

Mineralogy and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of orangeites (Group II kimberlites) from the Damodar Valley, eastern India

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ABSTRACT

A suite of ultramafic-mafic alkaline igneous rocks in the Damodar Valley, eastern India, contains carbonate, phosphate and titanate minerals that are not characteristic or common in minettes or lamproites, but are typical of orangeites (Group II kimberlite) from southern Africa. Phlogopite grains from the Damodar alkaline rocks yield mean $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 116.6 ± 0.8 Ma, 113.5 ± 0.5 Ma and 109.1 ± 0.7 Ma (1σ errors) using laser dating techniques. These ages are similar to the Rb-Sr ages of African orangeites, which lie mostly in the range 121 to 114 Ma. Prior to this study, only one possible occurrence of orangeite (the ~820 m.y.-old Aries pipe, Western Australia) was known outside the Kaapvaal craton and its environs. If the Damodar alkaline rocks are bona fide orangeites, it is likely that they were generated at depths of >150 km, within the stability field of diamond.

KEYWORDS: orangeite, Group II kimberlite, $^{40}\text{Ar}/^{39}\text{Ar}$, craton, diamond.

Introduction

UNDERSTANDING the origin of primary diamond deposits requires an accurate knowledge of the age, mineralogy and chemistry of their alkaline igneous hosts. Although a prolific source of diamonds, many kimberlite, orangeite (also known as 'Group II kimberlite') and olivine lamproite-hosted deposits are poorly characterised in terms of age and petrology. Other potentially diamondiferous occurrences have been insufficiently studied to determine whether they are similar in composition and derived from sources comparable to those of igneous rocks known to carry diamonds. This study presents new $^{40}\text{Ar}/^{39}\text{Ar}$ and mineralogical data for early Cretaceous ultramafic-mafic alkaline igneous rocks in eastern India. These rocks are mica-rich and have been termed lamproites, minettes, mica peridotites and glimmerites (e.g. Gupta *et al.*,

1983; Middlemost *et al.*, 1988; Rock *et al.*, 1992). They are of interest principally because of their mineralogical similarity to diamondiferous orangeites from the Kaapvaal craton, South Africa, and the diamond-rich (5 CM/100 t) Aries kimberlite, east Kimberley, Western Australia.

Ultramafic-mafic alkaline igneous rocks occur in all three Archaean cratons of India (Fig. 1; Scott-Smith, 1989). Emplacement ages of the four best-known lamproite occurrences, at Majhgawan and Hinota in the Aravalli craton, and at Chelima and Ramannapeta, Dharwar craton, range from 1391 Ma to 1044 Ma (K-Ar and Rb-Sr data; Rao *et al.*, 1996, and references therein). These ages are generally similar to those of kimberlites intruding the Dharwar craton (1363–1091 Ma, Rb-Sr and K-Ar data; e.g. Kumar *et al.*, 1993; Rao *et al.*, 1996). Phanerozoic 'lamproites' have been documented at only two localities in India: the

southeastern Raipur district of Madhya Pradesh (diamond-bearing diatremes at Buhradih, Payalikhand, Jangra and Kodomali) and the Damodar Valley, Bihar/West Bengal (Figs. 1 and 2). The Raipur intrusions have yet to be studied in detail, and their age and status are uncertain (see Newlay and Pashine, 1993; Chatterjee *et al.*, 1995). The status of the Damodar Valley intrusions also is unclear and we designate them 'Damodar alkaline igneous rocks' (DAIR) prior to adopting a mineralogical-genetic term later in this study.

Geological setting

The DAIR form part of a Cretaceous large igneous province on the eastern Indian margin that includes the Rajmahal basalts, carbonatites

and lamprophyres in the Shillong Plateau - Mikir Hills (Meghalaya/Assam), 'lamproites' in the Ramthi Valley, Darjeeling (Sikkim) and 'lamprophyres' at Sidhi (Madhya Pradesh) (e.g. Paul and Sarkar, 1984). A Shillong lamprophyre analysed by Sarkar *et al.* (1996) has a K-Ar age of 107 ± 4 Ma, outside analytical error of those obtained for the Rajmahal lavas and associated basic dykes (116-113 Ma, $^{40}\text{Ar}/^{39}\text{Ar}$ data; Kent *et al.*, 1997). A younger apatite fission-track age of 97 Ma reported for the Sidhi 'lamprophyre' (T. Ray Barman, in Paul and Sarkar, 1984) is best explained as the time of cooling below $\sim 110^\circ\text{C}$, the temperature at which track retention begins. In contrast, the Shillong carbonatites (149 ± 5 Ma, K-Ar age on phlogopite; Sarkar *et al.*, 1996) appear to be significantly older than the Rajmahal basalts.

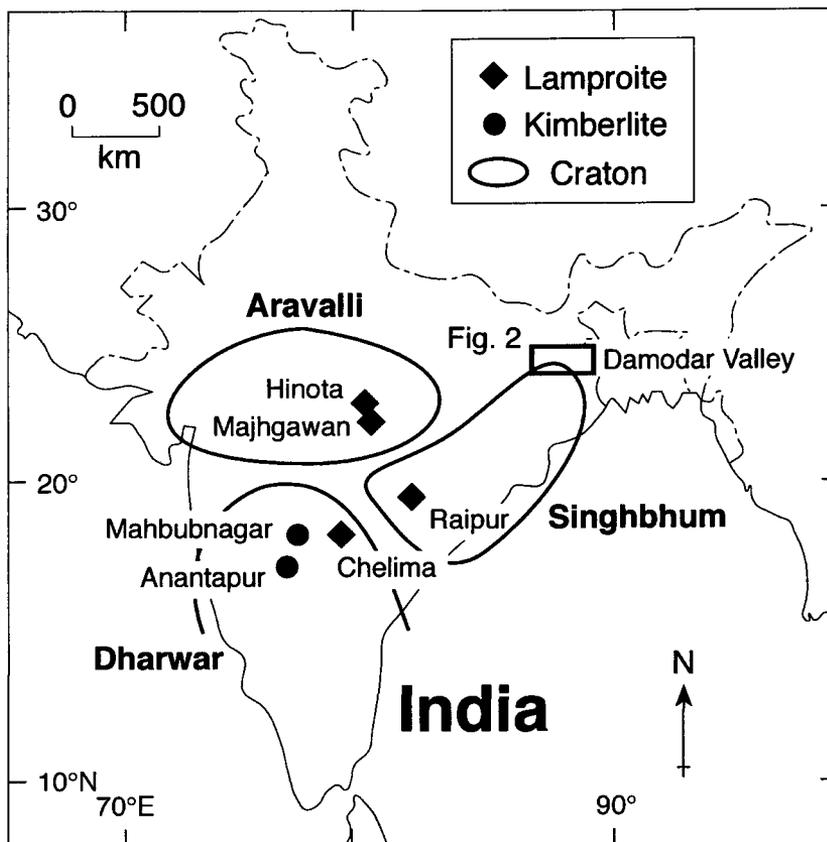


FIG. 1. Map showing the location of lamproites and kimberlites in India and their relationship to Archaean cratons (modified from Rao *et al.*, 1996). The Damodar Valley alkaline province lies on the northern margin of the Singhbhum craton.

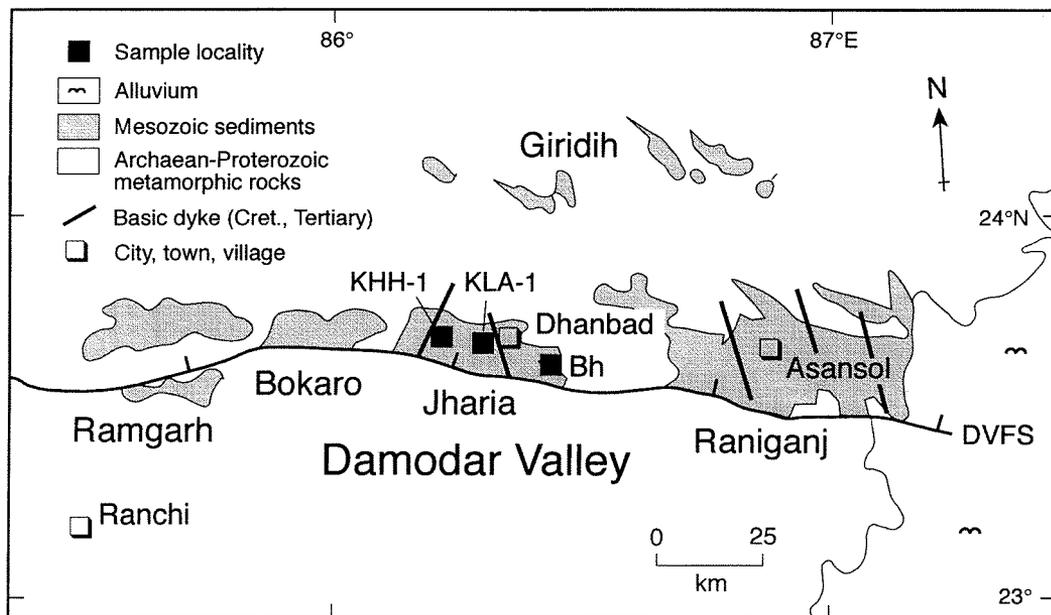


FIG. 2. Map of the eastern Damodar Valley, Bihar/West Bengal, eastern India, showing sample localities in the Jharia coalfield. Coalfields are designated by italics. Bh = Bhowrah, DVFS = Damodar Valley Fault System.

The DAIR intrude Mesozoic sediments and Proterozoic (>1.1 Ga) metamorphic rocks lying immediately to the north of the Singhbhum craton. The setting appears to be off-craton (*contra* Rock *et al.*, 1992), within the southern part of the Chotanagpur mobile belt. However, the northern boundary of the Singhbhum craton is not well-defined in subsurface and we cannot rule out the possibility that the craton extends beneath the Damodar Valley (see, for example, Radhakrishna, 1989). The DAIR occur as dykes, sills and cylindrical sill-like bodies up to 2 m in diameter (Fox, 1930; Gee, 1932; Kent *et al.*, 1992). Evidence of phreatomagmatic activity occurs at the brecciated termini of some cylindrical sills, but lavas and pyroclastic facies have not yet been identified. Magmas parental to the DAIR appear to have ascended to high crustal levels as dykes, facilitated by the presence of the Damodar Valley fault system (Fig. 2) and numerous minor faults cutting the Jharia and Raniganj coalfields. Although *bona fide* diatremes are not observed, flared dykes such as that exposed to the north of Bhowrah, Jharia coalfield (Fig. 2) have a structure analogous to that of 'blows' along orangeite dykes at Bellsbank and Newlands in

the Barkly West district, Cape Province, South Africa (see Mitchell, 1995, p. 21–27).

Sarkar *et al.* (1980) obtained K-Ar ages of 113–105 Ma for biotites from ultramafic–mafic alkaline rocks in the eastern Damodar Valley. The precision on the K-Ar ages (± 3 m.y.) of DAIR allows an age range for these rocks from 116 Ma to 102 Ma. If a 'lamprophyre' from Darjeeling is included (sample A818 RM-LP/4, K-Ar age of 121 ± 3 Ma; Sarkar *et al.*, 1980), the range becomes even greater (22 m.y.). Better resolution of this age span is desirable if we are to establish the timing of intrusion and compare the age of the DAIR with those of African orangeites.

Mineralogy

The petrography and mineralogy of DAIR samples have been described by several workers (see references of Rock and Paul, 1989; Rock *et al.*, 1992). Our studies, summarised in Table 1, reveal that the DAIR are similar to 'typical' orangeites (*sensu* Mitchell, 1995) in that they (i) contain microphenocrystal and groundmass phlogopite-tetraferriphlogopite (Table 2); (ii) contain microphenocrystal Sr-rich apatite; (iii) contain

TABLE 1. Comparison of the mineralogy of African orangeites and Damodar alkaline igneous rocks

Mineral	South African orangeites	Damodar alkaline igneous rocks
Olivine		
Macrocrysts	Common to rare - principally xenocrysts	Absent
Phenocrysts	Minor (F ₀₉₁₋₉₃) subhedral/euhedral/dog's tooth	Major (F ₀₈₅₋₉₂) euhedral, pseudomorphed by serpentine, Mg-calcite, ferromagnesite, talc
Mica		
Macrocrysts	Common, phlogopite cognate	Common, phlogopite xenocrystic?
Microphenocrysts	Common, phlogopite	Major, phlogopite-tetraferriphlogopite
Groundmass	Common, phlogopite-tetraferriphlogopite Poikilitic plates	Common, phlogopite Pseudo-hexagonal, castellated, or poikilitic plates
Spinel	Minor to rare, small (0.001–0.02 mm). Euhedral Mg-chromite common, rarely zoned to Ti-magnetite. Atoll spinel rare	Minor. Euhedral Mg-chromite common, sometimes as inclusions in apatite. Ti-magnetite in groundmass
Diopside	Microphenocryst. Common to rare. Commonly resorbed. Zoned to Ti- aegirine	Microphenocryst. Common. Ti- and Na- rich varieties, occasional aegirine. Often replaced by chlorite (diabazite)
Perovskite	Rare, subhedral to poikilitic SrO- (<1–6 wt.%) and (REE) ₂ O ₃ -rich (3–16 wt.%)	Rare, no quantitative data available
Apatite	Common euhedral prisms and poikilitic plates. SrO- (3–22 wt.%) and (REE) ₂ O ₃ - rich (<1–10 wt.%)	Major. Euhedral prisms and poikilitic plates. Strongly zoned, SrO- (up to 20 wt.%) and F-rich (up to 5 wt.%)
Carbonates	Common calcite. Common Sr-Mn-Fe dolomites, minor witherite, ancylite, strontianite, norsethite	Common breunnerite, dolomitic ankerite, Fe-dolomite, minor calcite, Sr-witherite, strontianite, alstonite
Serpentine	Common, secondary	Common, secondary
K-feldspar	Rare groundmass. Sanidine	Common to rare, primary? Sanidine, rare hyalophane
Amphibole	Rare groundmass. K-richterite	Minor groundmass. K-richterite or K- arfvedsonite
K-Ba hollandites	Common groundmass	Not identified
K ₂ Ti ₁₃ O ₂₇	Common	Common. Acicular aggregates previously identified as Ba-poor priderite
Ilmenite	Common, Mn-rich	Rare, Nb-rich
Rutile	Trace	Common, Nb-rich
Sphene	Absent	Rare
Zr silicates	Common, include zircon, wadeite, Zr-garnet	Common, zircon only?
Leucite	Rare pseudomorphs in poikilitic mica	Absent
REE phosphates	Minor monazite, daqingshanite. Sr-REE phosphate	Minor monazite
Baryte	Common	Common
Quartz	Minor, groundmass	Minor, groundmass
Glass	Absent?	Minor, devitrified
Macrocryst suite	Absent to extremely rare	Absent

rutile (Table 3) and prismatic crystals of K-triskaidecatitanate (K₂Ti₁₃O₂₇; Mitchell, 1995),

the latter misidentified as priderite by Gupta *et al.* (1983), Middlemost *et al.* (1988) and Rock *et al.*

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TABLE 2. Representative compositions of micas from the Damodar Valley alkaline igneous rocks

Wt.%	1	2	3	4	5	6
SiO ₂	38.08	38.26	38.68	36.94	38.23	39.29
TiO ₂	6.73	8.28	8.53	10.52	9.83	8.56
Al ₂ O ₃	12.33	9.28	9.38	12.09	10.99	10.48
Cr ₂ O ₃	0.46	n.d.	0.03	0.28	n.d.	n.d.
FeO*	5.83	15.58	13.43	6.27	9.02	9.69
MnO	0.19	0.26	0.17	0.21	0.01	0.10
MgO	17.69	12.51	14.22	18.19	17.41	17.95
CaO	n.d.	0.40	n.d.	0.36	0.34	0.14
Na ₂ O	n.d.	0.79	0.49	0.58	1.13	0.38
K ₂ O	10.46	10.43	9.88	10.50	10.32	10.37
BaO	4.73	n.d.	0.17	n.d.	n.d.	n.d.
NiO	n.d.	0.23	n.d.	0.10	n.d.	0.15
P ₂ O ₅	0.77	0.14	n.d.	n.d.	n.d.	n.d.
Total	97.27	96.16	94.98	96.03	97.28	97.09
Structural formulae based on 22 oxygens						
Si	5.585	5.786	5.822	5.361	5.528	5.680
Al	2.127	1.651	1.663	2.068	1.873	1.786
Ti	0.743	0.945	0.965	1.148	1.069	0.931
Cr	0.055	—	0.004	0.031	—	—
Fe	0.715	1.972	1.690	0.761	1.090	1.170
Mn	0.028	0.037	0.021	0.026	0.018	0.011
Mg	3.861	2.824	3.191	3.934	3.751	3.870
Ca	—	0.064	—	0.057	0.053	0.022
Na	—	0.229	0.143	0.161	0.315	0.106
K	1.953	2.008	1.897	1.945	1.904	1.912
Ba	0.275	—	0.010	—	—	—
Ni	—	0.028	—	0.012	—	0.018
P	0.092	0.018	—	—	—	—
mg#	0.844	0.589	0.654	0.838	0.775	0.768

Notes: Mineral compositions were determined by JEOL 8600S wavelength-dispersive electron microprobe at the University of Leicester. The electron beam was operated at an accelerating voltage of 15 kV and a probe current of 30 nA, with a diameter of 5 µm and a 60 s count time. X-ray intensities were standardised against a range of pure elements, synthetic oxides and natural minerals. Dead-time and background corrections were applied to X-ray intensities followed by ZAF correction to the measured oxide concentrations. FeO* = total Fe expressed as FeO. n.d. = not detected. mg# = atomic Mg/(Mg+Fe²⁺), where Fe₂O₃/FeO = 0.15. Numbers 1–6 refer to representative micas from sample KHH-1.

(1992); (iv) contain phenocrystal olivine and primary microphenocrystal and groundmass diopside or aegirine; and (v) characteristically contain sanidine, K-richterite, perovskite, monazite, ilmenite and a varied assemblage of carbonates (calcite, ferroan dolomite, strontianite, breunnerite, strontian witherite). The DAIR differ from archetypal orangeites in that the former

rocks contain Ti-rich micas and glass, and have a greater modal abundance of apatite and sanidine. The DAIR differ also from the Aries kimberlite, Western Australia, in that the DAIR contain amphibole, sanidine and perovskite.

Rutile and K-triskaidecatitanate are not characteristic or common in lamproites, and occur together only in the unusual Sisimiut lamproites

TABLE 3. Microprobe analyses of rutile from the Damodar Valley alkaline igneous rocks compared to data for rutiles from South African orangeites

Wt.%	KHH-1	M-3	Sover Mine	Lace (Crown)	Bellsbank
TiO ₂	97.88	97.30	93.33–95.20	97.28–97.67	91.91
Cr ₂ O ₃	n.d.	0.09	n.d.	n.d.	n.d.
Al ₂ O ₃	n.d.	0.06	—	—	—
FeO*	0.43	0.73	1.09–1.29	0.92–1.17	4.39
Nb ₂ O ₅	n.a.	n.a.	3.17–5.35	1.09–1.69	3.09
MnO	n.d.	0.06	0.25–0.48	0.23–0.41	n.d.
MgO	0.11	n.d.	—	—	—
CaO	0.23	0.09	—	—	—
Other#	1.23	0.16	—	—	—
Total	99.88	98.49	100.10–100.14	100.09–100.12	99.39

Notes: FeO* = total Fe expressed as FeO. n.d. = not detected. n.a. = not analysed.

Includes SiO₂, K₂O, Na₂O and NiO.

Sample KHH-1 is from this study, M-3 is from Gupta *et al.* (1983). The range in compositions of rutile from Sover Mine, Lace and Bellsbank (South Africa) is taken from Mitchell (1995, table 2.45). Data for Al₂O₃, MgO and CaO wt.% were not reported by Mitchell.

of central West Greenland (Scott, 1981; Thy *et al.*, 1987). Rutile in DAIR samples is exceptionally pure (98.0–98.8 wt.% TiO₂) but contains significant amounts of FeO* (0.4–0.7 wt.%) (Table 3; Gupta *et al.*, 1983). Chromium contents are similar to those of rutile in South African orangeites and the Aries kimberlite (i.e. ≤ 0.1 wt.% Cr₂O₃; Table 3). Niobium was not analysed during this study, but niobian rutile was reported in three DAIR samples by Rock *et al.* (1992).

Analyses of 'priderite' (actually K-triskaidecitanate) in samples from the Jharia and Raniganj coalfields indicate high TiO₂ and FeO* and variable BaO (76.2–84.5, 5.3–9.1, and 0.8–9.1 wt.%, respectively; analyses recalculated to 100 wt.%; Gupta *et al.*, 1983; Middlemost *et al.*, 1988; Rock *et al.*, 1992). These crystals have variable Ba contents but otherwise are similar in composition to K-triskaidecitanates from the Sisimiut lamproite (Scott, 1981) and the Lace, Sover North and Star orangeites, Orange Free State (Mitchell and Meyer, 1989; Mitchell, 1995).

Samples and methods

Fresh exposures of ultramafic and more-differentiated alkaline igneous rocks occupy an area of about 1 km² at Kustore colliery, 5 km southwest of Dhanbad, Jharia coalfield (Fig. 2; Kent *et al.*, 1992). These rocks occur as a clutch of cylindrical sills linked by thin (<0.5 m diameter) cross-

branches. Sample KLA-1 was obtained from Kustore, whereas sample KHH-1 was collected at 317 m depth in a borehole at Kharkharee, 22 km west-southwest of Dhanbad (Fig. 2). The intrusion intersected by the borehole is about 0.4 m thick and has chilled contacts with Barakar Formation sandstone, the sedimentary host rock.

A laser-probe Ar dating technique was used to analyse small phlogopite grains separated mechanically from KHH-1 and KLA-1. These grains show no visible signs of alteration and provide an excellent opportunity for precise ⁴⁰Ar/³⁹Ar dating due to their high K content (~10.5 wt.%). The grains were cleaned ultrasonically in methanol and de-ionised water, and then wrapped in aluminium foil prior to irradiation. Each sample received a total fast neutron dose of ~10¹⁸ n/cm² at the Ford reactor, University of Michigan. The MMhb-1 hornblende standard (520.4 Ma, Samson and Alexander, 1987) was used to monitor the neutron flux during irradiation. After irradiation, the samples were loaded into a ultrahigh-vacuum laser port and were placed under a heat lamp overnight to reduce atmospheric blank levels. Argon was extracted from small areas of each sample (laser pits between 50 and 100 μ m in diameter) using 1 ms to 200 ms pulses generated by a Spectron SL902 CW Nd-YAG laser and a computer-controlled external shutter. The laser beam was directed into a customised Leica Metallux 3 microscope and

focused at the sample surface via the objective lens. Gases released by the laser were collected for a minimum of 5 min and equilibrated via automated valves into a MAP 215-50 noble gas mass-spectrometer. Argon peaks (36 to 40) were scanned seven times and peak heights extrapolated back to the inlet time in order to take account of Ar build-up and memory effects. Blanks were analysed after every sample extraction. No significant neutron flux variation ($<0.5\%$) was apparent between the sample, and a J value of 0.006096 ± 0.00003 was calculated for both samples. Data presented in Table 4 were corrected for blanks, ^{37}Ar decay, and neutron-induced interference reactions. Correction factors used were $[\text{}^{39}\text{Ar}/\text{}^{37}\text{Ar}]_{\text{Ca}} = 0.000781 \pm 0.000053$; $[\text{}^{36}\text{Ar}/\text{}^{37}\text{Ar}]_{\text{Ca}} = 0.000205 \pm 0.000022$; and $[\text{}^{40}\text{Ar}/\text{}^{39}\text{Ar}]_{\text{K}} = 0.031 \pm 0.008$.

Phlogopite grains from sample KLA-1 were analysed also by laser total-fusion at the Free University, Amsterdam. Sample preparation methods, analytical techniques, blanks and corrections for neutron-induced reactions are similar to those described by Wijbrans *et al.* (1995). We accept a total-fusion age as an accurate estimate of the crystallisation age only if: (1) a well-defined, high-temperature age

spectrum plateau is defined by three or more contiguous steps, comprising $>50\%$ of the ^{39}Ar released and all concordant in age at the 95% confidence level; (2) a well-defined isochron exists for the plateau points, i.e. the F -variate statistic for the York2 regression, $SUMS/(N-2)$, is sufficiently small at the 95% confidence level; (3) the plateau and isochron ages are concordant at the 95% confidence level; and (4) the $^{40}\text{Ar}/\text{}^{39}\text{Ar}$ intercept on an isochron diagram does not differ from the atmospheric value of 295.5 at the 95% confidence level. The isochron age obtained for KLA-1 is preferred over the weighted mean plateau age because it combines quantitative estimates of analytical precision and internal disturbance of the sample (scatter about the isochron line) without making assumptions about the trapped argon component (cf. Singer and Pringle, 1996).

$^{40}\text{Ar}/\text{}^{39}\text{Ar}$ geochronology

Sample KLA-1 yielded a mean laser ablation age of 116.6 ± 0.8 Ma (1σ error) and KHH-1 yielded an age of 109.1 ± 0.7 Ma (Table 4). In Fig. 3, the laser-ablation age of KLA-1 is older (by 2.7%) than the $^{40}\text{Ar}/\text{}^{39}\text{Ar}$ total-fusion age of this sample

TABLE 4. $^{40}\text{Ar}/\text{}^{39}\text{Ar}$ laser probe data for phlogopites from Damodar alkaline igneous rocks

Sample	Position	$^{40}\text{Ar}/\text{}^{39}\text{Ar}$	$^{36}\text{Ar}/\text{}^{39}\text{Ar}$	Total ^{39}Ar	$^{40}\text{Ar}^*/\text{}^{39}\text{Ar}$	Age $\pm 1\sigma$ (Ma)
KLA-1 [†]	Rim	11.596	0.00209	55.33	10.978	116.9 ± 0.9
	Centre	10.949	0.00081	105.330	10.711	114.1 ± 3.2
	Centre	11.016	0.00035	54.30	10.914	116.2 ± 2.0
Weighted mean age						116.6 ± 0.8
KHH-1 [†]	Rim	11.672	0.00561	117.070	10.015	106.9 ± 1.1
	Centre	11.582	0.00383	12.86	10.451	111.4 ± 1.0
	Centre	10.746	0.00225	21.18	10.081	107.6 ± 1.5
Weighted mean age						109.1 ± 0.7

Sample	Age Spectrum		N	SUMS/N-2	Isochron Analysis	
	^{39}Ar (%)	Age $\pm 1\sigma$ (Ma)			Intercept $\pm 1\sigma$	Age $\pm 1\sigma$ (Ma)
KLA-1*	all	113.3 ± 0.3	6	0.49	246.4 ± 79.3	113.5 ± 0.5

Notes: Total ^{39}Ar = 'value $\times 10^{-10}$ cm³ STP'.

Key: [†] = laser ablation analysis. * = laser total-fusion analysis.

The total-fusion age of mica from sample KLA-1 is reported relative to the age of sanidine TCR 85G003 from the Taylor Creek rhyolite (27.92 Ma; e.g. Singer and Pringle, 1996). Relative to TCR 85G003, the hornblende standard MMhb-1 gives an age of 513.9 ± 1.0 Ma (references of Singer and Pringle, 1996).

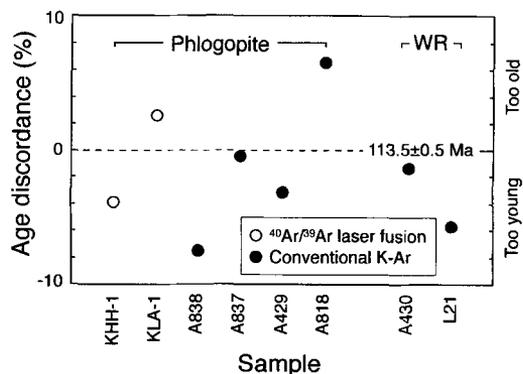


FIG. 3. Age discordance of laser-ablation $^{40}\text{Ar}/^{39}\text{Ar}$ ages (Table 4, this study) and conventional K-Ar ages (Sarkar *et al.*, 1980, 1996) shown as a percentage of the best estimate of the age of eastern Indian alkaline igneous rocks (113.5 ± 0.5 Ma; $^{40}\text{Ar}/^{39}\text{Ar}$ total-fusion age, Table 4). The K-Ar ages are generally less than the best age estimate, whereas laser $^{40}\text{Ar}/^{39}\text{Ar}$ ages can be higher or lower. WR = whole-rock sample.

(113.5 ± 0.5 Ma, Table 4). The isochron age has a $\text{SUMS}/(N-2)$ value of 0.49 and a $^{40}\text{Ar}/^{36}\text{Ar}$ intercept of 246.4 ± 79.3 , and is thus a reliable estimate of the crystallisation age. In comparison, the K-Ar age of mica from sample A818 RM-LP/4 (Sarkar *et al.*, 1980) is 6.6% older than the best age estimate for KLA-1. These age differences are unlikely to be real; the presence of clays and zeolites in DAIR samples suggests that the 'older' ages are the result of internal redistribution of ^{39}Ar and/or ^{37}Ar during low-temperature alteration. In contrast, the $^{40}\text{Ar}/^{39}\text{Ar}$ age of KHH-1 and previously reported K-Ar ages of DAIR samples are up to 8% younger than the best age estimate for KLA-1. These anomalously young ages could reflect Ca-derived ^{37}Ar recoil loss during and subsequent to irradiation, radiogenic ^{40}Ar loss from sites which have retained part of their K-derived ^{39}Ar or the presence of younger alteration minerals that were not completely removed during cleaning of the phlogopites. Thus, although they represent a significant improvement in precision over previous K-Ar ages, the laser ablation ages may be inaccurate and should be interpreted cautiously. Using the $^{40}\text{Ar}/^{39}\text{Ar}$ total-fusion age of KLA-1, we can say that the DAIR were intruded at ~ 114 – 113 Ma, supporting the view (Kent *et al.*, 1997) that alkaline volcanism accompanied eruption of the Rajmahal basalts.

Discussion

Radiometric age determinations on African orangeites have produced a range from 200 Ma to 110 Ma, with most samples having Rb-Sr ages of 121 Ma to 114 Ma (references of Skinner, 1989). A few orangeite intrusions near Prieska on the southwest margin of the Kaapvaal craton have imprecise Rb-Sr ages in the range 167–136 Ma (Skinner *et al.*, 1994; Smith *et al.*, 1994). A central thesis of Mitchell's (1995) monograph on kimberlites and orangeites is that members of the orangeite clan do not occur outside the Kaapvaal craton and its environs (cf. Skinner, 1989; Tainton, 1992). Orangeites are considered to be "genetically and mineralogically distinct from lamproites and thus represent a distinct magma type" unique to southern Africa (Mitchell, 1995, p. 380). The suggestion of a purely African provenance for orangeites is supported by the fact that while mineralogically similar to certain orangeites, samples from the Aries pipe, Western Australia, have chemical-isotopic compositions that resemble kimberlites (*s.s.*) from Koidu, Sierra Leone, and Kundelungu, Zaire, and thus probably are not true orangeites (cf. Edwards *et al.*, 1992; Taylor *et al.*, 1994).

We have drawn attention to close similarities between South African orangeites and the DAIR (e.g. Table 1). These similarities can be interpreted in three ways. Firstly, the DAIR could be *bona fide* orangeites. If so, this would be the first reported occurrence of orangeite outside southern Africa, with attendant implications for diamond potential (see below). A second possibility is that the DAIR are transitional rocks which cannot be unequivocally assigned to the orangeite, kimberlite or lamproite clans. In this case, their source would be unique, relative to the sources of alkaline igneous rocks in other cratons. A third interpretation is that the DAIR reflect a mineralogical convergence of two genetically different clans (orangeite and lamproite) that results from the differentiation of compositionally similar potassic liquids.

Mineralogical data, together with the new radiometric ages presented here, enable us to assess which of these interpretations is most likely to be correct. An important observation is that the DAIR conform to Mitchell's (1995, p.14) mineralogical definition of orangeite, but unlike African orangeites, are not typically peralkaline (i.e. most DAIR have molar $(\text{K}_2\text{O} + \text{Na}_2\text{O})/\text{Al}_2\text{O}_3 < 1$) or perpotassic (most DAIR have molar

$K_2O/Al_2O_3 < 1$). The DAIR contain several minerals (low-Na mica, Ti-poor richterite, K-triskaidecatitanate, monazite, ilmenite, rutile and carbonates) that are uncharacteristic of lamproite, but typical of orangeite. DAIR samples also differ petrographically from sanidine richterite lamproites (those closest to evolved members of the orangeite clan) in that the latter are rich in leucite, and typically lack olivine, perovskite, spinel and calcite (e.g. Mitchell and Bergman, 1991). However, some mica compositional evolutionary trends (notably Al_2O_3 versus TiO_2 ; Fig. 4a) are typical of lamproite; others, such as Al_2O_3 vs. FeO^* (Fig. 4b), resemble those of orangeite. Sodium contents in micas from DAIR samples are unusually low for lamproite (typically < 0.5 wt.% Na_2O), but comparable to those of orangeite micas. The compositions of other minerals are less useful discriminants. For example, K-richterites in the DAIR contain 0.03–0.24 atoms Ti/23 oxygens (Rock *et al.*, 1992), less than the proposed lower limit for richterite in lamproite (0.25 atoms Ti/23 oxygens; Mitchell and Bergman, 1991). Low Ti contents are characteristic of richterites from the Pniel orangeite (Tainton, 1992), but cannot be considered definitive of orangeite. For example, amphiboles in certain of the Sover North and Besterskraal orangeites have high, lamproite-like Ti contents of > 0.25 atoms Ti/23 oxygens (Tainton, 1992; Mitchell, 1995).

The DAIR were emplaced at ~ 114 – 113 Ma, during which time southern Africa and eastern India were several thousands of kilometres apart (see, for example, the plate reconstructions of Müller *et al.*, 1993). Clearly, the Damodar Valley is not an extension of the South African orangeite province and represents a magmatic field in its own right. Is the similarity in age therefore coincidental? The answer is likely to be 'yes', assuming a similar history for the mantle sources of African orangeites and the DAIR. Mitchell (1995) proposed that African orangeites might be derived from ancient (> 1 Ga) metasomatised lithospheric mantle. The DAIR are thought to have been produced from a deep lithospheric source similar in mineralogical character and mode (see Middlemost *et al.*, 1988) to that proposed for orangeites, suggesting that the root of the Singhbhum craton (or the cratonised Chotanagpur mobile belt) evolved in a manner similar to that of the Kaapvaal craton. Coincidentally, both cratons were heated by large mantle plumes (the Tristan and Kerguelen

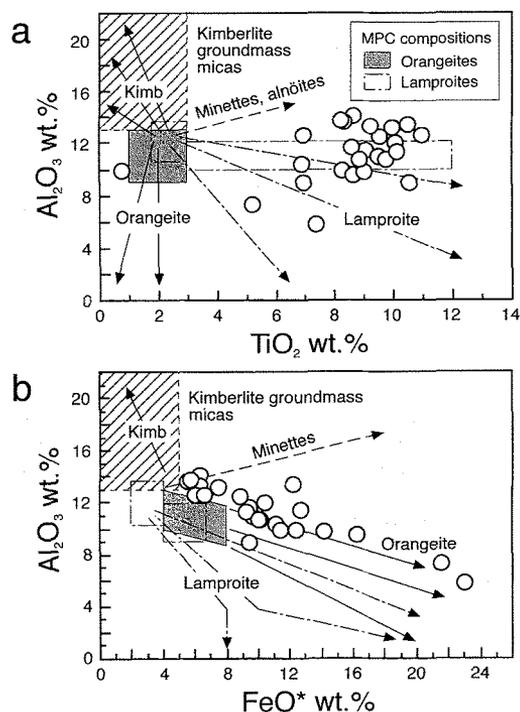


FIG. 4. (a) Al_2O_3 versus TiO_2 variation; and (b) Al_2O_3 versus FeO^* (total Fe) variation in phlogopite from the Damodar alkaline igneous rocks, compared to compositional evolutionary trends of micas from other ultramafic–mafic alkaline igneous rocks. Damodar mica data from this study, Middlemost *et al.* (1988) and Rock *et al.* (1992); South African orangeite and kimberlite mica fields from Mitchell (1995); lamproite, minette and alnöite mica fields from Mitchell and Bergman (1991). MPC = microphenocryst.

hot spots, respectively) during the early Cretaceous, permitting the generation of ultramafic–mafic alkaline magmas.

Summary and conclusions

In summary, the DAIR possess many of the features of archetypal orangeites and, coincidentally, are of a comparable age (114–113 Ma). The close mineralogical affinities to African orangeites support the contention that these rocks are members of the orangeite clan, rather than minettes or lamproites (*contra* Middlemost *et al.*, 1988; Rock *et al.*, 1992). This suggestion is also supported by new chemical and Pb–Nd–Sr

isotopic data (to be reported in detail elsewhere), which suggest a close kinship between DAIR samples and African orangeites. Thus, we conclude that orangeites are not unique to the Kaapvaal craton, as has been suggested by Skinner (1989), Mitchell (1995) and others. The implications of this finding are beyond the scope of this study, but it is noteworthy that the tectonic setting of the DAIR is directly comparable to that of the Prieska orangeites, which straddle the margin of the Kaapvaal craton (see Skinner *et al.*, 1994; Smith *et al.*, 1994).

Finally, if the DAIR are true orangeites, there is a high probability that they have interacted with diamond-bearing horizons in the deeper parts (>150 km) of the cratonised mobile belt surrounding the Singhbhum craton, or the root of the craton itself. Rock *et al.* (1992) considered the DAIR to be 'potentially diamondiferous', but to our knowledge, no sampling for diamond has been undertaken. Additional studies are underway to evaluate whether diamond xenocrysts occur in the DAIR, or have been lost through gravitational fractionation, resorption or graphitisation.

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