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A review of the Mesoproterozoic to early Palaeozoic magmatic and tectonothermal history of south–central Africa: implications for Rodinia and Gondwana

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Abstract: This paper provides a review of the tectonic evolution of central–southern Africa from Mesoproterozoic to earliest Palaeozoic times, using available geological information and a robust U–Pb zircon database. During the late Mesoproterozoic, the southern margin of the Congo–Tanzania–Bangweulu Craton was characterized by suprasubduction-zone magmatism and the accretion of arc and microcontinental fragments. Magmatism within the adjacent Irumide Belt formed by recycling of older continental crust. Ophiolite blocks, possibly part of an olistostromal mélange, are present in a Neoproterozoic sequence overlying the Irumide Belt, and the occurrence of high-pressure/low-temperature subduction-zone metamorphism and protracted Neoproterozoic suprasubduction-zone magmatism demonstrates that there was an ocean to the south (present-day coordinates) of the Congo–Tanzania–Bangweulu Craton until the amalgamation of Gondwana at 550–520 Ma, indicating that the Congo–Tanzania–Bangweulu Craton was not part of Rodinia. On the basis of their different ages and styles of magmatism, the Mesoproterozoic Kibaran Belt, Choma–Kalomo Block and Irumide Belt are not components of the same orogen, therefore precluding a sub-Saharan-wide, linked 'Kibaran' (*sensu lato*) orogenic event. Evidence is presented to illustrate that the Congo–Tanzania–Bangweulu and Kalahari Cratons developed independently until their final collision during the Pan-African Orogeny along the Damara–Lufilian–Zambezi Orogen at *c*. 550–520 Ma.

Keywords: Gondwana, Rodinia, Pan-African Orogeny, south–central Africa, tectonic evolution.

Like other continental regions, Africa comprises several Archaean cratonic nuclei that are 'stitched' together by younger orogenic belts. These belts record the history of break-up, accretion and collisional orogenesis that eventually resulted in the present configuration of the African continent. In this paper, we are concerned with the sub-Saharan region, which is known to consist of Archaean nuclei sutured by orogenic belts of Palaeoproterozoic, Mesoproterozoic and Neoproterozoic to earliest Palaeozoic age (Fig. 1, inset). It has been suggested that the Palaeo- and Mesoproterozoic belts may record the participation of these Archaean nuclei in the supercontinents Columbia (Rogers & Santosh 2002) and Rodinia (e.g. McMenamin & Schulte-McMenamin 1990; Hoffman 1991; Moores 1991; Dalziel 1992) respectively, and that the Neoproterozoic to earliest Palaeozoic belts record the formation and assembly of Gondwana (Shackleton 1996). However, details of several of these connections, especially those in pre-Gondwanan times, remain poorly constrained and locally disputed. The key to understanding the amalgamation history of this region lies in unravelling the complex tectonic history of the various orogenic belts, some of which have undergone penetrative structural and thermal reworking following their initial post-orogenic stabilization, rendering their interpretation especially cryptic. To improve understanding of these belts, the following first-order goals must be achieved.

(1) Discrimination between basement and juvenile material, such as sediments deposited in continental or oceanic basins, ophiolite suites, island arcs or continental-margin arcs.

(2) The robust dating of juvenile material (e.g. ophiolite formation and obduction, arc formation and accretion), and constraints on the timing of sedimentation.

(3) Defining the nature (collisional, accretionary, extensional) and precise timing of magmatism and peak orogenesis and the delineation of the late orogenic thermotectonic history (e.g. extensional collapse).

(4) The evaluation and backstripping of variably penetrative overprinting deformation and/or metamorphism.

Until recently, the understanding of most of the Mesoproterozoic and Neoproterozoic belts of southern Africa was at a rudimentary level. However, a flurry of recent publications, especially those presenting new, high-precision geochronology, *P–T* determinations and petrogenetic studies, have shed new light on various aspects of their tectonic evolution, prompting us to attempt a synthesis of the available data in the light of our own recent and continuing research.

Published geochronology

From the 1960s to the 1980s, the majority of published geochronological data for the African subcontinent was based on the Rb/Sr isotopic system (e.g. Cahen *et al*. 1984). Recent high-precision U–Pb dating by sensitive high-resolution ion microprobe (SHRIMP) and single-zircon isotope dilution-thermal ionization mass spectrometry (ID-TIMS) indicates that many of these Rb/Sr dates have been reset and are thus not amenable to straightforward interpretation. Therefore, this discussion is limited to U/Pb zircon or monazite ages, principally those determined by SHRIMP or ID-TIMS methods

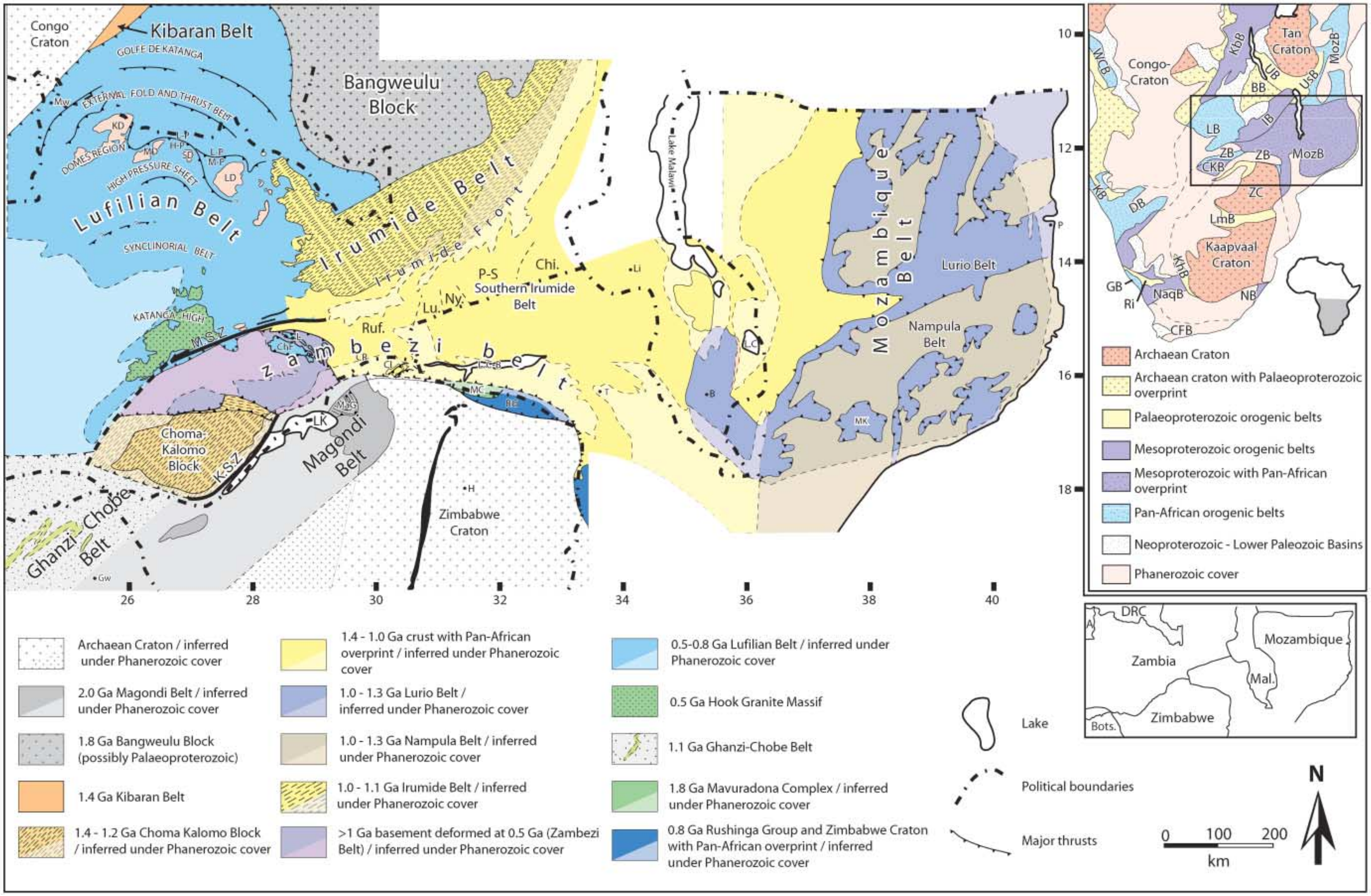


Fig. 1. Simplified geological map of central–southern Africa showing the distribution of tectonic features discussed in text. A tectonic map of sub-Saharan Africa is shown as an inset. The map of Africa is after Hanson (2003). Abbreviations for main map: B, Blantyre; Chi., Chipata Terrane; ChF, Cheta Formation; CI, Chewore Inliers; CR, Chowe River Region; Gw, Gweta; H, Harare; H-P, high-pressure thrust slice; KD, Kabompo Dome; K.S.Z, Kariba Shear Zone; L, Lusaka; L.C, Lake Chilwa; L.C.B, Lake Cabora Bassa; LD, Luwishi Dome; Li, Lilongwe; LK, Lake Kariba; L-P, low-pressure thrust slice; Lu., Luangwa Terrane; MaG, Makuti Group; MD, Mwombeshi Dome; MK, Mugeba Klippe; MC, Mavuradohna Metamorphic Suite; M-P, medium-pressure thrust slice; M.S.Z, Mwembeshi Shear Zone; Mw, Mwunilunga; Ny., Nyimba Terrane; P, Pemba; P-S, Petauke–Sinda Terrane; RG, Rushinga Group; Ruf., Rufunsa Terrane; SD, Solwezi Dome; T, Tete. Abbreviations for inset maps: A, Angola; BB, Bangweulu Block; Bots., Botswana; CFB, Cape Fold Belt; CKB, Choma Kalomo Block; DB, Damara Belt; DRC, Democratic Republic of Congo; GB, Garip Belt; GCB, Ghanzi Chobe Belt; IB, Irumide Belt; KB, Kaoko Belt; KbB, Kibaran Belt; KhB, Kheis Belt; LB, Lufilian Belt; LmB, Limpopo Belt; Mal., Malawi; MozB, Mozambique Belt; NaqB, Namaqua Belt; NB, Natal Belt; Ri, Richtersveld Terrane; UB, Ubendian Belt; UsB, Usagaran Belt; WCB, West Congo Belt; ZB, Zambezi Belt; ZC, Zimbabwe Craton.

(Tables 1 and 2). In addition, in the case of ages determined by the single-crystal Pb/Pb evaporation 'Kober' method, only those that are backed up by SHRIMP or TIMS analyses, which indicate they are concordant (e.g. Kröner *et al*. 2001), are used. Additional information on the timing of crust-forming events and high-grade metamorphism provided by whole-rock or mineral Sm–Nd and Lu–Hf isotopic ages is also discussed.

Mesoproterozoic tectonic history

In south–central Africa, three Mesoproterozoic terranes are recognized: the Kibaran Belt, Irumide Belt (and its reworked extension referred to here as the Southern Irumide Belt), and the Choma–Kalomo Block. These have traditionally been considered as components of a single, long-lived (*c*. 400 Ma) 'Kibaran-age' orogenic event, referred to as the Kibaran Orogeny, an interpreta-

Table 1. *List of 140 high-precision (errors at the 2σ level), Mesoproterozoic, igneous crystallization U–Pb ages used in text and Figure 3 to infer the tectonic evolution of central–southern Africa during the Mesoproterozoic*

		Dating method	Age (Ma)	Comment	Reference
	Kibaran Belt s.s.				
1	Kisele Monzogranite Gneiss	SHRIMP	1386 ± 8		Kokonyangi *et al*. 2004
2	Kisele Monzogranite Gneiss (metamorphic zircon)	SHRIMP	1079 ± 14	Upper intercept age	
3	FwiFwi Leucomonzogranite	SHRIMP	1372 ± 10		
4	Nyangwa Monzogranite	SHRIMP	1383 ± 5		
5	Kungwe–Kalumengonogo Monzogranite	SHRIMP	1377 ± 10		
6	Kabonvia Granodiorite	SHRIMP	1385 ± 7		
7	Mutanga Amphibole Norite (Musongati Massif, DB1)	SHRIMP	1374 ± 14		Tack *et al*. 2002
8	Rumeza Granite 63.865	SHRIMP	1383 ± 17		Tack, pers. comm.
9	Mugere Granite KI6684	SHRIMP	1379 ± 10		
10	Mugere Migmatitic Gneiss KI21	SHRIMP	1380 ± 8		
11	Kiganda Granite KI1	SHRIMP	1371 ± 7		
12	Muramba Granite KI14	SHRIMP	1380 ± 6		
13	Kilimbi–Muzimu Orthogneiss KI20	SHRIMP	1373 ± 6		
14	Bukirasazi Granite (A-type granite, LT7)	SHRIMP	1205 ± 19		
15	Kasika Granite (Sn granite, KI22)	SHRIMP	987 ± 6		
	Irumide Belt s.s.				
16	Lwakwa Granite	TIMS	1087 ± 11		Ring *et al*. 1999
17	Mwenga Granite	TIMS	1119 ± 20		
18	Mutangoshi Gneissic Granite (MTGG2, NE)	SHRIMP	1055 ± 13		De Waele, pers. comm, recalculated from De Waele *et al*. 2003*b*
19	Mutangoshi Gneissic Granite (MTGG1, NE)	SHRIMP	1029 ± 9		
20	Luswa Syeno-granite (LW2, NE)	SHRIMP	943 ± 5		
21	Porphyritic late tectonic granite (SER64, SW)	SHRIMP	1036 ± 13		
22	Porphyritic late tectonic granite (SER53, SW)	SHRIMP	1034 ± 5		
23	Mununga Quarry Granite (ZM36, central)	SHRIMP	1025 ± 10		
24	Serenje Quarry Granite (SQG, SW)	SHRIMP	1024 ± 9		
25	Sasa Granite (SASA2, SW)	SHRIMP	1016 ± 14		
26	Chilubanama Granite (MTG4, NE)	SHRIMP	1010 ± 22		
27	Chilubanama Granite (LW1, NE)	SHRIMP	1005 ± 21		
28	Kaunga Granite	TIMS	970 ± 5		Daly 1986*a*
29	Porphyritic late tectonic granite (CC5, SW)	SHRIMP	1038 ± 17		De Waele, pers. comm.
30	Porphyritic late tectonic granite (CC8, SW)	SHRIMP	1035 ± 12		
31	Porphyritic late tectonic granite (CHL5, central)	SHRIMP	1016 ± 17		
32	Biotite granite gneiss (CHT6, SW)	SHRIMP	1005 ± 7		
33	Porphyritic late tectonic granite (FW1, SW)	SHRIMP	1038 ± 58		
34	Porphyritic late tectonic granite (KK1, SW)	SHRIMP	1003 ± 31		
35	Porphyritic late tectonic granite (KN2A, SW)	SHRIMP	1031 ± 14		
36	Biotite granite gneiss (KN5, SW)	SHRIMP	1053 ± 14		
37	Porphyritic late tectonic granite (KN7, SW)	SHRIMP	1048 ± 10		
38	Biotite granite gneiss (KN8, SW)	SHRIMP	1022 ± 16		
39	Porphyritic late tectonic granite (MH4, SW)	SHRIMP	1017 ± 19		
40	Porphyritic late tectonic granite (MH9C, SW)	SHRIMP	1029 ± 14		
41	Porphyritic late tectonic granite (ND1, SW)	SHRIMP	1023 ± 7		
42	Porphyritic late tectonic granite (ND4, SW)	SHRIMP	1031 ± 5		
43	Porphyritic late tectonic granite (ND5, SW)	SHRIMP	1028 ± 7		
44	Fukwe River Migmatite (SER67)	SHRIMP	1021 ± 16		

(*Continued*)

Table 1. *Continued*

		Dating method	Age (Ma)	Comment	Reference
	Southern Irumide Belt				
45	Chewore Ophiolite plagiogranite (sample SJ106.1)	SHRIMP	1393 ± 22		Oliver *et al*. 1998
46	Kaourera Arc meta-dacite (sample SJ220)	SHRIMP	1082 ± 7		Johnson & Oliver 2004
47	Kadunguri Whiteschists	SHRIMP	1066 ± 21		Johnson, unpubl. data (reported by Johnson & Oliver 2004)
48	Chewore Inlier Granulite Terrane (sample ADC)	SHRIMP	1071 ± 8		Goscombe *et al*. 2000
49	Chewore Inlier Zambezi Terrane orthogneiss (sample AF)	SHRIMP	1083 ± 8		
50	Charnockite associated with Chipera gabbro/anorthosite	TIMS	1050 ± 20		Barr, unpubl. data (reported by Hanson 2003)
51	Garnet–spinel–cordierite gneiss (Chipata Gneiss)	TIMS	1046 ± 3	Monazite	Schenk & Appel 2001
52	ZM004 meta-dacite, Chowe River	SHRIMP	1070 ± 3		Johnson *et al*. 2004
53	ZM007 meta-dacite, Chongwe River	SHRIMP	1088 ± 20		
54	CH6 banded mafic gneiss, Chowe River	SHRIMP	1051 ± 12		
55	CH7 meta-tuff, Chowe River	SHRIMP	1064 ± 15	Xenocryst	
56	CH7 meta-tuff, Chowe River	SHRIMP	1037 ± 8	Xenocryst	
57	CH9 meta-dacite, Chowe River	SHRIMP	1040 ± 21		
58	CH9 meta-dacite, Chowe River	SHRIMP	1105 ± 22	Xenocryst	
59	CH10 Kspar augen gneiss, Chowe River	SHRIMP	1094 ± 2	Upper intercept age	
60	CH10 Kspar augen gneiss, Chowe River	SHRIMP	1105 ± 9	Xenocryst	
61	Porphyritic granite (EP26, Petauke–Sinda Terrane)	LA-ICP-MS	1125 ± 15		Cox, pers. comm.
	Zambian Zambezi Belt				
62	Kafue Rhyolite	TIMS	879 ± 19	Upper intercept age	Wilson *et al*. 1993
63	Mpande Gneiss	TIMS	1106 ± 19	Upper intercept age	Hanson *et al*. 1988*a*
64	Ngoma Gneiss	TIMS	820 ± 7	Upper intercept age	
65	Lusaka Granite	TIMS	846 ± 68	Upper intercept age	Barr *et al*. 1977
66	Granite intruding metasedimentary succession	TIMS	942 ± 30	Upper intercept age	Recalculated from Barr *et al*. 1977
67	Nchanga Granite	SHRIMP	877 ± 11		Armstrong *et al*. 1999
68	Munali Hills Granite	TIMS	1092 ± 4		Katongo *et al*. 2005
	Lufilian Belt				
69	Lapilli tuff, Mwashya Group (sample S11)	SHRIMP	1018 ± 27	Xenocryst	Rainaud *et al*. 2003
70	Lapilli tuff, Mwashya Group (sample S11)	SHRIMP	1056 ± 59	Xenocryst	
71	Lapilli tuff, Mwashya Group (sample S11)	SHRIMP	1267 ± 94	Xenocryst	
72	Lapilli tuff, Mwashya Group (sample S11)	SHRIMP	1273 ± 46	Xenocryst	
73	Mwunilunga Mafic Volcanics (samples 18C and 18C2)	SHRIMP	765 ± 5		Key *et al*. 2001
74	Mwunilunga Mafic Volcanics (sample 51D)	SHRIMP	735 ± 5		
75	Hook Granite (sample H1)	TIMS	559 ± 18	Upper intercept age	Hanson *et al*. 1993
76	Hook Granite (sample H2)	TIMS	566 ± 5	Upper intercept age	
77	Hook Rhyolite (sample H3)	TIMS	538 ± 1.5		
78	Post-tectonic granite (sample H4)	TIMS	533 ± 3		
79	Mwembeshi Rhyolite (sample H5)	TIMS	551 ± 19	Upper intercept age	
	Northern Kalahari Basement				
80	Masaso Metagabbro	TIMS	849 ± 2		Mariga *et al*. 1998
81	Mavuradona Met. Complex gabbro (unspecified)	TIMS	843 ± 7		Müller *et al*. 2001
82	Mavuradona Met. Complex gabbro (unspecified)	TIMS	837 ± 43		
83	Masoso Leucomigmatite (Zim-616)	TIMS	870 ± 1	Upper intercept age	Vinyu *et al*. 1999
84	Basal Rushinga Igneous Complex (Zim-610)	TIMS	805 ± 11	Upper intercept age	
85	Chironga Gneiss (ZIM97-53a)	TIMS	795 ± 2		Hargrove *et al*. 2003
86	Ocellar Gneiss (ZIM97-137b)	TIMS	1052 ± 1		
87	Pegmatite within the Ocellar Gneiss (ZIM97-135b)	TIMS	1058 ± 19		
88	Gnt-bearing augen orthogneiss in Ocellar Gneiss (ZIM97-138)	TIMS	870 ± 4		
89	Unspecified	TIMS	808 ± 4		Hanson, unpubl. data (reported by Hanson 2003)
90	Makuti Group (Makuti Gneiss)	TIMS	831 ± 6	Upper intercept age	Hanson *et al*. 1998
91	Makuti Group (sample Zim 20)	Pb/Pb evap.	794 ± 1		Dirks *et al*. 1999
92	Makuti Group (sample Zim 23)	Pb/Pb evap.	737 ± 1		
93	Makuti Group (sample Zim 24)	Pb/Pb evap.	764 ± 1		
94	Makuti Group (sample TS8)	Pb/Pb evap.	854 ± 1		

(*Continued*)

Table 1. *Continued*

		Dating method	Age (Ma)	Comment	Reference
95	Makuti Group (sample TS11)	Pb/Pb evap.	794 ± 1		
	Chomo Kalomo Block				
96	Siasikabole Granite (sample U10)	TIMS	1352 ± 14		Hanson *et al*. 1988*b*
97	Zongwe Orthogneiss (sample U5)	TIMS	1343 ± 6	Upper intercept age	
98	Chilala Orthogneiss (sample U12)	TIMS	1285 ± 64		
99	Semahwa Granite (sample U11)	TIMS	1198 ± 6		
100	Medium-grained two-mica granite (CK4)	SHRIMP	1174 ± 27		Bulambo *et al*. 2004
101	Foliated biotite granite (CK10)	SHRIMP	1188 ± 11		
102	Foliated biotite granite (CK12)	SHRIMP	1177 ± 70		
103	Foliated biotite granite (CK25)	SHRIMP	1181 ± 9		
104	Augen orthogneiss (CK13)	SHRIMP	1368 ± 10		
	Ghanzi–Chobe Belt				
105	Borehole CKP10a (sample CKP10A-1) Kwando Complex	TIMS	1107 ± 1		Singletary *et al*. 2003
106	Rhyolite Lava (sample 96-85) Ngezumbu River	TIMS	1106 ± 4	Upper intercept age	
107	Kavimba Granite	TIMS	1107 ± 2		
108	Borehole CKP 11 (sample CKP 11-8) granite	TIMS	1107 ± 0.5		
109	Borehole NG2 (sample NG2-1) Roibok Complex granite gneiss	TIMS	717 ± 2		
110	Kgwebe Rhyolite, Mabeleapodi Hills	TIMS	1106 ± 2		Swartz *et al*. 1996
111	Borehole CKP4-Ghanzi Group meta-sediments	SHRIMP	1104 ± 16		Kampunzu *et al*. 2000
	Southern Malawi				
112	Granitic gneiss (MA1)	Pb/Pb evap.	603 ± 1		Kröner *et al*. 2001
113	Diorite granulitic gneiss (MA2)	Pb/Pb evap.	645 ± 1		
114	Trondhjemite gneiss (MA3)	Pb/Pb evap.	583 ± 1		
115	Quartz monzonite gneiss (MA4)	Pb/Pb evap.	577 ± 1		
116	Charnockitic gneiss (MA6)	Pb/Pb evap.	590 ± 1		
117	Pelitic paragneiss (MA8)	Pb/Pb evap.	577 ± 1		
118	Pelitic paragneiss (MA8)	SHRIMP	576 ± 11		
119	Charnoenderbitic gneiss (MA9)	Pb/Pb evap.	1012 ± 1		
120	Biotite–hornblende gneiss (MA10)	Pb/Pb evap.	999 ± 1		
121	Biotite–hornblende gneiss (MA13)	Pb/Pb evap.	576 ± 1		
122	Biotite–hornblende gneiss (MA14)	Pb/Pb evap.	1040 ± 1		
123	Charnoenderbitic gneiss (MA15)	Pb/Pb evap.	555 ± 1		
124	Charnoenderbitic gneiss (MA15)	Pb/Pb evap.	744 ± 1	Xenocryst	
125	Biotite–hornblende gneiss (MA16)	Pb/Pb evap.	667 ± 1		
126	Biotite–hornblende gneiss (MA16)	SHRIMP	664 ± 27		
127	Biotite gneiss (MA17)	Pb/Pb evap.	710 ± 1	Xenocryst	
128	Biotite gneiss (MA17)	Pb/Pb evap.	769 ± 1	Xenocryst	
129	Biotite gneiss (MA17)	Pb/Pb evap.	775 ± 1		
	Lurio and Nampula Belts				
130	Mocuba Complex	SHRIMP	1028 ± 7		Costa *et al*. 1994
131	672 CJ 003	Pb/Pb evap.	1048 ± 1		Kröner *et al*. 1997
132	673 G J008	Pb/Pb evap.	1040 ± 1		
133	586 G C001	Pb/Pb evap.	1148 ± 1		
134	586 G C001	Pb/Pb evap.	1297 ± 1	Xenocryst	
135	587 G J003	Pb/Pb evap.	1067 ± 1		
136	618 G J007	Pb/Pb evap.	1114 ± 1		
137	Migmatitic granite gneiss (sample MS5)	SHRIMP	1094 ± 13		Kröner *et al*. 2001
138	Leucocratic granite (sample MS6)	SHRIMP	1009 ± 13		
139	Sample NHF, Nhansipfhe Megacrystic Granite Gneiss	SHRIMP	1112 ± 18		Manhica *et al*. 2001
140	Sample CVGN, Chimoio Granodiorite Gneiss	SHRIMP	1108 ± 12		

Specimen names are shown where available. The geochronological reference numbers shown to the left of the data are used in the main text as superscript numbers to refer to this table.

tion still favoured by some workers (e.g. Hanson 2003; Kampunzu *et al*. 2003). Furthermore, since the timing of Kibaran orogenesis (*sensu lato*; *s.l.*) has been considered to overlap with events associated with the formation of Rodinia, some workers have automatically related Kibaran (*s.l.*) events to amalgamation of this supercontinent (e.g. Kampunzu *et al*. 2003).

To simplify the discussion in this paper, the Mesoproterozoic belts of south–central Africa have been subdivided into 10 subregions on the basis of tectonic criteria (Fig. 2). For each tectonic subregion, the data are plotted as both histograms (50 Ma bin size) and probability density curves (Sircombe 2004), to produce a series of geochronological 'bar codes'

Table 2. *List of 32 high-precision (errors at the 2σ level), Neoproterozoic, metamorphic U–Pb dates used in the text and Figure 8 to infer the Neoproterozoic tectonic evolution of central–southern Africa*

		Dating Method	Age	Comment	Reference
	Zambezi Belt (Congo–Tanzania–Bangweulu Craton)				
141	Solwezi kyanite gneiss	ICP-MS	530 ± 2	Monazite	John 2001
142	Kaourera Arc meta-dacite (sample SJ220)	SHRIMP	517 ± 5		Johnson & Oliver 2004
143	Chewore Inlier Granulite Terrane (sample ADC)	SHRIMP	526 ± 17		Goscombe *et al*. 2000
144	Kalumbina metasediments, Kabompo Dome	SHRIMP	548 ± 8	Monazite	Steven & Armstrong 2003
145	Kalumbina metasediments, Kabompo Dome	SHRIMP	531 ± 21	Rutile	
146	Banded Mafic Gneiss, Chowe River (CH6)	SHRIMP	573 ± 1		Johnson *et al*. 2004
147	Meta-tuff, Chowe River (CH7)	SHRIMP	569 ± 5		
148	Meta-dacite, Chowe River (CH9)	SHRIMP	576 ± 13		
149	Fukwe River Migmatite, Irumide Belt (SER67)	SHRIMP	550 ± 20		De Waele, pers comm.
	Zambezi Belt (Kalahari Craton)				
150	Ocellar Gneiss (sample 97-135b)	TIMS	534 ± 24		Hargrove *et al*. 2003
151	Masaso Leucomigmatite (Zim-616)	TIMS	535 ± ?	Lower intercept	Vinyu *et al*. 1999
152	Sample Zim-615b	TIMS	536 ± ?		
153	Basal Rushinga Igneous Complex (Zim-610)	TIMS	521 ± 71	Lower intercept	
154	Makuti Gneiss	TIMS	542 ± 10	Lower intercept	Hanson *et al*. 1988*a*
155	Mavuradonha Complex (unspecified)	Pb/Pb	557 ± 1		Müller *et al*. 2000
156	Mavuradonha Complex (unspecified)	SHRIMP	539 ± 21		
	Southern Malawi				
157	Charnockitic gneiss (MA8)	Pb/Pb	575 ± 1		Kröner *et al*. 2001
158	Charnockitic gneiss (MA8)	SHRIMP	572 ± 9		
159	Felsic granulite (MA12)	Pb/Pb	549 ± 1		
160	Felsic granulite (MA12)	SHRIMP	547 ± 10		
161	Biotite–hornblende gneiss (MA13)	SHRIMP	564 ± 4		
162	Biotite gneiss (MA17)	Pb/Pb	551 ± 1		
	Lurio and Nampula Belts				
163	674 G T001	Pb/Pb	614 ± 1		Kröner *et al*. 1997
164	674 G T001	SHRIMP	615 ± 7		
165	674 G T002	Pb/Pb	613 ± 1		
166	Deformed pegmatite	TIMS	538 ± 2	Pb/Pb age	Jamal *et al*. 1999
167	Unspecified	TIMS	542 ± 5	Pb/Th age	
168	Metamorphic leucosome	TIMS	586 ± 3		
169	Unspecified	TIMS	523 ± 41	Lower intercept	
170	Interlayered paragneiss	TIMS	538 ± 46	Lower intercept	
171	Mafic granulites	TIMS	606 ± 34	Lower intercept	
172	Mafic granulites	TIMS	544 ± 96	Lower intercept	

Specimen names are shown where available. The geochronological reference numbers shown to the left of the data are used in the main text as superscript numbers to refer to this table.

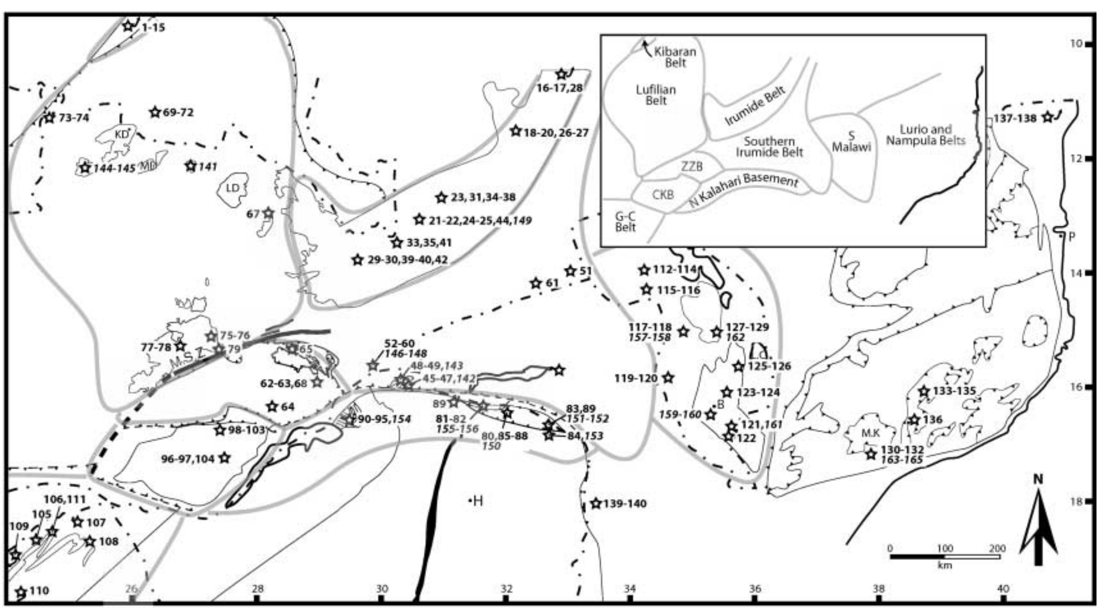
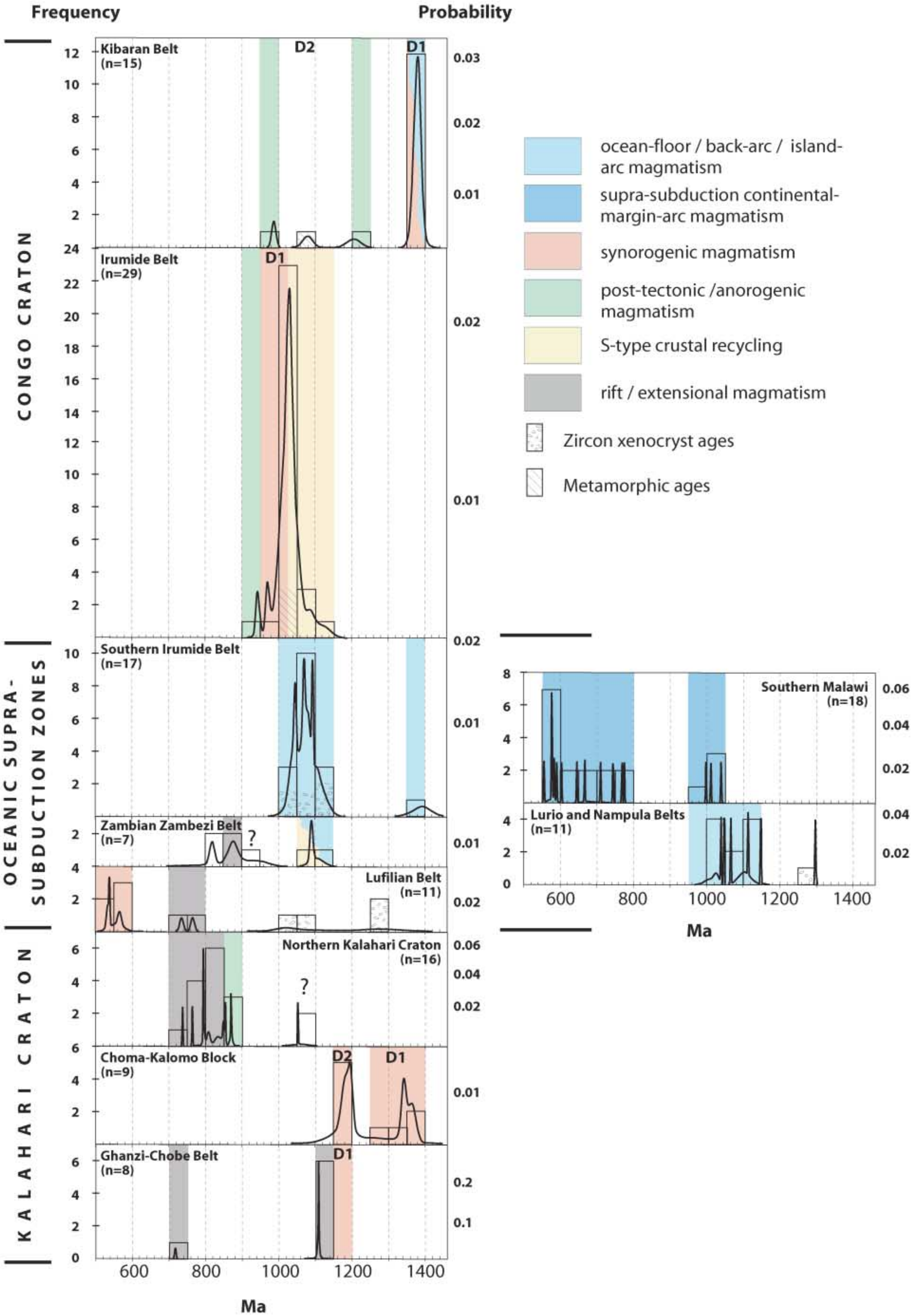


Fig. 2. Locations of high-precision U–Pb igneous crystallization (140) and metamorphic (32) ages. The corresponding age data are presented in Table 1, and superscript numbers that follow age ranges in the main text refer to these referenced geochronological data. CKB, Choma–Kalomo Block; G-C Belt, Ghanzi–Chobe Belt; ZZB, Zambian Zambezi Belt.


Frequency
Probability
CONGO CRATON
OCEANIC SUPRA-SUBDUCTION ZONES
KALAHARI CRATON
Kibaran Belt (n=15)
D2
D1
Irumide Belt (n=29)
Southern Irumide Belt (n=17)
Zambian Zambezi Belt (n=7)
Lufilian Belt (n=11)
Northern Kalahari Craton (n=16)
Choma-Kalomo Block (n=9)
Ghanzi-Chobe Belt (n=8)
600
800
1000
1200
1400
Ma
ocean-floor / back-arc / island-arc magmatism
supra-subduction continental-margin-arc magmatism
synorogenic magmatism
post-tectonic /anorogenic magmatism
S-type crustal recycling
rift / extensional magmatism
Zircon xenocryst ages
Metamorphic ages
Southern Malawi (n=18)
Lurio and Nampula Belts (n=11)

associated with specific types of magmatism for each subregion (Fig. 3).

Description of subregions

Kibaran Belt sensu stricto (s.s.). The southern tip of the NE-trending Kibaran Belt extends into the map area (Fig. 1). It is composed of deformed supracrustal rocks and orogenic granitoids (Rumvegeri 1991; Tack *et al*. 1994, 2002), and separates the Congo Craton to the NW from the Tanzania–Bangweulu Craton to the SE. A basement rise (2.2–1.9 Ga Rusizian Rise) subdivides the belt into an Eastern External Domain, underlain by the Archaean Tanzanian Craton, and a Western Internal Domain underlain by the Palaeoproterozoic Rusizian basement. The transition zone between the Western Internal Domain and Eastern External Domain is the site of numerous layered mafic complexes and associated A-type granitoids, dated at 1383–1371 Ma[7–13] in the northern part of the belt and at 1385–1370 Ma[1,3–6] farther south (superscripts refer to the geochronological reference numbers in Tables 1 and 2). Younger magmatism includes small-volume, A-type granitoid and mafic bodies that were emplaced along shear zones at *c*. 1205 Ma[14] (Tack *et al*. 1999, 2002) and intrusion of regionally widespread *c*. 980 Ma[15] Nb–Ta-bearing pegmatites and Sn-bearing granites (e.g. Brinckmann *et al*. 2001; Tack *et al*. 2002). Recent U/Pb dating of a low Th/U metamorphic zircon from the Kisele Monzogranite has yielded an upper intercept age of 1079 Ma ± 14 Ma[2], interpreted to date the time of sillimanite-grade D_2 deformation and metamorphism (Kokonyangi *et al*. 2001, 2004). The explanation for the *c*. 400 Ma magmatic and metamorphic history of the Kibaran Belt remains enigmatic. The *c*. 1370 Ma magmas have been interpreted as either (1) mantle and crustal melts resulting from lithospheric delamination following D_1 orogenic collapse (Tack *et al*. 1994) or (2) suprasubduction-zone mafic magmas associated with granites, formed by mixing of the mafic magmas with the products of dehydration melting of metasediments (Kokonyangi *et al*. 2004). Analyses of detrital zircon populations suggest that sediments forming the Eastern External Domain were deposited at *c*. 1.78 Ga and sourced from the Tanzanian Craton and Ubendian Belt, whereas parts of the Western Internal Domain sediments were deposited post-1.4 Ga by reworking of Eastern External Domain supracrustal rocks (Cutten *et al*. 2004).

Irumide Belt s.s.. The NE-trending Irumide Belt is predominantly underlain by granitoid rocks, but a Palaeoproterozoic platformal quartzite–pelite succession, the Muva Supergroup, is widespread in the NW (Daly & Unrug 1982). Deposition of the Muva Supergroup has been dated at *c*. 1880 Ma (De Waele & Fitzsimons 2004) and has recently been correlated with the Itremo Group in central Madagascar (Cox *et al*. 2004). In the NW of the belt, near the Irumide Front (Fig. 1), foliated granitoids (inferred to be part of the basement to the Muva Supergroup) have yielded ages in the range 2050–1930 Ma (De Waele *et al*. 2003*b*, 2004*a*, *b*). In contrast, granitoid magmatism elsewhere in the Irumide Belt occurred between 1660–1550 Ma and 1119–950 Ma[16–44]. These late Palaeoproterozoic to early Mesoproterozoic and late Mesoproterozoic to early Neoproterozoic granitoids have bulk-rock geochemical signatures and highly negative $\epsilon_{Nd}(t)$ values indicating that they formed entirely by recycling of continental crust (De Waele *et al*. 2003*a*). Polyphase deformation in the Irumide Belt was coeval with formation of the subassemblages cordierite–andalusite and sillimanite–K-feldspar in pelite (Mapani 1999), indicating relatively low-pressure–high-temperature (LP–HT) metamorphic conditions. Direct dating of the metamorphism in two areas has yielded U–Pb ages of 1046 ± 3 Ma on metamorphic monazite (Schenk & Appel 2001) and 1020 ± 7 Ma and 1004 ± 20 Ma on low Th/U zircon overgrowths (De Waele *et al*. 2003*b*), compatible with the ages of syn- to post-tectonic granitoids.

Tectonic windows in the Zambian Zambezi and Lufilian Belts. In the Zambezi Belt of southern Zambia, two tectonic windows through the Neoproterozoic cover expose the Irumide basement consisting of megacrystic K-feldspar augen granite (locally known as the Mpande Gneiss and Munali Hills Granite) that have been dated at 1106 ± 19 Ma[63] and 1092 ± 4 Ma[68], respectively. Whole-rock major, trace and rare earth element analyses of the Mpande Gneiss and Munali Hills Granite show aspects of both S-type and arc-type magmatic environments (Katongo *et al*. 2004, 2005). The granitoids may be juvenile products formed in an active continental-margin arc, or they may have inherited their arc-like signatures from older crustal material from which they were recycled.

In the adjacent SE Lufilian Belt, inherited zircon xenocrysts (mostly 1.1–1.6 Ga[69–72], but including grains with ages up to 3.0 Ga) are present in Neoproterozoic lapilli tuffs (Rainaud *et al*. 2003). These xenocrysts display a range of ages essentially indistinguishable from those of the Palaeoproterozoic basement in the central Irumide Belt (De Waele *et al*. 2004*a*), suggesting that this basement may underlie part of the Lufilian Belt.

Southern Irumide Belt. The Southern Irumide Belt (Fig. 1) in SE Zambia and Mozambique represents part of the Irumide belt *s.l.* that has been extensively reworked under upper-amphibolite-facies conditions during Pan-African orogenesis. Although there are few areas where reliable data are available, it appears that this part of the Irumide Belt is composed of a series of distinct terranes of various ages and tectonic setting (Fig. 1) (Mapani *et al*. 2004). The Rufunsa Terrane, including the Chewore Inliers and Chongwe River area, consists of a collage of accreted oceanic, island-arc and continental-margin-arc lithologies and associated sedimentary successions (Oliver *et al*. 1998; Johnson & Oliver 2000, 2004). The oldest dated lithology, at *c*. 1393 ± 22 Ma[45], is a plagiogranite from a marginal-basin ophiolite suite that that has been linked to an oceanic realm to the south (present-day coordinates) of the Tanzania–Bangweulu Craton (Oliver *et al*. 1998; Johnson & Oliver 2000, 2004). Suprasubduction-zone island-arc magmatism in this region occurred between 1105 and 1037 Ma[46–60] (Table 1, Fig. 3). On its NE margin, the Rufunsa Terrane is in contact with the Luangwa Terrane, which consists of polydeformed metasediments and dioritic gneisses with isoclinally folded amphibolite layers (former dykes). One of these dioritic gneisses has yielded an Archaean age of *c*. 2.6 Ga (Cox *et al*. 2002), suggesting that the Luangwa Terrane may be an accreted microcontinental fragment. The structurally overlying Nyimba Terrane is predominantly composed of marble and calc-silicate, and is itself overlain by the Petauke–Sinda Terrane, which consists of a suite of calc-

Fig. 3. Histogram and probability density curves for the 140 high-precision ages for each of the subregions shown in Figure 2. Scales on the ordinates of the histograms refer to the number of measurements and the calculated probabilities. The associated data are tabulated and referenced in Table 1.

alkaline plutonic rocks interpreted as an accreted island arc (Mapani *et al*. 2004). One of these lithologies has been dated at 1125 ± 15 Ma[61]. The easternmost terrane of the Southern Irumide Belt in Zambia, the Chipata Terrane, consists of retrogressed mafic and felsic granulites of supracrustal origin. Its tectonic setting is at present unknown. Mapani *et al*. (2004) suggested that terrane accretion was contemporaneous with deformation and metamorphism in the Irumide Belt.

Southern Malawi and Lurio–Nampula Belts (Mozambique). Calc-alkaline granitoids with gneissic fabrics and late Mesoproterozoic emplacement ages predominate in the Lurio and Nampula Belts (previously supergroups) of Mozambique (Pinna *et al*. 1993; Kröner *et al*. 1997; Jamal *et al*. 1999; Jamal & De Wit 2004) and southern Malawi (Kröner *et al*. 2001). In Mozambique, these granitoids were emplaced between 1148 and 1009 Ma[130–133,135–140] (Table 1, Fig. 3). The granitoids lack inherited Archaean or Palaeoproterozoic zircons, and show both positive and mildly negative $\epsilon Nd(t)$ values indicating derivation in a mature island-arc-type tectonic setting (Jamal & De Wit 2004). These rocks lack any evidence for Mesoproterozoic high-grade metamorphism that might be associated with continental collision, but are strongly overprinted by Pan-African tectonometamorphism. In southern Malawi, calc-alkaline granitic orthogneisses with crystallization ages of 1040–999 Ma[119,120,122] (Table 1, Fig. 3) have Nd isotopic signatures compatible with their formation as juvenile melts in a suprasubduction-zone, continental-margin-arc setting (Kröner *et al*. 2001). Neoproterozoic arc magmatism (775–555 Ma[112–118,121,123–129]) (Table 1, Fig. 3) has also been recognized in this region (Kröner *et al*. 2001).

Choma–Kalomo Block. The Choma–Kalomo Block (CKB in Fig. 1) on the NW margin of the Kalahari Craton is underlain by an unnamed supracrustal sequence consisting of pelitic schist with amphibolite layers and boudins cut by a batholithic granitoid complex. The oldest dated units in the batholith were emplaced between 1352 and 1343 Ma[96,97] (Table 1, Fig. 3), providing a minimum age for deposition of the supracrustal rocks. Younger intrusions have yielded ages of 1285 ± 64 Ma[98] and 1198 ± 6 Ma[99] (Table 1, Fig. 3), indicating a prolonged magmatic history. Ages of the two main magmatic pulses have recently been confirmed by SHRIMP dating of zircon, which has yielded a crystallization age of 1368 ± 10 Ma[104] for the main batholith, and ages between 1188 ± 11 Ma and 1174 ± 27 Ma[100–103] for cross-cutting intrusions. The batholith is inferred to have been emplaced as a series of 'syntectonic to late syntectonic plutons' (Hanson *et al*. 1988*a*), thereby also providing an approximate age for the main fabric-forming event in the Choma–Kalomo Block. Whole-rock Sm/Nd isotope data for the *c*. 1350 Ma granites suggest that, although the magmas interacted with sialic crust on ascent, they were not derived from an Archaean source (Hanson *et al*. 1988*a*).

Ghanzi–Chobe Belt. The present northwestern margin of the Kalahari Craton is the site of a late Mesoproterozoic intracontinental rift containing mafic and felsic volcanic rocks and small granitic intrusions, known as the Ghanzi–Chobe Belt (Fig. 1). The rift stratigraphy includes tholeiitic Kgwebe Volcanic series and rhyolite and granitic stocks (Kampunzu *et al*. 1998). The presence of Palaeoproterozoic zircon xenocrysts in volcanic units (Singletary *et al*. 2003) suggests that they were intruded through the Palaeoproterozoic Magondi Group. All dated felsic units have tightly constrained crystallization ages of 1107–1104 Ma[105–108,110,111] (Table 1, Fig. 3), and have been correlated with the volcanic rocks in the Upper Sinclair sequence (Hoal 1993) that occur along strike to the SW in Namibia and have a U/Pb age of 1094 ± 20 Ma (Hegenberger & Burger 1995). Greenschist-facies metamorphism of the Kgwebe Volcanic series and the unconformably overlying Neoproterozoic sediments in the Ghanzi–Chobe Belt took place during the late Neoproterozoic to Palaeozoic Damara Orogeny (Kampunzu *et al*. 2000). However, meta-sediments exposed in a borehole (CKP11), and in the Kwando Complex farther west have amphibolite-facies fabrics correlated with those that formed at 1.2–1.15 Ga in the Choma–Kalomo Block (Singletary *et al*. 2003).

Northern Kalahari margin. The northern Kalahari margin is a complex subregion of multiply tectonized Archaean basement gneisses interleaved with Neoproterozoic metasediments of the Rushinga Group, structurally overlain by the high-grade Zambezi Allochthonous Terrane. This terrane consists of granitoid gneisses, known as the Ocellar Gneiss and Masaso Metamorphic Suite, and the mafic Mavuradonha Complex (Barton *et al*. 1993; Vinyu *et al*. 1999; Hargrove *et al*. 2003). Components of the Ocellar Gneiss have been dated at *c*. 1050 Ma[86,87], but the majority of the unit has crystallization ages of *c*. 870 Ma[88]. The geochemical character of the Mesoproterozoic granitoids in the Zambezi Allochthonous Terrane is unknown. The Mavuradonha Complex (formerly Metamorphic Suite) is a Palaeoproterozoic metamorphosed layered gabbro–anorthosite intrusion (Müller *et al*. 2000, 2001; Hargrove *et al*. 2003), but other components of the Zambezi Allochthonous Terrane have yielded Neoproterozoic ages in the range 870–840 Ma[80–83], which are interpreted to represent the time of A-type magmatism and granulite-facies metamorphism (Hargrove *et al*. 2003).

A subcontinent-wide linked Kibaran (s.l.) orogeny?

Figure 3 shows the geochronological 'bar codes' for the various subregions described above, colour-coded for the type of magmatism. Although data for each subregion are limited in comparison with other well-studied Mesoproterozoic orogenic belts, they are sufficient to define and characterize the first-order signatures of the main magmatic and orogenic episodes. It is clear from Figure 3 that the timing of tectonothermal events and the nature of magmatism are different for the Kibaran Belt, Irumide Belt (*s.l.*) and the Choma–Kalomo Block, and thus we conclude that there was no Kibaran (*s.l.*) orogeny. Irrespective of whether *c*. 1370 Ma magmatism in the Kibaran Belt (*s.s.*) represents a post-D_1 delamination event or island-arc magmatism, it is clear that orogenesis, magmatism and post-orogenic collapse in both the Kibaran Belt (*s.s.*) and Choma–Kalomo Block had terminated some 100 Ma before initiation of magmatism in the Irumide Belt. In the Kibaran Belt (*s.s.*), D_2 medium-pressure and -temperature (MP–MT) Barrovian-style metamorphism (Kokonyangi *et al*. 2001*b*) at 1079 ± 14 Ma[2] was broadly coeval with the onset of magmatism in the Irumide Belt at 1087 ± 11 Ma[16], allowing a far-field tectonic linkage between the two belts. Nevertheless, both precede low-pressure–high-temperature (LP–HT) metamorphism in the Irumide Belt (D_1), which began at *c*. 1040 Ma.

Contemporaneous suprasubduction-zone, continental-margin-arc and island-arc magmatism between 1150 and 1000 Ma in the Rufunsa Terrane of the Southern Irumide Belt, and in the Lurio and Nampula Belts of Mozambique and in Southern Malawi, implies the existence of an oceanic basin to the south of the Congo–Tanzania–Bangweulu Craton. However, *c*. 1087–953 Ma magmatism in the Irumide Belt, which took place on the southern margin of the Congo–Tanzania–Bangweulu Craton and thus overlapped in time with suprasubduction magmatism elsewhere, formed by *in situ* melting of the continental margin. The tectonic processes involved are currently unknown, but we note that the youngest dated suprasubduction magmatic lithology in an accreted terrane in the Southern Irumide Belt is *c*. 1040 Ma[56,57], which is contemporaneous with LP–HT metamorphism in the Chipata Terrane at 1046 $\pm$ 3 Ma[51]. Thus cessation of arc magmatism occurred at about the same time as initiation of LP–HT metamorphism in the Irumide Belt and some parts of the Southern Irumide Belt, suggesting that magmatism was switched off as a result of accretion of arc and/or microcontinent terrane(s) to the Congo–Tanzania–Bangweulu Craton. A lack of MP or HP metamorphism of Mesoproterozoic age in the terranes of the Irumide Belt (*s.l.*) suggests that terrane accretion did not terminate in continental collision.

In the Kalahari Craton, Mesoproterozoic magmatism took place in the Ghanzi–Chobe Belt, the Zambezi Allochthonous Terrane, and the Choma–Kalomo Block (Fig. 3). Taking into account the type of magmatism, the bar codes for each subregion show significant differences, both between each other and in comparison with those from subregions adjacent to the Congo–Tanzania–Bangweulu Craton. Considering the presence of oceanic arc terranes to the south of the Congo–Tanzania–Bangweulu Craton at this time, this lack of correlation between the Kalahari and Congo–Tanzania–Bangweulu Cratons is not surprising and reinforces the interpretation that they evolved independently during the Mesoproterozoic.

Were the Congo–Tanzania–Bangweulu and/or Kalahari Cratons part of Rodinia?

There is no consensus on the configuration of the proposed Rodinia supercontinent, which is thought to have been produced by late Mesoproterozoic (*c*. 1.1–1.0 Ga) collisional events (e.g. McMenamin & Schulte-McMenamin 1990; Hoffman 1991). On the assumption that the Kibaran and Irumide Belts formed contemporaneously in this interval, as parts of an anastomosing orogenic network, previous workers have included both the Congo–Tanzania–Bangweulu and Kalahari Cratons in Rodinia reconstructions as: (1) components of a single southern African block (Dalziel 1997; Unrug 1997; Kampunzu *et al*. 2003; Torsvik 2003); (2) individual, but closely related cratons (Weil *et al*. 1998; Meert 2002; Meert & Torsvik 2003); or (3) independent cratons (Dalziel *et al*. 2000; Powell *et al*. 2001; Kröner & Cordani 2003; Loewy *et al*. 2003; Pesonen *et al*. 2003; Pisarevsky *et al*. 2003).

It is clear that: (1) formation of the Kibaran Belt (*s.s.*) preceded the assembly of Rodinia by at least 200 Ma and so it cannot have been directly related to the formation of the late Mesoproterozoic supercontinent; (2) evidence from the Southern Irumide Belt, Southern Malawi and the Lurio and Nampula Belts indicates an oceanic realm to the south of the Congo–Tanzania–Bangweulu margin at this time. Geological constraints on the other margins of the Congo–Tanzania–Bangweulu similarly lack evidence for collisional orogenesis at the time of Rodinia assembly (Kröner & Cordani 2003; Pisarevsky *et al*., 2003). Sparse palaeomagnetic data, mainly from the South American counterpart of the Congo–Tanzania–Bangweulu Craton, the São Francisco Craton, also suggest that the Congo–Tanzania–Bangweulu Craton was not a component of Rodinia, but existed as a separate distal cratonic entity (Fig. 4; Cordani *et al*. 2003; Kröner & Cordani 2003; Pisarevsky *et al*. 2003; Johnson & Oliver 2004).

Late Mesoproterozoic tectonic activity on the margins of the Kalahari Craton is well exposed in the Namaqua and Natal Belts. Although outside the geographical scope of this paper, both of these belts are relatively well known and a wealth of geochronological information is available (summarized by Hanson 2003), including a well-dated, robust, Rodinian palaeomagnetic pole from the Kalahari Craton itself (Powell *et al*. 2001). On the basis of this result, Powell *et al*. (2001) and Pisarevsky *et al*. (2003) concluded that the Kalahari Craton, along with its counterpart Grunehogna fragment in East Antarctica, may have been situated within Rodinia near the western margin of the Australian Craton until 800–750 Ma (Fig. 4). Furthermore, magmatic events associated with the break-up of Rodinia appear to have occurred in a similar time frame to those in the Kalahari Craton (outlined below).

Neoproterozoic to Early Palaeozoic tectonic history

The evidence presented above supports the interpretation that the Congo–Tanzania–Bangweulu and Kalahari Cratons evolved independently during the Mesoproterozoic, and that the Congo–Tanzania–Bangweulu–São Francisco Craton, with an oceanic realm on its present-day southeastern margin, was not part of the Rodinia supercontinent (Fig. 4). Thus the development of the Neoproterozoic belts separating the two cratons (i.e. the trans-continental Pan-African Damara–Lufilian–Zambezi Orogen) should record the collision between the Congo–Tanzania–Bangweulu and Kalahari Cratons and preserve evidence for Neoproterozoic ophiolites and suprasubduction magmatism.

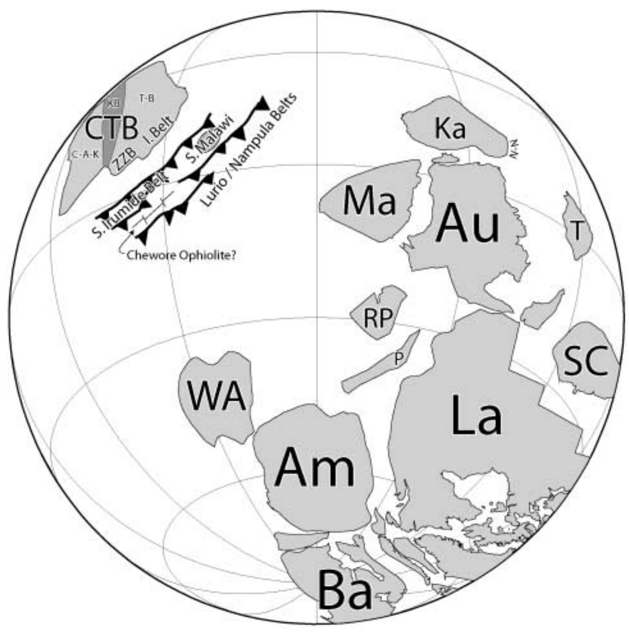


Fig. 4. Rodinia reconstruction after Pisarevsky *et al*. (2003), illustrating tectonic environments of the Congo–Tanzania–Bangweulu and Kalahari Cratons at *c*. 1.0 Ga. Am, Amazonian Craton; Au, Australia; Ba, Baltica; C-A-K, Congo–Angola–Kasai Craton; CTB, Congo–Tanzania–Bangweulu Craton; I. Belt, Irumide Belt; Ka, Kalahari Craton; KB, Kibaran Belt; La, Laurentia; Ma, Mawson Craton; N-N, Namaqua–Natal Belt; P, Pampean Terrane; RP, Rio de la Plata; SC, South China Craton; T, Tarim; T-B, Tanzania–Bangweulu Cratons; WA, West Africa; ZZB, Zambian Zambezi Belt.

Neoproterozoic magmatism, rifting and sedimentary basins

The Katangan Supergroup, Lufilian Belt–Congo–Tanzania–Bangweulu Craton. The Neoproterozoic Katangan Supergroup was deposited on the Congo–Tanzania–Bangweulu Craton margin as a passive margin sequence (Mendelsohn 1961; Binda 1994; Cailteux *et al*. 1994; Porada & Berhorst 2000). In the Lufilian Belt, the Katangan Supergroup is deformed into a north-verging, thin-skinned, low- to medium-grade, fold–thrust belt. The review by Porada & Berhorst (2000) presented the most comprehensive stratigraphic framework for the supergroup, which we have integrated with recent geochronological investigations (Armstrong *et al*. 1999; Key *et al*. 2001) to produce a revised stratigraphic column (Fig. 5). The supergroup, which has been defined for the Copperbelt region, has been divided into two lithotectonic sequences, an older sequence comprising the copper-bearing Roan Group, which is in tectonic contact with a younger sequence comprising the Mwashia and Lower and Upper Kundelungu Groups. The Roan Group rests unconformably on, and contains boulders of the 877 ± 11 Ma[67] (Master *et al*. 2002; Rainaud *et al*. 2002) Nchanga Granite. The Roan Group, interpreted as a rift-to-drift sequence, was divided into three subgroups (Binda 1994) that are inferred to be approximately contemporaneous, laterally equivalent units (Porada & Berhorst 2000; Wendorff 2000). To the south of the Mwembeshi Shear Zone (Fig. 1), rocks interpreted to be equivalents of the Roan Group (Porada & Berhorst 2000) are known as the Cheta Formation, and the base of the sequence is marked by a thin succession of intensely sheared meta-pelite (the Nazingwe Formation), which passes up into the voluminous and laterally extensive Kafue Rhyolite that has been dated at 879 ± 19 Ma[62]. Overlying the Kafue Rhyolite is a sequence of marble and pelite that contains thousands of exotic mafic and ultramafic blocks. The Cheta Formation is in tectonic contact with underlying basement rocks and it has yet to be conclusively proven that it is equivalent to the Roan Group and/or that it formed on the Congo–Tanzania–Bangweulu Craton. It is possible that it is an allochthonous sequence that was tectonically emplaced on the Congo–Tanzania–Bangweulu Craton.

In the Copperbelt, the Mwashia Group, overlying the Roan, is composed of platformal, shallow-water argillite and carbonate, with many horizons of rift-related bimodal volcanic rocks (Porada & Berhorst 2000), one of which has been dated at 765 ± 5 Ma[73]. The top of the Mwashia Group passes into glacial deposits (known as the Grand Conglomérat in the Democratic Republic of Congo (DRC)) of the Lower Kundelungu Group and its associated cap carbonate, which, in light of the *c*. 765 Ma age determination for the volcanic rocks in the Mwashia Group, probably correlates with the global Sturtian–Rapitan glacial event (Key *et al*. 2001). The remaining part of the Lower Kundelungu Group comprises a thick sequence (up to 5000 m) of shale. The overlying Upper Kundelungu Group consists of a second diamictite deposit at the base (equivalent to the Petit Conglomérat in the DRC and possibly recording the global 650–600 Ma Marinoan glacial event) overlain by a sequence of sandstone and siltstone, including the Cambro-Ordovician Luapula Beds (Vavrdova & Utting 1972). Deposition of the Mwashia and the overlying Kundelungu Groups is interpreted to have occurred in a rapidly subsiding extensional basin that opened some 100 Ma after initiation of the Roan rift basin (Porada & Berhorst 2000, and references therein).

Northern Kalahari margin. Sedimentary protoliths of the Neoproterozoic Rushinga Group were presumably deposited on the northeastern margin of the Kalahari Craton. In northeastern Zimbabwe, where metamorphic grade is in the amphibolite facies, the suite consists of basal quartzite or psammite overlain by a structural package of pelite, quartzite and marble (Barton *et al*. 1991; Hargrove *et al*. 2003), with volcanic and plutonic rocks dated between 808 and 795 Ma[84,85,89]. Farther west, abundant peralkaline sheets were emplaced along with mafic magmas, in a continental rift setting, presumably through the Palaeoproterozoic Magondi Belt (Munyanyiwa *et al*. 1997; Dirks *et al*. 1999). Igneous zircons from granitic gneisses have yielded a TIMS age of 831 ± 6 Ma[90] and Pb/Pb evaporation ages that range from 854 ± 0.8 to 737 ± 0.9 Ma[91–95].

The oldest Neoproterozoic magmatic rocks (870–840 Ma) on the Kalahari margin are the Ocellar Gneiss and the Mavuradohna

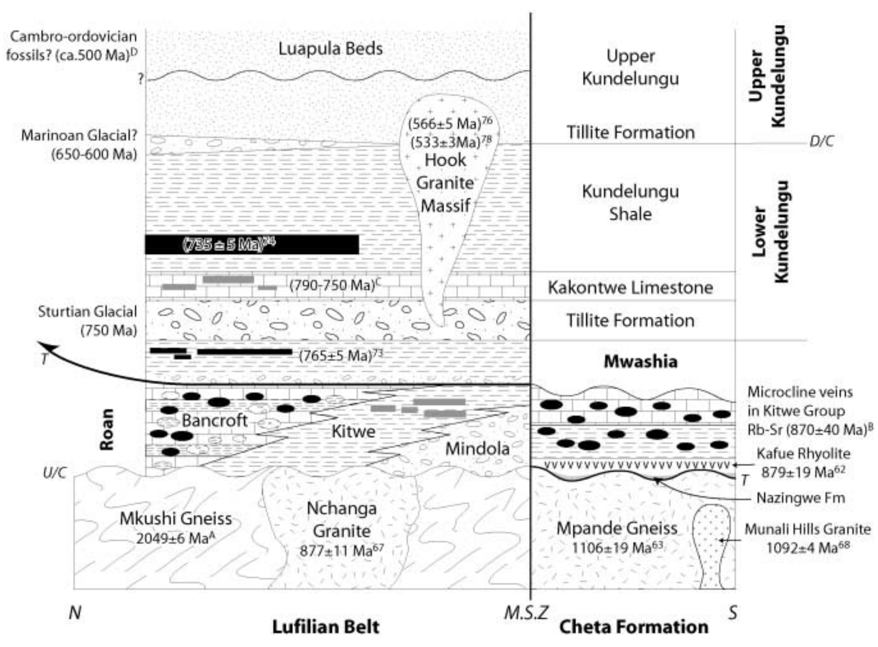


Fig. 5. Reconstructed lithostratigraphic section of the Lufilian Belt in Zambia, modified from Porada & Berhorst (2000) with additional age constraint information. Black ellipses within the Roan and Cheta Formation represent numerous mafic and ultramafic blocks (refer to main text). Superscript numbers following ages relate to Table 1; locations are illustrated in Figure 2. Other references (superscript letters): A, Rainaud *et al*. (2002); B, Cahen *et al*. (1984); C, Kampunzu *et al*. (1998); D, Vavrdova & Utting (1972). D/C, disconformity; M.S.Z., Mwembeshi Shear Zone; N, north; S, south; T, thrust; U/C, unconformity

Metamorphic Suite in the Zambezi Allochthonous Terrane; they have been interpreted to represent a period of mid-crustal A-type magmatic activity accompanied by granulite-facies metamorphism (Hargrove *et al*. 2003). However, these rocks occur in thrust slices, and it is not known whether they formed on the Kalahari margin or are allochthonous to it.

Contemporaneous rifting events on the Congo–Tanzania–Bangweulu and Kalahari margins? It is apparent that extension-related magmatism and sediment formation occurred on the northern margin of the Kalahari Craton and the southern margin of the Congo–Tanzania–Bangweulu Craton in the interval *c*. 880–750 Ma (Figs 3 and 6). The question therefore arises as to whether this provides credence to a model involving the attempted break-up of a previously assembled Congo–Tanzania–Bangweulu–Kalahari Craton and the formation of a 'Zambezi Aulacogen' (Daly 1986*a*, *b*; Hanson *et al*. 1988*b*, 1993, 1994, 1998; Unrug *et al*. 1998; Kampunzu *et al*. 2001; Hanson 2003; Hargrove *et al*. 2003). We argue that, although there is a broad similarity in the age of magmatism on both cratons, in detail the timing does not overlap. Rifting and associated extensional-related magmatism in the Congo–Tanzania–Bangweulu Craton occurred principally in two discrete pulses, some time after *c*. 890 Ma forming the Roan rift basin and at *c*. 765–735 Ma forming the Mwashia–Kundelungu basin (Porada & Berhorst 2000) (Fig. 6), although the Ngoma Gneiss (and possibly also the Lusaka Granite, although its crystallization age has a very large error of *c*. 68 Ma), were intruded between these two episodes. In contrast, rift-related magmatism on the Kalahari Craton occurred intermittently from 870 Ma (if the Zambezi Allochthonous Terrane formed on the Kalahari margin) to 750 Ma, with a peak at about 800 Ma associated with the deposition of the protoliths to the Rushinga Group (Fig. 6; Hargrove *et al*. 2003). However, as the Zambezi Allochthonous Terrane cannot be unambiguously linked to the Kalahari Craton on which it rests, extensional magmatic activity associated with the Kalahari margin may not have started until *c*. 830 Ma.

If the Kalahari Craton was a component of the Rodinia supercontinent, associated with Australia, Antarctica and South China (e.g. Dalziel 1997; Li *et al*. 2003; Loewy *et al*. 2003; Pisarevsky *et al*. 2003), then the A-type magmatism and associated rifting could have been related to supercontinent break-up. The timing of Neoproterozoic rifting events in this part of Rodinia was reviewed by Li *et al*. (2003, 2004) and a comparison of the timing of Neoproterozoic magmatism is shown in Figure 6. In this figure, the Kafue Rhyolite is shown as being associated with the Congo–Tanzania–Bangweulu Craton on which it sits at present, but this may be part of the Kalahari margin or allochthonous to both cratons. Li *et al*. (2003, 2004) proposed that there were two main rifting episodes, at 830–795 Ma and 780–745 Ma, which Pisarevsky *et al*. (2003) related to sequential break-up of the various cratons from the supercontinent. Apart from the Zambezi Allochthonous Terrane, which, as noted, may be allochthonous to the Kalahari Craton, the timing of rift magmatism in the Kalahari Craton is comparable with that in the South China and Australia Cratons; that is, the main magmatic activity switched off at 795 Ma.

Neoproterozoic ophiolites and continental-margin-arc magmatism

Plate tectonic models proposing the former existence of a major Neoproterozoic Zambezi Ocean (e.g. Barnes & Sawyer 1980), have been criticized in the past for the apparent lack of ophiolites

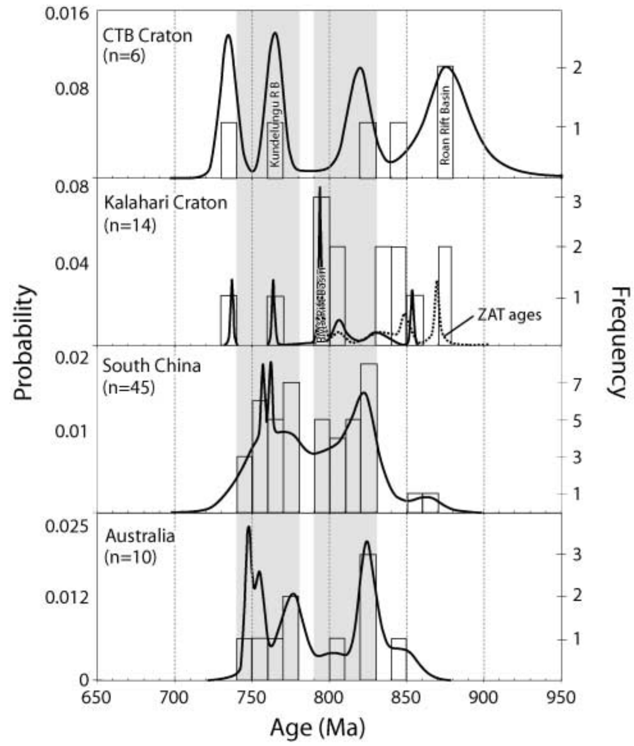


Fig. 6. Histogram and probability density curves comparing Neoproterozoic magmatic rifting episodes in the Congo–Tanzania–Bangweulu Craton with those for the Kalahari, South China and Australian Cratons that formed part of Rodinia. Data for the Congo–Tanzania–Bangweulu Craton are summarized in this paper; data for other cratons were summarized by Li *et al*. (2003). RMS, Rushinga Metamorphic Suite; ZAT, Zambezi Allochthonous Terrane.

and associated accreted oceanic fragments of appropriate age (e.g. Hanson *et al*. 1994; Munyanyiwa *et al*. 1997). However, Vrána *et al*. (1975) and Porada & Berhorst (2000) described the presence of thousands of metre- to kilometre-scale mafic and ultramafic blocks in the Roan Group in the Lufilian and Zambezi Belts, the tectonic significance of which has, until recently, largely been ignored. Furthermore, there have been recent reports of Neoproterozoic suprasubduction-zone, continental-margin-arc magmatism in the region (Kröner *et al*. 2001; Katongo *et al*. 2004).

Ophiolitic lithologies. The mafic and ultramafic blocks embrace a range of lithologies including basalt, amphibolite, gabbro, norite, picrite, peridotite, lherzolite, troctolite, eclogite and serpentinite (Vrána *et al*. 1975). Some of the mafic blocks in the Lufilian Belt have incompatible trace element ratios that are similar to those of magmas formed in an ocean-island setting or during the initial stages of continental rifting (Tembo *et al*. 1999). Gabbro, metagabbro and eclogite blocks from the Southern Irumide Belt have incompatible trace element and isotopic ratios comparable with those of mid-ocean ridge basalts (MORB) and have been interpreted as fragments of former ocean crust (John 2001; John *et al*. 2003, 2004*a*).

In addition to mafic and ultramafic blocks, the Bancroft Subgroup and Cheta Formation to the south of the Mwembeshi Shear Zone are composed of a shelf-like succession of dolomite and subordinate argillite and quartzite (Binda 1994; Porada &

Berhorst 2000) and contain blocks and fragments derived from the adjacent Mindola and Kitwe Subgroups (Porada & Berhorst 2000). There appear to be two possible explanations for the abundant mafic blocks in these sediments: (1) the unit is an ophiolitic mélange (Gansser 1974; Williams 1977), i.e. an olistostrome with blocks emplaced by sedimentary mass flow processes following the emergence above sea level of ophiolitic thrust slices; (2) the unit is a tectonic mélange and the mixing of ophiolitic and continental-margin lithologies occurred during thrusting in the subsequent Lufilian–Zambezi Orogeny. Discrimination between these two alternatives is not straightforward, as much of the Bancroft Subgroup and Cheta Formation has undergone polyphase deformation and amphibolite-facies metamorphism during the Lufilian–Zambezi Orogeny. An olistostrome origin is tentatively preferred because: (1) the blocks are apparently limited stratigraphically to the Bancroft Subgroup and Cheta Formation (Porada & Berhorst 2000); (2) there is no evidence of contact aureoles or skarns developed at the margins of any of the blocks (Mendelsohn 1961), suggesting they were not intrusive into the sediments; (3) many blocks of mafic and ultramafic igneous lithologies are undeformed; (4) the great variation in size, composition and texture of the blocks over small distances is more compatible with an olistostromal origin; (5) blocks of igneous lithologies are accompanied by blocks of the adjacent Mindola and Kitwe formations, implying that these units were also involved in thrusting associated with ophiolite emplacement. This interpretation implies that ophiolite obduction onto the margin of the Congo–Tanzania–Bangweulu Craton occurred after the deposition of the Roan Group; that is, some time after *c*. 880 Ma. Alternatively, if the mafic–ultramafic blocks developed in a tectonic mélange, the upper age limit of the ophiolite would be the time of the Zambezi Orogeny (i.e. *c*. 550 Ma; see below).

Neoproterozoic continental-margin-arc granitoid magmatism. In southern Malawi, juvenile calc-alkaline granitoids (Kröner *et al*. 2001) have crystallization ages of 710–555 Ma[112–118,121,123–129] (Table 1, Fig. 3). These were emplaced into *c*. 1040 and 999 Ma continental-margin-arc granitoids (Kröner *et al*. 2001) and suggest active-margin magmatism along part of the southern margin of the Congo–Tanzania–Bangweulu Craton from at least *c*. 710 Ma until terminal collision at *c*. 550–530 Ma.

Tectonothermal history and timing of the Zambezi–Lufilian Orogeny

Subduction-zone metamorphism? Some of the mafic blocks within the Roan Group and Cheta Formation of the Lufilian and Zambezi Belts display evidence for very high-pressure, moderate-temperature (HP–MT) eclogite-facies metamorphism. These blocks occur in a discrete HP thrust slice or slices some 200 km long and 40 km wide (Fig. 7), in which eclogites and associated mafic blocks typically crop out as isolated hills and contact relations with the surrounding metasedimentary rocks are rarely exposed. However, wherever present, the surrounding country rock metapelite contains kyanite (Vrána & Barr 1972; Vrána *et al*. 1975), suggesting that the metasedimentary matrix also underwent an HP metamorphic evolution. Peak *P–T* conditions have been estimated at 630–690 °C and 26–28 kbar, and are compatible with a subduction-zone setting (John & Schenk 2003; John *et al*. 2003). Sm/Nd whole-rock–garnet and Lu–Hf whole-rock–garnet–omphacite isochrons (John & Schenk 2003; John *et al*. 2004*a*) have yielded a range of ages between 659 ± 14 Ma and 595 ± 10 Ma, which are interpreted as the time of eclogite-facies metamorphism (John *et al*. 2004*a*).

If the blocks in the metasediments constitute a tectonic mélange, following eclogite-facies metamorphism of ocean crust in a subduction-zone setting, the HP rocks would have been incorporated into the metasediments of the Roan Group and Cheta Formation during collisional overthrusting (Porada & Berhorst 2000; John & Schenk 2003; John *et al*. 2003). Alternatively, if the blocks were originally part of an ophiolitic mélange, they would have been situated on continental crust prior to HP metamorphism, and their tectonometamorphic evolution would have involved subduction of the outer part of the Congo–Tanzania–Bangweulu Craton margin followed by its uplift and foreland-directed transport to mid-crustal levels. Attempted subduction of continental crust is known in several orogens (e.g. Alps, Appalachians, Scandinavian Caledonides, Variscides) and the presence of whiteschists in rocks of the Congo–Tanzania–Bangweulu margin (see below) provides additional evidence in favour of this model.

Collisional tectonometamorphism: Lufilian–Zambezi Orogeny. The Lufilian and Zambezi Belts are characterized by thin- and thick-skinned thrusting, respectively, and the polarity of thrusting in the Lufilian–Zambezi Orogen is illustrated in Figure 7. In the Zambezi Belt north of the Zambezi Valley, thrust transport was toward the NNE–NE (Wilson *et al*. 1993; Hanson *et al*. 1994; Johnson & Oliver 2004), whereas structures along the northern Kalahari margin are consistently SSE-verging (Barton *et al*. 1991; Dirks *et al*. 1999; Vinyu *et al*. 1999; Müller *et al*. 2001), indicating that the Lufilian–Zambezi Orogen was doubly vergent. Considering their spatial distribution, the orogen appears to have the architecture of an asymmetric, orogenic-scale structural fan. Orientations of the transport directions and the sinistral displacement on the Mwembeshi Shear Zone suggest that convergence between the Kalahari and Congo–Tanzania–Bangweulu Cratons was oblique.

In the Lufilian and Zambezi Belts, regional metamorphism reached amphibolite facies, except in discrete HP thrust slices, which are characterized by whiteschist-facies (kyanite–talc) assemblages (Johnson & Oliver 1998, 2002). *P–T* conditions in whiteschists peaked at *c*. 750 ± 25 °C at 13 ± 1 kbar and 600–900 °C at 8–15 kbar in the Lufilian and Zambezi belts, respectively, along a clockwise *P–T* path (Johnson & Oliver 1998, 2002, 2004; John 2001; John *et al*. 2004*b*). The development of these assemblages in both the reworked basement and cover sequences of the Congo–Tanzania–Bangweulu Craton implies that they originated by attempted subduction of the Congo–Tanzania–Bangweulu margin during the Zambezi Orogeny.

Figures 1 and 7 shows that the HP assemblages occur where the Damara–Lufilian–Zambezi Orogen is narrowest and the Congo–Tanzania–Bangweulu and Kalahari Cratons come into closest proximity at the current erosion level. This suggests that the HP metamorphism developed as a result of subduction of a promontory on the Congo–Tanzania–Bangweulu Craton and that the Congo–Tanzania–Bangweulu Craton formed the lower plate in the collisional orogen.

Age of collisional Pan-African tectonometamorphism. A total of 32 ages for the Pan-African metamorphism are available for this part of southern Africa (Table 2; Fig. 2) which has been divided into four subregions. Metamorphic ages for all four subregions overlap and are constrained between 615 and 517 Ma[141–172]. The oldest ages, between 615 and 586 Ma[163–165,168,171], are from the Lurio and Nampula Belts. However, metamorphic zircons from this subregion also display a younger peak at 544–

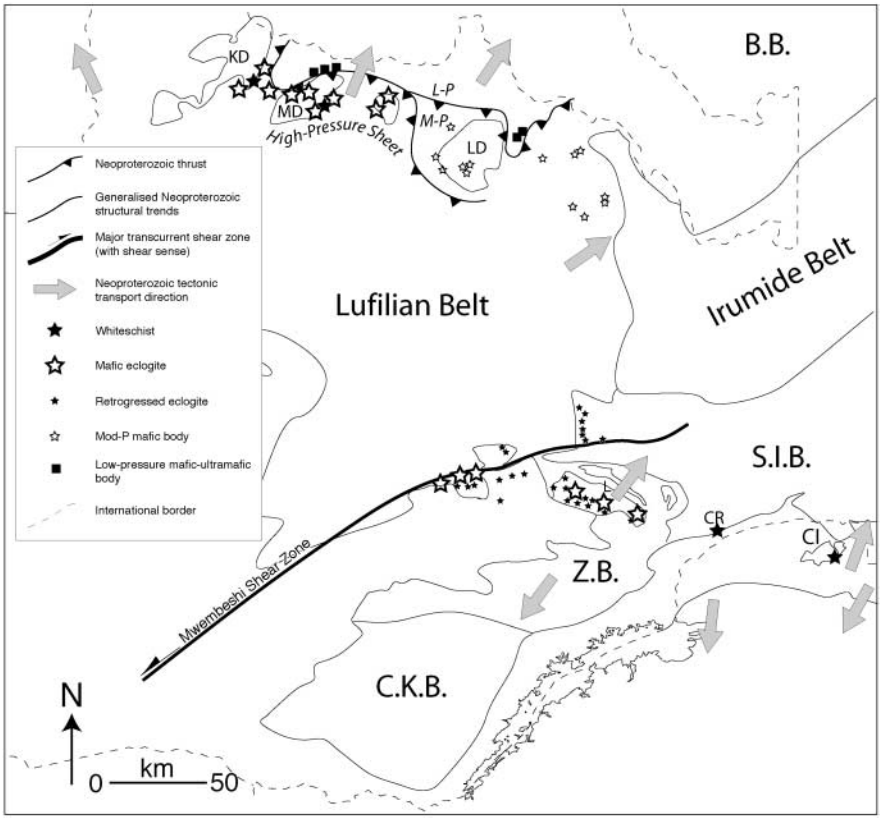


Fig. 7. Location of the numerous high-pressure mafic, ultramafic and whiteschist blocks in the Zambezi and Lufilian Belts (from Vrána & Barr 1972; Vrána *et al*. 1975; Cosi *et al*. 1992; John *et al*. 2003, 2004*b*). Abbreviations as in previous figures, and: B.B., Bangweulu Block; S.I.B., Southern Irumide Belt; Z.B., Zambezi Belt.

523 Ma[166,167,170,172], which overlaps with ages in some of the other subregions (Fig. 8). In southern Malawi the ages range between 575 and 547 Ma[157–162] with the older ages overlapping with suprasubduction-zone, continental-margin-arc magmatism (Kröner *et al*. 2001). In the Lufilian and Zambezi Belts in northern Zimbabwe and southern Zambia, the metamorphic ages are in the range 557–520 Ma[141–145,149–156], except for those from the Chowe River area (CR in Fig. 1), where HP metamorphism is dated at *c*. 570 Ma[146–148].

There is thus good evidence that Pan-African metamorphism in this region took place over a period of about 100 Ma from *c*. 615 to 520 Ma. The distribution of ages in Figure 8 shows that the interval *c*. 550–520 Ma is common to all four subregions, suggesting that this *c*. 30 Ma period was the time of the main collision between the Congo–Tanzania–Bangweulu and Kalahari Cratons. The presence of older metamorphic ages may indicate that collision was diachronous, or they may represent accretionary events between smaller terranes and blocks prior to the main continent–continent collision. Rb–Sr mineral isochrons and $^{40}Ar/^{39}Ar$ hornblende spectra from the Zambezi Belt (summarized by Vinyu *et al*. (1999) and Goscombe *et al*. (2000)) yield ages in the range 510–480 Ma and are interpreted to record unroofing of the orogen and cooling below *c*. 350 °C.

The *c*. 550–520 Ma timing of collisional metamorphism in the Zambezi Belt is similar to the 540–510 Ma range determined for the contiguous Damara Belt (Jung *et al*. 2001, 2003; Jung & Mezger 2003), suggesting the closure of a single ocean basin. The timing of peak metamorphism in the East African Orogen in Tanzania appears to have occurred significantly earlier, between 640 and 630 Ma (Muhongo *et al*. 2001; Kröner *et al*. 2003; Sommer *et al*. 2003), although the tectonic significance of these ages is disputed (e.g. Appel *et al*. 1998; Sommer *et al*. 2003). High-grade metamorphism in southern and central Madagascar occurred at *c*. 550 Ma (Kröner *et al*. 1996; De Wit *et al*. 2001; Kröner 2002), essentially coeval with that in the Damara–Lufilian–Zambezi Orogen. The spatial arrangement of the metamorphic ages and the differences in timing of peak metamorphism indicate that closure of the Mozambique and Zambezi–Adamastor oceans and amalgamation of Gondwana was not a simple collision between two large blocks or mega-cratons, i.e. East and West Gondwana, but rather a series of collisional events among several cratonic fragments, which took place over a period of about 100 Ma (Maas *et al*. 1992; Meert & Torsvik 1995; Fitzsimons 2000, 2003; Kröner 2002; Kröner & Cordani 2003; Meert 2003). Shackleton (1996) speculated on the location of suture zones that were integral to the formation and amalgamation of the Gondwana supercontinent. The data presented here show that the Damara–Lufilian–Zambezi Orogen has all the hallmarks of a suture zone that marks the site of a former Neoproterozoic ocean basin, and thus was integral to the amalgamation of Gondwana.

Conclusions

A careful review of available robust geochronological and geological data from central southern Africa shows the following.

(1) On the basis of their different ages and styles of magmatism, the Kibaran Belt, Choma–Kalomo Block and Irumide Belt are not part of the same orogen. There was no linked 'Kibaran' (*s.l.*) orogenic event in southern Africa.

(2) From NW to SE, the Irumide Belt is underlain by supracrustal rocks deposited on the Congo–Tanzania–Bangweulu Craton and a variety of accreted terranes including those of island-arc and continental-margin-arc affinity. Coupled with its

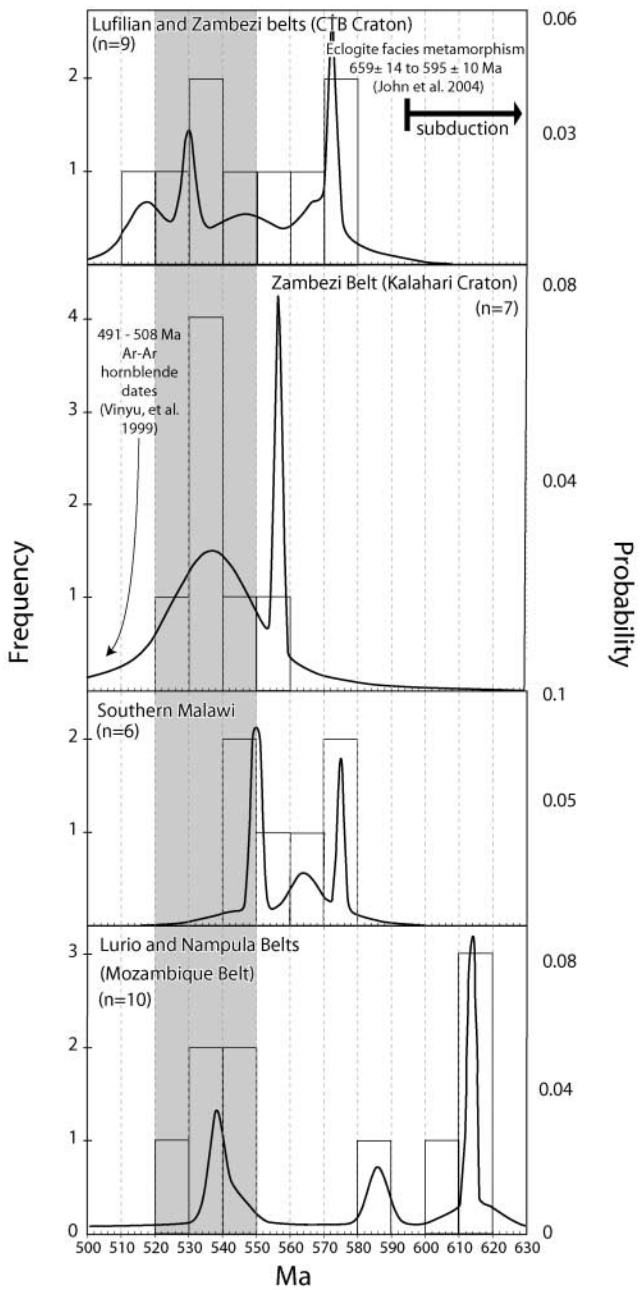


Fig. 8. Histograms and probability density curves for the 32 metamorphic ages presented in Table 2 and Figure 7.

LP–HT metamorphic signature, this suggests that it was an accretionary orogen open to an ocean to the SE (present coordinates).

(3) The thousands of exotic mafic and ultramafic blocks in the Neoproterozoic Roan Group and Cheta Formation on the Congo–Tanzania–Bangweulu Craton are tentatively inferred to represent an ophiolitic mélange, implying that these units mark the passage of ophiolite obduction onto the margin of the Congo–Tanzania–Bangweulu Craton some time after *c.* 880 Ma.

(4) The MORB-like chemistry of mafic blocks in the ophiolitic mélange and the presence of Neoproterozoic continental-margin-arc magmas in southern Malawi provide evidence for the presence of Neoproterozoic oceanic crust between the Congo and Kalahari Cratons.

(5) The presence, in the Lufilian and Zambezi Belts, of low-temperature eclogite and whiteschist assemblages in rocks attached to the Congo–Tanzania–Bangweulu Craton implies attempted subduction of the southern margin of the Congo–Tanzania–Bangweulu Craton during the Damara–Lufilian–Zambezi Orogeny. The peak of collision-related metamorphism in the Lufilian–Zambezi Belt at 550–520 Ma represents the time of collision between the Congo–Tanzania–Bangweulu and Kalahari Cratons.

(6) The Damara–Lufilian–Zambezi Orogen marks a Neoproterozoic suture, the site of a major ocean basin whose closure was an integral component of the assembly of Gondwana.

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