References and Notes

1. Mariner 6 and 7 were NASA missions managed by the Jet Propulsion Laboratory (JPL) and directed by Project Manager H. M. Schurmeier.
2. The spectrometers were designed and constructed by the authors in the Chemistry Department and Space Sciences Laboratory at the University of California at Berkeley. This instrument will be described in detail in a later publication.
3. Provided by Optical Coating Laboratories, Inc., Santa Rosa, California.
4. All times, latitudes, and longitudes are based upon JPL Pegasis calculations made on 8 August 1969 after the encounter. Because the Mariner 7 spacecraft experienced an orbit anomaly before the encounter, these latitudes and longitudes are subject to change as post-encounter tracking refines the actual orbit.
5. There was no trace of either of these absorptions in any of the 150 spectra recorded during the Mariner 6 flyby, which viewed an equatorial region between 13°N and 16°S and between 280°E and 95°E.
6. W. E. Osberg and D. F. Hornig, *J. Chem. Phys.* **20**, 1345 (1952).
7. D. A. Dows, *Spectrochim. Acta* **13**, 308 (1959).
8. E. D. Becker and G. C. Pimentel, *J. Chem. Phys.* **25**, 224 (1956).
9. In a separate experiment, conducted with Y. M. Huang, CH_4 was trapped in solid CO_2 at 20°K. When the sample was warmed to 77°K, the spectrum of CH_4 became more diffuse and somewhat weaker, but plainly most of the methane remained. After the solid was warmed to 100°K and recooled to 77°K, the amount of methane was reduced by approximately half. The experiment shows that methane can be trapped in solid CO_2 even at temperatures as high as 100°K and that near 3020 cm^{-1} the spectrum of methane suspended in solid CO_2 is not readily distinguishable from the Mars absorption.
10. G. Herzberg, *Infrared and Raman Spectra of Polyatomic Molecules* (Van Nostrand, Princeton, N.J., 1945).
11. B. R. Cairns and G. C. Pimentel, *J. Chem. Phys.* **43**, 3432 (1965).
12. B. R. Cairns, thesis, University of California, Berkeley (1964).
13. No methane absorption was observed at 1305 cm^{-1} in the martian spectra, but this does not constitute sufficient evidence to eliminate the methane interpretation. Spectra recorded in this spectral region depend upon the temperature difference between the planetary surface, which furnishes the spectral source, and the gaseous absorber. This provides a "thermal attenuation factor" that approaches zero as these temperatures approach each other. Thus a gas near the ground could not be observed at all. Furthermore, the feature at 1305 cm^{-1} is weaker than the band at 3020 cm^{-1} (under our resolution) and the path length is less than half as great (because the solar light at 3020 cm^{-1} passes through the atmosphere twice).

14. We acknowledge generous support by NASA for the development of the 1969 Mariner infrared spectrometers, as well as the endurance of the JPL Project staff and their courage as they turned on our instrument when it was thought to be the cause of the flight anomalies of the Mariner 7 spacecraft and a threat to the mission. We thank the personnel of the Chemistry Department Machine Shop and the staff of the Space Sciences Laboratory Electronic Shop for their enormous contribution to the fabrication of this instrument. We express our indebtedness to the Berkeley Group, led by P. B. Forney, J. L. Hughes, D. A. Waston, R. H. Weitzmann, and M. A. Carlson, for their dedicated efforts throughout the project.

29 September 1969 ■

Glacier Oxygen-18 Content and Pleistocene Ocean Temperatures

Abstract. *The mean oxygen-18 content of continental ice sheets during the last glacial maximum is estimated to $\delta O^{18} = -30$ per mille or less, and the consequent change in the isotopic composition of the oceans at that time to 1.2 per mille or more. This means that at least 70 percent of the oxygen-18 variations found in shells of planktonic foraminifera from deep-sea cores between times of glacial maximums and minimums are due to isotopic changes in ocean water, and at most 30 percent to changes in ocean surface temperature. Hence, Emiliani's "paleotemperature" curve rather depicts the amount of ice on the continents in excess of that present today. In this sense it may be renamed a "paleoglaciation" curve.*

The O^{18}/O^{16} ratio in shells of pelagic and benthonic foraminifera depends on the temperature of formation and on the isotopic composition of the ambient water (*1*). The empirical relation between these quantities was given by Epstein *et al.* (*2*) as

$$T = 16.5 - 4.3\,(\delta_f - \delta_{oc}) + 0.14\,(\delta_f - \delta_{oc})^2 \quad (1)$$

where δ_f and δ_{oc} are the relative per mille deviations of the O^{18}/O^{16} ratios from the SMOW (standard mean ocean water) standard (*3*) for foraminifera and ocean water, respectively. It is seen that temperatures derived from this equation depend critically on the isotopic composition of ocean water. During the Pleistocene this composition varied in response to the amount and the isotopic composition of excess continental ice.

Emiliani (*4–6*) has measured δ values of planktonic foraminifera from several deep-sea cores raised from the Caribbean and the equatorial Atlantic and has deduced a well-known generalized paleotemperature curve for Atlantic equatorial surface water, covering about 425,000 years. Paleotemperatures were calculated from Eq. 1 by assuming a mean isotopic composition of Pleistocene glaciers $\delta_g = -15$ per mille and a maximum variation in $\delta_{oc} = +0.5$ per mille. In Emiliani's measurements peak-to-valley variations of δ_f between glacial maximums and minimums averaged 1.65 per mille. With the above-mentioned assumptions this corresponds to a temperature variation of 5° to 6°C in Atlantic equatorial surface water.

This temperature variation has been questioned by a number of authors from various points of view (*7*). In the present report the mean isotopic composition δ_g of continental glaciers during the last glacial maximum, and hence the variation in δ_{oc}, is estimated on the basis of recent isotopic measurements on precipitation in presently and previously glaciated areas, on present ice caps, and even on ice actually formed during the last glaciation and still preserved in Greenland. These data have greatly improved the reliability of estimates of the isotopic composition of the large Pleistocene ice sheets at their maximum extension.

The disagreement between previous estimates of the isotopic composition of Pleistocene glaciers apparently is due not only to the scarcity of data, but also to a number of misunderstandings relating to ice-sheet dynamics. It may therefore be mentioned that a glacier is in a state of steady mass exchange, with accumulation of snow above the firn line and melting of snow and ice in the ablation zone. According to generally accepted theories of flow pattern in ice sheets (*8*), ice formed in the accumulation zone follows tracks that go deeper in the ice sheet (that is closer to the bottom) the closer to the ice divide the site of formation is (*8*). Ice formed in the periphery of the accumulation zone follows relatively superficial tracks and is melted in the ablation zone within a comparatively short time. The bulk of ice in an ice sheet, therefore, originates from precipitation far inland on the glacier (*9*). This also applies to an expanding glacier.

In agreement with this flow pattern it has been found that a considerable part of the ice in the ablation zone has a much lower δ value than precipitation in the source area of the upper layers. Isotope measurements on a core drilled to the rock bed in Antarctica, only 700 m from the coast at Adelie Land, and 86 m above sea level, showed this relation (*10*). The mean isotopic composition of the upper 31 m of the core (total length 98 m) corresponded to a $\delta = -20$ per mille, similar to the mean δ of precipitation in the area, whereas that of the lower 45 m corresponded to a $\delta = -44$

per mille. The mean value of the whole core was $\delta = -33$ per mille. A similar decrease of the δ values with increasing depth was found in a core drilled through the ice shelf at Little America (*11*), and ice in bottom layers at the very edge of the Greenland ice sheet showed δ's of the same order as that of precipitation in central Greenland today (*12*). This demonstrates that the use of δ values of surface layers as representative of the material below can yield only an upper limit for the true mean δ of an ice sheet.

The O^{18} concentration in atmospheric precipitation is mainly determined by its temperature of formation, particularly at high latitudes (*13*). Decreasing temperature of formation leads to decreasing δ values of rain and snow. An empirical equation relating the mean δ of precipitation with mean temperature t_a(°C) of surface air in northern latitudes has been given (*14*) as

$$\delta_p = 0.7\ t_a - 13.6 \text{ per mille} \qquad (2)$$

The equation is based on measurements on samples collected over a wide temperature range including North Atlantic coast stations and Greenland inland ice localities. It indicates a change of δ with temperature of 0.7 per mille per degree centigrade (*15*).

The distribution of surface temperatures and δ values on the Greenland ice cap is simple and well known. The δ values are linearly related to mean air temperature according to Eq. 2. The average δ of surface ice is -30 per mille (*13*, and Fig. 1). This is an upper limit for the mean isotopic

Table 1. Approximate distribution of continental excess ice during the last glacial maximum.

Excess glacial area (*18*) (10^6 km^2)	Mean thickness (water equiv., km)	Excess ice (10^6 km^3 water equiv.)
	North America	
14.8	2	30
	Scandinavia	
4.6	2	9
	Siberia	
3.1	1.5	4
	Central Asia	
1.4	1.5	2
	Antarctica } *Greenland* }	2
	Total	47

composition of the Greenland ice cap today. The isotopic composition of a 1390-m-long core drilled through the inland ice to the bed rock at Camp Century has recently been measured (*16*). In this core, ice from the past 100,000 years has been preserved. The upper part of the core had a $\delta = -29$ per mille. In layers from 10,000 to 70,000 years before the present, that is, from the last ice age, δ values were considerably lower than today. During the last glacial maximum, 20,000 to 25,000 years ago, average δ was below -41 per mille, that is, 12 per mille less than at present. This value must be representative for large parts of the northernmost segment of the North American and Greenlandic ice sheets at that time.

As to Antarctica, a number of measurements suggest that Eq. 2 is valid here too, except for the upper layers in the marginal zone (*14*). Considering the fairly well-known temperature distribution, these measurements suggest a mean surface δ for Antarctica lower than -40 per mille. Recent deuterium measurements from Adelie Land and some 1500 km inland, quoted above, suggest a mean surface δ of approximately -37 per mille for Antarctica (*10*). The three-dimensional average for Antarctica today must be lower than these values.

At the time of the last glacial maximum, approximately 20,000 years ago, sea level was lowered 130 m or more (*17*). At the climatic optimum, 5000 years ago, sea level was within a few meters of the present level. This range corresponds to an excess of continental ice during the last glacial maximum of 47×10^6 km^3 water equivalent. Assuming a maximum extension of ice sheets during the last glaciation as given by Flint (*18*), the distribution of excess ice will have been approximately as listed in Table 1.

Figure 1 shows measurements of the mean annual isotopic composition of precipitation in the North Atlantic sector. Stations were taken mainly from the I.A.E.A.–W.M.O. precipitation survey (*14*). From these measurements approximate iso-δ-lines have been drawn. In North America no survey stations are present north of the line for $\delta = -20$ per mille, but there is no doubt that δ values decrease further toward higher latitudes. A dashed line is entered on Fig. 1 to indicate the maximum extension, according to Flint (*19*), of the continental ice and pack ice during the last glacial maximum.

According to Flint (*20*), the ice divide for the Laurentide ice sheet (and presumably also for the Cordilleran ice sheet) ran very close to the present isoline for $\delta = -20$ per mille. With the Laurentide ice divide running mainly east-west and δ values decreasing north of the ice divide, the three-dimensional mean δ value of the whole sheet will be close to that of precipitation at the ice divide itself (see also 9). In view of this we consider it a conservative estimate if we assume that the three-dimensional mean δ value of the North American ice sheet at its maximum extension was that of precipitation in areas halfway between the present -20 and -15 δ lines, that is, close to the present line for $\delta = -17$ per mille.

During the last glacial maximum the surface of the ice sheet in these areas was about 2 km (*21*) above present

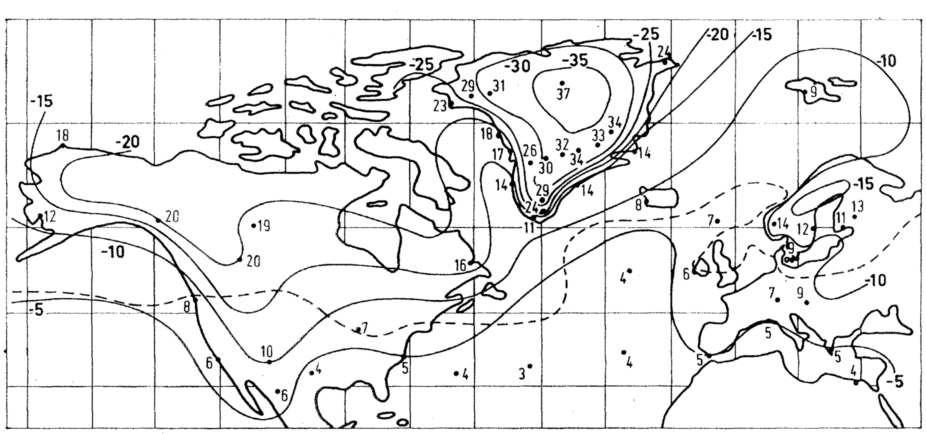

Fig. 1. Numerical mean δ values of annual precipitation, covering periods of 2 to 10 years, and approximate iso-δ-lines; δ values are relative per mille deviations of O^{18}/O^{16} ratios from Standard Mean Ocean Water (*3*). The influence of the present Gulf Stream is seen very clearly in the course of the isoline for $\delta = -10$ per mille. The dashed line indicates the maximum extension of continental ice and permanent pack ice during the last glacial maximum according to Flint (*19*).

ground level. With a normal lapse rate in the atmosphere of 0.65°C/100 m, the mean surface temperature was at least 13°C lower than the present annual mean at ground level, corresponding to a lowering of 9 per mille for mean δ values of precipitation in these areas. A similar, or somewhat larger, lowering of δ values is seen today between precipitation at coast stations in Greenland and inland stations elevated 2 km (see Fig. 1). During the Ice Age we further have to consider a general cooling of the glaciated areas. In the Camp Century core this effect was measured to −12 per mille for the period 20,000 to 25,000 B.P. (*16*). A third of this, that is, a lowering of 4 per mille, as an average for the North American ice sheet would seem a very conservative estimate, especially if the three-dimensional flow pattern is considered. In this way we arrive at a δ = −30 per mille as an upper limit for the mean isotopic composition of the North American ice sheet during the last glacial maximum.

The Scandinavian stations shown in Fig. 1 have higher δ values than the North American at the same latitude. This is due to proximity to open sea and to the Gulf Stream. With a lowering of the ocean level of more than 100 m, both the North Sea and the Baltic would be drained. If, at the same time, the Gulf Stream was essentially cut off from the Norwegian Sea and displaced 15° of latitude to the south in the eastern Atlantic (*22*), these stations would become as continental as those in Canada at the same latitude. Consequently, there is no reason to believe that the mean δ value for the Scandinavian ice sheet deviated significantly from that of the North American.

No information is available on present-day δ values of precipitation in localities representative for the North Siberian and the Central Asian ice sheets. However, today northern Siberia is the coldest part of the Northern Hemisphere, and the Central Asian ice sheets were located in mountain areas well above 1000 m above sea level. Even with a somewhat smaller thickness of these ice sheets (1.5 km), the mean δ values of surface precipitation on these glaciers can hardly have been higher than those of the North American and Scandinavian ice sheets. With the comparatively small amount of ice stored in these glaciers their composition will not have changed the average δ value of all Pleistocene excess ice.

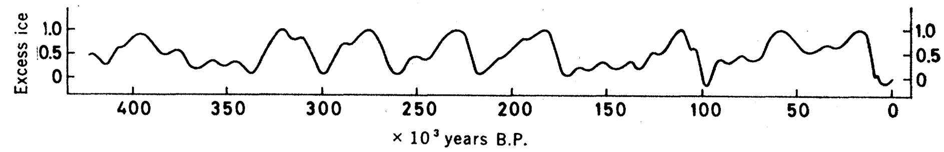

Fig. 2. Emiliani's generalized paleotemperature curve (*5, 6*) reversed and converted into a tentative paleoglaciation curve. The unit of the ordinate is the amount of excess ice during the last glacial maximum. The curve tends to underestimate the amount of excess ice in the initial phase of each glaciation, when the ice sheets have not reached their full height and thus the low oxygen-18 concentrations typical for ice sheets at their maximum extension. During the last 400,000 years interglacial phases were the unusual state and various degrees of glaciation the usual state.

The δ values of the excess ice on Antarctica cannot have been higher than the mean value for coastal ice on Antarctica today. At Adelie Land this is −33 per mille (*10*). Considering the flow pattern, δ values of the Antarctic excess ice would rather be expected to be equal to, or less than, the mean δ for the whole Antarctic ice cap today. Excess ice on Greenland must also have had δ values below −30 per mille.

We thus conclude that δ = −30 per mille is an upper limit for the mean isotopic composition of continental excess ice during the last glacial maximum. In arriving at this value conservative estimates were used throughout. The true mean δ of the continental excess ice was probably some per mille lower than −30 per mille. At any rate, a mean δ value of −15 per mille for the Pleistocene glaciers, as assumed by Emiliani (*4–6*), is impossible. This value is higher than the isotopic composition of precipitation in most parts of Canada today, when there is no ice age and no ice sheet with a surface 2 km above present ground level.

In considering the average δ value of Pleistocene glaciers it should also be remembered that the bulk of ice in the entire ice sheets on Greenland and Antarctica, with a water equivalent of 29×10^6 km³, was exchanged with ice of a much lower mean δ value than that of today. At Camp Century the average lowering, 20,000 years ago, of the whole column from the surface to bed rock was close to 10 per mille (*16*). Half of this, or 5 per mille, is considered a conservative estimate of the average change of the ice sheets on Greenland and Antarctica.

The present amount of ocean water is 1370×10^6 km³ with a mean δ = 0 per mille. If the continental excess ice (water equivalent of 47×10^6 km³) had a mean δ = −30 per mille or less, and the average lowering of the mean δ of the Antarctica and Greenland ice sheets (water equivalent of 29×10^6 km³) was 5 per mille relative to the present, the change in the mean δ value of ocean water during the last glacial maximum was at least

$$\Delta\delta_{oc} = \frac{(47 \times 30) + (29 \times 5)}{1370 - 47} = +1.2 \text{ per mille}$$

The average peak-to-valley variation of δ values of planktonic foraminifera between glacial maximums and minimums was 1.65 per mille (*5*). If δ of ocean water changed by at least 1.2 per mille, less than 27 percent of the measured change in shells is caused by variations in the temperature of Atlantic equatorial surface water. For the last glaciation alone the peak-to-valley variation in nine Caribbean and equatorial Atlantic cores averaged 1.75 per mille (*5*). Relative to this figure, at most 31 percent of the measured effect can be ascribed to temperature changes. We therefore conclude that only 30 percent or less of the δ variations found in shells of planktonic foraminifera are caused by changes in ocean temperatures, while 70 percent or more are due to changes in the isotopic composition of ocean water. This corresponds to a temperature variation between glacial maximums and minimums of Atlantic equatorial surface water of at most 2°C.

An upper limit for the isotopic composition of continental excess ice of δ = −30 per mille is close to a value of −33 per mille estimated by Olausson (*7*). Although this estimate was better founded than the somewhat conjectural estimate of δ = −15 per mille, it was criticized by Emiliani (*5*) on account of the arbitrary nature of the air temperatures assumed for the Pleistocene. In view of this criticism we have based our estimate entirely on measured δ values of precipitation and ice, combined with considerations of the flow pattern in ice sheets.

If the mean δ of ocean water increased by at least 1.2 per mille from interglacial to glacial maximum, and the oceans are well mixed within a time of a few thousand years, this increase should be found not only in surface water foraminifera, but also in shells of species living on the ocean bottom. In two cores from the eastern Pacific, Emiliani (*4, 5*) found an average isotopic change of only 0.5 per mille in shells of benthonic species and claimed this to be an upper limit for the isotopic change of ocean water. However, the correlation between measured levels and glacial maximums and minimums was not certain. Shackleton (*7*) has later measured the isotopic variation in benthonic species in another core from the eastern Pacific and in one from the Caribbean. In these cases the isotopic composition of benthonic species varied with about the same amount as in planktonic species. It may also be noted that Emiliani's own measurements on shells of benthonic species in cores from the equatorial Atlantic showed an average increase in δ from interglacial to glacial maximums of 1.23 per mille (*4*). This is identical with our estimate of the increase in the isotopic composition of ocean water, and thus the expected change in shells of benthonic species if no appreciable change in bottom water temperature occurred.

This reinterpretation of the signification of glacial-interglacial isotopic changes in shells of foraminifera does not detract from the immense importance of Emiliani's measurements on deep-sea cores. On the contrary, if the main part of the δ variations in foraminifera is due to changes in the isotopic composition of ocean water, caused by the building-up of continental ice sheets, we have a rigid correlation between marine isotopic variations and terrestrial glacial events. This was already pointed out by Shackleton (*7*). Emiliani's generalized "paleotemperature" curve can be reread as a curve depicting the amount of continental ice in excess of (or in deficit of) the amount present today. In this sense it may be named a "paleoglaciation" curve (Fig. 2). It should be noted that in Emiliani's generalized curve the glacial-interglacial variations have been reduced by 30 percent, as we would also do if the curve should be read as a paleoglaciation curve. In Emiliani's interpretation this was the correction to be applied due to the isotopic changes of ocean water. In our interpretation it is the maximum correction to be used due to temperature changes of Atlantic surface water.

If this interpretation is correct the paleoglaciation curve shows that within the last 425,000 years we have had seven or eight major, independent glaciations of the continents, each of a magnitude similar to the maximum of the last glaciation, and each with a time spacing of the order of 40,000 to 50,000 years. With a total length of the Pleistocene of 2 to 3 million years we should expect perhaps 40 major and independent glaciations during this period, rather than the four or six subdivided glaciations usually recognized.

W. DANSGAARD
H. C. Ørsted Institute, University of Copenhagen, Copenhagen, Denmark
HENRIK TAUBER
Carbon-14 Dating Laboratory, National Museum, Copenhagen

References and Notes

1. H. C. Urey, *J. Chem. Soc.* **1947**, 562 (1947).
S. Epstein, R. Buchsbaum, H. Lowenstam, H. C. Urey, *Geol. Soc. Amer. Bull.* **64**, 1315 (1953). In the original equation δ values meant relative deviations from the Chicago Peedee belemnite standard; δ values expressed in the SMOW scale differ approximately 0.1 per mille from δ relative to the PDB standard. The consequent change of Eq. 1 is insignificant for our considerations.
3. H. Craig, *Science* **133**, 1833 (1961).
4. C. Emiliani, *J. Geol.* **63**, 538 (1955).
5. ———, *ibid.* **74**, 109 (1966).
6. ———, *Science* **154**, 851 (1966).
D. B. Ericson, M. Ewing, G. Wolin, *ibid.* **146**, 723 (1964); E. Olausson, *Progr. Oceanogr.* **3**, 221 (1965); N. Shackleton, *Nature* **215**, 15 (1967).
J. F. Nye, *Proc. Roy. Soc. (London) Ser. A* **207**, 554 (1951); *ibid.* **239**, 113 (1957); *J. Glac.* **3**, 493 (1959); R. Haefeli, *ibid.* **3**, 1133 (1961).
9. According to the flow pattern calculated by Haefeli (*8*), approximately 80 percent of the ice in a large ice sheet with a symmetrical ice divide originates from precipitation in the middle 33 percent of the ice sheet.
10. L. Merlivat, C. Lorius, M. Majzoub, G. Nief, E. Roth, in *Isotopes in Hydrology* (International Atomic Energy Agency, Vienna, 1967), pp. 671–681.
11. S. J. Johnsen, W. Dansgaard, C. C. Langway: δ values measured on an ice core through the ice shelf varied from $\delta = -23$ per mille at the surface to $\delta = -35$ per mille near the bottom.
12. H. Oeschger, H. B. Clausen, B. Stauffer, data obtained on ice taken at Camp 3 (69°50′N) near Eqip Sermia, West Greenland, on E.G.I.G. expedition, 1967.
13. W. Dansgaard, *Medd. Grønland* **165**, No. 2 (1961).
14. ———, *Tellus* **16**, 436 (1964).
15. Equation 2 also appears to be valid for temporal variations of δ and temperature at a given locality due to climatic changes. This is suggested by a direct comparison, covering nearly 100 years, between δ of snow strata at Camp Century on the Greenland ice sheet and measured mean air temperatures at Upernavik on the West Greenland coast.
16. W. Dansgaard, S. J. Johnsen, J. Møller, C. C. Langway, Jr., *Science* **166**, 377 (1969).
17. F. P. Shepard, *Submarine Geology* (Harper & Row, New York, 1963), pp. 264–266; J. D. Milliman and K. O. Emery, *Science* **162**, 1121 (1968).
18. R. F. Flint, *Glacial and Pleistocene Geology* (Wiley, New York, 1957), p. 53.
19. ———, *ibid.*, plate 3.
20. ———, *ibid.*, pp. 154 and 315.
21. If the water equivalent thickness of the ice sheet was 2 km, the actual thickness was 10 percent higher. At the same time, sea level was approximately 100 m lower than at present. The additional surface elevation thus amounts to 2.3 km. Part of this was counterbalanced by isostatic subsidence of the underlying bed rock. In case of equilibrium this would probably have amounted to nearly 600 m. For a rapidly expanding glacier, equilibrium will not be attained, and the maximum postglacial isostatic rebound in central Canada is of the order of only 200 to 300 m (R. F. Flint, *ibid.*, pp. 247–255). An additional average surface elevation of 2 km relative to sea level thus seems reasonable. Close to the ice divide, surface elevation must have been several hundred meters higher than the average elevation. In order to be conservative we have disregarded the effect of this.
22. A. McIntyre, *Science* **158**, 1315 (1967).
23. Measurements of precipitation from stations at the I.A.E.A.–W.M.O. precipitation survey were made under contract No. 325/TC with the International Atomic Energy Agency, Vienna.

24 March 1969

Dredged Trachyte and Basalt from Kodiak Seamount and the Adjacent Aleutian Trench, Alaska

Abstract. *Blocky fragments of aegirine-augite trachyte (with accompanying ice-rafted gravels) were recovered from the upper slopes of Kodiak Seamount in several dredge hauls. An alkali basalt pillow segment was also dredged from a moatlike depression, at a depth of 5000 meters, near the west base of the seamount. These retrievals confirm the volcanic origin of Kodiak Seamount and further support the view of Engel, Engel, and Havens that the higher elevations of seamounts are composed of alkali basalts or related variants.*

Kodiak Seamount is located at latitude 56°52′N, longitude 146°15′W, 104 miles (167 km) southeast of Kodiak Island, in the Gulf of Alaska. It is of particular interest since it lies in the Aleutian Trench and is the northernmost seamount in the Pratt-Welker chain of seamounts and guyots. Earlier surveys had indicated that Kodiak Seamount had a flat top and was therefore a guyot or tablemount (*1*). The summit depth of Kodiak Seamount (2017 m) was considerably greater than that of adjacent guyots located beyond the south margin of the Aleutian Trench. These relations suggested to Menard and Dietz (*1*) that the formation of Kodiak Seamount and the truncation of its summit may have predated the subsidence of the Aleutian Trench.