1 Age and Implications of the phosphatic Birmania Formation, Rajasthan,
2 India

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4 Nigel C. Hughes^{a*}, Paul M. Myrow^b, N. Ryan McKenzie^{c,d}, Shuhai Xiao^e, Dhiraj M.
5 Banerjee^f, Daniel F. Stockli^d, Qing Tang^e

6
7 ^a*Department of Earth Sciences, University of California, Riverside, CA 92521, USA*

8 ^b*Department of Geology, Colorado College, Colorado Springs, CO 80903, USA*

9 ^c*Department of Geology and Geophysics, Yale University, New Haven, CT 06520,*
10 *USA*

11 ^d*Department of Geological Sciences, Jackson School of Geosciences, University of*
12 *Texas at Austin, TX 78712, USA*

13 ^e*Department of Geosciences, Virginia Polytechnic Institute, Blacksburg, VA, 24061,*
14 *USA*

15 ^f*Department of Geology, Delhi University, Delhi 110007, India*

16
17 * Corresponding author. *Email address* nigel.hughes@ucr.edu (N.C. Hughes)

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ABSTRACT

The Birmania inlier in western Rajasthan, India, contains important phosphate deposits, the depositional age of which is poorly constrained. Here we provide the first direct age constraints for the phosphate-bearing Birmania Formation using a combination of paleontological and detrital zircon U-Pb data. The occurrence of the multicellular algal fossil *Wengania exquisita* in phosphatic chert of the Birmania Formation suggest that it was deposited during the Ediacaran. Detrital zircon age distributions contain prominent populations of 1.7–1.9 Ga grains, with subordinate younger grains that range from 650–980 Ma. The distribution broadly resembles those from Neoproterozoic strata from both cratonic and Himalayan India but, like zircon age distributions from the Marwar Group, lacks 1.0–1.2 Ga grains, which may suggest that both areas shared similar local sources. The lack of zircon grains younger than ~650 Ma is consistent with an Ediacaran depositional age because almost all Cambrian or younger strata from India have yielded Cambrian or latest Neoproterozoic age grains. These findings raise the possibility of a previously unrecognized late Neoproterozoic episode of phosphogenesis on the India craton. The presence of *Wengania exquisita* further supports strong palaeogeographic affinity between the India and South China during the Neoproterozoic and Cambrian.

1. Introduction

India has an extensive and well-preserved suite of Proterozoic sedimentary rocks that have great potential for studies of the sedimentary and tectonic history of the

continent. Realizing this potential requires resolving uncertainties in the depositional age and correlations among the various outcrop belts that constitute the “Purana Basins” (a name applied to all Proterozoic successions across the Indian subcontinent) (Wadia, 1975), and recent studies have made considerable progress in this regard (e.g., Malone et al., 2008; McKenzie et al., 2011a, 2013; Patranabis-Deb et al., 2007; Pradhan et al., 2012). Ore-grade phosphorite deposits have attracted special interest for their economic potential their relatively rarity in the stratigraphic record, and their value as candidate marker horizons for inter-basinal stratigraphic correlation (e.g., Banerjee et al., 1986, 1997). However, uncertainty remains concerning the total number of phosphatic depositional events represented within the Indian stratigraphic succession. Improved resolution requires accurate dating of the depositional ages of individual phosphatic deposits. This paper provides such constraints for an important and enigmatic phosphate horizon, the Birmania Formation of the geographically isolated Birmania inlier of western Rajasthan.

2. History, Location and Geological Setting

The Birmania inlier is located west of the extensive outcrops of Proterozoic rock preserved in the Aravalli–Delhi Belt and the Vindhyan Basin (Fig. 1). It consists of a poorly exposed sequence of sedimentary rocks exposed on the desert floor and as occasional low hills about 70 km south of Jaisalmer (Fig. 2A), and has been mapped as two outcrop belts separated by a tract of sand (Mathur and Chauhan, 1995; Srikantan et al., 1969). The inlier contains sedimentary rocks that are assigned to three

68 formations: in ascending order the Rhanda, Birmania, and Lathi formations. The Lathi
69 Formation comprises undeformed sandstone (Pareek, 1984) reported to contain
70 fossilized wood and leaves (Mathur and Chauhan, 1995), and is known widely in
71 western Rajasthan (Roy and Jakhar, 2002). These early Jurassic (Roy and Jakhar,
72 2002) rocks rest unconformably on older strata. In the Birmania inlier the Lathi
73 Formation is mapped as unconformable overlying the strongly folded Birmania
74 Formation in the northern outcrop belt of the inlier known as the “Ladhu Singh ki Dhani”
75 block (Mathur and Chauhan, 1995; Muktinath, 1969; Roy and Jakhar, 2002, fig. 8.31)
76 (Fig. 1).

77 The Birmania Formation, which comprises both carbonate and siliciclastic rocks,
78 bears a horizon rich in pelleted and microcrystalline phosphate (Husain and Banerjee,
79 1986). This was one of the first potentially exploitable phosphate deposits to have been
80 recognized on the craton, now nearly fifty years ago by R. P. Sheldon in a still
81 confidential USGS report (Sheldon, 1966) and has an estimated reserve of phosphate
82 ore of ~3.5 million tons, hosted in carbonate strata with ~13% P_2O_5 (Deshmukh, 1979).
83 Despite its early recognition, the age and relationships of this phosphate have
84 remained enigmatic because of its geographic isolation from the major outcrops of
85 sedimentary rocks in northern cratonic India (Fig. 1). Furthermore, poor exposure, local
86 faulting, and intense recumbent folding (Fig. 2) have precluded accurate measurement
87 of the combined stratigraphic thickness of all units within the inlier, with estimates
88 varying from less than 600 m (Muktinath, 1969) up to 3084 m thick (Srikantan et al.,
89 1969). Maps of phosphate bed outcrop differ in many details (compare the sketch map
90 of Srikantan et al., [1969] with Mathur and Chauhan [1995, fig. 2]). Nevertheless, a

number of workers have described the basic stratigraphy of the formation, as well as its petrographic characteristics (Deshmukh, 1979; Husain and Banerjee, 1986; Mathur and Chauhan, 1995; Muktinath, 1969; Srikantan et al., 1969), many of them referring to unpublished reports compiled by officers of the Oil and Natural Gas Commission (ONGC) conducted in the late 1960's. All of these works have either advocated for, or accepted, a Palaeozoic age for the Birmania Formation phosphate, for reasons discussed below.

Stratigraphic logs repeatedly illustrate the Birmania Formation to unconformably overly the apparently only mildly deformed mature quartz arenite and red shale of the Randha Formation (e.g. Mathur and Chauhan, 1995, fig. 1; Maheshwari et al., 2002, fig. 2). However, an unconformable contact is not universally accepted, with Muktinath (1969) and Srikantan et al. (1969, table 1) questioning its veracity and Deshmukh (1979) indicating fault between the two formations. We have not observed the contact in the field, but do note, as did Pareek (1984, p. 26), that the degree of deformation of the Randha Formation at Randha village is considerably less than that of the Birmania Formation.

The Randha Formation is reported to unconformably overly a series of igneous and metamorphic rocks including the Jalore-Siwana granite, and the Malani rhyolites (Deshmukh, 1979; Husain and Banerjee, 1986; Muktinath, 1969; Srikantan et al., 1969), now collectively known as the Malani Igneous Suite. However, the contact between these units is not shown in the published maps, whereas the contact of the Birmania Formation with the Lathi Formation is mapped. Furthermore, discussions with authors of some of the earlier works (S.C. Mathur and G.P Deshmukh, pers. comm.

2014) suggest that no direct contact between the Randha Formation and the igneous/metamorphic rocks has been observed. On the other hand, Pareek (1984, p. 26) suggested that contact of the Randha Formation with the basement along the western boundary was faulted.

The Palaeozoic age determination for the Birmania Formation rests on indirect evidence. The Malani Igneous Suite has long been assumed to be Precambrian in age, a determination confirmed by recent U-Pb isotopic date of 771 ± 5 Ma (Gregory et al., 2009). Given that the Lathi Formation is lower Jurassic, early workers (e.g. (Muktinath, 1969; Pareek, 1984; Srikantan et al., 1969) appear to have simply interpolated a Palaeozoic age for the Randha and Birmania formations based on their putative stratigraphic occurrence between Precambrian and Mesozoic rocks. A similar argument was commonly applied to rocks of the Marwar Supergroup of eastern Rajasthan, a mixed clastic and carbonate suite that overlies the Malani Igneous Suite, with many authors considering the whole sequence to be Cambrian (e.g. Deshmukh, 1979; Muktinath, 1969; Pareek, 1984, p. 20). A general lithological comparison correlates the Randha Formation with the siliciclastic Jodhpur Group, and the Birmania Formation with carbonate of the Bilara Group (Deshmukh, 1979; Muktinath, 1969; Pareek, 1984) (Fig. 1). Oddly, a correlation between the carbonate-rich Birmania Formation and the siliciclastic Jodhpur Group has also been suggested recently (Pandey and Bahadur, 2009), although justification for this suggestion was not presented. While recent studies have confirmed that the uppermost siliciclastic units of the Marwar Supergroup are indeed early Cambrian (Kumar and Pandey, 2008; McKenzie et al., 2011a), much of the lower part of the Marwar Supergroup, including

the carbonate-rich Bilara Formation, is almost certainly late Neoproterozoic (Kumar, 2012; Mazumdar and Strauss, 2006) Thus, if the Bilara and Birmania formations are directly correlative, the latter is more likely to be Precambrian. Clearly then, the case for a Palaeozoic age for the Birmania Formation based on interpolation alone is weak.

Another argument for a Palaeozoic age has been the attempt to link the Birmania phosphates beds with a specific earliest Palaeozoic phosphate known elsewhere along the Himalayan margin in India and Pakistan (e.g., Husain and Banerjee, 1986). This argument recognizes that phosphorites occur sporadically in the sedimentary record, and thus can serve as valuable stratigraphic markers. The Chert-Phosphate Member of the Tal Group of the Lesser Himalaya near Mussoorie, India is now precisely dated as *Anabarites trisulcatus* – *Protohertzinia anabarica* Zone in age, the earliest of the three zones of the pre-trilobitic Cambrian based on small shelly fossils (~ 535 Ma) (Hughes et al., 2005). Similar phosphate bearing deposits also occur near Abbotabad in Pakistan (Mostler, 1980) although these are likely some 5 million years younger than those near Mussoorie (Hughes et al., 2005). Similarities between the Chert-Phosphate Member and the Birmania Formation are that both consist of stretched intraclasts of microgranular aphanitic phosphate, in a carbonate and quartz-rich matrix. However, the Chert-Phosphate Member is much richer in organic matter, pyrite, and clays, and is composed of francolite, whereas the Birmania Formation fluorapatite is depleted in carbonate (Husain and Banerjee, 1986). Accordingly, mineralogical grounds for invoking a genetic link rest on general, rather than specific, similarities. Nevertheless, this correlation was used to putatively place the position of the Precambrian-Cambrian boundary at the base of the phosphate within Birmania Formation at the initiation of a

positive carbon isotopic trend from values about $-4 \delta^{13}\text{C}$ to $+4 \delta^{13}\text{C}$ (Maheshwari et al., 2002).

There are several additional problems with these putative correlations. Firstly, as the nature of the contacts between the igneous rocks and the Randha Formation, and between the Randha and Birmania formations are poorly established (see above) the argument for both the Randha and Birmania formations accumulating sometime between 770 and 200 Ma based on stratigraphic position is undermined. Indeed, the difference in degree of deformation between the apparently mildly deformed Randha Formation and the intensely folded Birmania Formation suggests that a tectonic contact between these units is possible (see Pareek, 1984). Secondly, even if the contacts are stratigraphic, the grounds for interregional correlation based on lithology are weak. The Birmania Formation is primarily carbonate rich, whereas the lower Tal Group is almost exclusively siliciclastic. Likewise, the Jodhpur Group is much thicker than the Randha Formation, and the evaporitic Bilara Group differs markedly from the phosphate-bearing Birmania Formation. The suggestion that the trend in carbon isotopes locates the position of the Precambrian/Cambrian boundary within the Birmania Formation (Maheshwari et al., 2002) is unwarranted, as fluctuations of this magnitude occurred at least four times in the interval between 520 and 650 Ma, and quite commonly at other intervals in addition (Shields-Zhou and Och, 2011, fig. 2). Accordingly, the only firm existing age constraint on the age of the Birmania Formation is that it is pre-Jurassic. Clearly, direct evidence of the depositional age of the Birmania Formation is required. Here we present the first direct evidence of its antiquity.

3. Sedimentology:

Strata of the middle Birmania Formation consist of very thin to thin bedded, tan to white weathering, partly silicified, phosphatic, laminated pelletal grainstone and mudstone (Fig. 3a,b) (samples BA3 and BA4). Significant proportions of medium to coarse, well rounded quartz grains are dispersed within the carbonate. This rock is locally gently to isoclinally disharmonically folded on a meter scale (Fig. 3a). Overlying carbonate deposits consists of dolomudstone with low relief small-scale scours and abundant microbialite lamination (Fig. 3c). Associated thick calcareous sandstone beds are relatively structureless and ferruginous. Two sandstone beds were sampled for detrital zircon analysis, one from within the phosphatic facies of the Birmania Formation (BA2, N 26° 14.562', E 070° 55.663, 235m altitude) and one from a quartz arenite immediately below the phosphate (BA5, N 26° 14.207', E 070° 55.940, 260m altitude) a few hundred meters to the east. A sample of pure white quartzite from the underlying Randha Formation (RH1, N 26° 12.415', E 070° 57.701, 311m altitude) collected at Randha village yielded no zircon grains.

4. Micropaleontology

A carbonate sample, BA1, was collected from the Birmania Formation at N 26° 14.562' E 070° 55.662', 235 m elevation, and separate fractions were processed with formic acid and acetic acid, respectively: neither yielded fossil material. Phosphatic samples BA3 (N 26° 14.562', E 070° 55.663, 235m altitude) and BA4 (N 26° 14.212' E 070° 55.934' at 255 m altitude) were also processed for microfossils. These samples

were trimmed to make standard petrographic thin sections, which were examined under a transmitted light microscope.

Both samples BA3 and BA4 contain multicellular algal fossils that are identified as *Wengania exquisita* (Fig. 4a–h). Microscope slides of the illustrated material are housed at the Cincinnati Museum Center as specimens CMC IP72086 and CMC IP72087. The fossils are spheroidal in morphology, occasionally with a cylindrical extension (Fig. 4d–e). They consist of pseudoparenchymatous cells that are arranged linearly to form tightly packed filaments (Fig. 4c). When such parenchymatous thalli are cut perpendicularly to the filaments, cells appear to be unordered (Fig. 4f,h). Cells are elongate, with a long axis of 1.7–3.8 μm (average = 2.9 μm , s.d. = 0.6 μm , n = 30), a short axis of 0.9–2.1 μm (average = 1.3 μm , s.d. = 0.3 μm , n = 30), and an aspect ratio of 1.3–3.8 (average = 2.2, s.d. = 0.6, n = 30). Cell lumens are filled with organic matter, and phosphatic cements are present between cells. Cells are apparently degraded and shrunk to some degree prior to preservation. Phosphatic laminae, representing multiple generations of cementation, coat on the exterior surface of some specimens (Fig. 4f).

Numerous fragments of multicellular algal thalli are also present in BA3 and BA4 (Fig. 4i–l). Some of them are probably fragments of *Wengania exquisita* (Fig. 4i–j), given their similar cell size and arrangement. For example, cells in the specimen illustrated in Fig. 4i–j have a long axis of 1.0–2.1 μm (average = 1.4 μm , s.d. = 0.4 μm , n = 7), a short axis of 0.4–0.6 μm (average = 0.5 μm , s.d. = 0.1 μm , n = 7), and an aspect ratio of 2.1–3.9 (average = 2.9, s.d. = 0.6, n = 7), similar to measurements of *Wengania exquisita*. Others have more equidimensional cells and may represent additional taxa of algal thalli (e.g., Fig. 4k–l).

Wengania exquisita was first reported from the upper Doushantuo Formation at Weng'an of South China (Zhang et al., 1998). Subsequently, it has been documented widely at multiple Doushantuo sections in South China (e.g. Xiao et al., 2004; Xiao, 2004; Liu et al., 2014; Muscente et al., 2015; Xiao et al., 2014; Zhou et al., 2004), but only from that unit. At Baokang and Xiaofenghe, this species occurs in the upper Doushantuo Formation (Liu et al., 2014; Zhou et al., 2004). The current report is the first discovery of this species outside South China, although an unpublished form likely belonging to a different species occurs in unit A of the Krol Group of the Lesser Himalaya. Insofar as the depositional age of the Doushantuo Formation is constrained between 635 Ma and 551 Ma (Schmitz, 2012), the presence of *Wengania exquisita* in the Birmania Formation suggests a similar age. Current data suggest that *Wengania exquisita* is restricted to the upper Doushantuo Formation in South China (Liu et al., 2014; Zhang et al., 1998; Zhou et al., 2004), and phosphorite of the upper Doushantuo Formation at Weng'an has been dated at 599 ± 4 Ma (Barfod et al., 2002). Thus, the Birmania Formation may have been deposited between 599 Ma and 551 Ma.

5. Detrital Zircon Geochronology

All samples collected for detrital zircon U-Pb geochronology were processed and analyzed at the UTChron laboratory, Department of Geological Sciences, University of Texas at Austin. Zircon grains were extracted from sandstone samples using standard mineral separation techniques, which included crushing, passing samples over a density-separation water table, magnetic separation, and heavy liquids.

Whole individual zircon grains were fixed on a ~1 inch resin mount with double-sided tape for analysis via laser ablation inductively coupled mass spectrometry (LA-ICP-MS). Mounts were placed into a Helex 9 sample cell, volume c.30 cm³, and ablated with a 30 µm spot from a Photon Machines Analyte G2 ATLex 300si ArF Excimer laser. Ablated material was carried by helium gas to a ThermoFisher Element2 double-focusing magnetic sector ICP-MS for isotopic measurements. Data analysis was accomplished using Iolite (Igor Pro) (see supplementary materials for additional analytical details).

Distributions of detrital zircon ages (Fig. 5) for the two samples (BA2, BA5) show several distinct populations of grains. Sample BA2 contains a small population of 730–1000 Ma grains, a large population of 1.5–1.9 Ga grains, and scattered Archean to early Paleoproterozoic grains. The youngest single grain yielded a concordant ²⁰⁶Pb/²³⁸U age of 730 ± 5 Ma. Sample BA5 bears a closely similar age distribution, with a few grains with ages scattered between 650–1000 Ma, a large population of 1.5–1.9 Ga grains, and a small population of ~ 2.5 Ga grains. The youngest single grain yielded a concordant ²⁰⁶Pb/²³⁸U age of 652 ± 9 Ma.

Due to the relative low-abundance and spread of young grain ages within the populations, the four concordant young grains from BA5 were selected and additional spot analyses were conducted to assess the validity of the ages. All additional analyses yielded U-Pb ages similar to the initial ages derived (see supplement for details).

A total of nine grains with concordant ages between 0.6–1.0 Ga were found in both samples (n=187). This secures a maximum depositional age for the Birmania

Formation and excludes the possibility of correlating the Birmania Formation with any of the older Proterozoic phosphatic horizons known in the Lesser Himalaya, Vindhyan, or Aravalli outcrop belts (see McKenzie et al., 2011a, 2013). The single youngest grain (~652 Ma) provides only a likely maximum depositional age constraint. Cambrian and younger strata known from both the Indian craton and Himalaya quite consistently yield large populations of detrital zircon grains with ages around 500-650 Ma (e.g., Myrow et al., 2003, 2009, 2010, 2015; Hughes et al., 2010; Long and McQuarrie, 2010; McQuarrie et al., 2013) with the exception of one demonstrably Cambrian sample from the Himalaya that lacks grains younger than ~750 Ma (Myrow et al., 2010, fig. 3).

Both samples (BA2 and BA5) yielded a dominant sub-population of 1.5–1.9 Ga grains, with a prominent age-peak around ~1720 Ma (Fig 5). Grains of this age are common in siliciclastic rocks across northern India (McKenzie et al., 2011a); however they are especially common in siliciclastic strata of the Aravalli-Delhi Belt (Biju-Sekhar et al., 2003; Kaur et al., 2011, 2013; McKenzie et al., 2013) and the Marwar Supergroup (e.g., McKenzie et al., 2011a; Turner et al., 2014) (Fig. 6). Plutonic rocks with ages ~1.7–1.8 Ga are known from both the Northern and Southern Aravalli-Delhi Belts (Biju-Sekhar et al., 2003; Deb and Thorpe, 2004; Kaur et al., 2007, 2009), which have been interpreted to have been emplaced during Andean-type subduction (Kaur et al., 2011). Younger, ~0.8–1.0 Ga igneous sources are locally known from the Delhi Supergroup and the Erinpura granitoids (Choudhary et al., 1984; Crawford, 1970; Deb et al., 2001; Just et al., 2011; Pradhan et al., 2010) and 0.7–0.8 Ga rocks of the Malani Igneous Suite (Ashwal et al., 2013; Gregory et al., 2009; Just et al., 2011; Torsvik et

al., 2001). Siliciclastic rocks from the Aravalli-Delhi Orogenic Belt contain abundant
~1.6-1.9 Ga detrital zircon grains.

Source material for the few older Archean 3.1–3.2 Ga and Archean to earliest
Paleoproterozoic (ca. 2.5 Ga) grains are present across the Indian shield. These
include the Dharwar region of southern India (Jayananda et al., 2008; Nutman et al.,
1996; Nutman et al., 1992; Peucat et al., 1995), the Bundelkhand craton of central
India (Pradhan et al., 2010), and the locally exposed Banded Gneiss Complex (BGC-I)
in the southern Aravalli-Delhi Orogenic Belt (Wiedenbeck et al., 1996; Buick et al.,
2006). Other possible sources include Madagascar (Mojzsis et al., 2003; Cox et al.,
2004).

The lack of grains with ages from 1.0–1.2 Ga is striking. Grains of such ages are
common in late Neoproterozoic and early Paleozoic rocks throughout both the
Himalaya (e.g., Gehrels et al., 2011; Myrow et al., 2010) and cratonic India, notably in
the upper Vindhyan successions (Malone et al., 2008; McKenzie et al., 2011a; Turner
et al., 2014) and in the Delhi Supergroup of the Southern Aravalli-Delhi belt (McKenzie
et al., 2013), whereas grains of this age are generally sparse in the Marwar
Supergroup (Malone et al., 2008; McKenzie et al., 2011a; Turner et al., 2014) (Fig 6).
Potential igneous source rocks ranging from 1.0-1.2 Ga are presently only known from
various intrusions in the Bundelkhand craton (e.g., Gregory et al., 2006; Pradhan et al.,
2012) and from the Eastern Ghats Belt (e.g., DasGupta et al., 2013). While the overall
zircon age distributions vary in terms of relative abundance of grains, strata from both
the Marwar Supergroup and the Birmania Formation contain grains of the same age
populations. This includes the presence of abundant 0.7–1.0, 1.6–1.9, ~2.5 and 3.0–

3.3 Ga grains, all of which have known sources in the Aravalli sector (Fig. 6). Therefore both Birmania and Marwar strata are likely made up of locally-derived detritus from the northern Aravalli-Delhi Orogenic Belt (ADOB) where source rocks with 1.0–1.2 Ga zircon grains have not yet been identified, and perhaps only limited contributions of detritus from other regions of the Indian craton.

6. Interpretation and Implications

The ages of the youngest detrital zircon grains and of the fossils recovered are consistent with an Ediacaran depositional age for the Birmania Formation. Although this is our preferred interpretation, a younger age is also possible. This would require a stratigraphic range for *Wengania exquisita* that extends into younger rocks than currently known, and that no zircon grains younger than 650 MA were incorporated into a sedimentary rock that accumulated significantly later in time. These are unlikely for several reasons. Firstly, *W. exquisita* has yet to be described from any Phanerozoic rock, despite being common and widespread in the Ediacaran Doushantuo Formation of South China. Secondly, despite concerted efforts, no small shelly fossils, conodonts or other Phanerozoic microfossils (or macrofossils) have been recovered from the Birmania Formation to date. As shelly fossils are common in Phanerozoic phosphorites, notably in the Cambrian Tarnawai and Tal formations in the Himalaya (Mostler, 1980; Bhatt et al., 1983; Brasier and Singh, 1987), they would be expected to be present if the Birmania Formation is Cambrian or younger. Thirdly, although some Cambrian or younger sandstones from Rajasthan may lack Cambrian-aged zircon grains, out of

eight biostratigraphically constrained Cambrian sandstone samples with published detrital zircon ages (including one from Rajasthan) only one sample, from the Lesser Himalaya, is known to lack latest Neoproterozoic or early Cambrian age grains. Thus, although the lack of Cambrian aged grains in the Birmania Formation is not strong evidence for an Ediacaran age, it is consistent with it. Fourthly, the lithostratigraphic correlation with the Marwar Supergroup, also in Rajasthan, supports correlation of the Birmania Formation with the likely Neoproterozoic Bilara Formation. Together these data build a reasonably compelling case for an Ediacaran depositional age. If the age of the Birmania Formation is indeed Ediacaran, we predict that Doushantuo-style acanthomorph may yet be found that could definitively confirm this age, provided that the facies and taphonomic conditions were suitable for their occurrence and preservation (see Zhou et al., 2007).

Should it be accepted that the Birmania Formation is Ediacaran in age this episode of phosphate deposition cannot be correlated with the temporally restricted phosphate deposition that marked the first two small shelly fossil based stages of the early Cambrian represented in the Tal Group and deposits near Abbotabad because the two episodes of phosphate deposition were apparently separated by >10 Myr. Improved chronostratigraphic resolution of the Neoproterozoic and Cambrian of the Indian subcontinent now permits a more nuanced approach to temporal division than has been traditionally employed, and this will ultimately yield a more comprehensive understanding of the development of the northern Indian margin of Gondwana. Although the detrital zircon profiles share similar peaks with samples of similar

depositional age across the Indian craton and Himalaya, the distinct characteristics of the age distributions suggest a major influence from locally derived detritus.

The presence of *Wengania exquisita* on the Indian craton might have significant palaeogeographical implications because this is the first report of this taxon beyond the South China block. It is unclear whether this species is truly absent or simply has not yet been recognized elsewhere, despite intense research interest in rocks of this age around the globe in recent years. It is also not known certainly whether *W. exquisita* was pelagic or benthic, although a benthic habit is more consistent with its ~0.75 mm diameter, which is significantly larger than most phytoplankton. Given that no living red algae are pelagic (A.H. Knoll, pers. comm. 2015), there is no strong reason to expect a pelagic habit for this one. The presence of *W. exquisita* in the Birmania Formation thus provides a species-level biotic link between South China and the Indian craton and may suggest geographic proximity of the South China block to the Indian sector of core Gondwana at the time of deposition. This is consistent with a wide variety of geological evidence for linkage between the Indian craton, the Himalayan margin, and South China during the late Neoproterozoic/Cambrian, because rocks and biota of the Lesser Himalaya of that age are also closely allied to those of South China (Kumar, 1984; Jiang et al., 2003; Hughes et al., 2005; Tarhan et al., 2014), as are the Cambrian faunas of the Tethyan Himalaya (Jell and Hughes, 1997; Peng et al., 2009; Collette et al., 2011; McKenzie et al., 2011b; Popov et al., in press). This inference is further supported by similarities in detrital zircon signatures between Indian rocks of this age and temporal equivalents in South China (e.g. Wang et al., 2010; Xu et al., 2013, 2014).

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Figure Captions

Fig. 1. Bedrock outcrops of much of the northern part of the Indian craton, showing the position of the Birmania inlier at the western extent of this area, figure modified from McKenzie et al. (2013). (a) Detail of the western Vindhyan basin, the Aravalli-Delhi Orogenic Belt, and outcrops in western Rajasthan. (b) Positions of major tectonic units and depositional basins: ADOB = Aravalli-Delhi Orogenic Belt, RV = Rajasthan

Vindhyan, SV = Son Valley Vindhyan, CITZ = Central Indian Tectonic Zone, Ch = Chhattisgarh Basin, IB = Indravati Basin, Cu = Cuddapah Basin.

Fig. 2. (a) Outcrop of the phosphorite beds of the Birmania Formation in part of the Birmania inlier, showing different outcrop sectors previously designated as different “blocks”, and positions of the detrital and productive fossil samples analyzed herein. Inset shows geographic location of Birmania within Rajasthan. Modified from Srikantan et al. (1969) and Husain and Banerjee (1986). (b) Stratigraphic log of the Birmania inlier succession modified from Mathur and Chauhan (1995) and Muktinath (1969), showing position of the phosphatic bed from which samples BA1-BA5 were collected. Published thickness estimates for the Birmania Formation vary from less than 200 m to over 2000 m. v indicates volcanics, sh indicates shale, fs indicates sandstone, cs indicates coarse sandstone, cong indicates conglomerate, dolo indicates dolostone.

Fig. 3. Birmania Formation. (a-b) Site of BA2, BA3 collection. (a) Tightly folded phosphatic and silicified grainstone. (b) Detail of a phosphatic bed showing tan laminated grainstone with resistant dark layer of silicified phosphate. (c) Dolomudstone with flat lying cryptmicrobialite lamination.

Fig. 4. Multicellular algal fossils from the Birmania Formation. White scale bars are 100 µm long, black scale bars at 10 µm long. 6.7×141.0 etc. refer to the x,y coordinates of the fossil on the slide measured on an Olympus BX-51 microscope. (a–h) *Wengania exquisita* Zhang et al., 1998. (i–l) Fragments of multicellular algal fossils. (a) Fossil

(framed) among quartz grains (Qz), CMC IP72086, BA3, 6.7×141.0. (b–c) Successive magnifications of framed areas in (a–b), respectively, to show pseudoparenchymatous arrangement of cells. (d) Specimen with an extension (arrow), CMC IP72087, BA4, 7.3×116.2. (e–f) Successive magnifications of framed areas in (d–e), respectively, to show unordered cell arrangement (probably transverse section of pseudoparenchymatous thallus) and secondary phosphatic coating (arrow in f). (g–h) CMC IP72087, BA4, 13.8×106.7. (i–j) CMC IP72087, BA4, 13.5×134.5. (k–l) CMC IP72087, BA4, 14.8×105.6. (h, j, l) magnifications of framed areas in (g, i, k), respectively.

Fig. 5. Detrital zircon grain age spectra and concordia plots for two sandstone samples (BA2 and BA5) from the Birmania Formation. For locations see Fig. 2 and text.

Fig. 6. Pooled detrital zircon age spectra for the two Birmania Formation samples compared to pooled compilations of detrital zircon ages from the Marwar Supergroup, Aravalli–Delhi Orogenic Belt (ADOB), and Vindhyan successions. Data from the upper and lower Marwar Supergroup were partitioned by study to demonstrate regional variability in detrital zircon distributions (note the variation in relative abundances of distinct age populations between samples). Grey bar highlights the interval from 1.0–1.3 Ga (note the abundance 1.0–1.3 Ga zircon grains in the upper Vindhyan and upper ADOB, and the general lack of these grains in both Marwar and Birmania strata).

Figure 9

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