Department of the Interior
U.S. Geological Survey

Pliocene Climates: Scenario for Global Warming

Abstracts from USGS Workshop, Denver, Colorado, October 23 - 25, 1989

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Open-File Report 90-64

Introduction: USGS Workshop on Pliocene Climates

The U.S. Geological Survey (USGS) held a workshop on Pliocene climates in Denver, Colorado on October 23-25, 1989. The workshop brought together members of the USGS who are working on a long-term project to understand and map Pliocene climates and environments of the Northern Hemisphere with interested collaborators from the USSR, The Geological Survey of Canada, the National Center for Atmospheric Research, and the Institute for Arctic and Alpine Research, University of Colorado. Paleoclimate researchers from the USSR attended the workshop as part of an exchange and cooperative study of Pliocene climates that is organized under the auspices of Working Group VIII of the US-USSR Bilateral Agreement on Protection of the Environment.

The purposes of the workshop were to: 1) report and critique progress on individual research tasks during fiscal year 1989 (FY 1989); 2) exchange information and discuss methods for correlation, analysis and interpretation of proxy climate data, and data management; 3) plan and prioritize the group's research tasks for FY 1990; 4) continue dialogue with numerical climate modelers; and 5) expand efforts to involve the participation of researchers outside the USGS who can contribute to the Pliocene climate reconstruction.

The USGS integrated study of Pliocene climates began in FY 1988 as a research objective of the USGS Climate Change Program. During FY 1989, the USGS Pliocene research project became an element of the USGS contribution to the U.S. Federal Global Change Research Program.

The Pliocene Epoch is of particular interest in climate history because Pliocene deposits record the transition from relatively warm global climates, when glaciers and sea ice were absent or greatly reduced in the Northern Hemisphere, to the generally cooler climates of the Pleistocene with prominent high-frequency glacial-interglacial cycles and perennial ice-cover in the Arctic Ocean. Pliocene flora and fauna are very similar to modern flora and fauna, and Pliocene deposits are widespread and easily accessible in both continental and marine settings. In addition, most Pliocene sediments have undergone little diagenetic alteration. All these attributes facilitate quantitative estimates of environmental information and the development of regional and global patterns of climate data.

Better understanding of Pliocene climates, their evolution, and rates and causes of change will provide important clues to future earth systems changes and impacts that will occur as a result of a greenhouse-effect global warming. A major goal of the USGS Pliocene Project is to produce a synoptic map or "snapshot" of climate parameters during a time in the Pliocene representing conditions that are significantly warmer than modern climates. The Pliocene "Scenario for Global Warming" will provide a means for testing and validating results of general circulation models (GCMs) which attempt to model global warming.

Testing GCM results is crucial because these results are being used to predict future climate changes due to increasing atmospheric concentrations of CO_2 and other greenhouse gasses. Important questions to consider include: 1) are models capable of simulating warmer climates under different boundary conditions; 2) are regional responses to warming similar for all warm intervals, regardless of the boundary conditions or the cause of warming; and 3) what was the impact of substantial global warming on the biosphere during the Pliocene? Answering these questions will improve our ability to anticipate the effects and consequences of future warming.

A great deal of progress towards achieving the objective of a Pliocene paleoclimatic reconstruction for the Northern Hemisphere has been made as evidenced by results summarized in the following contributions. For example, several techniques have been developed to estimate sea-surface temperatures and the temperatures of inner shelf waters of Pliocene seas. In addition, progress has been made in quantifying temperature and precipitation ranges in terrestrial settings on the basis of fossil leaf assemblages and in the dating and correlation of Pliocene localities in the critical Arctic basin. Important work has begun on outlining the role of the carbon cycle in Pliocene climatic changes.

The accompanying map shows the locations of the studies included in this report. A list of workshop participants is in Appendix 1.

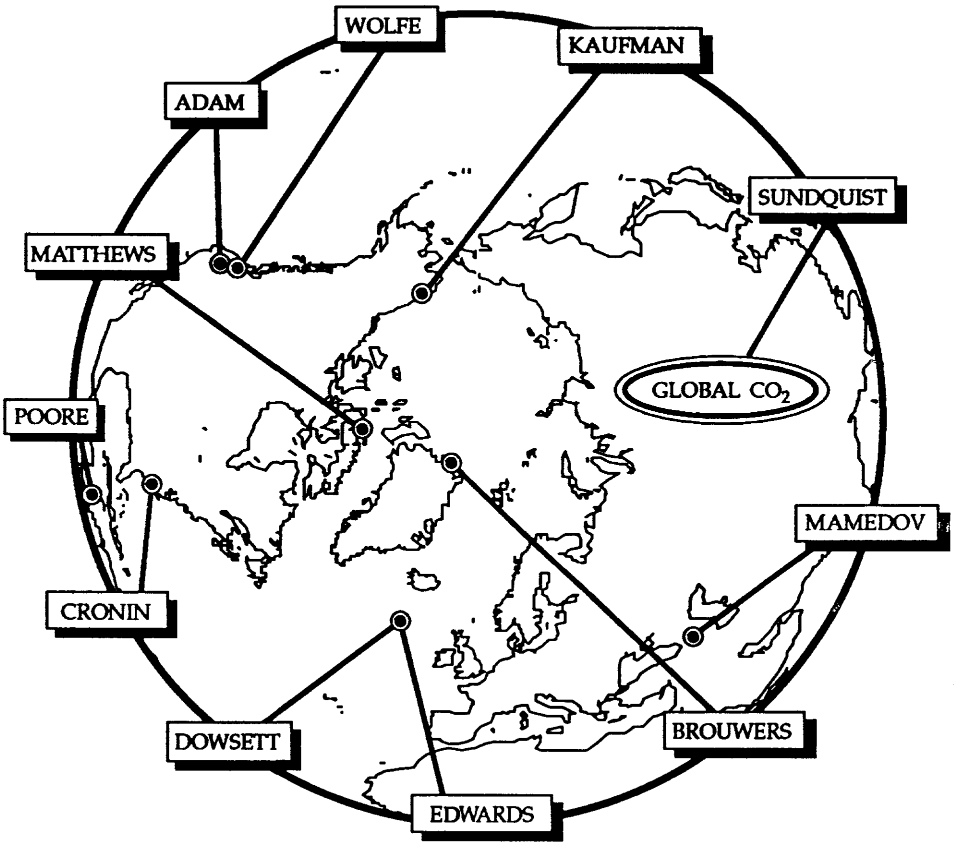

Figure 1. General location of Pliocene research sites discussed in this volume.

Pliocene Environmental Changes At Tulelake, Siskiyou County, California, From 3 to 2 Ma

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Diatoms and pollen grains deposited during the interval from 3 to 2 Ma at Tulelake, Siskiyou County, California, are well preserved, and their biostratigraphy documents limnological and environmental changes in this region. The entire 334m long sediment record is dated by paleomagnetic reversals and tephrochronology, and includes a continuous lacustrine stratigraphic record between 3 and 2 Ma. Changes in the relative abundance of pine pollen vs. pollen of the Taxodiaceae, Cupressaceae and Taxaceae (TCT) group reflect changes in the composition of the regional forest through time; proportional shifts between planktonic diatoms (*Aulacoseira solida* and large species of *Stephanodiscus*) and benthic diatoms (*Fragilaria* species) from the core can be interpreted in terms of past changes in lake depth, open water turbulence and seasonal thermal structure. These parameters are potentially under climatic control, and thereby may provide terrestrial paleoclimatic insights for this time interval that can be compared with data from marine records.

The Tulelake Basin (lat. 41°57.27'N, long. 121°28.23'W, elev. 1229m), which is part of the Modoc Plateau, is underlain by Tertiary and Quaternary volcanic rocks broken by prominent normal faults that strike mostly north and north-northwest. It is impounded by normal fault scarps to the west, by the Devils Garden and Medicine Lake volcanic fields to the southeast and south, by normal scarps to the east, and by irregular highlands to the north. Both volcanism and tectonism are active in the region today.

The Pliocene deposits consist of interbedded lake sediment and tephra. The lake sediment includes olive gray algal muds and marly units that are primarily massive but sometimes laminated, as well as darker gray to black silts and clays. The chronology for the Pliocene record is based on linear interpolation between six paleomagnetic data points. The sedimentation rate established for the Gauss Normal-Polarity Chron is in general agreement with rates observed higher in the section, but sedimentation rates varied greatly during the interval from 2.0 to 2.5 Ma. In particular, sedimentation was very slow near the base of the Matuyama Reversed-Polarity Chron (2.12 to 2.48 Ma; 219 to 229.68m depth). The low sedimentation rate inferred from the paleomagnetic data is corroborated by the lithology of the section, which shows a platy fracture pattern between 221 and 234m depth that is not present above or below that interval.

Within the Pliocene, pine and TCT pollen show a well-developed inverse relation, and the highest TCT frequencies commonly exceed 50% (fig. 1). By contrast, in the more recent part of the section the highest TCT frequencies are significantly lower, and other pollen types, particularly *Artemisia* (sagebrush), become more important. The strong inverse relationship between pine and TCT pollen evident in the Pliocene Tulelake deposits differs from the pattern at Tulelake during the past 1 Ma, but resembles patterns observed at Hodgdon Ranch in the Sierra Nevada (Adam, 1967; Latitude 38°N), and the cooler parts of the pollen record from Clear Lake, Lake County, California (Adam, 1988; latitude 39°N). We infer that regional environmental conditions at Tulelake (latitude 42°N) between about 2 and 3 Ma resembled conditions on the lower western slope of the Sierra today and around Clear Lake during the cooler parts of the last glacial cycle, rather than modern conditions near Tulelake. We interpret the pine and TCT curves in terms of a model in which the ratio of TCT to pine pollen increases with increasing temperature. According to this model, a significant warm interval occurred at Tulelake between about 2.8 and 2.6 Ma (fig. 1).

Four major diatom zones characterize the Tulelake paleolimnological record between 3 and 2 Ma: a basal zone, (3.0 to 2.9 Ma), dominated by *Stephanodiscus*, a lower intermediate zone, (2.9 to 2.65 Ma), characterized by *Aulacoseira solida* (= *Melosira solida* Eulenstein), an upper intermediate zone, (2.65 to 2.4 Ma), with abundant *Fragilaria*, and another *A. solida* zone, (2.4 to 2 Ma), that is interrupted by significant percentages of *Cyclotella bodanica* between 215 and 210m depth (estimated to date between 2.03 and 2.08 Ma). *Aulacoseira solida* returns to dominance by 2.02 Ma, but at the end of the Pliocene *Fragilaria* species become dominant once more.

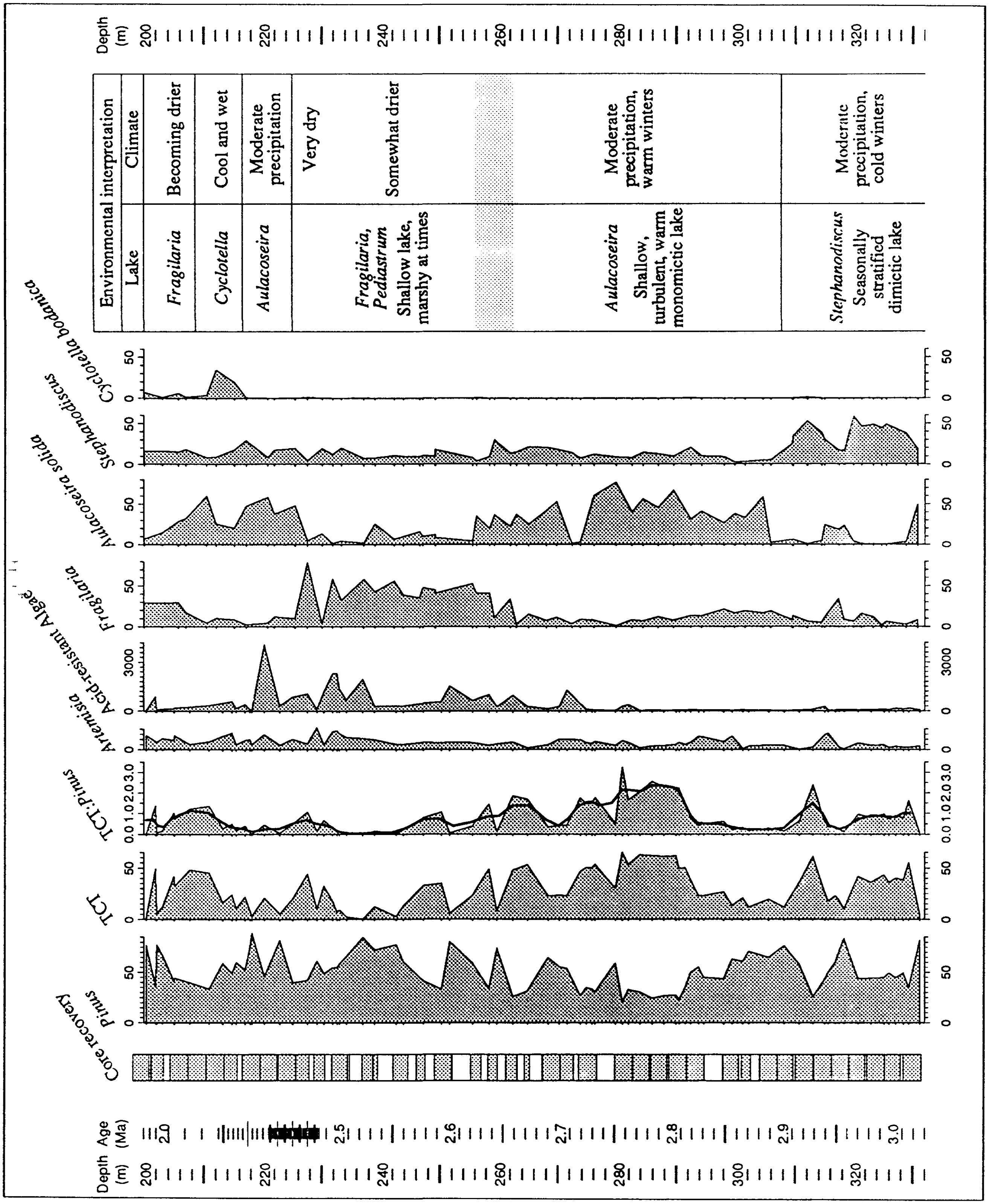

Figure 1. Selected variables plotted against depth and inferred age for the 3 to 2 Ma part of the Tulelake section. Acid-resistant algae curve is derived from pollen counts and expressed as percent of total pollen; diatom percentages are expressed as percent of total diatoms. Smoothed curve shown for TCT:Pinus is calculated as a weighted three-level moving average, with both TCT:Pinus and depth values smoothed.

Comparison of the pollen and diatom records for the Pliocene part of the Tulelake section indicate general agreement that a relatively warm interval occurred between about 2.9 and 2.6 Ma. The lake in the basin was fairly deep and monomictic, so winter temperatures must have been much milder than at present. The proposed modern and Pleistocene analogs for the fossil pollen spectra lie within a Mediterranean climate, several degrees of latitude south of Tulelake, and above the low-elevation band of high oak (*Quercus*) pollen frequencies. The general lack of spruce (*Picea*) pollen in the pollen spectra is also consistent with a Mediterranean climate, and indicates that summers were dry.

The warm interval between 3.2 and 2.4 Ma proposed by Loubere and Moss (1986; fig. 2) for DSDP Site 548 in the North Atlantic, based on $\delta^{18}O$ analyses of the *Globorotalia inflata* group, also affected the limnology of Tulelake and the regional forest composition of the surrounding area, as inferred from pollen and diatom stratigraphy. Although Loubere and Moss (1986) do not present a detailed chronology for their warm interval, the brief cool interval that appears at the top of their core 25 may well correlate with the relatively cool interval at about 2.7 Ma which is found at Tulelake in both the TCT:Pine and *Aulacoseira solida* curves.

The Tulelake warm period, which lies within the Gauss Normal Polarity Chron, can be correlated on that basis with the Reuverian deposits of the Netherlands. The Reuverian B deposits record a three-part warm interval, with two thermomers separated by a cryomer (Zagwijn, 1974; Suc and Zagwijn, 1983; fig. 2), that may be correlative with the similar sequence observed at Tulelake.

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Climatic Significance of the Ostracode Fauna from the Late Pliocene Kap København Formation, North Greenland

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The Kap København Formation occurs in the narrow coastal plain of northernmost Greenland at 82°N latitude. The Kap København is divided into two members: the basal Member A consists of glaciomarine facies and the overlying Member B includes shallow marine and terrestrial facies. Member A is homogeneous in lithology, dominated by finely laminated mud with lenses of diamicton. Member B has a more heterogeneous lithology and is subdivided into three units: the basal unit B1 consists of shallow marine sands, the middle unit B2 consists of bioturbated muds, and the upper unit B3 consists of estuarine and terrestrial sands.

The age of the Kap København Formation is bracketed by two faunal events. The maximum age is based on the presence of Pacific-derived mollusks and ostracodes, which indicate that the Kap København post-dates the opening of the Bering Strait (≤ 3.0 to 3.1 Ma). The minimum age is based on the co-occurrence in unit B3 of an extinct rabbit *(Hypolagus)* and an extant hare *(Lepus)*, whose stratigraphic ranges overlap between 2.4 and 2.0 Ma.

Forty-three ostracode species were recovered from seven localities of Member B; Member A did not yield ostracodes. A principal component analysis of 25 samples, based on species content, clustered most of the samples from units B1 and B2; samples from unit B3 formed an isolated cluster. The primary cluster, mostly representing unit B2, is interpreted to represent an inner sublittoral environment. The ostracode assemblage in unit B2 from the northern outcrop belt differs from the assemblage in unit B2 from the southern outcrop belt in species diversity, abundance, and composition. These differences are inferred to represent differences in water depth and paleotemperature.

The secondary principal component cluster, comprised of samples from unit B3, contains a low diversity ostracode assemblage dominated by genera characteristic of estuarine habitats. The dominance of estuarine taxa and the absence of inner sublittoral species indicates that this part of unit B3 is a nearshore environment with reduced salinity.

Ostracode assemblages were used to estimate the annual range of sea-bottom water temperatures (SBT) for the Kap København, using three different methods. The first approach plotted the known temperature range of 12 well-documented extant species (fig. 1).

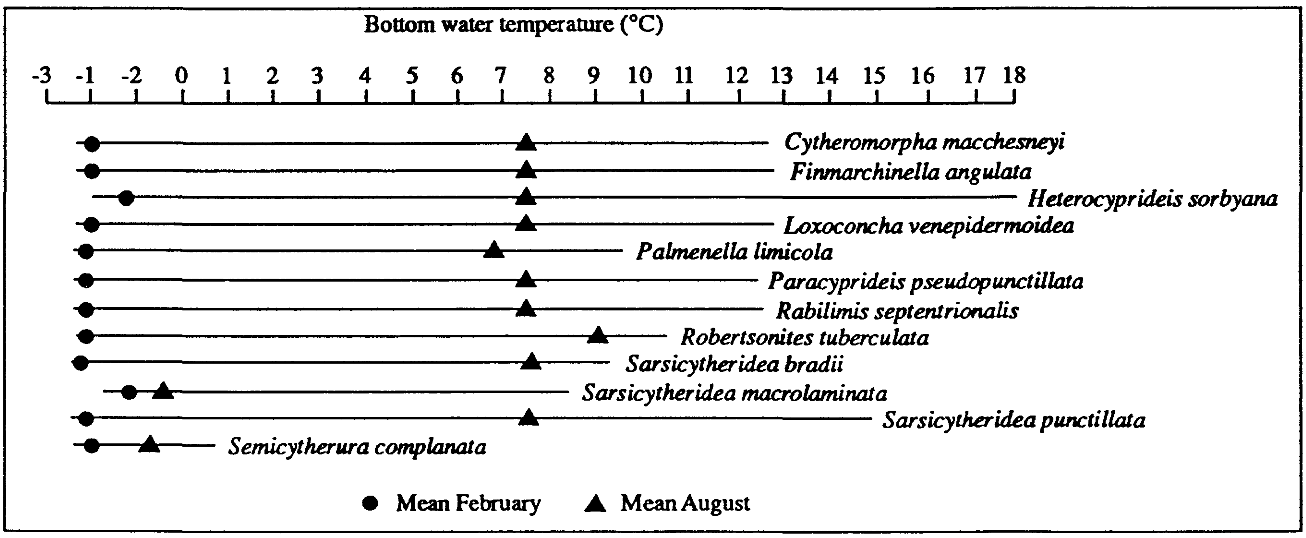

Figure 1. Mean and survival temperature range for 12 extant Kap København ostracode species based on latitudinal distribution. Solid line represents total temperature range of a species, defined by maximum and minimum survival temperatures. Circle represents average February sea-bottom temperatures (SBT); triangle represents average August SBT.

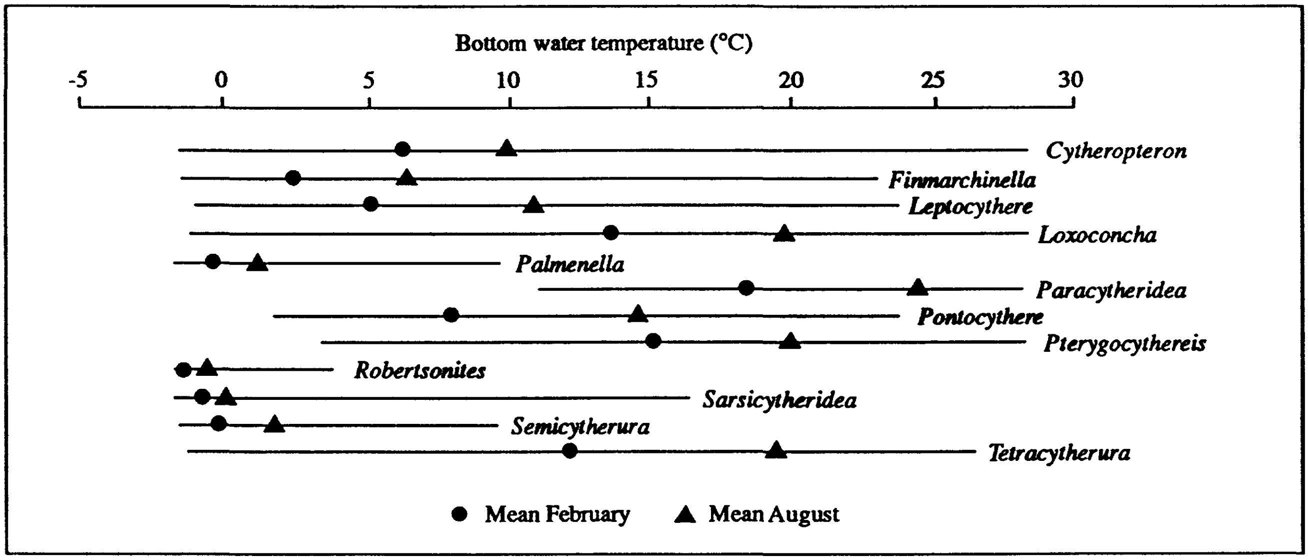

Figure 2. Temperature tolerance for 12 ostracode genera occurring in the Kap København Formation. Circle represents mean February sea-bottom temperature (SBT); triangle represents mean August SBT.

This method indicates minimum SBT of -2°C and maximum SBT of about 9°C for assemblages from units B1 and B2. The marginal marine assemblage from unit B3 suggests slightly higher (10 to 12°C) summer SBT.

The second method plotted the temperature tolerances of 12 genera, based on the modern data set of Cronin and Dowsett (in press) that was used to develop a factor analytic transfer function. The analysis at the genus level emphasizes the presence of six warmer water genera and highlights the anomalous mixture in the Kap København Fm. of arctic and temperate genera that today live in widely separated geographic-thermal regions (fig. 2). The analysis at the genus level also underscores the inferred temperature differences between the northern and southern localities of the Kap København. The genus level data indicates minimum and maximum SBT of 2 to 3°C and 9 to 10°C respectively for units B1 and B2. As with the species level analysis, the marginal marine assemblages from unit B3 suggest slightly higher summer SBT.

The third approach involves the application of the ostracode transfer function equations developed by Cronin and Dowsett (in press) to the Kap København assemblages. Based on this method, winter and summer SBT for the Kap København are estimated to average -0.8°C and 2.0°C. These estimates are considered to be inaccurate because of the anomalous thermal mixture of genera and a bias in the transfer function data set toward ostracode assemblages of the western North Atlantic. A separate eastern North Atlantic ostracode transfer function should prove more useful and is presently being developed.

The inferred SBT values for units B1 and B2 indicate a subfrigid marine climate, ranging from a minimum of -2°C to a maximum of 9 to 10°C. The marginal marine fauna in unit B3 indicates a cold temperate to subfrigid marine climate, with an estimated minimum SBT of -2°C and a maximum SBT of 12 to 14°C. These temperatures indicate that the Kap København winters were similar to, or 1 to 2°C warmer than today, and that summers were about 7 to 8°C warmer. In the estuarine environment, SBT during the summer were particularly warm, perhaps 10 to 12°C warmer than at present.

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Pliocene Marine Climates of the Western North Atlantic Ocean

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Fossiliferous Pliocene marine sediments common in coastal regions bordering the North Atlantic provide evidence for a period of high sea level and warm global climate. For example, along the eastern United States, biostratigraphic data from the Duplin Formation of North Carolina and South Carolina indicate that the Orangeburg Scarp, a linear shoreline feature now at 85m above sea level (ASL), was formed 3.5 to 3.0 Ma (Dowsett and Cronin, in press). By subtracting post-3.0 Ma uplift of the scarp, mid-Pliocene eustatic sea level was estimated to be about +30 to 35m ASL, a range supported by evidence from the Yorktown Formation (Virginia) and the Duplin Formation (Georgia), and generally by oxygen isotope evidence from deep-sea cores.

A transfer function was developed to estimate summer and winter paleotemperatures for Pliocene deposits of the western North Atlantic and adjacent seas by using fossil ostracodes (Cronin and Dowsett, in press). Q-mode factor analysis was run on ostracode assemblages from 100 modern bottom samples from continental shelves of North America, Greenland, and the Caribbean using 59 ostracode categories (57 genera, 14 of which are monospecific, plus two categories that include rare tropical and arctic species). Seven factors accounting for 80% of the variance define assemblages that correspond to frigid, subfrigid, cold temperate, mild temperate, warm temperate, subtropical, and tropical climatic zones. Multiple regression of the factor matrix was run against observed August and February bottom temperatures assembled from various sources of hydrographic data. The results of the regression gave an ostracode transfer function (OTF) having an accuracy of about ±2°C.

The OTF was used to reconstruct Pliocene shallow marine climates from 37 samples from 14 sites in the region between Massachusetts and southern Florida. These samples were selected on the basis of their stratigraphy and age indicating deposition at or near the maximum transgression. The samples and estimates of Pliocene February and August temperatures are shown in Table 1.

Three main inferences about the middle Pliocene hydrography and marine climatology can be made from the OTF results. First, the paleo-Gulf Stream was extremely influential during the period 3.5 to 3.0 Ma in causing winter temperatures 5 to 8°C warmer than those today off the coast between Virginia and Massachusetts. The peak warmth, which occurred just before a major regression signaling the onset of Northern Hemisphere glaciation, correlates with the closing of the Isthmus of Panama (Cronin, 1988). Second, marine thermal gradients during the Pliocene were not as steep as they are today. For example, steep gradients that exist today at Cape Hatteras, N.C. were not present. In addition, during the Pliocene summer temperatures off the coast of southern Florida were not as warm and winter temperatures off the coast of Massachusetts were not as cool as today. Third, onshore-offshore temperature gradients can be recognized in southeastern Virginia and South Carolina where middle-outer shelf assemblages of the Yorktown and Duplin Formations give OTF temperatures cooler than correlative inner shelf assemblages. The results indicate that separate OTFs can be developed for the Arctic, the Pacific, and the eastern North Atlantic for application to Pliocene deposits in these areas

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TABLE 1. Pliocene samples and ostracode transfer function temperature estimates from U.S. Atlantic Coast.

Site	#	Unit (Location)	lat	long	wt	st	comm	specim
I	1	Shelly sand	41°20'	70°30'	7.3	18.3	.85	278
	2	(Martha's Vineyard, Mass.)	"	"	11.6	22.4	.66	55
II	3	Yorktown	37°57.5'	75°29'	9.0	20.8	.69	101
III	4	(Delmarva Peninsula)	37°57'	75°31'	9.0	19.4	.73	188
IV	5	Yorktown (SE Va.)	37°17'	76°43'	11.9	18.5	.58	527
V	6	Yorktown (York R., Va.)	37°13.5'	76°29.5'	15.8	19.6	.69	316
	7	"	"	"	15.2	19.3	.63	464
	8	"	"	"	13.0	20.8	.69	380
VI	9	Yorktown (James R.,Va.)	37°02'	76°35'	11.2	17.4	.71	290
	10	"	"	"	12.4	17.9	.68	273
VII	11	Yorktown (NE N.C.)	36°28'	77°06'	13.2	23.8	.60	230
VIII	12	Yorktown (NE N.C.)	36°03'	77°19'	12.3	21.2	.60	230
	13	"	"	"	11.2	22.9	.66	193
IX	14	Yorktown	35°23'	76°45'	13.3	21.0	.67	378
	15	(Lee Creek Mine, N.C.)	"	"	12.9	22.8	.71	446
	16	"	"	"	14.1	23.7	.73	536
	17	"	"	"	13.2	23.6	.73	894
	18	"	"	"	13.6	23.7	.68	368
X	19	Duplin (SE N.C.)	34°36'	78°59'	15.4	22.2	.75	359
	20	"	"	"	13.1	21.9	.68	375
	21	"	"	"	12.1	23.5	.77	301
	22	"	"	"	13.8	21.8	.66	346
	23	"	"	"	14.3	23.3	.76	871
XI	24	Duplin (NE S.C.)	34°19'	79°58'	14.1	23.8	.73	292
	25	"	"	"	16.3	22.3	.46	452
XII	26	Duplin (Savanah R., Ga.)	32°34'	81°21'	15.7	23.3	.68	279
	27	"	"	"	15.2	25.9	.76	289
XIII	28	Pinecrest (Sarasota, Fla.)	27°20'	82°25'	14.7	23.2	.72	323
	29	"	"	"	12.9	21.9	.54	277
	30	"	"	"	16.3	26.4	.62	499
	31	"	"	"	16.6	24.9	.54	422
	32	"	"	"	16.6	26.4	.76	467
XIV	33	Pinecrest (SE Fla.)	25°45'	80°15'	15.9	25.8	.86	251
	34	"	"	"	19.5	23.2	.51	229
	35	"	"	"	17.1	27.0	.87	181
	36	"	"	"	18.0	22.1	.53	89
	37	"	"	"	20.0	29.9	.46	106

Abbreviations: # = sample number, lat = latitude N, long = longitude W, wt = winter temperature (°C), st = summer temperature (°C), comm = communality, specim = number of specimens in sample.

Pliocene Paleoceanography of North Atlantic Deep Sea Drilling Project Site 552: Application of Planktic Foraminifer Transfer Function GSF18

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We have developed a transfer function (GSF18) for planktic foraminifers based on a recombined and simplified modern core-top data set (Dowsett and Poore, in press) which preserves the primary environmental signals and allows direct comparison of Pliocene and earlier Pleistocene assemblages to modern assemblages. Q-mode factor analysis and regression techniques were used to write a set of equations (transfer function GSF18) which relate modern physical oceanography to 5 factors or faunal assemblages. Transfer function GSF18 has approximately the same accuracy as previous Pleistocene transfer functions (standard error of estimate for winter temperatures is 1.47°C, and for summer temperatures 1.36°C), but can be more readily applied to faunal samples ranging in age from Pliocene to Holocene.

Transfer function GSF18 has been applied to faunal data from Deep Sea Drilling Project Hole 552A (fig. 1) to produce a ~2.5 my sea-surface temperature (SST) time series (Dowsett and Poore, in press). A chronolgy for Hole 552A was obtained from paleomagnetic data, biostratigraphy, and stable isotope correlations (Dowsett and Poore, in press). Estimates derived from Site 552 between 4.6 Ma and 2.3 Ma show periods during when mean SSTs were several degrees warmer and other periods when SSTs were several degrees cooler than modern conditions (fig. 2). The warm intervals are of particular interest to us. Reconstruction of the last interglacial (isotope stage 5e) indicates that sea-surface temperatures in this area of the North Atlantic were essentially the same as modern conditions (CLIMAP, 1984). Our SST estimates identify intervals within the Pliocene that were substantially warmer than modern, isotope stage 5e, and judging from isotopic records, any other Pleistocene interglacial.

Between 4.6 and 4.0 Ma SST estimates are characterized by low amplitude and low frequency oscillations. Mean winter SST (6°C) was 3.0°C cooler than modern. At 4.0 Ma temperatures warmed significantly and remained elevated until about 3.4 Ma. Fluctuations about the mean were small from 4.0 to 3.8 Ma and then became more pronounced towards the end of the warm interval. Mean winter SST was about 2.0°C warmer than modern during this interval. A marked cool interval is seen between 3.4 and 3.2 Ma. Winter SSTs returned to levels seen prior to 4.0 Ma, but the amplitude of variation about the mean is greater. The end of the cooling event is marked by another sharp temperature increase to a peak at about 3.1 Ma. Mean winter SST at the peak warming was about 3.5°C warmer than modern. The temperature record after 3.1 Ma shows a trend towards cooler conditions with high amplitude variations about the declining mean. The coldest estimates coincide with the first unequivocal occurrence of ice-rafted detritus at about 2.45 Ma.

Oxygen isotope records from DSDP 552 are available down to approximately 3.5 Ma. Above 3.1 Ma, the SST and planktic and benthic oxygen isotope records exhibit a good general correlation although details of the curves differ, partly because of different sampling densities. The cooling trend indicated by faunal sea-surface temperature estimates is matched by a trend towards heavier values in the isotopic

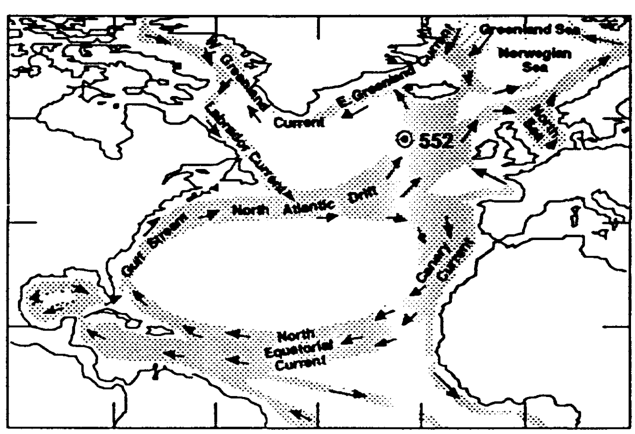

Figure 1. General circulation of the North Atlantic and location of DSDP Site 552.

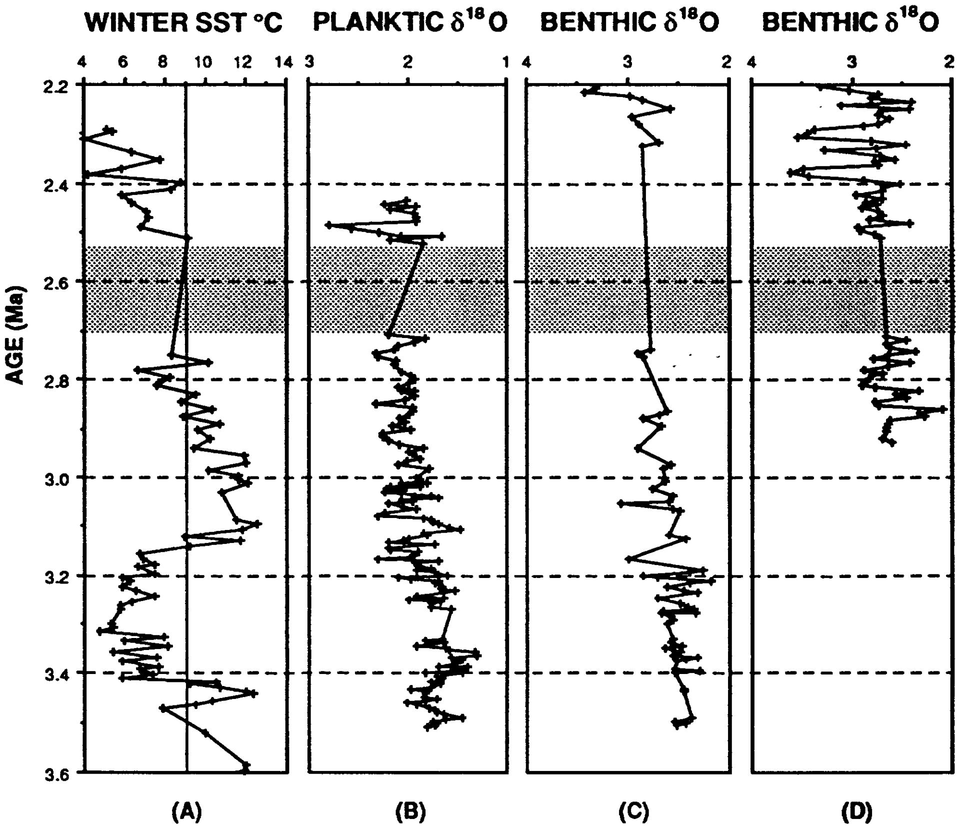

Figure 2. Comparison of winter SST with benthic and planktic stable isotope data from DSDP Hole 552a. (A), winter SST estimates. (B), planktic $\delta^{18}O$ from *Neogloboquadrina atlantica* analyses of Shackleton and Hall, 1985. (C), benthic $\delta^{18}O$ from *Cibicidoides* analyses of Shackleton and Hall, 1985. (D), benthic $\delta^{18}O$ from *Cibicidoides* analyses of Curry and Miller, in press. Shaded area indicates haitus.

records. High amplitude variability about the general trend is seen in all data sets. The warmest SST estimates at about 3.1 Ma correlate with a light excursion in the planktic isotopic data; the available benthic data at this point do not define a clear peak. These records indicate that following the thermal maximum at approximately 3.1 Ma, there was overall surface water cooling and gradual but continuous build up of global ice volume culminating in Northern Hemisphere glaciation. The glaciation is indicated by the presence of ice rafted detritus in 552A sediments at about 2.45 Ma. Between 3.1 and 3.5 Ma the isotopic records and SST estimates are not well correlated. The pronounced cool event seen in the SST estimates does not show up in either $\delta^{18}O$ record. Coarse fraction and percent benthic foraminifer indices and $CaCO_3$ data (Zimmermann and others, 1985) do not suggest that the faunal signal is related to a period of higher dissolution. In addition, the faunal change indicating cool temperatures in this interval of 552A is compelling both within this core and in other cores from the North Atlantic. Down core abundance data document the absence or very low abundance of *Globorotalia puncticulata* between 3.15 and 3.35 Ma. *Globigerina bulloides* shows well defined increases in abundance at the beginning and end, while "cold" *Neogloboquadrina* show a peak in the middle of the *Globorotalia puncticulata* gap. The same sequence of faunal events, including common occurrence of *Globorotalia puncticulata* above and below the "puncticulata gap" is a regional feature documented at several other sites in the North Atlantic (Sites 548, 606, and 607) at approximately the same time (Loubere and Moss, 1986; Raymo and others, 1986; Ehrman and Keigwin, 1986; Dowsett and others, 1988). We conclude that the cooling indicated by SST estimates in 552A between 3.15 and 3.35 Ma represents regional sea-surface cooling in the North Atlantic whereas the isotopic records from this interval of 552A are reflecting a complex signal that includes global ice-volume as well as local temperature near the surface and at the sea floor.

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Using Dinocysts To Calculate Winter Sea-Surface Temperatures, North Atlantic Ocean

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Although dinoflagellate cysts have considerable potential to contribute to our knowledge about past oceanographic conditions, they have rarely been used in paleotemperature reconstructions. This research is a pilot study to answer the basic questions: Can fossil dinocyst assemblage data be used to give quantitative estimates of past sea-surface temperatures, and how good are these estimates?

This study focuses on winter sea-surface temperatures and uses percentage abundance data of the 27 most common dinoflagellate cyst taxa from 103 modern North Atlantic samples in the data set of Wall and others (1977). The majority of these samples were taken along the coast and as transects along the shelf-slope-rise offshore of North America. A discontinuity in coverage - there are no sites with winter sea-surface temperatures between 10.6 and 15°C - reflects the steep latitudinal temperature gradient found off Cape Hatteras. Zoogeographic studies using molluscs (Hall, 1964) show the outer tropical province south of Cape Hatteras and the mild-temperate province north of Cape Hatteras. The warm-temperate province occurs in the eastern North Atlantic (off France, Spain, and Portugal) but is absent or unrecognized in the westernmost North Atlantic.

Analysis of the Wall and others (1977) data set shows that samples containing <5% total of the six key tropical species *Impagidinium aculeatum, I. strialatum, Operculodinium israelianum, Polysphaeridium zoharyi, Lingulodinium machaerophorum*, and *Tuberculodinium vancampoae* most likely represent winter sea-surface temperatures of <11 to 15°C. Within this cool subset, the percentage total of five key temperate species, *Nematosphaeropsis labyrinthus, Impagidinium paradoxum, I. sphaericum, I. patulum*, and *Tectatodinium pellitum*, is used to estimate winter sea-surface temperature. The temperature may be estimated quantitatively by the equation:

$$SST_{est} = 2.103 + 1.485 \sqrt{F}$$

where SSTest is estimated sea-surface temperature in °C and F indicates % composition of the five key temperate species. The standard error of this equation is 1.33°C.

The temperature may also be estimated semiquantitatively. Samples containing <5% key tropical forms and 10 to 30% key temperate forms most likely represent winter sea-surface temperatures in the 7 to 11°C range. Samples containing <5% key tropical and 1 to 10% key temperate forms most likely represent winter sea-surface temperatures in the 2 to 7°C range. Samples containing few or no key tropical forms and few or no key temperate forms could be from one of several specialized environments, cold or warm.

In order to extend and test the applicability of the above results, I added dinocyst data from 45 samples given in Turon (1984). Although not all of the counting categories are the same, and one must rely on the taxonomic identifications of a different author, these samples fill a very important gap in coverage: samples from the warm temperate province and samples that show smaller differences in summer and winter sea-surface temperatures (less seasonality) than the Wall and others (1977) data set.

Preliminary results allow semiquantitative estimates of winter sea-surface temperatures based on the combined data set. Samples containing >5% total of the six key tropical species given above are seen to represent winter temperatures of >10 to 11°C (warm-temperate or tropical). The warm-tropical dinocyst flora is relatively consistent over the range of 10 to 14°C winter sea-surface temperature. This flora can be separated from the tropical dinoflora on the basis of having >20% of six key temperate species: *Bitectatodinium tepikiensis, Nematosphaeropsis labyrinthus, Impagidinium paradoxum, I. sphaericum, I. aculeatum*, and *Tectatodinium pellitum*. The species, *Impagidinium aculeatum*, although useful in separating warm from cooler samples, can be seen to be a temperate species, with its peak abundance in the 10 to 12°C winter sea-surface temperature range.

Twenty productive samples from DSDP Hole 552A, located near the Rockall Plateau (latitude 56°02.56'N, longitude 23°13.88'W) were examined for dinoflagellate cysts (fig. 1). The present day winter sea-surface temperature in the area is just over 9°C. A total of 18 samples most likely represent either 6 to 10.5°C winter sea-surface temperature (cool and mild

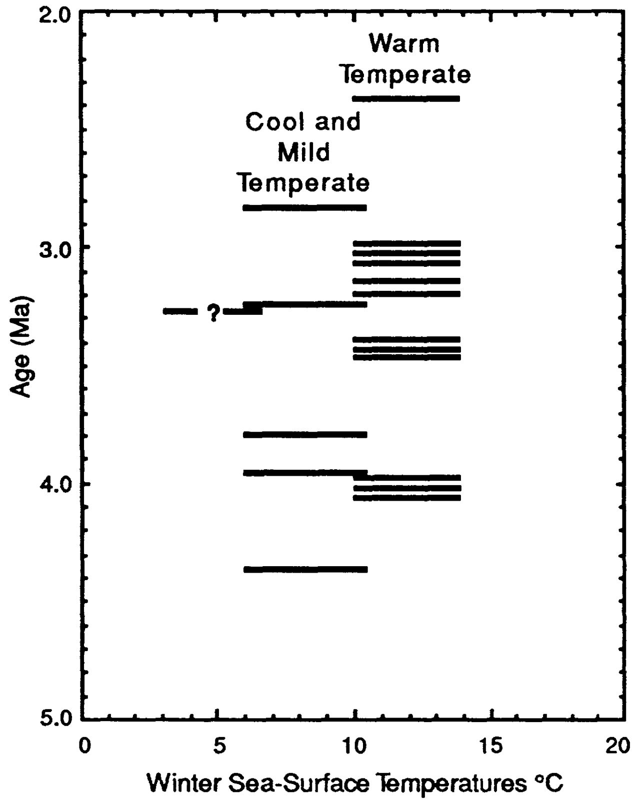

Figure 1. Estimated paleotemperatures (winter sea-surface temperature in °C) for 19 samples from Hole 552A. Estimates based on quantitative analysis of dinoflagellate cysts. Sample yielding poorly constrained temperature range because of poor preservation is questioned. Age is in Ma (million years before present) from Dowsett and Poore (in press).

temperate) or 10 to 14°C winter sea-surface temperature (warm temperate). One sample (at 3.26 Ma) possibly represents cooler temperatures, but this sample contains relatively few, poorly preserved specimens, and so this interpretation must be considered tentative. One sample (at 3.75 Ma), not plotted in figure 1, was dominated by two opportunistic species (*Spiniferites splendidus/mirabilis* and *Operculodinium centrocarpum*). These species do not help give an estimate of temperature, but may perhaps indicate a low level of nutrients. Figure 1 shows that during the Pliocene, winter sea-surface temperatures were for the most part slightly warmer than present day, although at certain times (2.83, 3.23, 3.26, 3.79, 3.95, and 4.36 Ma) the temperatures were about the same as or slightly cooler than the present. At 3.75 Ma, the dinoflora suggests a nutrient-poor environment (perhaps in the middle of an ocean gyre).

For Hole 552A, the paleotemperature estimates derived semiquantitatively based on the dinocyst assemblages are in close agreement with paleotemperature estimates produced from planktic transfer function GSF18 (Dowsett and Poore, in press). These two very different techniques allow us to recognize specific time intervals that were warmer or cooler than the present. Because the techniques independently produce estimates of such similar magnitude, we are more confident in the accuracy of our results.

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Strontium Isotopic Dating of Upper Cenozoic Marine Deposits, Northwestern Alaska

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High-precision strontium isotopic measurements were obtained for 24 late Cenozoic molluscan shells from northwestern Alaska to test whether such data can be used to date Arctic marine deposits. At present, the ages of geologic and climatologic events recorded by circum-arctic deposits older than the range of radiocarbon dating (>30 ka) are poorly known. Placing these events into a firm chronological framework will elucidate the interrelations between late Cenozoic global climate evolution and environmental changes in the Arctic, such as the inception of northern hemispheric ice sheets and perennial Arctic Ocean sea ice. Samples for this study were chosen from emerged and offshore marine deposits at Nome (Western Alaska), Skull Cliff, and the Colville River area (North Slope of Alaska), where ages are relatively well constrained on the basis of radioisotopic, paleontologic, paleomagnetic, and amino acid criteria (Brigham, 1985; Carter and others, 1986; Kaufman and others, 1989). An additional molluscan fossil was provided for analysis by J.V. Matthews (Geological Survey of Canada) from the Beaufort Formation on Meighen Island in the Canadian high Arctic.

Because a number of standards are currently in use, and there is no consensus on the $^{87}Sr/^{86}Sr$ value of any one standard, we report Sr isotopic compositions in delta (Δ) Sr units where, Sr = ($^{87}Sr/^{86}Sr_{(sample)}$ - $^{87}Sr/^{86}Sr_{(standard)}$) x 10^5. The measured $^{87}Sr/^{86}Sr$ for our standard, a Holocene marine carbonate of the mollusc *Tridacna* (EN-1), is 0.709180 with an external reproducibility of ±0.00001 (1 Δ unit), which we take to be the analytical reproducibility of our sample measurements.

Measured Δ Sr values in marine carbonates are converted to absolute ages using a calibration curve that defines the evolution of $^{87}Sr/^{86}Sr$ in seawater through time. High-precision Sr analyses have been performed on late Cenozoic foraminifera and nanno-

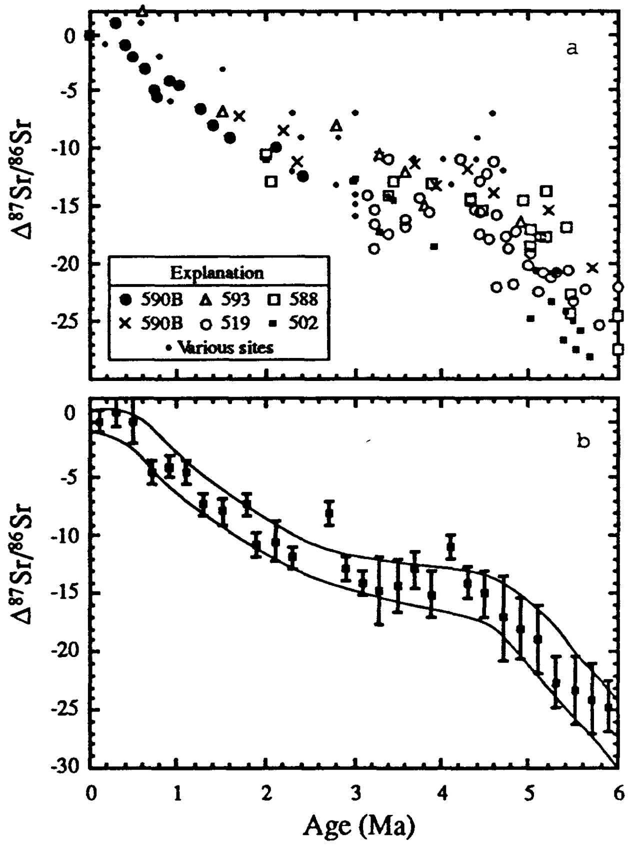

Figure 1. Evolution of $^{87}Sr/^{86}Sr$ in Late Cenozoic marine carbonate from several independently dated deep-sea cores reported as delta (Δ) Sr units, where Δ Sr = ($^{87}Sr/^{86}Sr_{(sample)}$ - $^{87}Sr/^{86}Sr_{(standard)}$) x 10^5. (a) Composite measurements from 5 cores as follows: Site 590B from DePaolo (1986) and Capo and DePaolo (1988); Site 593 from Hess and others (1986); Sites 519, 588, and 502 from Hodell and others (1989). Data from "various sites" (Koepnick and others, 1985) are of lesser precision and were excluded from the calibration curve. (b) Calibration curve used in this report to determine age ranges. Solid squares are mean isotopic values within 0.20 my time increments; error bars are ±1 standard deviation. See text for a description of procedure used to construct the curve.

Table 1. Results of high-precision strontium-isotopic analyses of Plio-Pleistocene molluscan fossils from northern Alaska and Canada.

Location	Unit	Genus	ppm	Strontium $\Delta^{87}Sr/^{86}Sr$ *
SKULL CLIFF AREA				
Sec 38	Walakpa	*Hiatella*	1580	0.0
Sec 20	Karmuk	*Hiatella*	3780	-3.0
Sec 20	Karmuk	*Hiatella*	2880	-2.5
Sec 20	Taupaktushak	*Hiatella*	2090	-4.2
Sec 20	Taupaktushak	*Mya*	2010	-4.1
Sec 65	Taupaktushak	*Hiatella*	1540	-5.6
Sec 65	Taupaktushak	*Mya*	1060	-5.9
Sec 24	Killi Cr	Unknown	2130	-7.2
Sec 67	Killi Cr	*Mya*	2600	-5.9
Sec 26	Nulavik	*Hiatella*	1790	-4.8
Sec 26	Nulavik	*Hiatella*	3100	-5.7
Sec 26	Nulavik	*Hiatella*	2410	-5.8
COLVILLE RIVER AREA				
Fish Cr	Fishcreekian	*Hiatella*	2450	-5.6
Fish Cr	Fishcreekian	*Hiatella*	2270	-3.7
Colville R	Bigbendian	*Hiatella*	2250	-13.0
Colville R	Bigbendian	*Hiatella*	2200	-12.1
Kikiakroruk R	Colvillian	*Hiatella*	3020	-12.7
Colville R	Colvillian	*Hiatella*	2290	-10.6
NOME				
Type loc	Anvilian	*Mya*		3.8
Type loc	Beringian	*Hiatella*		3.2
Type loc	Beringian	*Hiatella*	1550	23.3
Offshore		*Astarte*	1831	5.8
Offshore		*Mya*	3050	2.3
Offshore		*Mya*	4110	70.8
MEIGHEN ISLAND				
	Beaufort Fm	*Arctica*	-15.3	

* $\Delta^{87}Sr/^{86}Sr = (^{87}Sr/^{86}Sr_{(sample)} - ^{87}Sr/^{86}Sr_{(standard)}) \times 10^5$;
Analytical uncertainty is $\pm 1.0\ \Delta$ (± 0.00001)

fossils from a number of independently dated deep-sea cores (Koepnick and others, 1985; DePaolo, 1986; Hess and others, 1986; Capo and DePaolo, 1988; Hodell and others, 1989). Because the fine-scale structure of each calibration curve differs and we have no means of assessing their accuracy, we incorporated all published high-precision Sr determinations into a composite calibration curve for the late Cenozoic (fig. 1). We constructed the calibration curve by calculating the mean and standard deviation of all Sr measurements within time increments of 0.20 my and then used those values to guide a hand-drawn, best-fit envelope. The analytical precision of ±0.00001 reported by most laboratories was substituted for error limits when standard deviations fell below this value. We consider this sound because: (1) the Sr isotopic evolu-

Other age control	Estimated strontium age range (Ma)	Marine transgression	Lithostratigraphy (Gubik Fm)	Δ $^{87}Sr/^{86}Sr$ (range) Skull Cliff	Δ $^{87}Sr/^{86}Sr$ (range) Colville River area
	0.0-0.7	Pelukian	Walakpa	0.0	
N	0.3-1.1	Wainwrightian	Karmuk	-2.5 -3.0	
R	0.5-1.7	Fishcreekian	Tuapaktushak	-4.1 -5.9	-3.7 -5.6
R	1.9-4.7	Bigbendian	Killi Creek	-5.9 -7.2	-12.1 -13.0
	1.6-4.5	Colvillian	Nulavik	-4.8 -5.8	-10.6 -12.7
Bering Straits opens					
	2.8-5.1	Meighen Island Beaufort Fm		-15.3	

Figure 2. Summary diagram showing correlations of lithostratigraphic and marine transgressive sequence of the Gubik Formation, Alaska Arctic Coastal Plain (from Carter and others, 1986), and strontium age range in relation to other independent age control (N, normally magnetized; R, reversely magnetized). Strontium age ranges were determined using the composite calibration curve of Figure 1b as outlined in the text. Strontium values, reported in delta (Δ) units, where Δ Sr = ($^{87}Sr/^{86}Sr_{(sample)}$ - $^{87}Sr/^{86}Sr_{(standard)}$) x 10^5, are the maximum range of values measured from each unit. Values in shaded boxes are thought to erroneous, and were excluded from age estimates.

tion of seawater is not expected to conform to any *a priori* mathematical model; and (2) changes in the $^{87}Sr/^{86}Sr$ composition of seawater should be negligible over a time span that is short relative to the residence time of Sr (~4.0 my), so that the scatter is nonsystematic, but rather the result of analytical uncertainty, chronologic error and any diagenetic effects. The envelope encompasses 90% of the 101 isotopic measurements, taking into account the analytical precision of each measurement. The shape of the envelope is similar to that reported qualitatively by Jones and others (1989). It is characterized by two distinct inflections, at approximately 2.4 and 4.5 Ma, separated by a relatively flat zone in which there is little or no change in $^{87}Sr/^{86}Sr$ with time, implying that the technique has minimal age resolution within this time span.

We determined sample ages by applying the maximum range of measured Sr values for each stratigraphic unit (table 1), along with the analytical uncertainty of ±1.0 Δ Sr unit, to the calibration envelope and reading the estimated age range directly from the graph. Note that analyses of four shells of two genera which were collected from a single stratigraphic horizon (the informally named Tuapaktushak member of the Gubik Formation) over a distance of more than 100m have indistinguishable D Sr values within the range of analytical precision, confirming a lack of vital effects. Until the calibration curve is refined, we adopt the conservative approach that there is an equal probability that the true age of the sample, based entirely upon its Sr composition, lies anywhere within the estimated age range.

Although the age ranges are broad, Sr predicted ages are consistent with other age criteria for the majority of samples (fig. 2). The Sr age estimate (2.8 to 5.1 Ma) for the Meighen Island Beaufort Formation, based on a single analysis, is younger than the Late Miocene Sr age of 9 ±1 Ma or 6.5 Ma reported by Matthews (1989). The possible age range for this and the younger deposits can be narrowed using other independent age evidence such as the opening of Bering Straits at approximately 3.0 Ma, which is represented by an influx of marine molluscs of Pacific origin into the Arctic and North Atlantic Oceans. Thus, the utility of the Sr technique is enhanced when used in conjunction with other age-constraining criteria. Strontium isotopic ages for the

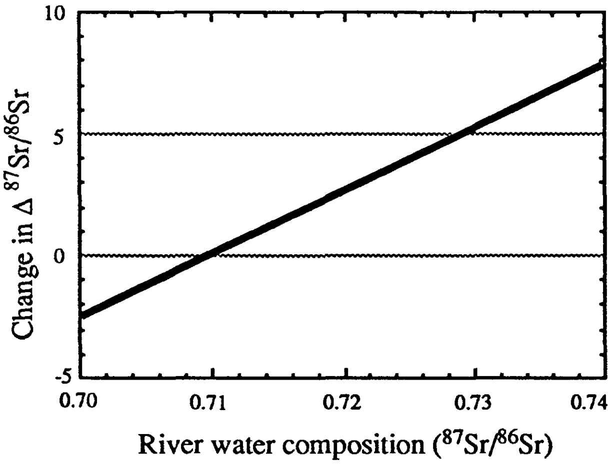

Figure 3. Simple strontium-mixing model used to estimate the change in seawater strontium composition assuming contamination by 20% fresh water (salinity = 28‰) containing 0.10 ppm strontium. Isotopic change given in delta (Δ) Sr units.

upper three units at Skull Cliff agree with other age evidence although the Sr age range for Fishcreekian deposits at Skull Cliff and the Colville River area (0.5 to 1.7 Ma) are younger than the 2.4 Ma age for Fishcreekian deposits in the Colville River area based on microtine rodent fossils (Repenning and others, 1987). Strontium isotopic ages for the lower two units at Skull Cliff are younger than expected. In contrast Δ Sr values in shells from deposits believed to be correlative with these two units in the Colville River area (Carter and others, 1986) are consistent with most other evidence. This indicates that the high Δ Sr values at Skull Cliff are probably the result of local effects and do not reflect the isolation and independent isotopic evolution of $^{87}Sr/^{86}Sr$ in the Arctic Ocean.

The Δ Sr values in five shells collected from on- and offshore deposits of the Nome area, ranging in age from late Pliocene to middle Pleistocene, are higher than modern ocean water (+2.3 to +70.8, table 1). The input of meteoric water and crustal Sr from river outflow into the shallow embayment of Norton Sound may have influenced the Sr isotopic composition of coastal water at Nome. However, a simple mixing model (fig. 3) predicts that Δ Sr of modern seawater would only increase by 1.2 Δ units if mixed with 20% river water containing a Sr composition similar to river water draining the area today (Sr concentration = 0.12 ppm; $^{87}Sr/^{86}Sr$ = 0.7120). Salinities of 28‰ are typical for bottom water of Norton Sound (USDI, 1988) and within the habitat tolerance of the molluscan genera found in marine deposits at Nome. The effect of meteoric water does not appear to be sufficient to account for the very high Δ Sr values measured in the Nome fossils. Possibly, the $^{87}Sr/^{86}Sr$ composition of rivers draining the Nome area was considerably higher in the past.

Although Sr isotopic data can provide useful age information from arctic sites, a better understanding of the processes controlling the Sr isotopic composition of coastal water will be necessary before Sr isotopes can be considered a reliable chronostratigraphic tool for the continental margins. Measurements of stable isotopes of oxygen and carbon in the fossils and Sr concentration and isotopic ratios in the river water draining the region may provide useful information. Furthermore, refinement of the calibration curve is needed to reduce uncertainties in age estimates.

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Climates and Paleogeography of Transcaucasia in the Pliocene

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Paleoclimate studies are receiving increased attention as scientists realize that information on past climates is an important component of modern climatology. Information on magnitudes, frequencies, and effects of past natural climate variability are crucial for creating a fundamental theory of climate, for predicting possible future changes, and for assessing the possible consequences of these changes for other components of the environment. Therefore, paleoclimatology is an important part of the general program of climate studies in the Soviet Union. At present, paleoclimate studies are being conducted in nearly all geographic regions, spanning a broad time range from the late Cretaceous to the Holocene.

This report outlines principal results from study of Pliocene climates in the area of Transcaucasia. The wide distribution and nearly continuous sequence of Pliocene sediments, some of which are very fossiliferous, make it an ideal place for reconstructing a comprehensive and detailed history of this region's climatic events.

The paleogeographic and paleoclimatic reconstructions summarized in this report are based on palynological, paleobotanical, faunistic, lithologic, and geochemical data. Numerous volcanic ashes throughout the sequence provide an important means for correlating and dating events.

Prior to the beginning of the Pliocene (within the Pontian), paleogeographic reconstructions show that a vast Transcaucasian intermountain depression was located between the Larger Caucausas to the north and the Smaller Caucausas to the south. Although many researchers consider the Pontian to be basal Pliocene, we have absolute dates that bracket the Pontian between 7 and 6 Ma. Therfore, the Pontian is Miocene and correlates with the late Tortonian and Messinian of the Mediterranean region.

A transgression during the Pontian connected the Aegean, Black, Dakian, and Caspian Seas and formed narrow, shallow inlets in the western and eastern parts of the Transcaucasian intermountain depression. Coarse terrigenous through clay-carbonate deposits up to 1000m thick were deposited in parts of the Transcaucasian depression during the Pontian transgression (Chumakov and Mamedov, 1988). Leaf floras and palynological data obtained from these sediments indicate surrounding land areas supported abundant and widespread forest vegetation. Palynological studies of Pontian sediments have identified 122 taxa representing 55 families and 90 genera (Mchedlishvili, 1963; Ramishvili, 1969). Over half of these plants are now found in Southeast Asia, North America, and Micronesia. The high diversity is explained, in part, by large vertical relief and resulting altitude controlled belts of vegetation in mountains surrounding the intermountain basin. The highest elevation was occupied by diverse coniferous forests, low- and middle-mountain belts were occupied by broad-leaved forests with evergreen elements, and the foothills and lowlands were occupied by evergreen forests and savanna type forest-steppes (fig. 1).

Arealograms and climatograms have been used to identify the following climatic characteristics for foothills and lowland territories of Eastern Transcaucasia: January temperatures of 12 to 14°C, July temperatures of 25 to 27°C, mean annual temperatures of 18 to 20°C, and precipitation of over 700 to 800mm/year. For equivalent elevations in the Western Transcaucasia, climatic characteristics were: January temperatures of 8 to 10°C, July temperatures of 22 to 25°C, mean annual temperatures of 15 to 17°C, and precipitation over 1000mm/year (Shatilova, 1967).

At the end of the Pontian, a trend towards dryer and cooler climates began that continued into the middle Pliocene. During this period, the area of dense coniferous forests (mostly composed of pines) significantly expanded, deciduous elements became more important in broad-leaved forests, and the most thermophilic and hydrophilic evergreen species disappeared (fig. 2). With regard to land faunas, taxa characteristic of dry savannas became more widespread.

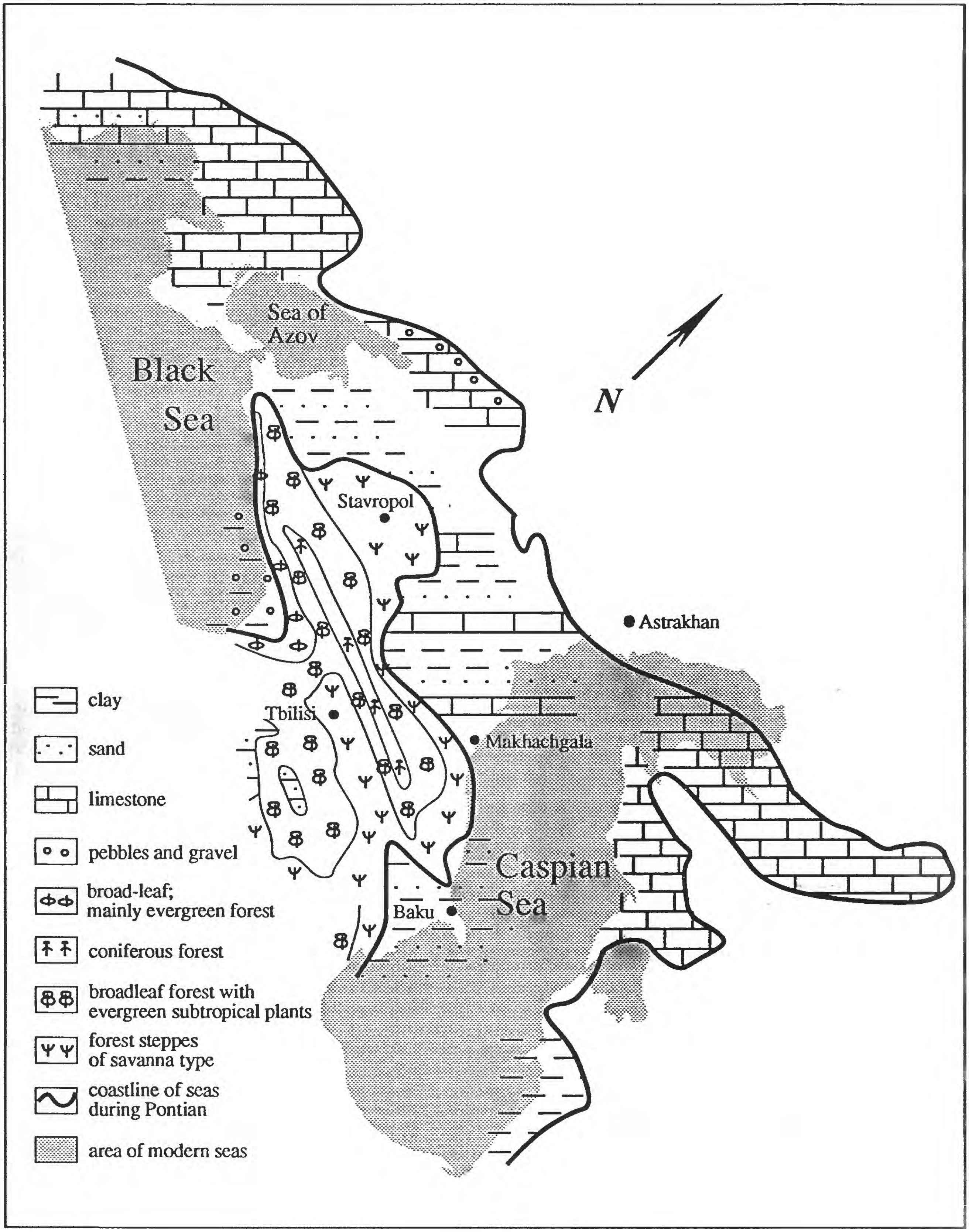

Figure 1. Reconstruction of Transcaucasus in Pontian (7 to 6 Ma).

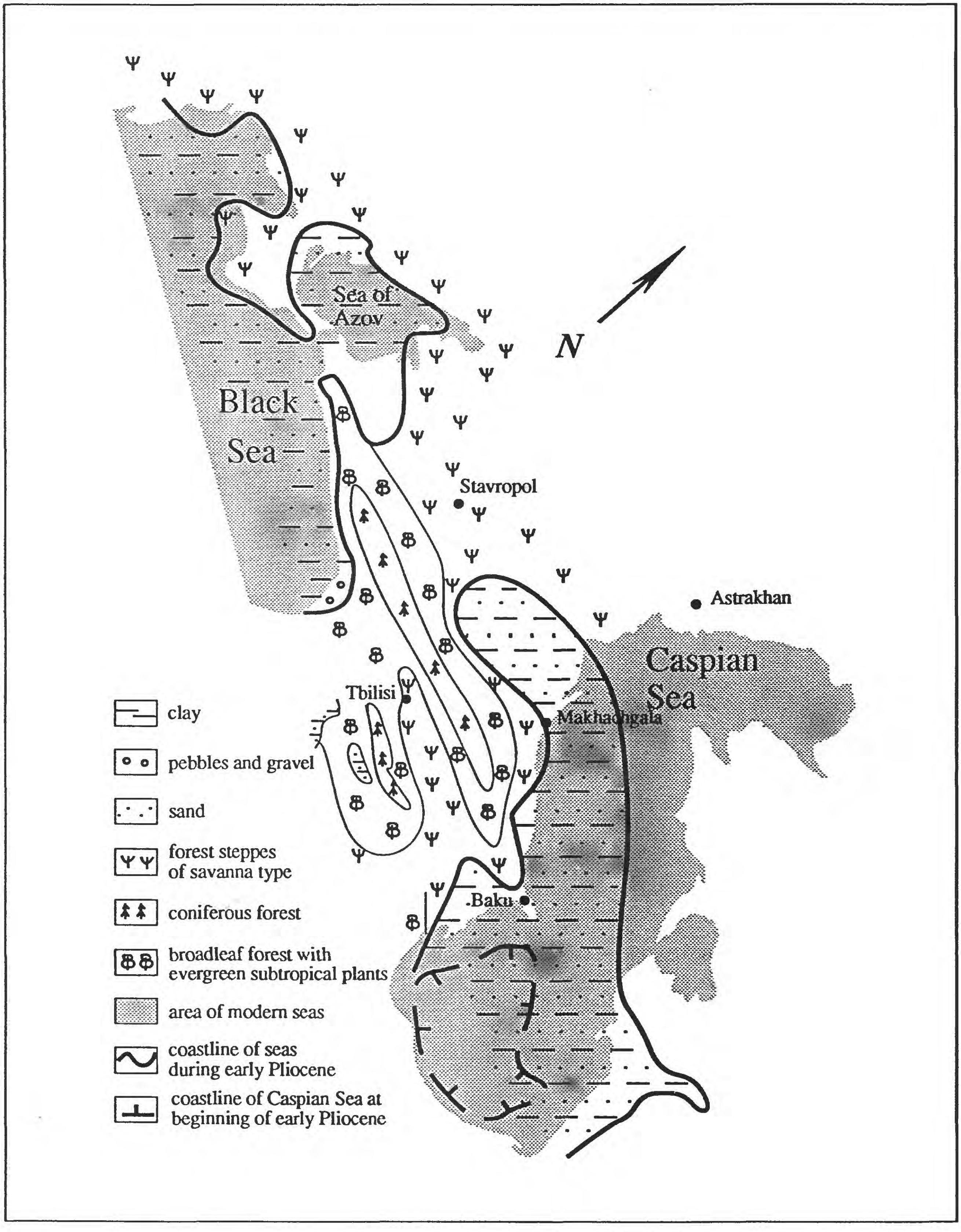

Figure 2. Reconstruction of Transcaucasus in early Pliocene (4.4 to 3.4 Ma).

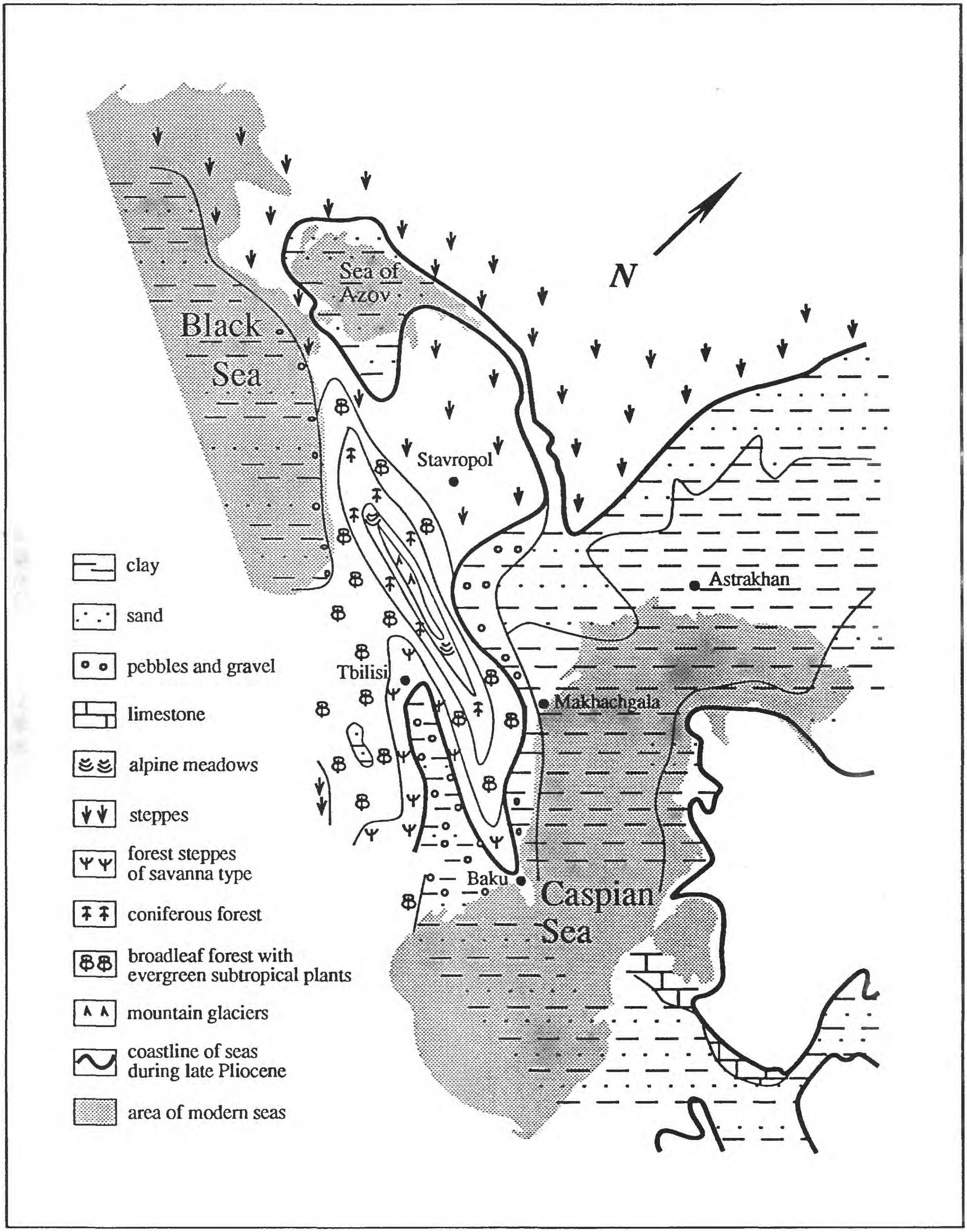

Figure 3. Reconstruction of Transcaucasus in late Pliocene (2.8 to 2.4 Ma)

The trend towards dryer and cooler climates at the beginning of the Pliocene coincided with a major regression of the Black and Caspian Seas. The level of the Caspian Sea fell 500m and the sea was transformed into a small lacustrine water body confined to the limits of the modern southern Caspian (fig. 2). The shoreline of the early Pliocene Black Sea did not extent beyond the limits of the modern Black Sea. Despite the sharp dramatic change of land/ocean ratio as compared to the Pontian, the climate remained mild, especially in the western part of Transcaucasia that was influenced by humid air streams from the Atlantic and the Mediterranean. There, over 30% of taxa in pollen assemblages from early Pliocene sediments represent evergreen taxa that currently occur in Southeast Asia in areas with January temperatures of 8 to 10°C, July temperatures of 24 to 26°C, and a mean annual temperature of 16 to 17°C.

In contrast, climatic conditions in Eastern Transcaucasia were influenced by the dry continental climates of the Iranian Highland and middle Asia. Evidence for hot and arid climates are inferred primarily from the common occurrence of gypsum and intermittent stream deposits. However, the presence of deltaic and lacustrine sediments along with a rise in the level and extent of the Caspian Sea towards the end of the early Pliocene indicates change to more humid conditions. At the present time, palynological studies of early Pliocene deposits of the Eastern Transcaucasia are incomplete and quantitative climate reconstructions of the area are not yet possible (Dzhabarova, 1979).

Palynological data indicate that the end of the early Pliocene (4.4 to 3.4 Ma) represents the climatic optimum of the Pliocene (Ramishvili, 1969). The paleoclimatic reconstruction for lowlands of the Western Transcaucasia based on palynological data indicate the following characteristics for the end of the early Pliocene: January temperatures of 12 to 14°C, July temperatures of 25 to 27°C, mean annual temperature of 18 to 20°C and precipitation of 1000 to 1500 mm/year.

During the late Pliocene, another sharp trend towards drying and cooling is seen (fig. 3). A considerable part of the subtropical and tropical genera disappear from the palynoflora. Pine forests and deciduous broad-leaved forests expanded (Shatilova, 1967;1974). However, continued presence of *Magnolia*, *Cinnamonium* and *Laurus* in the flora along with the occurence of rhinoceros, mastodon, and gazelles in the mammal fauna indicates that foothill and lowland areas had temperate-mediterranean climates with mean annual temperatures above 10°C and winter temperatures above 0°C. Figure 4 shows details of climatic fluctuations in foothill and lowland areas during the late Pliocene. During the interval between 3.4 and 1.8 Ma, three cooling and warming cycles are superimposed on the overall general cooling trend. Successive thermal maxima are progressively reduced in magnitude and maximum cooling occurred within the interval of 2.8 to 2.4 Ma, when glaciers were formed in the high-mountain zone of the Larger Caucasus. The cooling seen at 2.8 to 2.4 Ma is regarded as the major thermal minimum of the Pliocene. The Pliocene phases of cooling coincide, in general, with intervals of active volcanism in the Caucausus. (Most of the volcanic ashes in this area are found in upper Pliocene sediments).

The pattern of climatic change reconstructed from floristic analysis agrees well with the pattern of temperature changes in the Pliocene marine basins (Mamedov and Rabotina, 1984). Figure 5 shows that the warmest bottom-water temperatures (20 to 22°C) were in the early Pontian between 7 to 6.4 Ma and the late Cimmerian between 4.4 and 3.4 Ma. The coolest bottom-temperatures (8 to 10°C) occurred during the Akchaghylian at 2.8 to 2.4 Ma.

Our study demonstrates that the frequency and amplitude of climatic fluctuation changed during the Pliocene. From the beginning of the Pontian until the end of the Cimmerian, approximately 3.5 my, major climate cycles were on the order of 1 to 1.5 my with amplitudes of temperature change of about 2 to 3°C. In the late Pliocene, major climate cycles were on the order of 200,000 to 400,000 years with amplitude of temperature change ranging up to 10°C.

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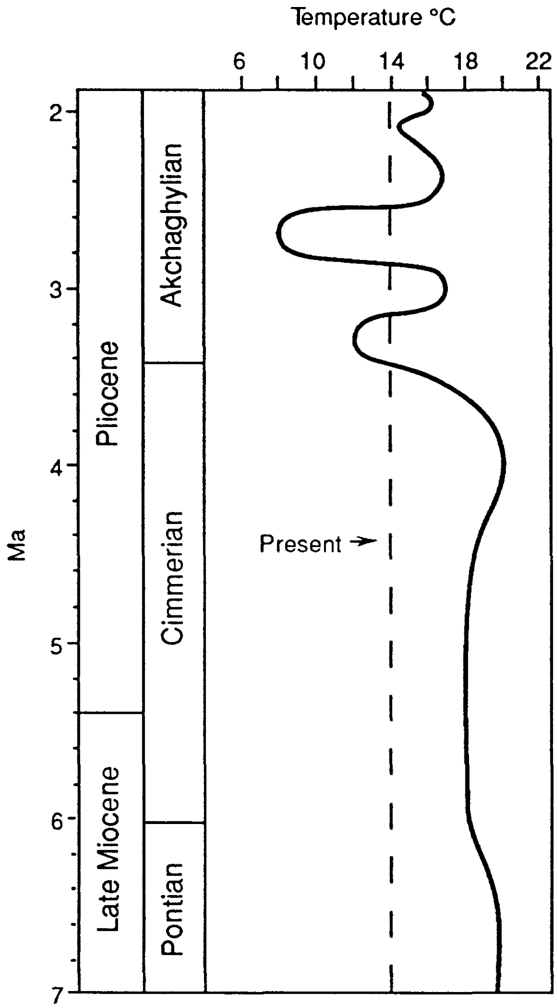

Figure 4. Average annual temperature in plains and foothills of Transcaucasus during the Pliocene. Reconstructed primarily from palynological studies in Western Caucasus.

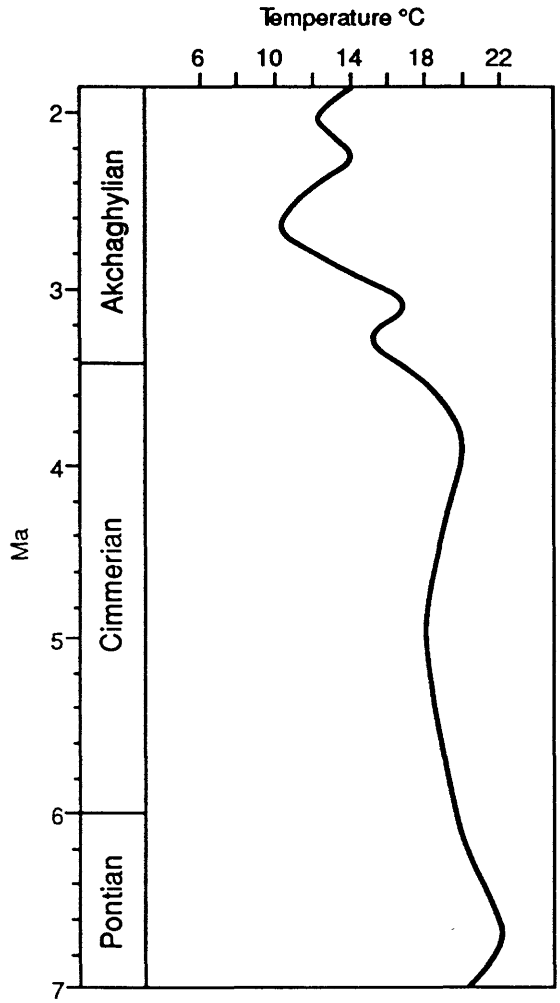

Figure 5. Temperature record of bottom water in the Pliocene seas of Transcaucasus.

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New Data on Pliocene Floras/Faunas from the Canadian Arctic and Greenland

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This report summarizes stratigraphic and paleontologic data from several sites of probable Pliocene and early Quaternary age in northern Canada and Greenland (fig. 1). An attempt is made to correlate the sites with the marine oxygen isotope record.

NOTES ON SITES

Worth Point Formation at Worth Point, Figure 1a (southern Banks Island, Northwest Territories)

Stratigraphy: The site is a marine shoreline section exposing a valley-fill sequence inset in the Beaufort Formation (Tertiary) and the Kanguk Formation (Cretaceous) sediments, and is overlain by deposits of the Banks Glaciation.

Paleontology: Exceptionally well preserved fossils of plants and insects are present, including those of a single tree type—the extant *Larix laricina* (table 1). All of the plant and insect fossils represent extant species; however, many of them do not occur on Banks Island today.

Age: The valley-fill sequence is stratigraphically below deposits known to predate the Bruhnes/Matuyama boundary. Preliminary results show that all Worth Point sediments are magnetically reversed (J-S. Vincent, pers. comm., 1989).

Comments: Fossils represent a cooling trend, involving a change from open larch woodland at the base of the sequence to low arctic tundra at the top. Fossils of the extinct plant *Aracites*, which is present in many Tertiary assemblages from the north, are conspicuously absent. This may mean that the deposits postdate extinction of that species in the Arctic. The latest dated record of *Aracites* in the north comes from the Lost Chicken peat (Matthews and

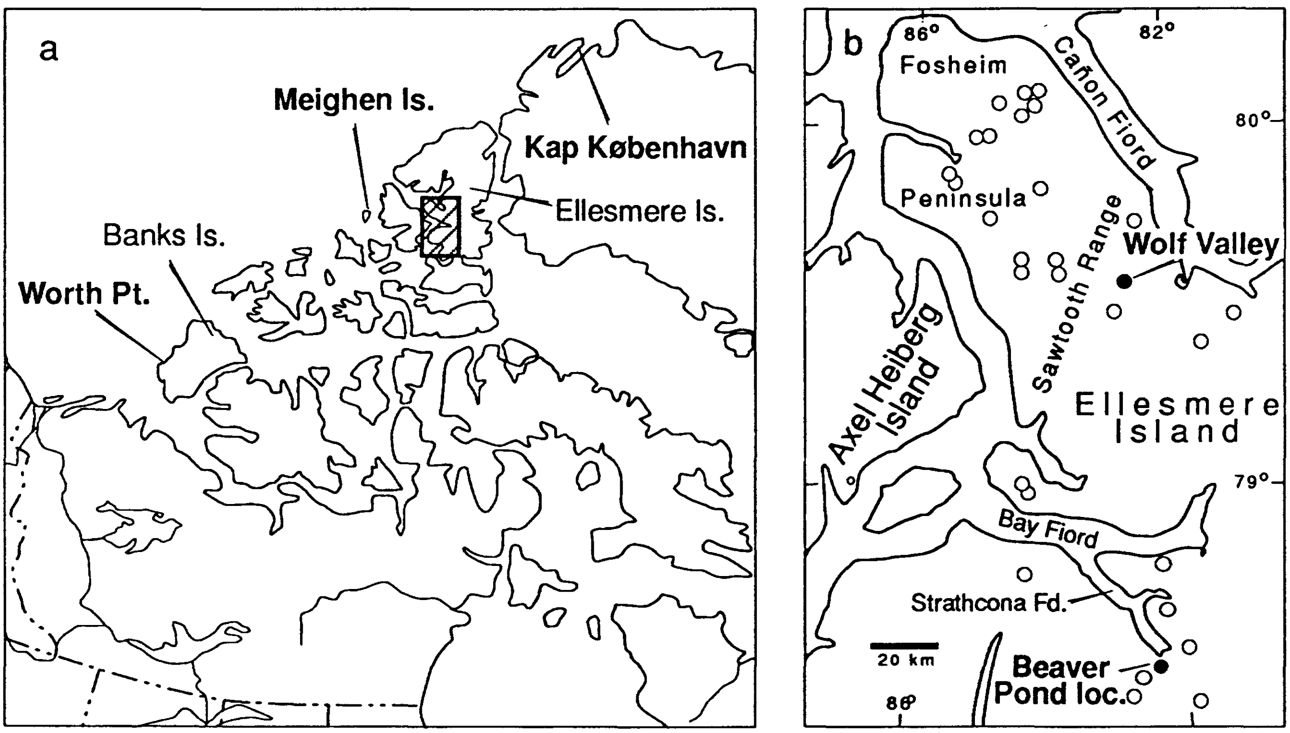

Figure 1. a: location map for probable Pliocene and early Quaternary sites in northern Canada and Greenland; b: detail of Ellesmere Island.

Table 1. Selected Plant taxa from Pliocene (?) sites in Canadian Arctic and Greenland arranged to emphasize floral differences.

	M.I.*	E.I. (b.p.)*	K.K.*	E.I (W.V.)*	W.P.*
Paliurus type **		+			
Decodon sp.		+			
Picea banksi Hills & Ogilvie	+				
Betula shrub type	+				
Pinus (5-needle type)	+	+			
Alnus tertiaria Dorof.	cf.	cf.			
Betula arboreal type	+	+			
Comptonia sp.	+	+			
Physocarpus sp.	+	+			
Larix sect. ***Multiseriales***	?	+	+		
Thuja occidentalis L.	+	+	+		
Aracites sp.	+	+	+		
Dryas sp.	+	+	+		
Myrica eogale V. Nikitin	?	cf.	cf.		
Scirpus microcarpus Presl.		+	+		
Picea mariana (Mill.) B.S.P.		+	+		
Menyanthes trifoliata L.	+	+	+	+	
Saxifraga oppositifolia L.	+	+		+	
Potamogeton spp.	+	+	+	+	+
Betula nana-glandulosa type		+	+	+	+
Alnus crispa (Ait.)Pursh	+	+	cf.	+	+
Oxyria digyna (L.)Hill.	+	+	+	+	+
Ranunculus lapponicus L.	+	+	+	+	+
Larix laricina (DuRoi)Koch.					+

*M.I.= Meighen Island; E.I.(b.p.) = beaver pond, Ellesmere Island; K.K.= Kap København; E.I. (W.V.) = Wolf Valley, Ellesmere Island; W.P. = Worth Point Banks Island, ** Bold text indicates taxa that are known or suspected to be extinct.

Ovenden, in preparation), which is dated at approximately 2.0 Ma (Naeser and others, 1982; J.A. Westgate, pers. comm., 1989).

Beaufort Formation on Meighen Island, Figure 1a (Queen Elizabeth Islands, Northwest Territories)

Stratigraphy: There are approximately 200m of sand with organic debris above current sea level and several hundred meters below the surface. On the west central side of the island the sands are interbedded with a unit of marine mud, the top of which is approximately 100m above sea level.

Paleontology: The marine unit contains shells of the boreal Atlantic clam, *Arctica*, and a small foraminifera assemblage which includes the Pliocene species *Cibicides grossus*. Terrestrial deposits above the marine unit contain plant fossils and insects indicative of forest-tundra, with 5-needle pine, spruce, larch, *Thuja*, and poplar (Matthews, 1987). Among herb/shrub fossils are those of the extinct plants *Aracites*, *Myrica* cf. *eogale*, a form of shrub birch, and *Alnus* cf. *tertiaria*. Some of the plants (*Physocarpus*, *Comptonia*) are now found far to the south of the forest-tundra zone; others (*Dryas*, *Oxyria digyna*) are typical of present arctic sites. Fossils of insects are abundant and include several that represent extinct species (Matthews, 1977).

Age: The most recent Sr isotope dates on *Arctica* suggest a Pliocene age (K. Miller, pers. comm., 1989) and the foraminifera an early Pliocene age (D.H. McNeil, pers. comm., 1988). Both estimates are younger than a previously published late Miocene age (Matthews, 1989).

Comments: The marine unit is tectonically disturbed, yet it probably represents an eustatic high sea level stand of some tens of meters. It may correlate with the Beringian transgression in Alaska (fig. 2). If so, a low Alle/Ile ratio on *Arctica* (Brigham-Grette and others, 1987) would be explained. Ice-wedge pseudomorphs in the marine sediments suggest existence of permafrost shortly after deposition of the marine unit.

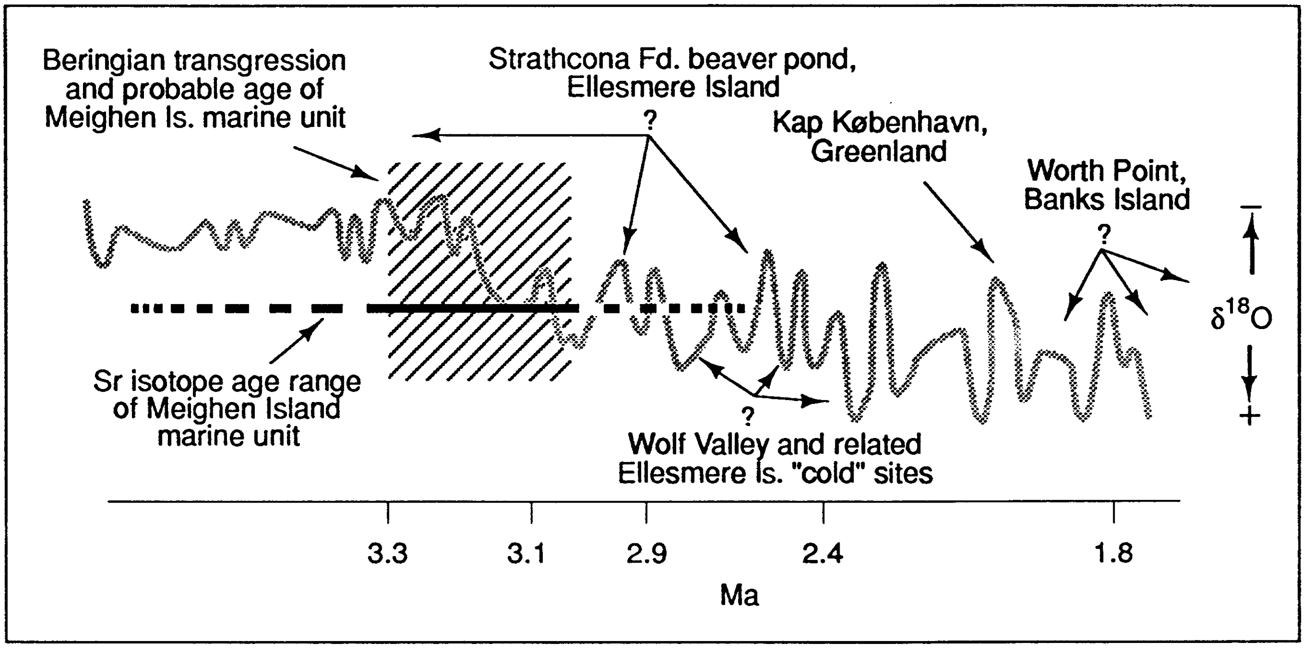

Figure 2. Possible correlation of northern Canada and Greenland Pliocene and Quaternary sites to benthic oxygen isotope record from equatorial Pacific core V28-179. Oxygen isotope record after Shackleton and Opdyke, 1977.

High level alluvial deposits, Ellesmere Island (Queen Elizabeth Islands, Northwest Territories)

Stratigraphy: Parts of Ellesmere Island are characterized by an elevated, dissected erosional surface (or surfaces) capped by a few to tens of meters of alluvial gravel with allochthonous and autochthonous organic zones. One of the latter, located near the head of Strathcona Fiord (fig. 1b) at an elevation of approximately 400m, represents the semi-autochthonous debris of a small beaver pond. Another is a continuous peat horizon exposed in Wolf valley, near the Fosheim Peninsula (fig. 1b).

Paleontology: Fossils from the beaver pond and associated deposits represent several types of trees, among them an extinct species of *Larix* sect. *Multiseriales* (O. Bennike, pers. comm., 1989), spruce, *Thuja occidentalis*, and at least one type of five-needle pine. Shrub and herb fossils include those of the extinct plants *Myrica* cf. *eogale*, *Aracites*, the extant sedge, *Scirpus microcarpus*, and the shrubs *Comptonia* and *Physocarpus*. The insect fauna is diverse, has a tree line character and is largely extant. In contrast with the beaver pond site, the fossils from the Wolf Valley peat (fig. 1b) represent a less diverse fauna and flora (table 1). The latter lacks extralimital shrubs like *Physocarpus* and *Comptonia*, and contains only rare conifer fossils (mainly wood, possibly rebedded).

Age: Some elements of the flora suggest the beaver pond is intermediate in age between Meighen Island and Kap København; however, the fossils of two other plants (*Paliurus* type and *Decodon*) from an adjacent locality may mean that the beaver pond sediments actually predate Meighen Island (Fyles, 1989; fig. 2) since neither plant has been seen in Meighen Island deposits. Vertebrate fossils occur in the beaver pond peat and represent a beaver that is an extict taxon similar to *Dipoides* (C. R. Harington, pers. comm.), rather than the modern *Castor*.

Comments: The beaver pond was probably surrounded by open larch woodland near regional tree line. The Wolf valley deposit, with an impoverished "modern" flora/fauna, appears to represent a cooler climate (fig. 2), though one warmer than at present. As indicated in figure 2, it may correspond to one of the cool intervals between 3.3 and 2.5 Ma.

Kap København Formation, Kap København (Greenland)

Stratigraphy: At Kap København (fig. 1a), fine grained, laminated marine sediments (Member A) are overlain unconformably by Member B, which consists of shallow marine sand (B1), marine silt (B2) with occasional dropstones, and sand (B3) with organic debris and wood.

Paleontology: Member B3 has yielded a number of plants and insects, many of which do not occur in Greenland today (Bennike and Böcher, in press). Among the trees are the same extinct species of *Larix* sect. *Multiseriales* (Bennike, in press) seen at the Ellesmere Island beaver pond site, *Thuja*, black spruce and *Taxus*. Shrubs and herbs include *Myrica* cf. *eogale* and *Scirpus microcarpus*.

Age: The molluscan fauna is comprised primarily of Pacific taxa, hence the deposits postdate opening of the Bering Straits, represented by the Beringian transgression(s) at around 3 to 3.3 Ma. *Cibicides grossus* is present in Member A. Paleomagnetic analyses suggest correlation with the lower part of the Matuyama chron, while fossils of two types of lagomorphs in unit B3 call for an age of about 2 Ma. (Bennike, in press).

Comments: The Kap København site is important partly because of its rich flora and fauna and partly because it is dated by several lines of evidence. In contrast with Meighen Island and the beaver pond site on Ellesmere Island, macrofossils of pine have not been found and judging from scant pollen data, pine was not growing in the Kap København region two million years ago.

DISCUSSION

Table 1 shows the occurrence of selected plant fossils from the sites discussed in this abstract. The arrangement of plants indicates floral distinctions which may be of chronologic significance. For example, note in the table that the Wolf Valley and Worth Point floras appear to consist entirely of extant species, which is verified when the entire floral list is examined. Note also, that only the beaver pond site contains *Paliurus* and *Decodon*, two taxa seen at Beaufort Formation sites on Prince Patrick Island, Banks Island, and at other Tertiary sites, but not at Meighen Island.

Except for the Worth Point site, the sites discussed here probably record climatic fluctuations in the high Arctic region during the Pliocene, possibly only during the late Pliocene. Although some evidence suggests it is early Pliocene in age, the Meighen Island marine unit may be correlative with the late Pliocene Beringian transgression. The overlying terrestrial deposits may represent the last Pliocene warm interval before start of the major climatic downturn of the late Pliocene (fig. 2). Kap København Member B3 deposits correspond to one of the warm intervals at around 2 Ma. The "modern" character of the Worth Point flora and fauna suggests that the Worth Point Deposits at Worth Point probably post date Kap København and hence are of early Quaternary age (fig. 2).

Chronologic placement of the various Ellesmere Island high level alluvium sites is more speculative because it is based solely on comparison of macrofloras. Some sites may represent the relatively warm peaks, whereas others represent the cold troughs in the marine isotopic record between 3.3 Ma and 2 Ma (fig. 2). The Wolf Valley flora is similar to that from Worth Point, but its stratigraphic context suggests it is older than than Worth Point.

There is hope that some of these dating problems will be solved by future analyses. A great need exists for paleomagnetic studies at Meighen Island and the Ellesmere Island sites. More complete studies of the marine microfauna and macrofauna in the Meighen Island marine unit and a concerted search for vertebrates will likely provide important new information on the age of the Beaufort Formation. Vertebrate fossils have already been found in the Ellesmere Island high level alluvium, and some of these undoubtedly will become important for dating those sites. However, it is already apparent from study of plant macrofossils that the high level alluvium is not all of the same age.

This is Geological Survey of Canada Contribution 33589.

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Potential of $^{87}Sr/^{86}Sr$ for Dating Pliocene Carbonates

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Strontium isotope ratios ($^{87}Sr/^{86}Sr$) show promise for high resolution stratigraphy and correlation of marine carbonates during much of the Cenozoic (e.g., DePaolo and Ingram, 1985; DePaolo, 1986). Successful application of the technique depends on establishing a standard or calibration curve. Strontium isotopes are measured in samples of known age to establish $^{87}Sr/^{86}Sr$ of marine carbonates, and thus presumably mean sea-water, through geologic time. Once a standard is constructed, $^{87}Sr/^{86}Sr$ can be used for high-resolution dating of marine carbonates of unknown age during intervals when the slope of the curve is steep and monotonic. As part of our study of Pliocene paleoceanography and climate, we have examined the potential of $^{87}Sr/^{86}Sr$ for high-resolution dating during the Pliocene.

A strontium isotope standard for the Pliocene has been compiled by Hodell and others (1989). The calibration curve is a composite of analyses from 5 different Deep Sea Drilling Project (DSDP) sites located in the South Pacific Ocean (DSDP 588, 590, and 593), the South Atlantic Ocean (DSDP 519), and the Caribbean Sea (DSDP 502). The composite calibration shows a great deal of scatter suggesting that $^{87}Sr/^{86}Sr$ has limited use for high-resolution stratigraphy during much or all of the Pliocene. We have found that ages of samples from the South Pacific Sites used to construct the standard are not well constrained. For example, dating of samples from DSDP 588 relied heavily on paleomagnetic data, but intensities of published paleomagnetic measurements throughout most of the Pliocene sediments from DSDP 588 are too low to be reliable (Barton and Bloemendal, 1986). Dating of samples from DSDP 590 rely on correlation to the suspect paleomagnetic record of DSDP 588. Samples from DSDP 593 were dated with limited biostratigraphic data and dating is further complicated by very poor recovery in several cores from the Pliocene sequence. We judge that analyses from South Pacific DSDP Sites 588, 590, and 593 should not be used to construct a standard.

Data used by Hodell and others (1989) from DSDP Sites 519 and 502 are shown separately in Figures 1a and 1b. Both data sets show a linear relationship between $^{87}Sr/^{86}Sr$ and age, although Site 519 data show a great deal of scatter. Regression lines for both data sets have very similar slopes, but they are offset. An adjustment of $3x10^{-5}$ in isotopic ratios or 800,000 years in the time scale is necessary to make the regression lines coincide. We think it is unlikely that dating of either core is in error by 800,000 years because reliable biostratigraphic and paleomagnetic control are available for both. Note also that several replicate analyses at Site 519 give very different values. For example, replicate analyses from the 3.4 Ma sample differ by $6x10^{-5}$. Therefore, the difference between the DSDP 519 and 502 data and scatter within each data set result from variance within the strontium data, not age assignments.

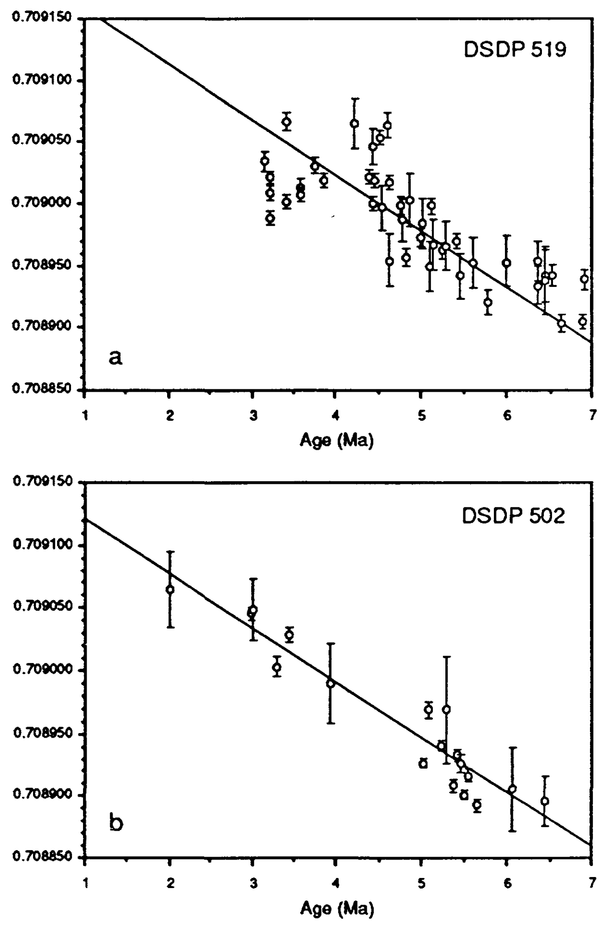

Figure 1. $^{87}Sr/^{86}Sr$ data for DSDP Sites 519 (a), and 502 (b). Data are from Hodell and others, 1989; table 1.

We are in the initial stages of generating a $^{87}Sr/^{86}Sr$ standard from DSDP Site 502 using monospecific samples of planktic foraminifers. Our preliminary results indicate different species from the same sample yield equivalent $^{87}Sr/^{86}Sr$. Similarly, different size fractions of the same species from a sample yield equivalent $^{87}Sr/^{86}Sr$. Our preliminary data also show a linear relationship between $^{87}Sr/^{86}Sr$ and age. However, our results are consistently offset from the Hodell and others' (1989) 502 data by about $5x10^{-5}$ (after adjusting all data to $^{87}Sr/^{86}Sr$ of NBS 987= .710250). Differences between our DSDP 502 results and the results of Hodell and others (1989) could be caused by interlaboratory calibration problems. Interlaboratory variation cannot be used to explain the differences between analysis of 502 and 519 that were done by Hodell and others (1989).

Our evaluation indicates $^{87}Sr/^{86}Sr$ must be used with caution for dating and correlating within the Pliocene until a better controlled calibration curve is constructed and tested. Problems may exist because of the limited data now available to construct a Pliocene standard. Alternatively, the change in $^{87}Sr/^{86}Sr$ during the Pliocene may be too small or complicated to be useful for dating given current analytical capability.

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Pliocene Atmospheric CO_2: Some Preliminary Observations

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It is not yet possible to estimate Pliocene atmospheric CO_2 concentrations with any degree of confidence. However, the deep-sea sediment record suggests that the Pliocene marine carbon cycle may have fluctuated in a manner similar to the late Pleistocene changes that have been linked to atmospheric CO_2 variations recorded in ice cores (Curry and Miller, 1989). Moreover, geochemical models (Budyko and Ronov, 1979; Berner and others, 1983; Lasaga and others, 1985) have simulated a long-term Cenozoic trend of decreasing CO_2 levels, suggesting the possibility of generally higher levels a few million years ago. Because these possible variations span such a broad spectrum of time scales and processes, the Pliocene represents a unique challenge and opportunity in carbon-cycle geochemistry.

It is useful to consider Pliocene mechanisms of CO_2 change as being of two types based on associations of particular processes and time scales. These two modes of change may be designated as intensive ("Pleistocene") and extensive ("Cenozoic"). Intensive changes are defined as redistributions of carbon within the system composed of the atmosphere, the oceans, the terrestrial biosphere and soils, and the marine carbonate sediments capable of buffering reactions on time scales of thousands to tens of thousands of years. Extensive changes are defined as additions or withdrawals of carbon and/or nutrients to or from the intensive system.

Intensive CO_2 changes, as defined here, are best exemplified by the fluctuations in atmospheric CO_2, deep-sea carbonate dissolution, and the distribution of carbon isotopes in oceanic dissolved carbon documented by the ice and sediment core records of the late Pleistocene. These changes generally occurred over time scales of thousands to tens of thousands of years, although the possibility of more rapid changes (perhaps on the order of decades) has been suggested (Broecker and others, 1985). Over these time scales, the intensive system can be considered a geochemically closed system. Marine carbonate sediments act primarily as a buffering mechanism with respect to changes in deep-sea chemistry. Because the amount of carbon available for redistribution is limited by constraints within the closed intensive system, intensive changes in atmospheric CO2 are not likely to exceed a factor of two (Sundquist, 1986).

It is theoretically possible for extensive changes in atmospheric CO_2 to be much larger. Through the processes of weathering, sedimentation, volcanism, and metamorphism, extensive changes can transfer carbon between the intensive system and the much larger carbon reservoirs of the earth's crust. Time scales on the order of hundreds of thousands to millions of years are necessary for these processes to affect atmospheric CO_2. The potential magnitude of such changes is exemplified by model calculations suggesting atmospheric CO_2 levels 13 times higher than present during the late Cretaceous (Lasaga and others, 1985). Over these long time scales, the sedimentation of marine carbonates is controlled primarily by mass and charge balances relative to global riverine fluxes to the oceans, rather than by responses to deep-ocean chemistry. It is also possible for extensive changes to influence atmospheric CO_2 through transfer of nutrients rather than carbon. For example, extensive changes in the oceanic phosphate reservoir could alter the distribution of CO_2 between the atmosphere and the nutrient-enriched deep sea.

The only published CO_2 estimates applicable to the Pliocene are the geochemical model results (referenced above) suggesting a long-term Cenozoic trend of decreasing atmospheric concentrations. These estimates must be viewed as long-term general approximations with limited relevance to a period as specific as the Pliocene. The models incorporate many very uncertain assumptions, and their application is limited by the very coarse stratigraphic resolution used in formulating inputs and boundary conditions.

Another approach to estimating long-term CO_2 trends is to calculate constraints on atmospheric CO_2 concentrations from the range of carbon isotope ratios observed in carbonate shells from coeval sediment samples. This approach, first suggested by Shackleton (1985) and Arthur and others (1985), is based on the reasonable hypothesis that the maximum range of carbon isotope ratios in oceanic dissolved inorganic carbon (DIC) should be inversely proportional to the average oceanic DIC concentration. Because extensive changes in atmospheric CO_2 must be associated with changes in oceanic DIC, any constraint on oceanic DIC also constrains atmospheric CO_2. However, Berger and Spitzy (1988) have shown that quantitative application of this constraint requires additional assumptions about oceanic alkalinity and circulation.

Both the long-term geochemical model estimates and the carbon isotope constraints apply only to extensive rather than intensive CO_2 changes. During the Pliocene, both types of changes may have occurred simultaneously, and the magnitudes of their respective effects on atmospheric CO_2 may have been comparable. Therefore, it is important to develop new models and constraints that address the complex interactions that controlled Pliocene CO_2 levels. One approach is to implement models of the intensive carbon-cycle system using a range of boundary and initial conditions representative of possible extensive changes. Another approach is to apply the carbon-isotope-range constraint to temporal intensive variations as well as spatial variations. The development of improved constraints using this approach will require a refined understanding of Pliocene oceanic alkalinity distributions and circulation patterns. Therefore, these methods will necessitate full utilization of the rapidly improving understanding of the Pliocene sediment record.

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Estimates of Pliocene Precipitation and Temperature Based On Multivariate Analysis of Leaf Physiognomy

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Many physiognomic characters of angiosperm (flowering plant) leaves are probably genetically controlled adaptations to environment and/or ontogenetic responses to environment (Bailey and Sinnott, 1916; Richards, 1952; Givnish, 1986). General correlations between some characters and particular environmental parameters (for example, between the percentage of entire-margined species and mean annual temperature; Wolfe, 1979) can be demonstrated and applied to obtain generalized estimates of paleoclimate (Wolfe and Poore, 1981). But for a variety of reasons, the precision and accuracy of these estimates can be debated (Dolph and Dilcher, 1979). These univariate correlations, moreover, fail to consider how particular physiognomic characters might be affected by other environmental factors. For example, while large leaf-size is considered to indicate high moisture levels, small leaf-size can result from either little moisture, low temperature, or combinations of both factors. Therefore, refinement of methodology is needed to obtain accurate and precise paleoclimatic inferences from fossil leaf assemblages.

The sampling of leaves in vegetation of known climatic parameters is in progress. The 44 samples obtained so far represent vegetation ranging from megathermal rain forests to megathermal desert to subalpine microthermal forests (megathermal = mean annual temperature (MAT) ≥20°C; mesothermal = MAT 13 to 20°C; microthermal = MAT <13°C.) Physiognomic characters are tabulated for each species in each sample; the percentages of each of the 34 character-states for each sample become the "species" in a correspondence analysis (Hill, 1973; 1979). Correspondence analysis compares the patterns of distribution of the physiognomic characters and produces scores for each character and sample on two or more axes (see Wolfe, in press, for description of methodology).

Correspondence analysis of modern samples indicates that about 62% of total variance in physiognomy is accounted for on the first two axes. The resulting sample plots (fig. 1) group together samples of similar physiognomy. The scores on axis I are readily apparent as representing optimal growing conditions: megathermal rain forest samples have high scores and moist microthermal and dry mesothermal samples have low scores. Axis II, on the other hand, can be readily interpreted as water stress: Sonoran Desert samples have high scores and humid to mesic forest samples have low scores. This analysis thus provides an objective method of comparing fossil leaf assemblages with modern vegetation.

The accuracy of the sample scores relative to climatic parameters can be evaluated by including a given climatic parameter as a physiognomic character in the database. For example, when mean annual temperature for each sample site is included, the relation of mean annual tempertature (MAT) to samples is shown by treating MAT as a loading vector and orthogonally projecting each sample score to this vector. A very general correspondence of sample scores to MAT results if all samples are included in the ranking. However, if dry vegetation samples are calibrated separately from humid to mesic vegetation samples, the two rankings result in calibrations that are accurate to ±0.5°C. The ranking of dry samples on the mean annual precipitation (MAP) vector results in a calibration accurate to ±3cm (fig. 2).

Both MAT and MAP produce moderately high scores relative to physiognomic characters (fig. 3). Runs with these climatic factors slightly raise eigenvalues and do not alter sample relations based only on physiognomy. This suggests that both MAT and MAP have strong relations to physiognomy. On the other hand, introduction of factors such as warm month mean temperature produces lower eigenvalues and very low scores, suggesting little correspondence to physiognomy.

Application of this technique (CLAMP: Climate-Leaf Analysis Multivariate Program) to late Pliocene leaf assemblages in western North America allows estimates of Pliocene climatic parameters (figs. 1, 2).

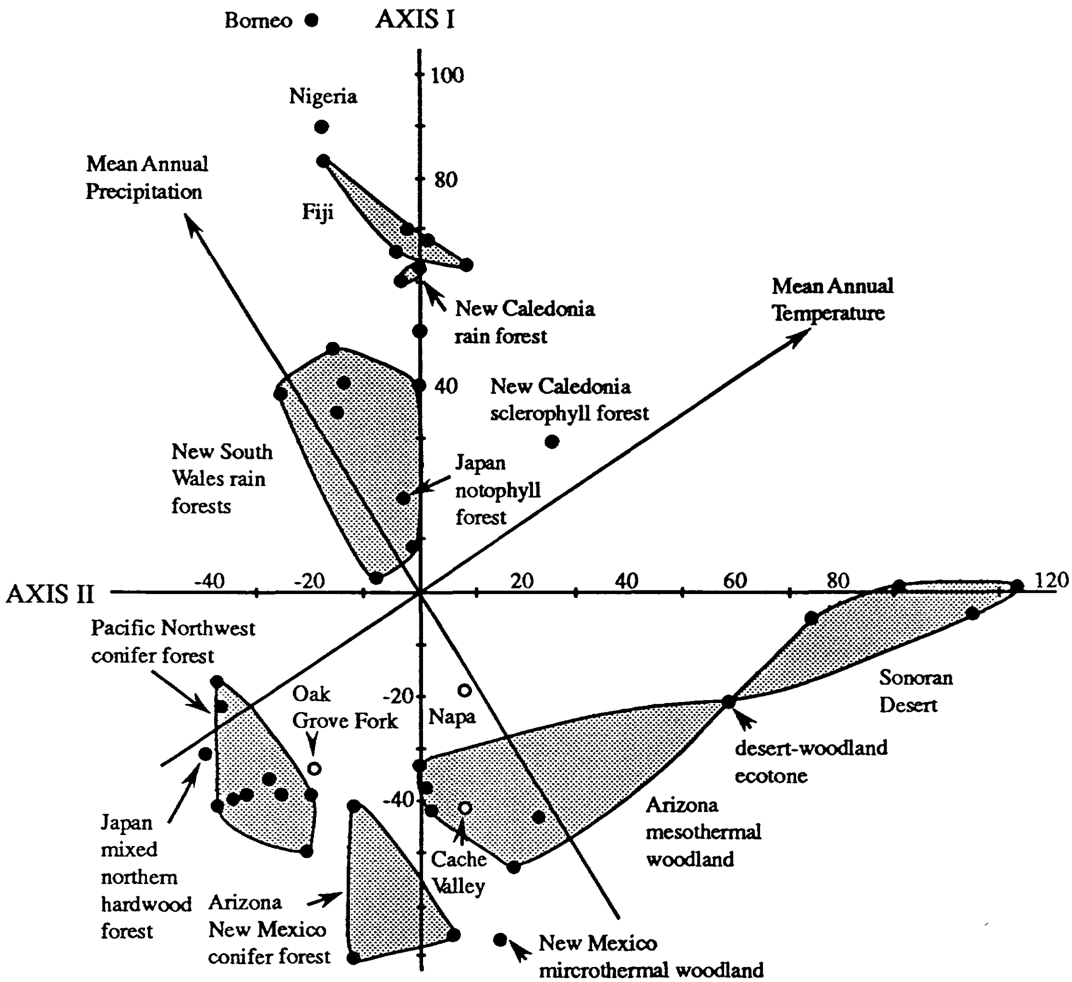

Figure 1. Sample plot based on foliar physiognomic characters. Based on collections made in 1988 and 1989, except for Borneo and Nigeria samples, which are based on Richards (1952, figs. 13 and 14). The average number of species/sample is 22. Also plotted are three late Pliocene leaf assemblages. Shown as vectors are mean annual temperature (MAT) and mean annual precipitaion (MAP) from figure 3. Open circle indicates Pliocene sites; closed circle indicates modern sites.

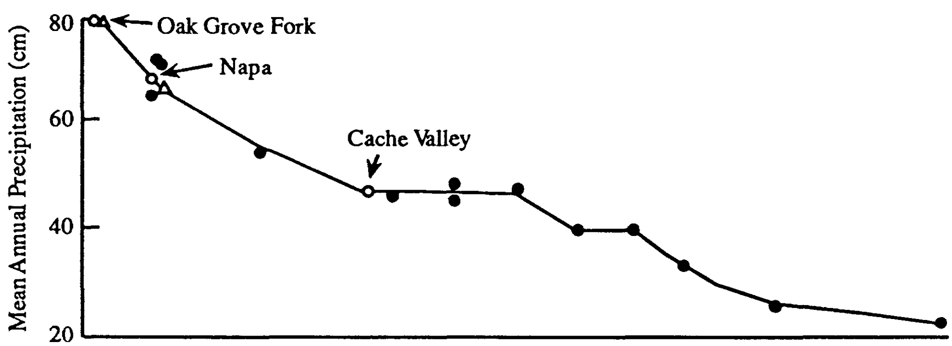

Figure 2. Calibration of dry vegitation samples with mean annual precipitation. Horizontal axis represents relative distance between sample scores when projected orthoganally onto MAP vector in figure 1. The plot is used to suggest values for three late Pliocene leaf assemblages. Open circle indicates Pliocene sites; closed circle indicates modern sites; triangles indicate two dry northwest sites.

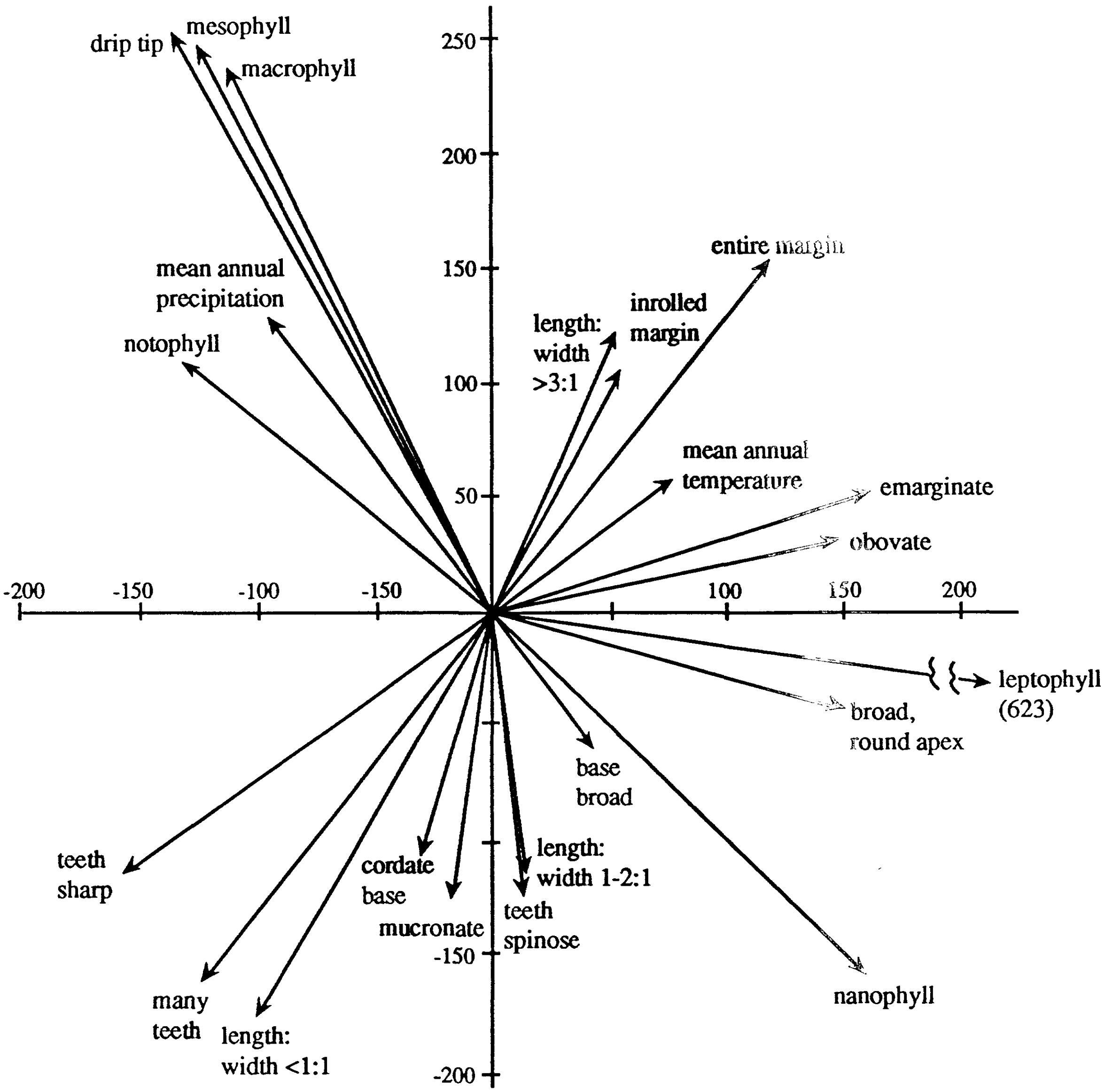

Figure 3. Physiognomic character plot based on samples shown in figure 1; characters with low scores on both axes I and II not shown. Characters are shown as vectors extended to score points. Introduction of values of mean annual temperature (MAT) and mean annual precipitation (MAP) minimally alter (typically no more than 1 to 2 units) character scores. Comparison of figures 1 and 3 shows the physiognomic characters that characterize the various samples.

The Napa assemblage of central California (Axelrod, 1950), the Cache Valley assemblage of northern California (Brown, 1949), and the Oak Grove assemblage of northwestern Oregon (Wolfe, unpublished data) all indicate that late Pliocene MAT was higher than present by 2 to 4°C. Relative to MAP, however, the Napa and Cache Valley assemblages had higher values (10 and 6cm, respectively) than now, but the Oak Grove Fork assemblage indicates about 100cm less MAP during the late Pliocene. These analyses illustrate how regional patterns of relative temperature and precipitation can be suggested for limited time periods; only the number of leaf assemblages and the adequacy of their age controls constrain details of the patterns.

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