Geochronology of the Zambezi Supracrustal Sequence, Southern Zambia: A Record of Neoproterozoic Divergent Processes along the Southern Margin of the Congo Craton


Institute for Research on Earth Evolution, Japan Agency for Marine-Earth Science and Technology, 2-15 Natsushima-cho, Yokosuka, Kanagawa-ken 237-0061, Japan
(e-mail: simon.johnson2@doir.wa.gov.au)

ABSTRACT

The Zambezi supracrustal sequence (ZSC) of southern Zambia comprises a metasedimentary package of clastics and carbonates, with a thick sequence of basal volcanics and lavas. The sequence has traditionally been interpreted as a Neoproterozoic continental rift succession, but the lack of reliable age constraints hinders any tectonic interpretation. In this article, we date magmatic and detrital zircons using the U-Pb SHRIMP method in order to better constrain the timing of rifting, volcanism, and basin deposition. The basal volcanoclastic Kafue Rhyolite and Nazingwe formations were erupted at ca. 880 Ma, and the sequence was intruded by the Lusaka Granite at ca. 820 Ma, providing lower and upper limits on the age of sedimentation. Whole-rock Nd isotopic signatures of these volcanics indicate that they formed as a result of assimilation and recycling of basement gneisses, probably during crustal thinning and extension. We uphold the correlation between the ZSC and the Roan Group in the Zambian Copperbelt and suggest that both successions formed in discrete rift basins along the southern margin of the Congo-Tanzania-Bangweulu (CTB) Craton; however, extension at this time probably did not result in complete continental separation. If the CTB Craton were an integral part of Rodinia, then rifting at ca. 880 Ma would represent one of the first known records of attempted breakup of the supercontinent.

Online enhancements: appendix, color figures.

Introduction

In the past decade there has been a considerable volume of publications and debate on the makeup, configuration, and existence of the Mesoproterozoic supercontinent of Rodinia (e.g., Hoffman 1991; Dalziel 1997; Weil et al. 1998; Pisarevsky et al. 2003; De Waele et al., forthcoming). The majority of data pertaining to these reconstructions come from only a limited number of cratons for which palaeomagnetic and geochronological data are better constrained (updated database of available global palaeomagnetic data in Pisarevsky 2005). Pertinent geological information may be obtained not only from the bounding orogenic belts, i.e., those that may have formed during the assembly of the supercontinent, but also from rift-related sedimentary and volcanoclastic deposits and igneous intrusions that formed during supercontinent breakup (e.g., Moores and Twiss 1995; Li et al. 2003).

Recent geochronological, geochemical, isotopic, and metamorphic studies from the Mesoproterozoic central southern African belts (De Waele 2005;
Johnson et al. 2005, 2006; De Waele et al. 2006a, 2006b) suggest that the southern margin of the Congo-Tanzania-Bangweulu (CTB) Craton experienced crustal thickening and compressional tectonics in a time frame similar to that of other collisions associated with Rodinia assembly. The identity of the opposing continental margin for the CTB Craton is not known, and the precise timing of rifting of this continental mass from the CTB, with the production of a Neoproterozoic ocean, the Zambezi Ocean (John et al. 2003), is also unknown. The presence of extensive early to mid-Neoproterozoic sediments and potentially rift-related volcanic and volcanoclastic deposits in southern Zambia and northern and northeastern Zimbabwe may provide essential data on the timing of these rifting events. However, because of folding and thrusting during Gondwana assembly, it is difficult to determine on which side of the Zambezi Ocean these basins were deposited (e.g., Armstrong et al. 2005). Dating of critical igneous and sedimentary units may help to resolve some of these issues.

The largest supracrustal sequence is located within the moderate- to high-metamorphic-grade Zambezi Belt in southern Zambia and will be referred to here as the Zambezi supracrustal sequence (ZSC; fig. 1a, 1b). Characteristics of the ZSC include the following:

1. The sequence is floored by a thick basal sequence of metarhyolites and minor metabasalts, which have been interpreted as bimodal, continental rift volcanics (the Kafue Rhyolite Formation) and for which only an unverifiable age has been available in the past (M. S. Wardlaw, unpublished data; reported in Wilson et al. 1993).

2. The sequence has traditionally been correlated with the lowermost Katangan sediments of the Copperbelt Region, the Roan Group (Annels 1974; Unrug 1988; Porada 1989; Binda 1994; Porada and Berhorst 2000; Johnson et al. 2005).

3. The upper part of the sequence, the Cheta Formation, contains thousands of randomly oriented, isolated blocks of variably metamorphosed, some up to eclogite facies (John and Schenk 2003), gabbro, mafic, and ultramafic blocks (Vrana et al. 1975). These blocks have normalized mid-ocean ridge basalt (N-MORB) chemistries (John et al. 2003, 2004) and are interpreted to represent a tectonically dismembered Neoproterozoic ophiolite (John et al. 2003) or a Neoproterozoic ophiolitic mélangé (Johnson et al. 2005).

4. The sequence is intruded by various Neoproterozoic within-plate-type granitoid bodies (Hanson et al. 1988, 1994; Wilson et al. 1993).

The ZSC sequence appears to record a complete tectonic cycle from continental rifting through ocean closure and subduction metamorphism, but so far there are no adequate constraints on the timing, tectonic setting, or provenance of these formations. In this article we aim to review the main sedimentological features of the ZSC succession and to provide critical new timing and tectonic constraints based on U-Pb SHRIMP (sensitive high-resolution ion-microprobe) zircon dating and whole-rock Sm-Nd isotopic analyses. Based on these, we propose a correlation between the ZSC and the Roan Group of the Copperbelt and at the same time point out inconsistencies in assigning a Kalahari affinity to the ZSC.

**Regional Geology and Tectonostratigraphic Relationships**

The Lusaka region was mapped in the 1960s by various geologists of the Northern Rhodesia Geological Survey (e.g., Simpson et al. 1963; Smith...
Parts of this work, including the stratigraphy and structural development of the region, were refined by Mallick [1963, 1966]. The region south of Lusaka is underlain by a variety of metamorphosed gneisses, metasediments, and granitoids that are often in tectonic contact with each other. So far, the oldest dated units are the Mesoproterozoic, variably deformed Mpande Gneiss [1106 ± 19 Ma; Hanson et al. 1988; sample U-7 in fig. 2a] and Munali Hills Granite [1090 ± 1.3 Ma; Katongo et al. 2004; samples MHG1, MHG2, and MHG9 in fig. 2a]. Metasedimentary rocks of possible Katangan age occur in two distinct lithotectonic packages, one to the southwest of the Munali Hills basement inlier around the town of Mazabuka and the other around the Mpande Dome and northward up to the Mwembe Shear Zone (figs. 1b, 2a).

As described by Smith [1963], the area southwest of the Munali Hills comprises two medium-grade metasedimentary formations (fig. 2a, 2b, fig. 3). At the inferred base, the dominantly siliciclastic Nega Formation consists of basal quartzites with intercalated metapelites and minor calcareous rocks. The upper part of the formation is dominated by extensive areas of mica schists, bearing retrogressed porphyroblasts of andalusite and kyanite [Brown 1967]. Because of poor exposure and intense deformation, the thickness of this unit is unknown. The Nega Formation is overlain, apparently in stratigraphic conformity, by the Muzuma Formation, dominated by calc-silicate rocks with minor pure quartzites, marbles, and mica schists. Again, the thickness of this formation is unknown. These two formations were tentatively correlated with the similar sequence of rock types in the Kafue-Lusaka area (Mulola to Cheta formations of Simpson et al. 1963).

Mineral exploration in the Munali Hills area in the 1970s led to the discovery of small, podlike gabbroic bodies within the marble units near the top of the Nega Formation, which bear a superficial resemblance to the gabbros and eclogites described by Vrána et al. (1975) in the Cheta Formation to the north. Mapping by the Anglo-American Company in the 1970s also appeared to show that the Munali Hills Granite is intrusive into the basal units of the Nega Formation (map shown in Hanson et al. 1988). This was used by Porada and Berhorst (2000) as evidence to suggest that the Nega Formation is of Mesoproterozoic age and is not equivalent to the Katangan sequence around Lusaka or in the Copperbelt. Our mapping of the area to the south of the Munali Hills, including drill core logs from work carried out during mineral exploration, indicates that the Nega Formation unconformably overlies the Munali Hills Granite because basal gritty quartzites and conglomerates contain angular granite clasts of the underlying igneous lithologies (figs. 2b, 4f). The intrusive contacts mapped and illustrated by Hanson et al. [1988] may suggest that there are vestiges of an older pre-Neoproterozoic sedimentary sequence or that parts of the Munali Hills Granite may be considerably younger and equivalent in age to the Ngoma Gneiss and Lusaka Granite (i.e., ca. 820 Ma). Field mapping and examination of drill core has identified massive to weakly banded metarhyolite or metarhyodacite and a pink leucocratic microgranite located in restricted areas between the base of the Nega Formation and the Munali Hills Granite (fig. 2b). The contacts between all these units are not exposed, and the relationships between them are not clear. However, on the map scale it is evident that the Nega Formation lithologies progressively truncate the metarhyolites and granites, and basal conglomerates themselves contain granite and metarhyolite clasts, indicating that the Nega sediments are unconformably overlying them. Although we have not been able to extract any zircons from these igneous units, we have dated zircons extracted from an extensive, homogenous, fine-grained portion of the matrix to a basal conglomerate of the Nega Formation [sample NgF47] in order to obtain a maximum age for sedimentation of the Nega Formation.

The base of the second lithotectonic package (fig. 3b) crops out to the north of the Munali Hills and to the south and west of the Mpande Dome (figs. 1b, 2a). The basal Kafue Rhyolite Formation (KRF)
is a 2500-m-thick sequence of folded, variably metamorphosed felsic volcanic flows and tuffs with subordinate tuffaceous sediments and extremely rare thin mafic horizons (Smith 1963; Mallick 1966). Coarse-grained volcanoclastics are rare, but an exceptional example is well exposed along a minor stream section centered on Grid Reference (GR) 35L 0642600-8250900. At this locality, a polymict conglomerate unit ∼50 m thick is composed of very coarse (up to 15 cm across), flattened, and deformed clasts comprising very fine-grained felsic volcanics, medium-grained equigranular granite, and mafic clasts set within a very fine-grained tuffaceous matrix (fig. 4a). This unit passes up, by the gradual loss of clasts, into ∼5 m of highly deformed tuffaceous schist (fig. 4b; sample KRF21) before the termination of exposure. The KRF passes up into the Nazingwe Formation (NF), a ∼500-m-thick sequence of tuffaceous semipelites with intercalated acid and minor mafic volcanic horizons (Smith 1963; Mallick 1966). Toward the top of the KRF, there are several horizons of semipelitic schist 30 cm to 5 m thick, and within the partly tuffaceous, semipelitic NF, there is at least one 5-m-thick rhyodacite flow (figs. 2d, 4c; sample NF30; GR 35L 0643909-8250495).

The nature of the boundary of the KRF with the basement rocks is uncertain because of a lack of exposure; however, Smith (1963) has interpreted it to be an unconformity. Everywhere along the Kafue Gorge Road section, between GR 35L 0642600-8250200 and GR 35L 0644400-8250900 (fig. 2d), the NF is intensely deformed and folded and carries a strong planar fabric that is at a high angle to primary lithological contacts (figs. 2c, 4c). We interpret the eastern boundary of the NF to be in tectonic contact with the underlying Mpande Gneiss basement and suggest that on a regional scale, the KRF and NF sequence forms a major, recumbent, east-verging fold with an overturned lower limb (fig. 2a). A single upper-intercept zircon crystallization age of 879 ± 19 Ma was obtained by thermal ionization mass spectrometry (TIMS) for the KRF (M. S. Wardlaw, unpublished data; reported in Wilson et al. 1993), but unfortunately these data have never been published in full, and it is thus impossible to determine their reliability. Furthermore, the location, geochemical composition, metamor-
Figure 4. Field photographs of the geochronological samples and other field relationships. 

a, Tuffaceous conglomerate that occurs 5 m below sample KRF21. $Gra =$ granite clast. 
b, Fine-grained tuffaceous sericitic schist KRF21. 
c, Rhyodacitic flow 5 m thick within the semipelitic Nazingwe Formation, sample NF30. Note the high angle of the main foliation to the bedding. 
d, Gray, plagioclase-rich granite (LG44) at the Geological Seismic Survey Station. 
A late high-strain shear zone (vertical) cuts through the main tectonic foliation (flat lying). 
e, K-feldspar–rich, porphyritic granite sample LG46 with 10-cm-long quartz-rich xenoliths. 
f, Nega Formation basal conglomerate containing flattened and deformed granite ($Gra$) and metarhyolite ($Rhy$) clasts. Sample NgF47 was collected from a matrix-rich portion of the conglomerate from a nearby outcrop.
phic grade, and other characteristics of the rock from which the zircon was extracted are also unknown, adding further to the uncertainty around the date. We have dated a rhyodacitic flow and a tuffaceous unit from the KRF and a rhyolite flow from the NF to determine the age of these volcanics.

The Kafue Rhyolite and Nazingwe formations are overlain by the quartzite-dominated Mulola Formation. At many localities, the formation preserves current bedding (Smith 1963; Mallick 1966), while the base of the formation is marked by a variable thickness of undeformed conglomerates that are in part composed of rhyolitic pebbles (Mallick 1966), illustrating that the formation is, at least in places, in depositional [rather than tectonic] contact with the underlying basement. On the map scale, the basal contact gradually oversteps the NF and KRF until it rests entirely upon Mpande basement gneisses (fig. 2a; Mallick 1966), indicating that it is an overstep unconformity. The Mulola Formation passes up into kyanite-bearing, biotite-rich schists and semipelites of the Chipongwe Formation and subsequently into the monotonous dolomitie marbles of the Cheta Formation (fig. 2a).

The Cheta Formation contains abundant gabbroic and ultramafic blocks that have N-MORB chemistries [Vrana et al. 1975; John and Schenk 2003; John et al. 2003, 2004]. It has been suggested that this succession is overlain with a structural break, possibly representing a thrust [Simpson et al. 1963; Porada and Berhorst 2000], this contact must be a thrust that juxtaposes these basement rocks upon the volcanic rocks. However, the base of the Mulola Formation is marked by a variable thickness of undeformed basal conglomerates

Our reconnaissance of the Lusaka Granite body indicates that it is composed of at least three different granitoid phases, a two-feldspar coarse-grained protomylonitic granite, a pink K-feldspar–dominant undeformed body, and a K-feldspar porphyritic unit with abundant quartz-rich xenoliths. The whole-rock major, trace, and rare earth element geochemistry of these various granitoid rocks was studied in detail by Katongo et al. (2004), who concluded that the granitoids all formed by melting of older silicic basement rocks. We have dated two of the three Lusaka Granite phases in order to constrain their age of intrusion into the sequence, thus providing a lower age limit on basin sedimentation.

In a recent review of the stratigraphic successions of the Lufilian and Zambezi belts, Porada and Berhorst (2000) presented a radically different interpretation of the lithostratigraphy around the Lusaka area. Basing their analysis entirely on the apparent differences in metamorphic grade, they combined the higher-grade parts of the Zambezi supracrustals [Mulola and Chipongwe formations] into a single formation, the Chunga Formation, and assigned this unit to part of the Mesoproterozoic basement. The remaining lower-grade parts of the Chipongwe Formation and the carbonate-rich Cheta Formation and Lusaka Dolomite were combined into a new Cheta Formation, which was interpreted as the only Neoproterozoic succession in the region. The Lusaka Granite is intrusive into these inferred basement schists and quartzites, but the potentially unconformable-thrust contact with the Lusaka Dolomite suggests that this unit might be younger, having been deposited after 846 ± 68 Ma, the current age of the Lusaka Granite (Barr et al. 1977). We disagree with the major parts of this subdivision on a number of grounds:

1. Apparent metamorphic grade and structural complexity have long been demonstrated as unreliable criteria for relative age determination and subdivision of stratigraphic sequences and rock units; moreover, the carbonates of the Cheta Formation are relatively pure and often lack adequate silicate minerals to demonstrate metamorphic grade.

2. On the 1 : 250,000-scale geological map of the Lusaka area [Anonymous 1984; fig. 2a], it is apparent that the Mulola Formation quartzites [Porada and Berhorst 2000] rest on the Kafue Rhyolite and Nazingwe formations. Therefore, in the model presented by Porada and Berhorst (2000), this contact must be a thrust that juxtaposes these basement rocks upon the volcanic rocks. However, the base of the Mulola Formation is marked by a variable thickness of undeformed basal conglomerates
that contain rhyolitic pebbles [Mallick 1966] probably derived from the underlying KRF, implying that the Mulola Formation was deposited after the eruption of the Kafue Rhyolites, i.e., it is an erosional overstep unconformity and not a thrust.

3. In order to explain the thousand of mafic blocks and lenses within the Cheta and Chipongwe formations, Porada and Berhorst (2000) invoked the presence of a major orogenic-scale thrust, with the mafic blocks representing a tectonic mélangé situated between the basement footwall and the Cheta Formation hanging wall. Although this is a valid interpretation, it is only one of a number of possible tectonic scenarios that can adequately explain the presence of these blocks [e.g., Johnson et al. 2005]. We certainly agree that some contacts between units are tectonic in origin, having been folded and thrust during the Pan-African orogeny, but we see no evidence for major orogenic-scale tectonic discontinuities or shear zones between the metasedimentary units.

In this article, we make use of the simplified stratigraphy of Mallick (1966; see our fig. 3), which does not contain any major structural or tectonic breaks and which is inferred to represent the evolution of a single, simple, rift-to-drift sedimentary basin. However, we note that the Lusaka Dolomite may be significantly younger than the main ZSC sequence, having been tectonically juxtaposed with the ZSC during Pan-African tectonism.

Sample Localities, Petrography, and Geochemistry

Several samples, including granitoids, metavolcanics, and metasediments, were collected for geochemical investigation. Some of these samples were also analyzed for their major and trace element and isotopic compositions. The analytical procedures are described in the appendix, available in the online edition and from the Journal of Geology office. The geochemistry of most of the major volcanic and granitoid rocks in the region has been presented by Katongo et al. (2004). Considering that most of these rocks have undergone upper-green-schist to lower-mid-amphibolite facies metamorphism, little emphasis has been placed on the fluid-mobile major-element analyses (Na₂O, CaO, and K₂O) or the large lithophile elements, such as Sr, Rb, and Ba (Humphris and Thompson 1978; Brekke et al., 1984).

Kafue Rhyolite and Nazingwe Formations. Sample KRF20 was collected from the main KRF along the Kafue Gorge Road (GR 35L 0642908-8250298; fig. 2d) and is representative of the felsic volcanics that dominate the KRF. The sample is relatively homogeneous but weakly deformed, has a rhyodacitic composition [SiO₂ = 68 wt%, table 1], and contains relatively high concentrations of high-field strength elements such as Nb, Th, Y, and Zr (table 1). The unit is dominated by 1–2-mm-diameter plagioclase phenocrysts set within a fine-grained (<0.1 mm), equigranular matrix of quartz, plagioclase, K-feldspar, and biotite. Some of the matrix plagioclase has been altered to muscovite. Even though the rock is weakly foliated in hand specimen, this fabric is hard to identify in thin section, and there appears to be little reworking or corrosion of the relict-igneous plagioclase phenocrysts. Fine-grained (0.5–1.0 mm diameter) Cu-bearing ore minerals such as chalcopyrite and bornite are randomly distributed throughout the rock and make up 1%–2% of the whole-rock volume.

Sample KRF21 is a pale gray, fine-grained, highly strained unit that occurs directly above the volcanoclastic unit (fig. 2a, 4a, 4b; GR 35L 0645671–8240290). The unit has a SiO₂ concentration of ~71 wt% and has extremely high K₂O, Zn, Pb, Rb, Sr, and Nb concentrations (table 1). In thin section, the unit is mineralogically banded. The dominant bands are up to 0.5 mm in thickness, make up 60%–70% of the section, are very fine grained (<0.1 mm), and are dominated by muscovite (90%–100% muscovite). The intermittent bands are thicker, up to 0.75 mm, and composed of equigranular quartz (60%) and plagioclase (40%). The muscovite is aligned and imparts a strong planar fabric parallel to the bedding. On the outcrop scale, it is evident that this planar fabric is crenulated with the production of recumbent, east-verging isoclinal fold hinges. These small-scale folds may be representative of the regional mesoscopic structures illustrated in figure 2a and 2d. The lack of relict-igneous plagioclase phenocrysts, fine-grained nature, and high muscovite content suggest either that this rock is highly tectonized and the plagioclase phenocrysts altered to muscovite or that this is a tuffaceous metasediment comprising both a volcanic and a sedimentary component.

Sample NF30 is from a 5-m-thick, weakly deformed felsic volcanic flow that occurs within the NF along the Kafue Gorge Road at GR 35L 0643909-8250495 (figs. 2d, 4c). The unit has a rhyodacitic composition [SiO₂ = 68 wt% ; table 1] and contains relatively high concentrations of high-field strength elements such as Nb, Th, Y, and Zr (table 1). In thin section, rare, mostly tabular, relict-igneous plagioclase phenocrysts up to 1 mm in length are set in a fine-grained (less than 0.1 mm) equigranular matrix of plagioclase, K-feldspar,
Table 1. Whole-Rock Major- and Trace-Element Data for the Dated Igneous and Metasedimentary Samples

<table>
<thead>
<tr>
<th>Sample</th>
<th>KRF20</th>
<th>KRF21</th>
<th>NF30</th>
<th>LG44</th>
<th>LG46</th>
<th>NgF47</th>
</tr>
</thead>
<tbody>
<tr>
<td>Major elements (wt%):</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>68.85</td>
<td>71.20</td>
<td>68.04</td>
<td>74.37</td>
<td>76.49</td>
<td>70.81</td>
</tr>
<tr>
<td>TiO₂</td>
<td>.92</td>
<td>.56</td>
<td>.91</td>
<td>.30</td>
<td>.35</td>
<td>.64</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>13.33</td>
<td>13.17</td>
<td>13.91</td>
<td>13.28</td>
<td>12.03</td>
<td>12.83</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>4.60</td>
<td>4.65</td>
<td>8.61</td>
<td>2.11</td>
<td>2.33</td>
<td>2.33</td>
</tr>
<tr>
<td>MnO</td>
<td>.06</td>
<td>.06</td>
<td>.01</td>
<td>.03</td>
<td>.04</td>
<td>.00</td>
</tr>
<tr>
<td>MgO</td>
<td>1.13</td>
<td>1.89</td>
<td>1.25</td>
<td>.24</td>
<td>.31</td>
<td>1.55</td>
</tr>
<tr>
<td>CaO</td>
<td>1.05</td>
<td>1.18</td>
<td>.28</td>
<td>1.20</td>
<td>1.29</td>
<td>1.01</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.65</td>
<td>3.33</td>
<td>.01</td>
<td>3.25</td>
<td>2.81</td>
<td>.19</td>
</tr>
<tr>
<td>K₂O</td>
<td>4.37</td>
<td>4.55</td>
<td>5.74</td>
<td>5.19</td>
<td>4.66</td>
<td>4.13</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>.18</td>
<td>.12</td>
<td>.21</td>
<td>.06</td>
<td>.07</td>
<td>.01</td>
</tr>
<tr>
<td>Total</td>
<td>99.94</td>
<td>99.71</td>
<td>98.95</td>
<td>100.03</td>
<td>100.39</td>
<td>100.32</td>
</tr>
</tbody>
</table>

| Trace elements (ppm): | | | | | | |
| Ba    | 580  | 1124 | 722 | 419 | 440 | 456 |
| Ni    | …    | …    | 11  | 12  | 5   | 42   |
| Cu    | 16   | 4    | 2   | …   | 2   | 6    |
| Zn    | 32   | 37   | …   | 17  | 26  | 0    |
| Pb    | <4   | 4    | 5   | 4   | 2   | 2    |
| Th    | 24   | 28   | 16  | 36  | 30  | 16   |
| Rb    | 210  | 165  | 266 | 282 | 240 | 170  |
| Sr    | 82   | 246  | …   | 62  | 57  | 9    |
| Y     | 55   | 68   | 52  | 41  | 38  | 24   |
| Zr    | 640  | 566  | 538 | 185 | 196 | 205  |
| Nb    | 49   | 47   | 52  | 22  | 21  | 16   |
| U     | n/a  | n/a  | n/a | n/a | n/a | 6    |
| Cr    | n/a  | n/a  | n/a | n/a | n/a | 231  |
| V     | n/a  | n/a  | n/a | n/a | n/a | 166  |

Note. Analytical procedures are outlined in the appendix, available in the online edition and from the Journal of Geology office.

quartz, biotite, and muscovite. It is possible that the muscovite is not primary, having replaced matrix plagioclase during metamorphism. Matrix K-feldspar displays a perthite structure, suggesting that the feldspar has partially exsolved during metamorphism. The weak fabric observed on the outcrop scale is hard to distinguish in thin section.

**Lusaka Granite.** Sample LG44 was collected from the Geological Survey Seismic Station (GR 35L 062591-8310913) and is a pale gray, two-feldspar, coarse-grained granitoid (wt%; SiO₂ = 74 wt%; table 1). The unit has been moderately deformed so that the feldspar phenocrysts have been flattened and crudely wrapped by biotite (fig. 4d), forming a protomylonitic planar fabric. This fabric is cut at a shallow angle by 10-cm-thick, west-dipping mylonite zones. These zones are roughly parallel to the axial planes of the recumbent folds observed in sample KRF21 and may represent localized thrust zones developed during regional-scale folding (fig. 2a, 2d).

LG46 is a pink, K-feldspar–rich porphyritic granitoid that contains abundant quartz-rich, flattened xenoliths (fig. 4e). The sample was collected from a small roadstone quarry at GR 35L 0626178-8311730. The rock has a chemical composition similar to that of LG44 (table 1).

**Nega Formation.** Sample NgF47 is a fine-grained, pale gray, quartz-feldspar-sericite-bearing schistose portion of the conglomerate (fig. 4f) at the base of the Nega Formation (fig. 2b; GR 35L 621638-8240886). The sericitic matrix contains rounded grains of bluish opalescent quartz (1–2 mm diameter), more abundant grains of altered feldspar, and abundant fine-grained hematite. It is possible that the opalescent quartz was derived from the underlying volcanic units, upon which the Nega conglomerates were deposited.

**U-Pb Zircon Geochronology and Whole-Rock Sm-Nd Isotopes**

**Kafue Rhyolite and Nazingwe Formations.** The zircons extracted from each of the three samples are relatively simple in structure. They are 100–150 μm in length, although sometimes broken, and show typical igneous terminations and a simple oscillatory-zoned structure under cathodoluminescence (fig. 5a–5c). Some of the grains are cracked, and some contain abundant silicate melt inclusions.
Figure 5. Cathodoluminescence images of zircons from the samples in this study.

typical of zircons in volcanic rocks (Thomas et al. 2003). From the textural features and U and Th concentrations (table A1, available in the online edition and from the Journal of Geology office), we conclude that these grains are igneous in origin. The results are shown in table A1 and in Tera-Wasserburg plots in figure 6a–6c.

From sample KRF20, six inclusion-free grains were analyzed. The U and Th concentrations range from 78 to 111 and from 50 to 88 ppm, respectively, with Th/U ratios between 0.59 and 0.82, typical values for igneous zircon. The six analyses yield a coherent group with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of $880 \pm 12$ Ma (MSWD = 0.83; fig. 6a). The sample yields an initial $\varepsilon_{\text{Nd}}(t)$ value ($t = 880$ Ma) of $-2.28$ and a depleted-mantle model age of 1.56 Ga (table A2, available in the online edition and from the Journal of Geology office). These isotopic values indicate that the felsic magma was derived either by a mixture of juvenile material and Paleo-prototozoic continental crust or by partial melting of a mainly Mesoproterozoic continental crust, a possible source being the underlying 1.1-Ga Mpande Gneiss basement. Unfortunately, a lack of inherited zircon within the sample and the fact that the isotopic characteristics of the Mpande Gneiss
Figure 6. Tera-Wasserburg U-Pb concordia diagrams for the three volcanic samples from the Kafue Rhyolite and Nazingwe formations and the two Lusaka Granite samples. The weighted mean $^{206}\text{Pb} / ^{238}\text{U}$ ages of the main age populations are graphically represented as insets within each U-Pb plot.
From sample KRF21, nine different inclusion-free grains were analyzed. The grains have U and Th concentrations of 73–282 and 39–147 ppm, respectively, and Th/U ratios of 0.47–0.81 (table A1). The analyses reveal two age populations. One grain (KRF21-4) gives a concordant $^{207}\text{Pb}^{206}\text{Pb}$ age of $1078 \pm 15$ Ma (table A1; fig. 6b). The remaining eight grains represent a single population with a weighted mean $^{206}\text{Pb}^{238}\text{U}$ age of $876 \pm 10$ Ma [MSWD = 0.99]. The younger population is interpreted to give the age of the volcanic component of this tuffaceous unit (fig. 6b), while the older Mesoproterozoic grain is either an inherited grain from the basement or a detrital grain intermixed with the tuffaceous layer. The sample yields an initial $\varepsilon\text{Nd}_{(t)}$ value [$t = 876$ Ma] of $-4.50$ and a model age of ca. 2.00 Ga (table A2; fig. 6d). The high Nd model age suggests that the whole-rock isotopic values of the tuffaceous rock have been altered by the input of older sedimentary material or that the magma interacted with or was derived from older continental basement rocks.

From sample NF30, eight clear, equant, inclusion-free grains were analyzed. The grains have U and Th concentrations of 78–131 and 39–76 ppm, respectively, and Th/U ratios of 0.49–0.66, similar to values for the two previous volcanic samples (table A1). The analyses give a weighted mean $^{206}\text{Pb}^{238}\text{U}$ age of $880 \pm 14$ Ma [MSWD = 1.3], which we interpret as the crystallization age of the volcanic unit (fig. 6c). The sample yields an initial $\varepsilon\text{Nd}_{(t)}$ value [$t = 880$ Ma] of $-2.78$ and a depleted-mantle model age of 1.61 Ga, similar to that of KRF20 (table A2; fig. 6d). This again indicates that continental crust, >1.6 Ga in age, was involved in the petrogenesis of the felsic magmas.

**Lusaka Granite.** Zircons were extracted from two samples of the Lusaka Granite. Zircons from both samples are clear and 100–150 $\mu$m in length, showing typical igneous terminations and a simple oscillatory-zoned structure under cathodoluminescence (fig. 5e, 5f).

From sample LG44, a total of 12 grains were analyzed, giving ranges of U and Th concentrations of 148–1984 and 48–1167 ppm, respectively, with Th/U ratios of 0.34–0.82 (table A1). Analyses 2c, 6, 8 and 10c plot well away from concordia [20%–81% discordant] and are characterized by significant proportions of common Pb ranging from 0.26% to 1.56% (table A1; fig. 6d). The remaining eight grains form a coherent age group with a weighted $^{206}\text{Pb}^{238}\text{U}$ mean age of $821 \pm 9$ Ma [MSWD = 0.7] that is interpreted as the age of crystallization of the granite (fig. 6d).

From sample LG46, a total of 12 grains were analyzed, giving ranges of U and Th concentrations of 115–841 and 55–518 ppm, respectively, with Th/U ratios of 0.34–0.81 (table A1). The data range from concordant to discordant (up to 28% discordant), with discordance related directly to the proportion of common Pb, which ranges from 0.01% to 1.31%. Four concordant grains form a coherent age group with a weighted $^{206}\text{Pb}^{238}\text{U}$ mean age of $821 \pm 13$ Ma [MSWD = 0.11] that is interpreted as
Figure 8. Cartoon reconstructions of the Rodinia supercontinent (after Pisarevsky et al. 2003; Li et al. 2004; De Waele et al., forthcoming) and the inferred tectonic evolution of the Zambezi supracrustal sequence (including the Roan Rift Basin) at ca. 880 Ma and the Nguba-Kundelungu rift basins at ca. 765 Ma.

The age of crystallization of the granite (fig. 6e, 6f). One grain that is 100% concordant gives a Late Archaean 207Pb/206Pb age of 2889 ± 4 Ma (fig. 7a). The Mesoproterozoic zircons with ages within 10% of concordia (i.e., those between 90% and 110% concordant) are plotted as a combined histogram (25-m.yr. bin size) and probability density distribution (fig. 7b; for details, see Sircombe 2004). The plot suggests the presence of two main age populations, one with a predominant age of 1090 ± 10 Ma (four grains; MSWD = 0.2) and a slightly younger population defined by four variably discordant zircon grains at ca. 1050 Ma (fig. 7a, inset). The discordant nature of these younger grains (although they are still less than 10% discordant) suggest that they have lost Pb during regional metamorphism, and a regression of this data yields an upper-intercept age of 1078 ± 13 Ma and a lower-intercept age of 468 ± 190 Ma (MSWD = 1.2). The older age is within error of the weighted mean age of the most concordant grains at ca. 1090 Ma, suggesting that all the zircons were originally of the same age population, with some grains having undergone partial Pb loss during regional Pan-African metamorphism.

Discussion

Age of the Nega Formation. The 1090 ± 10-Ma population of concordant detrital zircons extracted from the Nega Formation conglomerate is indistinguishable in age from the underlying Munali Hills Granite, dated at 1090.1 ± 1.3 Ma (Katongo et al. 2004). The conglomerate also contains clasts of the Munali Hills Granite and the weakly banded metarhyolites and microgranites (figs. 2b, 4f), indicating that the conglomerate was directly sourced from the underlying basement rocks. Detailed mapping of the conglomerate-basement contact, although the contact is poorly exposed, demonstrates that the Nega Formation lithologies progressively truncate the rhyolites, the microgranites, and the main granitoid body (fig. 2b), suggesting an unconformable relationship.

The single Archaean-aged grain dated at ca. 2890 Ma may suggest either that part of the underlying basement is of Archaean age or that there was limited Archaean crust exposed at the time of deposition, or it may represent an inherited detrital zircon derived from a distant source. Our data support the idea that the Nega and Muzuma formations were deposited unconformably on top of the Munali Hills Granite and Mpande Gneiss, sometime after ca. 1090 Ma, and are not older than the Munali Hills Granite, as suggested by Hanson et al. (1988) and Porada and Berhorst (2000). Considering the age constraints and the overall lithotectonic simili-
ties between this sequence and that around Lusaka [fig. 3], we suggest that the Nega and Muzuma formations are a contiguous part of the ZSC.

Age and Nature of Volcanism and Basin Evolution. The age of the three volcanic units collected from within the Kafue Rhyolite Formation and the overlying Nazingwe Formation are all within error of each other at ca. 880 Ma, suggesting that replacement of the thick volcanic/volcanoclastic pile occurred in a relatively short time frame. The age also refines the previously unpublished TIMS age of $879 \pm 19$ Ma for a volcanic unit within the KRF [M. S. Wardlaw, unpublished data; reported in Wilson et al. 1993]. The xenocrystic/detrital zircon from KRF21, dated at 1078 $\pm 14$ Ma, suggests either that Late Mesoproterozoic crust forms part of the basement from which, or onto which, these volcanics were emplaced or that Late Mesoproterozoic crust was exposed and eroded during the deposition of the tuffaceous unit. The predominance of felsic material (98%–99%) over basic material was volumetrically insignificant, and magma generation was due to infracrustal melting initiated by a rise in geothermal gradients during the onset of crustal thinning, or that juvenile mafic magmas ponded in the lower crust, leading to extensive partial melting and crustal assimilation, a scenario similar to that suggested for the origin of the Tobifera rhyolites in Argentina that formed as a result of extensional processes during Gondwana breakup [Gust et al. 1985; Kay et al. 1987]. Combined with their within-plate geochemical characteristics [Katongo et al. 2004], the KRF and NF volcanics have Nd isotopic signatures consistent with their generation by mixing/assimilation of juvenile material with older continental crust, and we interpret this magmatic event to mark crustal thinning and continental extension. This period of volcanism was followed by a depositional/volcanic hiatus with the production of a regional unconformity and the redistribution of Mesoproterozoic basement rocks in the Nega section and felsic volcanics in the Lusaka section as locally derived basal conglomerates. In both sections, the main phase of sedimentation is represented by a progressively deepening sequence of shelf and platform deposits, from coarse, shallow-water siliciclastics [Nega and Mulola formations] to thick stromatoloidal limestones [Muzuma and Cheta formations; fig. 3]. The final stages of basin development involved the intrusion of at least two discrete but geochemically similar [Katongo et al. 2004] granitoids into the Zambezi supracrustal succession [figs. 1b, 2c]. Both granitoids, the Lusaka Granite and Ngoma Gneiss, have melanocratic A-type geochemical characteristics [Katongo et al. 2004] and crystallization ages within error at ca. 820 Ma, thus providing an upper limit on the age of sedimentation. Although we cannot adequately explain the tectonic significance of these two granitoids, their crustal geochemical signature [Katongo et al. 2004] precludes the involvement of significant juvenile material.

Basin Provenance and Regional Correlations. The Nega and Lusaka lithotectonic packages are interpreted as part of a single progressively deepening basin of shelf and platform deposits. The basin was initiated at ca. 880 Ma by continental thinning and extension, and the main phase of sedimentation was complete by ca. 820 Ma with the intrusion of the Ngoma and Lusaka granitoids, although the Lusaka Dolomite could be considerably younger. The Late Mesoproterozoic xenocrystic/detrital zircon within the tuffaceous volcanic KRF21 indicates that the basement of this age was involved in the formation of, or was exposed at the time of, extensional volcanism, with the underlying Mpande Gneiss and Munali Hills Granite as an obvious source. Basal conglomerates of the overlying supracrustal sequence were locally sourced, either from the KRF and NF volcanics or from the Late Mesoproterozoic basement itself. Late Mesoproterozoic volcano-plutonic crust predominates the southern margin of the CTB Craton in the Irumide [De Waele et al. 2003, 2006a; De Waele 2005] and Southern Irumide belts [Johnson et al. 2005, 2006] but is known only within a relatively small, allochthonous klippe along the northern margin of the Kalahari Craton, the Zambezi Allochthonous Terrane [Barton et al. 1993; Vinyu et al. 1999; Hargrove et al. 2003]. Furthermore, abundant xenocrystic zircons, limited basement exposures, and the isotopic character of the magmatic units in the Irumide and Southern Irumide belts suggest that Palaeoproterozoic and Archaean crust form substantial parts of the basement of these belts [De Waele 2005; De Waele et al. 2006a, 2006b; Johnson et al. 2006], which can adequately account for the few older Archaean zircon xenocrysts extracted from the Nega Conglomerate and Lusaka Granite. All of these observations indicate that the ZSC was deposited on top of the Mpande-Munali granitoid basement (the CTB Craton) and does not represent an allochthonous nappe that may have originally been deposited on the Kalahari side of the Zambezi Ocean. This is contrary to the conclusions of Armstrong et al. (2005), who promoted a correlation between the ZSC and the Makuki Group [MG] of northern Zimbabwe. We disagree with this correlation for several reasons:
1. Supposed supracrustal units within the MG, previously interpreted as felsic metavolcanic units [Munyanyiwa et al. 1997], have later been shown to represent a series of deformed granitoid sheets [Dirks and Sithole 1999; Dirks et al. 1999].
2. These granitoids have a distinctly different geochemical composition from the granitoids and volcanics in the ZSC [Katongo et al. 2004].
3. Magmatism in the MG occurred at a distinctly different time than volcanism and magmatism within the ZSC (data summarized in Johnson et al. 2005). Johnson et al. [2005] also demonstrated that the timing of magmatism within the MG was more comparable to that of rifting events in South China and Australia, promoting a view that these three cratons formed an integral component within the Rodinia supercontinent [Dalziel 1997; Powell et al. 2001; Li et al. 2003; Loewy et al. 2003; Pisarevsky et al. 2003]. Based on broad lithological and stratigraphic similarities, the ZSC has in the past been correlated with the lowermost Katangan sediments, the Roan Group, in the Copperbelt Region [Annels 1974; Binda 1994; Unrug 1988; Porada 1989; Porada and Berhorst 2000]. A maximum age of deposition of the Roan Group is defined by the Nchanga Granite, which is unconformably overlain by the Roan. The Nchanga Granite has yielded a zircon U-Pb SHRIMP age of 883 ± 10 Ma [Armstrong et al. 2005], an age identical to that of the KRF, and it also has a geochemical composition similar to that of the KRF [Katongo et al. 2004], suggesting that the Nchanga Granite may represent a deeper-level equivalent to the KRF. Potentially equivalent felsic volcanic rocks have also been described from the Roan Group around the Luswishi Dome [Porada 1989], as part of the Roches Argilo-Talqueuses [Cailleux et al. 1994] and from the lower Roan rocks from southeast Shaba in the Democratic Republic of Congo [Lefebvre 1989], but precise geochronological constraints are needed to confirm that they form part of the same magmatic event.

The most striking similarity between the Roan Group and the ZSC is the presence of thousands of isolated mafic blocks that sit within and at the contact between the upper pelite and marble formations [Vrána et al. 1975]. In the Roan Group, these mafic bodies have tholeiitic intraplate geochemical signatures and have been interpreted as syn-rifting dikes and sills [Tembo et al. 1999], but in the ZSC they have N-MORB and depleted mantle–like geochemical and isotopic signatures [John et al. 2003, 2004]. A preliminary U-Pb SHRIMP zircon crystallization age of 852 ± 22 Ma [S. P. Johnson, unpublished data] for the Munali Gabbro intrusion in the Nega section suggests that ZSC mafic rocks may also represent rift-related rocks similar to those in the Roan Group [Tembo et al. 1999].

Apart from the Lusaka Dolomite, which may or may not be younger than the Lusaka Granite [Simpsonton et al. 1963; Porada and Berhorst 2000], there is little evidence to suggest that sedimentation occurred after 820 Ma in this region. Renewed rifting and sedimentation is recorded in the northwest arm of the Lufilian Belt at ca. 765 Ma [Key et al. 2001], with the extrusion of the Luakela mafic lavas and tuffaceous deposits. This rift basin rapidly developed into a major marine basin, the Nguba and Kundelungu basins, which accommodated tens of kilometers of sediments [Annels 1974; Unrug 1988; Porada 1989; Binda 1994; Porada and Berhorst 2000; Key et al. 2001], including the two globally recognized glacial horizons (the Sturtian and Marinoan).

The CTB Craton and the Rodinia Supercontinent. There is still considerable debate as to whether the CTB Craton was an integral part of Rodinia [e.g., Torsvik 2003; Li et al. 2004] or whether it acted as an independent cratonic entity [e.g., Kröner and Cordani 2003; Pisarevsky et al. 2003; Johnson et al. 2005; De Waele et al., forthcoming]. Although the data presented in this article do not provide a definitive answer, they do indicate that the southern margin of the CTB Craton experienced two periods of continental extension and rifting, the latter episode at ca. 765 Ma, that occurred in a time frame similar to that of other Rodinia rifting events [Li et al. 2003, 2004; Pisarevsky et al. 2003]. Our data suggest that the ZSC and the Roan Group are temporal equivalents, both successions having been deposited in a discrete rift basin(s) along the southern margin of the CTB Craton during crustal thinning and extension. The limited period of sedimentation (between 880 and 820 Ma) and lack of evidence for the production of oceanic crust suggest that this rifting event may not have culminated in full continental separation [fig. 8a]. A comparison of the palaeomagnetic data from the 795-Ma Gagwe Lavas and a new palaeomagnetic pole from the 765-Ma Luakela Lavas [Wingate et al. 2005] with the 748-Ma Mbozi Complex indicate that the CTB Craton underwent a major 90° rotation between 765 and 748 Ma. Along with evidence for Grenvillian-age (ca. 1020 Ma) compressional tectonometamorphism [De Waele 2005], these data are more consistent with the CTB Craton having formed an integral part of Rodinia, culminating in
rifting at ca. 765 Ma to form the Neoproterozoic Zambezi Ocean and accommodating the deposition of the Nguba and Kundelungu basins (fig. 8b). Rifting, extension, and sedimentation associated with the event at ca. 880 Ma would represent one of the earliest records for the attempted breakup of Rodinia, and rifting of this age has yet to be identified in any of the other Rodinian blocks. Alternatively, if the CTB Craton acted as an independent entity, then the Grenvillian-age collision must have occurred in response to the collision/accretion of another Rodinia-independent craton or microcontinental fragment, a scenario not too dissimilar to the peri-Gondwanan terranes along eastern Laurentia, i.e., the Grampian and Appalachian events (e.g., Oliver 2001; Murphy and Keppie 2005). Rifting at ca. 880 and 765 Ma may have been in response to slab-rollback processes initiated by complex plate motions.

Conclusions

Our new U-Pb SHRIMP zircon crystallization age data for various igneous units within the ZSC and limited Sm-Nd isotopic data lead us to several conclusions:

1. The base of the Nega Formation is marked by a conglomerate containing granitic and volcanic clasts derived from the underlying Mesoproterozoic granitoids. The zircons extracted from the matrix have a single-age population identical to the underlying granitoids and gneisses, indicating that the basal Nega Formation sediments were locally sourced. Some of these zircon grains have undergone variable Pb loss during regional Pan-African metamorphism.

2. Both sedimentary successions can be interpreted as a single succession that was deposited as a result of crustal thinning and extension that began at 880 Ma. Intrusion of the Lusaka Granite, dated here at 820 Ma, provides an upper age limit for deposition of the ZSC sedimentary succession, although the Lusaka Dolomite may be younger than ca. 820 Ma.

3. The Kafue Rhyolite and Nazingwe Formation volcanic rocks formed, to a large degree, as a result of crustal assimilation and recycling of basement granitoids and gneisses upon which they unconformably, and in places tectonically, rest.

4. The ca. 880-Ma date for the Kafue Rhyolite and Nazingwe formations is identical to the age of exposed basement in the Copperbelt region of Zambia, the Nchanga Granite, upon which a similar sedimentary succession rests, i.e., the Roan Group. We agree with previous correlations between the Roan Group and the ZSC and suggest that both Roan and ZSC were deposited on the CTB side of the Neoproterozoic Zambezi Ocean.

5. Should the CTB Craton have formed part of the 1.1–0.7-Ga Rodinia Supercontinent, the 880-Ma rifting event may form the first evidence for attempted breakup of the landmass. Full oceanic separation of the CTB Craton from Rodinia probably occurred during the formation of the Nguba and Kundelungu basins at 765 Ma.

Acknowledgments

All authors would like to thank R. Hanson, L. Ashwal and coworkers, and an anonymous reviewer for critical comments that improved this manuscript. Special thanks are given to T. Prave, who commented on an early version of the manuscript. We would also like to thank members of the Centre for Microscopy and Microanalyses at the University of Western Australia and various staff members of the Geology Department at the University of Zambia for their assistance. Special thanks go to W. Nundwe for his hammering skills, great driving, and excellent company in the field. This is a contribution to International Geological Correlation Program 418/419/440 and Tectonics Special Research Center manuscript 360.

References Cited


Barton, C. M.; Carney, J. N.; Crow, M. J.; Dunkley, P. N.; and Simago, S. 1993. Geological and structural framework of the Zambezi belt, northeastern Zim


Kay, S. M.; Makasev, V.; Moscos, R.; Mpodzis, C.; and

Kroener, A., and Cordani, U. G. 2003. African, south In-

Key, R. M.; Liyungu, A. K.; Njamu, F. M.; Somwe, V.;


Lefebvre, J. J. 1989. Depositional environment of copper-


Li, Z. X.; Evans, D. A. D.; and Zhang, S. 2004. A 90° spin on Rodinia: possible causal links between the Neo-


doctrine of amphibolites and
correlations with other continent: evidence for a
mante superplume that broke up Rodinia. Precam-

Munyanyiwa, H.; Blenkinsop, T.; Hanson, R.; and Tre-


Porada, H., and Berhorst, V. 2000. Towards a new un-
derstanding of the Neoproterozoic–Early Palaeozoic Lufilian and Zambezi belts in Zambia and the Dem-

Powell, C. M.; Jones, D. L.; Pisarevsky, S. A.; and Win-
gate, M. T. D. 2001. Paleomagnetic constraints of the position of the Kalahari craton in Rodinia. Precam-


Tani, K.; Kawabata, H.; Qing, C.; Sato, K.; and Tatsumi, Y. 2005. Quantitative analyses of silicate rock major and trace elements by x-ray fluorescence spectrome-

Tembo, F.; Kampunzu, A. B.; and Porada, H. 1999. Tho-

Thieme, J. G. 1968. Structure and petrography of the Lu-

