Abstract: Abstract

The Precambrian through Mesozoic geologic history of peninsular India covers nearly 3.5 billion years of time. India is presently attached to the Eurasian continent although it remains (for now) a separate plate. It is comprised of 5 major cratonic nuclei known as the 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, Prahnita-Godavari, Indravati, Bhima-Kaladgi, Kurnool and Marwar basins are the major Meso to Neoproterozoic sedimentary repositories on the subcontinent. The Paleozoic-Mesozoic age Gondwana sequence covers much of central India. Major episodes of 'trap' volcanism occurred during the Mesozoic with the eruption of the Rajmahal and Deccan traps.
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.
Tectonic Evolution of Peninsular India: A 3.5 Billion Year Odyssey

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Introduction

The Indian subcontinent covers approximately 5,000,000 square kilometers. Although India is connected geographically to the Eurasian continent, the subcontinent and Himalayan sectors make up a distinct lithospheric plate. Our goal in this paper is to describe the Precambrian through Mesozoic geology of Peninsular India. Such an effort would require a tome-length treatise (e.g. Naqvi and Rogers, 1987; Naqvi, 2005; Balasubrahmanyan, 2006; Ramakrishnan and Vaidyanadhan, 2008) and we apologize at the outset for any oversimplifications and omissions.

In this paper we summarize the geologic history of the subcontinent from oldest (the cratons) to youngest (Deccan Traps). We cover 3.5 billion years of geologic history and hope to give the reader an overview of this important continental block. Elsewhere in this volume, are accounts of the Cenozoic history of the Himalayan region (reference) and the surrounding Indian Ocean (reference). 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; figure #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. We use ‘stabilization’ in the same manner as Rogers an Santosh (2003) and consider a craton stabilized when it is intruded by undeformed plutons, closure of whole rock isotopic systems and the deposition of platform sediments on the newly formed basement. The Paleozoic and Mesozoic histories of the Indian sub-continent are described and limited to Gondwana sedimentation, Mesozoic sedimentation in the subcontinent and the eruptive histories of the Rajmahal and Deccan traps. 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.

**Aravalli and Bundelkhand Cratons**

The Aravalli-Bundelkhand protocontinent (figure 2) occupies the north-
central region of the Indian sub-continent. The Great Boundary Fault (GBF) divides
the protocontinent into two blocks: the Aravalli cratonic block to the west of the
GBF and the Bundelkhand-Gwalior block to the east of the GBF. These cratons are
bounded to the north-east by the Mesoproterozoic-aged Vindhyan basin and the
Indo-Gangetic 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 (to the west) by the Narmada Son lineament (Goodwin, 1991; Naqvi and Rogers, 1987).

Most of the Aravalli craton is underlain by the 3.5 Ga Banded Gneiss Complex
(hereafter BGC; Naqvi and Rogers, 1987). The BGC is composed of charnockites,
migmatites, gneisses, pelites and metasedimentary rocks. The BGC in the Aravalli
region is bounded by the Aravalli and Delhi fold Belts. Age constraints on both the
Aravalli and Delhi Belts are only poorly constrained to 2.5 - 1.9 Ga and 1.8 - 0.85 Ga
respectively (Kaur et al., 2007).

The age of metamorphism in the Aravalli craton are better constrained. Roy
et al. (2005) argue that the main pulse of metamorphism took place between 1725 -
1621 Ma at the outset of the Delhi Orogenic Cycle. Buick et al. (2006) also obtained
metamorphic ages of ~1720 Ma for part of the Aravalli Belt that was formerly
thought to be part of the BGC. Lastly, Kaur et al. (2007) determined two distinct
ages of granitoid intrusions in the Aravalli Range. The best constrained of these
magmatic events took place between 1711 - 1660 Ma supported by the $^{207}\text{Pb}/^{206}\text{Pb}$
and U/Pb isotopic data and appears to be coincident with the main phase of
metamorphism documented by both Buick et al. (2006) and Roy et al. (2005). The
ion microprobe $^{207}\text{Pb}/^{206}\text{Pb}$ zircon studies by Wiedenbeck et al. (1996), strongly
suggest a ~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
The younger Aravalli Supergroup sediments were unconformably deposited on this stabilized landmass.

Buick et al. (2006) also report a younger metamorphic (shear) episode that took place between ~950 - 940 Ma based on their $^{235}$U/$^{206}$Pb isotopic ratios on the metasedimentary rocks of the Mangalwar complex and suggested that these ages may be part of a larger metamorphic and igneous event that occurred during the 990 - 940 Ma interval. Support for this metamorphic event is also observed in detrital zircon spectra from the Sonia and Girbakhar sandstones in the Marwar Supergroup of Rajasthan (Malone et al., 2008). Subsequent tectono-metamorphic events in the Aravalli region are described in Deb et al. (2001) and Roy (2001).

These include a metallogenic event at around 990 Ma followed by a tectonothermal event between 990 - 836 Ma. This was followed by the onset of Malani felsic intrusive and extrusive igneous activity (820 - 750 Ma; Deb et al., 2001; Torsvik et al., 2001a,b Gregory et al., 2008; Van Lente et al., 2008). The Malani felsic province is overlain by the Neoproterozoic Marwar Supergroup.

The Bundelkhand craton is a less well studied region to the east of the Aravalli-Delhi fold belt (Figure 1). 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 of Soni and Jain (2001); (2) Undeformed multiphase granitoid plutons and associated quartz reefs – the Granite Suite of Soni and Jain (2001); and (3) Mafic dyke swarms and other intrusions – the Intrusive suite of 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). Three generations of gneisses are thought to have formed at 3.2 Ga, 2.7 Ga and 2.5 Ga respectively revealed from the $^{207}$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 (Tucker, personal communication). 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. The paleomagnetic directions on the other hand demonstrate at least three generations of dykes in the BGM (Pradhan et al., under review) 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 ages, but neither the ages nor the geochemistry are robust enough to confirm such a connection.

Overlying these rocks are the 1854 ± 7 Ma Hindoli Group (equivalent to the Gwalior Group) and the sedimentary sequences in the Vindhyan Basin (~1700 - 1000 Ma; Ray et al., 2002; Rasmussen et al., 2002; Sarangi et al., 2004; Ray et al., 2003; Malone et al., 2008). The Upper part of the Vindhyan basinal sequence is intruded by the 1073 ± 13.7 Ma Majhgawan kimberlite (Gregory et al., 2006). The Marwar basin, developed west to the Aravalli Range in northwestern Rajasthan, lies unconformably over the Malani Igneous Suite (MIS) and had been assigned an age of Neoproterozoic to Early Cambrian (Pandit et al., 2001; Mazumdar et al., 2006; Kumar and Pandey, 2008).

**Intrusive events in Aravalli-Bundelkhand cratons**

Two high grade Paleo-Mesoproterozoic tectono-thermal events in the Aravalli – Bundelkhand Craton took place around 2.5 Ga and 2.0 Ga. These events are recorded by the emplacement of several mafic intrusions. The oldest known mafic representatives intruding the greenstone sequence of Precambrian Aravalli crust are Mavli amphibolites having a Sm-Nd isochron age of 2828 ± 46 Ma (Gopalan et al., 1990; Upadhyaya et al., 1992).

A second suite of dykes that intrude the BGC show a chemical affinity to the island arc tholeiites and are thought to have formed during the ensialic opening of Aravalli basin.
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); dykes intruding the nepheline syenite pluton of Kishangarh; 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.

The MIS unconformably overlies Paleo- to Mesoproterozoic metasediments, and basement granite gneisses and granodiorites (Pandit et al., 1999); and is unconformably overlain by the latest Neoproterozoic to Cambrian Marwar Supergroup made up of red-bed and evaporite sedimentary sequences (Pandit et al., 2001; Torsvik et al., 2001a).

The Malani Igneous Suite (MIS) to the west of Aravalli Mountains form the largest felsic volcanic province of India (Figure 2). is the MIS is characterized by voluminous magmatism that occurred in three phases. The first phase commenced with basaltic and felsic flows, granitic intrusions dominate the second phase of igneous activity within the MIS. Predominately felsic and minor mafic dike swarms mark the third and final phase of the MIS cycle. The MIS is often explained as 'anorogenic magmatism' related either to crustal melting during extension or to an active hot spot (Bhushan, 2000; Sharma, 2004). Alternatively MIS magmatism can be interpreted in the context of an Andean-type active margin (Torsvik et al., 2001a; Torsvik et al., 2001b, Ashwal et al., 2002; Gregory et al., 2008). In their reconstructions, Gregory et al. (2008), relate the MIS activity to the nearby arc activity observed in the Seychelles islands and northeastern Madagascar.

Previously reported ages for the Malani magmatism ranges from 680-780 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. (2008) provide a more 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. (in press) provide U-Pb ages for the Sindreth felsic volcanics of $767 \pm 2.9$ Ma, $765.9 \pm 1.6$ Ma and $761 \pm 16$ Ma. The Sindreth felsic rocks in Rajasthan were thought to be correlative to the Malani felsic units and these new ages seem to cement their relationship. Van Lente et al. (in press) also report ages of $800 \pm 2$ Ma and $873 \pm 3$ Ma for tonalitic basement rocks in the region. Pradhan et al (in review) report an age of $827 \pm 8.8$ Ma for the Harsani granodiorite that also lies beneath the MIS. These data suggest that Malani volcanism was preceded by a protracted interval of granitic intrusion ranging from $\sim 875-800$ Ma.

The Bundelkhand craton in the Central Indian shield is also characterized by various extrusive and intrusive events during the Proterozoic. 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, Supa to the NE, Khajuraho on the SE and Tikamgarh to the SW (Figure 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) $1480 \pm 35$ to $1660 \pm 40$ Ma, (2) $1790 \pm 40$ to $1850 \pm 35$ Ma and (3) $1930 \pm 40$ to $2010 \pm 80$ Ma (after V.Finko cf Pati et al., 1997). The broad age ranges reported here testify to the need for more robust dating of these intrusive events.

There are numerous mafic dykes and dyke swarms intruding the BGM. A vast majority of these dykes trend NW-SE, although a few dykes including the ‘Great Dyke of Mahoba’ strike in a NE-SW direction (Fig. 1). Rao (2004) suggests that most of the 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. Paleomagnetic data suggest that at least three generations of dykes intrude the BGM (Pradhan et al., under review).

Sedimentary Basins
There are two large intracratonic basins of Mesoproterozoic – Paleozoic age that outcrop between the Aravalli and Bundelkhand Province of the North Indian shield: (1) The Vindhyan basin located in central peninsular India, and (2) The Marwar Supergroup in the western part of the Aravalli mountain range in Rajasthan (Fig. 2)

**Vindhyan Basin**

The Vindhyan basin is one of several “Purana” (ancient) sedimentary basins of the Indian subcontinent. The Vindhyan is a sickle-shape basin that outcrops between the Archaean Aravalli-Bundelkhand province to the north and east and the Cretaceous Deccan traps to the south and by the Great Boundary Fault to the west ((Mazumdar et al., 2000; Figure 1). 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 (Figure 4).

The trondhemitic 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 supports 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. (in press) looked at the geochemistry of the basal volcanic sequence (Khaimalia 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 tectonic zone (CITZ).

Stratigraphically, Vindhyan basin can be divided into two sequences: The Lower Vindhyan Sequence formed by Semri Group and the Upper Vindhyan Sequence sub-divided into the Kaimur, Rewa and Bhandar Groups, respectively (Chaudhari et al., 1999; Malone et al., 2008).

Age control on Vindhyan sedimentation is still the subject of considerable controversy as are the ages of the other Purana basins (Patranabis-Deb et al., 2007;
Basu et al., 2008; Azmi et al., 2008; Gregory et al., 2006; 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 the Semri Group. The Semri Group is made up of five alternating formations of shale and carbonates with areas of sandstones and volcanioclastic 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). Ages from the Semri Group 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 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 (Ray et al., 2003; Sarangi et al., 2004; Rasmussen et al., 2002a,b). Most authors (see Azmi et al., 2008 for an alternative view) agree with a Mesoproterozoic age for Lower Vindhyan sedimentation (~1750-1500 Ma). A basin wide unconformity between Rohtas limestone of the Semri Group and the Kaimur Group of the Upper Vindhyan separates those two sequences.

The Semri Group is separated from the Upper Vindhyan by a basin wide unconformity between the Rohtas limestone and the overlaying Kaimur Group. The Kaimur is intruded by the 1073 +/- 13.7 Ma Majhgawan kimberlite (Gregory et al, 2006), that cross-cuts both the Semri and Kaimur Groups and is currently exposed in the Kaimur Group (Baghain). 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 Ediacara fossils have been described in the Lakheri and Sirbu formations of the Bhandar Group and could indicate an age <635 Ma for the Bhandar (De, 2003; De, 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., in review). 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 Bhandar and Rewa Groups in the Rajasthan sector along with samples from the Lower Marwar Supergroup in Rajasthan. In that study, Malone et al. note that the youngest population of zircons from the Upper Bhandar is older than 1000 Ma. That observation, coupled with the similarity in paleomagnetic directions from the Upper Vindhyan and Majhgawan kimberlite led Malone et al. (2008) to conclude that Upper Vindhyan sedimentation was completed by ~1000 Ma a result consistent with recent data from another of the Purana basins to the south (Patranabis-Deb et al., 2007).

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 outcrops in western Rajasthan and the basin trends to the NNE–SSW with a slight westerly tilt. The type section of the Marwar Supergroup is represented by undeformed to mildly folded sediments up to 2 km in thickness (Roy, 2001). Lithostratigraphically, Marwar Supergroup overlies the basement rocks of Malani Igneous Suite and Delhi Supergroup and 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 (Raghav et al., 2005; Khan, 1973; Pareek, 1981). 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 (Vaidyanadhan. & Ramakrishnan. 2008; Chakrabarti et al, 2004). The nature of the Pokaran boulder bed is the subject of some debate, but given the recent age estimates for the Lower Marwar, it may represent a glacial deposit formed during either the Gaskiers (c. 580 Ma) or Marinoan (c. 635 Ma) glaciation.

The middle section contains evaporitic (boreholes only) and carbonate facies that are capped by sandstones. The age of the Marwar is considered to be of Ediacaran-Cambrian (~635-515 Ma; Naqvi and Rogers, 1987; Pandit et al., 2001) but there are no robust radiometric ages. 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 the earlier estimates.

Recently, the $^{238}$U/$^{206}$Pb isotopic analyses on the detrital zircon grains extracted from the Sonia and Girbakhar sandstone member 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; Pradhan et al., in review; Van Lente et al., in press).

**Singhbum Craton**

The Singhbum Craton (also called the Singhbhum-Orissa craton; Figure 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 comprised of three major geologic provinces: (1) a nucleus at the southern edge known as the Singhbhum nucleus; (2) the Singhbhum-Dhalbhum mobile belt; and (3) the Chhotanagpur terrain of gneisses and granite (Naqvi & Rogers, 1987).

**Older Metamorphic Group—Singhbhum Nucleus**

The Singhbhum nucleus is composed mainly of Archaean granitoid batholiths, including the Singhbhum Granite Complex. The batholiths are surrounded by several 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 orthoamphibolites (Naqvi & Rogers, 1987). Tonalite-trondhjemite gneisses (TTG’s) are found along the contacts with some of the amphibolites (Fig xx). 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 & Rogers, 1987). The most comprehensive attempt to date the OMG is described by Misra et al. (1999). Detrital zircons from the OMG yielded age ranges between 3.5-3.6 Ga and other zircons have peaks at 3.4 and 3.2 Ga. Misra et al. 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 discussion in this review paper.

**Singhbhum Granite**

The OMG is intruded by the approximately 10,000 km$^2$ Singhbhum Granite complex. The complex includes 12 domal magmatic bodies that are independent of one another. Whether these bodies were emplaced in one magmatic event or several remains a subject of debate; however recently acquired data suggests polyphase emplacement (Naqvi & Rogers, 1987).
The Singhbhum Granite 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). The other variety 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 3328 ± 7 Ma age for the Signhbhum 'phase II' granites and an 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. They report a discordia upper intercept for the Singhbhum granite of 3302 ± 14 Ma and a more robust 207Pb-206Pb age for the most concordant zircons of 3288 ± 8 Ma. The Sushin nepheline syenite body yielded the youngest ages from the SG complex of 922.4 ± 10.4 Ma (Reddy et al., 2008). The geochronological data from the Singhbhum granites therefore favors a 3-stage emplacement. The oldest intrusions of granites at ~3.3 Ga followed by a secondary emplacement at 3.1 Ga and the youngest intrusive event at ~0.9 Ga. Acharyya et al. (2008) report U-Pb ages from an 'earliest' phase of Singhbhum granite intrusion. 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).

Iron Ore Group

The cratonisation of the Singhbhum region appears to have occurred contemporaneously with 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 two sections, an Older and a Younger, with similar compositions but differing ages.

The Older IOG is comprised of clastic sedimentary rocks, proposed by Eriksson et al. (2006) to have a shallow marine depositional source, and of syndepositional volcanic rocks which imply large scale rifting (Eriksson et al., 2006). The Older IOG formed prior to 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 (Mondal et al., 2007; Eriksson et al., 2006). A recent geochronologic study of the IOG (Mukhopadhyay et al., 2008) yielded an age for a dacitic lava of 3506.8 ± 2.3 Ma and confirms that the IOG formed prior to the 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 cratonisation event and has a suggested depositional age >2.55 and <3.0 Ga. It is comprised of shallow or shelfal marine greenstone deposits with BIF (Eriksson et al., 2006).

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. The volcanic complex displays shows 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 & Rogers, 1987). This age appears to be too young as the Simlipal basinal rocks (including the Simlipal Complex) is intruded by the ~3.1 Ga Mayubhanj granite.

Mafic Dyke Swarms

Newer Dolerites

Dyke swarms found within the Singhbhum granitic pluton vary in rock type from mafic to intermediate. The largest suite of dykes are the so-called “Newer Dolerites” (Bose, 2008).

Nearly every group of rocks in the Singhbhum Nucleus has been 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 & Rogers, 1987; Bose, 2008; Srivastava et al., 2000). 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.

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 consist of a variety of textures including fine, medium and coarse grained specimens, though the majority show medium grain size. The main minerals seen in the slightly metamorphosed dolerites are augite and labradorite (Verma et al., 1974).

**Chotanagpur Terrain**

The Proterozoic Chotanagpur Granite Gneiss Complex (CGGC) 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). The 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 & Rogers, 1987).

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

The features displaying high-grade metamorphism characteristically contain intrusive pegmatites that display both concordant and discordant foliation. The pegmatites are thought to have occurred over a long time span and display indications of each of the different deformation episodes known 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 & Dey, 1942).

**Proterozoic Sedimentation**

**Paleo-Mesoproterozoic History**

Proterozoic sedimentation and volcanism found in the Singhbhum Craton is divided into several formations (Figure 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 (Eriksson et al., 2006; Mazumder, 2005). 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 (Eriksson et al., 2006; Mazumder, 2005). Both massive and schistose, amygdaloidal basalts are present in the formation. These low-Al tholeiites have been folded into a first order syncline (Naqvi & Rogers, 1987). Based on field and preliminary geochronologic results, Acharyya et al. (2008) propose a late-Archean to Early Paleoproterozoic age for the Dhanjori (~2.5 Ga).

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 a layer of volcanic tuff. Ultramafic-mafic intrusions and thin basaltic to komatiititic lava flows occur through the section (Eriksson et al., 2006). The depositional environment for the formation is thought to be largely fluvial and aeolian representing an overall terrestrial depositional history and 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 & 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 intracontinental rifting. A plume related origin also attracts many supporters, although no one theory has become generally accepted (Eriksson et al., 2006 and references therein).

The Chandil Formation exists as a belt of metamorphosed sedimentary rock and volcanic rocks that separate the Dalma sequence to the south from the Chotanagpur 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 (Eriksson et al., 2006; Mazumder, 2005). Whole rock Rb-Sr dating on a cluster of granites intruding the Chandil yielded an age of 1638 ± 38 Ma providing a poorly constrained younger limit for the Chandil Formation (Mazumder, 2005).

The Kolhan Group

In the southern part of the Singhbhum craton, there is a minor supracrustal suite known as the Kolan 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 subsequently deformed into a synclinal structure. Elongate domes and basins and dome-in-dome structures dominate the eastern part of the basin (Naqvi & 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
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 et al., 2004; Mukhopadhyay et al., 2006).

**Bastar Craton**

The Bastar Craton (Figure 7; also known as the Bhandara or Central Indian Craton) is bordered by the 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 & Rogers, 1987; Srivastava et al., 2004).

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

**Supracrustal Sequences**

The Bastar Craton contains three major supracrustal sequences of rocks (Figure 8), the Dongargarh, the Sakoli, and the Sauser suites. Of the three suites, only the Dongargarh Supergroup has been dated.

**Dongargarh Supergroup**

The Dongargarh Supergroup extends from the Chattisgarh 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. The group consists mainly of gneisses with secondary schists and quartzites (Naqvi & Rogers, 1987).
The Nandgaon Group contains two volcanic suites, the Bijli and Pitepani suites, are rhyolite dominated with secondary dacites, andesites, and basalts that show signs of fractionation (Neogi et al., 1996). The Bijli rhyolite has been 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 & Rogers, 1987). The Dongarhgarh 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 Namdgoan Group was developed ~2.5 Ga.

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 intertrappean shale, all conformably overlying each other (Naqvi & Rogers, 1987). It also contains two overlying volcanic suites, the Sitagota and Mangikhuta, that are dominated by more primitive tholeiitic basalts that are relatively unfractrated (Neogi et al., 1996). The volcanic suites are broken up by the Karutola sandstone (Naqvi & Rogers, 1987). The four volcanic suites within the Dongargarh Group erupted periodically between ca. 2462 and 1367 Ma (Neogi et al., 1996).

**Sakoli Group**

The Sakoli Group consists mainly of low grade metamorphic rocks of undetermined age in a large synclinorium. The Group is a significant volcano-sedimentary deposit comprised of (Youngest to oldest) slates and phyllites, bimodal volcanic suite and schists, metabasalts and cherts and conglomerates and Banded Iron Formations (BIF’s; Bandyopadhay 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 & Rogers, 1987). Unconformably overlying the Sakoli Group are the 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 $1295 \pm 40$ and $922 \pm 33$ Ma but the interpretation of these ages is fraught with difficulty (Bandyopadhyay et al., 1990).

**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 kilometers in length and 70 kilometers in width (Naqvi and Rogers, 1987). Detailed geochronologic studies are lacking within this belt; however, Roy et al. (2006) argue that the main phase of metamorphism (amphibolite-grade) took place between 800-900 Ma based on Rb-Sr and Sm-Nd geochronology. 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 $2672 \pm 54$ Ma. A mafic granulite within the southern belt yielded a Sm-Nd age of $1403 \pm 99$ Ma. The northern granulite yielded a Sm-Nd age of $1112 \pm 77$ 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 along the CITZ took place during the earliest Paleoproterozoic based on Re-Os ages from within the Sausar Belt (Malanjkhand granitoid batholith). They report ages of 2490 $\pm 2$ Ma for the granitoid that is nearly identical to U-Pb zircon ages of 2478 $\pm 9$ Ma and 2477 $\pm 10$ Ma (Ranigrahi et al., 2002). Mineralization ages associated with the intrusions ranged from 2446-2475 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 earliest Paleoproterozoic along the Sausar Belt (e.g. CITZ).

**Mafic Dyke Swarms**
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 (Subba Rao et al., 2008). The Bastar Craton is intruded by numerous mafic dyke swarms, spanning an area of at least 17,000 km$^2$, which 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 southern dykes trend NW-SE, paralleling the Godavari rift, and the 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 $\pm$ 0.1 Ma and 1883 $\pm$ 1.4 Ma and include boninite-norite and sub-alkaline mafic dykes, most of which display some degree of metamorphism (Srivastava, 2008; Srivastava et al., 2004; French et al., 2008). French et al. (2008) and Srivastava et al. (2008 and 2004) 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 the did not link the regions together paleogeographically and instead argued for a mantle upwelling on a global scale. 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). A different interpretation is presented for the Proterozoic dyke swarms. Hussain et al. (2008) postulate that these dykes were derived from subduction constituents that had been altered in the mantle lithosphere. A subduction related
genesis is theorized due to the increased incompatible lithophile elements seen in the geochemical analysis (Subba Rao et al., 2008).

Sedimentary Basins

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

Chattisgarh Basin

The 36,000 km² Chattisgarh Basin is comprised of an ~1500 meter thick sedimentary layer (the Chattisgarh Supergroup) of conglomerates, orthoquartzites, sandstones, shales, limestones, cherts, and dolomites (Naqvi & Rogers, 1987). The sedimentary sequence has been divided into a basal Chandarpur series and an upper Raipur series (Naqvi & Rogers, 1987; Patranabis-Deb et al., 2007). The Chandarpur Group consists of a shale-dominated sequence containing conglomerate and coarse arkose sandstone formed as coalescing fan-fan delta deposits, storm-dominated shelf deposits, and high-energy shoreface deposits. The Raipur Group, 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 group. 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 Chattisgarh 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
Chattisgarh basin by Das et al. (2009). In particular, zircon ages from the Khariar
tuff show a concentration of ages around 1455 Ma.

**Indravati Basin**

The 9000 km² Indravati Basin 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 Chattisgarh and it is postulated that at one point the two were
connected and later eroded into discrete basins (Naqvi & 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
(Maheshwari et al., 2005). Given the recent dating of the tuffaceous layers in the
Chhattisgarh basin, these K-Ar ages should be viewed with skepticism.

**Eastern Dharwar Craton**

The Dharwar craton is split into two separate cratons, an east and west, 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. The Closepet granite is a good
approximation of the western boundary and is used as such in this paper
(Vaidyanadhan and Ramakrishnan, 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
(Figure 9; Naqvi and Rogers, 1987; Vaidyanadhan and Ramakrishnan, 2008).

**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 (Vaidyanadhan and Ramakrishnan, 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 et al. (1990) used whole rock Pb/Pb dating to constrain the age of the Kolar schist belt between 2900-2600 Ma. Nutman et al. (1996) and Nutman (1998) used SHRIMP U-Pb zircon methods to obtain ages of ~2725-2550 Ma. Age trends in the belts indicate younging from west to east (Vaidyanadhan and Ramakrishnan, 2008). These schist belts are all, to some degree, intruded by syn- and post-tectonic felsic melts (Chadwick et al., 2000). Some of the more important greenstone-schist belts and supergroups that will be discussed briefly below include: Sandur schist belt, Ramagiri-(Penakacherla-Sirigeri)-Hungund superbelt (RPSH), Kolar-Kadiri-Jonnagiri-Hutti superbelt (KKJH), and Veligallu-Raichur-Gadwal superbelt (VRG).

The Sandur schist belt is characterized by dominant greenschist facies with amphibolites grade occurring 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 (Figure 9). Granites in the center of the belt were dated using SHRIMP U-Pb and found to be between 2600-2500 Ma (Vaidyanadhan and Ramakrishnan, 2008). Rhyolites from the Sandur greenstone belt have a SHRIMP zircon U-Pb age of 2658 ± 14 Ma (Nutman et al., 1996) and Naqvi et al. (1996) report Sm-Nd ages for basalts and komatiites of 2706 ± 84 Ma.

The RPSH consists of two discontinuous schist belts, the Ramagiri-Penakacherla-Sirigeri and the Hungund. The majority of metamorphism is at greenschist facies. The belts are intruded by a series of granites and gneisses that provide age constraints. Basalts from the Ramigiri 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 (Figure 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 metamorphism with various textures (granular, massive, tufted, and schistose). As with the other greenstone belts of the region, the KKJH is intruded by various felsic dykes that provide age constraints. Pb-Pb isochron data provides an upper estimate for the group at 2700 Ma. This age is consistent with SHRIMP U-Pb zircon analysis of granites and gneisses located in the belt (Vaidyanadhan and Ramakrishnan, 2008). A second age of ~2550 Ma was found in various units within the KKJH using SHRIMP U-Pb zircons and provides a younger limit for the supergroup (Rogers et al., 2007).

The VRG supergroup is located to the south of, beneath, and north of the Cuddapah basin (Figure 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 metabasaltic amphibolites. The northern portion contains pillow 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 Neoarchean in age as compared to those in the Western Dharwar craton described below.

Dharwar Batholith

The Dharwar Batholith is a term first used by Chadwick et al. (2000) to describe a series of parallel plutonic belts. 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 (Vaidyanadhan and Ramakrishnan, 2008). Peninsular gneisses also suggest an early Archaean origin, whereas the granitic gneisses of the Dharwar Batholith are much younger in age. The plutonic belts are approximately 15-25 km wide, hundreds of miles long and separated by greenstone belts (described above). They trend NW to SE except for in the south where the trend becomes predominately a N-S orientation. The belts are mostly mixtures of juvenile multipulse granites and diorites, and are wedge-shaped with steep granitic dyke intrusives (Chadwick et al., 2000; Vaidyanadhan and
Ramakrishnan, 2008). Dating for this unit comes from SHRIMP U-Pb zircon measurements and constrain the emplacement of the Dharwar Batholith to 2700-2500 Ma (Nutman et al., 1996; Nutman, 1998; Friend and Nutman, 1991; Krogstad et al., 1995). Ages for granitic units appear to decrease from west to east; however, gneissic protolith ages of >2900 Ma outcrop at various locations and correspond to the main thermal event of the WDC (Pradhan et al., 2008).

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 works suggest that the similar convexity of adjacent schist belts and granitic plutons may indicate that the Closepet is a ‘stitching granite’ formed during the suturing of the Eastern and Western Dharwar cratons (Figure; Vaidyanadhan and Ramakrishnan, 2008). The exposed rock is divided into northern and southern components by a portion of the Sandur Schist belt; however, both sections appear to be similar lithologically 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 Neoarchean phase of plutonism (Mojzsis et al., 2003) in both the eastern and western Dharwar cratons.

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 have been metamorphosed during the accretion. The Eastern Ghats mobile belt is thought to be the closure point during amalgamation of the EDC. Chadwick et al. (1996, 1999, and 2000) suggest a similar idea; however, their model involves the ‘Dharwar batholith’. This term indicates that separate island arcs or granitic plutons have already formed a solid landmass. The Dharwar batholith obliquely converged with the WDC causing sinistral transpressive shear systems at the margins of most belts. Greenstone belts developed as a result of intra-arc basins associated with the
batholith (Fig. 12). Proposed timing of this collisional history is between 2750-2510 Ma. These models suggest the Closepet granite has been accreted onto the WDC, in contrast with arguments 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 introduces melting to a colder and more depleted mantle. Induced melting from the plume is suggested to emplace juvenile magmas around 2500 Ma in the EDC (Figure 11). The greenstone belts are a result of inverse diapirism and resulting metamorphism. This model suggests that the Closepet granite is a batholith rather than an accreted island arc or a stitching pluton between the EDC/WDC.

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 et al., 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., 2006). Five major dyke clusters of the EDC, described below, include: (1) Hyderabad, (2) Mahbubnagar, (3) Harohalli/Bangalore, (4) Anantapur and (5) Tirupati (Figure 9).

The Hyderabad cluster is located to the north of the Cuddapah basin (Figure 9). Wide 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 (Figure 9), the Mahbubnagar dyke swarm intrudes local granitic gneisses with ages of 2.5-2.4 Ga and 2.2-2.1 Ga (Rb-Sr). The mafic dykes are predominantly gabbro; 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 aphyic or plagioclase-phycic 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 near-by dykes.

The Harohalli/Bangalore swarm is located between the southwestern portion of the Cuddapah basin and the southeastern limb of the Closepet granite (Figure 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 has been dated by Halls et al. (2007) and French et al. (2004) to be between 2366 and 2365 Ma using U-Pb methods. Initial dating of the Harohalli alkaline dykes constrained ages to 850-800 Ma through Rb-Sr whole rock measurements from Ikramuddin and Steuber (1976) and 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 (Figure 9). These two clusters are less studied than other areas; however, 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 1900-1700 Ma and 1500-1350 Ma respectively (Mallikarjuna Rao et al., 1995; Murthy et al., 1987). More recently, Pradhan et al., in review) dated the NE-SW trending ‘Great Dyke of Bukkapatnam’ using U-Pb methods to 1027 ± 13 Ma.
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 Rao 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 present in substantial amounts in four areas within the EDC (NW of, N of, SW of, and in the Cuddapah Basin; Kumar et al., 2007; Figure 9). They are characteristically potassic volcanic rocks, which sometimes bear diamonds. The main areas of kimberlite-lamproite intrusions are known as the Wajrakur, Narayanpet, Krishna and Mallamalai fields (Kumar et al., 2007). Each of these fields contains multiple pipes. There are excellent age constraints on many of these fields. The Wajrakur field is probably the best dated of the four. Rb-Sr ages on the Wajrakur field form a tight cluster between 1091-1102 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 Wajrakur) yielded a 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 for the Muligiripalle pipe of 1079 Ma, but analytical details were not provided. Rb-Sr ages on kimberlites from the Kotakanda and Mudabid kimberlite intrusions yielded ages of 1084 ± 14 and 1098 ± 12 Ma (Kumar et al., 2001). It should be noted that there are 40Ar/39Ar ages from the Kotakanda kimberlites that are much older. Chalapathi-Rao et al., (1999) obtained plateau ages of 1401 ± 5 Ma for a phlogopite separate from Kotakanda and 1417 ± 8 Ma from a lamproite at Chelima. The discrepancy in the Rb-Sr and 40Ar/39Ar ages from Kotakanda are addressed recently by Gopalan and Kumar (2008) who applied K-Ca dating to samples from the Kotakanda swarm and obtained ages of 1068 ± 19 Ma. Gopalan and Kumar (2008) argue that the 40Ar/39Ar results of Chalapathi Rao et al. (1999) are affected by excess argon and the
Kotakanda field is ~1100 Ma. It is unclear if the lamproite in Chelima represents an older suite of lamproitic intrusion.

Proterozoic Sedimentation

Cuddapah Basin

The Cuddapah basin, located in the eastern portion of the EDC, is one of the most well studied basins in India (Figure 10). 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 Eparchean Unconformity (a nonconformity associated with undisturbed contact to older Archaean rocks). Sediments and minor volcanic of the basin are estimated to be approximately 12 km thick and made up of two distinct groups. The Cuddapah Supergroup is the older unit and is present throughout. The Kurnool Group is deposited unconformably over the Cuddapah and is located 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 most recent presence of igneous activity in the basin is the kimberlite and lamproite field located in the center (Figure 10 Chakrabarti et al., 2006). Figure 12 provides a stratigraphic section of the Cuddapah basin.

Initiation of the basin deposition currently has two main hypotheses. Chatterjee and Bhattacharya (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 basic body in the southwestern portion of the basin that could have 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. The other idea is introduced by Chaudhuri et al. (2002). This hypothesis suggests that deep, basin margin faults have played a major role in controlling the evolution of the basin. Evidence for these faults comes from seismic sections and Bouguer anomaly data that indicate a possible rifting environment. A lower limit for the basin is given by a mafic dyke on the southwest border, dated by Chatterjee and Bhattacharya (2001), using $^{40}$Ar/$^{39}$Ar methods to obtain an age of $1879 \pm 5$ Ma. This age is similar to the 1882 Ma U-Pb ages on the
Pullivendla sill by French et al. (2008). Pradhan et al. (in review) note the similarity in paleomagnetic directions between the ~1.9 Ga Bastar dykes and the Cuddapah traps volcanics. The preponderance of the evidence suggests a thermal pulse of ~1.9 Ga for the initiation of basin formation in the Cuddapah basin. Cuddapah basin sedimentation was discontinuous with numerous unconformities within the Cuddapah Supergroup and a major unconformity between the Cuddapah Supergroup and the 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, but such a correlation is based more in tradition than in strong correlative evidence and radiometric dating.

**Prahnita-Godavari Basin**

The Prahnita-Godavari (P-G) basin is made up of two NW-SE trending subparallel basins sandwiched between the Dharwar and Bastar cratons (fig 7). 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. Extensive Paleozoic-Mesozoic ages Gondwana sediments lie 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 and aggregate thickness of ~6000 m (Ramakrishnan and Vaidyanadhan, 2008). The rocks are mildly deformed and weakly metamorphosed. Age constraints are lacking within the basin although Chaudhuri (2003) gives a range between 1330-790 Ma for the sequence.

There are numerous stratigraphic interpretations (and names) for the P-G sequence, but we present the version favored by Chaudhuri (2003). Chaudhuri (2003) presents the alternative stratigraphic groupings in his paper. 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 Mallampi is predominately limestone and quartzarenite 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 (see Chaudhuri, 2003 for a complete discussion).

The Proterozoic sedimentary sequence is unconformably overlain the the Paleozoic-Mesozoic aged Gondwana Supergroup.

**Bhima Basin**

The Bhima basin is located between the northern margin of the EDC and the Deccan Trap flows. The basin is much smaller than the Cuddapah and covers 5,200 km² 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 subformations do 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 ~2500 Ma (Sastry et al., 1999). It is currently under debate as to whether the basin formed during the Meso or Neoproterozoic (Malone et al., 2008; Patranabis-Deb et al., 2007).

**Western Dharwar Craton**

The Western Dharwar craton (WDC) is located in south-west India (Figure 13). 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 co-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 Closepet granite batholith is thought to be the border between the WDC and the EDC, and has been age dated by Friend and Nutman (1991) using SHRIMP U-Pb dating on zircon providing an age of 2513 ± 5 Ma.

Cratonization

The Archean tonalitic-trondhjemitic-granodioritic (TTG) gneisses are found throughout the WDC, 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 have also been published with ages ranging from 3.5 – 3.6 Ga. Three generations of volcanic-sedimentary greenstone 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 Gneiss 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 and amphibolite facies in the north and granulite facies in the south. The metamorphic increase corresponds to a paleopressure increase from 3 – 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 Archean crust that has been tectonically upturned and channeled by erosion is exposed in the western Dharwar Craton.

The Sargur Group

The Sargur Group greenstone belts display well-preserved volcanic-sedimentary sequences. Generally these comprise of ultramafic to mafic volcanic rocks (komatiitic to tholeiitic sources) that transition to felsic volcanic rocks upward, often interpreted to be related to a calc-alkaline source (e.g. Naqvi, 1981; Charan et al., 1988; Srikantia and Bose, 1985; Srikantia and Venkataramana, 1989;
Srikantia and Rao, 1990; Venkata Dasu 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 direct evaporation age of ~ 3.3 Ga. SHRIMP U-Pb analysis yielded ages between 3.1 – 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 ultramafic 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 helps to 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 lower limit to 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 ultra-mafic plateau sequences (Jayananda et al., 2008).

Two commonly suggested single stage methods of generating the observed rock in the WDC are suggested: 1) massive partial melting related to a plume event, and 2) magmatism related to subduction processes. The ultramafic rock in the greenstone belts shows 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, 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 et al., 1997; Arndt, 1994, 2003; 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, passed 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 the average, is andesitic in composition. Following this logic, many workers have suggested that subduction zones represent major sites of crustal 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 ratios of Nb-U, Nb-Th, Nb-La, Th-U that are not consistent with an arc setting (Jayananda et al., 2008). Also, even assuming the elevated Archean 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 ultra-mafic 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 (2008) conclude that ~ 50 Ma separate 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 (Jayananda et al., 2008). This thickened plateau crust, as well as the restite in the lithospheric mantle root beneath the plateau left over from the melting event, is unsubductable and likely to accrete against any continental margin.
encountered (Abbot and Mooney, 1995; Cloos, 1993). 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 (Jayananda et al., 2008). 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 (Figure 11). 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-tholiitic 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 (Jayananda et al., 2008) volcanism appears to have been penne-contemporaneous with the formation of the TTG basement, and provided hosting for the intrusion of the TTG protoliths (Jayananda et al., 2008). 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 (Boyet and Carlson, 2005). Instead, the komatiite magmas show the characteristic geochemical evidence of a depleted mantle source (Boyet and Carlson, 2005). This mantle depletion at 3.35 Ga is significant. It suggests that the earlier extraction of enriched materials from the mantle had depleted the upper mantle prior to 3.35 Ga (Jayananda et al., 2008). A mantle plume is the most likely culprit to form 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).

The Dharwar Supergroup

The Dharwar supergroup is exposed in 2 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 km². The base of this unit is represented by the Kartikere conglomerate, which extends
along the southern margin of the belt discontinuously for ~ 40 km. This unit grades into a quartzite unit. The detrital zircon population from the quartzite suggests that the main source of sediment was derived from the Chikmagalur granodiorite. The overlying formations typically consist of metabasalts with intercalated metasedimentary units, with occasional gabbroic sills, minor BLF, 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 km² 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 km²) NW trending belt separated from the previous two by outcropping TTG basement gneiss. The contact between these basement gneisses and the scist belt is observed as a zone of high grade metamorphism, often with kyanite and garnet phases present. Granitoid intrusions are also present in the north of the belt.

Proterozoic Dyke Swarms

Mafic dyke swarms intrude many areas of the WDC, showing a variety of orientations and variable density. 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 in the Western Dharwar craton of Proterozoic age known as the (1) Hasan-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 swam 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.

Proterozoic Sedimentary Basins
The Kaladgi-Badami basin is the only significant Proterozoic intracratonic basin of the Western Darwar craton, occupying an E-W trending position on the northern edge of the craton (Dey et al., In Press). This basin formed on TTG gneisses and greenstones of Archean age. The Kaladgi supergroup preserves the record of sedimentation in the basin, and consists of sandstones, mudstones and carbonates (Dey et al., In Press). 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., In Press). An angular unconformity between the two constituent groups (The lower Bagalkot and overlying Badami) suggests a period of uplift in the basin’s history (Jayaprakash et al., 1987). Deformation in Bagalkot group is significant, whereas the upper group only exhibited mild deformation (Kale and I’hansalkar, 1991).

Granitic Intrusions

Late to post-tectonic Dharwar potassic granite plutons (~ 2.5-2.6 Ga) that denote crustal reworking in WDC (Jayananda et al., 2006), occur as isolated discrete intrusions cutting across the foliation and banding of the Peninsular Genisses (~ 3.0 Ga). 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, which formed 500 Ma later (Jayananda et al., 2006). 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 K granites played in the cratonization of the WDC. Jayananda et al. (2006) suggests that they may be related either to a thermal event prior to the termination of craton stabilization, or that they actually represent part of long 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 isochrones yielding an age of ~ 2.6 Ga, as well as SIMS U-Pb zircon age dating of ~ 2610 Ma. The Jampalnaikankote Granite is roughly oval shaped pluton which intrudes into the Chitradurga schist belt. Rb-Sr ages put it at ~ 2.6 Ga. The Arsikere and Banavara Granites are thought to be from a single pluton that is connected at depth. The Arsikere granitic batholith is oval in shape of approximately 75 sq. km. The intrusion is primarily a potassic biotite granite which yields 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.

Southern Granulite Province

India’s Southern Granulite province (SG) consists of three late Archaean to Neoproterozoic, 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) (Figure 14).

Northern Block
The Salem Block (also known as the Northern 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; Figure 14). Lithologies present in the Salem include, pyroxene-bearing granites (charnockites), granite gneisses, and migmatites (Devaraju et al., 2007; Clark et al., in press). 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, however, a bivergent reflection pattern at the boundary between the NoB and the DC (with reflections dipping towards the boundary), a thicker crust (~46km) beneath the Salem Block, and a pattern of tectonically induced imbricate faulting in the lower crust and upper mantle of this region (Rao et al., 2006). These observations are indicative of a collisional environment in which the Salem Block was accredited onto the DC in the mid-Archaean.

Most recently, Clark et al. (in press) 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. (in press) 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.

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 Gondwanan orogenies on the Indian Peninsula (Harris et al., 1994). This shear zone has been correlated with the Bongolava-Ranotsara Shear Zone (De Wit et al., 1995; Clark et al., in press), the Madagascar Axial High-Grade Zone (Windley et al., 1994),
and the East Antarctic Napier and Rayner Complexes (Harris et al., 1994). Drury et al. (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 show ages of 485±12 Ma (Meissner et al., 2002).

**Nilgiri Block**

The Nilgiri 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 metamorphism as late as 2460±81 Ma (Raith et al., 1999). These dates 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 Peninsula; paleo-depths range from ~22 km in the southwest (6-7 kbar) to ~35km (9-10kbar) at the MSZ (Raith, 1999). The age of metamorphism is identical to that observed in the Salem Block to the north (Clark et al., in press)

**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, which are squeezed into a long, thin band of gneisses, migmatites presumably resulting from shearing in the PCSZ. Included in this area are the Maddukarai Supracrustals, which display broad doming, complex folding, and extreme hinge line variation caused by Neoproterozoic-Ordovician suturing along the PCSZ (Chetty and Rao, 2006). Retrograde amphibolite facies 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, 1969; Bernard-Griffiths et al., 1987). Electron microprobe analysis (EPMA) of zircons and monazites support these dates and reveal a second thermal
impulse at 2000-1700 Ma (Santosh et al., 2002).

**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 8-11kbar 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 (Braun et al., 2007). 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 a more constrained bimodal age range at 950-850 Ma and 600-450 Ma (Braun, 2007; Santosh et al., 2002). Braun et al. (2007) speculated that the bimodal distribution of monazite ages indicated two distinct high-grade metamorphic intervals one at c. 900 Ma and the other at c. 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.

**Anchankovil Shear Zone**

The ACSZ is a NW-SE trending, subrectangular shear zone stretching more than 120km NW-SE and up to 50km wide (Rajesh et al., 2006) that 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 et al., 2006). Model Sm-Nd ages from a selection of samples suggest a protolith age of 1700-1500 Ma, significantly younger than the protoliths for either of the adjacent granulite blocks (Brandon and Mean, 1995; Bartlett et al., 1998, Cenki et al., 2005). Individual zircons and monazites, analyzed by CHIME and SHRIMP show metamorphic events between 525 and 508 Ma (Collins and Santosh, 2004;
Santosh et al., 2004). Minerals analyzed through EPMA give chemical ages between 520 and 590 Ma (Braun and Broecker, 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) (Rajesh et al., 2004; Somesh et al., 1982). 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).

**Trivandrum Block**

Also known as the Kerala Khondalite belt, 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 (Santosh et al., 2006; Spear and Pyle, 2002; Montel et al., 2000; Bindu et al., 1998; Braun et al., 1998). Zircon cores dated as old as 3460±20 Ma are relicts of some of the earliest forming 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).

**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. The Closepet Granite separates the Eastern and Western Dharwar cratons and is considered to be a stitching pluton of Late Archean 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).

**Mobile Belts**

**Central Indian Tectonic Zone**

The Central Indian Tectonic Zone (CITZ) (Figure 15), also referred to as the Satpura Mobile Belt, is a Proterozoic complex orogenic belt that formed during the accretion of the Bastar-Singhbhum craton to the northern Bundelkhand craton (Radhakrishna, 1989). The mobile belt was originally named after the Satpur Hills and is bounded by the Narmanda-Son North Fault (NSNF) and to the south by the Central Indian Shear Zone (CISZ). The gneissic tectonic zone comprises three sub-parallel E-W trending supracrustal belts that are separated from each other by crustal scale shear zones (Blue India Book). 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.

**Central Indian Suture**

The Central Indian Suture (CIS) (Figure 15) 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 metasedimentary and granulite frocks to the north from the low-grade Proterozoicvolcanogenic 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 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 to result in the Malanjkhand and Dongargarh plutons; 3) at 2.1 Ga, the Sakoli and Nandgaon volcanics formed and 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 to 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.

**Mahakoshal Belt**

The Mahakoshal belt (MB) extends about 600 km from Barmanghat to Rihand Dam and trends in an ENE_WSW fashion. The NSNF delineates the northern boundary to the MB and separates it from the Vindyan basin to the north. The southern boundary, mostly covered by the Deccan Traps, is demarked by the NSSF that separates the MB from the Proterozoic granites of the CITZ (Indian Blue Book).

Quartzites, carbonitites, 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 recline, E-W striking folds with axial planes dipping to the south and pronounced crenulation cleavage. The third phase is evident by broad folds an N-S striking axial planes (India Blue Book).

**Betul Belt**

The Betul belt (BB) trends ENE-SWS and extends about 135 km from Betul to Chhindwara and has a width of about 15-20 km. 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 ~1500Ma syntectonic granites and ~850 Ma late-to-post tectonic granite indicate the BB and the MB bay be penecontemporaneous where the BB represent the arc and the MB represent s the back-arc at eh old continental margin (India blue book).

**Sausar Mobile Belt**

The Sausar Mobile belt is generally accepted to be part of the CIS, although it is widely debated that a sutureis present thorough the entire length of the CITZ (e.g., Mishra et al., 2000; Yedekar et al., 1990; Jain et al., 1991; Rao and Reddy, 2002; and Roy and Prasas, 2001. Rocks within the Sausar belt record a protracted history of convergent margin activities for ca. 1400 Ma to ca. 800 ma (Roy et al., ??). The Sausar metasedimentary rocks are metamorphose 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 width within the Tirondi gneisses (Jain et al., 1991).

Eastern Ghats Mobile Belt

The Eastern Ghats mobile belt (EGMB) is a Proterozoic granulite belt that 
extends for ~1000 km from the Bramani River in the north to Ongole in the south 
(Figure 15). 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; and Radhakrishna, 1989). The Eastern Ghats 
terrane comprises metapelitic 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).

Ramakrishna 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 consist of a mixture of lithologic unites 
belonging to both cratonic India and the EGMB. The SCZ comprises charnockites 
quarts-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 (sapphire-cordierite-spinel-
orthopyroxene). The CMZ is composed of migmatitic gneisses with intrusions of 
charnockite-enderbite, granite and anorthosite (Nanda and Pati, 1989).

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. Gneisses fabrics in the rocks were produced during 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 100-1100°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 amphibolites 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 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 Archean 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 40Ar/39Ar 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. These results imply a major discontinuity between these regions of the EGMB. The updated mineral ages of Mezger and Cosca (1999) indicate 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 present in Rayner Complex of Antarctica (Paul et al., 1990; Shaw et al., 1997) and thus lends credence to the SWEAT model of Moores (1991), Dalziel (1991), and Hoffman (1991) wherein India and Antarctica are juxtaposed.

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

Most recently Chatterjee et al. (2008) reported ages on the Chilka Lake anorthosite of 983 ± 2.5 Ma. 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.

**Gondwana Supergroup**

The Gondwana Supergroup outcrops in a series of small basins in eastern and central India (Figure 16). The age of the Gondwana succession is thought to range from early Permian to early Cretaceous (Dutta, 2002). There is considerable controversy regarding the stratigraphic age and sequences within and across the various Gondwana sub-basins (Dutta, 2002; Dutta, 1994; Kutty et al., 1987).

For simplicity, we accept the conclusions of Dutta (2002) for the general sequence of Gondwana sedimentation (Figure 16) with Facies A and B constrained to the Late Carboniferous to Permian and Facies C and D confined to the Triassic and Early Jurassic. The lower section (A & B facies) are composed of tillites, conglomerates shales, coal measures and sandstones. The Upper (C & D facies) are feldspathic sandstones, redbeds and quartz-rich sandstones.

**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 indicates 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 of these cratons to form ‘proto-India’ (with the exception of some blocks in the southern granulite province).

Major Proterozoic basin formation (the so-called “Purana” basins) appears to have occurred during three main pulses. These include a Mesoproterozoic phase of basin formation (Lower Vindhyan, Lower Chattisgarh and Cuddapah basins); a early Neoproterozoic phase of basin formation (Upper Vindhyan, Upper Chattisgarh basins) and a late Neoproterozoic phase of basin formation (Marwar basin, 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 baddelyite (see discussions above). Ultramafic intrusions in India are concentrated at about 1.1 Ga.

Peninsular India is thought to have been part of 5 different supercontinental configurations. The oldest of these “expanded-Ur” (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 Archean cratonic material in East Antarctica (fig 19-a). 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 Zegers et al., 1998 and Wingate, 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 19-b; Zhao et al., 1994, Rogers and Santosh, 2002). 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 Gondwana (fig 19-c). 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., 2001; Meert and Lieberman, 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 Squires et al. (2006) and Veevers (2004).

By the end of the Cambrian, India was part of the large southern continent of Gondwana (Figure 17-d) and remained a part of the Gondwana supercontinent until its breakup in Mesozoic times. The Gondwana basins were developed within India during the Paleozoic and early Mesozoic. Eruptions of the Rajhmahal and Deccan traps volcanic rocks took place during the breakup of Gondwana.

During the Cenozoic, India began its rapid journey at the trailing end of the Tethyan plate and was incorporated into the Eurasian continent. This concludes the most recent 3.5 billion years of history for the Indian subcontinent.

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**Figure Legends**

**Figure 1:** Generalized tectonic map of Indian subcontinent: Precambrian Cratons, Mobile belts and Lineaments. AFB = Aravalli Fold Belt, DFB = Delhi Fold Belt, EGMB = Eastern Ghat Mobile Belt, SMB=Satpura Mobile Belt, NSL = Narmada Son lineament, CIS = Central Indian Suture and BPMP = Bhavani Palghat Mobile Belt. (Modified from Vijaya Rao and P.R. Reddy; 2001).

**Figure 2:** Sketch map of the major units in the Aravalli-Bundelkhand craton, NW India (after Naqvi and Rogers, 1987 and Ramakrishnan and Vaidyanadhan, 2008).

**Figure 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).

**Figure 4:** Lithologies of the Vindhyan supergroup in the Rajasthan sector and the Son valley sector with previous age estimates (Malone et al., 2008).

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

**Figure 6:** Lithologies of the Singhbhum craton after Ramakrishnan and Vaidyanadhan, 2008.

**Figure 7:** Sketch map of the Bastar craton after Naqvi and Rogers, 1987 and Ramakrishnan and Vaidyanadhan, 2008.

**Figure 8:** Lithologies of the Bastar craton after Ramakrishnan and Vaidyanadhan, 2008.
Figure 9: Sketch map of the Eastern Dharwar Craton. Archaen assemblages associated with cratonization are shown here. Abbreviations for schist belts are as follows: Sa = Sandur, KKJH = Kolar-Kadiri=Jonnagiri-Hutti, RPSH+ Ramagiri-(Penakacheria-Sirigeri)-Hungund, and VPG = Veligallu-Raichur-Gadwal superbelt. The dotted and dashed line indicates possible location under the basin. Dashed lines represent Paleozoic to more recent sedimentary cover (Modified from Naqvi and Rogers, 1987). Dyke intrusions into the EDC are H&B=Harohalli and Bangalore swarm; A=Anantapur swarm; M=Mahbubnagar swarm; H=Hyderabad swarm.

Figure 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 melting of the lower lithosphere creates plutons that form the mainland of the EDC (Modified from Jayananda et. al., 2000).

Figure 11: Vertical tectonic plume model for the Eastern Dharwar craton after Jayananda et al. (2000), Chardon et al. (2002). The plume model suggests a large mantle plume situated just beneath the EDC/WDC boundary in an enriched mantle. Further east, the plume introduces melting to a colder and more depleted mantle. Induced melting from the plume is suggested to emplace juvenile magmas around 2500 Ma in the EDC. The greenstone belts are a result of inverse diapirism and resulting metamorphism. This model suggests that the Closepet granite is a batholith rather than an accreted island arc or a stitching pluton between the EDC/WDC.

Figure 12: Lithologies of the Cuddapah basin (eastern India) after Naqvi and Rogers, 1987 and Ramakrishnan and Vaidyanadhan, 2008.

Figure 13: Sketch map of the western Dharwar craton showing major lithologic boundaries after Naqvi and Rogers, 1987 and Ramakrishnan and Vaidyanadhan, 2008.

Figure 14: Sketch map of the southern granulite province blocks and associated shear zones after Naqvi and Rogers, 1987 and Ramakrishnan and Vaidyanadhan, 2008.

Figure 15: Major Precambrian sedimentary basins of India and fold belts (after Naqvi and Rogers, 1987 and Ramakrishnan and Vaidyanadhan, 2008).

Figure 16: (a) Map of the major Gondwana basins in central India (after Dutta, 2002) (b) Proposed correlations between Gondwana sediments in the various sub-basins (DB=Damodar Basin; PGB=Prahnita-Godavari Basin; SB=Son basin; SPB=Saptura Basin).
Figure 17: (a) Expanded “Ur” after Rogers and Santosh (2004); 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); (c) Rodinia after Li et al. (2008); (d) Gondwana after Gray et al. (2007) and Meert and Lieberman (2008).
Figure #1

LEGEND

- Himalayas
- Deccan Traps
- Rajmahal Traps
- Proterozoic Basins

CuB: Cuddapah Basin
Ba: Bastar Basin
ChB: Chattisgarh Basin
VB: Vindhyan Basin
CG: Closepet Granite
Figure #3
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<th>Son Valley</th>
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<tr>
<td>Bhandar Group (~1200 m)</td>
<td>Bhandar Ls. 750 Ma (Sr isotopic curve); 650+/-770 Ma Ediacaran (fossils?)</td>
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<td>Rewa Group (285 m)</td>
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<td>Kaimur Group (180 m)</td>
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<td>Kaimur Group (425 m)</td>
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<td>Khorip Group (475 m)</td>
<td>Rohtas Ls: 1601 +/-130 Ma</td>
<td>Rohtas Group (454 m)</td>
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<td>Lasrawan Group (272 m)</td>
<td>Rampur Sh: 1599 +/-12 Ma Porcellinite: 1628 +/-0.8; 1631 +/-5.4; 1630 +/-0.8 Ma</td>
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<td>Sand Group (210 m)</td>
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<td>Deonar Group (1200 m)</td>
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<td>Satola Group (835 m)</td>
<td>Kajrahat Ls: 1721 +/-90 Ma</td>
<td>Mirzapur Group (990 m)</td>
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Fig 4

Majhgawan Kimberlite - 1073 +/- 13.7 Ma
Figure 10a

WDC

Dharwar Batholith

Convergence Direction

Figure 10b

Cooling Rate

Kolar Schist Belt

Bangalore

Hoskote-Kolar

Closepet

Crust

Mantle

Depleted Mantle

Figure #10
Figure 11

Stage 1
3400-3350 Ma

- Holenarsipur
- Kalyadi
- Nuggihalli
- Ghattihosahalli
- J.C. Pura
- Banasandra Komatiites

Stage 2
3350-3300 Ma

- > 3400 Ma Microcontinent
- 3320 Ma TTG
- 3300 Ma Felsic Volcanics

Stage 3
3300-3200 Ma

- > 3400 Ma Microcontinent
- 3300-3200 Ma TTG
- 3250 Ma Felsic Volcanics
Figure #12

Nandyal Shale
Kolikuntala Limestone
Paniam Quartzite
Owk Shale
Narji Limestone
Bandanapalli Formation
Srisailam Quartzite
Cumbum Formation
Bairenkonda/Nagari Quartzite
Gandikota Quartzite
Tadpatri Formation
Pulivendla Quartzite
Vempalle Formation
Gulcheru Quartzite

Shaly Limestone
Limestone
Quartz Arenite
Silty Claystone and Quartzites
Limestone
Shale
Quartzite
Conglomerate
Quartzite
Alternating Dolomite and Slate
Quartzite and Slate
Shale and Quartzite
Dolomitic Shale
Conglomerate/Quartzose Sandstone
Dolomitic Mudstone
Quartzite and Conglomerate
Crystalline Basement
Figure #14

Legend
Adapted from Tamashiro, et al., 2004

- Interspersed Granites and Gneisses
- Granulite Terranes
- Granite-Greenstone Terranes
- Shear Zones
  - MBSZ: Moyar-Bhavani
  - PCSZ: Palghat-Cauvery
  - ACSZ: Achankovil
- Phanerozoic Sediment Cover

Study Area
Figure #15

LEGEND
CuB: Cuddapah Basin
ChB: Chattisgarh Basin
VB: Vindhyan Basin
CG: Closepet Granite
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
BPMP: Bhavani Palghat Mobile Belt
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<th>Craton</th>
<th>Oldest Age</th>
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Abbreviations: (t)=tectonothermal; (s) shear; (d) dyke intrusion (k) kimberlite/lamproite intrusion (g) granitic and/or mafic intrusions; (v) volcanism