Geol. Mag. **138** (*1*), *2001, pp. 31–38. Printed in the United Kingdom © 2001 Cambridge University Press*

The Mushandike granite: further evidence for 3.4 Ga magmatism in the Zimbabwe craton

M. H. DODSON*†, I. S. WILLIAMS* & J. D. KRAMERS‡

*Research School of Earth Sciences, The Australian National University, Canberra ACT 0200, Australia
‡Department of Geology, University of Zimbabwe, P.O. Box M.P.167, Mount Pleasant, Harare, Zimbabwe

(*Received* 29 August 2000; *accepted* 18 September 2000)

Abstract – The conflict between independently published ages for the Mushandike Granite, Zimbabwe (2.92 ± 0.17 Ga and 3.45 ± 0.13 Ga) has been resolved in favour of the older age by SHRIMP U–Pb analyses of zircon. Two samples yield indistinguishable estimates of 3374 ± 7 and 3368 ± 11 Ma for the crystallization age of the magma. Together with published data from elsewhere in southern Zimbabwe, the results imply a widespread magmatic event at about 3.35 Ga. A single zircon core giving 3.46 Ga, together with the granite's previously measured Nd model age, suggests that the Mushandike magma could have incorporated remobilized basement similar to the *c.* 3.5 Ga Tokwe gneisses which crop out 30 km to the west. The published Rb–Sr and Pb–Pb datasets show evidence of late Archaean disturbance of Sr and Pb isotope systematics. In the absence of exposed contacts between the Mushandike granite and the neighbouring Mushandike stromatolitic limestone, the new U–Pb emplacement age suggests that the limestone is unconformable on the granite.

1. Introduction

Greenstone belts in the Archaean granite–greenstone terrane of Zimbabwe span an age range from 3.5 to 2.7 Ga. Three main stages of development have been distinguished: Sebakwian (the oldest), Lower Bulawayan, and Upper Bulawayan (Wilson, 1979; Wilson *et al.* 1978; Wilson, Nesbitt & Fanning, 1995). Early fragmentary evidence that the Sebakwian was much older than the other components of the terrane (reviewed by Vail & Dodson, 1969) was supported by Hickman's (1974) Rb–Sr isochron age of 3.45 ± 0.13 Ga measured on the Mushandike granite (*sensu lato*), supposedly intruded into Sebakwian near the southern margin of the craton (Fig. 1). However, Moorbath *et al.* (1987), using both Rb–Sr and Pb–Pb whole-rock methods, obtained much lower ages of 2.92 ± 0.17 and 2.95 ± 0.13 Ga respectively, consistent with the Pb–Pb age of 2.84 Ga measured on the adjacent Mushandike stromatolitic limestone, which was considered, from field evidence, to pre-date the granite (Orpen & Wilson, 1981). Moorbath and co-workers suggested that the older age obtained by Hickman reflected chance sampling of early Archaean protolith material which was not fully homogenized during magmatism. The Nd model age of the granite, ~3.54 Ga (Moorbath, Taylor & Jones, 1986), they believed to be the age of the protolith. To resolve this conflict and to contribute to the ongoing debate about Archaean continental formation, we have determined SHRIMP U–Pb ages on zircons extracted from two samples of the Mushandike granite; our results strongly support an emplacement age close to 3.4 Ga, and are consistent with other evidence for coeval magmatism in southern Zimbabwe (e.g. Horstwood *et al.* 1999).

2. Geological setting and sample selection

The geological relationships between the Mushandike granite and adjacent rocks are obscured by poor exposure and tectonic reworking. The granite crops out mainly in the Mushandike National Park, 25 km west of Masvingo in south-central Zimbabwe (Fig. 1). It is overlain unconformably by the Gwenya Formation (mostly banded ironstones and ferruginous siltstones), which forms the base of the Upper Greenstone sequence of the Masvingo Greenstone Belt (Wilson, 1979). It is traversed by an array of shear zones of uncertain age. The Mushandike stromatolitic limestone, which is not in direct contact with the Gwenya Formation, was considered by Orpen & Wilson (1981), from field evidence, to pre-date the granite. Moorbath *et al.* (1987) determined a Pb–Pb isochron age of 2.839 ± 0.033 Ga for the limestone, which they believed to record the age of early diagenesis.

Zircon was extracted from two granite samples for the present study. MG14 was collected for Rb–Sr dating by Hickman (1974) near the northern margin of the granite [TN 515818]. The zircon proved to be of poor quality, however, and the yield was low. We there-

---

† Author for correspondence: mdodson@chevin.u-net.com; present address: School of Earth Sciences, University of Leeds, Leeds LS2 9JT, UK.
‡ Present address: Mineralogisch-Petrographisches Institut, Universität Bern, Erlachstrasse 9a, CH-3012 Bern, Switzerland.

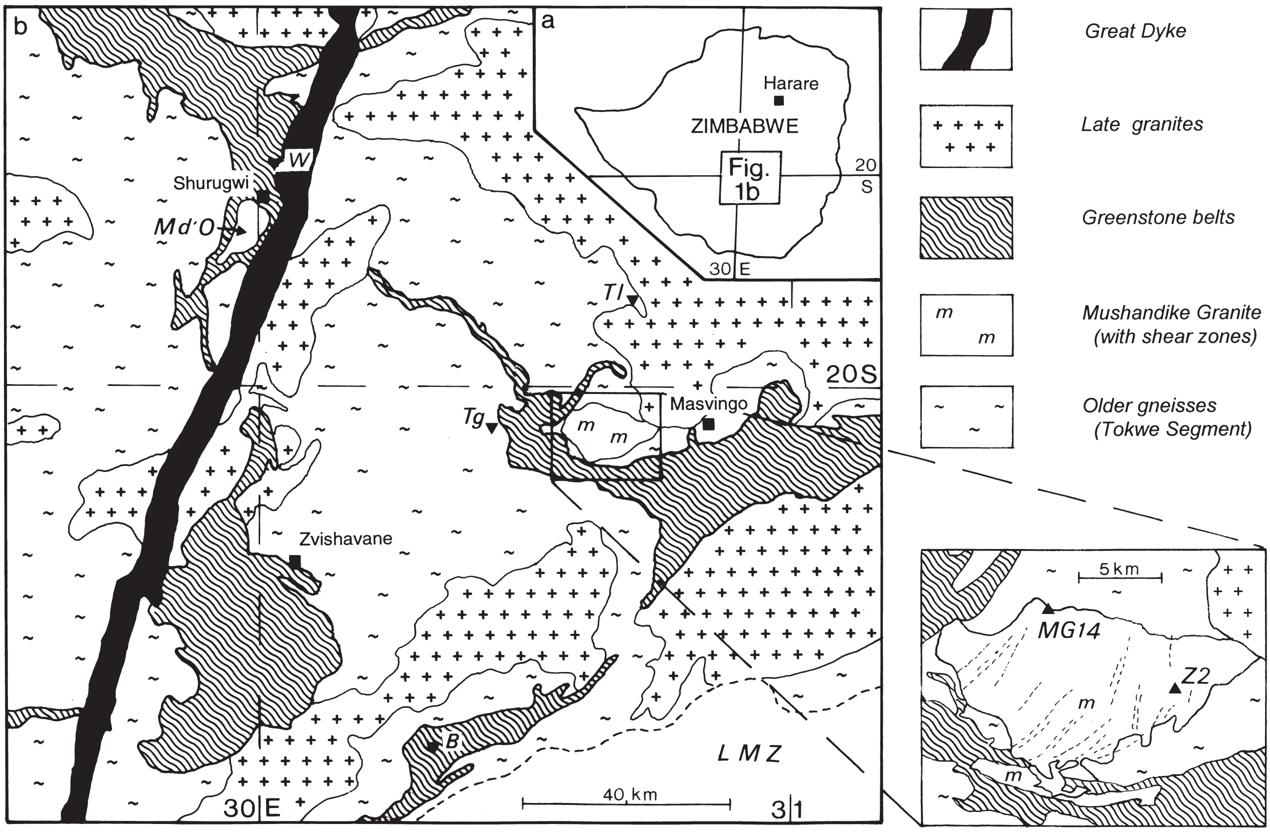

Figure 1. (a) Location of main map within Zimbabwe; (b) regional setting of Mushandike Granite (after Zimbabwe, 1985; Geological Survey, 1965). Sample locations: MG14, Z2, this paper; B – Buchwa; W – Wanderer (Dodson *et al.* 1988); Tg – Tokwe Gneiss; Tl – Tokwe leucosome (Horstwood *et al.* 1999). LMZ – Limpopo Mobile Zone. Md'O – Mont d'Or granite.

fore subsequently collected a second sample, Z2, from the interior of the pluton [TN 592773]. The granite at both localities appears to be unquestionably magmatic in character and is free of obvious xenolithic material. Neither locality is close to an exposed shear zone.

3. Analytical methods

The zircon was separated at Leeds and Canberra by standard Wilfley table and heavy liquid procedures, followed by hand-purification of the resultant heavy mineral concentrate. The grains were mounted in epoxy and polished to expose their centres. These mounts were then photographed in reflected and transmitted light under a high-power microscope to assist in targeting the ion microprobe analyses. The procedures for zircon geochronology using the ANU SHRIMP I have been described in detail elsewhere (Compston, Williams & Meyer, 1984; Williams & Claesson, 1987). In brief, a 10 kV beam of negative oxygen ions was focused to a ~ 30 μm diameter spot on the target surface, and the sputtered secondary ions transferred at 10 kV to a high (5000) mass resolution mass spectrometer, where they were counted by a single electron multiplier using sequential switching of the analyser magnet. The isotopic composition of the Pb was measured directly; inter-element fractionation was corrected by reference to standard zircon SL13, assumed to have an age of 572 Ma and U content of 238 ppm. Corrections for initial Pb were made using ^{204}Pb and, because of the large common Pb contents of some grains, an isotopic composition consistent with the age of the rock (Cummings & Richards, 1975). Isotopic data were processed by Isoplot 2.3 (Ludwig, 2000), using the decay constants recommended by Steiger & Jäger (1977).

Sites for analysis were selected on the basis of the photomicrographs. Because of the suggestion by Moorbath *et al.* (1987) that the Mushandike magma was a 2.9 Ga remelt of a 3.5 Ga protolith, a thorough search was made for datable, euhedrally zoned, melt-precipitated material. A few apparent cores also were targeted as possibly representing inheritance from an older protolith. To check this characterization, the analysed grains from sample Z2 were subsequently examined by cathodoluminescence, a technique that was not available when the isotopic work was done. The luminescence images revealed that the grains were actually dominantly melt-precipitated and that true older cores were very rare; most of the 'cores' identified from the photomicrographs were in fact crystal centres outlined by thin zones of high-U zircon made visible by their abnormal refractive index and reflectance.

Three types of luminescence response were distinguished: (1) relatively strong luminescence showing simple euhedral zoning, (2) weak luminescence showing no or weak zoning, and (3) weak luminescence showing a mottled texture. There proved to be a general, but not absolute, correlation between stronger luminescence and lower U content. Zoned zircon dominated most of the grains. Unzoned zircon occurred both within grains and as apparent overgrowths (Fig. 2). Mottled zircon formed only cores and was rare. The boundaries between the different zircon types were sharply defined, and commonly, but not always, parallel to the growth zoning. The zircon from MG14 was not suitable for cathodoluminescence imaging.

4. Zircon compositions

Isotopic analyses of 43 spots on 27 zircon crystals are listed in Table 1 and plotted on Figure 3. MG14 was analysed first as a reconnaissance. The poor quality of the zircon left few grains suitable for analysis, however, so those few were analysed using 14, rather than the normal 7, scans through the isotopes of interest to improve the analytical precision. The zircon from Z2 was of much higher quality, so was analysed using the normal seven-scan per set procedure. Uncertainty estimates for the ^{207}Pb/^{206}Pb ratios and the corresponding age ($t_{7/6}$) are based upon counting statistics and propagation of the uncertainties in common Pb correction, while those for the Pb/U ratios include in addition the estimated uncertainty in the Pb/U calibration (~1%).

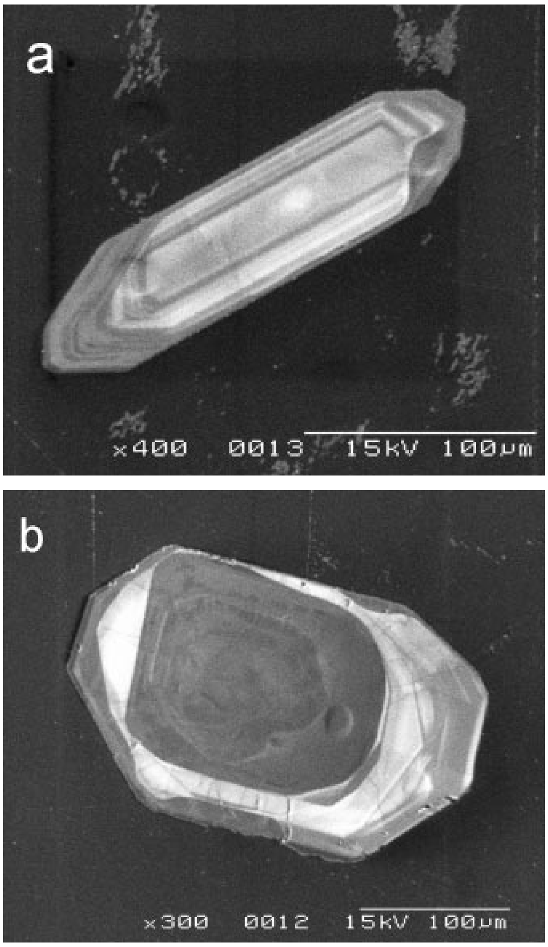

Figure 2. Cathodoluminescence images of zircon grains from sample Z2. (a) Grain 12, showing strongly luminescent euhedral zoning; (b) grain 5, showing a weakly luminescent, weakly zoned core, surrounded by broadly euhedrally zoned zircon, surrounded in turn by a partly unconformable, discontinuous, dark unzoned overgrowth.

The analyses from MG14 show a wide range in U content (38–2050 ppm) and consistently high Th/U (most > 1). There is also a very wide range in isotopic composition, but the analyses are dispersed along a discordance line (MSWD = 20), with the degree of discordance increasing with increasing U content (Fig. 3a, inset). The concordia intercepts of this line are 3390 ± 20 and 740 ± 80 Ma (95% confidence). Such extreme discordance of the highest U areas, with an apparent non-zero age of Pb loss, has been observed in other zircon studies on Archaean and early Proterozoic provinces in Southern Africa (e.g. Berger, Kramers & Nägler, 1995; Jaeckel *et al.* 1997; Kröner *et al.* 1999), in which the 'age' of projected lower intercepts ranges between 0 and 800 Ma, and has no geological significance. The mechanism of Pb loss (whether Neoproterozoic–Phanerozoic events, continuous diffusion or effects of weathering) remains a subject of speculation, but does not affect the interpretation of the upper intercept as indicating the age of zircon crystallization. Considering only the well-defined group of least discordant analyses (Fig. 3a), all have the same radiogenic ^{207}Pb/^{206}Pb within analytical uncertainty; the weighted mean value of the nine samples, 0.28196 ± 0.00057 (σ, analytical error) is equivalent to an age of 3374 ± 7 Ma (95% conf.). This is likely to be a more accurate estimate of the zircon age than the upper intercept of the regression line. None of the analyses shows any evidence for an older inherited component, so this also is the best estimate for the crystallization age of the granite.

The analyses from Z2 also show a wide range in U content (35–1550 ppm), but in general a lower Th/U (< 0.8). Like the MG14 zircon, they show a very wide range in isotopic composition and a correlation between the degree of discordance and U content. Their interpretation is assisted by reference to the different zircon types evident in the cathodoluminescence images. There is a well-defined discordance line (Fig. 3b, inset), with three exceptions. One highly discordant analysis (13–1) comes from the mottled core with the highest U content, and plots well to the low ^{207}Pb/^{206}Pb side of the discordance line defined by the other analyses. Two near concordant analyses on grain 14 are significantly higher in radiogenic ^{207}Pb/^{206}Pb (0.29821 ± .00090) than the analyses of all other grains. The character of the luminescence of this grain

Table 1. U–Pb data for Mushandike zircons

Spot	U (ppm)	Th/U	f_{206} (%)	$^{206}Pb/^{238}U$	$\sigma_{6/8}$	$^{207}Pb/^{206}Pb$	$\sigma_{7/6}$	$T_{7/6}$ (Ma)	$\sigma_{T7/6}$ (Ma)	Disc (%)	CL
Sample MG14											
1–1	80	1.91	0.6	0.67334	0.01097	0.28235	0.00208	3376.0	11.5	2.2	
1–2	636	1.00	1.3	0.38193	0.00550	0.24367	0.00095	3144.1	6.2	39.2	
2–1	594	0.50	0.8	0.33354	0.00480	0.23787	0.00093	3105.7	6.2	46.1	
2–2	1147	3.17	1.9	0.22397	0.00320	0.18461	0.00089	2694.7	7.9	56.8	
3–1	583	1.11	0.3	0.38893	0.00560	0.26641	0.00080	3285.0	4.7	41.5	
3–2	1257	0.98	0.8	0.16762	0.00239	0.14053	0.00079	2233.7	9.8	59.5	
4–1	72	2.38	0.8	0.66616	0.01082	0.28261	0.00189	3377.4	10.4	3.3	
4–2	73	1.96	1.3	0.64810	0.01064	0.28193	0.00233	3373.6	12.9	5.8	
5–1	114	2.94	1.1	0.68692	0.01065	0.28133	0.00179	3370.3	9.9	0.0	
5–2	324	1.75	1.4	0.45638	0.00660	0.25527	0.00103	3217.7	6.3	29.5	
6–1	170	1.31	1.1	0.66518	0.01431	0.27898	0.00199	3357.2	11.1	2.7	
7–1	160	1.69	0.7	0.65107	0.00980	0.28253	0.00131	3376.9	7.2	5.4	
8–1	382	0.86	2.3	0.64445	0.00939	0.28363	0.00107	3383.0	5.9	6.6	
9–1	58	1.34	3.5	0.46800	0.00799	0.27609	0.00387	3341.0	22.0	31.1	
10–1	939	1.14	2.4	0.26179	0.00375	0.18859	0.00098	2729.9	8.6	50.3	
11–1	2043	0.26	3.0	0.16289	0.00232	0.12165	0.00088	1980.6	12.8	54.7	
12–1	91	1.49	1.0	0.63699	0.01020	0.28006	0.00195	3363.3	10.9	7.0	
13–1	41	1.48	1.9	0.64567	0.01149	0.27665	0.00297	3344.1	16.8	5.0	
Sample Z2											
1–1	160	0.36	0.07	0.67084	0.00828	0.28274	0.00133	3378.1	7.3	2.6	S
1–2	1382	0.26	7.12	0.22539	0.00234	0.24081	0.00237	3125.3	15.7	63.9	S
2–1	347	0.29	0.32	0.54484	0.00596	0.28659	0.00104	3399.2	5.7	21.5	W
3–1	152	0.50	0.06	0.68458	0.00845	0.28292	0.00141	3379.1	7.8	0.7	S
3–2	416	0.73	0.23	0.68930	0.00755	0.28143	0.00087	3370.9	4.8	–0.3	S
4–1	42	1.10	0.13	0.70694	0.01256	0.28513	0.00298	3391.2	16.3	–2.1	S
5–1	43	0.79	0.09	0.66431	0.01117	0.28127	0.00246	3370.0	13.6	3.3	S
5–2	235	0.65	0.21	0.60090	0.00693	0.27499	0.00115	3334.7	6.5	11.3	W
6–1	198	0.81	1.12	0.37422	0.00428	0.27660	0.00182	3343.8	10.3	45.0	W
6–2	242	1.12	0.21	0.59043	0.00680	0.28102	0.00120	3368.6	6.7	14.0	S
7–1	193	0.46	0.23	0.69310	0.00817	0.27351	0.00119	3326.3	6.8	–2.6	W
7–2	231	0.49	0.44	0.67027	0.00778	0.27440	0.00126	3331.3	7.2	0.9	W
8–1	34	0.71	0.43	0.67201	0.01238	0.27446	0.00323	3331.7	18.4	0.7	S
8–2	76	0.68	0.12	0.64927	0.00920	0.27828	0.00190	3353.3	10.7	4.8	S
9–1	61	0.75	0.70	0.62100	0.00954	0.27327	0.00295	3324.9	16.9	8.0	S
10–1	757	0.28	0.16	0.57407	0.00736	0.27815	0.00181	3352.6	10.2	15.8	W
10–2	695	0.40	1.86	0.50241	0.00535	0.27061	0.00124	3309.6	7.2	25.1	W
10–3	1128	1.11	4.92	0.35090	0.00362	0.28610	0.00150	3396.5	8.2	49.4	MC
11–1	200	0.75	0.23	0.62478	0.00735	0.28173	0.00124	3372.5	6.9	9.1	W
11–2	163	0.79	0.19	0.67800	0.00834	0.27686	0.00138	3345.3	7.8	0.3	W
12–1	386	0.38	0.30	0.65548	0.00718	0.27908	0.00087	3357.8	4.9	4.1	W
12–2	210	0.77	0.72	0.69294	0.00806	0.27782	0.00134	3350.7	7.5	–1.7	S
13–1	1545	0.24	1.59	0.20561	0.00210	0.13746	0.00107	2195.4	13.5	49.3	MC
14–1	207	0.13	0.05	0.75298	0.00892	0.29631	0.00114	3451.0	6.0	–6.3	S
14–2	198	0.29	0.32	0.68664	0.00823	0.30138	0.00147	3477.3	7.6	4.0	S

CL – Cathodoluminescent grouping: S – strongly luminescent; W – weakly luminescent; MC – mottled core. f_{206}(%) is the percentage of common ^{206}Pb in the total ^{206}Pb. Disc (%) shows the discordance as the percentage difference between the $^{206}Pb/^{238}U$ and $^{207}Pb/^{206}Pb$ ages.

is not markedly different from that of the zoned, melt-precipitated zircon, but both analyses are within a large central region of the grain which is overgrown discordantly by a second generation of zoned zircon. These analyses sample the only truly older core (3460 ± 5 Ma (σ)) identified in either Mushandike zircon population.

The remaining analyses define a better discordance line than for MG14 (MSWD = 12), but in marked contrast, the lower intercept is indistinguishable from 0 Ma (170 ± 170 Ma; 95% conf.). The upper intercept of 3360 ± 14 Ma (95% conf.) is a possible estimate of the age of zircon crystallization, but reference to the cathodoluminescence classification shows this interpretation to be simplistic. Leaving out the highly discordant analysis, 10–3, from a mottled zircon 'core' with high U content, the remainder are distributed bimodally. Those of strongly luminescent, euhedrally zoned zircon plot in a relatively tight cluster close to concordia. The radiogenic $^{207}Pb/^{206}Pb$ shows some scatter, even when the more discordant analyses (1–2, 6–1, 6–2 and 9–1) are omitted, but there remains no obvious outlier. For the eight remaining analyses the weighted mean value of 0.28097 ± 0.00087 (σ_{obs}, the standard error based upon the observed scatter) is equivalent to an age of 3368 ± 11 Ma (95% conf.), indistinguishable from the zircon age of 3374 ± 7 Ma obtained for MG14. In contrast, the analyses of the

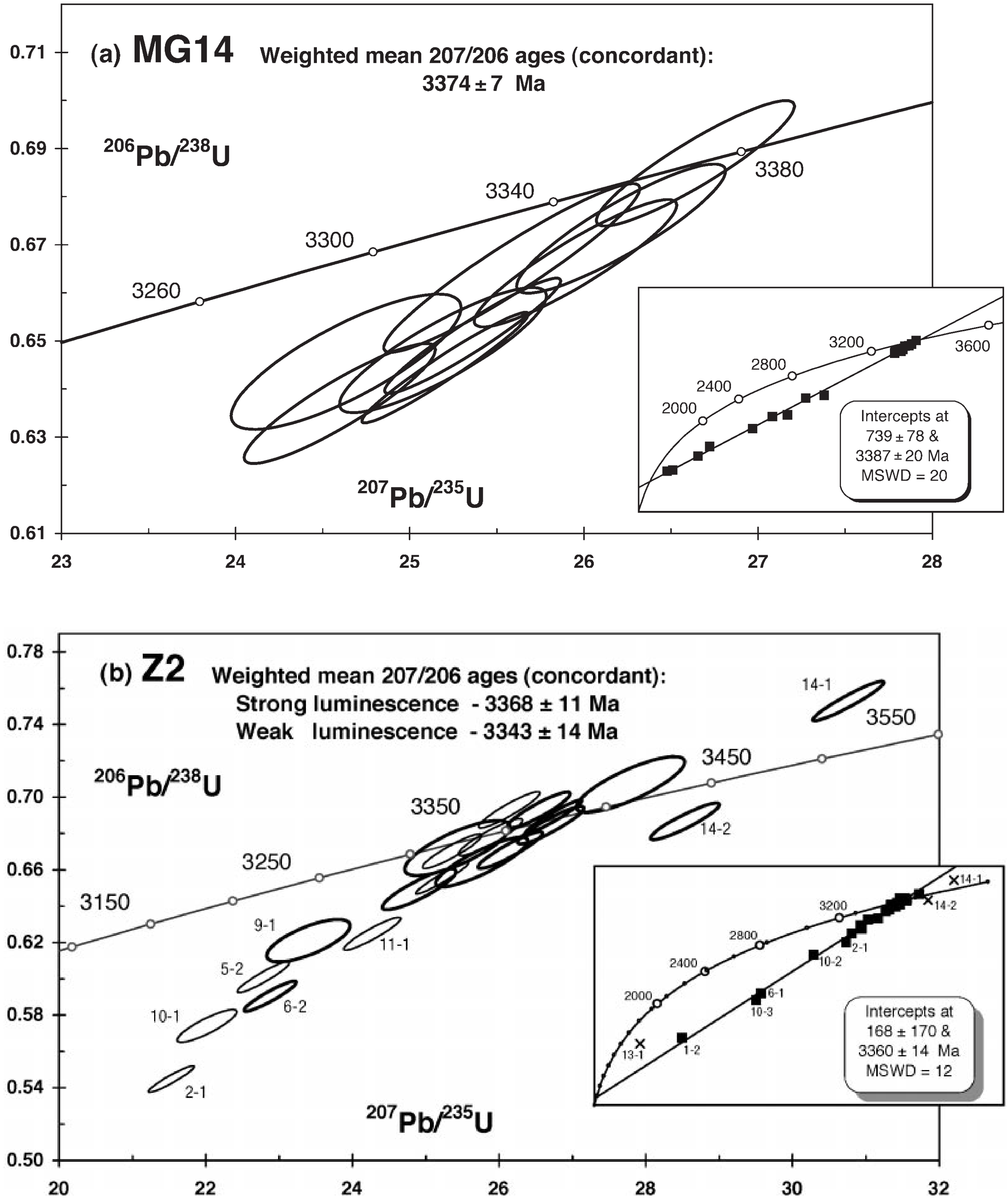

Figure 3. Concordia diagrams for U–Pb data on Mushandike zircons: (a) MG14; (b) Z2, crosses omitted from discordia fit, and bold ellipses correspond to strongly luminescent, well-zoned zircon grains identified by cathodoluminescence.

weakly luminescent zircon, (whether zoned or structureless, 'core' or 'overgrowth'), define a discordance line which is slightly offset from the other data. The analyses are widely dispersed, and with the exception of one (2–1), may be relatively well fitted (MSWD = 7) to a line with concordia intercepts of 3346 ± 44 Ma (95% conf.) and 150 ± 1040 Ma. This upper intercept is indistinguishable from the ^{207}Pb/^{206}Pb age of the strongly luminescent zircons. However, if only the most concordant of the analyses are considered (5–2, 7–1, 7–2, 10–1, 11–2 and 12–1), the weighted mean ^{207}Pb/^{206}Pb of 0.27639 ± 0.00097 yields 3343 ± 14 Ma (95% conf.), which is significantly younger than the strongly luminescent zircon. The difference probably reflects early Pb loss from the more structurally damaged zircon, so the age measured on the most luminescent zircon, 3368 ± 11 Ma, is preferred.

5. Geological implications

The evidence from both samples is that the Mushandike granite magma was emplaced at ~3.37 Ga. There is minimal evidence, in the form of the core of a single grain (no. 14), that the granite was derived from an older (~3.46 Ga) protolith or interacted with older material. Both samples show the effects of major post-crystallization isotopic disturbance, mainly variable Pb loss of apparent Neoproterozoic to recent age which cannot be interpreted uniquely in terms of the post-intrusion geological history. There is evidence for a much earlier (Archaean) Pb loss, but this is confined to a small but significant depression of the radiogenic ^{207}Pb/^{206}Pb in the most concordant of the weakly luminescent material. The strongly luminescent zircon may also have been affected, but to a much lesser extent.

Our conclusion that the Mushandike granite is ~3.37 Ga old is at variance with Moorbath and others' (1987) interpretation of their Rb–Sr and Pb–Pb data from the granite. Figure 4 shows their Rb–Sr analyses, plus the Rb–Sr measurements by Hickman (1974) that they considered to be aberrant. The emplacement age of the granite now being known, these data can be reassessed. The Rb–Sr whole rock analyses are widely scattered about a ~2.97 ± 0.19 Ga isochron (MSWD = 240). Much of the scatter, however, is contributed by only two analyses (MG3 and 29). Without those, the MSWD falls to 46 and the age rises to 3.17 ± 0.13 Ga, very little younger than the zircon age. There is still significant scatter about the isochron, but this, together with the slight lowering of the age below 3.37 Ga, is readily explicable by some Sr redistribution during later shearing. There is no need to invoke the presence of a randomly incorporated older component in a much younger magma.

The interpretation of the Pb–Pb isotopic data (Fig. 5) also is worth revisiting. Moorbath *et al.* (1987) used a subset of the analyses in calculating their age of 2.95 Ga, choosing to omit the least radiogenic samples and one relatively radiogenic outlier. If the least radiogenic samples are included, although the scatter increases, the isochron steepens significantly and the resultant age of 3.30 ± 0.12 Ga (MSWD = 44) becomes indistinguishable from the zircon age. This suggests that the scatter in the array of Pb isotopic compositions reflects isotopic disturbance of the more radiogenic samples, rather than inheritance of older material by the less radiogenic.

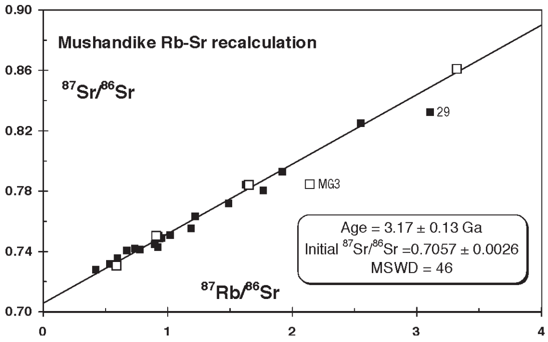

Figure 4. Published Rb–Sr data after Hickman (1974, open symbols) and Moorbath *et al.* (1987, filled symbols). Samples 29 and MG3 omitted from the new isochron calculation.

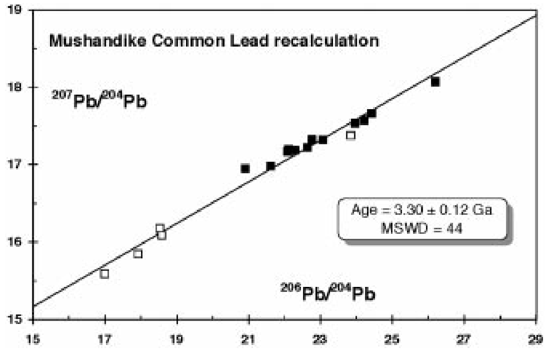

Figure 5. Recalculation of Pb–Pb data of Moorbath *et al.* (1987). Open symbols are data omitted from the 1987 age calculation. All samples have been included in the new line-fit calculation.

All three isotopic systems yield results consistent with the Mushandike granite having been intruded 3.37 Ga ago. All show clear evidence for isotopic disturbance. This disturbance might in part be related to fluid percolation associated with the shear zones traversing the granite. These zones show the effects of intense alteration: the granite is transformed into a quartz–sericite–andalusite schist. Much of the rock's K, Na and Ca appears to have been removed, leaving a peraluminous residue. The transition to unaltered granite is gradual, and it is likely that U, Pb, Rb and Sr were all mobilized, even at some distance from the shear zones. The complexity of the disturbance (for example, there is no correlation between ^{87}Sr/^{86}Sr and the ^{206}Pb/^{204}Pb) makes it difficult to infer its age with any precision. It could be related to any of several tectonometamorphic events that have affected the central Zimbabwe Craton. Folding and shearing in the Shurugwi area has been dated at 2.86 Ga (Nägler *et al.* 1997), and large scale shearing post-dating the *c.* 2.6 Ga Chilimanzi Granite suite is well documented (e.g. Treloar & Blenkinsop, 1995).

Our zircon data for the Mushandike granite complement the recent results of Horstwood *et al.* (1999) on gneisses from the Tokwe segment, a region bounded by the Masvingo, Shurugwi (Selukwe) and Mberengwa (Belingwe) greenstone belts (Wilson, Nesbitt & Fanning, 1995), and from the Midlands region further north. The age of emplacement for Mushandike is essentially identical with their 3368 ± 9 Ma age for zircons from a leucosome in Tokwe gneisses sampled 25 km north of Mushandike (Tl, Fig. 1). Similar ages have been published for the nearby Mont d'Or granite, south of Shurugwi: Moorbath, Wilson & Cotterill (1976) obtained 3.35 ± 0.12 Ga (whole-rock Rb–Sr), while Taylor *et al.* (1991) obtained 3.35 ± 0.05 Ga (Pb–Pb). A widespread magmatic event is suggested by these results. The T_{DM} age of 3.54 Ga for the Mushandike granite itself (Taylor *et al.* 1991) corresponds to a T_{CHUR} age of 3.50 Ga, a robust result due to the rock's very low Sm/Nd, and indicates that the magma incorporated a small amount of older crust, as does the one identified older (3.46 Ga) zircon core. The age by Horstwood *et al.* (1999) for the Tokwe gneiss, 3455 ± 2 Ma (sample Tg, Fig. 1) suggests a plausible source for that core. The presence of older crustal components in Southern Zimbabwe is further indicated by the high initial $^{87}Sr/^{86}Sr$ for the Mont d'Or granite (0.711 ± 0.001) and also by the abundant 3.8 Ga zircon grains in a conglomerate from the Wanderer formation, north of Shurugwi (Dodson *et al.* 1988). The 3.8 Ga age for sublithospheric mantle inferred by Nägler *et al.* (1997) from Re-Os systematics supports this picture. The further evidence of Horstwood *et al.* (1999) from the Kwe Kwe and Selukwe gneisses (3.46 and 3.57 Ga, respectively), leads them to suggest that the Tokwe crustal segment, together with the Midlands gneisses to the north, made up the 'Sebakwe protocraton', which after a lengthy tectonic history was stabilized during a widespread 3.35 Ga magmatic event.

The view that the Mushandike stromatolitic limestone pre-dates the Mushandike granite is contradicted by the new emplacement age for the granite. In contrast to the Rb–Sr and Pb–Pb ages from the granite, the 2.86 ± 0.03 Ga Pb–Pb isochron age for the limestone obtained by Moorbath *et al.* (1987) is well constrained and the data show no evidence for multiple later events separated widely in time. As there is no petrological evidence for metamorphism of the limestone, the interpretation of Moorbath *et al.* (1987) that this result records diagenesis is probably correct. The suggestion by Orpen & Wilson (1981) that the stromatolitic limestone is a Sebakwian sediment which pre-dated the Mushandike granite was based on (a) the apparent position of the limestone below the stratigraphic level of the Gwenya Formation, and (b) contact metamorphism observed in associated rocks. However, no intrusive contact between the granite and limestone has been seen. Furthermore, the unconformity at the base of the Gwenya Formation is locally intruded by a thick ultramafic sill belonging to the late Archaean Mashaba Igneous Complex (see Orpen & Wilson, 1981) which both obscures field relationships and could produce contact metamorphism. Our zircon age of 3.37 Ga for the Mushandike granite therefore does not imply that the Mushandike stromatolitic limestone is even older. The limestone is probably a member of the Gwenya Formation, forming part of a slice dislodged either by tectonism or through the intrusion of the Mashaba Complex sill. Its 2.86 ± 0.03 Ga Pb–Pb isochron age could well date the Gwenya Formation, the base of the Upper Bulawayan Sequence in the Masvingo Greenstone Belt.

Acknowledgements. We thank Professor Compston for his long-standing interest and encouragement and the staff of the ANU Electron Microscopy Unit for their asssistance with cathodoluminescence imaging. Zircon separations for MG14 were carried out by Ms C. M. Johnston. Travel funds for MHD to visit Canberra and Zimbabwe were provided by The University of Leeds, The Royal Society, The Australian National University, and the British Council. S. Moorbath and R. W. Nesbitt provided helpful reviews.

References

Berger, M., Kramers, J. D. & Nägler, Th. 1995. Geochemistry and geochronology of charnoenderbites in the Northern Marginal Zone of the Limpopo Belt, Southern Africa, and genetic models. *Schweizerische Mineralogische und Petrographische Mitteilungen* **75**, 17–42.

Compston, W., Williams, I. S. & Meyer, C. 1984. U–Pb geochronology of zircons from Lunar Breccia 73217 using a sensitive High Resolution Ion Microprobe. Proceedings of the 14th Lunar and Planetary Science Conference part 2. *Journal of Geophysical Research Supplement* **95**, B525–36.

Cummings, G. L. & Richards, J. R. 1975. Ore lead isotope ratios in a continuously changing earth. *Earth and Planetary Science Letters* **28**, 155–71.

Dodson, M. H., Compston, W., Williams, I. S. & Wilson, J. F. 1988. A search for ancient detrital zircons in southern Zimbabwe. *Journal of the Geological Society of London* **145**, 977–83.

Geological Survey of Southern Rhodesia. 1965. Geological Map of the country round Mashaba, 1:100,000.

Hickman, M. H. 1974. 3,500 Myr old granite in Southern Africa. *Nature* **251**, 295–6.

Horstwood, M. S. A., Nesbitt, R. W., Noble, S. R. & Wilson, J. F. 1999. U–Pb zircon evidence for an extensive early Archaean craton in Zimbabwe: A reassessment of the timing of craton formation, stablisation and growth. *Geology* **27**, 707–10.

Jaeckel, P., Kröner, A., Kamo, S. L., Brandl, G., & Wendt, J. I. 1997. Late Archaean to early Proterozoic granitoid magmatism in the central Limpopo Belt, South Africa. *Journal of the Geological Society, London* **154**, 25–44.

Kröner, A., Jaeckel, P., Brandl, G., Nemchin, A. A. & Pidgeon, R. T. 1999. Single zircon ages for granitoid gneisses in the Central Zone of the Limpopo Belt, Southern Africa and geodynamic significance. *Precambrian Research* **93**, 299–337.

Ludwig, K. R. 2000. *Users' Manual for Isoplot/Ex version 2.3.* Berkeley Geochronology Center Special Publication no. 1a.

MOORBATH, S., WILSON, J. F. & COTTERILL, P. 1976. Early Archaean age for the Sebakwian Group at Selukwe, Rhodesia. *Nature* **264**, 536–8.

MOORBATH, S., TAYLOR, P. N. & JONES, N. W. 1986. Dating the oldest terrestrial rocks—fact and fiction. *Chemical Geology* **57**, 63–86.

MOORBATH, S., TAYLOR, P. N., ORPEN, J. L., TRELOAR, P. & WILSON, J. F. 1987. First direct radiometric dating of Archaean stromatolitic limestone. *Nature* **326**, 865–7.

NÄGLER, TH. F., KRAMERS, J. D., KAMBER, B. S., FREI, R. & PRENDERGAST, M. D. A. 1997. Growth of subcontinental mantle beneath Zimbabwe started at or before 3.8 Ga: Re-Os study on chromites. *Geology* **25**, 983–6.

ORPEN, J. L. & WILSON, J. F. 1981. Stromatolites at c. 3,500 Myr and a greenstone-granite unconformity in the Zimbabwean Archaean. *Nature* **291**, 218–20.

STEIGER, R. H. & JÄGER, E. 1977. Subcommission on Geochronology: convention on the use of decay constants in geo- and cosmochronology. *Earth and Planetary Science Letters* **36**, 359–62.

TAYLOR, P. N., KRAMERS, J. D., MOORBATH, S., WILSON, J. F., ORPEN, J. L. & MARTIN, A. 1991. Pb/Pb, Sm–Nd and Rb–Sr geochronology in the Archaean craton of Zimbabwe. *Chemical Geology (Isotope Geoscience Section)* **87**, 175–96.

TRELOAR, P. J. & BLENKINSOP, T. G. 1995. Archaean deformation patterns in Zimbabwe: True indicators of Tibetan-style extrusion or not? In *Early Precambrian Processes* (eds M. P. Coward and A. C. Ries), pp. 87–108. Geological Society Special Publication no. 95.

VAIL, J. R. & DODSON, M. H. 1969. Geochronology of Rhodesia. *Transactions of the Geological Society of South Africa* **62**, 79–113.

WILLIAMS, I. S. & CLAESSON, S. 1987. Isotopic evidence for the Precambrian provenance and Caledonian metamorphism of high grade paragneisses from the Seve Nappes, Scandinavian Caledonides, II. Ion microprobe zircon U–Th–Pb. *Contributions to Mineralogy and Petrology* **97**, 205–17.

WILSON, J. F. 1979. *A preliminary reappraisal of the Rhodesian Basement Complex.* Geological Society of South Africa, Special Publication no. 5, 1–23.

WILSON, J. F., NESBITT, R. W. & FANNING, M. 1995. Zircon geochronology of Archaean felsic sequences in the Zimbabwe Craton: a revision of greenstone stratigraphy and a model for crustal growth. In *Early Precambrian Processes* (eds M. P. Coward and A. C. Ries), pp. 109–26. Geological Society of London, Special Publication no. 95.

WILSON, J. F., BICKLE, M. J., HAWKESWORTH, C. J., MARTIN, A., NISBET, E. G. & ORPEN, J. L. 1978. Granite-greenstone terrains of the Rhodesian Archaean craton. *Nature* **271**, 23–7.

ZIMBABWE GEOLOGICAL MAP. 1985. (reprinted) 1:1,000,000, 7th edition.