Reports

Eocene Volcanism and the Origin of Horizon A

Abstract. *A series of closely time-equivalent deposits that correlate with seismic reflector horizon A exists along the coast of eastern North America. These sediments of Late-Early to Early-Middle Eocene age contain an authigenic mineral suite indicative of the alteration of volcanic glass. A volcanic origin for these siliceous deposits onshore is consistent with a volcanic origin for the cherts of horizon A offshore.*

Widespread sedimentary deposits throughout the western North Atlantic area contain mineralogical suites indicative of altered pyroclastic material and can be interpreted as the result of a series of Late-Early to Early-Middle Eocene volcanic events (*1*). Subsequent drilling during legs of the JOIDES (Joint Oceanographic Institutions for Deep Earth Sampling) program has shown the presence of hard, siliceous beds of common age in many of the deep oceanic areas sampled. These radiolarian-diatom and chert deposits, which often inhibit drilling operations, have been dated as Early to Middle Eocene (*2*). We suggest that these essentially time-equivalent deposits have resulted from the marine diagenetic alteration of volcanic material coupled with an increase in the productivity and preservation of siliceous microplankton in response to an increase in both silica and the nutrient phosphorus that result from the ash alteration process. Similar chert deposits have been found to be seismic reflecting horizons in the ocean basins, and the evidence is compelling that horizon A in the western North Atlantic is a result of this Eocene volcanism.

We originally found siliceous material in Eocene sediments while examining dredge hauls off Long Island in the Hudson Canyon, in other nearby canyons, and on the continental slope to the east along the Atlantic margin of the United States. The rocks are primarily chalks and limestones containing as much as 54 percent SiO_2 (by weight); most is in the form of opal-cristobalite with little of terrigenous origin. The samples of latest Early Eocene age contain the largest amounts of silica, and those of Paleocene, earliest and latest Eocene, and Oligocene age in the same area contain considerably less silica (*3*). Additional mineralogical examination showed the presence of the zeolite clinoptilolite and clay minerals of the montmorillonite group. In addition to other textural considerations the authigenic nature of the zeolites is demonstrated in Fig. 1, where crystal growth of clinoptilolite has incorporated a coccolith. Although alternative explanations have been proposed (*4*), the presence of an authigenic mineral suite containing clinoptilolite-heulandite, opal-cristobalite, and montmorillonite in varying proportions has usually been considered indicative of the alteration of fine-grained volcanic glass (*5*).

Examination of more samples from the western North Atlantic region and of published reports reveals the widespread nature of this Eocene mineralogical suite. Evidence is found in New Jersey in the Manasquan formation (Ash Marl) from mineralogical and paleontological data (*6*). In a study of Eocene clay minerals in the coastal plain of South Carolina, Heron (*7*) reported opal-cristobalite and montmorillonite but discounted a volcanic source in the absence of zeolites. New analyses (*8*) on coarser fractions in which zeolites are more common showed the presence of clinoptilolite in these sediments. Weaver (*9*) has reported this suite from the JOIDES cores taken on the Blake Plateau off the coast of Florida. Reynolds (*10*) has reported this suite from the Eocene of the Alabama Coastal Plain, and it has been observed, together with bentonites, in Mississippi by Grim (*11*) and by Wermund and Moiola (*12*) in beds as much as 30 m thick. The siliceous Toledo member of the Eocene Universidad formation in Cuba averages 10 to 15 m in thickness (*13*). Samples of these Cuban rocks obtained from the collection of Brönnimann and Rigassi in the Museum of Natural History at Basel, Switzerland (*14*), were x-rayed and found to contain major quantities of clinoptilolite, opal-cristobalite, and montmorillonite. Although somewhat difficult to evaluate because of the dilution effect of terrigenous material in some areas, there appears to be a general decrease in the amounts of siliceous materials from south to north.

The age relations inferred from the stratigraphy and the associated microfossils attest to the relatively restricted and similar time interval involved over the entire geographic range. The planktonic foraminiferal assemblages can be referred to the *Globorotalia aragonensis–Globorotalia palmerae* zone, and the nannoplankton assemblages belong to the *Discoaster lodoensis* and *Marthasterites tribrachiatus* zones. Chert beds of horizon A encountered in JOIDES drilling operations have been assigned to similar Early to Middle Eocene ages (*2*).

The Hudson Canyon and nearby offshore samples are dated by both planktonic foraminifera (*15*) and nannoplankton (*16*) and fall within these zones. The New Jersey strata contain planktonic foraminifera belonging to this interval (*6*), as do the cores from the JOIDES holes off Jacksonville, Florida (*17*). Both foraminifera and nannoplankton assign the Cuban material to these zones (*13*), whereas the Tallahatta formation in Mississippi and Alabama contains sparse amounts of nannoplankton assigned to this interval (*18*). The Black Mingo, Warley Hill, and Congaree formations in South Carolina are Late Paleocene through Middle Eocene in age (*7, 19*), but, because of deposition under shallow, nearshore conditions, planktonic zonations are not available.

The geographically widespread and closely time-equivalent nature of the siliceous sediments from varied depositional environments is thus established. Mineralogical evidence indicates that much of the siliceous composition is derived from the alteration of volcanic material with probable additional contributions in the deep sea from abundant siliceous microfossils. The association of siliceous sediments with radiolarian-diatom deposits is a common one (*20*), and the relationship here deserves discussion. The explanation that the silica for these cherts and related rocks

is derived primarily from the dissolution of biogenic opals lacks, not only an extensive radiolarian-diatom source of silica, but also a reason for their increased productivity.

The explanation that the presence of abundant silica in the water can account for the increased productivity of siliceous organisms is inadequate in itself, although increased silica abundance should act to slow down the dissolution rate of dead tests, thus increasing their apparent concentration in sediments (*21*). This process could also be influenced by the presence of heavier tests in Eocene radiolaria relative to those found in the Quaternary (*22*). Pyroclastic volcanic material, however, would not only provide additional silica but also the nutrient phosphorus. The average phosphorus content of volcanic rock types ranges from 0.10 to 0.45 percent P_2O_5 (*23*). Although meaningful quantitative estimates would be difficult to make, the relatively high solubility of the volcanic glass coupled with its fine particle size would undoubtedly have made available additional phosphorus for biogenic consumption over the geologic interval under discussion. The increased productivity would account for high concentrations of radiolarian-diatom oozes containing more robust tests, both resulting from the concomitant increase in available silica.

It has been suggested (*24*) that the initiation of cold, deep water circulation in the North Atlantic as a result of seafloor spreading in the early Cenozoic might augment the distribution of nutrients and thus the productivity of siliceous organisms. Although such changes in oceanic circulation might help to explain the oceanic cherts, they cannot also explain the extensive and time-equivalent nearshore sediments of volcanic origin on the continents or the relatively restricted time interval involved.

Sources for the volcanic materials are unknown, but, in view of the apparent general decrease in the thickness of siliceous deposits in the northerly direction together with the absence of significant Eocene volcanic activity in this same direction, the most likely source appears to be the then active Middle American and Caribbean region (*25*). In addition to atmospheric dispersal, the northward movement of the Florida and Greater Antillean currents and the Gulf Stream would distribute volcanic material in this general direction along the east coast of the United States and into the oceanic areas with relative uniformity when considered over the geologic time interval involved.

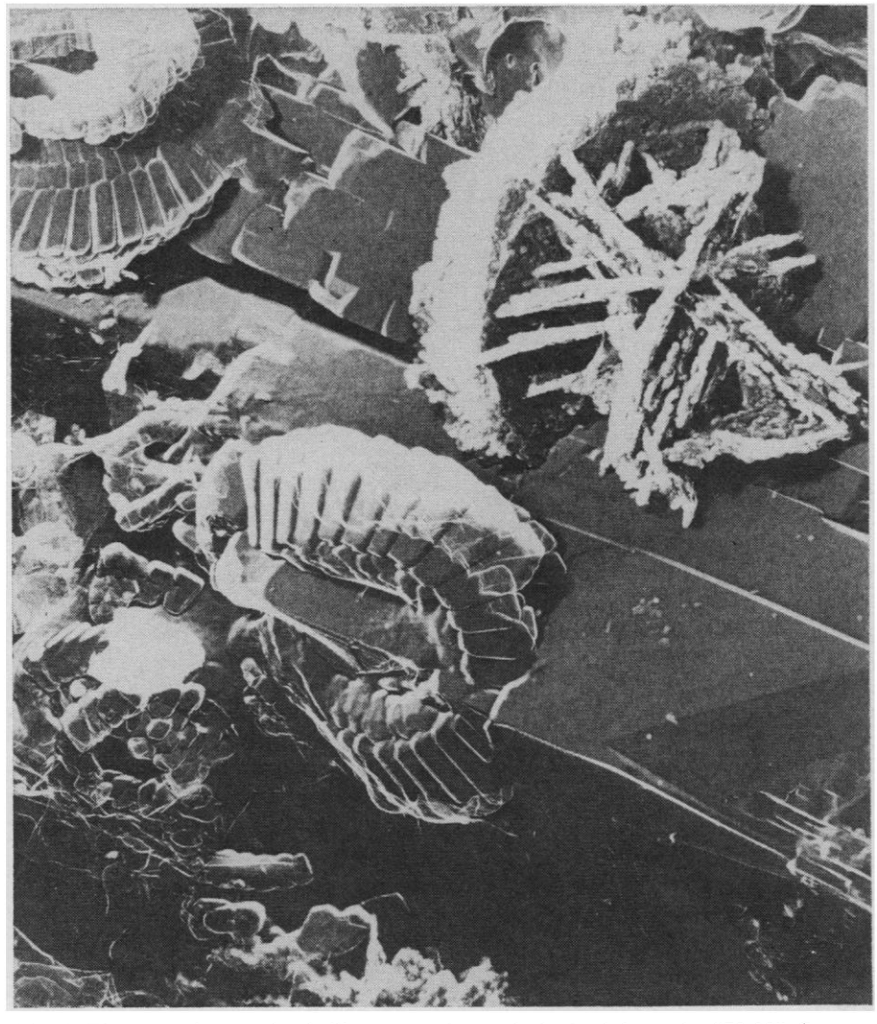

Fig. 1. Electron micrograph of siliceous marine rock dredged from the "70-30" Canyon in the Atlantic continental slope off Long Island. The rocks are the same age as seismic reflector horizon A in the western North Atlantic. The authigenic nature of the zeolite is shown by the clinoptilolite crystal that has incorporated a coccolith element during crystal growth (× 20,000).

Clear evidence of volcanism in the Late-Early and Early-Middle Eocene exists in the sediments all along the Atlantic and eastern Gulf coast of the United States. The presence of widespread siliceous deposits of similar age in the oceanic areas of the western North Atlantic cannot be considered fortuitous and of a different origin. Ewing *et al.* (*26*) have commented on the distribution and synchroneity of horizon A and speculated about the presence of correlative deposits on continental areas. The evidence presented here provides information on such deposits, information that bears on the origin of the cherts in question.

A volcanic origin provides a consistent explanation for the presence of these siliceous deposits in a wide variety of environments ranging from brackish and nearshore to deep marine over a wide geographic area. It also provides an explanation for the relatively restricted time interval involved.

THOMAS G. GIBSON
U.S. Geological Survey,
U.S. National Museum,
Washington, D.C. 20560

KENNETH M. TOWE
Department of Paleobiology,
Smithsonian Institution,
Washington, D.C. 20560

References and Notes

1. K. M. Towe and T. G. Gibson, *Geol. Soc. Amer. Annu. Meeting Program*, 1968, p. 299.
2. M. Ewing *et al., Initial Reports of the Deep-Sea Drilling Project* (Government Printing Office, Washington, D.C., 1969), vol. 1; M. N. A. Peterson *et al., Initial Reports of the Deep-Sea Drilling Project* (Government Printing Office, Washington, D.C., 1970), vol. 2; S. Gartner, Jr., *Science* **169**, 1077 (1970).

3. T. G. Gibson, *Geol. Soc. Amer. Bull.* **81**, 1813 (1970).
4. G. Brown, J. A. Catt, A. H. Weir, *Mineral. Mag.* **37**, 480 (1969).
5. M. N. Bramlette and E. Posnjak, *Amer. Mineral.* **18**, 167 (1933); K. S. Deffeyes, *J. Sediment. Petrol.* **29**, 602 (1959); J. C. Hathaway and P. L. Sachs, *Amer. Mineral.* **50**, 852 (1965); R. L. Hay, *Geol. Soc. Amer. Spec. Pap. 85* (1966).
6. R. L. C. Enright, Jr., personal communication (1968); thesis, Rutgers University (1969).
7. S. D. Heron, Jr., *Geol. Soc. Amer. Annu. Meeting Program* (1962), p. 71A; S. D. Heron, Jr., G. C. Robinson, H. S. Johnson, Jr., *S.C. State Develop. Board Div. Geol. Bull. 31* (1965).
8. We thank S. D. Heron, Jr., for restudying these sediments at our request and providing new x-ray data on their mineralogy.
9. C. E. Weaver, *Southeast. Geol.* **9**, 57 (1968).
10. W. R. Reynolds, *J. Sediment. Petrol.* **40**, 829 (1970)
11. R. E. Grim, *Miss. State Geol. Surv. Bull. 30* (1936).
12. E. G. Wermund and R. J. Moiola, *J. Sediment. Petrol.* **36**, 248 (1966).
13. P. Brönnimann and D. Rigassi, *Eclogae Geol. Helv.* **56**, 193 (1963).
14. We thank Dr. E. Gasche for providing these samples for study.
15. T. G. Gibson, J. E. Hazel, J. F. Mello, *U.S. Geol. Surv. Prof. Pap. 600-D* (1968), p. 222.
16. M. N. Bramlette, personal communication (1969).
17. JOIDES, *Science* **150**, 709 (1965).
18. M. N. Bramlette and F. R. Sullivan, *Micropaleontology* **7**, 129 (1961).
19. W. K. Pooser, *Univ. Kans. Paleontol. Contrib. Arthropoda*, Art. 8 (1965).
20. N. L. Taliaferro, *Calif. Univ. Dep. Geol. Sci. Bull.* **23**, 1 (1933); M. N. Bramlette, *U.S. Geol. Surv. Prof. Pap. 212* (1946).
21. W. R. Riedel, *Soc. Econ. Paleontol. Mineral. Spec. Publ.* **7**, 80 (1959).
22. T. C. Moore, Jr., *Geol. Soc. Amer. Bull.* **80**, 2103 (1969).
23. R. A. Daly, *Igneous Rocks and the Depths of the Earth* (McGraw-Hill, New York, 1933).
24. R. S. Dietz and J. C. Holden, *J. Geophys. Res.* **75**, 4939 (1970); W. A. Berggren and J. D. Phillips, *Symp. Geol. Libya*, in press.
25. H. J. MacGillavry, *Proc. Koninklijke Ned. Akad. Wetensch. Ser. B* **73**, 64 (1970).
26. J. Ewing, C. Windisch, M. Ewing, *J. Geophys. Res.* **75**, 5645 (1970).
27. We thank R. Cifelli, J. E. Hazel, and W. G. Melson for reviewing the manuscript. M. N. Bramlette provided valuable assistance in nannoplankton identification and biostratigraphy.

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Cosmic-Ray Tracks in Plastics: The Apollo Helmet Dosimetry Experiment

Abstract. *Counts of tracks from heavy cosmic-ray nuclei in helmets from Apollo missions 8 and 12 show variations caused by solar modulation of the galactic cosmic-ray flux. Specific estimates of the biological damage to certain nonreplaceable cells by track-forming particles during these space missions indicate that the fraction of deactivated cells could range from a lower limit of* 3×10^{-7} *to an upper limit of* 1.4×10^{-4}.

In passing through condensed matter, a heavily ionizing particle can produce unique effects by creating a narrow, roughly cylindrical region that is crowded with ionization and excitation, with atomic displacements and broken bonds (*1*). Unfortunately our minute knowledge of the biological effects of heavy ion irradiations stands in striking contrast to the extensive documentation that exists of the effects produced by more usual, randomly dispersed defects such as are caused by β or γ radiation. Although it has been demonstrated that heavy ions can have lethal effects upon colony-forming cells (*2*, *3*), for nonregenerative human cells such information is totally lacking, difficult to obtain, and hence unlikely to be available soon. Because of this gap in our knowledge and because of the probable loss by space travelers of irreplaceable cells by heavy ions in the cosmic rays, such particles have been monitored on the Apollo 8 and 12 missions by using Apollo helmets as heavy particle dosimeters.

The Apollo helmets consist of Lexan polycarbonate, a material that records the tracks of particles whose ionization level lies above that produced by neon at ~ 7 Mev per nucleon (*4*), essentially the same level as corresponds to inactivation with unit efficiency of human kidney T1 cells by a particle that traverses the cell nucleus (*2*). Because of the approximate identity of these two thresholds—for track formation and for cell destruction (at least for T1 cells)—Lexan is an appropriate detector material for assessing the dose of biologically destructive particles to which the astronauts were exposed.

The helmets used for this experiment include the one worn by astronaut Lovell on Apollo 8, all three of the Apollo 12 helmets, and a control helmet that was exposed to cosmic rays at a balloon altitude of 138,000 feet (41,000 m) at Fort Churchill. The helmets were stored in the dark to avoid ultraviolet enhancement effects (*5*). Later they were chemically etched to reveal tracks (*6*) with a stirred 1 : 1 solution (by volume) of ethanol and 6.25*N* NaOH at 23°C for 196 hours, a period sufficient to remove a veneer of ~ 80 μm of plastic (~ 4 percent of the initial thickness) and to develop tracks similar to those shown in Fig. 1. The helmets themselves were used as etching tanks; only ~ 600 cm^2 of the forward-facing portions of the helmets were etched to avoid damage to the headrest and other fittings. The etched portions were replicated with silicone rubber (*7*), and the gold-coated replicas [see figures 1 and 3 in (*7*)] were scanned in a stereomicroscope at a magnification of × 14. Features with vertical relief down to ~ 15 μm were readily visible under our viewing conditions, a height that for a track at 45° corresponds to a total etched length of ~ 135 μm. Although our cutoff for track observation is not perfectly sharp, we can use our estimate for effective value of 135 μm of minimum total etched length to show (*8*) that the tracks we observe correspond entirely to nuclei of atomic number (Z) ≥ 10 and that from the known abundances in this charge region (*9*) most of these will be iron or iron group ($24 \leq Z \leq 28$) nuclei.

The results summarized in Table 1 show that the Apollo 12 helmets experienced nearly three times the integrated flux of heavily ionizing cosmic rays as did the Apollo 8 helmet. Even when allowance is made for the different lengths of the two missions, the track formation rate on Apollo 12 is still significantly higher (by a factor of 2.0). The track formation rate in the balloon-flown helmet was 3.1 times higher than that on the Apollo 8 mission. On the Apollo 12 mission, helmets 504 and 506 were in the lunar module for 25 hours and were outside the lunar module, during the astronauts' extravehicular activity, for nearly 8 hours. During this 33-hour period the average thickness of shielding material surrounding the helmet is considerably reduced relative to the environment in the command module, where the remainder of the trip was spent. This different environment produced no statistically meaningful difference between the track densities in Gordon's helmet [1.41 (± 0.15)] and those in Bean's and Conrad's [1.51 (± 0.11)], although the possibility of as much as a 30 percent increase is allowed by standard deviation limits. Two relevant considerations are that the lunar module and extravehicular activity portions of the trip make up only 14 percent of the total time of 244.5 hours in space and that during the stay on or near the moon half of the incident cosmic rays