Geological Society of America
Special Paper 405
2006

The Limpopo Belt: A result of Archean to Proterozoic, Turkic-type orogenesis?

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ABSTRACT

Evidence is presented relating to the tectonic evolution of the Limpopo Belt of southern Africa. It is demonstrated that this evolution was protracted, lasting at least 700 m.y. between ca. 2.7 and ca. 2.04 Ga, and comprised the complex assembly of terranes along shear zones through processes involving compression, extension, and strike-slip strain (transpression and transtension). It is argued that this evolution is more akin to Turkic-type craton building than to Himalayan- or Alpine-type collision between the Kaapvaal and Zimbabwe cratons at either ca. 2.7 Ga or ca. 2.04 Ga, to which most earlier workers have subscribed. The steeply dipping shear zones separating terranes are tens of kilometers wide with gradational contacts, as opposed to narrow (<1 km) shear zones with relatively sharp contacts that characterize the Himalayas and Alps. Ductile deformation with both subvertical and horizontal transport directions is common in the Turkic-type orogenesis rather than thin skin, thrust tectonics. The shear zones are interpreted to be artifacts of long-lived subduction zones. Relics of calc-alkaline (I-type) magmatic arcs that formed both on continental and oceanic crust are identified and are related to subduction-accretion of the various terranes. In contrast, S-type magmatism (melts derived from sedimentary protoliths) characterizes deformation after continent-continent collision in the Alps and Himalayas. Individual terranes are characterized by distinct P-T-time (pressure-temperature) paths that show they experienced different tectonic histories under high-grade amphibolite- to granulite-facies conditions instead of lower-grade, greenschist- to amphibolite-grade conditions, which are more common in the Alps and Himalayas. Individual terranes also have distinct metallogenic signatures.

Keywords: Limpopo Belt, tectonics, terranes, P-T-time paths, Turkic orogenesis.

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Barton, J.M., Jr., Klemd, R., and Zeh, A., 2006, The Limpopo Belt: A result of Archean to Proterozoic, Turkic-type orogenesis?, *in* Reimold, W.U., and Gibson, R.L., Processes on the Early Earth: Geological Society of America Special Paper 405, p. 315–332, doi: 10.1130/2006.2405(16). For permission to copy, contact editing@geosociety.org.

INTRODUCTION

The Altaids of central Asia have been interpreted by Şengör and Natal'in (1996) as a wide complex of terranes formed by a subduction-accretion assemblage of crustal elements onto the Russian and Angara cratons over a period of several hundred m.y., lasting from the Vendian to the Permian (Table 1). Individual terranes evolved geographically independent of one another but were tectonically linked along one or more long-lived subduction zones, along which the strain changed among compression, transtension, or transpression and extension several times. The Russian and Angara (Siberian) cratons each increased in size by this mechanism before they were finally juxtaposed in their present orientation. The individual terranes comprising the Altaids therefore interacted with one another in diverse ways so that accretion in one area might have involved the welding of pieces of continental crust, while in others it might have involved the joining of oceanic magmatic arcs with continental nuclei. Some terranes may even be joined on a smaller scale along sutures similar to those in Alpine and Himalayan orogens, and individual terranes might also contain evidence for more than one period of tectonism (Şengör and Natal'in, 1996). Şengör and Natal'in (1996) argue that this type of continent building, termed by them *Turkic-type,* involves magmatic arcs on both continental and oceanic crust, bulk shortening of the subduction-accretion complexes (steeply dipping, ductile tectonics), and often high-grade metamorphism and was probably also the primary mechanism for forming Precambrian cratons. It is exemplified by the accretion process presently active along the western margin of North and South America.

The individual terranes within the Altaids are pieces of continental and, to a lesser extent, oceanic crust (ophiolites) that are bounded by steeply dipping shear zones, tens of kilometers wide with gradational contacts and that comprise rocks with: (1) different geological histories as outlined by isotopic signatures including ages in the absence of fossils; (2) distinct metamorphic P-T-time (pressure-temperature) path(s) that may be related to distinct tectonic environments; (3) unique mineral deposits, and (4) lithological units that cannot be traced across shear zones from one crustal element to another along their boundaries (criteria from Jones et al., 1983). The magmatic arcs on continental crust are recognized by the linear orientation of calc-alkaline granitoid intrusions, whose compositions may be directly related to subduction magmatic processes. The bulk shortening is recognized by relatively steeply plunging, ductile deformation in shear zones occurring at up to high-grade conditions and high-grade metamorphic rocks with upper-amphibolite to granulite-facies metamorphic assemblages.

These characteristics are in marked contrast to those of Alpine- and Himalayan-type orogenies (Table 1). In the Western Alps (the portion of the Alps that most modelers of the Limpopo Belt use as an analogue; Şengör and Natal'in, 1996; Stampfli et al., 2002; www-sst.unil.ch/research/alpes/cross_section.htm), the African Plate was thrust over the European Plate, resulting in tectonic polarity of northward-directed thrust sheets and crustal thickening through thrust stacking. Uplift during tecton-

TABLE 1. COMPARISONS OF CHARACTERISTICS OF ALPINE, HIMALAYAN, AND TURKIC OROGENS WITH THOSE OF THE LIMPOPO BELT

Characteristics	Alpine orogen	Himalayan orogen	Turkic orogen	Limpopo belt
Lateral subduction zone links among terranes before joining	No	No	Yes	Yes
Duration	Two episodes, one during the Cretaceous and the second ca. 40 m.y. ago	One episode beginning ca. 50 Ma and continuing until today	ca. 400 m.y. from ca. 650 to ca. 245 Ma	ca. 700 m.y. from ca. 2.7 to ca. 2.0 Ga
Granitoid magmatism associated with tectonism	Uncommon	Common in backthrust belt	Common	Common
I-type, calc-alkaline granitoid magmatism	Common, early before craton-craton collision	Common, early before craton-craton collision	Common, dominant	Common, dominant
S-type granitoid magmatism	Common, after plate-plate collision	Common, after plate-plate collision	Minor	None recorded
Ophiolite complexes recognized	Yes	Yes	Yes	No
Terranes involved	2, Africa and Europe in the Western Alps	2, Asia and India	Many	>10
Width of major boundaries between terranes	<1 km	<1 km	>10 km	>20 km
Horizontal tectonics	Dominant	Dominant on the Indian Plate	Occurs	Occurs
Craton-craton overthrustng	Yes	No	Locally occurring	No
Syntectonic, foreland sedimentary basins	Yes	Yes	Uncommon	None recognized
Uplift rates after peak metamorphism	Variable but generally ~10 mm/year	Variable but generally ~10 mm/year	Generally <1 mm/year	All determined <1 mm/year (e.g., 0.3 mm/year)

Note: For sources of information, see text ism was variable but during peak deformation it was on the order of 10 mm per year. Major foreland sedimentary basins formed syntectonically, both to the north and to the south. Magmatism was minor and exhumation of granulite-facies metamorphic rocks was restricted to the Ivrea Zone, the tectonically upturned, lower portion of the African Plate. The suture between the African and European Plates is situated at the base of the Ivrea Zone and is a few hundred meters wide. Alpine tectonism was rapid, occurring over a few tens of million years. Terranes were not involved in the tectonism of the Western Alps but are involved to the east.

In the Himalayas proper (Table 1 herein, Molnar and Tapponnier, 1975; Tapponnier et al., 1982; Searle et al., 1987; Şengör and Natal'in, 1996), tectonism began in the Eocene and is still going on. No overthrusting of cratons is involved, and the suture between the Asian and Indian Plates is narrow, a few hundred meters wide. The Indian Plate is being shortened by the southward stacking of thrusts, and the Asian Plate is being deformed by escape tectonism along wrench faults. Present uplift rates in the Himalayas are variable, but rates of 10 mm per year are not uncommon. Syntectonic sedimentary foreland basins occur both to the north and to the south. Granitic rocks in the Himalayas proper are predominantly S-type, formed from melting of sediments with large water contents. Calc-alkaline magmatism occurred on the Asian Plate as a result of subduction of oceanic crust beneath it prior to collision with the Indian Plate; i.e., in this instance calc-alkaline magmatism is pretectonic. Exposure of high-grade metamorphic rocks is uncommon, although it occurs in the east (e.g., Liu and Zhong, 1997).

The Limpopo Belt, although smaller, has long been described as either an Alpine- or a Himalayan-type orogenic belt caused by the oblique collision of the Zimbabwe and Kaapvaal cratons (e.g., Coward, 1976; Fairhead and Scovell, 1976; Coward and Fairhead, 1980; Fripp, 1982; Light, 1982; van Reenen et al., 1987, 1992; Roering et al., 1992; Treloar et al., 1992; Smit and van Reenen, 1997). This collision was interpreted to have occurred either at ca. 2.7 Ga or at ca. 2.04 Ga (e.g., Van Breemen and Dodson, 1972; Barton and van Reenen, 1992a; Barton et al., 1994; Kamber et al., 1995b; Barton and Sergeev, 1997; Jaeckel et al., 1997; Holzer et al., 1998, 1999). In these models, the continent-continent collision involved the thrusting of the Kaapvaal craton over the Zimbabwe craton, resulting in thickened continental crust. Only rocks belonging to the colliding cratons were involved; i.e., by implication the Limpopo Belt should contain no terranes other than the Zimbabwe and Kaapvaal cratons. Also in these models, the isostatic reaction to this crustal thickening was rapid uplift and erosion or "pop-up" (rapid, isothermal decompression; e.g., van Reenen et al., 1987; Roering et al., 1992), and the uplifted material was thrust outward onto the colliding cratons. The abundant granulite-facies metamorphic rocks presently exposed at surface supposedly formed at the bottom of the thickened crust and were uplifted and exposed during the final stages of isostatic adjustment. Despite the proposed rapid uplift, no foreland sedimentary basins related to it have been identified to the north or south. Sediments of the ca. 1.85 Ga Waterberg and Soutpansberg groups unconformably overlie the high-grade rocks of the Limpopo Belt (e.g., Watkeys, 1979, 1984; Barker, 1983; Watkeys et al., 1983; Cheney et al., 1990; Barton et al., 2003).

Alternative models for the Limpopo Belt involving terranes besides the Zimbabwe and Kaapvaal cratons have been proposed but not widely accepted. These recognized that the dominant strike-slip motions along the Palala-Zoetfontein and Triangle shear zones (Fig. 1) are incompatible with a continent-continent collision. Barton and Key (1981) argued, largely on the basis of available geochronological data, that the Central Zone (see next section for discussion of terminology) of the Limpopo Belt was exotic, unrelated to the adjacent cratons but incorporated between them along these strike-slip shear zones. McCourt and Vearncombe (1987, 1992) modeled an exotic Central Zone as an indenter, thrust from the east along these shear zones onto a combined Zimbabwe-Kaapvaal craton. Holzer et al. (1998, 1999) modeled the strike-slip motion along these shear zones as a necessary result of oblique collision between the Zimbabwe and Kaapvaal cratons, resulting from southeastward thrusting of the Magondi orogen to the west.

Several observations, however, many of which have been known for decades, are incompatible with a single craton-craton collision scenario. In this paper, we review these observations and suggest that the creation of the Limpopo Belt might be better interpreted in terms of Turkic-type rather than Alpine- or Himalayan-type orogenesis.

CRUSTAL ELEMENTS OF THE LIMPOPO BELT

As originally defined (Cox et al., 1965; Mason, 1973), the Limpopo Belt (Fig. 1) consists of three crustal elements: a Central Zone characterized by a unique succession of leucocratic, granitic, and high-grade, commonly migmatitic, quartzofeldspathic rocks, and supracrustal lithologies (quartzite and calc-silicate, carbonate, metapelitic, and para-amphibolitic rocks) separated by the wide Palala-Zoetfontein and Triangle shear zones from two marginal zones. The Southern Marginal Zone consists of high-grade, intensely deformed and metamorphosed, tonalitic, and trondhjemitic rocks with minor greenstones and was interpreted as reworked rocks of the Kaapvaal craton (e.g., Coward et al., 1976; van Reenen et al., 1987, 1992), whereas the Northern Marginal Zone is dominated by enderbite and enderbitic orthogneisses with very minor greenstones and a few granitic rocks (e.g., Mason, 1973; James, 1975; van Reenen et al., 1987, 1992; Kamber and Biino, 1995; Jelsma and Dirks, 2002). The rocks of the marginal zones are themselves thrust, respectively, southward and northward, onto the respective cratons along >10 km wide, steeply dipping shear zones, the Hout River and Umlali shear zones, respectively. All four shear zones mentioned above have complex strain histories and gradational contacts (see discussion below and Table 1). The three zones of the Limpopo belt were thought to have been

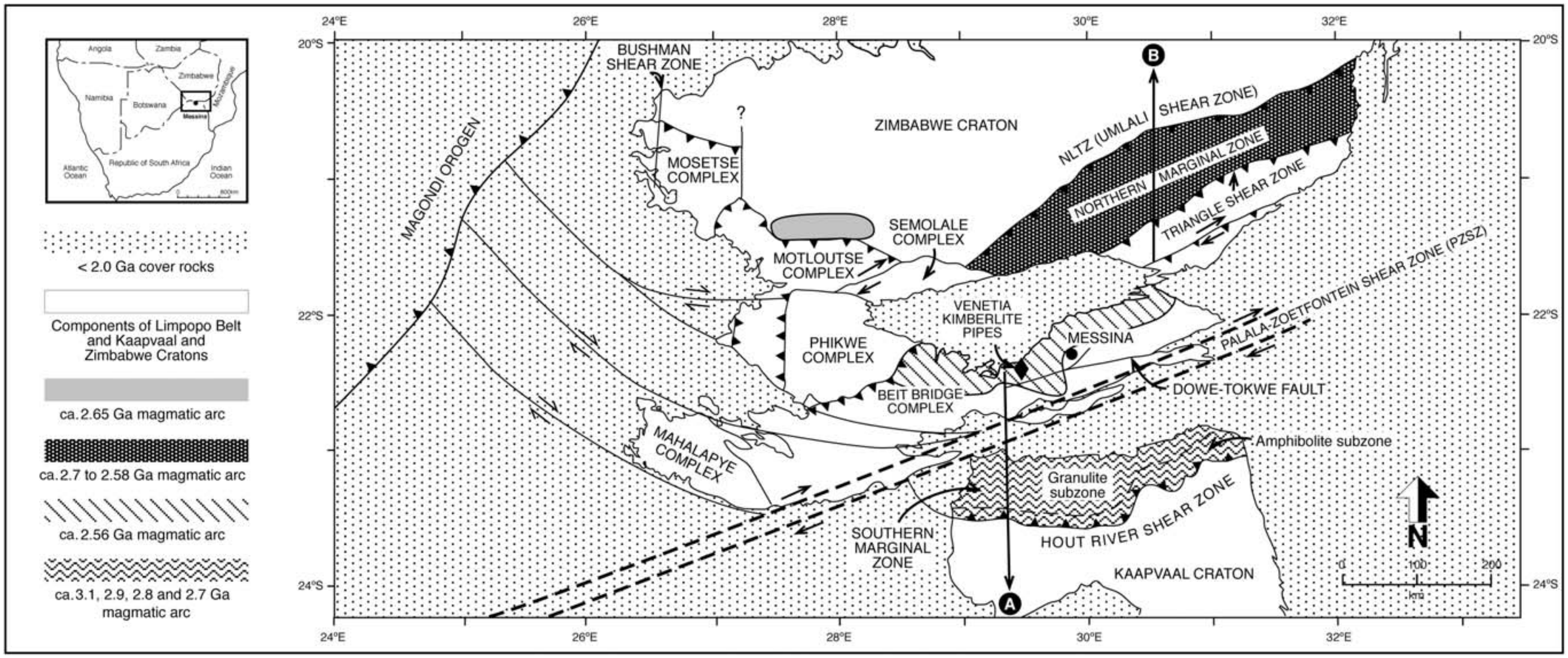

Figure 1. Geological map of the Limpopo Belt showing the terranes comprising it, rocks belonging to magmatic arcs and the Magondi orogen (modified from Aldiss, 1991; Carney et al., 1994; Barton and Pretorius, 1998; Holzer et al., 1999; Barton et al., 2003). NLTZ—northern Limpopo transition zone.

formed under granulite-facies conditions and subsequently overprinted under retrograde conditions (e.g., van Reenen et al., 1987, 1992). Their lateral extent to the east and west is poorly defined because they are covered by younger sedimentary and volcanic rocks.

Subsequent mapping, primarily in Botswana, delineated other shear zone–bounded crustal elements that had originally been considered part of the western Central Zone (e.g., Hickman and Wakefield, 1975; Aldiss, 1991; Brandl, 1992; McCourt et al., 2004). In addition, what was traditionally called the Central Zone proper was subdivided by Aldiss (1991) into (1) the Mahalapye Complex, which is dominated by ca. 2.02 Ga to ca. 2.0 Ga igneous rocks with minor metasedimentary rocks (Hisada and Miyano, 1996; Hisada and Paya, 1997; McCourt and Armstrong, 1998; Holzer et al., 1999; Chavagnac et al., 2001; Hisada et al., 2004); (2) the Phikwe Complex dominated by Archean, hornblende-bearing, tonalitic and trondhjemitic gneisses and igneous rocks (e.g., Brandl, 1992); and (3) the Beit Bridge Complex, with the infra- and supracrustal lithologies originally ascribed to the Central Zone. The granitic quartzofeldspathic rocks of the Beit Bridge Complex are now recognized as Archean, ca. 3.2 Ga orthogneisses (Boryta, 1988; Boryta and Condie, 1990; Barton and Pretorius, 1998; Barton et al., 2003; Barnett, 2005), whereas the supracrustal paragneisses may be divided into two, unconformity bounded successions, one ≥3.1 Ga and the other ≤2.66 Ga in age (e.g., Eriksson et al., 1988; Brandl, 1992; Barton et al., 2003; Buick et al., 2003). Quartzite from within the older unconformity-bounded succession contain zircons with U-Pb ages of ca. 3.2 Ga and ca. 3.8 Ga, suggesting that an unrecognized, very old protolith must exist within the Beit Bridge Complex (W. Compston, 1985, personal commun. And W. Compston and J.M. Barton Jr., 1985, personal commun.; Armstrong et al., 1987; Barton and Sergeev, 1997). The contact between the Phikwe and Beit Bridge complexes is clearly indicated on geological maps by changes in lithologies and roughly follows the valley of the Limpopo River (e.g., Aldiss, 1991; Brandl, 1992). However, because of poor exposure, the nature of this contact is uncertain.

Although the major shear zones separating the various crustal elements of the Limpopo Belt are easily observable on air photographs, satellite images, and geophysical surveys (e.g., Cox et al., 1965; Bahnemann, 1972; Mason, 1973), they are for the most part obscured by younger sedimentary rocks of middle Proterozoic age (Waterberg and Soutpansberg Groups) and Mesozoic (Karoo Supergroup) sediments and deep soils. Where exposed (Fig. 1; Table 2), they are tens of kilometers wide, anastomize, have long and complicated strain histories, and contain rocks of various metamorphic grades (see references below). As such, they are not technically shear zones with a coherent fabric formed in response to deformation in a single stress field but instead are wide zones of deformation in which reactivation in response to different stress fields has occurred. However, because the names of the "shear zones" are entrenched in the literature, the incorrect term "shear zone" is retained for them in this paper to avoid confusion.

TABLE 2. SUMMARY OF CHARACTERISTICS OF THE PRIMARY SHEAR ZONES WITHIN THE LIMPOPO BELT

Shear zones/characteristic	Hout River shear zone	Palala-Zoetfontein shear zone including the straightening zone of Bahnemann (1972)	Triangle shear zone or North Marginal transition zone	Umali shear zone or North Limpopo thrust zone
Orientation	Arcuate from west-northwest in the west to east-northeast in the east	Linear east-northeast	Linear east-northeast	Linear northeast
Separating	Kaapvaal Craton from Southern Marginal Zone	Separating Kaapvaal and Southern Marginal Zone from the Beit Bridge and Mahalapye Complexes	Separating the Beit Bridge Complex from the Northern Marginal Zone	Separating the Northern Marginal Zone from the Zimbabwe Craton
Orientation	Dipping steeply northward	Dipping both steeply northward and steeply southward	Dipping steeply southward	Dipping steeply southward
Width	>10 km	>22 km	~35 km	~5 km but deformation continues northward into the Zimbabwe craton
Displacement	Southwestward thrusting	Southward thrusting followed by right and left lateral strike-slip motion and then by normal faulting	Northward thrusting followed by right lateral strike-slip motion	Northward thrusting
Time(s) of movement	ca. 2.71 to ca. 2.67 Ga	ca. 2.05 to ca. 1.9 Ga; Post-Triassic	ca. 2.0 Ga	ca. 2.6 Ga
Primary lithologies involved	Calc-alkaline, quartzofeldspathic rocks of tonalitic to granodioritic compositions	Primarily calc-alkaline, quartzo-feldspathic rocks of tonalitic, enderbitic, and granitic compositions with minor meta-sedimentary and gabbroic rocks	Calc-alkaline, quartzofeldspathic gneisses and minor metasedimentary rocks	Calc-alkaline enderbitic to charnoenderbitic rocks with minor greenstone lithologies
Tectonics	Ductile	Ductile and brittle	Ductile	Ductile

Note: For sources of data, see text.

It is impossible to precisely demarcate the edges of these shear zones because strain along them was heterogeneous and decreased gradually outward over long distances. In any given shear zone, both brittle-deformed rocks (cataclasites and sometimes pseudotachylitic breccia; Brandl and Reimold, 1990) and ductile-deformed rocks (mylonites and sometimes steeply dipping sheath folds with straight gneisses) may be found (Table 1, see references below). However, because of limited exposure, only specific areas of any of them have been documented in detail. This limited documentation introduces uncertainty regarding sweeping regional pronouncements about the deformation histories illustrated in these possibly unrepresentative areas.

The arcuate, west-northwest– to east-northeast–trending Hout River and the northeast-trending Umlali steeply dipping shear zones have complicated strain histories, which are dominated by compression. The Hout River shear zone contains north-dipping shear foliations, whereas shear foliations in the Umlali shear zone dip to the south. For detailed descriptions of the Hout River shear zone that separates the Southern Marginal Zone from the Kaapvaal craton, see De Wit et al. (1992), Roering et al. (1992), Smit et al. (1992), Smit and van Reenen (1997), and Kreissig et al. (2001). For similar information for the Umlali shear zone or Northern Limpopo thrust zone separating the Northern Marginal Zone from the Zimbabwe craton, see Blenkinsop et al. (1995), Mkweli et al. (1995), and Holzer et al. (1998, 1999). In these shear zones, steeply plunging mineral elongation lineations predominate, and kinematic indicators in these zones show thrusting southwestward and northward, respectively, onto the adjacent cratons. Subhorizontal fabrics are minor. Geochronologic studies of the Hout River shear zone indicate that it was active between ca. 2.71 Ga and ca. 2.6 Ga (Kreissig et al., 2001), whereas motion on the Umlali shear zone occurred at or before ca. 2.6 Ga (Mkweli et al., 1995).

The east-northeast–trending, steeply southward dipping Triangle shear zone (Fig. 1, terminology of Kamber et al., 1995b) or Northern Marginal Transitional Zone (terminology of Holzer et al., 1999) is a deformation zone at least 35 km wide, in which a poorly constrained, early northward-thrusting event is overprinted by right-lateral transpression or transtension at ca. 2.0 Ga.

In the east-northeast–trending Palala-Zoetfontein shear zone exposed within the Koedoesrand window (McCourt, 1993; Schaller et al., 1999), rocks with specific strain senses are localized, and hence, the shear zone may be subdivided into domains that have different deformation histories at different metamorphic grades during deformation ranging from granulite to probably greenschist facies (e.g., McCourt, 1983; McCourt and Vearncombe, 1987, 1992; Brandl and Reimold, 1990; Broekhuizen et al., 1995; Schaller et al., 1999). For example, following the interpretations of McCourt (1983) and Schaller et al. (1999), in the ~4-km-wide southern domain, rocks belonging to the ca. 1.7 Ga to ca. 2.0 Ga Palala granite (e.g., Barton and McCourt, 1983; Walraven et al., 1990), the compositionally unique Koeodesrand sediments of uncertain age and ca. 2.04 Ga mafic rocks of the Bushveld Complex are all cataclastically deformed and subsequently mylonitized, indicating variations in P-T conditions during strain. Shear planes dip steeply to the south. The southern domain is terminated on the south across the Abbottspoort shear zone by undeformed igneous rocks of the Bushveld Complex. The >10-km-wide central domain is composed of lenses of >2.6 Ga paragneisses of uncertain origin occurring within calc-alkaline felsic, mafic, and enderbitic mylonites and ultra-mylonites. Shear planes dip steeply to the north with subhorizontal mineral elongation lineations. In the ~2.5-km-wide northern domain, calc-alkaline, quartzofeldspathic gneisses, retrogressed calc-alkaline, <2.1 Ga charnockitic rocks, and intermediate to felsic, calc-alkaline mylonites surround layers of calc-silicate rocks, compositionally similar to those within the Beit Bridge Complex to the north. On steeply northward-dipping shear planes, older, steeply dipping mineral elongation lineations associated with southward transport are overprinted by subhorizontal, amphibolite to greenschist facies mineral elongation lineations associated with kinematic indicators showing primarily right-lateral, with lesser left-lateral, strain. The northern boundary of the northern domain is defined as the Melinda fault, which offsets Mesozoic Karoo rocks, showing it was active at some point more recently than the Triassic and, hence, may be a young feature superimposed on the Palala-Zoetfontein shear zone. Deformation associated with the Palala-Zoetfontein shear zone extends at least 5 km further to the north into the Beit Bridge Complex in Bahnemann's (1972) straightening zone. Steeply northward-dipping shear planes show both subvertical and subhorizontal mineral elongation lineations in predominantly calc-alkaline, quartzofeldspathic rocks and minor paragneisses, including Proterozoic magnetite quartzite. Geochronologic studies (Schaller et al., 1999) indicate that the main movement of the Palala-Zoetfontein shear zone occurred between ca. 2.05 Ga and ca. 1.9 Ga.

In the west, the Palala-Zoetfontein and Triangle shear zones and their interpreted extensions have been interpreted to be truncated by the ca. 2.0 Ga to ca. 1.8 Ga Magondi orogen (Fig. 1, e.g., Hutchins and Reeves, 1980; Meixner and Peart, 1984; Carney et al., 1994; Holzer et al., 1998, 1999), and in some cases, movements along them control younger sedimentation and mafic volcanism, e.g., the rocks of the ca. 1.85 Ga Waterberg/Soutpansberg/Palapye Groups and those of the Permian to Triassic, Karoo Supergroup (e.g., Barker, 1983; Brandl, 1992).

CRUSTAL STRUCTURE BENEATH THE LIMPOPO BELT

Within the Limpopo Belt, three types of evidence exist with regard to deep crustal and upper lithospheric structure: (1) seismic, gravity, and resistivity geophysical data, (2) xenoliths from the Venetia kimberlite pipes, and (3) initial $^{207}Pb/^{206}Pb$ ratios of originally igneous and sedimentary rocks, which are related to μ values ($^{238}U/^{206}Pb$). These data exist only for the eastern portion of the Belt incorporating the Northern and Southern Marginal Zones and the Beit Bridge and Phikwe complexes.

Seismic, Gravity, and Resistivity Data

Refraction seismic data from a traverse from the Zimbabwe craton through the Northern Marginal Zone in Zimbabwe (Fig. 1, Traverse B) indicate that the Moho there shallows from ~40 km to 30 km going southward (Stuart and Zengeni, 1987). Similar data from a traverse from Pietersburg (now Polokwane) to Harare less precisely indicate the Moho but show a distinct, vertical, traveltime discontinuity at the position of the Palala-Zoetfontein shear zone (Durrheim et al., 1992). These data from Durrheim et al. (1992) also defined the deep crustal and upper mantle seismic structures used to produce the lithologic column under the Venetia kimberlite pipes (see next section). Data from a vibroseis traverse between Pont Drift and south of Pietersburg in South Africa (Fig. 1, Traverse A) come from insufficient depth (10 seconds two-way traveltime) to resolve the Moho under the Beit Bridge Complex, Southern Marginal Zone, and Kaapvaal craton (e.g., De Beer and Stettler, 1992; Durrheim et al., 1992; Roering et al., 1992). They do, however, indicate a reflector beneath the Beit Bridge Complex at ~10 km depth, underlain by a homogeneous zone with few if any reflectors. The reflection seismic data also do not resolve the Palala-Zoetfontein shear zone directly, indicating that it is a vertical to steeply subvertical feature.

On a more regional scale, Nguuri et al. (2001) interpreted refraction seismic data across the eastern Limpopo Belt to indicate an average depth to Moho of ~37 km beneath the Northern Marginal Zone and of ~40–42 km beneath the Southern Marginal Zone and the Kaapvaal craton. Data are ambiguous for rocks beneath the Beit Bridge Complex but suggest that the crust must be in excess of ~40–45 km thick, with possible discontinuities at ~35–30 km and between ~40 and 53 km. These data also indicate a dense lower crust beneath the Beit Bridge Complex. Deep seismic data (Fouch et al., 2004) indicate that the lithospheric keel beneath the Zimbabwe craton, Northern Marginal Zone, and Beit Bridge Complex extends to ~225–250 km, whereas the keel under the Kaapvaal craton and Southern Marginal Zone extends deeper to at least ~250–300 km. A discontinuity is indicated between these portions of crust at the position of the Palala-Zoetfontein Shear Zone.

Regional gravity data are interpreted to indicate that a zone of higher-density rocks extends from the Northern Marginal Zone southward under the Beit Bridge Complex, with its top at the same depth, ~10 km, indicated by seismic data (Fairhead and Scovell, 1976; Coward and Fairhead, 1980; De Beer and Stettler, 1992). No lower limit to the higher density rocks was modeled. Resistivity data also indicate a lithologic break at ~10 km depth beneath the Beit Bridge Complex (Van Zyl, 1978). The gravity data also show a distinct discontinuity, tens of km wide, at the position of the Palala-Zoetfontein shear zone (De Beer and Stettler, 1992).

Data from Crustal and Lithospheric Xenoliths

Crustal and upper mantle xenoliths brought to the surface in the ca. 530 Ma Venetia kimberlite pipes were studied by Pretorius (1996), Barton and Pretorius (1998), and Pretorius and Barton (2003a, 2003b). Lithologic and chemical compositions along with mineral-equilibrium pressure data and measured and calculated p-wave velocity data were combined with the seismic and gravity data (De Beer and Stettler, 1992; Durrheim et al., 1992) to construct a lithological section beneath the Venetia kimberlite pipes (Fig. 2). This approach of comparing geophysical data that reflect present structures with lithologies brought to the surface ~530 m.y. ago assumes that the crustal structure has not changed since the Venetia kimberlite pipes were emplaced. This assumption seems probable because this area is presently in isostatic equilibrium (De Beer and Stettler, 1992), and analysis of the kimberlite structures within the Venetia pipes suggests at most only a couple of kilometers of crust could have been eroded from the top since emplacement (Kurszlaukis and Barnett, 2003).

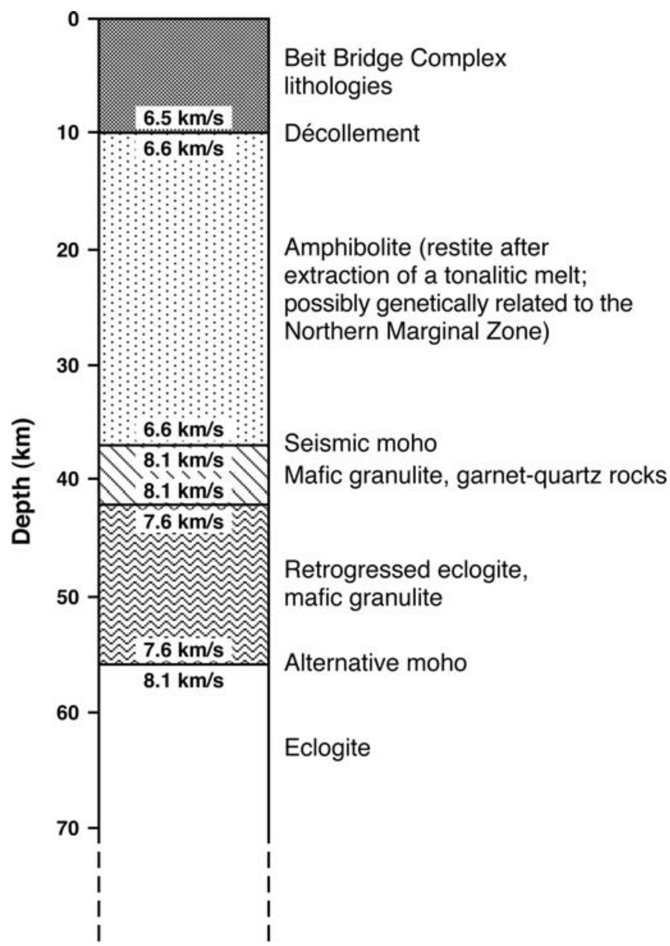

Figure 2. Lithological section beneath the ca. 530 Ma Venetia kimberlite pipes in the Beit Bridge Terrane constructed from pressure and P-wave velocity data and compositions of xenoliths from Venetia kimberlite (Pretorius, 1996; Pretorius and Barton, 2003a, 2003b).

Data from xenoliths extracted from beneath the Venetia kimberlite pipes indicate that lithologies of the Beit Bridge Complex extend only to a depth of ~10 km, where they terminate against amphibolite-facies, metabasitic rocks. This contact, also indicated by the seismic, gravity, and resistivity data, was interpreted as a décollement separating the supracrustal rocks from mafic rocks forming the lower crust (Barton and Pretorius, 1998). The mafic rocks, which extend to a depth of ~37 km, were interpreted as restites from which tonalitic magma has been extracted by partial melting and these restites correspond to the higher-density material indicated by the gravity and seismic data. These geophysical data, along with chemical data, were used to link the mafic rocks (mostly amphibolite) to the enderbitic rocks of the Northern Marginal Zone, suggesting that these lithologies may be genetically related (Pretorius, 1996; Pretorius and Barton 2003a) and, hence, might be the roots and top of an oceanic magmatic arc (see below, Fig. 3). However, the Triangle shear zone separates the Beit Bridge Complex from the Northern Marginal Zone, making this interpretation less tenable. The amphibolite is texturally homogeneous, which may explain the lack of reflectors observed in the vibroseis data.

Below ~37 km, p-wave velocities jump to ~8.1 km/s, caused by the increased density of garnet-quartz rocks and mafic granulite. The mafic granulite is chemically similar to the overlying amphibolite, suggesting that the seismic discontinuity is a metamorphic isograd. At ~42 km, p-wave velocities drop to ~7.6 km/s in mafic granulite and retrogressed eclogite that is again compositionally similar to the overlying amphibolite. P-wave velocities then increase again to 8.1 km/s at ~56 km depth where eclogite is encountered. Pretorius (1996) and Pretorius and Barton (2003a) suggested that this latter jump in p-wave velocities may represent the bottom of the continental crust, i.e., the lithological Moho. However, eclogite could also be a mafic component of continental crust metamorphosed to eclogite facies conditions, and so the base of the continental crust could be deeper. The various lithologic breaks between ~37 and 56 km may be those indicated in some of the regional refraction seismic data of Nguuri et al. (2001).

Pb-Isotopic Data

Large $^{207}Pb/^{204}Pb$ ratios compared to $^{206}Pb/^{204}Pb$ ratios (Fig. 4) characterize most igneous and meta-igneous rocks of the Beit Bridge and Phikwe Complexes and in the Northern Marginal Zone and Zimbabwe craton, suggesting that these complexes were derived from a source with a long-lived, high μ value (data from Barton et al., 1983; Taylor et al., 1991; J.M. Barton Jr., 1995, personal commun.; Barton, 1996; Kreissig et al., 2000). These large Pb isotopic ratios are in marked contrast to those for rocks exposed in the Southern Marginal Zone and the Kaapvaal craton, which were derived from sources with μ values close to bulk Earth (Fig. 4). Hence, the rocks of the Zimbabwe craton, Northern Marginal Zone, and Beit Bridge and Phikwe Complexes may be derived from the same mantle source, but they cannot be genetically related in an easy way to those of the Kaapvaal craton and Southern Marginal Zone.

A Cross Section Across the Limpopo Belt

Taking these geophysical and lithological data together, it is possible to create a schematic north-to-south cross section (Fig. 3) across the Limpopo Belt along the line of the vibroseis survey from south of Pietersburg on the Kaapvaal craton through the Southern Marginal Zone and Beit Bridge Complex to Pont Drift (De Beer and Stettler, 1992; Durrheim et al., 1992), and then northward along the seismic traverse of Stuart and Zengeni (1987) and Durrheim et al. (1992) through the Northern Marginal Zone and onto the Zimbabwe craton (Traverses A and B, Fig. 1). This cross section clearly shows the Palala-Zoetfontein shear zone as a wide crustal break, separating the Kaapvaal craton with the Southern Marginal Zone thrust upon it from the Beit Bridge Complex, and the Zimbabwe craton with the Northern Marginal Zone thrust upon it. The Beit Bridge Complex is shown as a, 10-km-thick veneer of rocks overlying mafic rocks that are separated from the Northern Marginal Zone by the Triangle shear zone. The cross section, as shown in Figure 3, is in

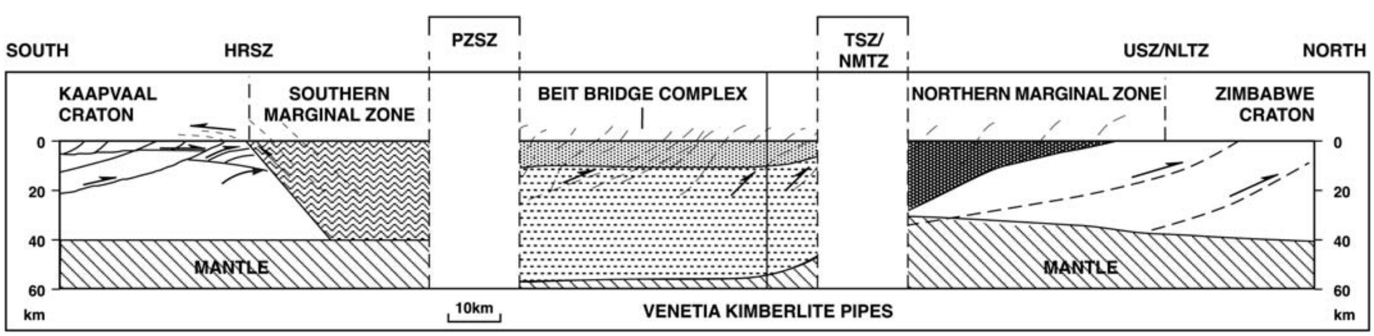

Figure 3. Cross section through the Limpopo Belt based on seismic and gravity data (modified from DeBeer and Stettler, 1992; Roering et al., 1992) and xenolith compositions from the Venetia kimberlite pipes (Pretorius, 1996; Pretorius and Barton, 2003a, 2003b). Note that the widths of the Palala-Zoetfontein (PZSZ) and Triangle (TSZ/NMTZ) shear zones are minimum estimates, and note that some authors show widths in excess of 35 km. HRSZ—Hout River shear zone, USZ/NLTZ—Umlali shear zone/northern Limpopo transition zone.

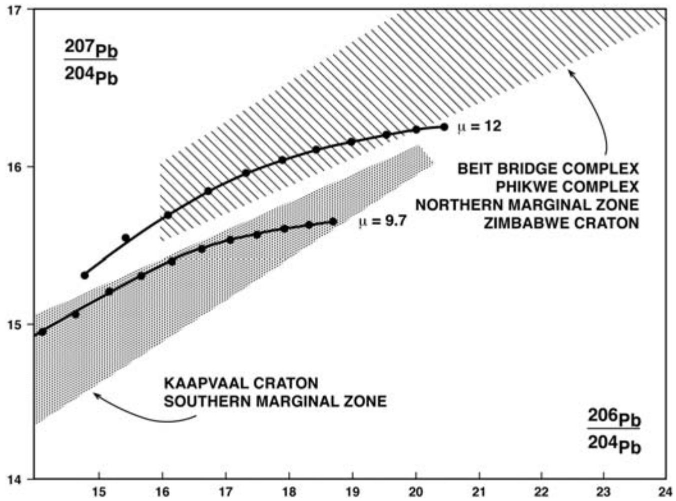

Figure 4. Plot of $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ data for rocks from different terranes in the Limpopo Belt and the Zimbabwe and Kaapvaal cratons (Barton et al., 1983; Taylor et al., 1991; Barton, 1995; and Barton, 2005, personal commun.; Kreissig et al., 2000). These data indicate that the rocks of the Kaapvaal craton and the Southern Marginal Zone are genetically different from those of the Beit Bridge Terrane, Phikwe Complex, Northern Marginal Zone and Zimbabwe craton and that the latter were derived from a source with a very long-lived, high μ composition.

marked contrast to that proposed by Roering et al. (1992, Fig. 2 herein), which was drawn to fit the "pop-up" model for the Limpopo Belt (e.g., van Reenen et al., 1987).

MAGMATIC ARCS

Calc-alkaline, tonalitic, granodioritic, and granitic lithologies, which are usually the result of melts formed and emplaced in magmatic arcs on continental crust in modern tectonic regimes, are extensive and are volumetrically significant within the Limpopo Belt, where they form orthogneiss complexes in linear bands (Fig. 1). Their presence implies that significant amounts of oceanic crust were subducted, dehydrated, and sometimes partially melted under continental crust during formation of the Limpopo Belt prior to ca. 2.0 Ga. The distribution of these rocks indicates where the subduction zones were located adjacent to them at the edge of the crustal blocks. However, ophiolites, blueschist-facies rocks, or extensive volumes of basaltic rocks with MORB (mid-oceanic-ridge basalt) signatures have not been identified within the Limpopo Belt, although one small occurrence of podiform chromite, often associated with sutures in continent-continent collisions, has been recognized within the Southern Marginal Zone (Smit, 1984). The lack of these subduction-related indicators could be used to argue that consumption of significant amounts of oceanic crust did not occur during formation of the Limpopo Belt and the Kalahari craton (see discussion in Light, 1982) or that this oceanic crust was completely subducted by a so-far unknown process.

The most obvious magmatic arc in the Limpopo Belt comprises the ca. 2.75 Ga to ca. 2.58 Ga charnoenderbitic to charnockitic igneous rocks making up ~90% of the Northern Marginal Zone (e.g., Berger et al., 1995; Kramers et al., 2001). As mentioned above, these rocks formed by partial melting of amphibolite and higher-grade rocks of an appropriate composition, similar to those preserved at a depth of >10 km beneath the Beit Bridge Complex (Pretorius, 1996; Pretorius and Barton, 2003a). McCourt et al. (2004) have also defined a ca. 2.65 Ga, granitic Francistown magmatic arc complex, probably related to subduction during the accretion of the Motloutse Complex onto the Zimbabwe craton. Magmatism in the Northern Marginal Zone and Motloutse Complex may be tectonically related to the emplacement of the upper greenstone succession in the Zimbabwe craton (Jelsma and Dirks, 2002).

Other examples of magmatic arcs within the Limpopo Belt are the ca. 2.63 to ca. 2.57 Ga tonalitic to granodioritic to granitic igneous rocks and orthogneisses within the Beit Bridge Complex, including the Bulai batholith and Singelele gneiss (e.g., Boryta, 1988; Boryta and Condie, 1990; Barton et al., 1994, 2003; Kröner et al., 1999; Boshoff et al., 2004; Barnett, 2005). These rocks are characterized by high U and Th contents that assist in their identification and in the delineation of their areal extent using radiometric remote sensing techniques (e.g., Andreoli et al., 2003; Barton et al., 2003; Barnett, 2005; Council for Geosciences, South Africa, 2003, personal commun.). In the Southern Marginal Zone and adjacent Kaapvaal craton, ca. 3.1–2.9 Ga tonalitic rocks (e.g., the Goud Plaats and Bavianskloof gneisses) and ca. 2.8 and 2.7 Ga granitic to granodioritic igneous rocks and orthogneisses (e.g., the Turfloop batholith and Matok pluton, respectively) probably indicate other magmatic arcs, although, with the exception of the Turfloop batholith, their limits have not been adequately defined (e.g., Barton et al., 1992; Brandl and Kröner, 1993; Henderson et al., 2000).

PRESSURE-TEMPERATURE-TIME PATHS

Pressure-temperature (P-T) paths throughout the Beit Bridge Complex and Southern Marginal Zone were initially interpreted to be parallel (identical), with significant isothermal decompression in retrograde segments followed by nearly isobaric cooling (e.g., van Reenen, 1983; van Reenen et al., 1987, 1988, 1990; Hisada and Miyano, 1996; Holzer et al., 1998; 1999; Kröner et al., 1999). It was thought that this was the result of a single orogenic event at ca. 2.7 Ga (e.g., van Reenen et al., 1987, 1990; McCourt and Vearncombe, 1987, 1992; Roering et al., 1992). Furthermore, it was suggested that all units within the Limpopo Belt underwent granulite-facies metamorphism, as supported by the results of petrologic studies, mostly from within a few tens of kilometers of Messina (now renamed Musina) within the Beit Bridge Complex (e.g., Windley et al., 1984; Droop, 1989; Perchuk et al., 2000) but also from the Mahalapye Complex (Hisada and Paya, 1997; Chavagnac et al., 2001), the Phikwe Complex (Tsunogae et al., 1989; Hisada and Miyano, 1996), the Northern Marginal Zone (Tsunogae et al., 1992; Kamber and Biino, 1995; Rollinson and Blenkinsop, 1995), and the Southern Marginal Zone (van Reenen, 1986; Stevens and van Reenen, 1992; Perchuk et al., 2000). These observations led to the conclusion that the high-grade metamorphic rocks of the Central and Marginal Zones of the Limpopo Belt were rapidly exhumed by "pop-up"–style tectonics (e.g., van Reenen et al., 1987; Roering et al., 1992; Holzer et al., 1998). Perchuk et al. (2004) presented a variation on this theme, whereby the pop-up was replaced by diapirism due to density differences among the rocks involved.

With the employment of quantitative phase diagrams such as pseudosections (e.g., Zeh et al., 2004b, 2005a, 2005b), variations in P-T paths of the Beit Bridge Complex in the Messina and Venetia areas have become apparent (Fig. 5A). Within the Beit Bridge Complex itself, different P-T-time paths are found within a few kilometers of one another in the Messina area and also in the area around the Venetia kimberlite pipes (Fig. 5A; compare and contrast results by, e.g., Droop, 1989; Klemd et al., 2003; Sachau, 2003; Zeh et al., 2003, 2004a, 2004b, 2005a, 2005b; van Reenen et al., 2004). One such example involves the P-T paths for the amphibolite-facies Endora and Venetia klippen near the Venetia kimberlite pipes that are separated by less than 1 km and the P-T path of the granulite-facies rocks of the Alldays area ~20 km to the west (Fig. 5A). The P-T vectors of most of these paths are similar to that obtained from the granulites of the Messina area but are shifted to lower pressure and temperature conditions. However, none of these paths displays significant isothermal decompression followed by isobaric cooling. Instead, they display simultaneous cooling and decompression after peak metamorphism and, thus, do not correlate with the paths from the Southern Marginal Zone, the Phikwe Complex, or the Mahalapye Complex (Fig. 5C; e.g., van Reenen, 1983, 1986; van Reenen et al., 1988; Stevens and van Reenen, 1992; Tsunogae et al., 1992; Kamber and Biino, 1995; Hisada and Miyano, 1996; Hisada and Paya, 1997; Hisada et al., 2000; Perchuk et al., 2000). Therefore, no single P-T path is characteristic of the Limpopo Belt as a whole.

When geochronological results are added to these P-T paths, further variations may be noted. In the Southern Marginal Zone, high-grade metamorphism occurred at or before ca. 2.7 Ga, as it is constrained by the intrusion at ca. 2.67 Ga of the Matok pluton through the retrograde, ortho-amphibole isograd (van Reenen, 1986; Barton et al., 1992). Much later, the Southern Marginal Zone was overprinted by greenschist- and amphibolite-facies metamorphism between ca. 2.0 and 1.9 Ga, as shown by Rb-Sr ages for biotite (Barton and van Reenen, 1992b) and U-Pb ages for apatite (Berger and Braun, 1997). At the same time, the Beit Bridge Complex of the Central Zone was affected by a major high-grade tectono-metamorphic event dated at ca. 2.04 Ga (e.g., Barton et al., 1994, 2003; Kamber et al., 1995a, 1995b; Jaeckel et al., 1997; Holzer et al., 1998, 1999; Kröner et al., 1999; Buick et al., 2003; Boshoff et al., 2004), with no evidence for a similar metamorphic event at ca. 2.7 Ga. Furthermore, no evidence for a ca. 2.7 Ga event was found in either the

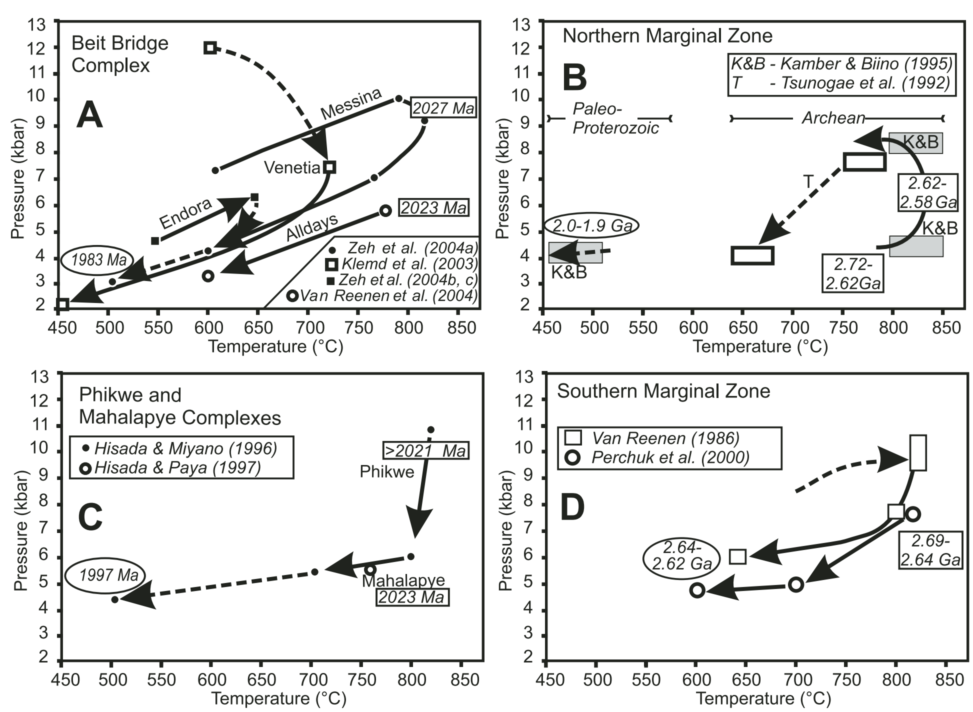

Figure 5. Contrasting pressure-temperature-time paths for various localities within the Limpopo Belt. For sources of the different age determinations, see discussion in text.

Mahalapye Complex (Holzer et al., 1999; Chavagnac et al., 2001) or the Phikwe Complex (Hickman and Wakefield, 1975; Holzer et al., 1999). A U-Pb age for apatite of 1983 ± 14 Ma (Holzer et al., 1998) and a Rb-Sr biotite age of 1970 ± 50 Ma (Barton et al., 1983) indicate that the Central Zone finally cooled below ~500 °C during the Paleoproterozoic.

In addition to the differences described above, the rocks of the Beit Bridge Complex show evidence of two high-grade metamorphic events, whereas those from other complexes do not (Barton et al., 1990). At the famous Sand River outcrop (Fripp, 1981, 1982, 1983; Buhlmann, 2003), migmatitic rocks of the Sand River Gneisses formed at >3.2 Ga and then were intruded by ca. 3.0 Ga mafic dykes under brittle conditions, as is shown by the local preservation of angular xenoliths of migmatite within them. Subsequently, these dykes underwent ductile deformation and granulite-facies metamorphism, resulting in partial melting of all other lithologies in the area. This granulite-facies overprint occurred at ca. 2.04 Ga (e.g., Barton et al., 1994; Jaeckel et al., 1997; Boshoff et al., 2004; Smit et al., 2004), much later than the intrusion of the Bulai batholith (~20 km to the west), which formed part of a ca. 2.57 Ga magmatic arc (see above). So far, there is no petrological, structural, nor geochronological evidence suggesting that the intrusion of the Bulai batholith was accompanied by an Archean high-grade regional metamorphic event (e.g., Barton and Sergeev, 1997; Jaeckel et al., 1997). Pb stepwise leaching ages of ca. 2.57 Ga, as reported by Holzer et al. (1998), were obtained from metamorphic garnet and sillimanite from xenoliths within the batholith. These ages, which are identical to that of the enclosing orthogneiss, are not indicative for a high-grade regional event at ca. 2.57 Ga but are more likely to reflect contact metamorphism. In fact, no isotopic or structural evidence for ca. 2.57 Ga metamorphism and deformation has been identified within the nearby Sand River Gneisses. The Sand River Gneisses commonly contain ca. 3.2 Ga zircons that only show metamorphic overgrowth at ca. 2.0 Ga, not at ca. 2.57 Ga (Barton and Sergeev, 1997). Furthermore, all structures can be correlated through their relationships with dated leucosomes with either the ca. 3.2 Ga or ca. 2.0 Ga events (e.g., Fripp, 1981, 1982, 1983; Barton et al., 1990; Jaeckel et al., 1997).

P-T paths for metapelitic rocks in the Messina area of the Beit Bridge Complex (Fig. 5A) indicate granulite-facies conditions were achieved along a clockwise metamorphic P-T path from 600 °C at 7.0 kbar to 780 °C at 9–10 kbar (the pressure peak) to 820 °C at 8 kbar (the thermal peak), followed by a decompression-cooling path to 600 °C at 4.0 kbar (Zeh et al., 2003, 2004b). A similar retrograde decompression-cooling path but with lower P conditions from 780 °C at 5.7 kbar (thermal peak) to 600 °C at 3.3 kbar was inferred by van Reenen et al. (2004) for high-grade gneisses from the central part of the Beit Bridge Complex northwest of Alldays. The peak P-T conditions derived by van Reenen et al. (2004) are similar to those inferred by Hisada and Paya (1997) for rocks from the Mahalapye Complex further southwest (>750 °C at ~5 kbar).

In contrast to the areas around Messina, northwest of Alldays and the Mahalapye Complex, metapelitic rocks of the Venetia klippe ~15 km north-northeast of Alldays and ~80 km west of Messina only underwent amphibolite-facies metamorphism, which is, so far, a unique feature within the Limpopo Belt (Klemd et al., 2003). Furthermore, two distinct P-T paths are distinguished, indicating that the area surrounding the Venetia kimberlite pipes forms part of a klippen complex. Rocks of the Venetia nappe provide evidence for a prograde decrease from 600 °C at 13 kbar (pressure peak) to 720 °C at 7–8 kbar (thermal peak), followed by retrogression to 350 °C at 2 kbar (Klemd et al., 2003). In contrast, metapelitic rocks and garnet-grunerite amphibolites from the Endora klippe, 1 km north of the Venetia klippe, provide evidence for a prograde P-T increase from 540 °C at 4.5 kbar to 650 °C at 6.5 kbar (metamorphic peak) followed by a P-T decrease (Zeh et al., 2004a, 2005a, 2005b). As already described, most P-T paths for rocks of the Beit Bridge Complex are similar but display varying pressure and temperature conditions as a consequence of horizontal tectonics. The only exception, so far, is the P-T path obtained by Klemd et al. (2003) from rocks from the Venetia klippe, which begins at much higher pressure. Its significance is not yet clear.

Despite their different early histories, P-T-time paths for rocks in the Beit Bridge, Phikwe, and Mahalapye complexes (Fig. 5) are similar, suggesting that the last deformation in them is related to a single orogenic event beginning at ca. 2.04 Ga that lasted no more than 30–60 m.y. (e.g., Hisada and Miyano, 1996; Zeh et al., 2004a, 2004b). The thermal peaks in rocks located ~10 km northwest of Messina occurred at ca. 2.03 Ga (Jaeckel et al., 1997; Holzer et al., 1998), in the area northwest of Alldays at ca. 2.02 Ga (Boshoff et al., 2004), in the Mahalapye Complex at ca. 2.02 Ga (Hisada and Miyano, 1996; Holzer et al., 1999), in the Phikwe Complex at ca. 2.08 Ga (Holzer et al., 1999), and in the Venetia klippe at ca. 2.02 Ga (Barton et al., 2003; Klemd et al., 2003). These ages are essentially the same within experimental error.

From a combined petrological, geochronological, and structural database, Zeh et al. (2004a, 2004b, 2005a, 2005b) concluded that the prograde evolution of the Beit Bridge Complex resulted from successive stacking due to horizontal nappe tectonics at ca. 2.04 Ga. Cooling from the thermal peak of this event is bracketed by a zircon age of 2027 ± 6 Ma (Jaeckel et al., 1997) and a U-Pb apatite age of 1983 ± 14 Ma (Holzer et al., 1998) for rocks from the Messina area. If it is assumed that the spherical zircon dated by Jaeckel et al. (1997) grew close to the thermal peak at 820 °C and 8 kbar and a closure temperature of ~500 °C is taken for the U-Pb system in apatite (e.g., Dahl, 1997), a linear cooling rate for these rocks of around 7 °C/Ma is indicated. Furthermore, a linear exhumation rate of ~0.3 mm/year is gained by assuming that the lithostatic pressure gradient was 270 bar/km and exhumation to 4 kbar at 500 °C. This slow exhumation rate and the retrograde P-T path involving simultaneous cooling and decompression argue against the pop-up model. The slow exhumation rate, mainly maintained by erosion (Zeh et al., 2004b), could, however, explain the general lack of reverse, dip-slip displacement fabrics associated with the ca. 2.04 Ga event.

In the Southern Marginal Zone, P-T paths (Fig. 5D) indicate peak metamorphic conditions of 820 °C at >9 kbar followed by decompression to 800 °C at 8.6–7.2 kbar and nearly isobaric cooling to 650–625 °C at >6 kbar (van Reenen, 1983, 1986; Stevens and van Reenen, 1992). Perchuk et al. (2000) derived a slightly different retrograde path, characterized by decompression and cooling from 870 to 850 °C at 8.2–7.2 kbar (thermal peak) to 700 °C at 5 kbar followed by isobaric cooling to 600 °C at 5–4.5 kbar. Kreissig et al. (2001) argued that this evolution occurred between ca. 2.69 and ca. 2.64 Ga.

Tsunogae et al. (1992, Fig. 5B herein) showed that rocks from the eastern portion of the Northern Marginal Zone experienced peak metamorphic conditions of 790–740 °C at 7.9–7.5 kbar followed by retrogression to 660–630 °C at 4.5–3.8 kbar. In contrast, Kamber and Biino (1995) inferred an anticlockwise P-T path from 850 to 800 °C at 4.5 kbar to 850–800 °C at 8.5 kbar, which took place at ca. 2.6 Ga, more recent than in the Southern Marginal Zone (Berger et al., 1995; Kamber and Biino, 1995; Mkweli et al., 1995).

The P-T paths described above for the Limpopo Belt are typical of many granulite-facies terranes worldwide (e.g., Harley, 1989). The anticlockwise P-T path of the Northern Marginal Zone involves isobaric cooling, whereas the clockwise P-T paths for most areas in the Southern Marginal Zone as well as the Phikwe and Mahalapye complexes involve a period of almost isothermal decompression followed by isobaric cooling. However, the slope of the clockwise P-T path for the Beit Bridge Complex indicates prograde simultaneous burial and heating during crustal thickening followed by contemporaneous decompression and retrograde cooling. The P-T paths of the Southern Marginal Zone and the Beit Bridge Complex are the result of tectonic thickening at different times, ca. 2.7 and ca. 2.04 Ga, respectively, followed by exhumation, which was dominated either by erosion (Beit Bridge Complex) or tectonic denudation (Southern Marginal Zone, Phikwe and Mahalapye complexes), producing a steeper-sloped decompression P-T path (e.g., Thompson and England, 1984). The prograde P-T paths of the Beit Bridge Complex and the Northern Marginal Zone are consistent with burial along a moderate geothermal gradient (intermediate P/T metamorphism) characteristic of continental collisions and, thus, in clear contrast to Alpine- or Himalayan-type metamorphism. The latter usually is associated with a steep geothermal gradient (high P/T metamorphism), which typically occurs in subduction zones (e.g., Ernst, 1988; O'Brien et al., 2001).

However, the most compelling difference between the tectonic evolution of most parts of the Limpopo Belt and Alpine- or Himalayan-type settings lies in the different post-peak metamorphic P-T evolutions. The Alpine or Himalayan types are characterized by a period of fast, almost isothermal, decompression (Ernst, 1988; Liu and Zhong, 1997; O'Brien et al., 2001), which is usually controlled by pure shear tectonics (e.g., Thompson and England, 1984; Ruppel et al., 1988). However, such tectonism is not supported by the retrograde P-T paths inferred from rocks of the Northern Marginal Zone and the Beit Bridge Complex, which either reflect isobaric cooling or contemporaneous cooling and decompression.

METALLOGENIC SIGNATURES

Although not particularly known as a major mineral-producing area, different crustal elements within the Limpopo Belt are characterized by different types and styles of metallogenesis, which formed prior to the final assembly of the Limpopo Belt at ca. 2.0 Ga. The Beit Bridge Complex was a significant source of alluvial corundum and also produced significant amounts of graphite, magnesite, and asbestos (Söhnge, 1945; Söhnge et al., 1948). It has also been a significant source and is a potential future source of magnetite iron ore (e.g., Söhnge, 1945; Du Plessis et al., 1997) derived from either hydrothermal alteration or weathering of banded iron formation that may be coeval with the ca. 2.6 Ga banded iron formation in the Transvaal succession (e.g., Walraven and Martini, 1995; Martin et al., 1998). The Cu and diamond mineralization in the Beit Bridge Complex is younger than ca. 2.0 Ga (e.g., Söhnge, 1945; Barton et al., 2003). The Phikwe Complex is unique in the Belt by containing a world-class Ni-Cu-Co deposit at Selebi-Phikwe with numerous minor occurrences around it (e.g., Van Breemen and Dodson, 1972; Gordon, 1973; Hickman and Wakefield, 1975; Wakefield, 1976; Lear, 1979; Gallon, 1986; Brown, 1987; Barton and van Reenen, 1992b). This mineralization appears to be associated with intrusion of one or more ultramafic sills that were subsequently subjected to intense ductile deformation at ca. 2.04 Ga (e.g., Wakefield, 1976; Barton and van Reenen, 1992b). Gold mineralization occurs in sheared rocks of the Southern Marginal Zone at Bochum (Wilson and Anhaeusser, 1998) and the Northern Marginal Zone at Renco Mine (e.g., Blenkinsop and Frei, 1996; Kisters et al., 1998; Blenkinsop et al., 2004). In the Bochum occurrence, mineralization is post–peak metamorphism (<2.7 Ga; Wilson and Anhaeusser, 1998), but in the case of Renco Mine, it apparently occurred during peak metamorphism (ca. 2.7–2.6 Ga) and was remobilized at ca. 2.0 Ga (e.g., Blenkinsop and Frei, 1996; Kisters et al., 1998; Blenkinsop et al., 2004).

ARE THE CRUSTAL ELEMENTS OF THE LIMPOPO BELT TERRANES?

From the evidence described above, the crustal elements recognized in the Limpopo Belt have all the characteristics of terranes (Jones et al., 1983). They are shear-zone bounded and have distinct geological histories as indicated by geochronology and P-T-time paths and different mineralization. They also contain lithologies that cannot be traced across the shear zones. Therefore, application of the word "complex" should be changed to the use of the word "terrane": e.g., the Beit Bridge Terrane.

ASSEMBLY OF THE LIMPOPO BELT

At this stage, knowledge is incomplete of the order and tectonic style in which the various terranes comprising the Limpopo Belt were assembled. Nevertheless, it is known that this assembly involved wide, steeply dipping, ductile shear zones and magmatic arcs over an extended period of time. The Mosetsi and Motloutse Terranes apparently were accreted onto the Zimbabwe craton at ca. 2.6 Ga (McCourt et al., 2004), and the Northern Marginal Zone apparently was accreted onto the Zimbabwe craton by ca. 2.58 Ga (Blenkinsop et al., 1995; Mkweli et al., 1995; Holzer et al., 1998, 1999). The Southern Marginal Zone had been accreted onto the Kaapvaal craton by ca. 2.7 Ga (see discussion above). The Beit Bridge Terrane, including a ca. 2.6 Ga magmatic arc complex, was accreted onto the Northern Marginal Zone along the Triangle shear zone at ca. 2.04 Ga and juxtaposed by transpression or transtension along the Palala-Zoetfontein shear zone with the Kaapvaal craton and Southern Marginal Zone at about the same time (e.g., Holzer et al., 1998, 1999). It is obvious, therefore, that the assembly of the terranes of the Limpopo Belt and Kalahari craton took place over a time span of ~700 m.y. and did not occur as a result of a simple and relatively rapid collision between the Zimbabwe and Kaapvaal cratons at either ca. 2.7 Ga or 2.04 Ga. The transtensional or transpressional stress field that existed along the Palala-Zoetfontein shear zone between ca. 2.04 and ca. 1.97 Ga could have produced regions of extension within the various terranes that it bounded and the Kaapvaal craton, allowing for the emplacement of the igneous Bushveld and Phalaborwa complexes at this same time (age constraints in Eriksson, 1984; Kruger et al., 1987; Walraven et al., 1990; Reischmann, 1995; Mapeo et al., 2004). It also seems unrealistic to think that the formation of the Limpopo Belt may be divorced from the evolution of the Magondi orogen that immediately followed it in the west and the formation of the Palapye, Waterberg, and Soutpansberg sedimentary basins on top of it shortly afterward.

CONCLUSIONS

Because the terranes comprising the Limpopo Belt have different tectono-metamorphic histories and were assembled independently onto the Zimbabwe and Kaapvaal cratons over a period of ~700 m.y. ending at ca. 2.04 Ga, it is possible and even probable that they evolved at widely spaced positions on the Earth's surface, as was the case with the Altaids (Şengör and Natal'in, 1996). The wide shear zones with gradational contacts separating them would be artifacts of major subduction zones connecting the terranes. This possibility is supported by the presence of magmatic arcs of different ages within the belt and the fact that the shear zones separating the terranes were reactivated at different times with different shear senses and at different metamorphic grades that could have resulted from either different crustal depths or different geothermal gradients. In addition, P-T-time paths for different terranes clearly reflect different tectonic regimes, even though activity within them sometimes overlapped in time. Because the Palala-Zoetfontein shear zone separates terranes with distinct Pb isotopic signatures, it probably represents the locus of final juxtaposition of the Zimbabwe craton with its associated terranes with the Kaapvaal craton and its associated terranes. Therefore, the assembly of the Limpopo Belt between ca. 2.7 Ga and 2.04 Ga was not similar to the formation of the Alps or the Himalayas. Instead it more closely resembles Turkic-type continental growth.

It is obvious from the above discussion of the currently available database that substantial further work is required in many parts of the Limpopo Belt to fully understand its tectonic evolution. Any new results will provide additional tests for the proposed Turkic-type model for the evolution of the Limpopo Belt.

ACKNOWLEDGMENTS

Inspiration for writing this paper came from a field trip to the Limpopo Belt in 2003 with Mike Searle, during which he took great pains to explain to JMB that he saw no Himalayan analogues there and from JMB listening to a series of lectures given by Kevin Burke, John Dewey, and Celâl Şengör in 2002 and again in 2003 at the University of the Witwatersrand, where the implications of transtension, transpression, and Turkic orogenesis were explained. Suddenly all the inconsistencies of applying an Alpine- or Himalayan-type model to the evolution of the Limpopo Belt became clear and a means of reconciling the inconsistencies was apparent. We also wish to thank all of our colleagues whose work we quote here for their contribution to the present understanding of the evolution of the Limpopo Belt. It is easier to construct models with limited data than to do so when all of the data are available and must be accommodated. We thank Paul Dirks, Roger Gibson, and Uwe Reimold for their critical and constructive reviews of an earlier version of this manuscript.

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Manuscript Accepted by the Society 4 October 2005

Printed in the USA