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Explosive Cenozoic Volcanism and Climatic Implications

Tectonic plate motion modifies the marine record of explosive volcanism and complicates its interpretation.

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The study of climate dynamics has received a strong impetus in recent years. Theories of the impact of volcanic dust on paleoclimate change have appeared sporadically in the past century. Recently, Kennett and Thunell (*1*) reported that greatly increased explosive volcanism in the past 2 million years, as recorded in deep-sea cores, coincided with greatly increased continental glaciation, and this was emphasized by Hammond (*2*). The possibility of such a correlation is certainly important enough to warrant a further hard look at the available pertinent data. To do this, we have examined both the volcanogenic material in all of the Lamont-Doherty piston cores and the published site data from all of the Deep Sea Drilling Project (DSDP) sites off east and southeast Asia. We selected this region because of its high volcanicity and the presence of a marine environment downwind from the volcanic sources.

Volcanogenic sedimentary data come from two sources, DSDP and piston cores. Core data from DSDP sites are sparse (as will be shown) and are best interpreted when compared to information on the geographic distribution of ash layers retrieved from piston cores that penetrate Recent and Pleistocene sediments. Piston core data are continuous and complete and come from numerous locations. But such data are limited by lack of deep penetration. The DSDP drilling penetrated to far greater depth, but only rarely was it interrupted for the purpose of taking complete cores from the hole. This leads to difficulties in interpretation. Also, all interpretations of history based on DSDP data, extending over some 60 million years, must include consideration of the sea-floor motion relative to the volcanic sources, a motion on the order of 100 kilometers per million years. Clearly, older volcanic material has been lost through subduction. Further, old sediments that do not contain ash layers have been carried into the ash-layer zone.

Data from Indonesia

Volcanic history of Indonesia. The present volcanic Indonesian Arc, apart from Sumatra, was formed in the early Tertiary (*3, 4*). The history of Sumatra extends back to the Paleozoic. The complete Cenozoic tectonic and magmatic history can be divided into two major phases: the first extended into the Early Miocene, and the second began in the Late Miocene and lasted until the present. Nearly complete submergence of the arc occurred during the Middle Miocene. Both phases were characterized by intense magmatic activity, except for the submergence period, when almost no activity occurred. Magmatism in the early phase was limited to intrusive activity until the Early Miocene, when extensive andesitic volcanism became active until the submergence.

The second magmatic phase in the Late Miocene began with granitic intrusion related to the uplift of sediments of Middle Miocene age. The associated explosive volcanism in the eastern section of the arc (Fig. 1) began and continued as andesitic eruptions. In westernmost Java and in Sumatra, volcanism began as rhyolitic eruptions and, as inferred from deep-sea cores, continued until the Late Pleistocene. The time history of explosive volcanism in the area of the Sunda Strait (between Sumatra and Java) was somewhat different from that in north Sumatra. Eruptions in the Sunda Strait area started in the Late Miocene and lasted until the Pleistocene (*3*), and in north Sumatra there was a single eruption about 70,000 years ago (*5*). The rhyolitic eruptions in Sumatra and westernmost Java were followed by recent andesitic volcanic activity. The recent andesitic eruptions were much less violent than the preceding rhyolitic activity, judged by their limited extent as observed in cores V19-150 and V19-151 from the trench south of the Sunda Strait (Fig. 1). Both terrestrial and marine observations indicate that the very violent rhyolitic ash–producing activity ceased about 70,000 years ago and was followed by quieter, less explosive andesitic eruptions. Because south Sumatra displays a series of many layers of rhyolitic tuffs beginning in the Late Miocene, compared to less numerous occurrences in north Sumatra, it appears that south Sumatra experienced more frequent explosive volcanism. This becomes important in core data interpretation.

Marine data from Indonesia. The locations of all of the Lamont piston cores taken off Indonesia are shown in Fig. 1 together with the DSDP sites (*6, 7*). All of these piston cores have been examined for the occurrence of volcanic ash

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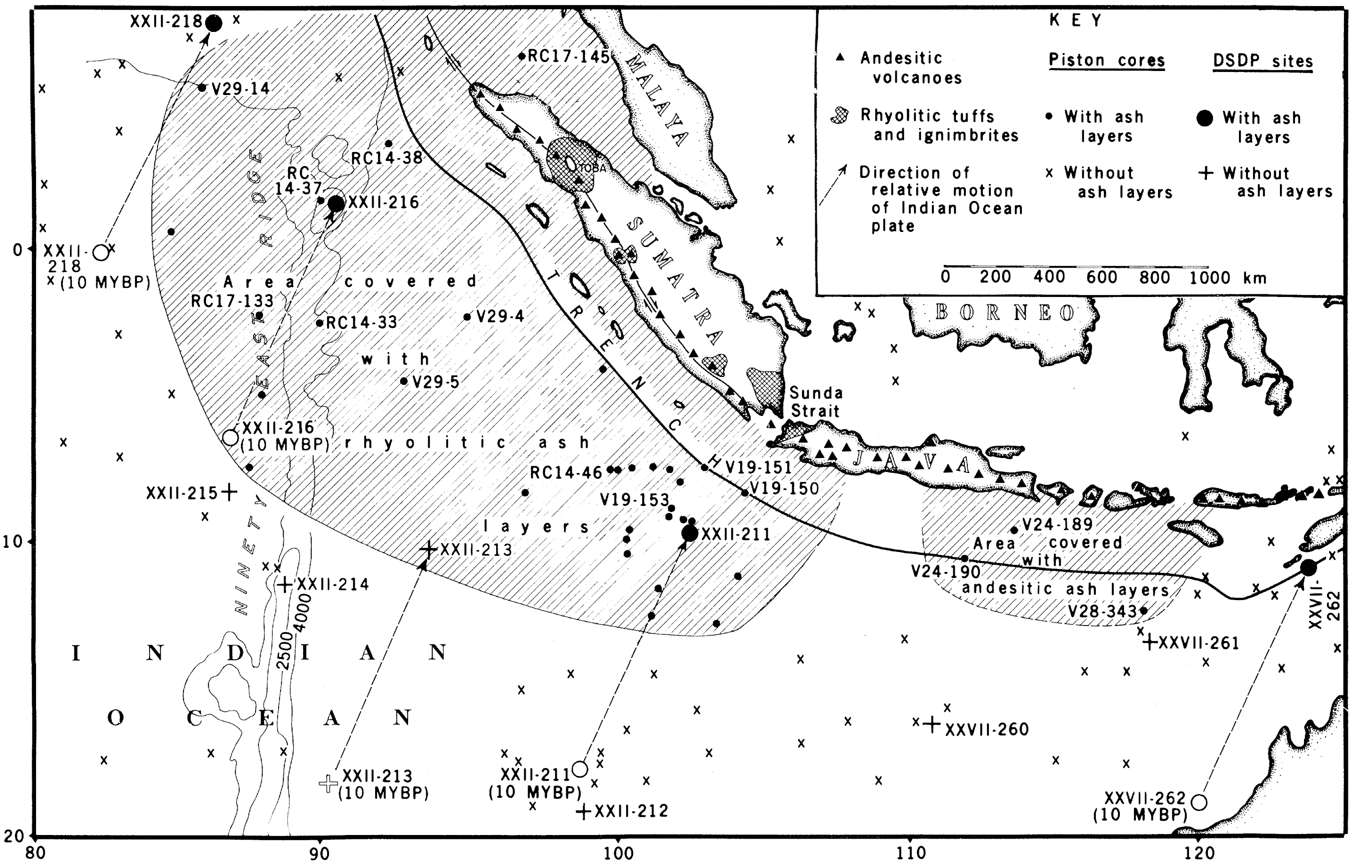


Fig. 1. Map of the western portion of the Indonesian Arc and adjacent seas, showing the distribution of rhyolitic tuffs and ignimbrites and andesitic volcanoes on the islands, and related ash deposits in deep-sea sediments. Locations of piston cores and DSDP cores with and without ash layers are denoted by different symbols (see Key). The shaded region defines the areas of Pleistocene ash layers on the basis of piston core data. The open circles show locations of the DSDP ash-layer sites about 10 million years ago, as relocated on the basis of sea-floor motion; *MYBP*, million years before present.

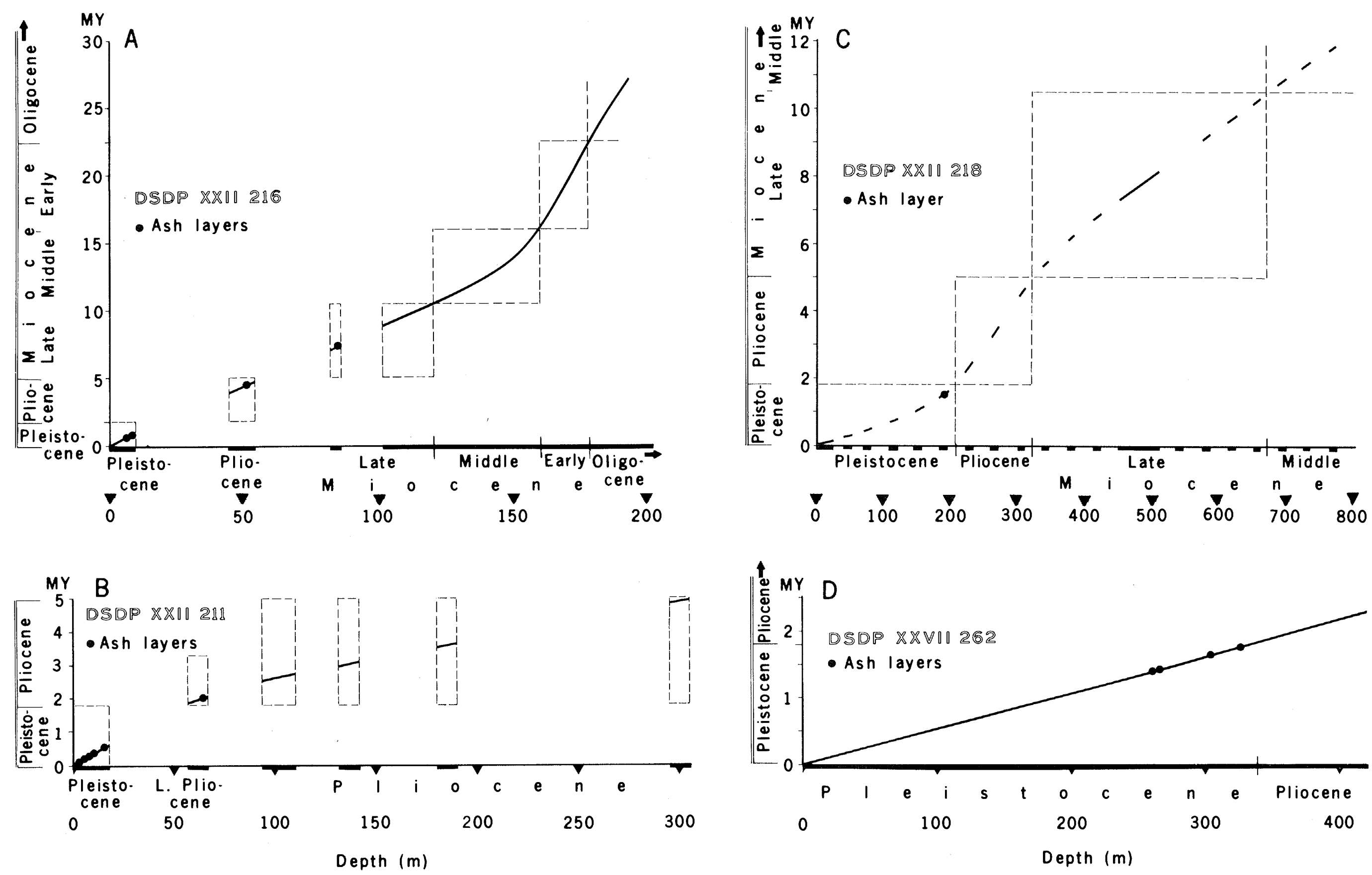


Fig. 2. Diagrams showing amount of core recovery and ash-layer distribution (•) for the four DSDP sites in Fig. 1. The boxes show the uncertainty in age for the sediment recovered.

layers. In the shaded zone to the west of Sumatra in Fig. 1 the ash is rhyolitic, with the addition of andesitic ash near the top of the two trench cores, V19-150 and V19-151. Cores from the shaded zone to the southeast of Java contain only andesitic ash.

The distribution of the ash zones appears to be related primarily to the tropospheric trade winds that prevail in the area. The composition of the ash in each of the shaded zones in Fig. 1 matches that observed on the islands, as described in the section above, and indicates that the known volcanic activity on the respective islands is the source of the adjacent marine ash layers. All of the piston core recoveries are limited to the Pleistocene.

To reach ages greater than Pleistocene, DSDP site observations are used. Interpretation of information from these sites must include consideration of the motion of the Indian Ocean floor relative to the Indonesian Arc because sediment and ash are continuously destroyed by subduction as plate motion carries this material to the trench bordering the arc. The dashed lines between the drilling sites and the open circles in Fig. 1 indicate the amount of plate motion (*8*) that occurred since the Late Miocene explosive volcanic phase began (about 10 million years ago). Note that the only site that would have been in an ash-layer zone continuously is site 216. Site 211 would have entered the ash-layer zone more recently. The other two sites are essentially out of the ash-layer zones.

The core recovery information from the four ash-bearing DSDP sites (*6, 7*) is shown in Fig. 2, in which stratigraphic depth is plotted against absolute age established by Berggren and Van Couvering (*9*). Sites 216 (Fig. 2A) and 211 (Fig. 2B) have short bottom extensions to the Cretaceous basement, which is not included here. Continuous core recovery was obtained for site 262 (Fig. 2D), which includes the entire Pleistocene and Late Pliocene. The other three sites were cored discontinuously, as indicated by the stratigraphic breaks. We will consider sites 216 and 211 together because of their position in the ash-layer zone.

Site 216 (Fig. 2A), the only site continuously in the ash-layer zone since the beginning of the rhyolitic explosive phase in the Late Miocene, shows two ash layers in the Pleistocene, one in the Pliocene, and one in the Late Miocene, interpolated to be about 7 million years old. From 9 million years ago to the Cretaceous basement no ash layers are found. No increasing frequency of volcanism can be inferred from this site, for it would be a remarkable coincidence if the small sections cored were taken from the only two ash layers in Pliocene and Late Miocene time. We would, on the contrary, expect as many ash layers in the missing sections as in the cored sections. Also, the absence of ash layers older than 9 million years is easily explained. The site had not yet entered the ash-layer zone. Site 211 shows five ash layers in the Pleistocene, one in the Late Pliocene, and none earlier. Again, there is no reason not to expect that other ash layers occurred in the missing section between Late Pleistocene and Late Pliocene time. Also, we cannot say whether any ash layers occurred before the Late Pliocene event and the underlying barren zone (~ 100 m in depth). But the lack of ash layers below the 100-m depth is understandable, because the site would not yet have entered the ash-layer zone. Of these two sites, only 216 is really representative of the explosive volcanic history since the Late Miocene. Site 211 gives only the later part of the history. We note that the number of Pleistocene ash layers at site 211 is greater than that at site 216 within the same ash-layer zone. This appears to be a consequence of the motion of site 211 close to the source of most frequent explosive volcanism, located, as noted earlier, in south Sumatra.

Site 218 is just beyond the northern edge of the ash-layer zone (Fig. 1). During its movement, the site would have barely touched the margin of the ash zone during the latter half of the interval since the Late Miocene. Core recovery data for this site are shown in Fig. 2C, which indicates that semicontinuous recovery was obtained to the Middle Miocene, the limit of drilling. Only one ash layer occurs at the beginning of the Pleistocene. From the site motion in Fig. 1, we would expect that any ash that does occur would be only in the upper part of the record. There are enough core data above and below this ash layer to indicate that there were no other significant ashfalls at this site for the time interval represented. This is also what would be expected from the motion of the site relative to the ash-layer zone.

Site 262 (Fig. 1), which was drilled in the Java Trench outside an ash-layer zone, was cored continuously through the Late Pliocene. Four ash layers occur in the Pleistocene (Fig. 2D). The data from this core are not particularly instructive about the volcanic history of the region because (i) the drilling operation did not penetrate below a depth corresponding to the Late Pliocene, and (ii) the section above the ash layers, when the site presumably reached the trench, shows thick sediments with nannofossils that include Cretaceous and Cenozoic forms (*7*). This indicates rapid sedimentation from turbidity currents and obscures the Pleistocene chronology of the ash layers.

At most of the sites located outside the ash-layer zone (Fig. 1) core recovery for the Cenozoic is excellent. But none of these sites contain ash layers. This indicates the importance of considering the positions of the DSDP sites relative to ash-layer zones in making historical interpretations.

The data and interpretation of Kennett and Thunell (*1*) for Indonesia are shown in the last column of figure 4 in their article. The curve showing an increase in the number of ash layers beginning 3 to 4 million years ago is similar to the data we have shown in Fig. 2. It is our contention, however, that this does not represent an increase in volcanism, as proposed by Kennett and Thunell (*1*), but is a consequence of the sea-floor plate motion relative to the ash-layer zone. Kennett and Thunell did not have all of the pertinent data on the areal distribution of ash layers available from our piston cores. Our interpretation might be questioned because of the extremely high, if narrow, peak exactly at 12 million years in Kennett and Thunell's Indonesian data. However, according to their figure 1 (*1*), these data apparently come from DSDP site 212, which is also in our Fig. 1. But we can find no record of any volcanic ash at this site in the DSDP report. Also, we find no evidence of any ash older than Late Miocene for any of the Indonesian DSDP sites (*6, 7*). This absence of ash older than 10 million years is in agreement with our interpretation based on sea-floor motion.

Our conclusion is supported by the work of Vallier and Kidd (*10*), who give a detailed volcanogenic analysis of all of the DSDP site core material from the entire Indian Ocean. Rather than recording only ash layers, they studied the distribution of volcanogenic material throughout the entire sediment series. The distribution of silicic glass shards in the sediments shows essentially continual production since the beginning of the Late Miocene. Although they analyzed all DSDP sites in the Indian Ocean, a particularly detailed analysis of glass shards was made for site 213 (Fig. 1), which contains no distinctive ash layers despite the continuous, disseminated shard content. It seems clear that the absence of discrete ash layers is a consequence of the location of the site relative to the ash-layer zone since the Late Miocene. A small amount of ash, not adequate to comprise a distinct layer, is always airborne to great distance. This constitutes the disseminated material.

Vallier and Kidd (*10*) conclude, from the detailed study of site 213 and other sites in the Indian Ocean, that pulses of explosive volcanism occurred rather uniformly in the last 10 million years. No evidence of a Late Cenozoic increase in such activity is found in their data.

Data from the Northwest Pacific

In examining the northwest Pacific region we recognize three areas with somewhat different histories: the Bering Sea north of the western Aleutians; the chain of arcs from Kamchatka through the Kurils and Japan; and the seas adjacent to the Marianas Arc. There is a large number of piston cores and DSDP data here, and all of the ash could have fallen into the sea, so that a good marine record is expected.

The Cenozoic volcanic history of the Japanese, Kuril, and Kamchatka arcs is very similar to that of Sumatra (*11*, *12*). A "green tuff" phase of activity that occurred in the Early Miocene was followed by a profound submergence of the Japan and Kuril arcs in the Middle Miocene. The Late Miocene emergence was associated with rhyolitic explosive activity, which continued into the Pleistocene. This, in turn, was followed by a more recent andesitic volcanic phase, which is currently active.

All available piston cores from the Lamont collection from these regions were analyzed for the presence of volcanic ash layers. The locations are shown in Fig. 3, which also indicates the DSDP sites that contained ash layers. The shaded area in Fig. 3 shows the distribution of volcanic ash layers in piston cores of Pleistocene age (*13*). The ash indicated in this region has been carried from the adjacent arcs by the westerlies. The Marianas (which lie in the trade winds) have no neighboring ash-layer zone. This is expected because the geological history of the Marianas (*14*) shows that the intense explosive volcanism of the region was limited to Oligocene and Miocene time. We will first examine the data from DSDP sites for the Marianas (*15*) and the Bering Sea (*16*), as these sites were stationary relative to adjacent arcs during the Cenozoic. Then we will consider data from the other sites from the northwest Pacific plate (*15*), which has been in motion relative to the adjacent arcs.

The Marianas region. As noted above, no Pleistocene ash-layer zone lies adjacent to the Marianas. There are two DSDP sites (Fig. 3) with ash layers to the west of the Marianas, in the Philippine Sea, and one to the east, between the arc and the trench. Data from these sites, compiled from the DSDP reports (*15*), are shown in Fig. 4. Note that core recovery is very sparse at these sites. For site 53, apart from a 20-m sediment section barren of ash on the top and a thin barren section on the bottom, all of the cored sediment described contains numerous ash layers with microfossils that indicate ages. The ash deposits do not occur above the Late Miocene and reach at least the Early Oligocene. Although only one of the three sites (site 53) has recovery from the top of the hole, the absence of volcanic ash in the upper 20 m correlates with a similar absence in the records of the piston cores in this region.

Sites 54 and 60 (Fig. 4) have an even poorer percentage of core recovery, and at both sites all the material recovered contained sediments interbedded with numerous ash layers of Middle Miocene age. From these data we can only conclude that explosive volcanism in the Marianas region extended from the Early Oligocene to the Late Miocene. No significant conclusions about volcanic frequency can be drawn from these data.

The tectonic history of the Marianas region (*17*) shows that these sites are located on portions of the sea floor that have been essentially stable relative to the Marianas throughout the Cenozoic, unlike the DSDP site locations described for the Indian Ocean. The composite of these sparse data gives a very rough, discontinuous history of most of the Cenozoic, and, for the Marianas at least, an absence of Late Cenozoic explosive volcanism seems evident.

Bering Sea. The three sites, 189, 190 and 191, in the Bering Sea (Fig. 3) are located on a plate that has been stationary relative to Kamchatka (*18*) and, like

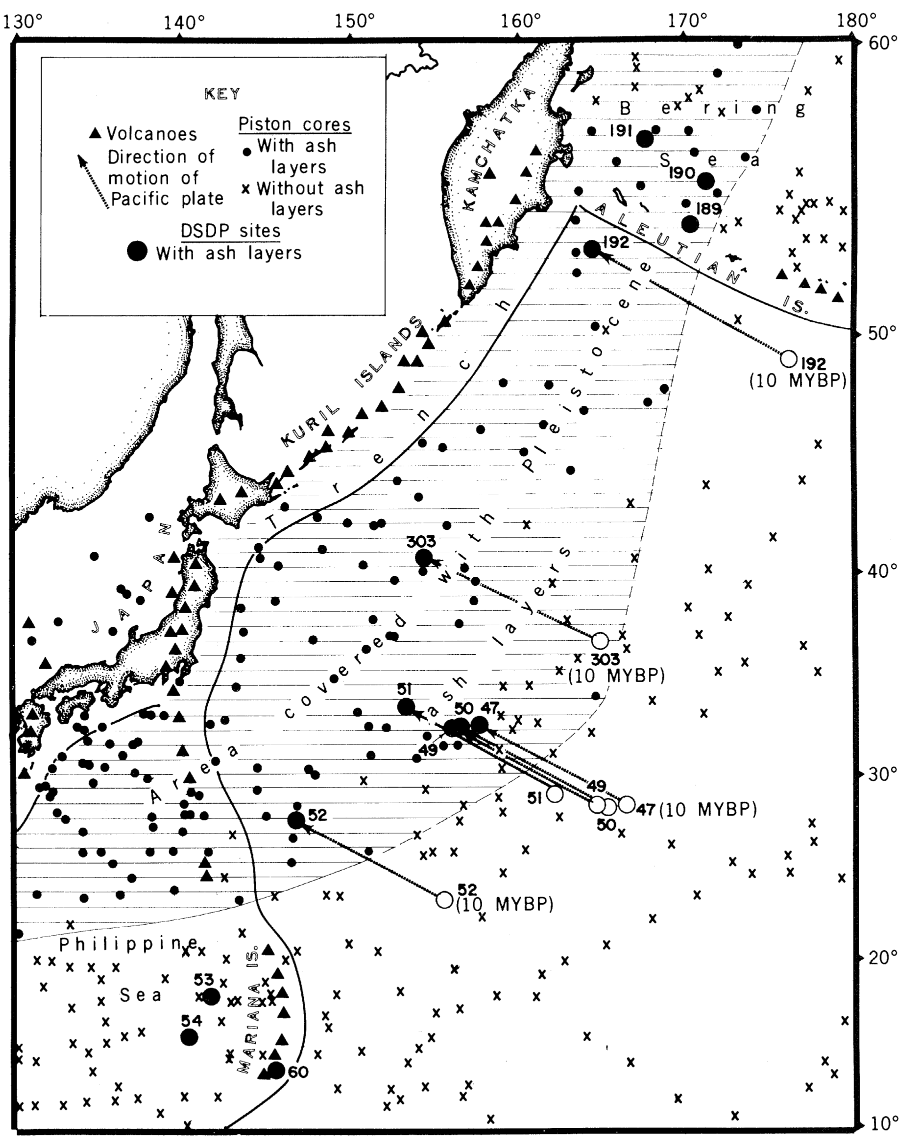


Fig. 3. Distribution of ash layers for the northwest Pacific. All symbols are defined as in Fig. 1. Note that north of the Aleutians and in the Philippine Sea, the DSDP sites are stationary relative to the arcs bearing the volcanic ash sources.

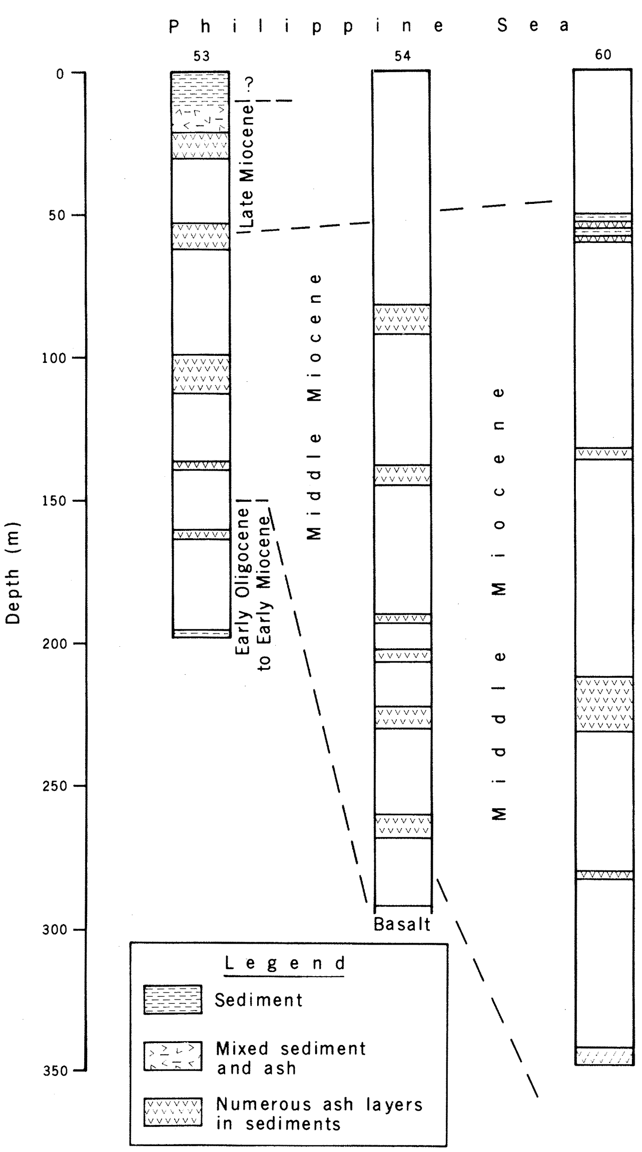
those for the Marianas, could give a fairly complete Cenozoic record of explosive volcanism in the neighboring arc.

Figure 5 shows the core recovery information for the three sites in the Bering Sea. Note again how poor actual core recovery is for these sites. Of the three, site 190 has the largest percentage of core information and penetrates sediment of Late Miocene age. The number of ash layers in the sediment of each recovered section is indicated in Fig. 5. Ash layers are present in all but one thin section. The number of ash layers is simply proportional to the thickness of the recovered section and indicates nothing, in this case, about the frequency of explosive volcanism. If anything, uniform activity is indicated from the Late Miocene to the present time. Recovery at sites 189 and 191 is so poor that little interpretation can be made, but the data generally agree with the above conclusion.

The Northwest Pacific plate. The Pacific Ocean plate is moving toward Asia and is subducted along the trenches marginal to the chains of arcs in the western Pacific (Fig. 3). The amount of plate motion (*8*) is shown in Fig. 3 by the broken arrows connecting present DSDP sites that bear ash layers with their locations 10 million years ago. We would not expect ash layers at these sites until the site vectors entered the ash-layer zone shown in Fig. 3, which was developed from the ash distribution in piston cores penetrating Recent and Pleistocene sediments.

Logs of the core recovery data for the DSDP sites east of Japan are shown in Fig. 6. Of the five sites east of Japan, site 47 has nearly continuous recovery of the Cenozoic geologic record. Only two ash layers are found, one in the Early Pleistocene and the second in the Late Pliocene. According to the site vector in Fig. 3, no older ash is expected because the site lay outside the ash-layer zone.

Core data from the adjacent sites 49 and 50 (Fig. 6) show Pleistocene sediment lying on Cretaceous sediment. Although a zone of intermixed ash and sediment is present in the Pleistocene section of each core, no historical interpretation can be made. At site 51, core recovery is very scanty and a thin sediment section of Early Pleistocene–Late Pliocene age is totally intermixed with sediment. Again, no interpretations can be made. For site 303, four thin core sections were recovered above the Cretaceous level (*19*). One zone of intermixed ash and sediment is present in the recovered part of the Pleistocene and one ash layer is present in the Late Mio-



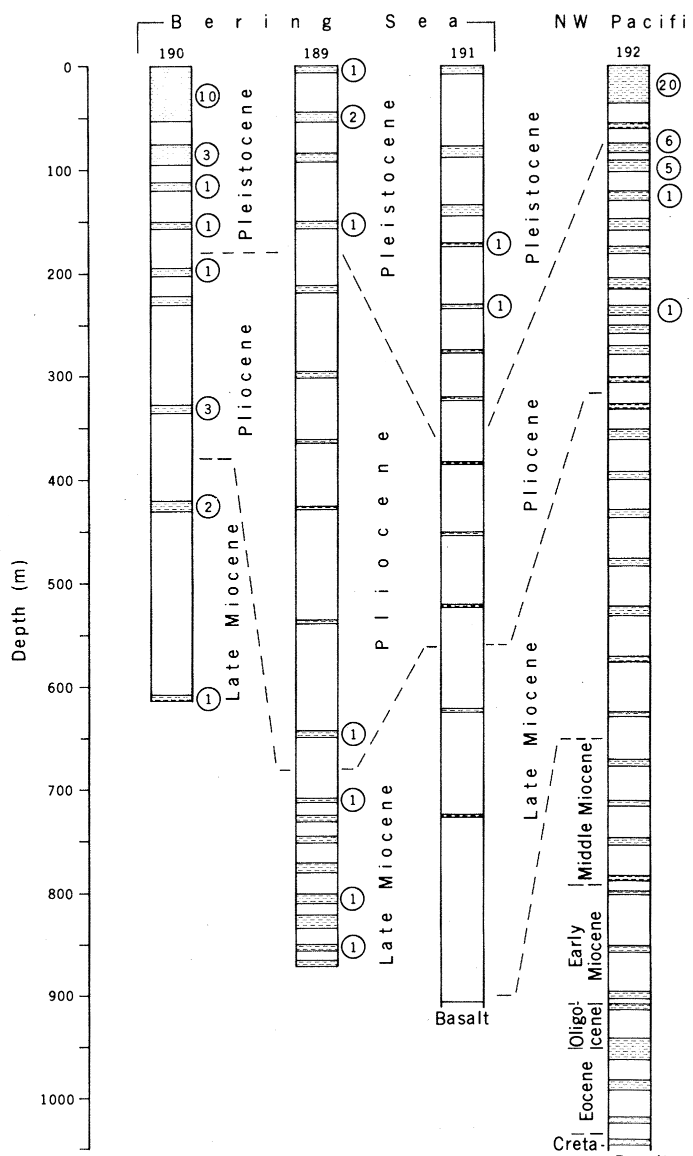


Fig. 4 (left). Columns showing the lithology in the relatively small core recovery zones for DSDP sites in the Marianas area. Fig. 5 (right). Columns showing the vertical distribution of sediment zones recovered in cores from DSDP sites east of Kamchatka. Circled numbers are the numbers of ash layers in the recovered sediment sections.

cene. The entire site vector for the last 10 million years lies within the ash-layer zone so that ash older than that at sites 47, 49, 50, and 51 is expected. All of the data from sites east of Japan are essentially useless in interpreting the history of explosive Cenozoic volcanism, but do not indicate any Late Cenozoic increase in such activity. We should note that we disregarded site 52 in Fig. 3 despite its good core recovery and numerous ash zones, because the record from Miocene to Recent time is reported (*15*) to be thoroughly mixed.

Site 192, far to the north but still on the Pacific plate (Fig. 3), lies close to the sites described for the Bering Sea. Because of the scarcity of piston cores (particularly with ash layers) for many degrees immediately south of the Aleutians, the eastern limit of the ash-layer zone cannot be located very precisely in these areas. Apparently the site vector entered the ash-layer zone more recently than about 5 million years ago. As with the Bering Sea sites, the ash in site 192 probably came from Kamchatka. But unlike these sites, site 192 does not contain ash older than Middle Pliocene (Fig. 5) because of the plate motion. After the Middle Pliocene there appears to be a numerical increase in ash layers for this site. This is reasonable, because the site moved close to the volcanic sources and could have received ash from relatively small eruptions. This increase in the number of ash layers as site 192 approaches the volcanic source is like the case of site 211 off south Sumatra as compared to site 216 off north Sumatra.

From the data and analysis given we draw the following conclusions:

1) The amount of data available from DSDP sites is very small, so that there is a potential for serious errors in interpretation.

2) Any interpretation that is made must be consistent with the data on sea-floor plate motion.

3) The data from Recent and Pleistocene piston cores provide a valuable reference for the time when ocean plates entered the marine ash-layer zone marginal to island arcs.

4) The best conclusions we can reach on the basis of information from Indonesia and the western Pacific area are that (i) the increase in the number of ash layers in the deep-sea cores of Late Cenozoic age, which was interpreted by Kennett and Thunell (*1*) as indicating a great increase in explosive volcanism, is more readily explained by motion of the sea floor and the data sites relative to the volcanic ash-layer zones, as well as motion toward the eruptive sources of volcanogenic sediments; and (ii) if anything, approximate uniformity of explosive volcanism is indicated by data from sites that were essentially motionless relative to volcanic sources, as in the western Bering Sea and the Marianas region.

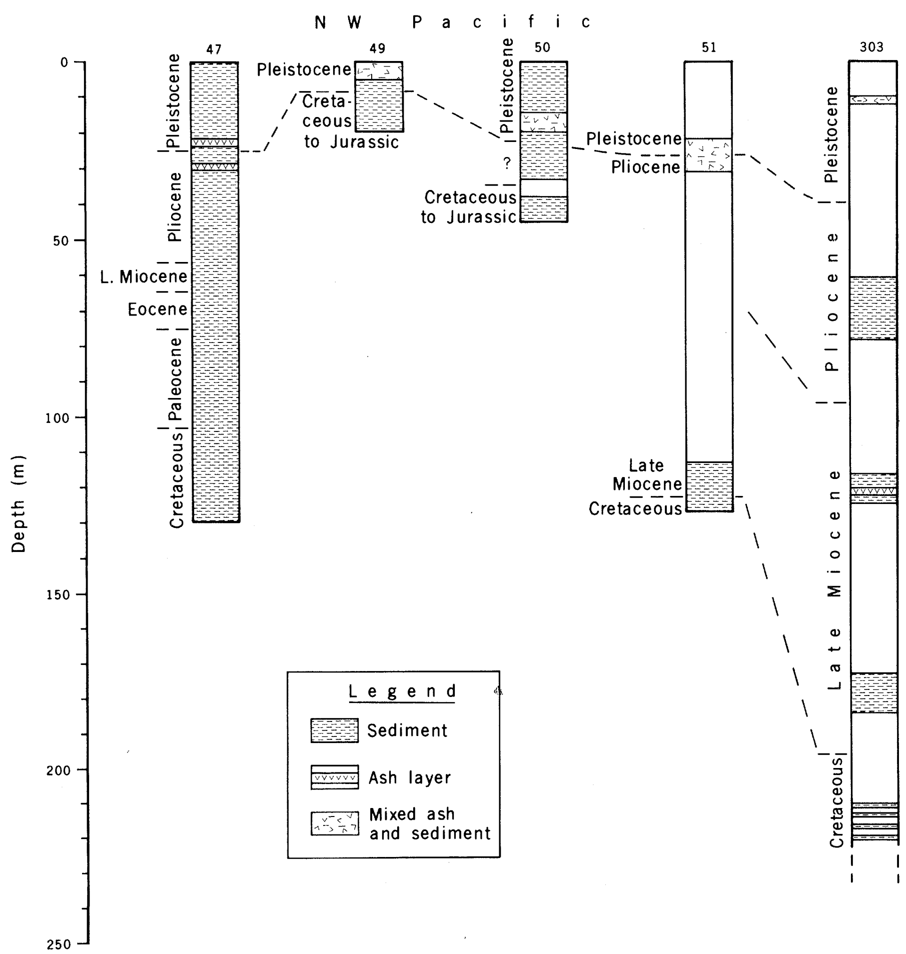


Fig. 6. Columns showing lithology and core recovery for DSDP sites east of Japan.

Volcanic Dust and Climate

Explosive volcanic activity can create a stratospheric dust veil that shields the earth's surface from a small percentage of direct solar radiation. However, the increase in diffuse sky radiation tends to offset much of this radiation loss. In the case of the eruption of Mount Agung on Bali, Indonesia, in March 1963, observations in Melbourne, Australia, showed that monthly total radiation was only reduced by 6 percent, while the solar beam suffered a 24 percent reduction (*20*). These values have been shown to vary with local and temporal meteorological conditions, but seem to indicate a general relationship.

The effect of a volcanic dust veil on climate has short- and long-range aspects. To assess the former we have recourse to historical meteorological records; for the latter we must examine the geological record.

The short-range effect. On the basis of temperature changes (*21*) following three great volcanic eruptions in the 19th century, Lamb (*22*) concludes that the middle latitudes suffer a decrease of 0.5° to 1°C in the year after a large eruption. Although two of these eruptions were within a 3-year period, temperatures quickly recovered in these as well as in all other cases of a temperature drop apparently following explosive volcanism. Also, in the Köppen temperature-time sequence, reported by Lamb, no significant effect seems to be related to the great Krakatau eruption of 1883, despite the pronounced atmospheric effects reported for years afterward (*23*).

There seems little doubt that many great eruptions close in time would charge the stratosphere with sufficient aerosols to seriously obscure total solar radiation, thereby causing a pronounced and possibly prolonged decrease in the global temperature. So far, there is little in the historical record to indicate that such frequent events have occurred. The little ice age (~ 1500 to 1850) is the most significant temperature change documented since the climatic optimum, but there is no ready correlation between this event and prolonged and frequent great eruptions, which averaged a few per century in historic times.

It may well be that the immediate consequences of global temperature decreases following explosive eruptions are climatic anomalies that are both local and temporary in nature. For example, recent droughts in both Africa and the middle United States have been ascribed by some to global cooling related to solar variations. There is no reason why they should not also have been brought on by cooling from volcanic dust veils. This may well be a direction for further detailed data analysis.

Long-range climate aspects. In reviewing the history of Cenozoic climate, we note that the distribution zone of tropical and subtropical fauna shows a progressive contraction during the Cenozoic era (*24*), which indicates progressive global cooling. Evidences of early glaciation of Antarctica are found in deep-sea sediments of Eocene and Oligocene age (*25*). Although the extent of this glaciation is uncertain, widespread continental glaciation of Antarctica seems certainly to have been established in the Miocene (*26, 27*). In the Northern Hemisphere strong glaciation of western Alaska was established at least 10 million year ago (*28*), and pronounced cooling marked by a decrease in temperature of about 10°C is indicated 13 million years ago (*27*). Although glaciation of Iceland and temperate high elevations occurred about 3 million years ago (*29*), the vast glaciations of the continents appeared to be delayed until some 700,000 years ago (*30*).

The progressive Cenozoic cooling that culminated in vast continental glaciation has been shown by Donn and Shaw (*31*), by use of a thermodynamic model, to be a natural consequence of continental drift and polar wandering. These effects carried continents to high latitudes and isolated the polar regions thermally from the rest of the world ocean. There is no good evidence that the progressive cooling of the Cenozoic was connected to an insolation decrease due to explosive volcanism because there have been no adequate global correlations of marine volcanic sediments and no good time resolution of the frequency of ash layers. There is also no evidence that the successive vast glaciations of the continents (particularly in the Northern Hemisphere) beginning in the late Pleistocene were triggered by cyclical events of protracted explosive volcanism. However, it may well be that when DSDP cores are taken more continuously, the resulting greater resolution will show some intervals of increased explosive volcanism.

The modulating effect may have been more important in the geologic past because the magnitude of individual explosive events appears to have been greater than in recent history. The greatest of the historic eruptions were relatively insignificant, as measured by their ash production observed in deep-sea cores. Another way of estimating the volume of ash production is to relate this value to the volume of the caldera formed after the eruption (*32*). On this basis, we can classify some of the greatest recent eruptions in order of increasing magnitude as Krakatau (1883), Katmai (1912), and Tombora (1815), with respective caldera volumes of about 5, 12, and 30 km^3 (*33–35*). Also, no ash has been recovered by piston-and-trigger-weight cores from the sea around Krakatau. Katmai produced an ash layer of only 2 to 5 cm at a distance of 300 km from the source (*36*). Tambora, the greatest known recent eruption, produced ash recovered only by bottom sampling (*37*), which does not indicate thickness.

Eruptions of much greater magnitude occurred in prehistoric times in the cases of Mount Mazama, Oregon (6600 years ago), where Crater Lake now lies; Santorini in the Aegean Sea (3500 years ago); the huge eruption of Toba in north Sumatra (70,000 years ago); and the eruptions of Taupo and Rotorua in New Zealand between 0.3 and 1 million years ago. The caldera volumes for Mount Mazama and Santorini are both about 70 km^3 (*37, 38*); that for Toba is 2000 km^3 (*3*). Because Mount Mazama is in the western United States, westerly winds distributed its ash to the east, entirely over land, and its thickness compared to our deep-sea core standard cannot be determined but would certainly have been large. In the case of Santorini, the thickness of ash in deep-sea cores is 2 m at a distance of 100 km. At a distance of 400 km it is 2 cm (*38*). Toba produced an ash layer of 15 to 20 cm at a distance of 1000 km (*39*). In the case of Taupo and Rotorua, whose caldera sizes lie between those of Mount Mazama and Toba, resulting ash layers 2 to 10 cm thick are found in piston cores 1500 to 2000 km to the east of New Zealand (*40*).

Most of the DSDP cores referred to here are on the order of 1000 km downwind from the volcanic arcs, but they show ash layers whose thicknesses vary from 2 to 20 cm. These eruptions must have been closer in magnitude to Toba rather than to eruptions of the recent past.

The radiation effect resulting from these intense events in the past could then modulate a global trend. For example, the general cooling trend in the Cenozoic could have been steepened temporarily by volcanic effects. The positive feedback resulting from an increase in snow cover would then have further intensified the cooling. The effect of volcanic dust on global temperature is real. Whether enough volcanic dust has ever been present for a long enough period of time is still unknown. By our reasoning, the effect may not be a first-order one, but if it is an important second-order, or modulating, climate factor, it should be studied in enough detail to establish the chronology and the atmospheric consequences of explosive volcanism.

Conclusions

The history of Cenozoic explosive volcanism has been studied by using data from DSDP and piston cores from the Indian Ocean off Indonesia and the western Pacific Ocean. Data from piston cores, which only penetrate Recent and Pleistocene sediments, are used primarily to establish the limits of the respective ash-layer zones. Core recovery from the DSDP sites has been sparse, but the drilling has penetrated much deeper, often through the entire Cenozoic, so some assay of the volcanic history is possible. Some DSDP sites are on plates that have moved toward the subduction zone; others are on plates fixed relative to the volcanic arc. Within the limits of the DSDP core information, data from the fixed zones do not indicate a significant change in the frequency of occurrence of ash layers. Data from sites on moving plates show an increase in the number of ash layers in the Late Pliocene–Pleistocene record. But this increase in frequency corresponds to the time when the sites entered the ash-layer zone and also moved close to the volcanic sources. Therefore, the higher frequency of ash layers in these regions cannot be taken to indicate a trend in the frequency of explosive volcanism.

Explosive volcanism can generate a stratospheric dust veil that cuts down the intensity of solar radiation reaching the earth's surface. During historic times atmospheric temperature quickly recovered after relatively large eruptions. Only an unusually high frequency of eruptions could have a long-term climatic effect. During historic times this has not occurred, and the deep-sea core record has not been adequately correlated on a global scale, nor adequately resolved on the vertical time scale, to determine whether such a high frequency of explosive volcanism occurred during the Cenozoic. It is true, however, that some explosive volcanism in the geologic past greatly exceeded in magnitude that in the historic past. Such events, when occurring at critical times of climate evolution, might have strongly modulated the intensity of climate change. We believe that a careful global chronology of explosive volcanism should be developed from available piston and DSDP core data, together with a related chronology of climate evolution. Only a study of this kind can settle the question of the possible influence of volcanism on climate.

References and Notes

1. J. P. Kennett and R. C. Thunell, *Science* **187**, 497 (1975).
2. A. L. Hammond, *ibid.* **191**, 208 (1976).
3. R. W. van Bemmelen, *The Geology of Indonesia* (Netherlands Government Printing Office, The Hague, 1949).
4. J. Westerveld, *Geol. Soc. Am. Bull.* **63**, 561 (1952).
5. D. Ninkovich, J. D. Hays, A. A. Abdel-Monem, *Geol. Soc. Am. Annu. Meet. Abstr.* **3**, 661 (1971).
6. C. C. von der Borch *et al.*, in *Initial Reports of the Deep Sea Drilling Project* (Government Printing Office, Washington, D.C., 1974), vol. 22, p. 13.
7. J. R. Heirtzler *et al.*, in *ibid.*, vol. 27, p. 89.
8. J. B. Minster, T. H. Jordan, P. Molnar, E. Haines, *Geophys. J. R. Astron. Soc.* **36**, 541 (1974).
9. W. A. Berggren and J. A. Van Couvering, *Paleogeogr. Paleoclimatol. Paleoecol.* **16**, 1 (1974).
10. T. L. Vallier and R. B. Kidd, *Geol. Soc. Am. Mem.*, in press.
11. M. Minato, M. Gorai, M. Hunahashi, Eds., *The Geologic Development of the Japanese Islands* (Shokam, Tokyo, 1965).
12. G. S. Gorshkov, *Volcanism and the Upper Mantle* (Plenum, New York, 1970).
13. D. R. Horn, M. N. Delach, B. M. Horn, *Geol. Soc. Am. Bull.* **80**, 1715 (1969); J. D. Hays and D. Ninkovich, *Geol. Soc. Am. Mem.* **126**, 263 (1970).
14. P. E. Cloud, R. G. Schmidt, H. W. Burke, *U.S. Geol. Surv. Prof. Pap. 280-A* (1956); J. L. Tracey, S. O. Schlanger, J. T. Stark, D. B. Doan, H. G. May, *U.S. Geol. Surv. Prof. Pap. 403-A* (1964).
15. B. C. Heezen *et al.*, in *Initial Reports of the Deep Sea Drilling Project* (Government Printing Office, Washington, D.C., 1971), vol. 6, p. 67.
16. D. W. Scholl *et al.*, in *ibid.* (1973), vol. 19, p. 325.
17. D. E. Karig, *Geol. Soc. Am. Bull.* **82**, 323 (1971).
18. W. J. Morgan, *J. Geophys. Res.* **73**, 1959 (1968); X. LePichon, *ibid.*, p. 3661.
19. R. L. Larson *et al.*, in *Initial Reports of the Deep Sea Drilling Project* (Government Printing Office, Washington, D.C., 1975), vol. 32, p. 17.
20. A. Dyer and B. Hicks, *Nature (London)* **208**, 131 (1965).
21. W. Köppen, *Meterol. Z.* **31**, 305 (1914).
22. H. Lamb, *Climate, Past, Present and Future*, vol. 1, *Fundamentals and Climate Now* (Methuen, London, 1972).
23. G. J. Symons, *The Eruption of Krakatoa and Subsequent Phenomena* (Trubner, London, 1888).
24. W. Durham, *Geol. Soc. Am. Bull.* **61**, 1243 (1950).
25. K. Geitzenauer, S. Margolis, D. Edwards, *Earth Planet. Sci. Lett.* **4**, 173 (1968); R. Rex and S. Margolis, *Antarct. J. U.S.* **4**, 168 (1969); S. Margolis and J. Kennett, *Science* **170**, 1085 (1970).
26. O. Bandy, *Am. Assoc. Petrol. Geol. Bull.* **50**, 643 (1966).
27. ———, E. Butler, R. Wright, *Science* **166**, 607 (1969).
28. G. Denton and R. Armstrong, *Am. J. Sci.* **267**, 1121 (1969).
29. T. Einarsson, D. Hopkins, R. Doell, in *The Bering Land Bridge*, D. M. Hopkins, Ed. (Stanford Univ. Press, Stanford, Calif., 1967), pp. 312–325; I. McDougall and H. Wensink, *Earth Planet. Sci. Lett.* **1**, 232 (1966).
30. J. D. Hays, T. Saito, N. D. Opdyke, L. H. Burckle, *Geol. Soc. Am. Bull.* **80**, 1481 (1969).
31. W. Donn and D. Shaw, *The Evolution of Climate* (Symposium on Long-Term Climatic Fluctuations, WMO No. 421, World Meteorological Organization, Aphora, Switzerland, 1975), pp. 53–62.
32. H. Williams, *Carnegie Inst. Washington Publ. 540* (1942), p. 162.
33. I. Yokoyama, *Bull. Earthquake Res. Inst. Univ. Tokyo* **35**, 75 (1957).
34. C. N. Fenner, *J. Geol.* **28**, 569 (1920).
35. M. Newmann van Padang, *Catalogue of the Active Volcanoes of the World*, part 1, *Indonesia* (International Volcanological Association, Napoli, Italy, 1951).
36. Y. R. Nayudu, *Mar. Geol.* **1**, 194 (1964).
37. G. A. Neeb, *Snellius Exped. East. Part Neth. E. Indies* **5** (No. 3), 265 (1943).
38. D. Ninkovich and B. C. Heezen, in *Colston Papers*, vol. 17, *Submarine Geology and Geophysics* (Butterworth, London, 1965), p. 413; *Nature (London)* **213**, 582 (1967).
39. D. Ninkovich, unpublished data.
40. ———, *Earth Planet. Sci. Lett.* **4**, 89 (1968).

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Biosynthesis and Function of Gangliosides

Gangliosides appear to participate in the transmission of membrane-mediated information.

Peter H. Fishman and Roscoe O. Brady

Gangliosides comprise a family of acidic glycolipids that are characterized by the presence of sialic acid. They are unusual compounds in that they contain both hydrophilic and hydrophobic regions, and they bear a strong negative charge. Gangliosides are primarily membrane components, and plasma cell membranes are highly enriched in these materials. The carbohydrate portion of gangliosides is made up of molecules of sialic acid, hexoses, and *N*-acetylated hexosamines. The hydrophobic moiety is called ceramide, and it consists of a long-chain fatty acid linked through an amide bond to the nitrogen atom on carbon 2 (C-2) of the amino alcohol sphingosine. Oligosaccharides are linked through a glycosidic bond to C-1 of the sphingosine portion of ceramide. The structure of one of the more common gangliosides called G_{M1} is illustrated in Fig. 1.

Gangliosides were first identified in brain more than 30 years ago by Klenk (*1*). In the ensuing years, much effort has been devoted to establishing the structures and physical properties of gangliosides. Within the past few years, the individual steps involved in the biosynthesis of gangliosides have been elucidated (*2–4*), and the reactions involved in the impaired catabolism of these compounds in heritable metabolic disorders such as Tay-Sachs disease have been established (*5*). Until recently, however, knowledge of the function of these components was extremely limited. Good evidence that these substances play significant roles in membrane-related phenomena is now available. Since the ceramide portion of the ganglioside molecule is embedded in the fluid phase of membranes and the negatively charged oligosaccharide chain is exposed to the external environment (*6*), these substances are well suited to participate in external signals and other events that occur on the surfaces of cells. In this article we briefly review current knowledge regard-

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