U-Pb age and Lu-Hf isotopic data of detrital zircons from the Neoproterozoic Damara Sequence: implications for Congo and Kalahari before Gondwana

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Abstract

The proximity of the Congo and Kalahari cratons during the Neoproterozoic breakup of the supercontinent Rodinia and during subsequent assembly of Gondwana is unclear. Neoproterozoic metasedimentary rocks from the rifted margins of Congo and Kalahari in the Damara Orogen yield distinctive detrital zircon U-Pb age distributions that correspond to the ages of prominent crustal components within the respective cratons. The most abundant zircons from Neoproterozoic strata deposited on the Congo margin give ages of 1150-1000 and 800-600 Ma, whereas, the most abundant zircons from the Kalahari margin strata range from 1350-1100 Ma. A 1350-1200 Ma detrital zircon population in the Kalahari margin strata is absent in the Damara-Congo strata. A prominent *c.* 1050-1000 Ma detrital zircon age population from Damara-Congo strata is nearly absent from the Damara-Kalahari strata, even though orogenic events of this age are found on both cratons. Damara strata on the Kalahari margin also lack detrital zircons with U-Pb ages of 900-600 Ma. The differences in detrital zircon age distributions are robust when comparing strata of the same age on both cratons, and remains so, even when younger, deeper water facies are excluded, which could have been

31 biased by other sediment sources. These data suggest that the Congo and Kalahari
32 cratons were not proximal in Rodinia, and did not establish their current relative positions
33 until the end of the Neoproterozoic when they were sutured together during the
34 collisional orogenies that formed Gondwana.

36 **1.1 Introduction**

37 The interval from the Neoproterozoic breakup of Rodinia to the Cambrian
38 assembly of Gondwana is one of the most critical periods in Earth history in terms of
39 geodynamics, paleoclimatic changes, and rapid development of biologic diversity
40 (Hoffman et al., 1998; Meert and Lieberman, 2008; Nance et al., 2014). The
41 configurations of continental fragments within Rodinia, the dispersal of these fragments,
42 and their subsequent amalgamation to form Gondwana, however, remain poorly
43 constrained (e.g., Li et al., 2008). In particular, the paleogeographic relationship between
44 two major components in the core of Gondwana, the Congo and Kalahari cratons (Fig. 1),
45 through the Neoproterozoic remains unclear (e.g., Tohver et al., 2006; Gray et al., 2008).

46 Diagnostic paleomagnetic data for Congo are limited and are absent for Kalahari,
47 (recent initial results of Hanson et al. (2011) suggest a high-latitude position for Kalahari),
48 which has led to contrasting hypotheses for the relationship between these two cratons.
49 Congo and Kalahari may have been located in different parts of Rodinia (or not part of
50 Rodinia) before suturing during the formation of Gondwana (e.g., Burke et al., 1977,
51 2003; Stanistreet et al., 1991; John et al., 2003; Kröner and Cordani, 2003; Tohver et
52 al., 2006; Gray et al., 2008). Alternatively, Congo and Kalahari could have been in close
53 proximity in Rodinia, separated by an intracratonic rift or Red Sea-like basin at Rodinia
54 breakup, and then sutured back together during amalgamation of Gondwana (e.g.,
55 Kröner, 1977; Martin and Porada, 1977; Miller, 1983; Hanson et al., 1994; Kampunzu et
56 al., 2000; Hanson, 2003; Jung et al., 2001). The Neoproterozoic configuration of the
57 Khomas Ocean and its closure to form the Damara-Zambezi Orogen (Prave, 1996; Gray
58 et al., 2008) are, therefore, critical to determining whether Gondwana was assembled
59 from large continents roughly resembling the modern continents or from numerous
60 smaller cratonic fragments, some of which were in Rodinia and some that were not. We
61 report U-Pb and Lu-Hf isotopic data from detrital zircons from metapsammites and Sm-

Nd isotopic data from metapelites from the Neoproterozoic Damara Sequence, which provide new constraints on the paleogeographic relationships between the Congo and Kalahari cratons.

1.2 Geologic Background

The components that formed the Gondwana supercontinent were sutured along a series of Neoproterozoic to Early Paleozoic orogenic belts (Fig. 1). These Pan-African/Brasiliano or Pan-Gondwanan orogenic belts are generally interpreted to have formed from ocean closure, accretion, and collision involving all of the continental fragments that comprised Gondwana (Burke et al., 1977; Veevers, 2003). Many of these continental fragments were previously part of Rodinia or located along relatively poorly defined positions around Laurentia, while others were probably scattered in the surrounding oceans (e.g. Hoffman, 1991; Weil et al., 1998; Dalziel et al., 2000; Powell et al., 2001; Meert and Torsvik, 2003; Kröner and Cordani, 2003; Pesonen et al., 2003; Tohver et al., 2006; Li et al., 2008). There are a variety of proposals for the locations of the Congo and Kalahari cratons within Rodinia, their relative positions with respect to Laurentia, as well as relative to each other (e.g., Powell et al., 1993, 2001; Wilson et al., 1997; Dalziel et al., 2000; Meert, 2003; Meert and Torsvik, 2003; Hanson 2003; Collins and Pisarevsky, 2005; Tohver et al., 2006; Gose et al., 2006; Gray et al., 2008). Congo and Kalahari have relatively similar tectonic and rift-drift sedimentary histories, which have been used as evidence that they were in roughly the same relative positions in Rodinia as they are now. The stratigraphic relationships, however, are not unique because Neoproterozoic rift to drift sequences on many other continents are similar in age, indicators of paleoclimate, and lithologic assemblage (e.g., Hoffman et al., 1998).

Neoproterozoic to Cambrian (Pan-African) orogenic belts in southern Africa (Fig. 2) record the closure of three ocean basins. The closure of the Adamastor Ocean between the Congo, Kalahari, and Rio de la Plata cratons (e.g., Brito Neves et al., 1999; Hoffman, 1999; Trompette, 1997; Kröner, 2001; Kröner and Cordani, 2003; Cordani et al., 2003; Goscombe et al., 2005; Foster et al., 2009) formed the Kaoko and Gariep belts (600-530 Ma) (Figs. 1 and 2). The East African Orogen (650-550 Ma) records the closure of the

93 Mozambique Ocean (e.g., Stern, 1994; Thomas et al., 2013; Satish-Kumar et al., 2013).
94 The Damara-Zambezi Orogen is among the younger of the composite Pan-African
95 orogenic belts and formed during closure of the Khomas Ocean and the suturing of the
96 Archean-Mesoproterozoic Congo and Kalahari cratons (Figs. 1 and 3). The Damara belt
97 joins the Kaoko and Gariiep belts in the west and continues into the Lufilian arc and
98 Zambezi belt (580-500 Ma) to the east, which in turns join with the East African Orogen
99 (e.g., Goscombe et al., 2000; Hanson, 2003; Hargrove et al., 2003; Meert, 2003; Gray et
100 al., 2008).

101 The Kalahari Craton contains large areas of crust older than 3.0 Ga, extensive
102 Paleoproterozoic orogenic belts (e.g., Magondi Belt), continental margin magmatic and
103 accretionary lithotectonic units (*c.* 1400 - 1200 Ma), a major Mesoproterozoic orogenic
104 belt (1240-1060 Ma; Namaqua-Natal), and a significant *c.* 1110 Ma largely mafic
105 magmatic province (Eglington and Armstrong, 2004; Becker et al., 2006; Jacobs et al.,
106 2008; Hanson et al., 2004; Hanson, 2003; Eglington 2006; Singletary et al., 2003;
107 Colliston and Schoch, 2013). In contrast, the Congo craton includes a smaller >3.0 Ga
108 component, large areas of Neoarchean, Paleoproterozoic, and early Mesoproterozoic
109 crust (De Waele et al., 2008; Thomas et al., 2013), and the distinctive *c.* 1375 Kibaran
110 magmatic province (Tack et al., 2010). Both Congo and Kalahari contain large, elongate,
111 late Mesoproterozoic orogenic belts (1300-1000 Ma) that overlap in age with the
112 Grenville Orogeny in Laurentia. Grenville temporal equivalents in southwestern and
113 southern Kalahari include the Namaqua-Natal belt of southwestern and southern Kalahari
114 (1240-1060 Ma), which; records the transition from accretionary to collisional events
115 over about 150 Ma (Singletary et al., 2003; Becker et al., 2006; Pettersson et al., 2007;
116 Jacobs et al., 2008; Bailie et al., 2007; Colliston and Schoch, 2013). The Choma-Kalomo
117 block and similar aged orogenic belts in Mozambique, along the northern margin of
118 Kalahari, range from 1350-1020 Ma (Hanson et al., 1988; Singletary et al. 2003; Manttari,
119 2008).

120 In the Congo craton the Irumide belt (1080-1000 Ma) extends along the southern
121 margin of the craton and possibly into the basement of the central Damara belt (Johnson
122 et al., 2007a; De Waele et al., 2006). The southwestern part of Congo also contains
123 1500-1400 Ma anorogenic igneous complexes and records granulite facies

metamorphism (Seth et al., 2003). Felsic igneous rocks related to rifting formed at c. 880 (e.g. Kafue Rhyolite) and c. 820 Ma along the southern margin (Zambezi belt) of Congo (Johnson et al., 2008; Hanson et al. 1994; Hanson, 2003). Extensive rhyolitic rocks with ages of about 746 Ma record rifting of the southern and western margins of Congo and initiation of Damara Sequence deposition (Hoffman et al., 1996, 1998). Along the southwestern Kalahari margin, post Namaqua-Natal rifting is recorded by 890-800 Ma alkaline igneous provinces (Frimmel et al., 2001; Hanson et al., 2011). Final rifting of Kalahari from Rodinia resulted in widespread rhyolitic magmatism at about 770-740 Ma (e.g., Frimmel et al., 1996, 2001; Borg et al., 2003). There is, therefore, significant overlap in the ages of Neoproterozoic felsic igneous rocks on both cratons.

Subsequent to this rifting, the 580-500 Ma Damara Orogen (Figure 3) formed as a doubly vergent orogen between the colliding Congo and Kalahari cratons with thrusting of carbonate platform and passive margin sequences to the north and south over the bounding cratons. The core of the orogen is dominated by granitic batholiths and high-T/low-P metamorphism. The location of the felsic magmatism, as well as the structural asymmetry, reflects primary subduction dipping beneath attenuated Congo craton (Barnes and Sawyer, 1980; Gray et al., 2008). The major components of the Damara Orogen include: 1) Paleo-Mesoproterozoic basement, including the Kamanjab Inlier of the Congo craton along the northern flank of the orogen, elongate basement-cored (Congo craton) structural domes in the Central Zone, and Kalahari basement in the Paleo-Mesoproterozoic Rehoboth Inlier in the south; 2) Neoproterozoic (c. 770-600) Damara Sequence shelf carbonates (Otavi Group) and deeper water turbidites (Swakop Group), 3) Neoproterozoic-Cambrian Mulden Group and upper Nama Group foreland basin (molasse) deposits of northern and southern Namibia, respectively (Miller, 1983, 2008; Prave, 1996; Hoffman et al., 1998; Hoffmann et al., 2004; Becker et al., 2006), and 4) widespread latest Neoproterozoic to Cambrian felsic plutonic rocks (e.g., McDermott et al., 1996).

Stratigraphically, the basal units of the Neoproterozoic Damara Sequence are rift-related coarse clastics of the Nosib Group (Fig. 4). In the core of the orogen the Nosib Group is overlain by turbiditic pelitic schists, psammites, and mafic schists of the Swakop Group, which contains two turbiditic carbonate formations that are correlated

with *c.* 750-735 Ma and *c.* 636 Ma diamictites on the Congo shelf (Hoffman et al., 1998; Hoffmann et al., 2004). Although the Damara Sequence was deposited over a similar time period on the Congo and Kalahari margins there are sufficient differences in facies to question direct stratigraphic correlation across conjugate margins (Kukla and Stanistreet, 1991; Miller, 2008).

Structurally, the Damara Orogen is asymmetric with the continental margin of Kalahari thrust under remnants of Khomas Ocean lithosphere and the Congo craton (Fig. 5) (Miller, 1983, 2008; Gray et al., 2006). A sliver of Khomas oceanic crust is preserved in the thin (200–300 m thick) shear zone-hosted, Matchless amphibolite, which extends along strike over most of the exposed part of the Damara belt (Figs. 3, 5) (Barnes and Sawyer, 1980). This unit consists of intensely deformed basalt, pillow basalt, and gabbro with tholeiitic geochemistry (Schmidt and Wedepohl, 1983), and is intercalated with banded chert (Killick, 2000). Discontinuous serpentinite boudins within the Uis Pass Line shear zone at the margin between the accreted turbidite fan sequences and passive margin sequences of the Kalahari Craton mark the Khomas suture (Barnes, 1983). The Kalahari flank of the former ocean basin is represented by craton-verging thrust systems within shelf carbonates (Hakos and Naukluft nappes in the Southern Margin zone) (Miller, 1983).

The deep water Khomas turbidite sequence is metamorphosed to form the Kuiseb schist and subdivided into two structural-metamorphic zones, referred to as the Central and Southern zones (Fig. 3). The Central zone (CZ) is underlain by attenuated Congo craton (Kukla and Stanistreet, 1991; Becker et al., 2006; Kisters et al., 2004). The Southern zone (SZ) schists and ophiolitic fragments make up the Khomas accretionary prism, which developed above the subduction interface during ocean closure between Congo and Kalahari (Kukla and Stanistreet, 1991; Miller 2008; Gray et al., 2007; Foster and Goscombe, 2013).

Rocks of the Khomas accretionary prism developed on oceanic crust of the Khomas Ocean, which presumably was connected to the Congo craton. There may, however, be some structural intercalation of turbidites scraped off the Kalahari side of the Khomas Ocean during development of the accretionary prism. The asymmetry of the Damara orogeny means that the leading edge of the Kalahari margin is beneath the

Khomas accretionary prism so that only part of the rift to drift sequence is exposed in the Southern Margin zone (SMZ) (shelf and rise facies) and Southern Platform (SP).

2.1 Methods

Zircons were separated from crushed sedimentary and metasedimentary rocks of the Damara Sequence using magnetic and density methods, mounted in epoxy, and then ground and polished to expose interiors. The polished grains were imaged using back-scattered electrons and cathodoluminescence (CL) to identify homogenous growth zones within grain interiors that were free of physical defects and inclusions. The U-Pb isotopic analyses were conducted on a Nu Plasma multicollector inductively coupled plasma source mass spectrometer (MC-ICP-MS) with a collector array optimized for U and Pb isotopes. Zircons were ablated in a He atmosphere with a New Wave 213-nm ultraviolet laser coupled to the MC-ICP-MS (Mueller et al., 2008). On-peak background measurements (20 s) were performed before each analysis in order to define the carrier gas blank, including ^{204}Hg contribution. Ablation proceeded for exactly 30 seconds. The isotopic data were acquired using the Nu-Instruments Time Resolved Analysis software. Data calibration and drift corrections were based on multiple ablations of the FC-1 zircon standard (Duluth Gabbro – Forest Center, 1098 Ma; Mattinson, 2010) bracketed between every 10 unknowns. Data were reduced using an in-house Excel spreadsheet. Representative age errors are based on the long-term reproducibility of FC-1 and are 2% for $^{206}\text{Pb}/^{238}\text{U}$ (2σ) and 1% for $^{207}\text{Pb}/^{206}\text{Pb}$ (2σ). Only grains with <10% discordance, and without a Damara Orogen metamorphic overprint (560-500 Ma), are considered in the following sections. Ages discussed below are based on $^{206}\text{Pb}/^{238}\text{U}$ if less than 1000 Ma, and $^{207}\text{Pb}/^{206}\text{Pb}$ for ages > 1000 Ma.

Lu-Hf isotopic analysis was performed on selected grains with concordant U-Pb data and growth zones (determined from CL images) large enough to contain the additional Lu-Hf laser pit. Lu-Hf analyses were also conducted with the Nu-Plasma MC-ICP-MS with each analysis consisting of a 20-second background measurement, one minute of measurement during ablation, and a one-minute purging period between analyses. The samples were calibrated against the reference zircon FC-1 (Black et al.,

2003; Woodhead et al., 2004), and the chondritic parameters of Bouvier et al. (2008) as described in Mueller et al. (2008).

Powdered whole-rock samples of metapelite were analyzed for Sm-Nd isotopic data using isotope dilution with a ^{150}Nd tracer added prior to dissolution in sealed Teflon vials. Isotopic compositions were determined by MC-ICP-MS methods, but otherwise followed the methods, including data reduction, in Ingle et al (2003).

3.1 Results

The sample suite is composed of Neoproterozoic sedimentary and metasedimentary rocks of the Damara Sequence from both the Congo and Kalahari passive margins (hereafter Damara-Congo and Damara-Kalahari, respectively). Samples include strata from the rift through the passive margin facies that record the separation of these two cratons from Rodinia or other continental fragments. Sample locations are included in Supplemental Table 1, and their stratigraphic position is noted in Fig. 4. The Damara-Congo samples include rift and passive margin strata from the shelf, slope and deep ocean floor (Fig. 4). Because of the abundance of carbonate-dominated lithologies, many of the Damara-Kalahari samples did not yield zircon. We were, however, able to extract zircons from a smaller number of rift to passive margin strata for the Kalahari shelf and slope (Fig. 4) that cover part of the same age range as the Congo margin samples (700-600 Ma).

A kernel density estimation plot (KDE) and histogram (plotted using DensityPlotter; Vermeesch, 2012) for zircon U-Pb data from each individual sample is included in Supplementary Figure 1. The two rift facies samples from the Central zone (Nosib Group: Etusis and Khan Formations) show identical detrital zircon age distributions. These samples feature a dominant age population at about 1030 Ma and a smaller number of grains with ages of about 2040 Ma (Fig 6). Figure 7 shows pooled KDEs for the post-rift, passive margin facies strata with depositional ages restricted to between about 700 and 600 Ma from individual structural zones of the Damara Orogen. Figure 8 compares all of the U-Pb zircon data from the Damara-Congo and Damara-Kalahari passive margin facies. The most abundant zircon populations from the Damara-Congo samples are composed of U-Pb ages between

1150-1000 Ma and 800-600 Ma, whereas the most abundant zircons from the Damara-Kalahari samples range in age from 1350 to 1100 Ma. KDE plots from both groups also show numerous zircons with late Paleoproterozoic and Neoproterozoic ages, with a gap between about 2100 and 2500 for the Damara-Congo distribution. Zircons from granitic gneiss that makes up the basement underlying the Etosis Formation in the Central zone give a concordia intercept U-Pb age of 1027 ± 10 Ma (sample DF09-43) (Fig. 9). This sample provides a basis for comparison with the detrital zircon populations in the overlying Etosis and Khan formations.

Lu-Hf isotopic data for zircons with concordant U-Pb data are included in Supplementary Table 2. The results from all grains are shown in a plot of initial ϵ_{Hf} (ϵ_{Hf_i}) and U-Pb age in Fig. 10. Lu-Hf data for the zircons from granitic gneiss sample DF09-43 from the Central zone basement are also included in Table 2 (data supplement). The ϵ_{Hf_i} values for the Archean and Paleoproterozoic grains from both the Congo and Kalahari margin samples range from -10 to +5, indicating an interaction between a juvenile, mantle component and an enriched, likely continental crustal component in all samples. The ϵ_{Hf_i} values of Mesoproterozoic grains from Kalahari and Congo also overlap and range from about -5 to +5, also indicating derivation from a mixture of mantle and crustal components. The majority of the Grenville-aged and Neoproterozoic zircons from the Congo margin samples give positive ϵ_{Hf_i} values from 0 to +12 (equal to the DM at that time), with the exception of a group of *c.* 1030 Ma grains from the rift facies (Etosis and Khan) samples in the Central zone and eight zircons with ages between 735 and 678 Ma.

The Sm-Nd isotopic data for metapelitic rocks is given in Supplementary Table 3. The nine samples north of the suture between the Damara-Congo (Southern Zone) and Damara-Kalahari (Southern Margin Zone) give a mean ϵ_{Nd} value of -8.9, whereas those to the south (*n*=3) give a mean of -13.9. With one exception, the depleted mantle model ages north of the suture range from about 1.4-1.6 Ga, and those south of the suture are >1.8 Ga (Fig. 11).

4.1 Discussion and Conclusions

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278 The pooled KDE plots for each structural zone (Fig. 8) include data from all of
279 the samples except the basal, rift-facies, Nosib Group strata from the Central zone,
280 and the single sample from the Klein Kuppe Formation in the Southern zone. The
281 Nosib Group (Etusis and Khan formations) samples show U-Pb detrital zircon age
282 distributions (Fig. 6) and Lu-Hf isotopic compositions (Fig. 10) identical to the
283 underlying basement in the Central zone (*c.* 1030 Ma with *c.* 2040 Ma inheritance;
284 Fig. 9; Supplementary Table 1) so were likely derived from local sources. The
285 Swakop, Khomas, and Hakos Group samples in Figs. 7 and 8 comprise passive
286 margin strata that range in depositional age from about 730 to 600 Ma. The
287 Damara-Congo passive margin strata from the NZ, CZ, and SZ show similar late
288 Paleoproterozoic detrital zircon age distributions with a group at about 2000 Ma and
289 a small number of zircons between 1750 and 1800 Ma. Only the SZ samples show a
290 significant Archean population centered at about 2600 Ma. The late
291 Paleoproterozoic detrital zircons as well as a small number of early
292 Mesoproterozoic zircons may have been derived from Congo craton basement
293 similar to that exposed in the Kamanjab and Kunene inliers (Becker et al., 2006;
294 Miller, 2008); the *c.* 2600 Ma grains could also be from the southern part of the
295 Congo craton (Angolian craton) (e.g., reworked Archean in the Kamanjab inlier gives
296 U-Pb ages of about 2600 Ma; Seth et al., 1998). The ϵ_{Hf_i} values for the
297 Paleoproterozoic and Archean grains are consistent with sources that include
298 reworked Archean crust older than about 2600 Ma and some juvenile crustal
299 materials.

300 There are many options in the Kalahari craton for the early Paleoproterozoic
301 and Archean zircons, including the Magondi belt and Rehoboth province (Fig. 2)
302 (Becker et al., 2004; van Schijndel et al., 2013). Late Paleoproterozoic zircons
303 (1850-1750 Ma) from the Hakos Group samples of the Damara-Kalahari margin
304 could also have provenance in the Rehoboth province or other belts of similar age in
305 the Kalahari craton (e.g., Becker and Schalk, 2008a; van Schijndel et al., 2013). The
306 abundant 1400-1200 Ma zircons in the SMZ and lower SF samples probably reflect

proximal sources in the Rehoboth inlier, Sinclair Supergroup, and related continental margin arc plutons (e.g., the c. 1380-1370 Ma Hammerstein tonalite and Aunis tonalite) (Becker et al., 2006; Becker and Schalk, 2008a; Miller 2008), and/or the Choma-Kalomo block (Hanson et al., 1988; Singletary et al., 2003).

There are abundant detrital zircons with U-Pb ages of 1100-1000 Ma in samples from all of the structural zones. The SZ, CZ, and NZ distributions are dominated by zircons with ages of c. 1080-1030 Ma, whereas the SMZ and SF samples are dominated by grains with ages around 1100-1080 Ma with a much smaller number of 1030-1000 Ma zircons. Ages for the SMZ and SF zircons are consistent with sources in the proximal and distal parts of the Namaqua belt, which includes the voluminous felsic plutons of the Gamsberg granite suite (1080-1060 Ma zircons; Becker and Schalk 2008b). The slightly younger Mesoproterozoic zircons in the SZ, CZ, and NZ passive margin strata are similar in age to the CZ basement and volcanic rocks (Miller, 2008). Zircons of this age also have widespread potential sources in the Irumide belt and in the Zambezi belt of the southeastern Congo craton (Fig. 2; e.g., Hanson, 2003; Hargrove et al., 2003; DeWaele et al., 2006). With the exception of the two Nosib Group samples, the ϵ_{Hf} values of the 1100-1000 Ma Damara-Congo zircons indicate a range of source rocks from essentially pristine mantle-derived magmas to magmas that interacted with older, more enriched sources, which suggests that they were not derived from the CZ basement.

The Damara-Kalahari Neoproterozoic sediments in the SMZ and SF lack detrital zircons with ages of c. 900-600 Ma, whereas these are nearly as abundant as the 1050-1000 Ma grains in the CZ and NZ (and SZ) of the Congo margin. The c. 880 and c. 750 Ma grains from the Damara-Congo strata are similar in age to the rift-related felsic plutons and rhyolites of the Zambezi belt and Lufillian arc, as well as the Nosib Group sediments in the NZ and Northern foreland (NF) of the Damara belt (Hanson et al. 1994; Hargrove et al., 2003; Johnson et al., 2008; Miller, 2008). The abundant c. 670-600 Ma zircons in the SZ, CZ, and NZ samples from the Congo-Damara suite may have been derived from the East African orogen, either by drainage across the craton or margin-parallel transport in the Khomas Ocean. The

Lu-Hf isotopic data for most of the Neoproterozoic zircons suggest a significant juvenile component in the host magmas, which would be consistent with either rift volcanism or long transport from Neoproterozoic juvenile terranes in northeastern Africa (Fig. 12)

While matching of the detrital zircon distributions with known components of the respective cratons is not completely definitive, it is striking that the main clusters of U-Pb ages are similar in the Damara-Congo NZ, CZ, and SZ samples, and then again in the Damara-Kalahari SMZ and SF samples. With the pooled number of grains analyzed from the Kalahari samples (n=129) the chance of not having found a representative of the 900-750 Ma zircon population in the samples analyzed, if they were present at the 5% level, is <1% (Dodson et al., 1988; Vermeesch, 2004). If individual samples are taken separately the chance of missing the younger populations is greater (9% chance of missing a population at the 5% level for the 47 or 48 concordant grains in the SMZ samples), but the similarity between all of the Damara-Kalahari samples suggests that pooling the data is reasonable.

The ϵ_{Nd} values and depleted mantle model ages of the pelitic schists suggest a more evolved crustal source for the fine-grained sediments in the Southern Margin zone (Damara-Kalahari) and a less evolved average source for the Southern zone (Damara-Congo). These results suggest that the average source of the detritus changes at the suture between these zones and likely reflects the generally more ancient lithosphere of the Kalahari craton compared to the Congo craton.

These independent results (U-Pb zircon age distributions and Sm-Nd isotopic data) for the metapelites and metapsammites strongly suggest that the passive margin sedimentary provenance of the Neoproterozoic Damara Sequence of Congo and Kalahari were different, at least from about 730-600 Ma. This is consistent with models suggesting that these two passive margins evolved separately and were not formed from rifting a once continuous area within Rodinia. It appears, therefore, that Congo was separate from Kalahari, and was not in proximity of Kalahari until the Neoproterozoic-Cambrian assembly of Gondwana. The fact that late Mesoproterozoic and Neoproterozoic basement rocks are widespread in Kalahari

(e.g., Namaqua-Natal) and that zircons from these provinces are not present in the detrital zircon distributions in the SMZ and SF samples means that alternative explanations are possible, including: 1. The Neoproterozoic zircons in the NZ, CZ, and SZ strata were likely derived from the East African orogen and transported across the Congo craton or around the Congo craton by long-shore currents only after Congo and Kalahari had rifted apart (i.e. after ~730 Ma). This would imply, however, that the rift between them had evolved into a large enough ocean basin that the sediment containing the Neoproterozoic detrital zircons did not cross the basin floor; or 2. The zircon provenance of the strata was so locally derived that the detrital zircon distributions reflect only the proximal sources of the respective margins and regions like Namaqua were not part of the drainage systems feeding the Kalahari shelf at this time. The likelihood of only local provenance influencing the distributions is small given that some of the Neoproterozoic detrital zircons do not reflect the ages of intrusions or crust within the southwestern part of the Congo craton and must have been derived from further east or northeast in Gondwana (present coordinates).

The sample from the Klein Kuppe Formation (Southern zone) is the one exception to the unique detrital zircon distributions within the Damara-Congo and Damara-Kalahari suites of samples. The Klein Kuppe psammites are distinct from the Kuiseb Formation turbidites in composition and assemblage. The detrital zircon distributions from sample DF09-01 are more similar to those of the Southern Margin zone and lack the Neoproterozoic zircons of the Kuiseb Formation samples. This formation makes up part of the southern-most section of the Southern Zone and so could have been derived from the Kalahari shelf-rise and then scraped off during subduction of the Khomas Ocean and intercalated into the toe of the Khomas accretionary prism (Kukla and Stanistreet, 1991; Gray et al., 2006) prior to Damara collision between the cratons.

Figure 13 shows a reconstruction at 750 Ma of cratonic nuclei that rifted from Rodinia, or were independent from Rodinia that is consistent with extant paleomagnetic data (Gray et al., 2008), and the detrital zircon data from the Neoproterozoic strata in the Damara belt. In this reconstruction the Congo and

Kalahari cratons are separated by some 30° of latitude with an intervening ocean of unknown dimensions. This reconstruction is not consistent with intracratonic rifting between these cratons based on, for example, correlation of the Nosib rhyolites of the Congo margin and the Rosh Pinah volcanics of the Kalahari margin (Frimmel and Frank, 1998). It is, however, difficult to constrain the paleogeography at this time because there are no reliable, published, late Neoproterozoic poles from the Kalahari craton and several alternative hypotheses have been proposed. Collins and Pisarevsky (2005), for example, argued for a Kalahari-West Australia connection based partly on the presence of overlapping Grenville-age events in the Northampton Block (Australia) and the Namaqua-Natal belts (Kalahari craton). Both the Kalahari craton and Australia have reliable palaeomagnetic poles of Grenville-age (1050-1100 Ma; Meert and Torsvik, 2003; Pesonen et al., 2003). These poles show a latitudinal offset between the West Australia and Kalahari of more than 30 degrees. For this reason Gray et al. (2008) reconstructed the Kalahari craton in proximity to the Congo-São Francisco craton, but detached from it. The position of the Congo and São Francisco cratons in Fig. 13 is based on the c. 755 Ma Mbozi complex pole (Meert et al., 1995).

The provenance indicators from the detrital zircon and Nd-Sm isotopic data from the Damara Orogen suggest that from the perspective of southern Africa assembly of Gondwana took place through suturing of microcontinental blocks and not a simple amalgamation of previously accreted East and West Gondwana cratons. The numerous isolated continental shelf environments that developed after breakup of Rodinia may have facilitated the gradual radiation of Ediacaran faunas leading to the “Cambrian explosion” (e.g., Meert and Lieberman, 2008). Final amalgamation of North (São Francisco-Congo-India) and South (Kalahari-Antarctica) Gondwana via closure of a Khomas-Southern Mozambique Ocean during the Damara-Zambezi Orogeny (580-530 Ma) (e.g., Gray et al., 2008) is consistent with the Damaran detrital zircon and Sm-Nd data, and suggests that final suturing of Gondwana occurred along the composite Kuunga through Damara orogens (e.g., Meert, 2003).

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Figure Captions

Figure 1. Map of Gondwana (after Gray et al., 2006) showing the locations of Pan-African orogenic belts in southern Africa. RP, Rio de la Plata craton; SF Sao Francisco craton.

Figure 2. Map of major geologic provinces and Precambrian orogenic belts in southern Africa (compiled after Goscombe et al., 2000 and Hanson, 2003).

Figure 3: Simplified geologic map of Namibia and the Damara Orogen (after Gray et al., 2006). Sample locations are indicated by the dots and numbers referenced to the sample name. Abbreviations: SF, Southern Foreland; SMZ, Southern Margin Zone; SZ, Southern Zone; CZ, Central Zone; NZ, Northern Zone; NF, Northern Foreland; OkSZ, Okahandja shear zone; OmSZ, Omaruru shear zone; AF, Autseib fault. A-B-C-D denote the endpoints of segments of the cross section in Figure 5.

Figure 4. Stratigraphic correlation diagram for Neoproterozoic Damara Sequence in the inland branch of the Damara Orogen based on data from Hoffmann et al. (2004), Halverson et al. (2002), Hoffmann (1983), and Miller (1983 and 2008) for the Congo province zones, and Bekker et al. (2004), Hoffmann (1989), and Miller (2008) for the Kalahari province zones.

Figure 5. Geological cross section of the Damara Orogen showing the underthrust margin of Kalahari under the Congo shelf and floor of the former Khomas Ocean (modified from Gray et al., 2006, Foster and Goscombe, 2013, and references therein).

Figure 6. Kernel density estimation curve and histograms of detrital zircon U-Pb data from the Neoproterozoic rift facies Nosib Group in the Central zone of the Damara Orogen.

Figure 7. Kernel density estimation curve and histograms of detrital zircon U-Pb data from the 730-600 Ma passive margin facies of the Damara Sequence from the NZ, CZ, SZ, and SMZ/SF. Data shown are for U-Pb analyses that are <10% discordant: $^{207}\text{Pb}/^{206}\text{Pb}$ ages for analyses >1000 Ma, and $^{206}\text{Pb}/^{238}\text{U}$ ages are analyses <1000Ma.

Figure 8. Kernel density estimation curve and histograms of detrital zircon U-Pb data from pooled Damara Sequence samples from the margins of the Congo and Kalahari cratons. Data shown are for U-Pb analyses that are <10% discordant: $^{207}\text{Pb}/^{206}\text{Pb}$ ages for analyses >1000 Ma, and $^{206}\text{Pb}/^{238}\text{U}$ ages for analyses <1000Ma. Shaded fields highlight some of the key populations from the Congo and Kalahari margins.

Figure 9. U-Pb Concordia plot for igneous zircons from a granitic gneiss that forms the basement in the Central zone.

Figure 10. Plot of initial ϵ_{Hf} against the U-Pb age of detrital zircons from the Damara Sequence.

Figure 11. Plot of whole rock Sm-Nd depleted mantle models ages for Neoproterozoic metapelites from the Southern Margin zone (Damara-Kalahari) and Southern zone (Damara-Congo).

Figure 12. Basement tectonic provinces of African showing the location of Archean-Proterozoic cratons and Pan-African orogenic belts (modified from Begg et al., 2009).

Figure 13. Reconstruction of the Congo and Kalahari cratons at about 750 Ma based on Gray et al. (2008). Continental fragments are AZ, Azania; SF, São Francisco; RP, Río de la Plata; WA, West Africa; Kal, Kalahari.