Rates and mechanisms of Mesoarchean magmatic arc construction, eastern Kaapvaal craton, Swaziland

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ABSTRACT

The mechanisms and time scales of magmatism and deformation during orogenesis are important for developing models of lithospheric growth. We summarize the first detailed study of a well-preserved Mesoarchean crustal section from the oldest portion of the Kaapvaal craton, southern Africa, which records the complex interactions between deformation and magmatism during craton assembly ca. 3.3–3.2 Ga. We use chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) U-Pb zircon geochronology and apatite thermal ionization mass spectrometry in conjunction with geological mapping to show that the tonalitic to granitic Usutu magmatic suite intruded into the ca. 3.66–3.45 Ga Ancient gneiss complex over a period of ~16 Ma ca. 3236–3220 Ma as discrete pulses of magma with variable intrusive styles. Usutu rocks retain magmatic fabrics that preserve a history of NW-SE regional shortening, consistent with synchronous deformation recorded in the adjacent Barberton greenstone belt. U-Pb zircon dates of ca. 3.28–3.23 Ga and apatite cooling dates from the Nhlangano gneiss SE of the Ancient gneiss complex reveal that they represent an older, largely undocumented period of crustal growth and modification. Synchronicity in magmatism and similarity in kinematics of deformation north and south of a previously suggested continental suture in the Barberton greenstone belt lead us to propose doubly vergent subduction zones that were active from at least ca. 3.28 to 3.22 Ga. Geochemistry of the Usutu suite reveals differences in magma sources from north to south, which, combined with Nd isotopic signatures, are consistent with their production in a subduction zone. The combination of <1 Ma precision on crystallization dates with field observations and geochemistry allows us to track the evolution of this magmatic system with temporal resolution unprecedented for Archean rocks.

INTRODUCTION

Studies of orogenic belts show that the generation, segregation, and emplacement of tonalitic to granodioritic magmas have been fundamental processes in building continental lithosphere since early in Earth history. An understanding of the relationship between magmatism and deformation as a function of time and space is therefore required to generate robust tectonic models for lithosphere construction. Changes in magmatic regimes, deformational styles, or strain partitioning through the crust (e.g., Axen et al., 1998; Hollister and Crawford, 1986; Miller and Paterson, 2001; Pavlis, 1996; Klepeis et al., 2003) occur on sub-million-year time scales and, thus, the efficacy of geochronology in unraveling a sequence of tectonothermal events is dependent on attaining equal levels of precision. High-precision geochronology is difficult in Archean rocks because complicated and protracted geological histories commonly result in open-system behavior of different geochronometers, compromising the accuracy and lowering the precision of calculated dates. Recent advances in sample preparation techniques, most notably the advent of chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) for U-Pb zircon analysis (Mattinson, 2003, 2005), have greatly improved our ability to isolate closed-system domains of Archean zircons. Once closed-system behavior is established, the 207Pb/206Pb date is the most precise and accurate approach for dating Archean rocks because absolute uncertainty actually decreases with the age of the mineral for a constant analytical error (whereas U/Pb date uncertainties increase roughly linearly with age; e.g., Mattinson, 1987). This in conjunction with improved laboratory blanks to <0.5 pg Pb has allowed for the generation of dates with uncertainties less than 1 Ma from single Archean zircons (Schoene and Bowring, 2007; Schoene et al., 2008). Such high-precision calibration of magmatic, metamorphic, and deformational processes leads to a much richer understanding of the tectonic events occurring during the construction of Archean cratons.

We examine a spectacularly preserved ca. 3.3–3.2 Ga sequence of magmatic rocks in the oldest portion of the Kaapvaal craton, in the vicinity of the Barberton greenstone belt (Fig. 1). Numerous studies of the structural, petrologic, sedimentary, geochronologic, and thermal history of the Barberton greenstone belt have led to a ca. 3.23 Ga subduction-accretion model. This phase of tectonic accretion coincides with high-pressure, low-temperature metamorphism SW of the Barberton greenstone belt, widespread syntectonic magmatism on all margins of the Barberton greenstone belt, pervasive NW-SE contraction, and a transition from shallow- to deep-marine sedimentation to upward-coarsening conglomeratic sequences adjoining a lithospheric suture within the belt (see reviews in Lowe and Byerly, 2007; Moyen et al., 2007; Stevens and Moyen, 2007). However, many of the studies focusing on the plate-tectonic assembly come from within or directly adjacent to the supracrustal rocks preserved in the Barberton greenstone belt and thereby lack a larger regional context (de Ronde and de Wit, 1994; Kamo and Davis, 1994; Lowe, 1994; Kisters et al., 2003; Schoene et al., 2008). Few studies have tested the subduction-accretion model by examining the plutonic and orthogneiss complexes NW and SE of the belt, for example, to test the vergence of the inferred
Figure 1. Geologic map of the eastern Kaapvaal craton, with important features from this study labeled. Black boxes denote areas of more detailed maps, in Figures 2A and 6. Sample locations are labeled here or in Figures 2A or 6. Map was compiled from this study and the literature (de Ronde et al., 1994; de Wit, 1982; Lowe and Byerly, 1999; Schoene et al., 2008; Wilson, 1982). Age information is from this study and the literature (Armstrong et al., 1990; Hegner et al., 1984; Hegner et al., 1994; Kamo and Davis, 1994; Kröner et al., 1989; Kröner and Tegtmeyer, 1994; Kröner et al., 1991b; Maphalala and Kröner, 1993). BGB—Barberton greenstone belt; SIFS—Saddleback-Inyoka fault system; AGC—Ancient Gneiss Complex.
subduction zone (e.g., Jackson et al., 1987). Here, we investigate several groups of rocks SE of the Barberton greenstone belt called the Usutu intrusive suite and the Nhlangano orthogneiss (Hunter, 1970; Hunter et al., 1978; Wilson, 1982). Little to no description of these rocks has been published, although some published geochronological data suggest that their generation and emplacement may be related to ca. 3.23 Ga craton assembly (Kröner et al., 1989, 1991b). An understanding of the timing and length scales of magmatism, mechanisms of intrusion, and geochemical evolution is essential for generating a regional-scale tectonic model.

This study integrates mapping, whole-rock geochemistry, and U-Pb geochronology and thermochronology to understand the tectono-thermal evolution of the eastern Kaapvaal craton at ca. 3.3–3.2 Ga. We summarize an investigation of outcrops of the Usutu intrusive suite, which extend from NW Swaziland, where they are in tectonic contact with the Barberton greenstone belt, to their southern extent in central Swaziland, where they intruded syntectonically into basement orthogneisses (Fig. 1). Recognition of primary magmatic fabrics on a regional scale in these rocks, in combination with crystallization ages, allows us to reconstruct paleostress during intrusion and compare this history to rocks within and flanking the Barberton greenstone belt. Dates from the Nhlangano gneiss reveal that this orthogneissic complex represents compositionally diverse ca. 3.3–3.2 Ga lower crust. U-Pb apatite thermochronology (closure temperature ~450–550 °C; Cherniak et al., 1991; Chamberlain and Bowring, 2000; Schoene and Bowring, 2007) is used to place thermal constraints on these rocks. We integrate these data with major- and trace-element geochemistry to develop a regional model for the early history of the Kaapvaal craton and to identify important areas of future research.

**GEOLOGIC BACKGROUND**

Rocks of the eastern Kaapvaal craton record a complex history during the Mesoarchean, culminating in its assembly and stabilization ca. 3.23–3.10 Ga in and around the Barberton greenstone belt (Fig. 1). A period of concomitant deformation, metamorphism, sedimentation, and magmatism in the Barberton greenstone belt at ca. 3.23 Ga has been inferred to be the result of the collision of two or more microcontinental blocks (de Wit et al., 1992; de Ronde and de Wit, 1994; Heubeck and Lowe, 1994a; Lowe, 1999). The closure of an ocean basin ca. 3.26–3.23 is recorded within the Barberton greenstone belt by a transition from deep- to shallow-marine sediment of the Fig Tree Group, deposited ca. 3.258–3.225 Ga (Kröner et al., 1991a; Byerly et al., 1996), to subaerially deposited conglomerates of the Moodies Group. The Moodies Group is interpreted to have been deposited syntectonically into distinct contractional and extensional basins (Heubeck and Lowe, 1994a, 1994b; Lowe, 1999). Although it remains poorly dated, the Moodies Group (unconformably) overlays the upper Fig Tree Group, which has an age of ca. 3.225 Ga (Kamo and Davis, 1994; de Ronde and Kamo, 2000).

Deposition and deformation of these portions of the Barberton greenstone belt were synchronous with deformation in supracrustal slivers included in the ca. 3.45 Ga basement orthogneiss complexes south of the Barberton greenstone belt, such as the Stolzburg complex (Fig. 1). Here, high-temperature deformation is bracketed by a U-Pb date on a syntectonic dike of 3229 ± 5 Ma (Dziggel et al., 2005), and the occurrence of a late synkinematic dike, dated by U-Pb at ca. 3213 Ma (Schoene et al., 2008). Pressure-temperature (P-T) paths from a series of rocks SW of the Barberton greenstone belt record high-pressure, low-temperature metamorphism consistent with a subduction-zone setting (Diener et al., 2005; Dziggel et al., 2005; Moyen et al., 2006). The timing of metamorphism is constrained by closure of metamorphic titanite, dated in two places at ca. 3.23 Ga (Diener et al., 2005; Dziggel et al., 2005), although preserved U-Pb apatite cooling dates from nearby outcrops of ca. 3.4 Ga suggest that metamorphism and the resultant thermal structure were spatially complex, and more work is required.

Structural and geochronological data indicate that a simple orthogonal convergence model at ca. 3.23 Ga may be oversimplified. Dextral transgression in the Phophonyane shear zone, along the eastern margin of the Barberton greenstone belt in the Pigg’s Peak inlier (Fig. 1), is constrained by U-Pb dating on syntectonic dikes to have occurred before 3223.4 ± 1.9 Ma (Schoene et al., 2008). Formation of the Steynsdorp core complex in the southern Barberton greenstone belt (Fig. 1) occurred just before ca. 3230 Ma (Schoene et al., 2008). Other strike-slip and extensional fabrics throughout the belt have been documented and may be related to prolonged transform boundary tectonics throughout the Barberton greenstone belt ca. 3.2–3.1 Ga (de Ronde and de Wit, 1994; Schoene et al., 2008; Westraat et al., 2005).

Despite a prolonged focus of research on the ca. 3.23 Ga history of the Barberton greenstone belt, the location or age of the inferred microcontinents and the vergence of inferred pre-subduction assembly are not well understood. De Wit et al. (1992) and De Ronde and De Wit (1994) suggested that the Saddleback-Inyoka fault system, which strikes NE-SW through the north-central portion of the Barberton greenstone belt, formed a major continental lithospheric suture between rocks north and south of the Barberton greenstone belt (Fig. 1). Lowe (1994) described a more complicated scenario involving at least four microcontinental blocks that were sutured during thrusting along numerous belt-parallel faults. Schoene et al. (2009) show that Sm-Nd isotopic evidence is consistent with at least two distinct lithospheric blocks across the eastern Kaapvaal craton.

In combination with the aforementioned evidence for ocean basin closure ca. 3.26–3.23 Ga, extensive ca. 3.23–3.22 Ga tonalite-trondhjemite-granodiorite (TTG) magmatism north of the Barberton greenstone belt has been used to infer a north-dipping subduction zone (de Ronde and de Wit, 1994; Lowe, 1994; Moyen et al., 2006, 2007). The most prominent features north of the Barberton greenstone belt are the Kaap Valley and Nelshoogte plutons, dated using concordant weighted mean 207Pb/206Pb zircon dates at 3272.7 ± 0.2 Ma and 3236.2 ± 0.3 Ma, respectively (Schoene et al., 2006, 2008), consistent with previous dates on the same intrusions (Layer et al., 1992; Kamo and Davis, 1994; de Ronde and Kamo, 2000). Concordant zircons from a banded tonalitic orthogneiss bounding the northern margin of the Barberton greenstone belt gave a weighted mean 207Pb/206Pb date of 3258.3 ± 0.3 Ma (Schoene et al., 2008), and a series of rocks from the Badplaas orthogeiss south of the Kaap Valley pluton records ca. 3.30–3.25 Ga ages (Kisters et al., 2006), which together indicate that the creation of an arc lithosphere started well before ca. 3.23 Ga. Crustal xenoliths of ca. 3.3 Ga age were also recovered from north of the Barberton greenstone belt within the ca. 3106 Ma Nelspruit Batholith (Kamo and Davis, 1994). Extensive published geochemistry (see following references) and focused experimental petrology (Clemens et al., 2006) of these northern 3.28–3.22 Ga tonalite-trondhjemite-granodiorite (TTG) complexes were interpreted by Moyen and Stevens (2006) and Moyen et al. (2007) to represent melting of mafic sources at variable depths in a subduction-zone setting, giving rise to the distinct adakitic signatures also seen in other Archean TTGs, such as high Sr/Y ratios, high Na contents, and steeply fractionated rare earth element (REE) patterns (Martin, 1993; Martin et al., 2005; Moyen and Stevens, 2006, and references therein).

The Ancient gneiss complex, located south of the Barberton greenstone belt in Swaziland, contains the oldest rocks in the Kaapvaal craton and is assumed to be one of the bounding blocks in a ca. 3.23 Ga continental collision (Jackson et al., 1987; de Wit et al., 1992; Lowe, 1994, 1999). The
Ancient gneiss complex is traditionally subdivided into three major units: the Ngwane gneiss, the Dwalile metamorphic suite, and the Tsawela gneiss (Hunter, 1970, 1979; Hunter et al., 1978; Jackson, 1984). The Ngwane gneiss consists of a series of multiply deformed tonalites, granodiorites, amphibolites, and pegmatites that are interlayered in heterogeneous bands and folds at centimeter to decimeter scale (Jackson, 1984). U-Pb crystallization ages from these rocks and associated homogeneous granites are older than 3.5 Ga; the oldest U-Pb dates of 3.66–3.64 Ga come from the Pigg’s Peak inlier in NW Swaziland (Compston and Kröner, 1988; Kröner et al., 1989; Schoene et al., 2008). The Dwalile suite is a series of actinolite-chlorite metasediments and mafic to felsic metavolcanics that resemble the basal portions of the Barberton greenstone belt both in rock type and age, and that have detrital and volcanic U-Pb zircon dates of ca. 3.55 Ga (Jackson, 1984; Kröner and Tegtmeyer, 1994). The Tsawela gneiss is a relatively homogeneous foliated hornblende tonalite for which no U-Pb dates have been published, but Pb-Pb dates are 3.45–3.43 Ga (Kröner et al., 1989; Kröner and Tegtmeyer, 1994). Throughout northern and central Swaziland, a series of younger tonalitic to granodioritic rocks called the Usutu intrusive suite crosscuts the Ancient gneiss complex (Hunter, 1970; Hunter et al., 1978; Wilson, 1982). The Usutu suite has never been described in detail; preliminary Pb-Pb zircon evaporation dates of ca. 3.23 Ga (Kröner et al., 1991b) suggest that these rocks may be related to deformation, magmatism, and inferred terrane assembly in the Barberton greenstone belt. A strongly foliated granodiorite from the Phophonyane shear zone in the Pigg’s Peak inlier, correlative to the Usutu suite, crystallized ca. 3226 Ma (Schoene and Bowring, 2007).

In south-central and southern Swaziland, series of mafic-silicic banded orthogneisses called the Nhlangano gneiss are similar in appearance and deformational style to the Ngwane gneiss. Field relations, however, suggest that these are younger than the Ngwane gneiss (Wilson, 1982; Jackson, 1984), although only one ion microprobe U-Pb date of a banded gneiss of ca. 2745 Ma has been published (Kröner et al., 1989). Detrital zircon from granulite-grade metapelites and quartzites from the Mahamba gneisses and Mkondvo metamorphic suite, which are tectonically interlayered with the Nhlangano gneisses, give dates of >3.4 Ga (Kröner et al., 1991b; Condie et al., 1996). The timing of metamorphism of these units is not well known, although Condie et al. (1996) suggested it to be ca. 2745 Ma based on a Pb-Pb zircon evaporation date from a melt leucosome from within the Mkondvo metamorphic suite. This date corresponds roughly to widespread granitic magmatism in central and southern Swaziland.

**GEOLOGY AND GEOCHRONOLOGY OF THE USUTU SUITE**

The present study examines the Usutu intrusive suite and Nhlangano gneiss from the SE margin of the Barberton greenstone belt through south-central Swaziland (Fig. 1). In general, Usutu rocks cover a wide range in composition from hornblende tonalite to megacrustic granodiorite with associated gabbronorites and diorites. Similar to the Ancient gneiss complex, the Usutu rocks are more easily eroded and fill the low-lying valleys of central Swaziland. Therefore, outcrop is generally confined to rivers or areas of high relief formed by the resistant ca. 2.7 Ga granites or younger diabase dikes. In mapping these rocks on a regional scale, we have redefined the Usutu suite as any intrusive rocks of ca. 3.23 Ga within northern and central Swaziland. Previous mapping and geochronological efforts (Wilson, 1982; Compston and Kröner, 1988; Kröner et al., 1989; Kröner and Tegtmeyer, 1994) grouped many rocks of this age with the Ancient gneiss complex, perhaps because deformed 3.23 Ga rocks are difficult to distinguish from the Ancient gneiss complex, and they are also heterogeneous on a fine scale. Thus, Figure 1 is modified from previous maps of Swaziland to include all rocks that are dated here as ca. 3.23 Ga as Usutu suite. This definition is somewhat inconvenient for field geologists, in that the Usutu suite cannot always be distinguished in the field. However, the small-scale complexity and poor outcrop of Usutu rocks makes defining units based on mineralogical or textural criteria somewhat premature at this stage. For simplicity and descriptive purposes in this section, we subdivide the Usutu intrusive suite into three units that are spatially distinct; later in the paper, we begin to subdivide rocks based on geochemistry. This section summarizes the field observations and U-Pb geochronology and thermochronology of the different lithologic units of the Usutu suite and Nhlangano gneiss. Analytical methods, detailed rock descriptions, geochronological interpretations, and U-Pb data are given in the online supplementary material (Table DR1).1

Here, we present concordia plots of the data with our preferred interpretation, as either a weighted mean representing the crystallization age of magmatic units or a range of possible dates for samples with more complicated zircon populations. All reported uncertainties are at the 95% confidence level and do not include tracer calibration or decay constant uncertainties, since only 207Pb/206Pb dates are compared quantitatively (Schoene et al., 2006; Schmitz and Schoene, 2007).

**The Malolotja Inlier—NW Swaziland**

Two exposures of Usutu-aged rocks crop out in NW Swaziland. One is an inlier of Ancient gneiss complex located near Pigg’s Peak (Fig. 1), described in detail by Schoene et al. (2008), where they are highly deformed by the Phophonyane shear zone. This transtensional shear zone, active at ca. 3.23 Ga and reactivated at ca. 3.14 Ga (Schoene and Bowring, 2007), was responsible for juxtaposing low-grade rocks of the Barberton greenstone belt against amphibolite-grade orthogneisses of the Ancient gneiss complex and Usutu suite. A second exposure of Usutu rocks is located along the Komati River near the border of the Malolotja nature reserve, where the Usutu rocks are in contact with Malolotja Group quartzites of the Barberton greenstone belt to the west (Lamb, 1984; Lamb and Paris, 1988) and are crosscut by the Pigg’s Peak Batholith on the east (Fig. 2A). We term this sequence of outcrops the Malolotja inlier. Several hundred meters east of the contact with the quartzites, the Usutu suite is very well exposed for several kilometers along the Komati River.

The main rock type present ~1.5 km east of the quartzite contact is a magmatic biotite granodiorite (sample EKC03–33; 3227–3232 Ma; Fig. 3C). This rock contains phenocrysts of K-feldspar up to 6 cm long and 2 cm wide that are aligned within a fine- to medium-grained matrix. There is no macroscopic evidence for subsolidus rotation of megacrysts into alignment, and matrix quartz and feldspar show little or no deformation. Thus, the megacryst alignment defines a magmatic fabric, and it strikes ~020° and dips 60°–80° to the west (Fig. 2A). The concentration of megacrysts varies between 0% and 50% of the rock, and this variability is in places gradational over tens of centimeters. Megacrystic zones are locally truncated sharply by compositionally (granitic to gabbroic) or texturally (coarse- to fine-grained) distinct units. Distinct gabbroic bands or schlieren have gradational contacts with the host granodiorite and commonly contain xenocrystic K-feldspar megacrysts (Fig. 3B). Rare ovoid mafic enclaves

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1GSA Data Repository item 2009219, U-Pb ID-TIMS and whole-rock major and trace element analytical procedures, sample descriptions and locations, and detailed geochronological interpretations; Table DR1, U-Pb ID-TIMS data; Table DR2, whole-rock major and trace element data, is available at http://www.geosociety.org/pubs/ft2009.htm or by request to editing@geosociety.org.
with fine-grained margins further suggest that the amount of solid-state strain in the rock was low (Fig. 3A). Multiple generations of fine-grained granitic to aplitic dikes that obviously crosscut the fabric in the rocks are also present.

The amount of megacrystic rock decreases toward the contact with the quartzites, and this transition correlates with an increase in the amount of mafic rock present (Fig. 2A). Also, solid-state strain in the rocks increases westward, grading into S-L tectonites within ~200 m of the contact with an amphibolite unit that separates them from the quartzites. Sample EKC03–23 represents a fine-grained weakly linedated tonalite from near the amphibolite contact, which gives a 207Pb/206Pb date of 3230.3 ± 0.4 Ma (Fig. 3C). Mafic bands within the granodiorite make up >10% of the rock and are flattened, with aspect ratios of ≤50:1 (Fig. 3B). Larger bands of mafic rock (1–10 m thick) occur as well. Megacrystic rock is also present, and K-feldspar megacrysts record the solid-state strain particularly well, behaving as mantled porphyroclasts (Passchier and Trouw, 2005). Most K-feldspar porphyroclasts have symmetric mineral trails, although some have slight sigmoidal geometry interpreted to reflect normal shear sense (top to the NW), which is consistent with poorly developed but steeply dipping stretching lineations in the rock (Fig. 2A). However, close to the contact with the quartzite, boudin trains trend roughly N-S (from which BS04–3 was sampled), suggesting a horizontal stretching direction, although no lineations are present. Sample BS04–3 represents one of these sheared dikes, and it gives a 207Pb/206Pb date of 3224.4 ± 0.4 Ma (Fig. 2B), though two slightly discordant zircons with 207Pb/206Pb dates of ca. 3160 Ma suggest that a younger timing of deformation is possible, and therefore requires more work.

The Mbuluzi River Section

The Usutu suite in central Swaziland north of Manzini outcrops sparsely in small river drainages or as highly weathered exposures in topographically steep areas. However, there is a good exposure along a 10 km stretch of the
Mbuluzi River that can be accessed in two spots by a series of dirt roads that cross the river north and west of Luve (Fig. 1). Access by the eastern crossing reveals two main lithologies of Usutu rocks, including megacrystic granodiorite (EKC02–23; 3229.8 ± 0.5 Ma; Fig. 4B) and homogeneous medium-grained biotite tonalite (EKC02–24; 3227.5 ± 0.3 Ma; Fig. 4B), which also contains more mafic bands within the megacrystic unit; crosscutting pegmatitic dikes are also present. The megacrystic unit records the interaction of mafic and silicic magmas during intrusion, as indicated by gradational contacts with mafic material. K-feldspar megacrysts are ≤3–4 cm long and ≤1–3 cm wide and are aligned such that they define a strong magmatic fabric (Fig. 4A) that varies locally in strike and dip by ±10°–15°. Other fabric-forming elements are present, such as compositional banding defined by abundance of biotite, hornblende, and feldspar, and by aligned mafic schlieren. Megacrysts also occur as xenocrysts in the more mafic units, and both single megacrysts and magmatic fabrics crosscut diffuse mafic-silicic contacts. Contacts between the tonalite and the megacrystic granodiorite in places are gradational, such that it is difficult to distinguish their relative age, although, in other places, the tonalite clearly crosses the megacrystic unit. There is local evidence of slight flattening of quartz and fine-grained feldspar that is inferred to indicate solid-state strain. The amount of strain is low, however, as megacrysts remain undeformed and show no mantles, which are commonly observed in higher-strain rocks of similar composition.

Usutu rocks at the western river crossing preserve intrusive relationships within the wall rocks, as well as mixing and mingling relationships between different magmas. Wall rocks are heterogeneously deformed and brecciated and are both mafic and silicic in composition. Their protolith resembles a series of bimodal volcanioclastic and volcanic rocks, but some may also be siliciclastic in origin. In places, host rocks are xenoliths, although in others, they may maintain continuity with wall rocks, perhaps as roof pendants. In sections with little intrusive material, the orientation of foliation in the metamorphic rocks strikes generally NW, at high angles to the magmatic fabrics in other Usutu rocks. Several magmatic phases crop out here, including a medium-grained leucotonalite (EKC02–32; 3231.3 ± 0.5 Ma; Fig. 5A), megacrystic granodiorite, and fine-grained hornblende gabbro. These rocks are cut by a series of undeformed granitic dikes and plugs (EKC03–18; 3186 ± 2.1 Ma; Fig. 5A), which are similar to rocks mapped by Wilson (1982) as Pigg’s Peak granite several kilometers north of the Mbuluzi River, although the Pigg’s Peak Batholith, dated by Schoene and Bowring (2007) at ca. 3140 Ma, is considerably younger. As with other Usutu rocks, the different phases preserve distinct crosscutting relationships and also more diffuse contacts where magma mixing is preserved (Fig. 5C). These outcrops have a much higher proportion of gabbroic rocks than elsewhere. In situ melts of fine-grained mafic material are also observed, and in places tonalitic leucosomes appear to form small dikes or veinlets (Fig. 5B). It is unclear if this in situ melting occurred synchronous with intrusion of the plutonic rocks.
Figure 4. The Usutu suite at the Mbuluzi east river crossing. (A) Photo of undeformed megacrystic granodiorite sample EKC02–23 with magmatic fabric defined by alignment of K-feldspar megacrysts. Late aplitic dike is in upper portion of photo. Pencil is for scale. (B) Concordia plot for megacrystic granodiorite EKC02–23 and tonalite EKC02–24 from the Mbuluzi east outcrop. Uncertainties are at the 95% confidence level. MSWD—mean square of weighted deviates.

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Figure 5. NE–SW fabrics (Fig. 5D) that are interpreted to be primarily magmatic in origin, based on the observation that bands defined by mineralogical and grain-size variability occur with no observable deformation in the grains. In some zones, however, there is mineral flattening and the development of several mete-scale open folds of compositional layers. Boudinage is also observed in mafic open folds within the tonalitic rocks. Further downstream (within several kilometers) from the road’s end, deformed metavolcanic/sedimentary rocks dominate the outcrop.

Central Swaziland

Exposure in the southern areas of the Usutu suite (here arbitrarily defined as outcrops south of Manzini) is also mainly confined to river outcrops, although the confluence of a number of major drainage systems makes for more continuous sections than in areas further north (Fig. 1). Usutu rocks in the area can be generally divided into three groups based on the style of intrusion, the ratio of intrusive rocks to host basement gneisses, and the amount of postintrusion deformation (Fig. 6).

The best exposures are south of Matsapha along the Usushwana River, where rocks contain a steeply dipping magmatic fabric defined by variations in grain size, modal distribution of major constituent minerals such as hornblende, biotite, and plagioclase, and other igneous textures such as alignment of elongate minerals (EKC03–21; 3232.1 ± 0.5 Ma; Figs. 7C and 8C). Magmatic compositional layering ranges in thickness from several centimeters to tens of meters. Mafic enclaves and xenoliths are concentrated within discrete layers several meters thick that strike parallel to magmatic foliations (Figs. 7A and 7B).

Further south on the Usushwana River, the proportion of xenoliths increases until older wall-rock pendants that were intruded by younger magma dominate the outcrops (near samples BS04–6, –7, and –8). These bimodal basement orthogneisses (sample BS04–7; ≥3300 Ma; Fig. 8B) are multiply deformed, as evidenced by refolded folds in mafic-silicic bands, and have a strong resemblance to the Ngwane gneiss from northern Swaziland, but they may be as young as 3300 Ma. Fold patterns in the orthogneiss are open to isoclinal with fold hinges that roughly parallel weak stretching lineations and boudinage trains. In places, hornblende leucotonalite (BS04–8; 3232.1 ± 0.4 Ma; Fig. 8A) is observed intruding boudin necks in the orthogneiss, suggesting it was intruded synchronously with deformation (Fig. 8C). Locally, alignment of hornblende warps around contacts with the older gneiss, although the overall magmatic fabric defined by aligned hornblende is subparallel to the foliation in the gneiss. Magmatic fabrics are also warped around xenoliths in places where continuous wall rock is not abundant. An undeformed medium-grained granodiorite cuts all fabrics in these outcrops (BS04–6; 3219.9 ± 1.0 Ma; Fig. 8A). This rock is similar to other massive granodiorite bodies in the map area, which were originally the only rocks within this southern area mapped as the Usutu granodiorite (Wilson, 1982).

Within several kilometers upstream of the confluence of the Usushwana and the Usutu Rivers, series of granodiorites and tonalites (hornblende) are strongly deformed whether they are associated with the bimodal orthogneisses or not. In most outcrops, the deformation is characterized by a steeply dipping foliation defined by flattened quartz and feldspar with subhorizontal lineations and open folds with moderately dipping axes. Foliation in bimodal gneisses near the confluence of the Ngwempisi and Usutu Rivers is subparallel to the foliation in the Usutu rocks, and the strike is subparallel to the magmatic fabrics exhibited further north. Bimodal gneiss in these outcrops (sample EKC02–64, discussed with the Nhlangano gneiss samples) is difficult if not impossible to distinguish from Usutu-aged rocks, if present. One outcrop examined along the Usutu River contained a granodioritic augen gneiss (BS04–12; 3212.8 ± 0.4 Ma; Figs. 9A and 9B) with strong foliation defined by flattened quartz and feldspar and alignment of biotite that dips steeply to the south. The foliation is cut by a late aplite dike that has a poorly developed foliation parallel to the host rock (BS04–11; 3221.6 ± 0.4 Ma; Fig. 9B). The southern extent of the Usuta suite is marked by the northern boundary of the undeformed ca. 2.7 Ga Hlatikulu granite (sample EKC02–65). Those rocks and their significance will be discussed later herein.

Although outcrop is sparse in the northern portion of the map area in Figure 6, several
samples were collected. This area is characterized by tonalite to granodiorite with variable amounts of hornblende. In the western side of the map area, the Usutu River exposes coarse-grained hornblende-rich tonalite (EKC02–35; 3236.1 ± 0.5 Ma; Fig. 8B). In places, a fabric defined by alignment of sub- to euhedral hornblende is well developed, striking NE-SW and dipping SE. A similar rock type occurs throughout central Swaziland, but it has a strong foliation; geochronology in the following section reveals that in at least one locality (EKC02–36; Fig. 8B), the foliated rock contains only zircons >3.4 Ga and can be grouped with the ca. 3.45 Ga Tsawela gneiss (Jackson, 1984; Kröner and Tegtmeyer, 1994).

The Nhlangano Gneiss and Younger Intrusions

The geology of south-central Swaziland can be characterized as a series of ca. 2.7 Ga granites that crosscut high-grade orthogneisses and paragneisses that are grouped into three main units: the Nhlangano gneiss, the Mahamba gneiss, and the Mkhondvo metamorphic suite (Wilson, 1982). These units are very difficult to distinguish in places, and the interrelationship between the paragneisses and orthogneisses is not well characterized.

Our field observation of the Nhlangano gneiss concentrated on one very well-exposed locality along the Mkhondvo River near an inlier of the Mkhondvo metamorphic suite (here called the Mkhondvo suite inlier; Fig. 1), which has been reported to contain granulite-facies metamorphic assemblages (Condie et al., 1996). This area contains rocks of the Mkhondvo suite that are infolded as

Figure 5. The Usutu suite at the Mbuluzi west river crossing. (A) Concordia diagram for leucotonalite sample EKC02–32 and a crosscutting granite EKC03–18. Uncertainties are at the 95% confidence level. MSWD—mean square of weighted deviates. (B) An example of a mafic host rock intruded by the Usutu suite, with in situ melt generation. Goat feces are ~5 cm in diameter. (C) Assimilation of mafic enclave into leucotonalite. (D) Stereonet showing structural information from the Mbuluzi east and west river crossings.
synformal remnants within Nhlangano orthogneiss. In general, the orthogneisses are similar to those intruded by the Usutu suite along the Usushwana and Usutu Rivers (e.g., BS04–7, presented earlier here; EKC02–64; 3263.6 ± 0.5 Ma; Fig. 10B), although in places they contain highly attenuated pink granodioritic augen orthogneiss (EKC03–36; >3240 Ma; Fig. 10B) in addition to highly attenuated mafic-silicic banded gneisses (EKC02–66 and EKC03–35; >3240 Ma; Fig. 10B) characterizing exposures described in the previous section. A prominent difference from exposures to the north, however, is that the NE-SW foliation trend in the orthogneiss observed north of the Hlatikulu granite changes to N-S foliations observed near the Mkhondvo suite inlier. More detailed observations of this transition zone are warranted, given this major change in structural trend of the region.

WHOLE-ROCK GEOCHEMISTRY RESULTS

The highly strained gneiss units in the Mkhondvo valley are cut abruptly by several plutonic rocks of unknown age. Some of these had been mapped originally as the Usutu suite (Wilson, 1982), presumably because of their granodioritic to tonalitic composition, although here we show them to be much younger (BS04–20; ca. 2734 Ma; Fig. 11). Another crosscutting unit, the medium- to fine-grained Hlatikulu granite (EKC02–36; 2728.9 ± 0.5 Ma; BS04–21; 2729.8 ± 0.4 Ma; Fig. 11), is present throughout the Mkhondvo valley and within the surrounding hills.

Sr/Y and REE patterns in particular have received much attention when discussing Archean TTG suites such as those present in the study area (e.g., Condie, 2008; Martin, 2008) (EKC02–31) and Schoene et al. (2008) (KPV99–94, KPV99–90). Analytical details are available in the GSA Data Repository material, and the data are given in Table DR2 in the GSA Data Repository (see footnote 1). For comparison with our data, we separate data from the literature into the Ancient gneiss complex, ca. 3.55–3.45 Ga TTGs, ca. 3.3–3.2 Ga TTGs from north of the Barberton greenstone belt, and ca. 3.1 Ga granite-monzonite-syenite suite (GMS; Hunter et al., 1978, 1984; Anhaeusser and Robb, 1983; Robb and Anhaeusser, 1983; Kröner et al., 1993; Condie et al., 1996; Yearron, 2003; Westraat et al., 2005; Moyen et al., 2007; Belcher et al., 2008, personal commun.).
Rates and mechanisms of Archean arc magmatism and deformation

The Usutu suite, similar to the TTGs and the GMS, shows increasing Sr/Y with decreasing Y (Fig. 12) and increasing SiO₂. The Ancient gneiss complex lacks Sr/Y ratios >50. High Sr/Y ratios (>50) in the Usutu suite are restricted to rocks south of Manzini (Figs. 1 and 6). The northern Nhlangano gneiss samples (Fig. 6), collected near the high-Sr/Y Usutu samples, also have high Sr/Y, whereas those from near the Mkhondvo inlier have Sr/Y <30.

The REE patterns determined in this study are highly fractionated and show an increase in La/Yb with decreasing Yb, similar to other rocks in the region except the GMS (Fig. 12). The northern Nhlangano gneiss samples (Fig. 6), collected near the high-Sr/Y Usutu samples, also have high Sr/Y, whereas those from near the Mkhondvo inlier have Sr/Y <30.

Figure 7. Igneous textures preserved in the Usutu suite along the Usushwana River, near sample locality EKC03–21 (Fig. 6). (A) Composite dike recording the interaction of mafic magmatic enclaves within tonalitic magma; no solid-state shearing is evident. (B) Unstrained hornblende gabbro enclaves within hornblende tonalite magma. (C) Magmatic layering in tonalite, defined by the modal proportion of hornblende versus plagioclase and variable grain size. This is a representative example of sample EKC03–21.

The northern and southern Usutu suite are lower than the 3.3–3.2 Ga TTGs and at the low end of 3.5–3.4 Ga TTGs (Fig. 12), but they overlap well with the Ancient gneiss complex and GMS. The Nhlangano gneiss samples are similar in both K₂O/Na₂O and Mg# to the TTGs.

In addition to these trends, the Usutu suite shows several distinct geochemical trends as a function of time between ca. 3236 and 3220 Ma, which also differ from south to north (Fig. 13). The southern Usutu suite shows a marked increase in incompatible elements such as Ba, Rb, Eu/Eu*, Ba/La, Ba/Nb, and Ba/Ta. This trend is also observed in northern rocks, but it is much less pronounced. The REE, such as Nd, decrease with increasing εNd(t) and do not share a common end member for their arrays (Fig. 14). Data from the Nhlangano gneiss show similar trends to the northern Usutu suite in some cases but are more scattered in others, which are interpreted below in terms of magma generation and evolution.

**DISCUSSION**

**Rates and Mechanisms of Magmatism in the Usutu Suite**

Our field work and geochronology are consistent with the Usutu suite being part of a largely undocumented period of ca. 3.23 magmatism.
south of the Barberton greenstone belt, and these results have significant implications for the evolution of the eastern Kaapvaal craton. The Usutu suite was emplaced over a period of ~16 Ma between 3236 and 3220 Ma, it preserves regional NE-SW–striking magmatic fabrics of variable intensity, and these records of strain can be used to infer paleostress conditions during intrusion (Fig. 15). Generally speaking, a magmatic fabric is defined as the preserved alignment of linear or planar minerals in the presence of melt (Hutton, 1988; Paterson et al., 1998). This definition purposely avoids any implications as to the way in which the fabric formed. Important factors to consider in the generation of these fabrics are the contributions of internal flow of magma within the confines of a rigid host rock versus the imposition of regional stress patterns on a potentially complicated crystal-rich/crystal-poor mush of variable composition and viscosity. One end-member model suggests that fabrics are generated by the competing interaction between density-driven or mechanically driven flow versus advective heat transfer, followed by a quenching of the magma to preserve snapshots of these dynamic processes (Barriere, 1981; Abbott, 1989; Wiebe, 1994, 1996). The other end-member model suggests that strain within the magma is imposed by regional stress fields, such that the orientation of elongate minerals, for example, may mimic the finite strain ellipsoid in the surrounding host rocks (Hutton, 1988; Neves et al., 1996; Brown and Rushmer, 1997; Brown and Solar, 1998b; Mahan et al., 2003). In any given intrusion, the dominant one of these two end members will dictate the final strain recorded; in a sequence of intrusions, the dominant strain-producing force, whether internal or external, may change over time.

Figure 8. Geochronology of the Usutu suite and orthogneisses in central Swaziland. See Figure 6 for sample locations. (A) Concordia plot summarizing zircon and apatite data from EKC03–21, BS04–6, and BS04–8. White ellipses with gray outline are BS04–6 apatite. Light-gray ellipse is the EKC03–21 apatite analysis. MSWD—mean square of weighted deviates. (B) Concordia plots for two orthogneisses, EKC02–36 and BS04–7, and one Usutu rock, EKC02–35. Uncertainties are at the 95% confidence level. (C) Photograph illustrating the intrusive relationship between orthogneiss BS04–7 and hornblende tonalite BS04–8. The younger tonalite is interpreted to have intruded syntectonically during boudinage of older banded gneiss. Black line is drawn to denote contact between the two units.
Evaluation of magmatic fabrics is usually approached by examining the relationship between fabrics or foliations in host rocks at different scales of observation relative to those within the intrusive rock (Paterson et al., 1998). Magmatic fabrics within the Usutu suite are consistent on a regional and outcrop scale, and when combined with geochronology, these data provide unique insight to the geologic history of the region and the processes of magma emplacement.

Usutu rocks in the Malolotja inlier preserve a magmatic fabric that is overprinted with increasing subsolidus strain toward the faulted contact with quartzites to the west (Figs. 2 and 3). The pristine magmatic fabric preserved furthest from the fault is defined by the alignment of elongate minerals such as megacrystic K-feldspar and biotite, mafic enclaves that are flattened in places, and compositional contacts within the rock. Evidence for subsolidus deformation, such as mineral flattening and mantled feldspar megacrysts, increases in intensity and abundance westward toward the fault, and the subsequent metamorphic foliation parallels the magmatic fabric, the fault plane, and the foliation of quartzites in the hanging wall. These observations suggest that fabric generation is closely linked to strain in the shear zone (i.e., a “coupled system” in the terminology of Paterson et al., 1998) and, therefore, was a result of either flow of magma emplaced as sheets parallel to the shear zone and/or through subhorizontal shortening perpendicular to the shear zone during intrusion. Because magmatic fabrics defined by aligned phenocrysts form when only ~20%–40% residual liquid remains (Paterson et al., 1998, and references therein), it is likely that contraction parallel to the fault played a role in the magmatic alignment of feldspars in these outcrops. The megacrystic sample EKC03–33 contains zircons with dates between ca. 3232 and 3227 Ma, and we suggest that feldspar alignment “locked in” near the end of that interval. The deformed tonalitic sample closer to the fault, EKC03–23, gives a coherent cluster of zircon dates of 3230.3 ± 0.4 Ma (Fig. 3C), showing that subsolidus deformation happened after that. Smaller-scale compositional banding features such as schlieren and mafic enclaves with and without visible subsolidus deformation argue for the existence of co-mingling silicic and mafic magmas, the latter of which were in some cases deformed during or shortly after intrusion. A constraint on deformation in the shear zone is given by E-W–directed shortening and boudinage (perpendicular to shortening) of the aplitic dike BS04–3 after ca. 3224.4 Ma (Figs. 2B, 2C, and 15).

The orientation of magmatic fabrics in the east Mbuluzi River outcrop is also NE-SW, although there is no exposed fault in this area to suggest a possible control on fabric formation. The two end-member units, the megacrystic granodiorite (EKC02–23) and the homogeneous tonalite (EKC02–24), contain zircon populations that crystallized 2 Ma apart, at 3229.8 ± 0.5 and 3227.5 ± 0.3 Ma, respectively (Figs. 4B and 15). Contacts between the units are gradational in spots and sharp in others, although xenocrystic feldspars in the tonalite show it to be younger, consistent with the geochronology. Magmatic fabrics cut compositional contacts when oblique to them, despite the apparent 2 Ma age difference between the two rock types. Recent geochronologic studies show variable durations of mineral growth (as inferred from zircon dates) from silicic plutonic and volcanic systems between <10 ka up to as much as millions of years, although this is likely highly dependent on the scale of observation (Glaszner et al., 2004; Matzel et al., 2006, 2009; Reid et al., 1997; Vazquez and Reid, 2002).

The west Mbuluzi outcrop contains several magmatic compositions that intruded metamorphic rocks of variable composition. These magmas also have NE-SW–striking magmatic fabrics, which strike at high angles to the variably oriented foliation in the host rocks and which are discordant to contacts (Fig. 5D). Such features are typical of magmatic fabrics generated by regional stresses where strain is partitioned into the magmatic phase as opposed to the previously deformed wall rock (i.e., an “uncoupled system” in the terminology of Paterson et al., 1998). While the intrusive and postintrusion deformational relationships here are much more complicated than the east Mbuluzi River outcrop, the predominant direction of magmatic fabrics is consistent with other outcrops observed in this study.

The section exposed along the Usushwana River in central Swaziland, south of Manzini...
(Fig. 6), is characterized by a southward increase in the proportion of basement gneiss as xenoliths toward a dominance of roof or wall pendants. In the NW portion of this section, steeply dipping magmatic fabrics are evident as compositional and textural layering, defined by features such as grain size, and the modal percentage and degree of alignment of hornblende (Fig. 7C). Additionally, horizons of mafic enclaves and xenoliths strike parallel to other magmatic fabrics, suggesting that these fabrics were generated when single batches of magma incorporated these inclusions and were then emplaced into their present position as steeply dipping dikes or sheets (e.g., Brown and Solar, 1998a; Snyder et al., 1997; Figs. 7A and 7B). Elongate horizons rich in mafic enclaves could also have intruded as composite mafic-silicic dikes, where dikes intrude as two co-mingling magmas (Snyder et al., 1997). An alternative model comes from similar observations in other magmatic systems that are interpreted to reflect accumulation of material on the floor of a magma chamber through gravity-driven transport of phenocrysts, mafic enclaves, or xenoliths (Wiebe, 1994, 1996; Cruden, 1998; Wiebe and Collins, 1998; Wiebe et al., 2002). Tilting of layers then comes from subsidence or downwarping within the chamber during magma accumulation through subsequent magma injections and crystallization. These systems are recognizable by “way-up” indicators such as load-casts or piping at mafic-silicic magma interfaces, unidirectional warping of enclaves or phenocrysts, or xenolith accumulation. Way-up indicators were not observed in the Usutu rocks, and the consistency of the strike of magmatic fabrics throughout Usutu rocks suggests that the steeply dipping fabrics did not result from

Figure 10. The Nhlangano gneiss. (A) U-Pb apatite thermochronology from the Nhlangano gneiss EKC03–35, –36, one Usutu rock EKC02–35, and an older orthogneiss EKC02–36 from central Swaziland. See Figures 1 and 6 for sample locations. Dates are interpreted to reflect partial resetting by adjacent ca. 2.73 granites and/or metamorphism. (B) U-Pb geochronology of the Nhlangano gneiss from magmatic zircons, indicating crystallization ages between ca. 3240 and 3280, based on extrapolation of discordant data back to concordia. Uncertainties are at the 95% confidence level. MSWD—mean square of weighted deviates. (C) Field photograph of mafic-silicic banded Nhlangano orthogneiss, indicating sample lithology of EKC03–35; actual sample location is several meters away. (D) Field photograph of Nhlangano augen orthogneiss, from which sample EKC03–36 was collected. Note ductile shear zones in both C and D, indicated by white arrows.
subidence of, for example, an ellipsoidal magma chamber. The alternating and conjugate dip angles in the fabrics (Fig. 6) and thus a lack of progressive steepening of the fabrics also argue against a floor-sinking model for the Usutu rocks.

There is a transition southward along the Usushwana River such that outcrop is dominated by older orthogneisses, which preserve intrusive contacts with the Usutu rocks. Consistent dates of ca. 3232.1 Ma from Usutu rocks through this transition (EKC03–21 and BS04–8; Figs. 8 and 15) suggest the sequential intrusion of a relatively homogeneous batch of magma in a short period of time. The interaction between intrusive rocks and host rock, combined with deformation patterns in the host rock show that the Usutu rocks intruded during subhorizontal NW-SE shortening and NE-SW stretching. This is evidenced by foliations in orthogneisses parallel to migmatic fabrics in intruding Usutu rocks, and the observation that Usutu rocks intrude boudinaged orthogneiss (Fig. 8C). The duration of NW-SE contractional deformation is bracketed in part by the date of the foliated megacrystic granodiorite of 3221.8 ± 0.4 Ma and the crosscutting dike of identical age (samples BS04–11 and BS04–12; Fig. 9B). The massive granodiorite intruded at 3219.9 ± 1.0 Ma (BS04–6; Fig. 8A), which cuts all migmatic and metamorphic fabrics and contains neither, suggesting that locally strain accumulation ceased in this area.

To summarize, migmatic fabrics observed in the Usutu rocks through north and central Swaziland are consistent with their intrusion into a NW-SE compressive regime between ca. 3236 and 3220 Ma. The only detailed studies of the structural history of the Ancient gneiss complex (Jackson, 1984; Jackson et al., 1987) also document a major period of NW-SE shortening coincident with regional uplift, although the timing of this event in the Ancient gneiss complex has been unknown. We have documented this event in >3.45 Ga host gneisses in central Swaziland at ca. 3.23 Ga (Fig. 6), and this may apply to the Ancient gneiss complex further north as well.

Finally, it is worth noting that the duration of high-volume magmatism is similar to periods determined in other continental arc settings, such as the North Cascades arc and the Tuolumne intrusive suite, United States, where large volumes of magma intruded over periods of ~5–15 Ma (Coleman et al., 2004; Glazner et al., 2004; Matzel et al., 2006). The full duration of subduction magmatism in those areas was, however, on much longer time scales (50–100+ Ma). If we include the intrusion of the Nhlangano gneiss (discussed later herein) with the Usutu data, the full range of magmatism may be as long as ~80 Ma, marking a much longer magmatic event in the region.

Formation and High-Grade Metamorphism of the Nhlangano Gneiss

The Nhlangano orthogneiss is similar in its lithologic variation to the Usutu suite, but it is ≤40 Ma older and has undergone heterogeneous noncoaxial subsolidus deformation typical of lower crustal gneiss terranes. The existence of this lower-crustal gneiss terrain has not been accounted for in current tectonic models. We interpret the weighted mean 207Pb/206Pb date of 3265.6 ± 0.5 Ma of concordant zircons from EKCO2–64 as the crystallization age of one portion of the Nhlangano gneiss (Fig. 10B). Samples EKCO2–66 and EKCO3–36 gave only discordant zircon dates. Of those analyses that were <3% discordant, all but four give 207Pb/206Pb dates that fall between 3239 and 3265 Ma. We have observed that magmatic zircons from other samples in this and related studies (Schoene and Bowring, 2007; Schoene et al., 2008) that are <3% discordant are not more than 15 Ma younger than the true crystallization age of that zircon, and this conjecture is consistent with upper-intercept dates of zircons from each sample (see supplementary material [see footnote 1]). Therefore, we conclude that the various samples of Nhlangano gneiss in this study likely crystallized between ca. 3240 and 3280 Ma. Two other grains from EKCO2–64 and EKCO3–35, which were significantly rounded compared to the others, gave 207Pb/206Pb dates of ca. 3329 and 3425 Ma, and these dates indicate that these rocks intruded through older crust. Based on Nd isotopic systematics, Schoene et al. (2009) showed that the Nhlangano gneiss formed through mixing of older crustal material, similar to that of the Ancient gneiss complex, with mantle-derived material. This is consistent with the >3.3 Ga xenocrysts that occur in this study as well as ca. 3.47 Ga xenocrysts reported from Nhlangano orthogneisses by Condie et al. (1996). Condie et al. (1996) reported Pb-Pb dates of ca. 2.73 Ga from melt leucosomes in the metasedimentary Mkhondvo metamorphic suite (Fig. 1), which they interpreted to date granulite-grade metamorphism concurrent with the intrusion of the Hlatikulu granite (samples EKCO2–65, BS04–20, and BS04–21). Our U-Pb apatite thermochronology from Nhlangano gneiss samples (EKC03–35...
Schoene and Bowring

and EKC03–36) several hundred meters from the sample locations of Condie et al. (1996) have 207Pb/206Pb dates between ca. 2730 and 3100 Ma (Fig. 10A). If granulite-grade regional metamorphism in the Mkhondvo metamorphic suite (Condie et al., 1996) had occurred ca. 2.73 Ga, it is likely that these apatites would have been fully reset, and their dates would reflect post–2.73 Ga cooling. Given that older dates come from the sample that is further away from the undeformed Hlatikulu granite (sample BS04–21), located ~400 and 800 m to the west of samples EKC03–35 and EKC03–36, respectively, it is possible that the data reflect partial resetting at ca. 2.73 Ga. Additionally, the syntectonic intrusion and deformation of samples BS04–11 and BS04–12 at ca. 3222 Ma (Figs. 9A and 9B) potentially bracket ductile deformation of the northern Nhlangano gneiss to have occurred before that, since they contain the same foliation trend as the nearby outcrop of ca. 3265 Ma Nhlangano gneiss (EKC02–64; Fig. 6). Thus, we tentatively conclude that high-grade deformation and metamorphism of the Nhlangano gneiss occurred synchronously with their igneous intrusion ca. 3.28–3.24 Ga. More detailed geochronology in combination with metamorphic petrology in the Mkhondvo metamorphic suite and structural mapping in this area will help deduce the timing of these events as well as the possibility of multiple high-grade deformational and/or metamorphic events ca. 3.3–3.2 and 2.73 Ga.

Geochemical Evidence for the Sources and Evolution of the Usutu Suite

Figure 12. Plots of Sr/Y versus Y, La/Yb versus Yb, SiO₂ versus Mg#, and SiO₂ versus K₂O/Na₂O for rocks from this study and also from the literature. AGC—Ancient gneiss complex; 3.5–3.4 Ga TTGs—tonalite, trondhjemite, granodiorites (see text); NBGB—ca. 3.3–3.2 Ga TTGs from north of the Barberton greenstone belt; 3.1 Ga GMS—granodiorite, monzonite, syenite (see text). Symbols are the same for both plots. N. Usutu are samples north of Manzini; S. Usutu are from south of Manzini (Fig. 1). Literature data are from: Anhaeusser and Robb (1983), Belcher et al. (2008, personal commun.), Condie et al. (1996), Hunter et al. (1978, 1984), Kröner et al. (1993), Moyen et al. (2007), Robb and Anhaeusser (1983), Westraat et al. (2005), and Yearron (2003).
Reviews, see Martin et al., 2005; Moyen, 2009). The term TTG has evolved to generally refer to Na-rich rocks with adakite-like signatures—namely high Sr/Y ratios and highly fractionated REE patterns, although numerous subseries have been identified (e.g., Champion and Smithies, 2007; Moyen et al., 2007; Moyen, 2009). There is still considerable uncertainty about the generation of rocks with adakite-like signatures, both in modern and Archean settings, although there is increasing experimental evidence that such signatures may be formed by partial melting of a metabasaltic source in the stability field of garnet, largely outside the stability field of plagioclase—above ca. 1.2 GPa—depending on temperature and composition (Barth et al., 2002; Foley et al., 2002; Clemens et al., 2006; Moyen and Stevens, 2006). Possible locations for such a process include a thick and basaltic lower crust or oceanic crust in a subducted slab (Martin, 1993; Smithies, 2000; Condie, 2005, 2008; Martin et al., 2005). However, adakite-like signatures can also be generated by nonbasaltic source compositions, melt interaction with a hydrated mantle, or melting rocks with moderate to high Sr/Y and La/Yb ratios (Martin et al., 2005; Moyen, 2009). Thus, Sr/Y and La/Yb are not unique indicators of any one melting regime.

Geochemistry of the 3.3–3.2 Ga TTGs from NW of the Barberton greenstone belt (Kaap Valley and Nelshoogte plutons and the Badplaas domain; Figs. 1 and 12), when combined with the abundant geologic evidence discussed here, suggests that these rocks were generated either in response to lower-crustal melting and/or slab melting during subduction-accretion (Moyen et al., 2007) or through dehydration of the mantle wedge (Kleinhanhs et al., 2003). Data from this study show that the ca. 3.23 Ga northern Usutu and southern Usutu suites (separated here by rocks north and south of Manzini) have different geochemical signatures; most notably, Sr/Y ratios >50 are found exclusively in the south, where rocks also have lower concentrations of REEs (Fig. 12). The southern Usutu suite also has strong positive Eu anomalies and ranges toward very high incompatible element contents (Ba, Rb, K2O, and K2O/Na2O; Figs. 13 and 14). Based on models proposed for the TTGs north of the Barberton greenstone belt, the differences in Sr/Y and La/Yb could represent contrasting depths of melt generation—deep (>1.2 GPa) in the south Usutu suite, and shallow (<1.2 GPa) in the northern rocks (Clemens et al., 2006; Moyen and Stevens, 2006; Moyen et al., 2007). It is worth noting, however, such high concentrations of incompatible elements and high K2O/Na2O distinguish the southern Usutu suite from other TTGs in the region, so assuming a similar petrogenesis based on Sr/Y may be misguided.

Figure 13. Selected geochemical data plotted as a function of isotope dilution–thermal ionization mass spectrometry (ID-TIMS) U-Pb crystallization age (from this study). Arrows are interpreted trends for the southern Usutu suite. The εNd(t) data are from Schoene et al. (2009). Symbols are the same as Figure 12.
Figure 14. Selected geochemical data (this study) plotted as a function of $\varepsilon_{\text{Nd}}(t)$ (Schoene et al., 2009). Conservative estimates for uncertainties on $\varepsilon_{\text{Nd}}(t)$ are ±0.5 (95% confidence; Schoene et al., 2009). Arrows indicate interpreted mixing lines in northern and southern Usutu suites, pointing to at least two different end-member source melt compositions (sources 1 and 2, S1 and S2), which mix with the Ancient gneiss complex (AGC). Symbols are the same as Figure 12.
A combination of the trace-element data with Nd isotopes (Schoene et al., 2009) helps to further constrain the magma sources and also the difference between northern and southern Usutu rocks (Fig. 14). Because Nd isotopic signatures are unaffected by source pressure and temperature, melt fraction, or fractional crystallization, geochemical trends on such diagrams likely represent mixing of different sources. Rare earth concentrations plotted versus \( \varepsilon_{Nd}(t) \) (the \( \varepsilon_{Nd} \) at the time of rock crystallization) reveal two distinct parallel trends with differing concentrations, corresponding to north and south (Fig. 14). Sr/Y increases with increasing \( \varepsilon_{Nd}(t) \) in southern rocks, whereas Sr/Y in northern rocks increases only slightly with increasing \( \varepsilon_{Nd}(t) \); these patterns are matched by Eu/Eu* anomalies. Opposite to the pattern with REE, other highly incompatible elements, such as Ba, increase with increasing \( \varepsilon_{Nd}(t) \). Ba/Ta and Ba/La also increase dramatically with \( \varepsilon_{Nd}(t) \) along different trajectories for the northern and southern Usutu suites (Fig. 14).

The trends outlined here can be explained if the northern and southern Usutu suites were derived from different “parental magmas” that interacted to variable degrees with pre-existing crust. One end-member magma in the south (S2 in Fig. 14) had \( \varepsilon_{Nd}(t) \geq 0 \), high Sr/Y, Ba/Ta, Ba/La, and K2O/Na2O and high concentrations of Ba corresponding to low REE content. Large positive Eu anomalies in the S2 end member are interpreted as the signature of breakdown of plagioclase in the source. The high \( \varepsilon_{Nd}(t) \) end member in the north (S1 in Fig. 14) has \( \varepsilon_{Nd}(t) \geq +4 \), Sr/Y of ~40, moderate Ba/Ta and Ba/La, and moderate concentrations of Ba and REE. A plot of \( \varepsilon_{Nd}(t) \) against La/Yb shows more scatter than the other elements. The low \( \varepsilon_{Nd}(t) \) end members from both north and south have similar characteristics, with the exception of higher REE concentrations in the north. An obvious candidate for this low \( \varepsilon_{Nd}(t) \) end member is the Ancient gneiss complex, which at ca. 3.23 Ga had \( \varepsilon_{Nd} \) values between about –3 and –6 (see summary in Schoene et al., 2009), low Sr/Y (Fig. 12), and variable REE contents. Alternatively, these rocks could have been generated by melting the Ancient gneiss complex with S1 and S2 as sources of contamination/mixing. In the case of the southern Usutu rocks, because high Sr/Y correlates with high \( \varepsilon_{Nd}(t) \) and because no high Sr/Y (>50) rocks have been measured south of the Barberton greenstone belt, it seems most likely that the high Sr/Y was generated in a depleted source rather than by melting of an enriched crust such as the Ancient gneiss complex. The Nhlangano gneiss is another potential assimilation source for southern Usutu magmas. However, a plot of \( \varepsilon_{Nd}(t) \) versus Nd also shows that our samples of Nhlangano gneiss are likely a mixture of multiple components, and the Nhlangano gneiss does not have adequately low \( \varepsilon_{Nd}(t) \) values to explain Usutu magmas.

It is possible that the S1 and S2 melts were generated from a source with similar composition but different Nd isotopic signature. It is worth noting that our sample with the highest \( \varepsilon_{Nd}(t) \) from the southern rocks has a value of ~0, and thus the S2 end member is \( \geq 0 \). It is not likely to reach values as high as those for the S1 end member (\( \geq +4 \)), because extrapolating the S2 trajectory in Figure 14 to an \( \varepsilon_{Nd}(t) \) of...
also lack any migmatic or subsolidus fabrics and thus correspond to the end of deformation and ca. 3.23 Ga magmatism in the area. It follows that there was a link between the termination of magmatism and deformation and a decrease in crustal assimilation and/or melting.

**Toward a Regional Geodynamic Model of Craton Assembly**

As outlined herein, current models for the eastern Kaapvaal craton focus on subduction and terrane accretion in and around the Barberton greenstone belt at ca. 3.23 Ga (de Wit et al., 1992; de Ronde and de Wit, 1994; Kamo and Davis, 1994; de Ronde and Kamo, 2000; Stevens et al., 2002; Dziggel et al., 2005; Schoene et al., 2008), and (3) potentially synchronous high-pressure, low-temperature metamorphism in the Stolzburg terrane (Diener et al., 2005; Dziggel et al., 2005; Moyen et al., 2006). Furthermore, studies based on syntectonic deposition and folding of sedimentary strata and paleocurrent directions in the ≤3.23 Ga Moodies Group from the Barberton greenstone belt are consistent with both the Ancient gneiss complex and rocks north of the Barberton greenstone belt as the overriding plates during subduction (Lamb, 1984; Jackson et al., 1987; Heubeck and Lowe, 1994a, 1994b).

In our revised model, generation of the Nhlangano gneiss occurred in the deep crust via subduction-zone magmatism and deep arc deformation ca. 3.28–3.24 Ga, followed by collision of the two continental fragments north and south of the Barberton greenstone belt at ca. 3.23 Ga (Fig. 16A). Continued NW-SE oblique contraction and partial to full delamination of the oceanic slab resulted in a migration of magmatism inward toward the suture ca. 3.236–3.224 Ma, thus generating the Usutu suite in a dominantly NW-SE contractual setting (Fig. 16B). The source magmas for the southern Usutu suite retain an adakite-like subduction signature and an \( \varepsilon_{Nd}(t) \) of ~0; the source for the northern Usutu suite was generated by melting a depleted mantle with \( \varepsilon_{Nd}(t) \) of about +4 by advective heating by and/or decompression of an upwelling mantle related to slab breakoff (Figs. 14 and 16C). We note that the interpretation of the geochemistry reported here is not unique. One could also suggest that there was a north-dipping subduction zone located to the south of the Nhlangano gneiss, resulting in the release of trench fluids (e.g., high Ba, Ba/La; Fig. 14) closer to the trench in the southern Usutu suite. However, given that complementary geologic evidence for subduction south of the Nhlangano gneiss has not been documented, we suggest the model in Figure 16 is the simplest explanation. A more detailed geochemical investigation of the Usutu suite and Nhlangano gneiss is certainly warranted, given our promising initial results here.

Other uncertainties exist regarding the geometries and mechanisms of ca. 3.23 Ga
ca. 3300(?)–3240 Ma

Figure 16. Idealized cartoon illustrating the geologic evolution of the eastern Kaapvaal craton. The subduction-accretion and its significance to the rest of the Kaapvaal craton. The Stolzburg and Steynsdorp complexes (Fig. 1) are labeled as terranes to signify that they are likely complex amalgamations of plutonic and volcano-sedimentary material (de Wit et al., 1992; Lowe, 1994). Their affinity to the Ancient gneiss complex suggest a common origin. For that reason, we have placed the subduction zone to the NW of the Stolzburg terrane. In reality, the complicated three-dimensional nature of the system, and potentially large strike-slip offset at ca. 3.2–3.1 Ga, means that the present geometry of the system may not reflect terrane positions at ca. 3.23 Ga accurately. The 3.2–3.3 Ga plutonic rocks from the central Kaapvaal craton, exposed in the Vredefort impact structure and the Johannesburg Dome (Robb et al., 1992; Barton et al., 1999; Hart et al., 1999; Poujol and Anhaeusser, 2001; Flowers et al., 2003), may imply the plate boundaries inferred here in the eastern Kaapvaal craton can be extended further west (de Wit et al., 1992). Expanding the model proposed here will require more geochronology in combination with structural investigation in the limited basement exposures north and west of the Barberton greenstone belt. Additional work in the Nhlangano gneiss will also provide a clearer picture into its role in ca. 3.23 Ga orogenesis, especially with regard to its distinct structural pattern (e.g., N-S foliation instead of NE-SW further north), and possible later reactivation at ca. 2.7 Ga (Condie et al., 1996).

ACKNOWLEDGMENTS

This work would not have been possible without the scientific discourse and logistic support of M.J. de Wit. Comments and reviews at various stages of manuscript preparation were provided by M. Chiaradia, J. Connelly, R. Miller, R. Parrish, U. Schaltegger, and an anonymous reviewer. Special thanks are due to J.-F. Moyen for use of his geochemical database, as well as for stimulating discussion and commentary on this work. Funding came in part from National Science Foundation grant EAR-9526702 to Bowring.

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