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Global Increase in Quaternary Explosive Volcanism

Volcanic ash in deep-sea sections indicates very high rates of explosive volcanism during the last 2 million years.

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The history of volcanism is usually regarded as being related to local or regional tectonism rather than demonstrating any global trends (*1, 2*). This is not surprising since the theory of plate tectonics relates orogeny to plate collisions involving one or two continents, and such collisions depend in part on the nature of plate geometry. As a result, two continents may collide during a time of subdued plate tectonic motions (*3*). The distribution of volcanic ash in deep-sea sequences can tell us much about the volcanic history in the source areas. Volcanic ash discharge from volcanic sources throughout the world could conceivably fall into any of several patterns. Ash distribution could be essentially random from region to region (*4*), it could reflect globally synchronous changes in explosive volcanism, or it could be similar in more restricted regions such as the circum-Pacific if it were related to motions of one or two plates. Cycles of volcanism on a short-term historical basis have been linked with paleoclimatic change (*5*), while in the longer term, episodes of diastrophism and related increased volcanism occur throughout the circum-Pacific orogenic belt (*6–8*). A chorus of intensified Quaternary volcanism has been noted by McBirney (*8*), who also pointed out that various episodes of igneous activity, reflected in styles of composition, intensity, and mode of expression, occurred in unison over large areas throughout the Mesozoic and Cenozoic with "utter disregard for the local setting" (*9*). During the Quaternary, similarities in mode of volcanism exist throughout the circum-Pacific region and the Antilles, especially with respect to current activity and present stage of evolution. Synchronism has also been suggested in lava extrusion from widespread hot spots, resulting perhaps from variable mantle plume activity (*3*).

Much difficulty is encountered in using the record on land to establish patterns of volcanism over large areas, especially because younger volcanics often obscure underlying sequences, making volume comparisons difficult, and because of the effort required in obtaining large enough numbers of radiometric dates to establish a detailed chronology. This is the first attempt to evaluate global volcanic history by using the distribution of volcanic ash in deep-drilled sequences obtained by the Deep Sea Drilling Project (DSDP). We have used a relatively simple semiquantitative approach to evaluate the tempo of explosive global volcanism by determining the distribution of coarse volcanic ash in all documented deep-sea drilled sequences (water depth > 1000 meters) now available in all oceans except the Arctic. Volcanic ash in deep-sea sediments is mostly produced by andesitic and acidic volcanism of island arc regions and continental margins, rather than by the more typically basaltic volcanism of oceanic islands. We have restricted analysis to sediments from the Neogene and Quaternary ($\simeq$ 20 million years ago to Recent) because, compared with older sediments, core recovery is good and little apparent silica mobilization has occurred to form cherts.

The sedimentary record of the cores drilled at the first 320 deep-sea sites has been examined to determine the distribution of volcanic ash layers and disseminated coarse ash from the earliest Miocene to the Recent (Fig. 1). Volcanic ash has been reported in 84 of these sites, most of which are close to island arcs and continental margins with present-day volcanic activity (Fig. 1). The data used in this article were derived from detailed sedimentary descriptions, based on visual examination and smear-slide analysis, published in the initial reports of 27 legs of the DSDP (*10*) and from much less detailed descriptions in the preliminary reports of six other DSDP legs (*11*). The descriptive procedures, including sampling intervals, used by the different teams of sedimentologists on the DSDP legs are highly standardized, so that the variability introduced in the data from this source is minimized and is not considered important in this investigation.

Ash layers occur in all sediment types from clays to biogenic oozes. Several problems exist in the recovery and observation of ash layers in deep-sea sediments; drilling may destroy fine ash laminations if sediment is relatively soft or soupy, as is often the case with the latest Tertiary and Quaternary, and burrowing by organisms may disturb ash layers, dispersing the ash to surrounding sediments. Because ash layers are often small-scale sedimentary features, few are reported in general lithological columns presented in the briefer preliminary reports (*11*). The DSDP sites represent a random data set as far as volcanic ash distribution is concerned; no sites were selected for the primary purpose of examining volcanic ash distribution. The following

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data were recorded for all ash occurrences: age; fossil zone; depth in core; approximate thickness of ash layer; whether the ash constituted a discrete bed, was in the form of multiple ash layers, or was dispersed through the sediments (*12*); and core recovery (length of cores recovered for a particular thickness drilled). It has been necessary to employ the most precise dating method possible, the standard Neogene planktonic foraminiferal zones (N Zones) (*13*), which have been chronologically calibrated by Berggren (*14*) (Fig. 2). Little difficulty was encountered in relating to the N zonation where only other biostratigraphic zonations were available.

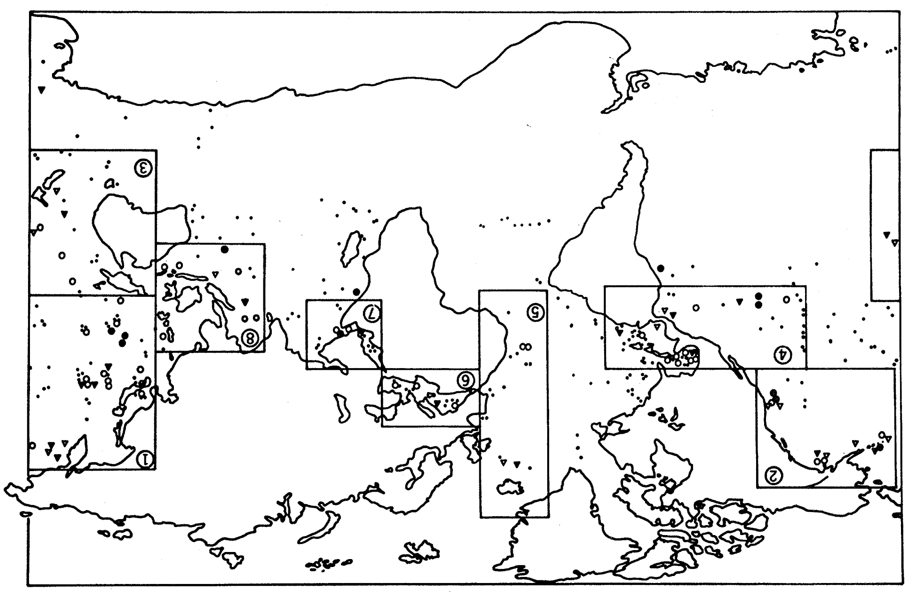

Fig. 1. Location of sites drilled during the first 34 legs of the Deep Sea Drilling Project. Small dots identify sites with no volcanic ash recorded in sediments of Miocene to Recent age (20 million years ago to Recent) or lack of sediment of this age. Other symbols identify the sites with Late Cenozoic volcanic ash reported: (○) single ash layers only; (△) multiple ash layers for Miocene to Quaternary intervals; (●) multiple ash layers confined to Miocene or Pliocene; (▲) multiple ash layers in Quaternary only. The numbered sectors are the areas whose volcanic history is shown in Fig. 4.

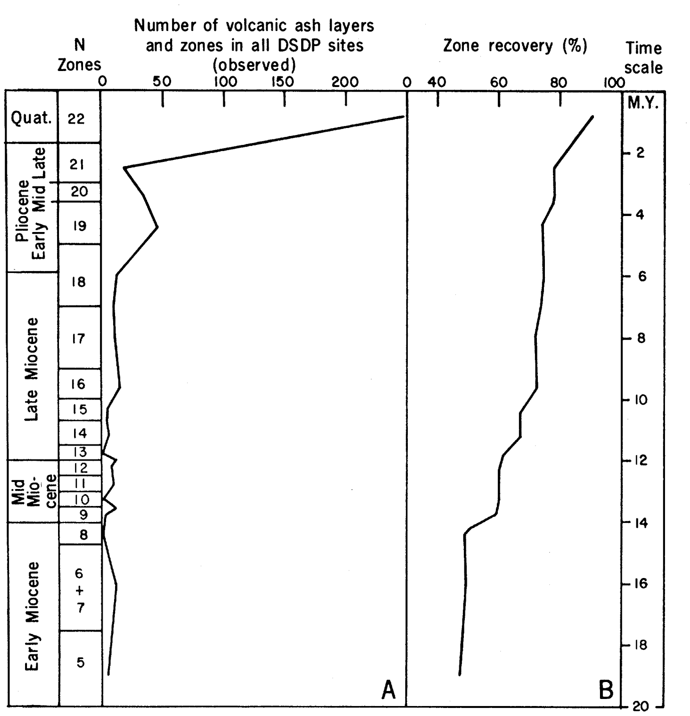

Fig. 2. (A) Total number of recorded ash layers and disseminated ash horizons within each N Zone [Neogene planktonic foraminiferal zone (*13*)] for the Early Miocene to the Recent (20 million years ago to Recent) in all DSDP sites from leg 1 to leg 34 (a total of 320 sites). The curve is based on a total of 84 sites that contain any ash of this age. (B) Percentage recovery of each N Zone for the total set of DSDP sites containing volcanic ash of the above age. For each site, a zone was recorded as present even if it was only partially represented because of incomplete core recovery or hiatus. The chronology at the right was calibrated to N zonation by Berggren (*14*).

Evidence for Exceptional Quaternary Volcanism and for Global Synchronism

Tabulation of observed volcanic ash horizons in each N Zone (Fig. 2) indicates that the number of horizons during the Quaternary was 5 times the number during any previous Neogene interval and in general 13 times the average number for all Neogene intervals. It is necessary to normalize these data, however, since the rate of core recovery is not equal between zones because of drilling differences and sedimentary hiatuses. Furthermore the zones are of different duration, which creates unequal opportunities for ash deposition if constant explosive volcanism is assumed. The percentage recovery (Fig. 2) for N Zones in the total set of cores containing ash decreases with increasing age as follows: Quaternary, 92 percent; Late Pliocene to Late Miocene, 70 to 75 percent; Middle Miocene, 60 percent; and Early Miocene, 50 percent. Recovery percentages and zone durations were normalized relative to the Quaternary, which is 1.8 million years in duration (*14*). The average length of cores recovered within specific Late Cenozoic intervals (Table 1) shows no significant difference throughout and therefore would not influence our results. After normalization of the raw data, the global curve (Fig. 3) clearly shows a peak in Quaternary volcanism (at a total of about 270 ash levels), approximately four times higher than the Neogene average and about twice as high as a Middle Pliocene peak. A peak also coincides with the Middle Miocene. The curve indicates that the Early Miocene, Late Miocene, and Late Pliocene were relatively quiescent volcanically.

To determine the extent of these trends, the sites have been grouped into eight areas (Fig. 1). The resulting curves (Fig. 4) show that a conspicuous Quaternary peak occurs in all regions, while other peaks occur during the Middle Pliocene in the Northwest Pacific, Southwest Pacific, and Central America and in the Early Pliocene in the Southwest Pacific, the Middle East, and Central America. Our curves also show that during the Middle Miocene volcanicity was widespread around the Pacific Ocean, supporting previous interpretations based on the land-based record (*9, 15*). The regions with greatest distribution of ash horizons are the Western Pacific, Central America, and the Northeast Pacific–Aleutians areas. Sampling is inadequate off western South America, an active volcanic region.

The percentage of distinct ash layers and disseminated ash horizons through time (Table 2) shows that, except for the Early Miocene, the frequency of distinct ash layers decreases with increasing age. This could reflect either a general decrease of highly explosive volcanism with age or postdepositional change within the sediment. In total, 74 percent of the ash occurrences are represented by distinct layers, while 26 percent are disseminated ash horizons.

Other Possible Effects on Ash Distributions

Island arc volcanoes, such as those found along the perimeter of the Western Pacific, are major contributors of volcanogenic material to deep-sea sediments. The volcanogenic material preserved in the sediments from the DSDP sites is virtually all airborne volcanic ash (glass shards). The range over which such material is distributed is a function of several factors, including the volume of material ejected, height of ejection (explosivity), particle size, and wind speed (*16*). Since this study is based mostly on the identification of coarse, sand-sized ash (*12*) the material mostly reflects deposition close to the volcanic source (within 1000 to 2000 km). Elimination of the fine fraction, which is very difficult to work with quantitatively (*17*), simplified the identification of periods of inceased volcanism. As a consequence, the global picture can only be determined by examination of ash data from worldwide marine stratigraphic sections.

No studies have yet considered the possibility of extensive diagenetic change of volcanic ash in deep-sea sediments of Middle to Late Cenozoic age. On the other hand, no evidence suggests that postdepositional chemical alteration to either cherts, zeolites, or montmorillonite clays has reduced the number of ash layers in the older levels. Alteration of siliceous sediments (possibly including volcanic ash) to form cherts largely occurs in sediments older than Miocene. Of the 84 sites containing ash, only 5 are reported to have cherty layers within the Late Cenozoic: one (site 84) in the Late Miocene, three (sites 33, 34, and 158) in the Middle Miocene, and one (site 288) in the Early Miocene. The zeolite phillipsite is often found in Miocene and younger sediments, but is confined to central areas of the sea floor and forms from the alteration of basic volcanic material (*18*). This alteration would have little effect on our data, as volcanic ash is largely of silicic origin. Another zeolite, clinoptilolite, does form from such material, but the considerable time for formation generally restricts it to Cretaceous to Eocene sediments (*18*). Substantial alteration of pyroclastic material has been reported in the Late Miocene to Recent San Blas Formation in the equatorial Eastern Pacific (*19*), forming large amounts of green montmorillonite (*20*). Thus alteration products of volcanogenic materials are generally conspicuous when they occur. Large decreases in volcanic ash in young intervals such as the Late Pliocene and the Late Miocene are further evidence that our curves are not caused by diagenetic alteration.

In some areas, the general increase of volcanic ash with decreasing Late Cenozoic age (Fig. 3) may have been partly affected by plate motions transporting ocean floor closer to the sources of volcanic ash in the areas of subduction. The sites most likely affected by such motions occur in the Northwest Pacific (Fig. 1) where a wedge of volcanogenic sediments within 1000 km of the Asiatic area reflect diachronous deposition as the plate moved closer to the source (*21*). About seven of the DSDP sites in this region (sites 47, 49 to 51, and 194 to 196) lack ash older than Late Pliocene, and thus the curve from this area (Fig. 4) may be biased. Nevertheless, this curve is very

Table 1. Average length of cores containing ash recovered at DSDP sites in seven regions for specific intervals of time, and overall averages for the same intervals of time. The maximum length of any drilled core is 9 m. Indonesian sites containing ash do not contain sediments older than Pliocene. Sediments of Middle Miocene age were not recovered in the Mediterranean because of insufficient penetration and disconformities.

Region	Average length of cores recovered (m)				
	Pleistocene	Pliocene	Late Miocene	Middle Miocene	Early Miocene
Eastern Atlantic	2.7	8.5	8.25	6	7.2
Central America	6.7	7.1	7.2	6.3	7.2
Eastern Pacific	5.0	4.3	4.95	6.25	3.2
Western Pacific	5.9	4.5	5.1	4.5	5.4
South Pacific	6.4	7.3	6.1	4.0	3.2
Mediterranean	5.1	5.5	2		1.5
Indonesia	7.4	6.3			
Average	5.6	6.2	4.8	5.4	4.6

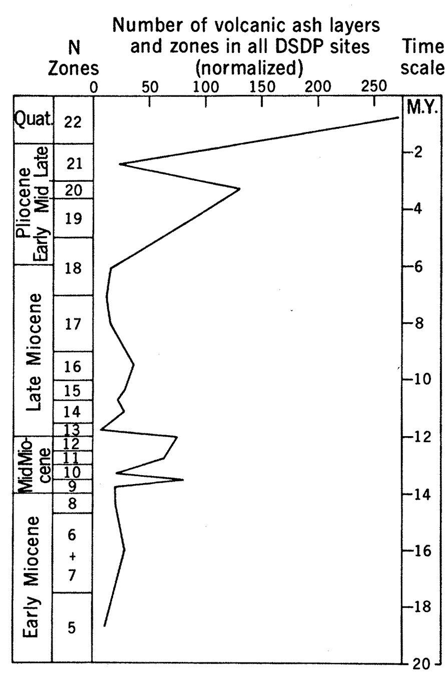

Fig. 3. Total number of volcanic ash layers and disseminated ash horizons within each planktonic foraminiferal N Zone (*13*) for the Early Miocene to the Recent in all DSDP sites from leg 1 to leg 34 (total, 320 sites). The curve is based on normalization for differences in sediment recovery during drilling of the 84 sites containing ash of this age and for age differences between N Zones. Volcanic ash distribution for each zone has been normalized to the percentage recovery of N 22 (Quaternary), as shown in Fig. 2. Ash distribution in each N Zone has also been normalized to the duration of N 22 (Quaternary), which is 1.8 million years. The chronology at the right was calibrated to N zonation by Berggren (*14*). The effect of normalization on the curve varies according to the duration of the individual N Zones. Although we have used the biochronological scheme of Berggren (*14*), optional schemes exist (*69*) which show similar duration of the N Zones in the Pliocene and Quaternary but may differ somewhat for the Miocene. However, whichever scheme is adopted, the resulting normalized curves for the Miocene vary only in detail, and all show the same basic pattern: a peak of volcanic activity in the Middle Miocene and low activity during the Middle and Late Miocene.

similar to those representing the volcanic history in the Northeast and Southwest Pacific (Fig. 4), which apparently were not affected by such plate motions. Sites north of the Aleutian Trench were unaffected by plate motions (*22*); only two sites (83 and 192) occur seaward of the subduction zone and their sedimentation record suggests a history of limited or no motion (*22, 23*). If the distribution of volcanic ash was controlled by plate motions, a decrease in thickness of ash layers would be expected with increasing age. No such trend exists (Table 2) and distinct ash layers have remained approximately constant in average thickness throughout the Late Cenozoic.

Supporting Evidence from the Land Record

Much evidence from land shows that the Late Cenozoic has been a period of structural rejuvenation, certainly throughout the Pacific (*6*) and probably globally (*3*). Our deep-sea record in general indicates that volcanicity has increased within the Late Cenozoic, reaching a peak of activity in the Quaternary. The land-based record also provides substantial evidence for increased and widespread Quaternary volcanism. In certain areas it is clear that Quaternary volcanism represents a maximum for the Neogene, while in others, although a distinct episode of Quaternary volcanism exists, earlier Neogene volcanic episodes are of greater intensity. For most regions knowledge of volcanic history is inadequate to determine relative intensities of Neogene volcanic episodes in any detail.

Hays *et al.* (*19*) have observed that the circum-Pacific area is marked by a "recent" phase of volcanism. In New Zealand, the recent phase of volcanism began with the eruption of the Taupo-Rotorua ignimbrites, which range in age from 0.8 to 0.2 million years (*24*). Ash layers of similar age range occur in marine sediments to the east of New Zealand (*19, 25*). In eastern Australia, volcanic activity seems to have occurred at nearly a constant rate over the last 20 million years, except for a distinct increase in lava field provinces during the last 3 million years (*26*). The intensification of explosive volcanic history in the North Pacific during the Quaternary recorded in the deep-sea drilled sites (*22*) and piston cores (*27*) is paralleled by the record on land, where some of the oldest rocks associated with the volcanoes of the Aleutian Arc are dated at 2 million years ago (*22, 28*). Increased volcanism in this region has been considered to be due to a quickened pace of relative motion between the North American and Pacific plates (*22*). Increased Quaternary volcanism is also reported for the Kurile-Kamchatka region (*29*). In Japan, increased volcanism is reported from the latest Miocene (Kitaura Stage) with andesitic and dacitic activity as well as large ignimbrite flows (*19, 30*). Andesitic and rhyolitic explosive volcanism in Japan has continued through the Late Pliocene and Quaternary with some indication of increased intensity in the Quaternary (*19, 30*). In Hawaii, where volcanism has occurred in episodes throughout the Cenozoic, the latest phase, which began 5 to 6 million years ago, is marked by increasing volumes toward the present (*3, 31*). In the Indonesian region, a major orogenic phase commenced in the Late Pliocene and has been continuing to the present day (*32*). Intense Early Miocene andesitic volcanism in the New Hebrides apparently was followed by a relatively quiescent cycle and then by basaltic volcanism during the Late Pliocene and Quaternary (*33*). The South American Andes developed during the Late Cenozoic with increased tectonism and volcanism during the Late Pliocene and Quaternary (*34*). Most of the volcanic rocks of

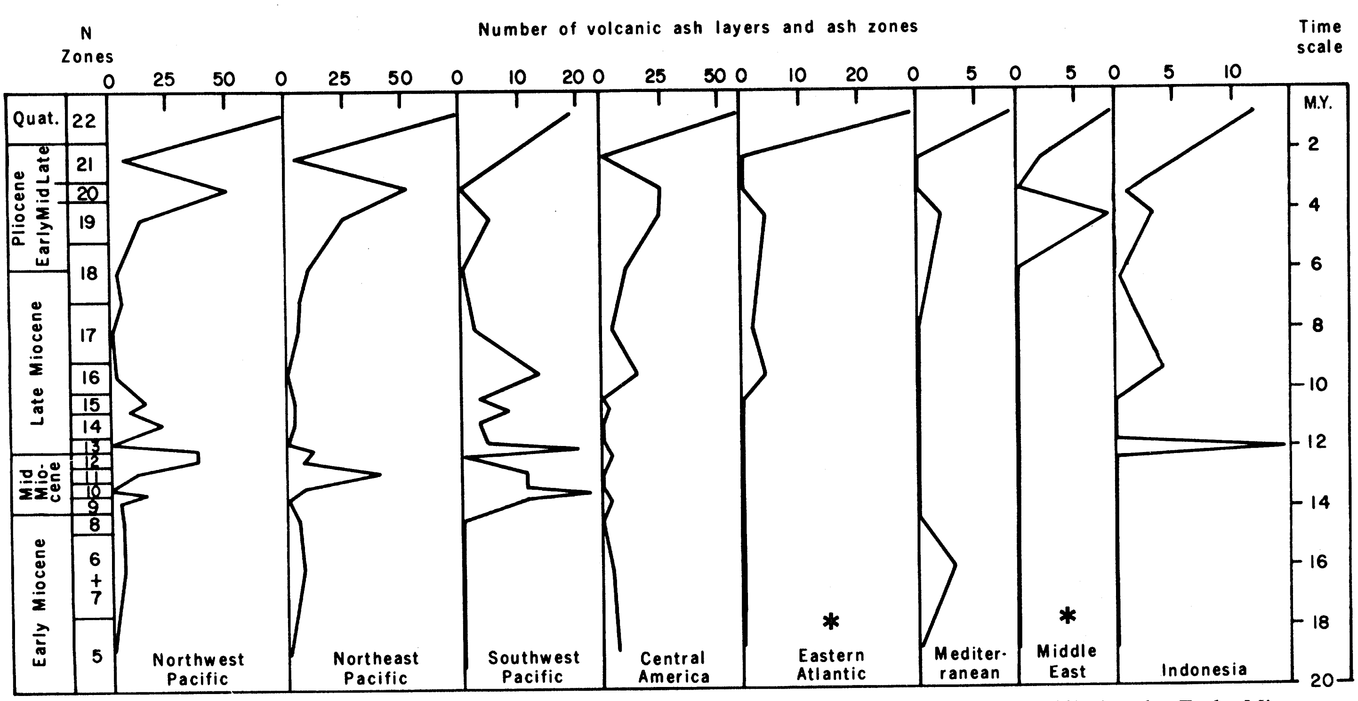

Fig. 4. Total number of volcanic ash layers and disseminated ash horizons within each N Zone (*13*) for the Early Miocene to Recent in DSDP sites located within the eight oceanic sectors shown in Fig. 1. Each curve is subject to the same normalization process described in the legend to Fig. 3. The chronology at the right was calibrated to N zonation by Berggren (*14*). Note that the Middle East and Eastern Atlantic curves (*) probably represent mainly volcanism associated with accreting plate margins and hot spots such as Afar, Iceland, and the Azores, Cape Verde, and Canary Islands. Other curves primarily represent island arc volcanism.

El Salvador are of Late Pliocene age and younger, representing similar increased volcanism (*35*). However, volcanism has been reported as being continuous in Nicaragua and Guatemala during the Late Cenozoic (*36, 37*), although large andesitic cones did not develop in Guatemala until the Quaternary (*37*). In central Mexico, major volcanic episodes have been recorded during the middle Late Miocene and latest Miocene to earliest Pliocene (*38*), although explosive volcanicity has been important during the Quaternary (*39*). Significant volcanism in the Canadian Cordillera suggests recent subduction of the Juan de Fuca Plate below the American Plate, as does the presence further inland of the Late Cenozoic volcanic belt of the Cascades (*40*).

Although the Atlantic region and Iceland have long volcanic histories (*41*), there is some evidence for increasing Late Cenozoic volcanism. Vogt (*3*) has presented evidence from topography and crustal thickness for accelerating Late Cenozoic discharge of basaltic lava toward the present in marine areas around Iceland. Furthermore, volcanicity in Iceland during the last 10,000 years has been three to five times more vigorous than during previous Late Cenozoic intervals (*42*). Volcanism in the Lesser Antilles has accelerated from 5 million years ago toward the Quaternary (*43*). In the Antarctic Peninsula major episodes of volcanism occurred during the Middle to Late Miocene and during the Quaternary (*44*), while elsewhere in Antarctica there have been considerable accumulations of volcanic material during the Quaternary (*45*). A major increase in tectonically related Late Pliocene to Quaternary volcanism has occurred in the East African rift valley, for more than 1500 km from Ethiopia to Tanzania (*46, 47*). At least in southern Ethiopia, this was preceded by a relatively quiescent period dating from about the Early Miocene (*47*). At Maruj Assuad, Libya, which is the largest volcanic province in North Africa, most flows presently exposed are younger than about 2 million years (*48*). In northern Israel, widespread plateau basalts were formed during the Quaternary; these followed phases of less frequent activity during the Early Pliocene and Middle Miocene (*49*). These references are by no means exhaustive, but represent supporting evidence from both "subduction" and "hot spot" areas for widespread Quaternary volcanism which in many cases was more intense than that during earlier Neogene intervals. This constitutes further evidence that the ash distribution in the deep-sea sequences is not generally controlled by diagenesis.

Table 2. Percentages of distinct ash layers and zones of disseminated ash and average thickness of distinct ash layers. Values are given for specific intervals of time for the total set of Late Cenozoic cores containing volcanic ash. Distinct ash layers represent 74 percent of ash occurrences; disseminated ash horizons are not included in thicknesses.

Time	Percentage of: Ash layers	Percentage of: Disseminated ash	Av. thickness of ash layers (cm)
Quaternary	81.6	18.4	13.6
Pliocene	72.8	27.2	10.7
Late Miocene	57	43	12.6
Middle Miocene	41.5	58.5	13.5
Early Miocene	70.5	29.5	14.9

Climatic Implications

Holmes (*2*, p. 726) stated in 1965 that geologically no correlation had yet been demonstrated between times of glaciation and periods of prolonged or exceptionally intense volcanism. Our work now shows that greatly increased volcanism during the last 2 million years closely coincides with that interval of earth history marked by major and rapidly oscillating climatic conditions related to glacial-interglacial cycles in the Northern Hemisphere (*50*). Shackleton and Kennett (*51*) dated the initiation of continental glaciations in the Northern Hemisphere at 2.4 million years ago. Since then there have been major and rapid fluctuations in Northern Hemisphere ice sheets, global climates, and sea level (*51, 52*). This general synchronism is almost certainly not coincidental, although our data are too general to discriminate among any of the cause and effect relations that have often been suggested (*46, 53*). Chappell (*54*) calculated that explosive volcanicity may need to be increased at least 40 percent within a period of a few thousand years for glaciation to be initiated by dust veil effects. Although our data show a general increase in volcanism much greater than this, the present resolution of the data does not permit us to deal with finer variations in volcanic output during the Quaternary.

Causes of Increased Volcanism

It is logical to relate increased Quaternary volcanism to increased orogenesis. Stille (*55*) considered that all orogenic phases are synchronous and worldwide in nature, although others, for example Gilluly (*56*), maintain that there is no periodicity to orogenesis and that orogenies have been continuous through geological time with no peaks in activity. Brookfield (*57*) considered that orogenies involving an oceanic plate underthrusting a continental margin or other oceanic plate (continental edge orogenies) should be synchronous. On the other hand, orogeny resulting from plate collisions involving one or two continents is in part dependent on the accidents of plate geometry and may occur during periods of subdued plate tectonic motions (*3*). Much of the volcanic ash in the DSDP sites is derived from areas of subduction and andesitic volcanism. In such areas the source of magma is commonly thought to be near the Benioff Zone, between 100 and 200 km in depth, with frictional heat playing an important role in magma genesis (*54, 58*). If this is the case, magma genesis and volcanism may vary with rate of plate convergence (*54*), and since one midocean ridge is related to several continental margins, synchroneity in increased subduction and hence increased volcanism might be expected (*57*).

Hays *et al.* (*19*) suggested that increased volcanism would occur when the underthrust oceanic plate reached a depth of between 100 and 200 km, where conditions are suitable for the production of andesitic magma. They suggested that a major change in the pattern of sea-floor spreading in the North Atlantic, the Pacific, and the Indian Ocean about 10 million years ago initiated a new regime of underthrusting, resulting in a Late Cenozoic phase in volcanism. Vogt (*3*), however, has suggested that for the last 60 million years the history of "plume" discharge has varied on a global basis (Fig. 5) and is related to the kinetics of crustal plates. His curve for Hawaiian discharge (Fig. 5) is in general similar to the curve for global volcanic ash output (Fig. 3), having a peak in activity at about 15 million years not significantly different from the peak in the ash curve at 13 million years, and showing increasing discharge commencing about 5 million years ago. The general similarity of the curves suggests that global synchronism in mantle plume activity, as determined by episodes of oceanic island (hot spot) volcanism (for example, Hawaii, the Middle East, the Eastern Atlantic), is in harmony with cycles of explosive andesitic volcanism centered mainly in the zones of subduction. Thus explosive volcanism in subduction areas might be expected to be synchronous with changes in sea-floor spreading and plate motions, rather than to show any delay as previously suggested (*19*).

If plate coupling is assumed, synchronous global volcanism should occur at least at certain times. Recent seismic evidence (*59*) suggests that the world's seismic belts (plate boundaries) are strongly coupled globally. Thus episodes of volcanism imply episodes in sea-floor spreading. Such episodic sea-floor spreading has been recognized by some (*60, 61*), while others (*62*) prefer to assume constant spreading. It would probably be difficult to detect any increase in sea-floor spreading rates during the last 2 million years from the study of oceanic magnetic anomalies. However, an acceleration proportional to the order of magnitude increases in volcanic discharge indicated by our results can be ruled out. An earlier discontinuity in sea-floor spreading rates has been reported at about 5 million years ago (between anomalies 3 and 4) in the Atlantic (*61, 63*) and the North Pacific (*64*) with some evidence of subsequent increased spreading rates (*61*). Future work may show that volcanic discharge rates represent sensitive "amplifiers" to small spreading rate variations. On the other hand, if rates of sea-floor spreading are subsequently shown to have been constant over the last 3 to 4 million years, it will be necessary to look for further possible fundamental mechanisms that control episodic volcanism over wide regions and yet have essentially no effect on rates of sea-floor spreading (*9*).

Another mechanism that may have played some role in increased Quaternary explosive volcanism is climatically related. Glacio-eustatic changes in sea level were large and relatively rapid throughout the last 2 million years because of fluctuations in the Northern Hemisphere ice sheets. Related changes in ocean volume caused the ocean basins to respond isostatically to eustatic loading and unloading (*65, 66*). This might have created periodic large-scale movements within the mantle, causing stress release and enhanced

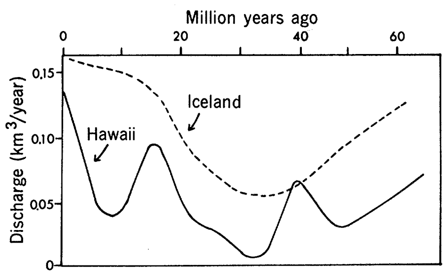

Fig. 5. History of volume discharge rate of basic volcanic rocks over last 60 million years for the Hawaiian-Emperor and Greenland-Iceland-Faeroe aseismic ridges; after Vogt (*3*). Vogt argued that the discharge curve for the Hawaiian hot spot mirrors the total world hot spot curve. In both curves note the increase in lava discharge toward the Pleistocene and Recent.

volcanism (*65*). Under this hypothesis a significant correlation was predicted between periods of unusual volcanic activity and Quaternary glaciation (*65*). This hypothesis is apparently not unreasonable when it is considered that arguments exist to show that volcanic eruptions may be triggered by lunar and solar tidal stress (*67, 68*). Indeed some workers have considered possible relations between volcanism and astronomical motion, with related feedback to paleoclimatic change (*54, 68*).

Summary

The worldwide distribution of volcanic ash has been determined from 320 deep-sea sections drilled during the Deep Sea Drilling Project. The ash distribution in the deep-sea sections, which span the last 20 million years, indicates that there has been a much higher rate of explosive volcanism from both island arc and hot spot volcanoes during the last 2 million years. This episode, and perhaps another in the Middle Miocene, correlates with previously reported widespread synchronism in increased volcanicity in certain oceanic islands. Increased Quaternary volcanism coincides approximately with that episode of the Cenozoic marked by major and rapidly fluctuating climatic change.

References and Notes

1. N. D. Newell, *Sci. Am.* **226**, 54 (1972).
2. A. Holmes, *Principles of Physical Geology* (Nelson, London, 1965).
3. P. R. Vogt, *Nature (Lond.)* **240**, 338 (1972).
4. Prevailing wind directions prevent volcanic ash deposits from being completely random in distribution.
5. H. H. Lamb, *Phil. Trans. R. Soc. Lond. Ser. A Math. Phys. Sci.* **266**, 425 (1970); J. R. Bray, *Nature (Lond.)* **248**, 42 (1974).
6. R. H. Dott, *Science* **166**, 874 (1969).
7. G. R. Heath, *Geol. Soc. Am. Bull.* **80**, 1997 (1969).
8. A. R. McBirney, *Comments Earth Sci.* **2**, 69 (1971).
9. ———, J. F. Sutter, H. R. Naslund, K. G. Sutton, C. M. White, *Geology* **2**, 585 (1974).
10. *Initial Reports of the Deep Sea Drilling Project* (Government Printing Office, Washington, D.C., 1969–1974 and in press), vols. 1–23, 25, 27–29.
11. Scientific Staff, *Geotimes* **17** (No. 9), 17 (1972); *ibid.* **18** (No. 3), 16 (1973); *ibid.* **18** (No. 9), 18 (1973); *ibid.* **18** (No. 10), 22 (1973), *ibid.* **18** (No. 12), 14 (1973); *ibid.* **19** (No. 3), 16 (1974); *ibid.* **19** (No. 4), 20 (1974).
12. In this analysis we do not consider volcanic ash reported in trace amounts. We have utilized only that ash occurring in large enough quantities to be recorded in the lithologic columns given in the DSDP initial reports. Furthermore, because very fine grained ash, smaller than several micrometers, is globally distributed by winds, it is probable that such ash is dispersed throughout all deep-sea sections. This ash is too small to be readily differentiated by sedimentologists and is not considered in this analysis.
13. F. T. Banner and W. H. Blow, *Nature (Lond.)* **208**, 1164 (1965).
14. W. A. Berggren, in *The Micropaleontology of Oceans*, B. F. Funnell and W. R. Riedel, Eds. (Cambridge Univ. Press, London, 1971), pp. 693–809.
15. A. K. Baksi and N. D. Watkins, *Science* **180**, 493 (1973).
16. T. C. Huang, N. D. Watkins, D. M. Shaw, J. P. Kennett, *Earth Planet. Sci. Lett.* **20**, 119 (1973); D. Ninkovich, B. C. Heezen, J. R. Connolly, L. Burckle, *Deep-Sea Res.* **11**, 605 (1964).
17. T. C. Huang, N. D. Watkins, D. M. Shaw, *ibid.*, in press.
18. K. Venkatarathnam and P. E. Biscaye, *Mar. Geol.* **15**, 11 (1973).
19. J. D. Hays, H. Cook, G. Jenkins, W. Orr, R. Goll, F. Cook, D. Milow, J. Fuller, in *Initial Reports of the Deep Sea Drilling Project* (Government Printing Office, Washington, D.C., 1972), vol. 9, pp. 909–931.
20. H. E. Cook and I. Zemmels, in *ibid.*, pp. 707–789.
21. B. C. Heezen *et al.*, in *ibid.* (1973), vol. 20, pp. 725–738.
22. D. W. Scholl and J. S. Creager, in *ibid.* (1973), vol. 19, pp. 897–913.
23. T. J. Fullam, P. R. Supko, R. E. Boyce, R. W. Stewart, in *ibid.*, pp. 887–896.
24. J. J. Stipp, J. A. Chappel, I. MacDougall, *Am. J. Sci.* **265**, 462 (1967); A. Cox, *Earth Planet. Sci. Lett.* **6**, 257 (1969); G. W. Grindley, H. J. Harrington, B. L. Wood, *N.Z. Geol. Surv. Bull. 66* (1959).
25. D. Ninkovich, *Earth Planet. Sci. Lett.* **4**, 89 (1968).
26. P. Wellman and I. McDougall, *Tectonophysics* **23**, 49 (1974).
27. J. D. Hays and D. Ninkovich, *Geol. Soc. Am. Mem. 126* (1970), pp. 263–290.
28. D. K. Bingham and D. B. Stone, unpublished report, Geophysical Institute, University of Alaska, Fairbanks, 1971.
29. I. V. Melekestsev, in *The Newest Volcanism and Glaciations* (Abstracts, Royal Society of New Zealand, Wellington, 1973), p. 234.
30. M. Minato, M. Gorai, M. Hunahashi, Eds., *The Geologic Development of the Japanese Islands* (Shokam, Tokyo, 1965).
31. H. R. Shaw, *Geol. Soc. Am. Bull.* **84**, 1505 (1973).
32. R. W. van Bemmelen, *The Geology of Indonesia* (Netherlands Government Printing Office, The Hague, 1949); M G. Audley-Charles, D. J. Carter, J. S. Milsom, *Nature (Lond.)* **239**, 35 (1972).
33. A. H. G. Mitchell and A. J. Warden, *J. Geol. Soc. (Lond.)* **127**, 501 (1971); B. P. Luyendyk, W. B. Bryan, P. A. Jezek, *Geol. Soc. Am. Bull.* **85**, 1287 (1974)
34. V. Auer, *Ann. Acad. Sci. Fenn. Ser. A3 No. 100* (1970); P. Gröber, *Rev. Cent. Estud. Doct. Cienc. Nat. Buenos Aires No. 1* (1936).
35. H. Williams and H. Meyer-Abich, *Univ. Calif. Publ. Geol. Sci. No. 32* (1955), pp. 1–54.
36. A. R. McBirney and H. Williams, *ibid., No. 55* (1965), pp. 1–65.
37. H. Williams, *ibid., No. 38* (1960), pp. 1–72.
38. N. D. Watkins, B. M. Gunn, A. K. Baksi, D. York, J. Ade-Hall, *Geol. Soc. Am. Bull.* **82**, 1955 (1971).

→ J. P. Kennett and P. Huddlestun, *Quat. Res.* **2**, 38 (1972).
40. R. A. Stacey, *Nature (Lond.)* **250**, 133 (1974).
41. S. Moorbath, H. Sigurdsson, R. Goodwin, *Earth Planet. Sci. Lett.* **4**, 197 (1968).
42. H. Sigurdsson, *Iceland and Mid-Ocean Ridges*, S. Bjornsson, Ed. (Societas Scientarium Islandica, Reykjavik, 1967), pp. 32–49.
43. ———, personal communication.
44. R. J. Adie, *Antarct. Geol. Geophys. Int. Union Geol. Sci. Ser. B No. 1* (1972), pp. 137–1141.
45. W. Dort, *ibid.*, pp. 645–652.
46. V. E. Fuchs and T. T. Peterson, *Geol. Mag.* **84**, 322 (1947).
47. P. A. Mohr, *Bull. Volcanol.* **32**, 5 (1968).
48. J. M. Ade-Hall, P. H. Reynolds, P. Dagley, A. E. Mussett, T. P. Hubbard, E. Klitzsch, *Can. J. Earth Sci.* **11**, 998 (1974).
49. G. Siedner and A. Horowitz, *Nature (Lond.)* **250**, 23 (1974).
50. W. A. Berggren, in *Initial Reports of the Deep Sea Drilling Project* (Government Printing Office, Washington, D.C., 1972), vol. 12, pp. 953–964.
51. N. J. Shackleton and J. P. Kennett, in *ibid.*, vol. 29, in press.
52. J. P. Kennett and P. Vella, in *ibid.*, vol. 29, in press.
53. W. J. Humphreys, *Physics of the Air* (McGraw-Hill, New York, 1940); H. Wexler, *Sci. Am.* **186**, 74 (1952); H. H. Lamb, *The Changing Climate* (Methuen, London, 1966); *Climate, Present, Past, and Future* (Methuen, London, 1972), vol. 1; J. P. Kennett and N. D. Watkins, *Nature (Lond.)* **227**, 930 (1970); J. C. Schofield, *N.Z. J. Geol. Geophys.* **13**, 737 (1970); A. J. Gow and T. Williamson. *Earth Planet. Sci. Lett.* **13**, 210 (1971).
54. J. Chappell, *Quat. Res.* **3**, 221 (1973).
55. H. Stille, *Grundfragen der Vergleichenden Tektonik* (Borntraeger, Berlin, 1924).
56. J. Gilluly, *Geol. Soc. Am. Bull.* **60**, 561 (1949).
57. M. Brookfield, *Earth Planet. Sci. Lett.* **12**, 419 (1971).
58. H. Kuno, *Geol. Surv. Can. Pap. 66* (1965), pp. 317–336; W. R. Dickinson, *J. Geophys. Res.* **73**, 2261 (1968); D. H. Green and A. E. Ringwood, *Contrib. Mineral. Petrol.* **18**, 105 (1968); T. Hatherton and W. R. Dickinson, *J. Geophys. Res.* **74**, 5301 (1969); D. L. Turcotte and E. R. Oxburgh, *Phys. Earth Planet. Interiors* **1**, 381 (1968); E. R. Oxburgh and D. L. Turcotte, *Geol. Soc. Am. Bull.* **81**, 1665 (1970); *J. Geophys. Res.* **76**, 1315 (1971).
59. K. Mogi, *Tectonophysics* **22**, 265 (1974).
60. P. R. Vogt, G. L. Johnson, T. L. Holcombe, J. G. Gilg, O. E. Avery, *ibid.* **12**, 211 (1971).
61. Tj. H. Van Andel and T. C. Moore, *Nature (Lond.)* **226**, 328 (1970).
62. R. J. Blakely, *J. Geophys. Res.* **79**, 2979 (1974).
63. J. R. Heirtzler, G. O. Dickson, E. M. Herron, W. C. Putman, X. Le Pichon, *ibid.* **73**, 2119 (1968); J. D. Phillips, *Science* **157**, 920 (1967); ———, G. Thompson, R. P. Von Herzen, V. T. Bowen, *J. Geophys. Res.* **74**, 3069 (1969); Tj. H. Van Andel and C. O. Bowin, *ibid.* **73**, 1279 (1968).
64. F. J. Vine, *Science* **154**, 1405 (1966).
65. R. K. Mathews, *Earth Planet. Sci. Lett.* **5**, 459 (1969).
66. N. Mörner, *Tellus* **24**, 586 (1972).
67. F. J. Mauk and M. J. S. Johnson, *J. Geophys. Res.* **78**, 3356 (1973).
68. W. L. Hamilton, *ibid.*, p. 3363.
69. W. A. Berggren, in *Symposium on Messinian Events in the Mediterranean*, C. W. Drooger, Ed. (North-Holland, Amsterdam, 1974), pp. 10–20; *Lethaia*, **5**, 195 (1972); Tj. H. Van Andel, G. R. Heath, T. C. Moore, *Geol. Soc. Am. Mem.*, in press.
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Environmental Mutagenic Hazards

Mutagenicity screening is now both feasible and necessary for chemicals entering the environment.

It has become increasingly clear in recent years that man is environmentally exposed to a wide variety of chemicals, some of which are mutagens (agents that induce mutations). The possibility therefore exists that human beings may be exposed, now or in the near future, to chemicals with individually or collectively powerful mutagenic effects. We describe here the ways in which environmental mutagens may be detected, possible methods of estimating the resulting adverse effects on human health, and regulatory principles that bear on the problem of preventing environmental mutagenesis.

Mutation consists of abrupt heritable changes in the composition or arrangement of genes, which are composed of deoxyribonucleic acid (DNA). Most mutations producing effects large enough to be observed are deleterious, although other mutations may produce effects of little or no consequence, and certain rare mutations may even be advantageous. The magnitudes of spontaneous mutation rates, the way selection acts on various gene combinations, and the size and structure of human populations are sufficient to maintain a rich source of genetic variability. An artificially increased mutation rate, however, is potentially capable of producing a general decline in genetic health unless balanced by increased selection against deleterious mutant genes; while such selection occurs extensively in most natural populations, the efficacy of modern medicine may increasingly tend to reduce selection against deleterious traits in many human populations.

Many geneticists believe that man's genes constitute his most precious heritage, and that a deterioration in gene quality can result in a corresponding decrease in the quality of life. Steady progress in the control of infectious diseases, lengthening human life spans, and improved procedures for identifying genetic disorders have revealed an important residue of genetic disease in human populations. An impressive proportion of hospital admittances, for instance, are now recognized as reflecting genetic disabilities (*1*). The prospects for directly curing the resulting genetic diseases, in contrast to merely alleviating their symptoms, are poor, and are not likely to improve in the near future. Furthermore, the wide variety of mechanisms by which radiations and chemicals induce mutations (*2*) make it very unlikely that generalized schemes can be devised to protect against mutagens, except by avoiding them in the first place.

Considerations such as these have recently led geneticists throughout the world to seek test systems capable of detecting environmental mutagens. A number of reports about the environmental mutagenesis problem have already appeared (*3*, *4*), and a special sec-

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