13. LATE PLIOCENE – PLEISTOCENE GLACIATION

W. A. Berggren, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

The discussion in this chapter is broken down into two parts: the first deals with glaciation in the North Atlantic as revealed in the data obtained on Leg 12; in the second part an attempt is made to provide a chronologic framework of Late Pliocene-Pleistocene glaciation and to correlate glacial/interglacial sequences as recorded in land and deep-sea sediments.

GLACIATION IN THE NORTH ATLANTIC

One of the most significant aspects of Leg 12 was the various results which were obtained regarding glaciation in the North Atlantic. Glacial sediments were encountered at all sites in the North Atlantic with the exception of Site 117 (for the purpose of this discussion the North Atlantic encompasses Sites 111 through 117; Sites 118 and 119 are referred to as the Bay of Biscay). The glacial sediments consist predominantly of quartz sand and silt, clay and various amounts of igneous and metamorphic rock and mineral fragments.

The natural gamma activity and lithology may serve as useful methods of distinguishing between pre-glacial and glacial sediments and estimating the approximate intervals and relative intensity of glaciation. The contact between glacial and pre-glacial sediments was cored at three sites: 111 (145 meters), 112 (115 meters) and 116 (71 meters). Above the glacial/preglacial boundary a sharp increase in gamma-ray activity (1500 to 2000±300 counts at Site 111) occurs. At Site 116 the gamma-ray activity increased just above 71 meters at the same level at which an increase in detrital mineral grains is seen. At Site 112 the exact depth cannot be picked by means of natural gamma activity because the boundary seems to lie between two cores (4 and 5). The boundary in this case has been determined paleontologically and lithologically.

From the data available on Leg 12, it seems that the significance of the natural gamma counts is as follows: Generally it may be said that the counts are a measure of the proportion of (non-carbonate + non-silica) in a core. This fraction contains detrital minerals and clays, both of which contribute toward the gamma count. This dual source of counts explains the generally poor correlation between the detrital minerals (excluding the clays) curve (see Figure 1) and the gamma curve (see Figure 2). This poor correlation may also be due to unrepresentative sampling. The increase in gamma-ray activity at the preglacial/glacial boundary can only be due to an increasing proportion of clays and detrital minerals. In absolute terms this means either (i) more detritals and clay deposited per unit time; or (ii) less organic carbonate and silica deposited per unit time; or (iii) both (i) and (ii); or (iv) more detritals, clay and carbonate deposited per unit time (that is, increased sedimentation rate) with the proportional increase in the former exceeding that of the latter; or (v) less detritals, clay and carbonate deposited per unit time (that is, decreased sedimentation rate) with the decrease in the latter exceeding the former. In view of the demonstrable increase in sedimentation rate above the preglacial/glacial boundary at Sites 111, 112 and 116 due to increased amounts of detrital minerals and the fact that glacial periods in high latitudes are characterized by a carbonate minimum (McIntyre *et al.*, in press) it can be seen that the correct explanation for the increase in natural gamma activity in the glacial part of the section is rather complex. Thin bands of carbonate were found at various levels intercalated with detrital-rich clays which indicates interglacial intervals, so that the correct explanation probably lies with (iii) above. In short, then, the increased natural gamma activity is an aid in determining the preglacial/glacial boundary but it is not entirely clear why this increase occurs.

Samples from Cores 116-1 to 11 and from Core 116-1 were treated with hydrochloric acid. The preglacial/glacial boundary can be picked at about 71 meters (116A-8-5, 79 to 80 centimeters; 116-1-3, CC) with the first (upward) influx of detrital minerals. In Core 116-1 this increases to about 45 per cent of the volume of the HCl insoluble, clay-free residue in Section 1 (69 to 71 centimeters). It should be pointed out that the residue volumes are quite small, ranging from about 0.001 to 0.24 cubic centimeters, and averaging about 0.06 cubic-centimeter. The approximate mean residue volume in the glacial sediments of Cores 116A-1-1 through 116A-8-5 is about 0.09 cubic-centimeter and in the preglacial sediments it is about 0.02 cubic-centimeter. These numbers are only estimates based upon the amount of residue occupying a 60 square micropaleontology faunal slide, and the quantities should be considered in relative terms only.

An examination of the data shown in Figure 1 is of interest to our discussion of glaciation in the North Atlantic.

If we place the Pliocene/Pleistocene boundary in Hole 116A at about 60 meters and assign an age of 2 million years to it we obtain an average sedimentation rate of 3 cm/1000 yrs for the Pleistocene. The 30-meter level would be about 1 million years old. Broadly speaking the data suggest two trends: 1) between 71 to 24 meters, or roughly between 3 to 0.8 million years, fluctuating percentages of detrital minerals ranging from 0 to 90 per cent with average peaks at about 60 per cent; 2) above 24 meters a large amount of detrital mineral grains, reaching a peak value of 95 per cent at about 15 meters tapering off to values of 40 per cent at the top of the hole. The significant decrease in detrital mineral and increase in radiolarians and sponge spicules in 116A-3-4, 28 centimeters (about 24 meters) suggests the presence of an interglacial period (?Yarmouthian).

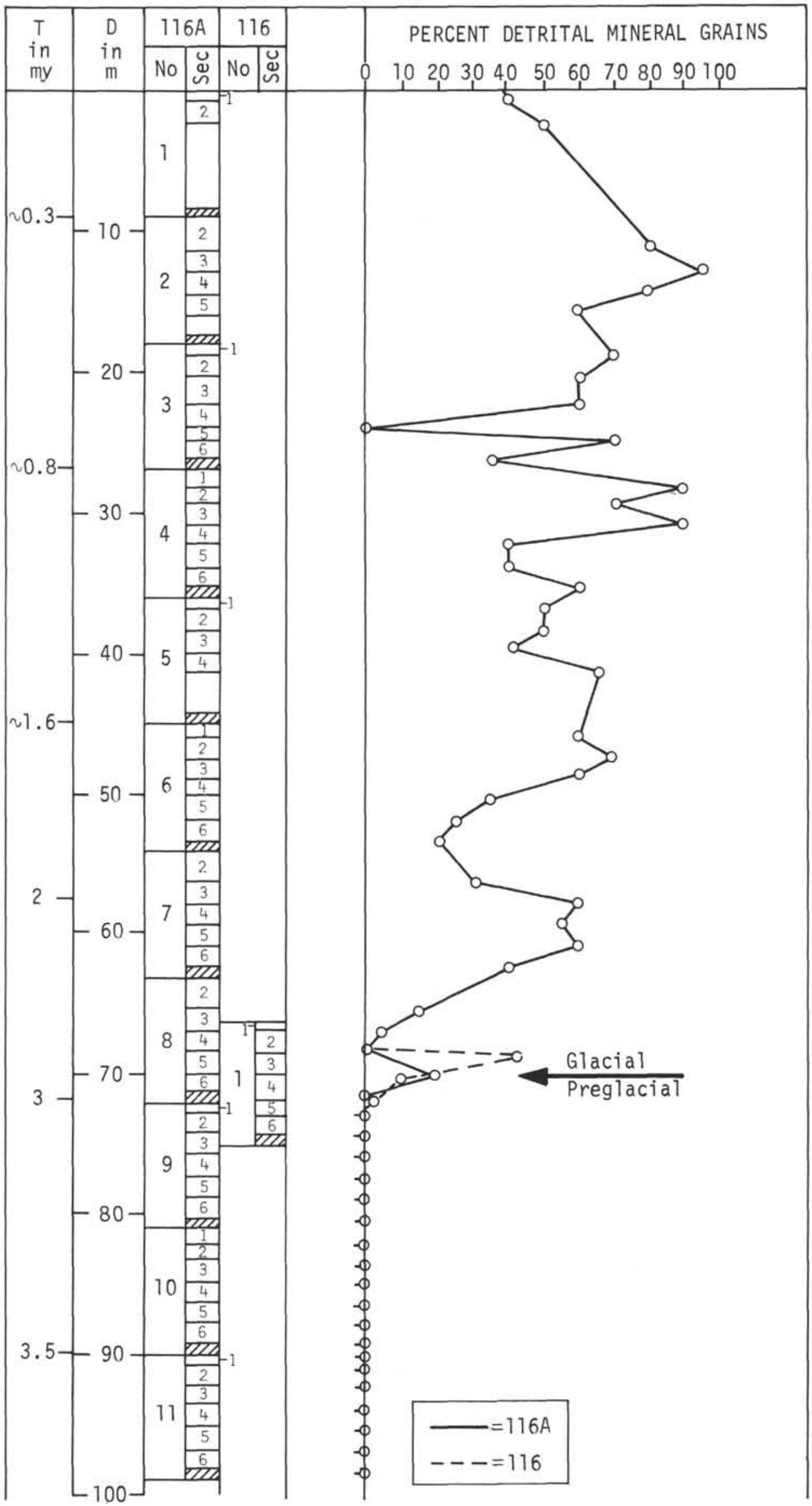

Figure 1. *Plot of percent detrital mineral grains in pre-glacial-glacial Pliocene of Site 116.*

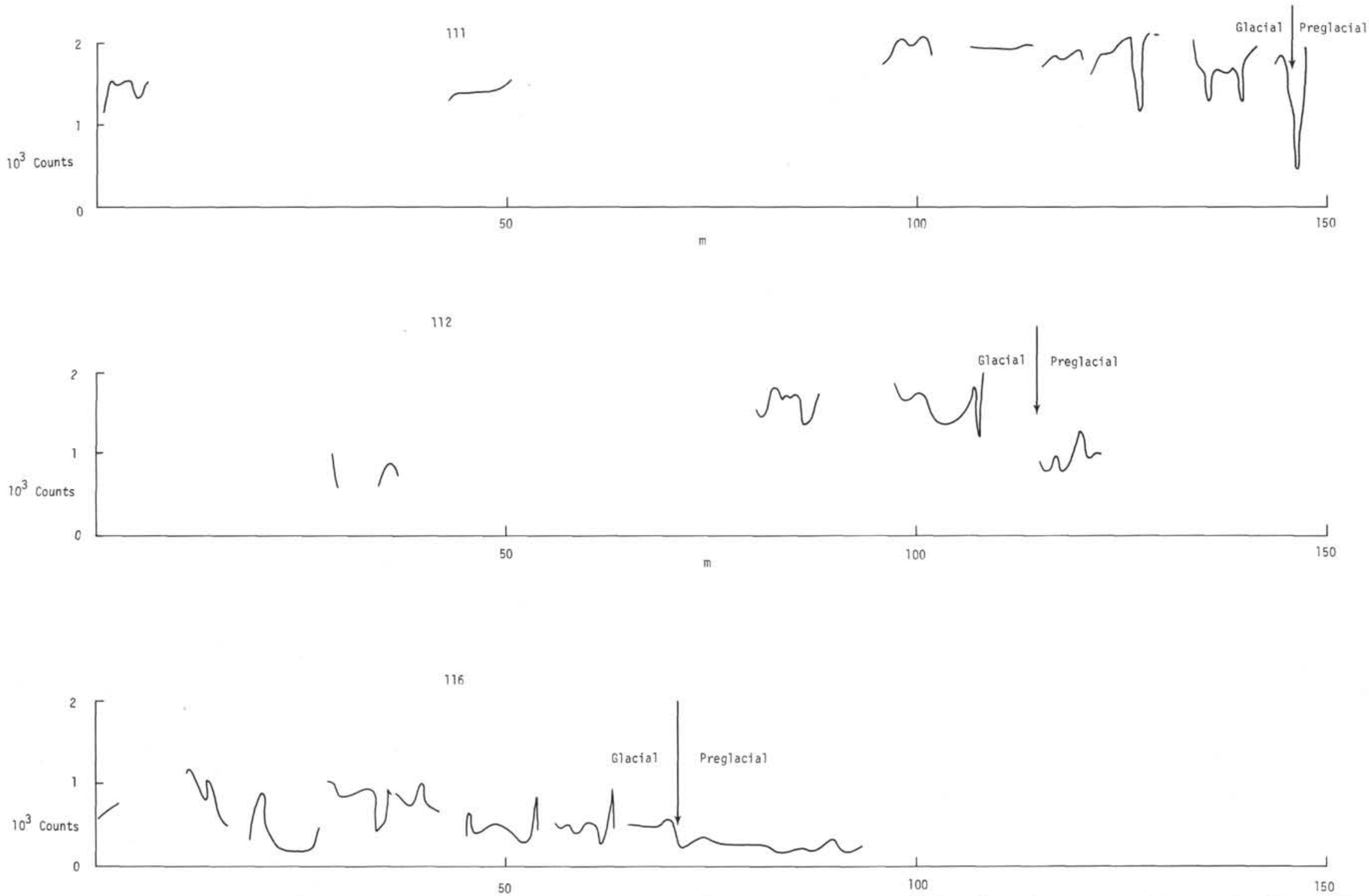

Figure 2. *Interpretation of natural gamma-ray plots in upper parts of Sites 111, 112, and 116. The pre-glacial/glacial boundary is denoted by a marked increase in natural gamma rays and is marked by an → in each case.*

If maximum percentages of mineral grains can be interpreted as indicative of major glacial advances, the data suggest the following interpretation:

Maximum % Mineral Grains	Estimated Age (m.y.)
10-15 m	0.4-0.6
~30 m	1
40-50 m	1.3-1.6
~60 m	2

There is a general tendency for the maximum percentage values of detrital mineral grains to increase upwards (from a maximum of about 60 per cent at about 60 meters to about 95 per cent at about 13 meters). All the values, and the estimates based upon them, should be considered as yielding information of primarily a qualitative rather than quantitative nature.

The most significant information about glaciation which was obtained on Leg 12 concerns the chronology and biostratigraphy of its initiation. Our data indicate that glaciation began in the North Atlantic about 3 million years ago. In terms of biostratigraphy this event occurs approximately at the *Reticulofenestra pseudoumbilica*/*Discoaster surculus* boundary (calcareous nannoplankton). It occurs at a level correlative with the lower part of Zone N21 (although the system of N-zones cannot be systematically applied to North Atlantic biostratigraphy). More precisely it occurs, as was seen at Site 111 in the Labrador Sea, just prior to the extinction of *Globoquadrina altispira, Sphaeroidinellopsis seminula, S. subdehiscens* and *Globorotalia multicamerata* (see Figure 3), and it would appear that the extinction of these forms is causally related to the onset of glaciation.

A branch of the Gulf Stream flowed along the eastern margin of Newfoundland–Labrador during the Cretaceous and Tertiary. This was suddenly stopped by the appearance of icebergs in the Labrador Sea about 3 million years ago. The Gulf Stream was then displaced southwards to its present position essentially south of latitude 45°N. The Labrador Current probably formed at this time as a response to the drastic cooling in this region and the Polar Faunal Realm was established with the development of a cold water fauna (characterized by the appearance of *Globigerina pachyderma* and *Globorotalia inflata*) in the North Atlantic. The development of a Polar Faunal Realm in the Late Pliocene is reflected in the scarcity of discoasters in our North Atlantic cores as well.

A CHRONOLOGY OF LATE PLIOCENE-PLEISTOCENE GLACIATION

The framework of time is the *sine qua non* of understanding geological phenomena, whether it be rates of evolution, sedimentation, rise and fall of sea-level, periods of orogeny, and so forth. This is no less true of glaciation. The evidence of glaciation is manifold and is reflected in both marine and non-marine sediments. Because of the nature of the evidence, however, the interpretation of the record has varied considerably, and this is no less true of attempts to erect a chronology for the glacial record. In this section I shall attempt to erect a basic chronologic framework for the Late Cenozoic glacial record. It is based upon an integration of the data we have obtained in the North Atlantic and the published literature dealing with marine (deep-sea cores) and non-marine sequences.

All too often the assumption is made that the "Glacial Period" is synonymous with the Quaternary. "Date the base of the earliest glaciation and you have dated the base of the Quaternary," or, from the other point of view, "date the base of the Quaternary and you have dated the earliest glaciation." This is bad stratigraphy, to say the least, if one is familiar with the Stratigraphic Code. On the contrary, the base of the Pleistocene has a clearly defined lithologic base and it is upon this base that the concept of a Pliocene/Pleistocene boundary must be founded. This boundary can be recognized by paleontologic means and has been dated by paleomagnetic stratigraphy at about 2 million years (Berggren *et al.*, 1967; Phillips *et al.*, 1968; Glass *et al.*, 1967). Current discussion about the identification of one or more polarity events between 1.6 to 2.0 million years and their terminology are of secondary importance to our discussion. Recent work in our laboratory at Woods Hole suggests that the extinction level of discoasters occurs near the top of the Olduvai (=Gilsa) at about 1.65 my and that the *G. tosaensis–G. truncatulinoides* transition appears to occur near the base of the Olduvai, between 1.8-1.75 my. The extinction of *Globorotalia miocenica* and *G. exilis* appears to occur at about 2.25 and 2.0 my, respectively. We prefer to use the base of the Olduvai in drawing the Pliocene/Pleistocene boundary and accordingly we place the Pliocene/Pleistocene boundary at about 1.8 my. The point is that this boundary has been narrowed down to a small time-span from estimates which formerly ranged from less than 1 million years to over 3 million years. Basic concepts relating to the Pliocene/Pleistocene boundary have been discussed more fully by Hays and Berggren (1971).

Climatic change has been the most fundamental and distinctive characteristic of the Pleistocene. However, it cannot be used as a criterion in determining or defining its limits. This idea has been held by most Quaternary specialists and persists in recent research (see Morrison, 1968). The most obvious flaw in this belief is the failure of 'climatic deterioration' or 'climatic change' or 'climatic cooling' to provide a basis for recognizing the boundary between the Pliocene and Pleistocene epochs. Rather, it was the recognition of paleontologic criteria which could be correlated with the Italian stratotype section, and subsequently dated by paleomagnetic methods (Berggren *et al.*, 1967; Phillips *et al.*, 1968), which placed the Pleistocene within the proper time perspective for the first time. Within this time framework climatic cycles could be, and are now, recognized. Indeed, the base of the Pleistocene as stratotypified in Italy is not characterized by a significant cooling.

The concept of equating the Quaternary with the ice age exists to the present day in the publications of Quaternary specialist and stratigraphic generalist alike. But there is ample evidence now that although climatic deterioration occurred in the world during the Cenozoic, in the Northern Hemisphere this was accentuated in the late Neogene and led to the development of continental glaciation during the Pliocene some 3 million years ago.

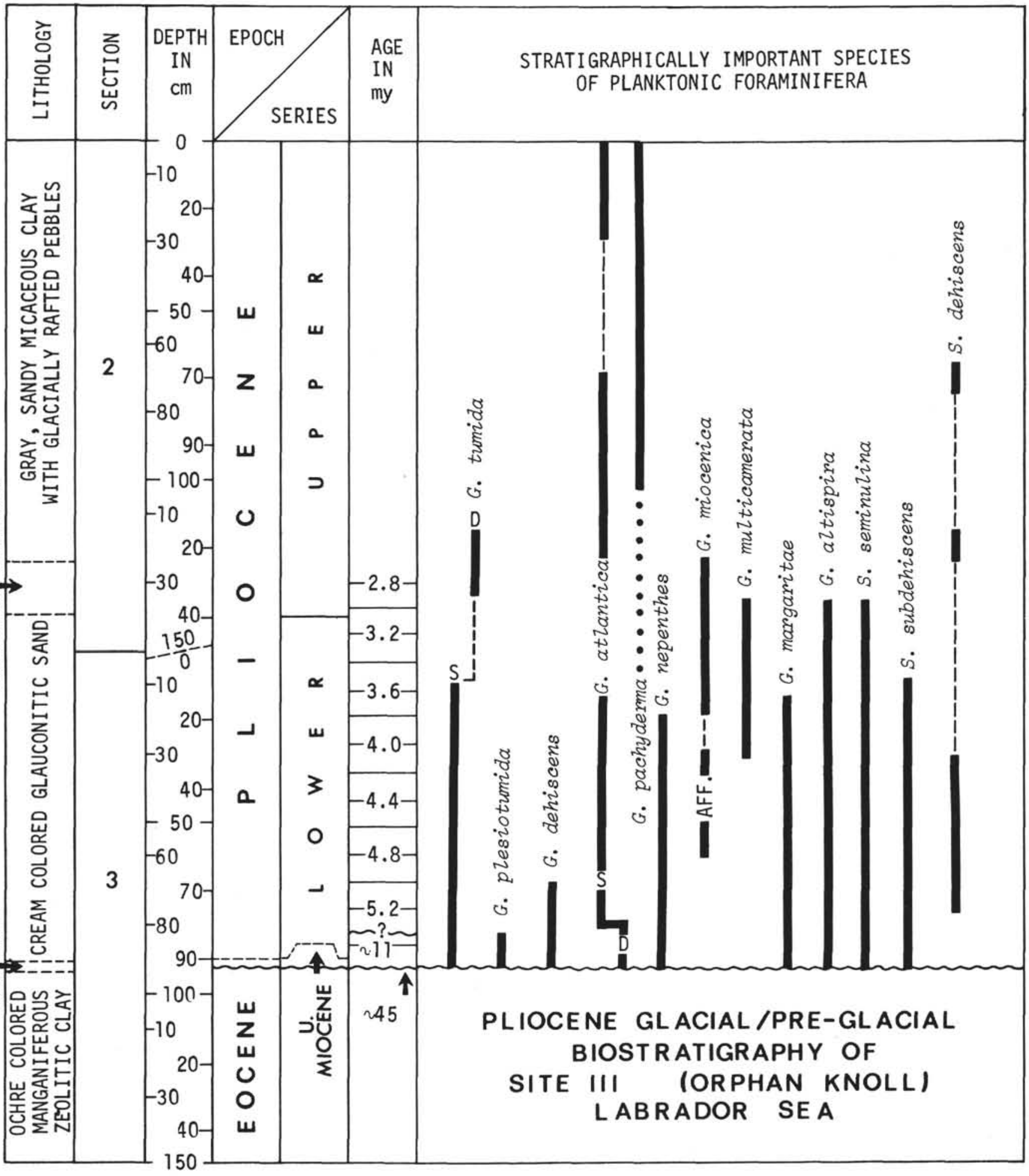

Figure 3. *Pliocene glacial/pre-glacial biostratigraphy of Site 111 (Orphan Knoll) Labrador Sea.*

In general it has been assumed by Quaternary specialists that the four classic glacial stages of the Alps can be correlated with those of North America. The older Donau has only rarely been correlated with an American glacial stage to my knowledge (Richmond, 1970, correlates it with the Nebraskan). At the same time two alternative schools of thought have arisen regarding the chronology of the Pleistocene glaciations (now seen to encompass late Pliocene time as well). These are the "short chronology" school (Emiliani, 1964, 1966a, b; Selli, 1967, Hays and Berggren, 1971) in which the four classic glaciations are compressed within the last 0.7 million years, and in the case of Emiliani, within the last 0.4 million years; and the "long chronology" school (Beard, 1969; and most Quaternary specialists) in which the four or five classic glaciations of the Northern Hemisphere are essentially equated with the Pleistocene Epoch and placed within a framework ranging roughly from 1.5 to 2.5 million years. The evidence cited above suggests that the "long chronology" interpretation is likely to be more consistent with the data now available.

There have been several attempts to relate the deep-sea climatic record to the classic glacial periods (Emiliani, 1961, 1964, 1966a, b; Ericson *et al.*, 1964; Ericson and Wollin, 1968; Beard, 1969). Emiliani's (1961) earlier estimates placed the base of the glacial Pleistocene at about 0.3 million years and later (1966a, b) at 0.42 million years, and the base of the Pleistocene Epoch at 0.6 to 0.8 million years (1961, 1964, 1966a, b). Ericson *et al.* (1964) estimated the base of the Pleistocene (and glaciation) at 1.5 million years. Following the establishment of a paleomagnetic time-scale, they extended the Pleistocene glacial chronology to 2 million years (Ericson and Wollin, 1968). Beard (1969) has recently extended the glacial chronology to 2.8 million years, suggesting a correlation of the earliest eustatic lowering of sea-level in the Gulf of Mexico and climatic deterioration based on foraminiferal data with the Nebraskan glaciation. Although I would disagree with the determination of the Pliocene/Pleistocene boundary at 2.8 million years (base Nebraskan) by Beard (1969, Figure 2), the chronology proposed in this paper is seen to be, in broad terms, quite close to that of Beard (1969) and Richmond (1970).

Although at present ice covers about one-tenth of the land surface, during the extensive glacial episodes of the Pleistocene as much as 30 per cent of the land surface may have been covered by ice (Holmes, 1965). Half of this was North America where the ice radiated from three main regions: a) Labrador; b) Hudson Bay; c) the Cordilleran ranges of the west.

In Central Europe four major glacial episodes have been recognized: Gunz, Mindel, Riss and Würm. A fifth, and older one, the Donau (Danube) is now generally accepted to be the oldest Alpine glaciation. There appears to be general agreement among glaciologists that these five Central European glacial episodes correspond to the episodes recognized in North America: Nebraskan, Kansan, Illinoian and Wisconsin. For an up to date summary the reader is referred to Richmond (1970).

To date there has been no general agreement on the exact correlation of the various glacial episodes on the continents of North America and Europe with the paleoclimatic cycles as reflected in the deep sea record. The difficulty rests in the nature of the data. With the exception of the loess sequences, the record on land is essentially a discontinuous one. The various "glacial stages" of the classic Alpine sequence are preserved in terraces which are a geomorphologic phenomena representing the growth and expansion of glaciers. As in the case of the glacial tills deposited by the North European and North American ice-sheets, they have been subsequently cut and denuded during the intervening interglacial intervals, as well as eroded by subsequent glacial advances. The post-Kansan glacial record on land indicates that glaciers alternatively advanced and retreated within recognized glacial cycles. The pre-Kansan record is not clear. Thus the land record is extremely patchy and the preserved part of a given "glacial stage" in many instances represents but a small fraction of the total time involved in the expansion and reduction of the glacial sheet. On the contrary, in the deep-sea record one sees cyclic changes of varying periodicities which reflect paleoclimatic changes on a relatively fine scale (Ericson *et al.*, 1961, 1963; Ericson and Wollin, 1968; Emiliani, 1955, 1961, 1964, 1966a, b; Broecker and Van Donk, 1970; Imbrie and Kipp, 1971). Indeed, the paleoclimatic cycles recognized within the late Pleistocene–when extended into older intervals–may prove, eventually, to be the best means of providing a chronology of the Late Pliocene–Pleistocene glacial record. Thus, whereas the record left on land by the glaciers has suggested the alternate advance and retreat of glaciers in a semi-quantitative manner, the deep-sea record suggests that within the last 0.7 million years alone an alternation of at least eight cold and warm cycles has occurred (Hays *et al.*, 1969) with periodicities ranging from 75,000 to 100,000 years. Indeed, it has been shown that the primary glacial cycle has a sawtoothed character, with gradual glacial buildups over periods of about 90,000 years followed by rapid deglaciations covering less than a tenth of that time (Broecker and Van Donk, 1970). It is because of these factors that the so-called classic "Pleistocene glacial stages" cannot be considered in time-stratigraphic terms, but rather as representative of intervals in which one or more glacial advances occurred. The purpose of the present discussion is primarily to attempt to place these various terms within a chronologic framework, not to suggest accurate correlation between them and the climatic record in the deep sea. In the following discussion reference should be made to Figure 4.

Our point of departure is the initiation of major glaciation in the Northern Hemisphere about 3 million years ago. Deposits of the McGee glaciation in the Sierra Nevadas, tentatively correlated with the Nebraskan continental glaciation (Blackwelder, 1931; Sharp and Birman, 1963), rest directly upon a basalt flow dated at 2.6 million years (Dalrymple, 1963). Glacial deposits between 2.7 to 3.1 million years are also known in the Sierra Nevadas (Curry, 1966). In Central Europe the Donau is the oldest recognized Alpine glaciation. In northern Europe the Praetiglian is recognized as the:

> oldest cold stage encountered so far in continental sediments of the Netherlands. The climate during this

> stage, called the Praetiglian, was subarctic. . . . It is probable that the oldest glacial stage in Europe has been recorded here (van Montfrans, 1971, p. 27).

The boundary between the Pliocene/Pleistocene is drawn here by most Dutch geologists, that is between the Reuverian and Praetiglian Stages. The boundary, drawn on the basis of palynology in nonmarine sediments (Zagwijn, 1960, 1963) has been shown to be essentially correlative with that drawn on the basis of the appearance of the arctic form *Elphidiella arctica* in marine sediments (van Voorthuysen, 1950).

Northeast of Castenedolo, in eastern Lombardy, in the pre-Alps area of northern Italy, morainic-fluvioglacial deposits of the Günz glaciation have been correlated with the marine upper Calabrian Stage and these are overlain in turn by lacustrine and morainic deposits of Mindel age (Venzo, 1968). The sequence is analogous to Lippe (Bergamo) where lacustrine deposits with a rich pollen flora are underlain by deposits of the Donau glaciation with four cold stadials alternating with three interstadials (*op. cit.*, p. 357). Venzo (1968) observed that the climate in Donau time was not as severe as during the Günz, and especially the Mindel stadials. This is corroborated by studies in the deep sea which suggest that glacial intensity increased with time. A point of importance for correlation pointed out by Venzo (1968) is that the earliest conglomerates of the Villafranchian exposed at the base of the Casnigo terrace dammed the Leppe Basin, so that "they must be correlated with the beginning of the Donau Glaciation (fluvioglacial facies)" (Venzo, 1968, p. 358). The base of the Villafranchian has been dated at about 3.4 million years at Etollaires (Auvergne, France) (Savage and Curtis, 1967). It would seem, then, that we have evidence to suggest a correlation between the Donau, Praetiglian and Nebraskan "glacial stages". They represent the first major glaciation of the Northern Hemisphere which began about 3 million years ago.

Richmond (1965, 1970) correlates the oldest glacial deposits in the Rocky Mountains—the Washakie Point—with the Donau and suggests that it was older than 1.2 million years.

McDougall and Wensink (1966) dated a normally magnetized lava flow in the Graue Stufe, Jökuldalar region, in northeastern Iceland at 3.10±0.10 million years, and placed it within the Gauss Normal Epoch. They have noted that the oldest glacial intercalation in the Graue Stufe in this area occurs just below the reversed polarity lavas of the Mammoth Event and immediately above the basalt dated at 3.10±0.1 million years. The fact that older glaciations have not eroded deep valleys in the basalts in contrast to the more recent glaciations suggests to the authors that the tillites are ground moraines from extensive ice sheets rather than moraines from small localized glaciers.

This agrees precisely with our evidence in North Atlantic cores where the first ice-rafted detritus appears at 3 million years. Glaciation in the North Atlantic began 3 million years ago, and it is probably valid to suggest that this was the time of the first major expansion of the Arctic ice sheets over Europe and North America, as well as, the mountain glaciations of the Sierra Nevadas, the Rockies, and the Alps.

The chronology and correlation of the Kansan "glacial stage" has presented problems no less taxing to stratigraphers than those involving the Nebraskan. The following data are pertinent to our consideration of the approximate chronology and stratigraphic correlation of the Kansan. The Blancan mammalian Stage of North America is correlated approximately with the Villafranchian (Durham *et al.*, 1954). The age limits of the Blancan Stage have been determined by Evernden *et al.* (1964): Blancan/Irvingtonian: 1.4 to 2.3 million years; Hemphillian/Blancan: 3.5 to 4.1 million years (within the Gilbert Reversed Polarity Epoch). A marked climatic cooling is recorded by the mammalian fauna in the middle part of the Blancan Stage (the upper part of which is correlated by Hibbard *et al.*, 1965) with the Nebraskan. The lower part of the Irvingtonian Mammal Stage is correlated with the subsequent Kansan Glacial Stage. *Mammuthus* bearing sediments—characteristic of the Irvingtonian—have been dated at about 1.36 million years (Evernden *et.al.*, 1964) in the Bruneau Formation of Idaho and in California at 1.5 million years (Hall, 1965). As Cox (1968, p. 120) points out:

> the evidence that the Irvingtonian is contemporaneous with continental glaciation is very convincing, while that for a warm interglacial interval between the late Blancan and the Irvingtonian much less so. Continental glaciation thus began no later than 1.5×10^6 years ago if these correlations are valid, with somewhat weaker evidence for yet an earlier glaciation.

But we have seen above that the evidence is clear for an earlier glaciation beginning at 3 million years. The conclusion I draw is that the Nebraskan (and its correlatives) began about 3 million years ago; the Kansan glaciation was occurring about 1.5 million years ago. Evidence from the deep sea record suggests a climatic cooling beginning about 1.6 million years (Gamper-Bravo, 1971) to 1.3 million years ago (Ruddiman, 1971) depending upon the interpretation of paleomagnetic data. Accordingly I have suggested a glacial interval—the Kansan—at about 1.6 million years in Figure 4. The reversely magnetized basalt upon which the Irvingtonian date is based (1.36 million years) in Idaho is within the Matuyama Reversed Polarity Epoch, and thus the Matuyama/Brunhes boundary (0.7 million years) is younger than the Kansan glaciation according to the paleomagnetic time scale. If the correlation of Kansan-Günz is correct, the Günz too, is older than 0.7 million years. (Incidentally, recent evidence suggests that the Rancholabrean mammalian Stage, above the Irvingtonian, is wholly of Wisconsin age or younger (Valentine and Lipps, 1970).

Richmond (1965, 1970) correlates the Alpine Günz with the Rocky Mountain Cedar Ridge glaciation and suggests that the latter ended about 0.7 million years ago. If the correlation of Günz = Cedar Creek is correct, the Gunz/Mindel interglacial may also have begun about 0.7 million years ago. I have suggested a slightly different chronology for this sequence of events (see discussion below and Figure 4). The Lower Eburonian in the Netherlands—a cold period—has been dated at 1.6 million years by the paleomagnetic method (van Montfrans, 1971) and I suggest that it, too, is correlative with a part of the Kansan=Günz (see Figure 4).

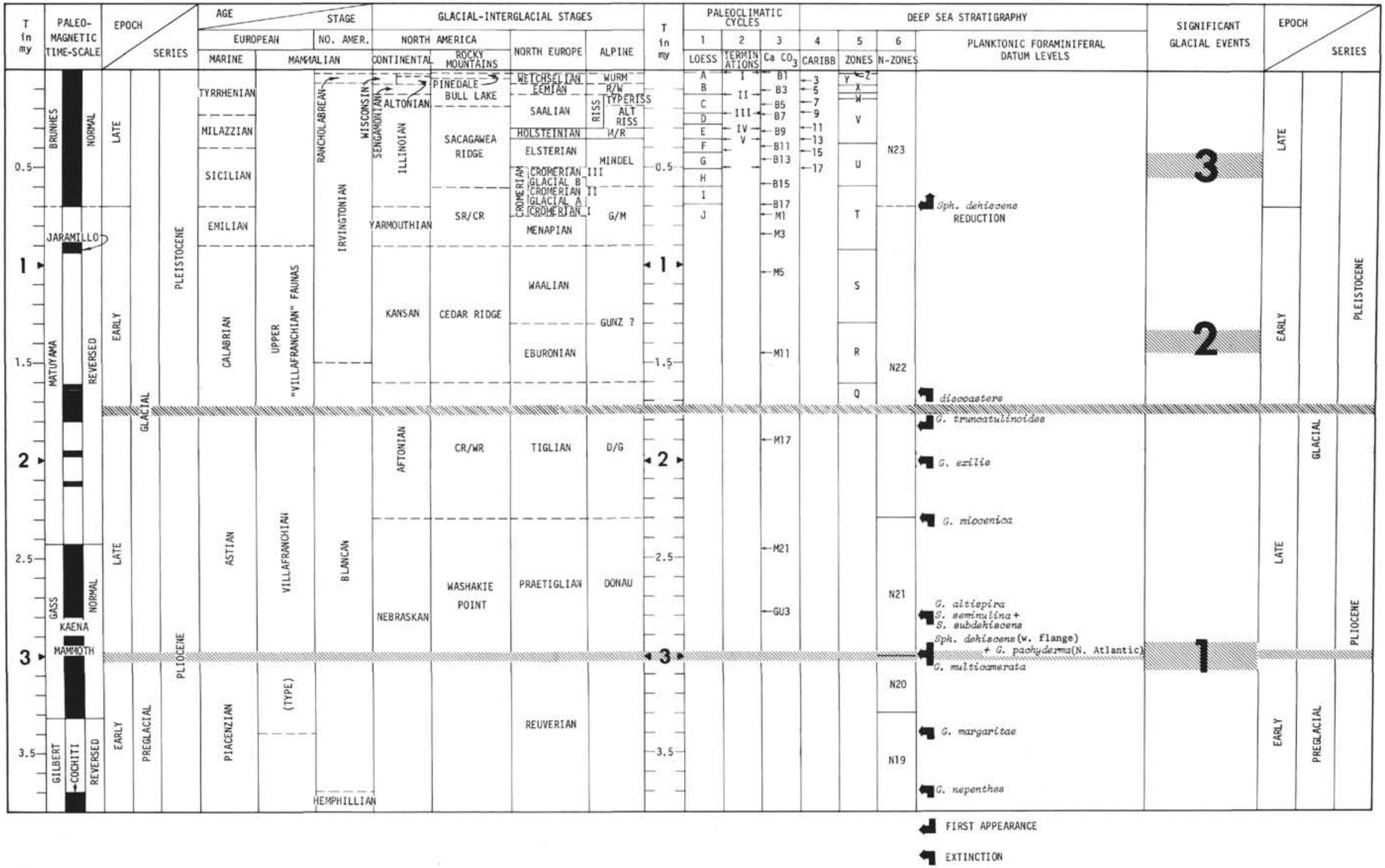

Figure 4. *A suggested chronology for Late Pliocene-Pleistocene glaciation. Three significant glacial events are shown on the right side. They are: (1) initiation of glaciation at 3 million years in Northerm Hemisphere; (2) major glacial event in Early Pleistocene about 1.5 million years; (3) major Late Plèistocene glaciation at about 0.4 to 0.5 million years.*

The data listed in the columns under Paleoclimatic cycles came from the following sources:

Column 1 (Loess) - Kukla (1970)
Column 2 (Terminations) - Broeker and van Donk (1970)
Column 3 (CaCo 3) - Hays et al., *(1969)*
Column 4 (Caribbean) - Emiliani (1961, 1964, 1966a, b)
Column 5 (Zanes) - Ericson et al., *(1963, 1968)*
Column 6 (Zanes) - Banner and Blow (1965)

The data dealing with the past 1 million years are numerous and difficult to synthesize short of monographic treatment. I shall attempt a brief summary of pertinent data insofar as they pertain to the correlations suggested in Figure 4. The marine record suggests that during the past 0.7 million years at least eight cold-warm cycles have alternated with each other (Emiliani, 1966a, b; Hays, *et al.*, 1969; McIntyre and Jantzen, 1969). Similar cycles are reflected in the record on land (Kukla, 1970; van Montfrans, 1971) and indeed Kukla (1970) has correlated his "Marklines"–denoting terminal points of glacial cycles–with the carbonate cycles found by Hays *et al.* (1969) and the terminations of Broecker and van Donk (1970).

Evidence in deep-sea cores suggests that the interval between 1 and 0.7 million years was, relatively speaking, warmer than the succeeding period (Gamper-Bravo, 1971; Ruddiman, 1971; Imbrie, personal communication).

Reversed polarity was found in the "glacial A", "Cromerian 1" and the Menapian of the Netherlands by van Montfrans (1971). The base of the Cromerian was correlated with the base of the Brunhes (at 0.7 million years) and the Waalian/Menapian boundary was tentatively drawn within the Jaramillo event. The Waalian and Menapian are correlated tentatively (Figure 4) with the upper part of the Kansan and the Yarmouthian.

Although correlation of the Würm, Weichsel and Wisconsin is generally agreed upon, the Riss and Mindel include moraines and terraces of multiple glacial cycles and their correlation with their North American counterparts are still hypothetical and speculative.

The base of the Mindel glaciation is placed at about 0.6 million years (within the lower part of the Brunhes). The Elsterian with which it is generally correlated, at least in part, is correlated with Loess cycle F or older by Kukla (1970, p. 162, Figure 10) which is approximately between 0.3 to 0.4 million years (Imbrie, personal communication). The Alpine Mindel is correlated with cycle G and/or older by Kukla (*loc. cit.*), which is about 0.4 million years (Imbrie, personal communication).

Richmond (1970, p. 11) states that the subsequent interglacial, equivalent to the Sangamonian, began at least 0.18 million years ago and ended about 0.12 to 0.13 million years ago. However, Kukla (1970, p. 162, Figure 10) indicates that the Saalian (which is generally correlated with the Riss) is correlative with his Loess cycle D and/or E. Cycle D and E span the time interval of approximately 0.22 to 0.35 million years (Imbrie, personal communication), which is considerably older than the time interval estimated by Richmond for the Riss (~0.12 to 0.75 million years). Richmond (1970, p. 12) discusses evidence for a separate glaciation between the type Mindel and type Riss. These deposits are generally referred to as late Mindel or Alt Riss by specialists. If we include them in the older part of the Riss, then they can probably be correlated with the Saalian.

Thus the Mindel, as envisaged here, corresponds to about half of the Brunhes Normal Epoch and encompasses several of the carbonate maxima observed in the Pacific cores by Hays *et al.* (1969). The "U" Zone of Ericson and Wollin (1963, 1968) contains a long significant cold faunal episode at about 0.4 to 0.5 million years (Ruddiman, 1971) and Imbrie (personal communication) has found that the U Zone has the coldest climatic extremes of the late Pleistocene and he is of the opinion that they correlate with the Mindel glaciation, at least in part. The Mindel is estimated to have had a duration of about 250,000 years, from 0.6 to 0.35 million years. The Holsteinian (=Mindel/Riss interglacial) is shown in Figure 4 to correspond to the lower, warmer part of Kukla's (1970) Loess cycle E.

The limits of the Riss have been the subject of considerable debate among glacial specialists (see summary in Richmond, 1970). The Riss is correlated with the Rocky Mountain Bull Lake glaciation, which is in turn correlated with the lower part of the Altonian substage of the Wisconsinian (Richmond, 1970). However, it should be remembered that we have correlated the older part of the Riss with the Saalian in this paper. Richmond (1970, p. 15) suggests an age span of *ca.* 0.12 my to some time older than 50,000 years for the Riss.

The marine record suggests that a positive or near positive sea-level stand occurred about 200,000 years ago (Broecker and van Donk, 1970). Following this there is a cold interval corresponding to the upper part of Kukla's (1970) Loess cycle C. This may correspond to the upper, or type Riss. Broecker and van Donk (1970, p. 185) point out that the ice-sheets that had built up by 130,000 years ago were largely destroyed during Termination II (~127,000 years). Although glaciers grew anew after maximum deglaciation ~124,000 years ago, their expansion was interrupted by two periods of reef-coral growth (Barbados Terrace II: ~103,000 years ago and Barbados Terrace I: ~82,000 years ago) (Mesolella *et al.*, 1969). This interglacial interval, preceding the most recent expansion of glaciers–the Wisconsin-Würm–is correlated with the high sea-level stand, Barbados Terrace III, at about 125,000 years ago and may represent the Sangamonian=Eemian (Broecker *et al.*, 1968; Mesolella *et al.*, 1969). In Figure 4 the Sangamonian is correlated with the Eemian and is estimated to have spanned the time interval of ~125,000 to 75,000 years ago.

Richmond (1970, p. 15) suggests that the Bull Lake glaciation began about 120,000 to 130,000 years ago (and was correlative with the Riss, exclusive of the Alt Riss). As I have discussed above the division between Alt Riss and type Riss is lowered to about 180,000 years (Figure 4) so that the type Riss corresponds to the cold period corresponding to the upper part of Kukla's (1970, Figure 10) Loess cycle C. Accordingly the base of the Bull Lake has been drawn at 180,000 years to correspond with the base of the type Riss in Figure 4.

The Würm is generally divided into two parts: an older, Alt Würm, and a younger, or Main Würm. The boundary is placed at about 28,000 years (Richmond, 1970). The Rocky Mountain Bull Lake-Pinedale interval is correlated with the Alt Würm, and the Pinedale glaciation with the Main Würm by Richmond (1970). In northern Europe the Weichsel is correlative with the Würm. Deep-sea sediments record two maximum cold periods in the Wisconsin=Würm: at about 70,000 years and about 18,000 years.

The final recession of the Main Würm was completed about 11,000 years ago and this event is dramatically recorded in the deep sea and on land by sea-level change as well as in floral and faunal assemblages. This level forms the boundary between the Pleistocene/Holocene, although for all practical purposes we are living within an interglacial interval within the Pleistocene and I prefer to retain the term Pleistocene for sediments deposited over the past 11,000 years as well.

Corroboration of this "long scale" chronology of glaciation in the Northern Hemisphere comes from studies on the stratigraphy of Tjörnes, northern Iceland (Einarsson *et al.*, 1967). A dramatic influx of Pacific boreal mollusks occurred in Iceland and the North Sea Basin either slightly more than 3.0 million years ago or more than 3.35 million years ago (depending upon the interpretation of paleomagnetic measurements made on the volcanic rocks of Tjörnes). Einarsson *et al.* (1967, p. 322) point out that this invasion occurred:

> at least half a million and perhaps more than a million years before the first Pleistocene glaciation was recorded there. The migration must have taken place at a time when the Arctic Ocean was warmer than at present, for some of the migrating taxa no longer range as far north as the Arctic Ocean.

Ten glacial cycles alternating with nine interglacial intervals are recorded in the Tjörnes sequences and indicate that:

> Pleistocene climatic history is much more complex than is suggested by the classical concepts evolved for the Alps and for the mid-continent of North America, which involve only four or five glaciations, and it casts serious doubt on intercontinental correlations of the older Quaternary glacial deposits (Einarsson *et al.*, 1967, p. 324).

This completes our discussion of a suggested chronologic framework for the late Pliocene-Pleistocene glacial sequence. We can summarize the basic conclusions of this discussion as follows:

1) Glaciation was initiated in the Northern Hemisphere about 3 million years ago.

2) Whereas glacial "stages" on land are represented more by gaps than by actual physical record, the deep-sea records a sequence of climate cycles with varying periodicity. A given continental glacial stage may, and in several instances probably does, encompass several deep-sea cycles of warm and cold climate. Direct correlation of the two chronologies will never be attained. The continental glacial stages should not be thought of as time-stratigraphic units.

3) The "long scale" school of glacial chronology, which interprets the glacial history of Europe and North America in a chronologic framework in excess of 2 million years is seen to reflect more accurately the data now available.

4) A general correlation of North American and European continental and mountain glacial sequences appears possible. In the younger part of the sequence it would appear that the glacial "stages" in one area may correspond to both glacial and interglacial intervals in other regions. An approximate correlation with the deep-sea paleoclimatic record is suggested.

5) Data suggest that successive glaciations increased in intensity reaching a maximum some 0.4 million years ago, in the Mindel (U Zone).

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