shimoga schist belt is a large (25,000 km^2) NW trending belt separated from the previous two by outcropping TTG basement gneiss. The contact between these basement gneisses and the schist belt is observed as a zone of high-grade metamorphism, often with kyanite and garnet phases present. Granitioid intrusions are also present in the north of the belt.

6.3. Proterozoic dyke swarms

Mafic dyke swarms varying in orientation and composition, intrude many areas of the WDC. Murthy et al. (1987) noted that the dykes are prevalent north of latitude 13°N and east of longitude 78°E, but that the dykes trend out towards latitude 12°N and are nearly gone south of latitude 11°N. All of the dykes post-date migmatitic activity in the host granitoids and are thus free of overprints of deformation and metamorphism.

There are three main dyke swarms of Proterozoic age in the Western Dharwar Craton known as the (1) Hassan–Tiptur dykes; (2) Mysore dykes and (3) "Dharwar" dykes (Radhakrishna and Mathew, 1993).

The Hassan–Tiptur dyke swarm contains two suites of dykes an older amphibolite and epidioritic swarm and younger and more widespread doleritic dykes. Age constraints are lacking on both suites of dykes. The Mysore dykes trend E–W and form a dense swarm near the town of Mysore. The dykes are not dated.

6.4. Proterozoic Sedimentary Basins

The E–W trending Kaladgi–Badami Basin is the only significant Proterozoic intracratonic basin of the Western Darwar Craton located along the northern edge of the craton (Fig. 12). This basin formed on TTG gneisses and greenstones of Archaean age. The Kaladgi Supergroup preserves the record of sedimentation in the basin, and consists of sandstones, mudstones and carbonates. The textural and mineralogical maturity of this basin increased over time, indicating that the regional relief surrounding the basin declined over time, with the clastic sediments being derived from the local gneiss and greenstone rock (Dey et al., 2009). An angular unconformity between the two constituent groups (The lower Bagalkot and overlying Badami) suggests a period of uplift in the basins history (Jayaprakash et al., 1987). Deformation in Bagalkot group is significant, whereas the upper group only exhibited mild deformation (Kale and Phansalkar, 1991).

6.5. Granitic intrusions

Late to post-tectonic Dharwar potassic granite plutons (~2.5–2.6 Ga) that are assumed to reflect crustal reworking in WDC, occur as isolated intrusions cutting across the foliation and banding of the Peninsular Genisses (~3.0 Ga; Jayananda et al., 2006). In many cases, these plutons occur as distinct types either separately or as parts of larger composite intrusions, likely related to the generation of melts at differing depths within the crust (Sylvester, 1994). Several classes of TTG's are present as well, broadly split into classical TTG and transitional TTG that formed 500 Ma later. These transitional TTG's are believed to be lower crustal derived melts, and share the garnet residue signal of the high K granites; however, this similarity may also indicate a mixing between these two melts (Jayananda et al., 2006). There is still uncertainty as to the role the late potassic granites played in the cratonization of the WDC. Jayananda et al. (2006) suggest that they may be related either to a thermal event prior to the termination of craton stabilization, or that they actually represent part of a longer term (~100 Ma) stabilization. Age data from Taylor et al. (1984) for the various intrusions range from 3080 ± 110 Ma (Rb–Sr) and 3175 ± 45 Ma (Pb–Pb isochron) for the Chikmagalur Granite to 2605 ± 18 Ma (Pb–Pb isochron). Much of the data is based on older, whole rock isotopic work.

The Chitradurga Granite is an elongate, lenticular body of late to post-tectonic granite, about 60 km long and 15 km wide. The granite is clearly intrusive into the Jogimaradi lavas of the Bababudan Group as well as into the TTG basement. The Chitradurga is biotite granite grading into granodiorite and quartz monzonite. Chadwick et al. (2007) dated the granite using Pb–Pb and Rb–Sr isochrons yielding an age of ~2.6 Ga, as well as SIMS U–Pb zircon age of ~2610 Ma. The Jampalnaikankote Granite is a ~2.6 Ga (Rb–Sr) roughly oval shaped pluton that intrudes the Chitradurga schist belt. The Arsikere and Banavara Granites are thought to be from a single pluton that is connected at depth. The Arsikere granitic batholith is approximately 75 km^2 and oval in shape. The intrusion is primarily potassic biotite granite that yielded a Rb–Sr age of ~2.6 Ga, and a SIMS U–Pb zircon age of ~2615 Ma. The Chamundi Granite is another potassic pluton, with associated radial and parallel dykes, that intrudes the Peninsular Gneiss. The granite has been dated via Rb–Sr at ~800 Ma.

7. Southern Granulite Province

India's Southern Granulite Province (SG) consists of three Late Archaean to Neoproterozeric, high-grade metamorphic blocks, joined together by a series of Neoproterozoic (?) shear zones. The Northern Block (NoB) is situated at the southern tip of the Dharwar Craton (DC) and is bounded to the south and east by the Moyar–Bhavani Shear Zone (MBSZ). The MBSZ is a complex network of mobile belts that form the northern boundary of the Central Block (CB), that is divided by a SW–NE trending branch of the MBSZ into the Nilgiri Block (NiB) in the west and the Madras Block (MaB) in the east. The MaB is bordered to the south by Palghat-Cauvery Shear Zone (PCSZ), marking the suture between the Archaean NoB and CB and the Proterozoic Madurai Block (MdB). The NW–SE trending Achankovil Shear Zone (ACSZ) separates the MdB from the southernmost Trivandrum Block (TB) (Fig. 13).

7.1. Northern Block

The Northern Block (also known as the Salem Block) of the Southern Granulites consists of a granulite massif at the southern edge of the Dharwar Craton. The block is located between the 'Fermor line' and the Palghat-Cauvery Shear Zone (PCSZ; Fig. 13). Lithologies present in the Salem include, pyroxene-bearing granites (charnockites), granite gneisses, and migmatites (Devaraju et al., 2007; Clark et al., 2009). Sm–Nd ages of 3.3–2.68 Ga suggest the DC as protoliths for these rocks (Devaraju et al., 2007). Geothermobarometric studies indicate that these rocks have formed at 700 ± 30 °C and 5 – 7 kbar (Harris et al., 1982). A gradational increase in metamorphic grade from the granite–greenstone belts (greenschist and amphibolite facies) of the DC to the granulite massif of the Salem Block, as well as a lack of a surficial structural break, call into question whether or not the Salem Block is a crustal block distinct from the DC. Seismic studies indicate a bivergent reflection pattern at the boundary between the NoB and the DC (with reflections dipping towards the boundary), a thicker crust (~46 km) beneath the Salem Block, and a pattern of tectonically induced imbricate faulting in the lower crust and upper mantle of this region (Rao and Prasad, 2006; Rao et al., 2006). These observations are indicative of a collisional environment in which the Salem Block was accreted onto the DC in the mid-Archaean.

Most recently, Clark et al. (2009) dated charnockitic rocks within the Salem block to 2538 ± 6 Ma and 2529 ± 7 Ma (SHRIMP Pb–Pb). SHRIMP-dated rims from the same zircons showed statistically distinct 2473 ± 8 Ma and 2482 ± 15 Ma ages. Clark et al. (2009) considered the Archaean-ages crystallization ages for the charnockitic protolith and the younger ages as partial melting that occurred during the accretion of the SB to the Dharwar Craton.

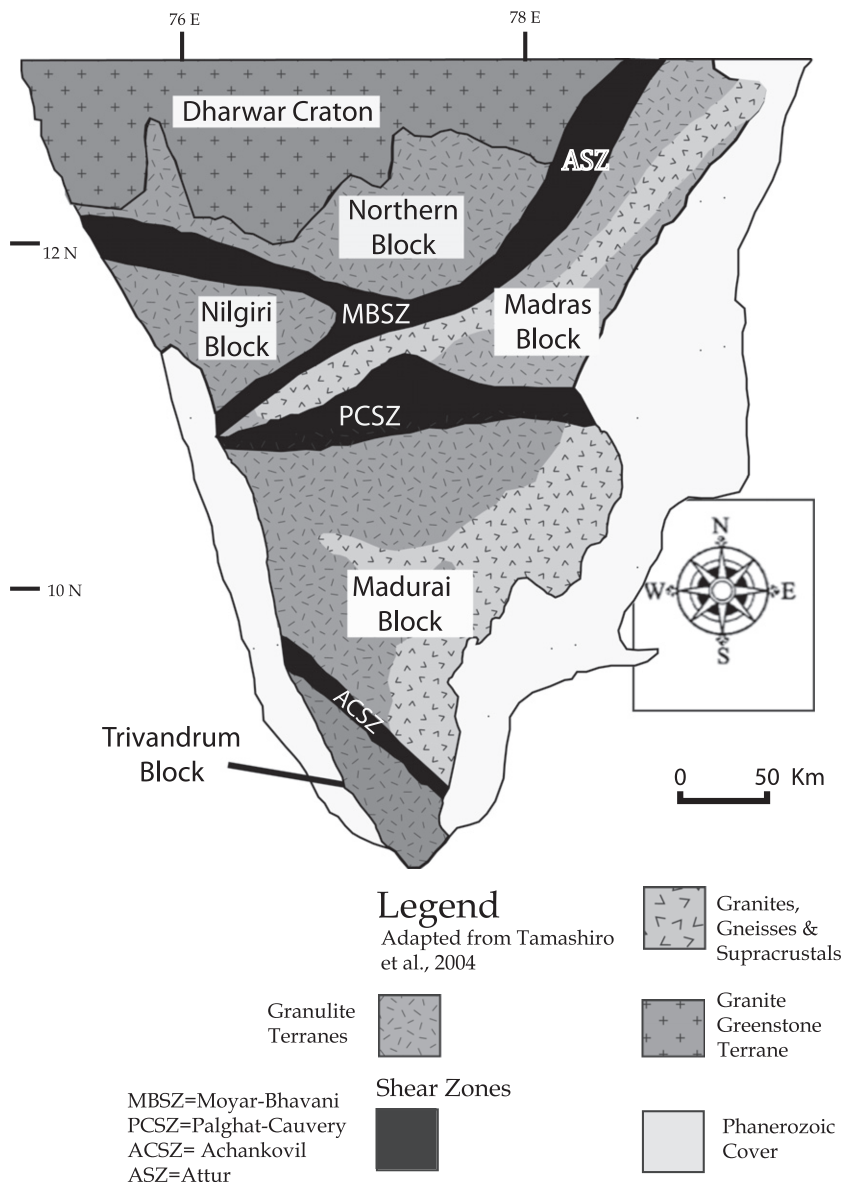

Fig. 13. Sketch map of the Southern Granulite Province blocks and associated shear zones after Naqvi and Rogers (1987) and Ramakrishnan and Vaidyanadhan (2008).

7.2. Palghat-Cauvery Shear Zone

The PCSZ is the suture between the Archaean granulite blocks to the north and the Neoproterozoic blocks to the south. It is the northernmost manifestation of Gondwana orogenies on the Indian Peninsula (Harris et al., 1994; Chetty et al., 2006; Naganjaneyulu and Santosh, 2010). This shear zone has been correlated with the Bongolava–Ranotsara Shear Zone (de Wit et al., 1998; Clark et al., 2009), the Madagascar Axial High-Grade Zone (Windley et al., 1994), and the East Antarctic Napier and Rayner Complexes (Harris et al., 1994). Drury (1984) and Meert (2003) have established the PCSZ as a strike-slip zone related to the final assembly of East Gondwana. Sm–Nd garnet pair samples yield isochrons of 521 ± 8 Ma, while biotite pair Rb–Sr data yield ages of 485 ± 12 Ma (Meissner et al., 2002). In contrast, Naganjaneyulu and Santosh (2010) also view the PCSZ as a major suture involving subduction-thickened crust to the north and collision and accretion of the Madurai Block with the Dharwar Craton to the north. In their model, the PCSZ is linked to Madagascar axial high-grade zone and the Achankovil shear zone to the south is correlated to the Bongolava–Ranotsara shear zone.

7.3. Nilgiri Block

The Niligiri Block (NiB) is a triangular block consisting mostly of garnetiferous, enderbitic granulites wedged between the segments of the Palghat-Cauvery Shear Zone. Kyanite-gneisses, quartzites, and gabbroic to anorthositic pyroxenites are also interspersed throughout the block (Raith et al., 1999). A preponderance of evidence, including whole rock Sm–Nd and Rb–Sr analyses indicates granulite facies metamophism as late as 2460 ± 81 Ma (Raith et al., 1999). These ages are supported by U–Pb data from overgrowths in detrital zircon found in enderbites that show metamorphic pulses at 2480–2460 Ma (Buhl, 1987). The NiB is believed to be the deepest exhumed crust on the Indian Penninsula; paleodepths range from ~22 km in the southwest (6–7 kbar) to ~35 km (9–10 kbar) at the Moyar Shear Zone (Raith et al., 1999).

The age of metamorphism is identical to that observed in the Salem Block to the north (Clark et al., 2009).

7.4. Madras Block

The MaB is situated between segments of the Palghat-Cauvery Shear Zone to the east of the NiB. The MaB consists of medium to high-pressure charnockites and gneisses that are squeezed into a long, thin band presumably resulting from shearing in the PCSZ. Included in this area are the Madukkarai Supracrustals, that display broad doming, complex folding, and extreme hinge line variation caused by Neoproterozoic–Ordovician suturing along the PCSZ extending beyond the Madras Block proper (Chetty and Rao, 2006). Retrograde amphibolite facies rocks appear at the boundaries of the shear zones (Santosh et al., 2002). A variety of methods, including U–Pb in zircons and whole rock Sm–Nd and Rb–Sr sampling have been used to constrain the age of granulite formation to 2600–2500 Ma (Vinogradov et al., 1964; Crawford and Compston, 1969; Bernard-Griffiths et al., 1987). Electron microprobe analysis (EPMA) of zircons and monazites support these dates and reveal a second thermal impulse with a broad age range from 2000 to 1700 Ma (Santosh et al., 2002).

7.5. Madurai Block

The MdB is largest of the Southern Granulite Blocks. The western part of the block, consisting almost entirely of charnockite massifs, displays UHP and UHT metamorphism at 8–11 kbar and 1000–1100 °C. These conditions are manifested in the field by the presence of sapphirine–spinel–quartz assemblages present in the rocks (Braun et al., 2007). The eastern part of block is composed of basement gneisses and related meta-sedimentary complexes. Zircons found in the gneisses have a bimodal age distribution of 2100–1600 Ma and 1200–600 Ma. Only the younger of these populations is found in the charnockites. Monazites display bimodal age range at 950–850 Ma and 600–450 Ma (Santosh et al., 2002; Braun et al., 2007). Braun et al. (2007) speculated that the bimodal distribution of monazite ages indicated two distinct high-grade metamorphic intervals one at ca. 900 Ma and the other at ca. 550 Ma.

Collins et al. (2007) dated a number of zircons from a quartzite unit and found detrital populations of 2700, 2260, 2100 and 1997 Ma and a metamorphic overprint at 508.3 ± 9.0 Ma. They attributed the Cambrian ages to high-grade metamorphism during the final stages of Gondwana assembly consistent with the model of Naganjaneyulu and Santosh (2010).

7.6. Achankovil Shear Zone

The ACSZ is a NW–SE trending, subrectangular shear zone stretching more than 120 km in a NW–SE direction and is up to 50 km wide (Rajesh and Chetty, 2006). The ACSZ forms the boundary between the MdB in the north and the TB to the south. The ACSZ contains a variety of lithologies, including highly migmatized biotite-garnet gneisses, cordierite-opx gneisses, aluminous metapelites, mafic granulites, and calc-silicates (Rajesh and Chetty, 2006). Model Sm–Nd ages from a selection of samples suggest a protolith age between 1700 and 1500 Ma, significantly younger than the protoliths for either of the adjacent granulite blocks (Brandon and Meen, 1995; Bartlett et al., 1998; Cenki-Tok et al., 2005). Individual zircons and monazites, analyzed by CHIME and SHRIMP indicate metamorphic events between 525 and 508 Ma (Santosh et al., 2004, 2005). Minerals analyzed through EPMA give chemical ages between 520 and 590 Ma (Braun and Brocker, 2004). Further U–Pb dating methods on zirconalite found in an ultramafic suite, as well as K–Ar studies in biotite show metamorphism continuing in the ACSZ until the Ordovician (478–445 Ma) (Soman et al., 1982; Rajesh et al., 2004). Broad overlap between ages given by a variety of minerals with dissimilar closing temperatures suggests rapid cooling in the shear zone (Rajesh et al., 2004).

This shear zone is considered by Naganjaneyulu and Santosh (2010) to be a continuation of the Bongolava–Ranotsara shear zone in Madagascar.

7.7. Trivandrum Block

Also known as the Kerala Khondalite Belt (KKB), the TB is comprised of an extensive array of UHT supracrustal rocks, including sillimanite granulites, garnet-opx granulites, two pyroxene granulites, garnet–biotite gneisses, and calc-silicates (Santosh et al., 2006). A large charnockite massif at the southern end of the TB is often referred to as the Nagercoil Block (NaG). Extensive studies of individual zircons throughout the TB have revealed significant information regarding the formation of continental crust and supercontinents (Bindu et al., 1998; Braun et al., 1998; Montel et al., 2000; Santosh et al., 2006). Zircon cores dated as old as 3460 ± 20 Ma are relicts of some of the earliest formed continental crust (Zeger et al., 1996). Ages of other mineral grain cores and ages of overgrowths on older grains cluster around 1600 Ma, 1000 Ma, and 600–400 Ma. These dates appear to be related to the assembly of the supercontinents Columbia, Rodinia, and Gondwana/Pangaea, respectively (Santosh et al., 2006).

8. Mobile Belts (North of the Southern Granulite Province)

Peninsular India comprises a mosaic of five Precambrian terranes, the Eastern Dharwar, Western Dharwar, Aravalli–Bundelkhand, and Bastar–Singhbhum Cratons in addition to the Southern Granulite terrane. These cratons are separated by suture zones, mobile belts, and rifts. In our opinion, the term 'mobile belt' should be abandoned. The original concept of a mobile belt was tied to geosynclinal theory and implies formation by a process other than plate tectonics. Nevertheless, the terminology has persisted in the literature and has morphed into something even more confusing. In some publications 'mobile belt' is synonymous with 'orogenic belt' and in other cases it retains a non-tectonic correlation. For the purposes of our paper, we use the term mobile belt for convenience with the express conclusion that they formed via normal plate tectonic processes.

The Closepet Granite separates the Eastern and Western Dharwar Cratons and is considered to be a stitching pluton of Late Archaean age (Friend and Nutman, 1991). The composite Dharwar Craton is separated from the Bundelkhand Craton to the north by the Central Indian Tectonic Zone (CITZ) and from the Bastar–Singhbhum Craton to the east by the Godavari Rift. The CITZ also separates the Bundelkhand Craton from the Bastar–Singhbhum composite Craton (Mall et al., 2008). The Bhavani–Palghat Mobile Belt separates the composite Dharwar terrane to the north from the Southern Granulite Terrane to the south (Reddy and Rao, 2000).

8.1. Central Indian Tectonic Zone

The Central Indian Tectonic Zone (CITZ) (Fig. 14a and b), also referred to as the Satpura Mobile Belt, is a complex Proterozoic orogenic belt that formed during the accretion of the Bastar–Singhbhum Craton to the northern Bundelkhand Craton (Radhakrishna, 1989; Acharyya, 2003). The mobile belt was originally named after the Satpura Hills and is bounded by the Narmanda-Son North Fault (NSNF) and to the south by the Central Indian Suture (CIS; Fig. 14b). The gneissic tectonic zone comprises three sub-parallel E–W trending supracrustal belts that are separated from each other by crustal scale shear zones (Ramakrishnan and Vaidyanadhan, 2008). The belts from north to south are the Mahakoshal Belt, the Betul Belt, and the Sausar Belt. The Narmada-Son South Fault (NSSF) separates the Mahakoshal and Betul Belts while the Tan Shear Zone (TSZ) separates the Betul and Sausar Belts.

Acharyya (2003) evisions a polyphase process for the assembly of north and south India and subsequent reactivation events concentrated in the CITZ. The CITZ began as the Mahakoshal Rift closed by southward subduction joining the northern India blocks with those from the south sometime between 1.8 and 1.7 Ga. Two major orogenic events followed wherein the Sausar and Chhotanagpur granite gneisses were metamorphosed between 1.6 and 1.5 Ga and again around 1.0 Ga. These latter two orogenic events are viewed as part of the global orogenic cycle leading to the formation of the Rodinia Supercontinent. As previously noted, this polyphase orogenic cycle for the CITZ is rejected by Stein et al. (2004) who consider that all 'events' younger than ~2.5 Ga are smaller scale reactivation events in this zone of crustal weakness.

8.2. Central Indian Suture

The Central Indian Suture (CIS) (Fig. 14a and b) is a brittle-ductile shear zone that delineates the southern boundary of the CITZ and forms the boundary between the Bundelkhand Craton to the north and the Bastar Craton to the south (Leelandandam et al., 2006). The suture zone separates the high-grade Sausar meta-sedimentary and granulite rocks to the north from the low-grade proterozoic volcanic rocks to the south. Silicified, brecciated, and mylonitized rocks typify the CIS which extends from the SE of Nagpur for almost 500 km to the ESE of Balaghat (Mishra et al., 2000). Yedekar et al. (1990) suggested that the CIS underwent the following tectonic history: (1) oceanic crust between the Bundelkhand and Bastar Cratons started to subduct at ca. 2.3 Ga; (2) initiation of the calc-alkaline plutonism formed the Malanjkhand and Dongargarh Plutons; (3) at 2.1 Ga, the Sakoli and Nandgaon volcanics formed an island arc in the southern block; (4) the two blocks collided from ~2.1 Ga to 1.7 Ga forming the CIS; (5) the Sausar Fold Belt developed during the 1.7–1.5 Ga interval; and finally (6) back-arc extension ensued resulting in the Khairagarh Group accompanied by bimodal volcanism continuing from 1.0 to 0.7 Ga.

8.3. Mahakoshal Belt

The Mahakoshal Belt (MB) extends about 600 km from Barmanghat to Rihand Dam and trends in an ENE–WSW fashion (Fig. 14b). The NSNF delineates the northern boundary to the MB and separates it from the Vindhyan Basin to the north. The southern boundary, mostly covered by the Deccan Traps, is bounded by the NSSF that separates the MB from the Proterozoic granites of the CITZ (Ramakrishnan and Vaidyanadhan, 2008). Quartzites, carbonatites, chert, banded iron formations, greywacke–argillite and mafic volcanic rocks dominate the MB. The MB exhibits evidence of three phases of deformation with an overall ENE–WSW trend. The first phase produced upright isoclinal folds with steep southward dipping axial planes. The second phase of deformation resulted in vertical to reclined, E–W striking folds with axial planes dipping to the south and a pronounced crenulation cleavage. The third phase produced broad folds and N–S striking axial planes (Ramakrishnan and Vaidyanadhan, 2008).

8.4. Betul Belt

The Betul Belt (BB) trends ENE–WSW and extends about 135 km from Betul to Chhindwara and has a width of about 15–20 km (Fig. 14b). The BB is located between the MB to the north and the SB to the south. The Betul Belt is thought to represent a continental arc setting based on the presence of bimodal volcanism with younger calc-alkaline granitoids. The occurrence of ~1500 Ma syntectonic granites and ~850 Ma late-to-post-tectonic granites indicate the BB and the MB may be pene-contemporaneous where the BB represents the arc and the MB represents the back-arc at the old continental margin (Ramakrishnan and Vaidyanadhan (2008).

8.5. Narmada-Son Lineament

Although the Narmada-Son lineament (fault; NSL) is not described as a mobile belt, it is one of the most prominent crustal features in Peninsular India and forms part of the larger CITZ (Fig. 14a and b). The lineament can be further subdivided into a NNSL (North Narmada-Son lineament) and SNSL (South Narmada-Son lineament). The NSL extends at least 1600 km in an ENE–WSW direction (Fig. 14a and b) and is thought to have originated in Archaean times. From a geophysical perspective, it divides India into southern and northern domains. The northern domains tend to have lower elevations and positive Bouguer gravity anomalies compared to the southern domain (Verma and Banerjee, 1992).

8.6. Sausar Mobile Belt

The Sausar Mobile Belt (SS; Fig. 14b) is generally accepted to be part of the CIS, although it is widely debated whether or not a suture is present thorough the entire length of the CITZ (e.g., Yedekar et al., 1990; Jain et al., 1991; Mishra et al., 2000; Roy and Prasad, 2001; Rao and Reddy, 2002). Rocks within the Sausar Belt record a protracted history of convergent margin activities from ca. 1400 Ma to ca. 800 Ma (Roy et al., 2006). The Sausar meta-sedimentary rocks are metamorphosed to upper amphibolite to granulite facies and have undergone migmatization. The CIS contains granulite lenses of ~0.5–0.7 km length and 0.2–2.0 km width within the Tirodi gneisses (Jain et al., 1991).

8.7. Eastern Ghats Mobile Belt

The Eastern Ghats Mobile Belt (EGMB) is a Proterozoic granulite belt that extends for ~1000 km from the Brahmani River in the north to Ongole in the south (Figs. 1 and 14a). The extent of the southern and northern margins of the EGMB is not well constrained and thus multiple interpretations have been proposed for the relationship to major orogenic belts to the north and south (e.g., Radhakrishna and Naqvi, 1986; Mukhopadhyay, 1987; Radhakrishna, 1989). The Eastern Ghats terrane comprises metapelite and enderbitic gneisses and enderbitic and charnockitic intrusions, two-pyroxene mafic and calc-silicate granulites (Stein et al., 2004) and contains two Phanerozoic rift valleys, the Mahanadi Rift in the north and the Godavari Rift in the south (Biswal et al., 2007).

Ramakrishnan et al. (1998) divided the EGMB into five lithotectonic units consisting of the Transition zone (TZ), the Western Charnockite zone (WCZ), the Western Khondalite zone (WKZ), the Eastern Khondalite zone (EKZ), and the Central Migmatitic zone (CMZ). The TZ consists of a mixture of lithologic units belonging to both cratonic India and the EGMB. The TZ comprises charnockites (quartz–feldspar–orthopyroxene), enderbites (quartz–plagioclase–orthopyroxene), mafic granulites (plagioclase–clinopyroxene–orthopyroxene–garnet) and banded iron formations. The WCZ and the EKZ consist of khondalites (garnet–sillimanite–graphite gneisses) intercalated with quartzite, calc-granulites (diopside–-garnet–plagioclase), and high Mg–Al granulites (sapphirine–cordierite–spinel–orthopyroxene). The CMZ is composed of migmatitic gneisses with intrusions of charnockite–enderbite, granite and anorthosite (Nanda and Pati, 1989).

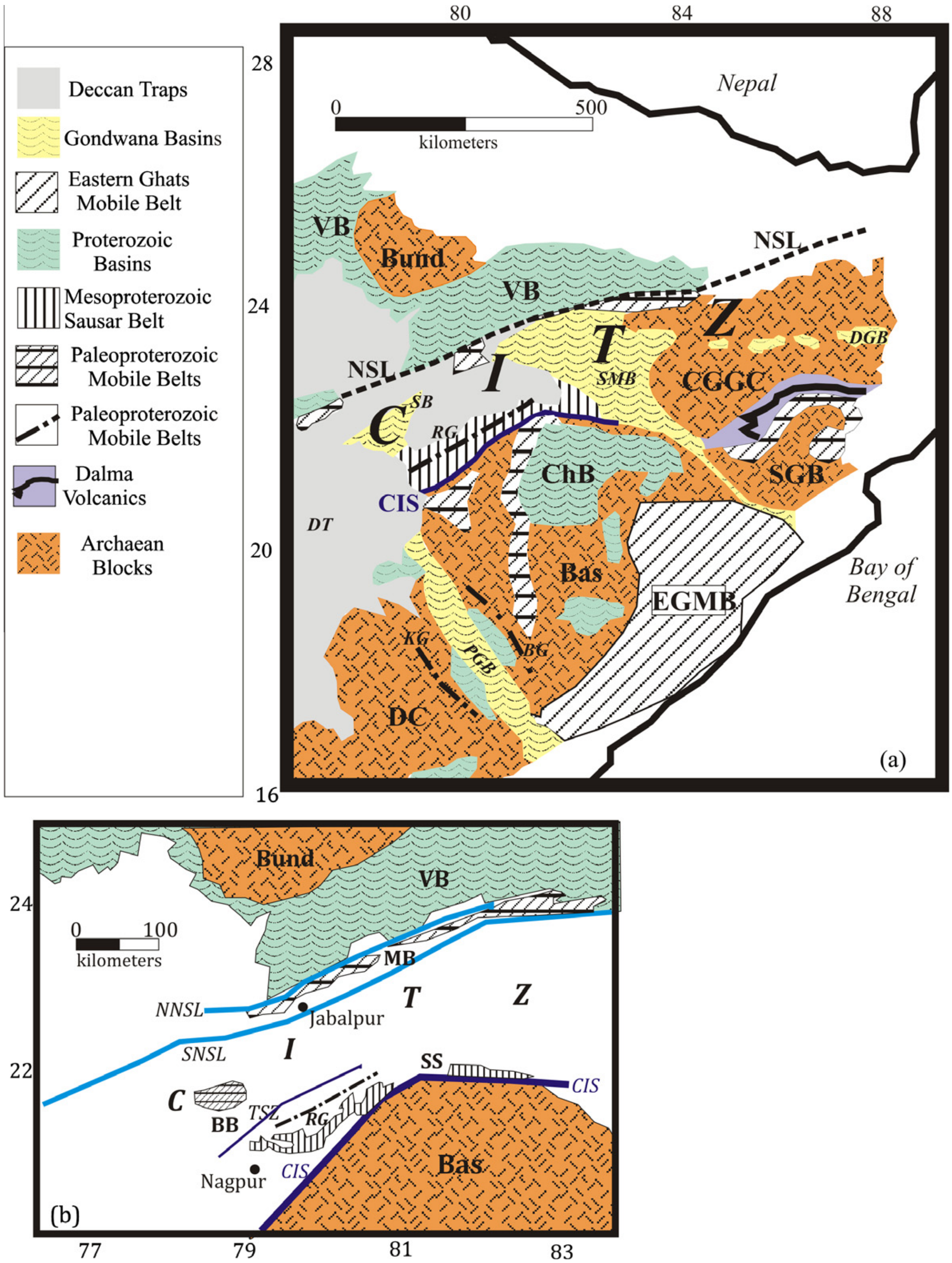

Fig. 14. Mobile Belts and tectonic elements of the Central Indian Tectonic Zone (CITZ; after Acharyya, 2003). Abbreviations: BG = Bhopalpatnam Granulite belt; CGGC = Chhotanagpur Granite Gneissic Complex; DGB = Damodar Gondwana Basin; DC = Dharwar Craton; DT = Deccan Traps; EGMB = Eastern Ghats Mobile Belt; KG = Karimnagar Granulite Belt; PGB = Pranhita Gondwana Basin; SMB = Son-Mahandi Gondwana Basin; SB = Satpura Gondwana Basin' SGB = Singhbhum Craton. (b) Expanded view of the central CITZ (after Acharyya, 2003) showing the BB = Betul Belt; MB; Mahakoshal Belt; NNSL = North Narmada-Son Lineament (fault); SNSL = South Narmada-Son Lineament (fault); SS = Sausar Supracrustals; TSZ = Tan Shear Zone.

The EGMB exhibits northeasterly structural trends attributed to early coaxial folding along a NE–SW axis (Murthy et al., 1971; Biswal et al., 1998). Dome and basin structures in conjunction with sheath folds of various sizes are readily observed in the EGMB (Natarajan and Nanda, 1981; Biswal et al., 1998). The mobile belt is characterized by several ductile and brittle-ductile shear zones, of which the Terrane Boundary Shear Zone (TBSZ) is the most prominent. The TBSZ delineates the tectonic boundary between the EGMB and the surrounding cratons (Biswal et al., 2000).

Three distinct metamorphic events occurred in the EGMB, the first of which is the UHT granulite facies metamorphism represented by sapphirine–spinel–orthopyroxene–garnet–quartz-assemblages in enclaves of khondalite. Gneissic fabric in the rocks was produced by metamorphic differentiation and partial melting during the early phases of folding and granulite facies metamorphism. P–T conditions for the first metamorphic event have been reported at 8–12 kbar and 1000–1100 °C (Lal et al., 1987; Kamineni and Rao, 1988; Rickers et al., 2001). The UHT metamorphism is followed by the second metamorphic event that represents granulite facies metamorphism at 8.0–8.5 kbar and 850 °C (Dasgupta et al., 1992). The third metamorphic event is characterized by retrograde amphibolite facies where P–T conditions are reported at 5 kbar and 600 °C (Dasgupta et al., 1994). Sm–Nd, Rb–Sr, and Pb–Pb data indicate that the rocks of the EGMB have crustal residence ages of 2.5–3.9 Ga. Rickers et al. (2001) interpret this to represent variable mixing of Archaean and Proterozoic crustal material within an active continental setting.

Mezger and Cosca (1999) provide a revised tectonothermal history of the EGMB, based on U–Pb zircon and monazite in addition to $^{40}Ar/^{39}Ar$ hornblende data, wherein the Western Charnockite Zone's tectonic history is distinct from that of the other units within the EGMB. The other regions (WCZ, EKZ and CMZ) were undergoing granulite facies metamorphism at ca. 960 Ma, while the WCZ had already cooled below the Ar diffusion temperature for hornblende (<450 °C). These results imply a major discontinuity between these regions of the EGMB. The updated mineral ages of Mezger and Cosca (1999) indicate that at least one of the high-grade metamorphic events preserved in the EGMB occurred during the late stage of the global 'Grenvillian' orogeny ca. 960 Ma. This Late Grenvillian orogenic episode is also recorded in the Rayner Complex of Antarctica (Paul et al., 1990; Shaw et al., 1997).

Mezger and Cosca (1999) also report the central units of the EGMB to have a major thermal overprint at ca. 500–550 Ma during the 'Pan-African' orogeny. A hornblende $^{40}Ar/^{39}Ar$ age of ca. 1100 Ma from an amphibolite in the Godavari Rift indicates that the Pan-African thermal event was weaker, if at all present, in the WCZ. Granulite facies metamorphism of 500–550 Ma age also occurs in Sri Lanka and Madagascar (Kröner and Williams, 1993; Kriegsman, 1995).

Most recently Chatterjee et al. (2008) reported 983 ± 2.5 Ma ages for Chilka Lake anorthosite. They correlated the emplacement of this anorthosite with charnockitic activity in the presumably adjacent Rayner province of East Antarctica. Dating of monazite cores and rims in the same region yielded ages of 714 ± 11 and 655 ± 12 Ma respectively. These ages were considered to reflect metamorphism during an oblique collision between India and NW Australia although new data from Gregory et al. (2009) and van Lente et al. (2009) argue against such a scenario until ~550 Ma.

9. Summary

The assembly of Peninsular India began with the cratonization process of the individual nuclei discussed above. The geochronologic constraints on this process need improvement, but the current data are consistent with amalgamation of Peninsular India by the end of the Archaean that nearly all of the nuclei had stabilized at about 2.5–2.6 Ga. Furthermore, it is also suggested that this same time interval is coincident with the welding together of most

Table 1
Summary of the Precambrian History of India.

Craton	Oldest age	Stabilization age	Metamorphic events	Igneous events	Sedimentary basins
Aravalli	~3.5 Ga	~2.5 Ga	~2.0 Ga (t) 1.7–1.6 Ga (t) 950–940 Ma (s) 990–836 Ma (t)	1711–1660 Ma (g) 820–750 Ma (g, d)	Marwar (<635 Ma)
Bundelkhand	~3.3 Ga	~2.5 Ga	3.3 Ga (t) 2.7 Ga (t) 2.5 Ga (t)	2.15 Ga (d) 2.00 Ga (d) 1.1 Ga (k)	Vindhyan 1.8–1.0 Ga
Singhbhum	~3.5–3.8 Ga	~2.5 Ga		3.3 Ga (g) 3.1 Ga (g) 3.5 Ga (v) 2.1 Ga (d) 1.5 Ga (d) 1.1 Ga (d) 900 Ma (v)	Dhanjori (~2.5 Ga) Kolhan (~1.1 Ga)
Bastar	~3.5 Ga	~2.5 Ga	2.5 Ga (t) 2.3 Ga (t) 1.1 Ga (t)	2.5 Ga (v) 1.9 Ga (d)	Chhattisgarh (~1.1 Ga) Indravati (1.1 Ga ?)
E. Dharwar	~2.7 Ga	~2.5 Ga	2.7–2.5 Ga (t)	2.5 Ga (g) 2.4 Ga (d) 2.2 Ga (d) 1.9 Ga (d) 1.2 Ga (d) 1.1 Ga (k, l) 1.0 Ga (d)	Cuddapah (~1.8–1.0 Ga) Kurnool (< 600 Ma?) Pranhita–Godavari (1.1 Ga) Bhima (1.1–0.6 Ga ?)
W. Dharwar	~3.6 Ga	~2.5 Ga	3.3 Ga (t) 3.1 Ga (t)	3.35 Ga (v) 2.6 Ga (g) 2.5 Ga (g) ?(d)	Kaladgi (1.1–0.6 Ga ?)
S. Granulite	~3.5 Ga	~2.5 Ga	~2.5 Ga (t) 800–900 Ma (t, s) 500–600 Ma (t, s)	500 Ma (g)	None

Abbreviations: (t) = tectonothermal; (s) shear; (d) dyke intrusion (k) kimberlite/lamproite intrusion (g) granitic and/or mafic intrusions; (v) volcanism.

of these cratons to form 'proto-India' (with the exception of some blocks in the southern granulite province). In our view, proto-India consisted of the Aravalli–Bundelkhand, Eastern Dharwar, Western Dharwar, Singhbhum and Bastar Cratons. We do note that controversies regarding the exact nature/age of tectonic events in the CITZ may indicate that 'proto-India' was not fully formed until the Mesoproterozoic (see above).

Major Proterozoic basin formation (the so-called "Purana" basins) took place during three main pulses. These include a Paleo- to Mesoproterozoic phase of basin formation (Aravalli – Delhi, Lower Vindhyan, Lower Chhattisgarh and Cuddapah Basins); an early Neoproterozoic phase of basin formation (Upper Vindhyan, Upper Chhattisgarh Basins) and a Late Neoproterozoic phase of basin formation (Marwar basin and the Kurnool Group of the Cuddapah Basin). Other basins in India have only poor age constraints though many (Bhima, Kaladgi and Indravati) are traditionally thought to be Late Neoproterozoic in age.

Major pulses of mafic dyke intrusions are also widespread in Peninsular India (see Table 1). Age constraints on many of the dykes are only poorly known, but several Paleoproterozoic and Mesoproterozoic pulses are now dated using U–Pb zircon and baddeleyite (see discussions above). Ultramafic intrusions in India are concentrated at about 1.0–1.1 Ga and may be part of a global ultramafic event.

Peninsular India is thought to have been part of five different supercontinental configurations. The oldest of these "expanded-Ur" (~3.0 Ga; Rogers, 1996) is composed of the Dharwar and Singhbhum Cratons in India, the Kaapvaal Craton of South Africa and the Pilbara and Yilgarn Cratons of Australia along with small blocks of Archaean cratonic material in East Antarctica (Fig. 15a). Tests of the "Ur" configuration are problematic as paleomagnetic data are mostly lacking and where data do exist (for example in the Kaapvaal and Pilbara Cratons), the proposed configuration of "Ur" does not hold although it is possible that a 'mega-craton' of Vaalbara existed (see Wingate, 1998; Zeger et al., 1998 for more complete discussion).

Depending on the exact model chosen, India is also placed adjacent to coastal East Antarctica, Madagascar, North China, Kalahari and Australia in a modified Gondwana fit in the "Columbia" supercontinent (Fig. 15b; Zhao et al., 1994; Rogers and Santosh, 2002).

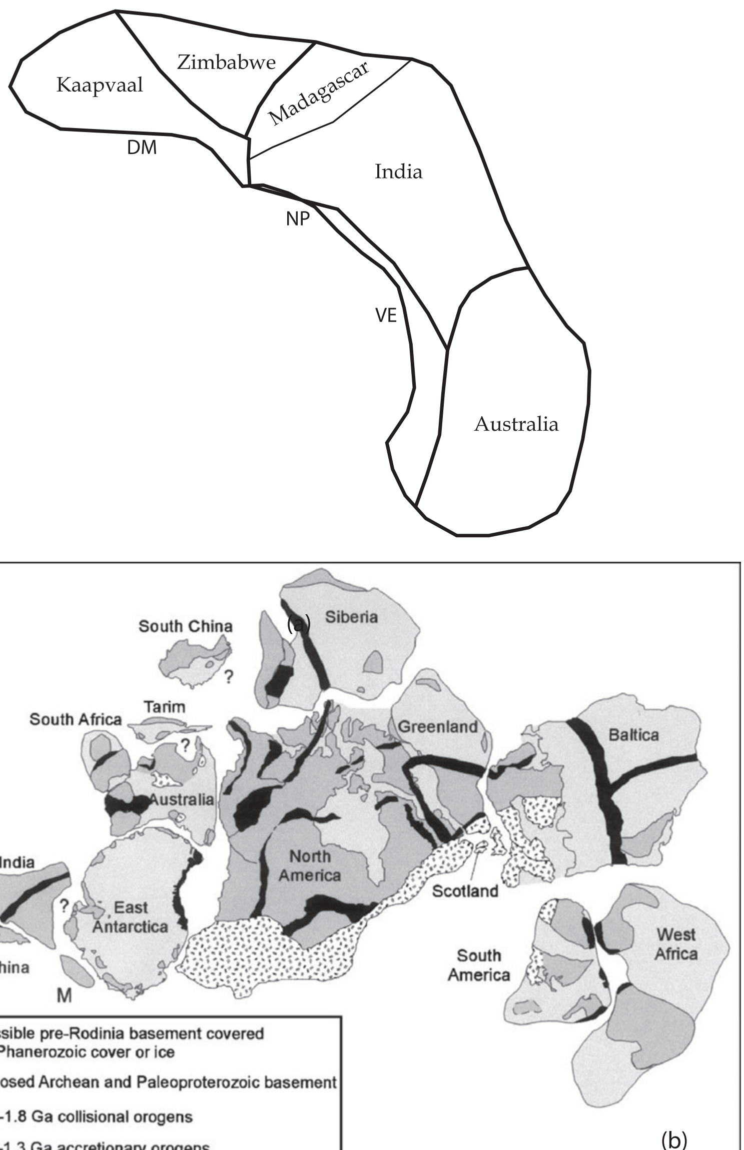

Fig. 15. (a) Expanded "Ur" after Rogers and Santosh (2002); DM = Dronning Maud land; NP = Napier Complex; VE = Vestfold Hills; (b) Paleoproterozoic supercontinent Columbia after Zhao et al. (2004) numbered orogenic belts can be found in Zhao et al. (2004).

Paleomagnetic tests of the Columbia supercontinent are also problematic although the model itself is based mainly on a preponderance of 2.1–1.8 Ga orogenic belts across the globe (Zhao et al., 1994; Meert, 2002; Pesonen et al., 2003).

The exact makeup of the Late Mesoproterozoic–Neoproterozoic Supercontinent of Rodinia is fluid (see Li et al., 2008 for a full review). In the 'archetypal' reconstructions of Rodinia, India is placed in a configuration nearly identical to its position within East

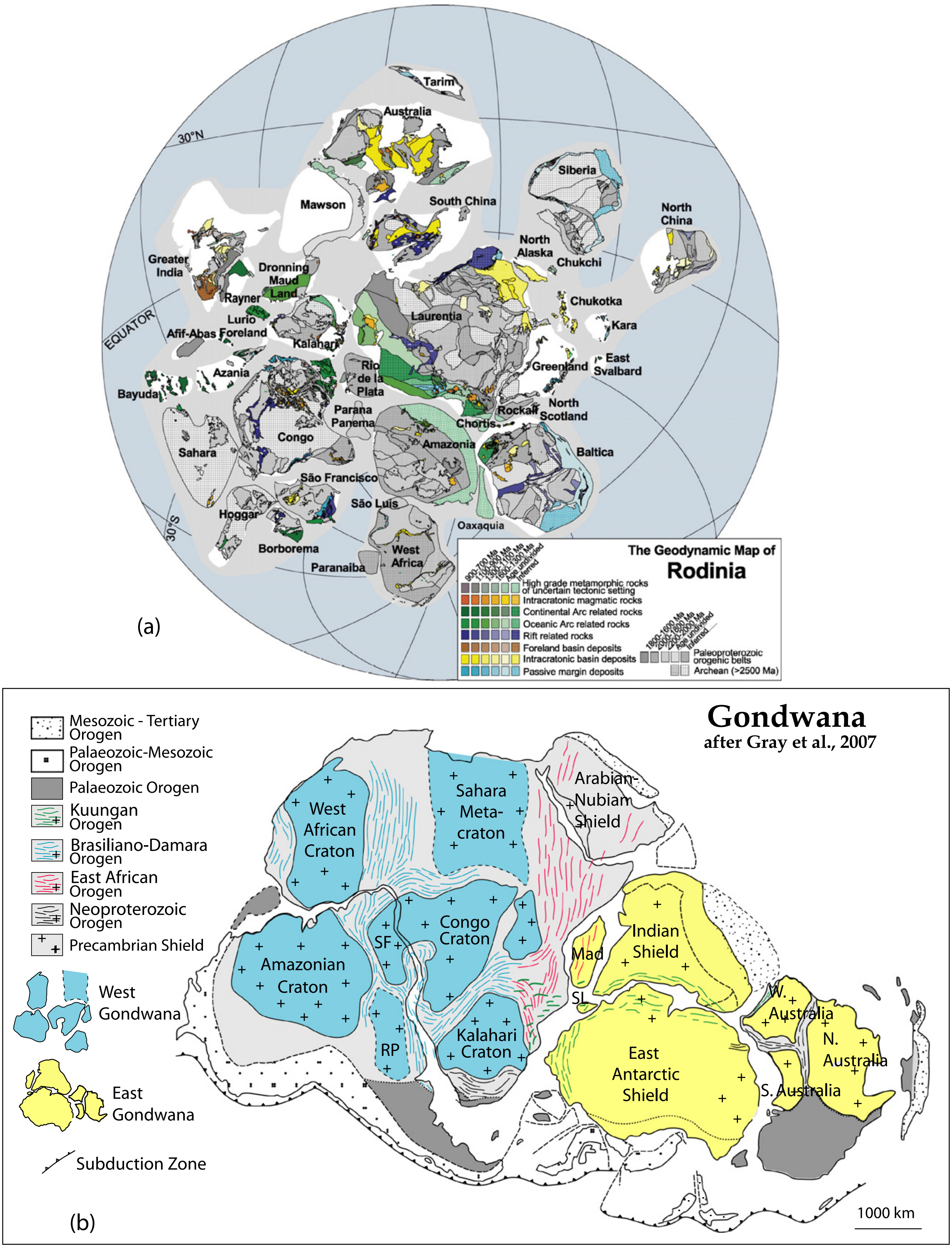

Fig. 16. (a) Rodinia after Li et al. (2008); (b) Gondwana after Gray et al. (2008) and Meert and Lieberman (2008).

Gondwana (Fig. 16a). Paleomagnetic data supporting such a position for India is non-existent and, in fact, the extant paleomagnetic data argue against such a configuration (Meert, 2003; Gregory et al., 2009; Torsvik et al., 2001a,b; Meert and Lieberman, 2004, 2008). There is also a considerable body of evidence indicating a polyphase assembly of East Gondwana during the Cambrian (see Meert, 2003 for a summary) and also negating the 'archetypal' Rodinia reconstructions (see Meert and Torsvik, 2003). Alternative views to the piecemeal assembly of eastern Gondwana can be found in Veevers (2004) andSquire et al. (2006).

By the end of the Cambrian, India was part of the large southern continent of Gondwana (Fig. 16b) and remained a part of the Gondwana supercontinent until its breakup in Mesozoic times.

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