ratios (sediment yield as a proportion of gross erosion) range from about 3 to 10 percent, with a weighted average of 4.7 percent (Fig. 1), a figure consistent with the results of a previous study (*6*).

The eroded material not transported from the basins has been deposited as colluvium and alluvium, with the alluvium as deep as 6 m in second to fifth order (small to medium-sized) streams (*7*). The formation of such deposits indicates a definite lack of the steady state. At present, as a result of declining agricultural land use and generally implemented soil conservation practices, streams have regained their transport ability, and the modern alluvial deposits in some are being dissected. That is, oversteepened gradients are readjusting and alluvium is migrating. However, this movement is limited by many large reservoirs, which act as efficient sediment traps. Thus, the sediment will not be exported within the foreseeable future (*8*).

By inference and according to the available evidence, many streams of the humid United States are analogous to those of the Piedmont and thus have not been in a steady state since European settlement. This means that "denudation rates" determined from downstream sediment yields are not synonymous with rates of erosion or with lowering of the land, as has frequently been assumed (*3, 9*).

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Weyl's Theory of Glaciation Supported by Isotopic Study of Norwegian Core K 11

Abstract. *Oxygen-18 analyses of pelagic and benthic foraminifera from core K 11 indicate that during the last glaciation Norwegian Sea bottom waters were warmer than in modern times and had the same physical parameters (temperature, oxygen isotope ratio, and salinity) as the North Atlantic deep water. This result indicates that the glacial Norwegian Sea was not a sink for dense surface water, as it is now, and that during glacial times North Atlantic deep water invaded the deep Norwegian basin.*

Oxygen isotopic analyses provide information about past ice-cap volumes and sea temperatures. The isotopic records of cores from the Atlantic (*1*), Pacific (*2, 3*), and Indian (*4*) oceans are highly correlated and have provided information about the worldwide evolution of oceanic climate. Numerous attempts have been made to correlate Pleistocene climatic cycles with variations in insolation calculated by Milankovitch (*5*), but thermodynamic calculations indicate that such variations may have been too small to trigger glacial climates (*6*). A possible mechanism for glaciation has been suggested by Weyl (*7*), who showed that minor long-term changes in the average behavior of the atmosphere might have triggered a decrease of the surface salinity of the North Atlantic, allowing sea ice to extend as far south as it does in the Pacific, that is, to almost 60°N. Such a southward extension of sea ice would have cooled northern Europe to temperatures similar to those of Alaska at present. The Norwegian Sea provides a striking test of Weyl's hypothesis. That sea is today a typical mediterranean basin (*8*) in which an inflow of light surface water is required to balance the outflow of deep, dense water formed by the action of the atmosphere at the sea surface within the basin. In a glacial period, according to Weyl's theory, the formation and sinking of dense water in the Norwegian Sea would have been suppressed. Newell (*9*) has related climatic fluctuations to coupling of the oceans and the atmosphere and suggested that the glacial periods resulted from variations in the poleward energy flux in the ocean-atmosphere system. Newell's theory also implies that the glacial Norwegian Sea was not a sink for dense, cold surface water. Ice ages are thus considered as periods of warming of the ocean.

During the cruise NESTLANTE I of the French R.V. *Jean Charcot* in the Norwegian Sea (August 1970), piston core K 11 was collected at 71°47′N, 01°36′E, from a water depth of 2900 m.

We analyzed $\delta^{18}O$ (*10*) for the pelagic species *Globigerina pachyderma* and the benthic species *Pyrgo depressa, Pyrgo oblonga*, and *Planulina wuellerstorfii* from core K 11. The experimental method has been described by Shackleton (*11*). A Micromass 602 C mass spectrometer was used. Each sample weight was about 0.4 mg of carbonate. Replicate analyses of the same carbonate give a standard deviation of 0.13 per mil for a single measurement. Pelagic samples were analyzed in triplicate, so that the precision of the mean is $\pm$ 0.07 per mil. Benthic specimens were much scarcer and generally were analyzed only once or twice.

The two species of *Pyrgo* give results indistinguishable from each other. *Planulina* has been already reported as isotopically lighter (*2, 12*). For core K 11, the mean difference in $\delta^{18}O$ values between *Planulina* and *Pyrgo*, deduced from a comparison of 25 levels in the core, was 1.15 $\pm$ 0.03 per mil. In obtaining mean values for each level in the core, we added 1.15 per mil to the *Planulina* values to correct for this effect. The $\delta^{18}O$ values for *Globigerina pachyderma* and the benthic species are shown in Fig. 1A. It was possible to correlate the *G. pachyderma* curve with stages 1 to 6 defined by Emiliani (*13*), which indicates that the sediment was not reworked. The benthic curve shows only minor variations all along the core and significant scattering (*14*). To suppress the scatter in the benthic curve, we smoothed the curve by calculating the running mean, using three adjacent samples for each level (*15*).

Figure 1B compares benthic curves obtained from Norwegian core K 11 and Pacific core V 28-238 (*2, 3*). From stage 1 to stage 2, benthic foraminifera become richer in ^{18}O in both cores, but the enrichment in the Norwegian core is much smaller (0.4 per mil) than that in the Pacific core (1.5 per mil). At the beginning of stage 3, the variations in both curves are well correlated and probably reflect minor variations in seawater ^{18}O. The small amplitude of the variations in the Norwegian benthic curve and the difference between the Norwegian and Pacific benthic curves can be explained in terms of bottom water temperatures.

The measured present temperature of the bottom water just above the site of core K 11 is -0.9°C; the water could not have been much colder, even in a glacial period. Hence, the temperature of the bottom water of the Norwegian Sea must have been either the same as at present, or warmer.

We first assume that the Norwegian Sea bottom water temperature has remained constant. Then, since the $^{18}O/^{16}O$ ratio of the benthic foraminifera depends on the temperature and on the $^{18}O/^{16}O$ ratio of

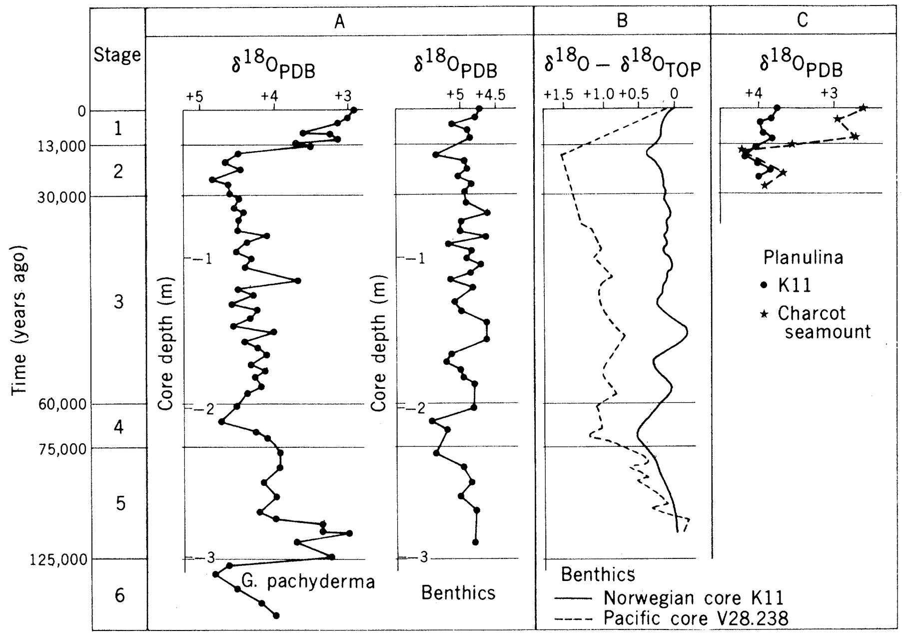

Fig. 1. (A) Variation of $\delta^{18}O$ (*10*) of *Globigerina pachyderma* and benthic foraminifera versus depth in core K 11 (71°47′N, 01°36′E). (B) Variation of $\delta^{18}O$ of benthic foraminifera from core K 11 after smoothing and from Pacific core V 28-238. To compare the past variations without taking into account the present $\delta^{18}O$ value in each core, we subtract at each level the $\delta^{18}O$ value of the benthic foraminifera of the top of each core. (C) Variation of $\delta^{18}O$ of *Planulina wuellerstorfii* in Norwegian core K 11 and an Atlantic core from the Charcot Seamount (45°19′N, 10°31′W) during stages 1 and 2. Stages (*13*) and time scale (*2, 3*) are shown at the left.

seawater, the benthic curve of core K 11 (Fig. 1B) must represent variations of the $^{18}O/^{16}O$ ratio of Norwegian Sea water in the past. Because the $^{18}O/^{16}O$ ratio of oceanic deep water is very uniform (*16*), this curve is valid for all the oceans and indicates that the oceanic water was isotopically heavier by about 0.4 per mil during stage 2. Figure 1B shows that the benthic foraminifera of the Pacific were isotopically heavier by about 1.5 per mil in stage 2. Thus, the difference, 1.1 per mil, indicates that the Pacific deep water was colder than today by about 4.5°C during stage 2. This is impossible, because the present temperature of these waters is about 1°C, and a cooling of 2.5°C at most would be possible.

We are then obliged to conclude that, during stage 2, the Norwegian deep water was warmer than it is today. This needs an oceanographic explanation. Figure 1C compares the $\delta^{18}O$ variations of the same species, *Planulina wuellerstorfii*, from core K 11 and from a core taken at the top of the Charcot Seamount (45°19′N, 10°31′W, at a depth of 2665 m) (*12*). At the top of the cores the difference in $\delta^{18}O$ values indicates that, at present, the North Atlantic deep water is about 4°C warmer than the Norwegian deep water, in good agreement with oceanographic data. In stage 2, both waters have the same physical parameters. This strongly suggests that in glacial stage 2 North Atlantic deep water invaded the Norwegian basin. The comparison in Fig. 1B between the Pacific benthic curve, which reflects mainly the $\delta^{18}O$ variations of seawater, and the benthic curve from core K 11 shows that this warming of Norwegian Sea deep water was taking place by the end of stage 5. Thus, the glacial circulation pattern we have described for stage 2 began during stage 5 and also occurred during stages 4 and 3.

From these observations, we infer the following oceanographic changes during the latest glacial episode.

1) Deep water did not form at the surface and sink in the Norwegian Sea. This result is in good agreement with the observations of Streeter (*17*) and Schnitker (*18*), who interpreted changes in benthic assemblages in terms of changes of water masses.

2) The deep currents in the Denmark Strait and the Faeroe Channel were opposite to their present directions (*19*). The water budget also implies a reversal of the surface currents flowing from the Norwegian Sea to the Atlantic Ocean. This circulation pattern is in good agreement with micropaleontological and sedimentological results of Kellogg (*20*), who concluded that the Norwegian Sea was covered with ice, probably during the whole year, during stages 2, 3, and 4 and the end of stage 5.

It would be of primary importance to evaluate the glacial temperatures of the Norwegian and North Atlantic deep waters, since both Weyl and Newell suggested that glacial deep water was warmer than present deep water during stage 2. However, such an evaluation requires a model giving the ^{18}O variations of seawater in the past. At present, we do not have a model precise enough to enable us to determine whether the temperature of North Atlantic deep water was warmer, colder, or the same as today.

In any case, our data confirm one aspect of the glacial circulation regime proposed by Weyl and by Newell, and thus support the theory that changes in atmospheric circulation and ocean-atmosphere interactions triggered the ice ages. As suggested by Weyl, the Milankovitch mechanism may have modulated atmospheric circulation; this would explain the fit between insolation and past climatic variations.

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10. We calculated $\delta^{18}O$ using the relation

$$\delta^{18}O = \left[\frac{(^{18}O/^{16}O)\ \text{sample}}{(^{18}O/^{16}O)\ \text{PDB}} - 1\right] \times 10^3$$

where PDB (Pee Dee belemnite) is the Chicago PDB-1 standard. In our measurements the isotopic compositions are referred to the PDB using the calibration given by N. J. Shackleton (personal communication) for the standard V MARBLE: $\delta^{18}O$(V MARBLE) − $\delta^{18}O$ (PDB) = −1.09.
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15. For the levels in which we have both *Pyrgo* and *Planulina*, we could calculate that the standard deviation of the difference between them is 0.23 per mil, but the standard deviation of the difference between the running mean for *Pyrgo* and *Planulina* becomes 0.10 per mil, confirming that *Pyrgo* is responsible for the scattering.
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