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Chapter 1

Introduction

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Grove Karl Gilbert, in his classic monograph on Lake Bonneville (1890, p. 1), succinctly described for all geologists the importance of the Quaternary:

> When the work of the geologist is finished and his final comprehensive report is written, the longest and most important chapter will be on the latest and shortest of the geological periods.

This book reviews the Quaternary geology of the contiguous United States beyond its glacial limits. Knowledge of the Quaternary has become increasingly important because it applies to many facets of paleoclimatology, engineering, and environmental geology, hydrogeology, and neotectonics.

We focus chiefly on Quaternary stratigraphy, not geomorphology. Geomorphic processes and systems that have operated in various regions of North America are discussed in Graf (1987). We also avoid discussing glacial geology as much as possible, although glacial relations are mentioned in areas adjoining glaciated ones. Richmond and Fullerton (1986) provide a recent synthesis of the glacial stratigraphy of the United States. Our volume also partly subordinates the Holocene and late Wisconsin records because Wright and Porter (1984) describe these records; also, Ruddiman and Wright (1987) cover part of this time span.

This book has two parts (see Table of Contents for details of coverage). The first part has short reviews of topics of general interest to students of the Quaternary: Quaternary paleoclimatology, dating methods applicable to the Quaternary, Quaternary volcanism, and Quaternary tephrochronology. Quaternary tectonism is treated in Slemmons and others (1991). Nonetheless, many of the regional chapters in this volume describe neotectonism within their region.

The second part of the book, more than three-quarters of its length, contains syntheses of Quaternary nonglacial geology of major regions of the contiguous United States. Chapters in this part are arranged to cover successive north-south strips from the Pacific to Atlantic coasts (Fig. 1 and Contents). The boundaries of these chapters generally conform to the outlines of the physiographic provinces of Fenneman (1930), with minor adjustments. The regional chapters provide general overviews for each region, augmented with summaries of detailed studies that give representative samples of the region. Consequently, the regional section has some gaps in its coverage of the unglaciated contiguous United States (Fig. 1). Most of this book covers regions west of the Mississippi River because of the generally better degree of accumulation and preservation of nonglacial Quaternary sediments there, and because fewer definitive studies have been made of equivalent deposits in the eastern United States. We have tried to synthesize the most up-to-date research, much of it previously unpublished, which results in conclusions often agreeing with, but in other places very different from, those given 25 years ago in Wright and Frey (1965).

WHY THE QUATERNARY PERIOD IS EXCEPTIONAL IN GEOLOGIC TIME

Climatic change is the outstanding characteristic of Quaternary time. Starting about 2.5 m.y. ago, the amplitude of climatic cycles increased greatly (Fig. 2), causing frequent large changes in the rates and types of deposition in both marine and terrestrial environments, to a degree that makes the better Quaternary stratigraphic records exceptional in geologic time. Fairbridge (1962, p. 111) commented:

> Seen from the vantage point of the whole geologic time scale . . . we must say: *the present climatic, oceanographic, structural, and sedimentological picture of the Earth is abnormal.* If we use the Lyellian philosophy of assuming the present is the direct key to the past we run a grave danger of being wrong. There is nothing wrong with that basic logic, but processes and relative factors are liable to great changes in velocity, scope, volume, etc.

Homo sapiens evolved in the late Pleistocene, and began to change from a hunting-gathering to a farming society 10,000 to 9,000 years ago, beginning human "civilization" soon after the start of the Holocene. From this perspective it is easy to understand why knowledge of the climatic, tectonic, and erosional/

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Figure 1. Landforms of the United States (by Erwin Raisz, published with permission of Kate Raisz). Dashed color line shows the maximum southern extent of the Laurentide and Cordilleran ice sheets; generally this is the pre-Wisconsin ice margin. The solid color lines outline the coverage by the various regional chapters in this book, designated by their chapter numbers. Dotted color lines indicate overlapping coverage between adjacent chapters.


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L. Superior
L. Michigan
L. Huron
L. Ontario
L. Erie
QUATERNARY
ICE
OF
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18/19
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21
OZARK PLATEAU
Gulf of Mexico
Atlantic Ocean
Chicago
Toronto
Montreal
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Charleston
Savannah
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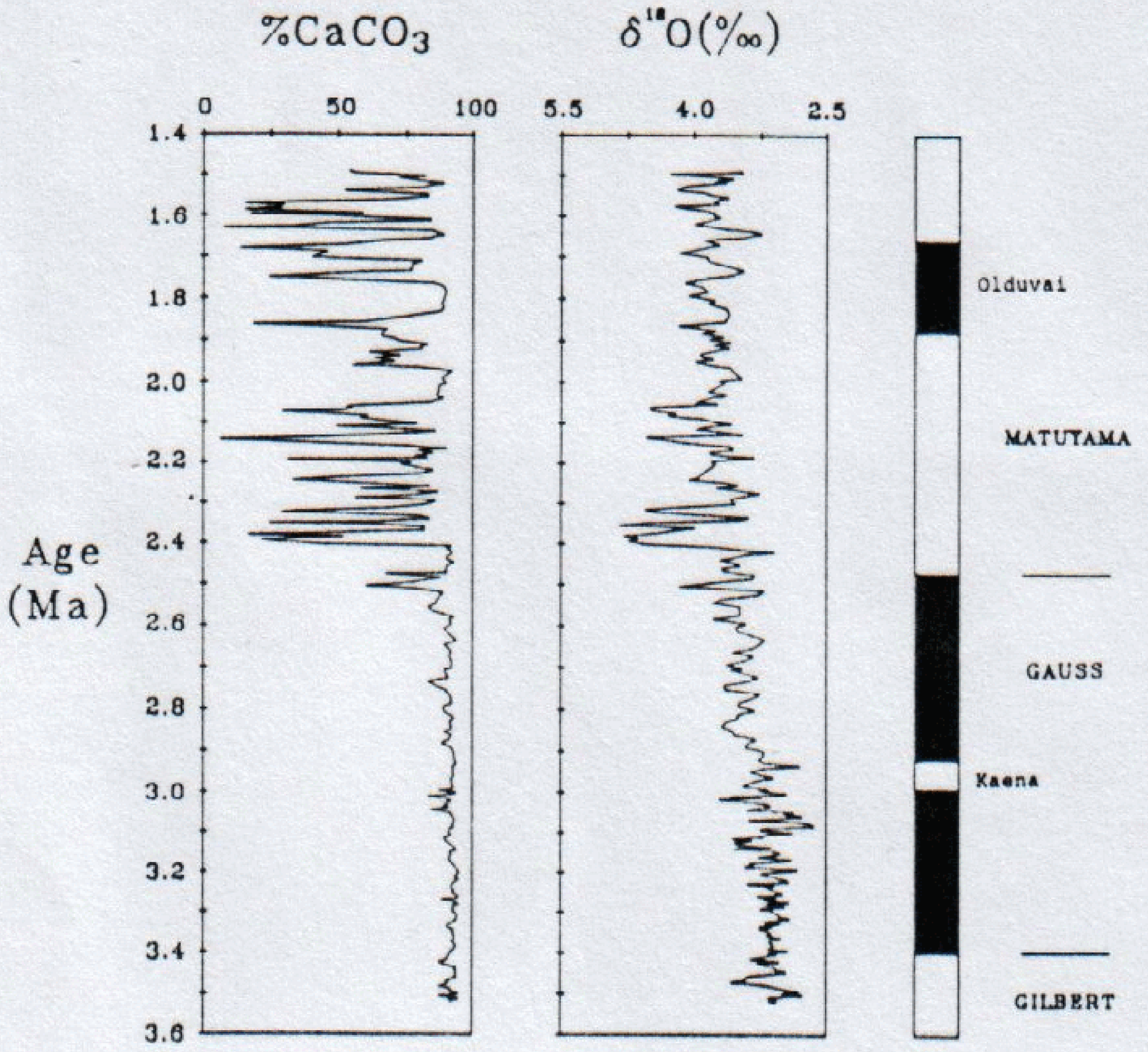


Figure 2. Late Pliocene and early Pleistocene records of percent $CaCO_3$ and $\delta^{18}O$ (from benthic foraminifera) in a deep-sea core from the North Atlantic Ocean. The abrupt increases in $CaCO_3$ near 2.55 Ma and especially 2.4 Ma mark the onset of ice rafting into the North Atlantic brought about by appearance of moderate-sized ice sheets in North America and Europe, after more than 1 m.y. of stable warm climate. (From Ruddiman and Wright, 1987a; after Shackleton and others (1984) and Zimmerman and others (1985); core site 552 at 56°03′N, 23°14′W.)

depositional history of the Quaternary Period is important in order to comprehend our changing environment and to predict our future.

Well-documented summaries of the development of concepts about the Quaternary and Pleistocene are given in Flint (1957, chapter 1; 1971, chapter 2). The Pleistocene was first defined (Lyell, 1839) on the basis of fossil mollusks. Later it became equated to widespread glaciation (the Great Ice Age) and also to the appearance of humanoids (the Age of Man) and certain other vertebrates. Modern research shows that these criteria (and biostratigraphic criteria such as microfossil assemblages) are too imprecise in chronostratigraphic definition to be internationally acceptable for marking the boundary between the Tertiary and Quaternary Periods (see Fig. 5). At present, two boundary levels are being considered, at about 1.65 and 2.5 Ma (see below), but a formal decision about this boundary and selection of an internationally acceptable boundary stratotype have not yet been achieved either by the International Association for Quaternary Research (INQUA) or by the International Geological Congress.

The deep-ocean core record

The most comprehensive record of late Cenozoic climatic change on a global scale is the oxygen-isotope ($\delta^{18}O/^{16}O$) record from deep-ocean cores (Figs. 2, 3, and 4) (Emiliani, 1955, 1967, 1970, 1972; Shackleton, 1969; Hays and others, 1969, 1976; Shackleton and Opdyke, 1973, 1976; Shackleton and others, 1984; Johnson, 1982; Imbrie and others, 1984; Ruddiman and Kidd, 1986; Ruddiman and Wright, 1987b). This record chiefly shows changes in the volume of ice stored on the continents during glaciations, and subordinately, temperature changes in the ocean-surface layer (Mix, 1987; Shackleton, 1969).

The oxygen-isotope record from deep-ocean cores has become a standard for Quaternary chronology, even among geologists who study terrestrial deposits, because the best deep-ocean cores provide far more complete sequences, with fewer time gaps than any terrestrial records. The deep-ocean record correlates strongly with long loessial records from central Europe (Fig. 3) (Kukla, 1975, 1977; Fink and Kukla, 1977) and China (Kukla, 1987, 1989; Kukla and others, 1988; Kukla and An, 1989). The marine oxygen-isotope cycles also correlate strongly with astronomical Earth-orbital cycles (Fig. 4), suggesting that Earth-orbital mechanisms were "pacemakers" for Quaternary climatic cycles (Hays and others, 1976; Johnson, 1982; Imbrie and others, 1984; Ruddiman and Wright, 1987a).

These records lead to the following four conclusions:

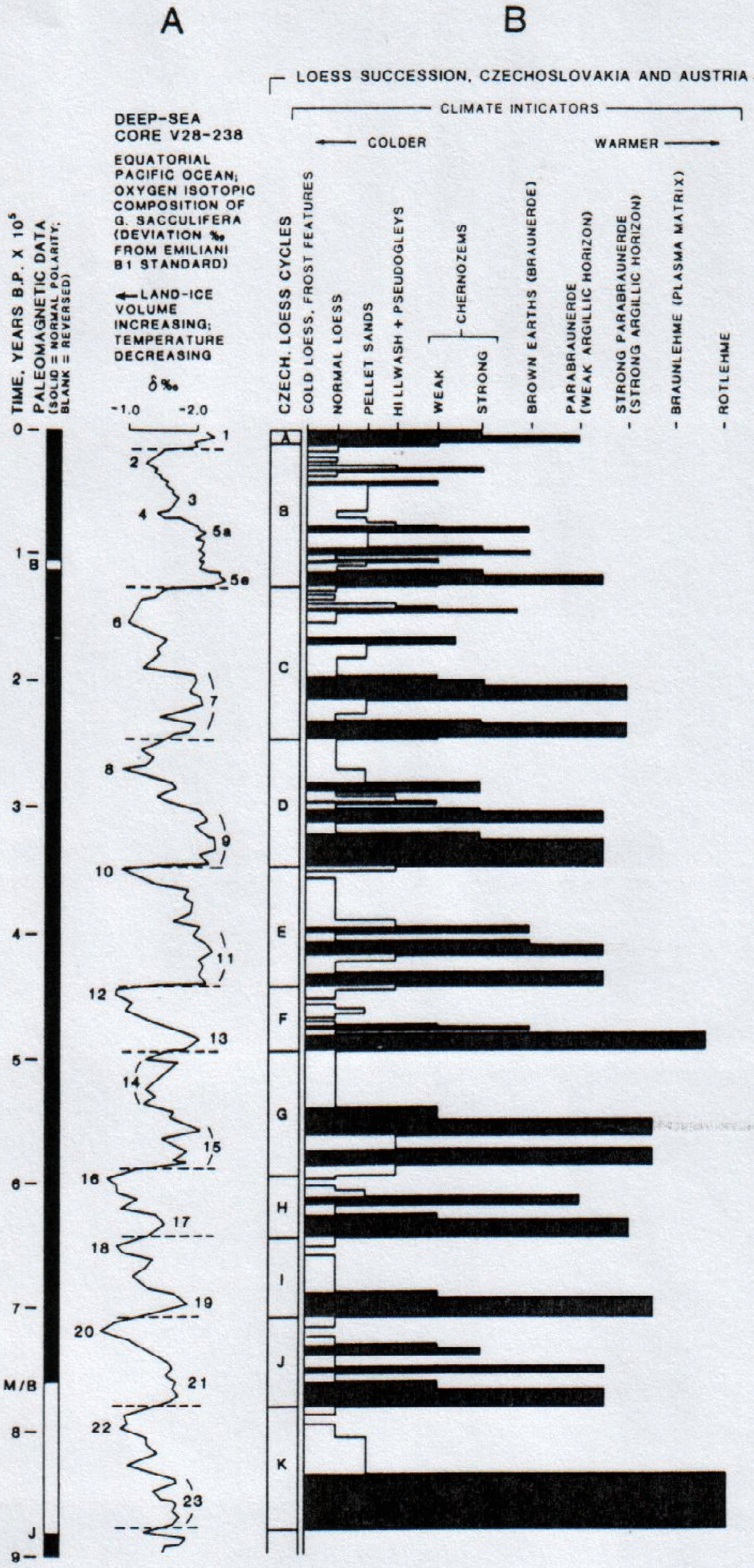
1. At least 17 complete interglacial-glacial cycles have occurred since the end of the Olduvai normal-polarity Subchronozone (about 1.65 Ma), and perhaps as many as 44 such cycles after the 2.48-Ma Gauss-Matuyama magnetic reversal (Gauss-Matuyama magneto-chronozone boundary) are recognizable in the loess sequences in China (Kukla and An, 1989; Kukla, 1989). The seven completed interglacial-glacial cycles during the past 620,000 years lasted between 88 and 118 k.y. apiece (Fig. 4, Table 1), an average duration of 100,000 years. Therefore, they were similar but not identical in duration. Also, they commonly differed in amplitude. Some cycles had stronger maxima and/or minima than others during their glacial and/or interglacial phases (Figs. 3, 4).

2. Five major interglacial intervals occurred during the past 529,000 years. Three of these were composed of two or three interglacial episodes separated by one or two cooler episodes (Table 1, Fig. 4). Interglacial episodes are defined here as times when the $\delta^{18}O$ values in deep-ocean-core data (from Imbrie and others, 1984, Table 7) were consistently at or below Holocene interglacial (since 9 ka) values; the $\delta^{18}O$ value of −0.05 is used here as a cutoff between interglacial and subglacial conditions.

On this basis, for the five completed interglacial-glacial cycles during the last 529,000 years, the interglacial episodes lasted from 28 to 49 k.y., or 27 to 42 percent of a given interglacial-glacial cycle (Fig. 4 and Table 1). This means that, including the Holocene, interglacial conditions prevailed for 39 percent of the last 529,000 years.

Interglacial and glacial intervals clearly have been somewhat erratic in occurrence and magnitude and have not been identical in duration. The last pre-Holocene interglacial interval, comprising all of O-isotope stage 5, lasted 55 k.y., albeit with two cool episodes (O-stages 5b and 5d; Figs. 3 and 4) in its later part. Its first part, O-stage 5e (Sangamon in the strict sense), was somewhat stronger than the Holocene and lasted about 18,000 years—longer than the Holocene. The three interglacial episodes within stage 5 occupied a total of 89 percent of stage 5 and 42 percent of the whole O-stage 2-5 interglacial-glacial cycle.

3. Ice accumulation accelerated greatly at the O-stage 5 to 4 transition; between 68 and 77 ka there was a 2-fold increase in the $\delta^{18}O$ ratio in deep-ocean cores. The Wisconsin/Würm glaciation peaked during oxygen-isotope stages 4 and 2 (stage 3 was a weak interstadial), and climaxed at 18 to 20 ka during O-stage 2. Deglaciation began about 15 to 14 ka and was almost complete about 9 ka, starting the interglacial phase of the Holocene.



4. Projecting this record into the future, it appears virtually certain that the Holocene interglacial will be followed by a long glacial phase. Such a cold phase will cause crises in food and energy supplies far more severe than any since civilized man developed. We are privileged to live in an exceptional time by paleoclimatic standards.

There are no certain paleoclimatic means of predicting exactly when the change from the present interglacial conditions to a much cooler climate will occur (note the erratic distribution and time spans of previous interglacials in Fig. 4 and Table 1). However, the past records provide a frame of reference that indicates possibilities. First, it is important to recognize that Earth now has progressed past an interglacial maximum into a somewhat cooler phase. If the Holocene resembles the last interglacial (which is rather unlikely), we Earthlings will soon (within hundreds to a few thousand years) come to the end of a warm episode like O-stage 5e, to enter a few thousand years of oscillating climate, with one or more subglacial cold episodes alternating with one or more near-Holocene-interglacial warm episodes. Then comes the onslaught of a long, severe major glacial phase lasting tens of thousands of years. The "Greenhouse Effect" warming is an aberration lasting perhaps several hundred years—and badly timed. It is too bad that it can't be postponed until it would help mitigate the inevitable next cooling phase, which will be beyond human control.

Correlations of deep-ocean oxygen-isotope records with other terrestrial sedimentary and paleoclimatic records

Although the deep-ocean oxygen-isotope record correlates strongly with loess records in central Europe and China (Fig. 4) (Kukla, 1975, 1977, 1987, 1989; Fink and Kukla, 1977; Kukla

Figure 3. Comparison of the deep-sea and central-European loess records for the last 900,000 years (from Morrison, 1978). A, the interglacial-glacial cycles represented by an important core from the Equatorial Pacific Ocean. Oxygen-isotope stage numbers are in color. The even numbers at the left indicate glacial maxima, and the odd numbers on the right denote interglacials. Also, in black horizontal dashed lines, are Terminations (T 1 to T 11) in the O-isotope record; these are the midpoints of the sudden transitions from glacial to interglacial conditions. B, a synthesis of the paleoclimatic record from key loess sequences in Czechoslovakia and Austria (Kukla, 1970, 1975, 1977). The horizontal axis represents semiquantitatively the indications of climate, from very cold glacial at far left to warm interglacial at far right, determined from snail faunas, sedimentologic, and pedologic criteria. The paleosols (shown in color) formed during the interstadials and interglacials. The black parts of the paleomagnetic data column represent normal polarity and the white parts, reversed polarity; B, indicates the Blake reversed polarity event; M/B, indicates the Matuyama-Brunhes Chronozone reversal (shown somewhat too young because it was based on 1977 data); J, indicates the Jaramillo normal-polarity Subchron.

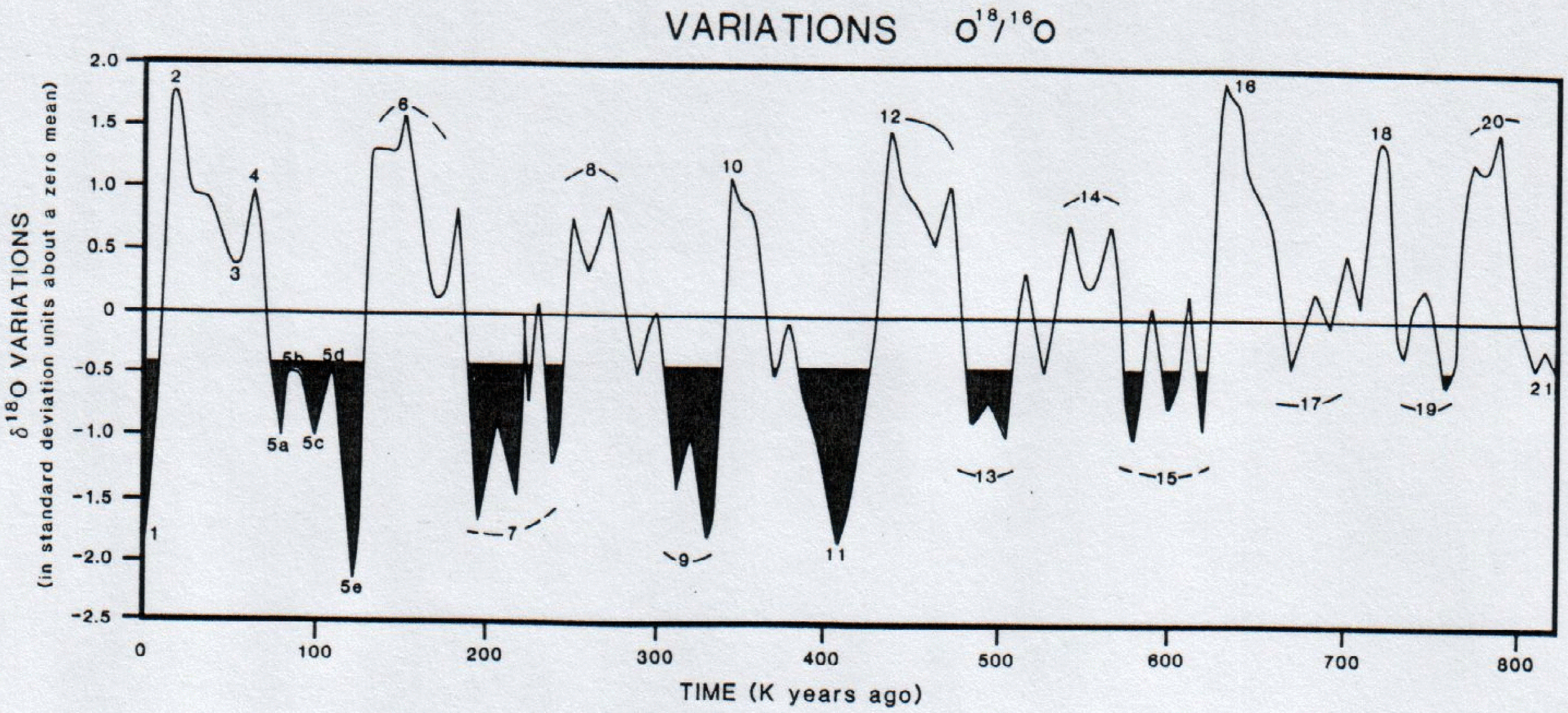


Figure 4. Record of $^{18}O/^{16}O$ variations in five deep-sea cores, tuned to each other and to earth-orbital parameters, as a function of time, plotted from Table 7 data in Imbrie and others (1984). Color shows the interglacial episodes, based on a cutoff at $-0.5\ \delta^{18}O$ oxygen-isotope values (equivalent to Holocene interglacial values). The numbers along the graph indicate oxygen-isotope stages; the even-numbered peaks (at top) are glacial maxima, and odd-numbered troughs are interglacial minima. Note: The portions of this graph older than 620 ka are here regarded as somewhat too young, and the portions older than 475 ka are believed to suppress excessively the oxygen-isotope minima for interglacial stages 13 through 21 (see text).

and others, 1988; Kukla and An, 1989); correlations with other terrestrial records commonly are less clear. Early investigators of terrestrial sequences tended to recognize only the more pronounced, larger amplitude manifestations of climatic change, essentially megacycle sets composed of more than a single interglacial-glacial cycle in the marine record.

The differences of opinion on correlation between marine and various terrestrial Quaternary sequences are due partly to the fragmentary character of nearly all terrestrial sequences (glacial, alluvial, lacustrine, volcanic, etc.) because of: (a) episodic deposition, with disconformities or diastems that represent small to large time-gaps in the geologic record; (b) disjunct records at various sites (natural/artificial exposures, drillholes, etc.) due to local erosion or nondeposition of units or concealment by younger deposits; and (c) many terrestrial records being diachronic/asynchronous within regional areas. Also, as paleoclimatic proxies to the deep-ocean record, the climate-cycle manifestations on land commonly were out of phase with the deep-sea record, being modified by time lags in the responses of various ocean/terrestrial systems to given climatic changes, by global differences in weather patterns, and by local orography, tectonism, and other factors.

The effects of glaciation were felt far beyond the glaciated areas. Along the coastal margins, glacioeustatic changes in sea level (commonly more than 100 m between glacials and interglacials) controlled not only the patterns of marine regressions and transgressions, as recorded by coastal marine terraces and platforms, but also controlled cycles of fluvial entrenchment and alluviation of coastal valleys, as well as episodic development of coastal dunes (Chapters 7, 19, and 21, this volume).

Unfortunately, some workers who make paleoclimatic models still propose "finger-matching" correlations between the deep sea and various terrestrial records. For example, many models date the last pleniglacial at 18 to 20 ka because it is the maximum of marine oxygen-isotope stage 2. However, the end moraines that record the maximum southward extent of the Laurentide ice sheet during the last glaciation range in age from about 22 to 14 ka in various places (Richmond and Fullerton, 1986). Also, the highest late Wisconsin pluvial-lake strandlines in the Great Basin range in age from about 18 to 13 ka. Earlier pluvial lakes in the Great Basin were at times out of phase with each other and with Sierra Nevada glaciations by tens of thousands of years (Morrison, Chapter 10, this volume).

Despite problems in sea versus land correlation for determining global or continental paleoclimatic models, there is no doubt that repeated climatic cycles of several orders of magnitude characterized the Quaternary. On land, these climatic cycles resulted in terrestrial-process cycles of similar orders of magnitude, causing repeated cyclic changes in the *rates* of all surficial processes (fluvial, eolian, mass-wasting, pedogenic, etc.), at times to

TABLE 1. COMPARATIVE DURATIONS OF INTERGLACIAL EPISODES VERSUS WHOLE GLACIAL-INTERGLACIAL CYCLES OF THE MARINE RECORD*

Oxygen-isotope interglacial-glacial cycle (O-isotope stages)	Age of whole cycle	Age of interglacial episode(s)	Percent of interglacial vs glacial time
2-5	10-128 (118)	73-85 (12) 91-110 (19) — (49) 110-128 (18)	42
6-7	129-245 (116)	188-219 (31) — (41) 233-243 (10)	35
8-9	245-338 (93)	302-337 (35)	38
10-11	338-426 (88)	391-424 (33)	33
12-13	426-529 (103)	478-506 (28)	27
14-15	529-620 (91)	572-580 (8) 592-602 (10) — (24) 612-618 (6)	37

*Ages listed are in thousands of years; durations of various episodes follow in parentheses. Imbrie and others' (1984) Table 7 is the source of the ages as well as the $\delta^{18}O$ values from which the various oxygen-isotope cycle boundaries, maxima and minima, and interglacial episodes are based.

Interglacial episodes are defined here as those with $\delta^{18}O$ values of minus 0.5 or less, comparable to or less than those of the interglacial part of the Holocene, after 9 ka (see Fig. 3).

such a degree as to cross an important geomorphic threshold, changing the dominant *type* of process in a given area. Nowhere on Earth have surficial processes acted at a steady state throughout the Quaternary!

Using the Great Plains as an example, I (Morrison, 1987) illustrate the results of the crossing of various levels of geomorphic thresholds, to induce erosion-deposition-landscape stability cycles of four orders of magnitude: (a) microcycles lasting in the 10- to 100-yr order of magnitude; (b) mesocycles of a 1,000- to 10,000-yr magnitude; (c) macrocycles lasting 95 ± 25 k.y.; and (d) megacycles that lasted 400 to 500 k.y. The macrocycles were approximately equivalent to, but not necessarily coeval with, the interglacial-glacial cycles of the marine record. The megacycles correspond to four or five interglacial-glacial cycles in this record and exhibit the strongest expression of four successive dominant-process stages: typically, (1) widespread downcutting, (2) lateral erosion, (3) alluviation, and (4) landscape stability and soil development.

CHRONOSTRATIGRAPHIC DIVISION OF THE QUATERNARY

Table 2 shows the chronostratigraphic divisions of the Quaternary used in this volume and the current best estimates of their boundary dates. The divisions are chiefly those used by Richmond and Fullerton (1986). The boundary dates are based mostly on correlations between oxygen-isotope data from deep-ocean cores and astronomical data on Earth-orbital variations (Berger, 1987), as discused below.

Formerly, the Pleistocene was divided on the basis of glaciations, the most striking manifestations of climatic change in the terrestrial stratigraphic record. The "classic" divisions in North America and Europe were based on the few then-recognized glaciations and interglaciations, and these divisions commonly were used akin to chronostratigraphic units. Now, as a result of more advanced research, many more glaciations (and interglaciations, stadials, and interstadials) are recognized throughout the Northern Hemisphere (Sibrava and others, 1986). Also understood is the fact that the boundaries of the physical units in glaciated areas (tills, outwash deposits, etc.) are strongly time transgressive (Richmond and Fullerton, 1986, p. 6, 8, 183–184, Chart 1).

Consequently, Quaternary workers are moving toward defining major chronostratigraphic boundaries on the basis of geologically isochronous units, such as tephra layers and geomagnetic reversals (Fig. 5). Tephra layers have limited areal extent, but magnetostratigraphic chronozone and subchronozone boundaries are recognizable throughout the world and therefore are now more frequently used internationally. This is illustrated by the recommendation of the INQUA 1987 Congress that the

TABLE 2. DIVISIONS OF THE QUATERNARY AND THEIR BOUNDARY DATES AS USED IN THIS VOLUME*

		Present
	Holocene (Oxygen-isotope stage 1)	
		10 to 12 ka
Late Pleistocene	Late Wisconsin (Oxygen-isotope stage 2)	
		~28 ka
Late Pleistocene	Middle Wisconsin of Richmond and Fullerton (1986) (O-isotope stages 3 and 4)	
		~71 ka
Late Pleistocene	Late Sangamon (Early Wisconsin and Eowisconsin of Richmond and Fullerton, 1986; O-isotope stages 5a-5d)	
		~115 ka
Late Pleistocene	Sangamon of Richmond and Fullerton (1986) (O-isotope stage 5e)	
		~128 ka†
Middle Pleistocene	Late-Middle Pleistocene (Illinoian of Richmond and Fullerton, 1986; O-isotope stages 6-8)	
		~300 ka
Middle Pleistocene	Middle-Middle Pleistocene of Richmond and Fullerton (1986) (O-isotope stages 9-15)	
		~620 ka§
Middle Pleistocene	Early-Middle Pleistocene (Richmond and Fullerton, 1986) (O-isotope stages 16-19)	
	(Matuyama-Brunhes Chronozone boundary)	750-775 ka**
	Early Pleistocene	
	Upper boundary of Olduvai Subcron	1.65 Ma
or	Gauss-Matuyama Chron boundary	2.48 Ma
	Pliocene	
		5.0-5.5 Ma‡
	Miocene	

Notes: *See text for supporting arguments.

†Corals from the highest strandlines of the last interglacial on Barbados and Curacao gave mean ages of 125 to 126 ka by high-precision uranium-thorium dating (Bard and others, 1989).

§Richmond and Fullerton (1986) use the Lava Creek B tephra layer, dated 620 ka by K-Ar and fission-track (G. A. Izett, U.S. Geological Survey, personal communication, 1987) to identify this boundary in much of the western U.S. This is the approximate age of the boundary between oxygen-isotope stages 15 and 16 (Figs. 1 and 2).

**INQUA's Subcommission on Subdivisions of the Pleistocene in 1987 recommended that the Matuyama-Brunhes paleomagnetic Chronozone boundary be adopted internationally as marking the boundary between the lower and middle Pleistocene.

‡The Miocene-Pliocene boundary currently is dated 5.0 to 5.5 Ma (Odin, 1982)

Matuyama-Brunhes Chronozone boundary be adopted internationally as the boundary between the lower and middle Pleistocene. Also, both the upper boundary of the Olduvai Subchronozone and (preferably) the Gauss-Matuyama Chronozone boundary currently are candidates for marking the Pliocene-Pleistocene (Tertiary/Neogene-Quaternary) boundary internationally (see below).

Age of the Pleistocene-Holocene boundary

Unfortunately, there is no magnetostratigraphic chronozone or subchronozone boundary at or close to the preferred position of the Pleistocene-Holocene boundary. Based on the deep-sea record, the Pleistocene-Holocene boundary should be placed at the boundary between O-isotope stages 2 and 1 (Termination I,

Fig. 3), commonly given as about 11 to 12 ka (e.g., Ruddiman and Wright, 1987a; Imbrie and others, 1984, Tables 6 and 7). However, in deep-sea cores from around the world this boundary is time transgressive between about 9 and 13 ka. Its various terrestrial litho- and biostratigraphic representations in North America and western Europe average about 10 ka (Fairbridge, 1972), albeit with much diachronism. Hopkins (1975) proposed an arbitrary age of 10,000 years as a compromise for divergent opinions based on land data. However, this proposal does not meet the requirement of the International Stratigraphic Code that a chronostratigraphic boundary of this rank be based on an internationally acceptable stratotype. Richmond and Fullerton (1986) accept 10,000 years as a provisional date for the Pleistocene-Holocene boundary; however, they note (p. 186) that it is a geochronometric boundary without a stratigraphic basis; it does not date the termination of continental glacial activity in the United States, and it has no significance in the overall record of glaciation in the United States. Neither INQUA nor the International Geological Congress have decided on a suitable stratotype and date for this boundary.

The Sangamon-Wisconsin boundary

For more than two decades, leading workers in Quaternary geology in the midwestern United States have placed the lower boundary of the Wisconsin(an) at 70 to 75 ka (e.g., Willman and Frye, 1970). This age corresponds to the boundary between marine oxygen-isotope stages 5 and 4, and would make O-stage 5 entirely interglacial. However, Richmond and Fullerton (1986) regard the O-stage 5d to 5a interval to be part of the Wisconsin glaciation and designate it "Eowisconsin" and "Early Wisconsin." This is because they correlate moderate glacial advances on the Yellowstone Plateau, Wyoming (as well as in Alaska and perhaps eastern Canada), with O-isotope stages 5b and 5d. Nonetheless, there is no reliable documentation in Europe or Asia of glacial advances correlative with O-stages 5b or 5d (Sibrava and others, 1986). Therefore, many students of terrestrial and marine records still regard the whole of O-stage 5 as a single, albeit modulated, interglacial.

Age of the Matuyama-Brunhes Chronozone boundary

The Matuyama-Brunhes (M/B) magnetostratigraphic Chronozone boundary has been proposed by the INQUA Commission on Subdivisions of the Pleistocene (1987) as the most acceptable marker for the boundary between the lower and middle Pleistocene. However, the age of this magnetostratigraphic boundary cannot be ascertained directly; like all geomagnetic reversals it must be determined by dating underlying and overlying strata by isotopic, fission-track, or other methods at many localities. The best approximation of the age of the M/B chronozone boundary appears to be about midway between the estimates of Mankinen and Dalrymple (1979), Imbrie and others (1984), and Johnson (1982) (730 ± 11, 734 ± 5 and 788 ka, respectively), for the following two reasons.

1. Johnson's (1982) date of 788 ka is somewhat too old, because it does not allow enough time between the M/B chronozone boundary and the end of the Jaramillo Subchronozone (well dated at 890 ka), as evinced by deposition rates in many deep-sea cores (G. J. Kukla, personal communication, 1989). Nonetheless, Richmond and Fullerton (1986) accept Johnson's date as a provisional age for the M/B boundary.

2. On the other hand, the ages for the M/B reversal used by Mankinen and Dalrymple (1979) and by Imbrie and others

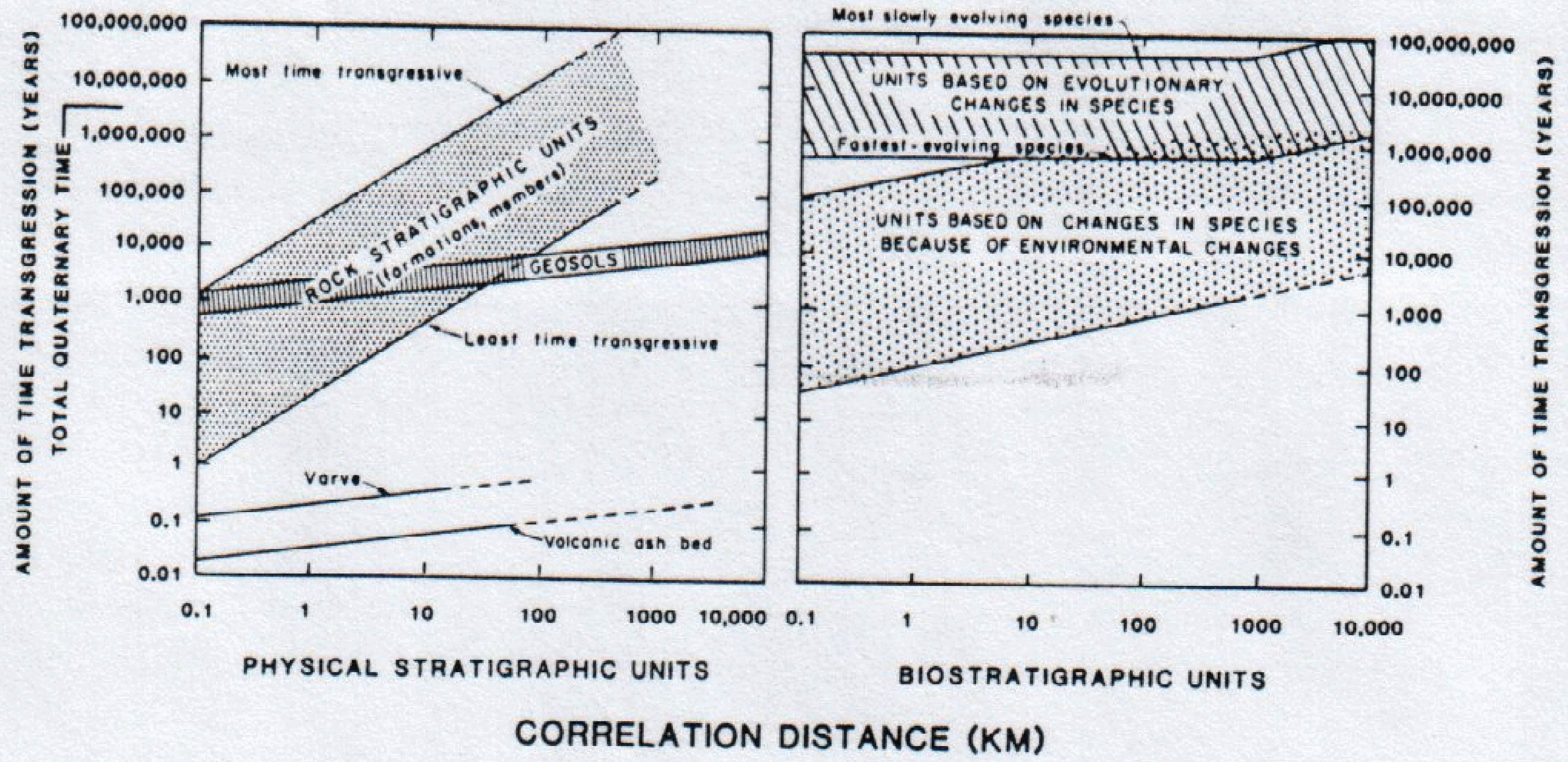


Figure 5. Amount of time transgression with correlation distance for various kinds of stratigraphic units (from Morrison, 1968). Note the logarithmic scales; also that the least time-transgressive units are varves, volcanic ash beds, magnetozone and chronozone/subchronozone boundaries/reversals, and geosols. Varves have the least geographic extent, but geomagnetic chronozone/subchronozone reversals are global. (The band designated "geosols" also represents these paleomagnetic reversals, which have similar time spans; the geosols considered here represent only a single interglacial or interstadial, not compound pedocomplexes.)

(1984) clearly are too young because they are younger than the Bishop Ash, which has normal polarity. Glen Izett (U.S. Geological Survey, personal communication, 1989; Izett and others, 1988) redetermined the age of this tephra layer, obtaining 738 ± 3 ka as a weighted mean of 15 dates on sanidine (14 ^{40}K-^{40}Ar dates and one ^{39}Ar-^{40}Ar date). The Bishop ash lies 3 m above the M/B reversal in a Lake Bonneville (Utah) sequence cored at the southern edge of the Great Salt Lake, and a strongly developed paleosol lies just below the ash layer; the M/B reversal is estimated from deposition rates to be between 15 and 40 k.y. older than the Bishop Ash, giving an age of 753 to 778 ka for the reversal (Eardley and others, 1973, Fig. 1 and p. 212). Also, in two borehole cores near Bakersfield, California, the M/B boundary was identified in lacustrine clay 3.7 and 4.9 m below the Bishop Ash; the average deposition rate including diastems is 11.7 cm/1,000 yr, making the approximate age of the M/B boundary about 775 Ma (Davis and others, 1977). Furthermore, on the basis of K-Ar ages from a volcanic sequence in the Jemez Mountains, New Mexico, the age of the M/B reversal is estimated to be about 770 ka (G. A. Izett, personal communication, 1984).

Because of the above considerations, the Matuyama-Brunhes Chronozone boundary is tentatively dated at 750 to 775 ka in this volume. The Bishop tephra layer is a key marker <1 m to rarely >3 m above the M/B reversal in remnants scattered widely over the western United States (Chapters, 5, 6, 7, 10, 13, 14, this volume).

Correlation of the deep-ocean-core and astronomical earth-insolation chronologies

Many deep-sea core-record chronologies were presented before Johnson (1982) published the first attempt to correlate the deep-sea oxygen-isotope and Earth-orbital records by statistical analysis, using oxygen-isotope data from a core from the central-western Pacific Ocean. Toward the same goal, Imbrie and others (1984) used more sophisticated statistical techniques to correlate data from this and four other deep-sea cores (from the Southern Atlantic, Indian, and Southern Oceans, and Caribbean Sea), and to correlate the core records with astronomical earth-orbital parameters (Berger, 1984). Three of the cores penetrated the M/B chronozone boundary. Imbrie and others initially used two calibration points: 127 ka for the O-stage 5/6 boundary, and 730 ka (from Mankinen and Dalrymple, 1979) for the M/B chronozone boundary. After the oxygen-isotope curves were "tuned" to the precessional parameters and averaged, the final ages of these calibration points were 128 and 734 ka, respectively.

However, the 734-ka age for the M/B chronozone boundary is too young, as explained above; a better estimate is between 750 and 775 ka. From the present back to about 620 ka the deep-sea O-isotope data not only are fine-tuned to close agreement with astronomical data but also agree well with terrestrial data; however, earlier than 620 ka the ages given by Imbrie and others become discordant, particularly with terrestrial data such as European and Chinese loess sequences (G. J. Kukla, written and oral communication, 1989). For example, O-isotope stage 19 is only 10,000 years long in Imbrie and others (1984), yet on land it is represented by a polygenetic paleosol comparable to all paleosols in the whole of O-isotope stage 5. Moreover, starting with interglacial O-stage 13 and continuing through interglacial stages 15, 17, 19, and 21, the $\delta^{18}O$ *minima* are not nearly low enough; these minima indicate only weak interglacial to subglacial conditions (compare Figs. 3 versus 4). In contrast, the central European and Chinese loess records (Fig. 2) show by degree of paleosol development and snail faunas that O-stage 13 was one of the Pleistocene's stronger interglacials, and that O-stages 15, 17, 19, and 21 were as strong or stronger than the Sangamon.

Because of these problems, I regard the portion of Imbrie and others' (1984, Table 7) data pertaining to ages older than 620 ka as somewhat too young; also, back beyond about 745 ka the oxygen-isotopic data strongly suppress the true amplitudes of interglacial minima.

Age of the Pliocene-Pleistocene boundary

Two quite different stratigraphic levels/ages currently are proposed for the Pliocene-Pleistocene boundary: (1) the end of the Olduvai normal-polarity Subchronozone, dated about 1.65 Ma; and (2) the 2.48-Ma Gauss-Matuyama magneto-chronozone boundary.

Placing the Pliocene-Pleistocene boundary at the end of the Olduvai Subchronozone. This is the provisional boundary selected in 1981 by joint resolution of the Working Group of the International Geological Correlation Program Project 41 (Neogene-Quaternary Boundary) and the International Union for Quaternary Research (INQUA) Subcommission 1-d on the Pliocene-Pleistocene Boundary (International Commission on Stratigraphy Working Group on the Pliocene-Pleistocene Boundary). Nevertheless, the end of the Olduvai subchron is seriously unsuitable as a candidate for this geologic-period boundary, for the following reasons.

1. Aguirre and Pasini (1985) propose that the international stratotype for the Pliocene-Pleistocene boundary be designated as the top of the Olduvai normal-polarity Subchronozone in the Vrica section, southern Italy. However, the proposed stratotype area is much deformed and faulted, with many tectonic and erosional hiatuses; even the Vrica section is truncated. Moreover, the paleomagnetic, tephrochronologic, biostratigraphic, and chronologic data are ambiguous and may be in serious error (Kukla, 1987, p. 214–216). Identification of the Olduvai Subchron here is questionable; the normal-polarity strata may represent an older subchronozone such as the Reunion (Arrias and Bonnadona, 1987).

2. The relatively short Olduvai Subchronozone cannot be identified paleomagnetically in many Pliocene-Pleistocene marine and terrestrial sequences, and identification of the precise position of its upper boundary is even less common.

3. Neither the Olduvai Subchronozone nor its upper boundary marks a substantial climatic event on a global basis; they are not marked by a distinctive worldwide litho- or biostratigraphic discontinuity.

Published comments adverse to placing the Plio-Pleistocene boundary at the top of the Olduvai Subchronozone include Richmond and Fullerton (1986, p. 186), who state,

> . . . there are no criteria by which the Pliocene-Pleistocene boundary thus defined can be located accurately in the stratigraphic sequences in the U.S.A. . . . The Pliocene-Pleistocene boundary thus defined has no significance in the stratigraphic and chronologic framework of glaciation in the United States. . . .It has no significance with respect to the dispersal of microtine rodents . . . or other vertebrate faunas . . . that distinguish the North American land mammal ages; . . . no clear significance with respect to climatic or environmental changes in North America based on biotic criteria.

In addition, G. I. Smith (Chapter 11, this volume) observes, regarding the deep-core record at Searles Lake, California,

> . . . the 1.6 Ma "beginning of Quaternary time" falls near the middle of a virtually uninterrupted intermediate hydrologic regime that lasted about 0.75 m.y.

Kukla (1987) comments that the proposed Pliocene-Pleistocene boundary at the top of the Olduvai Subchronozone has no lithostratigraphic or biostratigraphic representation in the loess sequences of China.

Geologists working with offshore core and seismic data in the Gulf of Mexico generally cannot recognize this boundary on stratigraphic or paleomagnetic grounds; they almost uniformly prefer a more distinguishable boundary.

The Gauss-Matuyama Chronozone boundary as a candidate for the Pliocene-Pleistocene boundary. Mounting evidence indicates that the 2.48-Ma Gauss-Matuyama Chronozone boundary would be a more suitable candidate for the Pliocene-Pleistocene boundary (Ruddiman and Wright, 1987b; Kuukla, 1987, 1989; Kukla and An, 1989). Arguments on behalf of placing the Pliocene-Pleistocene boundary at the Gauss-Matuyama magnetic reversal are as follows.

1. The Gauss-Matuyama magnetic reversal was close in time to a sudden cooling of the Earth and the initiation of moderate-sized ice sheets in North America and Europe, between 2.55 and 2.4 Ma (Fig. 2), after several million years without significant glaciation. This cooling ended a Pliocene warm period whose climatic cycles were much smaller in amplitude than those of the Pleistocene and never became colder than the Pleistocene interglacials, even at high latitudes (Matthews and Poore, 1981). This climatic shift—the true beginning of the "Great Ice Age"—is recorded in cores from the subpolar North Atlantic and the Labrador and Norwegian Seas by a marked increase in ^{18}O, a decrease in percent $CaCO_3$ (Fig. 3), and an increase in ice rafting (Backman, 1979; Shackleton and others, 1984; Zimmerman and others, 1985; Ruddiman and Kidd, 1986; Eldholm and Thiede, 1987; Srivastava and Arthur, 1987; Ruddiman and Wright, 1987a).

2. This drastic climatic change is well represented by lithologic and biostratigraphic discontinuities in marine and terrestrial sequences throughout the world (e.g., by the start of loess deposition in Europe and China (Kukla, 1987, 1989; Kukla and An, 1989).

3. The Gauss-Matuyama polarity reversal can be identified unambiguously in terrestrial and marine sequences throughout the world, much better than the top of the Olduvai Subchronozone.

4. Placing the Pliocene-Pleistocene boundary at the 2.48-Ma Gauss-Matuyama Chronozone boundary will accommodate the two classic concepts of the Quaternary, of it being "The Great Ice Age" and also that it is "The Age of Man" (the earliest humanoids evolved near this time).

Some significant revisions in Pleistocene terminology

Much of the "classical" terminology for designating the age of Pleistocene deposits is now revised throughout the Northern Hemisphere (Sibrava and others, 1986). Geologists in the United States should note the recommendations that terms such as Yarmouth(ian), Kansan, Afton(ian), and Nebraskan be abandoned (Richmond and Fullerton, 1986, p. 6–7, 183–184) because they have been widely misused as chronostratigraphic names. These names were originally based on litho- and pedostratigraphic units, but they oversimplify a complex stratigraphic record and have led to much miscorrelation of units. Other classical terms, including Sangamon, Illinoian, and parts of the Wisconsin are more narrowly redefined; Richmond and Fullerton (1986) also recommend that the names Wisconsinan and Sangamonian (which have chronostratigraphic connotations) also be abandoned.

REFERENCES CITED

Aguirre, E., and Pasini, G., 1985, The Pliocene-Pleistocene boundary: Episodes, v. 8, p. 116–120.

Arrias and Bonnadona, 1987: Ottawa, INQUA 12th Congress Abstracts, p. 67.

Backman, J., 1979, Pliocene biostratigraphy of DSDP Sites 111 and 116 from the North Atlantic Ocean and the age of the Northern Hemisphere glaciation: Stockholm Contributions to Geology, v. 32, p. 115–137.

Bard, E., Hamelin, B., Lao, Y., Anderson, R. F., and Fairbanks, R. G., 1989, Dating of the last interglacial period by U/Th mass spectrometry of Caribbean corals [abs.]: International Correlation Program 3 Meeting 1989, Cambridge, United Kingdom.

Berger, A. L., 1984, Milankovitch theory and climate: Reviews of Geophysics, v. 26, p. 624–657.

Davis, P., Smith, J., Kukla, G. J., and Opdyke, N. D., 1977, Paleomagnetic study at a nuclear power plant site near Bakersville, California: Quaternary Research, v. 7, p. 380–397.

Eardley, A. J., and 6 others, 1973, Lake cycles in the Bonneville basin, Utah: Geological Society of America Bulletin, v. 84, p. 211–216.

Eldholm, O., and Thiede, J., eds., 1987, Proceedings of the Ocean Drilling Program, Part A (Leg 104): Washington, D.C., U.S. Government Printing Office, 783 p.

Emiliani, C., 1955, Pleistocene paleotemperatures: Journal of Geology, v. 63, p. 538–578.

——, 1967, The Pleistocene record of the Atlantic and Pacific Ocean sediments, correlations with the Alaskan stages by absolute dating, and the age of the last reversal of the geomagnetic field: Progress in Oceanography, v. 4, p. 219–224.

——, 1970, Pleistocene paleotemperatures: Science, v. 168, p. 822–825.

——, 1972, Quaternary hypsithermals: Quaternary Research, v. 2, p. 270–273.

Fairbridge, R. W., 1962, World sea level and climatic changes: Quaternaria, v. 6, p. 111–134.

—— , 1972, Climatology of a glacial cycle: Quaternary Research, v. 2, no. 2, p. 283–302.

Fenneman, N. M., 1930, Physical divisions of the United States: U.S. Geological Survey Map, scale 1:7,000,000.

Fink, J., and Kukla, 1977, Pleistocene climates in central Europe; At least 17 interglacials afer the Olduvai event: Quaternary Research, v. 7, p. 363–371.

Flint, R. F., 1957, Glacial and Pleistocene geology: New : York, John Wiley and Sons, 553 p.

—— , 1971, Glacial and Quaternary geology, New York, John Wiley and Sons, 892 p.

Gilbert, G. K., 1890, Lake Bonneville: U.S. Geological Survey Monograph 1, 438 p.

Graf, W. L., ed., 1987, Geomorphic systems of North America: Boulder, Colorado, Geological Society of America, Centennial Special Volume 2, 643 p.

Hays, J. D., Imbrie, J., and Shackleton, N. J., 1976, Variations in the Earth's orbit; Pacemaker of the ice ages: Science, v. 194, p. 1121–1132.

Hays, J. D., Saito, T., Opdyke, N. D., and Burckle, L. H., 1969, Plio-Pleistocene sediments of the equatorial Pacific; their paleomagnetic, biostratigraphic, and climatic record: Geological Society of America Bulletin, v. 80, p. 1481–1514.

Hopkins, D. M., 1975, Time-stratigraphic nomenclature for the Holocene Epoch: Geology, v. 3, p. 10.

Imbrie, J., and 8 others, 1984, The orbital theory of Pleistocene climate; Support from a revised chronology of the marine delta ^{18}O record, *in* Berger, A. L., and others, eds., Milankovitch and climate, Part 1: Dordrecht, D. Reidel Publishing, p. 269–305.

Izett, G. A., Obradovich, J. D., and Mehnert, H. H., 1988, The Bishop ash bed (middle Pleistocene) and some older (Pliocene and Pleistocene) chemically and mineralogically similar ash beds in California, Nevada, and Utah: U.S. Geological Survey Bulletin 1675, 37 p.

Johnson, R. G., 1982, Brunhes-Matuyama magnetic reversal dated at 790.00 yr BP by marine-astronomical correlations: Quaternary Research, v. 17, p. 135–147.

Kukla, G. J., 1970, Correlations between loesses and deep-sea sediments: Geologiska Foreningen i Stockholm Forhandlung, v. 92, pt. 2, p. 148–180.

—— , 1975, Loess stratigraphy of central Europe, *in* Butzer, K., and Isaac, G. L., eds., After the Australopithecines: The Hague, Mouton, p. 99–188.

—— , 1977, Pleistocene land-sea correlations: Earth-Science Reviews, v. 13, p. 307–374.

—— , 1987, Loess stratigraphy in central China: Quaternary Science Reviews, v. 6, p. 191–219.

—— , 1989, Long continental records of climate; An introduction: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 72, p. 1–9.

Kukla, G., and An, Z., 1989, Loess stratigraphy in central China: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 72, p. 203–225.

Kukla, G., and 5 others, 1988, Pleistocene climates in China by magnetic susceptibility: Geology, v. 16, p. 811–814.

Lyell, C., 1839, Nouveaux elements de geologie: Paris, Pitois-Levrault, 648 p.

Mankinen, E. A., and Dalrymple, G. B., 1979, Revised geomagnetic polarity time scale for the interval 0–5 m.y. B.P.: Journal of Geophysical Research, v. 84, no. B2, p. 615–626.

Matthews, R. K., and Poore, R. Z., 1981, Tertiary delta ^{18}O record and glacio-eustatic sea-level fluctuations: Geology, v. 8, p. 501–504.

Mix, A. C., 1987, The oxygen-isotope record of glaciation, *in* Ruddiman, W. F., and Wright, H. E., Jr., eds., North America and adjacent oceans during the last deglaciation: Boulder, Colorado, Geological Society of America, The Geology of North America, v. K-3, p. 111–135.

Morrison, R. B., 1968, Means of time-stratigraphic division and long-distance correlation of Quaternary successions, *in* Morrison, R. B., and Wright, H. E., Jr., eds., Means of correlation of Quaternary successions; Proceedings INQUA 7th Congress, v. 8: Salt Lake City, University of Utah Press, p. 1–113.

—— , 1978, Quaternary soil stratigraphy; Concepts, methods, and problems, *in* Mahaney, W. C., ed., Quaternary soils: Norwich, United Kingdom, Geoabstracts, p. 77–108.

—— , 1987, Long-term perspective; Changing rates and types of Quaternary surficial processes; Erosion-deposition-stability cycles, *in* Graf, W. R., ed., Geomorphic systems of North America: Boulder, Colorado, Geological Society of America, Centennial Special Volume 2, p. 167–176.

Odin, G. S., ed., 1982, Numerical dating in stratigraphy, 2 vols.: New York, John Wiley and Sons, 1,040 p.

Richmond, G. M., and Fullerton, D. S., 1986, eds., Quaternary glaciations in the United States of America, *in* Sibrava, V., Bowen, D. Q., and Richmond, G. M., eds., Quaternary glaciations in the Northern Hemisphere: Oxford, Pergamon Press, p. 3–200.

Ruddiman, W. F., and Kidd, R., eds., 1986, Initial reports of the Deep Sea Drilling Project: Washington, D.C., U.S. Government Printing Office, v. 94, 1261 p.

Ruddiman, W. F., and Wright, H. E., Jr., 1987a, Introduction, *in* Ruddiman, W. F., and Wright, H. E., Jr., eds., North America and adjacent oceans during the last deglaciation: Boulder, Colorado, Geological Society of America, The Geology of North America, v. K-3, p. 1–12.

—— , eds., 1987b, North America and adjacent oceans during the last deglaciation: Boulder, Colorado, Geological Society of America, The Geology of North America, v. K-3, 501 p.

Shackleton, N. J., 1969, The last interglacial in the marine and terrestrial records: Royal Society of London Proceedings, series B, no. 174, p. 135–154.

Shackleton, N. J. and Opdyke, N. D., 1973, Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V27-238; oxygen isotope temperatures and ice volumes in a 10^5 and 10^6 year scale: Quaternary Research, v. 3, p. 39–55.

—— , 1976, Oxygen-isotope and paleomagnetic stratigraphy of Pacific Core V28-239; Late Pliocene to latest Pleistocene, *in* Cline, R. M., and Hays, J. D., eds., Investigation of Late Quaternary paleoceanography and paleoclimatology: Geological Society of America Memoir 145, p. 449–464.

Shackleton, N. J., and 16 others, 1984, Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region: Nature, v. 307, p. 216–219.

Sibrava, V., Bowen, D. Q., and Richmond, G. M., eds., 1986, Quaternary glaciations in the Northern Hemisphere; Quaternary Science Reviews, v. 5: Oxford, Pergamon Press, 511 p.

Slemmons, D. B., Engdahl, E. R., Blackwell, D., and Zoback, M. D., eds., 1990, Neotectonics of North America: Boulder, Colorado, Geological Society of America, Decade of North American Geology Map Volume (in press).

Srivastava, S., and Arthur, M. A., 1987, Proceedings of the Ocean Drilling program, Part A (Leg 105): Washington, D.C., U.S. Government Printing Office, 917 p.

Willman, H. B., and Frye, J. C., 1970, Pleistocene stratigraphy of Illinois: Illinois Geological Survey Bulletin 94, 204 p.

Wright, H. E., Jr., and Frey, D. G., 1965, The Quaternary of the United States: Princeton, New Jersey, Princeton University Press, 922 p.

Wright, H. E., Jr., and Porter, S. C., eds., 1984, Late Quaternary environments of the United States, 2 vols.: Minneapolis, University of Minnesota Press, v. 1, 407 p., v. 2, 277 p.

Zimmerman, H., and others, 1985, History of Plio-Pleistocene climate in the northeast Atlantic, Deep sea Drilling Project Hole 552A, *in* Roberts, D., and others, eds., Initial reports of the Deep sea Drilling Project: Washington, D.C., U.S. Government Printing Office, v. 81, p. 861–875.

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