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Pliocene Terrestrial Environments and Data/Model Comparisons

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Contents

Illustrations

(listed by first author and figure number)

Page

Introduction: USGS Workshop on Pliocene Terrestrial Environments and Data/Model Comparisons

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Over the last several years, the U.S. Geological Survey's PRISM (Pliocene Research, Interpretation, and Synoptic Mapping Project) project has accumulated information for a "snapshot" description of global marine and terrestrial conditions for the mid Pliocene, the last period in Earth history when globally warm conditions persisted over hundreds of thousands of years. PRISM researchers and collaborators are analyzing times-series of environmental changes both marine and nonmarine settings to determine the amplitude and periodicity of Pliocene climatic changes, and are assembling gridded data sets of vegetation cover to provide boundary conditions for General Circulation Model (GCM) simulations. Palynological and other terrestrial data will also being employed in validation exercises of Pliocene climate simulations provided by researchers using Goddard Institute for Space Studies (GISS) and NCAR (National Center for Atmospheric Research) GCMs.

As part of this effort, the U.S. Geological Survey Global Change and Climate History Program sponsored a workshop on "Pliocene Terrestrial Environments and Data/Model Comparisons," which was held in Herndon, Virginia, on May 22 and 23, 1993. This report presents the abstracts from the Herndon workshop. The primary objectives of this meeting were to review the available data on mid-Pliocene terrestrial environments, and to provide a forum for geological researchers and climate modelers to discuss the uses of palynological and other terrestrial environmental data in initializing and validating GCM simulations of past climates.

Terrestrial Palynological And Paleobotanical Records Of Pliocene Age From Alaska And Yukon Territory

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A joint research project initiated in 1990 by the U.S. Geological Survey and the Geological Survey of Canada, entitled "Late Cenozoic climate history of Alaska and Yukon," has conducted three field investigations thus far (1990, 1991, 1992). Many of the deposits investigated are of middle to late Miocene age, but several are of Pliocene age. One of the Pliocene deposits investigated is the Nenana Gravel in the Alaska Range in southern central Alaska, which forms a thick mantle on the north flank of the Alaska Range. Gravel deposits provide only rare opportunities for preserving fossil assemblages, but pollen and spores have been recovered from mudstone layers within the gravels. The Nenana Gravel is important because the gravels were deposited during or immediately following the period of rapid uplift of the Alaska Range, an event that greatly influenced the climate of the interior. Current evidence indicates that the major uplift of the Alaska Range began in early Pliocene time, *ca.* 4-5 Ma. Fission-track ages on granitic rocks now exposed in core of the Alaska Range date the initial uplifting and unroofing of the crystalline core. The timing of tectonic events leading to the uplift of the Alaska Range are highly relevant to understanding the regional versus global climate influences. The development of the arcuate, east-west trending high Alaska Range in early Pliocene time significantly altered the regional climate of interior Alaska. The rising mountains increasingly limited northward penetration of moist air masses from the north Pacific from reaching the interior. The climate of the interior is now highly continental, characterized by long, severely cold winters and warm summers. Precipitation is low. At Fort Yukon, for example, the modern mean annual temperature (MAT) is -6.4° C, mean January temperature is -28° C, mean July temperature is 16.4° C, and mean annual precipitation is only 168 mm.

Pollen and spores recovered from the Nenana Gravel include *Pinus*, Picea, Larix, Tsuga*, Abies*, Betula, Alnus, Diervilla-Weigela*,* and a number of herbaceous taxa that are rare or absent from Miocene deposits from the same region (taxa marked with asterisks in this abstract no longer grow in the Alaska Range or elsewhere in interior Alaska). Some herb taxa, such as *Polemonium,* did not evolve until Pliocene time, and others appear to have played only a very minor role in the regional vegetation prior to the Pliocene. The presence during the Pliocene of taxa such as *Tsuga* and *Abies* that are now missing from the interior region suggests that a highly continental climate was not yet fully developed. MAT in the lowlands of the interior was probably near 3° C, and permafrost was absent except at higher elevations. As mentioned above, deposition of the Nenana Gravel may have begun by ca. 5 Ma (based on the uplift history of the Alaska Range) and may have continued until ca. 2.8 Ma (based on the age of the Jumbo Dome igneous intrusion that deformed deposits of this gravel unit in one area of the Alaska Range at 2.8 Ma [minimum date]).

Another apparent Pliocene gravel deposit was found in ancient terrace deposits of the Yukon River near Circle, Alaska. It contains a pollen flora that matches closely that in the Nenana Gravel. The pollen flora does not match closely any known Miocene or Pleistocene floras from the region. The plant macrofossils identified from the gravel deposits near Circle contain taxa that range in age from late Miocene to Pliocene. Paleomagnetic samples from one exposure of the terrace gravel deposits have reversed polarity, which provides some small constraint on the age of the deposit. At present, I believe that the age of the gravels is probably between ca. 5 and 3 Ma. The gravel deposits contain wood (*Picea, Abies**, *Pinus**), cones (*Picea* and an undescribed species of *Larix*), needles (*Picea, Pinus*), and seeds (*Prunus**, *Epipremnum**, *Hypericum**, *Aracites**, *Weigela**, *Aralia**, and many others). This flora represents a transitional stage between the late Miocene floras, which are relatively rich in conifer taxa but poor in herbaceous taxa, and Quaternary interglacial boreal forest floras which are relatively depauperate in conifer and broadleaf taxa but rich in herbaceous taxa.

Another informative site is at Ch'ijee's Bluff, an exposure on the Porcupine River in northern Yukon. Unit 1 at the base of the bluff contains a flora that is probably mid-Pliocene age (ca. 3.5-2.5 Ma), but this age assignment is based mostly on the seed assemblage and needs to be refined through additional research. The unit 1 deposits are rich in wood, including beaver-chewed *Abies**, *Pinus** (Strobus group), *Larix*, and *Picea*. Cones in the deposit are identified as *Picea, Larix* cf. *L. minuta*, and *Pinus** (5-needle and 2-needle types). Seeds from the deposit include *Betula, Alnus, Larix**, *Comptonia**, *Sambucus**, *Carex, Sparganium*, and *Potamogeton*.

An additional very important Pliocene site is Lost Chicken Mine in east-central Alaska, where the sediments contain a tephra that has been fission-track dated at 2.9 $\pm$ 0.2 Ma (see Adam, this volume). This late Pliocene deposit contains a proto-boreal forest fossil assemblage of *Picea, Larix* (not *L. laricina*, the only larch species in Alaska today), *Pinus**, *Abies*?*, *Corylus**, *Betula, Alnus*, and Ericaceae. In contrast to the older Pliocene floras from the Nenana Gravel and the terrace gravels near Circle, the Lost Chicken Mine site has yielded only a sparse herbaceous flora. This may reflect a dense forest cover, with few openings for herbaceous taxa, and perhaps a mossy ground cover. The Lost Chicken flora lacks *Tsuga*, in contrast to late Miocene and older Pliocene floras from interior Alaska. The climate was warmer than the present (Holocene) interglacial in Alaska. MAT was probably ca. 2-3° C, vs. ca. -7° C today at the nearby community of Chicken.

On the North Slope of Alaska, the Fish Creek site provides some information on the early development of tundra or tundra-like vegetation during the late Pliocene, probably about 2.4 Ma. Tree and shrub elements of the older Fish Creek pollen flora include *Picea**, *Pinus**, *Larix**, *Abies**, *Betula, Alnus**, *Salix*, and Ericaceae. Herbs are well-represented, and include Cyperaceae (most abundant), Gramineae, *Artemisia*, Compositae (Tubuliflorae), Caryophyllaceae, and *Valeriana*. Pollen data within the Fish Creek site records a progressive loss of trees and shrubs, reflecting onset of colder glacial climates in northern Alaska.

The Pliocene history of climate and ecological changes in Alaska and northern Canada are still poorly understood, although progress is being made to identify and investigate such deposits. One of the major problems inhibiting progress has been inadequate age control for some of the most important fossil localities. Because the precise biostratigraphic ranges of many of the Pliocene taxa from the region are poorly known, the fossil assemblages alone can only provide broad-range age assignments. The best hope of reconstructing a detailed Pliocene terrestrial record from the region may be from the drilling project planned for 1994 at Fort Yukon, in interior Alaska. At that location we may obtain a long Neogene record (about 15 Ma) in about 300 meters of sediment core. A water well log from Fort Yukon suggests that the subsurface sediments are mostly lacustrine.

Palynological Records Of The Pliocene In Baffin Bay And Labrador Sea

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Pliocene deposits are virtually absent in eastern Canada, a region marked by strong glacial erosion during the late Quaternary. However, drilling in the adjacent marine basins, the Labrador Sea and the Baffin Bay (Leg 105 of the Ocean Drilling Program), has led to the recovery of nearly continuous Pliocene sequences. The palynology of these Pliocene sequences provides information about terrestrial climate on adjacent land (pollen and spores) and sea-surface conditions (dinoflagellate cysts and prasinophyte phycoma).

The palynological record of Labrador Sea is from the ODP Site 646 located on the southwest Greenland rise. The chronostratigraphy is well constrained by the marine microfossil biostratigraphy and the magnetostratigraphy. Sedimentation rates appear relatively uniform throughout the sequence and average ~9 cm/ka. Palynological analyses were performed in the Plio-Pleistocene sediments (480-0 m; 5.4-0 Ma) at 1.5 m intervals, which provided a time resolution of ca. 16 kyr. The Pliocene deposits are characterized by abundant prasinophytes (*Cymatiosphaera*) and dinoflagellate cysts (Protoperidiniales and Gonyaulacales). In the Gauss chronozone (3.4-2.47 Ma), the marine palynomorph assemblages reflect relatively low salinity (varying from 32 to 35‰), likely due to freshwater runoff, and cool conditions (*ca.* 8 to 11°C in summer) rather similar to present. Most samples are characterized by abundant pollen (>1000/cm^3), suggesting the existence of a dense vegetation over the source areas, the eastern Canada and Greenland. The assemblages are dominated by *Pinus* with common *Picea, Tsuga, Sciadopitys, Betula* and *Alnus,* along with Pteridophyte and *Sphagnum* spores. Such assemblages suggest influx from two main sources: (1) tree-pollen fluxes originating from a coniferous forest, where cool temperate and humid conditions prevailed according to the significant occurrence of *Sciadopitys* and *Tsuga*; and (2) inputs of shrub taxa pollen grains (*Betula, Alnus* and Ericaceae) originating from an open vegetation of forest tundra and shrub tundra related to a humid subarctic climate. The pollen and spores content of sediments probably relate to atmospheric influxes from southeastern Canada, where temperate conditions prevailed, in addition to fluvial inputs from southern Greenland, where subarctic climate can be inferred. The Pliocene interval, including the Gauss chronozone, is marked by small scale fluctuations in concentrations and taxa percentages, indicating changes in vegetational cover, hydrological conditions over southern Greenland and/or atmospheric trends across the Labrador Sea. A major shift in the Pliocene palynological record of the Labrador Sea occurs at about 2.5 Ma, coinciding with the first regional input of ice-rafted debris and the initiation of glaciations about the North Atlantic Ocean—a drastic drop in pollen and spore concentrations is accompanied by the decline of *Sciadopitys* and *Tsuga,* which indicates impoverishment of the vegetational cover and significant cooling in the source areas.

In the central Baffin Bay, the drilling of 1200 m of sediment at ODP Site 645 revealed a sequence spanning the entire Neogene. The marine microfossil records are poor, notably because of sparse productivity and because of biogenic carbonate dissolution. Nevertheless, the magnetostratigraphy and a few biostratigraphic datums allowed to set a reasonable chronostratigraphy. Sedimentation rates throughout the Plio-Pleistocene are relatively high, (13.5 cm/kyr). Palynological analyses were performed at 1.5 to 3.0 m interval in Pliocene sediments. Above the Gilbert-Gauss boundary (3.4 Ma), the marine palynomorph assemblages are dominated by prasinophytes and protoperidiniales, which indicate a low salinity (<30‰), estuarine type environment. Abundant pollen and spores (up to $50,000/cm^3$) were recovered in most Pliocene samples, which indicates a dense vegetational cover in adjacent terrestrial areas, the Baffin Island and Greenland. The assemblages are characterized by both coniferous trees (*Pinus* and *Picea*) and non-arboreal (*Alnus, Betula,* Ericaceae, *Sphagnum*) components. Such assemblages closely resemble those of the modern Hudson Bay: they suggest the existence of coniferous boreal forest to shrub tundra vegetation with extensive peatlands in terrestrial regions surrounding Baffin Bay. During the Pliocene, the transition between the subarctic and low Arctic bioclimatic zones was probably located in the southern Baffin Bay region, 5^{o} to 7^{o} north of the present limit. The Gauss interval is marked by fluctuations in pollen and spore concentrations and percentages. These variations may reflect changes in vegetation source, fluvial inputs, atmospheric trends and/or sedimentary processes. They probably respond to climatic fluctuations, which may also be related to recurrent growth and decay of ice caps in circum-Arctic areas as shown by ice-rafting deposition in Baffin Bay. However, strong increase in ice-rafting deposition related to major glaciation is dated back to *ca*. 2.5 Ma.

Pliocene Pollen Data Set Dynamics: Tulelake, California, And Lost Chicken Mine, Alaska

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Pollen data sets of Pliocene age from continental sites are relatively rare. Pollen work with a paleoecological slant has focused primarily on deposits of Holocene and Upper Quaternary age, which are common and readily sampled in many regions. Stratigraphic palynologists, in contrast, have focused on pre-Pliocene deposits; most of the pollen grains encountered in Pliocene deposits cannot be distinguished from modern forms, and the simple presence or absence of particular forms is not as informative as it is for older deposits.

Although many or most of the pollen types important in Pliocene deposits represent plants still extant, the associations formed by those plants in the past may well have been different than those of today. Even if Pliocene pollen assemblages have modern analogs, those analogs often represent geographical ranges far removed from the Pliocene sites. For example, poorly-preserved pollen samples from the Pliocene Santa Clara Formation near San Francisco contain up to 50% spruce (*Picea*) pollen; spruce does not occur in the area today (Adam *et al.*, 1983).

Pollen records from the Pliocene often span rather long time intervals; the longer the interval under study, the more work is required to achieve fine time resolution. Time series are thus often severely under sampled compared to Holocene studies, with sampling intervals measured in millennia rather than centuries or decades. Pollen zones then tend to be characterized not only by the frequencies of the various pollen types within them, but also by the nature of the variability within and between types.

Because pollen assemblages from Pliocene deposits often suggest climatic conditions significantly different from those of today, it is important to understand not only the nature of the assemblages, but their underlying dynamic structure. Assuming that the sequence of pollen samples through time is known, the dynamics can be investigated by plotting the samples in stratigraphic order in some appropriate phase space, chosen so as to clarify the patterns present in the data. This approach to past vegetation dynamics is illustrated below using data from Tulelake, California, and Lost Chicken mine, Alaska.

Tulelake

The Tulelake record is from a 334-m core that spans the past 3 Myr (Adam *et al.*, 1989, 1990). Initial inspection of the pollen diagram suggested that the behavior of pine pollen vs. TCT (Taxodiaceae, Cupressaceae, and Taxaceae) pollen varied through time in a significant way. A plot of pine *vs.* TCT pollen for the entire section (6A) was not helpful, but a plot of the Pliocene part of the core (6B) indicated a well-developed pattern of variability between the two types. When stratigraphically adjacent samples are connected with lines (Figure 1C), the nature of the underlying dynamics of the data set emerge. Nearly all of the connecting lines have a negative slope, indicating that when pine increases, TCT decreases and *vice versa.* Only rarely do both types increase or decrease together.

Comparison of the pattern of pine *vs.* TCT pollen discussed above with modern data from central California suggested that the late Pliocene climate at Tulelake had much in common with the present climate at middle elevations along the western slope of the Sierra Nevada near Yosemite, and with the climate that prevailed at lower elevations in the northern California Coast Ranges during the cooler parts of deep-sea oxygen isotope Stage 5 (Adam *et al.*, 1990). Comparison of the various sites was greatly facilitated by the phase-plot technique illustrated here.

Lost Chicken Mine

The Lost Chicken Mine is a working placer mine along the Taylor Highway in east-central Alaska (see also Ager, this volume). It includes exposures of plant- and mammal-bearing fluvial sediments and peats of upper Pliocene age, as indicated by the Lost Chicken tephra layer, with a glass isothermal fission-track age of 2.9±0.2 Ma (John Westgate, oral communication). Thirty-seven pollen samples were analyzed from various stratigraphic units; age relationships between samples are sometimes clear but sometimes inferred. The Pliocene samples apparently represent a warm interval at or near the end of the Gauss Normal paleomagnetic chron; the Holocene samples all predate the White River Ash, which has an age of 1200 years (Péwé, 1975).

The pollen counts were reduced to a condensed data set and subjected to a detrended correspondence analysis (DCA; Gauch, 1982). The DCA reduced the data set to four primary orthogonal axes that summarize the variability within the data. Because each axis reflects patterns of variability that apply to the entire data set rather than just to individual variables, shifts in the behavior of the data with respect to these axes are likely to reflect changes in the dynamics of the underlying climate system than simple plots of one taxon versus another.

The pollen samples are plotted against the first two DCA axes in 7, with samples from the same stratigraphic unit in known stratigraphic order connected by lines. The Holocene samples (29-33) are clearly set apart from the other samples, which are all Pliocene in age. In addition, inspection of the sequence of samples through time suggests that the Pliocene samples represent two separate dynamic regimes: (1) Axis 1 and Axis 2 scores positively correlated, and (2) Axis 1 and Axis 2 scores negatively correlated.

Discussion

Changes in the vegetational dynamics recorded in the Tulelake and Lost Chicken data sets are reflected in the phase plots shown in figures 1 and 2 by changes in the slopes of the lines connecting stratigraphically adjacent points. Particular dynamic regimes are represented by elongate clouds of points in the phase space. These clouds are roughly linear in the examples selected, but curvilinear clouds may also occur. The character of a cloud representing a single regime should be independent of the time intervals between samples; this property is useful in dealing with Pliocene data sets that usually lack detailed time control. Shifts in the location or orientation of the cloud are best identified through inspection of the sequence of lines connecting the points in stratigraphic order; the shifts show up as abrupt changes in slopes of the lines.

The method illustrated here is intended to augment rather than supplant standard methods of interpretation. It provides a way to characterize groups of samples that covary, and to contrast them with adjacent groups. When samples are spaced relatively far apart compared to the frequencies of the underlying climate signals, such characterizations may provide useful insights into the behavior of the record.

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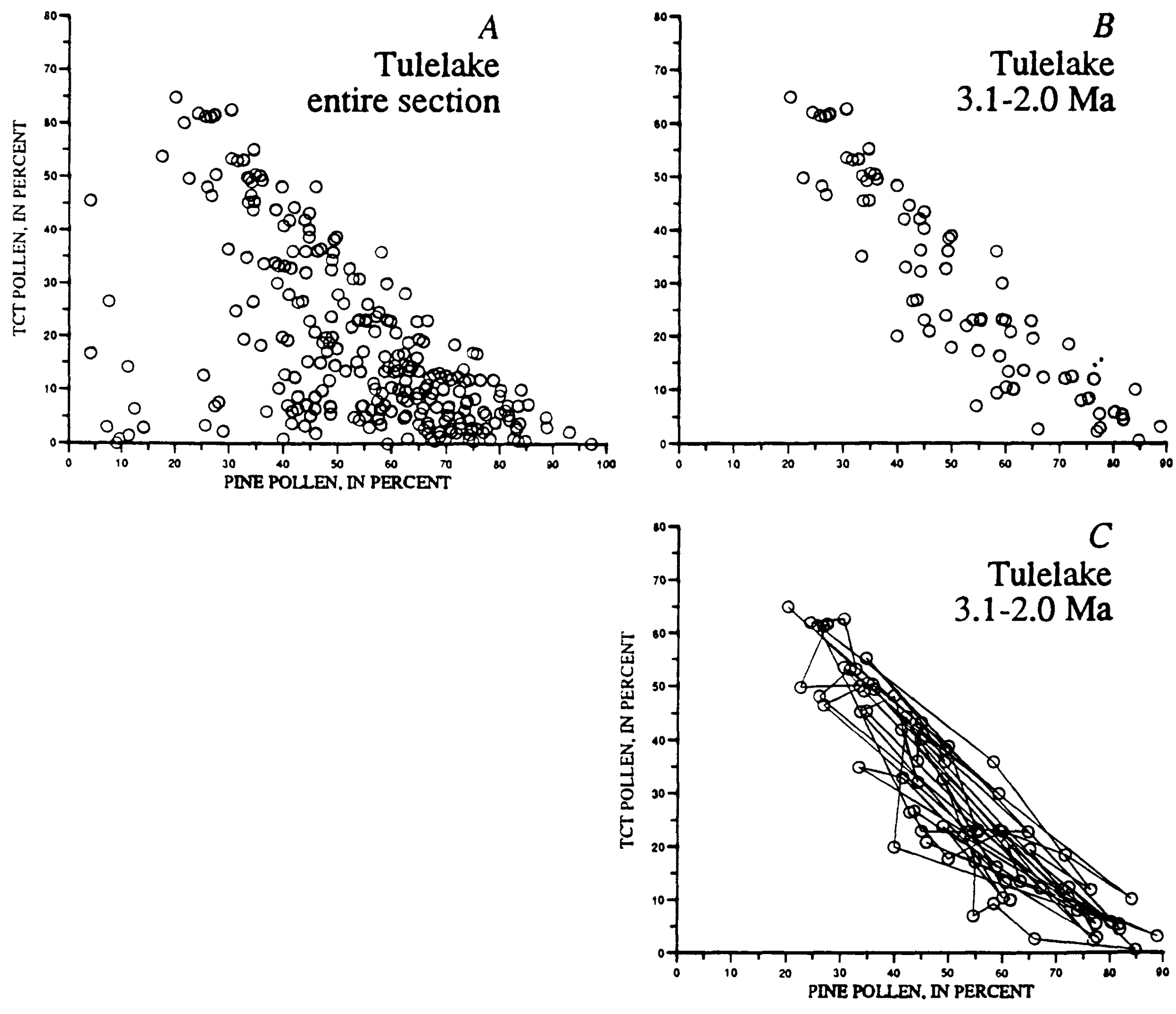

Figure 1. Plots of pine pollen percentage vs. TCT (Taxodiaceae, Cupressaceae, and Taxaceae) pollen from Tulelake, Siskiyou County, California. A) data for entire section; B) data for samples with ages from 2.0-3.1 Ma; C) same data as in B, but with stratigraphically adjacent samples connected by lines. Data are from Adam and Vagenas (1990).

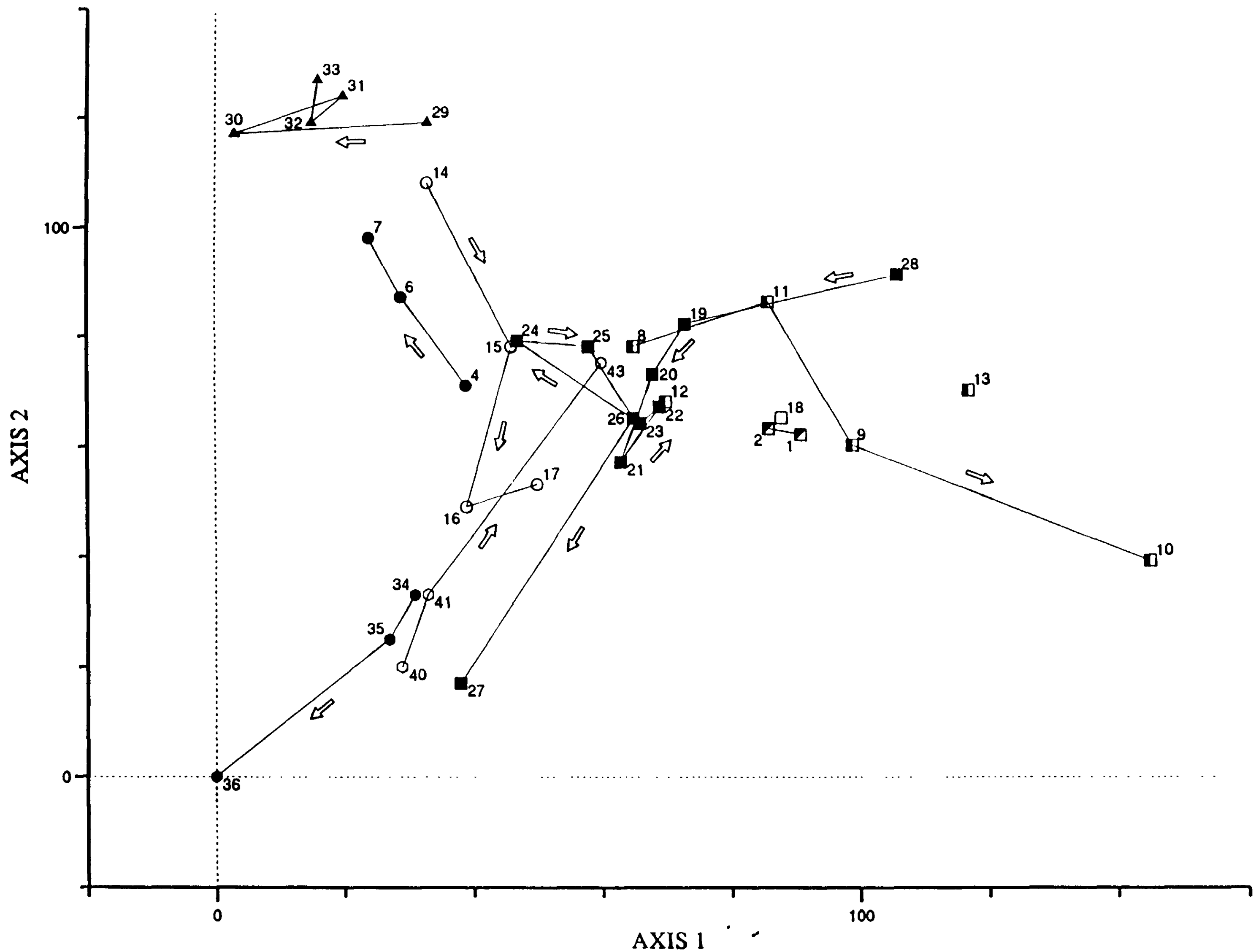

Figure 2. Plot of Lost Chicken, Alaska, pollen samples against Detrended Correspondence Analysis axis 1 and 2. Plot symbols represent various stratigraphic units, numbers are sample numbers (some numbers are "missing"), and arrows indicate direction of stratigraphic succession (older to younger).

Palynological Records From Pliocene Sediments In The California Region: Centerville Beach, DSDP Site 32, And The Anza-Borrego Desert

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Localities along the west coast of North America have been selected for palynological analysis to provide a basis for evaluating terrestrial vegetation patterns along the Pacific Coast of North America during the Pliocene. Preliminary results from three of these localities (Figure 1) demonstrate the nature of the material that is available, the types of information that can be obtained, and the problems associated with extracting paleoclimate estimates in this area.

DSDP Site 32

Leg 5 of the Deep Sea Drilling Project (DSDP) drilled Site 32 (37° 7.63' North, 127° 33.38' W) near the seaward margin of the submarine Delgada Fan off the coast of California. This fan contains a mixture of terrigenous and clastic sediments derived mostly from areas now drained by the San Joaquin and Sacramento Rivers. During the Pliocene, the Delgada Fan presumably received sediment from the same general area, although a marine embayment filled much of the Great Valley of California.

Previous biostratigraphic studies based on diatoms, planktonic foraminifera, and calcareous nannofossils provide a temporal framework for Site 32 (Barron, 1992). Samples for palynology were collected from a nearly complete sequence from about 4.5 Ma to about 3 Ma (younger Pliocene sediments were not recovered). Age estimates for these samples were determined by extrapolating from the existing biostratigraphic framework based on other microfossils.

Pollen of Taxodiaceae-Cupressaceae-Taxaceae (TCT) accounts for more than 50% of the pollen sum in most palynological assemblages from Site 32, although some assemblages are dominated by *Pinus* (e.g., at ~3.75 Ma and ~3.00 Ma; see Figure 2). *Pinus* in Pliocene sediments of Site 32 is less abundant than amounts reported by Heusser and Balsam (1977) for modern sites from the northeast Pacific Ocean. Adam *et al.* (1990) developed a model for the Tule Lake, California, core record in which increased *Pinus* and decreased TCT indicate cooler conditions in Pliocene time, though perhaps still warmer than today. Other elements of the modern California flora present in Site 32 samples include *Quercus*, *Artemisia*, Chenopodiaceae, and Asteraceae. Exotic taxa that may be relictual from the California Miocene include *Carpinus*, *Carya*, *Ilex*, *Taxodium*, *Tilia*, *Pterocarya*, and *Ulmus*.

Sequoia pollen is significantly reduced in all Site 32 samples, relative to today. One of the most important factors controlling the distribution of *Sequoia* is fog. Low amounts of *Sequoia* pollen at Site 32 may reflect climatic conditions less favorable for the formation of fog. The presence of *Taxodium* suggests that the summers may have been wetter than today.

Results of DECORANA (Detrended Correspondence Analysis) reveal patterns in the overall data matrix from Site 32 that may relate to overall paleoclimatic variability with fluctuations apparently occurring on a 200-400 k cycle. Inspection of the data matrix reveals that these fluctuations are in part related to changes in the relative abundances of mixed conifers (*Picea*, *Tsuga heterophylla*, and *Abies*). Because these taxa along the Pacific Northwest coast are controlled to a large degree by moisture, the fluctuations may reflect variations in precipitation during the Pliocene.

Preliminary analysis of the palynological data from Site 32 reveals patterns that suggest departures from modern climatic conditions during the Pliocene. Some taxa have increased or decreased relative abundances when compared to their relative abundance in modern marine core top samples. Quantification of these differences in terms of temperature and precipitation awaits acquisition of reliable modern pollen data sets from marine core tops in the northeast Pacific.

Centerville Beach

The Centerville Beach section is located in the Eel River basin of northern California (40° 32.5' N, 124° 22.5' W). This section contains approximately 1800 m of marine mudstones, siltstones, and sandstone that were deposited in the Humboldt basin along the western edge of the North American Plate.

Previous workers have developed a temporal framework for the Centerville Beach section based on microfossil biostratigraphy, paleomagnetism, and radiometric dating of volcanic ashes (Barron, 1992; McCrory, 1990). Additional paleomagnetic studies are underway to refine the magnetostratigraphy in the Gauss paleomagnetic chron. Palynological samples from the Centerville Beach section have proven to be productive; additional sampling may be necessary to obtain a complete time series and to verify preliminary results.

Palynological results from the Centerville Beach section are only preliminary but suggest that the variability observed in DSDP Site 32 may also be present at Centerville Beach. Plots of the mixed conifer group (*Picea, Tsuga heterophylla, Abies*) display fluctuations on about the same frequency as at Site 32 and appear to be approximately in phase with these. Coincidence of these patterns between Centerville Beach and Site 32 suggests that the variability may reflect a climatic, as opposed to a taphonomic, signal. Around 2.5 Ma, the mixed conifer group shows an increase in relative abundance, possibly reflecting a southern shift of the mixed conifer forests of the Pacific Northwest. This is perhaps related to the pre-Pleistocene glaciation at approximately this time.

Anza-Borrego Desert

The Anza-Borrego Desert State Park is located in the Salton Trough of southern California. Badland exposures in the park include a thick sequence (~4600 m) of Pliocene and Pleistocene sediments. Much of this sediment was derived from the ancestral Colorado River as it deposited its sediment load into the Gulf of California. Detailed paleomagnetic studies have been integrated with radiometric dates on volcanic ashes and provide an excellent temporal framework for Pliocene rocks at Anza-Borrego (Johnson *et al.*, 1983; Opdyke *et al.*, 1977).

Palynology samples were collected throughout the 2-4.5 Ma interval, with closely spaced samples above and below the 3 Ma horizon. Unfortunately, samples processed from about 3 Ma are barren or only marginally productive. However, some samples from about 3.5 Ma and older are productive. Although palynological recovery from the Anza-Borrego section is apparently inadequate for evaluating paleoclimate at 3 Ma, reworked fossils in the section may provide important insight into climatic conditions of the Colorado Plateau during the Pliocene.

Pollen assemblages recovered from the lower part of the Anza-Borrego section contain several species of Cretaceous pollen and spores. Some of these species are distinctive and have restricted biostratigraphic and paleobiogeographic ranges in the Western Interior of North America. *Proteacidites* spp. first appear in the Coniacian and range through the Maastrichtian; they are present in sedimentary rocks throughout the Western Interior. *Aquilapollenites* spp. and *Mancicorpus* spp. first appear in the Campanian and also range through the Maastrichtian; they are present in the northern part but are absent from rocks in the southernmost part of the Western Interior. The southernmost extent of the latter taxa forms a line that approximately parallels the Utah-Arizona and Colorado-New Mexico boundaries.

The stratigraphic distribution of these Cretaceous taxa in Pliocene sediments of Anza-Borrego suggests that erosion of Cretaceous rocks containing *Proteacidites*, but lacking *Aquilapollenites* and *Mancicorpus*, had begun by the early Pliocene. Erosion into Cretaceous rocks containing *Aquilapollenites* and *Mancicorpus* did not begin until about 3.7 Ma and supports the hypothesis that the cutting of the Grand Canyon occurred during the Pliocene, accompanied by extensive and rapid erosion of Cretaceous and older rocks. Cretaceous pollen and spores were transported as detritus down the Colorado River to the ancestral Gulf of California. Rapid erosion and transport of this amount of material requires significantly increased precipitation on the Colorado Plateau during the Pliocene. This is consistent with climate models involving wetter summers and wetter winters induced by relatively recent uplift of the Colorado Plateau.

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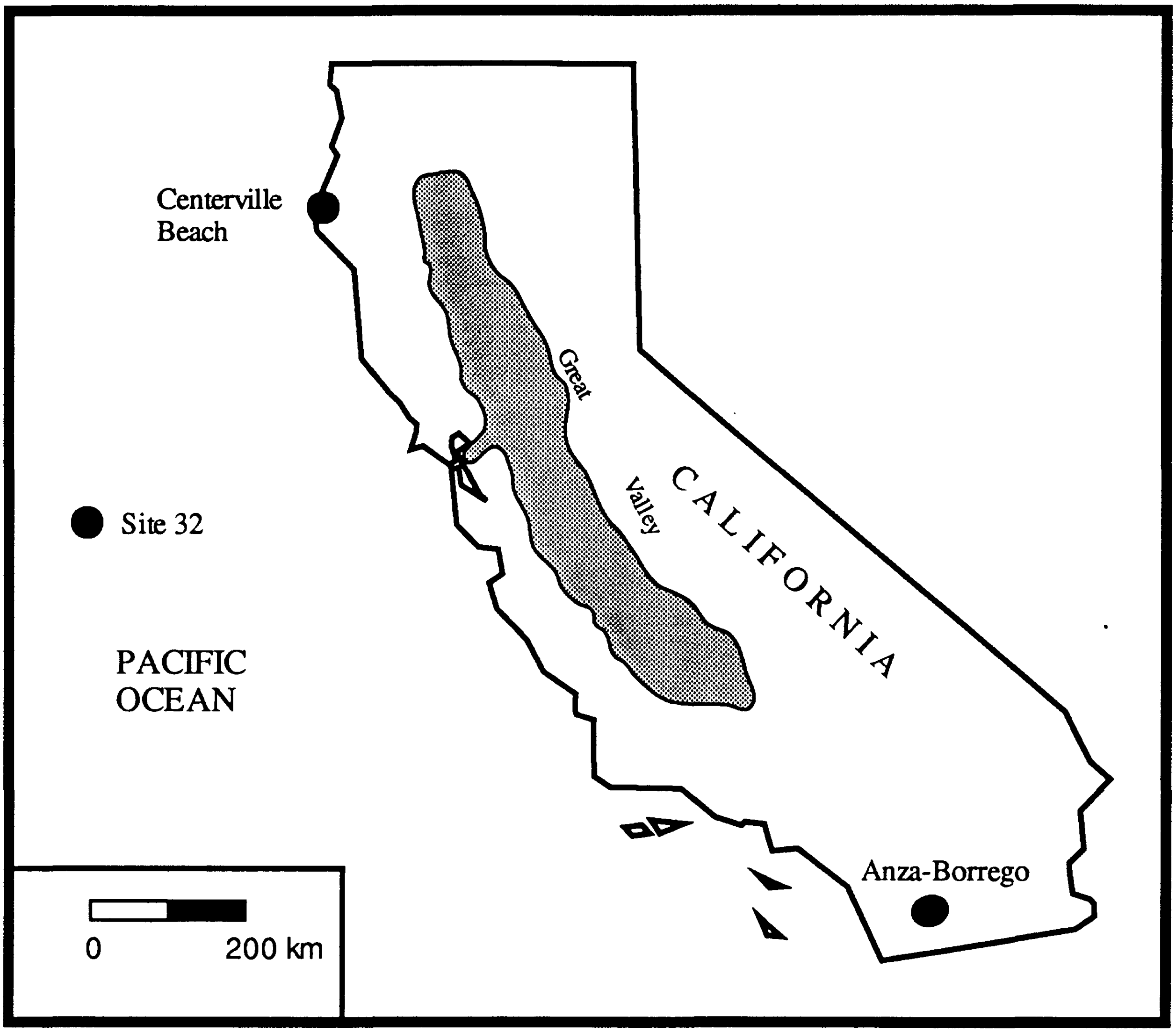

Figure 1. Sketch map showing position of Centerville Beach, DSDP Site 32, and Anza-Borrego.

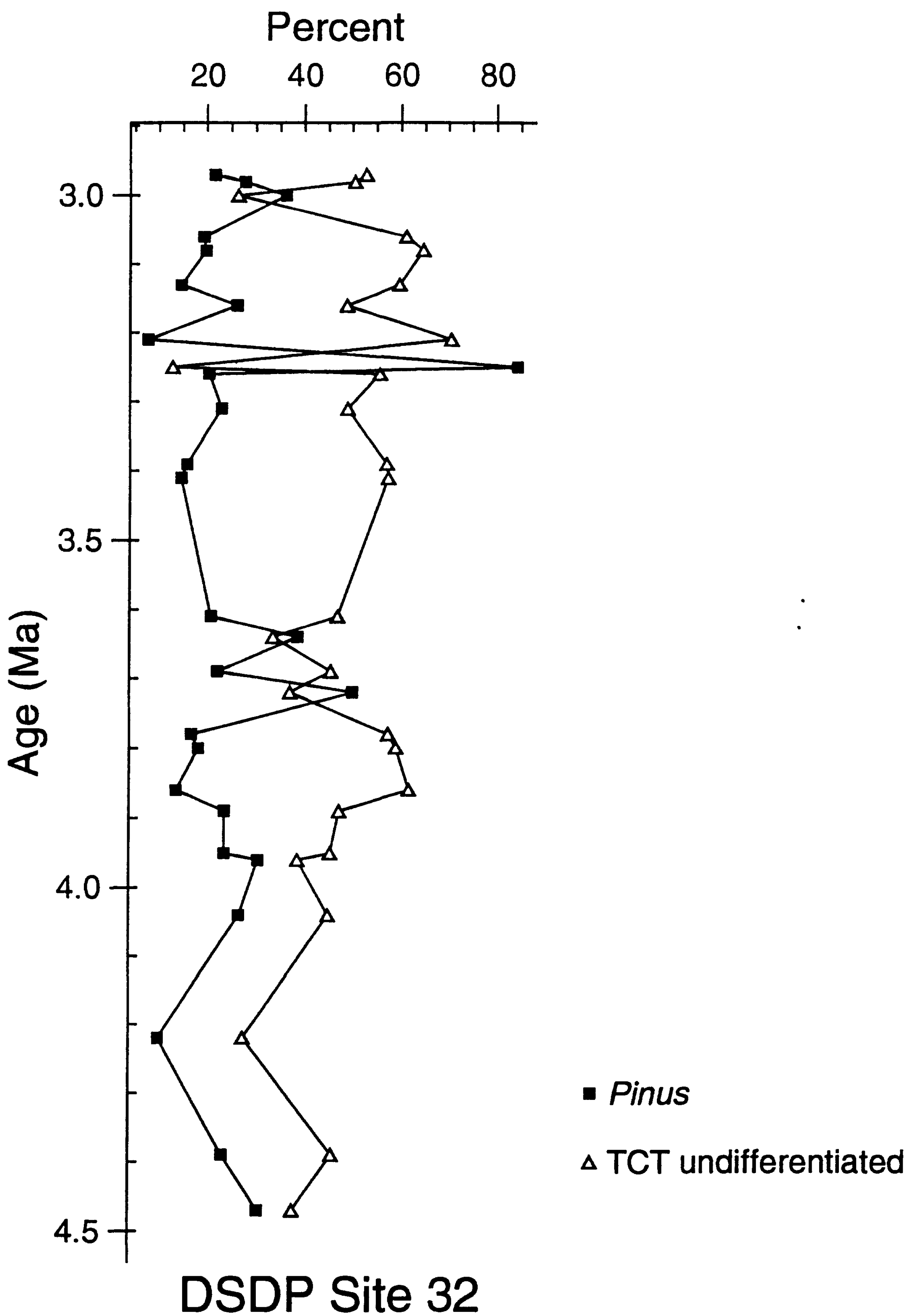

Figure 2. Relative abundance profile of *Pinus* and TCT pollen from DSDP Site 32.

Mid-Pliocene Vegetation, Environment, And Climate In The Western Interior Of The United States

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Tectonic basins in the western interior of the United States contain thick sedimentary sequences that preserve records of long-term climatic changes. However, the same forces that built these basins led to mountain building and the development of rain-shadows that impose local signals on the records of climatic variations. The modern climate in the northern portion of this region is dominated by winter precipitation associated with the westerlies, while the southeastern sector receives significant summer rainfall from subtropical monsoonal sources. Except for higher mountains, the entire region is semi-arid to arid, due to the rain-shadows of major mountain masses and to the distance from the oceans. Seasonal temperature variations are extreme, especially under the highly continental climate of the northern interior.

Vertebrate paleontological and paleobotanical data suggest that temperature extremes were less than today during much of the Pliocene, and paleohydrological and palynological records from this region (figure 1) indicate that conditions were also much wetter. Lacustrine conditions occurred in now-arid settings at Searles Lake (Smith, 1984) and other sites in southern California and adjacent Nevada (*e.g.* Amargosa Desert — Hay *et al.*, 1986; Hoover, 1989); central Arizona (Verde Valley — Nations *et al.*, 1981), and southern Arizona (St. David Formation — Wang *et al.*, 1991; Safford and San Simon Valleys — Gray, 1961; Tomida, 1987); central Utah; and southern Idaho and adjacent Oregon and Wyoming. The most detailed chronology is from Searles Lake, where sedimentary and geochemical data indicate sustained moisture from 3.2 to 2.6 Ma, followed by aridity until 2.0 Ma. Deuterium isotopes from calcite veins at Furnace Creek, California (near Death Valley — Winograd *et al.*, 1985) indicate that high Pliocene moisture levels continued until the last million years, when (presumably) the uplift of the Sierra Nevada and Transverse ranges blocked the incursions of moisture from the Pacific into the interior. Geological studies from the high central Sierra Nevada suggest that as recently as 3 Ma this mountain mass was significantly lower than today (Huber, 1981), and much of the modern aridity of the West must be tied to the development of regional rain shadows since the mid Pliocene.

In southwestern Idaho, the Glenns Ferry Formation appears to represent a deep (>300 m?) long-lived Pliocene lake in the modern Snake River drainage. Pollen data from a ~3.7 to ~3.4 Ma section at the eastern margin of this system (Leopold and Wright, 1985) record a near-modern flora with a sequence of from steppe to forest and back to steppe. Sedimentological variations between lake-margin sands and lignites suggest shorter term, lower-amplitude climatic fluctuations also occurred that are not reflected in the pollen record.

Farther west, a record from the deeper water sediments from near the town of Bruneau, in the east-central part of the Glenns Ferry Formation, apparently covers portions of the period from ~3.0 to ~2.48 Ma (Thompson, 1992). Very few exotic Tertiary elements are present in this diagram, and the vegetation fluctuates between three forest periods and three steppe periods before the lake becomes shallow (still within Gauss Chron). Unlike at Fossil Gulch, no sedimentological variations from this deeper-water environment correspond with the pollen fluctuations. Younger pollen assemblages from the Bruneau Formation at this site indicate a cold steppe environment sometime in post-Olduvai Matuyama time.

The undated Glenns Ferry pollen assemblages from a well site in the town of Vale, Oregon, contain palynological assemblages similar to the forest periods at the Bruneau site, and the longer (but also undated) Mountain Home Idaho core records a forest to steppe oscillation similar to those in the Bruneau core. The West Weiser outcrop, near the western extent of the Glenns Ferry Formation domain, is presumably >4 Ma and is strongly dominated by conifers and contains more Tertiary elements than sediments post-dating this time.

To the east of the Glenns Ferry deposits, the INEL (Idaho National Engineering Laboratory) 2-2a core from southeastern Idaho contains episodic lacustrine deposits separated by thick sequences of basalts. As with the Glenns Ferry palynological data, this site records fluctuations between steppe and forest dominance through the Pliocene. Assemblages thought to date ~4 Ma are dominated by conifers and are more diverse than younger pollen spectra (Thompson, 1991). Pollen assemblages dated from ~3.0 Ma are dominated by steppe taxa, which give way to conifers by ~2.9 Ma. Spectra dated to ~2.4 Ma contain low abundances of coniferous and other arboreal taxa and resemble late Pleistocene glacial-age pollen assemblages from southern Idaho. Coniferous taxa are somewhat more abundant by 2.0 to 1.8 Ma, but did not recover their pre-2.4 Ma levels.

Figure 2 illustrates the long-term Pliocene/early Pleistocene vegetation history of southern Idaho and adjacent Oregon: diverse early Pliocene coniferous forests with Tertiary elements gave way to modern steppe dominance over this period (presumably due in part to developing rain shadows). However, this trend was not monotonic, and apparent Milankovitch band high-amplitude fluctuations were recorded in forest/steppe cycles. With the exception of the apparent glacial-age steppe period (~2.4 to ~2.0 Ma), the palynological record suggests that conditions were wetter than today through the Pliocene and early Pleistocene in this region.

The information presented here provide only a sketch of Pliocene conditions in the western interior. Much more work is required to segregate the influences of global climate change from the regional aridification caused by mountain-building through the late Neogene and Quaternary.

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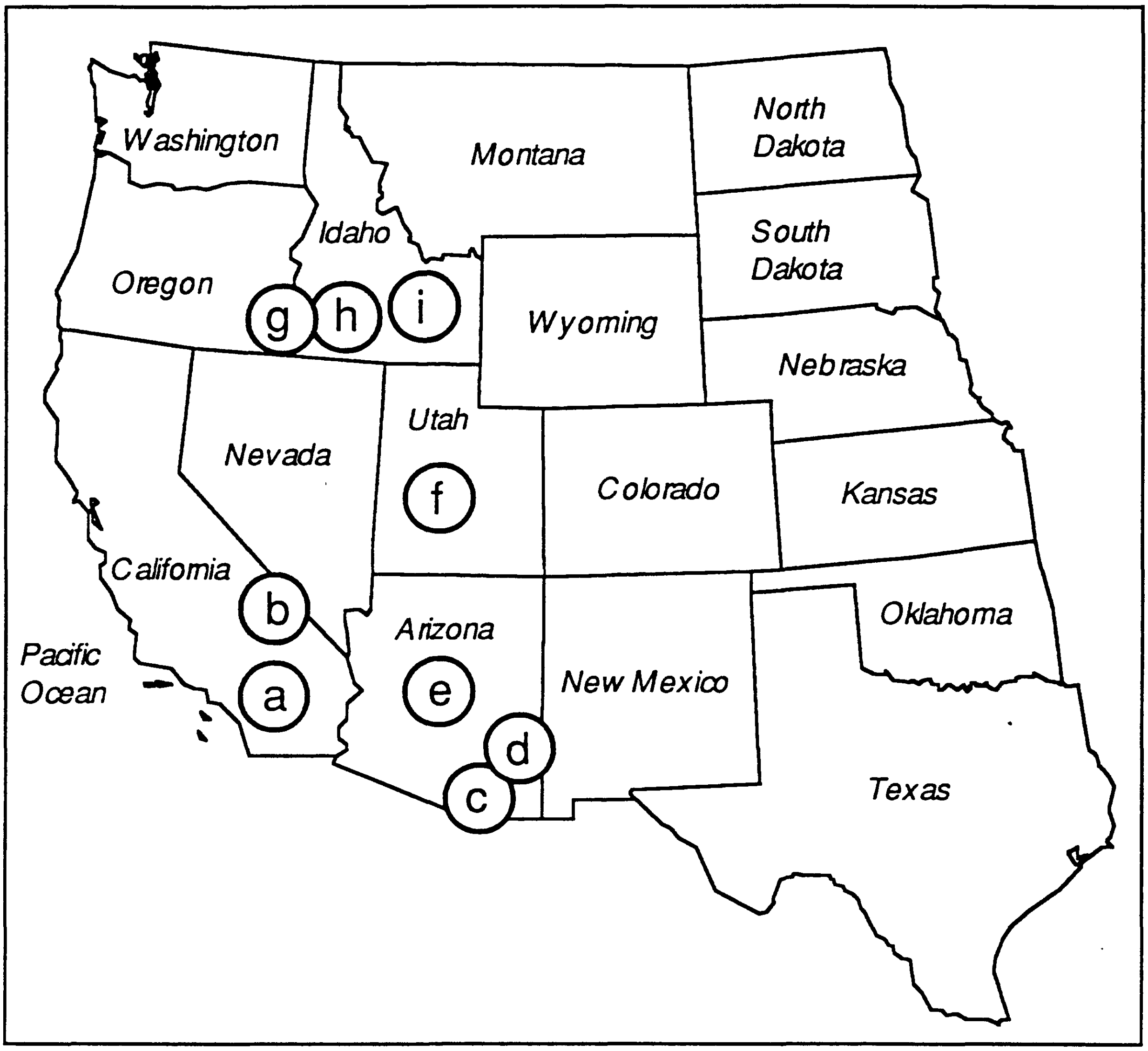

Figure 1. Sites mentioned in text. Key to sites: a) Searles Lake, b) Amargosa Desert and Furnace Creek, c) San Pedro Valley (St. David Formation), d) Safford and San Simon Valleys, e) Verde Valley, f) Central Utah, g) West Weiser and Vale, h) Bruneau, Mountain Home, and Fossil Gulch, i) Idaho National Engineering Laboratory (INEL) core site.

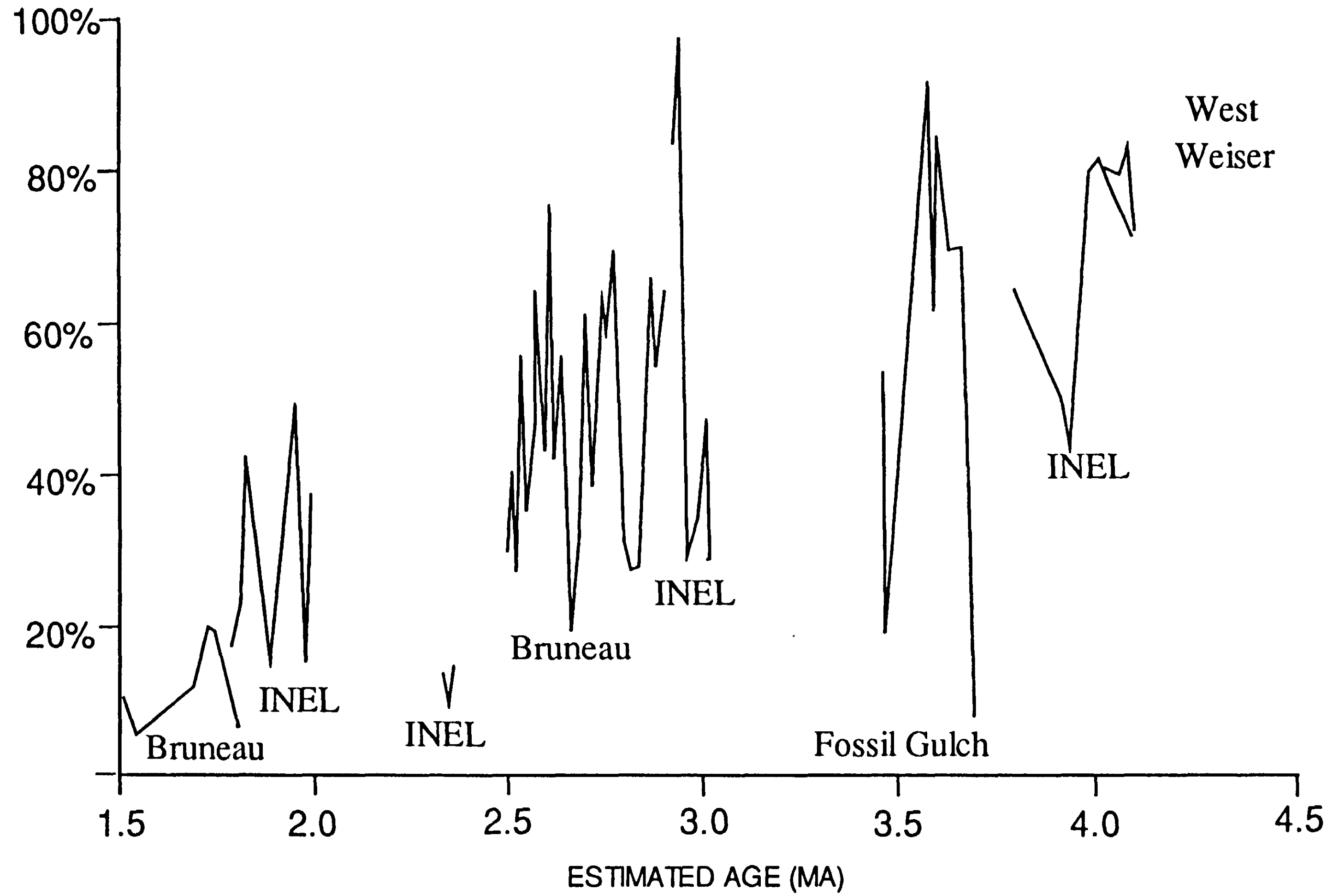

Figure 2. Changes in the relative abundance of conifers through the Pliocene and earliest Pleistocene in southern Idaho and adjacent Oregon.

Palynological Record From The North Atlantic Region At 3 Ma: Vegetational Response To A Period Of Global Warmth

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Middle Pliocene pollen assemblages from sites from 27° - 67° N (Fig. 1) were analyzed to study the latitudinal distribution of vegetation during the middle Pliocene "warm interval" around 3 Ma and to determine what mechanism for warming corresponds best to climatic estimates based on vegetation data. The samples at the five sites were selected to bracket the time interval around 3 Ma as closely as possible, based on the best available age data.

Pollen assemblages from ODP Site 642 in the Norwegian Sea and Pliocene strata from the Tjörnes Peninsula of Iceland are dominated by *Pinus* and *Picea* pollen. *Quercus, Alnus, Betula,* and *Ilex* pollen also are present at both sites, as well as *Castanea* (Tjörnes), *Sciadopitys*, and *Pterocarya* (Site 642). The presence of pollen from plants presently found in more temperate regions today suggests that Pliocene temperatures were warmer than today at both sites; Tjörnes probably had winter temperatures at least 4° C warmer than those of today.

Samples from deep-ocean sediments from ODP Site 646 in the Labrador Sea) are dominated by *Pinus* pollen, with *Picea* subdominant and *Abies, Betula, Alnus, Quercus, Sciadopitys,* and *Tsuga* present in most samples (see also de Vernal, this volume). These assemblages were compared to modern samples using the Modern Analog Technique (MAT) of dissimilarity coefficients, and the closest modern analogs are from Newfoundland and Nova Scotia. If the source area for pollen at Site 646 was Labrador or Quebec, as suggested by present wind patterns, then these data indicate a northward expansion of the northern hardwood forest into Labrador and Quebec and indicate that Pliocene temperatures in those regions were at least 3.5° C warmer than today with higher precipitation levels.

Pollen assemblages from the shallow-marine sediments of the Yorktown Formation in southeastern Virginia are dominated by *Pinus* with *Quercus, Carya,* and *Sciadopitys* pollen commonly present. The closest modern analogs are from North Carolina and South Carolina, indicating Pliocene temperatures about 2° C warmer than today and higher precipitation levels. The Pinecrest Beds of southwestern Florida also are dominated by *Pinus* pollen, with *Quercus* and taxodiaceous pollen abundant. These assemblages are analogous to modern assemblages from sites along the coast of Florida that represent the longleaf/slash pine forests that cover much of the Atlantic Coastal Plain and indicate no significant differences between Pliocene and modern temperatures or precipitation in this subtropical region.

The presence of vegetation characteristic of temperatures much warmer than present at high latitudes grading southward to present-day temperatures at low latitudes is similar to patterns shown by marine invertebrates in both degree and timing of change. Comparison of the marine and terrestrial records indicate that they responded to climate forcing mechanisms in phase with each other. The relatively greater amplification of temperatures at subarctic sites is consistent with predictions by general circulation models using increased meridional heat transport as the mechanism for middle Pliocene warming in the northern hemisphere. Further research on lower latitude and southern hemisphere sites should help determine whether Pliocene circulation patterns in southern oceans differed appreciably from those of today and whether they exerted similar controls on terrestrial vegetation.

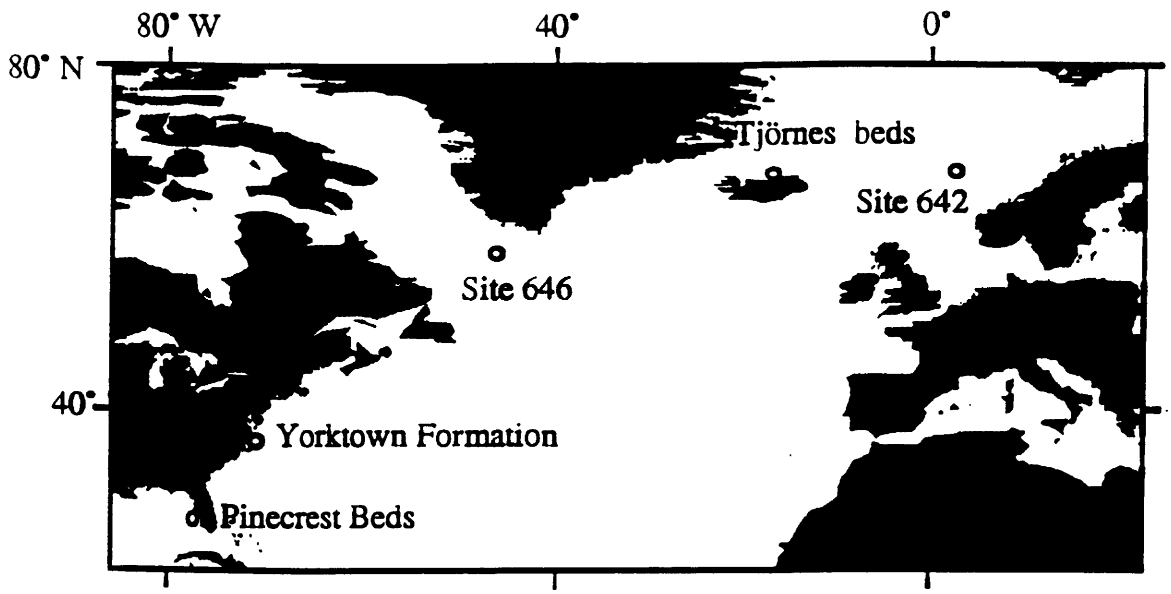

Figure 1. Location of sites analysed for pollen, North Atlantic transect.

Neogene Palynofloras And Terrestrial Paleoenvironments In Northern Latin America

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The database for estimating Tertiary terrestrial paleoenvironments in northern Latin America (Mexico, Central America, The Antilles), based on palynological evidence, is shown in Fig. 1 (see Note 1). The localities are scattered, and collectively provide only a low-resolution picture of environmental change. Therefore, several floras must be used to establish trends leading into the Pliocene. The two most relevant to the PRISM workshop are from the Gatun Formation, Canal region of Panama, and the Paraje Solo Formation near Coatzacoalcos, southern Veracruz state, Mexico. The chronological control on the pollen/spore-bearing part of the Gatun Formation is adequate for age assignment, but not precise. In a recent review of the marine record Coates *et al.* (1992) note (p. 827) that nannofossils from the middle part of the formation indicate an age range of 8.2-5.6 Ma. The 'terrestrial' phase (coastal mangrove zone) is subsurface and lacks marine palynomorphs (e.g., dinoflagellates) sufficient for tying it to the marine chronology. However, the pollen/spore-bearing lignitic material is from the upper part of the formation (top of solid rock; cores SL-49, SL-103). Coates *et al.* (1992, p. 827) state that "Gatunian" facies persist in Costa Rica during the Pliocene, but are absent from Panama after the earliest Pliocene. Thus, the maximum age of the plant microfossils is considered latest Gatun time (5.6 Ma) or slightly younger.

The age of the Paraje Solo Formation is better constrained through marine faunal assemblages in the overlying Agueguexquite and underlying Upper Concepcion Formations within zone N20 (Fig. 2) (see Note 2). Machain-Castillo (1985, p. 123) notes that "the Paraje Solo may be in part contemporaneous to the Agueguexquite. The Agueguexquite strata contain the youngest (middle N20), most abundant, and diverse fauna, indicating an inner neritic environment of deposition and representing a local marine transgression of short duration in the northern part of the basin. By upper Agueguexquite time brackish and continental conditions returned." The Paraje Solo palynoflora contains an abundant brackish *Rhizophora* (mangrove) and diverse inland to upland continental component, suggesting deposition at the time and under the conditions described for the upper Agueguexquite sediments. Akers (1979) presented a similar stratigraphy and age assignment of the Lower Concepcion through Agueguexquite formations (Figs. 2 and 3) based on calcareous nannofossils and planktic foraminifera. He concluded (p. 4) that "the Agueguexquite Formation must be considered to be of middle Pliocene age, and it probably belongs just above the precise middle of zone N20." Thus, the Paraje Solo plant microfossils are Pliocene, and probably middle Pliocene from that part of the formation equivalent to the Agueguexquite, ~mid-N20, ~3-4 Ma.

The microfossils include wind-blown and water-transported forms, and preserve an environmental record for areas delimited mostly by physiography. For Panama (Gatun Formation) the reconstructions are probably applicable to most of southern Central America where land surfaces were primarily east-west trending and where physiographic relief was moderate. For Mexico (Paraje Solo Formation) the reconstructions are probably valid for the eastern escarpment of the Sierra Madre Oriental within southern and central Veracruz state where latitudinal extent and physiographic diversity was greater.

The palynological data can provide a general quantitative estimate of climate at individual sites, but only a broad picture of regional and long-term changes. This is due more to the scattered occurrence of the material than to the imprecise nature of the technique.

The available data suggests that from the Eocene (Chapelton, Gatuncillo floras) through the lower Miocene (Uscari, Culebra, Cucaracha, and La Boca floras) current tropical conditions and relatively low relief characterized much of northern Latin America south of Mexico. In southern Central America conditions had changed by Gatun time. Pollen of the Gramineae (grasses), which was virtually absent in the lower Miocene, increased to 7.5%. Maximum estimated elevations increased from ~1200-1400 m to ~1700 m. Greater environmental and habitat diversity resulted in greater biotic diversity, with the number of taxa increasing from 44, 55, 21, and 54 types in the lower Miocene floras noted above, to 110 types in the Gatun Formation. In addition to individual elements, there was also a greater diversity of communities. All elements in the lower Miocene floras can be accommodated in four forest types- tropical wet, tropical moist (including mangrove forest), premontane wet, and possibly premontane moist. By Gatun time there were seven forest types with 11 or more taxa that can occur in each community- tropical moist (38 taxa), tropical wet (31), premontane wet (27), premontane moist (21), lower montane moist (12), premontane rainforest (11), and tropical dry forest (11). The first six of these are similar to the modern vegetation of lowland to moderate-altitude habitats on the northern (Atlantic) side of Panama where the following climates prevail (Barro Colorado Island, Gatun Lake): average annual rainfall 2750 mm/yr (range 1900-3600 mm, strongly seasonal, dry season January through April); MAT 27^{o} C (annual range 21-32^{o} C). The presence of the tropical dry forest suggests altitudes had reached a point where moist Atlantic winds were deflected to create the drier Pacific side as at present (1700 mm/yr at Balboa). The close relationship between the Mio-Pliocene Gatun vegetation and the modern lowland to mid-altitude communities suggests that the primary forcing mechanism for vegetational change at latitude 9^{o} N was physiographic relief. Coastal climates were similar to those of present during the late Miocene/earliest Pliocene, and probably since the middle to late Eocene (Gatuncillo flora).

The situation at comparable low to mid-altitudes in Veracruz (latitude 18^{o} N) by Paraje Solo time (mid-Pliocene) was different. The present-day tropical rain forest was poorly developed to absent, mid-altitude (presently 1000-2000 m) temperate communities were prominent in the lowland basin of deposition, and *Picea* (spruce) was present. Spruce occurs today in Mexico only in the mountains 1000 km or more to the north. The common factor in these changes is lower temperature. As an estimate, a MAT lower by ~2-3^{o} C than at present would likely produce the observed differences between modern and the Paraje Solo vegetation. The present MAT at Coatzacoalcos is 25.3^{o} C (range 22-27^{o} C), and annual rainfall is 2726 mm (range 50-525 mm, driest season January through March with some rain falling in every month). The mid-Pliocene MAT is thus estimated at ~23^{o}.

The lack of other fossil floras in the region make it impossible to detect when the time changes from lower Miocene to Paraje Solo conditions began. However, some estimate can be gained by extrapolating from two other lines of evidence. Preliminary study of the Oligo-Miocene Simojovel flora (Chiapas, Mexico) shows a mangrove forest, with oak-pine woodland grading upwards into more temperate communities on the adjacent slopes. Since this is similar to much of the modern local vegetation, it seems reasonable to conclude that the Neogene cooling trend had not set in there by late Oligocene-early Miocene time. Also, the benthic marine temperature curve (Fig. 4) suggests the middle Miocene as a likely time, which is consistent with the limited paleobotanical evidence.

If it is assumed that temperatures at 18^{o} N latitude were ~3^{o} cooler than at present (Paraje Solo flora), and that at 9^{o} N latitude contemporaneous to somewhat older lowland climates were comparable to the present (Gatun flora), it is tempting to reconstruct intervening terrestrial MATs by lapse rate methodologies (yielding ~0.3^{o} C/1^{o} latitude).

The figure may ultimately prove to be approximately correct, but there is little corroborative evidence. Also, the Paraje Solo communities were growing under or adjacent to continental conditions in an area of significant physiographic relief. Thus, any exogenic change in climate read from the vegetational history would be magnified by the response of communities growing along steep altitudinal gradients. In contrast, the Gatun communities were growing on islands and peninsulas of low physiographic relief and any exogenic change in climate would be dampened. Some general estimate of temperatures in the uplands can be gained from the current worldwide mean altitudinal lapse rate of 5.0-6.0^{o} C/km.

At present it is not possible to determine more precisely the range of terrestrial paleoclimatic variability in northern Latin America during the Pliocene from the floras available. Eventually, data from the Guastatoya and Herreria Formations of Guatemala may prove useful.

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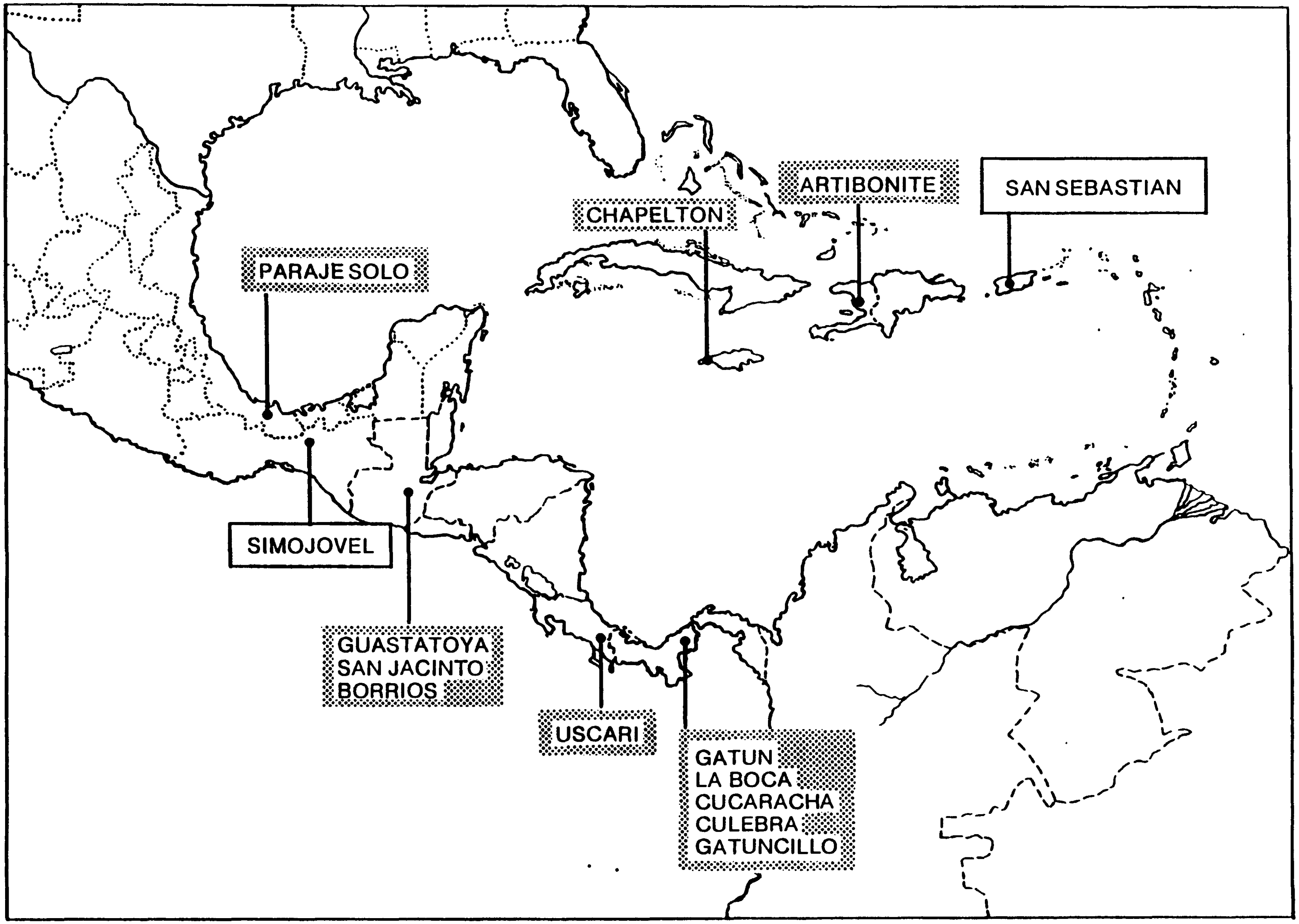

Figure 1. Geographic distribution of principal palynofloras of the southern Gulf-Caribbean region. Stippling indicates study completed; open areas indicate floras under study or revision.

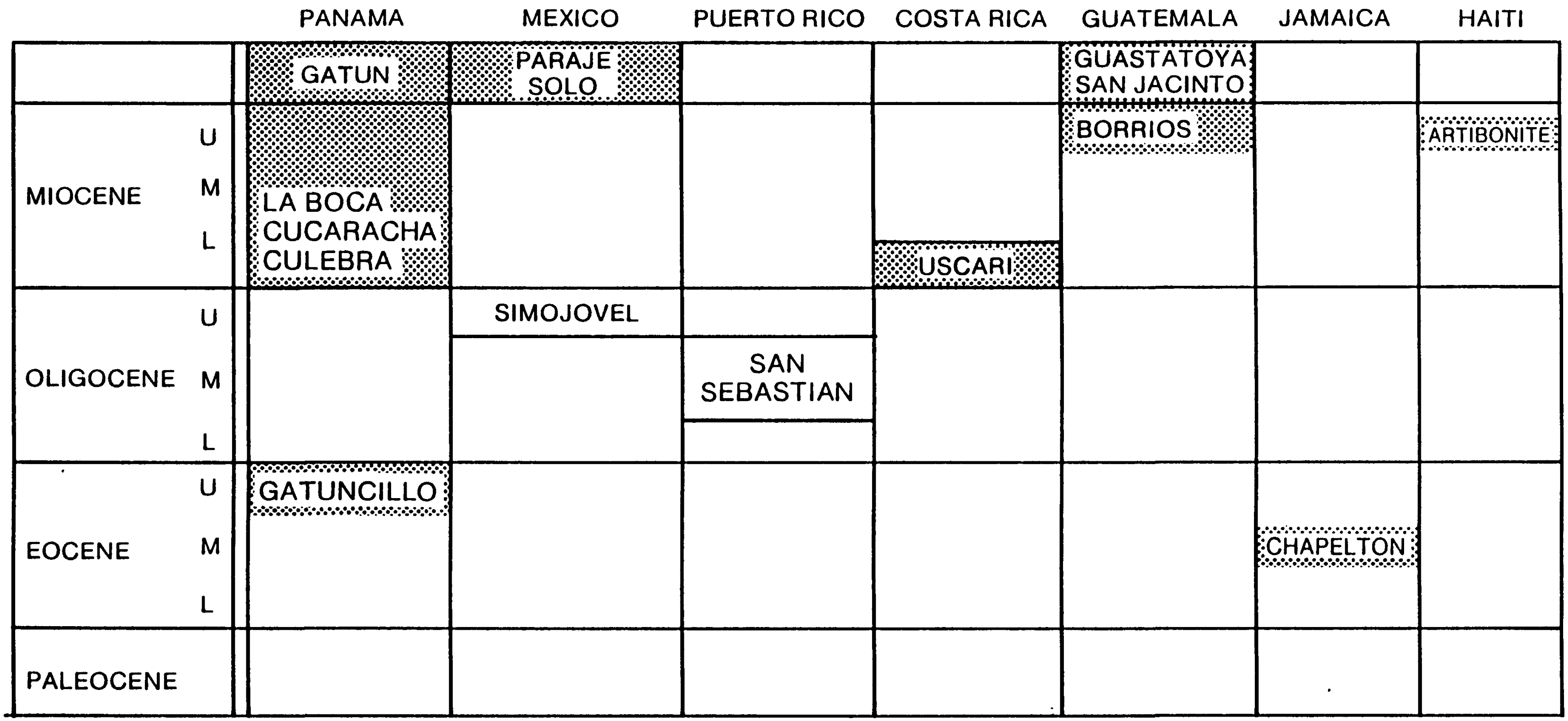

Figure 2. Stratigraphic distribution of principal palynofloras of the southern Gulf-Caribbean region. Stippling indicates study completed; open areas indicate floras under study or revision.

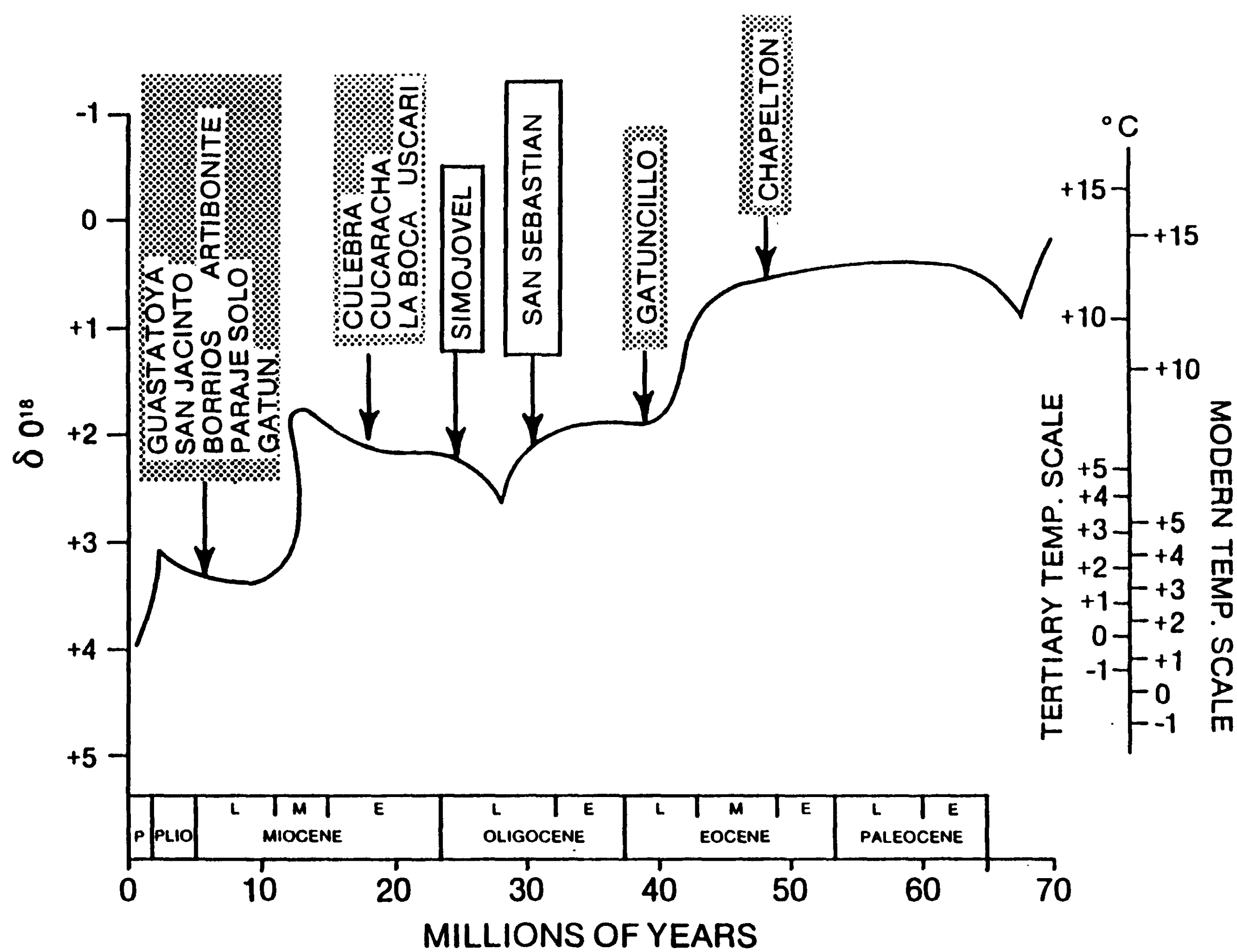

Figure 3. Palynofloras of the southern Gulf-Caribbean region plotted on the benthic paleotemperature curve.

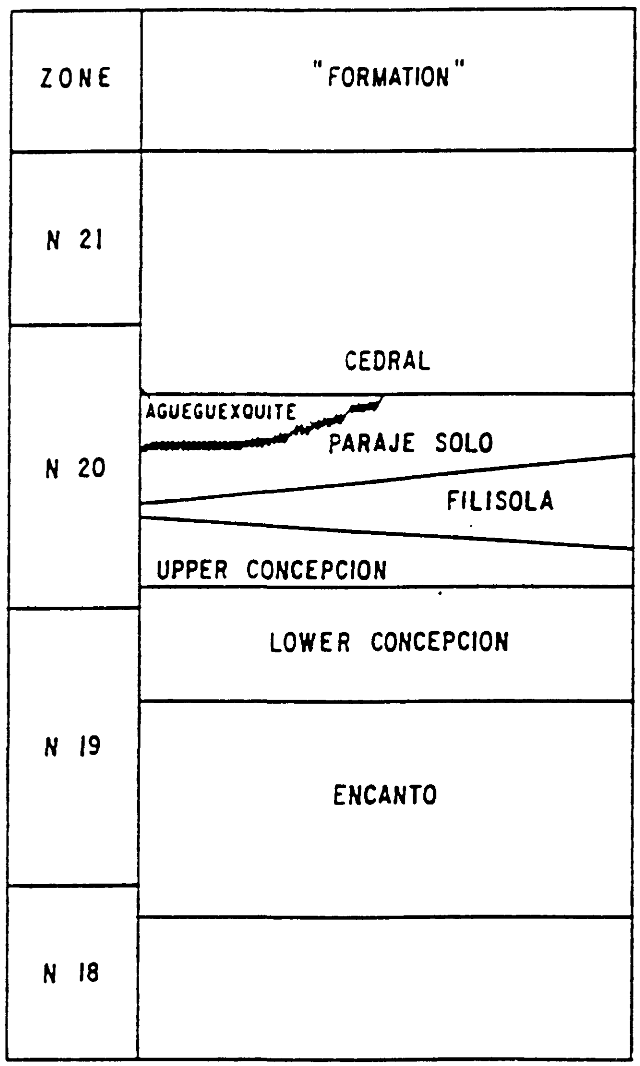

Figure 4. Position of the Paraje Solo Formation in relation to the nannofossil zones (from Machain-Castillo 1985).

Notes

Note 1

Notes on the charts: 1) There are no extensive, well-preserved, recently studied Tertiary macrofossil floras known from the region. The only sizable flora (the Oligocene San Sebastian flora of Puerto Rico) was last studied in 1928. All terrestrial paleoenvironmental estimates presented here are based on plant microfossil evidence.

2) The markings indicate different stages in the completion of the studies (stipple--study complete and results published; others in progress). The Chapelton study is in press as part of GSA Memoir 182.

3) Samples from Guatemala that have yielded palynomorphs (fair to poor) are from the Padre Miguel Group (now excluding the younger San Jacinto Formation; radiometric data indicates the Padre Miguel ignimbrites erupted between 19 and 14 Ma - middle to late Miocene), Barrios sequence (possibly Barrios Formation, Miocene), the Guastatoya Formation (age unknown, possibly Pliocene), and the Herreria (Pliocene, and possibly equivalent to the upper part of the Guastatoya). Information on the Barrios Formation fide Burkart (pers. comm.), all other from Donnelly *et al.* (1990). The Guatemalan lignites are not coastal brackish-water deposits containing *Rhizophora* (mangroves) as are the others, but were deposited in swamps that periodically became saline through evaporation, as indicated by gypsum crystals (Burkart, pers. comm.).

Note 2

Notes on Fig. 2. 1) Ostracodes listed for Lower and Upper Concepcion bed (N19-20) are *Actinocythereis vineyardensis, Touroconcha lapidiscola,* and abundant *Hulingsina* sp. 1, *Henryhowella* ex. gr. *asperrima,* and *Puriana* spp. In addition, Upper Concepcion species include abundant *Cyprideis* spp., *Perissocytheridea* spp., *Basslerites*? sp., *Malzella conradi,* and *Echinocythereis margaritifera* absent from the Lower Concepcion (Machain-Castillo, 1985).

2) From the Agueguexquite Formation Akers (1979) lists the following planktic foraminifera-*Globigerina bulloides apertura, G. bulloides bulloides, G. juvenilis, Globigerinoides obliquus extremus, G. obliquus obliquus, Globigerinoides quadrilobatus quadrilobatus, G. ruber, Globorotalia (Globorotalia) cultrata limbata, G. (Turborotalia) acostaensis acostaensis, G. (T.) acostaensis humerosa, Hastigerina (Hastigerina) siphonifera siphonifera, Orbulina universa,* and *Sphaeroideinella dehiscens dihiscens* forma *immatura;* and the following calcareous nannofossils (in addition to those listed in Akers and Koeppel, 1973, p. 83) - *Ceratolithus cristatus, Coccolithus doronicoides, Cyclococcolithina leptopora, C. macintyrei, Discoaster brouweri, D. pentaradiatus, D. surculus, D. variabilis, Gephyrocapsa caribbeanica, G. reticulata, Helicopontosphaera kamptneri, H.* sp., *Lithostromation perdurum,* and *Pseudoemiliania lacunosa.*

Paleoclimatic Conditions Around 3 Million Years BP: Pollen Evidence From Colombia

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Introduction

In the Eastern Cordillera of Colombia, the high plain of Bogotá (ca. 25x40 km in size) represents the bottom of a former lake that occupied a subsiding intermontane basin. Pollen records from deep bore holes represent the period from the Late Pliocene to latest Pleistocene. Climatic changes are well documented by the pollen rain that is conserved in the slowly accumulating lake sediments of the Bogotá basin. The tropical Andes especially seem to be in a favorable position to register climatic change. A change in climatic conditions results mainly in a vertical shift of vegetation belts over the mountain slopes (Fig. 1). The different vegetation belts stay in the vicinity and are registered continuously by their intercepted pollen. The sediments of the Bogotá basin (2550 m alt.) accumulated in an elevation that lies halfway between the highest position of the upper forest line during interglacial conditions (c. 1800 m alt.), rendering the Bogotá sediments a sensitive recorder of paleoclimatic change in northern Latin America.

Pollen Data And Time Control

A new 586 m long core Funza II (Fig. 2) was recovered from the sediments of the high plain of Bogotá. Results of the palynological analysis of 430-540 m core depth, representing the estimated time interval 3.2-2.36 Ma, with sample distances of 100 cm are presented (Fig. 3). The interval 586-540 m of the sediment core was poor in pollen recovery or barren. Time control of the Funza II core is based on zircon fission-track datings of intercalated volcanic ash horizons. In addition, the upper part of pollen record Funza II could be correlated in detail with pollen record Funza I; the latter has been graphically correlated with the oxygen isotope record of ODP Site 677 (E. Pacific). It is expected that the pollen record can be correlated with the deep-sea record up to about oxygen isotope stage 110 when samples are available at 20 cm distance (Hooghiemstra, in prep.)

Several phases in the evolutionary history of the montane forests and paramo vegetation in the Eastern Cordillera of Colombia can be recognized. Late Pliocene and Pleistocene climatic change and the closure of the Panamanian Isthmus had a substantial influence on the development of these montane ecosystems.

Paleoclimatic Conditions At Ca. 3 Ma

The interval 540-465 m core depth (3.2-2.7 Ma) shows warm climatic conditions and pollen spectra have no late Quaternary analogs. The basin had just started to accumulate lacustrine and river sediments, after a period in which sediment only accumulated in the perifere valleys. The upper limit of the subandean forest belt was situated at some 500 m lower elevation than today. In the Andean forest belt *Podocarpus* rich forest, *Hedyosmum-Weinmannia* forest (a precursor of the modern *Weinmannietum*), and *Vallea-Miconia* forest, respectively, were the main constituents with increasing elevation. *Hypericum* and *Myrica* played an important part in the timberline dwarf forests, which possibly constituted a substantial transitional zone from the early Andean forest belt (upper montane forest belt) to the open grassparamo belt. The contribution of herbs to the paramo vegetation, dominated by Gramineae and Compositae, seems less diverse than during the late Quaternary. The late Pliocene (upper) Andean forests were more open than during the middle and late Quaternary, as heliophytic elements, such as *Borreria*, were abundant (Fig. 3). The composition of forests on the high plain was dynamic: arboreal taxa with pioneer qualities (*Dodonaea, Eugenia*) and other taxa (*Symplocos, Ilex*) constituted seemingly at irregular intervals azonal forests in the basin. The upper forest line oscillated most of the time from 2800 to 3600 m elevation. The average annual temperature on the high plain was c. 22.5-16.5°C.

Using composite pollen records Van der Hammen also documented warmer conditions in sediments from the area of the high plain of Bogotá during Pliocene time (Van der Hammen *et al.*, 1973). Recently additional evidence for warmer late Pliocene climatic conditions was collected on the basis of plant macrofossils from this area (Wijninga and Kuhry, submitted). Although these fragments of pollen records most probably evidence phases of the final upheaval of the Eastern Cordillera, which has a 'cooling' effect on the pollen records, the continuous Funza II pollen record does evidence climatic change. It is expected that part of the variability documented in the Funza II summary diagram (Fig. 3) is associated with the initial stage of development of the large tectonic sedimentary basin of Bogotá.

Potential Use Of Pollen Data As Modern/Future Analogs

The late Pliocene (3.2-2.8 Ma interval) pollen spectra differ significantly from the late Quaternary ones. It is inferred that the late Pliocene Andean forests were of a more open character (e.g. presence of *Borreria*, originally a savanna element) whereas important elements of the present-day forest belt (*Alnus* and *Quercus*) were absent. Estimation of the altitudinal position of the upper forest line, therefore, is less accurate than during more recent intervals of the pollen record. However, accumulating evidence and better understanding of these late Pliocene ecosystems will enable to provide quantitative paleoclimatic estimates concerning temperature ranges and climatic variability (Hooghiemstra, in prep.).

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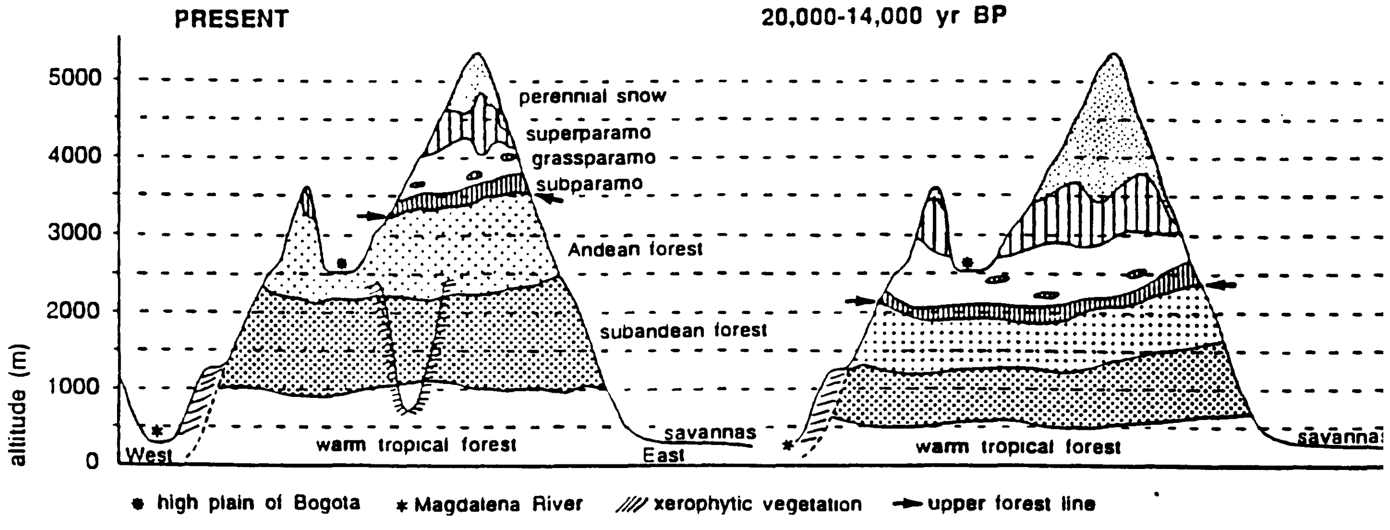

Figure 1. Altitudinal distribution of vegetation belts in the Eastern Cordillera of Colombia at present and during the last glacial maximum. Vertical shifts of the vegetation belts are mainly related to changes in temperature and form the basic mechanism of registration of climatic change (after Van der Hammen, 1974).

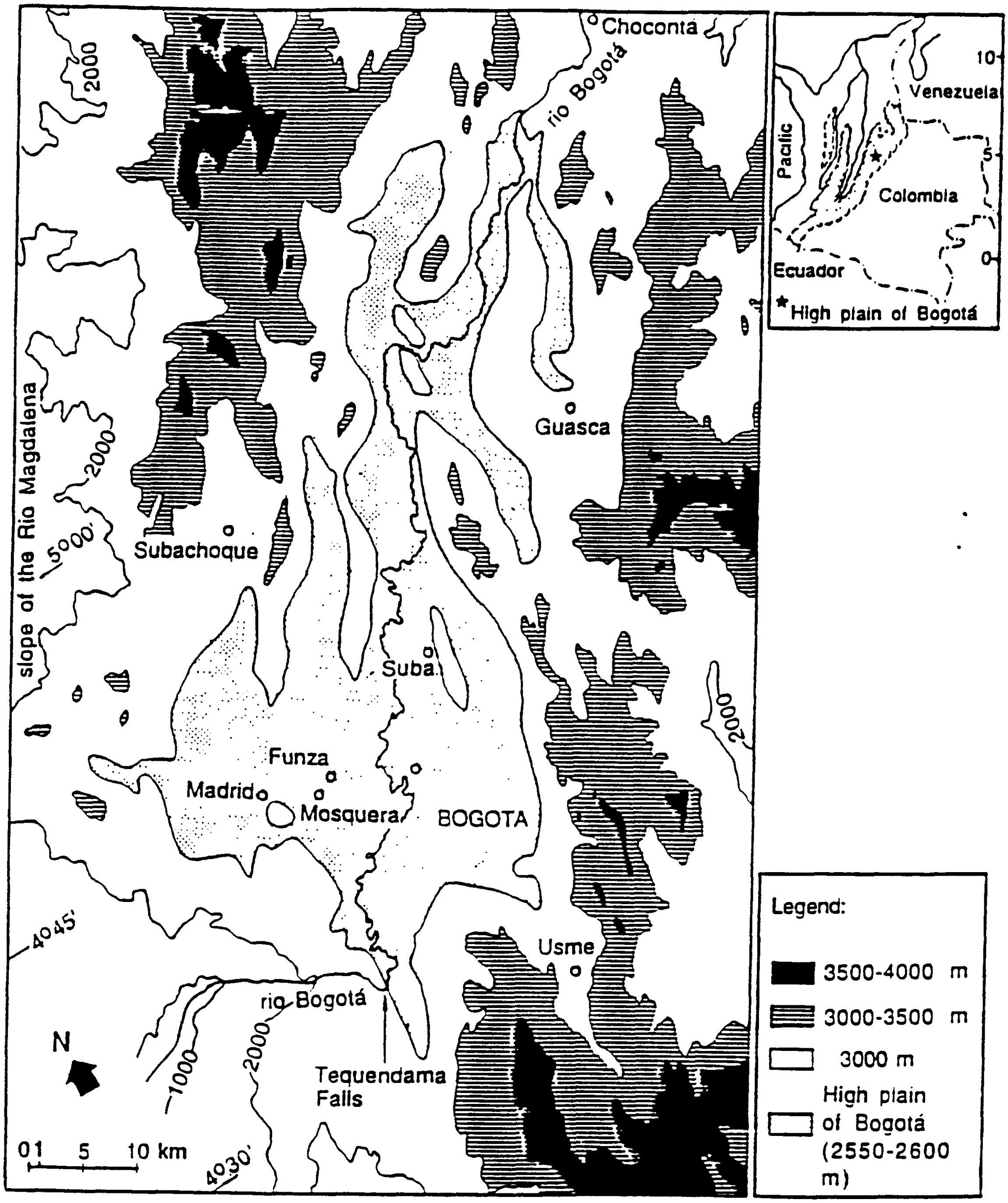

Fig. 2. Map of Colombia and the high plain of Bogotá in the Eastern Cordillera, indicating the geographical setting of the sites of the long continental pollen records near the village of Funza.

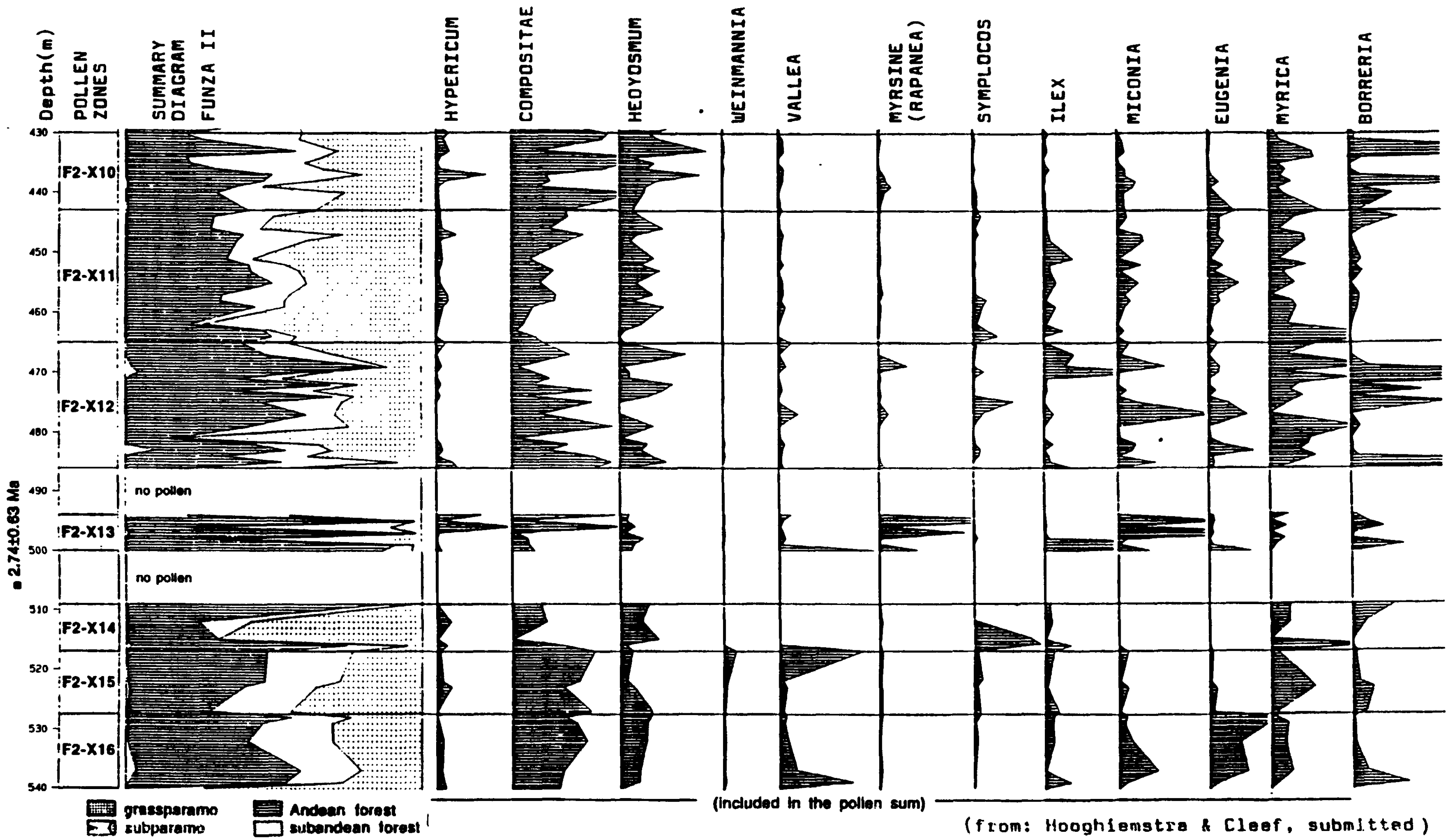

Figure 3. Pollen percentage diagram of the 540-430 m interval of the Funza II core (Eastern Cordillera, Colombia, 2550 m alt.) analyzed with 100 cm sample distance. Records of selected pollen taxa are presented. The pollen zones are provisional and will be defined more precisely after the present time resolution of c. 6000 years has been increased (Hooghiemstra, in prep.). Time control of the lower part of the Funza II core is provided by a zircon fission-track dating of an intercalated volcanic ash horizon at 506 m core depth (2.74 ± 0.63 Ma; Andriessen *et al.*, submitted). (The upper part of the Funza pollen record is correlated with the ODP Site 677 oxygen isotope record (Shackleton and Hooghiemstra in Andriessen *et al.*, 1993). Note the absence of *Alnus* and *Quercus* in this part of the record (northern hemisphere taxa which arrived in the Eastern Cordillera during middle Pleistocene time after having passed the Panamanian landbridge).

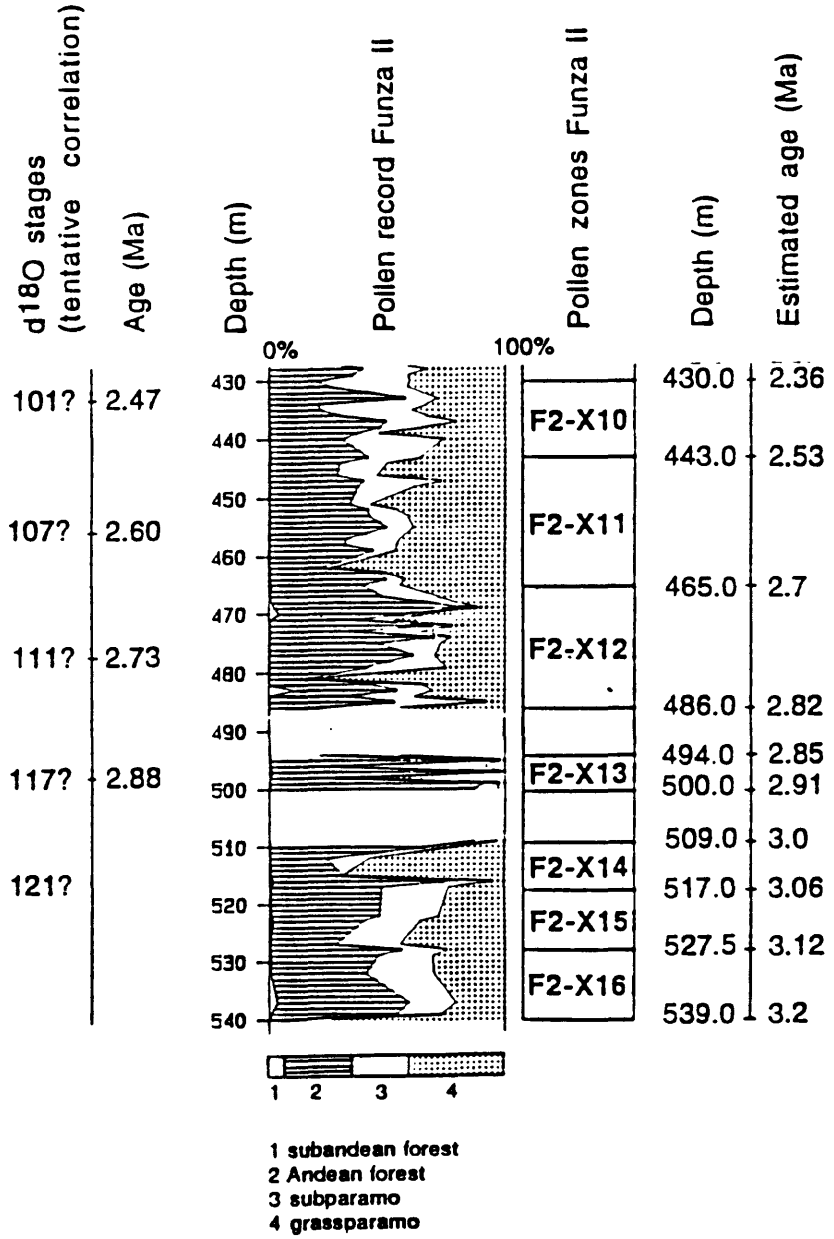

Figure 4. Summary pollen diagram Funza II (540-430 m interval) with pollen zones and tentative correlation with the deep-sea oxygen isotope stratigraphy (Ages of d18O stages after Imbrie et al., 1984).

Mediterranean Pliocene Vegetation And Climate: How To Quantify The Climate Parameters?

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The present vegetation of the Mediterranean region is controlled by physiographic and climatic peculiarities:

- the longitudinal length of the sea and the presence of West-East oriented mountains (Pyrenees, Alps, partly Carpathians, Caucasus), which constitute barriers for plant migrations;
- the southward extension of areas of high relief associated with peninsulas (Spain, Italy, Greece) which provides a penetration of boreal environments into the warm Mediterranean region;
- a double seasonal rhythm (cool winter, warm summer; dry summer, humid spring and autumn) of the climate which shows a high variability according to the Atlantic air mass effects, desert influence and local orientation.

The factors above result in a very complex mosaic structure of the vegetation which contrasts with the latitudinal (thermic) zonation of Central and North Europe.

The Mediterranean early-middle Pliocene climate reconstruction expresses the evolution from tropical/subtropical to warm-temperate conditions before the onset of the glacial-interglacial cycles (Suc *et al.*, 1992).

Pollen Data

Pollen records and their chronological controls. More than 60 sections (boreholes and outcrops) have been investigated (the 45 most significant ones are located on figure 1). They represent more than 1,300 pollen spectra (at least 600,000 counted pollen grains) and more than 250 identified taxa. Synthetic pollen diagrams are constructed after the ecological arrangement of the taxa presented in detailed pollen diagrams (Suc, 1984).

Most of the localities are coastal marine deposits in which the chronostratigraphic control is precisely provided by foraminifer and nannoplankton records, sometimes associated with magnetostratigraphy and radiometric measurements (fig. 1). The age of some sections is known from mammals in association with magnetostratigraphy, and only a few sections are not well-dated (fig. 1). The large number of well-dated pollen records provides a suitable basis for vegetation reconstruction through the entire Pliocene (fig. 2).

5.2 to 3.2 Ma. Pollen localities in this age range are very numerous because of the "sudden" inundation of the Messinian canyons by the lowermost Pliocene transgression. Very diverse landscapes are recorded, reflecting both different latitudes and local/regional features. These include:

A. *The Atlantic face* where subtropical (Taxodiaceae) and warm-temperate trees dominated and Ericaceae were abundant. The north-south thermic gradient promoted differences in the frequency of thermophilous elements. Humid conditions prevailed.

B. *The southern Mediterranean open xeric vegetation* which began south of Barcelona. Herbs, including subdesertic taxa (*Lygeum, Calligonum, Nitraria, Neurad*, Agavaceae, etc.) were predominant. Tropical (megathermic) plants were still scarcely represented in lower latitudes (Southern Spain, Tunisia, Sicily, Crete, Egypt, Israel). The modern Mediterranean xerophytes were important in some places (Catalonia, Sicily). A xeric-thermic gradient controlled the plant distribution in this area. The large latitudinal distribution of such assemblages supports the theory that seasonally dry climates occurred in this region as far back as the Miocene (Suc *et al.*, 1992).

C. *The northwest Mediterranean region* (from the Pyrenees to mid-Italy) was characterized by high relief, which induced the formation of altitudinal forest belts. The lower belt was mainly occupied by *Sequoia*-type, the middle belt with *Cathaya* and *Cedrus*, and the upper belt with *Abies* and *Picea*. Mediterranean xerophytes were relatively scarce, and on the whole, the pollen records indicate cooler and more humid conditions than the lower Mediterranean latitudes.

D. *The northeast Mediterranean region* which appears to have been cooler than the previous region. Warm-temperate forests alternated with grasslands (Gramineae). The influence of mountains is obvious, chiefly in Bulgaria, and Mediterranean xerophytes increased southward.

E. *The Nile area* was almost exclusively inhabited by a savanna-like vegetation, including desertic elements. Tropical plants were present.

F. *Local swamp environments* are represented by assemblages that contain high percentages of *Taxodium*-type, *Nyssa, Symplocos, Myrica*, Cyrillaceae-Clethraceae, etc. They occurred only in Portugal (Rio Maior), Catalonia (Garraf 1), and Rumania (Ticleni).

3.2 to 2.6 Ma. The same environmental and geographic subdivisions existed as during the previous period. Only some localities record changes at 3.2 Ma (Rio Maior, Garraf 1, Nice area, Ticleni), with the disappearance or the decrease of the most thermophilous elements only the most humid places (local marshes and steep slopes). It has been interpreted as the emergence of cool winters, i.e. a well-marked seasonal thermic rhythm to be superimposed to the preexisting pluvio-metric one. Therefore, the rather constant climatic conditions of this period are considered as nearly similar (except a slightly higher level in temperature) to the present conditions. So, it can be used as a suitable reference in modeling the forthcoming warming due to the greenhouse effect.

After 2.6 Ma. The earliest glacial-interglacial cycles are characterized in the north Mediterranean area by steppe-like vegetation (with *Artemisia* and *Ephedra*) alternating with forest environments (mainly constituted by warm-temperate elements). These steppe assemblages appear obviously warmer than those of the late Pleistocene glacials. There is no evidence of change in the southern Mediterranean region, except in the Nile area, where the Compositae increased within the herb group. In middle and northern oceanic Europe, these cycles corresponded to temperate forest and tundra-like (Gramineae, Cyperaceae, Ericaceae) alternations. In the Alps, the climate fluctuations are marked by temperate-altitudinal forest replacements.

Quantitative Paleoclimatic Estimates

It is difficult to estimate the peri-Mediterranean Pliocene climate because of the little latitudinal variation of plants on the one hand, of the non-existence of analogues in the present vegetation on the other hand. Thus a climate reconstruction is only possible if one takes into account some of the climatically-significant taxa and their respective relative occurrence in modern pollen spectra. A first attempt of estimation of the climate parameters is in progress, initially considering only the open vegetation environments. This approach will be progressively applied to forest assemblages.

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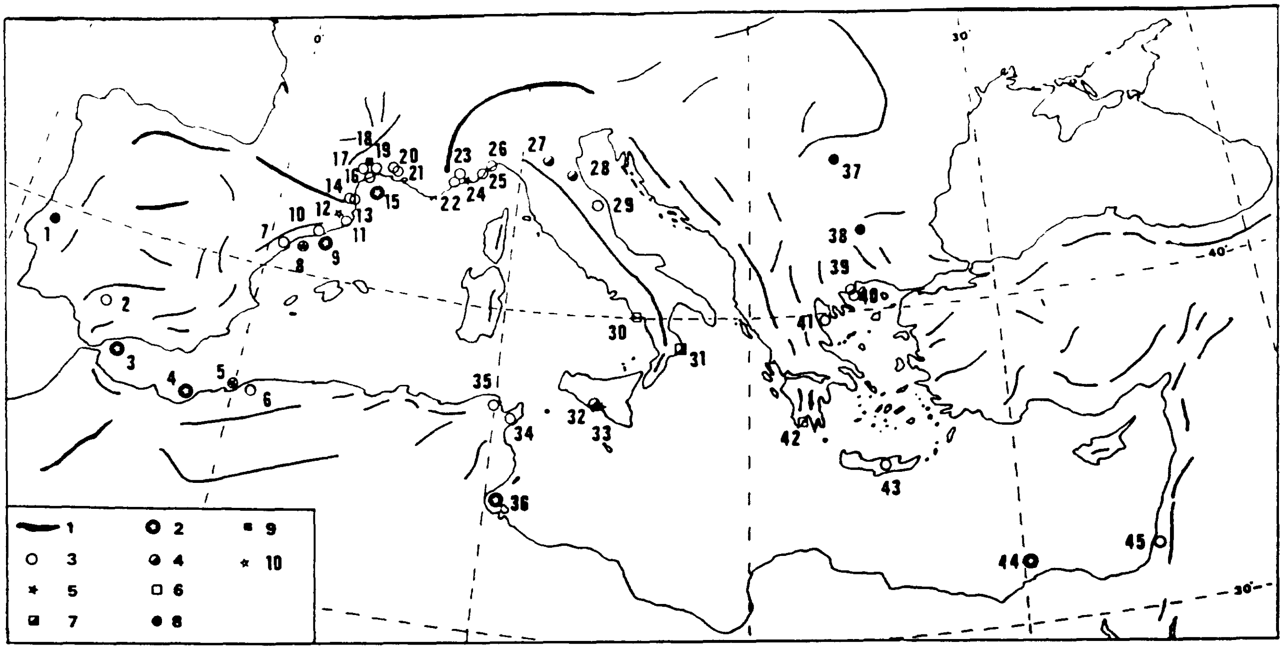

Fig. 1. Selected Pliocene pollen localities in the Mediterranean region.

<u>Legend</u>.

1, Relief.

2, Continuous sections interesting the entire Pliocene (chronostratigraphic control provided by foraminifers and/or nannoflora).

3, Sections comprised between 5.2 and 3.2 Ma (chronostratigraphic control provided by foraminifers and/or nannoflora).

4, Sections comprised between 5.2 and 3.2 Ma (chronostratigraphic control provided by foraminifers and/or nannoflora, and by magnetostratigraphy).

5, Sections comprised between 3.2 and 2.6 Ma (chronostratigraphic control provided by foraminifers and/or nannoflora).

6, Sections younger than 2.6 Ma (chronostratigraphic control provided by foraminifers and/or nannoflora).

7, Sections younger than 2.6 Ma (chronostratigraphic control provided by foraminifers and/or nannoflora, and by magnetostratigraphic and radiometric datations).

8, Lower to Middle Pliocene sections with unsatisfactory chronologic control.

9, Section younger than 2.6 Ma dated by paleomagnetic and radiometric measurements.

10, Section younger than 2.6 Ma dated by mammals and magnetostratigraphy.

<u>Localities</u>.

1, Rio Maior (Diniz, 1984). 2, Carmona (Suc & Ferrier, unpublished). 3, Andalucia G1 (Bessais, unpublished). 4, Habibas 1 (Suc, 1989). 5, Arzeu 1 (Abdelmalek, unpublished). 6, Sidi Brahim (Abdelmalek, unpublished). 7, San Onofre (Bessais & Cravatte, 1982). 8, Tarragona E2 (Bessais & Cravatte, 1982). 9, Garraf 1 (Suc & Cravatte, 1982). 10, Papiol (Suc & Cravatte, 1982). 11, Ciurana (Suc & Cravatte, 1982). 12, Bañolas - Bobila Ordis (Julia & Suc, 1980; Leroy, 1990). 13, Canet 1 (Cravatte *et al.*, 1984). 14, Le Boulou (Suc *et al.*, 1992). 15, Autan 1 (Cravatte & Suc, 1981). 16, Cap d'Agde 1 (Suc, 1989). 17, Cessenon (Suc, 1981; Suc & Drivaliari, 1991). 18, Bernasso (Suc, 1978; Leroy, 1990). 19, Celleneuve (Suc, 1973). 20, Pierrefeu 1 (Suc, unpublished). 21, Iscles 1 (Suc, unpublished). 22, Cagnes-sur-mer - La Combe (Zheng, 1990). 23, Saint-Martin du Var (Zheng, 1990). 24, Saint-Isidore (Zheng, 1990). 25, Castello d'Appio (Zheng, 1990). 26, Cava di Villanova (Zheng, 1990). 27, Stirone (Bertini, 1992). 28, Monticino 87 (Bertini, 1992). 29, Maccarone (Bertini, 1992). 30, Camerota (Brenac, 1984). 31, Crotone (Combourieu-Nebout, 1990; Combourieu-Nebout & Vergnaud Grazzini, 1991). 32, Capo Rossello (Suc & Bessais, 1990). 33, Punta Piccola (Suc, unpublished). 34, Oued Tellil (Suc, unpublished). 35, Oued Galaa (Suc, unpublished). 36, Jeriba (Suc & Ferrier, unpublished). 37, Ticleni (Drivaliari, 1993). 38, Lozenec (Drivaliari, 1993). 39, Ravno-Polé (Drivaliari, 1993). 40, Nireas 1 (Drivaliari, 1993). 41, Nestos 2 (Drivaliari, 1993). 42, Kremmidia (Combourieu-Nebout, unpublished). 43, Aghios Vlassios (Drivaliari, 1993). 44, Naf 2 (Drivaliari, 1993). 45, Gan Yavne 5 (Drivaliari, 1993).

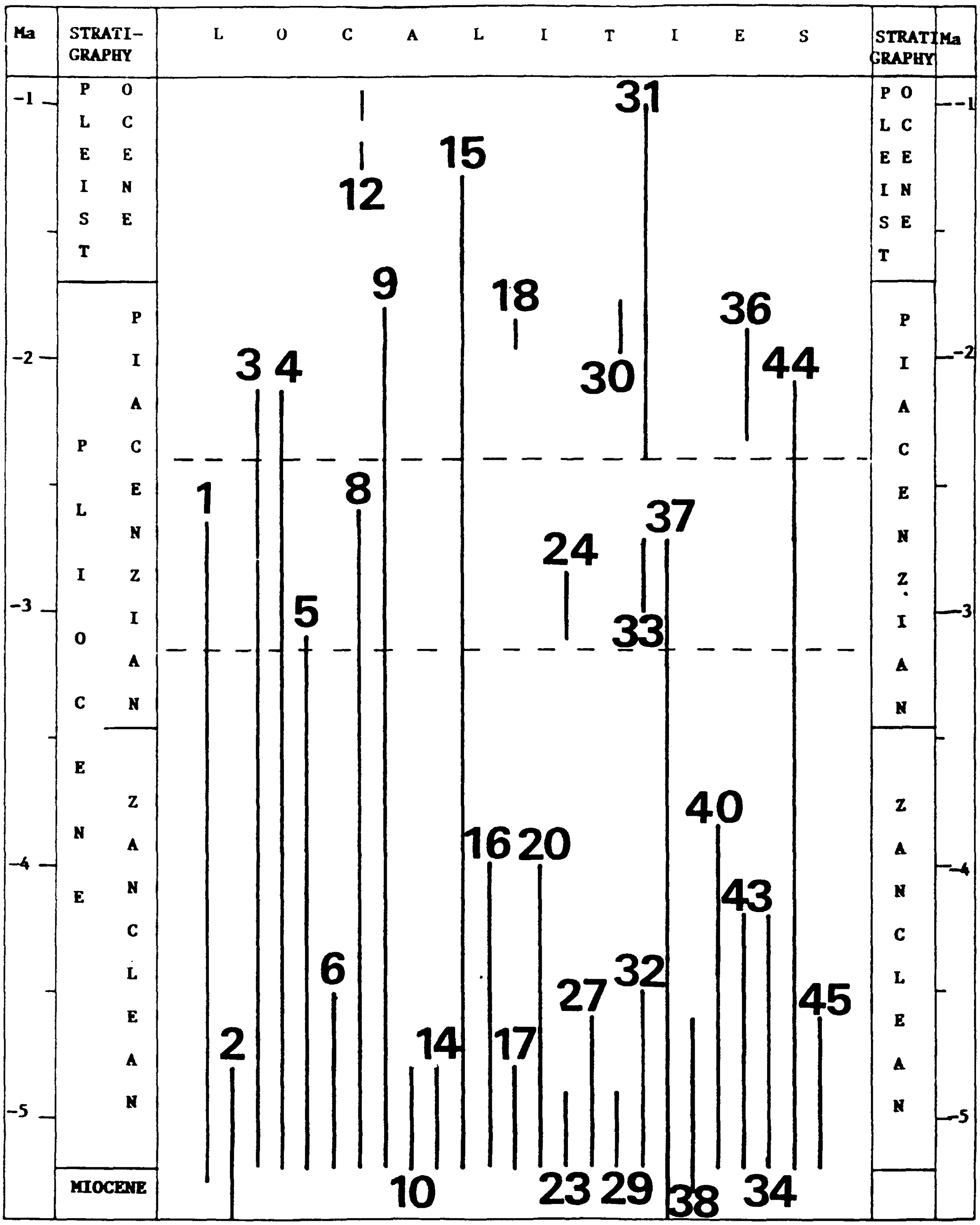

Fig. 2. **Chronological assignment of some selected Pliocene sections in the Mediterranean region.**
Locality numbers, see fig. 1.

Steps Toward Drier Climatic Conditions In North-Western Africa During The Upper Pliocene

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A 200 m long marine pollen record from ODP 658 (21°N, 19°W) reveals cyclic fluctuations and long-term variation in vegetation and continental climate of north-western Africa from 3.7 to 1.7 Ma.

ODP Site 658 (Ocean Drilling Program, leg 108) is situated north-west of Africa (Fig. 1) at a water depth of 2263 m on the continental slope 160 km west off Cape Blanc. It is located below an important near-shore up welling cell induced by the trade winds. The position at a terrace of the continental slope between two major canyon systems restricts disturbance of the sediment record by lateral down-slope transport of particles to a minimum (Ruddiman *et al.*, 1988). The upper 100 meters of sediment covering the Brunhes Normal Polarity Chron and the lower 200 meters representing the Upper Pliocene are separated by a hiatus spanning the Lower Pleistocene (Sarnthein and Tiedemann, 1989). The high sedimentation rate, due to high organic production in the up welling zone combined with high Saharan dust influx, provides a Plio-Pleistocene record of high quality in which bioturbation hardly obscures the fine-scale resolution up till 1000 years (Tiedemann *et al.*, 1989).

The time-scale of the sequence (Fig. 2) is provided by biostratigraphy, palaeomagnetism, and oxygen isotope stratigraphy (Ruddiman *et al.*, 1988; Sarnthein and Tiedemann, 1989; Tiedemann, 1991). The ages of the isotope stages of ODP 658 are derived by comparison with ODP 659 (18.05'N, 21.02'W; Tiedemann, pers. comm. 1992). That time-scale reaches back 5 Ma (Tiedemann and Sarnthein, in press). The ages of the ODP 659 time-scale are similar to the independent calibrations of Shackleton *et al.* (1990) on Pacific sediments (ODP 677) and of Hilgen (1991) on Mediterranean sapropels. However, the ages of ODP 659, and thus of ODP 658, are consistently 0.13 Ma older than those obtained by Raymo *et al.* (1989) from the N Atlantic (DSDP 607). Correlation of ODP 658 with ODP 659 and DSDP 607 revealed a hiatus at 154 mbsf and several coring gaps (Tiedemann, 1991). The time resolution of the pollen record of ODP 658 (401 samples) exceeds 1 sample per 5 ka except for the mentioned gaps and spans the period from 3.7 to 1.7 Ma.

Nowadays, at latitudes between 19°N and 21°N, the climatic sensitive Sahelian vegetation gives way to the desert. In the adjacent East Atlantic, at 21°N, ODP 658 is located where north-easterly trade winds are overlain by the mid-tropospheric African Easterly Jet (Fig.1; AEJ, summer maximum of the Saharan Air Layer). Trade winds transport pollen from their source areas in the Mediterranean and the Sahara to the marine site (Hooghiemstra *et al.*, 1986). Dust and pollen from the Sahel and the southern Sahara are carried into altitudes of the AEJ (1000-5000 m) by strong, heat induced squall lines. Then, the AEJ carries pollen from latitudes between 16°N and 20°N westwards and northwards over the Atlantic. Modern transport of pollen grains by river discharge and ocean currents is insignificant (Hooghiemstra, 1989; Dupont and Agwu, 1991).

For the Pliocene and the Pleistocene, the sedimentology of ODP 658 shows dust transport into the Atlantic by winds (trades and African Easterly Jet) as well as clay transport by rivers (Tiedemann *et al.*, 1989; Tiedemann, 1991). On the one hand, quartz content and siliciclastics (eolian dust > 6 micron) indicate wind vigor, while on the other hand, the clay content illustrates the importance of river discharge to the formation of the sediment. The pollen record of ODP 658 confirms the conclusion drawn from sedimentary analysis (Tiedemann, 1991) of persistent river discharge prior to 3.4 Ma. Afterwards, river discharge subsequently ceased. River-borne pollen seems abundant until 2.97 Ma.

An estimation of trade wind vigor is given by the sum of those pollen taxa that have their main source areas in the northern Sahara and North Africa: *Ephedra, Artemisia,* and *Pinus,* plus Asteraceae Liguliflorae for periods after 2.9 Ma. Generally, the strength of the trades was much lower during the Pliocene than during the late Pleistocene (Fig. 3). Trade winds were very weak until 3.17 Ma resulting in low transport of pollen from north of the Sahara. However, at 3.26 Ma, trade-wind vigor probably increased for a short period. Stepwise increase of the level of trade-wind strength is found at 3.17, 2.76, and 2.61 Ma. During the final Pliocene, trade winds are rather strong with exception of the period between 1.87 and 1.85 Ma. The record of trade-wind indicators corroborates the estimates of wind strength by grain size analysis of ODP 658, ODP 659 and DSDP 397 showing an increase of the trade winds between 3.2 to 2.6 Ma (Tiedemann *et al.*, 1989; Tiedemann, 1991).

Studies on time-slices of the Holocene optimum and the Last Glacial Maximum revealed the latitudinal stable position of the AEJ during the last glacial-interglacial cycle and the increasing strength of the trade winds during glacial times (Hooghiemstra, 1988). During the Brunhes Chron, latitudinal shifts up to 10° of latitude for the vegetation zones of Sahara, Sahel, and savanna were registered through the AEJ carrying pollen grains from those vegetation zones, which occurred at latitudes between 16°N and 20°N. These shifts are reflected in the progression of percentage maximums of Cyperaceae, Poaceae, CCA (sum of Caryophyllaceae, Chenopodiaceae, and Amaranthaceae), *Artemisia,* and *Ephedra* (Fig. 2H, arrows; Dupont and Hooghiemstra, 1989; Dupont *et al.*, 1989). Each of them is interpreted as a southward extension of dry vegetation or even deserts and, therefore, as a reflection of drier climate. Few of the shifts have been recorded before 3 Ma, but they regularly occur from 2.61 Ma onwards.

The pollen record of ODP 658 shows a large number of short-time (< 50 ka) fluctuations (Fig. 2) in vegetation and climate from 3.74 to 1.71 Ma corresponding to the oxygen isotope record: dry periods correlate with large global ice-volume and humid periods with small global ice-volume. On top of these cycles, several periods, at ca. 3.5, 3.2, and 2.5 Ma, mark irreversible changes in the development of vegetation and climate.

A humid, probably warm, climate with weak trade winds prevailed from 3.74 to 3.48 Ma (Zone I). According to percentages of *Rhizophora* pollen exceeding 10%, mangrove swamps were growing near Cape Blanc around 3.70 Ma, probably in connection with a paleoriver. Percentages of the sum of pollen from Sudanian and Guineian vegetation, i.e. from wooded savanna, woodland and tropical forest, repeatedly exceed 5%. It indicates that forest and savanna had a distribution at least as north as 21°N. The period may be correlated in northwestern Europe to Brunssumian C of the Dutch palynostratigraphy (Zagwijn, 1960) and Zone Plc of the north-western Mediterranean area (Suc, 1984).

The period from 3.48 to 3.25 Ma (Zone II) shows five dry phases. Three of them, at 3.48, 3.35, and 3.26-3.27 Ma, show high percentages of Amaranthaceae-Chenopodiaceae pollen (50%) indicating arid conditions. Two other dry phases, at 3.44-3.40 and 3.31 Ma, are less prominent: they show high percentages of Caryophyllaceae but the percentages of Amaranthaceae-Chenopodiaceae hardly exceed 40%. High percentage values (> 20%) are found for Asteraceae Liguliflorae between 3.44 and 3.40 Ma, and at ca. 3.27 Ma presenting a no-analogue situation. The youngest and most arid phase shows a percentage maximum of *Ephedra* of 4% at 3.26 Ma. This phase could correspond to a drop in winter sea surface temperatures in the Mediterranean Sea (Zachariasse *et al.*; 1990) and a decline of thermophilous elements in pollen records of coastal swamps and mountain slopes of the Mediterranean area (Suc *et al.*, 1992). From the sedimentary record of ODP 658, Tiedemann (1991) concluded an increase of eolian activity, especially trades, around 3.26 Ma. The period from 3.48 to 3.25 Ma records through the course of several oscillations a trend toward more aridity and an extension of open vegetation types in north-western Africa probably in relation to the onset of the trade winds and the first Northern Hemisphere's glaciations.

Humid conditions re-established between 3.25 and 3.19 Ma (beginning of Zone IIIa) indicated by high percentages of tropical forest (> 2%) and Cyperaceae (> 15%). Afterwards, the climate progressively becomes drier again and percentages of Cyperaceae decline (end of Zone IIIa). During the next period from 2.97 to 2.61 Ma (Zone IIIb), prevailing percentages of grass pollen (Poaceae up to 70%) are followed by percentage maximums of CCA (> 50%), at 2.73 and 2.69 Ma, *Ephedra* (> 5%) at 2.69 Ma, and the first maximum of *Artemisia* (2%) at 2.66 Ma. A slight increase in trade-wind strength occurred at 2.76 Ma. The period may be correlated to the Reuverian in the Netherlands and to the Mediterranean pollen zone PII (Zagwijn and Suc, 1984).

The isotope Stages 104 (2.60 Ma), 100 (2.53), and 98 (2.49 Ma) record high percentages of *Ephedra* (ca. 3%), *Artemisia* (> 2%), and CCA (> 50%) indicating severe dry periods. They start a climatic regime in north-western Africa resembling glacial to interglacial cycles that result in arid cold and humid warm phases. The period is correlated with the Praetiglian and the lower part of the Lieth-series (Ekholt Glacial; Mencke, 1975). Within the period between 2.6 and 1.7 Ma (Zone IV), only two extended humid periods occur corresponding to the weakly developed isotope Stages 76 and 68.

Long-term variation (Fig. 3) indicates a first step toward drier climate between 3.5 and 3.2 Ma and a second, stronger one starting at about 2.6 Ma. The percentage maximums of *Rhizophora* (> 5%) indicate that until 1.9 Ma, mangrove swamps irregularly occurred near Cape Blanc (i.e. 5° north of their present northern limit). Prior to 3.5 Ma, and between 3.25 and 2.6 Ma, mean percentages of tropical forest elements (> 2%) indicate a northern extent of forests that probably shifted southwards after 2.6 Ma. After 2.8 Ma, a declining trend in mean percentages of Poaceae indicates a reduction of savanna vegetation (including wooded savanna and dry open forest), probably as a result of the development of a desert vegetation in north-western Africa. Percentages of Asteraceae Liguliflorae reach high values prior to 3.2 Ma, but decline to low values afterwards. From 2.7 Ma on, mean percentages of CCA reach high values (around 40%), comparable to those for the Brunhes indicating aridity in north-western Africa. Mean percentages of *Ephedra* and *Artemisia* of the Pliocene are five times lower than those of the late Pleistocene indicating that arid periods were still less prolonged and/or less severe during the Pliocene.

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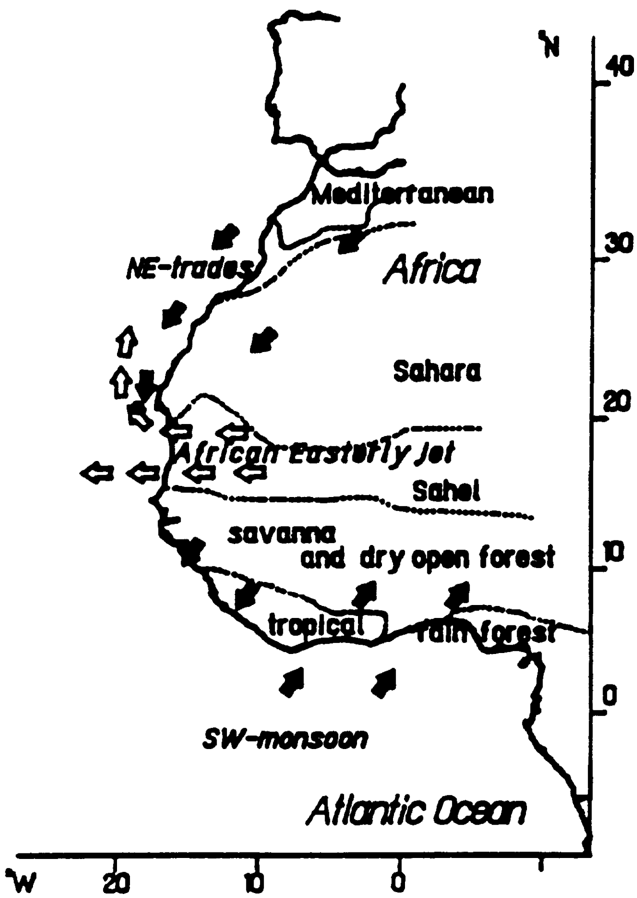

Figure 1. Location of ODP Site 658 (21°N, 19°W; black dot), surface winds (trades and monsoon; black arrows) and mid-tropospheric winds (African easterly Jet; open arrows), modern vegetation zones (Mediterranean, Saharan, Sahelian, savanna and dry open forest, tropical rain forest).

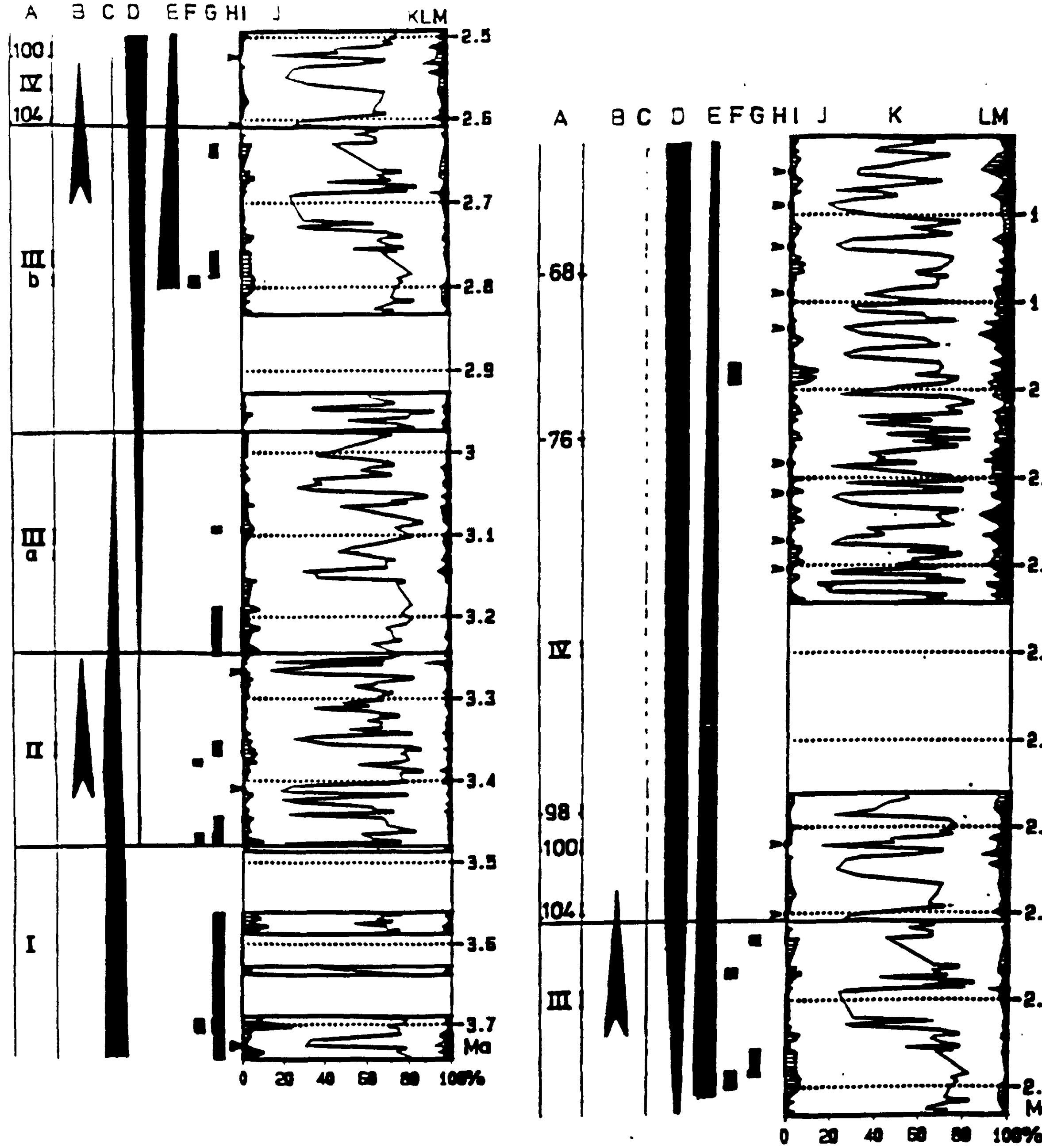

Figure 2. From left to right: (A) Pollen zones in roman numbers, Arabic numbers refer to oxygen isotope stages. (B) Arrows indicate the steps toward drier climatic conditions. (C) Declining trend in river discharge. (D) Increasing trend in trade-wind vigor. (E) Declining trend in grass-rich vegetation types like savanna and open forest. (F) Occurrence of mangroves near Cape Blanc. (G) Extent of the tropical forest to 21°N. (H) Arrows denote levels that show a progression of percentage maximums of pollen taxa indicating a southward shift of dry vegetation. (I) Percentages of *Rhizophora* plus tropical forest elements (hatched left). (J) Percentages of Cyperaceae plus Poaceae (open left). (K) Percentages of CCA (sum of Caryophyllaceae and Amaranthaceae-Chenopodiaceae) plus Asteraceae (open right). (L) Percentages of *Artemisia* plus 'Mediterranean' elements (hatched right). (M) Percentages of *Ephedra* (black). Time-scale in Ma on the vertical axis after Tiedemann *et al.* (in press). Oxygen isotope taxonomy after Tiedemann (1991).

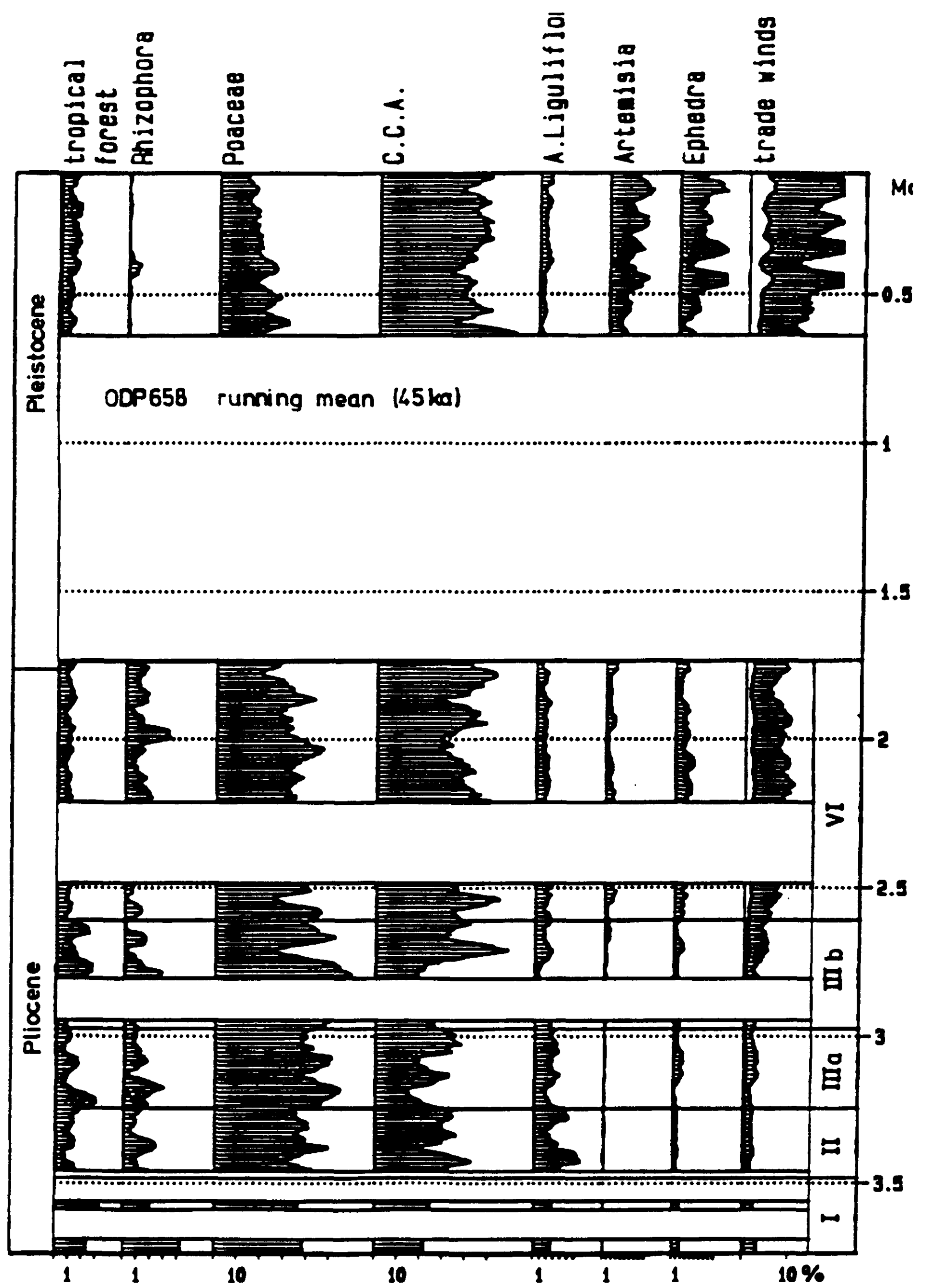

Figure 3. Running mean (9 equidistant samples 5 ka apart) of percentages of the following pollen taxa: tropical forest, *Rhizophora*, Poaceae (grasses), CCA (sum of Caryophyllaceae and Amaranthaceae-Chenopodiaceae), Asteraceae Liguliflorae, *Artemisia*, *Ephedra*, and trade-wind indicators. Time-scale in Ma on the vertical axis.

An Attempt To Reconstruct Temperature And Rainfall From The Pliocene Pollen Record In Ethiopia

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The deposits of the Hadar Formation have recently been redated using single-crystal laser-fusion $^{40}Ar/^{39}Ar$ technique (Walter and Aronson, in press) to between 3.4 ± 0.04 Ma and 2.92 ± Ma (Walter, 1993). The best palynological data from this formation is from an organic clay below the Kadamoumou Basalt (Bonnefille *et al.*, 1987). Based on an estimated mean sediment accumulation rate of 80 cm/kyr (Walter, in press), these sediments were apparently deposited over a few tens-of-thousands of years (20 to 30,000 years?) that began a few thousand years after the deposition of the volcanic ash SHT, which is precisely dated at 3.4 ± 0.04 Ma (Walter and Aronson, in press).

The new presentation of the results in this report focuses on a comparison of the fossil data with an extensive modern data set (118 spectra) covering all the vegetation types encountered on an altitudinal gradient from 500 to 4000 meters elevation. Statistical Factorial Analysis was carried out both on the modern (118) and the fossil pollen spectra (18), with the fossil data being considered as supplementary elements. In this procedure, local aquatics and grasses are excluded, and the statistical treatment is applied to the pollen counts for the 189 identified taxa. The statistical results show a distribution of the modern samples into four distinct altitudinal vegetation zones, with the fossil samples from Hadar linked to modern samples taken from vegetation existing now at altitude above 500 meters. Therefore, the statistical analysis confirms previous interpretation of the past vegetation interpreted as similar to that presently known in the Ethiopian mountains. However, the interpretation of the pollen record should take into account possible rifting that might have occurred since the deposition of the Hadar Formation, and could produce, in the pollen assemblages, the same changes as those indicating a noticeable cooling. Although the Pliocene pollen data can be interpreted as indicating conditions much more humid than the arid climate that prevails in the Hadar region today, the factorial analysis indicates close affinities with modern pollen data from the dry forests of northeastern Ethiopia but not with the wettest forests of southwestern Ethiopia. This is a clear indication of the eastern limit of humid lowland and highland forests, and that they did not reach the 40° longitude at that time.

Palynologists have developed a statistical methodology which enables the quantitative reconstruction of climatic parameters. We have been successful in reconstructing such parameters for the Holocene and Late Glacial pollen data in Central Africa. Because the Pliocene pollen can be identified by comparison with the modern taxa, an attempt was made to apply such a quantitative reconstruction of climatic parameters to the Pliocene record, using the modern pollen data set from Eastern Africa. The results of this essay show that a possible short cooling event associated to more humid conditions could have occurred around 3.4 Ma ago. But quantitative estimate of this cooling would need a better evaluation for the amplitude of the Ethiopian plateau uplift in the Hadar area. The modern pollen data set from Ethiopia expresses a clear pattern for the decreasing temperature gradient with increasing altitude, nevertheless a more reliable result can only be achieved when pollen data from the Last Glacial Period in Ethiopia are obtained, in order to constrain the coefficients for evaluating the climatic parameters.

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Some Manifestations Of Pliocene Warming In Southern Africa

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Pliocene vegetation types in southern Africa, especially those of the Middle Pliocene, are not well documented. They can be reconstructed in part on the basis of palynological data from the interior and the western coastal and offshore regions. Pollen spectra from marine deposits off the northern Namibian coast (van Zinderen Bakker, 1984) which can broadly be assigned to the middle Pliocene, contain relatively prominent Chenopodiaceae pollen, and are interpreted here as reflecting a tropical climate, possibly associated with more intense evaporation. Early Pliocene pollen accompanying fauna from the Varswater Formation at Langebaanweg, in the "Mediterranean" southwestern Cape Province, indicates that swampy conditions existed in a fynbos (macchia) environment, with a suggestion of slightly more tropical characteristics than at present (Scott, in press). Middle- to late-Pliocene (ca. 3 Myr) cave breccias at Makapansgat in the interior of the subcontinent (Cadman and Rayner, 1989), and a coprolite from these layers (this report), contain "Bushveld" pollen types, but exotic *Pinus* contaminants suggest that the spectra were probably derived from the modern environment. Pollen from terminal Pliocene to early Pleistocene travertines in the Sterkfontein and Kromdraai sites of the highveld grassland region, tentatively indicates open *Protea* savanna at that time. In comparison with older strata from Sterkfontein, these deposits contain a fauna indicative of relatively open vegetation (Vrba 1985). It can be assumed therefore, that the preceding mid-Pliocene vegetation contained denser woodland, although there are no independent palynological data to support this.

Geomorphological and sedimentological data provide independent support for some of these conclusions and provide further insights into the amplitude and timing of Pliocene changes in the subcontinent. High marine terraces at elevations up to 90 m are associated with an extensive terrestrial fauna at Langebaanweg on the western Cape coast, and provide important proxy evidence of early - mid Pliocene deglaciation. The upper terraces of the Vaal River contain sparse Pliocene fossils; their massive gravel armoring is indicative of semi-arid alluviation, probably towards the end of the Gauss chron. At about the same time a shift is evident within the hominid-bearing cave deposits of the interior from sub aqueous accumulation to of predominantly fine clastic sediments to an influx of courser colluvial elements under the influence of episodic sheetfloods. Major carapaces of the Ghaap escarpment, which accumulated between 2.4 Myr and the middle Pleistocene can also be linked to cycles of spring activity under semi-arid climatic conditions. In sum, a growing body of evidence is coming to hand which documents an important shift from relatively warm, mesic conditions during the Pliocene to the recurrent cool, relatively dry cycles which typified much of the Quaternary.

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Landscape And Climate Of The Southwestern Russian Plain In The Pliocene

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Along the northern coast of the Black sea on the southwestern Russian Plain, interbedded continental and marine sediments cover the entire Pliocene. The fossil floras and faunas of these deposits have been studied in detail, providing the basis for paleoclimatic interpretations and biostratigraphic correlations. This report discusses the most complete sections from this region, concentrating on sites where there are distinct boundaries between stages and horizons and where there are abundant paleontological data from marine mollusks, gastropods, ostracodes and terrestrial mammals and palynological data. The Pliocene of this region is divided into three stages (based on paleontological assemblages) which are (from older to younger) the Pontian, Kimmerian, Kuajlnician. Nine sections were studied from fluvial terraces above the flood-plains of the rivers Byk and, Kuajlnik; from the coastal terrace scarps above Khadzibejsky bay and from the Kertch peninsula of Crimea (Fig. 1).

Paleomagnetic studies for each of these sections were carried out in Laboratory of the Moscow University following the method of A. Khramov and L. Cholpo (1967). Paleoclimatic reconstructions are based on paleofloristic data from the above mentioned sections studied by S.V. Siabraj and on published materials from adjacent regions (A. Negru (1986), S. Medianik (1985), and N. Shchekina (1979) The methodology used to estimate paleoclimatic conditions on the basis of fossil floras was developed by V. Grichuk (1987), following the concepts of J. Iversen (1941) and modified to be applied to Tertiary floras. In the following text I present paleogeographical reconstructions developed from our paleomagnetic and paleofloristic research, combined with previously published information.

Pontian Stage (5.4-4.7 Ma)

During the Novorossijskoe substage, at the beginning of the Pontian age, the Zanklian transgression united the Black and Caspian Seas with the Mediterranean. By the end of early Pontian time (~5.0 to 4.9 Ma?) the sea retreated southwestward and the greater part of the southwestern Russian Plain was exposed.

Marine faunas indicate that the early Pontian sea was closed and brackish. The mollusks assemblages from these deposits include *Unio* and *Anodonta,* while the ostracode fauna includes *Caspiella, Bakunella* and *Leptocythere.* From these data Ivchenko (1986) estimated that the mean annual water temperature was near +12° to +15°C (today near the city of Odessa the mean annual water temperature is +9.6°C). In the early Pontian forest-steppe prevailed in the western part of the region under investigation — pine forests grew in the river valleys, and the northern part was occupied by mixed conifer-broadleaved forests.

By the middle of Novorossijskoe time some aridification had taken place in the east and the role of open vegetation communities increased. Chenopodiaceae were important in these communities and possibly these included halophytes, which may have occupied the recently exposed Black Sea shelf.

In the west such changes of the climate were not registered and forest-steppe vegetation continued to grow. By the end of this period forests dominated in both the east and in the west. These were primarily broadleaved formations with *Ulmus, Zelkova,* and *Fagus.* The climate was similar to the contemporary climate of the southern Crimea coast, but the presence of exotic taxa suggest that it was something warmer than today

(January mean temperature ~ +5°C and July ~ +25°C)

The Pontian stage of the history of the Hipparion fauna is poorly represented in the former USSR. Only scattered small occurrences of mammals bones have been found in southern Ukraine. This "Tauric complex" (Korotkevich, 1988) includes the first occurrence of *Paracamelus*, while typical members of the Hipparion fauna were not present at this time. The increased abundance of ostriches suggests greater availability of open dry areas. However, the presence of certain water-swamp birds and tortoises points to the development of marshlands and warm-water environments, whereas the abundance of *Cervus* and *Proboscidea* indicate the presence of wooded vegetation. On the whole, this faunal assemblage reflects moderately warm and relatively dry climate.

Kimmerian Stage (4.7-3.4 Ma)

In the middle Pliocene more land was exposed and sea waters were confined to within the limits of the modern Black Sea and Caspian Sea depressions. The Kimmerian Black Sea basin was closed with brackish water. The marine fauna of this period included Dreissenidae, Cardiidae, endemic gastropods, *Viviparus, Melanopsis, Planorbis* and the freshwater ostracodes *Caspiela* and *Caspiocypris*. The majority of the marine fauna had a boreal character with few Mediterranean taxa. At this time the broad-leaved forests were composed mostly of heat-loving and moderately heat-loving species: *Acer, Quercus, Juglans*, and less common *Castanea, Morus, Carya, Fraxinus*, and *Tilia*. By the end of this period the proportion of steppe elements increased and steppe and forest-steppe landscapes appeared. Paleobotanic data from middle Kimmerian sediments indicate a warmer and more humid climate than previously. During the first half of the Kimmerian, the percentage of subtropical genera in the flora (*Pittosporum, Parrotia, Nyssa, Aralia, Magnolia, Staphylea, Taxodium*) increased over the levels seen during late Pontian, suggesting that temperatures had increased. The maximum temperature increase occurred during deposition of the sediments of the Kamyshburun horizon (4.2 - 4.0 Ma). Mean January temperature rose by almost 2°C (to 6°C) and mean July temperature rose from 22°C to 23°C.

The Kuchurganian and Moldavian faunal complexes that developed during the Kimmerian time indicate the progressive aridification. The earlier (Kuchurganian) included in its composition the main elements of the Hipparion fauna: *Deinotherium, Zygolophodon, Hipparion*, and *Tapirus*, and is largely dominated by forest and forest-steppe species. The Moldavian faunal assemblage reflects the progressive drying of the climate: taxa characteristic of humid environments disappear and inhabitants of open areas such as Paracamelus and Equus (which makes the first appearance here) and more moderate climate appear (Korotkevich, 1986). The climate was apparently humid and moderately warm (January - above +4°C; July ~ +22°C).

Kuajlnician Stage (3.4-2.3.Ma)

In the late Pliocene this region was an elevated plain on which the modern river network was formed. the sea basin remained only at the south of the area. It was a closed basin with brackish water , but its salinity was less than in Kimmerian time. The marine fauna was inherited the main features of Kimmerian fauna, but was less diverse. The mollusks *Limnocardium*, and *Dreissena* appeared during the Kuajlnician, and gastropods were represented primarily by the fresh-water forms *Planorbis, Valvata*, and *Melanopsis*. Ostracodes include the typically fresh-water and brackish-water taxa *Cypria kurlaevi* and *Cyprideis*.

Mixed conifer/broad-leaved forests grew along the northern Black Sea coast at the beginning of this time, suggesting more humid conditions than earlier. *Acer, Tilia, Fagus*, and *Quercus* grew with more thermophilous trees such as *Pterocarya, Zelkova, Fraxinus, Morus*, and *Juglans*. To the east relatively moist steppe vegetation was present. Compared with that of the

Kimmerian, the terrestrial vegetation of Kuajlnik time had lower representations of broad-leaved trees and higher levels of steppe taxa. The climate was rather warm and humid, with mean January temperature near +4°C and July mean temperature near +22°C.

A substantial reduction of the forest area occurred at the end of the Kuajlnik time, and among the remaining trees pine and dark-coniferous types (*Picea, Abies, Tsuga,* etc.) prevailed. The diversity and number of broad-leaved trees and other warmth-loving species decreased. Pollen spectra of Kuajlnik deposits are close in composition to those of the early Pleistocene and only rare pollen grains of Myricaceae, Moraceae, Rutaceae and *Taxodium* allow us to assign these deposits to the Pliocene. Collectively, these data point to an intensification of aridification and a decrease in temperature.

This is the time of the appearance of the Chaperon faunal assemblage which includes many modern elements. Warmth-loving animals (giraffes, *Tapirus, Deinotherium*) disappeared by this time and the warm-moderate types (apart from *Archidiscodon, Equus, Cervus, Dicerorinus etruscus,* etc.) occured more widely (Korotkevich,1988). Fauna data indicate arid climatic conditions and small mammal assemblages indicate steppe vegetation. The climate was apparently dry with moderately warm temperatures (January about +4°C; July +22°C).

Conclusions

In the beginning of the Pliocene forest-steppe vegetation covered most of the southern Russian Plain. By middle Pliocene time temperate broad-leaved forests begin to prevail, and by the late Pliocene increasingly arid and cold conditions led to more open landscapes with steppe and forest-steppe vegetation. The warmest climates of the Pliocene occurred in middle Kimmerian time (~4.2 to 4.0 Ma), but warmer than modern conditions continued until approximately 2.5 Ma.

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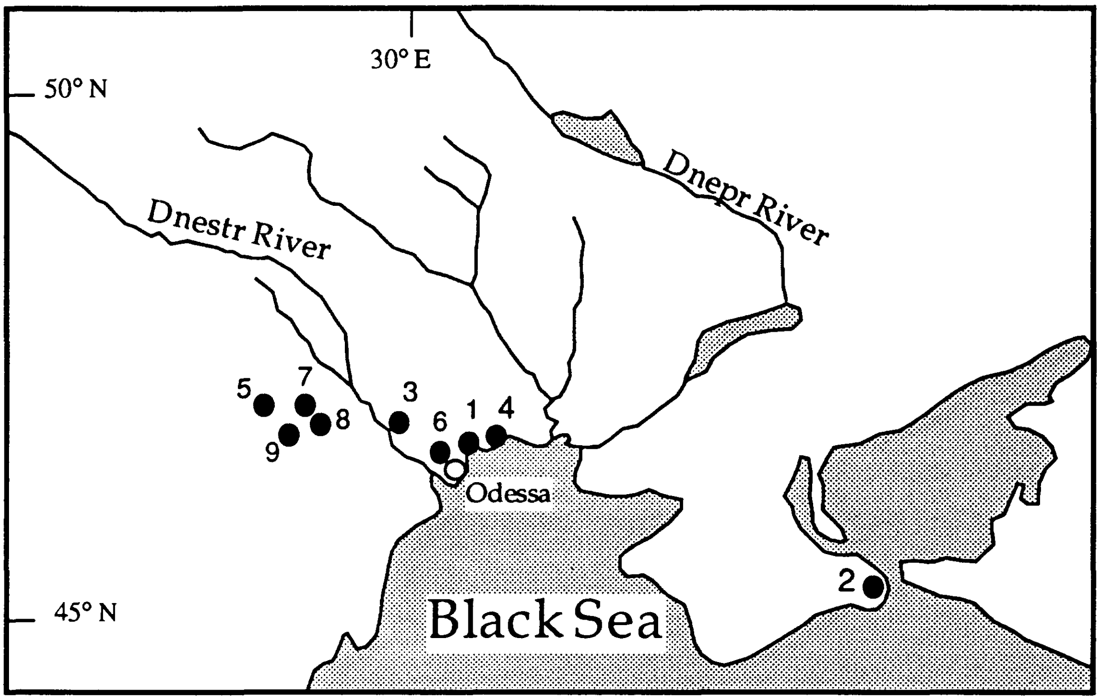

Figure 1. Site Locations, Southwestern Russian Plain

1. Big Fontan	2. Eltigen	3. Grebenniki
4. Cherevychny	5. Vinogragovka	6. Novoelizavetovka
7. Ripa Skorcelskaya	8. Dolinskoe	9. Borisovka

Landscape And Climate Of The South-Central And Southeastern Russian Plain In The Pliocene

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Pliocene conditions on the Russian Plain can be reconstructed from data provided by many researchers over the past 50 years. Palynological studies were carried out on the Russian Plain by Drs. E. Ananova, V. Grichuk, N. Shchekina, S. Syabray, N. Kovalenko, Yu. losifova and many others; plant macrofossils were studied by A. Negru; and faunal research was done by E. Korotkevich (large mammals), V. Topachevsky and R. Krasnenkov (small mammals), G. Karmishina (ostracodes), L. Il'ina (gastropods), L. Nevesskaya and A. Chepalyga (mollusks). Paleoclimatic estimates were obtained from the modern climatic ranges of Pliocene plant taxa following the climagram methodology of V. Grichuk (1985).

Pliocene deposits of the south-central and southeastern Russian Plain are correlated with the stratigraphical sequence of the circum-Black Sea region, which in turn is correlated (on the basis of faunal and paleomagnetic data) with the Mediterranean stratigraphical framework. Pliocene stratigraphy of the Don Plain is not worked out in detail, as the sediments of the age are usually of fluvial origin, and thus their distribution is discontinuous, and detailed inter-correlations are impossible. Generally the Pliocene deposits of this region are subdivided into the Usmanskaya and Krivoborskaya suites, which in turn include a number of subdivisions. The sediments are mainly sands, loams and clays. The deposits of three fluvial cycles can be found within the Krivoborskaya suite, the middle one corresponding to the Caspian Akchagylian stage and the upper one to the Caspian Apsheronian stage (Nalivkin, Sokolov, *et al.*, 1986).

Pliocene deposits in the south-astern Russian Plain are more widespread than on the Don Plain. Sediments of Pontian age (early Pliocene, approximately 5.4 - 4.7 Ma) are found in the southwest of Astrakhan district and in Kalmykiya, where they are represented by limestones and clays with interbedded sands and shells (the Novorossiysk substage). These sediments contain rare shells of *Congeria* sp. and a variety of ostracodes: *Cyprideis torosa, Bakunella guriana* and others. These deposits were probably formed in a brackish water shallow inlet of the Pontian marine basin.

Continental deposits of the Pliocene on the southeastern Russian Plain (the Kushumskaya suite) are more widespread and accumulated from late Miocene through the early and middle Pliocene time. These layers overlie eroded deposits of varying ages and origins and underlie Akchagylian marine layers.

Upper Pliocene deposits in this region are mainly of marine origin and are related to a transgressions of the Caspian Sea. The Akchagylian stage (poorly dated, but estimated to range from 3.5 - 2.0 Ma) is subdivided into three substages on the basis of mollusk and ostracode faunas and on palynological data. The Apsheronian stage (about 2.0 - 0.7 Ma) is thought to be a part of Pliocene in this region, although some researchers regard it as a part of Eopleistocene. Apsheronian deposits are very widespread and are represented both by marine and continental facies.

At the beginning of Pontian time, the Zanklian transgression united all the marine basins of southern Europe, including the Caspian Sea. At the end of the early Pontian (~5.0 Ma) there was a major regression, and the Caspian Sea lost its connection to the Black Sea. During the early Pontian in the southwestern region, forest and forest-steppe vegetation dominated under warm and rather dry climatic conditions (estimated mean January temperature was 3° C, and that of July 24° C). In the Don River basin steppe vegetation was predominant at this time, and only riparian forests existed in this region (*Populus tremula, Alnus incana, A. glutinosa,* etc.). Unfortunately the modern climatic ranges of these species are too broad to estimate paleotemperatures from this area. The southeastern part of the Russian Plain was occupied by forest-steppe and steppe vegetation. The climate was rather warm, as indicated by the presence of thermophilous mollusks and ostracodes. Caspian Sea surface temperatures were estimated from strontium isotopes on *Unio* and *Valvata* shells to range between 11° and 22° C, which is characteristic of the subtropical zone (N. Yasamanov, pers. comm.).

During the middle Pliocene (the Kimmerian stage of the Black Sea regional stratigraphy), marine basins occupied only the depressions of the Black and Caspian Seas. In the circum-Black Sea region broad-leaved forests with thermophilous elements were widespread. The reconstructed temperatures for the interval from 4.0 to 4.2 Ma are the warmest of the Pliocene, and at the end of Kimmerian the area of forest cover was reduced due to drier climatic conditions. Faunal assemblages also changed, as animals typical of open landscapes, such as *Equus* and *Paracamelus,* appeared.

Middle Pliocene deposits are scattered on the south-central Russian Plain, and palynological data are sparse. The lower Kimmerian fauna in this region includes steppe species (*Cricetulus, Spalax,* etc.). Later in the Kimmerian faunal assemblage became more heat-loving and included some tropical elements. Its composition also included characteristic forest inhabitants (for example, *Pliopetaurista*), which may imply more moist conditions. Terrestrial mollusk faunas also point to warm climatic conditions: thermophilous species of the forest-steppe zone were present, along with several subtropical taxa (*Striatura, Hawaia,* and five species of Gastropoda). Based on these data, the lower Kimmerian can be regarded as the warmest part of the Pliocene.

At the beginning of the middle Pliocene steppe landscapes dominated by Chenopodiaceae were prevalent in the southeastern Russian Plain. During late Kushum time, steppe vegetation was partly replaced by forest-steppe and forest. Along with coniferous forests, broad-leaved vegetation was widespread, with different species of *Ulmus, Celtis, Juglans, Pterocarya, Liquidambar* and others. The climate remained warm (mean January temperature = ~ 1° C, July temperature = ~ 20° C) and became more humid. Farther to the south on coast of the Caspian Sea, shells of the heat-loving mollusks *Unio* and *Valvata* were found in the layers of the same age, confirming the interpretation of a warm middle Pliocene.

The Caspian Sea regressed to near its modern level in the beginning of the late Pliocene (a continuation of the lowering of sea level that began in Kimmerian time), and a river pattern similar to that of today was established. A reduction of forest cover took place during the Kujalnik stage in the northern Black Sea region, and forest-steppe vegetation was replaced by steppe. The flora was very similar to that of today, and only the presence of rare pollen grains of exotic taxa allow us to consider these layers to be of Pliocene age. Clearly the climate became drier and cooler during the late Pliocene.

During the early Akchagylian substage (corresponding to the Kujalnik) in the Don river basin, steppe vegetation existed with patches of riparian forests. The middle Akchagylian is marked with cooling which is reflected in a spread of *Betula* (arboreal pollen spectra are dominated by *Betula* in these deposits), and micromammal faunal assemblages confirm the existence of dry open landscapes. A new stage of forest expansion indicates that warming took place later in the middle Akchagylian. More humid conditions allowed broad-leaved trees such as *Quercus* and *Tilia* to grow in this region, and thermophilous mollusks, including *Striatura, Hawaia, Parmacella* and others, were identified in deposits of this age.

Profound cooling occurred in late Akchagylian time, when exotic thermophilous taxa died out. The northern part of Don River basin was occupied by forest-steppe vegetation, while the southern part had steppe vegetation. These interpretations from pollen data are confirmed by fauna assemblage that contain the steppe taxa *Equus stenonis*, *Paracamelus* and *Struthio*. The final part of late Pliocene on the Don Plain is characterized by temperate climate and forest-steppe vegetation.

On the southeastern Russian Plain, late Pliocene vegetation was dominated by steppe and forest-steppe. Coniferous trees were more common than broad-leaved ones in the forests, and the flora included a small number of subtropical elements than in Kushum time. The paleobotanical data suggest that mean January temperature was about 0°C, and that of July was about 20°C. The ostracode assemblages include paleoarctic taxa such as *Candona* and *Iliocypris*, indicating cooler water conditions than occurred earlier in the Pliocene.

The middle part of the Akchagylian was the warmest within the late Pliocene; vegetation was dominated by mixed coniferous-broadleaved forests, and heat-loving taxa of mollusks were abundant in the Caspian Sea. Mean January temperature was about 2°C, that of July 21°C.

In the late Akchagylian a new cooling took place, marked by the spread of coniferous forests of *Pinus*, *Picea* and *Tsuga*. The flora became more depauperate than earlier, and increasing aridity is interpreted from a spread of the steppe vegetation at the end of Akchagylian time. During the final part of the Pliocene (the Apsheronian stage) further cooling and drying occurred on the southeastern Russian Plain, and steppe vegetation became dominated by *Artemisia* and Chenopodiaceae associations.

In general, for any stage within the Pliocene the same pattern can be observed: greater continentality and aridity in the east than in the west. Overall, since the time of maximum warmth during the middle Pliocene Kimmerian stage a similar pattern toward greater continentality and aridity through time can be seen in any given subregion of the southern Russian Plain. The climatic oscillations on the background of this general trend can be seen more clearly by time sequence of landscape change in the southeastern Russian Plain. The middle Pliocene (early Kimmerian) time can be regarded as the climatic optimum of the Pliocene in the southern Russian Plain.

For the warm global climates of the Pliocene, the highest amplitude climatic fluctuations were expressed in the eastern, more continental, part of the southern Russian Plain. In contrast, under the cold global climates of the late Quaternary, climatic fluctuations were greater in western Europe than in the east.

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Pliocene of the Southern Russian Plain

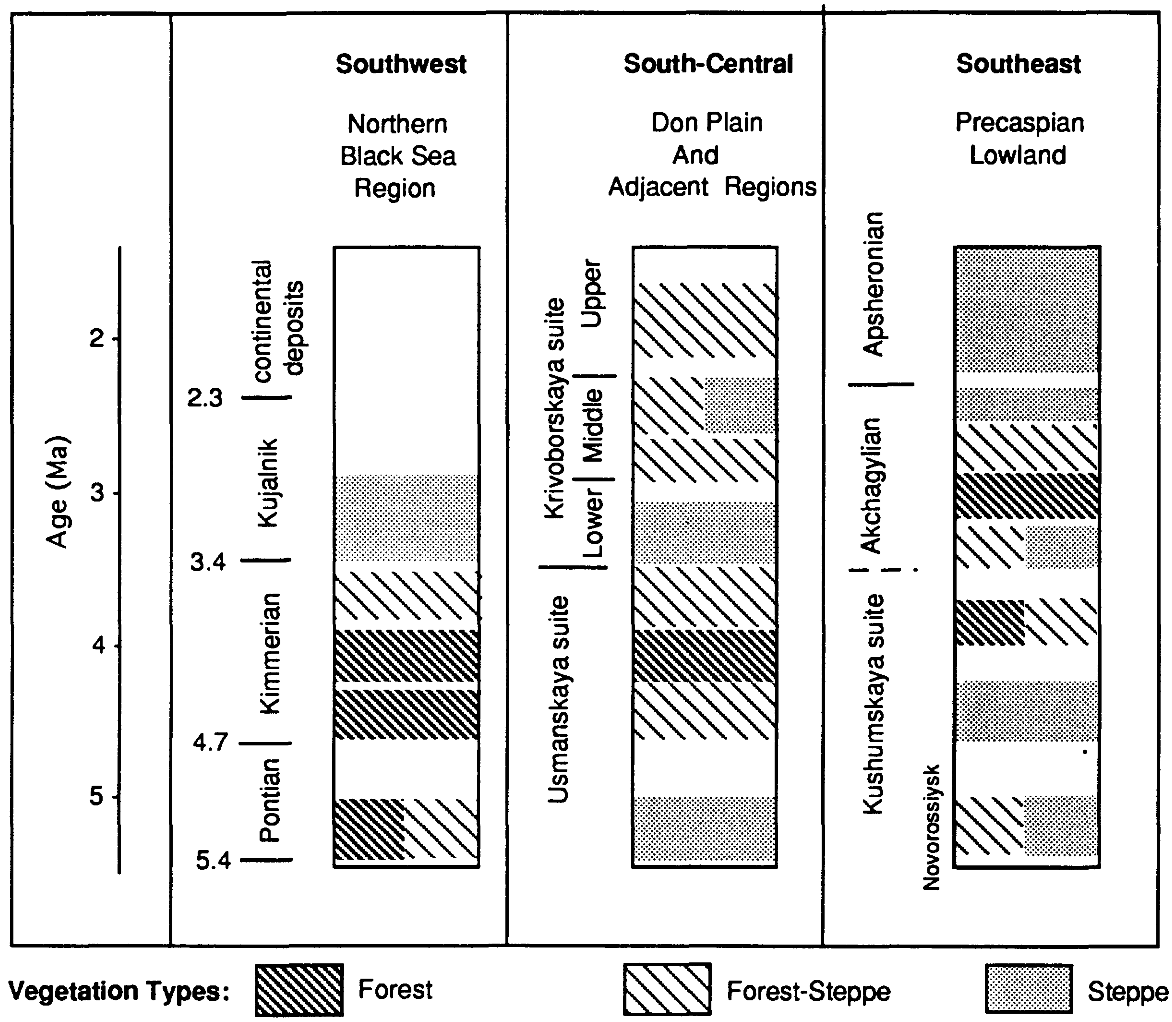

Figure 1. Pliocene stages and inferred vegetation for three regions of the southern Russian Plain.

GCM Simulations Of The Pliocene Climate: Feedbacks, Ocean Transports, And CO_2

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Estimates of sea surface temperatures (SSTs), based on marine microfauna, reveal the existence of a middle Pliocene warm period between about 3.15 and 2.85 million years ago (Dowsett *et al.*, 1992). Terrestrial pollen records, although not as well dated, also show evidence for a warmer climate at about this same time in the Pliocene and, further, indicate that continental moisture levels varied significantly from the present day. The cause of the altered climate is not known with certainty, but sensitivity experiments, conducted using the Goddard Institute for Space Studies General Circulation Model (GISS GCM), have indicated that warmer climates, such as those of the Pliocene, can be simulated by using increased ocean heat transports (Rind and Chandler, 1991) [figure 1]. Dowsett *et al.* (1992) suggested that this might be the case for the Pliocene, based on the distribution of North Atlantic SSTs.

One test of this hypothesis is to supply Pliocene SSTs, together with an estimate of the terrestrial vegetation coverage, as boundary conditions in a GCM simulation, and to examine the temperature feedbacks and moisture changes that result. Consistency between the palynological estimates of climate and the simulation results provide one level of validation for the GCM; the GCM then provides a method for investigating the atmospheric processes involved in maintaining the warmer climate.

Using the GISS GCM together with PRISM Northern Hemisphere ocean surface and vegetation boundary conditions (Chandler *et al.*, submitted; Dowsett *et al.*, submitted) we found both consistencies and inconsistencies between model and data-generated paleoclimate estimates. Generally, temperature estimates show the greatest consistency, with both model and data indicating significantly warmer temperatures at high latitudes and diminished warming nearer to the equator. The GCM yields temperature increases up to 10°C along the Arctic coasts and shows greatest warming in the winter. Although the original temperature increase is driven by warmer SSTs, much of the continental interior warming is generated by an ice-albedo feedback, as reduced snow cover in the warmer climate increases the absorption of solar radiation at the surface during winter months [figure 2]. Further warming at high latitudes comes from the increased levels of atmospheric water vapor (a greenhouse gas) that are a result of the warm, ice-free ocean conditions. Despite the generally warmer climatic conditions, some areas show overall cooling. Notably, East Africa cools by 2 to 3° C due to increased low-level cloud cover which reflects large amounts of incoming solar radiation back to space. This result is consistent with the single palynological record that exists for that region.

Model-data moisture estimates show far less consistency than do the temperature estimates, not a surprising result given the complexities involved in modeling hydrologic processes using coarse-grid numerical models like the GISS GCM. The most common discordance seems to be an underestimation of the increased wetness suggested by pollen records at several localities throughout the Northern Hemisphere. The model's simple ground hydrology responds to the warmer summer ground temperatures by drying out while the diminished intensity of the atmospheric circulation (a result of reduced latitudinal temperature gradients) decreases the amount of moisture advected from over the oceans to the continents. In the Arctic, where modern tundra environments were replaced by Pliocene boreal forests, the altered boundary conditions required that wetter soil moisture conditions be specified.

The Pliocene Arctic soils remained wetter than the present day, indicating that the specified wet conditions were in equilibrium with the simulated climate.

In addition to the above experiment, several simulations were conducted using increased levels of atmospheric carbon dioxide; higher CO_2 amounts have also been proposed as a potential cause of the warmer Pliocene climates (Crowley, 1991). Rind and Chandler, (1991) pointed out that SST patterns such as the one seen in the Pliocene are inconsistent with CO_2 generated warming, however, it is possible that some combination of CO_2 increase and ocean heat transport change could have resulted in the warmer Pliocene surface temperatures. Figure 3 shows the various levels of ocean heat transport required to generate the PRISM SSTs given various atmospheric CO_2 increases. The graph indicates that with modern ocean heat transports (0% increase) CO_2 levels must have been at least 1400 ppm (4.5 times the modern value) in order to generate the global warming of the Pliocene. So far, estimates based on carbon isotope measurements by Raymo and Rau (1992) suggest that Pliocene CO_2 levels were, at most, 100 ppm greater than today.

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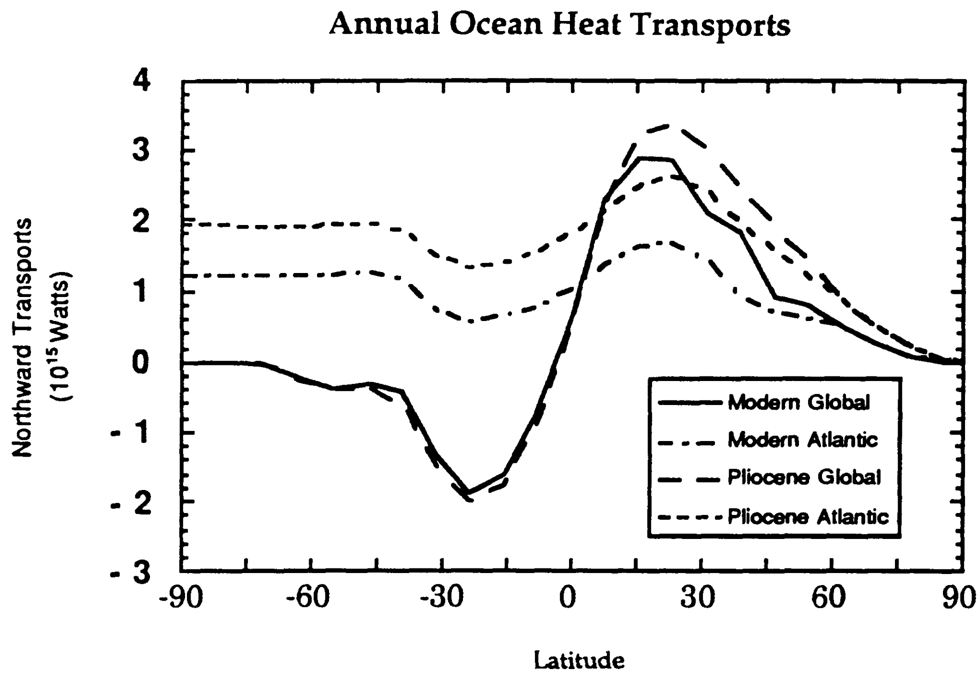

Figure 1. Pliocene and modern annual ocean heat transports for the Atlantic Ocean.

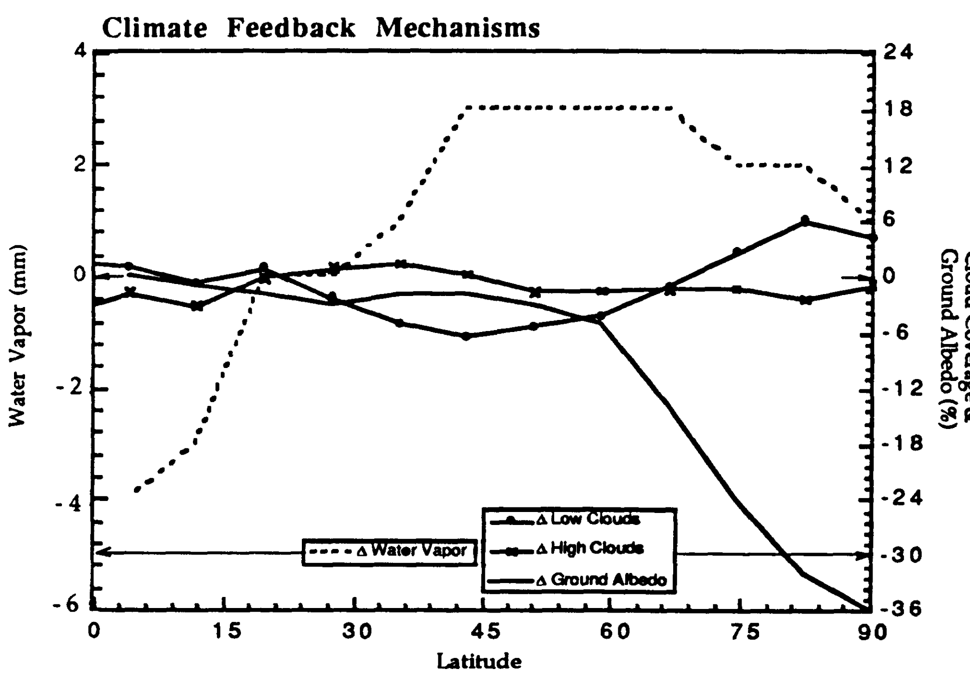

Figure 2. Water vapor, cloud coverage, and ground albedo in the northern hemisphere as a function of latitude in the GISS pliocene GCM simulation.

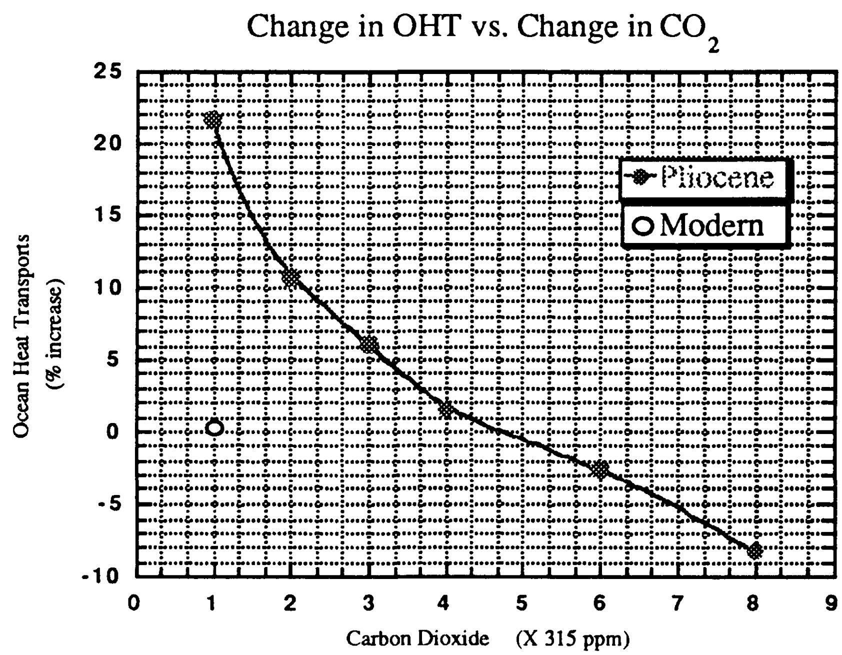

Figure 3. Levels of ocean heat transport required to generate PRISM sea-surface temperatures at varying levels of atmospyheric CO_2 concentrations.

Model-Model Comparisons And Data-Model Comparisons: Considerations For The PRISM Paleoclimate Study

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Because plans in the PRISM collaboration call for Pliocene climate simulations to be produced by two different climate models (GISS and NCAR), a comparison of the two models is necessary at the start. Both models are three-dimensional general circulation models (GCMs) of similar design, but there are important differences in specific climatic processes parameterized by the models, and in spatial resolution, among other factors. Therefore, before Pliocene results from the two GCMs are compared in this project, the models must be compared at a present day baseline climate. The strengths and weaknesses of each GCM should be documented, and differences in model performance under present day conditions between the models and observations must be considered before the global Pliocene results are examined. Some of the model characteristics have been documented by other model users, but we need to compare the model versions that will actually be used at each phase of the PRISM project. We should also keep in mind how the individual model parameterizations relate to factors thought to be important to Pliocene climate.

A hierarchy of comparisons of the model results should be carried out. First-order comparison will be of global mean annual values, at least for temperature, precipitation, top of the atmosphere energy elements, and general model sensitivity to forcing (*e.g.*, CO_2 doubling). Second-order comparisons of zonal averages will include comparison of model output fields of temperature (annual and seasonal), precipitation (annual and seasonal), sea ice, snow, zonal wind speed at near and mid troposphere levels, vertical velocity at midtroposphere level, and energy components. Third-order comparisons should include global (map view) distributions of minimum, mean annual, and annual amplitude of surface temperature, mean annual and seasonal precipitation, soil moisture, runoff, and total moisture (precipitation-evaporation). Statistical differences for these fields may prove useful. Other climate parameters that may be of specific relevance to Pliocene climate in general and the PRISM data base in particular should also be examined for these present day cases.

Comparisons between model results and data are an integral part of the long-term goals of PRISM. Comparisons at various stages of the project will serve as the means of evaluating both model results and the synoptic mapping efforts. These comparisons will also be used to plan for future interactions of climate experiments. There are several factors to consider when comparing model results to proxy data interpretations. First, model spatial and temporal resolutions will almost certainly not correspond to the resolutions reflected in the paleoclimate data. This must somehow be taken into account in comparisons. Second, the GCMs parameterize many processes instead of explicitly modeling them, and this may affect the results in relation to data interpretations of climatic factors. Third, careful consideration must be given to the question of what the proxy data have actually recorded; do pollen or macrofaunal assemblages most closely reflect mean annual temperature, growing season temperature, or some other quantity(ies)? This is critical when different types of proxy climate information are related to each other and are compared to climate model results. Last, multiple independent proxy indicators for a given climate parameter are useful to obtain because they yield multiple signals for a given system. This again is useful information for understanding the Pliocene climate system and its possible forcing factors.

Proxy climate data can be compared to model results on the basis of quantitative or qualitative information. Qualitative information includes wet/dry and hot/cold indicators which provide a first-order test of model results. Quantitative information includes minimum temperature, mean annual temperature, mean annual temperature range, mean annual precipitation, and seasonal precipitation, can be used as a second-order test of model results.

Use Of A High-Resolution Atmospheric Model For Simulations Of Paleoclimate

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To demonstrate the potential for using high resolution atmospheric models to reconstruct past climates, we here discuss briefly simulations of 18 ka BP perpetual January and July climate over the western U.S. that have been conducted with a high-resolution (mesoscale) atmospheric model (RegCM). RegCM, the climate version of the National Center for Atmospheric Research-Pennsylvania State University mesoscale model MM4, was run with a grid spacing of 60 km, a resolution that resolves regional atmospheric circulations and terrestrial features (*e.g.*, coast lines, mountains, lakes) that exert strong forcings on the climate of the West. The model includes a surface physics package (Biosphere Atmosphere Transfer Scheme) that couples the biosphere with the atmosphere, and a fully interactive lake model to simulate lake-atmosphere feedbacks associated with Lakes Bonneville and Lahontan.

A series of 90-day 18 ka BP and O ka BP simulations were conducted to allow estimates to be made of the general climate and of the effect of feedbacks between the lakes and the atmosphere. Initial and lateral boundary conditions (SSTs and vertical profiles of temperature, humidity, wind velocity) for the simulations were obtained from 90 days of output from the paleosimulations conducted by Kutzbach and Guetter (1986) with the NCAR Community Climate Model (CCM0). The surface of RegCM included estimated distributions of 18 ka BP vegetation, montane glaciers, the continental ice sheet, and the area and depth of Lakes Bonneville and Lahontan. SSTs were fixed at the values used by Kutzbach and Guetter.

Results indicate that over the west 18 ka BP January air temperatures were about 2^{o}K colder than the 0 ka BP control, whereas 18 ka BP July temperatures were >5^{o}K colder than the July control. January precipitation over the region was substantially greater at 18 ka BP; however, 18 ka BP July precipitation was not much different than 0 ka BP, except around Lakes Bonneville and Lahontan.

Analyses indicate that the hydrologic "signal" associated with the position and strength of the 18 ka BP jet stream dominate the hydrologic budgets of the Bonneville and Lahontan basins. Lake-atmosphere feedbacks isolated over the Bonneville basin indicate that lake-effect precipitation is a substantial component of the hydrologic budget in both January and July; however, such feedbacks are minor components of the hydrologic budget of Lake Lahontan. These results may help to explain the relative difference in the sizes of the lakes at 18 ka BP.

The modeling system used in this simulation of 18 ka BP climate can readily be applied to studies of pre-Pleistocene climates to assess, for example, the effects of uplift of the Rocky mountains and Colorado Plateau or the growth and influence of large paleolakes such as "Lake Idaho". Application of the model to the Pliocene depends on our ability to provide meaningful boundary conditions (e.g., sea level and SSTs, topography, inland water, atmospheric composition). Global modeling now being conducted at the Goddard Institute for Space Science and NCAR ultimately will result in a "consensus" Pliocene climate simulation that could be used to provide initial and lateral boundary atmospheric conditions for the regional model. Because estimates of Pliocene features such as topography and the extent of inland water are likely to remain uncertain, the appropriate use of the regional model would be to conduct a series of sensitivity tests by varying, for example, model topography to provide a range of output with which to compare with geologic evidence. Comparisons with the data would allow further refinements of the boundary conditions in the model, and thus allow some degree of convergence between the geologic data and simulated climate. In this way, it may be possible to establish plausible Pliocene climatic conditions at a regional scale.

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The Forward-Modeling Approach In Paleoclimatic Analysis: Middle-Pliocene Vegetation Distributions In North America

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Introduction

Paleoclimatic analysis can be defined as the joint examination of the *history* and *causes* of past climatic variations. As such, it is an activity distinct from the simple reconstruction or documentation of past climatic variations in the absence of a specific conceptual or mathematical model of the behavior of the climate system. Paleoclimatic analysis can therefore provide the information necessary for understanding how perturbations of the large-scale controls of the climate system--including those produced by humans--govern the regional and local responses.

There are two general approaches in paleoclimatic analysis: an inverse-modeling, or reconstruction approach, and a forward-modeling, or hypothesis-testing approach. The first approach, which might also be called a "bottom-up" approach, is the classical one in paleoclimatic analysis. In this approach, paleoclimatic evidence, in the form of geological or paleoecological records, is interpreted in climatic terms by one means or another. These reconstructions are then used to infer (usually subjectively) the nature of the changes in the controls of climate responsible for the observed patterns. The second, or "top-down" approach has been less frequently applied in paleoclimatic analysis. In this approach, ideas about the changes in the large-scale controls of climate are applied to one or more models (conceptual as well as numerical) of the climate system, and the resulting simulations are then compared with paleoclimatic observations. This approach is synonymous with "paleoclimatic data-model comparison," or "hypothesis-testing."

The properties of the forward-modeling approach that favor its application in paleoclimatic analysis (of Pliocene climates in particular) will be discussed in the next section. Following this, an example of the application of the forward-modeling approach applied to the middle-Pliocene distribution of several plant taxa in North America will be schematically illustrated.

Inverse- and forward-modeling approaches

The inverse-modeling approach. The "bottom-up" or inverse-modeling approach has been the usual one applied in paleoclimatic analysis. The approach is called "inverse-modeling" because conceptually, the flow of cause and effect (i.e., climatic variations cause responses in paleoclimatic indicators) is inverted in order to infer from the responses the nature of the causes. In practice, individual or groups of paleoclimatic records are interpreted in climatic terms, producing a set of climate reconstructions. These reconstructions can then be used by themselves (or in combination with other information) to, for example, make inferences about the proximate or ultimate controls of the climatic variations recorded by the data. The reconstructions may also be used to compare paleoclimates simulated by a model with the "observations" of paleoclimate at a particular time. For example, Bartlein *et al.* (1984) reconstructed the temporal and spatial patterns of July temperature and annual precipitation for the midwestern United States from fossil-pollen data, and from those reconstructions described a set of changes in atmospheric circulation that were consistent with the data. Other examples of the quantitative interpretation of fossil-pollen data in climatic terms are provided by Huntley and Prentice (1988), Prentice *et al.*

(1991), and Bartlein and Whitlock (1993). The comparison of paleoclimatic simulations with inferred past climates ("data-model" comparison) is illustrated for eastern North America over the past 18 kyr by Webb *et al.* (in press), and for North America during the Eocene by Sloan and Barron (1992).

There are a number of conditions and assumptions both statistical and ecological that must be satisfied when making inferences from paleoclimatic data; these have been discussed in detail by Sloan and Barron (1992), and Bartlein and Webb (1985). One major problem that arises in the reconstruction approach, even if the technical assumptions are satisfied, is the indeterminacy of the paleoclimatic data itself: the specific controls of a particular paleoclimatic record cannot always be uniquely or unambiguously determined from the data alone. In other words, even if the paleoclimatic reconstructions are correct, they alone may not reveal the controls of the past climates.

Indeterminacy of paleoclimatic records. Indeterminacy of paleoclimatic records arises from two sources: one intrinsic to the paleoclimatic indicators themselves, and one related to the nature of the climate system. The intrinsic source of indeterminacy arises because rarely will a particular biological or geological indicator of past climate reflect a single climatic variable. Vegetation, for example, is determined by a number of climatic variables, and interactions among these variables is typical. Vegetation is often limited by its moisture requirements as well as by temperature, and "effective moisture" reflects both precipitation and evapotranspiration. Evapotranspiration, in turn, is determined by a number of distinct climatic variables, as well as by the properties of the soil and vegetation. Consequently, it may be difficult to identify the particular individual or group of climatic variables that may have contributed to a particular feature of a paleoecological record. Derivations of the response functions (either empirical or theoretical) of the different paleoclimatic indicators can illustrate how a particular indicator depends on one or more controlling variable. Such information can be used to indicate when an observed response of a particular indicator may not be unambiguously interpretable in terms of a single climatic variable.

The climate system itself also contributes to indeterminacy. Two examples illustrate the problem (Fig. 1). In the western United States at 18 ka and at 9 ka (as well as 6 ka) similar patterns of reconstructed effective moisture anomalies (relative to present) exist (Thompson *et al.*, in press): drier than present in the Pacific Northwest and wetter than present in the Southwest (see also COHMAP Members, 1988). The likely controls of these similar patterns can be inferred from paleoclimatic simulations (e.g. Kutzbach and Wright, 1985), and differ considerably from one another (see Thompson *et al.*, in press). At 18 ka, the large Laurentide ice sheet split the jet stream in the simulations (particularly in winter), with the southern branch of the jet crossing the west coast south of where it does at present. At the surface in the simulations, a glacial anticyclone developed. In the Pacific Northwest, these circulation changes combined to weaken the westerly (from the west) winds at the surface, and consequently to reduce precipitation. In the Southwest, the southward-displaced jet stream created more onshore flow, and consequently greater precipitation.

In the simulations for 9 ka, greater summer insolation than present heated the center of the continent and increased the land/ocean temperature contrast in summer. This thermal anomaly in turn affected circulation, strengthening the east Pacific subtropical high pressure system in summer, and producing a "heat low" over the continent (Kutzbach, 1987). The stronger subtropical high, and the greater regional-scale subsidence associated with it, consequently suppressed precipitation in the Pacific Northwest, while the greater onshore flow (than present) into the heat low increased summer precipitation.

The same regional paleoclimatic anomaly pattern--drier than present in the Pacific Northwest and wetter than present in the Southwest--apparently can result from two different sets of ultimate controls, and the reconstructed pattern of effective moisture alone cannot discriminate between the two. This situation exists despite the likelihood that the precipitation anomaly was best expressed in winter in one case and in summer in the other. Further paleoclimatic information, such as the status of the levels of former lakes (Guiot *et al.*, 1993), or the variation in abundance of plant taxa with specific seasonal climatic requirements (Thompson, 1988) might be used to differentiate between the controls, but the effective moisture pattern alone cannot provide an unambiguous identification of the ultimate controls in each case.

A second example illustrates a case in which the ultimate control of a response observable in a particular paleoclimatic record is clear, but there are more than one proximate controls, and it is not apparent which or all are important. In this example, the specific paleoclimatic observation is the replacement of steppe vegetation by conifers in the basins in the interior western United States during the Pliocene, a vegetation change unambiguously interpretable as indicating a change toward drier conditions (Thompson, 1991). The ultimate control of this trend toward dryness clearly is the uplift of western North America. But there are at least two different pathways by which this large-scale control can produce drier conditions within the region, however, and there are intra-regional (positive) feedbacks that can reinforce the trend as well, adding a third pathway. At the continental scale, the uplift of western North America over the past several million years gradually strengthened the ridge in the upper-level westerly winds that now prevails over western North America (Ruddiman and Kutzbach, 1989; Raymo and Ruddiman, 1992). This circulation change in turn would have increased subsidence over western North America, and consequently suppressed precipitation. At the subcontinental scale, the uplift or formation of individual ranges such as the Sierra Nevada and Cascade Range would have increased the regional-scale rainshadows, by increasing orographic precipitation on the west side of the ranges and increasing subsidence on the east side. This mechanism would also reduce precipitation. Within the region, decreased precipitation would likely have reduced the size of the very large (relative to present) Pliocene lakes, further reinforcing the drying trend by reducing lake-effect precipitation. In this second example, although the ultimate control of the change recorded by various paleoclimatic indicators is clear, the manner in which the regional response is actually determined is ambiguous. Both examples show how the bottom-up, inverse-modeling approach may be inadequate when the objective is to discover how changes in the large-scale controls of climate govern regional climatic responses.

The forward-modeling approach. The "top-down" or forward-modeling approach in paleoclimatic analysis begins with some information or assumptions regarding the state of the large-scale controls of climate (i.e. boundary conditions") and then applies these to a climate model. The model in turn makes projections of the potential response to this particular configuration of the controls. The projections made by the climate model can then be passed to an environmental "sub-model" (e.g. a hydrological (Hostetler and Giorgi, 1993), or a vegetational (Prentice *et al.*, 1992a) model). The output of these succeeding models is then compared with the available paleoclimatic indicators.

The climate model, as well as the environmental sub-models, do not need to be numerical models; conceptual models may also be used (e.g. Bartlein *et al.*, 1991), although these are generally viewed as inferior to numerical models. One objective of the forward-modeling approach is to expose inadequacies in any model. This information could be then used, for example, to advocate replacement of a conceptual model with a numerical one, or to revise elements of the model or the assumed state of the controls.

In an example of this approach, Webb *et al.* (1987) used output from the paleoclimatic simulations produced by Kutzbach and Guetter (1986) to simulate the changing distribution of vegetation in eastern North America over the past 18,000 years, as represented by fossil-pollen data. The paleoclimatic simulations were based on the CLIMAP Project Members (1981) reconstruction of sea surface temperatures (SSTs) and ice-sheet size for the last glacial maximum (LGM). Although there was broad-scale agreement between the simulations and observations, for some locations and time intervals (i.e. the southeastern United States during the interval between 18 ka and 12 ka) there was a serious mismatch between the two. There are three potential sources of the mismatch: 1) misinterpretation of the fossil data, 2) model inadequacy, or 3) the specification of inappropriate boundary conditions. In this case, the discrepancy seems assignable to the third source, in specific through: a) the adoption of the CLIMAP SSTs for the ocean adjacent to North America--these show little difference from present, particularly in the Gulf of Mexico; and b) the inclusion of a very large ice sheet in the model, which resulted in the simulation of relatively high temperatures in the southeastern U.S. produced by adiabatic heating of air that descended from the broad dome of the ice sheet. Consequently the paleoclimatic simulations in the southeastern U.S. also differed little from present, leading to the particular mismatch that was observed. In this example, the comparison of simulated paleoecological responses with the observed provides information that can be used to enhance our understanding of the how the patterns recorded in the paleoclimatic data were generated.

Forward-modeling can also be iteratively applied in order to resolve apparent inconsistencies among different sources of paleoclimatic data. For example, Prentice *et al.* (1992b) were able to resolve the apparent inconsistency between the observations in the Mediterranean region at the LGM of steppe vegetation (signaling dry conditions), and high lake levels (signaling wet conditions). They combined a simple numerical model of the regional-scale water balance with a conceptual model of atmospheric circulation changes (based in part on numerical GCM simulations, Harrison *et al.*, 1992). By perturbing the present climate of the region in a set of sensitivity tests with a water-balance model, they were able to find a plausible configuration of precipitation, evapotranspiration and runoff that was consistent with both the paleoclimatic data and simulations.

The indeterminacy problem is greatly reduced in the forward-modeling approach. By requiring the use of models (that usually must be sufficient to explain the present climate and in turn its influence on geological and biological indicators), the approach implicitly recognizes the multivariate nature of the controls of the different indicators upon which paleoclimatic reconstructions are based. For example, trade-offs between such large-scale controls of regional climates as atmospheric circulation and insolation might be evaluated using regional- or mesoscale-climate models. Similarly, the application of a hydrological model might reveal the relative importance of increased precipitation as opposed to decreased evapotranspiration in producing an inferred increase in effective moisture.

Another advantage of the forward-modeling approach is its potential for allowing comparisons between model simulations and paleoclimatic evidence in regions of sparse data coverage. Kutzbach and Guetter (1980) examined the role that the distribution of sites that register the response to large-scale controls has in reconstructing those controls. In their analysis, they examined the degree to which (controlling) sea-level pressure patterns could be estimated using a network of sites with observations of temperature and precipitation responses. They found that the level of explained variance of the pressure patterns was dependent on the uniformity, density and extent of spatial coverage of the temperature and precipitation sites. At "low" site densities (below those of late-Quaternary fossil-pollen sites in eastern North America or western Europe, and roughly equal to the density of late-Quaternary sites from western North America), less than half of the variance of January sea-level pressure patterns could be explained. These results imply that where the density of sites with paleoclimatic evidence is low, it may not be feasible to recover by inverse modeling the nature of the proximate controls of the climatic variations recorded at those sites.

Forward-modeling, in contrast, permits the simulation of paleoenvironmental responses at a resolution determined only by the practical limits of application of the associated environmental sub-models. In other words, the forward-modeling approach could be used to generate, for example, a very high-resolution spatial pattern of vegetation types across a region, regardless of how dense the network of fossil-pollen sites in that region is. Point-by-point comparisons between the simulations and observations could be made where fossil data is available. A more telling comparison would likely result from a subjective evaluation (in light of the observed fossil-pollen data) of the predicted vegetation pattern.

The forward-modeling approach has the advantages of reducing the indeterminacy problem, and allowing paleoclimatic hypotheses to be tested even in regions of sparse coverage of paleoclimatic evidence. The inverse-modeling approach, although based on "observed" or "real," as opposed to simulated data, suffers from indeterminacy, and problems of applicability in regions of sparse coverage of fossil data. In summary, the forward-modeling approach can be applied over the entire domain of interest (e.g. North America), generating predictions of, for example, the past distributions of different taxa that can be tested wherever data does exit. The inverse-modeling approach can be usefully applied in those regions where data is sufficiently dense, multiple paleoclimatic are available, and the assumptions that underlie the techniques used are not violated. In such regions, the approach can provide quantitative reconstructions of past climates with relatively low uncertainties.

Middle-Pliocene Vegetation Distributions in North America

One issue that may be addressed using the forward-modeling approach is the comparison of different climate model simulations, in order to identify which model or particular simulation among a group is "best." Such an analysis could be used to discriminate among a number of different models, in order to evaluate their ability to correctly simulate particular patterns in the data, or to determine the appropriate set of boundary conditions for a particular model that will produce a simulation consistent with observations. This second application is illustrated here through simulations of the distributions of three plant taxa from North America.

Dowsett *et al.* (n.d.) have compiled a data set of boundary conditions for the middle Pliocene that includes paleogeography, sea-surface temperatures, sea-ice and land-ice extents, and surface cover. These boundary conditions were then used by Chandler *et al.* (n.d.) as input to the GISS (Goddard Institute of Space Science) general circulation model (GCM), to produce a simulation of middle-Pliocene climates. In the simulation of Chandler *et al.*, the concentration of carbon dioxide in the atmosphere was kept at modern values, because they were not certain that CO_2 concentrations during the Pliocene were greater than present. This "Pliocene" simulation can therefore be considered to represent a "non-elevated-CO_2" portrayal of middle-Pliocene climates. Another simulation with the GISS model, performed by Hansen *et al.* (1984), can be used to illustrate the response of the simulated climate to a doubling of carbon dioxide alone (i.e. without any other changes in boundary conditions). This "2xCO_2" simulation can therefore be used to represent a Pliocene climate that differs from the present one only because of variations in the composition of the atmosphere. The extent to which either simulation can correctly reproduce the observed difference in vegetation between the middle Pliocene and present can then be used to differentiate between the two controls--altered boundary conditions as opposed to elevated CO_2--in explaining the nature of the middle Pliocene climate. This example, although rather forced, does illustrate the general approach of using the forward-modeling approach to analyze the causes of past climates.

Vegetation-Climate Relationships And The Simulation Of Vegetation Distributions

In the example presented here, the distributions of three plant taxa were simulated under three different climatic "scenarios." These scenarios were 1) the observed modern climate, 2) the simulated "Pliocene" climate, and 3) the simulated "2xCO_2" climate. Three taxa were selected: *Picea mariana* (black spruce), representing the boreal forest, *Quercus alba* (white oak), an important constituent of the midlatitude deciduous forest, and *Artemisia tridentata* (sagebrush), a steppe taxon that at present is abundant in the interior of western North America. The ranges of these three taxa were digitized from the range maps of Little (1971) and overlaid on a 25-km equal-area grid covering North America. Climate data (January and July temperature and precipitation) were interpolated onto the same grid using a procedure that incorporated the effects of elevation on climate.

Relationships between the incidence of different plant taxa and climate are summarized here using response surfaces (Bartlein *et al.*, 1986; Prentice and Solomon, 1991). Response surfaces for presence-absence data (as obtained from range maps) illustrate the probability of occurrence of a particular taxon at different locations in the climate space defined by the four climatic variables. Response surfaces represent a type of equilibrium (as opposed to dynamic) vegetation model (Prentice *et al.*, 1993; Prentice and Solomon, 1991). Such models are appropriate for simulating the broad-scale distributions of plants under slowly changing climates.

The response surfaces are fitted here using a locally weighted-averaging approach (Prentice and Solomon, 1991). The environmental preferences of the different taxa can be illustrated by plotting the predicted probabilities of occurrence of the taxa on several "slices" through the climate space (Figs. 2-4). *Artemisia*, for example, reaches its highest probabilities of occurrence (indicated by shading on the figures) at locations with relatively low July precipitation, medium July temperatures, and low January temperatures--the general conditions that prevail in the lowland areas of the intermountain region.

The response surfaces may be used to simulate the distributions of the different taxa under a particular climate by "plugging in" the values of the climate variables for each of the 25-km grid cells and then mapping the estimated probabilities of occurrence. In the examples here, the "Pliocene" and "2xCO_2" climates that were input to the response surfaces consisted of the simulated anomalies (paleoclimatic experiment minus control) applied to the observed modern values on the 25-km grid.

Results

The digitized range maps of each of the three taxa are shown in the upper-left panels of Figs. 5-7, and the simulated distributions under the present climate are shown in the upper-right panels. These sets of maps indicate that the response surfaces are able to reproduce the present distributions of the taxa adequately. The simulated distributions of the three taxa under the "Pliocene" and "2xCO_2" climates are shown in the lower left and lower right panels for Figs. 5-7, respectively.

The reconstructed vegetation patterns for the Pliocene (i.e. those inferred from the paleoecological evidence) have the following features (Dowsett *et al.*, n.d.): Evergreen forests (represented here by *Picea*) advanced northward to the Arctic coast, while maintaining their modern dominance in subarctic regions. In parts of the western interior, *Picea* was common, but not dominant. Deciduous forests (represented here by *Quercus*) were largely at their present range limits during the Pliocene, but may have extended farther north than today. Steppe (represented here by *Artemisia*) was present across the interior of western North America during the Pliocene, but was possibly less abundant than at present.

Which of two climate simulations best reproduces these patterns? The simulated distributions of *Picea* for the "Pliocene" and "2xCO_2" simulations are quite similar to one another. The distributions of *Picea* under both simulations advance at their northern edges, and there is a slight increase (although probably not significant) in *Picea* in the Northern Rocky Mountain. region in the "Pliocene" simulation. Both simulations seem consistent with the reconstructions (Dowsett *et al.*, n.d.), and there is little to distinguish between them.

The simulated patterns of *Quercus* in eastern North America are quite similar under the two simulations as well. Both show a northward shift in the range of the *Quercus*, with the northward retreat of the southern range margin slightly greater under the 2xCO_2 simulation. West of the Mississippi valley, however, the simulations are markedly different, with the "2xCO_2" simulation showing a greater northward advance relative to present on the northern Great Plains, and the "Pliocene" simulation showing a greater incidence on the southern Great Plains and in the interior basins of the intermountain west, where steppe vegetation prevails at present. In eastern North America, as was the case with *Picea*, both simulations seem consistent with the reconstructions. In western North America, however, the high probabilities of *Quercus* under the "Pliocene" simulation seem inconsistent with the reconstructions.

The simulated patterns of *Artemisia* differ more overall between the two simulations than do the other taxa. Both patterns do show a decrease in the incidence of *Artemisia*, consistent with the reconstructions of less steppe and grassland during the Pliocene. Under the "2xCO_2" simulation, the range of Artemisia is approximately the same as at present, while under the "Pliocene" simulation, the range and incidence maximum shift to the east.

Of the two simulations, the "2xCO_2" simulation produces distributions of these tree taxa that seem more consistent with the vegetation reconstructions than does the "Pliocene" simulation. It should be pointed out that neither simulation (one with Pliocene surface boundary conditions, but no increase in CO_2, and the other with increased CO_2, but without Pliocene surface boundary conditions) should be considered to be a "full" simulation of Pliocene climates. It is also true that only three taxa have been examined, and simulations of other taxa could change the picture. In any case, this example shows the utility of the forward-modeling approach in comparing two climate-model simulations.

Summary

In the forward-modeling approach to paleoclimatic analysis, specific hypotheses about the large-scale controls of climate or the climatic states that they produce are proposed, and used to simulate estimate the responses of different paleoenvironmental indicators, such as vegetation to those climates. These simulated responses are then compared to the paleoenvironmental record to test those hypotheses. Because this approach requires the specification of an explicit model that links the climatic controls to the environmental responses, the indeterminacy of the paleoclimatic indicators is reduced.

In addition to testing specific paleoclimatic hypotheses, the forward-modeling approach can be used to discriminate among different climate models, or among different sets of boundary conditions for those models. In such an application, the sets of paleoenvironmental responses produced by the approach would be compared to syntheses of paleoclimatic data in order to identify which simulation was most consistent with the data. This type of application was illustrated here through an example of the simulation of three plant taxa by two different simulations of Pliocene climates for North America.

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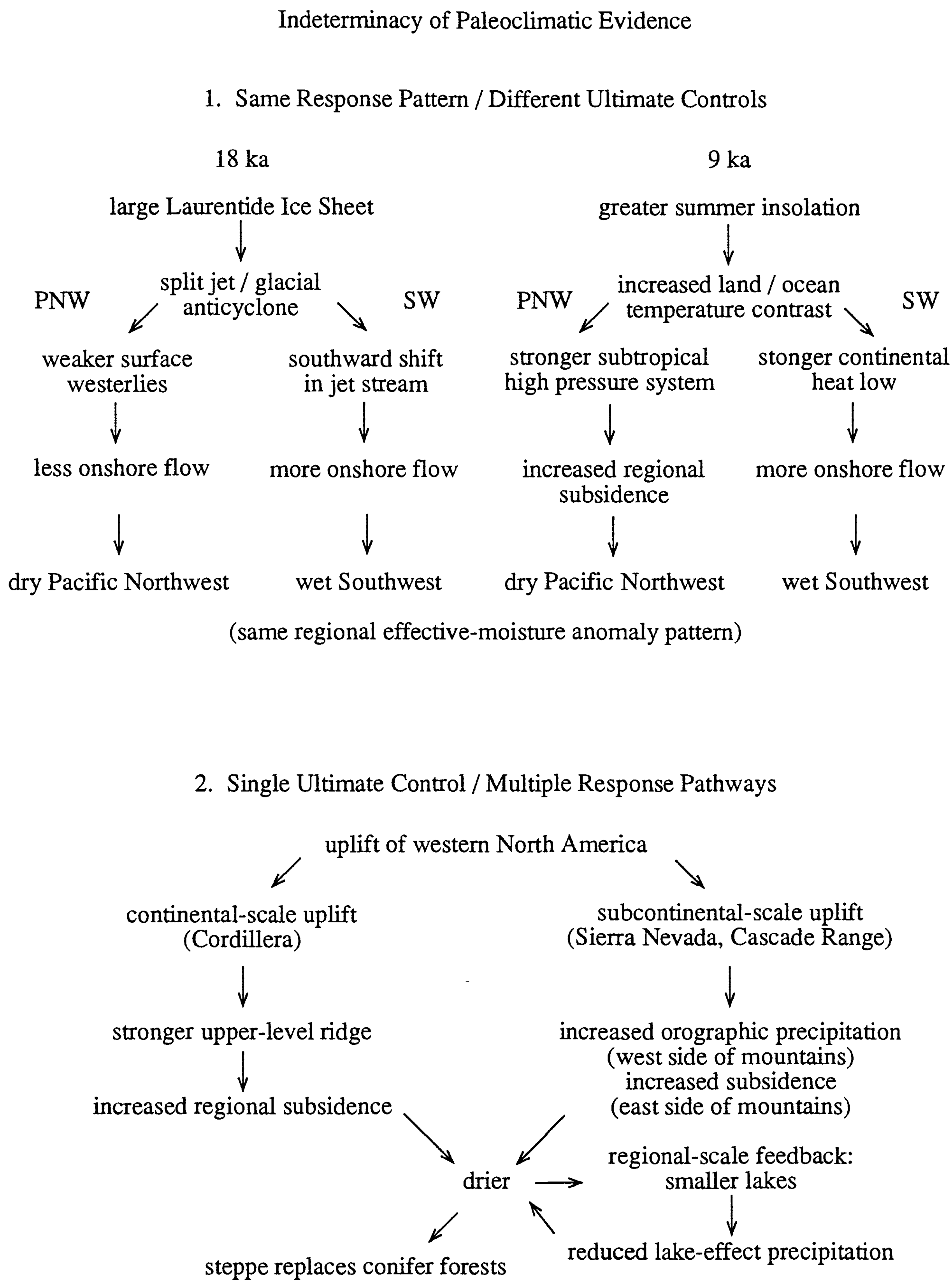

Figure 1. Examples of indeterminacy of paleoclimatic evidence. Example 1 shows how the similar regional patterns of effective moisture (dry in the Pacific Northwest, wet in the southwest) can result from two different large-scale controls. See Thompson *et al.* (in press) for further discussion. Example 2 shows how the multiple pathways by which a single large-scale control (uplift of western North America) can produce a signal in a regional paleoclimatic indicator (i.e. steppe replacing forest in the interior western United States; see Thompson (1991) for further discussion).

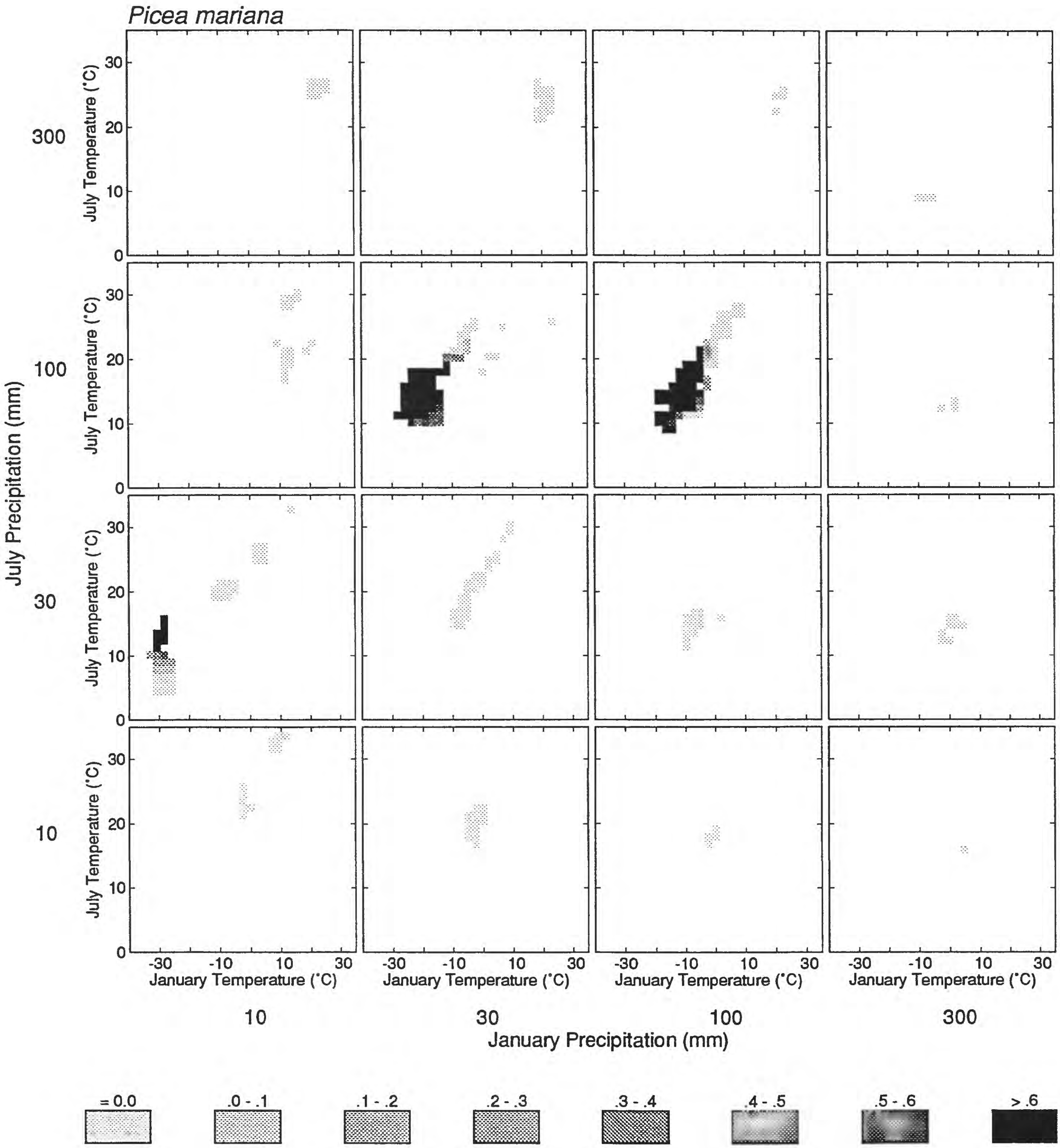

Figure 2. Response surface for *Picea mariana* (black spruce). The diagram shows sixteen two-dimensional slices through a four-dimensional climate space. Each slice shows the probability of observing *P. mariana* as a function of January and July temperatures for the specific values of January and July precipitation given in the margin. The probabilities are shown by shading, with the lightest shade indicating zero probability. Unshaded areas represent combinations of the four variables that do not occur over North America.

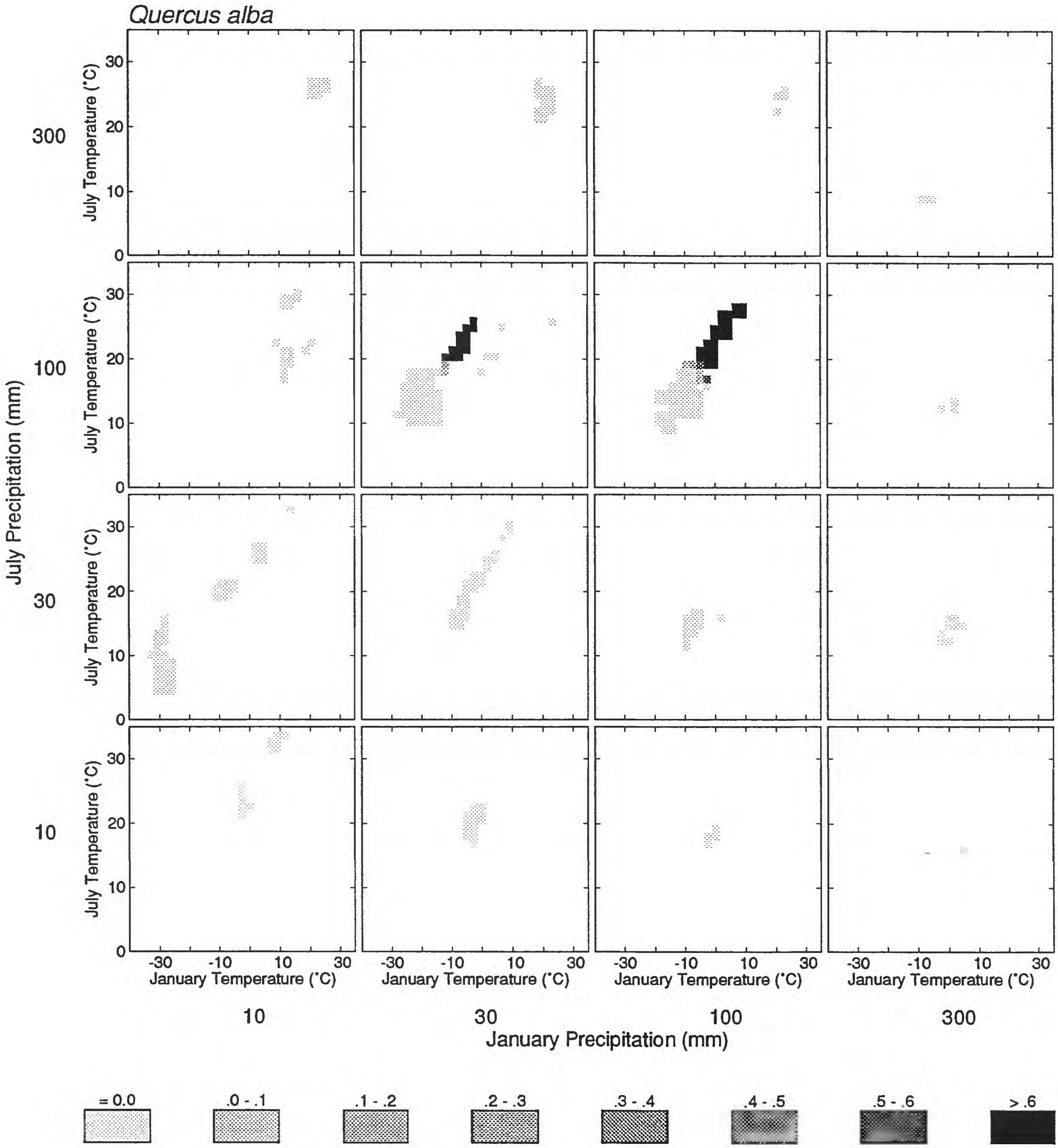

Figure 3. As in figure 2, except for *Quercus alba* (white oak).

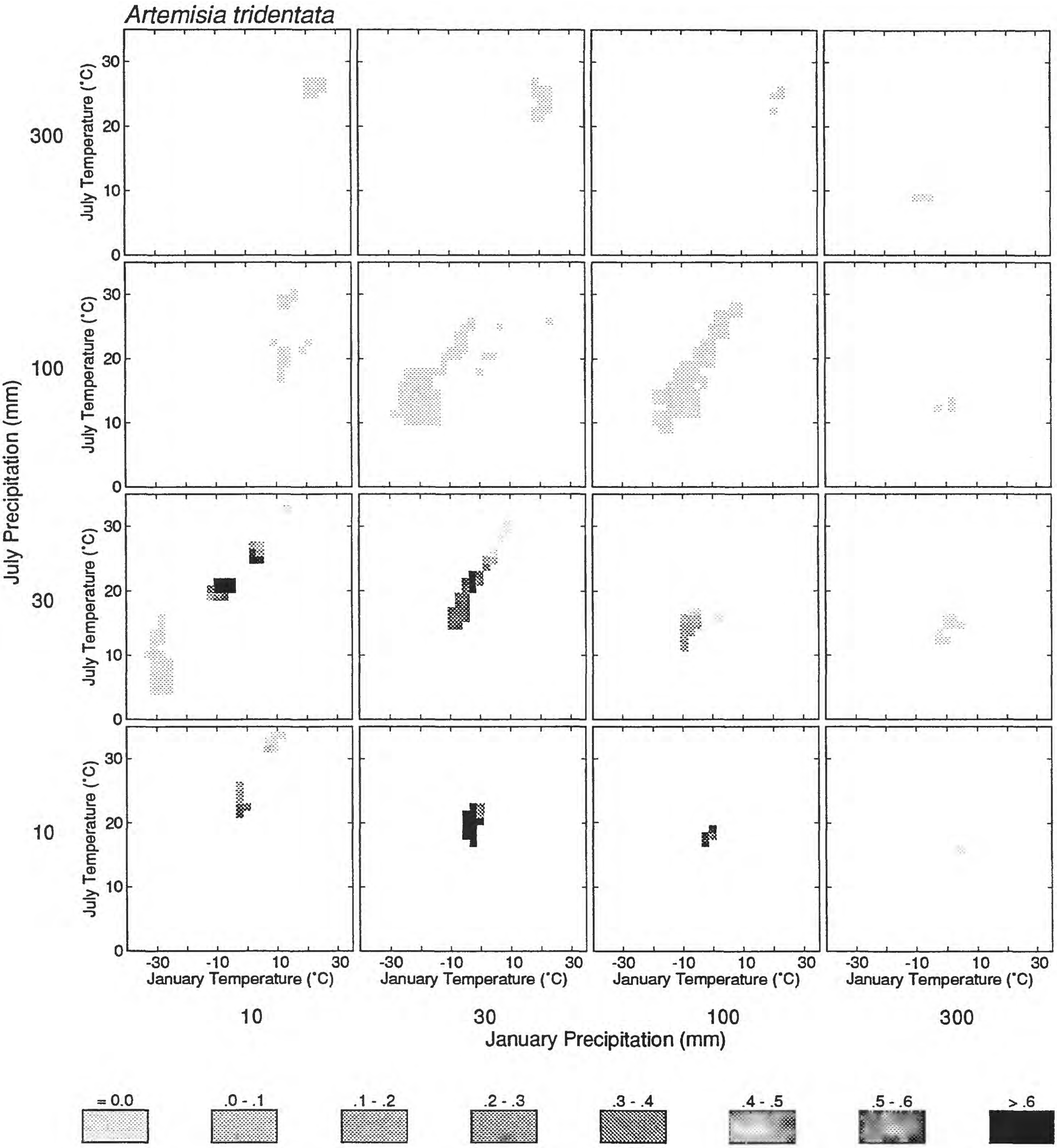

Figure 4. As in figure 2, except for *Artemisia tridentata* (sagebrush).

Picea mariana

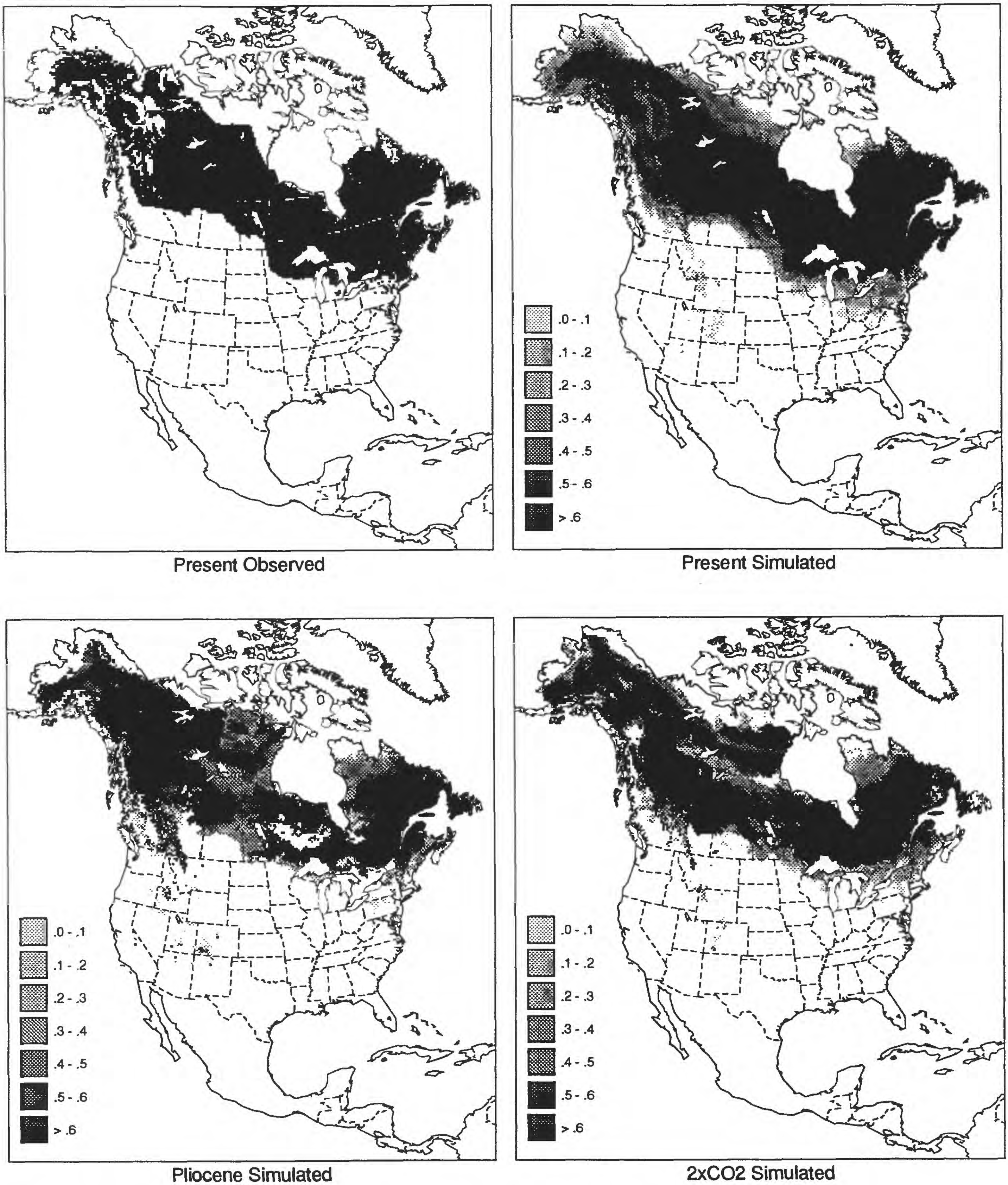

Figure 5. Distribution maps for *Picea mariana* (black spruce). The upper left-hand map shows the present distribution of *P. mariana* as digitized onto a 25-km grid from the range maps in Little (1971). The upper right-hand map shows the present distribution of *P. mariana* as simulated by plugging in the observed values of the four climate variables that define the response surface for this taxon. The probability of occurrence of this taxon at each grid point is shown by shading. The lower left-hand map shows the probability of occurrence of *P. mariana* as simulated by plugging in the climate values generated by the "Pliocene" simulation, and the lower right-hand map shows the probability of occurrence of *P. mariana* as simulated by plugging in the climate values generated by the "2xCO_2" simulation. See text for details.

Quercus alba

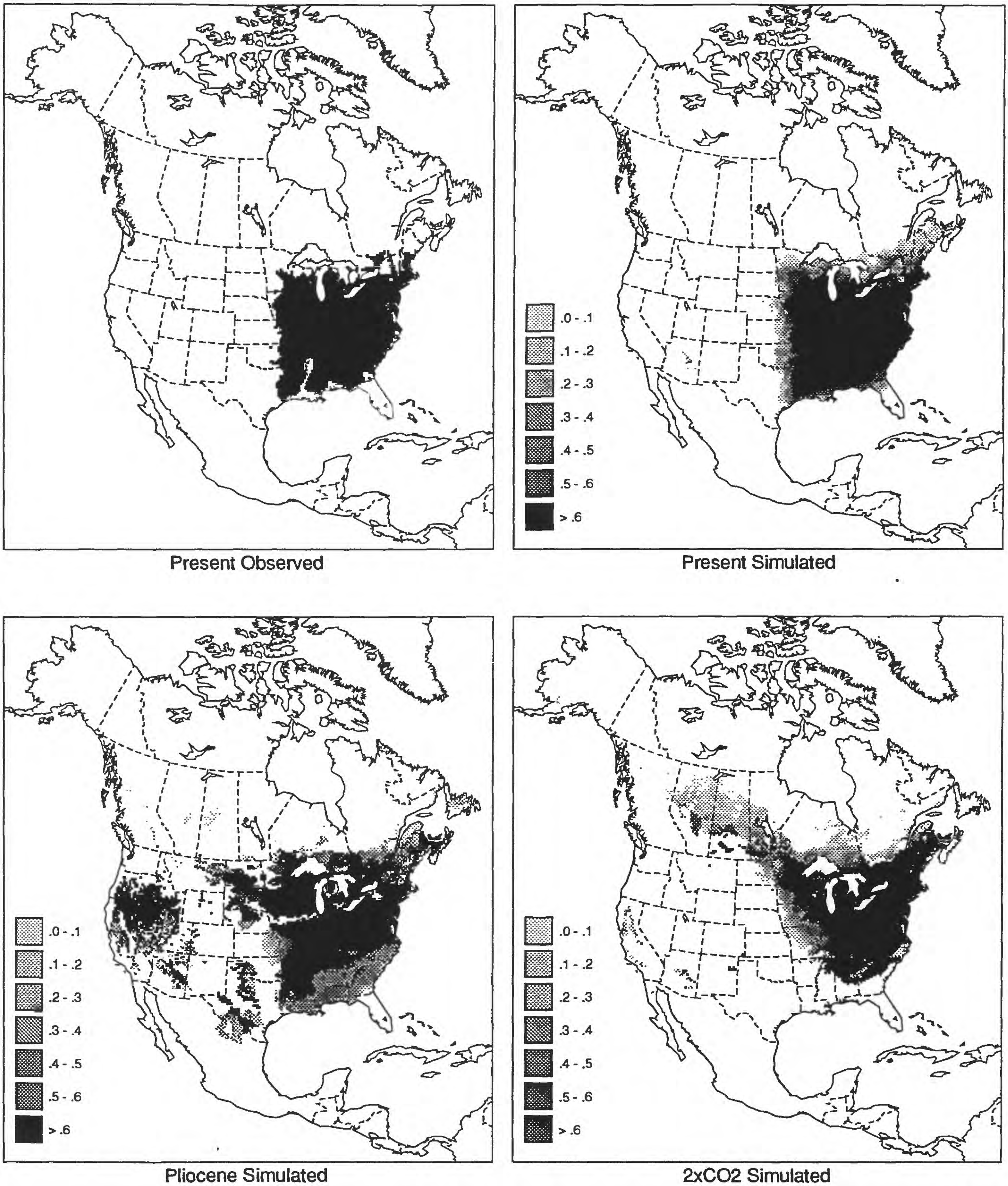

Figure 6. As in figure 5, except for *Quercus alba* (white oak).

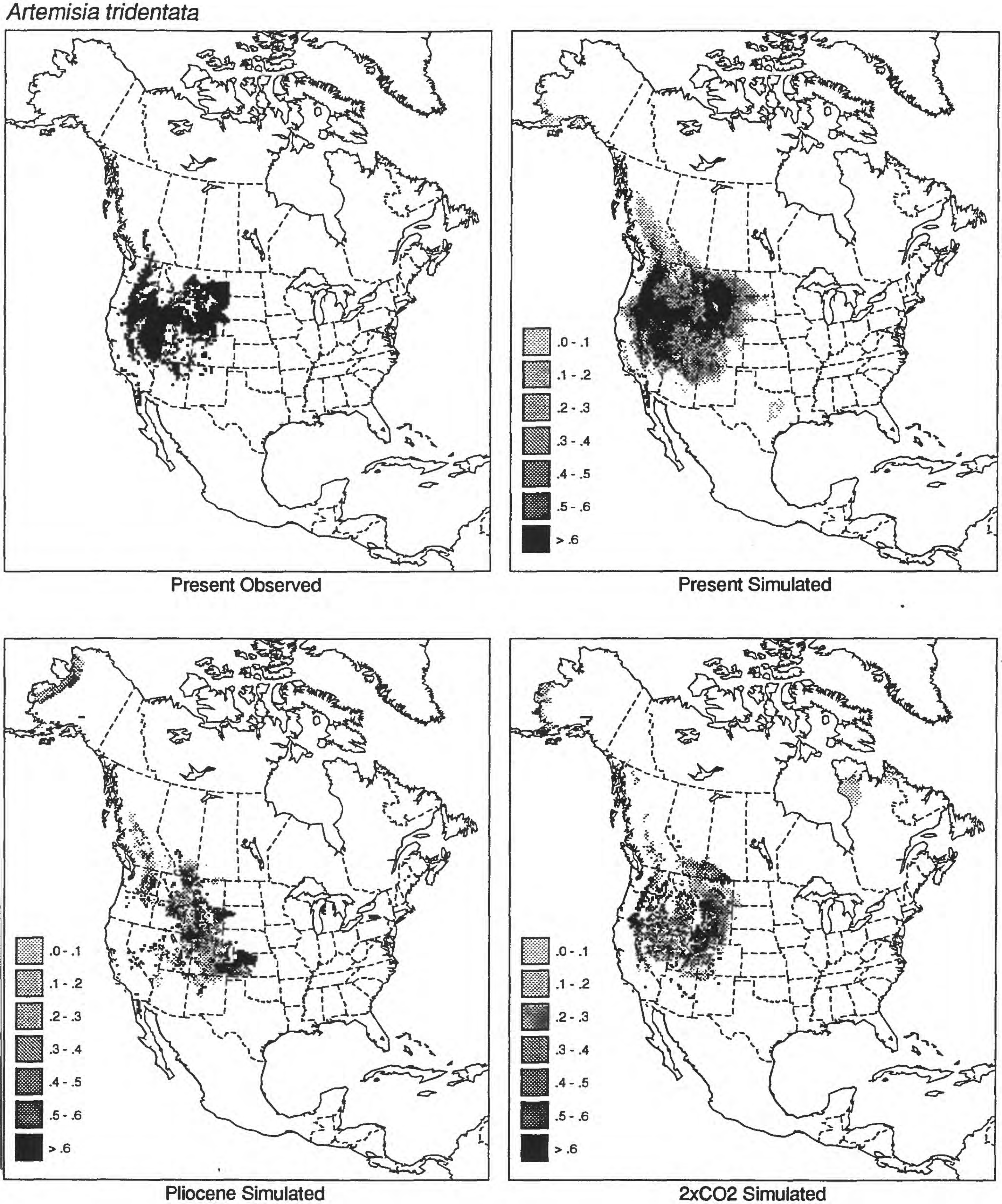

Figure 7. As in figure 5, except for *Artemisia tridentata* (sagebrush).

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