There are ongoing debates and speculations about the origin of the oldest crust on Earth, as well as about the crust–mantle evolution during Hadean to Archaean times (e.g. Armstrong 1991; Amelin et al. 1999, 2000; Kamber et al. 2003, 2005; Condie et al. 2005; Harrison et al. 2005, 2008; Tolstikhin et al. 2006; Valley et al. 2006; Kramers 2007; Blichert-Toft & Albarède 2008; Nutman et al. 2008; Zeh et al. 2008, 2009; Kemp et al. 2009, 2010; Hawkesworth et al. 2010). Some of these speculations result from the scarcity of information about Hadean to Early Archaean rocks, owing to their limited exposure worldwide, but also from crust alteration processes, which can cause resetting or disturbance of certain isotope systems (e.g. Sm–Nd) during post-Hadean–Early Archaean volcanic and sedimentary events, and decoupling of certain isotope systems during multiple geological events (e.g. the Sm–Nd whole-rock zircon system; Vervoort et al. 1996; Moorbath et al. 1997).

Despite these uncertainties, much progress has been made over the past 10 years, in particular by the application of combined U–Pb and Lu–Hf isotope analyses to zircons from (meta)sedimentary rocks and orthogneisses from many Archaean gneiss terranes worldwide (e.g. Patchett et al. 1981; Vervoort et al. 1996; Griffin et al. 2004; Davis et al. 2005; Halpin et al. 2005; Hartlaub et al. 2006; Zeh et al. 2008, 2009; Kemp et al. 2009, 2010). In this context it is surprising to note that little is known so far about the crust–mantle processes that led to the formation and reworking of the oldest basement units of the Kaapvaal Craton, exposed in the Ancient Gneiss Complex of Swaziland (Fig. 1). This basement complex consists of tonalite–trondjemite–granodiorite (TTG) gneiss and granite samples from Swaziland reveal that the oldest rocks of the Ancient Gneiss Complex in southern Africa formed by reworking of Early Archaean or perhaps Late Hadean crust at 3.66 Ga, and that new crust was extracted from a depleted mantle source during Palaeoarchaean events between 3.54 and 3.32 Ga. This interpretation is supported by $\epsilon_{Hf}$ of $-1.6 \pm 2.0$ obtained from 3.66 Ga TTG gneisses, corresponding to hafnium model ages between 3.77 ± 0.18 Ga, for a presumed Hadean–Early Archaean chondritic mantle, and 4.08 ± 0.18 Ga, for a presumed Hadean depleted mantle reservoir, with the first model age being the most likely in the light of recent data from worldwide sources. Furthermore, it is reflected by superchondritic $\epsilon_{Hf}$, up to +2.2 ± 2.0 for TTGs formed at 3.54, 3.45 and 3.32 Ga. The new datasets additionally show that the Palaeoarchaean crust formed between 3.54 and 3.32 Ga was intensely reworked afterwards, without significant addition of depleted mantle derived material, during orogenic and intracratonic melting processes at 3.23, 3.1 and 2.7 Ga. This is well reflected by an array of decreasing $\epsilon_{Hf}$ from +2.2 to $-7.2$ between 3.3 and 2.7 Ga, which can be forced by $^{176}$Lu/$^{177}$Hf of 0.0113, which is similar to that of present-day average continental crust, and might result from lower crust zircon fractionation during Archaean crust reworking.

Abstract: Combined U–Pb and Lu–Hf isotope analyses of zircons from 16 tonalite–trondjemite–granodiorite (TTG) gneiss and granite samples from Swaziland reveal that the oldest rocks of the Ancient Gneiss Complex in southern Africa formed by reworking of Early Archaean or perhaps Late Hadean crust at 3.66 Ga, and that new crust was extracted from a depleted mantle source during Palaeoarchaean events between 3.54 and 3.32 Ga. This interpretation is supported by $\epsilon_{Hf}$ of $-1.6 \pm 2.0$ obtained from 3.66 Ga TTG gneisses, corresponding to hafnium model ages between 3.77 ± 0.18 Ga, for a presumed Hadean–Early Archaean chondritic mantle, and 4.08 ± 0.18 Ga, for a presumed Hadean depleted mantle reservoir, with the first model age being the most likely in the light of recent data from worldwide sources. Furthermore, it is reflected by superchondritic $\epsilon_{Hf}$, up to +2.2 ± 2.0 for TTGs formed at 3.54, 3.45 and 3.32 Ga. The new datasets additionally show that the Palaeoarchaean crust formed between 3.54 and 3.32 Ga was intensely reworked afterwards, without significant addition of depleted mantle derived material, during orogenic and intracratonic melting processes at 3.23, 3.1 and 2.7 Ga. This is well reflected by an array of decreasing $\epsilon_{Hf}$ from +2.2 to $-7.2$ between 3.3 and 2.7 Ga, which can be forced by $^{176}$Lu/$^{177}$Hf of 0.0113, which is similar to that of present-day average continental crust, and might result from lower crust zircon fractionation during Archaean crust reworking.

Supplementary material: Results of in situ U–Pb and Lu–Hf isotope zircon analyses and concordia diagrams are available at www.geolsoc.org.uk/SUP18465.
worth noting that the gneisses of the Ancient Gneiss Complex underwent a complex history between 3.66 and 3.0 Ga, as is reflected by complex age patterns obtained from single zircons using in situ U–Pb sensitive high-resolution ion microprobe (SHRIMP) dating (Compston & Kröner 1988; Kröner et al. 1989). At present, it is unclear whether the different zircon ‘SHRIMP ages’ reflect new zircon formation or just alteration, and if the different events were accompanied by the addition of new crustal material.

Most of the problems mentioned above can be circumvented using recent developments in analytical techniques, in particular by applying combined cathodoluminescence (CL) imaging and in situ U–Pb and Lu–Hf isotope analyses to single zircon grains or well-defined zircon domains. This approach has been proven to be extremely useful in unravelling the magmatic–metamorphic history of complex polymetamorphic Archaean gneiss terranes (e.g. Gerdes & Zeh 2009; Kemp et al. 2009, 2010; Millonig et al. 2010; Zeh et al. 2010a). Besides the advantage of spatial resolution, combined U–Pb and Lu–Hf isotope patterns additionally allow distinction between zircon zones formed or altered during distinct metamorphic events (Gerdes & Zeh 2009), even if primary features such as initial U–Pb ages or zoning patterns were erased during later alteration events (Zeh et al. 2010a,b). In this study we will apply this method to zircons from 16 TTG gneisss and granite samples from Swaziland to set unambiguous constraints for the crust–mantle evolution of the oldest rocks of the Kaapvaal Craton, and their subsequent Archaean evolution until 2.7 Ga. The sample locations and coordinates are shown in Figure 1 and Table 1.

**Analytical techniques**

**Sample preparation**

Zircon grains were selected from 2–5 kg rock samples by means of standard crushing techniques (jaw crusher, disc mill), a Wilfley table and heavy liquids. Subsequently, zircon grains of the respective samples were selected by hand-picking under a binocular microscope, mounted in epoxy resin (up to seven samples per 1 inch block), and ground down to expose their centres. After polishing, but prior to isotope analyses, all grains were imaged using a Jeol JSM-6490 scanning electron microscope (SEM) with Gatan MiniCL at Goethe University Frankfurt (GUF) to obtain information about their internal structure (Fig. 2). Based on the CL images, between 12 and 41 U–Pb and Lu–Hf analyses were obtained from selected zircon domains of each sample (see Table 2). In a first step, U–Pb analyses were carried out, and during a later session Lu–Hf isotope analyses were obtained from the same domains, by setting the Lu–Hf spot either directly ‘on top’ of the U–Pb spot or immediately beside it but within the same growth zone as obtained by CL (see Fig. 2).

**LA-SF-ICP-MS U–Pb dating**

Uranium, thorium and lead isotopes were analysed by laser ablation sector field inductively coupled plasma mass spectrometry (LA-SF-ICP-MS) using a Thermo-Finnigan Element 2 SF-ICP-MS system coupled to a New Wave Research UP-213 laser...
system with a teardrop low-volume cell at GUF following Gerdes & Zeh (2006, 2009) and Frei & Gerdes (2009). Laser spot-sizes were 30 μm with a typical penetration depth of c. 15–20 μm. A common-Pb correction based on the interference- and background-corrected $^{204}$Pb signal and model Pb composition (Stacey & Kramers 1975) was carried out where necessary. The necessity of the correction was usually based on the $^{206}$Pb/$^{204}$Pb value ($<10 \times 000$). However, in case the interference-corrected $^{204}$Pb could not be precisely detected (e.g. <20 counts per second (c.p.s.)), this was applied only when the corrected $^{207}$Pb/$^{206}$Pb was outside the internal errors of the measured ratios and yielded more concordant results (about 1–2% better concordance). The interference of $^{204}$Hg (mean = 255 ± 17 c.p.s.) on the mass 204 could not be precisely detected (e.g., 10 000). However, in case the interference-corrected $^{204}$Pb ($^{204}$Hg) was applied only when the corrected $^{207}$Pb/$^{206}$Pb ($^{204}$Hg) was $<10 \times 000$. However, in case the interference-corrected $^{204}$Pb could not be precisely detected (e.g. <20 counts per second (c.p.s.)), this was applied only when the corrected $^{207}$Pb/$^{206}$Pb was outside the internal errors of the measured ratios and yielded more concordant results (about 1–2% better concordance). The interference of $^{204}$Hg (mean = 255 ± 17 c.p.s.) on the mass 204 was estimated using a $^{204}$Hg/$^{206}$Hg ratio of 0.2299 and the measured $^{202}$Hg. All data were normalized relative to the GJ-1 zircon Plesˇovice was analysed to check quality and accuracy of the obtained data. Fifteen analyses yield a concordia age of 2.49 (Ludwig 2001).

LA-MC-ICP-MS $\text{Lu}–\text{Hf}$ isotope analyses

Hafnium isotope measurements were made with a Thermo-Finnigan Neptune multicollector (MC)-ICP-MS system at GUF coupled to a New Wave Research UP-213 laser system with a
Table 2. Compilation of U–Pb and Lu–Hf isotope results obtained during this study

| Sample | Age ± 2σ (Ma) | Type | c/s/z | Core (Ga) | Minor domain (Ga) | 176Hf/177Hf<sub>int</sub> (mean) | ±2σ | εHf<sub>int</sub> (mean) | ±2σ | No. | Age (Ma) |
|--------|---------------|------|-------|-----------|-------------------|---------------------------|------|----------------|------|-----|-----------|
| AG7    | 3644 ± 7      | conc. | 23/40/38 | 3.64      | –                 | 0.28037                   | 6    | –1.6          | 2.0  | 23/26 | 3663 ± 0.5 |
| AG6c   | 3662 ± 17     | u.i.  | 02/40/38 | 3.19      | –                 | 0.28053                   | 4    | –6.8          | 1.5  | 2/26 | 3223 ± 1.9 |
| AG6a   | 3219 ± 12     | conc. | 02/49/29 | 3.66      | –                 | 0.28037                   | 7    | –1.3          | 2.6  | 20/36 | c. 3660 |
| AG6b   | 3321 ± 12     | conc. | 06/23/19 | 3.32      | –                 | 0.28069                   | 1    | –0.4          | 0.4  | 4/36 | –         |
| AG6b   | 3230 ± 9      | conc. | 10/34/32 | 3.23      | –                 | 0.28069                   | 6    | –0.3          | 2.1  | 13/16 | –         |
| SW26   | 3539 ± 10     | conc. | 03/12/11 | 3.54      | –                 | 0.28052                   | 3    | +1.6          | 1.2  | 4/4  | –         |
| DA16   | 3433 ± 24     | conc. | 13/25/23 | 3.43      | –                 | 0.28054                   | 6    | –0.7          | 2.3  | 5/25 | –         |
| DA17   | 3456 ± 8      | conc. | 22/27/23 | 3.46      | –                 | 0.28058                   | 3    | +1.1          | 1.0  | 27/27 | –         |
| DA18   | 3436 ± 8      | conc. | 18/27/21 | 3.44      | –                 | 0.28059                   | 4    | +1.3          | 1.6  | 25/25 | –         |
| MB30   | 3436 ± 8      | conc. | 18/25/20 | 3.44      | –                 | 0.28057                   | 5    | +0.5          | 1.8  | 12/12 | –         |
| SW25   | 3323 ± 7      | conc. | 22/27/23 | 3.32      | –                 | 0.28070                   | 3    | +2.2          | 1.2  | 26/26 | –         |
| MB31   | 3237 ± 7      | conc. | 19/29/25 | 3.24      | –                 | 0.28069                   | 5    | –0.3          | 1.6  | 23/23 | –         |
| SW29   | 3217 ± 8      | conc. | 17/25/29 | 3.28      | –                 | 0.28069                   | 4    | –0.5          | 1.4  | 16/16 | –         |
| SIN25  | 3067 ± 12     | conc. | 05/23/10 | 3.07      | –                 | 0.28073                   | 6    | –2.8          | 2.0  | 14/14 | 3074 ± 4 |
| MB32   | 2722 ± 7      | conc. | 14/38/27 | 2.72      | –                 | 0.28085                   | 5    | –6.6          | 1.7  | 34/34 | 2691 ± 2 |
| SIC19  | 2722 ± 10     | conc. | 12/21/20 | 2.72      | –                 | 0.28083                   | 3    | –7.2          | 1.1  | 18/18 | 2723 ± 7 |
| NG22   | 2727 ± 9      | conc. | 16/28/24 | 2.73      | –                 | 0.28084                   | 5    | –6.9          | 1.8  | 15/15 | 2722 ± 4 |

1 Intrusion age; conc., concordia age; u.i., upper intercept age.
2 c/s/z, c, number of zircon analyses used for age calculation; s, number of spot analyses per sample; z, number of zircon grains analysed per sample.
3 Zircon cores and dominant zircon population.
4 Minor zircon population.
5 Average 176Lu/177Hf<sub>int</sub> calculated by applying the ages given in column 2.
6 Average εHf<sub>int</sub> calculated by applying the age in column 2.
7 Number of zircon spots used for mean 176Lu/177Hf<sub>int</sub> and εHf<sub>int</sub> calculation/Lu–Hf spot analyses per sample.
8 Published U–Pb zircon age for the respective plutons: (1) Maphalala & Kröner (1993); (2) Layter et al. (1989); (3) Schoene et al. (2008); (4) Kroener et al. (1989).

Teardrop-shaped, low-volume laser cell following Gerdes & Zeh (2006, 2009). Multiple analyses of Lu- and Yb-doped JMC 475 solutions show that results with a similar precision and accuracy can be achieved also if Yb/Hf and Lu/Hf are 5–10 times higher than in most magmatic zircons. All data were adjusted relative to the JMC 475 176Hf/177Hf ratio of 0.282160 and quoted uncertainties are quadratic additions of the within-run precision and the reproducibility of the 40 ppb JMC 475 solution (2SD ≈ 0.0033%, n = 5 during the analytical session). The correctness of the adjusted application is reflected by the data for the standard zircon GJ-1 (0.282011 ± 0.000032 2SD, n = 21), which were obtained during the same session as the sample zircon data.

For calculation of εHf, the chondritic uniform reservoir (CHUR) was used as recommended by Bouvier et al. (2008; 176Lu/177Hf = 0.0336 and 176Hf/177Hf = 0.282785), and a decay constant of 1.867 × 10⁻¹¹ (average of Scherer et al. 2001; Söderlund et al. 2004). Initial 176Hf/177Hf and εHf for all analysed zircon domains were calculated using the apparent Pb–Pb ages obtained for the respective domains, and for all co-genetic zircon domains by using the intrusion ages of the respective granitoids. Mean values for 176Hf/177Hf and εHf (and related errors ±2σ) for the respective granitoids are summarized in Table 2. Depleted mantle hafnium model ages (TDM) were calculated using values for the depleted mantle as suggested by Blichert-Toft & Puchtel (2010), with 176Hf/177Hf = 0.283294 and 176Lu/177Hf = 0.03933, corresponding to a straight DM evolution line with εHf<sub>today</sub> = +18 and εHf<sub>558Ga</sub> = 0.0. TDM ages for all data were calculated by using the measured 176Lu/177Hf of each spot for the time since zircon crystallization, and a mean 176Lu/177Hf of 0.0113 for the Palaeoproterozoic–Archaean crust (mean of average continental crust as suggested by Taylor & McLennan 1985 and Wedepohl 1995). In addition, TDM ages were calculated using 176Lu/177Hf = 0.02 as recommended for Hadean–Archaean mafic (protocrust) crust (Kemp et al. 2010), in combination with depleted mantle parameters of Blichert-Toft & Puchtel (2010). These model ages are on average between 0.07 and 0.38 Ga older than those calculated with 176Lu/177Hf = 0.0113. Because the existence of a depleted mantle reservoir during the Hadean–Early Archaean is highly speculative (see Hawkesworth et al. 2010; Kemp et al. 2010; and discussion below), we additionally calculated T<sub>CHUR</sub> model ages for the oldest zircons from Swaziland, using the CHUR parameters of Bouvier et al. (2008) in combination with 176Lu/177Hf = 0.02 for mafic crust. The T<sub>CHUR</sub> ages are on average 0.3 Ga younger than the corresponding TDM ages. It must be noted, however, that there are many other options to obtain model ages. For example, a lower 176Lu/177Hf<sub>0.007</sub>, which is typical for present-day upper crust (Taylor & McLennan 1985; Wedepohl 1995), would decrease the TDM of our zircon analyses by about 0.01–0.2 Ga, relative to 176Lu/177Hf = 0.0113, whereas a higher 176Lu/177Hf of 0.015, as proposed by Rudnick & Gao (2003) for the bulk continental crust, would increase the
T_{DM} by 0.12 Ga (on average). Additional uncertainties with similar influence on the calculated T_{DM} result from the $^{176}$Lu/$^{177}$Hf heterogeneity of depleted mantle melts (e.g. Chauvel & Blichert-Toft 2001). The application of hafnium model ages to Archaean–Hadean rocks is discussed below in the section ‘Crustal evolution’.

**Results and interpretations**

For data interpretation we use the combined set of CL images, and U–Pb and Lu–Hf isotope analyses (Figs 2–4; Table 2). The CL images of zircons from most granitoids revealed typical oscillatory magmatic zoning and monophase growth patterns (not shown), except for zircons of the TTG gneisses from the Piggs Peak inlier (samples AG7 and AG6a–c) and from the Mankanye area (sample DA16). These commonly show core–rim relationships (Fig. 2). It is important to note that multiple analyses of the ‘monophase’ zircons always yielded, within error, identical $^{176}$Hf/$^{177}$Hf, even if the Pb–Pb ages were significantly different (e.g. zrc4 in sample DA17, zrc11 in sample SW25, zrc11 in sample MB32), and that a large number of analyses of ‘monophase’ zircons from single samples ($n = 12–23$) show relatively small $^{176}$Hf/$^{177}$Hf variations, mostly between $\pm 1.0$ and $\pm 2.2$ epsilon units when all data are calculated back for the time of intrusion (Fig. 3a, Table 2). These variations are in most cases well within the range obtained by multiple analyses of the standard zircon GJ-1 ($\pm 1.1$ epsilon units, 2SD). This indicates that all zircons (and zircon domains) in these samples formed during magma crystallization, that the respective magmas had (in most cases) a relatively homogeneous Hf isotope composition ($^{176}$Hf/$^{177}$Hf variations smaller than $\pm 1.3$ epsilon units), and that some zircon domains were affected by multiple Pb loss after crystallization, without changing the Hf isotope composition significantly (equal to the case 1 and case 2 scenarios of Zeh et al. 2009; see Fig. 3a).

This implies that the oldest U–Pb zircon ages reflect the time of granite intrusion, or are nearest to it. Taking this into account, we obtained intrusion ages of 3.54 Ga (sample SW26), 3.46–3.44 Ga (samples DA17, DA18, MB30), 3.32 Ga (sample SW25), 3.21–3.24 Ga (samples DA16, MB31, SW29), 3.07 Ga (sample SIN23), and 2.72–2.73 Ga (samples MB32, SIC19, NG22), which are similar to ages obtained during previous studies from similar rocks of Swaziland (Table 2). Combination of the intrusion ages with the Lu–Hf isotope analyses yielded superchondritic $^{176}$Hf/$^{177}$Hf for the granitoids that intruded between 3.54 and 3.32 Ga ($\epsilon^{176}_{Hf} = +1.6 \pm 1.2$; $\epsilon^{176}_{Hf} = +1.3 \pm 1.6$ to $+0.5 \pm 1.8$; $\epsilon^{176}_{Hf} = +2.2 \pm 1.8$), and successively more subchondritic $\epsilon^{176}_{Hf}$ for all younger granitoids ($\epsilon^{176}_{Hf} = -0.3 \pm 1.6$; $\epsilon^{176}_{Hf} = -2.8 \pm 2.0$; $\epsilon^{176}_{Hf} = -7.2 \pm 1.1$) (Fig. 3, Table 2).

In contrast to the ‘monophase’ zircons, the zoned zircons as observed in samples AG7, AG6a–c and DA16 yielded significantly different concordant U–Pb ages and $^{176}$Hf/$^{177}$Hf for their cores and rims (Figs 2 and 3), far outside the external reproducibility of the method (equal to the case 4 scenario of Zeh et al. 2009). In situ U–Pb analyses of zircon cores from the TTG gneiss sample AG7, taken near the southern boundary of the Phophonyane shear zone (Fig. 1), gave a concordant U–Pb age of 3.191 ± 0.009 Ga, and a few rims gave a younger upper intercept U–Pb age of 3.191 ± 0.009 Ga. Similar ages were also obtained from samples AG6a–c, which were taken from an outcrop at the road near ‘the falls’, where dark banded tonalite gneisses are transected by dykes and bodies of granodiorite composition.

Zircons from the dark banded tonalite gneiss (sample AG6c) yielded three ages at 3.662 ± 0.017 Ga ($n = 10$), 3.219 ± 0.013 Ga ($n = 2$), and 3.127 ± 0.009 Ga ($n = 7$). The oldest age was obtained from zircon cores, and the two younger ages from zircon rims, as well as from needle-like zircons (Fig. 2). In contrast, zircons from a granodiorite dyke (sample AG6a) and body (sample AG6b), which cross-cut the dark gneiss, yielded concordant ages of 3.221 ± 0.012 Ga and 3.230 ± 0.009 Ga, respectively, and only a few zircon cores gave older ages of 3.52–3.56 Ga (Table 2).

Apart from CL images and U–Pb ages, zircon growth events in the rocks from the Piggs Peak inlier can also be discriminated on the basis of $^{176}$Hf/$^{177}$Hf (Table 2). The oldest zircons (or zircon domains) in samples AG7 and AG6c yielded the lowest $^{176}$Hf/$^{177}$Hf of c. 0.28037, which corresponds to $\epsilon^{176}_{Hf}$ of $-1.6 \pm 2.0$ (AG7) and $-1.3 \pm 2.6$ (AG6c), and to $T_{DM}$ model ages between 3.95 ± 0.18 and 4.08 ± 0.18 Ga. T_{CHUR} ages of the oldest zircons are 3.77 ± 0.21 Ga. In contrast, zircons (or domains) formed at c. 3.2 Ga yielded higher $^{176}$Hf/$^{177}$Hf of c. 0.28069 ($\epsilon^{176}_{Hf} = -0.4$), and the youngest zircon population formed at 3.127 Ga yielded the highest $^{176}$Hf/$^{177}$Hf of 0.28077 ± 0.00004 ($\epsilon^{176}_{Hf} = -0.1$) (Table 2).
Discussion

Age and hafnium isotope data

The presented CL images and U–Pb–Lu–Hf isotope data indicate that the igneous protoliths of the oldest gneisses of the Ancient Gneiss Complex, which occur in the Piggs Peak inlier, were emplaced at 3.66–3.64 Ga, and were affected by two subsequent events at 3.22 and 3.13 Ga. The two older ages are in agreement (within error) with those previously obtained by high-precision U–Pb single-grain dating on zircons from other banded tonalitic gneisses samples (3.663 Ga) and related granite dykes (3.223 Ga) (Schoene & Bowring 2007; Schoene et al. 2008), and by U–Pb SHRIMP dating of zircon cores (3.644 Ga : Compston & Kro¨ner 1988; Kro¨ner et al. 1989). In contrast to our dataset (and to Schoene & Bowring 2007), Compston & Kro¨ner (1988) and Kro¨ner et al. (1989) reported a large number of additional U–Pb zircon SHRIMP ages of 3.58, 3.50, 3.43, 3.20, 3.0 and 2.99–2.87 Ga, which they interpreted to reflect different post-intrusive thermal–metamorphic–tectonic events. However, the meaning of the different ages in a geological context is still unclear. An unambiguous interpretation of these data is hampered by the lack of CL or SEM images for these zircons, by very complex Pb loss patterns, and by the negative concordance of many analyses. It is likely that the 3.58–3.50 Ga event reported by Kro¨ner et al. (1989) correlates with granitic intrusions in the Piggs Peak inlier at 3.546 Ga (Phophonyane granite: Schoene & Bowring 2007), where there is no evidence for magmatic or metamorphic events at 3.43, 3.0 or <3.0 Ga.

The 3.23 Ga U–Pb zircon ages obtained from the granodiorite dykes and bodies of the Piggs Peak inlier during this study (AG6a–c, AG7) agree well with ages of granite dykes dated by Schoene et al. (2008; 3.223 ± 0.002 Ga), with emplacement ages obtained from granitoid rocks from the Usutu Suite in the southern part of Swaziland (3.236–3.220 Ga; Zeh et al. 2009; Schoene & Bowring 2010; this study), but also with those from Barberton Mountain Land to the north and south of the Saddleback–Inyoka fault system (Fig. 1), comprising the Dalmain granite and the Kaap Valley, Nelshoogte and Stentor plutons (Kamo & Davis 1994; Schoene et al. 2009; Zeh et al. 2009). Furthermore, the 3.23 Ga zircons from the Piggs Peak inlier have, within error, identical \(^{176}\text{Hf}/^{177}\text{Hf} = 0.28069 ± 0.00006\) to the TTGs near Manzini \(^{176}\text{Hf}/^{177}\text{Hf} = 0.28064 ± 0.00003\) (Zeh et al. 2009), from the Usutu Suite (sample SW29: 0.28069 ± 0.00005), near Mbabane (sample MB31: 0.28069 ± 0.00004) and west of Mankanye (sample DA16: 0.28069 ± 0.00003), and to the zircons of the Dalmain granite (0.28070 ± 0.00003; Zeh et al. 2009), which intruded the southern part of Barberton Mountain Land at c. 3.2 Ga (Fig. 1, Table 2). In contrast, zircons from the contemporaneous Kaap Valley and Nelshoogte plutons, which are exposed north of the Saddleback–Inyoka fault system, yielded significantly more radiogenic \(^{176}\text{Hf}/^{177}\text{Hf} = 0.28076 ± 0.00003\) (Zeh et al. 2009). In fact, the Lu–Hf isotope data of this study support previous conclusions that the Saddleback–Inyoka fault system (or the central part of the Barberton Greenstone Belt) represents an important terrane boundary, which separates a southern, less radiogenic terrane (Barberton South) from a northern, more radiogenic terrane (Barberton North). This separation is reflected not only by the Lu–Hf isotope data (Zeh et al. 2009, this study; Fig. 4f), but also by Sm–Nd isotope data (Schoene et al. 2009). Schoene
et al. (2009) interpreted this isotope difference (together with many other field observations and time constraints) to result from magmatism above a doubly vergent subduction zone at c. 3.2 Ga, whereby the slab-induced magmas formed beneath the southern terrane (comprising the Ancient Gneiss Complex, and Stolzburg and Steyndor terranes of Swaziland) assimilated higher amounts of older crust than the magmas formed beneath the northern terrane. Assimilation of pre-3.23 Ga gneisses during magmatism at 3.23 Ga in the southern terrane is well reflected by zircon xenocrysts with ages between 3.65 and 3.43 Ga (and with lower radiogenic $^{176}\text{Hf}^{177}\text{Hf}$) in rocks from the Pigg's Peak inlier, and in the area west of Mankyane (samples AG6a, AG6b and DA16; this study) and around Manzini (samples AGC1 and AGC2; Zeh et al. 2009). The 3.13 Ga event detected in the zircons from sample AG6c from the Pigg's Peak inlier correlates well with emplacement ages obtained from the potassium-rich granite batholiths, which form voluminous, sheet-like bodies in Swaziland and around the Barberton greenstone belt, comprising the Pigg's Peak batholith (3.10 Ga), Mpuluzi batholith (3.08 Ga) and Nelspruit batholith (3.11 Ga), and the Boesmanskop syenite (3.10 Ga) (age data from Zeh et al. 2009). In further agreement, the 3.13 Ga magmatic zircons from sample AG6c have, within error, identical $^{176}\text{Hf}^{177}\text{Hf}$ (0.28077 ± 0.00004) to the zircons from the batholiths (0.28075 ± 0.00003 to 0.28078 ± 0.00004; data from Zeh et al. 2009). This relationship hints that the 3.66 Ga banded tonalitic gneisses of the Pigg's Peak inlier (sample AG6c) were infiltrated by younger magma, which fed the nearby Pigg's Peak batholiths at 3.10 Ga (Fig. 1), and caused the crystallization of new zircon of this age as discrete crystals and rims around older cores. It is worth noting, however, that clear evidence for infiltration has not been observed in sample AG6c, although the occurrence of tiny, banding-parallel younger layers cannot be excluded.

Zircons from the Sinceni granodiorite yielded a U–Pb zircon age of 3.067 ± 0.012 Ga, which is, within error, identical to the age obtained by Maphalala & Kröner (1993). This age confirms previous conclusions that the Sinceni granodiorite was emplaced later than the large batholiths in Swaziland. Although the emplacement ages of the Sinceni granodiorite and the batholiths are significantly different, their $^{176}\text{Hf}^{177}\text{Hf}$ values overlap (sample SIN23: 0.28073 ± 6; batholiths: 0.28075 ± 0.00003 to 0.28078 ± 0.00004; data from Zeh et al. 2009) (Table 2). In contrast, the youngest granites of Swaziland, which intruded at 2.72 Ga, show much higher $^{176}\text{Hf}^{177}\text{Hf}$ than the Sinceni granodiorite. This is well reflected by the data from the Mbabane granite (sample MB32: 2.722 ± 0.007 Ga; $^{176}\text{Hf}^{177}\text{Hf}$ = 0.28085 ± 0.00005), Sicunusa granite (sample SIC19: 2.72 ± 0.010 Ga; $^{176}\text{Hf}^{177}\text{Hf}$ = 0.28083 ± 0.00003), and Ngwempisi granite (sample NG22: 2.727 ± 0.009 Ga; $^{176}\text{Hf}^{177}\text{Hf}$ = 0.28084 ± 0.00005). The U–Pb ages for the Sicunusa and Ngwempisi granites are identical, within error, to those previously obtained by Maphalala & Kröner (1993), but significantly older than the age of 2.691 ± 0.002 Ga reported by Layer et al. (1989) for the Mbabane granite. Furthermore, it should be noted that the Mpageni granite, which intruded the northern part of the Barberton Mountain Land (Fig. 1) at nearly the same time (2.69 Ga), shows very similar $^{176}\text{Hf}^{177}\text{Hf}$ of 0.28088 ± 0.00004 (Zeh et al. 2009).

Crustal evolution

The new datasets reveal that the protoliths of the oldest TTG gneisses of Swaziland were emplaced at 3.66 Ga by reworking of an even older crust, as is reflected by the subchondritic $^{176}\text{Hf}^{177}\text{Hf}$ of the oldest zircons. However, putting an exact age on this older crust is difficult and involves many uncertainties, which amongst others result from the choice of the parameters used to calculate appropriate hafnium model ages (see section ‘LA-MC-ICP-MS Lu–Hf isotope analyses’), and especially from the problem that the Hf isotope evolution of the Earth’s mantle during the Hadean–Early Archaean is not well constrained (see the discussions by Valley et al. 2006; Zeh et al. 2008, 2009; Hoffmann et al. 2010; Kemp et al. 2010). In fact, it is still a matter of debate whether or not a voluminous depleted mantle reservoir, which formed in response to important continental crust extraction, already existed during the Hadean to Early Archaean. Although the existence of a depleted reservoir (with respect to hafnium isotopes) was suggested on the basis of zircon solution Hf isotope data from Greenland and the Jack Hills (Vervoort et al. 1996; Vervoort & Blechert-Toft 1999; Blechert-Toft & Albarède 2008), and by the early zircon laser ablation data from the Jack Hills (Harrison et al. 2005), its existence could not be verified by more recent in situ laser ablation U–Hf–U–Pb zircon isotope studies (see Fig. 5a), which have revealed only chondritic to subchondritic Hf (Harrison et al. 2008; Kemp et al. 2009, 2010). Kemp et al. (2009, 2010) argued that the systematically higher radiogenic hafnium values obtained by the zircon solution Hf data from Greenland and the Jack Hills result from younger zircon overgrowths that could not be removed prior to dissolution, and that the large laser spots used by Harrison et al. (2005) sampled both cores and rims of the complexly zoned Jack Hills zircons. Taking this into account, the most reliable (trustworthy) zircon dataset available at present, as compiled by Kemp et al. (2010), provides no evidence for the existence of a depleted (global-scale) mantle reservoir during the Hadean–Eoarchaean. In contrast, it points to the existence of a volumetrically insignificant ‘KREEP’-like protocrust (KREEP = high potassium, REE and phosphorus) that was continuously reworked by remelting over nearly 400 Ma (without addition of new depleted mantle material), and finally disappeared with the onset of new, juvenile crust formation during the Archaean at <4.0 Ga (also see Kamber et al. 2003, 2005; Zeh et al. 2008). Based on the most reliable zircon datasets, Kemp et al. (2009, 2010) also argued that most of the juvenile Early Archaean crust was extracted directly from a chondritic mantle reservoir rather than from a depleted mantle source. This final interpretation is, apart from the most reliable zircon datasets (Kemp et al. 2010), also supported by solution Hf isotope data obtained from 3.72–3.75 Ga Isua pillow basalts and metasediments (Polat et al. 2003; Hoffmann et al. 2010), which all scatter around CHUR (Fig. 5a and b). Nevertheless, highly positive Hf values up to +13.5 obtained from pristine 3.75 Ga Isua boninites (Hoffmann et al. 2010) provide evidence that the Early Archaean mantle was heterogeneous, and contained spatially restricted, highly depleted (long-term isolated) reservoirs (Fig. 5b). Similar mantle reservoirs have also been suggested to be the source for some Barberton komatitites (eHf45Ga = +2.3 to +7.5), which extruded contemporaneously with much less depleted tholeiitic basalts (eHf45Ga = +2.3 to −0.5) in the Barberton Greenstone Belt, to the north of Swaziland (Blechert-Toft & Arndt 1999).

$^{176}\text{Hf}^{177}\text{Hf}$ model ages of 3.77 ± 0.18 Ga obtained from the oldest magmatic zircons from Swaziland during this study (sample AG6, AG7) are in agreement with the conclusions of Kemp et al. (2010); thereafter abundant mafic crust (with $^{176}\text{Lu}^{177}\text{Hf} = 0.02$) was directly extracted from a CHUR mantle during the Early Archaean. In fact, the 3.66 Ga Swaziland TTGs plot on the same crust evolution trend as meta-igneous, granitic and metasedimentary rocks from Greenland and the Jack Hills (see Fig. 5b).
Nevertheless, despite this coincidence, it cannot be ruled out completely that a (global-scale) depleted mantle source already existed during the Late Hadean–Eoarchaean, although there is no clear evidence for it so far. Taking this into account, the TDM model age of 4.08 $\pm$ 0.18 Ma obtained from the oldest zircons (Fig. 5b) might be considered as a maximum age for the protolith of the Swaziland TTGs. In addition, there is the possibility that the 3.66 Ga TTGs resulted from reworking of an Eoarchaean or even Hadean TTG crust, which was admixed with juvenile, mantle-derived magmas at 3.66 Ga. This final option, however, seems to be less likely in the light of the latest Hf isotope data of Kemp et al. (2009, 2010), which refute the existence of a Hadean TTG crust, as was originally suggested by Harrison et al. (2008) and Blichert-Toft & Albarède (2008). None the less, the existence of a (global) depleted mantle reservoir at 3.66 Ga is very likely, and is in fact required by the zircon Hf isotope data from Greenland, the Jack Hills and the 3.54–3.32 Ga granitoids of Swaziland (see Fig. 5b), as well as by the Hf isotope solution data for Archaean mafic rock, comprising those from the 3.45 Ga Barberton greenstone belt (tholeiitic basalts + some komatites; Blichert-Toft & Arndt 1999), the 3.75 Ga Isua basalts (Polat et al. 2003), the 2.82 Ga Kostomuksha komatites (Blichert-Toft & Puchtel 2010), and the 2.7 Ga Superior province basalts (Polat & Münker 2004). Most of these data (excluding the abnormally depleted rocks from Greenland and Barberton) can be forced to fit a straight depleted mantle array between 4.0 Ga and today using a present-day $^{177}$Lu/$^{176}$Hf of 0.04017, and an average present-day $^{177}$Hf/$^{176}$Hf of 0.283294 Ma (Vervoort & Blichert-Toft 1999) (see Fig. 5b). This line is only slightly steeper than that proposed by Pietranik et al. (2009) based on the compilation of zircon hafnium isotope data.

Formation of abundant (juvenile) crust in the Swaziland terrane during the Palaeoarchaean is well reflected by super-

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**Fig. 5.** (a) Compilation of Hadean and Archaean Lu–Hf–U–Pb isotope data from the Jack Hills and Greenland, comprising zircon solution Hf isotope data and laser ablation zircon data obtained before 2009 (Harrison et al. 2005, 2008; Blichert-Toft & Albarède 2008; grey symbols), and the latest laser ablation data from Kemp et al. (2009, 2010) (K’2009, K’2010) on zircons from metasedimentary and meta-igneous rocks, and whole-rock solution Hf isotope data from Hoffmann et al. (2010) (H’2010). (b) Synopsis of Hf isotope data obtained from granitoid rocks from Swaziland (this study) and other parts of the Kalahari craton (average values from Zeh et al. 2007, 2009, 2010a; Millonig et al. 2010 (Z’2007, 2009, 2010 and M’2010)), combined with zircon data of Kemp et al. (2009, 2010) (K’2009, 2010). In addition, the diagram shows the composition of Archaean mafic and ultramafic rocks (B&A’1999, Blichert-Toft & Arndt 1999; P’2003, Polat et al. 2003; P’2004, Polat & Münker 2004; B&P’2010, Blichert-Toft & Puchtel 2010), and of Isua boninites (H’2010, Hoffmann et al. 2010). SCT, Swaziland crustal trend with $^{176}$Lu/$^{177}$Hf = 0.0113; MC, mafic crust with $^{176}$Lu/$^{177}$Hf = 0.02; TCHUR and TDM define the respective model ages (average) obtained from the oldest Swaziland zircons. DM, depleted mantle evolution using the parameters of Blichert-Toft & Puchtel (2010); DM1, evolution of the depleted mantle between 4.0 Ga and today.
chondritic εHf of +0.5 to +2.2 obtained from nearly all TTGs that were emplaced between 3.54 and 3.32 Ga (Fig. 3). At present, the reason for this c. 200 Ma period of juvenile granitoid magmatism is unclear. One explanation could be that several juvenile island arcs were successively accreted between 3.54 and 3.32 Ga, although there is little structural control that supports this scenario. Alternatively, the same pattern could also be achieved by periodic slab melting during more or less steady subduction of hydrated oceanic crust beneath a relatively stationary proto-craton. However, because of the lack of additional geochronological data, this scenario remains speculative too. The formation of new continental crust (by extraction from a depleted mantle source) obviously continued until 3.23 Ga, as can be concluded from εHf of −0.3 to −0.5 (+1.5) obtained from the granitoids of the Pigg’s Peak inlier, as well as from those of the Manzini and Mankyne areas. The lower εHf values of these granitoids, compared with the 3.54–3.32 Ga TTGs, could be explained by the assimilation of a significant amount of older crust at 3.23 Ga. This conclusion is well supported by zircon xenocrysts found in many of the 3.23 Ga granite gneisses (Table 2, and data of Zeh et al. 2009). Schoene et al. (2009) suggested that the 3.23 Ga magmatism in Swaziland was the result of a southward-directed subduction of an oceanic crust, which vanished by c. 3.2 Ga as a result of the amalgamation of a southern terrane (comprising the Swaziland + Stolzburg + Steyndorp terranes = Barberton South of Zeh et al. 2009) with a northern terrane (comprising the northern part of the Barberton greenstone belt = Barberton North of Zeh et al. 2009).

Following the 3.23–3.22 Ga collision event, the eastern part of the Kaapvaal craton was affected by voluminous magmatism at 3.1 Ga, which caused the formation of numerous potassium-rich, sheet-like granite batholiths with slightly subchondritic εHf of −0.1 ± 1.4 to −1.7 ± 2.2 (this study; Zeh et al. 2009). Zeh et al. (2009) suggested that these batholiths formed by the reworking of older crust, with the addition of minor components from a depleted mantle source. It is likely that the magmas of the batholiths originate from lower crustal melting at 3.1 Ga, triggered by incubational heating owing to K–U–Th decay, in combination with rift- or transtension-related mantle upwelling (see Schoene et al. 2009). Crustal recycling is also in agreement with the data from the 3.07 Ga Sinceni granodiorite with εHf = −2.8 ± 2.0, and in particular with those from the 2.73 Ga potassium-rich granites (Mbahane, Sincunusa and Ngwempisi), which show highly subchondritic εHf between −6.6 and −7.2.

In general, the combined U–Pb and Lu–Hf isotope analyses of zircons reveal that the oldest TTGs of Swaziland, which were emplaced at 3.66 Ga, have subchondritic εHf of −1.6 ± 2.0. These data indicate that these TTGs formed by reworking of an even older crust, which most probably was derived from an Early Archean chondritic uniform mantle reservoir at about 3.8 Ga, immediately prior to TTG formation, although an older crustal source of Hadean age cannot be completely excluded, but would require complex mixing models.

During the following 200 Ma, new juvenile crust was added to the Ancient Gneiss Complex, as is well reflected by super-chondritic εHf of +2.2 to +0.5 obtained from TTG gneisses with ages between 3.54 and 3.32 Ga. The mechanism that formed this new crust is not well understood. It could be explained either by successive accretion of primitive island arcs or by successive subduction and periodic slab melting beneath a relatively stationary proto-craton.

The new datasets presented here additionally reveal that during subsequent magmatic processes the Palaeoarchaean crust formed between 3.54 and 3.32 Ga was intensely reworked, as is reflected by a linear array of decreasing εHf from +2.2 to −7.2 between 3.32 and 2.72 Ga. This reworking process started with the amalgamation of Swaziland’s Ancient Gneiss Complex with terranes of the Barberton Greenstone Belt at 3.23 Ga, and continued during subsequent intracratonic melting processes at 3.1, 3.06 and 2.72 Ga. The derived array with an average 176Lu/177Hf of 0.0113 can be explained either by zircon fractionation during lower TTG(?) crust melting or by minor addition of
mantle-derived melts during the granite formation events, or by a combination of both.

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Special Publication 355

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Reservoir Compartmentalization

Edited by S. J. Jolley, Q. J. Fisher, R. B. Ainsworth, P. J. Vrolijk and S. Delisle

Reservoir compartmentalization, the segregation of a petroleum accumulation into a number of individual fluid-pressure compartments, controls the volume of moveable oil or gas that might be connected to any given well drilled in a field, and consequently impacts on reserves ‘booking’ and operational profitability. This is a general feature of modern exploration and production portfolios, and has driven major developments in geoscience, engineering and related technology. Given that compartmentalization is a consequence of many factors, an integrated subsurface approach is required to better understand and predict compartmentalization behaviour, and to minimize the risk of it occurring unexpectedly. This volume reviews our current understanding and ability to model compartmentalization. It highlights the necessity for effective specialist discipline integration, and the value of learning from operational experience in: detection and monitoring of compartmentalization; stratigraphic and mixed-mode compartmentalization; and fault-dominated compartmentalization.

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Sedimentary Basin Tectonics from the Black Sea and Caucasus to the Arabian Platform

Edited by M. Sosson, N. Kaymakci, R. A. Stephenson, F. Bergerat and V. Starostenk

This wide area of the Alpine-Himalayan belt evolved through a series of tectonic events related to the opening and closure of the Tethys Ocean. In doing so it produced the largest mountain belt of the world, which extends from the Atlantic to the Pacific oceans. The basins associated with this belt contain invaluable information related to mountain building processes and are the locus of rich hydrocarbon accumulations. However, knowledge about the geological evolution of the region is limited compared to what they offer. This has been mainly due to the difficulty and inaccessibility of cross-country studies. This Special Publication is dedicated to the part of the Alpine–Himalayan belt running from Bulgaria to Armenia, and from Ukraine to the Arabian Platform. It includes twenty multidisciplinary studies covering topics in structural geology/tectonics; geophysics; geochemistry; palaeontology; petrography; sedimentology; stratigraphy; and subsidence and lithospheric modelling. This volume reports results obtained during the MEBE (Middle East Basin Evolution) Programme and related projects in the circum Black Sea and peri-Arabian regions.