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# The Source of the Great Dyke, Zimbabwe, and Its Tectonic Significance: Evidence from Re-Os Isotopes

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## ABSTRACT

Re-Os data for chromite separates from 10 massive chromitite seams sampled along the 550-km length of the 2.58-Ga Great Dyke layered igneous complex, Zimbabwe, record initial  $^{187}\text{Os}/^{188}\text{Os}$  ratios in the relatively narrow range between 0.1106 and 0.1126. This range of initial  $^{187}\text{Os}/^{188}\text{Os}$  values is only slightly higher than the value for the coeval primitive upper mantle (0.1107) as modeled from the Re-Os evolution of chondrites and data of modern mantle melts and mantle derived xenoliths. Analyses of Archean granitoid and gneiss samples from the Zimbabwe Craton show extremely low Os concentrations (3–9 ppt) with surprisingly unradiogenic present-day  $^{187}\text{Os}/^{188}\text{Os}$  signatures between 0.167 and 0.297. Only one sample yields an elevated  $^{187}\text{Os}/^{188}\text{Os}$  ratio of 1.008. Using these data, the range of crustal contamination of the Great Dyke magma would be minimally 0%–33% if the magma source was the primitive upper mantle, whereas the range estimated from Nd and Pb isotope systematics is 5%–25%. If it is assumed that the primary Great Dyke magma derived from an enriched deep mantle reservoir (via a plume), a better agreement can be obtained. A significant contribution from a long-lived subcontinental lithospheric mantle (SCLM) reservoir with subchondritic Re/Os to the Great Dyke melts cannot be reconciled with the Os isotope results at all. However, Os isotope data on pre-Great Dyke ultramafic complexes of the Zimbabwe Craton and thermal modeling show that such an SCLM existed below the Zimbabwe Craton at the time of the Great Dyke intrusion. It is therefore concluded that large melt volumes such as that giving rise to the Great Dyke were able to pass lithospheric mantle keels without significant contamination in the late Archean. Because the ultramafic-mafic melts forming the Great Dyke must have originated below the SCLM (which extends to at least a 200-km depth), the absence of an SCLM signature precludes a subduction-related magma-generation process.

## Introduction

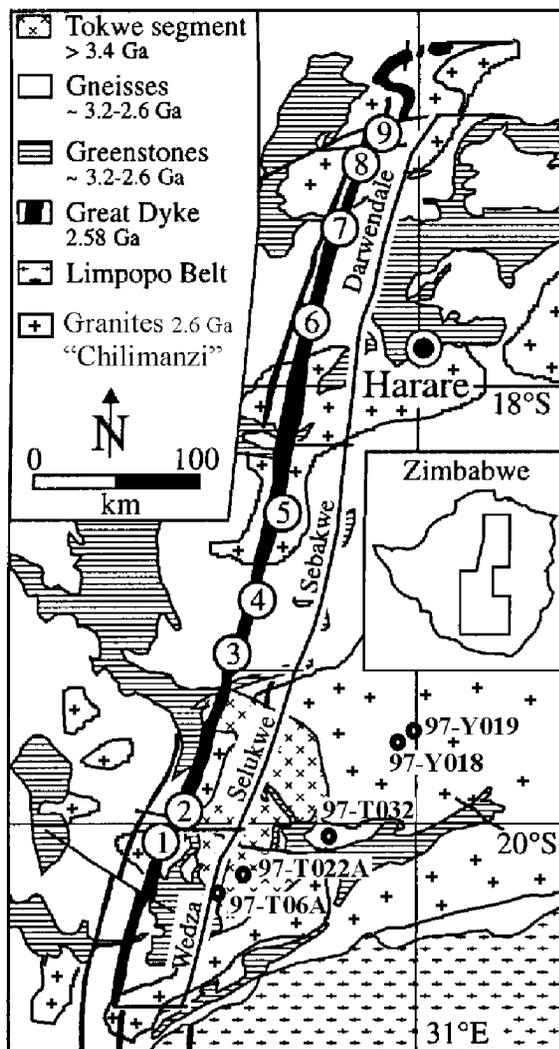
The Great Dyke of Zimbabwe is a very elongated mafic-ultramafic layered igneous complex (Wilson and Prendergast 1989), approximately 550 km long and 3–11 km wide (fig. 1). It intruded across the entire Archean Zimbabwe Craton through a major NNE-SSW-trending fracture system. Several recent geochronological studies (Mukasa et al. 1998; Armstrong and Wilson 2000; Wingate 2000; Collerson et al. 2002) have constrained its age close to 2580 Ma. A very precise set of dates fixes the age of intrusion at  $2576 \pm 1$  Ma (Oberthür et al. 2002) and also shows that emplacement and cooling oc-

curred rapidly. For comparison, U-Pb zircon and Rb-Sr whole rock dates with errors less than  $\pm 20$  Ma for the almost craton-wide Chilimanzi suite of granitoids range from  $2601 \pm 14$  to  $2574 \pm 14$  Ma (Hickman 1978; Wilson et al. 1995; Frei et al. 1999) and U-Pb dates for the Razi granite suite in the northern marginal zone of the Limpopo Belt from  $2627 \pm 7$  to  $2590 \pm 7$  Ma (Frei et al. 1999). These large intrusions document widespread lower crustal melting (Hickman 1978). Nevertheless, the mode of emplacement of the Great Dyke along craton-wide fractures implies that the crust of the Zimbabwe Craton including the Chilimanzi granitoids was rigid at 2576 Ma. This provides a rare opportunity to compare cratonization processes that occurred in the crustal part of the lithosphere (recorded by the well-documented metamorphic and magmatic history of the craton) with concomitant mantle processes (recorded as geochemical

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**Figure 1.** Geological sketch map of part of the Zimbabwe Craton, showing the Great Dyke (Wedza, Selukwe, Sebakwe, and Darwendale subchambers marked), chromite seam sample locations (*encircled numbers 1–9*), and granitoid-gneiss sample locations (*small circles [identified by 97 numbers]*).

fingerprints of the Great Dyke). Thus, two central questions are of major concern: (1) What was the source of the Great Dyke magmas? (2) Is there a possible connection between the craton-wide episode of granite formation at ~2600 Ma and the intrusion of the Great Dyke ~24 m.yr. later?

Studies of Rb-Sr, Sm-Nd, and Pb-Pb isotope systematics (Mukasa et al. 1998; Oberthür et al. 2002) established that the Great Dyke was either a slightly crustally contaminated asthenospheric mantle-derived intrusion or an uncontaminated intrusion from an enriched mantle source. Mukasa

et al. (1998) considered rather uniform initial Sr, Nd, and Pb isotope ratios to favor the latter, and viewed the Great Dyke in a subduction and continent collision context. Oberthür et al. (2002), examining a larger Sm-Nd data set, found variable amounts of crustal contamination, which they regarded as being possible only during emplacement. Emplacement-related contamination would allow a primary magma of pure mantle origin as the source for the Great Dyke, most probably of intra-plate character, and possibly produced by a mantle plume.

In resolving such questions, Os isotope systematics can usefully complement Pb, Sr, and Nd data. On the one hand,  $^{187}\text{Os}/^{188}\text{Os}$  ratios of mantle-derived magmas can be raised by crustal contamination, as shown in the case of the Bushveld layered igneous complex (McCandless et al. 1999; Schoenberg et al. 1999). On the other hand, they can be lowered by interaction with the mantle portion of the lithosphere. The subcontinental lithospheric mantle (SCLM) has normal mantle Os concentrations but is depleted in Re. Being long-lived under old cratons, it evolves, over time, to characteristically low  $^{187}\text{Os}/^{188}\text{Os}$  (Nägler et al. 1997; Pearson 1999) compared with the asthenospheric mantle, for which the Re-Os evolution is modeled from chondrite data (Allègre and Luck 1980; Walker et al. 1988; Meisel et al. 2001). Magmas originating from an ancient SCLM would thus have significantly lower  $^{187}\text{Os}/^{188}\text{Os}$  ratios than those from the asthenosphere (Lambert et al. 1995). Furthermore, as melts are Os depleted relative to their mantle source, the Os isotope ratio of a magma that originated in the asthenosphere may be strongly affected by the unradiogenic Os isotope signature of the SCLM through which it passes (Nägler et al. 1997).

This study reports new Re-Os isotope data for the Great Dyke and for gneisses and granitoids of the Zimbabwe Craton that can place limits on the amount of crustal contamination as well as help deciphering the involvement of the SCLM in the genesis of the Great Dyke.

### Geological Setting and Sampling

The NNE-SSW trending “Great Dyke” (the term is a misnomer) consists of four boat-shaped subchambers (fig. 1), each comprising a lower ultramafic sequence and an upper mafic sequence (Wilson and Prendergast 1989). Gravimetric studies revealed a trumpet-shaped cross section of these subchambers, with a feeder dike up to 1 km thick (Podmore and Wilson 1987). The layering of sequences con-

sistently dips inward toward the center, although not necessarily parallel to the contact of the complex with its host rocks. Two parallel fractures to the Great Dyke host further intrusions, the East Dyke and the Umvimeela (West) Dyke, which did not develop into differentiated layered complexes but whose gabbroic magma may be regarded as an undifferentiated equivalent to that of the Great Dyke itself (Wilson and Prendergast 1989).

The ultramafic sequences of the Great Dyke are characterized by repeated cyclic units, which are typically separated by a basal chromitite seam. In the lower parts of the sequences (lower Dunite Succession), cyclic units are dominated by olivine cumulates (dunitites and harzburgites). In the upper parts of the sequences (upper Pyroxenite Succession), the units grade from dunitites or harzburgites through olivine-bronzitites into orthopyroxenites. Chromitite seams typically contain between 70 and 95 vol % cumulus chromites with a grain size of 0.5–5 mm and interstitial serpentinized olivine and orthopyroxene (<1-mm size). Local small-scale shearing and mylonitization of cumulate layers could be the result of (1) subsidence of the layered intrusion during cooling (Wilson 1992) and (2) later craton-wide tectonism (Holzer et al. 1998), which probably also caused the offsets of the intrusion seen on the map (fig. 1).

The mafic sequences are characterized by the appearance of cumulus plagioclase in olivine-gabbros, norites, and gabbro-norites. A few meters below the transition from the ultramafic to the mafic sequence, the uppermost orthopyroxenite hosts the Main Sulfide Zone, a several-meter-thick layer containing disseminated sulfides locally enriched in platinum group elements (PGEs) and Ni (Prendergast and Wilson 1989; Prendergast 1990; Oberthür et al. 1997).

For this study, a variety of chromitite seams from different stratigraphic levels were sampled along the entire length of the Great Dyke (fig. 1) to provide both stratigraphic and geographical resolution. Because Re and platinoids are most likely to be associated with sulfides in such rocks, care was taken to sample only extremely fresh material in which no oxidation was apparent. Gneiss and granitoid sampling localities are also given in figure 1.

Chromite is an ideal mineral for Os isotope studies. First, it has relatively high osmium contents (5–400 ppb), which allow precise determination of the present-day Os isotopic composition. Second, the very low Re/Os ratios of chromites minimize uncertainties of in situ Re decay correction. Disturbance of chromite Re/Os ratios seems to be rare, and initial Os isotope ratios can commonly be de-

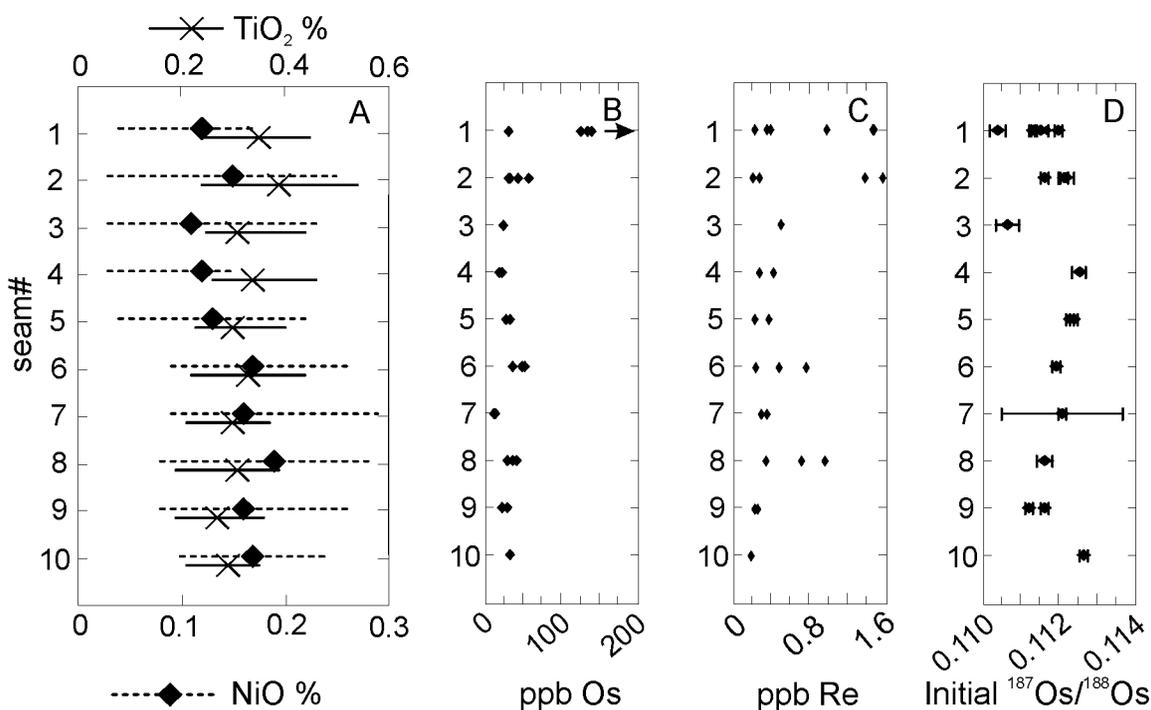
termined accurately. Third, chromite is very resistant to surface weathering and hydrothermal alteration with respect to Re-Os isotope systematics.

Many layered intrusions contain multiple horizons of massive chromite layers in their stratigraphy. The crystallization of such rhythmically occurring massive chromitites has been proposed to reflect repeated supplies of primitive magma into chambers containing more evolved magma (Irvine 1977). Mixing of these magmas drives the composition of the resulting hybrid into the stability field of chromite, which leads to chromite crystallization and deposition of massive chromitite layers on the chamber floor. Geochemical indicators, such as  $Al/(Al + Cr + Fe^{3+})$ ,  $Mg\# (= Mg^{2+}/[Mg^{2+} + Fe^{2+}])$ , or Ni and Ti concentrations of chromites reflect this process and are therefore useful to constrain the evolution of the magma by crystal fractionation and crustal contamination as a function of stratigraphy (Wilson 1982; Eales et al. 1990; Scoon and Teigler 1994).

### Analytical Techniques

Preconcentration of chromite grains from chromitite seams involved crushing and sieving, followed by magnetic and heavy liquid separation. Final mineral separates were handpicked. The separates were leached with 3 M HCl to remove possible minor magnetite overgrowths. This is an important step because if magnetite is present, there is a possibility that it was affected by weathering. This commonly results in a disturbance of the Re/Os ratio by redox processes (Frei et al. 1998).

Sample dissolution using Carius tubes with reverse aqua regia as a dissolving agent and chemical separation of Os followed the protocol of Nägler and Frei (1997). The U.S. Bureau of Standards mentioned incomplete dissolution of platinum group minerals (PGMs) after Carius tube digestion using reverse aqua regia in their experiments in the 1940s. Since it is nearly impossible to filter potentially undissolved submicron PGM inclusions from the dissolving agent, reproduction in initial  $^{187}Os/^{188}Os$  of different splits from the same sample was attempted as a control of this problem. In addition, a recent Os-isotope study on single osmiridium grains at the isotope laboratory in Bern revealed that the initial size of these grains of 200–400 microns decreased to undetectable sizes within two days in Carius tube digestion with reverse aqua regia. Re was separated using solvent extraction as described by Birck et al. (1997). Loading technique and analytical conditions for the negative thermal ionization mass spectrometry (N-TIMS) analysis of



**Figure 2.** Chemical and Re-Os characteristics of chromites along the sampled stratigraphy (seam 1, uppermost; seam 10, lowermost). *A*, Averages and ranges of TiO<sub>2</sub> and NiO concentrations. *B*, *C*, Os and Re concentrations; all analyses. *D*, Initial <sup>187</sup>Os/<sup>188</sup>Os ratios of samples shown with filled symbols in figure 3A.

sample 5/56-1 were identical to those applied by Schoenberg et al. (1999). Analytical methods for Re and Os isotope determination on the Nu-Instruments multicollector inductively coupled plasma-mass spectrometry (MC-ICP-MS) are described in Schoenberg et al. (2000). For the chromite separates of the Great Dyke, all Os isotope determinations were performed in static mode on Faraday collectors. For the Chilimanzi suite of granitoids, ion beams were measured on three parallel multipliers using ion counting detection and a two-cycle peak-jumping mode (Schoenberg et al. 2000). External reproducibility of the <sup>187</sup>Os/<sup>188</sup>Os ratio of an ICP Os standard solution during the course of this study was 0.07% ( $2\sigma$ ;  $n = 10$ ) for static measurements on Faraday collectors and 0.16% ( $2\sigma$ ;  $n = 5$ ) for ion counting. Os concentration and the <sup>187</sup>Os/<sup>188</sup>Os ratio were determined during the same run on spiked sample aliquots. Analytical blanks were 0.3 pg for Os and 0.5 pg for Re.

Microprobe data on unleached chromite samples were collected on a Cameca SX-50 microprobe using a wavelength-dispersive system (15 kV, 20 nA) and a calibration with standards of well-known composition. Ti, V, and Zn were measured for 30 s on peak and background, and all other elements

for 20 s. Oxide correction was performed assuming ideal unit cell structure for spinel crystals ( $X^{2+}Y_2^{3+}O_4$ ) and charge balance.

## Results

**Chemistry and Re-Os Isotope Data of Great Dyke Chromitites.** Electron microprobe analytical data for chromite samples from the 10 seams are accessible from *The Journal of Geology's* Data Depository free of charge upon request. Trends for Ti and Ni contents are shown in figure 2A. Ti shows a slight but steady increase with stratigraphic height, whereas the compatible element Ni decreases upward through the sequence. Fe and Cr (not shown) behave as Ti and Ni, respectively. Such trends are qualitatively expected in a fractionating mafic magma, and there is thus no indication that a change in magma source occurred, or that magmas from different sources contributed in varying proportions to the Great Dyke.

Re and Os contents of chromite separates from the Great Dyke's massive chromitite seams range from 0.167 to 1.46 ppb and from 10.6 to 462 ppb, respectively (table 1). Apart from some high concentrations in the two uppermost seams, there is

**Table 1.** Re-Os Isotope Data of Chromites

Sample	Re (ppt)	Os (ppb)	$^{187}\text{Re}/^{188}\text{Os}$	$^{187}\text{Os}/^{188}\text{Os}$	$\pm 2$ SE	$^{187}\text{Os}/^{188}\text{Os}$ (2.58 Ga)
5/Sb-1 <sup>a, b</sup>	1460	462	.0153	.11230	.00040	.1116
5/Sb-1 <sup>a, b</sup>	380	133	.0138	.11210	.00020	.1115
5/Sb-1 <sup>a</sup>	970	124	.038	.11200	.00020	.1103
5/Sb-1 <sup>b</sup>	337	138	.0119	.11180	.00010	.1113
6/Da-1 <sup>b</sup>	209	30.3	.0335	.11339	.00010	.1119
2/Sl-2	1559	31.4	.2416	.11309	.00010	.1025
2/Sl-2	371	28.9	.2306	.11261	.00046	.1025
3/Sb-2 <sup>b</sup>	259	55.6	.0227	.11256	.00010	.1116
6/Da-2 <sup>b</sup>	183	41.5	.0215	.11307	.00020	.1121
6/Da-2 <sup>b</sup>	189	42.5	.0216	.11304	.00010	.1121
3/We-3 <sup>b</sup>	482	22.8	.1029	.11512	.00032	.1106
3/Sb-4 <sup>b</sup>	408	20.4	.0975	.11677	.00019	.1125
6/Da-4	253	17.7	.0696	.11805	.00010	.1150
4/Sb-5 <sup>b</sup>	358	26.0	.0669	.11529	.00010	.1123
4/Sb-5 <sup>b</sup>	203	31.6	.0312	.11361	.00010	.1122
7/Da-6	469	47.1	.0484	.11204	.00025	.1099
7/Da-6	754	50.3	.0729	.11211	.00020	.1089
7/Da-6 <sup>b</sup>	220	34.3	.0312	.11326	.00010	.1119
7/Da-7 <sup>b</sup>	332	12.0	.1345	.11796	.00160	.1120
7/Da-7 <sup>b</sup>	279	10.6	.1285	.11768	.00010	.1120
8/Da-8	949	34.4	.1341	.11379	.00027	.1079
8/Da-8	699	40.1	.0849	.11260	.00039	.1089
8/Da-8a <sup>b</sup>	323	27.1	.0578	.11410	.00021	.1116
9/Da-9 <sup>b</sup>	206	27.6	.0363	.11318	.00010	.1116
8/Da-9 <sup>b</sup>	233	21.0	.0539	.11353	.00010	.1112
8/Da-10 <sup>b</sup>	167	31.1	.0261	.11377	.00010	.1126

Note. First figure of sample number indicates locality (fig. 1). Subchambers: Sb = Sebakwe, Sl = Selukwe, Da = Darwendale, We = Wedza. Last figure indicates seam (1 = highest, 10 = lowest).

<sup>a</sup> Measured by N-TIMS (all others by multicollector ICP-MS).

<sup>b</sup> Included in regression (filled symbols in fig. 3A).

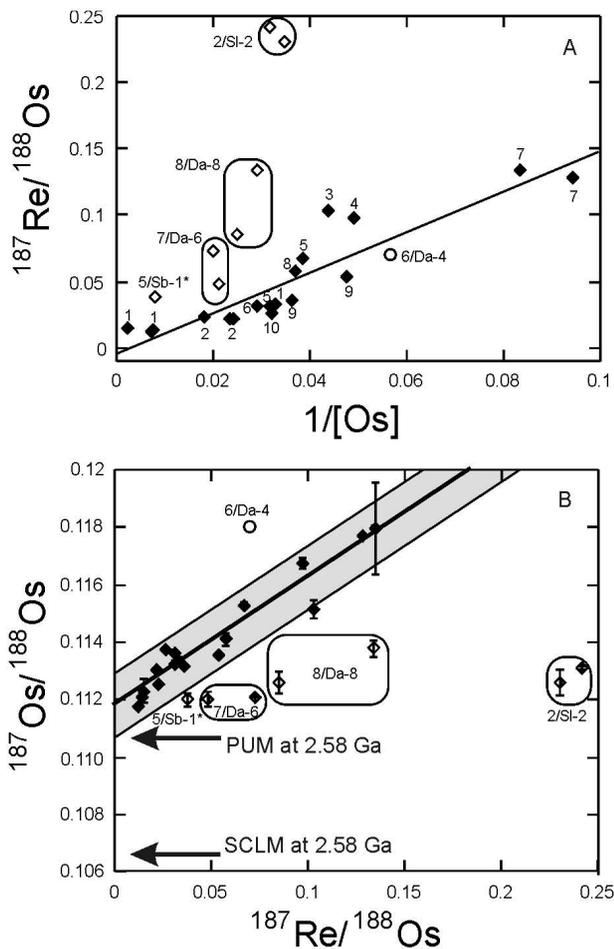
no discernible trend of Re and Os content with stratigraphic height (fig. 2B, 2C). Relatively large variations in Re and Os contents (and Re/Os ratios) were found in duplicate analyses of chromite splits from the same sample (e.g., seam 1). Such variations in chromite samples have been documented before (Marcantonio et al. 1993) and are thought to reflect microinclusions of PGE alloys. Germann and Schmidt (1999) observed tiny inclusions (lowest to submicron scale) of PGE alloys in chromite grains from the Great Dyke by high-resolution scanning electron microscope imaging. In  $^{187}\text{Re}/^{188}\text{Os}$  versus  $1/\text{Os}$  space (fig. 3A), most of the data define a positive linear trend with intersections close to the origin. This might reflect variable mixtures of two mineral phases with different Re/Os ratios, in accord with the observation of Germann and Schmidt (1999).

As expected for chromite, the  $^{187}\text{Re}/^{188}\text{Os}$  ratios are low (0.0119 to 0.1345), but the correction of Os isotope ratios for 2.58-Ga ingrowth of  $^{187}\text{Os}$  is nevertheless significant. For all duplicated sample splits, some of which indicate a very variable Os concentration and Re/Os ratio, the initial  $^{187}\text{Os}/^{188}\text{Os}$  ratios are reproduced. It is therefore concluded that the PGE alloy inclusions and host chromites

are cogenetic and contemporaneously crystallized at the time of intrusion of the Great Dyke. There is no trend of initial  $^{187}\text{Os}/^{188}\text{Os}$  with stratigraphic height (fig. 2D), nor is there a correlation with geographic location (table 1). In the  $^{187}\text{Os}/^{188}\text{Os}$  versus  $^{187}\text{Re}/^{188}\text{Os}$  plot (fig. 3B), 18 of the 26 data points fall on an array, which yields an apparent errorchron age of  $2.58 \pm 0.5$  Ga (MSWD = 57). The average initial ratio for this errorchron is 0.1118, while the range of individual initial ratios is 0.1106 to 0.1126 (table 1). Eight of the data points deviate from this main array, with seven plotting below and one above it (fig. 3B).

Sample 6/Da-4, which deviates above the main array in figure 3B, does not significantly deviate in figure 3A. This sample has the highest  $^{187}\text{Os}/^{188}\text{Os}$  ratio measured, whereas its  $^{187}\text{Re}/^{188}\text{Os}$  ratio is within the range of the other samples with similar Os concentrations (fig. 3A). It is difficult to determine whether the anomalously high initial  $^{187}\text{Os}/^{188}\text{Os}$  ratio of 0.1150 of sample 6/Da-4 is the result of Re loss, incomplete dissolution of PGM microinclusions or some other analytical artifact, since no replicate analysis of this chromite split was measured. In any case it is a lone outlier.

All other samples deviating from the main  $^{187}\text{Os}/$



**Figure 3.** A,  $^{187}\text{Re}/^{188}\text{Os}$  versus  $1/[\text{Os}]$  (ppb) plot of analyzed chromite separates. Most samples (filled symbols) define a main array probably reflecting mixing of phases with different Re/Os ratios. B,  $^{187}\text{Os}/^{188}\text{Os}$  versus  $^{187}\text{Re}/^{188}\text{Os}$  evolution diagram. Filled symbols define "errorchron" date of  $2.58 \pm 0.5$  Ga (MSWD = 57). For discussion of open symbols, see text.  $^{187}\text{Os}/^{188}\text{Os}$  ratios of primitive upper mantle (PUM; Meisel et al. 2001) and subcontinental lithospheric mantle (SCLM; Nägler et al. 1997) at the time of intrusion of the Great Dyke magma shown for comparison.

$^{188}\text{Os}$  versus  $^{187}\text{Re}/^{188}\text{Os}$  array plot below it. Although their present-day  $^{187}\text{Os}/^{188}\text{Os}$  ratios are well within the range of the general data population, their age-corrected  $^{187}\text{Os}/^{188}\text{Os}$  ratios are in part impossibly low (down to 0.1025). Two of these samples (8/Da-8 and 2/SI-2, both in duplicate) have relatively high Re/Os ratios. Together with sample 7/Da-6 and 5/Sb-1 they define an almost horizontal array with an  $^{187}\text{Os}/^{188}\text{Os}$  intercept of 0.112, in the range of initial  $^{187}\text{Os}/^{188}\text{Os}$  ratios of the main array. In the  $^{187}\text{Re}/^{188}\text{Os}$  versus  $1/[\text{Os}]$  plot (fig. 3A)

these samples deviate upward from the main array. Such features may be best explained as a result of a (sub-) recent Re gain. Scanning electron microscopy of unleached splits of chromite separates revealed that some have cracks containing partly serpentinized olivine and orthopyroxene. Re is highly mobile in the near-surface environment (Frei et al. 1998; Peucker-Ehrenbrinck and Blum 1998), and Re addition could have resulted from fluid circulation most probably affecting the silicates that fill cracks rather than the chromite crystals themselves. This view is supported by further sample data: the two outliers of sample 7/Da-6 belonged to the first batch of three digests to which the HCl pre-leach step was not applied. Data for a third, pre-leached split of sample 7/Da-6 lie within the trend of the other chromite samples in the isochron plot. The two outliers of sample 8/Da-8 were sampled at a surface outcrop, whereas sample 8/Da-8a, which plots in the middle of the main array, was sampled from the mined seam, approximately 5 m below the surface. Although no exact criteria can be developed for identification of outliers, the discussed data points plotting below the array are not considered in further discussion.

**Re-Os Data for Gneisses and Granitoids of the Zimbabwe Craton.** Re and Os concentrations and  $^{187}\text{Os}/^{188}\text{Os}$  ratios of the granitoids and gneisses are listed in table 2. Os concentrations of these samples (3–9 ppt) are much lower than the values proposed for a crustal average (53 ppt: Esser and Turekian 1993; 31 ppt: Peucker-Ehrenbrinck and Jahn 1999). Further, the Os is surprisingly unradiogenic with  $^{187}\text{Os}/^{188}\text{Os}$  ratios ranging from 0.167 to 0.297 with the exception of sample 97/T022-A, which has an  $^{187}\text{Os}/^{188}\text{Os}$  of 1.008, and there is no significant difference between the Chilimanzi granites (late Archean crustal remelts; Hickman 1978) and the ancient gneisses, thus showing no apparent Re/Os fractionation in intracrustal melting. The elevated  $^{187}\text{Re}/^{188}\text{Os}$  ratios of the samples cannot be a primary feature, as their model (mantle derivation) ages vary from 744 to 11 Ma. Thus these rocks cannot have acted as closed systems for Re-Os systematics. Inconsistently high  $^{187}\text{Re}/^{188}\text{Os}$  ratios of similar magnitude have been reported in previous studies (Esser and Turekian 1993; Burton et al. 2000). They could in principle result from sub-recent loss of Os, or addition of Re (as discussed previously), or both. Loss of Os cannot be a priori excluded but is unlikely in fresh rock samples in which sulfide grains are intact. Further, the model ages correlate inversely with the highly variable Re concentrations of the samples, rather than with their much more uniform Os contents (table 2), and

**Table 2.** Localities and Summary Descriptions and Re-Os Data of Gneiss and Granitoid Samples

Sample	Locality	Description	Re (ppt)	Os (ppt)	$^{187}\text{Re}/^{188}\text{Os}$	$^{187}\text{Os}/^{188}\text{Os}$	$\pm 2$ SE	Model age (Ma)
97/T06-A	20°16'26"S/ 30°08'04"E	Mashaba tonalite; medium to coarse grained, slightly foliated tonalitic gneiss with small patches of leucosome crossing the foliation	26.2	9	14.2	.2811	.0020	670
97/Y018	19°40'10"S/ 30°45'36"E	Chilimanzi granite; no obvious deformation, homogeneous and medium-grained (~5 mm) matrix, large (up to 20 mm) K-feldspar crystals	29.3	3.1	45.5	.2487	.0003	162
97/Y019	19°37'58"S/ 30°58'59"E	Chilimanzi granite, as previous	42.5	5.2	40.1	.1675	.0011	62
97/T022-A	20°12'52"S/ 30°14'01"E	Mafic enclave in Shabani tonalitic gneiss; homogeneous, slightly foliated	6788	6.6	5011	1.008	.012	11
97/T032	20°03'09"S/ 30°36'20"E	Mushandike granodiorite; medium grained (~5 mm) and unfoliated	301	4.2	353	.2966	.0028	29

Note. Re contents corrected for 0.5-pg analytical blank. Os contents and isotope composition corrected for 0.3-pg analytical blank with  $^{187}\text{Os}/^{188}\text{Os}$  ratio of 0.15 (measured from larger quantities of reagents).

if the extremely young model age of sample 97/T022-A were entirely due to loss of Os, this rock would originally have had at least 1.6 ppb Os, which is higher than the Os concentration expected from a mantle melt and thus highly unlikely. We therefore consider that subrecent Re enrichment is the primary cause for the impossibly young model ages.

The increase in Re/Os ratios, although subrecent in geological terms, could have affected the  $^{187}\text{Os}/^{188}\text{Os}$  ratios of the rocks; in particular, in sample 97/T022-A with its very low model age of 11 Ma, most of the radiogenic  $^{187}\text{Os}$  might have ingrown since the time of Re-addition. Thus the measured  $^{187}\text{Os}/^{188}\text{Os}$  ratios of all samples must be regarded as maxima for the true Os-isotope composition prior to Re/Os increase.

Nevertheless, the data can be used to place a limit on the possible Os budget of Zimbabwe cratonic crust at the time of intrusion of the Great Dyke. Samples 97/T06-A, T022-A, and T032 are from a region where Nd model ages are about 3.5 Ga (Taylor et al. 1991). Zircon U-Pb evidence of ca. 3.5-Ga continental crust abounds (Horstwood et al. 1999). The Chilimanzi granites (samples 97/Y018 and Y019), although intruded at around 2.6 Ga, are clearly derived from much older crust (Hickman 1978; Berger et al. 1995). Assuming a mantle derivation age of 3.5 Ga, maximum  $^{187}\text{Os}/^{188}\text{Os}$  ratios of these rocks at 2.58 Ga are found by interpolating between a mantle  $^{187}\text{Os}/^{188}\text{Os}$  ratio at 3.5 Ga (0.10183) and the present-day, measured values (ta-

ble 2). An average value of 0.14 is obtained when the obviously unrealistic data point for 97/T022-A is disregarded.

## Discussion

**Re-Os Isotope Systematics and Their Constraints on Crustal Contamination.** The range of initial  $^{187}\text{Os}/^{188}\text{Os}$  ratios of the chromites defining the error-chron in figure 3A is slightly above the value of the chondritic model mantle at 2.58 Ga. The lack of correlation with stratigraphic level (fig. 2D) is in strong contrast to observations in the Bushveld Complex (McCandless et al. 1999; Schoenberg et al. 1999), where  $^{187}\text{Os}/^{188}\text{Os}$  ratios are seen to increase upward. The absence of systematic differences in initial Os isotope signatures between the different subchambers along the strike of the Great Dyke means that if contamination affected the  $^{187}\text{Os}/^{188}\text{Os}$  ratio of the magma, this did not vary regionally.

To place limits on the amount of required crustal contamination of the primary Great Dyke magma, a simple bulk assimilation model can be used. The limited constraints available do not warrant more sophisticated modeling. Mass balance considerations yield the well-known relation:

$$F = \frac{(R_{\text{mix}} - R_1)C_1}{(R_{\text{mix}} - R_1)C_1 + (R_2 - R_{\text{mix}})C_2}, \quad (1)$$

where  $F$  is the fraction of the contaminant in the

final mixed melt,  $R_s$  are isotope ratios,  $C_s$  are concentrations of an element, and 1 and 2 refer to the mantle melt and the bulk crustal contaminant, respectively. The admixture  $A$  of the contaminant as a fraction of the amount of original intruding melt is given by

$$A = \frac{F}{1 - F}. \quad (2)$$

To assess ranges of possible crustal contamination of the Great Dyke magma, we have used conservative estimates of parameters. The crustal contaminant at 2.58 Ga is assumed to have 10 ppt Os (the highest value found is 9 ppt; see table 2) with a  $^{187}\text{Os}/^{188}\text{Os}$  ratio of 0.14, discussed above to be the maximum estimate corresponding to a crustal age of 3.5 Ga. The mantle melt was assigned an Os concentration of  $\sim 50$  ppt (a low estimate for a melt resulting from 5%–10% mantle melting; Schoenberg et al. 1999). If the primary melt is assumed to be derived from a primitive upper mantle (PUM) source with a  $^{187}\text{Os}/^{188}\text{Os}$  ratio of 0.1107 at 2.58 Ga (from parameters of Meisel et al. 2001), the amount of crustal admixture  $A$  to the intruding melt needed to produce the initial Os signatures of the Great Dyke chromitite seams would vary between 0% and 33%. For a magma from the SCLM ( $^{187}\text{Os}/^{188}\text{Os}$  at 2.58 Ga = 0.1065; Nägler et al. 1997), the corresponding figures are 66%–100%. An exception is sample 6/Da-4, which would appear to reflect about 80% crustal admixture for a PUM source or 160% for an SCLM source. Because of the conservative parameter values used, all these calculated admixtures are minimum estimates. Before drawing conclusions from this, it is of interest to compare these results with contamination as calculated from other isotope systematics.

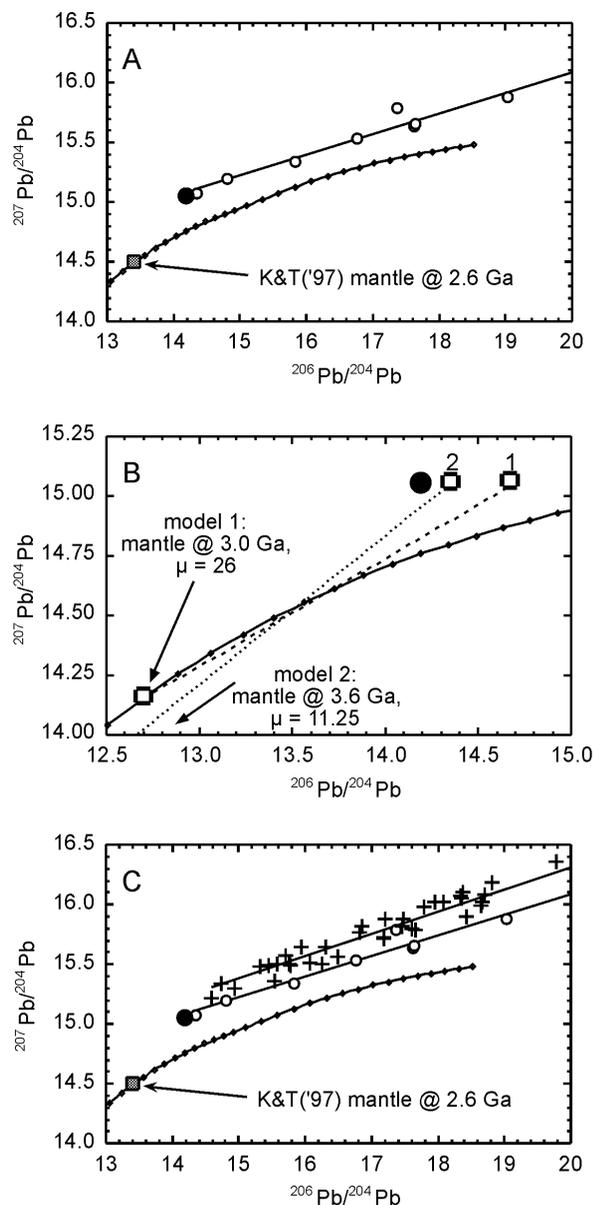
**Crustal Contamination Estimates from Other Data.** Nd, Sr, and Pb isotope data are available for the Great Dyke (Mukasa et al. 1998; Oberthür et al. 2002). The enrichment of Pb in the continental crust relative to the depleted mantle is ca. 100-fold (Taylor and McLennan 1995); that of Nd and Sr is ca. 40-fold. Nd is the most reliable indicator, as its isotopic evolution in crustal rocks is reasonably predictable. The use of Pb as an indicator, while sensitive, is often impaired by crustal inhomogeneity and uncertainty of average crustal isotope compositions. Sr has both unpredictable crustal evolution patterns and is readily mobilized in fluids. We consider Nd and Pb data alone.

Nd isotope systematics of Great Dyke rocks (Mukasa et al. 1998; Oberthür et al. 2002) yield reliable  $\epsilon_{\text{Nd}}$  values between  $-2$  and  $+2.2$  for  $T = 2.58$  Ga.

The data of Hahn et al. (1991) on the West (Umvimeela) Dyke yield  $\epsilon_{\text{Nd}}$  (2.58 Ga) values of 3, 4, and 4.2, which are even higher than the average depleted mantle  $\epsilon_{\text{Nd}}$  value of Nägler and Kramers (1998;  $+2.5$  at 2.58 Ga). As the West and East Dyke gabbros are considered to reflect the parental magma of the Great Dyke, this clearly shows that the magma source itself was not crustally contaminated. The amount of crustal contamination within the Great Dyke, as reflected by these results, can be estimated using Nd isotope data from the Zimbabwe Craton. Six gneiss suites intruded by the Great Dyke or close to it (Rhodesdale, Chingezi, Tokwe, Shabani, Mont d'Or, and Mushandike) have  $\epsilon_{\text{Nd}}$  (2.58 Ga) values ranging from  $-2.6$  to  $-12$ , averaging  $-8$  (Taylor et al. 1991). The average Nd concentration of these suites is 20 ppm, close to the average Archean crustal Nd concentration of 25 ppm proposed by Taylor and McLennan (1995). A magma derived by 10% melting of average upper mantle has about 8 ppm Nd. If an  $\epsilon_{\text{Nd}}$  value of  $+2.5$  is assumed for the depleted mantle at 2.58 Ga, these parameters yield Nd crustal contamination for the Great Dyke ( $A$  values; see eqq. [1], [2]) between 8% and 19% with a single outlier at 49%. If a source magma  $\epsilon_{\text{Nd}}$  value of  $+4$  is assumed (following the data of Hahn et al. 1991), crustal contamination estimates between 15% and 26%, with an outlier at 64%, result. The outlier is a quartz gabbro sample, considered contaminated on petrological grounds (Wilson and Prendergast 1989; Oberthür et al. 2002).

The whole rock and mineral Pb isotope data of Mukasa et al. (1998) yield a  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  isochron age of  $2596 \pm 14$  Ma (fig. 4A). This is close to the more precise zircon and rutile U-Pb ages, indicating that post-emplacement disturbance of U-Pb systematics was not very significant. As noted by Mukasa et al. (1998), this isochron lies above the Pb growth curve of Stacey and Kramers (1975) in  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  space. It is well above the Pb evolution model curve for the upper mantle of Kramers and Tolstikhin (1997). Mukasa et al. (1998) have suggested that these elevated  $^{207}\text{Pb}/^{204}\text{Pb}$  ratios reflect an ancient enriched mantle source for the Great Dyke magma. Figure 4B shows two possible scenarios experienced by "normal" mantle, starting with the isotope composition of the Kramers and Tolstikhin (1997) model: the U/Pb ratio of the mantle source would have to be about  $3.5 \times$  that of average mantle if the enrichment dated from 3 Ga, and  $1.5 \times$  if it dated from 3.6 Ga.

High initial  $^{207}\text{Pb}/^{204}\text{Pb}$  ratios are not a characteristic of Archean mantle-derived volcanics (Kra-



**Figure 4.** Pb isotope systematics of the Great Dyke and Zimbabwe Craton. *A*, Whole rock and mineral Pb isotope data of Mukasa et al. (1998) (*open circles*) yielding a  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  age of  $2596 \pm 14$  Ma, compared with upper mantle development curve of Kramers and Tolstikhin (1997), showing their elevated  $^{207}\text{Pb}/^{204}\text{Pb}$  ratios. *Closed circle* = maximum possible initial ratios of Mukasa et al.'s (1998) sample suite. *B*, Two putative scenarios of "enriched mantle" dating to 3.0 or to 3.6 Ga, yielding initial Pb close to (but still more radiogenic than) the initial ratio of *A*.  $\mu$  denotes the  $^{238}\text{U}/^{204}\text{Pb}$  ratio of the reservoirs. *C*, Great Dyke data compared with back-corrected Pb isotope data from the North Marginal Zone of the Limpopo Belt (*crosses*; Berger and Rollinson). Pb data from the Zimbabwe Craton (Taylor et al. 1991) portray similarly high  $^{207}\text{Pb}/^{204}\text{Pb}$  ratios.

mers and Tolstikhin 1997). However, they are a typical regional feature of the Zimbabwe Craton (Taylor et al. 1991). Also, Pb isotope ratios from the Northern Marginal Zone of the Limpopo Belt, which is probably a high-grade metamorphic equivalent of the Zimbabwe Craton, form an array with anomalously high  $^{207}\text{Pb}/^{204}\text{Pb}$  values (Berger and Rollinson 1997). In figure 4C, these arrays and their likely initial ratios are shown and compared with the model upper mantle of Kramers and Tolstikhin (1997) at 2.6 Ga. From these data, the amount of crustal contamination can be assessed, if it is assumed that the Great Dyke magma was derived from a model upper mantle. The average Pb concentration of continental crust is in the range of 10–20 ppm (Taylor and McLennan 1995; Wedepohl 1995). That of a mafic melt derived by 10% mantle melting in the Archean is around 1 ppm. Given these concentration ranges and the Pb isotope ratios for mantle, crust, and Great Dyke magma at 2.6 Ga shown in figure 4C, the amount of crustal contamination (*A* values; see eqq. [1], [2]) needed to produce the observed pattern is between 5% and 25%.

In summary, the crustal contamination estimates for the Great Dyke magma obtained from Pb and Nd systematics are in reasonable agreement with each other, and on the whole do not appear excessive from the petrography. The samples studied by Mukasa et al. (1998) are websterites and (mainly) gabbros and norites from a stratigraphic level well above that of the chromitite seams, which are hosted by dunites to pyroxenites. The Nd data from Oberthür et al. (2002) are mainly from pyroxenites at the chromite seam levels, and gabbros. If crustal contamination progressed during differentiation, then lower stratigraphic levels would be expected to be less contaminated than higher ones (as is seen in the Bushveld Complex; Schoenberg et al. 1999). Crustal contamination at the chromite seam level should thus be within the lower part of the range found from their combined data sets.

Contrary to this expectation, the range of contamination levels (0%–33%) calculated from the Os data for a PUM-sourced primary Great Dyke magma extends to higher values than that derived above from Nd and Pb data. Bearing in mind that the contamination estimates from the Os data are minima, a homogeneous PUM-type source for the Great Dyke magma is unlikely. The Os data can be reconciled with the Nd and Pb results if it is assumed that the mantle was slightly heterogeneous with respect to  $^{187}\text{Os}/^{188}\text{Os}$  and that the Great Dyke magmas originated from mantle portions

with slightly more radiogenic Os than PUM. Indeed, work on present-day plumes has revealed such heterogeneity. Some high- $\mu$ -sourced basalts have significantly elevated  $^{187}\text{Os}/^{188}\text{Os}$  ratios, attributed to recycled oceanic crust (Hauri and Hart 1993). Basalt provinces considered to be generated from plume heads also show heterogeneous  $^{187}\text{Os}/^{188}\text{Os}$  values ranging slightly above those of PUM, and combined with high Os concentrations, are features that cannot be readily explained by crustal contamination or recycled oceanic crust and may reveal inherent  $^{187}\text{Os}/^{188}\text{Os}$  heterogeneity in the deep mantle (Brooks et al. 1999; Walker et al. 1999; Schaefer et al. 2000).

A more definite conclusion from the calculations is that the crustal admixture range obtained for an SCLM-sourced magma (66%–100%) is in total contradiction with the Nd- and Pb-derived results. It is also unreasonable from a petrographic point of view. Hence, any significant SCLM contribution to the Great Dyke primary magma would aggravate the discrepancy already existing for a PUM source. Thus the SCLM cannot have contributed significantly to the Great Dyke magma.

***How Can the Absence of a Significant Subcontinental Lithospheric Mantle Component Be Explained?*** The absence of an SCLM Os isotope signature in the Great Dyke magmas is somewhat surprising because this reservoir has significantly contaminated (or sourced) ultramafic magmas intruding greenstone belts in the Zimbabwe Craton from 3.5 to about 2.7 Ga (Nägler et al. 1997), although not 2.7-Ga komatiites in the Belingwe Greenstone Belt (Walker and Nisbet 2002). As discussed in "Introduction," Os isotopes are a highly sensitive indicator of such contamination. One possible solution to the problem of lack of SCLM contamination is to suggest that the SCLM below the Zimbabwe Craton was removed by delamination before the Great Dyke intrusion. Frei et al. (1999) and Kamber and Biino (1995) have pointed out that such a process could have provided the heat necessary to form the Chilimanzi suite of granites. However, there are several reasons why removal of the SCLM from below the Zimbabwe Craton at ca. 2.6 Ga is improbable.

First, there is geophysical evidence that a lithospheric keel exists underneath the Zimbabwe and Kaapvaal Cratons today. Seismic tomography shows a high-velocity zone extending to a depth greater than 200 km (James et al. 2001) and also, today's heat flow values for both Cratons are low compared with the surrounding Proterozoic mobile belts (Jones 1998). This author interpreted this as a result of a low heat flux to the base of the cratonic

crust, caused by diversion of the advective heat in the mantle to the edges of the cratons. Such a diversion of advective heat in the mantle requires a refractory heat shield at the base of the cratons, which can only be an SCLM.

Second, thermal modeling of the Zimbabwe Craton with heat production calculated for 2.6 Ga shows that this craton could have been solid to the base of the crust at this time only if the steady-state basal heat flux was approximately 30 mW/m<sup>2</sup> or lower (Kramers et al. 2001). As total mantle heat production at 2.6 Ga was about twice today's value (and today, 20 mW/m<sup>2</sup> is considered an anomalously low mantle heat flux), this certainly implies the presence of the SCLM as a heat shield at that time. Some thermal pulse must have led to the generation of the Chilimanzi suite magmas, but the new dates for the Great Dyke at 2576 Ma indicate cooling of the lower crust of the Zimbabwe Craton from supersolidus conditions in the time range 2.62–2.58 Ga (when the Chilimanzi granite melts were produced) to conditions allowing the formation of crustal-scale fractures at 2.576 Ga. If the thermal pulse was caused by delamination of the SCLM, the convecting asthenospheric mantle would then have provided a heat source of a depth scale probably in excess of 100 km to the base of the Zimbabwe Craton. Thermal equilibration over such a depth scale would require >300 m.yr. and thus such rapid cooling would not have been possible.

We therefore conclude that the Great Dyke magma intruded through the SCLM of the Zimbabwe Craton without being significantly contaminated by it. This has been shown for other large layered complexes: initial Os signatures of the Bushveld and Stillwater Complexes (Marcantonio et al. 1993; McCandless et al. 1999; Schoenberg et al. 1999) also do not reveal involvement of SCLM to their magma sources, although the existence of such keels underneath the Kaapvaal and Wyoming Cratons at the time of their intrusions is well documented (O'Brien et al. 1995; Pearson 1999). Hence, it is suggested that large magma volumes such as those of layered intrusions are able to ascend through SCLM keels without melting significant amounts of rock from them.

***Implications for the Geotectonic Setting.*** The thickness of the SCLM keel underneath southern Africa is reliably estimated from various lines of evidence to be >200 km (Boyd 1989; Jones 1998; James et al. 2001; Nguuri et al. 2001). A direct conclusion from the previous data and argumentation is that the Great Dyke magma must have originated at a depth below this range. Since slab

dehydration in subduction zones occurs at depths less than about 100 km, this effectively precludes an origin of the Great Dyke magma in a subduction zone setting as proposed by Mukasa et al. (1998).

An alternative hypothesis is that the Great Dyke resulted from a large mantle plume in a process comparable to failed rifting. In this context, the almost coeval emplacement of the large suite of Chilimanzi granites could be entirely coincidental or might have been caused by the spreading of early Great Dyke—related magma at the base of the crust (underplating). As discussed earlier, thermal modeling for 2.6 Ga of the Zimbabwe Craton (Kramers et al. 2001) suggests that its crustal base was close to the vapor-absent tonalite solidus at 2.6 Ga if the basal heat flux was about 30 mW/m<sup>2</sup>, and a thermal pulse caused by underplating could have caused widespread partial melting in the lower crust in a relatively short time. The resulting Chilimanzi magmas, spreading out in large sill-like intrusions at shallow depth, could then have transported heat (and heat-producing elements) rapidly to shallow crustal levels. Subsequently, tension fractures could have developed in the craton as a result of the thermal bulge, coinciding in time with further plume magma intrusion forming the Great Dyke and its satellites. The lack of apparent SCLM contamination could be explained either by the very large magma volume, or by postulating intrusion through the SCLM following the same conduits used by earlier magma pulses. Crustal contamination is expected to result from roof melting rather than conduit walls, so that its effect should be more noticeable in the main intrusion than in the satellites. This could account for the uncontaminated Nd signature of West Dyke magma found by Hahn et al. (1991). Assuming a plume source could also imply some heterogeneity in source <sup>187</sup>Os/<sup>188</sup>Os ratios, with values somewhat above those of PUM as observed in present-day large plumes. This would reduce the amount of crustal contamination required as discussed previously.

**Contrast with the Bushveld Complex.** The contrasting Os isotope systematics of the Great Dyke and the 2.0-Ga Bushveld intrusive complex (McCandless et al. 1999; Schoenberg et al. 1999) need to be discussed for two reasons. First, crustal contamination is clearly apparent in the Os isotope signatures of the latter, with <sup>187</sup>Os/<sup>188</sup>Os ratios increasing upward in the stratigraphy from one chromite seam to the next and reaching ~0.175 in the Merensky Reef. Second, this effect was successfully modeled by Schoenberg et al. (1999) using average crustal Os concentrations and Re/Os ratios of Esser

and Turekian (1993). As shown earlier, Esser and Turekian's (1993) average crustal Os concentrations and Re/Os ratios do not apply to rocks of the Archean Zimbabwe Craton and may also not approximate the Archean crust of the Kaapvaal Craton. Our unpublished data of extremely fresh granitoid samples from the eastern Kaapvaal Craton reveal similarly low Os concentrations and unradiogenic <sup>187</sup>Os/<sup>188</sup>Os ratios to those found in the Zimbabwe Craton. Using such data, the Os isotope systematics of the Bushveld chromite sequence cannot be reconciled with crustal contamination, even taking into account that it is considerably younger than the Great Dyke.

Nevertheless, there is no other known way than crustal contamination to raise the <sup>187</sup>Os/<sup>188</sup>Os ratios of mantle-derived melts. A solution may lie in the difference in geological settings. Unlike the Great Dyke, the Bushveld Complex is intruded into a vast sedimentary complex, the Early Proterozoic Transvaal Supergroup, which contains abundant shale sequences. For the present earth, Peucker-Ehrenbrinck (2000) has emphasized the disproportionate importance of black shales as a reservoir of radiogenic Os. If Re and Os were already similarly concentrated in the Early Proterozoic black shales of the Transvaal Supergroup, the progressively more radiogenic Os contamination of the Bushveld chromitite seams could be explained in this way. Importantly, the scarcity of black shales in Archean times excludes any contamination with very radiogenic <sup>187</sup>Os/<sup>188</sup>Os from these sediments in the case of the Zimbabwe Craton and the Great Dyke.

### Conclusions

Chromite samples from the Great Dyke have initial <sup>187</sup>Os/<sup>188</sup>Os ratios in the range of 0.1106–0.1126, which is above the value of the PUM at 2.58 Ga (0.1107) and far above that inferred for the SCLM at that time (0.1065). There is no geographic or stratigraphic correlation of the samples with their Os isotope composition.

Gneiss and granitoid samples from the Zimbabwe Craton have extremely low Os concentrations (3–9 ppt). This Os is also surprisingly unradiogenic (<sup>187</sup>Os/<sup>188</sup>Os = 0.16 to 1), and the Os concentration and <sup>187</sup>Os/<sup>188</sup>Os ratio of Zimbabwe Craton crust are not quite sufficient to account for the Great Dyke chromite <sup>187</sup>Os/<sup>188</sup>Os ratios by crustal contamination, if the mantle source of the Great Dyke magma had a PUM Re/Os ratio. It is suggested that a reservoir within a heterogeneous mantle, which had an Re/Os ratio on average some-

what higher than PUM, acted as source to the Great Dyke magma. Certainly, any significant contribution to the Great Dyke magma by SCLM is excluded by the data.

Thermal considerations show that the SCLM was in place at the time of formation of the Great Dyke, and the data exclude a subduction zone or continent collision setting for generating the Great Dyke magma because in this case, magma would have been generated within the SCLM depth range. It is proposed instead that the Great Dyke was the result of a large mantle upwelling, or "plume," akin to a failed rift setting, and that the magma escaped SCLM contamination either by sheer volume or by propagation in conduits already lined by previous intrusions. At the present stage, there still is no

proof for a relationship between the heat source generating the Chilimanzi and Razi suites of granites and the source of the Great Dyke melts.

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