The Brunhes Epoch: Isotopic Paleotemperatures and Geochronology

Abstract. *Oxygen isotopic analysis of a long piston core from the western equatorial Pacific has produced a record for the entire Brunhes epoch. This record can be correlated point by point with the isotopic records of previously analyzed Atlantic and Caribbean cores, leading to the construction of a generalized temperature curve for the entire Brunhes epoch.*

Oxygen isotopic analysis provides an accurate and detailed method for assessing the climatic evolution of the geological past. Analytical work on numerous cores from the Atlantic and the Caribbean, and noise minimization by accurate intercore correlation, has produced a well-established oxygen isotopic curve for the past several hundred thousand years (*1–3*). This record has been extended past the base of the Brunhes magnetic epoch by using core V28-238 from the western equatorial Pacific (*4*). Agreement between the record of this Pacific core and the record obtained from Atlantic and Caribbean cores is exceptional (Fig. 1), particularly in view of the great geographical distance between the areas from which the two sets of data were obtained. The character and fine structure of the isotopic record evidently permits extremely refined long-distance correlations.

The following stratigraphic markers have been used for intercore correlations: (i) Matuyama-Brunhes boundary, stage 19, near carbonate minimum B17 of Hays *et al.* (*5*); (ii) extinction of *Pseudoemiliania lacunosa*, stage 13/12 catathermal (*4, 6*); (iii) extinction of *Globoquadrina pseudofoliata*, stage 7, Pacific (*7*); (iv) speciation of *Emiliania huxleyi*, stage 7; (v) local extinction of *Globorotalia menardii flexuosa*, top of stage 5 in the Atlantic. In addition, the coiling direction changes in *Globorotalia crassaformis* and *G. truncatulinoides* are useful in the Atlantic-Caribbean region, and the size distribution of *Coscinodiscus nodulifer* is useful in the Pacific (*8*).

The isotopic records of the cores show quasi-periodic oscillations which increase slightly in amplitude toward the more recent stages. Within each core, every minimum, except perhaps that of stage 14, reaches an almost identical isotopic value; as to the maxima, those of stages 3 and 17 are lower, and those of stages 1 and 5 are highest, with that of stage 5 to a small but significant extent the higher of the two.

Within each stage, the isotopic record displays fine structure which constitutes a real record of worldwide events. The pattern is not simply sinusoidal, sawtooth, or square wave with added noise, and any model based on any one of these alternatives is automatically invalid.

The isotopic record obtained from fossil foraminiferal shells in deep-sea cores represents the cumulative effects of temperature change and change in the isotopic composition of ocean water (*1, 9*). The latter is proportional to the quantity of water extracted during a glacial age, and to the isotopic composition of that water. For a glacio-eustatic decrease in sea level of 120 m, the following extremes may be established for the isotopic composition of seawater during glacial maxima:

1) +0.4 per mil, if the water removed had the oxygen isotopic composition of water vapor in isotopic equilibrium with the subtropical sea surface [−11 per mil (*10*)]; and

2) +1.6 per mil, if the water removed had an oxygen isotopic composition similar to that of snow in the interior of Antarctica [−50 per mil (*11*)].

Clearly, neither of these two extremes is applicable to the natural situation at the glacial maxima. Therefore, three independent methods have been used in an effort to obtain a better estimate of this important factor.

First, one may estimate the oxygen isotopic composition of ice directly, taking into account the known isotopic distribution in modern ice sheets. Olausson (*12*) estimated the isotopic composition of snow which might have formed the major Pleistocene ice sheets, making use of temperature reconstructions suggested by Flohn (*13*). Shackleton (*9*) made a three-dimensional estimate which took into account the flow within the ice sheet but did not incorporate the assumption of reduced temperatures. Dansgaard and Tauber (*14*) based their estimate on the isotopic composition of modern precipitation in formerly glaciated areas and on the measured isotopic composition of the last glacial ice still present in the core of the Greenland ice sheet. These estimates place the isotopic composition of the stored ice between −30 and −35 per mil, leading to a change of about 1.1 per mil in the oxygen isotopic composition of the ocean.

Second, the glacial/interglacial change in the oxygen isotopic composition of the ocean may be estimated by analyzing benthic foraminiferal shells from a core taken where the modern bottom temperature is so low that further lowering, as during a glacial age, would be unlikely. In this case, the only contribution to the glacial/interglacial change measured would be that of the isotopic composition of seawater (*9*). The isotopic records of benthic Foraminifera from equatorial Atlantic and equatorial Pacific cores are shown, respectively, in figure 5 in (*15*) and figure 4 in (*9*). Although Shackleton (*9*) reasoned that temperature changes in the hadal environment were probably insignificant, there is now evidence that there were in fact important changes. For instance, the relative abundance of certain benthic species in the North Atlantic core 280 changes in step with the isotopic variations (Fig. 2); and in core V28-238 there are changes in both the relative abundance of benthic species and the mean size of Uvigerinids (*4*). Glacial/interglacial changes in the benthic foraminiferal faunas of

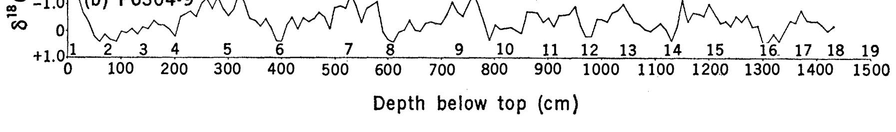

Fig. 1. Cores (a) V28-238 (western equatorial Pacific, 1°1′N, 160°29′E) and (b) P6304-9 (Caribbean sea, 14°57′N, 69°20′W); oxygen isotopic record of *Globigerinoides sacculifera* during Brunhes epoch [see (*4*) for pre-Brunhes record of core V28-238]. The horizontal scale of core V28-238 has been adjusted to make the peak of stage 5 coincide with that of core P6304-9. The numbers above the abscissa identify deep-sea core stages.

Table 1. Estimate of the glacial/interglacial change in the oxygen isotopic composition of seawater based on nonisotopic methods.

Region	Temperature change (°C)	$\delta^{18}O$ (per mil)		
		Equivalent	Observed	Difference
Caribbean	7.5	1.7	1.9	0.2
Equatorial Atlantic	5.5	1.3	1.8	0.5
Equatorial Pacific	3.5	0.8	1.1	0.3
			Average for ocean	~ 0.4

Table 2. Glacial/interglacial temperature variation in surface seawater based on estimates of the isotopic composition of glacial seawater.

Method	$\delta^{18}O$ (per mil)		Temperature change (°C)		
	Ocean	Ice	Caribbean	Atlantic	Pacific
Ice sheet reconstruction	+ 1.1	− 35	4	3	0
Benthic foraminiferal analyses					
Minimum	+ 0.5	− 16	7	6	3
Maximum	+ 1.1	− 35	4	3	0

the North Atlantic have been discussed also by Streeter (*16*). The measured isotopic change between the minimum of age 2 and today in the equatorial Pacific, as obtained from isotopic analysis of monospecific samples of benthic foraminiferal shells, amounts to about 1.4 per mil (*4*). If the present temperature of 1.5°C decreased to − 1.5°C to be in complete equilibrium with freezing surface water at − 1.8°C, the temperature component of the measured change would be 0.9 per mil, setting a lower limit of 0.5 per mil for the isotopic change of seawater. The upper limit, assuming no glacial/interglacial temperature change in the surface water of the equatorial Pacific is set at 1.1 per mil, equivalent to the total amplitude measured (*4*). A similar argument based on benthic Atlantic foraminifera sets a lower limit of 0.2 per mil and an upper one of 1.1 per mil [data from (*15*)].

Third, one may estimate the glacial/interglacial temperature change in the surface water by nonisotopic means, and then use this estimate in conjunction with the isotopic data from planktonic foraminifera to estimate the isotopic change of the seawater. Using this method, Emiliani (*17*) estimated a glacial/interglacial temperature change of 7° to 8°C for the Caribbean, 5° to 6°C for the equatorial Atlantic, and 3° to 4°C for the Pacific. These figures lead (Table 1) to an estimate of about 0.4 per mil for the glacial/interglacial change in the oxygen isotopic composition of seawater. Table 2 summarizes the results of the three approaches just discussed.

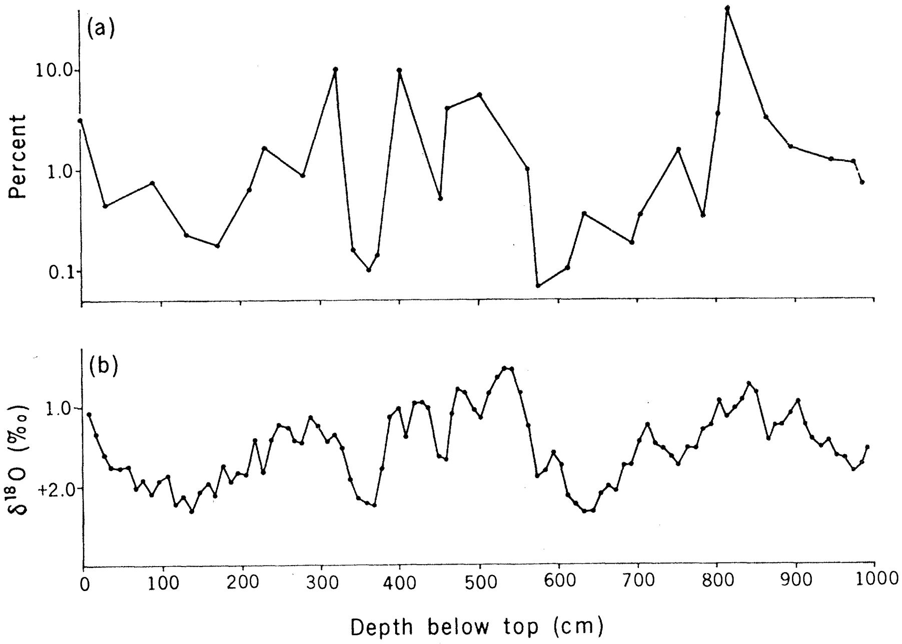

Fig. 2. Core 280 (North Atlantic, 34°57′N, 44°16′W). (a) Ratio of *Epistominella umbonifera* to *E. exigua*; (b) oxygen isotopic record of *Globorotalia inflata*; the data are from (*30*) and (*15*).

The first absolute age measurements on deep-sea cores were made by Piggott and Urry (*18*) who used the excess ^{230}Th, measured through its daughter product ^{226}Ra, to estimate the age of Caribbean cores and of the paleoenvironmental changes described by Cushman (*19*). By this method, the maximum of what is now known to be stage 5 was found to have an age of about 100,000 years. Following the initial ^{18}O work on deep-sea cores, Rubin and Suess (*20*) used the ^{14}C method to date the minimum of stage 2 at about 22,000 years. Emiliani (*1*) extrapolated this age to obtain an age of about 100,000 years for the maximum of stage 5. The $^{230}Th/^{231}Pa$ method was applied to several Atlantic and Caribbean cores (*21, 22*) and again an age of about 100,000 years was found for the peak of stage 5, although ages up to 25 percent higher were found in two cores (*21, 23*). While the ages are approximately correct, the universal validity of the $^{230}Th/^{231}Pa$ method remains unproved. As a result, the oxygen isotopic record of the deep-sea sediments may be related with confidence to absolute ages only within the range of the ^{14}C method and at the Brunhes/Matuyama boundary (700,000 years ago, stage 19).

Interpolation between 0 and 700,000 years ago for core V28-238 yields an average rate of accumulation of 1.74 cm per 10^3 years, and an age of 120,000 years for the temperature peak of stage 5. This age compares favorably with that obtained by extrapolation from the top of the core, assuming that the temperature minimum of stage 2 at 35 cm has an age of 20,000 years. It also compares favorably with the age of 120,000 years for the fossil coral reefs dated by various authors (*23, 24*) and interpreted as representing high stands of sea level. Chronological coincidence between high sea levels and maxima in the oxygen isotopic curve is to be expected (*25*) although perhaps not strictly necessary.

It has been emphasized many times [for example, see (*2, 3*)] that the validity of deductions on paleoenvironmental changes based on deep-sea sediments must rest on a proof of the continuity of the sedimentary record of the deep-sea floor. Continuity for the Caribbean and Atlantic cores has been demonstrated (*2*). Continuity for the Pacific core V28-238 is virtually proved by the extremely close correlation with the Caribbean and Atlantic cores (Fig. 1) in terms of the oxygen isotopic records. Furthermore, the same number of cycles may be seen in the isotopic record of core V28-238 and in the carbonate percentage records of the Pacific cores described by Hays *et al.* (*5*) and Arrhenius (*26*) covering the entire Brunhes epoch. It is therefore possible to construct an interpolated time scale for stages 7 to 19 of greater reliability than that previously available (*2*), which was based on extrapolation alone.

Figure 3 displays the previously published generalized paleotemperature curve [figure 3 in (*3*)] together with four additional time scales which may be constructed in the light of the new ^{18}O evidence from Pacific core V28-238. Time scale A (*2*), which was based on extrapolation, can now be improved. Time scales B and C represent modifications of time scale A incorporating the additional information that the Matuyama/Brunhes boundary, 700,000 years old, falls within stage 19. Time scale B is simply interpolated between ages 0 and 700,000 years, assuming a uniform rate of sedimentation of 2.3 cm per 10^3 years in the Caribbean. This time scale dates the peak of stage 5 at about 150,000 years and the minimum of stage 2 at about 30,000 years, both of which ages are inconsistent with all available radiometric data. Time scale C incorporates the additional assumption that the peak of stage 5 dates from 100,000 years. This assumption is based on the $^{18}O/^{16}O$ measurements and $^{230}Th/^{234}U$ age determinations on speleothems from Orgnac, France (*27*) and West Virginia (*28*). These analyses appear to identify a high temperature episode at about 100,000 years ago which must correlate with the peak of stage 5 (*29*). Time scale C leads to two different rates of sedimentation in the Caribbean, 3.5 cm per 10^3 years above, and 2.1 cm per 10^3 years below the base of stage 5.

Time scales D and E in Fig. 3 are based on core V28-238. Time scale D assumes uniform sedimentation throughout the core, and implies a Caribbean sedimentation rate of 2.8 cm per 10^3 years from stage 1 to stage 5, a rather uniform rate of 2.0 cm per 10^3 years between stages 6 and 14, and a rate of 2.7 cm per 10^3 years from stage 13 to stage 19. Time scale E incorporates the additional assumption, based on the speleothem dates, that the age of the peak of stage 5 is 100,000 years, resulting in a sedimentation rate of 3.5 cm per 10^3 years from stage 1 to stage 5, a rate of 1.9 cm per 10^3 years between stages 13 and 16, and a rate of 2.7 cm per 10^3 years from stage 16 to stage 19.

Time scales D and E indicate that the sedimentation rate in the Caribbean during the early Brunhes (stages 13 to 19) was similar to that of the late Brunhes (stages 1 to 5). This is paralleled by the pattern of changing carbonate content in Pacific cores, where the carbonate percentage is consistently higher near the top of the cores (stages 2 to 6) and just above the base of the Brunhes (stages 15 to 19) (*5, 26*). The similar sedimentation rates and carbonate percentages in stages 1 to 5 and 16 to 19 are apparently the result of lower temperatures and more prolonged glacials than during the intermediate portion of the Brunhes epoch. This is well supported by the isotopic record (Fig. 1).

Time scales D and E differ by at most 20 percent in the vicinity of the temperature maximum of stage 5. The two time scales rapidly converge both above and below. The exact age of the temperature peak of stage 5 will come from correlation with dated continental deposits, because the precision of $^{230}Th/^{234}U$ ages in appropriate materials is well within 10^4 years, and the problems of biological, sedimentological, or geochemical reworking which beset deep-sea sediments are eliminated or greatly reduced. The exact age of the peak of stage 5 has ceased, in any case, to be the cornerstone for absolute dating in the time range between 10^5 and 10^6 years.

A new generalized paleotemperature curve, based on all available isotopic core records and on time scales D and E, is shown in Fig. 4. This curve covers the entire Brunhes epoch.

It is clear that the climatic record of deep-sea cores shows a quasi-periodic variation with a wavelength approaching 10^5 years. Such a wavelength is far too long to be explained on the basis of air-sea interaction or of atmospheric and oceanic resonances. It is far too short to be explained in terms of mechanisms deriving from the earth's interior (continental drift,

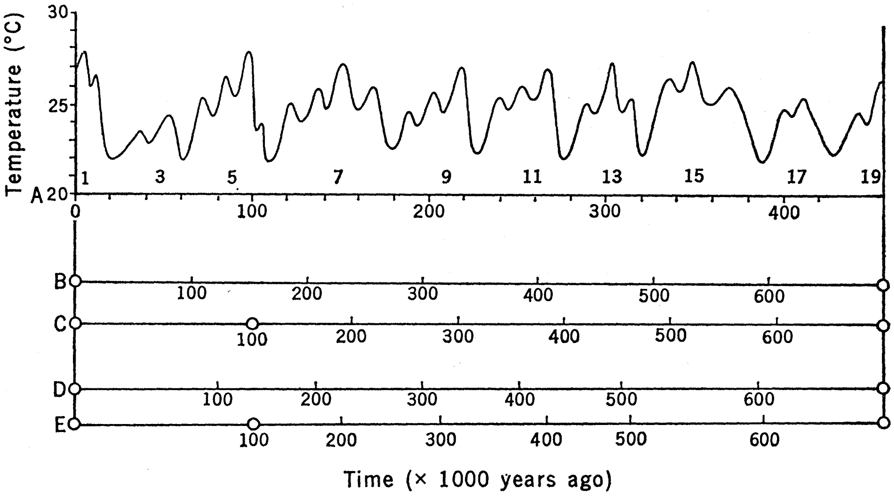

Fig. 3. Generalized paleotemperature curve of 1972 (*2*) with four possible time scales (B to E). The numbers above the abscissa for curve A identify deep-sea core stages.

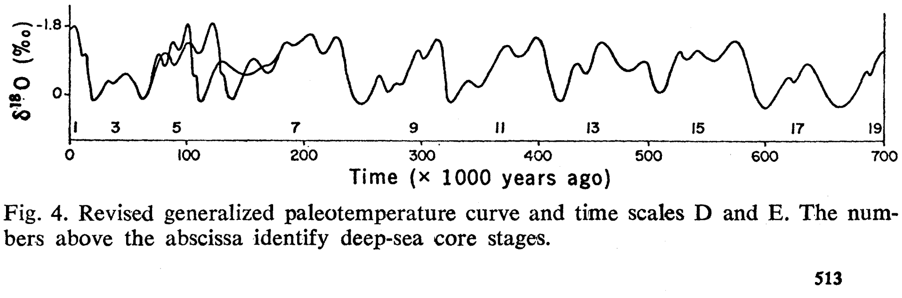

Fig. 4. Revised generalized paleotemperature curve and time scales D and E. The numbers above the abscissa identify deep-sea core stages.

mountain building, and so forth), nor has it yet been possible to derive it from solar activity. On the other hand, the wavelength in question is of the right order of magnitude to be related to the basic astronomical motions of the earth. Now that a reliable time scale for the entire Brunhes epoch has been established, it should be possible to determine appropriate weightings for the various factors involved, so that the hypothesis that the astronomical motions of the earth are the basic cause of climatic change during the Pleistocene may be tested with more rigor than hitherto possible.

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References and Notes

1. C. Emiliani, *J. Geol.* **63**, 538 (1955).
2. ———, *ibid.* **74**, 109 (1966).
3. -→—, *Science* **178**, 398 (1972).
→ N. J. Shackleton and N. D. Opdyke, *Quat. Res.* **3**, 39 (1973).
5. J. D. Hays, T. Saito, N. D. Opdyke, L. H. Burckle, *Geol. Soc. Amer. Bull.* **80**, 1481 (1969).
6. S. Gartner, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **12**, 169 (1972).
7. T. Saito and P. R. Thompson, personal communication.
8. C. Emiliani, *Science* **154**, 851 (1966).
→ N. Shackleton, *Nature* **215**, 15 (1967).
10. H. Craig and L. I. Gordon, *Mar. Geochem. Proc. Symp. Univ. R.I. Narragansett Mar. Lab. Occas. Publ. No. 3-1965* (1966), p. 277.
11. W. Dansgaard, *Tellus* **16**, 436 (1964).
→ E. Olausson, *Progr. Oceanogr.* **3**, 221 (1965).
13. H. Flohn, personal communication to E. Olausson.
14. W. Dansgaard and H. Tauber, *Science* **166**, 499 (1969).
15. C. Emiliani, *J. Geol.* **66**, 264 (1958).
→ S. S. Streeter, *Quat. Res.* **3**, 131 (1973).
17. C. Emiliani, *Science* **168**, 822 (1970).
18. C. S. Piggott and W. D. Urry, *Amer. J. Sci.* **240**, 1 (1942).
19. J. A. Cushman, *ibid.* **239**, 128 (1941).
20. M. Rubin and H. E. Suess, *Science* **121**, 481 (1955).
21. J. N. Rosholt, C. Emiliani, J. Geiss, F. F. Koczy, P. J. Wangersky, *J. Geol.* **69**, 162 (1961); *J. Geophys. Res.* **67**, 2907 (1962).
22. E. Rona and C. Emiliani, *Science* **163**, 66 (1969).
→ W. D. Broecker, D. L. Thurber, J. Goddard, T.-L. Ku, R. K. Matthews, K. J. Mesolella, *ibid.* **159**, 297 (1968).
24. H. H. Veeh, *J. Geophys. Res.* **71**, 3379 (1966); ——— and J. Chappell, *Science* **167**, 862 (1970).
25. C. Emiliani, *Science* **166**, 1503 (19→ N. J. Shackleton, *Proc. R. Soc. Lond. Ser. B Biol. Sci.* **174**, 135 (1969).
26. G. Arrhenius, *Rep. Swed. Deep-Sea Exped. 1947–1948* **5**, 1 (1952).
27. J. C. Duplessy, J. Labeyrie, C. Lalou, H. V. Nguyen, *Quat. Res.* **1**, 162 (1970).
28. P. Thompson, H. P. Schwarcz, D. C. Ford, *Science*, in press.
→ C. Emiliani, *ibid.* **171**, 571 (1971); N. J. Shackleton, *Geol. Soc. Lond. Spec. Publ.* **5** (1971), item 390, p. 106.
30. F. B. Phleger, F. L. Parker, J. F. Peirson, *Rep. Swed. Deep-Sea Exped. 1947–1948* **7**, 1 (1953).
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Transvaal Stromatolite: First Evidence for the Diversification of Cells about 2.2 × 10⁹ Years Ago

Abstract. *The well-preserved fossil remnants of filamentous blue-green algae have been found in petrographic thin sections of a dolomitic limestone stromatolite in the Transvaal Sequence of South Africa. Some of these filaments contain enlarged cells which are interpreted as akinetes. A new species and genus,* Petraphera vivescenticula, *is proposed for this microfossil, which is morphologically similar to the living cyanophyte genus* Raphidiopsis. *This would constitute the first known occurrence of cell diversification in the Precambrian with an age of about* 2.2×10^9 *years.*

Evidence is presented for the oldest known occurrence of the diversification of cells in the Precambrian with an age of approximately 2.2×10^9 years. Some of the microfossils are morphologically similar to certain recent cyanophytes.

A geological milestone in the study of early life-forms is the appearance of stromatolites (*1*). The oldest known stromatolite structures occur in a limestone of the Bulawayan Group in Southern Rhodesia (*2, 3*). It is assumed that this stromatolite was formed by algal activity although microscopic examination has revealed

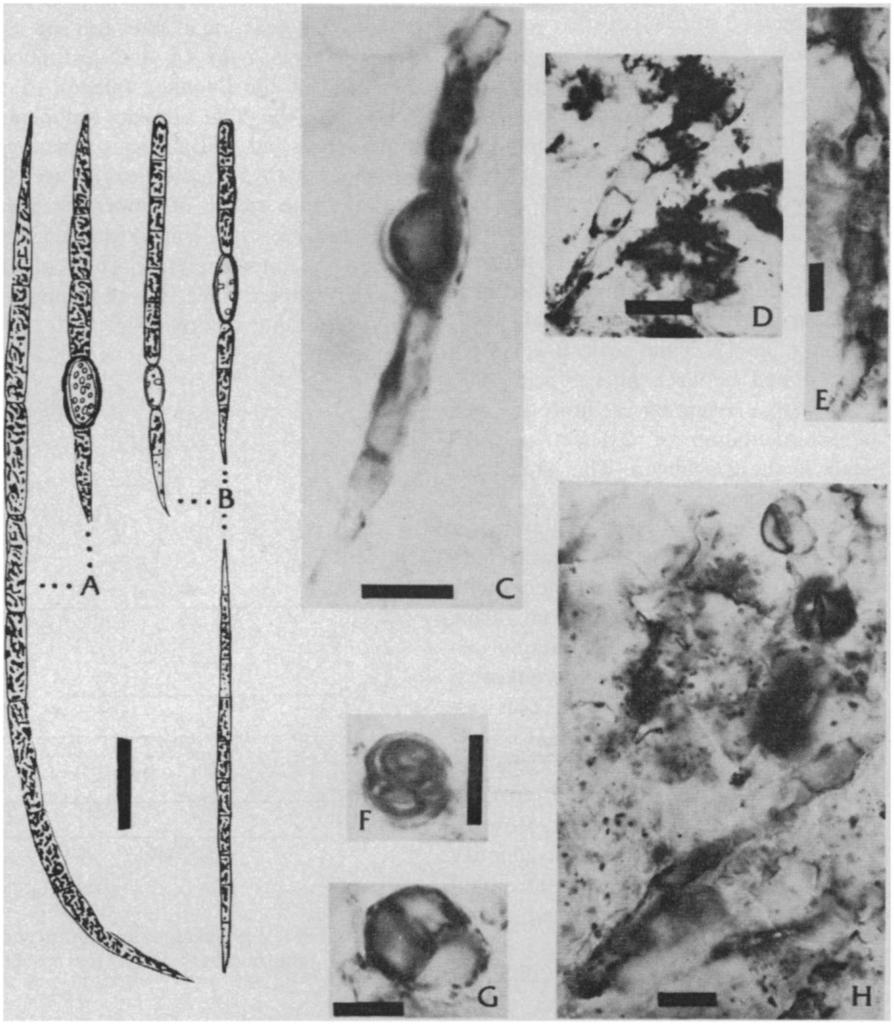

Fig. 1. (A and B) Comparison drawings modified from Desikachary (*17*) of the recent blue-green algae *Raphidiopsis mediterranea* and *Raphidiopsis indica*, respectively; scale bar, 10 μm. (C) New microfossil, *Petraphera vivescenticula*, showing enlarged akinete-like structure; scale bar, 10 μm. At the completion of this investigation the slide (No. TR1A) containing the type specimen will be deposited in the algal herbarium collection of the Academy of Natural Sciences of Philadelphia. (The specimen is marked with a diamond scribe.) (D and E) Broken filamentous forms; scale bar, 10 μm. (F and G) Simple round forms which could represent coccoid blue-green algae; scale bar, 5 μm. (H) Assemblage of a broken filament and round forms; scale bar, 10 μm. All the Transvaal photomicrographs are from petrographic thin sections.