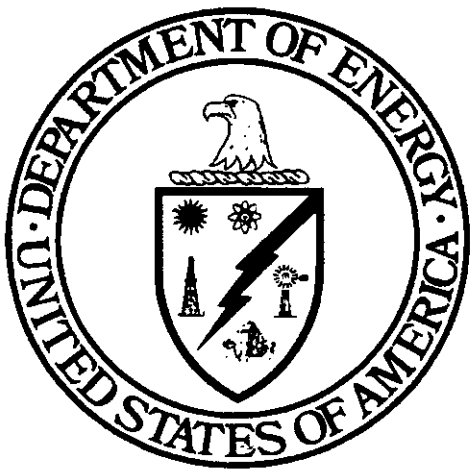
Title 40 CFR Part 191
Compliance Certification
Application
for the
Waste Isolation Pilot Plant

Appendix CLI



United States Department of Energy
Waste Isolation Pilot Plant

Carlsbad Area Office
Carlsbad, New Mexico

Long-Term Climate Variability at the Waste Isolation Pilot Plant

CONTRACTOR REPORT

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Long-Term Climate Variability at the Waste Isolation Pilot Plant, Southeastern New Mexico, USA

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Prepared by Sandia National Laboratories Albuquerque, New Mexico 87185 and Livermore, California 94550 for the United States Department of Energy under Contract DE-AC04-76DP00789

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Long-Term Climate Variability at the Waste Isolation Pilot Plant, Southeastern New Mexico, USA*

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ABSTRACT

The United States Department of Energy is developing the Waste Isolation Pilot Plant (WIPP) in southeastern New Mexico for the disposal of transuranic wastes generated by defense programs. Because changes in climate during the next 10,000 years (10 ka) may affect performance of the repository, an understanding of long-term climate variability is essential for evaluating regulatory compliance.

Fluctuations in global climate corresponding to glaciation and deglaciation of the northern hemisphere have been regular in both frequency and amplitude for at least 780 ka. Coolest and wettest conditions in the past have occurred at the WIPP during glacial maxima, when the North American ice sheet reached its southern limit roughly 1200 km north of the WIPP and deflected the jet stream southward. Average precipitation in southeastern New Mexico during the last glacial maximum 22 to 18 ka before present (BP) was approximately twice that of the present. Driest conditions (precipitation approximately 90% of present) occurred 6.5 to 4.5 ka BP, after the ice sheet had retreated to its present location. Wet periods of unknown duration have occurred since the retreat of the ice sheet, but none have exceeded glacial conditions. Global climate models suggest that anthropogenic climate changes (i.e., warming caused by an increased greenhouse effect) will not result in an increase in precipitation at the WIPP. The climate of the last glacial maximum is therefore suitable for use as a cooler and wetter limit for variability during the next 10 ka.

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Figures

1. Introduction

The Waste Isolation Pilot Plant (WIPP), located 42 km east of Carlsbad, New Mexico (Figure 1), is being evaluated by the United States Department of Energy (DOE) for disposal of transuranic wastes generated since 1970 by defense programs. The repository is excavated approximately 655 m below the ground surface in bedded halite of the Late Permian Salado Formation, deposited approximately 255 million years before the present (255 Ma BP).

Before the WIPP can be used for long-term disposal of transuranic waste, the DOE must demonstrate compliance with the United States Environmental Protection Agency's (EPA) *Environmental Standards for the Management and Disposal of Spent Nuclear Fuel, High-Level and Transuranic Waste* (40 CFR 191) (U.S. EPA, 1985), hereafter referred to as the standard. Although the standard was vacated by a Federal Court of Appeals in 1987 and is undergoing revision, by agreement with the State of New Mexico, the DOE is continuing to evaluate repository performance with respect to the regulation as first promulgated until a new version is available (U.S. DOE and the State of New Mexico, 1981, as modified in 1984 and 1987).

The standard requires that the DOE consider "all significant processes and events that may affect the disposal system" during the 10,000 years (10 ka) following decommissioning. The performance assessment being conducted for the DOE by Sandia National Laboratories is therefore examining, among other things, the likelihood and consequences of long-term changes in climate. Climatic changes have the potential to affect repository performance directly, by altering groundwater recharge and flow in the region, and indirectly, by changing human land-use patterns in the region. Increases in precipitation are of primary concern because they may result in increased groundwater flow and, in the event of a breach of the repository, increased transport of radionuclides to the accessible environment.

2. Modern Climate at the WIPP

At present, the climate at the WIPP is arid to semiarid. Mean annual precipitation at the WIPP has been estimated to be between 28 and 34 cm/yr (Hunter, 1985). At Carlsbad, 100 m lower than the WIPP, 53-yr (1931-1983) annual means for precipitation and temperature are 32 cm/yr and 17.1°C (University of New Mexico, 1989). Short-term variation about the annual means can be considerable, and historic weather data cannot be used to predict long-term climatic shifts. For example, the 105-yr (1878 to 1982) precipitation record from Roswell, 135 km northwest of the WIPP and 60 m

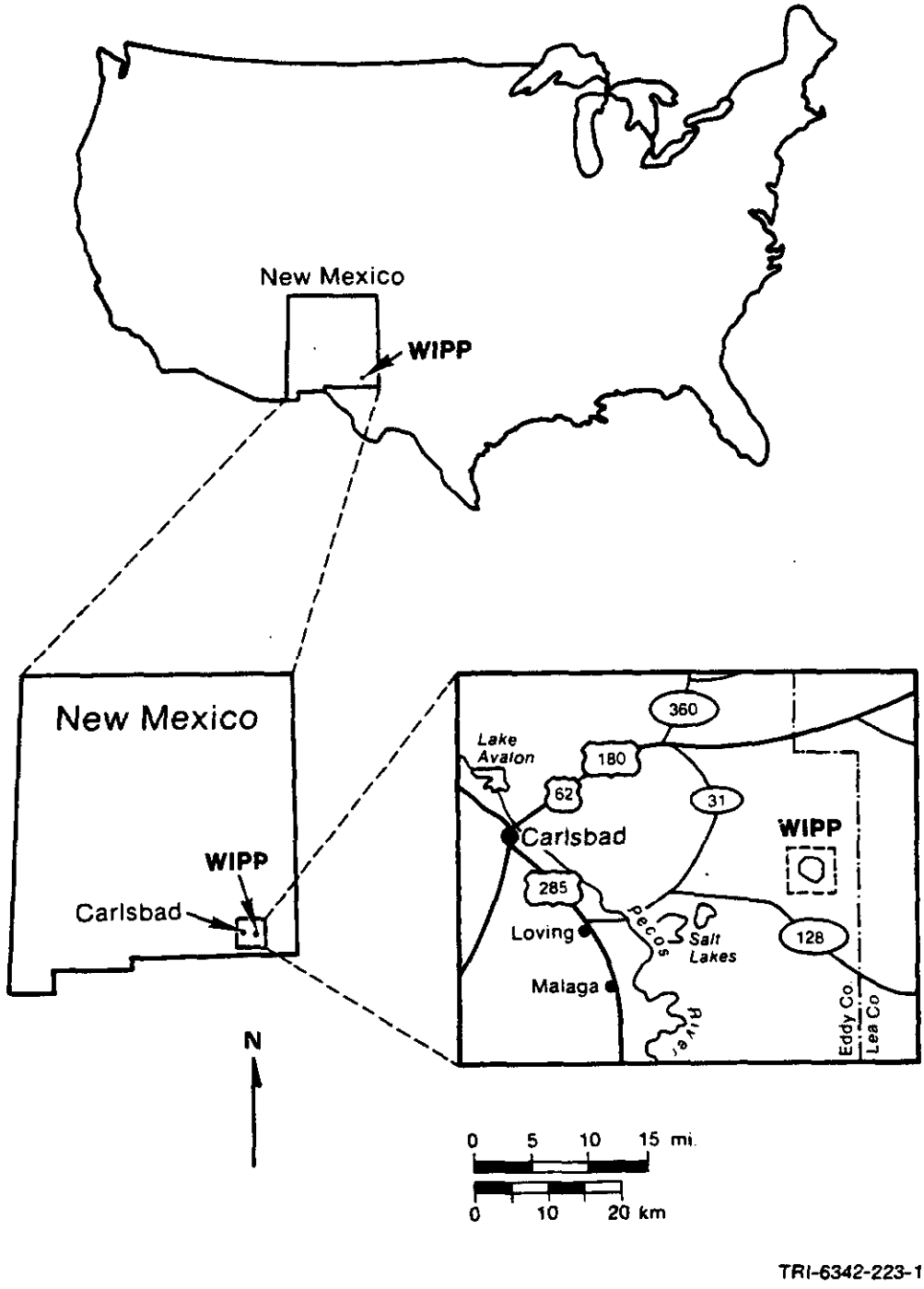

Figure 1. Location of the WIPP (after Bertram-Howery and Hunter, 1989).

higher, shows an annual mean of 27 cm/yr with a high of 84 cm/yr and a low of 11 cm/yr (Hunter, 1985).

The climate of southeastern New Mexico is monsoonal: most of the precipitation falls in late summer, when solar warming of the continent creates an atmospheric pressure gradient that draws moist air inland from the Gulf of Mexico (Cole, 1975). The coincidence of precipitation and temperature maxima is typical of a monsoonal climate (Figure 2). Much of the rain falls during localized and often intense summer thunderstorms, and winters are cool and generally dry. Both temperature and precipitation are dependent on elevation, and local climates vary with topography. At lower elevations throughout the region, including the vicinity of the WIPP, potential evaporation greatly exceeds precipitation. Freshwater pan evaporation in the region is estimated to exceed 274 cm/yr (Hunter, 1985). Surface runoff and infiltration of rainwater into the subsurface are limited. Hunter (1985) concluded from a literature review that within the vicinity of the WIPP an average of 96 percent of precipitation is lost to evapotranspiration. Evapotranspiration values may be significantly higher or lower locally.

3. Climatic Change

Presently available long-term climate models are incapable of resolution on the spatial scales required (e.g., Hansen and others, 1988; Mitchell, 1989; Houghton and others, 1990), and it is not realistic to predict the climate of southeastern New Mexico for the next 10 ka. Instead, this report reviews evidence of past climatic changes in the region, and establishes limits on future precipitation based on known and modeled past extremes. Much of the available paleoclimatic data only record long-term average levels of precipitation, and these limits do not reflect the high variability apparent in the modern short-term data. The precipitation record presented here primarily reflects gradual shifts in long-term mean values.

A fundamental assumption, analogous to that made by Spaulding (1985) in a study of climatic variability at the Nevada Test Site, is that climatic extremes of the next 10 ka will not exceed those associated with glaciations and deglaciations that have recurred repeatedly in the northern hemisphere since the late Pliocene (2.5 Ma BP). The possibility that human-induced changes in the composition of the earth's atmosphere may influence future climates complicates projections of this cyclic pattern into the future, but, as presently modeled (e.g., Mitchell, 1989; Houghton and others, 1990), such changes do not appear likely to have a negative effect on the performance of the WIPP. The highest past precipitation levels in southeastern New Mexico,

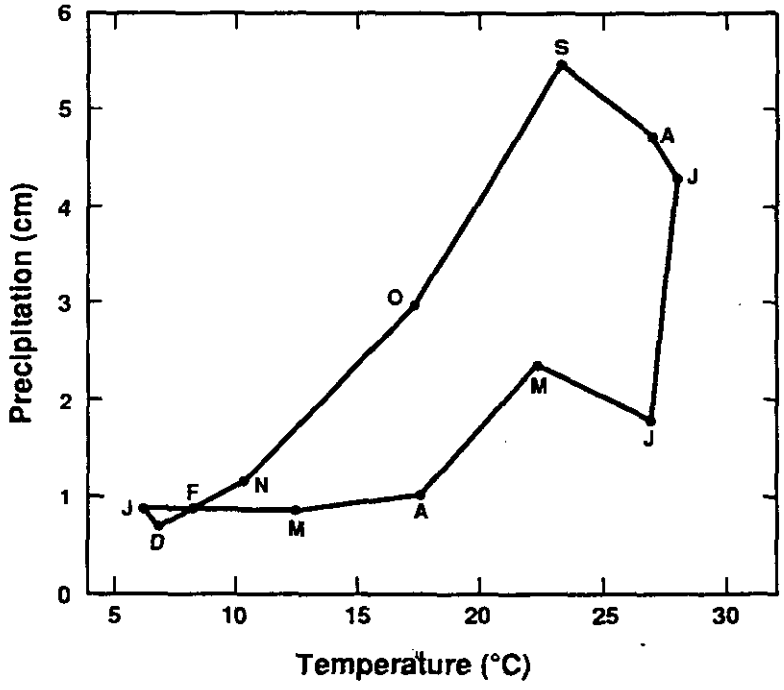

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Figure 2. Climatograph showing 30-yr (1951-1980) monthly precipitation and temperature means recorded at the Carlsbad, New Mexico airport, approximately 45 km west of the WIPP and 50 m lower (data from NOAA, 1989).

up to twice those of the present, occurred during full-glacial conditions associated with global cooling (e.g., Van Devender and others, 1987; other sources cited below). Presently available greenhouse models, however, predict average equilibrium global warming of 1.8 to 5.2°C for carbon dioxide concentrations twice present levels (Mitchell, 1989; Houghton and other, 1990), a condition that could delay the start of renewed glaciation. Published model predictions of precipitation trends accompanying greenhouse warming are less consistent and less reliable than temperature predictions, but none suggest significantly higher levels of precipitation in southern New Mexico than those of the present (Washington and Meehl, 1984; Wilson and Mitchell, 1987; Schlesinger and Mitchell, 1987; Houghton and others, 1990). Because long-term increases in recharge are improbable without increases in precipitation, the highest-risk climatic change that will be considered here is, therefore, a return to the glacial extremes of the past.

Data that can be used to interpret paleoclimates in southeastern New Mexico and the surrounding region come from a variety of sources, and indicate an alternation of arid and subarid to subhumid climates throughout the Pleistocene. Prior to 18 ka BP, radiometric dates are relatively scarce, and the record is incomplete. From 18 ka BP to the present, however, the climatic record is relatively complete and temporally well constrained by radiocarbon dates. This report cites extensive floral, faunal, and lacustrine data from the region that permit reconstructions of precipitation and temperature during the late Pleistocene and Holocene. These data span the transition from the last full-glacial maximum to the present interglacial period, and, given the global consistency of glacial fluctuations as described below, they can be taken to be broadly representative of extremes for the entire Pleistocene.

4. Variability in Global Climate Over the Last 2.5 Million Years

Core samples of datable marine sediments provide a continuous record that reveals as many as 50 glaciation/deglaciation events in the last 2.5 Ma. Specifically, correlations have been made between major glacial events and variables such as the ratio of $^{18}O/^{16}O$ measured in the remains of calcareous foraminifera and past sea-surface temperatures determined from planktonic assemblages (Ruddiman and Wright, 1987). In addition, glacial cycles have been observed in the past composition of the earth's atmosphere preserved in polar glacial ice (Langway and others, 1985; Jouzel and others, 1987; Barnola and others, 1987) and in a $^{18}O/^{16}O$ record from calcite vein fillings in Nevada (Winograd and others, 1988).

Oxygen isotope ratios from oceanic foraminifera provide the most direct evidence, because they reflect past volumes of glacial ice (Imbrie and others, 1984). Evaporation fractionates ^{18}O and ^{16}O isotopes in water, producing a vapor relatively enriched in ^{16}O and residual seawater relatively enriched in ^{18}O. Glacial ice sheets store large volumes of ^{16}O-enriched meteoric water, thereby preventing the remixing of the two isotope fractions and significantly altering $\delta^{18}O$ values in the world's oceans.* Foraminifera preserve samples of past $\delta^{18}O$ values when they extract oxygen from seawater and incorporate it into calcareous body parts. Abundant fossil remains permit the construction of detailed records such as that shown in Figure 3, covering the last 780 ka. High positive values of $\delta^{18}O$ reflect glacial maxima, and negative values reflect warm interglacial periods. Because the largest volumes of glacial ice were incorporated in the North American sheet, $\delta^{18}O$ fluctuations can be interpreted directly as a first order record of North American glaciation and deglaciation (Mix, 1987; Ruddiman and Wright, 1987). Because the correlation is quantitative, the isotopic record indicates that the most recent glacial event was as severe as any within the last 780 ka. It also indicates that the present value is at or near that of a glacial minimum.

Sea-surface temperature records, although not as closely tied to glacial events, show the same alternating pattern. Temperatures at the surface of northern hemisphere oceans, as determined from the fossil assemblages of planktonic foraminiferal species, were measurably colder during glaciation and warmer during interglacial periods (Ruddiman, 1987).

Samples from the ice sheets of Greenland and Antarctica and calcite vein fillings in Nevada provide independent confirmation of the oceanic data (Langway and others, 1985; Jouzel and others, 1987; Barnola and others, 1987; Winograd and others, 1988). Glacial ice preserves $\delta^{18}O$ and δD values of the precipitation that formed the ice, and, because fractionation of the isotopes is temperature dependent, fluctuations can be interpreted quantitatively as changes in local mean temperature. Bubbles of air trapped within the ice can also be sampled to give a measure of past CO_2 concentrations in the atmosphere, which, because of the importance of CO_2 in the earth's greenhouse effect, correlate well with the isotopic temperature record. Figure 4 shows

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* By convention, $^{18}O/^{16}O$ ratios are reported as:

$$\delta^{18}O = 1000 \times \frac{(^{18}O/^{16}O_{sample} - {}^{18}O/^{16}O_{reference})}{{}^{18}O/^{16}O_{reference}}$$

Deuterium/hydrogen ratios (D/H) are similarly reported as δD values.

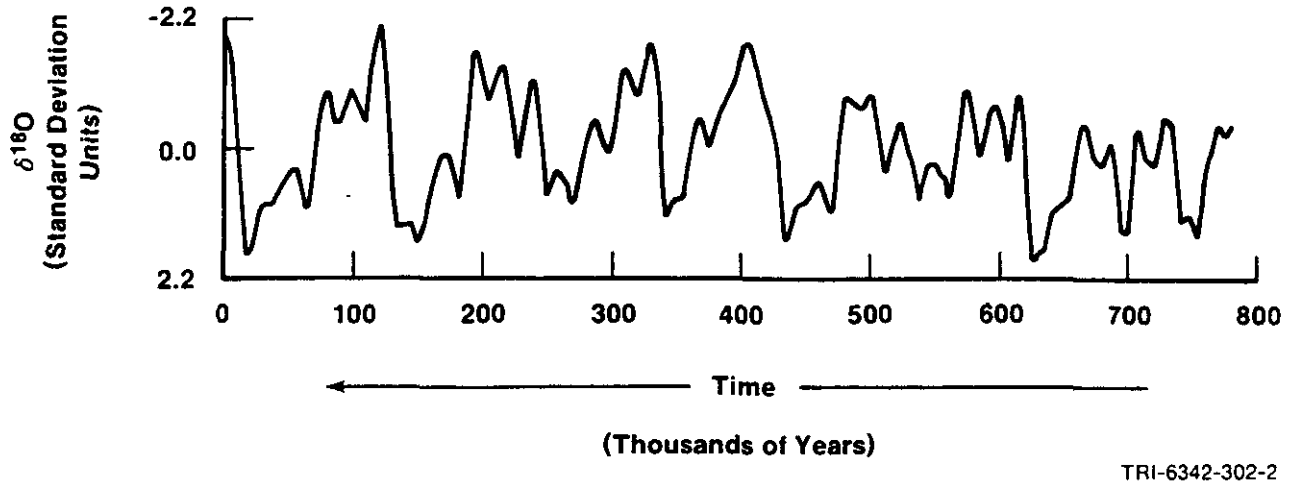

Figure 3. Foraminiferal $\delta^{18}0$ record of the last 780,000 years. Curve reflects $\delta^{18}0$ variations from five deep-sea core samples. Data have been normalized, stacked, and smoothed with a nine-point Gaussian filter (Imbrie, and others, 1984).

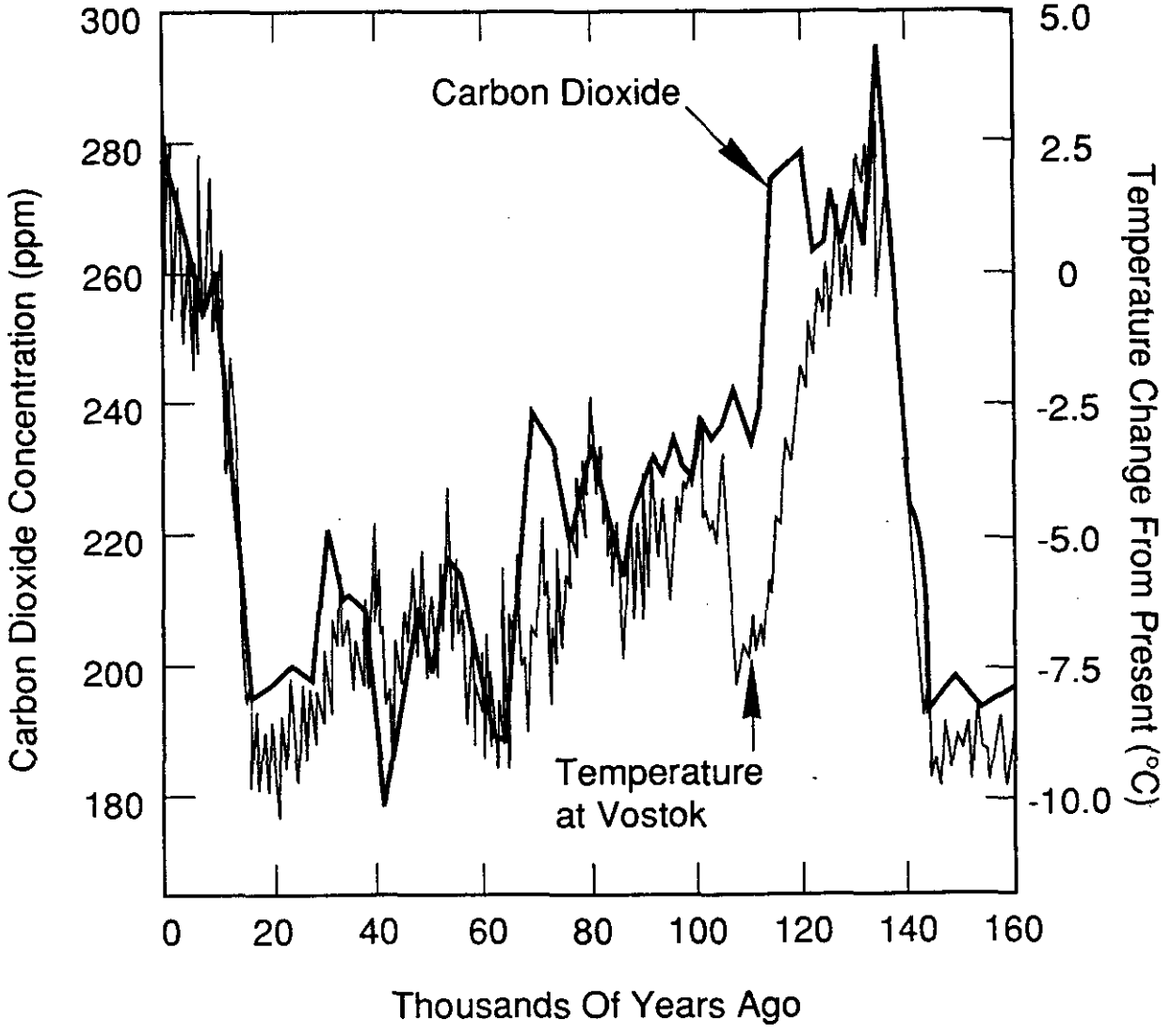

Figure 4. Atmospheric CO_2 concentration and δD temperature record from the Vostok, Antartica, ice core for the last 160 ka (modified from Schneider, 1990; after Barnola and others, 1987).

the 160-ka record of CO_2 concentration and temperature as determined from the D/H ratio at Vostok in East Antarctica. With some minor discrepancies, the two curves reveal the same basic glacial chronology visible in the last 160 ka of the oceanic $\delta^{18}O$ record shown in Figure 3. The $\delta^{18}O$ record from 50 ka to 310 ka BP from calcite vein fillings in Devils Hole, Nevada, reflects changes in local surface temperature as well as other, less well quantified, factors, including groundwater travel time. This record also shows a chronology similar to that of Figure 3, although peaks in the calcite curve are shifted toward increasingly older times earlier in the record relative to the oceanic curve. The reason for this phase shift, which reaches 28 ka at 272 ka BP, is not known (Winograd and others, 1988).

5. Stability of Glacial Cycles

The causes of glaciation and deglaciation are complex and not fully understood (Ruddiman and Wright, 1987), but the strong periodicity of the isotopic record indicates that climatic alternations have been systematic in the past. Spectral analysis of the foraminiferal $\delta^{18}O$ curve for the last 780 ka shows that within that time the primary control on the periodicity of glacial events has been variation in global insolation caused by irregularities in the earth's orbit (Figure 5). Observed periods of 19, 23, 41, and 100 ka in the oceanic $\delta^{18}O$ curve correspond to calculated periods of northern hemisphere summer insolation minima of 19 and 23 ka related to the precession of the earth's axis, 41 ka related to the tilt of earth's axis, and 94, 125, and 413 ka related to the eccentricity of the earth's orbit (Milankovitch, 1941; Hays and others, 1976; Imbrie and others, 1984; Imbrie, 1985). Calculations based on astronomical observations indicate that orbital parameters have not changed significantly in the last 5 Ma (Berger, 1984), and geological evidence suggests they may have been stable for as long as 300 Ma (Anderson, 1984; Heckel, 1986).

Longer-term global climatic changes, such as the beginning of the present pattern of glaciation and deglaciation 2.5 Ma BP, have been attributed to changes in the configuration of the earth's continents, which in turn controls both the potential distribution of ice sheets and global circulation patterns (e.g., Crowell and Frakes, 1970; Caputo and Crowell, 1985; Crowley and others, 1986; Hyde and others, 1990). Continental masses move at plate-tectonic rates of centimeters per year, several orders of magnitude too low to affect glacial processes within the next 10 ka. Vertical uplift or subsidence of large continental regions may also affect global climate by changing circulation patterns (e.g., Ruddiman and Kutzbach, 1989), but, again, maximum uplift rates are at least an order of magnitude too low to change present circulation patterns within the next 10 ka.

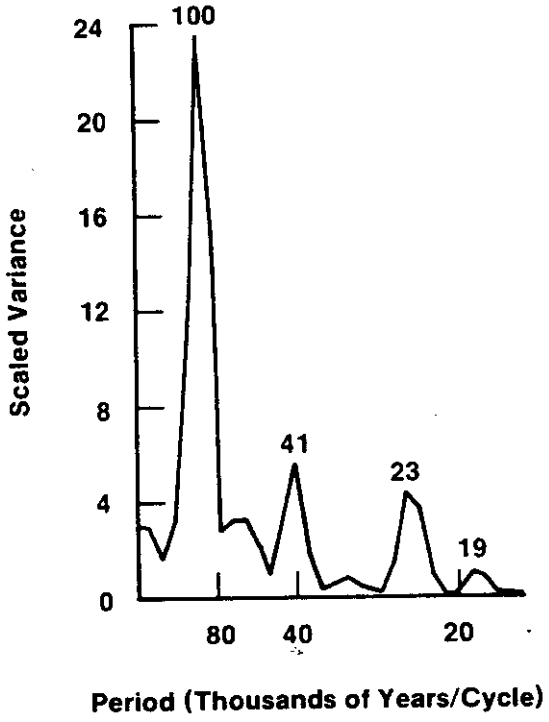

Figure 5. Spectral analysis of the foraminiferal $\delta^{18}O$ record presented in Figure 3, showing periodicity of glaciation and deglaciation (after Imbrie, 1985).

The long-term pattern of the cycles of glaciation and deglaciation provides the basis for concluding that climatic extremes of the next 10 ka will remain within past limits. The relative amplitudes of past glacial cycles (Figure 3) imply that future glaciations will be no more severe than the last one. The periodicity of the pattern indicates that although glacial minima such as that of the present are relatively brief, glacial advances are, in general, slow and the next full maximum will not occur for many tens of thousands of years. Predictions about the precise timing of future glacial events are not straightforward, however. Higher resolution records, such as those available from polar ice cores, show that some glacial advances can occur relatively rapidly. Modeling of glacial processes is complicated by uncertainties about feedback processes involved in the growth of ice sheets, but extrapolation of the isotopic curve of Figure 3 using a relatively simple model for nonlinear climate response to insolation change suggests that, in the absence of anthropogenic effects, the next full glacial maximum could occur in approximately 60 ka (Imbrie and Imbrie, 1980). These observations, combined with the climatic data discussed below, justify the choice of the late Pleistocene full-glacial climate as a conservative upper limit for precipitation during the next 10 ka.

6. Pleistocene and Holocene Climates of Southeastern New Mexico

Early and middle Pleistocene paleoclimatic data for southeastern New Mexico and the surrounding region are incomplete, and permit neither continuous reconstructions of paleoclimates nor direct correlations between climate and glaciation prior to the last glacial maximum 22 to 18 ka BP. Stratigraphic and pedologic data from several locations (Figure 6), however, indicate that cyclical alternation of wetter and drier climates in southeastern New Mexico had begun by the early Pleistocene. Fluvial gravels in the Gatuña Formation exposed in the Pecos River Valley of eastern New Mexico suggest relatively wetter conditions 1.4 Ma BP and again 600 ka BP (Bachman, 1987). The Mescalero caliche, exposed locally over much of southeastern New Mexico, has been interpreted as indicating relatively drier conditions 510 ka BP (Lambert and Carter, 1987), and loosely dated spring deposits in Nash Draw west of the WIPP imply wetter conditions again later in the Pleistocene (Bachman, 1981, 1987). The Blackwater Draw Formation of the southern High Plains of eastern New Mexico and western Texas, time-correlative to both the Gatuña Formation and the Mescalero caliche, contains alternating soil and eolian sand horizons that show at least six climatic cycles beginning more than 1.4 Ma BP and continuing to the present (Holliday, 1989a). The duration, frequency, and total number of Pleistocene climatic cycles in southeastern New Mexico have not been established.

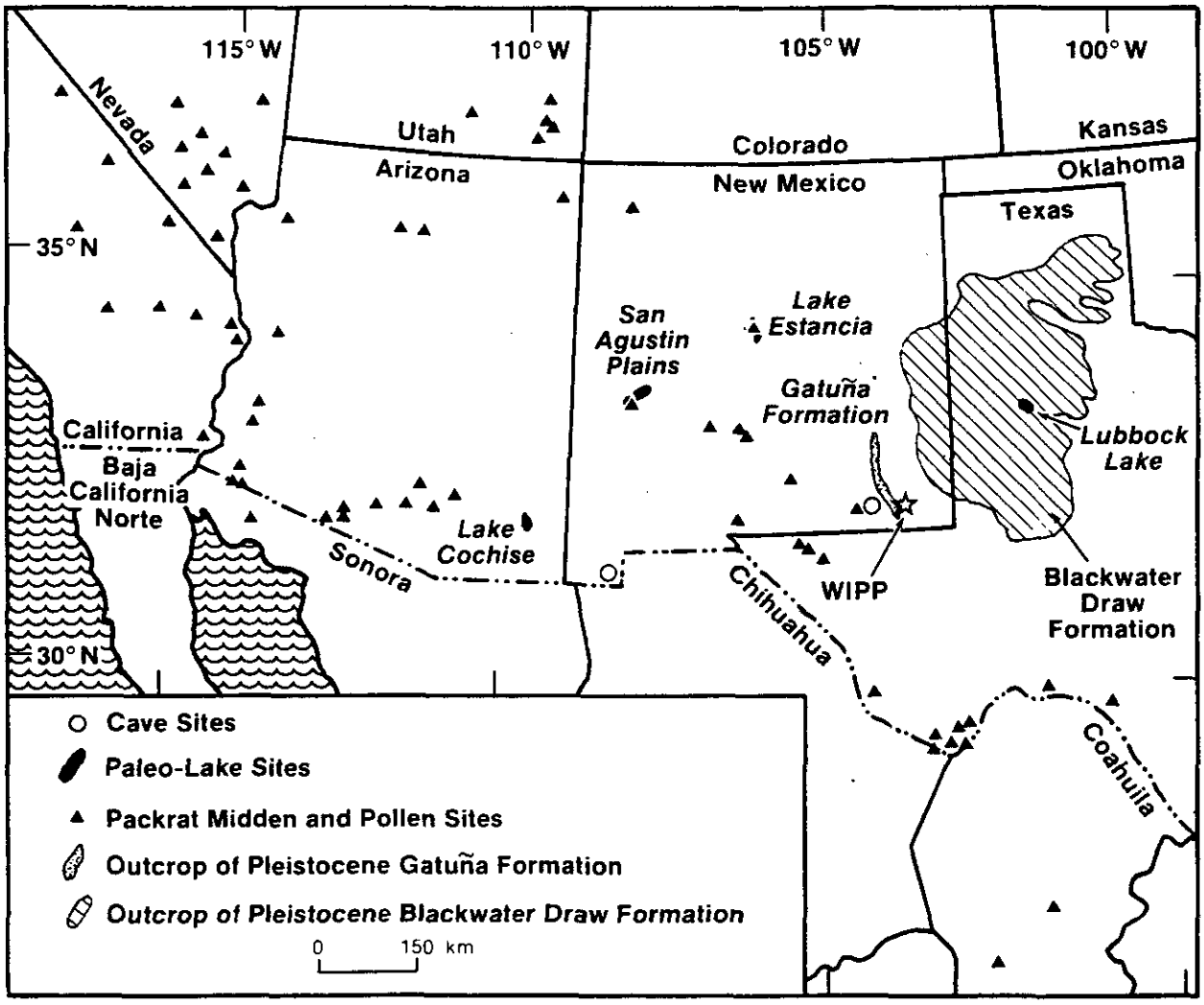

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Figure 6. Location map for paleoclimate data. Data from Bachman (1981); Markgraf and others (1983); Harris (1987); Pierce (1987); Van Devender and others (1987); Waters (1989); Bachhuber (1989); Holliday (1989a); Van Devender (1990); Allen (1991); Phillips and others (1992).

Data used to construct the more detailed climatic record for the latest Pleistocene and Holocene come from multiple lines of evidence dated primarily using carbon-14 techniques. Packrat middens examined at sites throughout the southwestern United States, including locations in southeastern New Mexico, preserve local plant communities and, in some cases, insect remains (Van Devender, 1980, 1990; Van Devender and others, 1984, 1987; Elias, 1987; Elias and Van Devender, 1990). Pollen assemblages have been analyzed from lacustrine deposits in western New Mexico, western Texas, and other locations in the southwestern United States (Martin and Mehringer, 1965; Markgraf and others, 1984; Bryant and Holloway, 1985; Van Devender and others, 1987). Faunal data come from gastropod assemblages from western Texas (Pierce, 1987), ostracode assemblages from western New Mexico (Markgraf and others, 1984; Forester, 1987; Phillips and others, 1992), and vertebrate remains from caves in southern New Mexico (Harris, 1987, 1988). Stable-isotope data are available from ostracodes in western New Mexico (Phillips and others, 1992) and groundwater samples in northwestern New Mexico (Phillips and others, 1986). Paleo-lake level data are available from sites throughout the southwestern United States (Reeves, 1973; Smith and Anderson, 1982; Markgraf and others, 1983, 1984; Benson and Thompson, 1987; Holliday and Allen, 1987; Bachhuber, 1989; Waters, 1989; Wells and others, 1989; Enzel and others, 1989; Benson and others, 1990; Allen, 1991). Figure 6 shows the locations of key sites discussed here and in the references cited.

Because decreases in temperature and increases in precipitation produce similar environmental changes, not all data cited uniquely require the paleoclimatic interpretation presented in this report (Figure 7). For example, lake-level increases can, in theory, result solely from decreased evaporation at lower temperatures. Interpretations drawn individually from each of the data sets are consistent with the overall trends shown in Figure 7, however, and the pattern of change is confirmed by global general circulation climate models (Kutzbach and Guetter, 1986; COHMAP Members, 1988). Furthermore, specific floral and faunal assemblages are sufficiently sensitive to precipitation and temperature effects to distinguish between the two (e.g., Van Devender and others, 1987; Pierce, 1987; Van Devender; 1990). The paleoclimates described here are those that best explain data from all sources.

Prior to the last glacial maximum 22 to 18 ka BP, evidence from mid-Wisconsin faunal assemblages in caves in southern New Mexico, including the presence of extralimital species such as the desert tortoise that are now restricted to warmer climates, suggests warm summers and mild, relatively dry winters (Harris, 1987, 1988). Lacustrine evidence confirms the interpretation that conditions prior to and during the glacial advance that

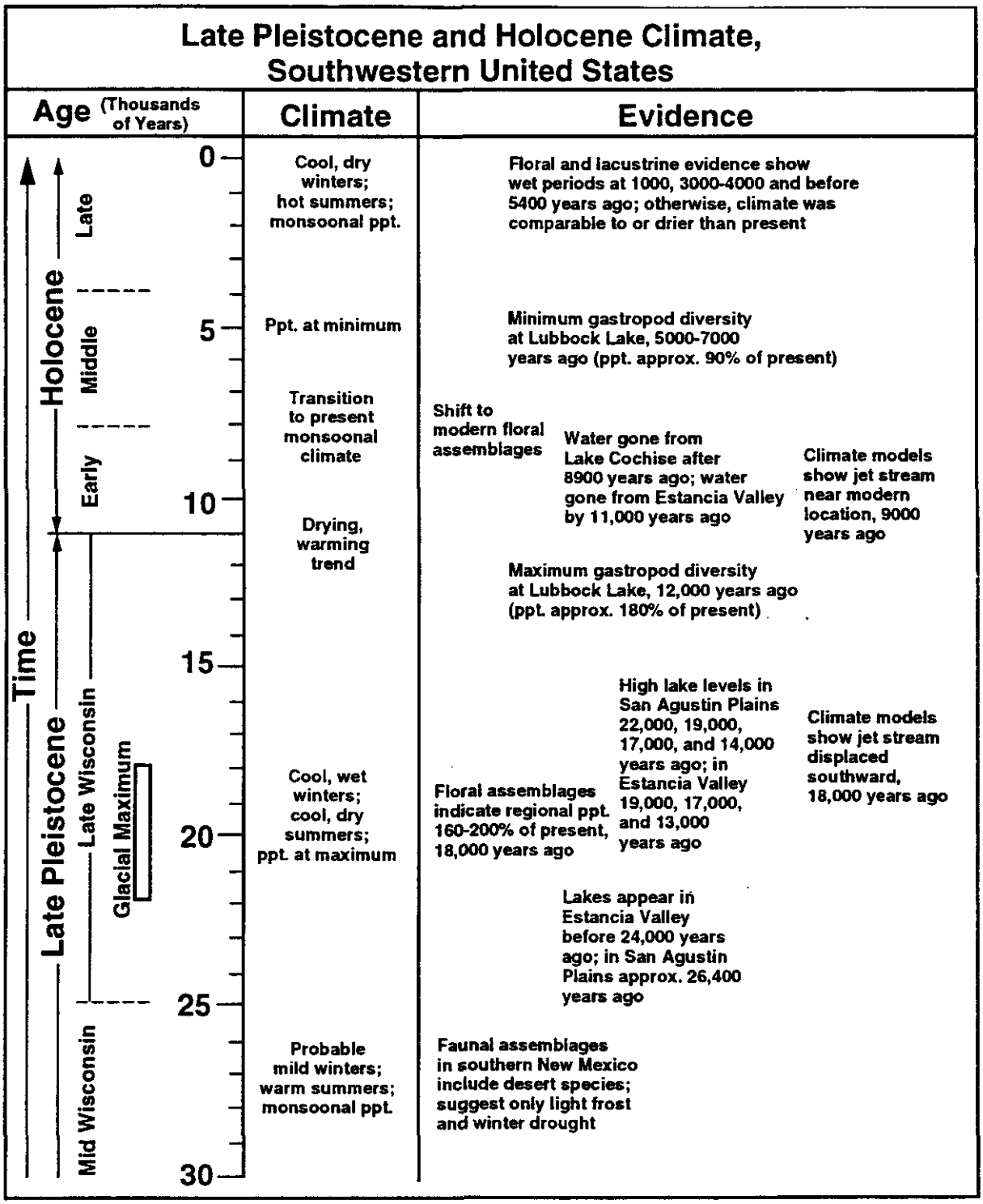

Figure 7. Late Pleistocene and Holocene climate, southwestern United States. Time scale after Van Devender and others (1987). Climate references cited in text.

were generally drier than those at the glacial maximum. Permanent water did not appear in what was later to be a major lake in the Estancia Valley in central New Mexico until sometime before 24 ka BP (Bachhuber, 1989). Late-Pleistocene lake levels in the San Agustin Plains in western New Mexico remained low until approximately 26.4 ka BP, and the $\delta^{18}O$ record from ostracode shells suggests that mean annual temperatures at that location did not decrease significantly until approximately 22 ka BP (Phillips and others, 1992).

Ample floral and lacustrine evidence documents cooler and wetter conditions in southeastern New Mexico and the surrounding region during the glacial peak (Van Devender and others, 1987; Pierce, 1987; Bachhuber, 1989; Allen, 1991; Phillips and others, 1992). These changes were not caused by the immediate proximity of glacial ice. None of the Pleistocene continental glaciations advanced farther southwest than northeastern Kansas, and the most recent, late Wisconsin, ice sheet reached its limit in South Dakota, roughly 1200 km from the WIPP (Andrews, 1987). Discontinuous alpine glaciers formed at the highest elevations throughout the Rocky Mountains, but these isolated ice masses were symptoms, rather than causes, of cooler and wetter conditions, and had little influence on regional climate at lower elevations. The closest such glacier to the WIPP was on the northeast face of Sierra Blanca Peak in the Sacramento Mountains, 220 km to the northwest (Richmond, 1962).

Global climate models indicate that the dominant glacial effect in southeastern New Mexico was the disruption and southward displacement of the westerly jet stream by the physical mass of the ice sheet to the north (Figure 8) (Manabe and Broccoli, 1985; Kutzbach and Guetter, 1986; COHMAP members, 1988). At the glacial peak, climate models show that major Pacific storm systems followed the jet stream across New Mexico and the southern Rocky Mountains, and winters were wetter and longer than either at the present or during the previous interglacial period.

Field evidence does not support the suggestion (Galloway, 1970, 1983; Brakenridge, 1978) that higher lake levels and changed faunal and floral assemblages at the glacial maximum could have resulted solely from lowered temperatures. Plant communities indicate that, regionally, the decrease in mean annual temperatures below present values was significantly less than the 7 to 12°C required by cold and dry climate models (Van Devender and others, 1987; Van Devender, 1990). Interpretation of stable-isotope data from groundwater samples from northwestern New Mexico suggest mean annual temperatures were 5 to 7°C colder than at present at that location (Phillips and others, 1986). Isotopic data from ostracode shells in the San Agustin Plains suggest mean annual temperatures there may have been 8.3°C colder than

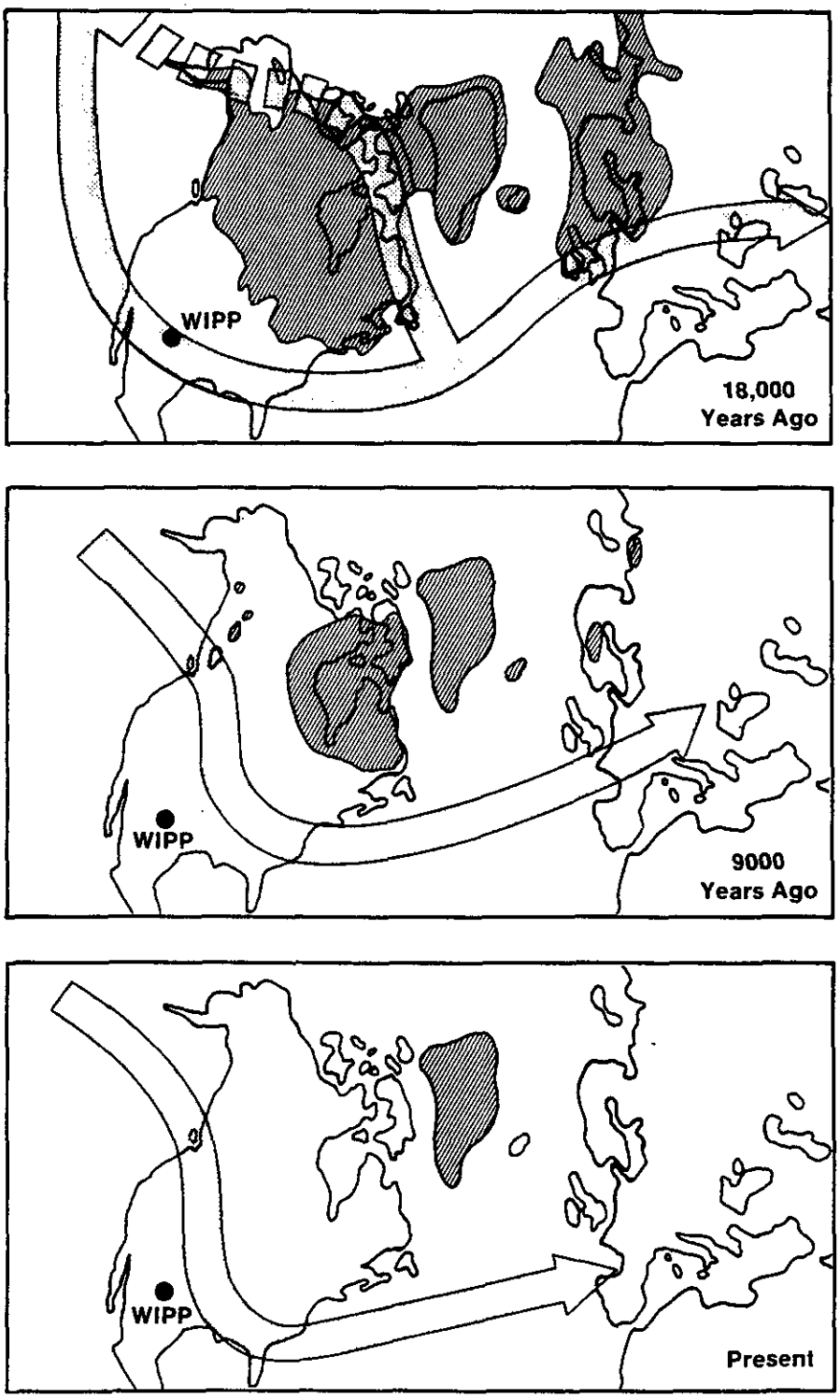

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Figure 8. Distribution of northern hemisphere ice sheets and modeled position of jet stream at 18 ka BP, and present (from COHMAP Members, 1988). Ice shown with dark pattern, jet stream shown with arrow (broken where disrupted or weak).

at present at approximately 21 ka BP, but hydrologic modeling of the basin indicates lake levels were controlled by precipitation rather than evaporation throughout the glacial advance and retreat (Phillips and others, 1992). Gastropod assemblages at Lubbock Lake in western Texas suggest mean annual temperatures 5°C below present values (Pierce, 1987). High water levels in playas in western Texas reflect an increase in runoff best explained by substantially higher levels of precipitation (Reeves, 1973).

Floral and faunal evidence taken together indicate that mean annual precipitation throughout the region at the last glacial maximum was 60 to 100 percent more than today (Spaulding and Graumlich, 1986; Pierce, 1987; Van Devender and others, 1987). Floral evidence also suggests that winters may have continued to be relatively mild, perhaps because the glacial mass blocked the southward movement of arctic air. Summers at the glacial maximum were cooler and drier than at present, without a strongly developed monsoon (Van Devender and others, 1987). Piñons, oaks, and junipers grew at lower elevations throughout southern New Mexico (Van Devender and others, 1987; Van Devender, 1990), probably including the vicinity of the WIPP.

According to climatic modeling, the jet stream shifted northward following the gradual retreat of the ice sheet after 18 ka BP (Figure 8), and the climate responded accordingly. By the Pleistocene/Holocene boundary approximately 11 ka BP, conditions were significantly warmer and drier than previously, although still dominated by winter storms and still wetter than today (Van Devender and others, 1987). Major decreases in total precipitation and the shift toward the modern monsoonal climate did not occur until the margins of the ice sheet had retreated into northeastern Canada in the early Holocene.

Evidence for the late Pleistocene and early Holocene drying trend comes from several sources. In contrast to the northern Great Basin, where lake levels continued to rise until between 15 and 13.5 ka BP (Benson and others, 1990), lake levels in southern New Mexico and the surrounding region decreased following the glacial maximum. Permanent water disappeared from late-Pleistocene lakes in the Estancia Valley between 12 and 11 ka BP, following a series of fluctuations in lake levels with progressively lower high stands at approximately 19, 17, and 13 ka BP (Allen, 1991). Lake Cochise (the modern Willcox Playa) in southeastern Arizona was dry after 8.5 ka BP, following two high stands prior to 14 ka BP and a third between 14 and 13 ka BP (Waters, 1989). Modeling of lake levels in the San Agustin Plains shows high stands at progressively lower elevations at approximately 22, 19, 17, and 14 ka BP (Phillips and others, 1992). Water remained in lakes in the San Agustin Plains until 5 ka BP, but ostracode assemblages suggest an increase in salinity by 8 ka BP, and the pollen record shows a gradual shift

at that location from a spruce-pine forest 18 to 15 ka BP to a juniper-pine forest by 10 ka BP (Markgraf and others, 1984). Packrat middens in Eddy County, New Mexico, indicate that desert-grassland and desert-scrub communities predominated at lower elevations between 10.5 and 10 ka BP (Van Devender, 1980). Soil studies indicate drier conditions at Lubbock Lake after 10 ka BP, although marshes and small lakes persisted at the site until the construction of a dam and reservoir in 1936 (Holliday and Allen, 1987). Based on a decrease in diversity of both terrestrial and aquatic gastropod species, Pierce (1987) estimated a drop in annual precipitation at Lubbock Lake from a high of 80 cm/yr (nearly twice the modern level at that location of 45 cm/yr) at 12 ka BP to 40 cm/yr by 7 ka BP.

Coincident with this decrease in precipitation, evidence from vole remains recovered from caves in southern New Mexico (Harris, 1988) and from plant communities throughout the southwestern United States (Van Devender and others, 1987) indicates a rise in summer temperatures. Mean annual temperatures interpreted from the isotopic composition of groundwater samples from northwestern New Mexico also show a sharp rise in the early Holocene (Phillips and others, 1986).

By middle-Holocene time, the climate was similar to that of the present, with hot, monsoon-dominated summers and cold, dry winters. The pattern has persisted to the present, but not without significant local variations. Soil studies show the southern High Plains were drier from 6.5 to 4.5 ka BP (Holliday, 1989b) than before or since. Gastropod data from Lubbock Lake indicate the driest conditions from 7 to 5 ka BP (precipitation approximately 90 percent of present, mean annual temperature 2.5°C higher than present), with a cooler and wetter period at 1 ka BP (precipitation approximately 145 percent of present, mean annual temperature 2.5°C lower than present) (Pierce, 1987). Plant assemblages from southwestern Arizona suggest steadily decreasing precipitation from the middle Holocene to the present, except for a brief wet period around 990 years ago (Van Devender and others, 1987). Stratigraphic work at Lake Cochise shows two mid-Holocene lake stands, one near or before 5.4 ka BP and one between or before 3 to 4 ka BP, but both were relatively short-lived, and neither reached the maximum depths of the late-Pleistocene high stand that existed before 14 ka BP (Waters, 1989).

Precipitation maxima during these Holocene wet periods were less in both magnitude and duration than those of the late Pleistocene. Enzel and others (1989) observed comparable Holocene wet periods recorded in playa deposits in the Mojave Desert 3620 ± 70 and 390 ± 90 years ago, and related them to short-term changes in global circulation patterns that resulted in increased winter storm activity in the region. Historical records over the last several hundred years indicate numerous lower intensity climatic

fluctuations, some too short in duration to affect floral and faunal assemblages, which may also be the result of temporary changes in global circulation (Neilson, 1986). Sunspot cycles and the related changes in the amount of energy emitted by the sun have been linked to historical climatic changes elsewhere in the world (e.g., Lamb, 1972), but the validity of the correlation is uncertain (Robock, 1979; Stuiver, 1980). Correlations also have been proposed between volcanic activity and climatic change (Robock, 1979; Palais and Sigurdsson, 1989; Bryson, 1989). In general, however, causes for past short-term changes are unknown, and it is impossible at present to predict the amplitude or frequency of recurrence. Despite this uncertainty, the past record does support the conclusion that future short-term fluctuations in southeastern New Mexico will not be as severe as the long-term climatic changes created by major ice sheets in the northern hemisphere. Full-glacial conditions remain a conservative upper limit for mean annual precipitation at the WIPP during the next 10 ka.

7. Climatic Implications of Data from WIPP Groundwater Samples

Isotopic data from groundwater samples collected in the vicinity of the WIPP from the Late Permian Rustler and Dewey Lake Formations that overlie the Salado Formation are generally consistent with the climatic changes described above. Lambert (1986) and Lambert and Harvey (1987) concluded that although deuterium/hydrogen and $^{18}O/^{16}O$ ratios indicate a meteoric origin for water in the confined aquifers, they are sufficiently distinct from modern surface water values to suggest that the contribution of modern recharge to the system is slight. Chapman (1986) disagreed with this interpretation, noting similar ratios in the presumably young waters of the Roswell Artesian Basin immediately to the north, and she concluded that stable-isotope data from the WIPP area do not permit definitive interpretations about the age of the groundwater. Tritium data are less ambiguous. Low tritium levels in all WIPP-area samples indicate minimal contributions from the atmosphere since 1950 (Lambert, 1987; Lambert and Harvey, 1987). The four internally consistent radiocarbon analyses currently available for water samples from the Rustler and Dewey Lake Formations support this interpretation. Modeled minimum ages in each case are between 12 and 16 ka, suggesting that both units have had little recharge since the period immediately following the late Pleistocene glacial maximum (Lambert and Harvey, 1987). Lambert and Carter (1987) presented uranium isotope data that also support this interpretation: observed high $^{234}U/^{238}U$ activity ratios require a conservative minimum residence time in the Culebra Dolomite of several thousands of years and more probably reflect minimum ages of 10 to 30 ka. Chapman (1988) questioned the validity of equating isotope residence times with groundwater age, but agreed that high $^{234}U/^{238}U$ activity ratios occur in

regions of low transmissivity, where flow is presumably slower and residence times are longer.

Lambert (1991) used groundwater isotope data, along with supporting evidence from $^{87}Sr/^{86}Sr$ ratios in vein fillings, to argue that the Rustler Formation has been essentially a closed hydrologic system for the last 12 ka. In his interpretation, significant recharge last occurred during the late Pleistocene, and the present flow field reflects the slow draining of the aquifer. If this interpretation is correct, recharge may not occur again until precipitation levels are substantially higher than at present.

Other data suggest that, isotopic evidence notwithstanding, some recharge may be occurring at the present. Anomalous increases in water levels have been observed at seven WIPP-area wells since 1988 (Beauheim, 1989). Recharge from the surface cannot be ruled out as a cause for these rises, although no specific link to precipitation events has been demonstrated. Other possible causes include decreases in discharge, changes in reservoir volume related to incomplete recovery from the transient pressure changes associated with the pumping test itself, changes in reservoir volume related to external changes in the regional stress field, or undetected recharge from other aquifers or from the surface through existing boreholes (Beauheim, 1989). Numerical modeling of groundwater flow in the WIPP area indicates that, although it is hydraulically possible for present flow to reflect late Pleistocene recharge (Davies, 1989), some component of modern vertical recharge is also compatible with observed conditions (Haug and others, 1987; Davies, 1989). Major ion chemical analyses of groundwater samples support the interpretation of vertical recharge south of the WIPP, where low salinities may be the result of mixing with fresh surface water (Chapman, 1988). Lambert (1991) suggests instead that water chemistry has remained essentially unchanged from the late Pleistocene and is a function of host rock composition, noting that groundwater salinity correlates well with the distribution of halite in the Rustler Formation.

Questions about recharge to the Rustler Formation and the true age of WIPP-area groundwater remain unanswered. In the absence of definitive data, this report makes no assumptions about groundwater age.

8. Discussion and Conclusions

Speculation about future climate variability must be based on observed past fluctuations. The largest global climatic changes in the last 2.5 Ma have been those associated with glaciation and deglaciation in the northern hemisphere. The high degree of consistency in both frequency and intensity

displayed in the glacial record indicates that an accurate interpretation of past climatic cycles does provide a useful guide for estimating future changes.

Geologic data from southeastern New Mexico and the surrounding region show repeated alternations of wetter and drier climates throughout the Pleistocene. Floral, faunal, and lacustrine data permit detailed and quantitative reconstructions of precipitation that can be linked directly to glacial events of the late Pleistocene and Holocene. Figure 9 shows estimated mean annual precipitation for the WIPP for the last 30 ka, interpolated from the composite regional data cited above and based on present average precipitation at the site of 30 cm/yr (Brinster, 1991). This plot should be interpreted with caution, because its resolution and accuracy are limited by the nature of the data used to construct it. Floral and faunal assemblages may change gradually and show only a limited response to climatic fluctuations that occur at frequencies higher than the typical life span of the organisms in question. For long-lived species such as trees, resolution may be limited to hundreds or even thousands of years (Neilson, 1986). Sedimentation in lakes and playas has the potential to record higher frequency fluctuations, including single-storm events, but only under a limited range of circumstances. Once water levels reach a spill point, for example, lakes show only a limited response to further increases in precipitation. Dry playas generally show little response to decreases in precipitation. A more complete record of precipitation would almost certainly show far more variability than that implied by the plot presented here. Specifically, Figure 9 may fail to record abnormal precipitation lows during the Holocene, it may underestimate the number of high-precipitation peaks during the same period, and it may underestimate the magnitude of relatively brief precipitation maxima during the late Pleistocene. Although the magnitude of the long-term shift in precipitation is adequately documented by the data reviewed here, the amplitude of the higher-frequency fluctuations in both the late Pleistocene and the Holocene is not well constrained, and the climate may have been wetter or drier than shown for some intervals.

With these observations in mind, three significant conclusions can be drawn from the climatic record of southeastern New Mexico and the surrounding region. First, maximum precipitation in southeastern New Mexico in the past coincided with the maximum advance of the North American ice sheet. Minimum precipitation occurred after the ice sheet had retreated to its present limits. Second, past maximum long-term average precipitation levels were roughly twice present levels. Minimum levels may have been 90 percent of present levels. Third, short-term fluctuations in precipitation have occurred during both the glacial maximum and the present, relatively dry,

interglacial period, but fluctuations during the present interglacial period have not exceeded the upper limits of the glacial maximum.

It would be unrealistic to attempt a direct extrapolation of the precipitation curve of Figure 9 into the future. Too little is known about the relatively short-term behavior of global circulation patterns, and it is at present impossible to predict the probability of a recurrence of a wetter climate such as that of approximately 1000 years ago. The long-term stability of patterns of glaciation and deglaciation, however, do permit the conclusion that future climatic extremes are unlikely to exceed those of the late Pleistocene. Furthermore, the periodicity of glacial events suggests that a return to full glacial conditions is highly unlikely within the next 10,000 years.

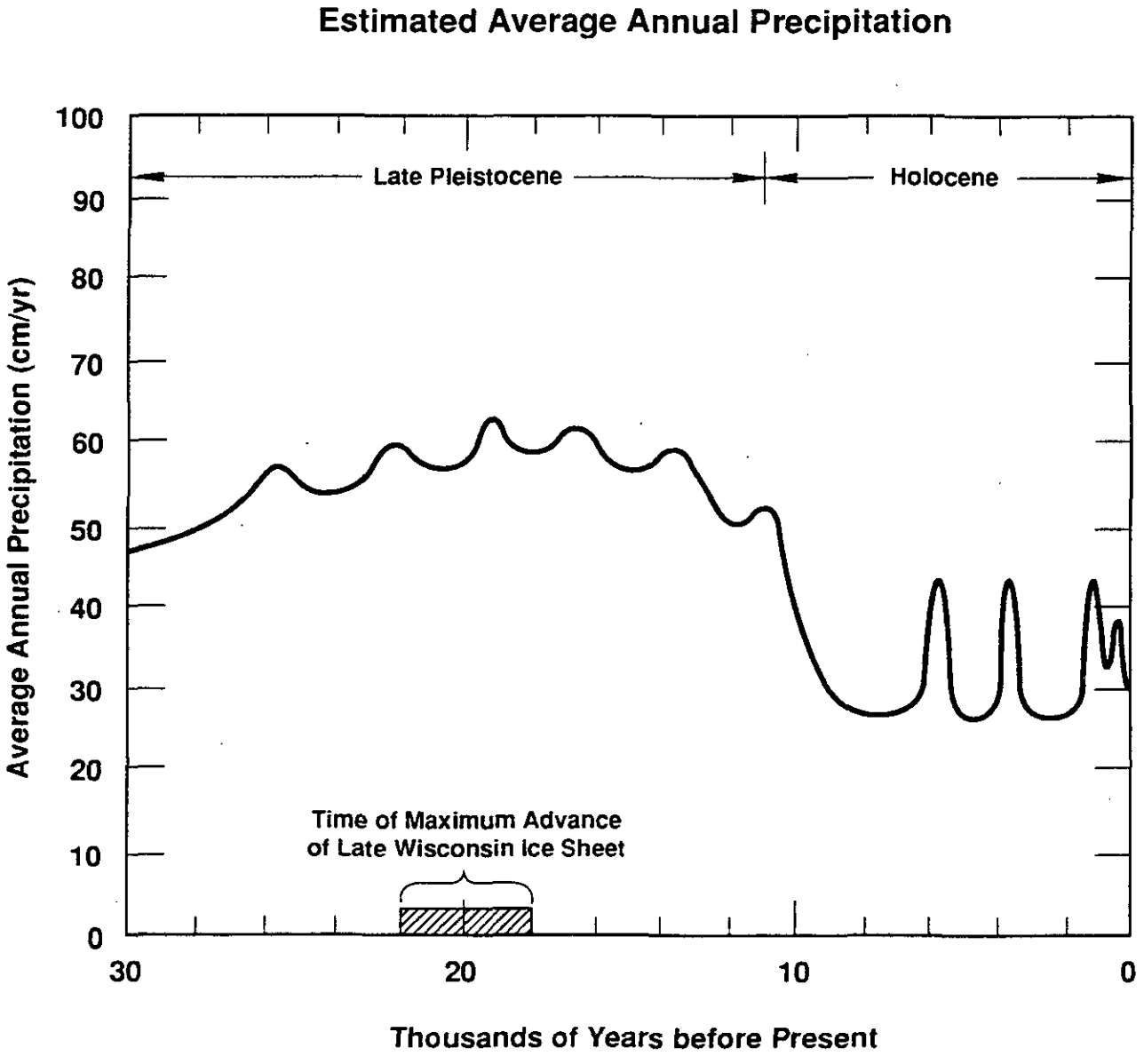

Figure 9. Estimated mean annual precipitation at the WIPP during the late Pleistocene and Holocene. Amplitudes of relatively high-frequency fluctuations are less well constrained than the amplitude of the shift from glacial conditions of the Pleistocene to interglacial conditions of the Holocene. Data from Van Devender and others (1987); Pierce (1987); Waters (1989); Allen (1991); Phillips and others (1992); and other sources cited in text.

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