progressively in alternating magnetic fields. From the demagnetization curves the field ($H_{1,000}$) required to destroy 90% of the VRM was determined. The field needed to destroy a VRM is proportional to the logarithm of the acquisition time[15,16]. Therefore, it is possible to deduce the field ($H_{0\text{-}69M}$) that would be required to destroy a VRM acquired since the beginning of the Brunhes normal polarity epoch from $H_{0\text{-}69M} = H_{1,000} \times \log(0.69 \text{ m.y.})/\log(1{,}000 \text{ h}) = H_{1,000} \times 3.26$ (see Table 1).

The data indicate that both the intensity and the stability of the natural remanent magnetization of the hole 57 basalt can be accounted for by a magnetization which is almost entirely viscous, acquired since the beginning of the Brunhes epoch. Viscous remanences such as these, requiring alternating fields of one or two hundred oersted for erasure, have previously been reported in other titanomagnetite-bearing igneous rocks[16]. It is interesting that in spite of the viscous nature of their remanences the samples had Königsberger ratios ranging from 2 to 7, contradicting the suggestion that ratios greater than 1 in igneous rocks imply single domain thermoremanent magnetization[17]. The presence of an appreciable coarse-grained titanomagnetite fraction suggests that the remanence is a multidomain type VRM.

In the Brunhes epoch there are several postulated short reversals, some of which may only represent excursions of the geomagnetic field rather than true reversals[18–22]. Some of these short events may be different recordings of the same event. Their durations are thought to be only 2,000–6,000 yr. The effect on the intensity of VRM that may have been acquired during the predominantly normal epoch would be to reduce it by only a few per cent, and their possible existence does not alter drastically the computed VRM for that period of time.

Available information for oceanic ridge systems that have associated linear magnetic anomaly patterns indicates that the underlying basalts possess stable magnetic properties that account for these anomalies. Most submarine basalt magnetic studies have been carried out on fine-grained ocean ridge tholeiites obtained from dredge hauls or deep sea drilling. The unstable character of the basalt from hole 57 is quite different from that found in ocean ridge basalts. The basalt is also different in that it is probably an oceanic island type (I. Ridley *et al.*, unpublished), more differentiated than ocean ridge tholeiites. The iron oxides have a slightly higher titanium content than usual and are coarser in grain size.

The basalts that form oceanic islands, island chains, seamounts and ocean bottom lava flows arising from volcanic activity distant from ridge systems are genetically different from oceanic ridge tholeiites. If the hole 57 basalt is typical of this group then unstable magnetic behaviour may be more common in oceanic basalts than previously supposed. Unstable and very low coercivity NRM has been observed in other DSDP basalts from the Eastern Atlantic[5] and VRM has been detected in DSDP basalts from the Caribbean Sea[6]. Where such basalts form the oceanic basement magnetic anomalies would only reflect topographic influence and local variations in magnetization contrast. A correlatable linear magnetic anomaly pattern would be absent.

Several explanations have been offered for the existence of regions of uncorrelatable magnetic anomalies, or quiet zones[23], and it may be that no unique explanation will serve for all these areas. None of the explanations is based upon measured rock magnetic properties. The results of this study of the magnetic instability in hole 57 basalts suggest another possible explanation which may be applicable to some quiet zones. The lack of significant anomalies in these quiet zones may be due to viscous remanent magnetization of the underlying crust.

This work was supported by NSF and ONR grants. I thank N. D. Opdyke, R. Larson, J. M. Ade-Hall and D. Kent for comments.

WILLIAM LOWRIE

Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York 10964

Received January 19, 1973.

1 Vine, F. J., and Matthews, D. H., *Nature*, **199**, 947 (1963).
2 Cox, A., and Doell, R. R., *J. Geophys. Res.*, **67**, 3997 (1962).
3 Lowrie, W., Løvlie, R., and Opdyke, N. D., *Earth Planet. Sci. Lett.* (in the press).
4 Phillips, J. D., and Forsyth, D., *Geol. Soc. Amer. Bull.*, **83**, 1579 (1972).
5 Lowrie, W., and Opdyke, N. D., in *Initial Reports of the Deep Sea Drilling Project*, **14**, 777 (US Government Printing Office, Washington, 1972).
6 Lowrie, W., and Opdyke, N. D., in *Initial Reports of the Deep Sea Drilling Project*, **15** (US Government Printing Office, Washington, in the press).
7 Ade-Hall, J. M., Aumento, F., Brooke, J., McKeown, D. L., Ryall, P. J. C., and Gerstein, R. E., *Canad. J. Earth Sci.* (in the press).
8 Fischer, A. G., *et al.*, *Initial Reports of the Deep Sea Drilling Project*, **6**, 493 (US Government Printing Office, Washington, 1971).
9 Melson, W. G., *et al.*, *Initial Reports of the Deep Sea Drilling Project*, **6**, 1119 (US Government Printing Office, Washington, 1971).
10 Wilson, R. L., and Watkins, N. S., *Geophys. J.*, **12**, 405 (1967).
11 Thellier, E., *Ann. Inst. Phys. Globe*, **16**, 157 (1938).
12 Shimizu, Y., *J. Geomag. Geoelect., Kyoto*, **11**, 125 (1960).
13 Néel, L., *Ann. Géophys.*, **5**, 99 (1949).
14 Néel, L., *J. Phys. Rad.*, **12**, 339 (1951).
15 Rimbert, F., *Rev. Inst. Franç. Pétrole*, **14**, 1 (1959).
16 Biquand, D., and Prevot, M., *Zeit. für Geofys.*, **37**, 471 (1971).
17 Stacey, F. D., *Earth Planet. Sci. Lett.*, **2**, 67 (1967).
18 Smith, J. D., and Foster, J. H., *Science*, **163**, 565 (1969).
19 Bonhommet, N., and Zahringer, J., *Earth Planet. Sci. Lett.*, **6**, 43 (1969).
20 Mörner, N. A., Lanser, J. P., and Hospers, J., *Nature Physical Science*, **234**, 173 (1971).
21 Barbetti, M., and McElhinny, M. W., *Nature Physical Science*, **239**, 327 (1972).
22 Bucha, V., *J. Geomag. Geoelect.*, **22**, 253 (1970).
23 Vogt, P. R., Anderson, C. N., Bracey, D. R., and Schneider, E. D., *J. Geophys. Res.*, **75**, 3955 (1970).

Comments on "Is the African Plate Stationary?"

Burke and Wilson[1] suggested that the African plate became stationary some 25 m.y. ago after having moved northeast from the Mid-Atlantic ridge since the opening of the South Atlantic. Like Rhodes[2] they believe that the continent drifted over hot spots or convective plumes in the mantle and that Walvis Ridge and the Discovery and Meteor seamount chains represent trails of hot spots marking the relative motion of Africa.

The west coast of Southern Africa is characterized by linear belts of plutonic and volcanic complexes; if these rocks were generated over plumes and the African plate has moved laterally over them they should appear progressively younger in the same direction, from east to west.

The hypothesis may therefore be tested by examining the isotopic age data for the lineaments. This has been done by Marsh[3] for the Upper Mesozoic alkaline igneous complexes of Angola and South West Africa (Fig. 1), demonstrating that no sensible pattern in the variation of ages exists. The available data range from 123 m.y. to 164 m.y. (refs. 4, 5) and do not support the mantle plume theory. Similar results were obtained for two E–W lines of alkaline complexes in Brazil[3] where the ages range from 132–80 m.y. and 80–51 m.y. respectively[6], showing no directional pattern where, according to the

hot spot hypothesis, younger ages should be expected from W to E.

Instead, Marsh[3] correlated the lineaments of alkaline igneous activity with transform faults now active, offsetting the Mid-Atlantic ridge, and concluded that the Upper Mesozoic complexes reflect the initial rifting of the Africa–South America plate some 130 m.y. ago.

The oldest lineament of igneous activity along the west coast of southern Africa consists of the Kuboos–Tatasberg–Bremen–Garub line (Fig. 1) and is lower Palaeozoic. The Kuboos Granite was dated at 550 ± 20 m.y. (ref. 7) and provisional data indicate an age of 560 m.y. for the Bremen complex further northeast. A pyroclastic sill associated with the carbonate-bearing diatremes at Garub yielded 500 ± 5 m.y., but it is uncertain whether this age represents the time of extrusion or a later regional thermal event (A. K., H. Allsop and H. Welke, unpublished). Nevertheless, no systematic change in age following the 250 km long lineament is apparent.

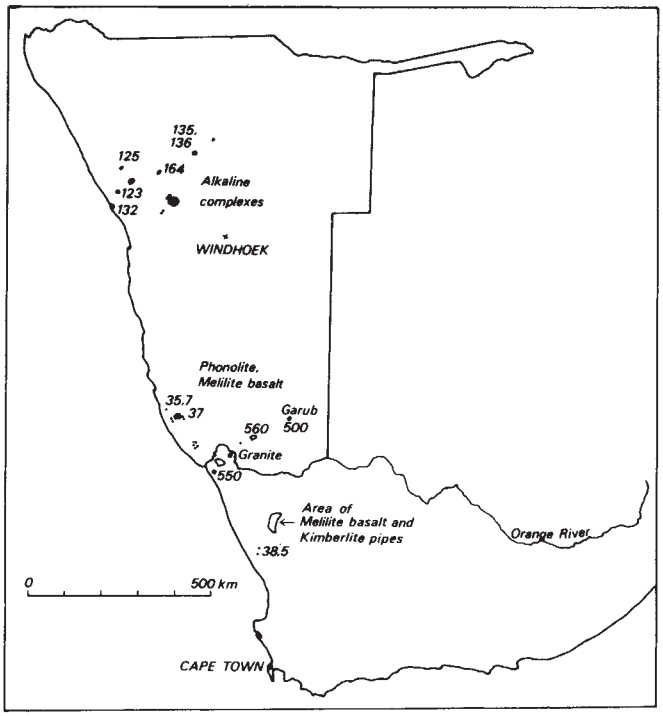

Fig. 1 The early Palaeozoic, Upper Mesozoic and Tertiary lines of volcanic complexes along the west coast of southern Africa. Numbers are radiometric ages in m.y.

Burke and Wilson[1] maintain that such surface structures can only be related to those of the underlying asthenosphere if a plate is temporarily at rest. The alkaline complexes of South West Africa clearly follow a NE trend which reflects the regional structures of the underlying Palaeozoic Damara Orogen, yet the African plate had already begun to drift at the time of alkaline igneous activity[8,9].

During the Oligocene the coastal areas of southern South West Africa and Namaqualand were affected by volcanic activity over a length of some 450 km (Fig. 1). The lines of phonolite and melilite basalt plugs are clearly oriented in a NNW direction and follow structural trends of Late Precambrian origin. Their ages range between 35.7 and 38.5 m.y. (J. A. S. Adams, personal communication) when, according to Burke and Wilson, the African plate was still in motion.

Clearly, lines of volcanic activity in southern Africa are related to older structures in the crust without the African plate necessarily being at rest. I believe that none of the lines mentioned here represents trails of hot spots and I am not aware of any line of volcanoes in Africa that shows a consistent age pattern in any direction.

A. KRÖNER

Precambrian Research Unit,
University of Cape Town

Received January 8, 1973.

1 Burke, K., and Wilson, J. T., *Nature*, **239**, 387 (1972).
2 Rhodes, R. C., *Tectonophysics*, **12**, 111 (1971).
3 Marsh, J. S., *Earth Planet. Sci. Lett.*, **18**, 317 (1973).
4 Siedner, G., and Miller, G. A., *Earth Planet. Sci. Lett.*, **4**, 451 (1968).
5 Manton, W. I., and Siedner, G., *Nature*, **216**, 1197 (1967).
6 Amaral, G., Bushee, J., Condani, U. G., Kawashita, K., and Reynolds, J. H., *Geochim. Cosmochim. Acta*, **31**, 117 (1967).
7 De Villiers, J., and Burger, A. J., *Annals. Geol. Surv. S. Afr.*, **6**, 83 (1967).
8 Dietz, R. S., and Holden, J. C., *Scient. Amer.*, **223**, 3, 30 (1970).
9 Francheteau, J., and Le Pichon, X., *Bull. Amer. Assoc. Petrol. Geol.*, **56**, 991 (1972).

Removal of Xenon and Radon from Contaminated Atmospheres with Dioxygenyl Hexafluoroantimonate, O_2SbF_6

RADIOACTIVE noble gases, such as ^{133}Xe, ^{135}Xe, ^{85}Kr, and ^{88}Kr, are formed in uranium fission and are released to the atmosphere by nuclear power plants and fuel reprocessing plants. Studies of the US Public Health Service have shown that a modern boiling-water reactor, for example, releases about 20 Ci of ^{85}Kr and 13,000 Ci of ^{133}Xe per year under normal operating conditions[1,2]. Under adverse operating conditions, the release rates may be much higher[1]. Of the physical methods for reducing these emissions[3–7], cryogenic distillation and charcoal adsorption are being developed for large-scale use.

Soon after the first noble gas compounds had been discovered[8–12], Pomeroy[13] suggested that chemical methods might be useful for trapping krypton and xenon isotopes. Slivnik[14] has studied reactions of krypton-xenon mixtures with fluorine for waste-gas treatment and has demonstrated that xenon can be separated from krypton by such reactions. Fluorine is not, however, a convenient reagent for this purpose, as it must be heated with the process gas, and the excess fluorine must be removed afterwards. I have shown that radon, the heaviest noble gas, can be collected by oxidation with liquid bromine trifluoride and solid halogen fluoride-metal fluoride complexes, such as ClF_2SbF_6, BrF_2SbF_6, BrF_2BiF_6, and $IF_4(SbF_6)_3$ (refs. 15 and 16). I report here further experiments with a dioxygenyl salt, O_2SbF_6, which reacts with radon and xenon at 25° C and which appears very promising as a reagent for removing both of these gases from contaminated atmospheres.

Dioxygenyl hexafluoroantimonate was prepared in 0.5–10 g amounts by photochemical reaction of oxygen, fluorine, and antimony pentafluoride[17]. Samples (0.5 g) of the compound, a white, crystalline powder, were exposed to samples of radon, xenon and krypton in 300 ml. 'Pyrex' bulbs on a metal vacuum line. Tracer ^{222}Rn (8.1 mCi) reacted with the powder immediately at 23–25° C, forming a nonvolatile radon compound. Xenon at 10–200 mm pressure also reacted immediately, forming a pale-yellow, translucent xenon compound. Oxygen was shown to be liberated in the xenon reaction by the fact that the residual gas was not completely condensible at −195° C. No reaction of krypton at 485–700 mm pressure was observed from 23° to 150° C.

Flow experiments were carried out with O_2SbF_6 powder and samples of air (0.33–0.76 l. at STP) which had been artificially contaminated with ^{222}Rn and ^{133}Xe. In each experiment, a rare gas/air mixture was passed through a glass U-tube packed with the powder, then through a trap cooled with liquid