$p)^{0.286}$, where T is temperature in kelvin. Potential temperature is conserved under adiabatic conditions. For the low temperatures characteristic of the winter polar vortices, Θ = 465 K corresponds to about p = 50 hPa and Θ = 585 K corresponds to about p = 22 hPa.
18. Potential vorticity (PV) is defined as PV = $-g(f + \zeta)\, d\Theta/dp$, where g is the acceleration due to gravity, f is the Coriolis parameter, p is pressure, Θ is potential temperature (*17*), and ζ is the component of relative vorticity orthogonal to the Θ surface. For adiabatic, frictionless flow, PV is conserved, and contours of PV on potential temperature surfaces comprise the same air parcels.
19. M. R. Schoeberl and D. L. Hartmann, *Science* **251**, 46 (1991).
20. The MLS pointing geometry and the inclination of the UARS orbit lead to measurement coverage from 80° on one side of the equator to 34° on the other. The UARS orbit plane precesses in such a way that all local solar times are sampled in about 36 days, after which the spacecraft is rotated 180° about its yaw axis. Thus, 10 times per year, MLS alternates between viewing northern and southern high latitudes.
21. J. B. Kumer, J. L. Mergenthaler, A. E. Roche, *Geophys. Res. Lett.* **20**, 1239 (1993).
22. M. Loewenstein, J. R. Podolske, K. R. Chan, S. E. Strahan, *J. Geophys. Res.* **94**, 11589 (1989); *Geophys. Res. Lett.* **17**, 477 (1990).
23. D. W. Fahey *et al.*, *J. Geophys. Res.* **94**, 16665 (1989).
24. J. M. Rosen, S. J. Oltmans, W. F. Evans, *Geophys. Res. Lett.* **16**, 791 (1989); P. Hamill and O. B. Toon *ibid.* **17**, 441 (1990); G. Hübler *et al.*, *ibid.*, p. 453; S. R. Kawa, D. W. Fahey, L. C. Anderson, M. Loewenstein, K. R. Chan, *ibid.*, p. 485; D. W. Fahey *et al.*, *Nature* **344**, 321 (1990); D. J. Hofmann and T. Deshler, *J. Geophys. Res.* **96**, 2897 (1991).
25. R. J. Salawitch, G. P. Gobbi, S. C. Wofsy, M. B. McElroy, *Nature* **339**, 525 (1989).
26. G. C. Toon *et al.*, *J. Geophys. Res.* **94**, 16571 (1989).
27. We cannot specify exactly when the decrease in H_2O commenced because the MLS 183-GHz radiometer used to measure H_2O was not operational for most of the second half of June and the first part of July 1992 because of a UARS solar array drive problem.
28. K. K. Kelly *et al.*, *J. Geophys. Res.* **94**, 11317 (1989); K. K. Kelly *et al.*, *Geophys. Res. Lett.* **17**, 465 (1990).
29. M. Bithell, L. J. Gray, J. E. Harries, J. M. Russell, A. Tuck, *J. Atmos. Sci.* **51**, 2942 (1994).
30. Climate Analysis Center, National Meteorological Center, *Southern Hemisphere Winter Summary, 1992, Selected Indicators of Stratospheric Climate* (National Oceanic and Atmospheric Administration, Washington, DC, 1992).
31. G. L. Manney *et al.*, *Geophys. Res. Lett.* **21**, 2405 (1994).
32. G. L. Manney *et al.*, *ibid.*, p. 813.
33. W. H. Brune *et al.*, *Science* **252**, 1260 (1991); J. Austin, N. Butchart, K. P. Shine, *Nature* **360**, 221 (1992).
34. The ClO abundances are much smaller on the "night" side of the orbit because of a lack of photolysis of the dimer Cl_2O_2, formed from ClO recombination.
35. We thank M. R. Schoeberl for encouraging us to pursue the MLS HNO_3 measurement; A. E. Roche and the CLAES team for N_2O data and for discussions and comparisons of HNO_3 data; National Meteorological Center (NMC) Climate Analysis Center personnel, especially M. E. Gelman and A. J. Miller, for making NMC data available to the UARS project; two anonymous referees for helpful comments; and our MLS colleagues, especially T. Lungu, R. P. Thurstans, and E. F. Fishbein. M.L.S. gratefully acknowledges a Resident Research Associateship from the National Research Council. The work at the Jet Propulsion Laboratory, California Institute of Technology, was sponsored by the National Aeronautics and Space Administration, and the work at Edinburgh University and Heriot-Watt University was sponsored by the Science and Engineering Research Council.

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Timing of Hot Spot–Related Volcanism and the Breakup of Madagascar and India

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Widespread basalts and rhyolites were erupted in Madagascar during the Late Cretaceous. These are considered to be related to the Marion hot spot and the breakup of Madagascar and Greater India. Seventeen argon-40/argon-39 age determinations reveal that volcanic rocks and dikes from the 1500-kilometer-long rifted eastern margin of Madagascar were emplaced rapidly (mean age = 87.6 ± 0.6 million years ago) and that the entire duration of Cretaceous volcanism on the island was no more than 6 million years. The evidence suggests that the thick lava pile at Volcan de l'Androy in the south of the island marks the focal point of the Marion hot spot at ~88 million years ago and that this mantle plume was instrumental in causing continental breakup.

Many of Earth's flood basalt provinces occur at the rifted margins of continents and can be linked to the initiation of oceanic hot spot tracks (*1*, *2*). Flood basalt provinces have been interpreted as signaling the arrival of a new mantle plume (*3*) or, alternatively, as forming by extension over a preexisting shallow region of hot spot mantle (*2*), developed by the steady upwelling and accumulation of plume material beneath continental lithosphere (*4*). Debate also surrounds the question of the role that mantle plumes play in causing continental breakup (*5*).

The widespread Cretaceous flood basalts of Madagascar can be related to the track of the Marion hot spot (Fig. 1) (*6*); however, the precise timing of the volcanism is poorly known. Potassium-argon dates range from 31 to 97 million years ago (Ma) (*7*), whereas paleontological evidence (*8*, *9*) suggests a Maastrichtian [65.4 to 71.3 Ma (*10*)] to Turonian age range (88.7 to 93.3 Ma) for the igneous activity. Possible correlative rocks are the numerous Cretaceous mafic dikes that are found in southwest India (*11*). In this report, we present high-precision $^{40}Ar/^{39}Ar$ age determinations for the Cretaceous volcanic and dike rocks of Madagascar. Plate-tectonic considerations favor an active, rather than a passive, role for the Marion plume in the generation of this flood basalt province and the breakup of Madagascar and India.

The Cretaceous volcanic and intrusive rocks of Madagascar crop out semi-continuously along the 1500-km length of the east coast, which marks the rifted margin, and in the Majunga and Morondava basins in western Madagascar (Fig. 2). The rocks include basalt flows and dikes and some rhyolite flows. Along the rifted margin the flows lie mainly on Precambrian basement, whereas most of the dikes are parallel to the coast. Whether there are any seaward-dipping basalt flows offshore is unknown; however, the steepness of the rifted margin precludes the existence of a large volcanic wedge of the type that typifies most other volcanic passive margins, such as East Greenland (*12*). The sedimentary basins in western Madagascar record episodes of rifting and subsidence since the late Paleozoic (*9*, *13*). Basalt flows crop out for 700 km in the northwest and 200 km in the southwest and are interbedded with Upper Cretaceous sedimentary rocks. In the Majunga Basin the flows are reported to occur between post-Cenomanian (93.3 to 98.5 Ma) sandstone and fossiliferous pre-Upper Turonian limestone (*8*). A dike swarm crops out adjacent to the basalt flows in the northern part of the Morondava Basin. The Volcan de l'Androy complex in southern Madagascar contains the thickest sequence of Cretaceous volcanic rocks exposed on the island; this massif is about 50 km wide by 90 km long and consists of more than 1.5 km of interbedded basalt and rhyolite, with microgranite outliers (*14*). To the west is the Ejeda-Bekily dike

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swarm, which trends toward the Volcan de l'Androy massif.

We dated rocks from along the length of the rifted east coast margin, from flows in the Majunga and Morondava sedimentary basins, the Volcan de l'Androy massif, and the associated Ejeda-Bekily dike swarm. The $^{40}Ar/^{39}Ar$ age determinations were carried out by both conventional whole-rock step heating and the laser fusion method on single and multiple feldspar grains (*15*) (Table 1). Seven whole-rock samples gave well-defined age plateaus in the step-heating experiments (*16*) (Fig. 3). The proportion of ^{39}Ar released in the plateau steps varied from 69 to 100%. Samples MJ92-10 and SAM92-1 had anomalously young ages for the lowest temperature step, suggesting some alteration-related loss of radiogenic ^{40}Ar. Samples with suspected excess ^{40}Ar, such as MAN90-37, are indicated by saddle-shaped age-spectra patterns (*17*) and isotope correlation (inverse isochron) diagrams in which the best fit, least squares line has an $^{40}Ar/^{36}Ar$ intercept greater than the atmospheric value of 295.5. Despite the presence of excess ^{40}Ar, however, the whole-rock plateau ages for MAN90-37 and MAN90-47 are concordant with the inverse isochron ages. The other whole-rock sample with excess ^{40}Ar, TAM92-21A, yielded a well-defined age plateau (85.9 ± 0.9 Ma) but an imprecise inverse isochron age, due to the close grouping of Ar compositions for individual steps in the isotope correlation plot.

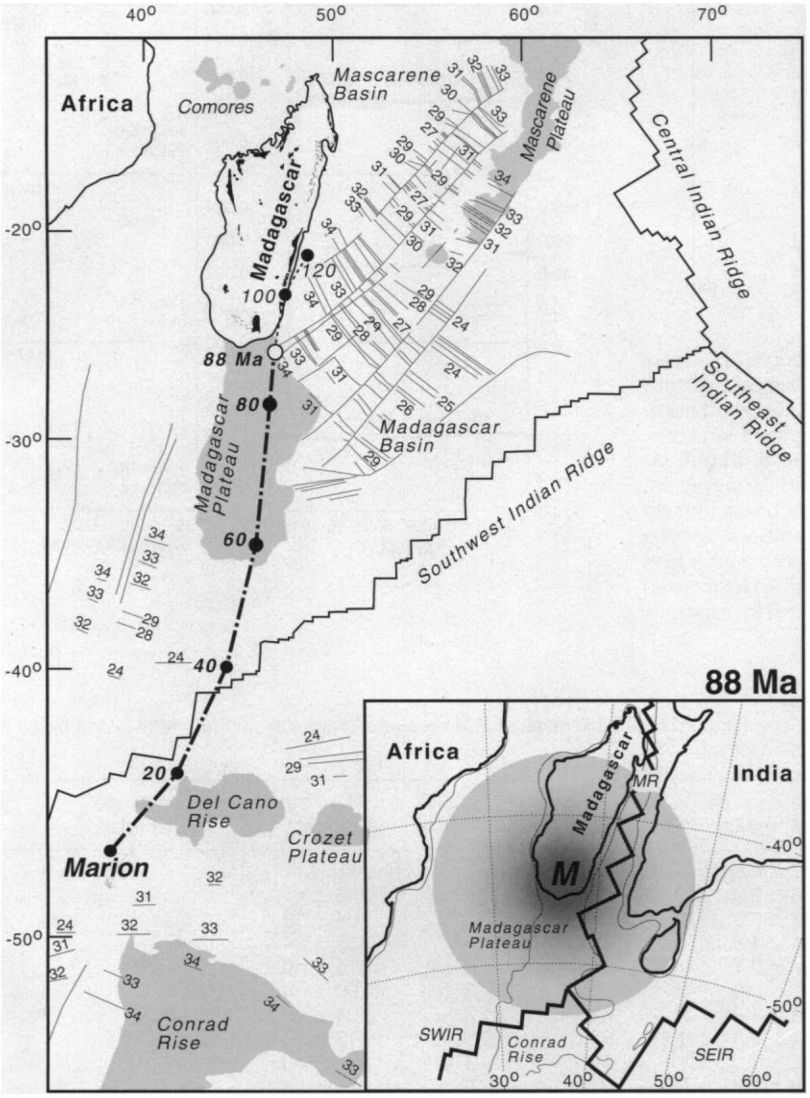
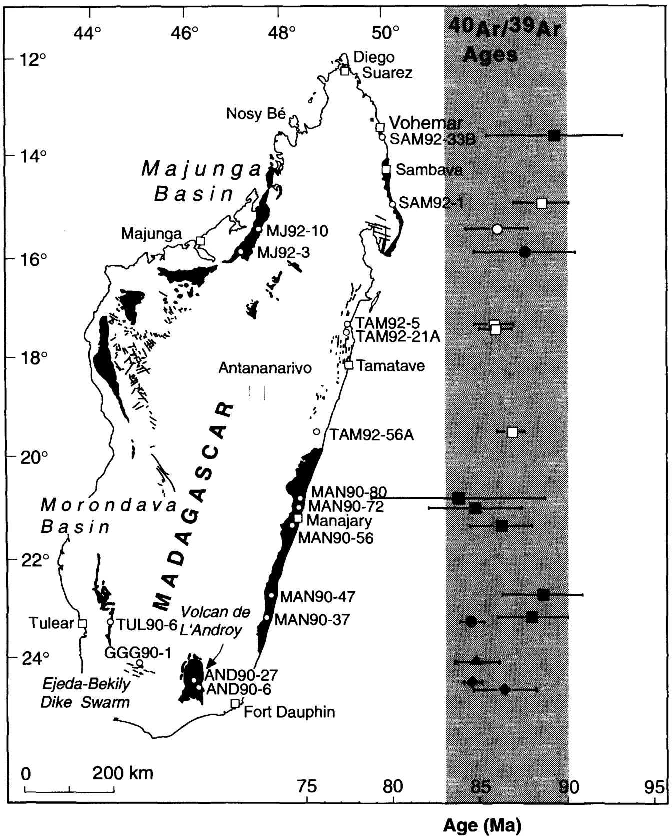


Fig. 1. Madagascar and the principal structures in the southwestern Indian Ocean. Magnetic lineations in the Mascarene and Madagascar basins are from Dyment (*18*). We calculated the Marion hot spot track relative to the African plate using the model of Müller *et al.* (*23*). The hypothetical track (88 to 120 Ma) is shown by the dashed line. The longitudinal error in the model for the hot spot track after 84 Ma is on the order of several hundred kilometers. The uncertainties increase for the older portion of the track. (Inset) A plate reconstruction for 88 Ma showing the relative positions of Africa, India, and Madagascar and the paleoridge system: Mascarene Ridge (MR), Southwest Indian Ridge (SWIR), and Southeast Indian Ridge (SEIR). The shaded circle illustrates the possible extent of the Marion plume, with M marking the proposed center of the hot spot beneath Volcan de l'Androy.

The $^{40}Ar/^{39}Ar$ age data reveal that the Madagascar Cretaceous volcanic province formed over no more than ~6 million years, a much shorter time interval than indicated by the published K-Ar dates. The volcanic rocks of the rifted margin show virtually no statistically significant differences in age (Fig. 2); the weighted mean of the inverse isochron ages is 87.6 ± 0.6 Ma. Two tholeiitic basalt flows (MJ92-3 and MJ92-10) from the Majunga Basin gave ages of 87.6 ± 2.9 Ma and 88.5 ± 1.3 Ma, respectively, which are within error of the age determinations for the rifted-margin rocks. A rhyolite dike (AND90-6) intruded into the base of the Volcan de l'Androy lavas, and a small dolerite intrusion (AND90-30) from near the top of the massif gave $^{40}Ar/^{39}Ar$ ages of 86.3 ± 1.9 Ma and 86.3 ± 1.0 Ma, respectively, statistically indistinguishable from the mean of the rifted-margin and the Majunga Basin dates. A slightly younger age of 84.4 ± 0.4 Ma was given by sanidine phenocrysts from a rhyolite flow (AND90-27) at the top of Volcan de l'Androy. Similar young ages were also given by a basaltic member of the Ejeda-Bekily dike swarm (GGG90-1 = 84.8 ± 1.3 Ma) and by a basalt flow from the southwestern part of Morondava Basin (TUL90-6 = 84.5 ± 0.7 Ma). These data suggest that Madagascan volcanism ceased first in the north.

Sea-floor spreading in the Mascarene Basin in the Late Cretaceous resulted in the separation of Madagascar and Greater India. The oldest magnetic lineations recognized in the southern Mascarene Basin are anomalies 33, 33r, and 34 (*18*). They trend to the northwest and are displaced by a series of northeast-southwest–trending fracture zones (Fig. 1). Anomaly 34 is less than ~50 km to as far as ~200 km from the continental edge of southeastern Madagascar, as determined from satellite gravity data. The linearity of the rifted east coast margin is suggestive of strike-slip faulting before the opening of the Mascarene Basin (*18*). This would have provided a zone of weakness, as would the strong anisotropy in the basement fabric (*9*), during the breakup of Madagascar and India. Because breakup occurred in the Cretaceous quiet zone, the precise age of the rift-drift transition cannot be determined from magnetic lineations; however, sea-floor spreading was clearly well organized in the southern Mascarene Basin shortly before 84 Ma, the age of anomaly 34 (*19*). The establishment of regular sea-floor spreading in the Mascarene Basin, therefore, appears to have been essentially



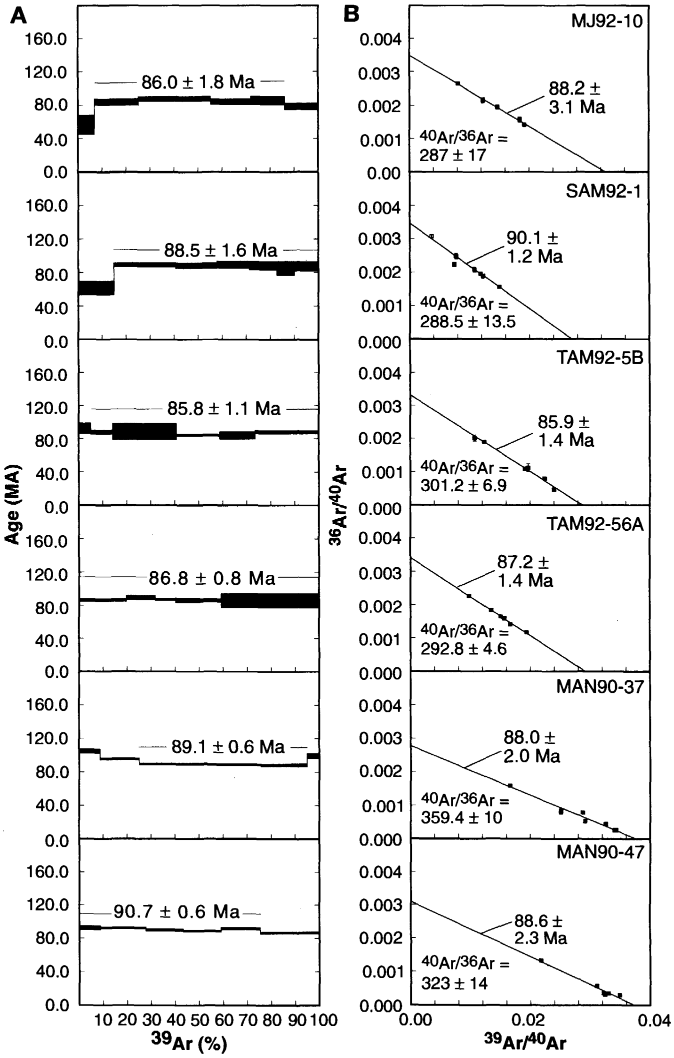


Fig. 2. (**Left**) Outcrop of Madagascar Cretaceous volcanic rocks (black) and new $^{40}Ar/^{39}Ar$ age determinations with analytical uncertainties. Sample symbols: squares = rifted eastern margin; circles = Majunga and Morondava basins; diamonds = Volcan de l'Androy; triangle = Ejeda-Bekily dike (filled symbol = inverse isochron age; open symbol = plateau age). **Fig. 3.** (**Right**) (**A**) Whole-rock age spectra showing apparent age as a function of the cumulative fraction of ^{39}Ar released. The vertical width of the horizontal boxes indicate estimated analytical error ($\pm 1\sigma$) about each step age. Note that the age variation between some individual "plateau" steps for MAN90-47 is slightly greater than $\pm 1\sigma$. (**B**) Isotope correlation diagrams ($^{36}Ar/^{40}Ar$ versus $^{39}AR/^{40}Ar$) for the step Ar compositions measured. The inverse isochron age is calculated from the best fitting line through selected step gas compositions.

Table 1. $^{40}Ar/^{39}Ar$ ages for Madagascar basalts and rhyolites. The standard FCT-3 biotite at 27.8 Ma is equivalent to the standard Mmhb-1 at 520.4 Ma (*27*). MSWD, mean square of weighted deviates (*28*). N.A., not applicable; see text.

Sample number	Sample type	Inverse isochron age $\pm 1\sigma$ (Ma)	*n*	MSWD	$^{40}Ar/^{36}Ar$ intercept	Plateau age $\pm 1\sigma$ (Ma)	^{39}Ar (%)
		Phenocryst total fusion ages by laser ablation (standard = Mmhb-1)					
MJ92-3	Plagioclase (basalt)	87.6 ± 2.9	8	0.8	295 ± 5		
SAM92-33B	Plagioclase (basalt)	89.3 ± 3.9	5	1.2	404 ± 19		
MAN90-56	Plagioclase (basalt)	86.2 ± 1.8	4	0.4	332 ± 34		
MAN90-72	Plagioclase (basalt)	84.7 ± 2.7	6	1.5	414 ± 40		
MAN90-80	Plagioclase (basalt)	83.7 ± 5.0	6	0.6	290 ± 14		
AND90-6	Plagioclase (rhyolite)	86.3 ± 1.9	7	1.2	304 ± 28		
AND90-27	Sanidine (rhyolite)	84.4 ± 0.5	6	0.5	312 ± 8		
AND90-30	Plagioclase (dolerite)	86.3 ± 1.0	5	0.3	280 ± 60		
GGG90-1	Plagioclase (basalt)	84.8 ± 1.3	9	1.0	290 ± 21		
TUL90-6	Plagioclase (basalt)	84.5 ± 0.7	4	2.5	Tied to air		
		Whole-rock step heating experiments (standard = FCT-3 biotite)					
MJ92-10	Basalt	88.5 ± 1.3	4	0.4	286 ± 3	86.0 ±1.8	79
SAM92-1	Basalt	90.1 ± 1.2	5	0.02	289 ± 14	88.5 ± 1.6	85
TAM92-5B	Basalt	85.9 ± 1.4	6	2.1	301 ± 7	85.8 ± 1.1	95
TAM92-21A	Basalt	N.A.	N.A.	N.A.	N.A.	85.9 ± 0.9	74
TAM92-56A	Rhyolite	87.2 ± 1.4	6	0.4	293 ± 5	86.8 ± 0.8	100
MAN90-37	Basalt	88.0 ± 2.0	7	5.9	359 ± 10	89.1 ± 0.6	69
MAN90-47	Basalt	88.6 ± 2.3	6	10.3	323 ± 14	90.7 ± 0.6	75

synchronous with the eruption of the Madagascar flood basalt province.

Extension-driven melting (*2*) of the Marion plume provides an explanation for the magmatism along the long-rifted eastern margin of Madagascar and the possibly equivalent activity on the southwestern coast of India. The apparent lack of a prominent basalt wedge offshore may be due to the fact that the Marion plume was offset to the south of the rifted margin during early sea-floor spreading in the Mascarene Basin. Rare outliers of Cretaceous volcanic rocks in central Madagascar suggest that the basalts in the Majunga and Morondava basins may have been erupted from vents along the continental rift of east Madagascar and flowed 300 to 400 km westward. Geochemical similarities between some of the east coast and west coast basalts (*20*) and the fact that flood basalt lavas may travel great distances (>750 km) (*21*) also provide some support for this idea. This interpretation implies that the rift structure must have been strongly domed in the east. Alternatively, the lavas in the basins of western Madagascar may reflect areas of thin lithosphere that made possible local melting of the plume (*22*).

The occurrence of by far the thickest lava sequence (~1.5 km) at Volcan de l'Androy appears paradoxical. This massif is located more than 100 km west of the continental rift in an area that has undergone little crustal extension. The flows appear to have been derived locally, as indicated by micrograniten, which partly encircle the massif; the overall structure resembles a wide "sag" caldera. The plate motion model of Müller *et al.* (*23*) places the Marion hot spot about 100 km south of Madagascar at 88 Ma (Fig. 1). We propose that Volcan de l'Androy marks the center of the Marion hot spot at this time—that is, within the errors of the model position—and conclude that the plume generated flood basalt magmas in both intraplate and continental-rift environments beneath Madagascar.

Between the opening of the Mascarene Basin and the time of anomaly 34, the Marion plume also seems to have interacted with the triple junction of the Southwest Indian Ridge (SWIR), the Southeast Indian Ridge, and the nascent Mascarene Ridge, as indicated by the presence of large oceanic plateau edifices (*24*) that are flanked by anomaly 34 (Fig. 1). The Conrad Rise and the western and southern part of the Northern Madagascar Plateau are conjugate with respect to the SWIR; the eastern part of the Northern Madagascar Plateau and the bathymetric high on the west side of the southern tip of India (*25*) are conjugate with respect to the Mascarene Ridge, as suggested by the structural trends of these features (*18*). Collectively, the Madagascar flood basalts and these oceanic plateaus constitute a single large igneous province (*26*) with a probable original area of more than 1×10^6 km^2.

Flood basalt provinces may signal the arrival of new mantle plumes beneath the continental lithosphere; however, there must be a period of thinning and removal of material above the mechanical boundary layer before the plume head can undergo significant melting (*4*). Therefore, distinguishing between the plume initiation and plume incubation models for the origin of continental flood basalts is problematical, because the two models are not entirely mutually exclusive. A more pertinent question is whether plumes play an active, rather than a passive, role in flood basalt volcanism and continental breakup. In the passive scenario (*2*), continental breakup allows plume mantle to upwell and melt, causing flood basalt volcanism; however, this model does not account for the voluminous eruption of lavas at Volcan de l'Androy, through Archean lithosphere well to the west of the continental rift. In the active scenario, the trigger for breakup itself may have been the northward propagation of the Mascarene Ridge as a consequence of the capture at ~88 Ma of the spreading ridge system to the south by the Marion hot spot, as the trailing edge of the Madagascar lithosphere passed north of the plume head's axis. Whether the Marion hot spot existed before 88 Ma is unknown; however, the postulated track of the hot spot (88 to 120 Ma) closely parallels the rifted eastern margin of Madagascar (Fig. 1), along which significant strike-slip faulting probably occurred before breakup (*18*). Thus, even if the plume existed well before ~88 Ma, it could have played an active role in the breakup of Madagascar and India by providing a preferred path for propagation of the Mascarene Ridge.

REFERENCES AND NOTES

1. W. J. Morgan, in *The Sea*, C. Emiliani, Ed. (Wiley, New York, 1981), vol. 7, pp. 443–487.
2. R. S. White and D. P. McKenzie, *J. Geophys. Res.* **94**, 7685 (1989).
3. M. A. Richards *et al.*, *Science* **246**, 103 (1989).
4. R. W. Kent, *Geology* **19**, 19 (1991); ______, M. Storey, A. D. Saunders, *ibid.* **20**, 891 (1992); A. D. Saunders *et al.*, in *Magmatism and the Causes of Continental Break-Up*, B. C. Storey, T. Alabaster, R. J. Pankhurst, Eds. (Spec. Publ. no. 68, Geological Society of London, London, 1992), pp. 41–60.
5. R. I. Hill, *Earth Planet. Sci. Lett.* **104**, 398 (1991).
6. J. Mahoney, C. Nicollet, C. Dupuy, *ibid.*, p. 350.
7. J. Dostal, C. Dupuy, C. Nicollet, J. M. Cantagrel, *Chem. Geol.* **97**, 199 (1992); K. M. Storetvedt *et al.*, *Afr. Earth Sci.* **15**, 237 (1992).
8. H. Besairie and M. Collignon, *Ann. Geol. Madagascar* **1972**, 35 (1972).
9. J. Boast and A. E. M. Nairn, in *The Ocean Basins and Margins*, A. E. M. Nairn and F. G. Stehli, Eds. (Plenum, New York, 1982), pp. 649–696.
10. J. D. Obradovich, *Geol. Assoc. Can. Spec. Pap.* **39**, 379 (1993).
11. A. Kumar *et al.*, *Proc. Indian Acad. Sci.* **97**, 107 (1988); T. Radhakrishna *et al.*, in *Mafic Dykes and Emplacement Mechanisms*, A. J. Parker, P. C. Ringwood, D. H. Tucker, Eds. (Balkema, Rotterdam, 1990), pp. 363–372; T. Radhakrishna *et al.*, *Earth Planet. Sci. Lett.* **121**, 213 (1994).
12. H. C. Larsen *et al.*, in *Early Tertiary Volcanism and the Opening of the NE Atlantic*, A. C. Morton *et al.*, Eds. (Spec. Publ. no. 39, Geological Society of London, London, 1988), pp. 95–114.
13. H. Besairie, *Serv. Geol. Madagascar Rep. 172* (1966).
14. R. Battistini, *Bull. Geol. Soc. Fr.* **7**, 187 (1959).
15. For the whole-rock step heating experiments, approximately 0.5 g of each sample was sealed in an evacuated quartz tube and irradiated in the core of the Oregon State University TRIGA reactor. The flux gradient and the efficiency of the conversion of ^{39}K to ^{39}Ar by neutron capture were monitored with samples of the biotite standard FCT-3 [27.8 ± 0.2 Ma; A. J. Hurfurd and K. Hammerschmidt, *Chem. Geol.* **58**, 23 (1985)]. We measured argon isotopic compositions on an AEI MS-10s mass spectrometer using procedures detailed in R. A. Duncan, *Proc. Ocean Drilling Program Sci. Results* **121**, 507 (1991). Feldspar grains (355 to 600 μm) for the laser heating experiments were leached in 4 M HCl (in an ultrasonic bath) and then washed in distilled water. Samples and the standard Mmhb-1 [520.4 ± 1.7 Ma; S. D. Samson and E. C. Alexander, *Chem. Geol.* **66**, 27 (1987)] were wrapped in Al foil, stacked in a quartz tube, and irradiated at the Ford reactor, University of Michigan, Ann Arbor. After an initial low-temperature heating step, using a defocused laser beam, we extracted Ar through fusion of one or more feldspar grains by firing short pulses of the laser (continuous Nd:yttrium-aluminum-garnet, maximum 17 W). The Ar isotopic composition was measured on an MAP215-50 equipped with a Johnson electron multiplier and ion-counting electronics.
16. A plateau is taken to consist of three or more sequential step ages with calculated ages that are statistically indistinguishable (±1σ) and was calculated from the weighting of each step age by the inverse of its variance to arrive at a weighted mean [R. A. Duncan, *J. Geophys. Res.* **89**, 9980 (1984)].
17. M. A. Lanphere and G. B. Dalrymple, *Earth Planet. Sci. Lett.* **32**, 141 (1976).
18. J. Dyment, thesis, Université Louis Pasteur de Strasbourg (1991).
19. W. B. Harland *et al.*, *A Geologic Time Scale* (Cambridge Univ. Press, Cambridge, 1990).
20. M. Storey *et al.*, unpublished data.
21. T. L. Tolan *et al.*, *Geol. Soc. Am. Spec. Pap.* **239**, 1 (1989).
22. R. N. Thompson and S. A. Gibson, *J. Geol. Soc. London* **148**, 973 (1991).
23. R. D. Müller *et al.*, *Geology* **21**, 275 (1993).
24. J. Goslin *et al.*, *Bull. Geol. Soc. Am.* **91**, 741 (1980); M. C. Sinha, K. E. Louden, B. Parsons, *Geophys. J. R. Astron. Soc.* **66**, 351 (1981).
25. A. S. Laughton, *General Bathymetric Chart of the Oceans* [Canadian Hydrographic Service (1975)], sheet 5-05 (map).
26. M. F. Coffin *et al.*, *Rev. Geophys.* **32**, 1 (1994).
27. M. A. Lanphere *et al.*, *Eos* **71**, 1658 (1990).
28. G. A. McIntyre *et al.*, *Geophys. Res.* **71**, 5459 (1966).

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