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SITE CHARACTERIZATION PLAN

Chapter 5.0

CLIMATOLOGY AND METEOROLOGY

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Chapter 5

CLIMATOLOGY AND METEOROLOGY

5.0 INTRODUCTION

This chapter describes the recent climatology and meteorology of the Hanford Site and what is known about the past climates (paleoclimates) of the area. Meteorological information will be used in designing facilities and in planning operations at the repository during construction and loading. Paleoclimatic information will be used in conceptual modeling of past and future climates, and in estimating the future geologic stability of the site.

The Hanford Site is being considered as a repository for nuclear waste because it may provide a very long-term (100,000 yr) isolation for high-level radioactive waste. Such isolation comes from various barriers to the migration of nuclides during the most hazardous period. Principal among these are the geologic barriers. All components of the geologic isolation system should remain effective during the next 10,000 yr and must be evaluated for comparative purposes with respect to their behavior over the next 100,000 yr (DOE, 1984, p. 47757).

Groundwater flow is expected to be the major source for transport of radionuclides out of the repository; climatic changes could modify the amount of precipitation and runoff that are available as recharge to the groundwater system and that could reach the repository. Increases in precipitation may lead to increased recharge rates to the groundwater. Decreased temperatures also may increase recharge by lowering evaporation. Of course, it is also possible that precipitation could decrease or that temperatures could increase. To predict changes in the recharge rates, it will be necessary to understand possible climatic changes.

Purpose of the climatology and meteorology program

The overall purpose of the climatology and meteorology program is to establish an information base adequate to allow the following:

- The planning of safe and efficient repository construction and loading operations.
- An understanding of the various paleoclimatic factors (glaciation, changes in sea surface temperatures, atmospheric circulation, and temperature patterns) that have controlled the site's climatic evolution, and that will likely influence climatic change in the future.
- The development of reliable models to forecast long-term local and regional climatic change affecting the site.
- Most importantly, to predict groundwater recharge patterns at the site.

Sources of climatological and meteorological information

Detailed knowledge of the recent climate of the Hanford Site is based on local meteorological measurements. Since 1945, an extensive meteorological measurement program has been conducted at the Hanford Meteorology Station. The station is located within the boundaries of the controlled area study zone. Atmospheric temperature, wind direction, and wind speed are measured at several levels on the station's 122 m (400 ft) instrumented tower. Other meteorological parameters, such as atmospheric pressure, precipitation, humidity, sky cover, incoming solar radiation, and atmospheric stability, are measured at, or observed from, the station. From 1982 until the present time, additional data on area winds have been obtained from 21 sensors deployed throughout the Hanford Site and in the surrounding area.

Understanding of potential climatic change is based on consideration of the climatic changes that have occurred in the Pacific Northwest throughout the Quaternary Period (the last 1.67 m.y.). Significant changes (glacial cycles) have occurred over this time span. Climatic extremes in the Pasco Basin that accompany a glacial age are not well understood. Almost all available information has been obtained from palynological studies (i.e., the use of fossil plant pollen to reconstruct past environments). Only a single site east of the Cascade Range (Carp Lake) has yielded information about the climates of the last glaciation. Many sites that have been studied do not have associated radiometric age dates. Thus, the available data include items that are unlikely to yield the needed information for a license application.

Plans for obtaining additional climatological information

Of greatest concern in the barrier system is the possibility of groundwater transport of radionuclides. Under present conditions, the groundwater system appears to provide the isolation time desired. To ensure that this isolation time does not change, variations in controls on the groundwater system must not result in rates of radionuclide release above the mandated EPA limits (EPA, 1985). An important control on the groundwater system is the climate. The magnitude of changes in the groundwater system that could result from climatic variations over the next 100,000 yr remains unknown at this time (January, 1987). Such changes could be estimated by sensitivity analyses using groundwater models and, if such estimates were available, the paleoclimatic studies and future climatic modeling that are required could be more clearly defined. Unfortunately, this information is not available. Work required to provide that information is described in Section 8.3.1.3; however, it may be several years before the information is available. Therefore, the Basalt Waste Isolation Project (BWIP) will proceed with a conservative plan, assuming that the studies described in Section 8.3.1.5 will be required.

The extent of areas of potential recharge to the groundwater system under climatic regimes that could be operative in the next 100,000 yr is not currently known. To reflect a conservative approach, the BWIP will assume that any area underlain by basalts of the Columbia River Basalt Group is an area of potential recharge in the next 100,000 yr. Research during site characterization will be designed to more accurately identify potential recharge areas.

Additional work is required to characterize the effect of climatic changes upon repository stability. Planned studies will emphasize the coring of lake sediments to obtain fossil pollen data for quantitative paleoclimatic reconstructions. These studies will allow estimates of the variation in recharge rates that can be expected during the repository lifetime.

The information obtained from such paleoclimatic studies will be used in modeling to define a set of possible climatic scenarios for future climates at the site. These will define the boundary conditions of groundwater flow for modeling transport of radionuclides under changed conditions of recharge. Details and timing of such future climatic states will be derived from computer models of the local climate and global climatic change.

The conceptual models of such local climatic change include an expression of the external influences on global climate. The dominant influence is the variation in the orbit of the earth in a quasi-cyclic manner. These variations influence the quantity and distribution of solar radiation received by the earth, which in turn modifies the atmospheric and oceanic circulation patterns. Global climate can be further perturbed by other natural events (such as volcanism) and by anthropogenic influences. Local climate is a response to such global influences, which are further modified by factors such as the orographic effects of the Cascade Mountains.

The paleoclimatic reconstructions that are to be obtained will be used to refine the conceptual model of climatic change, which is described in this chapter. These reconstructions also will form a critical test of the ability of computer models to characterize climatic change adequately.

Uses of climatological and meteorological information

The Hanford Meteorology Station has established a comprehensive meteorological data base for the Hanford region. This data base details the recent climate of the region and can be used to predict the probability of meteorological extremes, aid in routine weather forecasts, model atmospheric dispersion, or simulate the impact of atmospheric processes in hydrological models. The results of these analyses can be used for such purposes as designing facilities, preparing operation plans, and predicting environmental impacts.

The most important use for paleoclimatological information about the site and surrounding region will be the incorporation of this information into a model that can represent hydrological cycles and predict future groundwater recharge rates and patterns at the repository. This model will consider atmospheric changes, glaciation patterns, oceanographic (sea surface temperature) changes, and regional topography.

Quality of climatological data and models

Presently available data on paleoclimates of the Columbia Plateau are mostly limited to the latest Quaternary and the Holocene Periods. These consist of sediment cores from lakes that have been used for qualitative pollen studies. Only one site from the Columbia Plateau (Carp Lake) has yielded a core with material dating from the last glacial maximum. To understand the spatial variability of climate during a glacial stage, and to allow extrapolation of the reconstructed climate, further cores are needed. These cores should be subjected to quantitative statistical analyses.

Although there is substantial data available that describes the glacial history of the Pacific Northwest during the Quaternary, the BWIP has not completed its evaluation of that data. Some information is available that helps to constrain the rate and timing of advance of the Cordilleran ice sheet in the Late Wisconsin. The dynamic behavior of that ice sheet and the effects of the ice sheet on the surface hydrologic system are less well known. Additional information is required to define the characteristics of Missoula floods in sufficient detail to allow tests of models of future floods. Careful study of the effects of Missoula floods on the groundwater system of the Pasco Basin is still needed. No information on this phenomenon is now available.

Very little information on the variations in sea surface temperatures of the northeast Pacific Ocean during the Quaternary is now available. Knowledge is limited because very few ocean cores from this area have been studied. Even the published data are largely limited to the time of the last glacial maximum (about 18,000 yr ago). At a minimum, additional currently available cores should be analyzed, and analyses of all cores should be extended to times other than the last glacial maximum. The end of the last interglacial (oxygen isotope stage 5e) is of great interest, because it is the closest analog to current climatic conditions.

Models of the local climatic variations of the Columbia Plateau at other times in the Quaternary are not currently in existence. The only model known to the BWIP that has been applied to the reconstruction of paleoclimates for a comparable area was created for application to the Nevada Nuclear Waste Storage Investigations and covered the drainage area of Death Valley, California. That model is a statistically based computer algorithm that computes monthly values of temperature, precipitation, evapotranspiration, and runoff. The runoff data form input to a model of the surface water system, which computes the resulting lake configuration. This model will be modified for application to the Columbia Plateau.

Needed improvements to the local climate model include computation of the local wind field as it is affected by topographic barriers and the presence of the Cordilleran ice sheet. The model must be thoroughly tested, and this will require development of new testing procedures not previously available with the existing model for the southwestern United States.

A model of ice sheet dynamics capable of describing the future behavior of the Cordilleran ice sheet in the next glacial age is not currently in existence. Dynamic ice sheet models are available, but they do not account for the influence of steep topography, such as occurs in the Pacific Northwest. An existing two-dimensional dynamic model of ice sheet behavior will be modified to represent the influence of topography on ice sheet dynamics.

Large-scale Missoula floods have been modeled with two-dimensional unsteady flow algorithms. Available solutions are limited to the region of the ice dam. These programs must be extended to allow solutions of flood dynamics within the Pasco Basin.

Global scale modeling involves three components: the atmosphere-ocean system, the cryosphere, and the external influences. Existing models of the global atmosphere system are three-dimensional prognostic primitive equation numerical solutions using spectral methods. These models have already been applied to paleoclimatic reconstructions and to future modeling of anthropogenic influences. Such models will be applied to the forecasting of future climates of the Pacific Northwest. Additional improvements are anticipated that will allow computation of the coupled behavior of the oceanic-atmospheric system at periods of extreme climatic changes. The models are limited in their direct applicability to forecasts of climates in the Columbia Plateau because of the coarse grid size necessitated by their computational load. For this reason, a local climate model is needed to transfer the global forecasts into locally meaningful estimates.

Solutions of the governing equations for hemispheric ice sheets are now available. These are two-dimensional unsteady flow models with the isostatic effects of the ice sheet incorporated and with the climatic boundary conditions parameterized with an orbital driver. Thermodynamic considerations are either parameterized or ignored. Such models, with slight improvements, will provide adequate forecasts of the behavior of future ice sheets during the next 100,000 yr.

The future climatic states of the earth for the next 100,000 yr can be estimated, but these estimates will include great uncertainty. To understand the uncertainty and to represent its effects on the various components of the climatic system will require considerable research effort. Such a modeling exercise has not previously been achieved and many of the techniques must be developed in this program.

Organization of chapter

The organization of this chapter can best be discussed from a chronological perspective. The first section discusses the recent meteorological characteristics of the site. The second section presents the past climatic record of the site and surrounding region, then explains methods by which these data can be used to forecast future climates affecting the repository. The third section summarizes these data, relates them to the repository design, and identifies additional information that will be needed to complete this study. Throughout these discussions, the method and models through which data have been and will be obtained and interpreted are presented.

The general climatic and meteorological phenomena of the Hanford Site discussed in Section 5.1 include winds, temperature, precipitation, sky cover and solar radiation, atmospheric pressure, humidity, severe storms, and atmospheric dispersion. Jet stream patterns and topographical features that affect the climate of the region are reviewed. Finally, the current program for collecting meteorological data at the site is presented.

Section 5.2 consists of discussions of (1) the paleoclimatology of the site and the Pacific Northwest region, (2) modeling approaches by which future climates at the site may be forecast, and (3) the program by which additional paleoclimatic data will be gathered and interpreted for site characterization.

Existing studies of the paleoclimatic system of the Pacific Northwest discussed in this chapter are grouped as follows:

- o Reconstructions of the paleoclimate of the Columbia Plateau (Section 5.2.1.1).
- o Reconstructions of the history of glaciation of the Cordilleran ice sheet and the Cascade Range (Section 5.2.1.2).
- o Reconstructions of oceanographic change during the Quaternary period that reflect variations of the sea surface temperatures of the northeast Pacific Ocean (Section 5.2.1.3).
- o Modeling future global climatic change that can influence the Pacific Northwest (Section 5.2.2.2).
- o Modeling future climatic change at the Hanford Site (Section 5.2.2.3).

Additional studies planned in these areas are described in Section 8.3.1.5.

In addition to summarizing what is currently known about the meteorology of the Hanford Site and the paleoclimatic record of the Columbia Basin, Section 5.3 identifies information still needed to forecast climatic changes adequately in the future and understand climatic changes that have occurred in the past. The relationship of both the existing data and data still to be gathered to the design of the repository is also discussed.

5.1 RECENT CLIMATE AND METEOROLOGY

In this section, the recent climate (1912 through 1980) of the Hanford Site region is characterized. A knowledge of average and extreme meteorological conditions is required for designing facilities and planning and scheduling site characterization activities, facility construction, and repository operations. Climatological data are also required for environmental impact studies; these data are used to model groundwater, surface water, and atmospheric processes such as aquifer recharge and the transport of radionuclides.

In Section 5.1.1, the general climate of the Hanford Site region and important influences on the climate are briefly discussed. In Section 5.1.2, specific climatological parameters are examined. These include wind direction and speed, atmospheric temperature, precipitation, sky cover, and humidity. Average and extreme values, based in some cases on up to 75 yr of measurements, are presented. Severe-weather phenomena and atmospheric dispersion are then discussed, and a joint frequency distribution of wind direction, wind speed, and atmospheric stability is presented. In Section 5.1.3, the network of instruments used to collect meteorological data for the Hanford Site region is discussed.

Much of the data presented in this section were obtained from Stone et al. (1983), the definitive report on the recent climate of the Hanford Site area. More detailed information on aspects of the climate at the Hanford Site can be obtained from that report.

5.1.1 CLIMATE

The Hanford Site is located in the Pasco Basin in south-central Washington State (Fig. 5.1-1). The climate of this area can be classified as midlatitude semiarid (DOE, 1982) or midlatitude desert, depending on the climatological classification scheme being used. Summers are warm and dry with abundant sunshine. Large diurnal temperature variations are common during this season, resulting from intense solar heating and nighttime cooling. Daytime high temperatures in June, July, and August can exceed 40 °C (104 °F). Winters, on the other hand, are cool with occasional precipitation. Outbreaks of cold air associated with modified arctic air masses can reach the area and cause temperatures to drop below -18 °C (0 °F). Overcast skies and fog occur periodically in this season.

Topographic features have a significant effect on temperature, wind, and precipitation (Berry, 1984). All air masses that reach the Pasco Basin undergo some modification, resulting from their passage over the complex topography of the Pacific Northwest (DOE, 1982). The climate of south-central Washington is strongly influenced by the Pacific Ocean and the Cascade Range to the west. The Rocky Mountains to the east and the north are also an important influence on the climate of the region. Locally the climate of the Hanford Site is influenced by the Yakima Ridge, Rattlesnake Hills, and Horse Heaven Hills to the west and south, and Saddle Mountains to the north (see Fig. 5.1-1). All air masses that reach the Hanford Site area undergo some modification resulting from their passage over this complex topography. The relatively low annual average rainfall (161 mm (6.3 in.)) at Hanford is, in large part, due to the rain shadow created by the Cascade Range. These mountains limit much of the maritime influence of the Pacific Ocean, resulting in a more continental climate than would exist if the mountains were not present. Maritime influences are experienced in the Hanford Site area during the passage of strong synoptic (large-scale) systems. Maritime air also penetrates into the region through gaps in the Cascade Range (such as the Columbia River Gorge). Continental influences are limited by the mountain ranges to the north and east of the Hanford Site. These mountains play a key role in protecting the region from the more severe winter storms and the extremely low temperatures associated with the modified arctic air masses that move southward through Canada.

In the cooler months, the position of the jet stream tends to direct storm systems into Washington State. These systems cause changes in cloudiness, wind velocity, and atmospheric pressure as they pass through the Hanford Site area. The majority of precipitation at Hanford is produced by these systems. The cold or occluded fronts associated with these storm systems have their origins in maritime polar, continental polar, or arctic air masses. Warm fronts occur as warmer maritime air flows over colder continental air. An average of 10 warm fronts and 52 cold fronts pass through the region each year (DOE, 1982), with the majority of these passages occurring during the cooler months. During the cool season, cold core lows in the upper troposphere periodically become centered southwest of the state; the resulting atmospheric circulation increases the frequency and intensity of precipitation in the region. Persistent high-pressure ridges in the upper troposphere also occur periodically in the cool season. The resulting circulation tends to trap cold air near the surface, resulting in extended periods of wintertime stagnation.

In the warmer months, the jet stream tends to direct most storm systems north of Washington State. Therefore, frontal passages are fewer and weaker. High pressure with stable, subsiding air is the dominant meteorological condition. Precipitation, when it occurs, tends to be associated with convective activity and the advection of moister maritime air into the region.

5.1.2 LOCAL AND REGIONAL METEOROLOGY

Meteorological parameters of interest include wind direction and speed, atmospheric temperature, precipitation, sky cover, solar radiation and visibility, and atmospheric pressure and moisture. These parameters have been measured at or near the Hanford Meteorology Station since 1945. The meteorology station is located within the controlled area study zone (Fig. 5.1-2), between the 200 West and 200 East operating areas. From 1912 to 1945, meteorological measurements were made at the old Hanford townsite, 16 km (10 mi) to the east-northeast of the Hanford Meteorology Station. Measurements at the townsite ended with the evacuation of the town and the establishment of the Hanford reservation during World War II. The Hanford Meteorology Station and other meteorological measurement facilities are described in Section 5.1.3.

5.1.2.1 Surface winds

The general surface airflow patterns in the vicinity of the controlled area study zone are significantly influenced by the local topography (Elliott et al., 1976). Winds at many locations in the western half of the Pasco Basin show a tendency to blow parallel to the northwest-southeast oriented ridgelines of the Rattlesnake Hills. Close to the hills, drainage flows periodically produce winds that are oriented perpendicular to the ridgeline. At the northern and eastern borders of the Hanford Site, winds show a tendency to follow the Columbia River Valley (east-west flow along the northern border and north-south flow along the eastern border of the Hanford Site). The average distribution of wind direction for a number of Hanford area locations is presented in Figure 5.1-3.

The prevailing near-surface wind (at 15.2 m (50 ft) above the ground) within the borders of the reference repository location has a strong westerly component. (Wind directions are always referred to by the direction from which the wind is coming.) The average wind speed within the reference repository location is 3.4 m/s (7.6 mi/h), as determined from measurements at 15.2 m (50 ft) above the ground at the Hanford Meteorology Station.

In the reference repository location, seasonal changes in the average wind direction are not very large, but seasonal changes in the average wind speed can be fairly significant. June has the highest average monthly wind speed (4.1 m/s (9.2 mi/h)), and the prevailing wind direction is from the west-northwest. In November and December, average wind speeds fall to a minimum of 2.7 m/s (6.0 mi/h), and the prevailing direction is from the northwest (Table 5.1-1).

Average diurnal changes (i.e., changes within a 24-h period) in both wind direction and speed can be large. These changes are especially significant during the summer months. In July, hourly average wind speeds range from a low of 2.5 m/s (5.6 mi/h) between 0900 and 1000 Pacific Standard Time (PST) to a high of over 5.8 m/s (13.0 mi/h) between 2100 and 2200 PST. In contrast, the corresponding wind speeds in January are 2.5 and 2.9 m/s (5.6 and 6.5 mi/h), respectively. Warm-season diurnal winds are typically associated with large-scale drainage circulations from the Cascade Range and local drainage winds from nearby terrain (Stone et al., 1983, pp. V3-V11).

In the Pasco Basin, high-speed, gusty winds are caused by frontal passages, squall lines, thunderstorms, and strong pressure gradients. In the winter, a chinook (warm and dry) wind from the southwest can result in a sudden large temperature rise, rapid melting of snow, and strong gusty winds. Spring and summer winds may have higher average speeds than fall and winter winds, but, unless reinforced by frontal activity or strong pressure gradients, they will not gust to very high speeds. June, the month with the highest average wind speed, has fewer instances of hourly average wind speeds exceeding 14 m/s (31 mi/h) than does December, the month with the lowest average wind speed.

The maximum recorded peak gust at 15.2 m (50 ft) above the ground at the Hanford Meteorology Station is 36 m/s (80 mi/h). (Table 5.1-1 contains information on the highest recorded peak gust during each month of the year.) Using 35 yr of wind data, a 100-yr return period peak gust of 38 m/s (86 mi/h) has been calculated. However, the elevation above the ground that is commonly used by engineers to determine the peak near-surface winds for design purposes is 9.1 m (30 ft) and not 15.2 m (50 ft). Because of friction generated between the wind and surface features, wind speed typically increases with increasing elevation above the ground in the height range being considered. This increase in wind speed can be approximated using Equation 5.1-1.

$$U_1 = U_2 \cdot \frac{\ln(Z_1/Z_0)}{\ln(Z_2/Z_0)} \qquad (5.1\text{-}1)$$

where

U_1 = wind speed at level 1

U_2 = wind speed at level 2

Z_1 = height above ground of level 1

Z_2 = height above ground of level 2

Z_0 = surface roughness, determined to be 0.03 m (0.1 ft) for the area around the Hanford Meteorology Station

Applying this equation, using the calculated value for the 100-yr return peak gust at 15.2 m (50 ft) for U_1, it is estimated that the 100-yr return peak gust at 9.1 m (30 ft) is 35 m/s (79 mi/hr).

Even during brief periods, such as on the order of one minute, wind speeds can fluctuate significantly. At weather stations, the sampling of wind velocity typically involves a measurement of 1- or 10-s duration. Longer term averages of wind speed are computed from a continuous sequence of these short duration measurements. Because of the large volume of data that are collected from this continuous measurement program, data storage is an important problem. To reduce the volume of data archived, only hourly averaged wind velocity and the peak speed (wind gust) measured during an hour are stored. Studies indicate that the maximum of the sixty 1-s speed measurements recorded during a minute is typically a factor of 1.3 larger than the average value of the wind speed for that minute (ERDA, 1975, pp. II-3 to II-21). This factor is reduced to a value of 1.1 if measurements of 10-s duration are considered rather than 1-s. Using this technique, the "fastest mile of wind"* at the Hanford Meteorology Station can be approximated from the maximum recorded wind gusts. The "fastest mile of wind" at 15.2 m (50 ft) above the ground at the Hanford Meteorology Station is estimated to be in the range of 30 to 35 m/s (67 to 78 mi/h).

Periods of relatively low wind speed are also part of the wind regime of the Hanford region. During the winter, approximately 28% of hourly average wind speeds are less than 1.8 m/s (4 mi/h) (Stone et al., 1983, pp. X-27 to X-30). During the entire year, wind speeds are less than 1.8 m/s (4 mi/h) only 19% of the time. Episodes of fog are typically associated with low wind speeds; wind speeds are in excess of 1.8 m/s (4 mi/h) during only 27% of the hours in which fog is present.

5.1.2.2 Elevated winds

The frequency distribution of wind direction at heights of 61 and 122 m (200 and 400 ft) on the Hanford Meteorology Station meteorological tower is fairly similar to the distribution at 15.2 m (50 ft) (see Stone et al., 1983, pp. X-34 to X-54). With increasing elevation, a slightly higher percentage of

*The term "fastest mile of wind" is based on the old measurement technique of measuring how long it took for an anemometer to repeatedly be spun about its axis by the wind to simulate 1 mi of travel in the wind field. Today, this term generally refers to the highest recorded wind speed averaged over a 1-min period.

winds with a northwesterly component is seen. While the average difference in the annual distribution of wind direction with height may be small, major differences between instantaneous measurements do occur.

The wind speed relationship between selected levels from the Hanford Meteorology Station meteorological tower are presented in Table 5.1-2. The values are based on 35 yr of wind data.

Strong seasonal and diurnal variations in wind speed occur at all elevations. As is the case near the surface, diurnal variation in wind speed is greatest in the summer and is very small during the winter months. Diurnal variation also appears to increase with increasing height between 15.2 and 122 m (50 and 400 ft) (Stone et al., 1983, pp. X-9 to X-12).

5.1.2.3 Temperature

The mean surface air temperature averages approximately 11.7 °C (53 °F) at the Hanford Meteorology Station. In the period between 1912 and 1980, the mean annual temperature varied between 10.1 and 13.4 °C (50.2 and 56.1 °F). Annual and monthly averages and extremes of temperature are presented in Table 5.1-3. Normal and extreme daily temperatures are plotted in Figure 5.1-4. Facility design and planning engineering weather data are presented in Table 5.1-4.

July tends to be the warmest month of the year, with temperatures averaging 24.7 °C (76.5 °F). The average daily minimum and maximum temperatures for July are 16.2 and 33.3 °C (61.2 and 91.9 °F), respectively. The highest temperature ever recorded at the Hanford Site was 46.1 °C (115 °F) on July 27, 1939. The highest temperatures during each summer between 1912 and 1980 and the probability that the highest temperature during a year will not exceed a given temperature are presented in Figure 5.1-5.

January tends to be the coolest month of the year with temperatures averaging -1.4 °C (29.5 °F). The average daily minimum and maximum temperatures for January are -5.5 and 2.7 °C (22.1 and 36.9 °F), respectively. The lowest temperature ever recorded at the Hanford Site was -32.8 °C (-27 °F) on December 12, 1919. The lowest temperature ever recorded in January was -30.6 °C (-23 °F). The lowest temperatures during each winter between 1912 and 1980 and the probability that the lowest temperature during a year will be lower than a given temperature are presented in Figure 5.1-6.

Notable periods of very cold temperatures, each lasting from 14 to 22 d, occurred in 1919, 1957, and 1959. These three periods account for 7 of the 10 recorded days during which the maximum temperature was below -18 °C (0 °F) and 9 of the 16 recorded days for which the minimum temperature was below -29 °C (-20 °F). Three other notably cold periods (1916, 1930, and 1937) account for the remaining 7 d in which the minimum temperature did not exceed -29 °C (-20 °F).

An average of 174 d/yr at the Hanford Meteorology Station are free of freezing temperatures, with the recorded range lying between 142 and 215 d. The average last day of frost is April 23, with the earliest day being March 19 and the latest being May 16. The average first day of frost is October 15, with the earliest day being September 11 and the latest being November 12.

5.1.2.4 Precipitation

The mean annual precipitation at the Hanford Meteorology Station is approximately 161 mm (6.3 in.). Detailed information on averages and extremes of precipitation are given in Table 5.1-5. Over a period of roughly 80 yr, the annual precipitation has varied from a low of 76 mm (3 in.) to a high of 291 mm (11.5 in.). Annual precipitation totals and the probability that precipitation totals will not exceed a given amount are presented in Figure 5.1-7. On average, approximately 42% of the annual precipitation falls during November, December, and January. January is the wettest month, with an average of nearly 100 h of precipitation producing just over 23 mm (0.9 in.) of water. Only 11% of the annual precipitation falls during July, August, and September. July is the driest month, with an average of only 10 h of precipitation producing less than 4 mm (0.15 in.) of water. Even though precipitation is less frequent during the summer months, summer rainfall, when it does occur, is on average twice as intense as winter precipitation (Stone et al., 1983, pp. IV-8).

The maximum number of consecutive days without precipitation is 102 d, recorded in the summer of 1919. Extended dry periods are not restricted to the summer. In 1976, less than 4 mm (0.15 in.) of precipitation fell during the 128 d from August 26 through the end of the year. Even during notable "wet" periods at the Hanford Site, precipitation totals are low compared with the amounts routinely encountered west of the Cascade Range. One notably wet period for the Hanford Site started on the last day of October 1973. During this period, there was precipitation on 31 of 37 d, with a total accumulation of over 80 mm (3.15 in.) of water.

Total annual snowfall, which includes all frozen precipitation, has varied from a low of 8 mm (0.31 in.) to a high of more than 1,100 mm (43.3 in.). The average annual snowfall is 336 mm (13.2 in.). Snowfall accounts for approximately 38% of all precipitation (by water content) from December through February. The largest depth of snow on the ground at any one time was 622 mm (24.5 in.), recorded in February 1916. This depth of snow is quite unusual; during no other winter has the snow depth at the Hanford Site exceeded even half of the record accumulation. The probability that the greatest depth of snow on the ground will exceed various potential values once during a winter can be obtained from Figure 5.1-8. On average, the depth of snow on the ground will exceed 150 mm (5.9 in.) in about only one winter out of eight.

Hail is precipitation in the form of balls or lumps of ice that is typically formed within cumulonimbus clouds during thunderstorms. Hail is an unusual phenomenon in the Pasco Basin; it has never been recorded at the Hanford Meteorology Station more than 2 d during any year of record. Hailstones that have fallen near the Hanford Meteorology Station have tended to be rather small (with a maximum diameter in the 5- to 8-mm (0.2- to 0.3-in.) range). The largest hailstone measured at the Hanford Meteorology Station had a diameter of 10 mm (0.4 in.), which is small compared with the massive hailstones observed in other parts of the United States. Glaze ice is a coating of ice formed when rain or drizzle freezes on contact with surfaces that have a temperature below freezing. Glaze ice occurs an average of six times a year at the Hanford Site, always between November and March. Rime ice is formed when supercooled droplets freeze on contact with solid objects. Rime formation in the Hanford Site area is generally associated with supercooled fog either at higher elevations in the nearby hills or along the banks of the Columbia River.

Evapotranspiration is not measured at the Hanford Meteorology Station. The closest sites at which evaporation data are collected are two locations at Washington State University's Irrigated Agriculture Research and Extension Center near Prosser, Washington. The center is approximately 39 km (24 mi) to the southwest of the Hanford Meteorology Station. Evaporation measurements at or near the center have been conducted daily between April 1 and November 1 since 1924. Data are available from Washington State University in the form of monthly summaries or floppy disks.

Evapotranspiration processes can play a key role in groundwater recharge. Evapotranspiration levels can vary greatly with location at Hanford due to differences in soil properties and the type and density of local vegetation. Some evapotranspiration measurements have been made at Hanford, but a detailed study has not been made in the controlled area study zone or at locations with similar vegetation and soil characteristics. Wallace (1977, p. 19) used Hanford Meteorology Station data to compute an average potential evapotranspiration figure for the Hanford Site using three different methods: the Penman, Thornthwaite-Maither, and Mortion methods. The potential evapotranspiration levels calculated by Wallace were from five to nine times the mean annual precipitation. To gauge the actual evapotranspiration levels within the controlled area study zone, hydrologists are currently planning a measurement program. Preliminary work has indicated that during significant precipitation events, water can move below the vegetation root zone and escape evapotranspiration processes (Gee and Kirkham, 1984). This indicates that although potential evapotranspiration is estimated to be several times larger than annual precipitation, local precipitation is still contributing to groundwater recharge.

5.1.2.5 Sky cover and solar radiation

Sky cover is recorded at the Hanford Meteorology Station hourly between sunrise and sunset and is determined by estimating the number of tenths of sky that are obscured by clouds or other phenomena. The average daily sky cover is computed by averaging hourly values. On average, 192 d/yr can be categorized as sunny or mostly sunny (Stone et al., 1983, pp. VI-1 to VI-2). Cloudy conditions tend to dominate during the winter and sunny days during the summer.

Incoming solar radiation averages about 367 langleys per day. Incoming solar radiation is at a maximum not in June, when the days are longest, but in July, when the average cloud cover is somewhat less (Stone et al., 1983, pp. VI-3 to VI-6). In July, incoming solar radiation averages 647 langleys per day. The darkest days of the year are in December; during this month, incoming solar radiation averages only 115 langleys per day. On average, incoming solar radiation is 70% to 80% of its potential value in the summer and only 40% to 50% of its potential value in the winter.

Visibility at the Hanford Meteorology Station is reduced to less than 10 km (6.2 mi) an average of just under 4% of the time. Fog accounts for just over three quarters of these occurrences. Precipitation and blowing dust account for nearly all the other periods of low visibility.

5.1.2.6 Atmospheric pressure, moisture, and mixing height

The height above sea level at the Hanford Meteorology Station is approximately 220 m (720 ft). The average atmospheric pressure, corrected to sea level, is 981.3 mbars (29.21 in. of mercury). In general, the atmospheric pressure is higher in the winter than in the summer, although both the highest and lowest recorded pressures at the Hanford Site occurred during the winter. Averages and extremes of atmospheric pressure are presented in Stone et al. (1983, pp. VIII-1 to VIII-2).

The annual mean relative humidity within the reference repository location is approximately 54%, with the highest relative humidity (80%) occurring in December and the lowest (32%) in July. The annual mean dew point is 1 °C (34 °F), and the annual mean wet-bulb temperature is 6.6 °C (44 °F). Facility design and planning engineering weather data are presented in Table 5.1-4. More detailed information on atmospheric moisture is available in Stone et al. (1983, pp. VII-1 to VII-13).

The mixing height is defined as the height above the ground that corresponds to the upper boundary of the well-mixed surface layer of the atmosphere. Under an unstable atmospheric stratification, the mixing height may be on the order of several thousand meters above ground level. Unfortunately, the 122-m (400-ft) tower and the acoustic sounder at the Hanford Meteorology Station do not probe high enough into the atmosphere to accurately measure all mixing heights. When the mixing height exceeds instrument measuring capabilities, data collected during pilot balloon (noninstrumented balloon) releases from the Hanford Meteorology Station and regional radiosondes (instrument packages carried aloft by balloons that are launched twice a day from selected National Weather Service stations) are used to supplement the lower-level Hanford Site area measurements. The frequency of occurrence of mixing heights at the Hanford Site within various ranges is given in Table 5.1-6.

5.1.2.7 Severe weather phenomena

A study of severe weather phenomena typically focuses on such events as hurricanes and thunderstorms. At the Hanford Site, the severe weather phenomenon that occurs most frequently and has the greatest impact on local activities is duststorms.

Tornadoes are very rare in the vicinity of the Pasco Basin. On average, the State of Washington experiences fewer than one tornado each year. Tornadoes and funnel clouds have been observed three times from the Hanford Site since 1961 (Stone et al., 1983, pp. IX-6 to IX-7). During this period, 23 tornadoes or funnel clouds were reported within a 161-km (100-mi) radius of the Hanford Meteorology Station. The nearest reports of tornado damage were at Yakima, Washington (April 30, 1956), approximately 72-km (44-mi) west of the reference repository location, and at Wallula Junction, Washington (June 26, 1958), approximately 81-km (50-mi) southeast of the reference repository location. However, no loss of life or extensive property damage has ever been reported from the tornadoes observed in this region.

The Pasco Basin is too far north and inland for hurricanes or active tropical storms to reach the area. Moisture and cloudiness from the remnants of tropical storms can be advected into the Hanford Site region under unusual circumstances. Rain or thunderstorms can be produced by these systems.

The Pasco Basin is not a major thunderstorm area. On average, only about 10 thunderstorm days per year are recorded at the Hanford Meteorology Station, although this number has varied from a low of 3 to a high of 23 thunderstorm days per year. Thunderstorms can theoretically occur during any month of the year, although none have yet been observed in either November or January. They occur most frequently from April through September. Data on the number of monthly and annual thunderstorm days are presented in Stone et al. (1983, pp. IX-1 to IX-2). The largest number of thunderstorm days recorded in a single month, 8 d, has occurred in both June and August.

Two main categories of thunderstorms are the "air mass" thunderstorm and the "squall line" thunderstorm. Both types of thunderstorms occur under warm, moist, and unstable conditions; however, the more intense squall line thunderstorms are associated with organized frontal systems. Thunderstorms are classified as severe when wind gusts exceed 25 m/s (56 mi/h) and (or) hail size equals or exceeds 19 mm (0.75 in.) in diameter. Such severe storms are exceedingly rare in the Hanford Site region.

Large differences in electric potential can occur during thunderstorms. Lightning occurs when the difference in electric potential increases to a point where the atmosphere can no longer sustain it. In general, less than 25% of lightning strikes are cloud-to-ground discharges; most are cloud-to-cloud discharges. Lightning strikes in the summer have occasionally ignited grass fires that have burned thousands of acres in the Hanford Site region. These fires play an important role in the ecological cycle of the desert. Towers and other elevated structures have increased probabilities of being hit by lightning. The 122-m (400-ft) meteorological tower at the Hanford Meteorology Station has been struck several times during its 40-yr existence. Lightning-strike frequency can be estimated from climatological data on thunderstorms (Cianos and Pierce, 1972, pp. 23-24).

Duststorms or sandstorms occur periodically in the Pasco Basin. They are recorded at the Hanford Meteorology Station when visibility is restricted to under 10 km (6.2 mi) by either dust or blowing dust. Dust can be carried into the area from a distant source and may occur with very little wind (as low as 1.8 m/s (4 mi/h)). Blowing dust occurs when strong winds (higher than 8 m/s (18 mi/h)) pick up local material. Blowing dust is the more common of the two phenomena at the Hanford Site.

Studies of duststorms at the Hanford Site were conducted in the 1970s by Orgill et al. (1974, pp. 214-219), Sehmel (1976, pp. 99-101), and Jenne (1978, pp. 1-136). These studies indicate that an average of eight duststorms a year at the Hanford Meteorology Station decrease visibility to below 10 km (6.2 mi). These duststorms last an average of just over 3 h but have lasted as long as 18 h. The sand and dry soil of the Pasco Basin, local construction activities, and the surrounding agricultural land (for dryland wheat and irrigated crops) are all sources of airborne dust for the Hanford Site region. During severe duststorms, one of the primary sources of airborne dust is the dryland wheatfields in the Horse Heaven Hills south of the Hanford Site.

Duststorms occur most frequently from March through May and also in September. The spring maximum occurs while the soil is disturbed by spring agricultural activity and gusty winds. September duststorms appear to be related to ground breaking for winter wheat in the east-central part of the state (north and northeast of the Hanford Site). Duststorms are least frequent in August, November, and December (Jenne, 1978, p. 8).

Dust devils, which transport small amounts of dust, are rotating columns of air, typically about 10 m (33 ft) in diameter, that are made visible by the dust suspended in them. They are similar in appearance to waterspouts or tornadoes but are much smaller and weaker. Dust devils occur frequently at the Hanford Site on sunny days with light winds and seldom last for more than a few minutes. Characteristics of dust devils and information on their occurrence at the Hanford Site are discussed in further detail in Orgill and Schwendiman (1975, pp. 1-2).

Significant increases in the ambient concentration of pollutants can occur during stagnation periods. Maximum background concentrations for many pollutants are recorded during such periods. Stagnation conditions occur when low wind speeds and shallow mixing layers (low-level inversions) persist for extended periods of time. Significant stagnation episodes occur only in the winter when incoming solar radiation is of insufficient strength to break up low-level inversions. In a study by Jenne (1963, p. 29), a stagnation episode was considered to be an uninterrupted period when the daily average wind speeds (for one or more days) were 1.8 m/s (4 mi/h) or less and (or) peak gusts were 7 m/s (16 mi/h) or less. In this study, Jenne examined 15 yr of data for the months November through February. Jenne concluded that stagnation periods lasting for 20 d can be expected only 1 season in 20, but a 10-d stagnation period can be expected every other season. On the average, only one season in three will not have a stagnation period of at least 8 d.

5.1.2.8 Atmospheric dispersion

Evaluations of the atmospheric dispersion of air pollutant releases from the controlled area study zone are based primarily on measurements and observations made at the Hanford Meteorology Station. In the controlled area study zone, conditions favorable for dispersion occur a majority of the time. Extremely and moderately stable atmospheric conditions exist only 18.4% of the time. These conditions are coupled with wind speeds of less than 1.5 m/s (3.4 mi/h) approximately 3% of the time (Table 5.1-7). Although 3% is not a large percentage of the time, it still corresponds to a significant number of hours during the year when low wind speed and very stable atmospheric conditions produce poor dispersion conditions.

Preliminary estimates of long-term air quality impacts can be computed using a simple computer model. For this and other applications, the use of several years of hourly data for meteorological input involves a level of complexity that is often unwarranted. Joint frequency distributions of meteorological conditions are often used to provide meteorological input without processing extensive data files. These distributions present the frequency of occurrence of specific combinations of meteorological conditions.

To illustrate a joint frequency distribution, consider three key atmospheric dispersion parameters that are recorded for the Hanford Site: atmospheric stability, wind direction, and wind speed. Atmospheric stability is often categorized using the 7 Pasquill-Gifford classes, wind direction is frequently classified using 16 direction sectors (N, NNE, NE, ENE, etc.), and wind speed can be indicated using 8 speed classes. There are 784 different atmospheric dispersion situations that can arise when all of the possible combinations of these parameters (7 by 16 by 8) are considered. The specific frequency of each atmospheric dispersion situation can be determined from the meteorological data base. Using data on the frequency of occurrence of each specific dispersion situation (a joint frequency distribution), a computer model can determine the long-term air quality impact of a chronic pollutant release without processing hourly data.

A joint frequency distribution of wind speed, direction, and stability for the Hanford Meteorology Station is presented in Table 5.1-7. Additional joint frequency distributions for the Hanford Site, involving the consideration of time of year and using wind data from different measurement heights, are presented in Stone et al. (1983, pp. X-76 to X-91).

The U.S. Nuclear Regulatory Commission (NRC) guidelines allow some latitude as to the specific dispersion models that can be used for official dispersion analyses. Of NRC-developed and NRC-approved models (NRC, 1977, pp. 1.111-7; 1979, p. 1.145-2 for a full listing of these models), the MESOI model (Ramsdell et al., 1983) is specifically designed to use both Pasco Basin terrain data and data from all the available wind measurement locations to simulate the spatial and temporal variation in atmospheric dispersion.

In addition to the models developed for the NRC, other atmospheric dispersion and dose models have been developed over the years for use at the Hanford Site. These models are summarized in documents by Strenge et al. (1976, pp. 1-10), Napier (1981, pp. 1-34), and Strenge and Peloquin (1981, pp. 1.1-3.1).

The Hanford Site is one of the few sites in the country where extensive experiments on dispersion processes have been conducted. Several hundred diffusion experiments using various atmospheric tracers (e.g., fluorescent particles and rare gases) were conducted at the Hanford Site from the mid-1950s through 1985 (Ramsdell et al., 1985). A significant number of these experiments also examined dry deposition. Although wet deposition is a significant removal mechanism at many locations, dry deposition plays a more important role in the semiarid climate of the Hanford Site region.

5.1.3 SITE METEOROLOGICAL MEASUREMENT PROGRAM

The collection of meteorological data in the Hanford Site area began in 1912. From 1912 to 1945, cooperative observers for the U.S. Weather Bureau collected meteorological data at the Hanford townsite. In 1945, the location of meteorological measurements was moved 16-km (10-mi) west-southwest to the Hanford Meteorology Station.

The mission of the Hanford meteorology program is to meet the meteorological and climatological needs of the DOE and the Hanford Site contractors. In particular, the meteorological program calls for the measurement, observation, and recording of various climatological data; continuous monitoring of regional weather; forecasting of weather for site operations; and collection of data for environmental reports, atmospheric dispersion models, and emergency response purposes. In fulfilling their mission, the staff of the Hanford Meteorology Station have amassed a data base that characterizes in detail the recent climate of the Hanford region. The current data base and monitoring program should fulfill the requirements of the BWIP for recent climatological data. Significant additions or modifications to the existing meteorological monitoring program should not be required.

A wide range of meteorological variables is measured at the Hanford Meteorology Station and approximately 500-m (1,640-ft) east on a 122-m (400-ft) tower. Parameters measured near the surface include temperature, humidity, atmospheric pressure, precipitation, visibility, incoming solar radiation, and subsoil temperatures. On the tower, measurements of wind direction, wind speed, and air temperature are made at elevations of 9.2, 15.2, 30.5, 45.7, 61.0, 76.2, 91.5, 107.0, and 122.0 m (30, 50, 100, 150, 200, 250, 300, 350, and 400 ft) above the ground. Additional temperature measurements are made at heights of 0.9 and 6.1 m (3 and 20 ft), and additional wind measurements are made at a height of 2.1 m (6.9 ft).

At the Hanford Meteorology Station, upper-level wind measurements have been made for more than 25 yr using the single-theodolite pilot-balloon method. Data are currently taken twice daily at 1200 and 2400 PST. Some of these observations have been summarized for climatological and case study purposes. Measurements in the middle and upper atmosphere using a rawinsonde-type system are not conducted routinely, although the capability to conduct such measurements does exist.

A monostatic acoustic sounder was operated at the Hanford Meteorology Station from 1975 through mid-1985. Mixing-layer depths, the formation and breakup of temperature inversions, and characteristics of low-altitude gravity waves are some of the physical phenomena that could be monitored by this instrument system. In 1985, the system was replaced with a more advanced doppler acoustic sounder. This device provides a more accurate estimate of the temperature and turbulence structure of the atmosphere. It also can estimate wind velocities up to a height of more than 400 m (1,300 ft) above the surface.

In addition to the Hanford Meteorology Station data, near-surface (9.1-m (30-ft)) wind measurements are made at 21 locations on the Hanford Site and in the surrounding communities (Fig. 5.1-9). This network of wind measurement stations provides a detailed assessment of atmospheric transport in the Hanford Site vicinity. Three of the wind measurement stations are within the controlled area study zone. Data from all 21 stations are available in real time at the Hanford Meteorology Station.

In 1987, the Hanford meteorological monitoring network is scheduled to expand with the deployment of three 61-m (200-ft) towers and four acoustic sounders at additional locations on the Hanford Site. Additional wind measurement stations will also be located offsite. These additions will raise the number of wind measurement sites within 65 km (40 mi) of the Hanford Site to more than 45. With the addition of these stations, sufficient data on onsite and offsite winds should be available to more than adequately model the transport of effluents released during repository-related activities.

The data collected at the Hanford Site since 1955 have been archived on magnetic tape. The data are periodically analyzed and published as a report. The most recent version of the report is by Stone et al. (1983); previous versions are Stone et al. (1972) and Jenne and Kerns (1959). As mentioned earlier, much of the climatological analysis presented in this chapter is excerpted from Stone et al. (1983).

All instruments on the Hanford Meteorology Station tower were replaced in 1982 with equipment meeting NRC guidelines (NRC, 1972, pp. 23.1-23.6). Instrumentation is calibrated at least once a year.

5.2 LONG-TERM CLIMATIC ASSESSMENT

This section reviews the current knowledge of past climate change that affects the Hanford Site and discusses models by which past and future climate change at the site can best be understood and predicted. A knowledge of past climate change is necessary, because the Earth's current and future climate must be viewed in the context of the sequence of glacial cycles that characterize the Quaternary. Such glacial cycles typically last millions of years. The current major control on the continuation of such glaciation is the variation of the Earth's orbit. The projections of future orbital characteristics and the record of great length of glacial cycles suggest that these cycles will continue for the next 100,000 yr.

Glaciation, atmosphere, and oceans will largely define the climate change expected at the Hanford Site over the next 100,000 yr. Therefore, as each factor is reviewed, the discussion also will indicate the importance of each in controlling climatic change at the Hanford Site. Further, the study indicates gaps in our understanding of the mechanisms of climatic change and suggests remedies. This discussion sets the stage for the description of additional studies found in Section 8.3.1.5.

There are problems with these extrapolations to forecast the future climates of the Earth (Section 5.2.2). The orbit of the Earth will not vary in the next 100,000 yr in a manner totally analogous to the past record (Berger, 1978a). Other factors influence in important ways the climatic state of the Earth. The existing total ice volume is one such control, as is the chemical composition of the atmosphere (especially for gasses, such as CO_2) (Broeker, 1985). It is likely that many such factors will differ in the next 100,000 yr, so that the record of the Quaternary will not provide a close estimate of all future climates. These problems are discussed in more detail in this section. The plans for accounting for each of these factors in forecasting future climates are described in Section 8.3.1.5.

5.2.1 PALEOCLIMATOLOGY

Section 5.2.1.1 summarizes the current understanding of climate change as it affects the Hanford Site, and describes the record of climatic change in the Columbia Basin. Given first are the available reconstructions of climate over the past few hundred years using dendroclimatic procedures. These will form the basis of a definition of a "standard modern climate."

Fluctuations of climate in the Pacific Northwest throughout the Holocene and late Pleistocene are inferred from the fossil-pollen records. Such pollen records provide firm evidence that greater climatic change is possible over a 10,000-yr period than is represented in the historic record.

More extreme climate changes occurred during the last advance of continental glaciers into northern Washington. This advance began approximately 30,000 yr B.P., with the Cordilleran ice sheet reaching its maximum extent approximately 15,000 yr B.P. Few data are available to describe that climate extreme. The single pollen record extending back to the beginning of the last glaciation is described.

Fluctuations in earlier Quaternary climates in the Pacific Northwest are discussed next. The record is scanty and must be supplemented by observations from other parts of the globe. This record suggests that glaciations as extreme as the most recent one are typical of the entire Quaternary and can be expected to continue throughout the containment period of radioactive waste within a repository.

A brief discussion of very long-term (millions of years) trends in global climate change is also given. Although pre-Quaternary climates cannot be used as direct guides to expected climates in the next 100,000 yr, they suggest the maximum variability of such future climate.

Three major components of the climate system combine to produce the changes that have been observed throughout the Quaternary: glaciers, oceans, and atmosphere. Each of these components is discussed separately.

Section 5.2.1.2 summarizes glacial history. Over 100,000-yr time spans, change in the volume and configuration of ice on the continents forces modifications in the oceanic and atmospheric circulation. Thus, an understanding of the potential for glaciation is needed to forecast climate change.

Section 5.2.1.3 presents variations in the oceanic system, especially in the northeast Pacific Ocean. Sea surface temperatures directly control the supply of precipitable water to the atmosphere. Specific patterns of sea surface temperatures arise from oceanic currents. Such currents are a major mechanism for the redistribution of the energy balance of the Earth. That energy balance, patterns of its redistribution, and sea surface temperatures all change through time, producing important climatic changes. The atmosphere responds to these changes in the ocean.

Section 5.2.1.4 presents available evidence about the character of the atmosphere at various points in the glacial-interglacial cycles. The atmosphere directly produces the weather and climate of an area. The history of the atmosphere and its response to changes in the other components of the climate system (glaciers and oceans) is only poorly preserved. Our understanding of changes in atmospheric behavior comes mostly from modeling experiments. Salient features of those experiments and the resultant understanding of long-term atmospheric dynamics are presented.

5.2.1.1 Climatic change

Future climates of the Pasco Basin can influence the groundwater system of the repository; therefore, these future climates must be estimated. A fundamental guide to the future climates of the Pasco Basin is provided by an understanding of the past climates of the area. This section reviews the evidence about climates and climate change in the Pacific Northwest. There are three time spans that will be discussed: the last 400 yr, the late Quaternary (0.125 m.y.B.P.), and the Quaternary as a whole (1.67 m.y.B.P.).

Section 5.2.1.1.1 summarizes geologic evidence concerning the stability of climates in the last few centuries. This evidence is used to define the extent to which the modern instrumental records can be considered representative of the future climate of the area. Most of the evidence presented in this discussion is derived from tree-ring records. Section 5.2.1.1.2 describes the existing record of the late Quaternary climate of the region. This discussion, covering the past 10,000 yr, relies heavily on palynological evidence. This evidence provides the basis for forecasts of the extent of climate change in the future. Section 5.2.1.1.3 provides an overview of the climates that have been observed throughout the Quaternary and earlier times. This long-term perspective gives a framework within which the current fluctuations are more readily interpreted and indicates the high probability of future climate change.

5.2.1.1.1 Modern climatic record

Understanding the groundwater system of the repository site is essential to site characterization. If our ideas about modern precipitation patterns are not firmly established, the resulting groundwater models may be unreliable. Thus, we must understand the modern climate to understand the groundwater system.

The extent of sensitivity by the groundwater system to climatic changes in unknown. Even under modern climatic conditions, the extent to which interannual variations in recharge could affect groundwater flow at the horizon of the proposed repository is not known. It is possible that no significant changes in the groundwater system could be considered <u>anticipated</u> (or even unanticipated) under the magnitude of climatic variability that is representative of modern climate. Whether or not this is the case should be considered during site characterization, and one way of doing that is by sensitivity analysis. Data needed for such a sensitivity analysis would include an estimate of the variability of modern climate at a sufficient resolution to identify unanticipated events. Such information is expected to be available from the dendroclimatic studies described here and in Section 8.3.1.5.

Records from climate stations are too short to provide a representative picture of modern climate. To estimate the climate to be observed over a 30-yr period requires extrapolating from a record 10 times that length. The most useful means available to reconstruct climate over the last several centuries is through the study of tree-ring widths. Appropriate calibration techniques allow quantitative climate reconstructions; this is called dendroclimatology.

High-quality tree-ring records sensitive to climatic variations can provide climatic information on time scales as short as one season to as long as a century. Reliability of such dendroclimatic data is maintained by sampling high-quality tree-ring materials (LaMarche et al., 1982), dating and processing them carefully, and subjecting only the best materials to rigorous dendroclimatic analysis (Fritts, 1976; Brubaker and Cook, 1983).

A grid of 65 western North American tree-ring chronologies (Fig. 5.2-1) from semiarid sites has been calibrated with climatic data. The calibration has been used to reconstruct spatial variations of past temperature, precipitation, and atmospheric pressure over a large sector of North America and the north Pacific Ocean (Blasing and Fritts, 1976; Fritts et al., 1971, 1979). Similar tree-ring records are used to reconstruct climate in other areas of the world (e.g., Fritts, 1982; Briffa et al., 1983).

Special collecting, measuring, and analyzing techniques remove the nonclimatic variations from the climatic information in tree-ring-width data (Fritts, 1976; Graybill et al., 1982). Such techniques were applied in the selection of the 65 tree-ring chronologies from western North America. The chronologies were selected on the basis of the greatest number of trees sampled, the statistical characteristics of the data, the longest records, and the spatial distribution of the sampled tree sites (Fritts and Shatz, 1975). All chronologies spanned the 1600 to 1963 period, but were best replicated after 1700 and had a geographic coverage extending from the north Pacific coastal states to the Black Hills of South Dakota and from the Canadian Rocky Mountains to Durango, Mexico. This range of geographic coverage can be seen in Figure 5.2-1.

The density of the 65, tree-ring chronology data set is approximately seven sites per million square kilometers (Cropper and Fritts, 1982), nearly three times the "high-density" network considered by Kutzbach and Guetter (1984a) to be adequate for evaluating large-scale spatial variations in climate. Large-scale atmospheric circulation anomalies over the north Pacific Ocean and North America tend to move from west to east (Bryson and Hare, 1974) beyond the margins of the calibration grid used. Thus, the results of Kutzbach and Guetter (1980) support the use of the larger tree-ring data set for paleoclimate reconstruction for the general region of the Pacific Northwest.

Data from particular tree-ring chronologies can be used to reconstruct climatic variations in specific areas, but even data from the dense North American data set are likely to emphasize variations over hundreds of kilometers rather than local variations that are unique to individual stations. Cropper and Fritts (1986) evaluated several different approaches to reconstructing the climate history of the Pasco Basin. In their preliminary models they used two different, mutually exclusive sets of tree-ring data: the 65 tree-ring chronology set previously described and a smaller 10 tree-ring chronology set on deposit in the International Tree-Ring Data Bank. In addition, they varied the number of climatic stations used. These stations are listed in Table 5.2-1. They chose the best models, based on each model's ability to predict climatic data withheld from the model-building process. These final models predict temperatures at three stations (Baker, Oregon, Spokane and Walla Walla, Washington; Fig. 5.2-2) and precipitation at three stations (Baker, Oregon, Colville, Washington, and Lewiston, Idaho; Fig. 5.2-2) on the basis of the 65 tree-ring chronology set.

Reconstructions of annual temperature and precipitation for the Pasco Basin are shown in Figure 5.2-3. Seasonal values are shown in Figures 5.2-4 and 5.2-5. The statistics reconstructed for the prior three centuries are given in Table 5.2-2. The reconstructed temperature averaged over the year was generally warmer and more variable in the three prior centuries. The average annual temperature for 1602 to 1900 was only 0.09 °C warmer, which is 22% of the standard deviation for annual temperature reconstructed in the 20th century. However, the summer and autumn average values are reconstructed to have been somewhat cooler in the prior three centuries.

The annual precipitation reconstructed from tree rings was, on the average, 8.13 mm higher in the prior three centuries; the standard deviation of the reconstructions was 29% higher. The precipitation reconstructed for spring and summer was 3.81 and 2.97 mm lower in the prior three centuries, respectively.

Other tree-ring based reconstructions for Washington show significant differences from the findings of Cropper and Fritts (1985). Graumlich and Brubaker (1986) reconstructed annual temperature back to 1590 for Longmire, Washington, a climatic station approximately 200 km (125 mi) east of the Pasco Basin. They based their reconstruction on six tree-ring chronologies collected at upper tree lines along the crest of the Cascade Range. Their results indicate that annual temperatures between 1590 and 1900 were approximately 1 °C colder than those during the 20th century. Their estimate of pre-20th century cooling is corroborated by evidence that glaciers of the Cascade Range extended beyond their modern termini during the Little Ice Age (Miller, 1969; Burbank; 1981, 1982; Heikkinen, 1984). The differences between the estimates of Cropper and Fritts (1985) and those of Graumlich and Brubaker (1986) for pre-20th century temperatures may be attributed to the differing sensitivities of subalpine versus semiarid trees in responding to long-term changes in temperature. Further collection and analysis of subalpine tree-ring chronologies from areas surrounding the Pasco Basin would be necessary to ascertain the spatial extent of the cooling documented in Graumlich and Brubaker (1986).

Long-term estimates of annual precipitation for the Columbia Basin, based on tree-ring chronologies from within the basin (Graumlich, 1985, 1987), show differences in the timing of wet and dry episodes as compared with the estimates of Cropper and Fritts (1985). Both reconstructions, however, show an increase in the standard deviation of annual precipitation prior to the 20th century. Differences between these two reconstructions could be due to the different spatial scales of the primary data sets. The Cropper and Fritts tree-ring data encompass the entire western United States, while Graumlich's reconstructions are based on a localized grid of tree-ring sites in the Columbia Basin.

A commonality to all the results is the indication that the most recent meteorologic records cannot be considered fully representative of the variability of climate that can be expected in the next decades to centuries at the Hanford Site. The lack of agreement between the climatic estimates of Cropper and Fritts (1985) and others has implications for the potential reliability of tree-ring-based reconstructions of the climate history of the Pasco Basin. Only limited tree-ring data from the Pacific Northwest were available to Cropper and Fritts at the time of their analyses (Fig. 5.2-1 shows the sites used). While approximately 40 new tree-ring chronologies from the Pacific Northwest have become available in the past several years (Graumlich, 1985), the tree-ring data coverage for areas adjacent to the Pasco Basin is still sparse. High site-to-site correlations between tree-ring chronologies elsewhere in the Pacific Northwest (Graumlich, 1985; Brubaker, 1980) indicate that the tree-ring data from this region are faithful recorders of regional macroclimatic conditions.

The reliability of the tree-ring estimates of past climate at the Pasco Basin can be increased by (1) enhancing the tree-ring data network from areas in and adjacent to the Pasco Basin and (2) carefully selecting sites in which the sensitivity of the trees to the climatic variable of interest is maximized. The degree of resolution that is required by the BWIP for the purposes of site characterization is not known. Plans to develop this information are described in Section 8.3.1.5. As described in Section 8.3.1.5, the BWIP intends to make use of the longer record of dendroclimatic reconstructions to characterize the short-term climate most representative of the Pasco Basin.

5.2.1.1.2 Late Quaternary climatic record

Vegetation change is time transgressive (Wright, 1976), and it is necessary to define certain time-stratigraphic units to which the observed chronology of environmental change may be related. The last glacial age of

North America is termed the Wisconsinan, and the current interglaciation is termed the Holocene. The Wisconsinan/Holocene boundary (equivalent to the Pleistocene/Holocene boundary; Fig. 5.2-6) is set at 10,000 yr B.P. by international convention (Olausson, 1982, pp. 10-15).

The middle Wisconsinan interstadial, a period of generally warmer global temperatures and reduced ice sheet volume, ended approximately 23,000 yr B.P. (see Fig. 5.2-6) (Dreimanis and Raukas, 1975). The subsequent late Wisconsinan stadial included the last period of maximum glaciation and greatest lowering of global temperatures. The Wisconsinan maximum of global ice volume, or the last full glacial, is dated at 18,000 yr B.P. (CLIMAP, 1976).

The period after the onset of deglaciation but before the beginning of the Holocene is termed the late-glacial or latest Wisconsinan and, for the purpose of this document, is assigned the inclusive dates of 12,000 to 10,000 yr B.P. Informal subdivisions of the Holocene are as follows: the early Holocene is set from 10,000 to 8,000 yr B.P., the middle Holocene from 8,000 to 3,500 yr B.P., and the late Holocene from 3,500 yr B.P. to present (VanDevender and Spaulding, 1979).

5.2.1.1.2.1 Palynological methods of reconstructing late Quaternary climates

The palynological record of any given site is dominated by that of wind-pollinated terrestrial plants and of aquatic plants. Insect-pollinated plants are poorly represented in pollen-stratigraphic sequences, because their pollen is not dispersed into the atmosphere and thus comprises an insignificant fraction of the total pollen rain. Fortunately, most of the plants that characterize the bioclimatic subdivisions of the Columbia Basin and vicinity are wind-pollinated species and are well represented in the pollen record. These include the conifers of the surrounding mountains and most of the shrubs and grasses of the Columbia Basin itself.

Because the dispersal capabilities of some types of pollen vary considerably among taxa, caution must be exercised in interpreting any pollen record. For example, at sites in the central Columbia Basin, pine pollen may comprise between 20% and 50% of the total pollen encountered in subrecent samples (Mack and Bryant, 1974), despite the fact that the nearest, upwind pine populations are more than 100 km distant. The sparse vegetation cover and relatively low pollen production of steppe ecosystems mean that an important component of the pollen rain will originate in distant montane forests. Thus, a knowledge of the differential production and dispersal capabilities of pollen types is important for adequate reconstruction of vegetation and climatic conditions.

Basin size and local topography are also important considerations in the analysis of pollen-stratigraphic records. The larger the basin, the more effectively its sediments integrate pollen from a wide area. Smaller basins have been shown to provide records of primarily local conditions, while larger basins provide a more regional picture (Prentice, 1985). The fallout of pollen into any basin is affected by local wind patterns. Hence, care should be taken in comparing pollen records from a crater lake with a distinct rim and from an exposed lake on the Columbia Plateau.

Pollen taxonomy is an important consideration in paleoclimatic reconstruction. Identification of the plant species comprising the local paleovegetation is desirable, since each species differs in its climatic tolerances and ecological characteristics. Species-level determinations are normally impossible for the pollen of principal plant taxa of the Columbia Basin and vicinity.

For example, there are great differences in the climatic regime of areas where species of Artemisia grow. Also, some species are herbaceous, while others are woody shrubs. However, the pollen of these plants can only be discriminated to the genus level (Artemisia). Similarly, the pollen of grasses can only be identified to the family level (Gramineae), and the pollen of pine can only be identified to subgenus (Pinus subgen, Haploxylon, or subgen Diploxylon). On occasion, the pollen of two genera may be indistinguishable, such as that of Douglas-fir (Pseudotsuga menziesii) and larch (Larix). Hence, the pollen-type designation is Pseudotsuga/Larix.

Table 5.2-3 presents the scientific and common names for the principal pollen taxa and vascular plant species considered in this review. For simplicity's sake, common names will be employed in the text wherever possible.

The above considerations add uncertainty to the determinations of palynology; however, palynology is the only means of paleoclimatic reconstruction that has been widely applied in the Columbia Basin. This is the best alternative to gaining an understanding of actual long-term climatic variability in the vicinity of the proposed repository site.

Pollen-stratigraphic records of climatic change

C. J. Heusser (1983, 1985), Mehringer (1985a), Barnosky (1984), and R. G. Baker (1983) have reviewed the late Pleistocene and Holocene palynological studies of the late Pleistocene and Holocene from the Pacific Northwest, including the Columbia Basin. Such palynological records allow reconstruction of climatic changes inferred from the fossil pollen preserved in sediment of that age. A major drawback of the records from east of the Cascade Range in Washington is that, with one exception, they are too short to provide data on the last glacial maximum. Scouring of the land surface by glacial ice in the north (see Waitt and Thorson, 1983) and by late-glacial Missoula floods in the main part of the basin (Baker, 1973) removed most of the preexisting Quaternary record.

There are detailed records of paleoenvironmental conditions that span the last 10,000 to 12,000 yr and one site that provides a fossil record spanning 33,000 yr. These sites yield data on glacial climatic conditions and on present interglacial climatic conditions. The means and methods of pollen-based environmental reconstructions employed in this region are examined next.

Chronological control

Accurate dating of sedimentary sequences providing proxy climate data is essential. To establish the chronology of paleoclimatic regimes and describe rates of major climatic change, the timing of stratigraphic succession at any site must be well controlled. The time-series approach employed in pollen stratigraphic studies requires accurate age determinations of sediments at regular intervals in the stratigraphic column. Radiocarbon dating and tephrochronology have been used in the Columbia Basin and vicinity. The assay of $^{12}C/^{14}C$ ratios in organic material is a well-established method of dating organic samples younger than 60,000 yr. Radiocarbon dating is performed directly on sediment samples of known depth. Palynological studies rarely specify the details of dating, but the common approach is to date the insoluble organic fraction of the sample.

Since a radiocarbon date is a statistical measure of the ^{14}C activity in a sample, it carries an uncertainty value expressed by convention as one standard deviation. Standard deviations on radiocarbon dates are normally less than 200 yr for samples younger than 5,000 yr B.P., less than 500 yr for samples younger than 15,000 yr B.P., and less than 1,000 yr for samples younger than 30,000 yr B.P. The practical dating range (unless accelerators are used) of most radiocarbon laboratories is approximately 30,000 yr.

Tephrochronology is a particularly useful technique when used in conjunction with radiocarbon dating to establish time control of stratigraphic sequences in the Columbia Basin and vicinity. Eruptions of Cascade Range stratovolcanoes periodically deposit blankets of volcanic ash (tephra) in the Columbia Basin, which are often preserved in the sediments of lakes and marshes as discrete layers. Sufficient research has been conducted to determine the age of the most widespread of the late Wisconsinan and Holocene tephra layers of the Pacific Northwest and to provide mineralogic data necessary to identify different tephras (Porter, 1978; Mehringer et al., 1984; Sarna-Wojcicki et al., 1983).

Tephras not only provide time lines useful in correlation, they also provide an independent check on the accuracy of radiocarbon dates. In certain areas, depositional basins contain groundwater rich in dissolved carbonate. This carbonate is usually derived from older geologic formations and is deficient in ^{14}C. Submerged organisms that synthesize dissolved carbonate often have an apparent radiocarbon age many centuries or millenia in excess of their true age (Riggs, 1984). Radiocarbon dates bracketing a known-age tephra provide a convenient test of the reliability of radiocarbon dating at a given site and allow the elimination of one potential source of error in paleoclimatic reconstruction.

The tephra layers reported in eastern Washington sediment cores and their approximate ages (in years) are noted in Table 5.2-4 (Sarna-Wojcicki et al., 1983).

Reconstruction of local environments

Large-amplitude climatic changes occur infrequently, and climatic records spanning millenia rather than decades are required to assess their amplitude and impact properly. The paleobiotic record provides the most comprehensive data on which to base reconstructions of past climatic conditions. The following is a synopsis of the available literature on late Quaternary climatic change in the Columbia Basin, focusing in particular on pollen-stratigraphic analyses.

The pollen-stratigraphic record at any given site combines indicators of local vegetation and environment with those of regional significance. Reconstruction of local environmental conditions is a valuable aspect of these studies, because they provide inferences on the effectiveness of aquifer recharge, changes in salinity and alkalinity, and rates of erosion in the catchment basin. Lowering of lake levels can be inferred from a change in sedimentation rate, sediment type, or paleosalinity. A change in the abundance of the pollen of rooted aquatic plants also can be evidence of changing water depth. A hiatus in a near-continuous sedimentary record may suggest a significant climatic event.

When undiscerned, discontinuities also may be a major source of error in calculation of the age-depth relationships essential to establishing an accurate chronology of climatic change. Desiccation of a lake may be recorded by the development of an incipient soil profile, which can be identified in sediment cores. A soil horizon often features an increase in degraded or corroded pollen.

The fossils of algal colonies are frequently encountered in the course of palynological studies, and they have considerable potential for augmenting pollen-based environmental reconstructions. Of particular importance are the fossils of the algae Pediastrum and Botryococcus, which appear to fluctuate in abundance in response to changes in water chemistry (Mehringer, 1985a). The pollen of rooted aquatic plants also can be used as an indicator of lake status and salinity. Increases in the percentage of the pollen or such plants can mean an increase in the extent of shallow-water or brackish habitats suitable for their establishment.

Reconstruction of regional environments

Definition of the pollen source area is critical for the proper interpretation of palynological records. In steppe regions, the source area can be vast. The source area for pollen records in the Columbia Basin proper is the entire basin and the mountains that form its western rim. This consideration plays an important role in the elucidation of the climate signal provided by fossil records.

For example, sagebrush is characteristic of relatively warm, dry environments, while pine is characteristic of relatively cool, moist conditions. Increases in the percentages of sagebrush and decreases in the percentages of pine in a pollen-stratigraphic profile could be interpreted as an indication of increasing aridity.

This is not necessarily the case, since sagebrush is a component of the local vegetation and pine pollen falling in the Columbia Basin has its origins in the distant Cascade Range coniferous forests. Under normal circumstances, increased effective moisture in a steppe ecosystem would lead to denser stands of shrubs and grasses in the vicinity of the site. The resultant increase in locally produced sagebrush and grass pollen production would overwhelm the input of pine pollen from its distant source areas.

Thus, higher values of sagebrush relative to pine pollen actually could mean increased effective moisture. Similarly, in Columbia Basin records, an increase in pine pollen could well indicate increased local aridity and consequent reduction in the pollen production of steppe species. Calculations of pollen influx are valuable in this regard, because they identify which pollen types contribute to a fluctuation in relative percentages.

The fossil record provides the best available source of proxy climatic records for long time spans. Organisms, particularly micro-organisms and plants, are often restricted to a relatively narrow range of temperature and precipitation conditions. Assuming these limits have not changed through time, fossils can be used as evidence of a climatic regime similar to that where their modern counterparts occur today.

Reconstruction of vegetation from fossil pollen data is usually based on the comparison of modern pollen rain with present-day vegetation. This relationship provides modern analogs useful in interpreting the fossil assemblages. An analog approach is necessary, because there currently exists no direct correlation between the abundances of a given pollen taxon and the local abundance of the plant taxa that produced that pollen type.

In the Pacific Northwest, the modern pollen rain has been described for several regions (Fig. 5.2-7): the Columbia Basin (Mack and Bryant, 1974), the Snake River Plain (Davis, 1984b; Bright, 1966), the Okanogan Highlands (Mack et al., 1978d), parts of the Cascade Range (Heusser, 1973, 1978a, 1978c; Barnosky, 1981), the Puget Trough (Davis, 1973; Heusser, 1978b; Heusser and Heusser, 1981; Barnosky, 1981), the Olympic Peninsula (Heusser, 1969), and the southwestern Oregon Coast Range (Heusser, 1983). The type of sample used as well as the presentation and documentation of the pollen spectra vary among these studies, and several points need to be clarified before published, modern, surface samples can be compared from region to region or with the fossil record.

1. The taxonomic level of pollen identifications is inconsistent from study to study. Some studies, for example, identify different species of Alnus, Picea, and Abies (e.g., Heusser, 1969), whereas others do not. The identification of certain important taxa, crupressaceae (cedar family) most notably, is highly variable in these studies.

2. The percentages are calculated on a variety of pollen sums.

3. The modern pollen spectra were collected largely from soil samples and moss polsters. Because different types of samples have different pollen trapping characteristics (Birks and Birks, 1980, pp. 183-187), the modern pollen spectra may not be comparable to the fossil spectra contained in lake sediment cores.

4. The present-day vegetation in many areas of the Pacific Northwest has been greatly modified by agriculture and forest clearance. Disturbance of the vegetation means that pollen spectra from surface samples may not represent the pollen rain of the natural vegetation.

Before relationships between pollen rain and vegetation and climate can be established, the degree of difference between postsettlement and presettlement pollen assemblages needs to be assessed. Such information is obtained by examining the pollen stratigraphy in the uppermost sediments of natural lakes and ponds. In the Columbia Basin, for example, a study of a short core from Wildcat Lake revealed a significant increase in Chenopodiineae and composite pollen as a result of agricultural disturbance in the last century (Davis et al., 1977). Similar short-core studies are needed for all major vegetation types in the Pacific Northwest to determine whether the present data base of surface spectra needs to be expanded or replaced.

Although satisfactory analogs usually can be found in the modern or presettlement pollen rain to reconstruct vegetation during postglacial times, the approach requires caution when dealing with older pollen records that may represent vegetation with no modern counterpart. As will be discussed, there exists some uncertainty as to the exact nature of vegetation in the Columbia Basin during the last glacial age. This uncertainty results, in large part, from a lack of suitable modern pollen analogs. Despite this shortcoming, it is still possible to make general statements concerning the nature of the vegetation and probable climatic conditions that existed.

Climatic indicators in the pollen-stratigraphic record

Pollen taxa useful in reconstructing paleoclimatic regimes and in describing climatic change are those that can be associated with modern plant species that have distinct climatic affinities or that are restricted to discrete, environmentally distinct habitats. They are divisible into terrestrial and aquatic species, and the former are further divisible into the pollen of trees (arboreal taxa), shrubs, grasses, and herbs.

Arboreal taxa include conifers and hardwood species. Upland trees important in the pollen record of the Columbia Basin are primarily conifers, and have distinct distributions in response to changes in gradients of temperature and effective moisture. Mountain hemlock (Tsuga mertensiana), spruce (Picea engelmannii), whitebark pine (Pinus albicaulis), and subalpine fir (Abies lasiocarpa) are among those arboreal taxa typical of the highest elevation forest of the Columbia Basin.

The presence of spruce (Picea) pollen in glacial-age sediments, particularly at Carp Lake and Bonaparte Meadows, is an indication of lowered thermal regimes. The absence of mesophytic high-altitude conifers (fir, mountain hemlock) indicates generally dry conditions. The spruce of the interior mountains (Picea engelmannii) and whitebark pine (Pinus albicaulis) are drought tolerant relative to most other high-elevation conifers. Whitebark pine is considered to be the Haploxylon pine pollen recorded at most sites on the northern periphery of the Columbia Basin in latest Wisconsinan time (Mack et al., 1978b).

Haploxylon pines generally grow at higher elevations and in wetter sites than either of the Diploxylon pines common in the interior Pacific Northwest today (i.e., ponderosa pine and lodgepole pine). Both Diploxylon species are relative thermophiles, and lodgepole pine also is an important seral species. That is, it occupies habitats disturbed by fires or forest clearing. Pollen studies in the Pacific Northwest also indicate that lodgepole pine was an early invader of deglaciated substrates (Baker, 1983).

Spruce, true fir, Douglas-fir, or larch are moderately important in the late-glacial and the late Holocene interval at sites on the northern periphery of the Columbia Basin. These are middle- and high-elevation conifers whose presence suggests somewhat cool and moist conditions. Their late-glacial occurrence is usually attributed to long-distance sources, but in the late Holocene they were part of the local vegetation.

Other conifers important in the record are the mesophytic "maritime" species more typical of the coastal forests than the interior. Western hemlock (Tsuga heterophylla) and western red cedar (Thuja plicata), particularly, are restricted to areas of relatively high annual precipitation in the Pacific Northwest. These become prominent in the pollen record of the periphery of the Columbia Basin in the late Holocene.

Hardwood trees important in the pollen-stratigraphic record include alder, birch, and willow. The particular species involved in the fossil record are unknown in the absence of plant-macrofossil data. It is generally thought that these trees grew in riparian settings and probably around the margin of each study site. Alder also may have been an important seral species and thus provides a record of disturbance. Oak pollen is also present in the fossil record. It is the most drought-tolerant tree in the Pacific Northwest and is confined to the lower forest edge.

The three most important pollen taxa of the Columbia Basin (Artemisia, Gramineae, and Chenopodiineae) have very ambiguous climatic signals. The reason for this, as suggested earlier, is that they include a large number of species of widely varying climatic tolerances, but with indistinguishable pollen. Artemisia pollen may have come from one of many herbaceous alpine species (e.g., A. scopulorum, A. norvegica, A. michauxiana, and A. trifurcata) or from steppe shrubs (e.g., sagebrushes such as A. tripartita, A. tridentata, and A. arbuscula). The environmental affinity of the sagebrushes is not straightforward; some species range to high elevations in the Cascade Range.

In interpreting the fossil record, the entire assemblage must be considered as a whole before deciding whether alpine or steppe species of *Artemisia* were the likely contributors to the pollen rain. Grass (Gramineae) pollen abundance is well correlated with *Artemisia* pollen in most of the Columbia Basin records, but, unfortunately, grasses have even wider climatic amplitude than *Artemisia*. They also are important in northern latitude steppe communities as well as in tundra. Chenopodiineae pollen may represent disturbance-adapted taxa like goosefoot (*Chenopodium*) or tumbleweed (*Salsola*), or desert shrubs like saltbush (*Atriplex*) or greasewood (*Sarcobatus*). Their distribution is often localized to the most xeric and saline substrates.

Aquatic plant pollen reflects primarily changes in the local environment of the depositional basin. Barnosky (1985) has pointed to the near complete lack of aquatic taxa in glacial-age sediments of Carp Lake. This also applies to the latest Wisconsinan sediments from other sites on the periphery of the Columbia Basin. Attributing their absence to colder temperatures during the Wisconsinan glacial age, their reappearance during the early Holocene is thought to be a response to postglacial warming.

Sedge pollen frequencies reflect the same general trend from lake to lake on the margin of the Columbia Basin. The late Holocene is a time of increased relative frequencies of sedge pollen, which probably occurred as a result of the progressive filling in of bog and lake environments.

5.2.1.1.2.2 The paleoenvironmental and paleoclimatic record from the Columbia Basin and vicinity

The Carp Lake site is the oldest, nearly continuous record from the Columbia Basin, with pollen data extending back to 33,000 yr B.P. (Barnosky, 1985). This record is discussed in detail in Section 5.2.1.1.7. Figure 5.2-8 shows the location of this and other sites of palynological records discussed herein. Thus, information is available on the nature of paleoclimates from the middle Wisconsinan interstadial, through the last glacial maximum, and in the current interglaciation. Several pollen records spanning the last glaciation, however, have been described from the mesophytic conifer forests west of the Cascade crest (Heusser, 1983; Barnosky, 1984). These data show some clear climatic signals, which will form an important basis of comparison with planned work (see Section 8.3.1.5) in the Columbia Basin.

The Hanford Site lies in part of the Pasco Basin. There are three reasons for considering paleoclimatic records from the entire physiographic Columbia Basin and its periphery in an assessment of climate variability in the vicinity of the proposed repository site:

- The present mesoscale climate of the entire Columbia Basin is similar to that of the Pasco Basin, and climatic changes would presumably take place over the entire Columbia Basin. Through examination of the impacts of the same climatic change at different sites, better perspective on the nature of that climatic change can be gained.

- A sufficient number of sites for reconstructing climatic change cannot be obtained from the immediate vicinity of the Pasco Basin.

- Global climatic changes often are the cause of local variations of temperature and precipitation. To model properly the potential future climatic change in the vicinity of the proposed repository site, the nature of the boundary connections between the Columbia Basin and the global climatic system must be examined.

This review of the available literature on the late Quaternary paleoenvironments of the Columbia Basin and vicinity concentrates on those pollen-stratigraphic records that span at least the last 10,000 yr with no major hiatuses. This reivew is supported by published reports, including those of the U.S. Department of Energy's Pacific Northwest Laboratory (Cropper and Fritts, 1986).

Specific site records are discussed in this section to provide as accurate a portrayal as possible of the late Quaternary paleoenvironmental conditions of the Columbia Basin and vicinity. There are approximately two dozen sites in the Columbia Plateau and surrounding regions from which palynological records have been obtained. These locations were shown in Figure 5.2-8 (not all are discussed in this document). Procedures used to screen reports discussed in this section place emphasis on the duration and completeness of the record and the availability of final research results. Late Quaternary paleoecological studies in the interior Pacific Northwest that were not considered (Table 5.2-5) were rejected because of the following:

- Fossil record less than 10,000 yr long or accompanied by a hiatus exceeding 2,000 yr.

- Site is outside the limits of the Columbia Basin.

- No or inadequate time control of the stratigraphic section.

Summaries exist that compare the original interpretations of paleoclimatic investigations in the Columbia Basin and vicinity (Heusser, 1983; Baker, 1983; Mack et al., 1978d; Mehringer, 1985a). Table 5.2-6 presents a list of the six sites selected for detailed assessment. All published sites in eastern Washington were shown in Figure 5.2-8. Of the six sites discussed herein, four are located on the northern periphery of the Columbia Basin (in the Okanogan Valley of north-central Washington and the Selkirk Range of northeastern Washington); one is located in the eastern foothills of the Cascade Range (on the southwestern periphery of the Columbia Basin); and one, Wildcat Lake (Mehringer, 1985b), is near the center of the Columbia Basin within 100 km of the Hanford Site.

The following paragraphs provide a summary of published information about paleoclimates of the Columbia Basin obtained from palynological studies. Each site is discussed separately and the information is given in the same format for each site to facilitate comparisons. Data reported include the following:

- o Original reference.
- o Geographic location (see Fig. 5.2-8).
- o Description of the site.
- o Modern vegetation.
- o Present climate.
- o Tephra layers.
- o Radiocarbon dates.
- o Interpretations of pollen evidence in chronologic sequence.
- o Discussion of the record.

The first five items above provide the basic data about each site; the sixth and seventh give some measure of chronologic control. Chronologic control is important to measure the uncertainty in estimates of the climates of a time period and rates of change of climate. The eighth and ninth provide the interpretations and references that the BWIP will apply based on the record presented.

A recently devised, more objective approach to reconstructing past climates from fossil-pollen data offers promise in the clarification of the paleoclimatology of the region. This method relies on the construction of statistical relationships between modern pollen data and climate in a given bioclimatic region. These relationships are then applied to known-age fossil-pollen data from a geographic network of sites to obtain quantitative estimates of paleoclimate at specified time horizons (Howe and Webb, 1983; Bartlein et al., 1984). Applications of this new methodology to the characterization of long-term climatic variability in the vicinity of the Hanford Site must await additional data and construction of pollen-climate transfer functions. As will be demonstrated, there is considerable need for this approach to diminish the uncertainty that remains concerning the nature of past climates in the Columbia Basin.

<u>Big Meadow Fen</u>

<u>Reference</u>: Mack et al., 1978a, pp. 956-965.

<u>Location</u>: Central Selkirk Range, Pend Oreille County, Washington; NW1/4, sec. 7, T. 37 N., R. 42 E.; 1,040 m elevation (see Fig. 5.2-8).

<u>Description</u>: A fen that lies in part of a depression formed by glacial erosion and subsequent deposition of a glacial moraine. The site is in an upland setting near the headwaters of Big Muddy Creek in a valley approximately 150 m deep and approximately 2 km wide, cutting across the north-south-trending Selkirk Range. This valley is inferred to be a glacial feature due to its U-shaped cross section and the presence of glacial till on

the flanks and bottom. The depression holding Big Meadow Fen was free of ice approximately 13,500 yr B.P., and the stratigraphic section has a maximum age of approximately 12,500 yr B.P.

Modern vegetation: Western hemlock and mountain lover shrub dominate in the immediate vicinity of the fen. Grand fir, Engelmann spruce, Douglas fir, and western white pine are important trees in the surrounding forest.

Present climate (as reported): Extrapolated from the climate station at Newport, Washington, approximately 64 km to the south-southeast of the site. Mean July and January temperatures for the year 1975 are 21 and -5 °C, respectively. Annual precipitation is approximately 700 mm.

Tephra layers: Mount Mazama Layer O; Glacier Peak, undetermined.

Radiocarbon dates: Eight.

Major features and interpretations of the paleoenvironmental record at Big Meadow Fen

Five major pollen zones are recognized in this record.

- Approximately 12,500 yr B.P. Time of formation of Big Meadow Fen depositional basin, initiation of sedimentation, and beginning of pollen zone 1. The pollen spectrum is characterized by high percentages of sagebrush and grass and low percentages of arboreal taxa in which the pollen of Haploxylon pines (probably western white pine or whitebark pine) is important (Fig. 5.2-9). This late-glacial assemblage may represent:

  (1) An alpine or subalpine parkland.

  (2) An early pioneer community invading a freshly deglaciated substrate.

  (3) A tundra-like community such as has been described south of the Laurentide ice margin in the Midwest (Mack et al., 1978d).

  The third alternative was chosen because of the particularly low pollen influx at Big Meadow Fen and the traces of mesophytic conifer pollen. However, no known suitable modern analog exists. Increases in the percentages of willow and birch pollen at approximately 11,500 yr B.P. coincide with the marked increase in total pollen influx at this time.

- Approximately 9,700 yr B.P. Beginning of pollen zone 2. Comparatively high frequencies of Diploxylon pine (ponderosa or lodgepole pine), alder, and grass typify this zone. Pond lily and Douglas fir/larch pollen appear for the first time. Continued moderate percentages of *Artemisia* pollen suggest either bunchgrass-sagebrush steppe with ponderosa pine in the vicinity, or an open ponderosa-pine parkland with heliophytic grasses and sagebrush. A change from cool-moist conditions during zone 1 time to a climate warmer and drier than today (zone 2) is inferred. The prominence of relative thermophilous aquatic plants, such as pond lily and cattail, also indicates a marked rise in temperatures in zone 2 time.

- Approximately 7,000 yr B.P. Beginning of pollen zone 3. Percentages of grass pollen decline relative to the preceding period, and there is a significant increase in Douglas-fir pollen (Fig. 5.2-10). The pollen influx diagram, however, shows a reduction in the abundance of all pollen types (see Fig. 5.2-9). Within zone 3, sagebrush pollen declines to negligible amounts approximately 5,000 yr B.P., while percentages and influx of pine, spruce, fir, and sedge pollen begin to increase. Reference to the pollen-influx diagram shows an increase in absolute numbers of pollen of these taxa (see Fig. 5.2-9). An establishment of a mesophytic conifer forest and consequent increase in effective moisture is implied.

- Approximately 3,300 yr B.P. Beginning of pollen zone 4 at the approximate 2.8-m depth. A decline in alder pollen and a step-wise increase in influx of spruce and fir pollen are evident at this time horizon. Accompanying these changes is a marked increase in the influx of western hemlock pollen (see Fig. 5.2-9), suggesting continued increase in effective moisture.

- Approximately 2,400 yr B.P. Beginning of pollen zone 5. This date marks an increase in western hemlock pollen and a decline in that of spruce and fir. Conditions are generally similar to zone 4, although with reduced influx values for all pollen types (see Fig. 5.2-9).

Discussion of the Big Meadow Fen record

In summary, the Big Meadow Fen record suggests cool, relatively moist conditions prior to 9,700 yr B.P. when tundra-like communities prevailed. It was followed by a period, from 9,700 to 3,300 yr B.P., when the climate was warmer and drier than today's and steppe vegetation was more widespread. Independent indications of dry conditions are the abrupt changes in sediment type and in the rate of sediment accumulation, both of which suggest possible hiatuses and lake desiccation. As early as 5,000 yr B.P., and especially after 3,300 yr B.P., trees invaded this landscape, presumably in response to the introduction of the cooler, more humid conditions of the present day.

Waits Lake

Reference: Mack et al., 1978d, pp. 499-506.

Location: The southwest flank of the Selkirk Range, Stevens County, Washington; sec. 17 and 20, T. 31 N, R. 40 E; approximately 2 km west of the Colville River Valley, approximately 540 m elevation (see Fig. 5.2-8).

Description: Waits Lake is an oval depression approximately 2 km long that covers approximately 180 ha. It is borderd by bedrock and glacial deposits that rise approximately 60 m above the surface of the lake. It lies in an upland setting, approximately 120 m above the Colville River flood plain. The retreat of the Colville lobe of the Cordilleran ice sheet from the vicinity of the site took place approximately 13,500 yr B.P.

Modern vegetation: Douglas-fir and the shrub mallow-ninebark typify the local vegetation. Other common shrubs are snowberry and ocean-spray. Western red cedar, grand fir, and western larch are conifers common near the lake.

Present climate (as reported): Extrapolated from the climate station at Chewelah, Washington, 13 km northeast of Waits Lake. Average July temperatures ranged between 18 and 21 °C, and average January temperatures ranged between -5 and -4 °C for the years 1972 to 1975 inclusive. Annual precipitation for these years varied from 480 to 660 mm.

Tephras: Mount Mazama Layer 0; Glacier Peak, undetermined.

Radiocarbon dates: Eight.

Major features and interpretations of the paleo-environmental record at Waits Lake

The pollen record is divided into four major zones; zone 1 is divided into two periods.

- Approximately 12,500 yr B.P. Beginning of pollen zone 1a. This zone is characterized by a moderate percentage of *Artemisia* (either sagebrush or more likely herbaceous species), grass, willow and buffalo-berry pollen (Fig. 5.2-11). Haploxylon pine (western white pine or whitebark pine) pollen is well represented in pollen zones 1a and 1b. A largely treeless, tundra-like vegetation under a climate that was considerably cooler and moister than that of today is inferred.

- Approximately 11,500 yr B.P. Beginning of pollen zone 1b, marked by an abrupt decline in buffalo-berry and pine pollen, and a distinct rise in *Artemisia* pollen. Arboreal taxa that appear consistently at about this time include spruce, true fir, and birch (see Fig. 5.2-11). No climatic interpretations are offered by the authors to account for these changes. A gradual increase in pine pollen and decrease in *Artemisia* pollen begins at approximately 11,200 yr B.P., indicating a further increase in the regional extent of forest and a decline in tundra-like vegetation. Concomitant with these changes is an increase in the pollen percentages of Diploxylon pine (ponderosa pine or lodgepole pine) (see Fig. 5.2-11).

- Approximately 10,000 yr B.P. Beginning of pollen zone 2. Changes that began at about the time of deposition of the Glacier Peak tephra reach a maximum early in this zone, with generally low percentages of *Artemisia* pollen and high percentages of pine (mostly Diploxylon) pollen (see Fig. 5.2-11). The authors infer somewhat warmer and drier conditions than in zone 1. Within this zone there is a modest trend toward declining pine pollen frequencies and increasing percentages of *Artemisia*, grass, and saltbush pollen, indicating increasingly dry conditions. In this case, the *Artemisia* pollen is probably attributable to sagebrush. The zone 2/3 boundary lies at the culmination of this trend, immediately above Mount Mazama ash.

- Approximately 6,700 yr B.P. Beginning of pollen zone 3. A surge in sagebrush frequencies follows the fall of Mount Mazama ash, but shortly after a decline occurs, first in sagebrush then in saltbush and ragweed pollen (see Fig. 5.2-11). The pollen frequencies of these xerophytic plants decline at the same time as percentages of pine increase. The high values of sagebrush suggest this period as the warmest, driest interval of the Holocene. A gradual retreat and probable local reduction of steppe vegetation in the Colville River Valley and a consequent increase in forest is indicated by the end of the zone.

- Approximately 5,000 yr B.P. Beginning of pollen zone 4. Percentages of sagebrush and other steppe and grassland pollen types decline, while pine percentages increase. Forest vegetation developed in the vicinity of Waits Lake at this time and has persisted with little apparent change over the last five millenia. Variations in the percentages of alder, birch, and aquatics occured at approximately 2,500 yr B.P., and probably indicate shallowing water and local hydroseral development.

Discussion of the Waits Lake record

Because Waits Lake lies in an area of carbonate rocks, it seems likely that the radiocarbon dates are too old. Mount Mazama ash, which is well dated elsewhere as 6,700 yr B.P., has bracketing dates of 8,000 ±120 yr B.P. and 7,910 ±100 yr B.P. at Waits Lake. The discrepancy between these dates and the true age of Mount Mazama ash is c.a. 1,300 yr. The authors assumed that ancient carbon deficient in ^{14}C was incorporated into the Waits Lake sediments at a constant rate. On that basis, they suggested that all their radiocarbon dates be corrected by -1,300 yr to give the true age.

The pollen record at this site suggests three main climatic episodes. The late-glacial period (prior to 10,000 yr B.P.) featured cool, moist conditions and tundra-like vegetation. Progressive warming took place from 10,000 to 5,000 yr B.P. with the period of maximum warmth and aridity occurring just after the eruption of Mount Mazama. At this time, steppe was most extensive in the Okanogan Highlands. A return to cooler, moister conditions after 5,000 yr B.P. led to the development of forest at the expense of steppe. Modern conditions and climate were established after 2,500 yr B.P.

Okanogan Valley

NOTE: The pollen-stratigraphic records from two sites, Mud Lake and Bonaparte Meadows, are treated collectively in this study. They lie 46 km apart on opposite sides of the Okanogan Valley (see Fig. 5.2-8).

Reference: Mack et al., 1979, pp. 212-225.

Mud Lake

Location: West side of the Okanogan Valley in the Pine Creek drainage, NE1/4, SE1/4, sec. 24, T. 38 N., R. 30 E.; approximately 10 km northeast of Conconully, Washington; 655 m elevation (see Fig. 5.2-8).

Description: A small pond in the Pine Creek drainage; a "bedrock controlled depression" left by the wasting of a stagnant block of ice in glacial drift.

Modern vegetation: The site lies within the Douglas-fir-snowberry habitat type described by Daubenmire and Daubenmire (1968). Ponderosa pine and Douglas-fir forest dominates the north-facing slopes within 100 m of the site. Sagebrush-bunchgrass steppe occupies drier, south-facing slopes in the vicinity.

Present climate (as reported): Extrapolated from the climate station at Conconully, Washington (elevation 707 m), for the years 1972 through 1975. Average annual precipitation is 424 mm; average July and January temperatures of 20 and -6 °C, respectively.

Tephras: Mount St. Helens, Washington; Mount Mazama Layer O.

Radiocarbon dates: Five.

Bonaparte Meadows

Location: To the east of the Okanogan Valley; NW1/4, SE1/4, sec. 20, T. 38 N., R. 30 E.; 28 km northwest of Republic, Washington; 1,021 m elevation (see Fig. 5.2-8).

Description: A large (112-ha) fen on the floor of Bonaparte Creek Valley, a narrow valley inferred to have been a glacial meltwater channel.

Modern vegetation: The upland forest surrounding the fen is typified by Douglas-fir and ponderosa pine with lodgepole pine and western larch. Widely scattered paper birch, spruce, and aspen are restricted to the margin of the fen.

Present climate (as reported): Extrapolated from the climate station at Republic, Washington (800 m elevation), for the years 1972 through 1975. Average annual precipitation is 393 mm, and mean July and January temperatures are 18 and -8 °C, respectively.

Tephras: Mount Mazama Layer O (two layers).

Radiocarbon dates: Fourteen.

Major features and interpretations of the paleoenvironmental record from the Okanogan Valley

Four major pollen zones have been designated and one hiatus has been recognized.

- Greater than 11,000 yr B.P. to approximately 10,000 yr B.P. Pollen zone 1. A period dominated by the pollen of *Artemisia* and grass (Fig. 5.2-12) suggests open vegetation. The pollen of western white pine or whitebark pine and sedges also is abundant in this zone (see Fig. 5.2-12). The low pollen influx suggests a sparsely vegetated landscape, although no exact modern analog has been found. The occurrence of Haploxylon pine pollen argues for cool, moist conditions.

- Approximately 10,000 yr B.P. Beginning of pollen zone 2. Marked increase in percentages and influx of grass and sagebrush pollen are accompanied by increases in birch, alder, and willow pollen. There is a contemporaneous decline in pine pollen (see Fig. 5.2-12 and 5.2-13), most of which is Diploxylon type (either lodgepole or ponderosa pine). The rise in birch pollen occurs much earlier at Mud Lake (by approximately 11,300 yr B.P.; Fig. 5.2-14) than it does at Bonaparte Meadows (approximately 10,000 yr B.P.; see Fig. 5.2-12). A sagebrush-bunchgrass steppe with trees confined to moist habitats is inferred for the regional vegetation during this pollen zone (approximately 10,000 to 6,900 yr B.P.). The climate is inferred to be warmer and drier than that of today.

- Approximately 8,300 to approximately 6,900 yr B.P. Desiccation at Bonaparte Meadows and inferred hiatus in the sedimentary record.

- Approximately 6,900 yr B.P. Beginning of pollen zone 3. Inferred vegetation and climate are similar to that of zone 2 time.

- Approximately 4,800 yr B.P. Major change in the pollen stratigraphy and the beginning of pollen zone 4. Sagebrush percentages and influx values decline at this zone boundary, to be replaced primarily by pine pollen (see Fig. 5.2-12 and 5.2-13). Grass, willow, birch, and alder pollen decline at about this time, while Douglas-fir or larch pollen increases. A major vegetation change, from steppe to forest, and a climatic change from warm and dry conditions to cooler and moister conditions are indicated. An increase in the amounts of spruce, true fir, Douglas-fir or larch, and pine pollen at approximately 2,600 yr B.P. (see Fig. 5.2-13) indicates progressively cooler, moister conditions approaching those of the present day.

Discussion of the Okanogan Valley records

Of the two sites reported in this document, the Bonaparte Meadows record is by far the most useful. Radiocarbon dates at Mud Lake were obtained only on the bottom section of the core, from approximately 11,500 to approximately 8,000 yr B.P. (452 to 257 cm) (see Fig. 5.2-14). The two prominent tephra units reported for the Mud Lake section (Mount Mazama ash at approximately 175 cm and Mount St. Helens ash at approximately 40 cm) are unaccompanied by any bracketing radiocarbon dates. Thus, the accuracy of radiocarbon dates on sediments of this lake cannot be tested, and deposition rates for the top part of the core cannot be calculated.

There are two pronounced episodes of vegetation and climatic change recorded in the pollen-stratigraphic sequences. The first occurs at approximately 10,000 yr B.P. and, vegetatively, can be characterized as a change from open tundra-like vegetation with some trees to sagebrush steppe similar to that which occurs today in the Columbia Basin. The second fluctuation took place approximately 4,800 yr B.P., when the regional vegetation changed from sagebrush steppe to widespread forest. Before this time, vast areas of the Okanogan Valley supported steppe-shrub vegetation. After this time, sagebrush communities were a relatively minor component of the vegetation, occupying restricted habitats such as the south-facing slopes near Mud Lake.

Wildcat Lake

Reference: Mehringer, 1985b.

Location: Near the eastern edge of the Channeled Scablands, approximately 3 km south of the Palouse River, 5 km south-southwest of Hooper, Washington; 342 m elevation (see Fig. 5.2-8).

Description: A steep-sided plunge pool formed during the Missoula floods in the Columbia River basalts. The last of the Missoula floods traveled through the Cheney-Palouse tract between 14,000 and 13,000 yr B.P. (Mullineux et al., 1978; Baker, 1983). Initiation of sedimentation at Wildcat Lake must postdate flooding. Recent observed water depth is 2 to 5 m. Sediment depth is 25 m. The depth of the stratigraphic column yielding fossil data reported is 23.8 m. The period of reported record is 10,600 yr B.P. to the present with a hiatus from approximately 6,900 yr B.P. to approximately 5,400 yr B.P.

Modern vegetation: Bunchgrass-fescue grassland, approximately 50 km east of the sagebrush-bunchgrass steppe of the central Columbia Basin, and approximately 40 km to northwest of the nearest stands of ponderosa pine in the foothills of the Blue Mountains. The site is approximately 200 km east of the present eastern limit of Cascade Range coniferous forests.

Present climate (as reported): No report. This information is not available to the BWIP, however, it is required for correct interpretation of the record at this site and will be obtained during site characterization (Section 8.3.1.5). The record will be reviewed when that information is available.

Tephras: Mount St. Helens Set Y; Mount St. Helens Set W; Mount Mazama Set O; three unidentified ashes more than 10,000 yr old.

Radiocarbon dates: Seventeen.

Major features and interpretations of the paleo-environmental record from Wildcat Lake

Three below-Mount Mazama ash-pollen (BA) zones are described for Wildcat Lake.

- Approximately 10,600 yr B.P. Beginning of BA pollen zone 1. Grass pollen was important relative to sagebrush and long-distance transported pine pollen (Fig. 5.2-15). Grassy steppe is the reconstructed vegetation. The beginning of the zone features relatively slow sediment accumulation, inferred deep water, and moist climatic conditions. The greatest abundance of grasses (see Fig. 5.2-15) occurs approximately 9,700 yr B.P., suggesting a period of enhanced effective moisture.

- 9,000 yr B.P. Beginning of BA pollen zone 2. Decline in pollen concentration suggests well-vegetated slopes, reduced sedimentation, and deep water. A marked increase in abundance of *Pediastrum* colonies (Fig. 5.2-16) suggests enhanced input of fresh water. The inferred sagebrush-grass steppe implies a period of maximum effective moisture.

- Approximately 8,000 yr B.P. Beginning of BA pollen zone 3. This zone begins a period of rapid sedimentation and inferred shallow water caused by increased slope wash due to decreased vegetation cover. A trend toward less grass pollen relative to transported pine and sagebrush, both of which are interpreted to have come from distant sources (see Fig. 5.2-15), implies less dense and more xerophytic local vegetation. A marked decrease in the abundance of *Pediastrum* algae colonies (see Fig. 5.2-16) suggests decreased input of fresh water. Accompanying indications of decreasing effective moisture is a significant rise in juniper pollen (see Fig. 5.2-16), probably that of the thermophile western juniper (*Juniperus occidentalis*), and a decline in the pollen of the hydrophytic taxon willow. A brief but distinct reversal of the trend toward increasingly xeric vegetation conditions occurs between approximately 7,400 to 7,200 yr B.P. The BA pollen zone 3 ends with approximately 1,500 yr of the record missing. This was due to the mechanical difficulty of recovering the great thickness of Mount Mazama ash and tephra-rich sediments.

Seven above-Mount Mazama ash-pollen (AA) zones are described for Wildcat Lake.

- o <u>Approximately 5,400 yr B.P.</u> Beginning of AA pollen zone 1. Conclusion of rapid deposition (Fig. 5.2-17). A slowing sediment-accumulation rate indicates increased effective moisture and better vegetated, more stable slopes. Accompanying these changes are declines in the relative percentages of ragweed and sagebrush pollen and an increase in grass pollen (Fig. 5.2-18) that suggest increasing effective moisture.

- o <u>Approximately 4,400 yr B.P.</u> The AA pollen zone 2 begins at about this time. Very slow deposition (deeper water) begins (see Fig. 5.2-17), indicating well-vegetated slopes and cooler and effectively moister conditions. A decline in ratios of sagebrush to grass pollen (Fig. 5.2-19) also suggests increased effective moisture in a transition to sagebrush-grass vegetation.

- o <u>Approximately 2,400 yr B.P.</u> Beginning of AA pollen zone 3 (see Fig. 5.2-18). An inferred decrease in local vegetation cover caused sediments to accumulate at rates comparable to those of 8,000 to 5,400 yr B.P. (see Fig. 5.2-17).

- o <u>Approximately 1,750 yr B.P.</u> Beginning of AA pollen zone 4. No marked changes in pollen frequencies or ratios are noted at this horizon (see Fig. 5.2-18 and 5.2-19).

- o <u>Approximately 1,100 yr B.P.</u> Beginning of AA pollen zone 5. High ratios of pine to grass (see Fig. 5.2-19) indicate maximum development of distant Cascade Range coniferous forests.

- o <u>Approximately 550 yr B.P.</u> Beginning of AA pollen zone 6. This period features drought from 550 to 400 yr B.P.

- o <u>Approximately 130 yr B.P.</u> Beginning of AA pollen zone 7. Pollen concentration and sediment-accumulation rates markedly increase (Fig. 5.2-20). Pollen of introduced species are present, and an abrupt rise in <u>Chenopodiineae</u> (saltbush, tumbleweed, goosefoot) pollen (see Fig. 5.2-18) indicates extensive, anthropogenic alteration of plant communities in the Columbia Basin.

<u>Discussion of the Wildcat Lake record</u>

Wildcat Lake is far removed from any vegetation zone boundaries and all vegetation changes that took place in the period of record occurred in sagebrush or sagebrush-grass communities. Therefore, the pollen record is complacent relative to those from sites on the periphery of the Columbia Basin, where major changes are recorded by the shifting position of the forest/steppe border and by the changing composition of the local forest. The relative complacency of the Wildcat Lake record also is due, in part, to the low taxonomic resolution of the major pollen taxa in steppe vegetation.

The pollen-stratigraphic record from Wildcat Lake is well dated and is technically the most detailed paleoenvironmental study from the Columbia Basin. Rapid deposition (18.8 m, excluding the Mount Mazama tephra, in 10,600 yr compared to 8 m in 33,000 yr at Carp Lake) and close interval sampling allow high resolution of the biostratigraphy of the basin. Many inferences of changing climatic conditions are based on changes in sedimentation rates (see Fig. 5.2-17) or on the ratios of different pollen taxa, some assumed to be long-distance transported and others assumed from local sources. Mehringer (1985b) assumes that the ecotone between sagebrush-dominated communities and grass-dominated communities is governed by today's climate. On the basis of the relative amounts of grass and sagebrush pollen, these two communities are identifiable in the fossil record at Wildcat Lake. Shifts in pollen percentages are interpreted as past shifts in the position of the ecotone. Mehringer has used weather station data to estimate the magnitude of the climate change to effect the shift. He suggests that all vegetation changes recorded by the Holocene data can be attributed to no more than a 100-mm variation in annual precipitation or to less than a 2-°C variation in mean annual temperature.

Based on cluster analyses, the most pronounced periods of vegetation and climatic change below Mount Mazama ash occurred at approximately 9,000 yr B.P. and at 8,000 yr B.P. (the pollen zone 1/2 and 2/3 boundaries for the below-Mount Mazama ash-pollen zones) (Fig. 5.2-21). These are interpreted as a pronounced increase in effective moisture and then a marked increase in aridity, respectively. In the stratigraphic section above Mount Mazama ash, cluster analysis indicates that major episodes of environmental change occurred at approximately 2,400 and 500 yr B.P. (the pollen zone 2/3 and 5/6 boundaries for the above-Mount Mazama ash-pollen zones, respectively) (see Fig. 5.2-21). The two episodes are interpreted as the initiation of periods of effective aridity. On sedimentary grounds, another important transition occurred at approximately 4,400 yr B.P. (see Fig. 5.2-17). A change in sedimentation rate is indicated by the changed slope of the line, which began about that time. The occurrence of the mineral struvite, the slow rates of sediment accumulation and pollen accumulation (i.e., pollen influx), and the abundance of the algae Pediastrum and Spirogyra zygospores are used as evidence of deeper water. Mehringer (1985b) explains the change in water depth as the result of cooler, moister conditions beginning approximately 4,400 yr B.P.

Carp Lake

References: Barnosky (1983, 1984, pp. 619-629, 1985, pp. 109-122).

Location: Yakima folds section of the Columbia Basin, approximately 30 km north of the Columbia River, 15 km north of Goldendale, Washington; 714 m elevation (see Fig. 5.2-8).

Description: A lake covering 11 ha in the bottom of a volcanic crater, with walls that vary from 54 to 73 m above the lake surface. Maximum observed water depth is 2 m. Depth of stratigraphic column yielding fossil data reported is 8 m. Period of reported record is approximately 33,000 yr B.P., with an inferred hiatus in deposition from approximately 9,000 to approximately 8,300 yr B.P.

Modern vegetation: Douglas fir/ponderosa pine forest with lodgepole pine, grand fir, and dwarf juniper. The site is approximately 5 km northwest of the forest-steppe boundary. In this region, a distinct zone of oak woodland separates conifer forest from grassland and steppe.

Present climate (as reported): Mean annual precipitation of 510 mm/yr, contrasting with values of less than 250 mm/yr in nearby steppe habitats. Mean July temperatures range between 18 and 21 °C, while average January temperatures have been slightly below freezing for 70 yr of record (station of record is Goldendale, Washington). More than 90% of the annual precipitation falls during the winter half-year. The average length of the growing season is 260 d.

Tephras: Mount St. Helens, Mount Mazama Layer O; unidentified ash, approximately 29,000 yr B.P.

Radiocarbon dates: Thirteen.

Major features and interpretations of the paleo-environmental record from Carp Lake

Four pollen zones are identified at Carp Lake.

- Approximately 33,000 yr B.P. Beginning of pollen zone C4 at base of the stratigraphic column. *Artemisia* and grass are the dominant pollen types (Fig. 5.2-22), indicating widespread sagebrush-grass steppe vegetation and general lack of coniferous forest. Vegetation is inferred to be more similar to the Columbia Basin than today, with distant conifer populations. Low pollen influx that continues into the next zone (Fig. 5.2-23). The progressive decline in pine pollen to approximately 23,500 yr B.P. (see Fig. 5.2-22) indicates increasing dry and (or) cold conditions.

- Approximately 23,500 yr B.P. Beginning of pollen zone C3. Lowest values of pine pollen are encountered in this zone (see Fig. 5.2-22). The elimination of most aquatic plant pollen suggests the beginning of very low-temperature conditions. A periglacial steppe vegetation prevailed at this time, as evidenced by high relative percentages of sagebrush and grass pollen (see Fig. 5.2-22) and low pollen influx.

  The lake shallowed, and clasts derived from the margins of the crater were deposited in the center of the lake at approximately 21,000 yr B.P., confirming pollen evidence for very low effective moisture at this time. By 17,000 yr B.P., extremely low pollen influx values were reached, suggesting a very barren landscape (see Fig. 5.2-23). The lake shallowed to a marsh at approximately 13,500 yr B.P., and effective moisture was much lower than at present. Low pollen influx values continue. A modest increase in relative percentages of spruce and pine (see Fig. 5.2-22) suggests an expansion of distant montane forests. Higher temperatures probably accompanied increased drought, and thus contributed to marsh development locally, as well as to increased forest in the Cascade Range.

- <u>Approximately 10,000 yr B.P.</u> Beginning of pollen zone C2. Sagebrush, grass, and spruce frequencies generally decline throughout this zone. An abrupt increase in goosefoot or saltbush pollen indicates locally alkaline conditions and (or) mud flats. A sharp rise in indeterminate pollen types suggests periods of intermittent drying that allowed degradation of pollen prior to their burial (see Fig. 5.2-22). The precise dating of this pollen zone boundary is uncertain, but Barnosky (1985) places it at approximately 10,000 yr B.P. Douglas fir or larch pollen, juniper or cedar (<u>Cupressaceae</u>) pollen, and bracken fern spores make a sudden appearance at approximately 9,500 yr B.P. The presence of these taxa, as well as in temperate aquatics, suggests increased temperatures (see Fig. 5.2-22). The appearance of bracken spores may indicate increased temperature and higher precipitation. A hiatus in the record is noted beginning approximately 8,500 yr B.P., when the lake completely dried. During the hiatus, a pine forest was established locally.

- <u>Approximately 8,500 yr B.P.</u> Beginning of pollen zone C1. The lake basin once again contained water, and sediment deposition recommenced. Arboreal pollen percentages and taxonomic diversity increased greatly. Western hemlock and oak pollen appear for the first time (see Fig. 5.2-22). The climate shifted from warm-dry to temperate-mesic conditions. Lodgepole or ponderosa pine pollen frequencies decline at approximately 4,000 yr B.P., while Douglas-fir or larch pollen and western hemlock pollen increase (see Fig. 5.2-22). The pollen changes suggest a period of increased effective moisture. Marked water-level fluctuations at Carp Lake were over by approximately 1,500 yr B.P.

<u>Discussion of the Carp Lake record</u>

The pollen-stratigraphic record from Carp Lake is the only sequence from the Columbia Basin and vicinity that provides data from the middle and late Wisconsinan stadials, and it includes a complete record of the full glacial. Four distinct climatic episodes are hypothesized at the Carp Lake site.

A period from 33,000 to 23,500 yr B.P. featured a temperate but drier climate than today, which allowed steppe communities to spread to higher altitudes. This was succeeded by a very cold dry phase, when the temperate steppe was replaced by a sparsely vegetated periglacial steppe. At approximately 13,500 yr B.P., there is some suggestion of a climatic warming, as evidenced by the increase in spruce pollen (probably from distant sources) and by the sedimentary changes that suggest a lower lake level. Warming was well established by approximately 10,000 yr B.P., when the steppe assumed a temperate aspect and the lake became very shallow. By 9,000 yr B.P. the basin dried completely, and a hiatus of 500 to 1,000 yr is recorded. Cooler, moister conditions are implied after approximately 8,300 yr B.P. by high percentages of pine pollen. Further cooling occurred approximately 4,000 yr B.P., as indicated by the increase in Douglas-fir, true fir, western hemlock, oak, and alder.

5.2.1.1.2.3 Summary of climatic records of the Holocene and late Pleistocene

An understanding of changing patterns of vegetation and climate in the Columbia Basin is a critical element in the overall characterization of the Hanford Site. In addition, knowledge of the paleoclimatic history of southern British Columbia provides important indicators of future glacial patterns that may potentially affect the repository. Each region is discussed individually in this section.

Columbia Basin

Major patterns of vegetation and climatic change in the Columbia Basin and vicinity are evident in the late Quaternary pollen-stratigraphic record. Prior to approximately 10,000 yr B.P., the Columbia Basin and surrounding highlands were treeless. The nature of the vegetational and climatic reconstruction, however, vary considerably among authors.

Carp Lake is the only site that provides data on environmental conditions during the middle and all of the late Wisconsinan. For both periods, steppe vegetation appears to have prevailed across the southwestern Columbia Basin. Temperatures were higher during the middle than during the late Wisconsinan stadial, but aridity was pronounced during both periods. Low pollen influx values typified the last half of the late Wisconsinan at the Carp Lake site. Sparsely vegetated steppe, rather than temperate shrub-steppe, has been the vegetation type proposed to account for the pollen spectra and influx values of this time. These vegetation conditions imply pronounced aridity and extreme cold.

How arid and how much colder was it during the glacial maximum? The pollen-stratigraphic studies done to date offer no quantitative information as to the degree of change. The qualitative reconstructions available do not speak to the absolute amount of cooling or decrease in precipitation that typified the climates of the last glacial age. However, it appears likely that the decline in average annual temperatures in the southwestern Columbia Basin was extreme (Barnosky, 1985).

Average annual precipitation may have been even lower than today's, based on the extremely low pollen influx values obtained from late Wisconsinan sediments and the expansion of steppe far beyond its current limits in the Columbia Basin (Barnosky, 1985).

The reconstruction of the late-glacial vegetation and climate near the periphery of the ice sheet is more difficult (Mack et al., 1978a, 1978b). Records from the Okanogan Highlands and northeastern Washington indicate cool, humid conditions in most valleys until approximately 10,000 yr B.P. The paleoclimatic inference is based on small amounts of spruce, Haploxylon pine, and fir in the pollen spectra. According to this model, a periglacial fringe existed close to the ice sheet and the Okanogan region was covered by tundra-like communities as a result of locally cold, humid conditions. These particular conifers are not present farther south at Carp Lake, which reinforces the argument for aridity there.

To some degree, the early pollen records north of the Columbia Basin also must reflect the pioneer species that were available to colonize a deglaciated landscape (Mack et al., 1976, 1978a, 1978c, 1978d, 1979). This may explain the presence of weedy species with broad ecological amplitude (e.g., Chenopodiineae, buffalo berry). An alternative explanation, not yet addressed in the literature, is that the area was covered by a periglacial steppe just as at Carp Lake. Such a reconstruction suggests cold, arid conditions throughout eastern Washington, even close to the ice-sheet margin.

The conflicting interpretations of the late-glacial climate highlights the need for additional long records at some distance from the former ice sheet. Additional data will clarify the extent and character of the periglacial zone, so that the regional glacial climate can be distinguished from the possibly more-localized conditions at the northern margin of the basin.

Nearly all the pollen data in the Pacific Northwest suggest a major climatic warming between 11,000 and 9,500 yr B.P. Barnosky (1984) notes that western Washington records, as well as those at Carp Lake, show initial climatic warming at approximately 13,500 to 12,500 yr B.P. However, the period of maximum warmth and aridity did not begin until approximately 10,000 yr B.P., both west of the Cascades and in the southwestern Columbia Basin. Pollen data from the Okanogan Highlands and northeastern Columbia Basin show climatic warming between 10,000 and 9,000 yr B.P. The only exception is Wildcat Lake, which lies some distance from any ecotone and, therefore, may not have been palynologically sensitive to this climatic change.

The nature of subsequent climatic events varies from site to site and has been summarized in separate studies (Mack et al., 1978d; Barnosky, 1984; Mehringer, 1985a, 1985b). The climatic events are depicted in Figures 5.2-24 and 5.2-25 (see Fig. 5.2-8 for locations). Some sites, such as Carp Lake, Goose Lake, Bonaparte Meadows, and Big Meadow, suggest that the early Holocene was the warmest and driest interval of the present interglaciation. At most of these sites, this warm/dry episode ends just before the eruption of Mount Mazama, although at Carp Lake it ends as early as 8,300 yr B.P.

An early Holocene warm period is consistent with the climatic reconstruction from western Washington, British Columbia, southern Idaho, and even the northwest Arctic (Barnosky, 1985; Mathewes, 1973; Alley, 1976; Davis, 1984a; Ritchie et al., 1983). An interpretation of early Holocene warming is in apparent conflict, however, with evidence for renewed glacial activity in the Cascade Range at about the same time (Beget, 1980; Waitt et al., 1980). This evidence is discussed in Section 5.2.1.2.2.

On the other hand, a few pollen records from the northeastern Okanogan Highlands and the eastern Columbia Basin show the period of maximum warmth as the middle Holocene, either bracketing or just following the eruption of Mount Mazama (Waits Lake, Wildcat Lake, Williams Fen). This reconstruction is similar to that from sites in the Rocky Mountains area and the Great Basin, and is in keeping with the classical Altithermal sequence of Antevs (1948).

The timing of maximum warmth and aridity in eastern Washington is a major discrepancy in the published climatic interpretations. It appears that sites just west and north of the Columbia Basin show an early Holocene warming, whereas those to the east and southeast record a middle Holocene event. Regardless of the chronology, the effects are clear: steppe was greatly increased and the forest/steppe border was shifted several kilometers (as much as 100 km) to the north, east, and west. Many lakes were at their lowest level and some dried completely.

Barnosky (1985) suggested that the apparent asychroneity in the pollen response may reflect slightly different plant communities at each site. These differences may be the result of nonclimatic factors (e.g., bedrock type, topography, rates of species dispersal and propagation, and competitive interactions among species). Alternatively, they may be the result of subtle climate changes within a region brought about by shifts in the position of the prevailing westerlies. Because local topography modulates the penetration of maritime air in eastern Washington, such shifts would create very different climates in close proximity, disrupting the spatial pattern of vegetation and, thus, the pollen source area. Changes in the position of the prevailing westerlies or an attenuation in their intensity also would alter the provenance of pollen transported long distances. This effect would have been of special importance when the productivity of the local vegetation was low (e.g., during periods of tundra, steppe, or desert).

The pollen records show a return to cooler, moist conditions in the late Holocene, but again the timing and complexity of the climatic change vary from site to site. Carp Lake shows an early increase in effective moisture with the expansion of pine forest at approximately 8,300 yr B.P. At Big Meadow Fen, Goose Lake, Simpson's Flats, and Bonaparte Meadows, the cooling trend is introduced about the time of the Mount Mazama eruption (approximately 6,900 yr B.P.). At all these sites, further cooling is inferred between 5,000 and 3,300 yr B.P. from the spread of mesophytic conifers. At Williams Lake, Waits Lake, and Wildcat Lake, this interval also marks the beginning of cooler, moister conditions.

A few records (Simpson's Flats, Big Meadow Fen, and Hager Pond in Idaho) suggest another stepwise cooling approximately 2,500 to 1,500 yr B.P., leading to the establishment of the modern climate. The record from Wildcat Lake is an exception to the generally three fold Holocene climatic sequence noted in eastern Washington. The pollen data are interpreted largely on the basis of changing ratios between major taxa, and the result is a complex sequence of warm/dry intervals alternating with cool/humid periods that is unmatched in other well-dated records.

British Columbia

Paleoclimatic reconstructions in British Columbia have been attempted for the Olympia interglacial interval, Fraser Glaciation, and Holocene (i.e., for that part of the Quaternary with good radiometric dating control). The Fraser Glaciation in the Pacific Northwest is the approximate correlative of the late Wisconsinan glaciation in the midcontinent of the United States (Clague, 1981, p. 5). The Olympia interglacial preceded the Fraser Glaciation. Studies of plant microfossils in middle Wisconsinan nonglacial sediments and of oxygen-isotope fractionation in spleothems dated by uranium-series methods have provided information on the climate of southern British Columbia during the Olympia interval (Clague, 1981, p. 12). The data suggest that the climate generally was cooler than at present, but that there were periods during which air temperatures attained present-day values.

Palynological investigations of early Fraser Glaciation sediments in south coastal British Columbia indicate a gradual deterioration of climate after approximately 29,000 yr B.P., marked by the expansion of subalpine plants into lowland areas. In the southern interior, the first clear paleo-ecological signal of climatic deterioration dates to approximately 25,000 yr B.P.

Although the climate during the Fraser Glaciation obviously was cool and moist enough to support a large ice sheet over British Columbia, it apparently was not as severe as has been previously assumed. There is little or no evidence of permafrost or large expanses of tundra at the southern and western margins of the Cordilleran ice sheet during the Fraser Glaciation. Sedimentological data and the apparent absence of large-scale subglacial thrusting suggest that the Cordilleran ice sheet was mainly warm based (i.e., at the pressure-melting point).

Paleoecological studies of 18,000-yr-old sediments near Vancouver, British Columbia, and 15,000- to 16,000-yr-old sediments on the Queen Charlotte Islands indicate the existence of diverse floras in coastal areas of British Columbia at the height of the Fraser Glaciation. In the vicinity of Vancouver, British Columbia, approximately 18,000 yr B.P., Abies lasiocarpa - Picea cf. engelmannii forest and parkland grew under cold, humid continental conditions near sea level. This plant assemblage is similar to that occurring today at subalpine elevations (900 to 2,250 m) in the interior of British Columbia.

Thus, climate at 18,000 yr B.P. in Vancouver, British Colubmia, probably was characterized by long, cold, wet winters and very short, dry, frost-free summers. Mean annual temperature was depressed approximately 8 °C, and the tree line was 1,200 to 1,500 m lower than today. Approximately 15,000 to

16,000 yr B.P., the lowlands of the Queen Charlotte Islands supported a varied, nonarboreal terrestrial and aquatic flora indicative of a moderately oceanic cold climate, but obviously one not so severe as to preclude the survival of biota.

It thus appears that the climate at the western margin of the Cordilleran ice sheet at the climax of the last glaciation, and probably also during earlier glaciations, was somewhat milder than might be expected given the large expanses of ice at those times. This is undoubtedly related to the moderating influence of the Pacific Ocean. The extent to which interior areas of British Columbia were similarly affected by changes in the Pacific Ocean during glaciations remains unknown.

The climate immediately following the Fraser Glaciation in southern British Columbia was cool and moist. In some areas there were minor resurgences of glaciers at that time, but the climate rapidly ameliorated and probably was as warm as, or warmer than, the present from 10,500 yr B.P. until at least 6,600 yr B.P. This warm period was followed by a generally cooler, moister period that has persisted until the present (Clague, 1981, p. 1).

During this cool interval, relatively minor advances of alpine glaciers occurred between 5,000 and 4,000 yr B.P., 3,100 and 2,300 yr B.P., and within the last several centuries. During the 17th and 18th centuries, most glaciers in British Columbia achieved their greatest late Holocene size, but these advances are insignificant when compared to the great advances of the Pleistocene (Clague, 1981, p. 24).

5.2.1.1.3 Climate change through the Quaternary

Climates of the Quaternary (past 1.67 m.y.) are likely to represent the climates to be expected over the next 100,000 yr. Although studies of late Quaternary (the last 125,000 yr) climates do provide some estimate of the magnitude of possible climate change, that period cannot be considered an exact analog of the future. To obtain a conservative estimate of the potential range of climate change, long records spanning the Quaternary must be considered whenever available.

As would be expected, knowledge of the paleoclimates of earlier times in the Quaternary is less complete than knowledge of the late Quaternary. Several factors are responsible.

- o Resolution and precision of the paleoclimatic record decrease in proportion to the antiquity of that record.
- o Dating procedures, both absolute and relative, tend to be more widely applicable to the last 50,000 yr.

- The record is incomplete, because it is reworked or removed by erosion. Loss of record is roughly proportional to the time elapsed since original deposition.

- A natural bias exists; researchers tend to concentrate effort on the records that are most productive.

Our knowledge of the Quaternary as a whole does not reveal the detail that is known from the late Pleistocene and the Holocene, such as discussed previously.

The information about paleoclimates throughout the Quaternary that has been obtained tends to support theories developed for more recent times. It also is most reasonable to assume that the shorter term variability that is known from the more recent record also occurred in the earlier parts of the Quaternary. This section reviews Quaternary climate change prior to the last glaciation, and provides a basis for the modeling of future climate change described later in this section and in Section 8.3.

Few data are available from the Columbia Basin itself that can provide evidence of climatic conditions prior to the latest glacial episode. One exception is the clear evidence from flood deposits dated at older than 200,000 yr (Tallman et al., 1978, pp. 78-81) that glacial conditions have existed prior to the last glaciation. Loess deposits on the Columbia Plateau (the Palouse loess) record multiple glacial-interglacial cycles. This inference is based on the assumption that the loess itself accumulated during, or immediately following, glacial episodes, and that the intervening soils represent times of interglacial climates.

Multiple glacial deposits have been recognized in northern Washington and southwestern British Columbia. Some of these may represent pre-Fraser glaciations (Hicock and Armstrong, 1983, pp. 1232-1247; Rutter, 1977). Porter (1976, pp. 61-75) discusses the evidence of multiple glaciations in the Cascade Range. The paleoclimatic implications of these data are summarized by Clague (1978, pp. 95-100) and Alley (1979, pp. 213-237). The last interglacial (oxygen-isotope stage 5) is thought to be represented in northern Washington by the Whidbey Formation (Easterbrook et al., 1982, p. 161; Easterbrook and Rutter, 1981, p. 444) and in southwestern British Columbia by the Muir Point Formation (Armstrong and Clague, 1977, p. 1474). Evidence of the gross climatic conditions of the Pacific Northwest over at least the past 75,000 yr (Stuiver et al., 1978) appears to be in agreement with the record of western North America and the northern hemisphere in general. Interpretation of the data suggests that this synchroneity and amplitude correspondence have been typical of the entire Quaternary and can be expected to remain the case for the next 100,000 yr. Understanding of longer term variability requires the analysis of records obtained from regions far removed from the Hanford Site.

On a global basis, three major forms of evidence are available to reconstruct the paleoclimatic record of the entire Quaternary:

- o Lake sediment cores have been retrieved from sites of nearly continuous sedimentation throughout the Quaternary.
- o A relatively new source of information about paleoclimates of the past 150,000 yr is available from the cores of ice sheets in Greenland or the Antarctic.
- o The extensive collection of cores of deep-sea sediments has been obtained over the past few decades.

Of the three forms of evidence, the deep-sea sediment cores are the most useful.

Each of these data sources is summarized below. Major conclusions about long-term climate change that are relevant to site characterization are at the end of this section.

Five lake sediment cores, each from a different continent, are available and offer climatic records for much of the Quaternary. The shortest, from Grande Pile, France, extends only to approximately 130,000 yr B.P., but presents a highly detailed record of climates over that period (Woillard, 1979). Spectral analysis (i.e., the determination of frequency components in a time series that have significant amplitude) of the record (Molfino et al., 1984, pp. 391-404) reveal significant cycles of climatic variation at frequencies corresponding to the major precessional periodicities (23,000 and 19,000 yr B.P.) of the Earth. Harmonics of those frequencies also are present.

At Lake Biwa, Japan, a 500,000-yr paleotemperature record has been obtained by Fuji (1976, pp. 357-421). Spectral analysis of this record reveals significant 104,000-, 44,000-, 25,000-, and 12,700-yr B.P. periods (Kanari et al., 1984, pp. 405-414). The first three correspond reasonably well with the known quasi-periods in the orbital parameters of the Earth (Berger, 1978b, pp. 2362-2367); the latter may be a harmonic of the precessional term.

The spectral analysis by Kanari et al. (1984) included an autoregressive prediction model (autoregressive models employ regression techniques to determine relations between a series and values of itself at previous time steps) that was extended to estimate relative global climatic variations for the next 50,000 yr (Fig. 5.2-26). Calibration of that forecast on the assumption that the past temperature variations were 5 °C leads to a forecast of global cooling by 2.4 °C in the next 20,000 yr (Fig. 5.2-27). Uncertainty in that forecast increases rapidly after 15,000 yr. The overall forecast of future global climate differs with those of Imbrie and Imbrie (1980, pp. 943-953), Kukla et al. (1981, pp. 295-300), and Berger (1980, pp. 103-122), as discussed in Section 5.2.2.

A third core has been obtained from Macedonia. There, the pollen record extends for approximately 600,000 yr. It has been correlated with evidence of sea surface temperature variations in the ocean core K708-7. Ruddiman and McIntyre (1976) concluded that the record there shows the influence of the same global controls as shown in other deep-sea records.

The fourth and longest continental record is from Funza, near Bogota, Colombia. The Funza record of paleoclimate is based on palynological methods. It extends back 3.6 m.y. and has been correlated with the records of deep-sea cores V28-239 (Shackleton and Opdyke, 1976, pp. 449-464) and V16-205 (van Donk, 1976, pp. 147-163). Although spectral analysis has not been reported for the paleoclimatic record of the Funza core, 26 glacial cycles can be recognized. A distinct change in the frequency of the signal is reported at 2.8 m.y. B.P. (Hooghiemstra, 1984, pp. 371-377).

On the North American continent, Stuiver and Smith (1979, pp. 68-75) and Smith et al. (1983) reported a record from Lake Searles, California, that extends nearly as far as the Funza core. The Lake Searles record is interrupted by numerous diastems, which apparently reflect the desiccation of that lake (Fig. 5.2-28). Alternate cycles of fillings and emptyings of the lake are clearly shown in that record, and Smith (1984) suggested that these correlate with the longest eccentricity period (440,000 yr) of the Earth's orbit.

Cores taken from the ice caps of Greenland and Antarctica provide records that generally go back to the last interglacial (125,000 yr B.P.). Their glacial implications are further discussed in Section 5.2.1.2. The major value of these ice cores is that they demonstrate the same long-term fluctuations as do the continental and deep-sea cores. This shows that the entire cyrosphere-atmosphere-ocean system was involved in the climatic fluctuations of the Quaternary. Thus, we conclude that defensible models of climates to be expected at the Hanford Site, or at any of the proposed nuclear waste repository sites, must be based on a comprehensive representation of the entire climate system at a global scale.

The most fundamental source of information about paleoclimate on a global scale has been obtained from the deep-sea piston cores, of which more than 16,000 have been obtained (Ruddiman, 1985, p. 198; Fig. 5.2-29). Although there are a number of complications to the method, paleoclimate signatures can be extracted from these cores with a resolution of a few thousand years (Ruddiman, 1985, p. 203). The various paleoclimatic indicators used and the reconstruction procedures applied are summarized by Ruddiman (1985, pp. 203-209) and include oxygen isotopes, calcium carbonates, carbon isotopes, Foraminifera, Diatoms, and Radiolaria. Limitations imposed by the coring techniques and the oceanographic setting rule out all but a few hundred cores for use in paleoclimatic reconstructions.

Emiliani (1955) showed that the ratio of two isotopes of oxygen ($^{18}O/^{16}O$) contain a record of the climatic history of the Earth. Later, Shackleton and Opdyke (1973) demonstrated that this record mostly reflects the amount of ice stored (as glaciers) on the continents. Mix and Ruddiman (1982) have shown that deciphering that record is made difficult by the nonlinear interaction of several factors. This interaction can introduce a lag in the oxygen isotope signal of as much as 3,000 yr. The true amplitude of the ice volume fluctuations also may be misrepresented.

From core V28-239, a record extending back over 2.1 m.y. has been obtained (Shackleton and Opdyke, 1976, pp. 449-464). This is shown in Figure 5.2-30. Evidence of glacial-interglacial cycles is pervasive throughout the core. Prior to approximately 800,000 yr B.P., the amplitude of the fluctuations is considerably reduced (Ruddiman, 1985, p. 223). This also appears in core DSDP-502B from the Caribbean (Prell, 1982). Although the timing is not certain, the latest sequence of Northern Hemisphere glaciations probably began sometime between 3 and 2.4 m.y. B.P. in the late Pliocene (Ruddiman, 1985, p. 223). The date of 2.4 m.y. B.P. from deep-sea drilling site 552 is now generally accepted (Shackleton and Hall, 1984).

The work of Shackleton and Opdyke (1973, pp. 39-55) demonstrated fairly conclusively that the oxygen isotopic variations preserved in the oceanic cores (e.g., V28-239; see Fig. 5.2-30) record the changes in global ice volume due to continental glaciation. These data are discussed more fully in the review of the long-term glacial history (Sections 5.2.1.2.4 and 5.2.1.2.5).

5.2.1.2 Glaciation

This section will document that the most extreme response to climatic change that has been recorded in the Pacific Northwest during the Quaternary is that accompanying glaciation. The presence of large ice sheets influences the climate in the immediate area and may perturb large-scale circulation systems. Glaciers also exert direct effects on the surrounding region, including isostatic depression of the crust and disruption of surface drainage systems.

The history of glaciation provides a basis for forecasts of glacier activity. The most recent glacial fluctuations are documented for the Pacific Northwest (Section 5.2.1.2.1). These include the readvances of the "Little Ice Age" in the past few centuries. Such fluctuations are the latest pulse in the late Holocene glaciations (Section 5.2.1.2.2).

Continental glaciation characterized the late Quaternary when most of British Columbia was covered with the Cordilleran ice sheet. At its maximum, that ice sheet extended into northern Washington. The chronology of that glaciation throughout its southern extent is summarized in Section 5.2.1.2.3.

Although evidence of earlier glaciations in the Pacific Northwest is scanty, information suggests that glacial advances in this area were broadly synchronous with those throughout the world (Section 5.2.1.2.4). Thus, on the basis of global records, it is inferred that the area has been subjected to glaciation at least 5 and perhaps as many as 27 times throughout the Quaternary. The best record of this long-term ice volume change is derived from the oceans.

Glacial cycles are known from other times in the geologic past, and that record is briefly reviewed (Section 5.2.1.2.5). Of interest is the time of initiation of the latest sequence of glaciations. This shows the high probability that such cycles will continue for the next 100,000 yr.

5.2.1.2.1 Modern glaciers in the Pacific Northwest

Presently, most glaciers in the Pacific Northwest are confined to a few higher mountain peaks of the Cascade Range and Olympic Mountains and to higher ranges in British Columbia (Fig. 5.2-31). In the North Cascades, there are 756 glaciers having an area of 0.1 km^2 or more (Post et al., 1971, pp. A2-A4). A number of ice fields with multiple outlet glaciers can be found in British Columbia. The largest ice mass in southern British Columbia is the Kunaklini glacier. This ice field is located in the Pacific Ranges of the Coast Mountains.

The elevations of glaciers in the Pacific Northwest are generally higher in the south due to increased temperature and lower in the west due to increased precipitation from the Pacific Ocean (Meier, 1961). Alpine glaciers in the northern Cascade Range do not extend below 1,500 m elevation. On the Olympic Peninsula in the Olympic Mountains, a number of active glaciers are found at considerably lower elevations. Many glaciers exist in the southern portions of British Columbia, especially along the coast. More northerly glaciers actually extend to sea level.

As with other glaciers throughout the world, those of the Pacific Northwest have fluctuated in size during historic times (Post et al., 1971, p. A7). Many glaciers in the northern hemisphere have contracted considerably since reaching an historic maximum 100 to 500 yr B.P. during the so-called "Little Ice Age." Glaciers in the Cascade Range extended down valley at least 1 km farther at that time. Evidence from British Columbia is more scanty, but it appears that that area experienced the same advances. This episode is but one example of the short-term climatic fluctuations that glaciers are responsive to.

Factors responsible for glacial fluctuations were diverse, both spatially and temporally. Miller (1969, p. 63) concludes that certain glacial advances were controlled by temperature decreases. An example is the expansion of the South Cascade, Dana, and LeConte glaciers in the 16th century.

Other advances were the result of precipitation increases, although temperature changes also may have been a factor. The general advances in the 19th century are an example. Certain advances must have involved simultaneous changes in temperature and precipitation. This was probably the case for the 12th and 13th century advance of Chickamin glacier (Miller, 1969, p. 62).

More study of this phenomenon is likely to yield valuable information about changes in temperature and precipitation. The most reasonable approach would be a survey of existing relevant glaciological, geological, and meteorological literature.

5.2.1.2.2 Glacial change during the Holocene

The Holocene is defined as the period from 10,000 yr B.P. to the present (Burke and Birkeland, 1983, p. 3). This epoch includes the time of melting of the great ice sheets that covered much of Canada and northern Europe during the last glaciation (Wright, 1983, p. xi). Wasting of these ice sheets had begun well before the beginning of the Holocene in most areas of the world.

In the Pacific Northwest, glacial retreat had begun soon after a maximum stillstand at 15,000 yr B.P. (Clague et al., 1980, p. 14). This was approximately 3,000 yr later than the maximum extent of the Laurentide ice sheet in eastern North America. By 10,000 yr B.P., the ice front had receded to the higher elevation, isolated regions that had served as the center of advance prior to the glacial maximum (Fig. 5.2-32). Even so, during the Holocene the Cascade Range has remained "one of the most heavily glaciated areas in the western United States" (Burke and Birkeland, 1983, p. 8).

The early Holocene was a time of general warming in those regions of the world that had suffered the extremes of glacial conditions. Within the Cascade Range, however, there is evidence of minor readvances of alpine glaciers (Beget, 1980; Waitt et al., 1980). Dates on these readvances range from 8,500 to 6,700 yr B.P.

By approximately 6,000 yr B.P., the climate in many areas had reached its post-glacial "optimum" (Denton and Karlen, 1973). Since that time, climate has continued to change to a character more like that of the last glaciation in an episodic fashion. This optimum was probably not synchronous around the globe. It has been referred to in various ways, including the Altithermal (Antevs, 1948, p. 176) 7,500 to 4,000 yr B.P. and Hypsithermal (Deevey and Flint, 1957) 9,000 to 2,500 yr B.P. Additional terms such as the "climatic optimum" may be encountered in the literature. All refer to aspects of the same phenomenon.

Even within the Holocene, climatic change has not been simple. The general trend of warming followed by neoglacial cooling is not observed throughout the world. Moreover, the changes have been interrupted by excursions and oscillations. If examined on a time scale of decades to hundreds of years, the trends may not be obvious at a given locale.

Neoglaciation is a catch-all term suggesting the fact that glaciers have increased in size following their maximum shrinkage during the Altithermal. These glacial fluctuations have not been synchronous throughout the world. Even in the North Cascades, glaciers have reached maximum areal extent anywhere from 4,900 yr B.P. to sometime within the past 100 yr (Miller, 1969).

In the Cascade Range, at least two advances are recognized during neoglaciation. The earliest is dated at approximately 5,000 yr B.P. (Miller, 1969) at Dome Peak. Younger deposits suggest another advance in the last 1,000 yr. Another intermediate advance is recognized from deposits on Mount Rainier. These have been dated between 3,500 and 2,000 yr B.P., using the Mount Rainier Set C and the Mount St. Helens Set Y tephra (Crandell and Miller, 1964). A similar situation may exist for the Wallowa Mountains in Oregon (Kiver, 1974).

Times of certain ice advance in other parts of North America are not represented by moraines in the North Cascades. For example, the advance of glaciers that occurred throughout much of the western United States at approximately 2,800 and 2,600 yr B.P. is not recognized in the North Cascades (Miller, 1969, p. 63).

It can be expected that glaciers in the area will continue to fluctuate significantly in size but in a nonsynchronous pattern over time spans of decades to thousands of years. The general trend of glacier advances during the neoglaciation are summarized in Figure 5.2-33. The limitations of the curve must be understood. That part of the curve from approximately 2,600 to 400 yr B.P. is based largely on (1) indirect climatic evidence rather than radiocarbon dates associated with known glacier fluctuations and (2) data of rather restricted geographic distribution. The curve was constructed on the assumption that major climatic and glacier fluctuations were broadly synchronous, at least throughout the Northern Hemisphere. However, some data contradict the curve; for example, that part of the curve for the period 1,800 to 500 yr B.P. is based mainly on information from the North Atlantic region and depicts glaciers in a retracted position. Yet, several glaciers in northwestern North America were advancing during this period. It is risky to assume that glaciers in one geographic region all respond in the same way to climatic change.

The main conclusion that can be drawn from the curve is that neoglaciation in the Northern Hemisphere was characterized by two main intervals of glacier growth, the first culminating approximately 2,800 to 2,600 yr B.P., and the second spanning the last several centuries and culminating near the middle of the last century. More systematic and higher amplitude patterns of fluctuations in size can be expected over longer time spans, as addressed in the following sections.

Regardless of the number of glacial advances during the Holocene, it is not safe to assume that glacial activity was synchronous everywhere (Burke and Birkeland, 1983, p. 10).

5.2.1.2.3 Glaciation during the late Pleistocene

The Pleistocene in the Pacific Northwest was characterized by the episodic growth and decay of the Cordilleran ice sheet, a complex of valley and piedmont glaciers that enveloped almost all of British Columbia as well as parts of the Yukon Territory, Alaska, Alberta, Montana, Idaho, and Washington (Fig. 5.2-34). The Cordilleran ice sheet in Canada was largely confined between the high mountains of the western and eastern cordillera. Areas west of the Coast Mountains and, to a lesser extent, east of the continental divide also were covered by ice.

Glaciers in several ranges, such as the Ogilvie and Vancouver Island Ranges and the St. Elias, Wernecke, Mackenzie, Rocky, and Queen Charlotte Mountains, were more or less independent of the Cordilleran ice sheet, even at the climax of glaciation. In the northwestern conterminous United States, the ice sheet was composed of a series of large lobes fed, in large part, by ice streaming down large, south-trending valleys in southern British Columbia (Waitt and Thorson, 1983, pp. 53-70).

5.2.1.2.3.1 Character and pattern of glaciation

The character and pattern of glacier growth and decay during each major Pleistocene glaciation were strongly influenced by physiography. The first stage in the formation of the Cordilleran ice sheet was the expansion of valley and cirque glaciers in high mountains of western Canada. During this formative period, which for some glacials lasted thousands of years, plateaus and lowlands of British Columbia remained ice free (Clague et al., 1980; Fig. 5.2-35).

Hydrologic and vegetation changes altered the equilibrium of streams flowing across plateaus and lowlands, and as a result many interior and coastal valleys became aggraded with sediment. With continued cooling, valley glaciers advanced beyond mountain fronts to coalesce as piedmont complexes over plateaus, plains, and coastal lowlands (Porter et al., 1983, pp. 71-111). Eventually, piedmont complexes from separate mountain source areas joined to cover most of British Columbia and adjacent areas.

Throughout many or most glaciations, the major mountain systems remained the principal sources of glaciers, and ice flow was controlled largely by topography. However, at the climaxes of the great Pleistocene glaciations, ice thickened to such an extent over the British Columbia interior (approximately 2.5 km) that one or more domes became established with surface flow radially away from their centers (Mathews, 1955). Figure 5.2-36 shows the configuration of such an ice sheet at its greatest extent. Flow lines on that figure indicate glacier motion at that time.

Each glaciation terminated with rapid climatic amelioration. Deglaciation occurred by complex frontal retreat in peripheral glaciated areas and by downwasting accompanied by widespread stagnation throughout much of the British Columbia interior. Along the western periphery of the ice sheet, glaciers calved back in contact with eustatically rising seas (Armstrong et al., 1985; Waitt and Thorson, 1983). In areas of moderate relief in the interior, the pattern of deglaciation was complex, with uplands becoming ice free first and dividing the ice sheet into a series of valley tongues that decayed in response to local conditions (Clague, 1981, pp. 16-17) (see Fig. 5.2-32).

During each glacial cycle, the time of deglaciation varied from region to region. The first areas to become ice free were those near the periphery of the ice sheet. Active glaciers probably persisted longest in some mountain valleys; however, these glaciers may have coexisted with large masses of dead ice in the plateaus of the Cordilleran interior. Geologic evidence indicates that periods of ice sheet decay were much briefer than periods of ice sheet growth (Fulton, 1984, p. 40).

5.2.1.2.3.2 <u>Glaciation in Southern British Columbia</u>

The major sources of Pleistocene glaciers in southern British Columbia are the Coast, Cariboo, Monashee, and Selkirk Mountains; secondary sources include the Cascade Range and Purcell and Rocky Mountains. During the early phase of each glaciation, glaciers advanced down major valleys in these mountain ranges, while ice caps developed at high elevations. These glaciers expanded and coalesced to form ice sheets that gradually enveloped lowland areas in the British Columbia interior (Fulton, 1984, p. 43).

Ice from the Coast Mountains advanced along a complex digitate front eastward across the Fraser and Thompson plateaus, whereas ice from the Cariboo and Monashee Mountains moved west and south down large valleys transecting the Quesnel and Shuswap highlands. Much of the ice in the Selkirk and Purcell Mountains flowed south by way of the Columbia River valley and the Purcell and Rocky Mountain trenches. During later phases of glaciation, ice from the Coast Mountains coalesced with, and was diverted south by, ice from the Caribou and Monashee Mountains, giving rise to strong southerly flow into the northwestern United States.

During periods of deglaciation, the snowline rose above the levels of the ice sheet surface and most mountain source areas. As a result, the ice sheet rapidly downwasted and receded. Piedmont lobes at the southern periphery of the ice sheet in the northwestern United States retreated rapidly across the international boundary, and a complex network of stagnating ice tongues developed in southern British Columbia valleys.

5.2.1.2.3.3 <u>Fraser Glaciation</u>

Knowledge of the timing of late Pleistocene glaciation in the Pacific Northwest is limited by the lack of suitable techniques for dating events more than 50,000 yr old and by the generally poor preservation of sediments deposited prior to the last continental glaciation. Those events within the range of the radiocarbon method (less than 50,000 yr) are the most precisely dated; however, even for the last glaciation, the exact number and timing of glacial advances remain uncertain.

The last continental glaciation, referred to in the Pacific Northwest as the Fraser Glaciation (late Wisconsinan), began approximately 30,000 to 25,000 yr B.P. (Fulton, 1984, p. 40; Clague, 1981, p. 13). Glacier growth was slow at first, with ice confined to mountain ranges until 25,000 to 20,000 yr B.P., depending on locality. Some parts of southernmost British Columbia remained ice free until 17,000 yr B.P. (Fig. 5.2-37), and the Cordilleran ice sheet did not attain its maximum extent in the Puget Lowlands and probably on the Columbia Plateau until 14,500 yr B.P.

The ice sheet began to decay around 14,000 yr B.P. Parts of the coastal lowlands of southwestern British Columbia were ice free by 13,000 yr B.P., and the ice sheet probably had vacated areas south of the international boundary by approximately 12,000 to 11,500 yr B.P. The Cordilleran ice sheet had completely disappeared by 10,000 yr B.P. or shortly thereafter (Clague, 1981).

Detailed stratigraphic studies in southwestern and southeastern British Columbia just north of the international boundary have shown that ice sheet growth and decay were complex (Armstrong et al., 1985; Clague, 1975). In these areas and presumably elsewhere at the periphery of the Cordilleran ice sheet, there were intervals of retreat prior to the climax of the Fraser Glaciation. There also were stillstands and readvances during deglaciation.

Stratigraphic evidence from southern British Columbia (Fig. 5.2-38) indicates that the Fraser Glaciation was preceded by a long nonglacial period, known as the middle Wisconsinan Olympia interglaciation. The pattern of temperature fluctuations for the Pacific Northwest throughout the past 70,000 yr is shown in Figure 5.2-39. This pattern is inferred from palynological investigation of sediments on the Olympic Peninsula. In this figure, the time spans of the Holocene and late Wisconsinan (Fraser) glaciation are based on the data of Heusser (1972, 1977). The time spans of the mid-Wisconsinan interval are based on stratigraphic successions in western Canada studied by Fulton (1977). The Olympia interglaciation began sometime before 59,000 yr B.P. and, thus, was more than 30,000 yr in length. Throughout this period, glaciers were confined to the major mountain ranges of British Columbia, and environmental conditions and the geomorphic framework were broadly similar to those of the present interglacial (i.e., the Holocene) (Clague, 1981, p. 8).

5.2.1.2.3.4 Postglacial paleohydrology

Landye (1973) described an abundant fossil assemblage from a late-glacial lake in the lower Grand Coulee system. The lake, which he named Lake Bretz (Landye, 1969), occupied the modern basins of Soap, Lenore, Park, and Sun Lakes and was approximately 30 km (19 mi) long. At maximum extent, this Pleistocene lake reached an elevation of 353 m (1,159 ft) above sea level, resulting in a surface area of approximately 50 km^2 (18 mi^2) and a depth of 250 m (800 ft).

The Lake Bretz fossil assemblage included freshwater sphaeriid bivalves, gastropods, ostracods, fish, duck, amphibian, and plant remains. The mollusks, in particular, indicate the presence of cool-water habitats in large lakes. The lake sediments also contain abundant volcanic ash from Glacier Peak, originally described by Fryxell (1965). Fryxell interpreted the ash's age as 12,000 yr B.P. More recent work by Mack et al. (1983) and Mehringer et al. (1984) reinterpreted the age of Glacier Peak tephra (layers B and G) as 11,250 yr B.P..

Based on studies of mollusk populations and sizes in Lake Bretz sediments, Landye (1973) estimated that the lake must have been near its maximum extent for at least 35 yr and that it probably persisted for no more than a few hundred years. It was probably fed by cool, freshwater springs related to high water-table conditions that occurred immediately after the diversion of significant surface water flow from the Grand Coulee system. Aquifers in the basalt apparently could not maintain the lake for longer than a few hundred years, since recharge was cut off by the surface water flow diversion. The presence of Glacier Peak ash indicates that this diversion occurred no earlier than approximately 11,500 yr B.P.

Freshwater mollusk assemblages similar to those in Lake Bretz also have been recognized in exposures of fine-grained, rhythmically bedded sediments throughout the Columbia Basin in association with the Mount St. Helens Set S tephra. The original interpretation of these sites was that they indicated relatively persistent lacustrine sedimentation (Fryxell, 1972; Gustafson, 1976), clearly associated with cooler, wetter environments than occur in the region today. Waitt (1980) described many of these same exposures and ascribed their origin wholly to cataclysmic flooding. This interpretation has yet to be reconciled with the fossil evidence for lacustrine conditions (Baker and Bunker, 1985).

The transition from cool, wet, late-glacial conditions to drier, postglacial conditions is well illustrated by lake sedimentation and palynology in the Columbia Basin. At Wildcat Lake, near Hooper, Washington, the early Holocene (10,600 to 8,000 yr B.P.) record indicates relatively deep lake water and enhanced effective moisture relative to modern conditions (Mehringer, 1985b). Increasing sedimentation rates and vegetation changes from 8,000 to 5,400 yr B.P. indicate relatively shallow lake water and less effective moisture (Mehringer, 1985b). Effective moisture then increased from 5,400 to 2,400 yr B.P. and decreased again from 2,400 yr B.P. to the present (Mehringer, 1985b). A significant hydrologic event seems to have occurred late in the record, with drought from approximately 550 to 400 yr B.P. (Mehringer, 1985b) and possible deeper, fresher, lake water after deposition of the Mount St. Helens ash, approximately 450 yr B.P. (Davis et al., 1977).

A somewhat similar interpretation has been made for Williams Lake, a cataclysmic flood plunge pool located 19 km south of Cheney, Washington (Nickman, 1979). The cool, wet conditions of the late Wisconsinan and early Holocene were followed by warm, dry conditions approximately 7,500 yr B.P. A return to cool, wet conditions began approximately 4,100 yr B.P. (Nickman, 1979).

The paleohydrologic variations implied by the pollen records are paralleled by physical depositional systems in the Columbia Basin. Glacier Peak ashes (layers B and G) are most commonly found in fluvial or lacustrine deposits that are immediately younger than cataclysmic flood deposits (Landye, 1973; Moody, 1978).

Similarly, the Mount St. Helens Set J ash, emplaced approximately 8,600 yr B.P. at Lind Coulee (Moody, 1978), occurs in a fluvial context. In contrast, the Mount Mazama ash, with an age of approximately 7,000 to 6,700 yr B.P. (Sarna-Wojcicki et al., 1983), is found generally in eolian or alluvial fan deposits and only rarely in fluvial sediments. Episodes of soil formation between 8,000 and 4,000 yr B.P. indicate fluctuating landscape stability in this period.

Hammatt (1977) described the late Quaternary stratigraphy of the lower Snake River canyon south of Pullman, Washington. The last major phase of scabland flooding swept up these canyons close to the time of emplacement of the Mount St. Helens Set S tephra, approximately 13,000 yr B.P. Snake River flows, approximately 10,000 yr B.P., were much more dynamic than today. Occasional floods inundated the canyon floor to at least 23 m above the modern channel level. Between 10,000 and 8,000 yr B.P. the river stabilized, and point bars developed by vertical and lateral accretion. By 8,000 yr B.P., the canyon bottom was heavily vegetated. Landscape stability at approximately 7,500 yr B.P. is indicated by a phase of pedogenesis.

Snake River flood magnitudes clearly declined after 8,000 yr B.P.. Eolian activity was initiated on the upper terrace and bar surfaces that were inundated during the great early Holocene floods. Riparian vegetation disappeared from the terraces, perhaps associated with lower groundwater levels. The Mount Mazama ash fall (7,000 to 6,700 yr B.P.) occurred during this relatively dry period.

Hammatt's (1977) study indicates an increase in frequency and heights of seasonal Snake River flooding between 5,000 and 4,000 yr B.P. This period was contemporaneous with pedogenesis and stability on higher terrace surfaces. Maximum stability seems to have been established approximately 2,000 yr B.P.. Like the period from 8,000 to 7,000 yr B.P., sedimentation rates slowed and alluvial stability increased. A major paleosol was developed. After approximately 1,500 yr B.P., the landscape developed to its present state. Flooding and eolian sedimentation increased, but not to the high levels of instability that occurred prior to 8,000 and approximately 4,000 yr B.P. In general, this physical record parallels the interpretation of pollen records described previously.

Changing late Holocene flood frequency on the Columbia River has been documented from a study of paleoflood sediments at a site 8 km downstream of Grand Coulee Dam. Using techniques described by Baker et al. (1983), Chatters and Hoover (1985) demonstrated that the average recurrence interval for Columbia River floods exceeding 532,000 ft^3/s was 84 yr between 1870 and

980 yr B.P. However, it decreased to 34 yr between 980 and 560 yr B.P. and increased to 137 yr thereafter. The record implies a climatically induced nonstationarity in the flood series.

5.2.1.2.4 Glaciation through the Quaternary

Studies of ice-contact volcanic landforms and glacial deposits interbedded with lava flows and pyroclastic rocks have shown that there was extensive mountain glaciation in British Columbia during early Quaternary and perhaps late Tertiary time. One, or more, of the largest expansions of the Cordilleran ice sheet is known to be older than 1 m.y. Unfortunately, the deposits of these early glaciations are fragmentary; thus, little is known about the character of these events.

Reasonable stratigraphic and landform evidence is available only for the last three glaciations (Illinoian, early Wisconsinan, and late Wisconsinan of midcontinent terminology), all of which have occurred in the last 150,000 yr. By far, the most is known about the last glaciation, because its deposits and landforms occur at the surface throughout British Columbia. In contrast, deposits of older glaciations are less common in the province and are covered by younger sediments nearly everywhere.

Evidence of glaciation for periods prior to the late Pleistocene is scanty on all of the continents. Problems include the following:

- Preservation of records for earlier times is considerably less complete.
- Methods to date such records are less robust.
- The ^{14}C method is rarely usable for materials older than 50,000 yr.
- Few depositional basins have been identified that yield continuous, high-resolution records older than the late Pleistocene.

Five especially long sedimentary cores, from Colombia, Japan, Greece, France, and the United States, have been analyzed. These cores do not provide evidence of glaciation directly. They do tend to support conclusions drawn from other sources that climatic variability over time periods on the order of hundreds of thousands of years is dominated by orbital influences. These cores were discussed in Section 5.2.1.1.3.

5.2.1.2.4.1 Evidence of Quaternary glaciations from ice cores

Stable isotopic studies in ice cores and marine carbonates have provided estimates of changing temperatures at high latitudes and measures of the global volume of glacial ice. In ice-core studies, the empirical relationship observed between cloud temperature and the oxygen isotopic composition of snow can be used to estimate temperature changes from isotopic analysis of the water in ice cores (Dansgaard et al., 1984).

Direct evidence of Quaternary fluctuations in high-latitude temperature is obtained from oxygen isotopic ratios of ice cores taken from Greenland (Fig. 5.2-40) and Antarctica. The Greenland ice-cap record has been extended back over 120,000 yr at Camp Century (Dansgaard et al., 1984, p. 290), a high-latitude site, and to 100,000 yr at Dye 3, a more southerly location (Dansgaard et al., 1982).

The oxygen isotopic record from the Camp Century ice core clearly shows the transition from the last interglacial (125,000 yr B.P.) into the Wisconsinan glaciation and back to the present interglacial conditions during the past 15,000 yr (see Fig. 5.2-40). Dansgaard et al. (1984, pp. 292-295) performed a spectral analysis of the isotopic data and identified a quasi-periodicity of length approximately 2,550 yr.

Measurement of CO_2 concentrations in the Dye 3 core (location shown on Fig. 5.2-40) has revealed a sizeable variation of CO_2, which occurs in phase with the oxygen isotopic variations (Stauffer et al., 1983). No physical interpretation of these observations has been offered. Since the isotopes measured reflect at least hemispheric and probably global variations in these values, it must be assumed that significant fluctuations in atmospheric chemistry, glacier volume, and, hence, global climate can occur at these time scales.

5.2.1.2.4.2 Evidence of Quaternary glaciation from the ocean

The oxygen isotopic ratio of marine carbonates reflects the composition of the water in which the carbonate is formed and the temperature of precipitation. The oxygen isotopes, obtained from analysis of planktonic foraminifera (single-cell animals that live on the surface waters of the ocean and whose shell is made of $CaCO_3$) and benthic foraminifera (animals that live at the sea floor), provide a record of changing local temperature and changes in the isotopic composition of the ocean. The global nature of isotopic records obtained from a large number of sediment cores suggests that the dominant processes recorded in these sediments are changes in the global isotopic composition in the ocean and local temperature changes are relatively small.

The major mechanism that controls the isotopic composition of the ocean on time scales of a few tens of thousands to hundreds of thousands of years is the removal of isotopically light water from the ocean by glacial ice sheets. Thus, the most extensive record of global volume change can be obtained from cores of deep-sea sediments. Many thousands of such cores have been obtained to date, and cores that contain suitable carbonate microfossils can be used for oxygen isotopic analysis.

Cores have been obtained that extend the ice-volume estimates to over 4 m.y.B.P. A 1-m.y. record from core V28-238 (Shackleton and Opdyke, 1973), from the western equatorial Pacific Ocean, is illustrated in Figure 5.2-41. The detailed record of glaciation provided by the isotopic record allows the division of the last million years into stages (Emiliani, 1955). Periods of large global ice volume are even-numbered stages and periods of limited ice volume are odd-numbered stages. Detailed analysis of these records indicates that the stages, first identified by Emiliani (1955), can be further subdivided (Prell et al., 1986). Thus, the present interglacial is stage 1. The last time the ice-volume level was comparable to today's current level is isotopic substage 5e, 125,000 yr B.P.

It is now widely recognized that the dominant variance observed in the oxygen isotopic records and, by inference, the dominant controls on continental ice volume and, thus, climate has been the changes in the orbital parameters of the Earth. These orbital parameters vary in only a quasi-periodic fashion.

Over the past million years, the principal periods have been 19,000 and 23,000 yr for the precession, approximately 41,000 yr for the obliquity, and 100,000 and 440,000 yr for the eccentricity (Berger, 1978a). All but the last cycle is observed in many of the deep-sea cores that have been analyzed. Most paleoclimatic records are too short to allow detection of that longer cycle.

Imbrie et al. (1984) have concluded that the orbital theory (usually referred to as the Milankovitch Theory) is now demonstrated. They used the mathematically calculated, orbitally derived insolation curves to calibrate the deep-sea record. Imbrie and Imbrie (1980) have used the insolation curve for 65° N. latitude to reconstruct a best-fit estimate of the volume of glacial ice on the Earth for the past million years. The same equations have provided one prediction of the volume of glacial ice that can be expected over the next 100,000 yr (Fig. 5.2-42). These calculations will be described in more detail in Section 5.2.2.1.

It is concluded that the record of glaciations on a global scale provides clear evidence that glacial-interglacial cycles are the rule throughout the Quaternary. We are presently nearing the end of another interglaciation. Indeed, the most recent interglacial period reached its maximum approximately 6,000 yr B.P. Since that time, the Earth has spasmodically moved toward the next glacial maximum, which could occur in the next 20,000 to 25,000 yr (Imbrie and Imbrie, 1980; Kukla et al., 1981). The next major ice age--one as extreme as the last--is not likely to occur until approximately 50,000 yr from the present (Oerlemans and van der Veen, 1984, pp. 200-201).

5.2.1.2.5 Pre-Quaternary glaciation

A review of glaciation over geologic history is important, because it shows that, although periods of glaciation are relatively rare and not well understood, they are relatively longlasting. They usually require a certain amount of "stage setting" through alteration of global atmospheric and oceanic circulation by continental drift and tectonic uplift (Powell and Veevers, 1986). The slow tectonic processes that produce such a global configuration will naturally act slowly to bring the world back out of a glacial period. This serves to reinforce the impression, from Quaternary records, that the world is not yet likely to be beyond Pleistocene glaciations.

The deep-sea coring program has shown the present glacial episode to be much more complex than once thought. A recent oxygen isotopic chronology suggests that there have been up to 22 glacial events in continuous succession over the past 2 m.y. Over this period of the Earth's history, glaciation has been the rule rather than the exception. Glacial cycles are on the order of 90,000 to 100,000 yr in length; intervening interglaciations are relatively brief. The last interglacial period lasted only approximately 10,000 yr.

Figure 5.2-43 shows a slow and steady decline into aridity, cold, and seasonality from high temperatures and ice-free conditions (Axelrod, 1984, p. 135) in early to middle Eocene times (Keller, 1983, pp. 92-93; Dorf, 1970). The decline includes many important steps. Ocean stratification changed during Eocene-Oligocene cooling, probably related to the buildup of polar ice (Keller, 1983) and to the separation of Antarctica from South America and Australia, which permitted the onset of the circum-Antarctic current. This blocked the transfer of heat to Antarctica and, thereby, allowed its refrigeration (Diesielski et al., 1982, pp. 20-21). Ice rafting began shortly thereafter in the Ross Sea (Hayes and Frakes, quoted in Mercer and Sutter, 1982, p. 197), but buildup of the east Antarctic ice cap began 15 to 14 m.y.M.P. (Woodruff et al., 1981). Glaciation of the west Antarctic sea was delayed until approximately 8 m.y.B.P., and northern hemisphere glaciation did not begin until 2.4 m.y.B.P. (Prell, 1984; Shackleton et al., 1984).

There is very little evidence for Mesozoic glaciations, although Harland and Herod (1975, p. 192) report possible Jurassic or younger tillites in Antarctica, and the significant late Cretaceous fall in temperature was probably partly, but not totally, caused by increasing volcanism (Axelrod, 1981, pp. 22-23).

In late Paleozoic times, Gondwanaland suffered major glaciation. No more than five stadials are seen in any one locality, but the locus of glaciation shifted over time as the southern continents shifted, and glaciations waned whenever they drifted off the south pole (Powell and Veevers, 1986). This glacial episode lasted approximately 100 m.y. and affected South America, Africa, Australia, Antarctica, India, and Madagascar; although it should be noted that, because of continental drift, the Carboniferous glaciation in South America and Africa was over before the Permian Australian glaciation had begun (Kummel, 1970, p. 392). Then, as now, the position of the continents relative to the pole seems to have been very important, and broad uplift may have intensifying or triggering effects (Powell and Veevers, 1986).

A late Ordovician-early Silurian glaciation is seen principally in the Sahara, which was then over the pole. A separate but synchronous glaciation has been proposed for northern Spain and southern France (Fortuin, 1984, p. 245). This episode may have lasted 30 m.y. (Crowell, 1986).

There were at least three late Precambrian glacial episodes. Details are poorly known and correlations are very difficult, but fairly distinct episodes seem to have occurred approximately 960, 770, and 615-600 m.y.B.P. in Gnesjo, Spartian, and Varengian times, respectively (Urrutia-Fuccagauchi and Tarling, 1983, pp. 325-328). There are some possible intermediate episodes (Cook and Shergold, 1984, Fig. 1) and some earlier ones (e.g., the 2,288-m.y.B.P. Huronian Gowganda tillite north of Lake Superior (Harland and Herod, 1975, p. 202)). It suffices to say here that the Varengian, Spartian, and Gnesjo glaciations each lasted tens of millions of years and that each showed interbedding of glacial and interglacial deposits (up to a maximum of 21 advances and retreats at one site). One unique and now well-established, but not well understood, aspect of the Varengian glaciation is the product of some near-sea-level and very-low-latitude tills, possibly deposited only 8° to 15° from the equator (Urrutia-Fuccagauchi and Tarling, 1983, p. 328).

5.2.1.3 Oceanographic change

Procedures for reconstructing paleotemperatures of the ocean are now well developed, and an extensive record is available. The record of temperature changes in the oceans is consistent from site to site, and is closely correlated with the record of continental glaciations. A clear record of the correlation of orbital variations with the changes in these components is also available. The major evidence for these connections is reviewed in this section. Emphasis is placed on measured variations in sea surface temperatures. This forms the basis of models of future climate variability discussed in Sections 5.2.2 and 8.3.1.5.

The Pacific Ocean is the source of the moisture that ultimately falls on the Pasco Basin. Thus, the nature of oceanographic change in the North Pacific Ocean and its effect on the climate of western North America is of interest. Deep-sea sediments provide the opportunity to define nearly continuous climate and oceanographic records. The combination of chemical and paleontologic studies provides a means of examining local changes in oceanographic conditions and their relationship to global climate change. The continuous nature of sedimentary processes in many of the ocean basins allows detailed studies of climate change on time scales of hundreds of thousands of years with a resolution of a few thousand years.

5.2.1.3.1 Oceanographic change during the late Pleistocene

To date, only a limited number of studies have been published that examine the nature of oceanographic change in the northeast Pacific Ocean during the late Pleistocene. Reconstruction of sea surface temperatures 18,000 yr B.P., during the last maximum in global ice volume, shows that summer sea surface temperatures decreased by approximately 4 °C over much of the central-north Pacific Ocean and by at least 2 °C throughout the Gulf of Alaska (Moore et al., 1980).

The decrease in sea surface temperatures along the eastern margin of the Pacific Ocean, south of 45° N., was less than 2 °C. Winter-estimated temperatures show a 2- to 4-°C decrease throughout the central North Pacific Ocean with a decrease of at least 4 °C on the eastern margin off the Washington-Oregon coast. In general, these temperature changes represent a southward shift in sea surface temperature isotherms of approximately 15° N. latitude (Fig. 5.2-44).

The reconstruction of the glacial ocean and atmospheric models of glacial boundary conditions has provided important insight into climate change. However, these studies do not provide information about regional climate variability. Two studies of North Pacific Ocean sediment sections provide information about the oceanographic variability of the North Pacific Ocean during the late Pleistocene.

Published data from the eastern margin of the North Pacific Ocean provide a record spanning only the last 36,000 yr. Moore (1973) has shown that sea surface temperature at a site off the northern California coast increased by approximately 4 °C during the transition from the last major glaciation, 18,000 yr B.P. to the present (Fig. 5.2-45 for location of Moore study). Superimposed on this increase in temperature are variations on the order of a few degrees on time scales of a few thousand years.

Sancetta (1983) has proposed a descriptive model of the effect that glaciation might have had on subarctic Pacific Ocean oceanography and climate. She suggested that the combination of extensive sea-ice cover in the Bering and Okhotsk Seas and substantial input of water from melting icebergs in the northeast Pacific ocean would act to produce and sustain a density contrast in the upper water column, with a cold, fresh, surface layer overlying a more saline, lower layer. This would prevent vertical and lateral advection of heat into the surface waters of the northeast Pacific Ocean and result in less evaporation, with a corresponding decrease of precipitation in northwestern America. It should be noted that this latter conclusion applied primarily to Beringia (northern Alaska and Siberia) and might not be applicable to areas farther south (i.e., southern British Columbia and Washington). The degree of change in precipitation (if any) in this region is perhaps the most important problem yet to be resolved.

5.2.1.3.2 Oceonograhic change during the last interglacial

The other period for which a reconstruction of sea surface temperatures has been accomplished is that of the last interglacial interval. This time is of great interest, because it represents the most recent analog to the present (Holocene) interglacial. The maximum of the last interglacial lasted for approximately 8,000 yr (from 130,000 to 122,000 yr B.P.) and was followed by a short but intense period of glaciation, after which there was a return to relatively milder climate (Kukla et al., 1972; Imbrie et al., 1984). The Holocene interglacial has already lasted for 10,000 yr. Therefore, argument by analogy suggests that climatic deterioration may occur within the next few thousand years (Kukla et al., 1972). The last interglacial period and the subsequent glaciation, thus, represent the best analog available for forecasting climatic conditions to be expected at the Hanford Site over the next 10,000 yr. A major study of the characteristics of the oceans of the world at the maximum of the last interglaciation has been completed (CLIMAP, 1984).

The Climate Long-Range Mapping Project reconstruction shows that the mean global sea surface temperatures during the last interglacial (oxygen isotopic stage 5e) were generally approximately 2 °C higher than at present. The pattern of surface circulation was very similar to the present. Differences are notable, especially in the equatorial regions. Ruddiman and McIntyre (1976, p. 143) infer, from the position of the polar front in the North Atlantic Ocean during the last interglaciation (approximately 120,000 yr B.P.) that ocean temperatures in those areas were warmer than at present by 3 °C in the winter and 4 °C in the summer. Thus, the last interglacial period was one of the warmest climatic cycles the Earth has experienced in the past 600,000 yr. At that time, sea level was as high or higher than today, as evidenced by the Barbados III high sea level (Mesolella et al., 1969, pp. 250-274).

5.2.1.3.3 Oceanographic change at longer time scales

Studies spanning a greater time interval in the northeast Pacific Ocean include those of Sachs (1973), von Huene et al. (1976), and Sancetta and Silvestri (1984, 1986). The longest record close to the Hanford Site is that of ocean core RC10-216 (Fig. 5.2-46), which spans the last million years (Sancetta and Silvestri, 1984, 1986). Ice-rafted detritus occurs throughout the core, suggesting that there has been substantial glaciation of northwestern North America for at least the past 1 m.y., and that oceanographic or climatic conditions at present are atypical in the relative lack of ice. This conclusion is supported by a study of the diatom fossils in this core and others in the subarctic Pacific Ocean. The species recorded are presently most common in the Sea of Okhotsk, subarctic areas, or transitional areas (see Fig. 5.2-46). Sancetta and Silvestri (1986, p. 178) concluded that "in a geologic perspective the Holocene subarctic Pacific Ocean is abnormally 'warm,' and conditions similar to the Holocene have been attained only for brief intervals during peak interglacial periods of the late Pleistocene. The most common state for the late Pleistocene has been similar to conditions now existing in the Central Sea of Okhotsk; cold, fresh, and highly stratified." These authors interpret shifts in the diatom fossil assemblage to indicate periodic advances and retreats of the subarctic water mass for the last 1.5 m.y., with the greatest extremes having occurred since approximately 0.35 m.y.B.P. These changes are believed to be linked to shifts in the overlying atmospheric circulation, with warmer temperatures and greater precipitation occurring during interglacial intervals and cold dry conditions occurring during glacial times.

Unpublished results from a record spanning the last 200,000 yr show that oceanographic change in the eastern margin of the North Pacific Ocean does not follow changes in global ice volume as observed in the central North Pacific Ocean. Estimates of sea surface temperatures at this core site (43° N. 126° W.) are in progress. However, changes in the relative abundance of fossil Radiolaria species provide useful insights to the nature of oceanographic changes in this area. Radiolaria are single-celled animals that are part of the zooplankton and form a siliceous test that is preserved in deep sea sediments.

Over longer time spans embracing the entire Quaternary, the history of the oceans is known from only a limited number of cores. In core K708-7, Ruddiman and McIntyre (1976, pp. 134-135; Fig. 5.2-47) identify seven major cycles (exceeding 20,000 yr) over the past 605,000 yr. These range in length from 56,000 to 113,000 yr. In all (including the shorter episodes), 11 major incursions of polar water into the North Atlantic Ocean are recognized. Such incursions can exceed 20° latitude and represent a cooling of 12.5 °C in the winter and 13 °C in the summer. These values agree with those of McIntyre et al. (1972, pp. 590-591).

Based on the occurrence of ice-rafted terrigenous debris, McIntyre et al. (1972) estimate that there have been at least 20 significant ice sheet advances within the past 1.2 m.y. Major transitions from glacial to interglacial conditions recognized in the oxygen isotopic record are called terminations, and they provide the dates of eight such terminations (Table 5.2-7; for more recent data, see Imbrie et al., 1984). Ruddiman and McIntyre (1976, p. 136) find a close correspondence between the significant ice sheet advances recorded in core K708-7 from the North Atlantic Ocean and that of core V28-238 from the equatorial Pacific Ocean (Shackleton and Opdyke, 1973, pp. 39-55).

That correspondence is repeated in other cores, such as V16-205 from the equatorial Atlantic Ocean (van Donk, 1976, pp. 147-159). In that core, 21 glacial and interglacial stages are recognized in the past 2.3 m.y. Prior to approximately 1 m.y.B.P., the glacial stages are less pronounced. Major periods of glaciation are recognized at 750,000, 530,000, 240,000, 145,000 yr B.P., and the last glaciation. The oldest corresponds to the date of the Sherwin glaciation in the Sierra Nevada (Sharp, 1968, pp. 351-364). In

Yellowstone National Park, the Bull Lake glaciation is dated at 145,000 yr B.P. High sea levels occurred 12 times in the past 1.9 m.y., most recently at 490,000, 280,000, and 120,000 yr B.P.

The record of ice-rafted detritus at deep-sea drilling project site 178 (Fig. 5.2-48) supports the conclusion that glaciation has occurred in southern Alaska for at least the past 0.8 m.y. (von Huene et al., 1976, p. 412). In that core, the relative amount of sand and pebble-size particles were determined as a measure of ice-rafted debris. At sites that have their supply of sediments from continental sources cut off by deep-sea trenches (as in the case for the Gulf of Alaska), the only mechanism for transporting coarse-grained material is by ice rafting. Thus, these records reflect changes in the amount of ice rafting of sediments and, by interference, the amount of continental glaciation. The results of this study (Fig. 5.2-49) are in broad agreement with the record of ice advances in northern Europe as reconstructed by Morner (1972, pp. 551-557). The results show that the number of glaciations experienced around the northeast Pacific Ocean was similar to the rest of the Northern Hemisphere and indicate periods of ice advance at roughly 12,000- to 15,000-yr intervals.

Sachs (1973) also found evidence that the greatest extremes in sea surface temperatures have occurred during the last 400,000 yr. His interpretations were based on changes in the assemblage of fossil radiolarians in a core from the central North Pacific Ocean (44° N., 163° W.), spanning 800,000 yr. Estimated temperature change is on the order of 8 °C for the entire record, with the maximum temperature occurring approximately 400,000 yr B.P., and the minimum approximately 375,000 yr B.P. The general trend of sea surface temperature change follows changes in global ice volume as recorded by oxygen isotopic analysis from other core records, but the absence of calcite in the core studied by Sachs (1973) precluded a direct study of the relationship between local sea surface temperatures and changes in global climate.

A quantitative transfer-function methodology was developed by Geitzenauer et al. (1976, pp. 423-448) to estimate paleosea surface temperatures from six assemblages of coccoliths recovered in cores of late Pleistocene sediments in the Pacific Ocean. Such transfer functions will be used to allow additional reconstructions of sea surface temperatures in the northeast Pacific Ocean (Section 8.3.1.5).

An additional tool that will be important for stratigraphic correlation between the various cores as more data are made available is the SPECMAP time scale for the past 780,000 yr (Imbrie et al., 1984, pp. 269-305; Fig. 5.2-50). This time scale was developed from oxygen isotopic measurements of planktonic foraminifera in five different cores from throughout the world. The time scale was tuned and calibrated using the insolation curves of Berger (1978a, pp. 139-167). This "tuned" oxygen isotopic curve provides the best available model of the timing of ice volume fluctuations on the glacial-interglacial scale throughout the last 780,000 yr. It makes clear the quasi-periodic changes in global climate states that can be expected to continue throughout the next 100,000 yr in the absence of major anthropogenic influences.

5.2.1.4 Atomospheric change

Because of the difficulty of obtaining geologic records of atmospheric behavior, most information prior to the modern observations comes from the application of models. Thus, this section begins with a description of modeling procedures and how they are used to reconstruct the atmospheric circulation patterns of previous times.

There are a variety of approaches to climate modeling in use today. One way to differentiate between them is in terms of their complexity and another is in terms of what they focus on in the climate system. Meehl (1984) divides climate models into four general categories: energy-balance models, radiative-convective models, statistical-dynamic models, and general circulation models.

General circulation models are the most complex of the four model types, striving to model climate and circulation for the entire Earth system. The other three model types describe particular aspects of the general climate system and are generally used to investigate these aspects. As observed by Simmons and Bengtsson (1984), the simpler models also can contribute to the design of the more complex general circulation models by providing a better understanding of the processes they model.

5.2.1.4.1 General circulation models

General circulation models are those that numerically describe the processes and elements that comprise a climate system. In such models, a horizontal grid is used to describe the surface of the Earth, with the atmosphere represented by a number of vertical levels. Depending on the specific model used, the horizontal grid resolution and the number of vertical levels can vary. Thus, any location on the Earth can be described in terms of the simulated climate, at least to the level of detail that model resolution allows.

General circulation models can generate an entire reconstructed climate beyond the scope of any other branch of paleoclimate research (although it is only simulated climate). A general circulation model can generate wind fields, as well as pressure and temperature fields, and it can generate these values for any or all points on the Earth. For example, one could locate an ice sheet on North America and investigate the effect of that ice sheet upon several (vertical) levels of wind fields. At the same time, any (related) changes occurring in precipitation at locations of interest could be studied. Characteristics of general circulation models are discussed further in Section 5.2.2.2.

General circulation models can be used to investigate the known and unknown relationships, between climate elements and environmental factors of interest. For example, a pertinent research question to the study of the reference repository location is: What effect could a shift in ocean-current patterns of the North Pacific Ocean have on inland climate? A general circulation model, with an active oceanic component, can provide estimates of the winds, temperatures, and precipitation amounts that could occur in equilibrium with such currents.

A general circulation model can explore the climate system of the northwestern United States as a unique geographic configuration of ocean-land-ice elements. No modern analog of the ice age conditions in this area exists. The possibility of ice sheet growth on the land mass also can be investigated with such a model.

Model simulations of past or future climate can be used to supply information not available otherwise or supply additional information such as cloudiness, pressures, or upper level wind patterns that generally cannot be inferred from other sources.

The simulations also can provide a means of evaluating the quality of the inferred climatic conditions from paleoclimatic data and can provide a method of testing the model and calibrating it for predictions of future climates. An example of a paleoclimatic reconstruction of atmospheric circulation patterns over North America is given in Figure 5.2-51.

The models also can be used to predict future climates, especially for a variety of possible future scenarios (e.g., increased atmospheric CO_2, presence of glacial ice sheets, sea-level change, changes in oceanic circulation, and changes in the orbital parameters).

There are two methods presently used to describe the horizontal structure of the Earth in general circulation models: finite-difference (or grid point) general circulation models and spectral general circulation models. Finite-difference general circulation models represent all prognostic variables on a system of overlapping grids. These grids are generally at a spacing of anywhere from 2° by 2° to 4° by 7° in resolution. Variables representing climatic processes (e.g., temperature, precipitation, evaporation, and zonal and meridonal winds) are solved at every grid point. More complete explanations of grid point general circulation models are found in Gates (1976b) and Gates et al. (1971).

Spectral methods represent variables in terms of truncated expansion of spherical harmonics. Reviews of spectral general circulation models are provided by Bourke et al. (1977), McAvaney et al. (1978), Meehl (1984), and Simmons and Bengtsson (1984).

There is some question as to the accuracy of climate predictions from general circulation models when considered at the single-cell level. Area averaging of several cells may be found to yield a more reliable climate estimate, although it represents a larger area.

In any general circulation model, boundary conditions must be supplied. These conditions describe the environment of the global system modeled. Boundary conditions generally include the following:

- Surface character of each cell (land, water, continental, or sea ice).
- Elevation for each land or continental ice cell.
- Temperature for the remaining surface cells.
- Solar radiation amount (a function of the orbital parameters for the Earth) at the time simulated.

There are two basic categories of general circulation models. The first category, atmospheric general circulation models, models climate processes over the Earth, but treats the oceans as fixed-temperature surfaces. Atmospheric-oceanic general circulation models, the second category, are more complex than the atmospheric models, because the processes considered may include ocean circulation (Fig. 5.2-52) and interaction within the land-atmosphere-cryosphere system (Fig. 5.2-53). There are presently several ways in which the oceans are described in such models, ranging from an ocean described only with a surface energy balance to a fully computed oceanic circulation system (Meehl, 1984).

5.2.1.4.2 Uses and capabilities of general circulation models

Processes described by the general circulation models include radiation, convection, precipitation, evaporation, and advection. Pertinent elements in the described global system include land, continental and sea ice, vegetation, oceans, surface temperatures, and soil moisture.

When using a general circulation model, one attempts to arrive at equilibrium conditions for the climatic state specified by the boundary conditions. The model solves for the climatic variables at each time step. Time steps are generally several minutes in length. A certain length of simulated time is necessary for convergence of the variable values to equilibrium conditions. For different general circulation models, this time to equilibrium depends on all of the processes included in the model.

After equilibrium is reached (or at least after variables begin to converge), the variables are time averaged to provide estimates of the various climate conditions. The climatic processes are expressed for each cell as wind patterns, temperatures, pressure, and other processes (e.g., cloudiness, precipitation, and evaporation).

General circulation models are most commonly used to examine present-day and future climate concerns. Work is being done to better understand relationships between climatic variables and the Earth-atmosphere system, as this will provide improved climate model components. As these models are changed to incorporate better descriptions of the climate system, they provide better tools for predicting past and future climates.

5.2.1.4.3 Existing general circulation modeling studies

The potential climatic change for the Hanford Site and other sites includes the possibility of increased glaciation and the possibility of warming due to increased CO_2. Previous general circulation model experiments for Pleistocene and Holocene conditions and for doubling and quadrupling atmospheric CO_2 levels are of interest for understanding possible future climatic change.

The glacial conditions for 18,000 yr B.P. have been simulated with general circulation models at the National Center for Atmospheric Research, the Princeton-National Oceanic and Atmospheric Administration Geophysical Fluid Dynamics Laboratory, and the Rand Corporation, resulting in major publications (Williams et al., 1974; Gates, 1976a, 1976b; Manabe and Hahn, 1977; Manabe and Broccoli, 1984) and a number of other limited instances.

Williams et al. (1974) used a nearly grid point general circulation model with sea surface temperatures derived prior to Climate Long-Range Mapping Project compilations. Gates (1976b) used a two-level general circulation model and Manabe and Hahn (1977) a higher resolution grid point model, both with Climate Long-Range Mapping Project sea surface temperatures. Manabe and Broccoli (1984) use a spectral general circulation model for the first time to predict sea surface temperatures with a mixed-layer ocean (Fig. 5.2-54). Such a mixed-layer model treats the ocean as consisting of a number of distinct layers, each of which has a separate horizontal velocity (u and V), density defect (σ), etc. Other simulations (e.g,. with the National Center for Atmospheric Research spectral community climate model) are as yet unpublished.

The models differ in many respects: from specified to computed snow, specified versus computed clouds, no surface hydrology to incorporated surface hydrology, differences in resolution, and differences in specified boundary conditions. The primary results of these simulations include the following:

- How ice sheets displace or modify circulation (e.g., displacement of zones of cyclonic activity south of the Northern Hemisphere ice sheet margin).
- Modifications to the monsoons and increased continental aridity (general decrease in intensity of the hydrologic cycle).

- o A dependence of the intensity of the hydrologic cycle and strength of the Hadley circulation on the specified sea surface temperatures.

- o Reduced westerlies and strengthened flow in January for the Northern Hemisphere.

Further, Manabe and Broccoli (1984) reproduce Northern Hemisphere sea surface temperatures from the Climate Long-Range Mapping Project (minus changes due to shifts in currents), but achieve little Southern Hemisphere ocean-surface temperature change in contrast with Climate Long-Range Mapping Project data.

Climate sensitivity to changes in orbital parameters (solar insolation) at intervals during the last 18,000 yr, including some investigation of different boundary conditions (presence of ice), has been completed by Kutzbach and Guetter (1984a) and illustrates the importance of annual cycle amplitude and land-sea thermal contrast on monsoonal precipitation regimes. This work highlights an additional cause of model sensitivity.

Kutzbach (1983a) has done a series of investigations beginning with conditions 18,000 yr B.P. Later studies have included more recent climate states at 3,000-yr intervals (18,000, 9,000, and 6,000 yr B.P.). Kutzbach also has investigated climate at the last interglacial, approximately 115,000 yr B.P. His studies use the National Center for Atmospheric Research community climate model.

These studies have included analysis of the sensitivity of late Holocene glacial climates to changes in orbital parameters. Such changes are defined by changes in the solar radiation (Kutzbach, 1981, 1983a, 1983b; Kutzbach and Otto-Bliesner, 1982). In addition, they have examined changes in some boundary conditions, including ocean surface temperatures and ice sheet extents, and their effects on the resulting climate (Kutzbach and Guetter, 1984b).

Gates (1976b) used a two-level fixed-ocean model with the CLIMAP (1976) boundary conditions. Peteet (in Lamont-Doherty, 1984) is investigating last glacial maximum reconstructions with the Goddard Institute for Space Sciences general circulation model, comparing model results to palynological data for the same time period. (The investigators listed are only the most widely published; they are probably not the only investigators in this field.)

The climatic response to increased CO_2 levels in the atmosphere is perhaps the most extensively studied climate change using climate models. Schlesinger (1983, pp. 20-23) lists 20 general circulation model experiments for a CO_2 doubling, each with different characteristics. The surface temperature response in these models varies from 0.2 to 3.9 °C for a doubling of CO_2. The solutions are apparently extremely model dependent, although the extreme limits to the range in model sensitivity are each characterized by some unrealistic aspects.

The assumption that present sea surface temperature is fixed provides the lowest model sensitivity. The largest sensitivity response occurs in a model with strong cirrus cloud feedback; however, the present seasonal variation in observed cirrus is much smaller than present-day seasonal model response. Although the large differences in model characteristics have severely hampered intermodel comparisons, the elimination of unrealistic characteristics in model simulations reduces the range in surface temperature sensitivity for a CO_2 doubling to 1.5 to 3.5 °C.

In the Geophysical Fluid Dynamics Laboratory and National Center for Atmospheric Research model versions, the cooler the control simulation the greater the sensitivity. These results suggest that surface air temperature in the control, as it influences albedo feedback, is a major factor in model sensitivity. Models with and without ocean heat transport show similar sensitivities, if the control model without ocean heat transport is a good representation of the present climate. This consideration may result in greater convergence of sensitivity estimates in the more realistic models.

In addition to the degree of warming, the surface hydrologic balance has received considerable attention in CO_2 sensitivity experiments. Most model results show substantial regional variations in surface hydrology, and the regional changes have a strong seasonal component.

The studies cited above can provide the Hanford Site paleoclimate study with indications of (1) what is important to study most extensively, (2) information on which model version(s) are most useful, and (3) a general (starting) idea of climate variability in the Quaternary to establish whether the climate of the Hanford Site region has been significantly different in the recent past.

5.2.1.4.4 Data bases available for climate modeling studies

The following data bases have been used for model boundary conditions to aid in reconstruction of boundary conditions, or for comparison with model simulation results.

- CLIMAP (1976)--Described geography, albedo, ice extent and elevation, and sea surface temperatures. Sea surface temperatures for 18,000 yr B.P. were reconstructed from data taken from deep-sea sediment cores. Derived temperatures for the ancient ocean surfaces were the result of applying transfer functions to planktonic fossil foraminifera, coccolith, and radiolaria recovered from core sediments. The transfer functions related the fossil assemblages to modern assemblages, which were in turn related to water-mass surface temperatures.

- CLIMAP (1981)--Project members expanded the 1976 study using more deep-sea cores in the reconstructions. Results included conditions for two seasons and biogeographic zones, vegetation, sea surface temperatures, albedo, and ice elevation and extent.

- CLIMAP (1984)--This project focused on the last interglacial period, centered around 122,000 yr B.P. Results apparently include only sea surface temperature reconstructions. This may not yet be available as a data base.

- Peterson et al. (1979)--This was the predecessor to the Cooperative Holocene Mapping Project, with the continental record of environmental conditions at 18,000 yr B.P. This was not synthesized to a worldwide extent and only compiled existing data.

- Deep-Sea Drilling Project--This is still an ongoing project. Deep-sea sediment cores yield continuous geologic records, but chronological control is still not always good. Deep-Sea Drilling Project data have been vital to advances in worldwide paleoclimate studies.

- Cooperative Holocene Mapping Project--The objective for this project is global-scale climate mapping, reconstructing climate for every 3,000 yr since 18,000 yr B.P. (Kerr, 1984).

- Data bases describing present conditions are also available (NOAA, 1980, 1983).

5.2.2 FUTURE CLIMATIC VARIATION

The Annotated Outline (DOE, 1985) calls for this section to include a discussion of what can be expected in the future with respect to the following:

- Changes and rates of change of precipitation and air temperature.
- Windflow and precipitation patterns.
- Potential for glaciation.
- Fluctuations in sea level and the cryosphere.

No estimates of future values for these factors have been published that are specific to the Hanford Site. Estimates of future values of the first two items have not been attempted for the time span of interest for nuclear waste disposal. For the last two items a few estimates of long-term variation have been published that include estimates of global ice volume up to 100,000 yr in the future. These will be discussed in this section.

Because of the paucity of information available to answer the questions posed by the Annotated Outline (DOE, 1985), emphasis in site characterization will be placed on the development of methods capable of providing such information and using those methods to generate the needed data. Available methodologies that can be adapted to the purpose of site characterization are discussed in this section to rationalize the plans for characterization that are recommended in Section 5.2.3 and presented in detail in Section 8.3.1.5.

This section discusses methods that are available to help forecast the future behavior of the climate system and estimate the climates that should be expected at the Hanford Site. Because of the multitude of factors that influence such climates (Fig. 5.2-55), no unified theory of climate change is available. Thus, the models available for the various components are discussed separately.

First, methods to forecast the future global climate state are reviewed (Section 5.2.2.1). Methods to compute the response of the atmosphere to changed global boundary conditions are then reviewed (Section 5.2.2.2). Since the cryosphere forms one critical boundary condition, modeling methods to characterize its dynamics are reviewed (Section 5.2.2.3).

To provide a description of the local response of the climate system to the broader regional or global changes, a more detailed model of climate at the Hanford Site is required. Available modeling procedures, their limitations, and needed improvements are described (Section 5.2.2.3). This includes a description of modeling procedures appropriate to the definition of the response of the local surface water (hydrologic) system to these climate changes. Included among these responses are the major flood events associated with the draining of proglacial lakes.

5.2.2.1 Forecasting future climates

This section examines the known and hypothesized controls on climate change at three different time scales. Based on the current wording of 10 CFR 960 (Federal Register, 1984, pp. 47761-47762), relative to the question of favorable conditions of climate change, climatic variability over the next 100,000 yr must be considered during site characterization. Postclosure stability assessment must consider climates from 100 to 100,000 yr in the future. Thus, the time scale of concern spans three orders of magnitude: (1) 100 to 1,000 yr, (2) 1,000 to 10,000 yr, and (3) 10,000 to 100,000 yr. The paleoclimatic record of the past 1 m.y. shows that each of these time scales is characterized by different controls on climate and by a different magnitude of variability of climate (Fig. 5.2-56). Even the shortest period (i.e., 100 to 1,000 yr) extends beyond the range of all but a very few human records of climate. Thus, the primary evidence for climate change at each of these time scales is geologic in nature.

These different time scales are characterized by distinct quality and extent of evidence of climatic change. Obviously, the best evidence about climatic change at the time scale of 100 to 1,000 yr comes from the most recent 1,000 yr of the record. However, during site characterization, concern must be placed on the variability to be expected in each 1,000-yr period considered, 100 of them in all. It is not at all certain, indeed it appears unlikely, that the variability of climate in each 1,000-yr period is the same. It is not certain that the variability of climate in the ninth 1,000-yr period in the future will be the same as that of the eighth, and so on; therefore, it will be necessary to characterize the variability of climate in a typical 1,000-yr interval and the amount by which that variability may change from one such interval to another. The same questions arise for the intermediate time scale (1,000 to 10,000 yr). A slightly different form of the question must be addressed for the longest time period of interest. That point will be further discussed in Section 8.3.1.5.

It will be assumed during site characterization that climatic changes over shorter time spans (i.e., years to decades) do not affect site stability, regardless of the time in the future at which such changes may occur. Only those climate conditions that persist over longer periods, 100 yr or more, can affect the Hanford Site. It is assumed that the principal means by which climatic change could affect Hanford Site stability is its influence on the groundwater system. Because of the low permeability of the repository rocks, the low recharge rates, and the large transport distances involved, climatic changes that are very brief (100 yr or less) will not produce measurable changes in repository stability. Therefore, only climatic conditions integrated over 100 yr or more will be considered during site characterization.

This does not mean that the characteristics of shorter term events will not be considered. Climate is merely the integration of shorter term meteorologic events. Thus, some representation of the types of precipitation events that typify a given 100-yr (or longer) time period must be made. Whether storms tend to be long, gentle rains (in which case a great deal of recharge to the groundwater system may be possible) or short, intense events (in which the same amount of rain may fall but most would go to runoff) has a big influence on the hydrologic system. Characteristics of meteorologic phenomena in each time interval will be summarized (as described in Section 8.3.1.5.1) using statistical measures including frequency distributions (Chervin and Schneider, 1976) and stochastic process models (Box and Jenkins, 1976, pp. 1-19; Jenkins and Watts, 1973, pp. 209-257).

5.2.2.1.1 Global climatic change

Climatic change in the Quaternary can be divided into two theories: (1) those in which the changes are due to perturbations of various components of the terrestrial system and (2) those in which the changes are external in nature. Included in the former are such phenomena as:

- o Extreme volcanic events.
- o Fluctuations in the amount of CO_2 in the atmosphere.
- o Influences of ice sheets on the global heat balance.

Sources of climatic variation external to the Earth include changes in the solar constant due to either of the following:

- o Variations in the orbit of the Earth around the Sun.
- o Change in the energy balance of the Sun.
- o Masking influence of dust in the orbit of the Earth.

The first of these theories has found wide acceptance in the scientific community.

It is widely accepted that no single theory explains all of the variation in climate that has occurred throughout the history of the Earth. A number of differing mechanisms have operated, and continue to operate, on the climatic system. The dominant mechanism of concern depends on the time period of interest and the resolution that is needed. This section will review the various controls on climatic change that have been discussed in the literature. Most modern theories of climatic change on this time span involve computer models. A discussion will be given of the major aspects of such models that lend themselves to use in forecasting the stability of the proposed repository.

Over extremely long time spans (a billion years or more), the chemical and physical evolution of the Sun dominates the energetics of the solar system. On the scale of hundreds of millions of years, it is the chemical evolution of the Earth's atmosphere that has exerted the most pronounced effects on the climates of the continents. On the time scale of tens of millions of years, the crustal dynamics of the Earth have rearranged continental configurations, modifying the oceanic and atmospheric circulation patterns, and thereby changing the organization of the heat balance of the Earth. Over the latest several millions of years, subcontinental-scale tectonics, including the closing of the interocean circulation at the Isthmus of Panama and the continued uplift of several major mountain ranges, have exerted the strongest influence on the changing climate of the Earth. All of these processes act slowly relative to the time scales of interest for nuclear waste isolation.

Where the time scales of interest include periods from 100 to 100,000 yr, the dominant controls on climatic change are of a different nature. Principal among these controls, and the one that lends itself most readily to precise estimation, is the variation in the characteristics of the Earth's orbit. The orbital parameters of interest are the precession, obliquity, and eccentricity. There is reasonable agreement in the scientific community that approximately 60% (±10%) of the variability of climate--at time scales ranging from 2,000 yr to nearly 1 m.y.--is correlated with variations in these three parameters (Berger et al., 1984, p. x). Physical mechanisms capable of explaining the correlation have not been convincingly demonstrated; at present the connection is a statistical one.

Whatever the actual connection, it is clear that it must be a highly nonlinear one. Reconstructions that suggest the form of the nonlinear relation are of two major types: (1) energy balance and (2) those with ice sheets and (or) a CO_2 cycle. Although numerical models of atmospheric and oceanic circulation have increased in sophistication, they are not able to integrate the critical equations of continuity over the needed time spans, nor are they able to fully represent the complete system of the cryosphere and the chemistry of the atmosphere. Empirically calibrated, quasi-physical models are the most sophisticated now available to represent climate change over time spans of thousands of years or more.

Energy-balance models are zonally averaged equilibrium representations of the thermodynamic balance of the Earth. Early models failed to represent the nonlinear response of the Earth to the orbital forcing. Because of the significant phase lags observed in the oxygen isotopic record as compared to the orbital solutions, it is unlikely that simple, equilibrium-type energy-balance models will provide an adequate representation of climatic change at the time scales of interest. The most successful models to date are those that represent the seasonal fluctuations of insolation, usually including several atmospheric layers.

More useful results have been obtained with models that represent the feedback effects of the large ice sheets, especially those that grow in the Northern Hemisphere. Such ice sheets exert a number of indirect influences on climate. The ice ties up huge quantities of water, resulting in a lowered sea level. This allows further growth of the ice, especially in coastal areas where that growth is nurtured by the proximity of the maritime moisture sources. During growth of the ice sheet, a further feedback arises from the increased albedo of the ice. Because of this, the energy balance at high latitudes becomes more negative and temperature gradients become accentuated.

The weight of the ice isostatically depresses the continent below. When melting of the ice begins, it is accentuated by the calving at the coast. The crust remains depressed, recovering only very slowly (crustal relaxation time is on the order of 4,000 yr). The Pollard model (Pollard, 1982, pp. 30-48; 1984, pp. 541-564), by incorporating this effect, has successfully reproduced the observed, approximate 100,000-yr cycle in ice volume change as reconstructed from the oxygen isotopic record.

More recently, evidence of significant fluctuations in the CO_2 content of the atmosphere over the last glacial cycle has prompted the development of models capable of representing the infrared heat flux. Such models appear to be capable of explaining the temperature extremes documented through the Quaternary. Several atmospheric levels and an interactive oceanic component, with at least a diffusive transport term, are required in these models. Linking of the ice dynamics and the chemical cycle models has been achieved. Results of these models appear to be quite reasonable.

All such models are driven by the variations in orbital parameters. Thus, adequate modeling of climate change at these time scales requires that solutions of those parameters be available. Such solutions cannot be achieved in an exact sense. There is no analytic solution to the orbital equations, only numerical approximations. At least seven different solutions have been published in the past 20 yr. Each differs in the degree and order of the approximation (Table 5.2-9).

The most exhaustive solution has been achieved by Berger (1978b). Berger's solutions have been extensively applied to a number of recent modeling efforts. They appear to provide acceptable accuracy for at least 1 m.y. in the future as well as the past (Berger, 1984, pp. 3-39). Thus, Berger's solutions can be used for the assessment of long-term climatic change at the Hanford Site.

Simple energy-balance models have been found unacceptable in representing long-term climatic variability. Observed temperature fluctuations are not reproduced. Seasonal energy-balance models, especially those with a multilevel atmosphere, yield improved estimates of temperature fluctuations. No equilibrium energy-balance models are capable of reproducing the approximate 100,000-yr oxygen isotopic record.

Nonlinear amplification of the eccentricity term can be achieved using more realistic ice sheet models. Critical components of these models are the albedo feedback of mid-latitude ice sheets and the isostatic lags. The latter is accentuated when iceberg calving is allowed (Denton and Hughes, 1983). Even these relatively sophisticated models are incapable of completely reproducing the extreme temperature fluctuations of a glacial cycle, and they do not completely explain the coupling of the response in the Northern and Southern Hemispheres. The addition of terms describing the changes in CO_2 content of the atmosphere due to its increased solubility in the cooler oceans provides a satisfactory description of climatic variability on the time scales of several thousand to 100,000 yr.

Exact analytic solutions of the governing equations of the cryosphere-atmosphere-ocean system are not possible, even for modern conditions. This difficulty is magnified for other times for which the relevant boundary conditions are poorly understood. Even numerical approximations can introduce important errors due to truncation errors and computational instabilities (Birchfield and Weertman, 1984, pp. 605-606). Uncertainties arise from the form of the approximation, the boundary conditions, the initial conditions, and the computational characteristics. However, it is not possible to provide general rules for the final level of precision of reconstructions or forecasts made using these procedures.

It can be assumed that the ultimate precision of the procedures will not exceed the 60% observed match between the orbital forcing function and the oxygen isotopic record of the climatic response for the past 500,000 yr. The greatest uncertainty arises from the form of the model used. It can be expected that a seasonal, thermodynamic, two-dimensional model, having a dynamic ice sheet with an isostatic adjustment and a calving margin, and linked to a two-level atmosphere with a CO_2 cycle will provide a defensible estimate of long-term global climatic change. Such a model will set the general global conditions that define the character of the more specific global climatic system and the detailed local response to climatic change.

5.2.2.1.2 Short-term climatic change

Over time spans ranging from 100 to 1,000 yr, climatic change in a single area can be considerable. In the past 1,000 yr, the Northern Hemisphere has seen a series of minor readvances of glaciers that lasted for 100 yr or more. Other periods of similar length were characterized by mild or even warmer climates. Some examples of such short-term excursions are discussed below.

Bryson et al. (1970, p. 53) suggest that frequent, rapid changes from moist to dry conditions occurred numerous times throughout the Holocene in the north-central United States. Dean et al. (1984, p. 1193) found such a pattern at Elk Lake, Minnesota, where fluctuations of clastic input to the lake occurred within 200 yr. These are explained by changes in lake levels due to climatic variations.

Brubaker and Cook (1983, pp. 226-28; see also Fritts et al., 1979) find that "temperatures in the 18th century were generally above the 20th century mean" in the western United States. They also state that the Pacific Northwest was wetter before 1900 than it is at present. LaMarche (1974) documents several periods in the late Holocene when climates switched between warm and cool over time spans from hundreds to several thousand years. More detailed data for the past 1,000 yr show fluctuations of temperature and precipitation over time spans from 100 to 300 yr in length (Fig. 5.2-57).

Fritts (1985a) data from tree rings of the Columbia Basin show clear differences in the variance of temperature from one century to another since the year 1600 (Fritts, 1985a, p. 13; Fig. 5.2-58). In the Pasco Basin, Cropper and Fritts (1985, p. 4) conclude that temperatures were warmer and more variable over the last three centuries than they are today. Precipitation was greater and more variable also. This suggests that modern instrument data based on the most recent 30 yr of record will produce a biased estimate of recharge to the groundwater system and should not be used to the exclusion of other data for groundwater modeling. Such oscillations of climate at these time scales (100 to 1,000 yr) must be considered in the assessment of Hanford Site stability.

Few physical explanations of these oscillations have been offered. Indeed, it is only over the past decade or so that the phenomenon has been recognized. Certain links in the climatic system are capable of varying over these time scales. These include a vegetation-albedo feedback (Rind, 1984, pp. 73-91), intense volcanic activity (Porter, 1981, pp. 39-142), and changes in the concentration of CO_2 in the troposphere (Broecker and Peng, 1984, pp. 327-336; Broecker and Takahashi, 1984, pp. 314-326).

There is a known relationship between the dynamic response of vegetation to climatic change and the resulting feedback to the climatic system, since albedo is dependent on vegetation (Hartmann, 1984). Inertia in the vegetative system means that several hundred to 1,000 yr may pass between the initial climatic change and the later feedback on albedo (Solomon et al., 1981, p. 154). Vegetation also exerts an effect on the soil moisture, the roughness of the surface, and the effect of snow cover on the albedo. Thus, adequate understanding of climatic changes in an area at the time scales of 100 to 1,000 yr requires that the response and the influences of vegetation be considered.

Intense volcanic activity can affect climate through the injection of CO_2 and other aerosols into the atmosphere. Such gases can lead to an increase of temperature worldwide (Sarna-Wojcicki et al., 1983, p. 52). Increases in sulfuric acid or the volume of fine glass shards in the upper atmosphere can actually produce a long-term cooling trend following an initial slight warming (Davitaya, 1969; Lamb, 1970, 1977, pp. 195-196; Pollack et al., 1976). It has even been suggested that volcanic activity may be a triggering mechanism for glaciation (Bray, 1974, 1977; Kennett and Thunell, 1975; Porter, 1981). Because the location and timing of such volcanic activity cannot be accurately predicted, climatic change due to this phenomenon must be considered a random variable.

Changes in the CO_2 composition of the atmosphere do occur over time spans shorter than 1,000 yr. Indeed, anthropogenic changes in CO_2 have occurred in just 100 yr. The principal evidence for such changes that has been observed in the geologic record suggests a longer scale of operation. Therefore, that phenomenon will be discussed in the next section.

The climatic fluctuations throughout the past several thousand years are complex in nature and still poorly understood. At this time scale (1,000 yr) and level of resolution (decades), much of climatic variability must be considered to be a stochastic process.

5.2.2.1.3 Intermediate-term climatic change

All of the phenomena discussed in the previous section are at work over the longer time spans considered here. Other factors also come into play when considering the range of climate that may be experienced over a 1,000- to 10,000-yr period. Included among these factors are various dynamic behaviors of specific components of the climatic system.

Mixing periods of the oceans following an external forcing are typically on the order of 1,000 to 1,500 yr (Gordon, 1975; van Donk, 1976, p. 147). Deep-ocean water circulation is critical in the redistribution of the energy budget of the Earth. During the glacial epochs, the North Atlantic Ocean deep-water pattern of circulation changes in velocity and depth. New circulation patterns may take thousands of years to become established. Other intermediate response times arise from chemical factors. Equilibration of the CO_2 concentration in the atmosphere and the oceans may require over 1,000 yr or more (Broecker and Peng, 1984, p. 327).

The influence of ice can produce various lags and oscillations in the climatic system. Complete disintegration of an ice sheet can take 5,000 to 10,000 yr (Barry, 1985). Other periodicities also may be present in the climatic system. Dansgaard et al. (1984, pp. 288-298), through a spectral analysis, have identified a 2,550-yr periodicity in paleoclimatic data from the Dye 3 and Camp Century ice cores of the Greenland ice sheet. No mechanism has been suggested for that phenomenon.

Barry (1985, pp. 259-290) discusses other snow- and ice-related factors of climatic change at the time scales of 100 to 100,000 yr and longer. Several distinct thresholds exist in this system. Snow fields may require hundreds or thousands of years to build up to a depth sufficient to flow under their own weight. At that point, these glaciers become an active component of the climatic system.

Similarly, the freezing temperature of ice is a threshold for two important phenomena. If the basal temperature of ice is at or below the pressure-melting point, the glacier bed will be frozen fast to the bedrock below, and no ice movement will occur at the bedrock contact. Above the pressure-melting point, basal sliding will occur. This can be an important component of the total ice velocity.

Thus, the temperature of the ice, in relation to the pressure-melting point, will impart an important control on the dynamics of this part of the climatic system. This temperature, in turn, reflects the integrated response of very long-term trends in mean surface temperature. The Earth's crust responds to increased loading by ice. Depression of the crust occurs with a time constant of 4,000 to 5,000 yr. This depression will lower the elevation of the ice surface with respect to the snow line. The snow-line elevation is itself a function of incident solar radiation. Such interactions are summarized in recent model results for the last 120,000 yr in North American ice sheet growth (Fig. 5.2-59).

Another influence of the freezing-point threshold is the growth of ice shelves. As sea surface temperatures cool during a glacial cycle, they may reach the freezing point of the (saline) sea water. As ice shelves form, they create a barrier to evaporation of sea water, removing an important source of precipitable water for the growth of glaciers. The ice shelves also affect albedo and surface friction. With great thermal inertia (high heat capacity), and because of mixing, ocean waters may require more than 1,000 yr of cooling temperatures before significant ice shelves will form.

Although the last deglaciation began 18,000 to 15,000 yr B.P. (it began at different times at different locations), it was a relatively sudden occurrence that may have involved several steps (Ruddiman and Duplessy, 1985, p. 15). Calving of ice shelves and ice sheets exposed to marine conditions, because of the combination of rising sea levels and continental bedrock still isostatically depressed, may have been a major factor in the rapid wasting of the ice sheets in the early stages. Because of this, the sea level rose faster than the isostatic rebound of the crust (Bloom, 1983). The relative rates depend on the geographic location. The coast of Washington is located in zone II (Fig. 5.2-60).

A number of authors have referred to the self-forcing features of the global climatic system. Among these features are the nonlinear feedbacks between components that have differing response times. Such connections can produce quasi-stable oscillations of the climatic system, even when it is near a stable point. Thus, the climatic state may not actually display a stable point. Rather, it will be in a constant state of flux in which the stabilization of one component forces others to oscillate more. The more sophisticated of these formulations suggest that the integrated climatic state of the Earth will display a red-noise spectrum because of these complex interactions (Hasselmann, 1976; Nicolis, 1984; Saltzman, 1982).

The precession of the seasons is defined by the change in the longitude of perihelion of the Earth's orbit. It is usually expressed as the sine of that value. The half-period of the precession term is approximately 11,000 yr. Over that time span, the Earth will experience a reversal of the seasonal distribution of insolation. At present, the Earth nears perihelion in January. Thus, summer insolation is now near a relative minimum. Eleven thousand years ago (or in the future), summer insolation in the Northern Hemisphere was at a maximum (Ruddiman and Duplessy, 1985, p. 13).

A continuous change between the two extremes occurs over each 11,000-yr half-period. The influence of the precession term depends on the concurrent value of eccentricity. For periods in which the eccentricity is low (approximately 25,000 yr after present), the precession effect is negligible (see Fig. 5.2-58, top). When eccentricity is at its maximum (approximately 100,000 yr after present), the variation of the seasonal distribution of insolation over the half-period of the precession term is most extreme.

Over a longer period (approximately 40,000 yr), the tilt (obliquity) of the Earth's axis relative to the plane of the ecliptic also changes by as much as 2° (Fig. 5.2-61, bottom). Obliquity exerts an influence on the amount of solar insolation received at various latitudes at different times of the year. This also results in changes in climate. Such effects are documented in the following section.

5.2.2.1.4 Long-term climatic change

A single hypothesis for climatic variability at the 10,000- to 100,000-yr time scale has come to be accepted as clearly explaining a significant (and major) fraction of the observed variance. It is believed that variations in the seasonal distribution of solar insolation at the Earth's atmosphere, due to changes in the Earth's orbit, lead to the cycles of glaciation that have been observed during the Quaternary; this is known as the Milankovitch Theory. The concept had been discussed by Croll (1867) and by Adhemar (1842), but Milankovitch (1941) was the first to demonstrate quantitatively the reasonableness of the link. Exact physical mechanisms for the effect have not been established.

Hays et al. (1976, pp. 1121-1132) showed that oxygen isotopic fluctuations have strong coherence with the orbital variations of the earth over the past 780,000 yr. It is now seldom disputed that the Milankovitch Theory correctly explains a significant fraction of climatic variability over the past 1 m.y. at a scale of resolution of approximately 2,000 yr (Imbrie and Imbrie, 1980, pp. 943-944; Imbrie et al., 1984, pp. 269-270).

A key paleoclimatic signature that has been analyzed is the ratio of two isotopes of oxygen (Ruddiman, 1985; pp. 197-257). This ratio is affected by a number of processes; the most important of these is the preferential incorporation of the lighter isotope of oxygen in precipitable water. This water falls as snow and ultimately becomes incorporated into ice sheets. The oxygen isotopic ratios observed in the tests of micro-organisms preserved in deep-sea sediments are a surrogate measure of the volume of ice on the continents. Thus, the signal of global-scale ice volume changes varies coherently with orbital variations and has been sufficient to establish the Milankovitch Theory.

The orbital variations of the Earth can be computed from orbital theory. There are three parameters that must be considered: precession, obliquity, and eccentricity. Although the equations describing their variation over time can be specified, they cannot be solved analytically; solution is by numerical approximation. Usually, the equations are expanded in infinite trigonometric series, and the series is truncated at a given number of terms. The order of the approximation and the degree of the polynomial terms will fix the accuracy of the resulting solution. To date, at least seven such solutions have been published (see Table 5.2-8).

Solutions are of value for times symmetrically displaced from the initial point at which the expansion was made. This is usually taken to be 1950 A.D., a date for which the initial values are known. Thus, it is possible to predict accurately the orbit of the Earth for the next 100,000 yr. The best solutions to date may be usable for 1.5 m.y. or more (Berger, 1984, p. 3). On the other hand, certain published solutions have been found to yield unreliable answers even within a 100,000-yr period.

The solutions of Berger (1978a) show that the variations in the Earth's orbit are quasi-periodic with the major spectral power concentrated in certain specific bands. Eccentricity varies at two different cycles: approximately 100,000 and 410,000 yr. Obliquity varies at one fairly tight band centered at 41,000 yr. Precession also shows power concentrated at two bands that are quite close: 23,000 and 19,000 yr. Thus, it is not at all uncommon to see the precession term associated with a single 20,000-yr cycle. It is important to realize that even these five frequencies are but abbreviations of an extremely complex pattern of variation that, strictly speaking, is not at all periodic.

Despite the fact that the link between climate and the Earth's orbit is now well established, the mechanism of that link has not been resolved. The precession of the Earth will lead to a seasonal variation in insolation of approximately 7% over a complete cycle. The eccentricity term, the only one that actually can produce a true change in the total insolation of the Earth, can only affect that insolation by approximately 0.3% over its maximum range.

In the absence of an exact mechanism, a number of strategies have been employed to express the relation in a manner that successfully reproduces the observed record of Quaternary climate. Such attempts include conceptual and mathematical models. Because of their greater potential as a predictive tool, discussion will be focused on the latter.

No empirical analysis of the relation between climatic variation and the orbital forcing function has suggested that the Milankovitch Theory is capable of explaining more than approximately 85% of the variance of climate. That explanation only extends, at best, to variability at time scales greater than approximately 3,000 yr. Thus, any model of climate change at the 10,000- to 100,000-yr time scale must recognize the portion of climatic variance that is not explained. Such variance might be treated as a white-noise term (i.e., a random variation without a temporal structure of its own). Alternatively, the uncertainty could be modeled as a red-noise term.

5.2.2.2 Modeling future global climatic change

The forecasting of future climates will be based on models of anticipated atmospheric circulation patterns. General circulation models provide the most comprehensive approach to global climatic modeling (see Section 5.2.1.4). The discussion in this section is limited to general circulation models. Useful discussions of all types of climatic models and of climatic modeling in general can be found in Smagorinsky (1974), Gates (1982), Meehl (1984), and Simmons and Bengtsson (1984).

5.2.2.2.1 Models of global atmospheric and oceanic circulation

Climatic models provide a mathematical representation of the physical processes important for determining climate. Such mathematical models replace the complex natural system by a simplified one that can be used as a quantitative tool. Climatic models range from highly simplified calculations, for which globally averaged surface temperature is the only predicted variable, to three-dimensional time-dependent models that are capable of predicting variables such as temperature, pressure, winds, precipitation, and evaporation.

General circulation models represent the most complex and realistic extreme of this climatic model spectrum. These models are based on a set of physical laws (first law of thermodynamics, conservation of mass, equations of motion, continuity equation, and equation of state) that are transformed into a predictive model. The characteristics of general circulation models can vary considerably depending on computational requirements, differences in degree of parameterization, and manner in which the more slowly evolving components of the climatic system (e.g., hydrosphere, cryosphere) are incorporated.

The numerical scheme of general circulation models is usually either finite-difference (grid point) or spectral (truncated spherical harmonics). Spectral models are computationally more efficient than grid point versions. The model resolution may vary from hemispheric to global domains, from simplified geography to realistic geography, from 1° to 10° horizontal resolution, and from 2 to 12 or more vertical levels. The computational requirements generally vary as the cube of the spatial resolution.

The parameterized components of atmospheric general circulation models that may differ include cloudiness, snow cover, rainfall, subgrid scale processes (e.g., convection, boundary layer processes), surface hydrology, vegetation, and radiation. For example, cloudiness may be specified or computed based on humidity and (or) static stability. The radiative properties of specified or computed clouds also may differ.

Ocean processes are incorporated in one of several ways:

- o Sea surface temperatures may be prescribed.
- o Sea surface temperatures may be computed from energy-balance considerations (no heat storage or motion).
- o Heat storage may be incorporated in a mixed-layer ocean (no motion).
- o Oceanic processes may be computed with a fully resolved oceanic general circulation model.

Sea ice may be specified and is computed based on thermodynamics alone or with some dynamics included.

The need to simplify the climatic system by limiting spatial resolution, parametering processes, or specifying variables introduces uncertainties. Models can be verified only against observations of the present-day climate and, in a limited sense, with paleoclimatic data if the climatic forcing factors are well known. Although climatic models are necessarily simplified, the complexity of general circulation models and the number of differences between models severely hampers intermodel comparisons. These limitations and the nature of the problems to be investigated guide the philosophy of general circulation model use.

The nature of the problems investigated with climatic models generally entail one of four objectives:

- o Identification of physical mechanisms of climate change.

- o Identification of conditions that maintain circulation or climate characteristics.

- o Reconstruction of a global climatic pattern.

- o Assessment of model responses to changes in process or boundary conditions.

The latter two objectives are the most applicable to questions of climatic change at the Hanford Site and at other proposed nuclear waste sites.

In the reconstruction of a global climatic pattern, there exists a range of possible simulations. All boundary conditions can be specified, including the following:

- o Sea surface temperature.
- o Sea ice.
- o Permanent snow cover.
- o Vegetation.
- o Surface albedo.
- o Geography.
- o Solar insolation.

The climatic model then produces a physically consistent solution within model limitations.

The accuracy of the model simulation is judged primarily through its ability to simulate the present climate. General circulation models at the major United States climatic modeling centers do an excellent job of reproducing present atmospheric characteristics.

The number of boundary conditions specified also may be limited to incorporate the prediction of some characteristics (e.g., sea surface temperatures) that can then be used independently in model verification.

In the reconstruction of a global climatic pattern, success depends on knowledge of the factors that influence the climate (e.g., whether the climatic forcing factor is solar variation or sea surface temperatures). Obviously, in future climatic predictions and most paleoclimatic modeling, many of the boundary conditions are unknown or subject to uncertainties. Consequently, much research centers on the assessment of model response or climatic sensitivity. There are two items of concern in climatic sensitivity analyses:

- o The quantitative change of climatic variables in response to external conditions (e.g., extent of a polar ice insolation change) or a change in a model process (e.g., a predictive cloudiness scheme).

- o An assessment of whether the climatic change in response to some variable is significant.

Sensitivity analysis uses a control simulation (e.g., present data), modifies a single variable (if possible) in an experimental simulation, and compares the results of the two simulations. Multiple experiments are performed over a range of possible variation. Extreme variations are frequently examined, because large perturbations will give the clearest model response and will clarify whether the suite of smaller variations is likely to be significant. The greatest advantage of sensitivity experiments is to gain insight into the potential for climatic variability and to understand the model response to changes in specific variables that may influence climate.

Sensitivity analysis, which is essential if boundary conditions are unknown or uncertain, benefits from multiple experiments. This requirement tends to result in computational limits. For example, a seasonal mixed-layer ocean atmospheric general circulation model requires enormous computational expense to reach equilibrium (3,000 d), and research would be limited to one or two experiments per year. For the same computational expense, sensitivity experiments with an atmospheric general circulation model with fixed insolation (January or July) can be completed in approximately 300 d per simulation for a range of variables (e.g., sea surface temperatures, solar insolation, glacial extent).

An assessment of the significance of a model response generally involves analysis of whether the model signal is detectable above the level of meteorologic or model noise. The most frequently applied statistical technique used in general circulation model simulations involves calculating standard deviations for time-averaged model means for key variables. Significance testing is based on the classical student's t-test. These techniques help establish the confidence by which a difference in an experimental simulation can be ascribed to a specific prescribed change in comparison with a control.

5.2.2.2.2 Problems of applying climatic models to studies of future climates

There are several fundamental problems in using general circulation models for such climatic studies, and each problem could have an impact on resulting simulated climate. These problems apply to any target site investigated with a general circulation model and are not unique to the Hanford Site.

Uncertainty in the boundary conditions specified, such as the estimated sea level lowering at the last glacial maximum or the generalized vegetation cover for the Earth during the same period, affects the simulated climatic estimates. Uncertainty is unavoidable in such reconstructions, yet it cannot be disregarded in the final estimates of climate from models requiring boundary conditions.

It is also assumed that processes that are generalized in the model (e.g., hydrology) may introduce some uncertainty into the climatic estimates due to the abstractions involved.

Uncertainty due to the boundary conditions is another type of variability to note. Data supplied for use as boundary conditions may not be expressed at the same resolution required by the model. In most cases, the model resolution is several times coarser than the data provided. In this case, interpolation to the desired spacing must be done. When this occurs, some accuracy is lost and uncertainty is added.

For example, with the Rand grid point two-level model, one 4° by 5° model grid cell must be resolved to one surface type from five 2° by 2° data cells, which may be of various surface types. Choice of values is a serious problem with regard to uncertainty, especially in continental margin regions. Such averaging can alter the character of surface cells in the model boundary conditions, which changes the continental configurations and affects the resulting simulated climate.

Climatic models still possess a "black-box" status: there is an element of uncertainty in the model sensitivities. It is difficult to evaluate model uncertainties and weaknesses due to practical constraints on model simulations. These constraints include the amount of time necessary to run general circulation model simulations of climate and the cost of the required computing.

The use of coupled ocean atmospheric general circulation models is still in the early stages. This type of model has not been used for paleoclimatic work to date (as far as known) and so the margin of uncertainty for model performance must be added to the uncertainty of the results.

5.2.2.2.3 Models of future global glacial change

Two approaches have been used in computer simulations of ice sheets: one approach constructs mechanical models, the other constructs geomorphic models. Each approach is discussed in this section.

5.2.2.2.3.1 Mechanical models

Input to mechanical models includes the temperature and ice accumulation-ablation pattern over the surface, and topography at the bed. In the time-dependent versions, dating from Mahaffey (1976), input also includes Milankovitch Theory variations of insolation over the ice sheets. These variations are used to parameterize temporal and spatial variations in the surface equilibrium line and the "elevation desert effect" observed over the Greenland and Antarctic ice sheets today and to show precipitation patterns over former ice sheets predicted by atmospheric general circulation models for ice-age boundary conditions. This input permits the simulated ice sheet to respond to variations in rates of ice accumulation and ablation with altitude, latitude, longitude, and time, all of which are specified at grid points of a finite-difference numerical scheme. Points range from 10 to 200 km (6.25 to 125 mi) apart in these models, depending on what resolution is desired. A 10-km (6.25-mi) spacing is not practical for simulating advance and retreat of continental ice sheets, however, and a 100- or 200-km (62.5- or 125-mi) spacing has been employed. Isostatic rebound data are used to fit empirical equations that predict crustal adjustments at each grid point during advance and retreat. Output from these time-dependent mechanical models includes variations of ice sheet elevation, thickness, and area over time, which is a full 100,000-yr glaciation cycle in many simulations.

The first mechanical model of this type was developed by Budd and Smith (1981), who used it to simulate advance and retreat of North American ice sheets over the past 120,000 yr. By tuning the numerous parameters within reasonable limits, they were able to reproduce most of the main events of the last glaciation. Because of the coarse grid (200 km (125 mi) between points), however, the model nucleated ice sheets that covered all of Alaska and the Cordillera, as far south as Mexico. The coarse grid also prevented the model from simulating ice streams (fast currents of ice that develop near the margins of ice sheets), which field studies in Greenland and Antarctica have shown to be the most dynamic components of an ice sheet. The model also allowed isostatic rebound to put Foxe Basin and most of Hudson Bay above sea level 120,000 yr B.P., so that Laurentide ice could nucleate in Baffin Island and advance over dry ground onto the North American mainland. It is not known if that was the case, and limiting the model to a simulation of terrestrial ice sheets advancing across dry land prevented any role for marine ice sheets and floating ice shelves. Marine ice and ice shelves are important dynamic components of the Antarctic ice sheet today, and they were probably important components of Northern Hemisphere ice sheets during the last glaciation (Denton and Hughes, 1981; Hughes et al., 1985).

Time-dependent, two-dimensional, mechanical models that constructed ice thickness along flow lines or cross sections through an ice sheet have attempted to include the effects of marine ice, ice shelves, ice streams, and theoretically based glacio-isostatic adjustments. The history of these developments can be followed from the work by Weertman (1974, 1976), Thomas and Bentley (1978), Hughes et al. (1980), Stuiver et al. (1981), Birchfield et al. (1981), Bindschadler and Gore (1982), Pollard (1983), Oerlemans and van der Veen (1984), Fastook (1984, 1986), McInnes and Budd (1984), and Lingle and Clark (1985), among others.

Three-dimensional, steady-state, mechanical models have also been developed. These models are most appropriate for studying the present-day Greenland and Antarctic ice sheets, but they have been used by Hughes (1973) and Sugden (1977) to study North American ice sheets at a glacial maximum. These models date from the one developed by Budd et al. (1971) to derive physical characteristics of the present-day Antarctic ice sheet. Input includes ice thickness and elevation, mean annual temperature and rates of ice accumulation and ablation at the surface, and topography and geothermal heat flux at the bed. Output includes the temperature and velocity field in the ice sheet and the temperature or rates of melting and freezing at its bed. From these, such things as ice trajectories, time lines, and oxygen isotope stratigraphy can be derived. The most advanced version of these steady-state mechanical models is the one by Radok et al. (1982), which was used to derive climatic and physical characteristics of the Greenland ice sheet.

Most purely mechanical models have employed the finite-difference method of numerical modeling. The most useful model for forecasts of future ice sheet behavior is a fully three-dimensional, time-dependent, mechanical model; one that will produce time variations of internal temperature and velocity and basal temperature and freeze-thaw rates as an ice sheet advances and retreats. Budd et al. (1984), among others, have moved in this direction.

5.2.2.2.3.2 Geomorphic models

A different modeling approach constructs geomorphic models. These differ from mechanical models, in that glacial geology is included as model input. Since terminal moraines are one of the most prominent glacial geologic features, and because they have been mapped and dated along much of the margins of all former ice sheets especially during the last deglaciation, geomorphic models generally construct flow-line profiles of these ice sheets from their ice margins to their ice divides. Mechanical models construct flow-line profiles from ice divides to ice margins. If glacial geology beneath a former ice sheet is interpreted correctly, zones where the bed is frozen, thawed, freezing, or melting also can be specified as input, as well as the location and length of former ice streams, the location of ice shelf grounding lines, and the areal extent of marine ice (parts of former ice sheets that were grounded well below sea level all the way to their margins). Glacial geology constitutes a valuable source of data and should not be ignored as input to models that attempt to reconstruct the dynamic history of former ice sheets. The payoff is an accuracy in simulating the areal and, therefore, vertical extent of former ice sheets that cannot be matched by purely mechanical models. This is because geomorphic models include all the input included in mechanical models, plus glacial geology.

The pioneering geomorphic model was the one developed by Hughes (1981). The model was used by Hughes et al. (1981) to reconstruct global ice sheets 18,000 yr B.P. for the Climate Long-Range Mapping Project maximum glaciation experiment. Stuiver et al. (1981) simulated partial deglaciation of the Antarctic ice sheet 125,000 yr B.P. for the Climate Long-Range Mapping Project maximum deglaciation experiment. This model reconstructed, as output, ice thickness and elevation along ice flow lines prescribed as input. Flow lines had constant width and constant accumulation rates along their length, but topography and the degree of isostatic equilibrium could be specified at the bed for each finite-difference step along a flow line. Steps were from 50 to 100 km (31.25 to 62.5 mi) long. Fastook (1984) made a significant improvement in the model by (1) changing the flow lines into flow bands having variable widths and (2) allowing variable rates of accumulation and ablation along the flow bands. The next step is a scheme in which the following occurred:

- The equilibrium line elevation changed to conform with changes in Milankovitch Theory insolation variations.
- The pattern of accumulation changed with the size of the growing or shrinking ice sheet.
- Isostatic adjustments varied exponentially with changes in the ice load.

These improvements would make the time-dependent model quasi-three dimensional.

Reeh (1982) made a major breakthrough in geomorphic models by developing a perfect-plasticity solution for creep of ice sheets in which flow lines were output, instead of input, to the model. His model reproduced all the ice stream drainage basins of the present-day Greenland ice sheet. He then employed it to reconstruct the largely marine Innuitian ice sheet over the Queen Elizabeth Islands of Arctic Canada during the last glacial maximum (Reeh, 1984). Variable bed conditions, as determined by glacial geology, and variable temperature and precipitation conditions at the surface were treated by varying the plastic yield stress to simulate these conditions. Reeh et al. (1986) used their model to reconstruct the Laurentide ice sheet at the last glacial maximum.

A major problem with geomorphic models is that quite different interpretations of glacial geology are possible. Boulton et al. (1985) produced two quite different reconstructions of the Laurentide and Scandinavian ice sheets, based on two quite different interpretations of glacial geology. Both interpretations were different from those employed by Sugden (1978) and Hughes et al. (1981).

One interpretation assumed that isochrones for ice sheet margins during deglaciation could be drawn perpendicular to glaciogenic longitudinal and radial lineations, mostly drumlins and striations. Dated recessional moraines aligned normal to these lineations provided a means for assigning dates to the isochrones. The major assumption here was that the pattern of flow lines for the ice sheets during their closest approach to steady-state flow at the glacial maximum did not change substantially during deglaciation. The other interpretation places emphasis on the dispersal of erratics, with the pattern of dispersal used to reconstruct flow lines during nearly steady-state flow at the glacial maximum. Again, no major shifts in flow lines during deglaciation were a primary assumption. Moreover, Boulton et al. (1985) assumed that ice moved from a rigid bed on the Canadian and Baltic shields onto a deformable bed south of these shields. The pattern of flow lines and ice elevation contour lines was quite different for these two different interpretations of glacial geology.

As an added complication in interpreting glacial geology, it now seems that ice streams presently draining the Antarctic ice sheet can exist on a rigid bed that is largely frozen (Scofield and Fastook, 1986) and on a bed that is thawed and deformable (Alley et al., 1986).

A prudent research strategy is one which makes use of mechanical and geomorphic models and ultimately combines the two into a single model that can simulate the full range of possible boundary conditions, both in space and in time. That is described in Section 8.3.1.5. A two-pronged strategy seems appropriate: (1) use low-cost, finite-difference models to investigate a wide range of spatial and temporal boundary conditions for simulating the future advance and retreat of ice sheets worldwide and (2) use two- and three-dimensional, time-dependent, finite element models to make the most exacting simulations of the future ice sheets on local scale.

5.2.2.3 Modeling future local climatic change

While global climatic change models provide valuable insight into past, current, and future climatic patterns for the entire Columbia Basin, they lack precision in forecasting specific conditions over time at the Hanford Site itself. This section surveys existing climatic models, glacial change models, and flood models that hold potential for long-range predictive capability on a localized basis.

5.2.2.3.1 Local climatic models

Although detailed models of the atmosphere capable of representing the local climate of the area are available, they can only be applied for relatively short (days to months) time spans and only for modern conditions. Solutions for other climatic conditions are precluded by the lack of information about the boundary conditions. Integration of the equations for very long time spans is not possible because of the computational demands it creates.

Long-term solutions can be achieved at a considerable cost in resolution, both temporal and spatial. As an example, the current atmospheric general circulation models have a spatial grid resolution of approximately 5° latitude and 7° longitude. These are equilibrium models. Time-varying models of climate dynamics are considerably coarser (approximately 10° on a side). Thus, the combination of the demands of spatial resolution, temporal latitude, and computational limitations precludes the application of "classical" meteorologic methods to forecasting the climate at the Hanford Site for the next 100,000 yr.

Mesoscale atmospheric systems are those with scale ranges from a few kilometers to a few hundred kilometers horizontally, tens of meters to the depth of the trophosphere vertically, and 1 to 12 hr temporally (Pielke, 1984). A more formal definition is given by Pielke (1984) as having the following characteristics:

- o Having a horizontal scale sufficiently large so that the hydrostatic equation can be applied (approximately 10 km (6.25 mi)).
- o Having a horizontal scale sufficiently small so that the Coriolis term is small relative to the advective and pressure gradient forces, resulting in a flow field that is substantially different from the gradient wind relation even in the absence of friction effects.

Pielke (1984) also notes that because the Coriolis force is a function of latitude, the maximum horizontal scale also is a function of latitude.

Orlanski (1975) proposed a spatial classification for atmospheric scales. He defines mesoscale atmospheric phenomena as those whose size ranges from 2 to 2,000 km. He further subdivided the mesoscale into three divisions:

1. Meso-gamma - 2 to 20 km.
2. Meso-beta - 20 to 200 km.
3. Meso-alpha - 200 to 2,000 km.

Each of these divisions is based on the maximum probability that the interval would contain certain phenomena (e.g., meso-gamma would contain thunderstorms and urban effects, meso-beta would contain squall lines and mountain disturbances, and meso-alpha would contain fronts and hurricanes).

Pielke (1984) assigns Orlanski's meso-gamma to the microscale and meso-alpha to the macroscale. Consequently his definition of mesoscale corresponds roughly to Orlanski's meso-beta scale. Pielke (1984) also considers two types of mesoscale (Orlanski's meso-beta) atmospheric systems:

1. Terrain-induced mesoscale systems.
2. Synoptically induced mesoscale systems.

Terrain-induced mesoscale systems are those controlled by variations in topography as a parcel of air moves through an area. Examples include sea and land breezes, mountain-valley winds, and forced flow over rough terrain. Synoptically induced mesoscale systems are those controlled by large atmospheric disturbances such as fronts, squall lines, and hurricanes.

Numerical methods may be used to model both of these systems. Numerical mesoscale models can be classified into six basic types (Drake et al., 1981):

1. Uniform wind field.
2. Area of influence.
3. Empirical interpolation.
4. Objective analysis.
5. Simplified physics.
6. Full physics or primitive equation models.

Uniform wind field, area of influence, and empirical interpolation models are generally used to model wind fields. These models can provide initial values required by objective analysis and simplified physics models. Similarly, objective analysis and simplified physics models are generally used to model winds and provide initial data for full physics models. Full physics models are prognostic; that is, they can be used to forecast forward in time.

The emphasis on winds in this classification reflects the extensive modeling of airborne pollutant dispersal at the mesoscale level. Diagnostic (steady-state) wind models are used to provide quick evaluations of wind in the area when station data are sparse.

5.2.2.3.1.1 Uniform wind field models

As the name implies, uniform wind field models assume a uniform wind direction throughout the area. Such models are most appropriate for modeling simple terrain or small areas where the assumption of a uniform wind field is appropriate. A uniform wind field model was used for a larger area by Roberts (1981). He derived orographic variables using a uniform wind field model for use in a statistical model of climate for the southwestern United States.

5.2.2.3.1.2 Area of influence models

When more than one station with wind data is available in the area, each station may be assigned an area of influence. Within each area of influence the winds are considered uniform, just as in a uniform wind field model. As a parcel of air moves across the boundary between two areas of influence, it abruptly changes to the new uniform wind direction. Hinds (1970) applied such a technique to the coastal mountains of southern California.

5.2.2.3.1.3 Empirical interpolation models

Empirical interpolation models carry the concept of area of influence one step farther. Rather than simply defining an area of influence, the influence of a station on a point is a function of the distance to the point.

Wendell (1972) applied an empirical interpolation model to data from 21 stations over an area of 12,420 km^2 of the Snake River Plain, Idaho. He used a weighted average of the velocity components from several stations nearest each of the grid points. The weighing factor was a function of the inverse of the squared distance to the station.

Empirical interpolation has been used by Phillips (1979) and Drake et al. (1981) to obtain initial grid point values for input to objective analysis models.

5.2.2.3.1.4 Objective analysis models

The term "objective analysis" was used for the first time by Panofsky (1949). While a subjective analysis would include human analysis (e.g., drawing isopleths on a map), objective analysis uses an objective mathematical expression for fitting data.

Pielke (1984) includes uniform wind, area of influence, and empirical interpolation models under the heading of objective analysis. Indeed, Wendell (1972) referred to his empirical interpolation model as an "objective interpolation." However, Gandin (1965) notes that objective analysis is not simply an interpolation of data. That is only a part of objective analysis. The model also should consider geostrophic or static effects, for example.

Drake et al. (1981) categorize objective analysis models much as Gandin (1965). Their criteria include that the divergence of an air mass is zero or, in other words, there is conservation of mass. Given this requirement, the terrain influence on the flow field is calculated. Such models provide for efficient estimates of the effect of complex topography on winds.

Terrain influence can be calculated in different ways. Patnaik and Freeman (1977) and Sherman (1978) adjusted interpolated mean winds in a weighted least-squared sense (MATHEW model). The theoretical basis for the model was developed by Sasaki (1970). Phillips (1979) developed a model (NOABL) based on physical reasoning. He calculates a perturbation velocity potential that results from any initial errors in the wind field and terrain influences. An elliptic equation is developed and solved by iteration. Another method is used by Drake et al. (1981). They adjust an interpolated mean wind field using basis and amplitude functions developed from station data (GRAMA model).

5.2.2.3.1.5 Simplified physics models

Simplified physics models use a simplification of the Navier-Stokes equation (Drake et al., 1981). An example of a simplified physics model is the WINDS model, developed by Fosberg et al. (1976) based on earlier work by Anderson (1981). The WINDS model is a two-dimensional, slab grid model based on divergence and vorticity equations and neglecting advection. It adjusts the winds to variations in terrain, surface roughness, temperature, and a nondimensional pressure field (Childs and Marlatt, 1981). Dietrich and Marlatt (1981) applied the model to a 8,600-km^2, topographically complex area in San Diego County, California, and obtained realistic results.

5.2.2.3.1.6 Full physics models

Full physics or primitive equation models are based on the full Navier-Stokes equations. They are three-dimensional and prognostic; consequently, they can be slow running. Drake et al. (1981) indicate that the SIGMET model (Traci et al., 1977) runs at half real time on a CDC 7600 computer. Consequently, these models are unusable when studying long-term (years) mesoscale changes.

Pielke (1984, Appendix B) provides an overview of groups active in prognostic modeling or whose work is of direct relevance to mesoscale modeling. Selections of primitive equation models from his listing are given in Table 5.2-9. Anthes (1983) provided a similar review of regional-scale models. His paper limits regional-scale models to those that have a horizontal resolution of 50 to 250 km or, alternately, having a horizontal scale of 250 to 2,500 km. Orlanski (1963) would classify this as meso-alpha scale, and Pielke (1984) would classify this as macroscale.

The dynamic initialization technique uses full physics models and can be used instead of objective analysis models to obtain initial values for a full physics experiment. The full physics model is integrated over a period of time, so that values in the observed field that are not representative of mesoscale processes are minimized (Pielke, 1984). The model equations adjust to an equilibrium state. Hoke and Anthes (1976) describe the technique and applied it to a primitive equation model of the 1962 hurricane Alma (Hoke and Anthes, 1977). However, dynamic initialization techniques are slow and can require 12 h of computer time for a 24-h experiment (Pielke, 1984).

Statistical methods can be used to develop meteorologic or climatic models. This methodology is reviewed in texts such as those by Essenwanger (1976) and Murphy and Katz (1985). The advantage of statistical models is that they do not require the computer time of full physics models. However, they are not as portable as full physics models--new equations must be developed when applying the model to a different area.

Houghton (1979) developed a multivariate model of monthly and annual orographic precipitation in the north-central Great Basin. Roberts (1981) used a similar technique to model modern and last glacial temperature and precipitation of the southwestern United States. Koa and Kau (1981) developed a statistical-empirical model of wind flow over complex terrain.

The first model of long-term climatic variability at the Hanford Site employed an index of the climatic state of the world at 100-yr intervals to infer the local climatic conditions. Global climatic state was based on the orbital characteristics of the Earth (precession, obliquity, and eccentricity). These were combined into a single value, the Astronomical Climate Index (Kukla et al., 1981). This index includes an additional random term, representing nonsystematic perturbations due to such phenomena as volcanic eruptions (Fig. 5.2-62).

Local influences of this climatic state included fluctuations in glacier margins, sea level, catastrophic floods, and groundwater recharge rates. This model synthesizes many of the processes that must be considered in site characterization. It suffered from a large number of weaknesses and, thus, cannot be directly applied to site characterization (Foley et al., 1982, pp. 3.1 through 3.15).

The model of Foley et al. (1982) is a zero-dimensional model; that is, it represents system behavior at a number of distinct points that do not have an explicit spatial organization. Since it is incapable of representing the spatial variability of climate, important controls on site conditions (e.g., recharge areas and recharge rates) are grossly simplified and the entire climatic system must be assumed to behave linearly. Such an assumption is unwarranted by modern meteorologic observations.

It is further unlikely that the climate of the Pacific Northwest varies in unison with that of the rest of the globe. Long-term variations in insolation are latitudinally controlled. Furthermore, the local climate is clearly controlled by the configuration of the Cordilleran ice sheet. That ice sheet varied in size out of phase with the global ice volume changes by at least 3,000 yr. Thus, a defensible model of local climate must reasonably portray the global and local controls on climate.

No such model has been prepared for the Hanford Site. A model similar to that required has been created for the Nevada Test Site (Craig and Roberts, 1983, 1984; Craig, 1984). That model is based on a set of regression equations to forecast mean monthly values of temperature and precipitation. The equations are solved for a grid of points surrounding the Nevada Test Site.

Independent variables in the equations represent the orographic influences (effects produced by the rise and fall of air masses as they cross mountainous areas) on temperature and precipitation and the effects of changed boundary conditions, including sea surface temperature patterns, sea level, and atmospheric circulation. Sea surface temperatures and atmospheric circulation patterns are specified as mean values for the months of February and August. Atmospheric circulation patterns must be specified from atmospheric general circulation models run with the appropriate boundary conditions.

Limitations of that model include its static representation of climate. At present, useful solutions have been obtained only for the last glacial maximum, since that is the only time for which global reconstruction of sea surface temperatures were available. Since that time, sea surface temperature reconstructions for the last interglacial have become available. Methods of reconstructing local sea surface temperature patterns at other times would extend the applicability of the model greatly.

Adaptation of the model to the Hanford Site would require developing some means of representing the influence of the proximity of the Cordilleran ice sheet during extreme glacial events. The model's winds are grossly simplified. Improvements in modeling the winds, including the topographic effects on wind direction and speed may provide a basis for the representation of the effects of the ice sheet.

For the Nevada Test Site model, tests of the local climatic model focused on the response of the surface hydrologic system (Singer and Craig, 1983). This response is extensive and well documented in the geologic record. At the Hanford Site, such a test will not be as useful. In this case, some other methods must be devised to certify the accuracy of the model. To provide an independent test of the ability of the model to reproduce modern climate, an estimate of runoff could be made that can be compared to modern gaging station data. Estimates of runoff must include the computation of evapotranspiration and infiltration.

For validation of the model's ability to reconstruct paleoclimates, comparison to the vegetation record could be used. The vegetation record could provide a comprehensive representation of paleoclimates in this area.

Another validation could be based on a test of the ability of the model to reconstruct the size and configuration of glaciers in the Cascade Range. These points are discussed more in Section 8.3.1.5.

5.2.2.3.2 Glacial change models

Ice sheets are large bodies that have ice flow from the center radially out to the margins (Paterson, 1981, p. 153). This flow is due to internal deformation of the ice crystals and basal sliding of the ice along the bedrock floor (Nye, 1959, p. 493). Various studies concerning the details of these components have been made (Glen, 1955; Lliboutry, 1968; Colbeck and Evans, 1973). Basal sliding is only a concern where the base of the glacier is at the melting point. Glaciers whose base is below the melting point should not experience sliding (Paterson, 1981, p. 112).

Using a series of simplifying assumptions, cross-sectional profiles along flow lines can be constructed for ice sheets and ice caps (Paterson, 1981, pp. 153-161). This procedure assumes the ice sheet is in steady-state and has a uniform mass balance and bed roughness. The ice sheet also is assumed to be isothermal, a condition only met by temperate glaciers (Paterson, 1981, p. 83). Ice profiles developed from this method are useful for the reconstruction of past ice sheets from ice margin location data. However, since these reconstructions assume a steady-state condition, they do not allow analysis of the ice dynamics.

Nonsteady-state models, such as the one presented by Mahaffy (1976), allow the investigation of ice flow through time. This model computes ice flux for a two-dimensional matrix of points representing a plan view of the ice sheet. The ice flux and net mass budget are used to calculate the ice thickness at each point.

Input to the model includes initial ice height, mass balance for each point, basal sliding velocity, bedrock topography, and required constants for the flow law. This type of model allows the study of the advance and retreat of the ice as the mass balance of the glacier changes. Although bedrock topography is input, the model assumes a relatively smooth bedrock profile. This assumption is not likely to be valid for most of the Cordilleran ice sheet.

Other models that include the calculation of the temperature distribution in the glacier also have been published. Grigoryan et al. (1976) document a three-dimensional model for nonisothermal glaciers. For this model they have developed a system of equations to solve for the stress, temperature, and velocity distributions in a nonisothermal glacier. Primary assumptions are that the length and width of the glacier are much greater than its height and that any large features in the bed have a large radius of curvature, so can be considered insignificant.

Jenssen (1977) also has developed a three-dimensional model that relates the thermodynamics and flow properties of the ice sheet. Computations in this model used finite-difference forms of three-dimensional heat conduction and continuity-of-mass equations along with an empirically derived flow law. Ice shelf growth and mass loss from calving also were considered. The model was then run using input data intended to represent the Greenland ice sheet. The results indicate that a finer grid spacing may be necessary to obtain reliable results.

Some of the above methods for glacial modeling have been used in the reconstruction of paleo ice sheets. Andrews and Mahaffy (1976) used Mahaffy's unsteady flow model, discussed previously, to study the maximum growth rate of the Laurentide ice sheet. The focus of their reconstruction was the sea level low, which occurred at 115,000 yr B.P. Calculation of the rate of ice sheet growth was intended to test theories of rapid glaciation and sea level lowering. Parameters for the model in this reconstruction were chosen to maximize the total ice accumulation. Ice shelves and calving were not considered, and zero basal sliding velocity was assumed. The authors conclude that the general pattern of ice movement and configuration agree with previous information.

Ice sheet reconstructions using ice profiles along flow lines have been done for the major ice sheets present during late Wisconsinan time (Hughes, 1981). These steady-state profiles are intended to represent the maximum ice configuration and were constructed from available ice margin data. The profiles also account for the change in the basal shear stress, which occurs when there are changes in the nature of the ice-bed interface.

Multiple profiles were constructed for each of the major ice sheets, including the Cordilleran ice sheet that was represented by a total of eight flow lines for the maximum reconstruction. From these profiles, a three-dimensional reconstruction of the ice sheet was made by connecting the ice and bedrock elevations, which are radially disposed around ice domes and divides. The present landform topography was used to represent the bed of these ice profiles.

After the initial ice profile was developed, the weight of the overlying ice is allowed to isostatically depress the underlying bedrock. Final flow line profiles were then computed over the new bedrock elevations. Isostatic adjustment was assumed to take place by crustal and mantle deformation with no change in crustal thickness (Hughes, 1981, p. 254).

This method of ice sheet reconstruction takes into account factors such as ice shelves, ice streams, ice-bed coupling, and isostatic loading of the bed. However, since this is a steady-state model, it cannot be readily used for nonsteady-state conditions (e.g., during ice sheet growth and decay).

5.2.2.3.3 Flood models

As stated earlier, it is quite certain that future glacial events will occur during the next 100,000 yr. Such glaciation may affect the northern portions of Washington and may be more extreme than those in the recent past. During such glaciations, multiple ice advances from the north blocked the Clark Fork River in Idaho, leading to the formation of Lake Missoula. Multiple failures of these ice dams produced immense floods over the Columbia Plateau in Washington, including the Pasco Basin. Should such glaciation occur, it is likely that floods, analogous to the Missoula floods of the last glaciation, will again occur (Fig. 5.2-63). The precise nature of those floods cannot be specified, since an exact analog could not be expected, even if the character of past floods was known precisely. Craig and Hanson (1986) have shown already that it is not likely that such floods can ever be reconstructed from field evidence because of the time-transgressive nature of the evidence. Indeed, the precise nature of the floods is unknown even in the case of the latest floods. It will be important to construct models that can describe the behavior of these floods in sufficient detail that potential disruptive effects of the floods can be recognized.

Several types of disruptive events from Missoula floods are considered possible. It is not known whether any will disrupt the site even if a Missoula flood were to occur. It is also not certain that additional floods will occur in the next 100,000 yr. These points must be evaluated during site characterization. Possible disruptive events include the following:

- Severe erosion of basalts.
- Pressurized recharge of the groundwater system.
- Rearrangement of the surface drainage.
- Modification of recharge and discharge areas of the groundwater system.

To evaluate the significance of their effects on the stability of the proposed repository, sensitivity analysis must be performed. Such sensitivity studies should be based on estimates of the magnitude of disruptions under anticipated and unanticipated flood scenarios.

This section will review some of the characteristics of Missoula floods and the factors that influence the occurrence of such floods. This information will form the basis of estimates of the probability of such floods in the future and of the impact of such floods. Plans for further evaluation of the risk due to extreme flooding are defined in Section 8.3.1.5.

Potential impacts of the floods include erosion and transportation of the unconsolidated and poorly consolidated sediments that now lie on the Columbia Plateau basalts. This could produce a long-term change in recharge rates to the basalts. The same floods also could deposit large quantities of loose sediment on top of the existing basalts. The presence of these sediments also could modify long-term recharge rates.

For certain times during a flood, very high fluid velocities probably occur in the Pasco Basin. Such velocities may be sufficient to erode portions of the basalt flows at the surface. This would especially be the case where the basalts are highly fractured.

Because flood flows may reach depths of nearly 200 m in the Pasco Basin, and because flood flows may last for several weeks (although not at that great a depth), it appears possible that the high water pressures at the base of that mass of water may be sufficient to produce pressurized recharge into the basalts. Such recharge could lead to change in hydraulic head and groundwater flux rates through the repository. The possibility of pressurized recharge will be investigated during site characterization. A plan to study this is defined in Section 8.3.1.5.

To provide the information about the typical and extreme characteristics of a flood that will be required to assess these questions, a model of flood behavior is required. Several modeling studies of different aspects of the Missoula floods have already been accomplished. Also, there are a number of different modeling methods that have been developed to examine more typical floods, such as are encountered today. To set the stage for the selection of the methods to be applied in site characterization, the models that have been applied or are now available will now be reviewed.

V. R. Baker (1973) used the contracted-opening (Matthai, 1967) and slope-area methods (Dalrymple and Benson, 1968) to compute the maximum discharges of flood waters through a number of channelways of the Scablands see (Fig. 5.2-63). Baker also applied other quantitative procedures to reconstruct the characteristics of the flood. At Wallula Gap, Baker found that the maximum discharge was approximately 9.6×10^6 ft^3/s. Several other authors have applied similar techniques to estimate the dynamic behavior of the flood waters.

Baker (1978) discusses a number of sources of error in his calculations. Improved estimates can be achieved using the computer procedures for step-backwater analysis (Feldman, 1981). Baker et al. (in press) used step-backwater analysis to revise the estimated maximum flood discharge at Rathdrum Prairie from 21.3×10^6 to 17.5×10^6 m^3/s^{-1}.

Clarke et al. (1984) used a computer simulation model to estimate the discharge of Lake Missoula at the location where the ice dam failed. Their model includes the time-dependent thermal erosion of the ice dam by the waters flowing through the opening. It also accounts for the decreasing head upstream of the ice dam as the lake empties. Clarke et al. (1984) conclude that there is a large uncertainty in the possible maximum discharges at the ice dam, ranging from a low of 12×10^6 m^3/s to a high of 26×10^6 m^3/s. Such uncertainties arise from the lack of knowledge about the frictional terms of the flow equations.

The Clarke et al. (1984) model has several acknowledged shortcomings. It is strictly one dimensional; thus, it fails to represent the effects of the complex channel geometry on the flood flows. It also fails to allow for backwater effects in Rathdrum Prairie. It models the lake basin with a simplistic geometry: a closed-tube draining a rectangular box.

The Clarke et al. (1984) computer simulations of Missoula flooding are based on model assumptions derived from observed jokulhlaups for modern ice-dammed lakes (Mathews, 1973; Nye, 1976). The simulations generate outflow hydrographs whose peak discharges range from 2.7 to 14 x 10^6 m^3/s.

As pointed out by Baker and Bunker (1985), major physical extrapolations are required in the Clarke et al. (1984) computer simulations.

Beget (1986) pointed out that paleoflood discharges estimated by Baker (1973) were larger than the discharges achieved in the Clarke et al. (1984) computer simulations. Arguing that the largest Missoula flood flow phenomena should be compared to a linear extrapolation from much smaller historic jokulhlaups, Beget (1986) noted that the Clarke et al. (1984) estimates fit the extrapolation, while the Baker (1973) estimates greatly exceeded the extrapolation. This is not surprising, since the Clarke et al. (1984) models were derived from observations of the same relatively small historic jokulhlaups from lakes that were at least three orders of magnitude smaller in volume than Lake Missoula. Since the Baker (1973) estimates are based on physical evidence (high water marks) rather than on assumptions concerning the mode of ice dam jokulhlaup release, it is possible that the anomalously large discharges reflect a different mode of failure for the maximum flood of Lake Missoula than for the much smaller historic jokulhlaups. More research on this anomaly is clearly indicated.

The Pacific Northwest Laboratory geologic simulation model of the Hanford Site (Foley et al., 1982) included a representation of the effects of Missoula floods on the stability of a nuclear waste repository at a 1-km depth. The geologic simulation model is stochastic in nature, using random draws from a frequency distribution to determine whether events occur or not. Flood characteristics considered include whether floods occur within a given time horizon, the number of such floods that occur, whether or not hydraulic ponding occurs at Wallula Gap, the amount of erosion that such a flood produces, and the amount of sediment that is deposited by a flood (if any).

Difficulties with the geologic simulation model of the Hanford Site have been described by Craig in Foley et al. (1982, Chapter 3). Major improvements in the model would be required before it could be applied to site characterization. Certain of the required improvements have been implemented by Underberg (1983a, 1983b) and include a more complete analysis of the probability of hydraulic ponding behind Wallula Gap given that a flood does occur.

Underberg (1983a) also studied the likely depth-discharge relationship that would apply during a similar flood. Study results suggest that the use of the step method by Baker in his earlier analysis led to underestimation of the discharge corresponding to a given flood level. Underberg's (1983a) results suggest that future floods are likely to lead to increased sediment deposition in the Pasco Basin. Such results suffer from the fact that computations of discharge are steady-state values and that the sedimentation estimates are strictly stochastic in nature.

More recently, the probability that additional floods will occur in the next 100,000 yr has been reexamined (Craig, 1985; Craig and Singer, 1984; Singer and Craig, 1984). These studies found that there is a high probability that such floods will occur. At least one, and as many as 68, episodes of flooding can be expected in this time period. Thus, it is concluded that more detailed modeling of the characteristics of those floods is necessary.

Such modeling should take advantage of the capabilities of numerical methods that are now available. Several implementations of the two-dimensional, unsteady-flow, Navier-Stokes equations for hydrologic systems have been prepared. For example, Leendertse et al. (1981) have created a computer code applicable to tidal basins. Simons et al. (1979) used a similar approach to model certain reaches of the Mississippi River. This model includes representations of the sediment transport equations. A similar model is being developed by Craig and Underberg (1984). Thus, the capability exists to obtain a comprehensive model of the influence of Missoula floods on the stability of the Hanford Site. The actual methodology to be applied is described in more detail in Section 8.3.1.5.

5.2.3 SITE PALEOCLIMATIC INVESTIGATION

A number of questions remain open concerning the climatic changes that have occurred during the Quaternary in the Pasco Basin. Certain questions should be resolved during site characterization and are of two types: on the one hand, more detailed and quantitative estimates of the climatic change that has occurred are required; on the other, questions concerning the mechanism of change must be answered to forecast more accurately the extremes in the future climatic states of the area. This is because future climates are not expected to be exact analogs of the late Quaternary climates.

Summaries of the major open questions and problems that have been recognized in the existing record are given below. These are grouped according to the different areas of concern:

- Short-term climatic change.
- Long-term climatic change.
- Glaciation.
- Characteristics of changes in the northeastern Pacific Ocean.
- Variability of atmospheric circulation patterns.

Following these summaries of open questions in each of these fields, methods that can be used to answer these questions are briefly listed. Based on the list of important questions, data needs and work items are identified in Section 5.3. These form the basis of the plans that are described in Section 8.3.1.5.

5.2.3.1 Short-term climatic change

In the following discussion, the various limitations in the understanding of the history of short-term climatic change are itemized first, then remedies are suggested. In this discussion, "short-term" implies changes occurring within hundreds of years.

- The reconstructions used in the Pacific Northwest for the Pasco Basin region climatic reconstruction project (Cropper and Fritts, 1986) were based on 65 tree-ring chronologies that began in 1600, or earlier, and extend to 1963. Many of the chronologies are based on a small number of trees during the first half of the 17th century, and the standard error is high.

- The chronologies also are widely scattered over a large geographic area and were calibrated with an even larger grid of temperature and precipitation stations. The canonical regression used in the calibration related only the largest scale variations in tree growth to the very largest scale variations of climate. That project (Cropper and Fritts, 1986) used the combined records of four individual station estimates closest to the Pasco Basin region. These reconstructions appear to contain about one half of the climatic variance of the Pasco Basin region (although this value was not measured directly) that was probably dominated by the large-scale variations throughout the region. The remaining unexplained variance represents unassessed climatic variations, especially those of a local nature, and noise or error.

- There probably were climatic records closer to the Pasco Basin region that were not analyzed. This was largely because of the limitations of time and resources. These data should be examined to relate the mean values of the Pasco Basin to those for the adjoining stations that were reconstructed for the same time period.

- The climatic analysis of Cropper and Fritts (1986) was limited in scope. Future analyses should consider circulation features that are revealed in the pressure reconstructions, and the analyses should be closely related to the climatic modeling effort.

- The filtered annual reconstructions for a region appear to be the most reliable available. The statistics that represent century-long and regional averages can be used with a high degree of certainty, because the averaging has reduced the noise in the individual station seasonal climatic estimates. Considering smaller time or space scales, such as the results for individual stations (the variance over the region) and an annual or a seasonal value, the error or noise in the record attains much less acceptable levels and the reconstructions appear to be relatively inaccurate.

Methods to improve the knowledge of short-term climatic change include the following:

- To obtain reliable estimates of anticipated and unanticipated variability of the modern climate, a number of climatically sensitive tree-core collections should be obtained from the mountains and plateaus around the Pasco Basin area. These should include 15 to 20 chronologies, 2 cores per tree, and 25 to 40 trees per site. The ages of these trees will be expected to range from 150 to 450 yr. For temperature reconstruction, high-altitude species that may extend back 500 yr or more will be collected. This should include 10 additional sites. The reconstructions should then be recalibrated to include these new chronologies.

  The recalibration will not be restricted to these 30 sites. They will include the 10 new chronologies studied by Cropper and Fritts (1986) and many, if not all, of the 65 chronologies that have proven to be climatically sensitive.

- Calibration models will include analysis of the biological system and autoregressive moving-average features of the tree-ring series (Box and Jenkins, 1976, pp. 46-84). They also should include ridge-regression analysis, which is a possible replacement of response-function analysis. Such an approach would provide the information needed to determine the best strategy of calibration. It will help answer the following questions: Should the climate of different seasons be calibrated separately? What seasons should be used for temperature and precipitation? Should the high and low frequencies in growth variation be calibrated separately? If so, what are the optimum models to use? Perhaps by separating these two frequency extremes, higher frequencies can be calibrated more effectively.

- The available climatic data will be examined and those series most useful to this study identified. The data will then be obtained in preparation for analysis. Any lack of homogeneity of data will be identified and eliminated and missing values estimated.

- The relationships of these data to atmospheric pressure anomalies over the North Pacific Ocean also will be studied, and the reconstructions of the pressure data will be related to the anomalies of temperature and precipitation. Relationships that become evident from this study will be used to suggest ways of improving the climatic modeling.

5.2.3.2 Long-term climatic change

Deficiencies in the understanding of long-term climatic change include the following:

- There is a need for quantitative approaches to extracting the detailed climatic signals obviously present in the fossil-pollen data to characterize climatic variability over a glacial-interglacial cycle. Modern climatic data should be obtained for all sites where fossil-pollen data have been obtained.

- Analysis of available high-quality pollen-stratigraphic data from the Columbia Basin demonstrates another pressing need. At this time Carp Lake is the only site that provides an appreciable record of glacial age climatic conditions in the Columbia Basin and vicinity. To assess the reliability of those data, other sites should be located that can provide additional data on the climate of the last glacial age. These sites should be south of the limit of the Cordilleran ice sheet and above the upper limit of Missoula flooding. Preferably, one should be located within 50 to 100 km of Carp Lake to provide some replication of the record there and some additional insight into the nature of the bioclimatic changes evident in the Carp Lake record.

- Almost all Quaternary paleoclimatic data for British Columbia come from studies of sediments deposited during the last 50,000 yr. Little work has been done on older sediments that, at present, are poorly dated.

- Even for this relatively short period, there is little paleoclimatic information for the British Columbia interior. Most paleoecological work in the province has been conducted on the south coast; work in the interior, until recently, has been sporadic and very limited in scope. Such information is needed to better understand controls on the growth of the Cordilleran ice sheet,which is important if forecasts of future behavior of that ice sheet are to be available.

Before fossil-pollen data can be used to quantify the paleoclimatic history of the Columbia Basin, the following areas of controversy should be addressed:

- The nature of the full glacial vegetation. The currently accepted reconstruction of a full glacial steppe throughout the Columbia Basin needs to be supported by additional long records. Ideally, long pollen records should be obtained from all quadrants of the Columbia Basin (and along major bioclimatic boundaries).

- The reconstruction of late glacial vegetation and climate. The existence of a localized periglacial environment at the northern border of the Columbia Basin is unresolved in the literature. Plant macrofossils in the context of pollen-stratigraphic data, more rigorous morphologic studies to identify major pollen taxa to species level, and a search for suitable vegetational and climatic analogs in the modern landscape will all be helpful. Resolution of this question has impact on the assumed magnitude of full glacial climatic change at the site.

- The metachronous Holocene climatic sequence. Most records show a three-part climatic record, but the direction and timing of the signal are quite variable. The record closest to the Hanford Site shows no major changes, but the numerous minor pollen fluctuations are ascribed a great deal of climatic significance. In other records, there may be important pollen-stratigraphic events that would correlate with those at other sites. Spatial or temporal patterns may emerge when the data base is considered as a whole. The pollen identifications, the selection of samples for radiocarbon dating, and the method of zonation are presently not, but should be, standardized.

- The sedimentary record as an indicator of climatic change. The sedimentary record is given uneven attention and the evidence is interpreted differently. Possible lake level fluctuations should be studied at all sites to develop a regional picture of paleohydrology. This regional picture will be important in understanding the relationship between major climatic change and variations in the groundwater system.

- The importance of local versus regional environmental factors in shaping the pollen record. The role of the Columbia River Gorge in determining the local climate at Carp Lake, the impact of humans on vegetation, and the effect of volcanic eruptions on the landscape all need further investigation to sort out how much of the pollen record reflects climate and how much does not.

5.2.3.3 Glaciation

Although the general pattern and timing of glacier expansion in British Columbia during the last glaciation is known, details are lacking. The configuration of the ice sheets determines whether major floods from Lake Missoula will occur. It also influences the atmospheric circulation and the local climate. The exact pattern of deformation of the crust by the Cordilleran ice sheet also should be refined and quantified.

There is a need to understand the dynamics of the late Quaternary ice sheets in sufficient detail to use that data to test the computer models to be used in forecasting their future behavior.

5.2.3.4 Oceanographic change

Deficiencies in our knowledge of oceanographic change include the following:

- The studies cited in Section 5.2.1.3 have been based on records from widely separated locations using different methods and time intervals of sampling, and extending over varying intervals of time. Concepts and techniques also have changed since the earlier studies. Therefore, the conclusions drawn by the various workers cannot be compared easily, and there is no coherent description covering all aspects of the northeastern Pacific Ocean conditions during the Pleistocene.

- It is now generally recognized that a full understanding of climatic change must consider several different time scales, so that short-term trends can be seen in the context of much broader changes. The two studies mentioned in Section 5.2.1.3 covered only the last glacial to interglacial transition (Moore, 1973) or the most recent cycle, while the studies covering a longer interval (Sachs, 1973; Sancetta and Silvestri, 1984, 1986) were all located very far from the area of interest. Still lacking is a local record of long-term change over several climatic cycles. The studies of Sachs and Sancetta and Silvestri suggest that this long-term record should extend beyond 400,000 yr B.P.

- Only two of the previous studies were geographically near the Hanford Site (Y6910-2 (Moore, 1973) and Y7211-2 (Heusser and Shackleton, 1979; Pisias et al., 1981)). This is insufficient to provide information on regional patterns. The climatic models require maps of sea surface temperatures, which must be constructed from a larger number of data points.

- The only detailed local study published to date concentrated on the transition from glacial to interglacial conditions (Moore, 1973). However, the present climate is that of a full interglacial maximum. Any future change, therefore, will almost certainly be the opposite (i.e., a transition from interglacial to glacial conditions (Kukla et al., 1972 and references therein)). Transitions between glacial and interglacial extremes are not symmetrical (deglaciations occur much more rapidly than glaciations), and have different forcing functions. For the purpose of site characterization, therefore, it is essential to study the analog of an interglacial to glacial transition.

Methods to improve our knowledge of oceanographic change include the following:

- As a first step, the published work must be evaluated and, if necessary, reinterpreted to produce a unified, coherent description of current knowledge of oceanography and climate of the North

Pacific Ocean during the Pleistocene. The evaluation should identify weaknesses and contradictions in the existing data and (or) interpretations, and define specific questions to be addressed in any further studies.

- One or two cores that span the last 500,000 yr and located relatively near the site (i.e., just west of Washington or Oregon) must be analyzed. Analyses of all available cores should be extended back to the time of oxygen isotopic stage 5e. These cores should be sampled at short enough time intervals to permit identification of climatically relevant frequencies (e.g., 41,000 yr and 23,000 yr B.P.). Time-series records should be based on at least two different types of data (e.g., different microfossil groups, stable isotopes) to cross-check on each other and to distinguish possible changes in several parameters (e.g., temperature, precipitation, intensity of upwelling).

- An adequate reconstruction must be based on a grid-like pattern of data points, so that a two-dimensional map of sea surface temperatures can be developed. Because the site lies near the latitude of a major climatic boundary and near an eastern boundary current, north-south and east-west trends in surface temperatures must be included. At a minimum, this would require one north-south transect and one east-west transect of deep-sea cores, and preferably two of each.

- The cores in the transects described above must be sampled at closely spaced time intervals (between 3,000- and 5,000-yr intervals) and must include the last full interglacial to glacial transition (isotope stages 5 and 4). At least one core should have an oxygen isotopic record to establish adequate time control and to allow correlation with records elsewhere. The estimation of sea surface temperatures should be based on at least one and preferably two types of data to cross-check each other. Either radiolaria or diatoms would be appropriate bases for the estimates; planktonic foraminifera also would be useful if they are preserved in sufficient numbers throughout the interval.

5.2.3.5 Atmospheric change

The previous discussion on status of general circulation model research, philosophy of model usage, and the brief review of previous model experiments for glacial and CO_2 warming conditions (see Sections 5.2.1.4 and 5.2.2.3) provides a perspective for needed research related to the Hanford Site. This research will fall into two categories: (1) attempts to reconstruct past climates with general circulation models in cases where that climate may repeat in the future and (2) attempts to assess model sensitivity to specific factors that may influence future climate.

The global paleoclimate and the paleoclimate for the northwestern United States and the eastern Pacific Ocean must be investigated using a general circulation model with an interactive ocean component. Use of a model with the dynamic ocean component would be necessary to study most accurately the coastal climate of Washington and the effects of the Pacific Ocean on the inland climate. The influence of factors such as the positions of the Aleutian Low and the Subarctic Current could be best studied in this manner.

The possibility of using a less complex type of climatic model (such as those listed in Sections 5.2.1.4 and 5.2.2.3) also must be considered not only to investigate the paleoclimate of the Hanford Site, but also to evaluate the influence of boundary conditions used by the general circulation model and even to evaluate the general circulation model results themselves.

Some attention must be given to lessening or smoothing the disparity between microscale paleoclimatic evidence and coarser macroscale modeling results. This may be accomplished by developing standardized methods of measuring or expressing paleoclimatic data in terms that can be more easily related to variables generated by the general circulation model. Another aspect to consider is further investigation to resolve the levels of uncertainty outlined above.

The BWIP plans to execute a set of modeling experiments such as Kutzbach and Guetter (1984a) is doing for the Holocene (simulating climate conditions for every 3,000 yr), but extending back to the last glacial maximum and for the last interglacial. Also, comparative studies of the last interglacial to the present interglacial will provide useful insight to the changing climate. A set of predictions of future climates should be undertaken, using most probable future conditions for the boundaries. Future conditions to consider include altered ocean surface temperatures, atmospheric CO_2 concentrations, and orbital parameters.

Computational efficiency to allow investigation of an as-broad-as-possible range of variables is considered to be an important consideration. The following classes of experiments should be completed:

1. Perform January and July fixed-insolation experiments at 18,000 yr with Climate Long-Range Mapping Project-specified boundary conditions (computationally efficient, physically consistent simulations with model intercomparison possible and sense of seasonal variability).

2. Assess limitations to the above fixed-insolation experiments by considering sensitivity experiments, based on reasonable variations in Climate Long-Range Mapping Project boundary conditions.

   Based on previous research, ice cap extent and sea surface temperatures are the two most important variables to consider. With fixed boundary conditions, the experiments are computationally efficient and provide a range of plausible climatic variability for glacial surface conditions.

3. Assess possible extreme warmth conditions as a result of a CO_2-induced climatic change. Ocean temperatures must be computed with a coupled oceanic model. The choices of simulations are mean annual insolation or annually varying insolation. To be consistent, a seasonally varying experiment with a mixed-layer ocean should be used to define a surface boundary condition for January and July. This will be computationally expensive, but is necessary to produce a consistent, comparable set of experiments. These conditions can be used to define sea surface temperature experiments similar to the fixed-insolation experiments (1.). The seasonal and fixed-insolation experiments also can be used to estimate the solution dependence on the fixed-insolation assumptions (e.g., perpetual January).

4. Complete sensitivity experiments for variations in ice extent and sea surface temperatures for a warm climate scenario in a similar fashion as for glacial climates described in the second experiment (2.), based on the results from the extreme warmth condition assessment (3.).

The above experiments provide a consistent set of experiments that are comparable and bracket the plausible range in boundary conditions applicable to future climates on the time scale of 100,000 yr. The potential climatic variability can be determined from these experiments within a computationally efficient framework, providing insight into the sense of climatic change produced by specific factors. Although a few of the experiments have been completed, it is desirable to use the identical model for all experiments and then use the intermodel comparisons (although limited in possibility) to help define the extent to which solutions may be model dependent.

Furthermore, the modeling efforts should be sufficiently flexible to learn from the data of the Cooperative Holocene Mapping Project and the modeling efforts of Kutzbach and Guetter (1984a) for intermediate time scales between 18,000 yr B.P. and the present, and the paleoclimatic data to be reconstructed specifically in association with Hanford Site investigations.

5.3 SUMMARY OF CLIMATOLOGY AND METEOROLOGY

This section provides a brief summary of the points developed in Chapter 5. Section 5.3.1 is a recap of the current climate and of the range of past climates as it is now known. This includes a few forecasts of future climate for the next 100,000 yr. Section 5.3.2 discusses the considerations that should be given to proper design of the repository in light of climate changes that can be anticipated. The third section, Section 5.3.3, lists the major studies that are still necessary to understand probable climate changes at the site. Section 5.3.4 shows how this chapter satisfies the information required by Regulatory Guide 4.17 (NRC, 1985).

5.3.1 SUMMARY OF SIGNIFICANT RESULTS

The available paleoclimatic data will allow definition of the basic factors that appear to control climatic variability over the long time spans of interest for nuclear waste isolation. Existing data from the Columbia Plateau, in the form of fossil pollen analyses, demonstrate that climatic change in the northern portions of the Columbia Plateau during the last glaciation involved significant increases of precipitation as well as drastic cooling. During this time, a major ice sheet covered nearly all of British Columbia and extended into northern Washington, initiating major floods that affected the Pasco Basin. Atmospheric general circulation model experiments show that the wind patterns of the area probably suffered reversal of the dominant flow; during the glacial maximum, high-level winds came from the northeast rather than the west, as is the case today. Oceanic circulation patterns were deflected by ice shelves in the Gulf of Alaska and cooler sea surface temperatures probably contributed to intensified circulation systems and more frequent incursions of arctic air, which reached further to the south. Such long-term changes appear to be a response to changes in global insolation patterns that reflect the varying orbit of the earth.

5.3.1.1 Meteorology of the Hanford Site

The recent climate of the Hanford region is generally classified as mid-latitude semiarid or mid-latitude desert. The Pacific Ocean, Cascade Range, Rocky Mountains, and local terrain features significantly impacted the climatic conditions of the Hanford region. Summers are warm and dry; winters are cool with periods of overcast skies and occasional precipitation. Hurricanes, tornadoes, intense thunderstorms, hailstorms, and other severe weather events occur infrequently.

An extensive meteorological measurement program has produced a detailed data base of the recent climate of the Hanford region. Formal meteorological measurements in the Hanford region began in 1912; the Hanford Meteorology Station and its 122-m (400-ft) instrumented tower began operation in 1945. Atmospheric pressure, precipitation, humidity, sky cover, incoming solar radiation, atmospheric stability, and other parameters are measured at, or observed from the station. Atmospheric temperature, wind direction, and wind speed are measured at several levels on the instrumented tower. In 1982, a network of automated sensors began the measurement of near-surface wind directions and speeds at key locations scattered throughout the region.

The mission of the Hanford meteorology program is to meet the meteorological and climatological needs of the DOE and the Hanford Site contractors. In particular, the program calls for the measurement, observation, and recording of climatological data; continuous monitoring of regional weather; forecasting of weather for site operations; and collection of data for environmental reports, atmospheric dispersion models, and emergency response purposes. In fulfilling their mission, the staff of the Hanford Meteorology Station have amassed a data base that characterizes in detail the recent climate of the Hanford region. Because the station is located within the controlled area study zone, the current data base and monitoring program should fulfill the requirements for recent climatic data. Significant additions or modifications to the existing meteorological monitoring program should not be required.

Much of the detailed meteorological information presented in this report is excerpted from the most recent edition of "Climatological Summary for the Hanford Area" (Stone et al., 1983).

5.3.1.2 Climatic change in the Columbia Basin

Climate in the Columbia Basin has varied in the past several hundred years. This variation includes changes in temperature and precipitation. Temperature in the winter, when precipitation is highest, has decreased by over 0.3 °C, and precipitation has decreased in the autumn by over 1.3 mm (a 13% increase). Modern climatic records, which extend back only a few decades, are insufficient to capture the variability of temperature and precipitation in the region over the past 300 yr. Nevertheless, though this variability is more random than predictable, these factors will need to be included in hydrologic models used for site characterization.

Over longer time spans and at all scales of observation, climate has exhibited measurable variations. Both temperature and precipitation changes have occurred. These changes have been sufficient to produce advances of glaciers within the Cascade Range of a few kilometers in the past few thousand years. Several such episodes of advance and retreat have occurred in the past 10,000 yr.

Within the past 30,000 yr, climate has changed on a regional and global scale. That change was sufficient to produce an ice sheet in British Columbia covering nearly all of that province and extending into northern Washington State. In the past million years, such changes have occurred many times in a predictable pattern. That pattern of change can be expected to continue for at least the next 100,000 yr. Several extremes of glaciation can be expected in that time.

As these changes occur, the climates of the Columbia Basin also change. Most of the evidence for this comes from studies of fossil pollens. At present, only one pollen record extends back through the last glaciation. That record, only a 100-km distance from the Hanford Site, suggests that during extremes of glaciation, the Columbia Basin was a cold and relatively arid periglacial steppe. Other areas, closer to the glacier margin, may have been relatively humid at that time. It is clear that climates within this region can change much more than would be expected from extrapolations of modern instrumental records.

Local climatic change probably reflects the regional patterns of sea surface temperatures in the northeast Pacific Ocean. Such temperatures decreased greatly during the last glaciation. The present pattern has existed for only a short portion of the Quaternary and may be atypical. Other important controls are exerted by the regional windflow patterns and the proximity of the Cordilleran ice sheet. Some information on paleowindflow patterns has been obtained through the use of general circulation models of the atmosphere.

Forecasts of future climatic change must recognize the global nature of such changes. Important control of that change is given by changes in the orbital configuration of the earth. Solutions of the required predictive equations are available to allow such forecasts; however, the uncertainty in these forecasts is poorly understood.

5.3.2 RELATION TO REPOSITORY DESIGN

The magnitude of climatic change is unlikely to be sufficient to lead to erosion that can expose the repository. Some erosion is likely, and this will decrease the thickness of the geologic column at the repository. Floods of large magnitude will accompany future glaciations. Such floods will erode surficial materials in the Pasco Basin and will subsequently deposit additional unconsolidated sediments above the remaining material. Erosion by the floodwaters will probably be more significant than normal erosion events. This will also decrease the thickness of the geologic barrier.

Deposition of flood-transported sediments will mantle the region with additional materials of high porosity and permeability. This could lead to enhanced infiltration. Such increased infiltration could modify the configuration of the groundwater system. Floods can affect the groundwater system in another way. Such floods would last approximately 10 to 14 d (Craig and Hanson, 1986). During the flood, depths of water may exceed 200 m. Pressures at the base of the water column may be sufficient to induce recharge of the groundwater in the basalts. This likelihood is enhanced by the scouring of pre-existing sedimentary cover and by the fractured nature of the basalts at the surface.

Independent of the occurrence of floods, climatic changes within the Pasco Basin could lead to changes in the groundwater system. Depending on the nature of that change, infiltration of precipitation and recharge of the groundwater system may increase or decrease. Changes in both directions are likely both in time and space. Thus, recharge may, at a given point, decrease for a period of time then increase later. At any given time, and relative to a previous time, recharge may have increased in one area and decreased in some other area. Thus, changes in the groundwater system may be complex. Locations of recharge and discharge areas may shift over time.

Either independent of, or in conjunction with, climatic changes, the discharge of the Columbia, Yakima, and Snake Rivers may vary over time. Such variations can influence the erosion/deposition regime of the rivers. Variations in discharge of these rivers might also lead to modification of the location of recharge and discharge zones of the groundwater system.

Virtually all climatically induced changes that are of interest could impact the repository through modifications of the following:

- Rates of recharge and (or) discharge.
- Areal extent of recharge and (or) discharge zones.
- Locations of recharge and (or) discharge zones.
- Distance to the accessible environment.

Thus, repository design and assessment of repository performance must identify the potential directions and magnitudes of such changes. The objective of that analysis will be to determine if the credible range of these changes is sufficient to produce significant changes in system performance (see Section 8.3.5.2.4.4).

5.3.3 IDENTIFICATION OF INFORMATION NEEDS

Information needs can be recognized in two distinct areas. First, there is a need to develop a more systematic estimate of future climates and the uncertainty in those estimates. These estimates will require the use of computer models capable of representing the factors that can be expected to influence the local climate. A second need is a greater understanding of the paleoclimate and the Quaternary history of the Pacific Northwest. Each of these needs will be discussed in turn the following sections.

The basis of the identification of these information needs was provided in Section 5.2, and plans for satisfying them were summarized in Section 5.2.3. A more detailed formulation of those plans is provided in Section 8.3.1.5. Section 8.2 provides a synopsis of the information needs in the context of other information needs of site characterization.

5.3.3.1 Future climatic change

Forecasts of the future climate of the Pasco Basin must rely on computer models. Such models will allow representation of controlling climatic factors that can be expected to vary in the next 100,000 yr. Climatic change in the Pasco Basin will be controlled by global variations in the atmosphere, oceans, and cryosphere. These, in turn, will be strongly influenced by the variations in the configuration of the Earth's orbit.

5.3.3.1.1 Future global climates

Two distinct modeling efforts are required to estimate the global climatic changes that control local climate at the Hanford Site:

- o Models to forecast orbital influences.
- o Models of the atmospheric and oceanic responses.

Such models exist but must be adapted to the specific needs of site characterization, which are validation of the computer code, scenario analysis, and estimation of uncertainties.

Global controls such as orbital changes and anthropogenic factors must be represented to adequately characterize the climatic changes that can occur at the site. Since they are global in nature, any site chosen for disposal of high-level radioactive waste will be influenced by the same global changes. There is a clear need for all sites to use consistent scenarios to describe the global changes that are to be expected in the climate. This requires a common program coordinated by the DOE.

5.3.3.1.2 Future local climates

The local climate will respond in distinctive ways at each site considered. Modeling such local changes will require a separate set of computer programs. For the Hanford Site, the local climatic model must take into consideration the following factors:

- o Atmospheric circulation in the Pacific Northwest.
- o The proximity of glaciers and their disruption of the atmospheric circulation.
- o Sea surface temperature patterns in the northeast Pacific Ocean.
- o The local topography of the Columbia Basin, Cascade Range, and other neighboring mountain ranges.

An example of such a model is that developed for a similar application at the Nevada Test Site (Craig, 1983, 1984; Craig and Roberts, 1983, 1984).

Forecasting the local climatic response in the Pasco Basin will also require the ability to represent the response of the hydrologic system and cryosphere. A computer code will be required to represent the surface water runoff, infiltration, and river flow within the Columbia River Basin. This code can be patterned after a similar code that was developed for the Nevada Test Site (Craig, 1983; Singer and Craig, 1983). Enhancements of the Nevada Test Site code will be required to allow detailed representation of the hydrologic cycle, especially evapotranspiration and infiltration.

Representation of the cryosphere will require development of a computer code representing the two-dimensional flow characteristics of glaciers and ice sheets. Such codes have been developed for other areas but must be adapted to the particular topographic and climatic setting of the Pacific Northwest.

Because of the concern for future glacially induced flooding within the Pasco Basin, it is necessary to link the glacier flow model to a model of the local surface water system. This will allow forecasting the boundary conditions of such floods in the future. The floods themselves must be modeled with a two-dimensional dynamic representation. This is needed to estimate the potential for disruption of the local hydrogeologic system within the Pasco Basin.

5.3.3.2 Past climatic change

Information about past climatic changes may be used in two ways. It can be used to provide direct estimates of the range of conditions that should be anticipated in the future. The information also can be used to calibrate and test more quantitative models used to forecast climatic change. Both of those applications are required during site characterization.

Four major areas of study are required to characterize important paleoclimatic information:

- Reconstruction of late Quaternary paleoclimates within the Columbia Basin, especially using fossil pollen data.
- Quaternary glaciation in the Pacific Northwest.
- Oceanography, especially the history of sea surface temperature variations, in the northeast Pacific Ocean.
- Eolian deposits of the Columbia Plateau.

Results of those studies are needed to estimate the range of future climates and test models to be used for forecasts.

5.3.3.2.1 Late Quaternary paleoclimates

Needed studies for paleoclimate characterization include the following:

- A more comprehensive assessment of the most recent climatic history of the area.
- A systematic characterization of paleoclimates in the late Quaternary are needed for paleoclimatic characterization.

The first of these needs must be satisfied with dendroclimatic analyses, and the second must be satisfied primarily by pollen analyses. These analyses must be supplemented by additional data sources, which will include geochemical analyses, diatom studies, and other biotic and nonbiotic climatic proxy data.

Dendroclimatic work is required to reduce the standard error of the existing estimates of paleoclimates of the last 300 yr. That standard error is high because of the small number of trees that have been used for the existing reconstructions. Increased density of the tree-ring grid in the neighborhood of the Pasco Basin is also needed to describe more of the variance of climate in that local area. This dendroclimatic reconstruction must use climatic data that is currently available but that was not used for the existing reconstructions. Quantitative estimates of paleo-atmospheric pressure patterns should also be obtained. Quantitative estimates of the uncertainties in these reconstructions are also required.

Paleoclimatic reconstructions employing fossil pollen data are needed to provide quantitative estimates of climate and estimate the uncertainty in these reconstructions. Additional sites having a record at least as long as the Carp Lake record should be studied to test the replicability of that record. These are also required to provide a basis for the interpolation of the spatial variability of climate. Additional paleoclimatic data should be collected to characterize the response of the groundwater system to fluctuations in climate. The sensitivity of the groundwater system to climatic changes that can be resolved with the pollen record should be investigated. Reconstructions of climate at 3,000-yr intervals, from 21,000 yr ago, should be made to calibrate and test the climatic models to be used for future forecasts.

5.3.3.2.2 Quaternary glaciation

Details of the pattern and timing of the last glacier expansion in British Columbia are needed. These will be used to understand the relation between climatic change and growth of the ice sheet. This relation will form the basis of the model of future glacier growth. The reconstructions of the glacial history of the region should emphasize the same time horizons as the paleoclimatic reconstructions. This will provide a common basis for interpretation of the climate/glacier relation. This information is also necessary to validate the model of glaciation.

If the paleoclimatic studies using fossil pollen data are not fully successful at reconstructing climate, alternative methods can supplement that information for use in testing the local climatic model. Alpine glaciation in the Cascade Range accompanied climatic change in the Pacific Northwest during the last glaciation (see Section 5.2.1.2.3) and can be expected to do so in the future. Such glaciation integrates climatic changes, including temperature, precipitation, and wind velocity. The local climatic model can be extended to provide estimates of the mass balance of alpine glaciers. This information, linked to an alpine glacier model, can be used to calibrate and test the local climatic model. Such tests can be accomplished by comparing model reconstructions for specified time periods to the glacial geologic record of the Cascade Range and other nearby mountain ranges that suffered alpine glaciation (Bitterroot Mountains, Blue Mountains, etc.). This strategy is further defined in Section 8.3.1.5.

5.3.3.2.3 Oceanography

Information is needed about two distinct aspects of the history of the northeast Pacific Ocean. First, the spatial pattern of sea surface temperatures should be reconstructed for the late Quaternary. Second, the long-term record of sea surface temperature variations in at least one location in the northeast Pacific Ocean should be compiled.

Sea surface temperatures must be reconstructed at the same 3,000-yr intervals of the late Quaternary as described previously for paleoclimates and glaciers. These three data sets, together with topographical information about the region, will allow an accurate model of climatic change in the Pacific Northwest to be finalized. These data are also needed to complete the testing of the model of local climatic change to be used in forecasting future climates at the Hanford Site.

The long-term history of sea surface temperatures in the northeast Pacific Ocean must be obtained at a site as close to the Washington State coast as practical. This information is needed to determine the relation between the local oceanographic changes and the more global controls on those changes. The information also will be needed to estimate the uncertainties in that relation, to determine the form of the model to be used for future climatic forecasts, and to estimate the reliability of the final forecasting models to be used.

5.3.3.2.4 Eolian deposits of the Columbia Plateau

There is a considerable volume of eolian sediment preserved on the Columbia Plateau. Such sediments can include loess, eolian silt, and sand dunes. Loess deposits may have been formed throughout the Quaternary during each major climatic cycle and, because of this, those deposits may provide the best available measure of the long-term variability of climate. Few studies have been conducted on these deposits. However, the importance of this potential source of information is recognized by the BWIP, and plans are described in Section 8.3.1.5 to study these deposits to extract the long-term paleoclimatic signal.

The loess record of the Pacific Northwest is more detailed than the glaciation record of this area and nearly as complete as the more distant marine record of the northeast Pacific Ocean. The eolian record contains information on paleowind directions, which express boundary layer effects directly. This will provide an important test and control for the modeling program. Wells (1983) has tied the full glacial paleowind direction information of the Pacific Northwest into a reconstruction of the continental scale boundary layer circulation pattern.

5.3.4 RELATION TO U.S. NUCLEAR REGULATORY COMMISSION REGULATORY GUIDE 4.17

The Regulatory Guide 4.17 (NRC, 1985, Chapter 5) suggests items that the DOE should consider with respect to climatology and meteorology. They are divided into two parts: Part 5.1, Recent Climate and Meteorology, and Part 5.2, Long-term Climatic Assessment.

The information requested by Parts 5.1 and 5.2 of NRC (1985, Chapter 5) was summarized in Sections 5.1 and 5.2, respectively.

Part 5.2 of the Regulatory Guide 4.17 (NRC, 1985) provides an extensive list of information that is requested by the NRC as part of site characterization. In general, the requested information has been provided. Where the needed data are not available, plans have been defined to provide it as part of the site characterization. There are some specific items that have been requested in the Regulatory Guide 4.17 that are not appropriate for site characterization at the Hanford Site. Another item of information that has been requested requires interpretation of NRC intentions. These points of digression are discussed in the following paragraphs.

Items requested by the NRC (1985) that are inappropriate for the Hanford Site will be presented first.

1. The NRC requests an analysis of the paleoclimatology of the candidate area for the Quaternary. Geologic evidence is not available for the paleoclimates of the candidate area for the entire Quaternary. Nearly all geologic records in the immediate vicinity of the site extend to no more than 33,000 yr ago. There is little likelihood that information for the entire Quaternary can be reconstructed during site characterization. Instead, the similarity of the available record to that of global climatic change is discussed. The more extensive record of global climatic change is provided as a general guide to the pattern of change at the site. Some effort to use the record preserved in loess deposits to correlate with the global record is expected under the "geologic" task (see Chapter 1 and Section 8.3.1.2).

2. The NRC requests an identification of the windflow patterns of the Quaternary. In this case, it is not reasonable to expect to reconstruct any systematic description of windflow patterns, even for the late Quaternary. A few local indicators of windflow patterns are available; however, these are very difficult to date accurately. In lieu of that information, modeling of paleowindflow and future windflow patterns will be done. This will include the use of models of the global atmospheric circulation patterns and models to describe the local patterns of windflow as influenced by local topography.

3. The NRC requests information about sizes of glaciers and on accumulation and ablation rates. Considering that there are thousands of glaciers in the Pacific Northwest, attention is limited to only the largest ones. In particular, there is a discussion on the characteristics of the Cordilleran ice sheet and Piedmont glacier of the Cascade Range during the last major glaciation in the area. Information for other glaciations is so sparse that it is not useful for exact reconstructions.

4. The NRC requests information regarding all assumptions used in the procedures and models. It is not possible to enumerate all assumptions; however, disagreements between assumptions used in the procedures and models and those applied in the literature will be identified and clarified.

The NRC (1985) specifically requests information about rates of change of past and future climates. Such estimates of rates must recognize the need to specify the time period of resolution of the estimates. Very rapid climatic changes may occur but may be limited to only short durations. For example, temperatures may change by 30 °C within a single day. For the purposes of site characterization, it is assumed that the resolution of interest concerns the rates of climatic change that are maintained over long enough time spans to significantly affect the groundwater system. It is also assumed that climatic changes must be maintained over at least 100 yr to produce such an effect.

The NRC (1985) has requested information about maximum and minimum rates of change in climatic variables. Strictly speaking, a minimum rate of change would always be zero. For the purposes of site characterization, it is assumed that "minimum" rates of change are those that lead to a maximum rate of decrease of the absolute or relative values of a variable. Maximum rates of change will be interpreted to be those that increase the value of that variable.

5.4 REFERENCES

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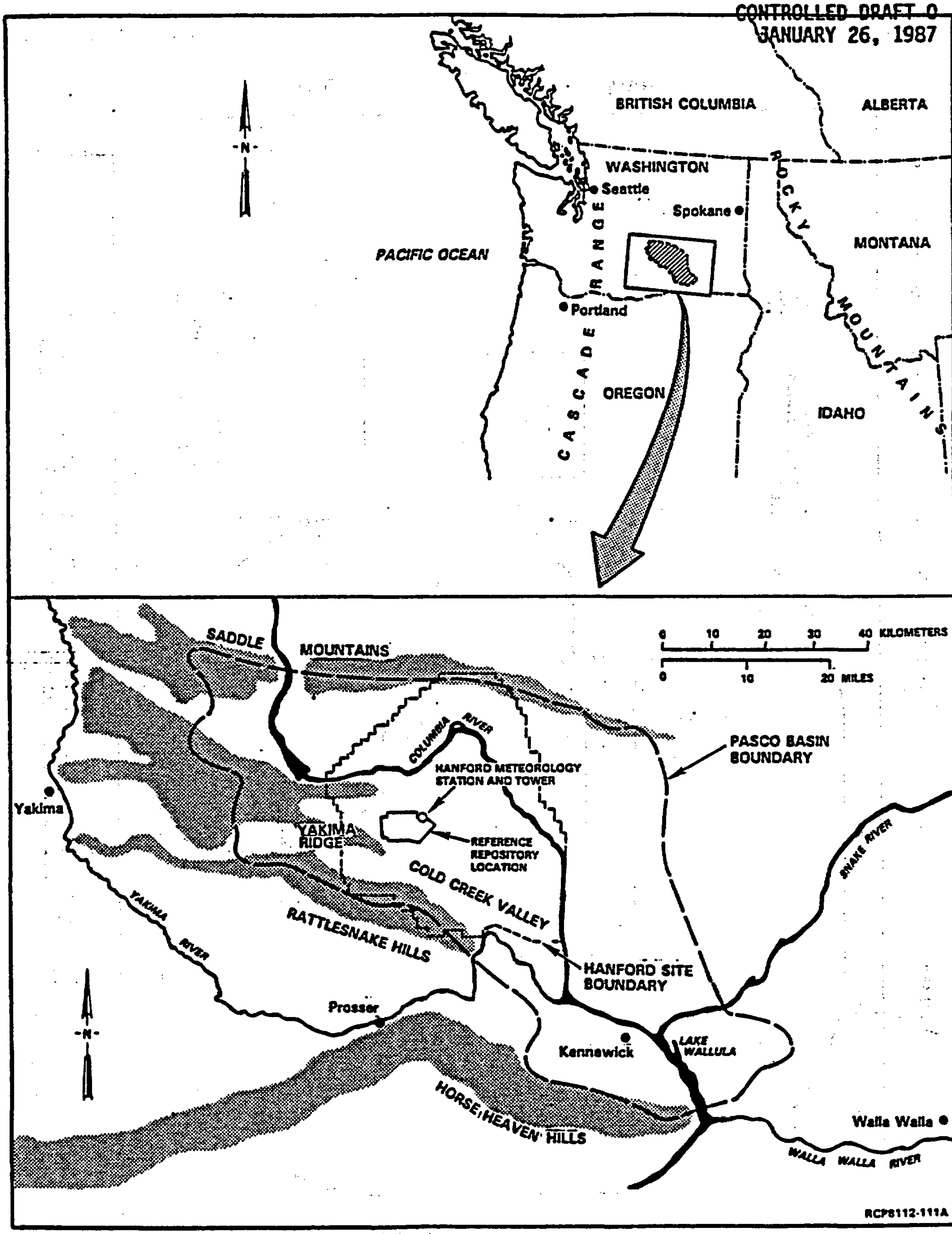


Figure 5.1-1. The Hanford Site in relation to surrounding terrain.
RCP8112-111A

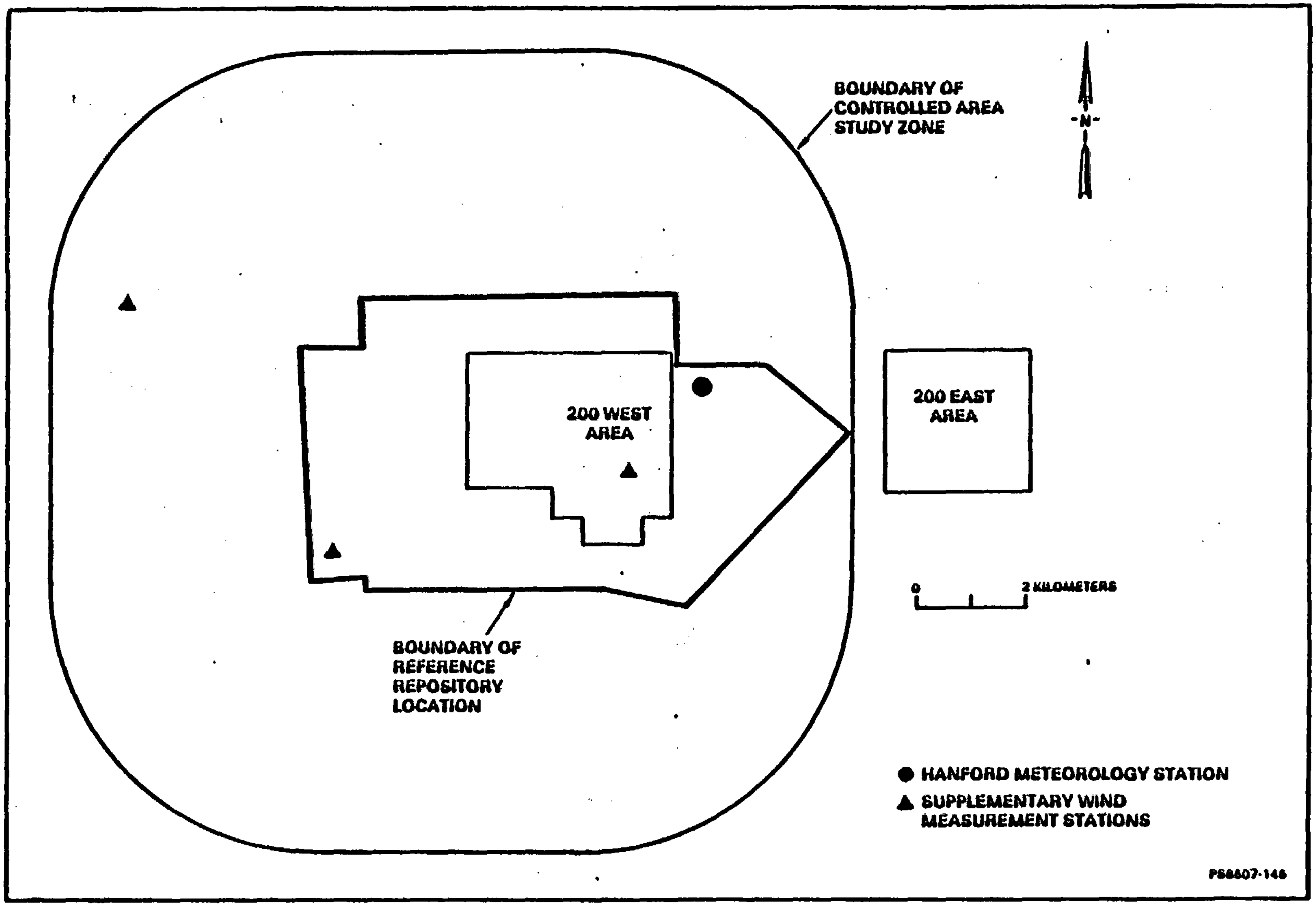


Figure 5.1-2. The locations of the Hanford Meteorology Station and three supplementary wind measurement stations within the controlled area study zone. PS8607-146

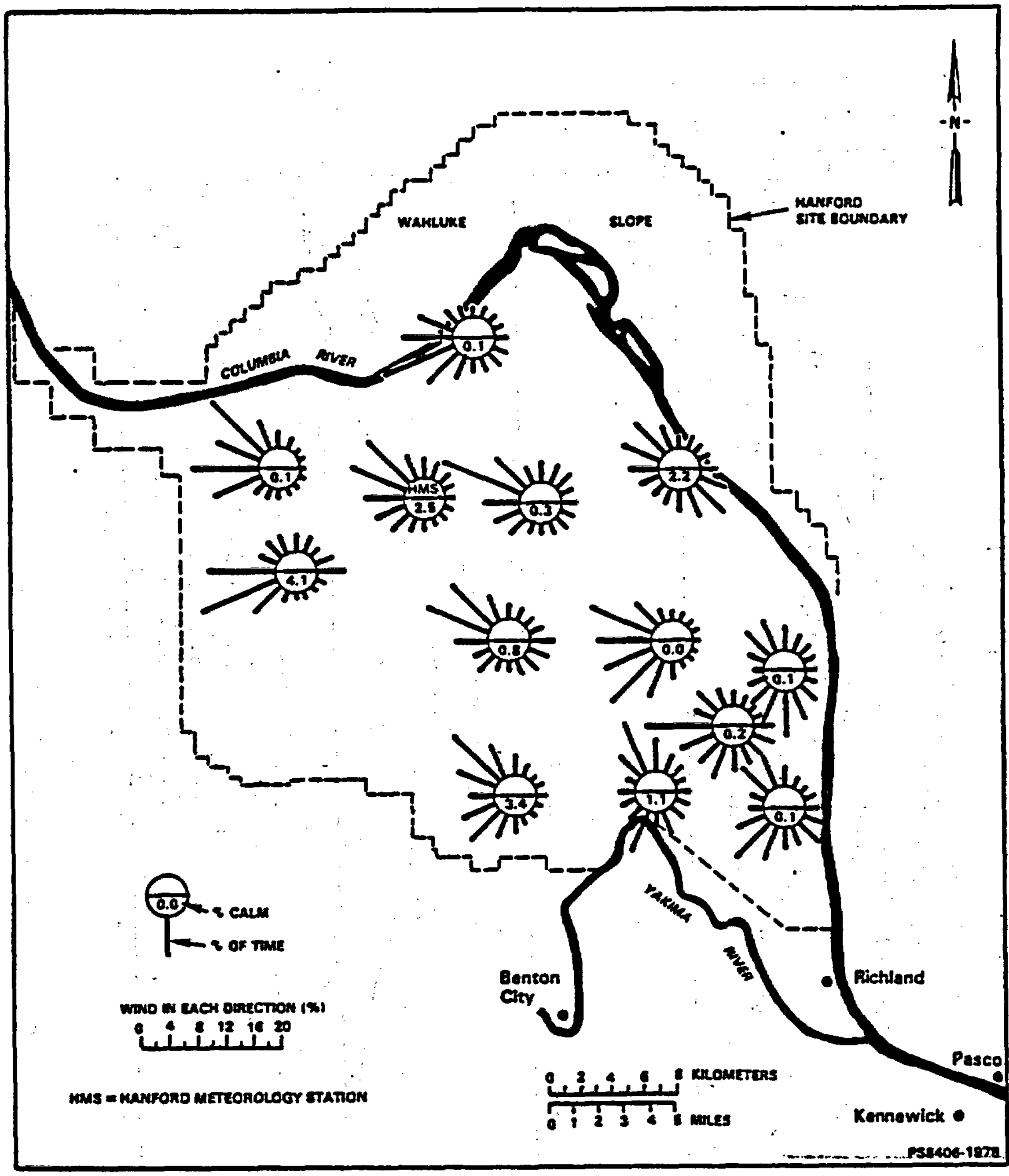


Figure 5.1-3. Wind roses for the Hanford Meteorology Station and 12 supplementary wind measurement stations (based on 3 years of hourly data). Wind roses are a convenient tool for graphically showing the frequency distribution of wind direction at a station. The wind roses in this figure consist of a small interior circle with 16 solid lines, or "paddles," running perpendicularly into the circle. The length of each of these paddles represents the average percentage of the time during the study period that the wind blows from the paddles direction category towards the station.
PS8406-1978

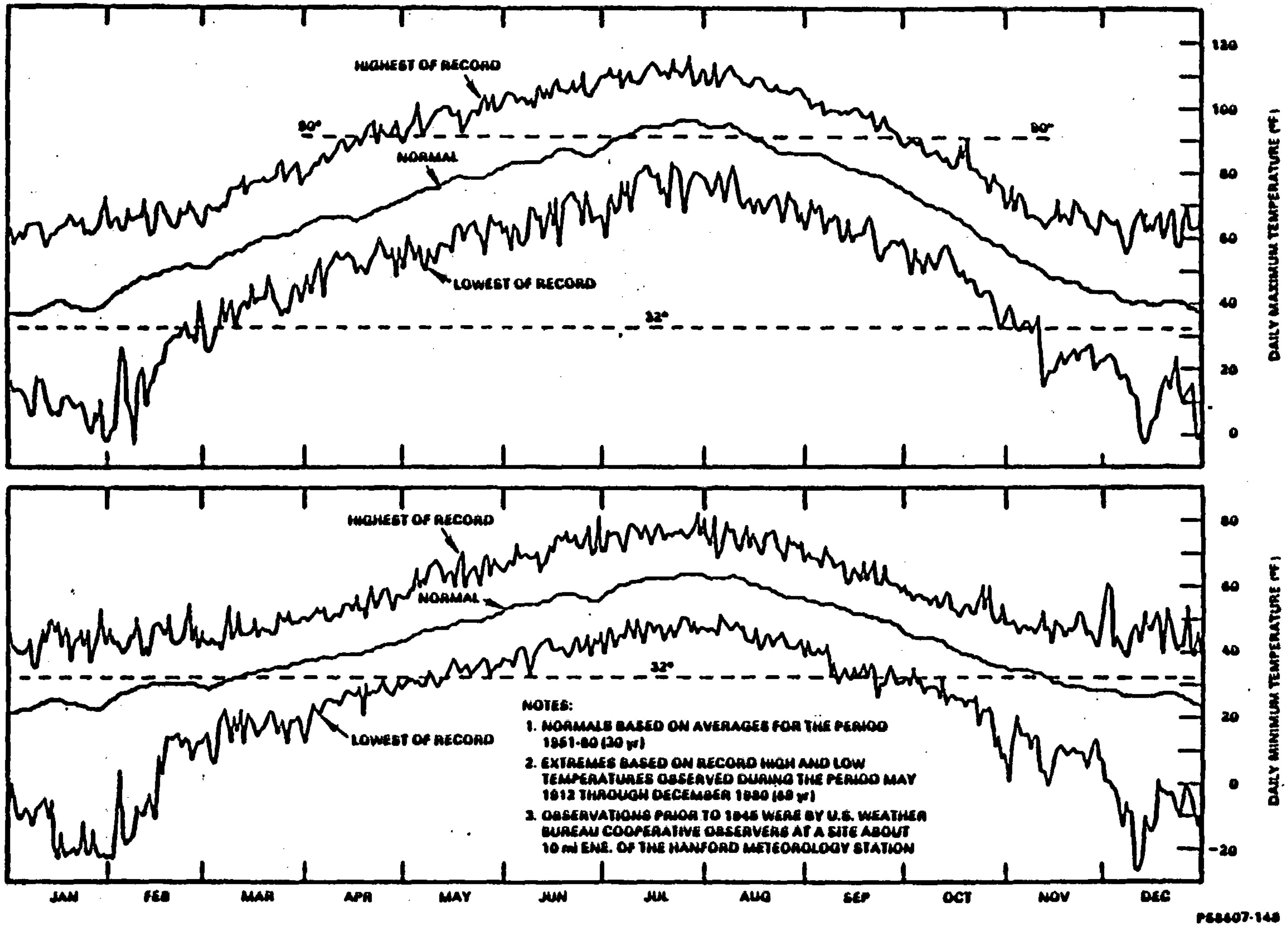


Figure 5.1-4. Normal and extreme daily temperatures at the Hanford Site.
PS8607-148

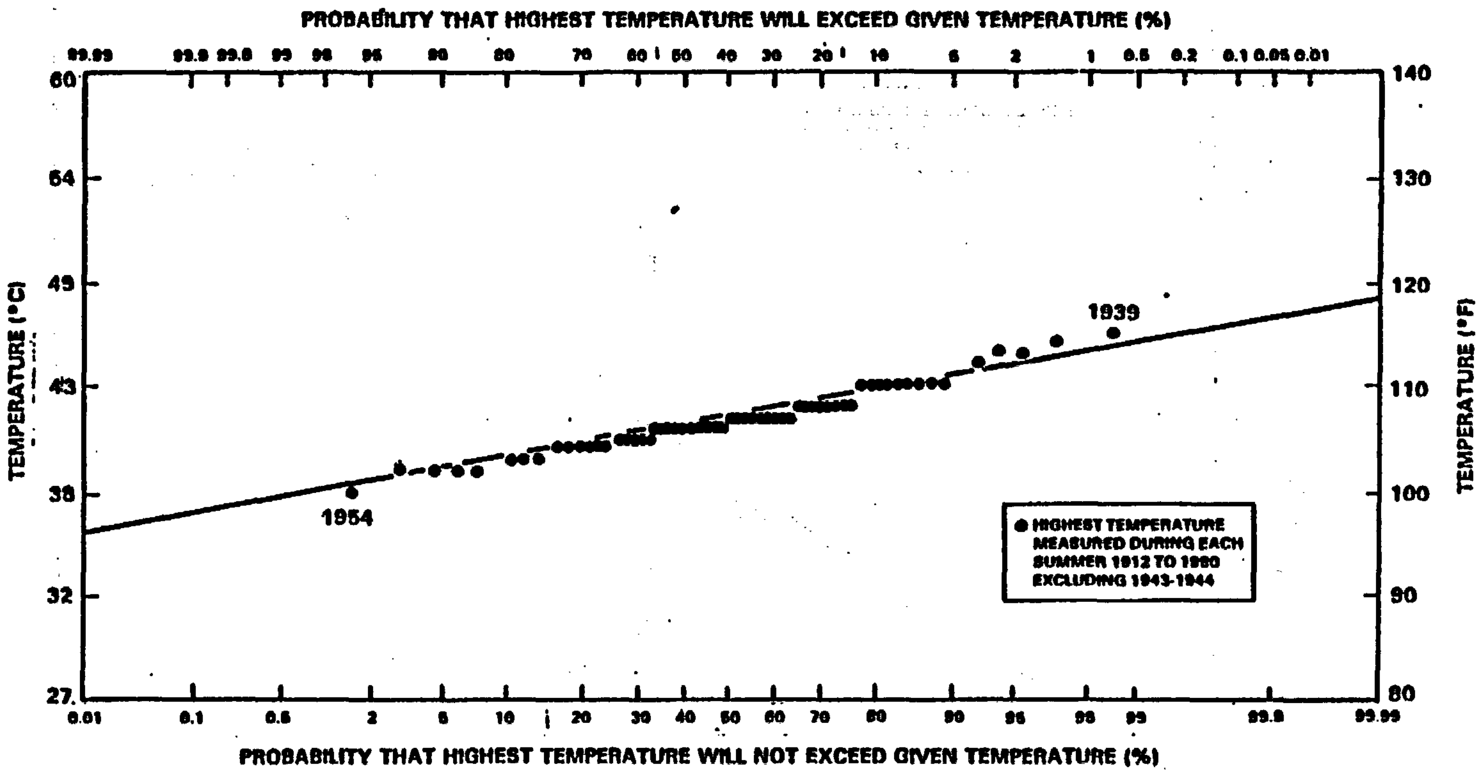


Figure 5.1-5. Plot of the highest temperatures measured during each of 66 summers of record at the Hanford Site: 1912 to 1980 (1943-44 missing). The maximum and minimum highest temperatures recorded are lableled with their year of occurrence. The straight line through the data can be used to estimate the probability (%) that the highest temperature during a future summer will exceed a particular temperature. PS8201-73A

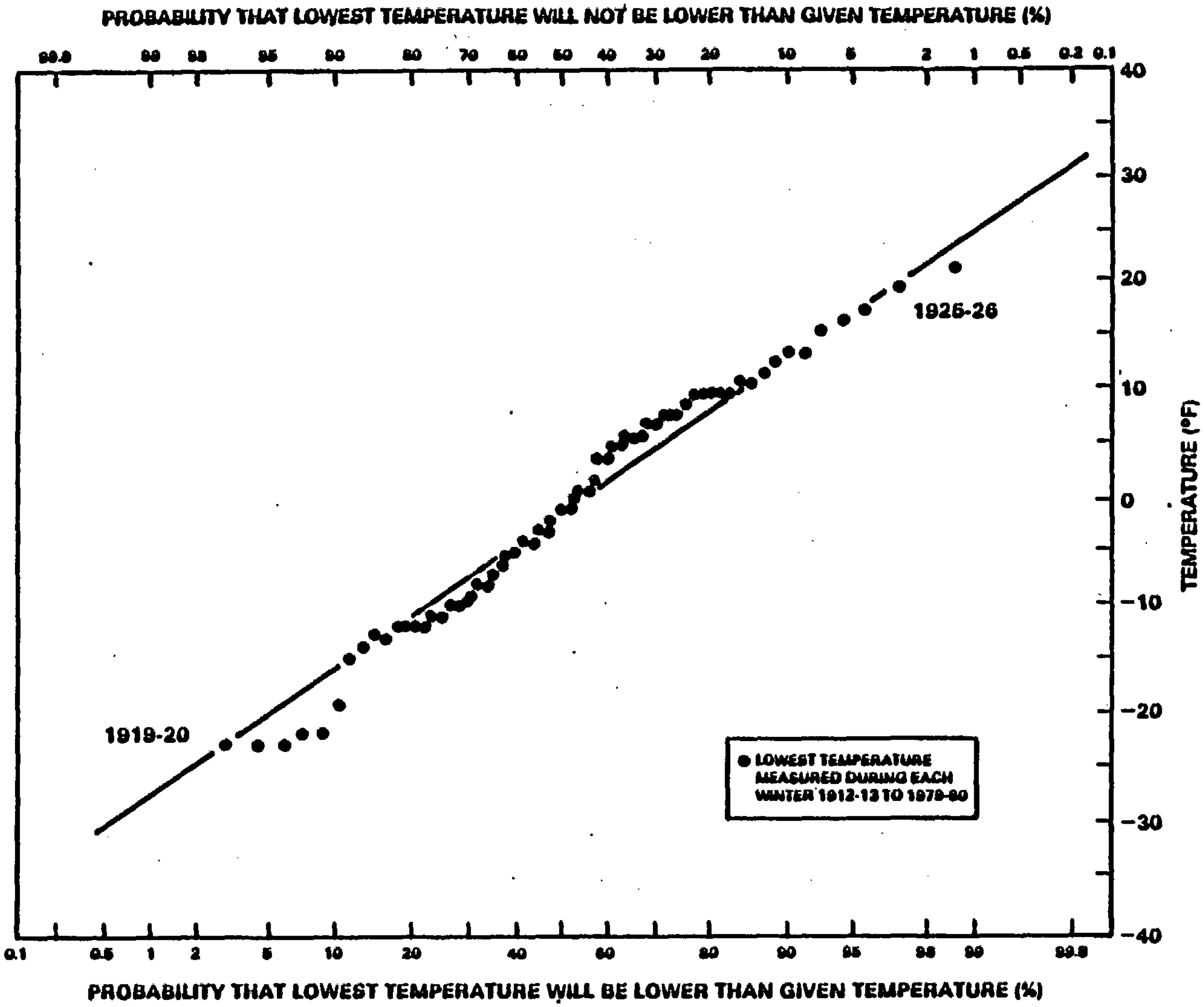
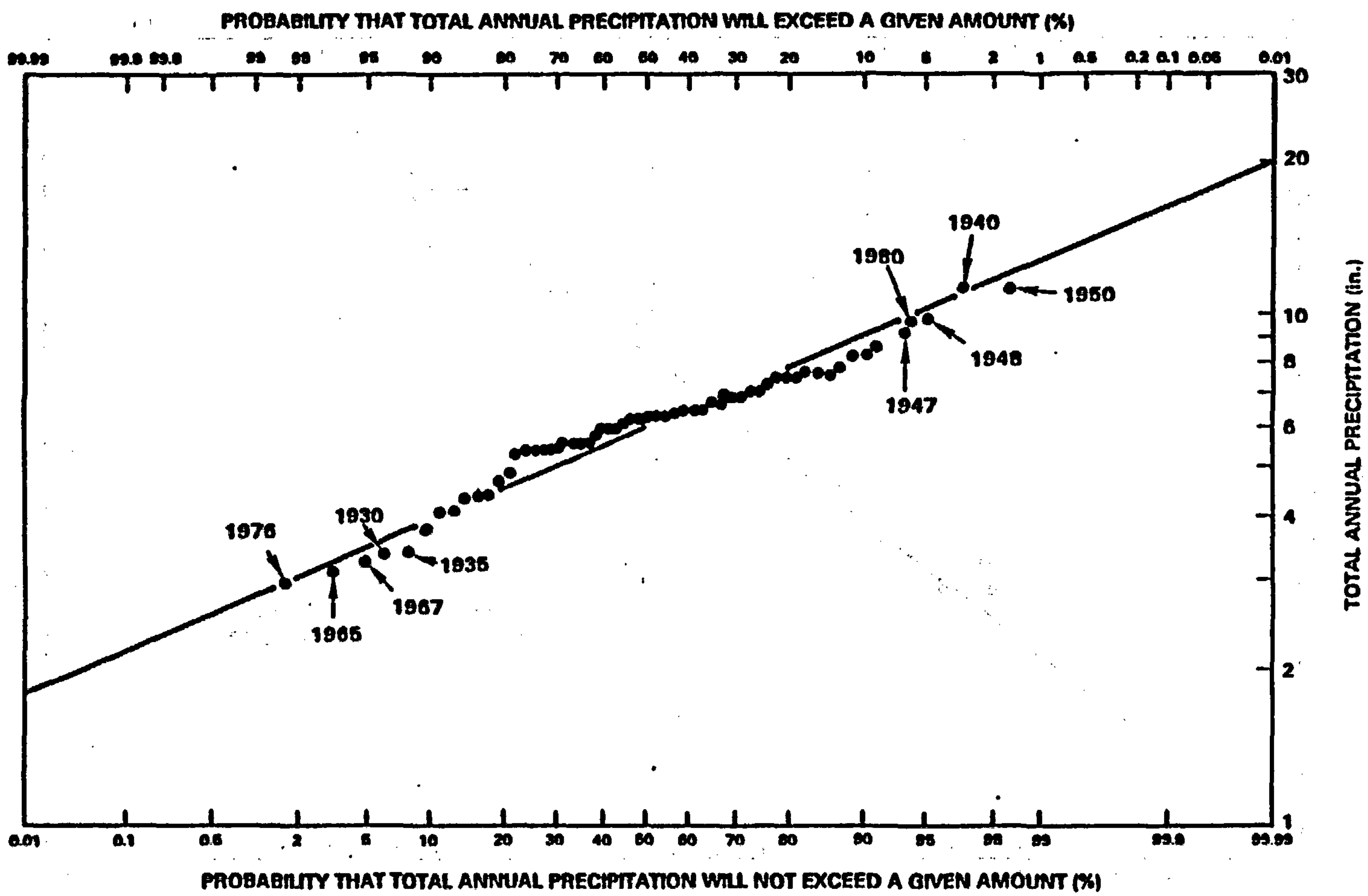


Figure 5.1-6. Plot of the lowest temperatures during each of 68 winters of record at the Hanford Site: 1912-13 to 1979-80, and the probability (%) that the lowest temperature will exceed a given temperature. The maximum and minimum of these lowest temperatures are labeled with their years of occurrence. The straight line through the dots can be used to estimate the probability (%) the lowest temperature during a future summer will exceed a particular temperature. PS8607-150



PS8607-151

Figure 5.1-7. Total annual precipitation at the Hanford Site: 1913 to 1980 (1943 to 1946 missing) and the probability (%) that the total annual precipitation will exceed a given amount. PS8607-151

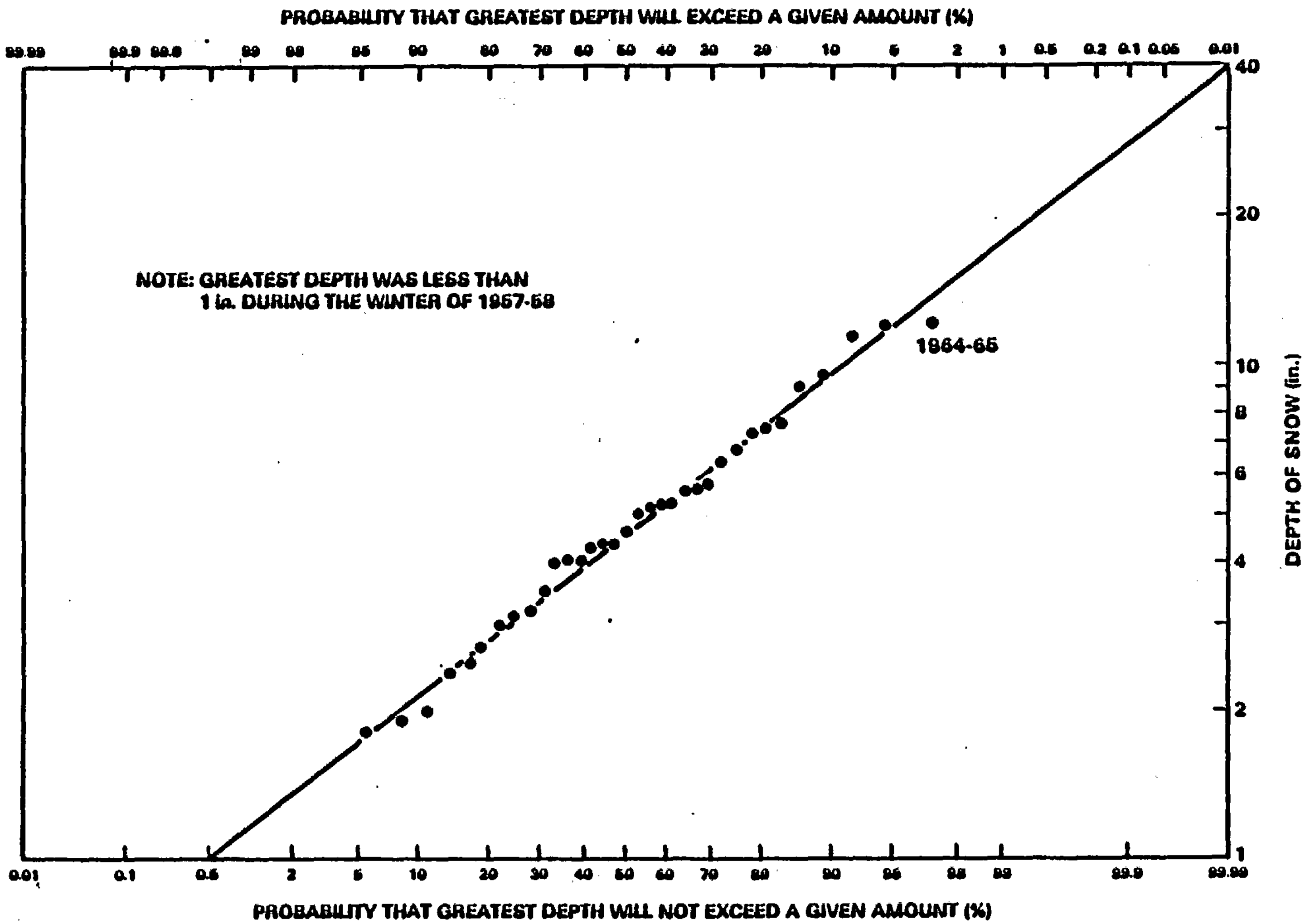


Figure 5.1-8. Greatest depth of snow on ground during winters of 1946-47 to 1980-81 and the probability (%) that the greatest depth of snow will exceed a given amount. PS8607-152

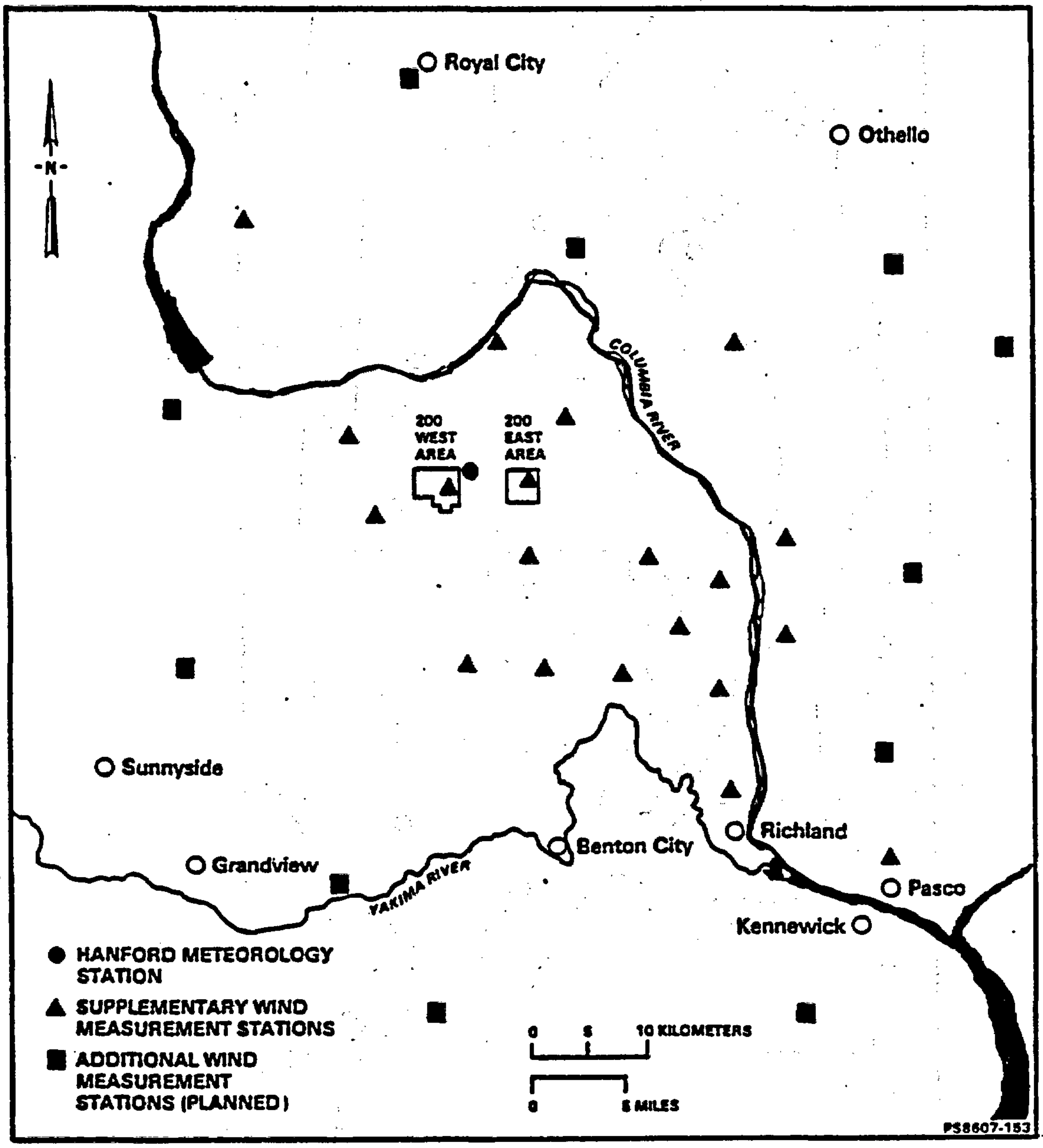


Figure 5.1-9. The location of the Hanford Meteorology Station, supplementary wind measurement stations, and some of the additional wind stations that will be added to the network. PS8607-153

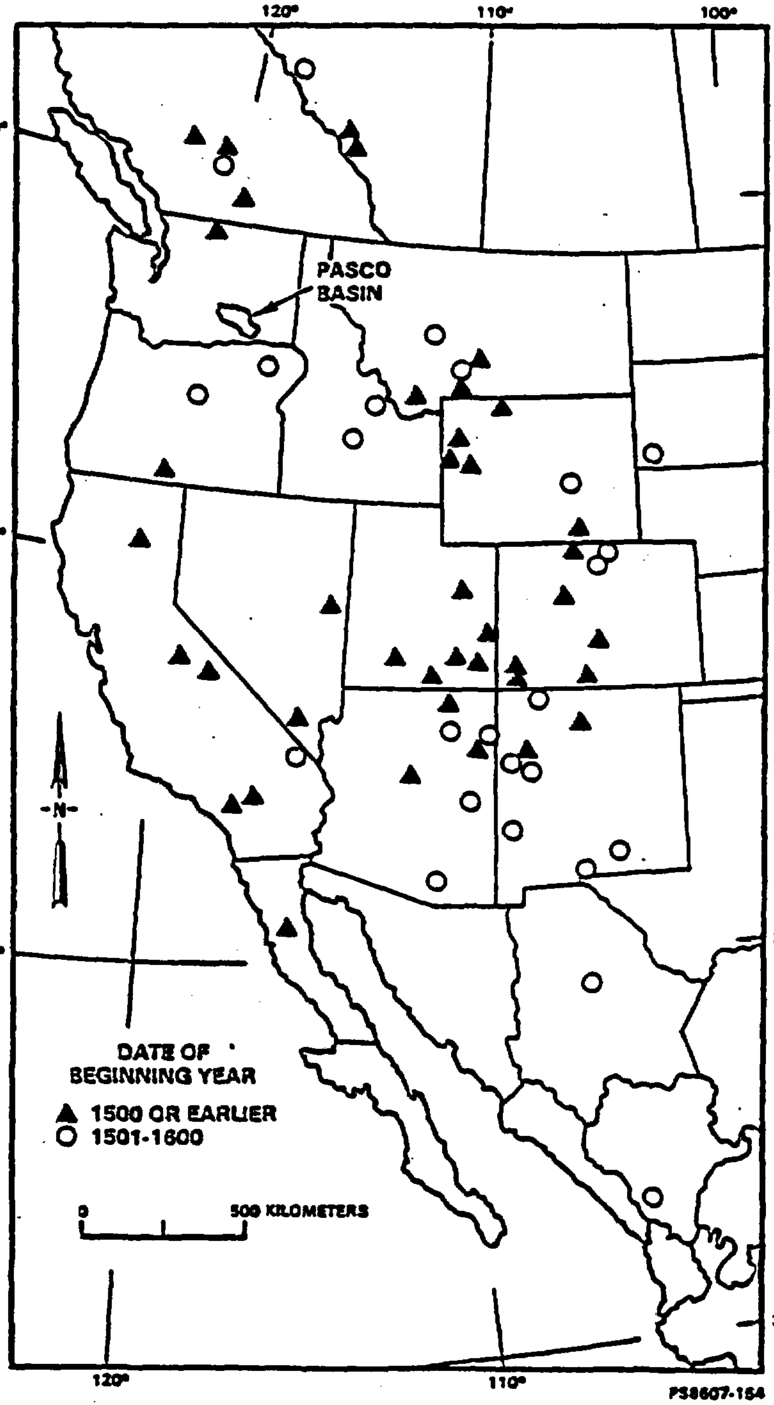


Figure 5.2-1. Locations in western North America at which tree-ring chronologies have been developed for dendroclimatic purposes (Cropper and Fritts, 1986, Fig. 1). PS8607-154

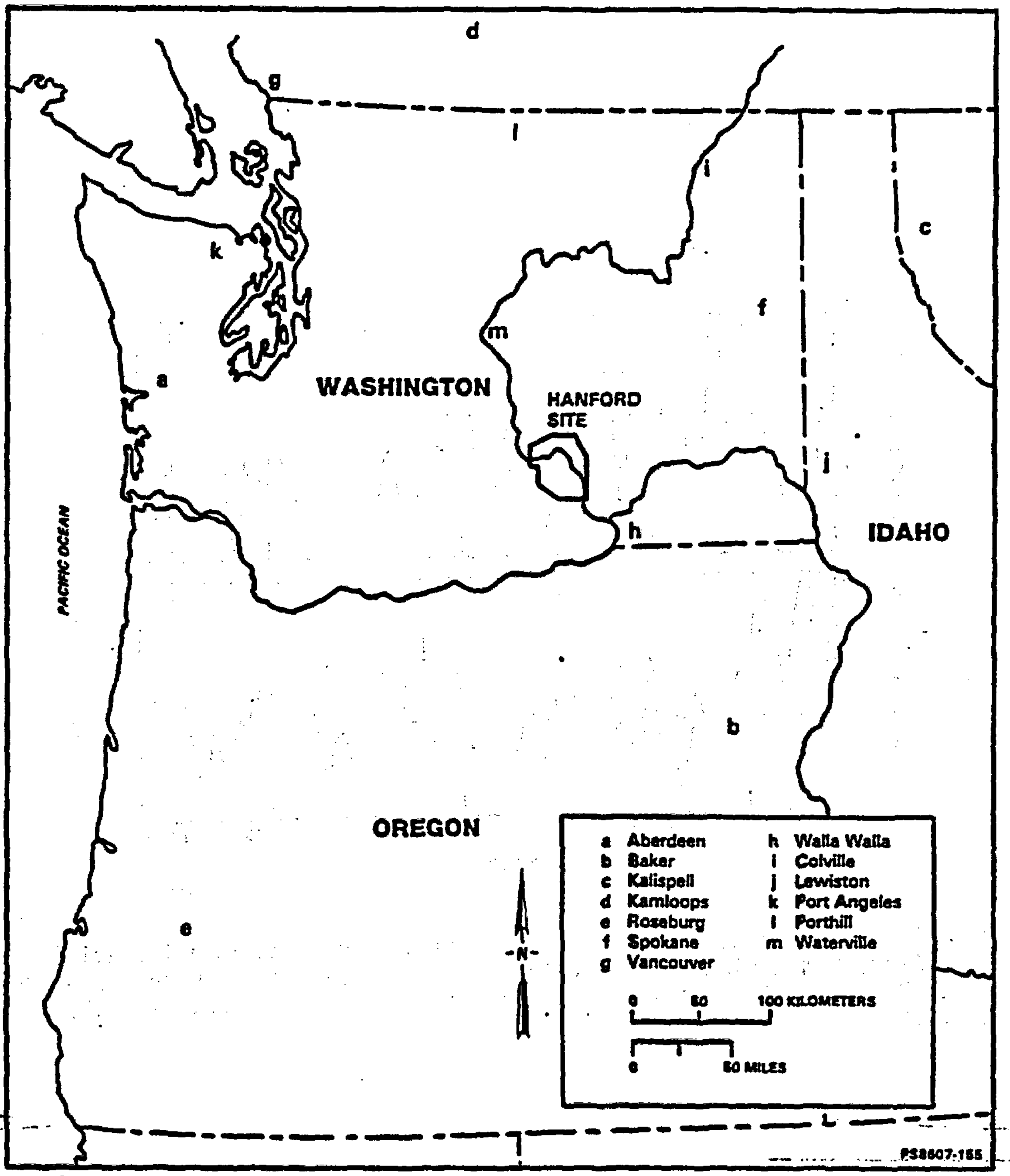


Figure 5.2-2. The locations of climate stations whose records were used in research on the dendroclimatology of the Pasco Basin (Cropper and Fritts, 1986). The station names are given in Table 5.2-1, with their corresponding letters. PS8607-155

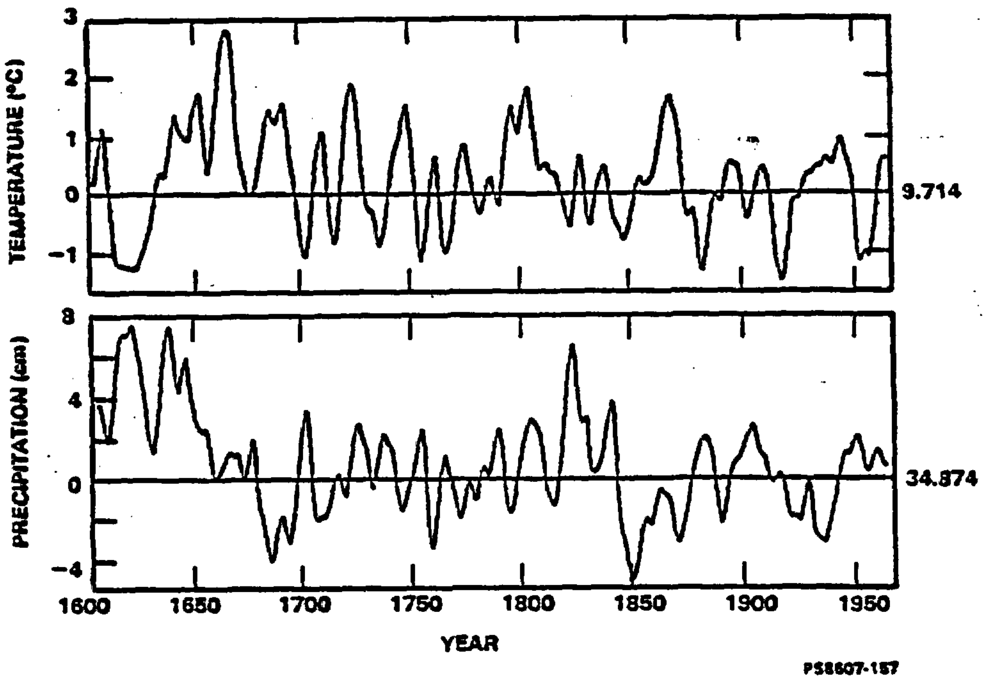


Figure 5.2-3. Reconstructed mean annual temperature and annual precipitation of the Pasco Basin since 1600. Values are expressed as departures from the recent 30-year instrumental record (Cropper and Fritts, 1986, Fig. 4). PS8607-157

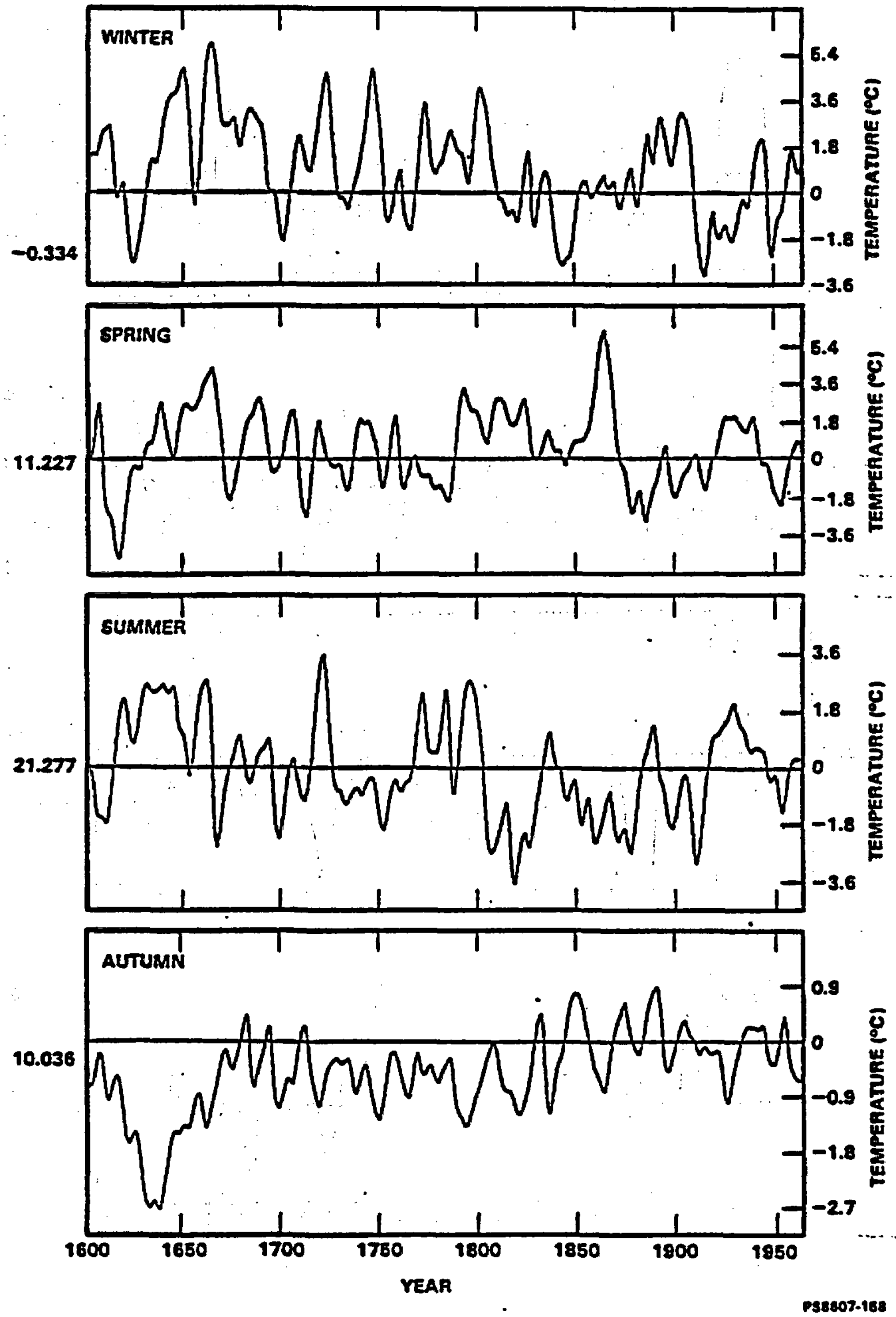


Figure 5.2-4. Reconstructed seasonal temperatures of the Pasco Basin since 1600. Values are expressed as departures from the most recent 30-year instrumental record (Cropper and Fritts, 1986, Fig. 5). PS8607-158

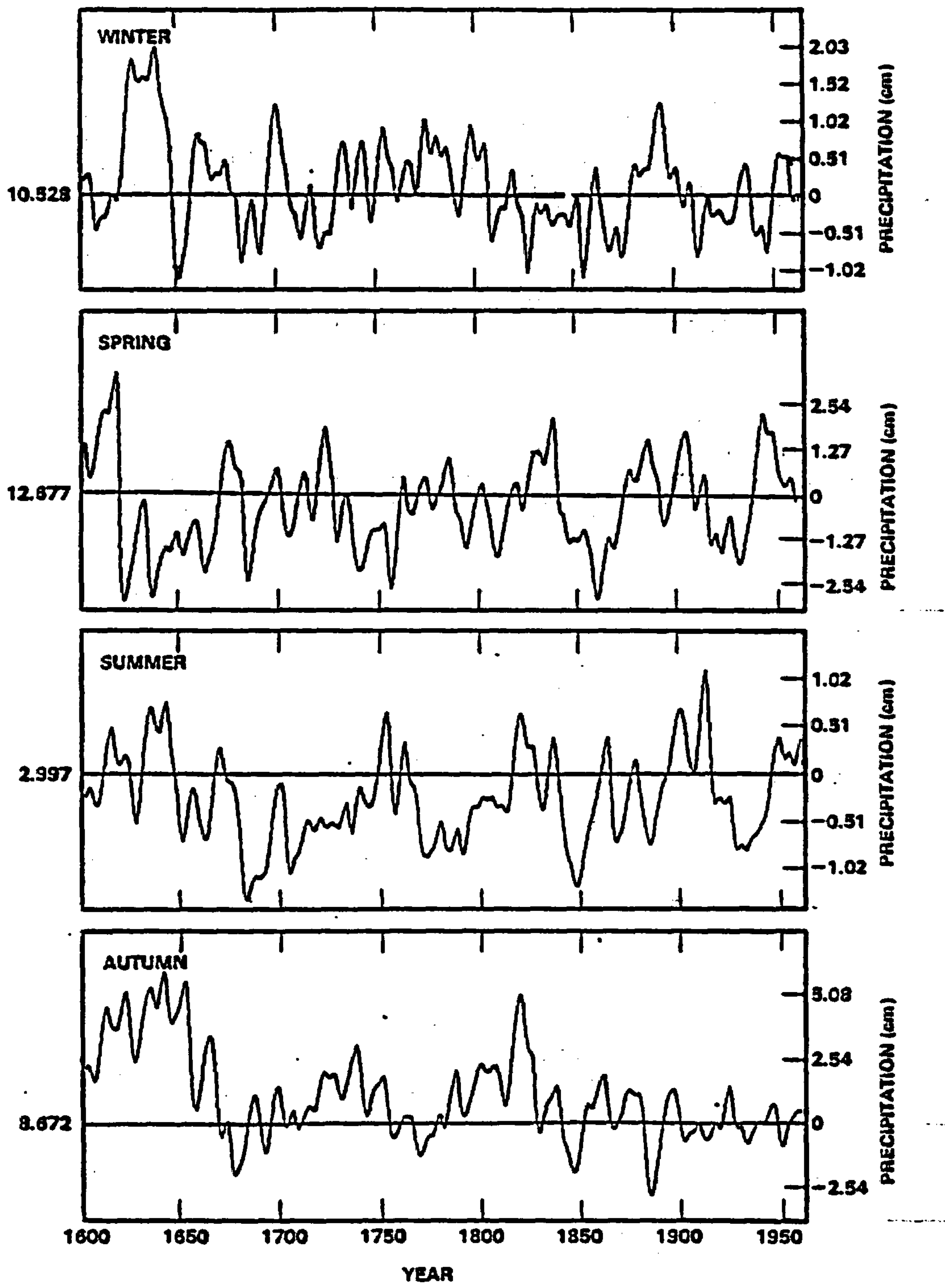


Figure 5.2-5. Reconstructed seasonal precipitation of the Pasco Basin since 1600. Values are expressed as departures from the most recent 30-year instrumental record (Cropper and Fritts, 1986, Fig. 6). PS8607-159

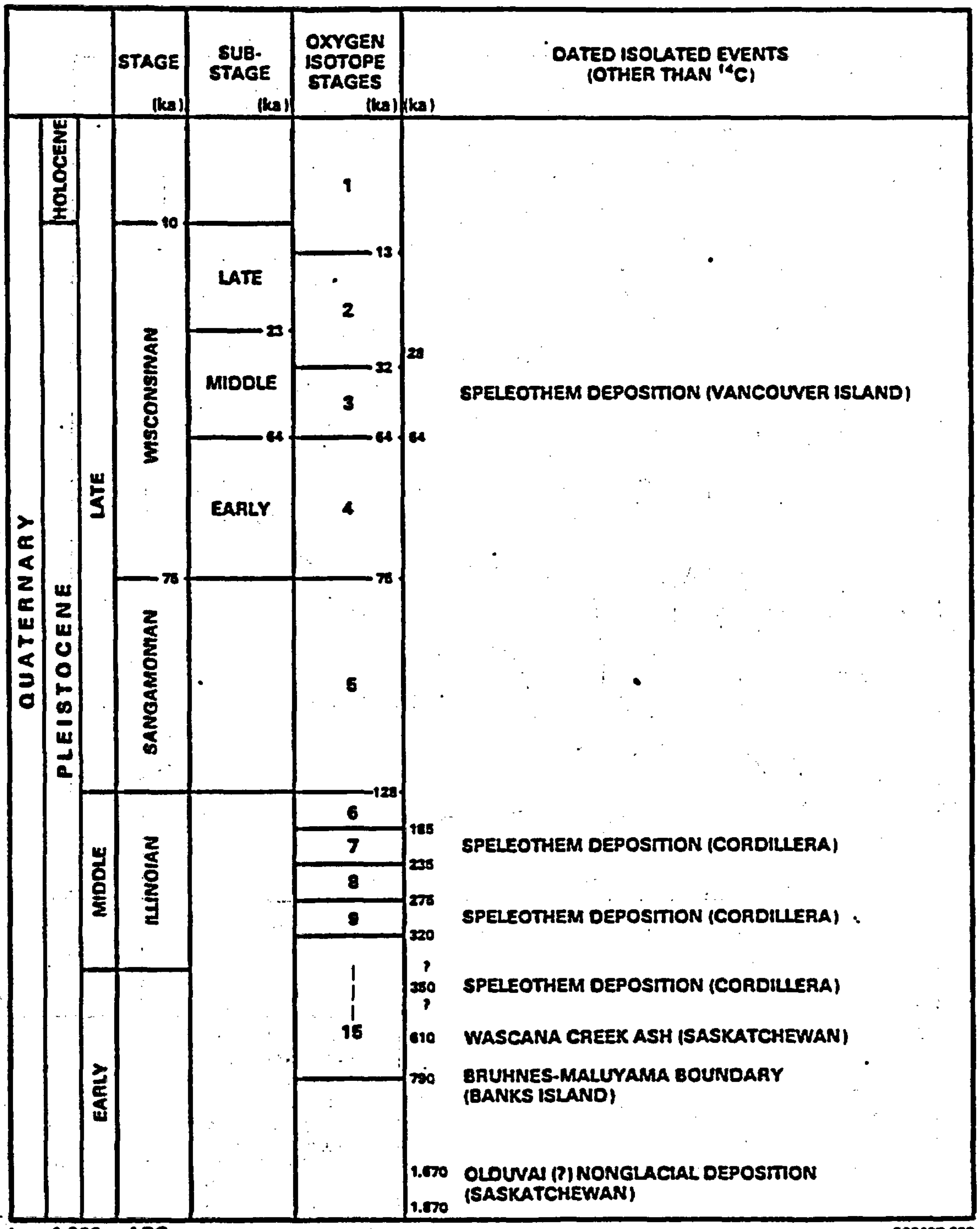


Figure 5.2-6. Time stratigraphic terms, chronology of the Quaternary Period, and some associated events in the Pacific Northwest (Fulton, 1984, Table 1, p. 3). PS8607-223

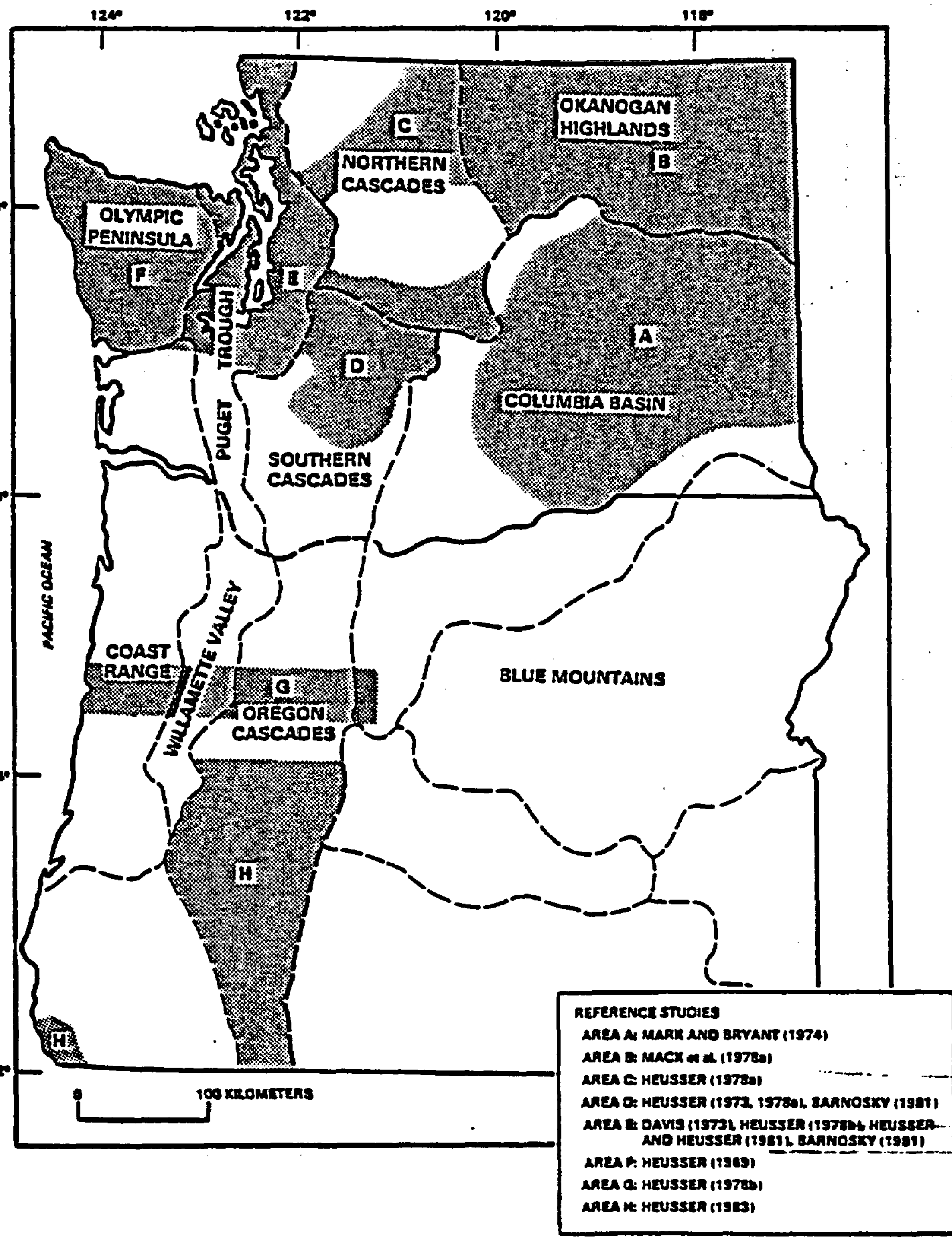


Figure 5.2-7. Physiographic provinces of the Pacific Northwest (Franklin and Dyrness, 1973). Stippled areas show location of pollen rain studies. Not shown is the Snake River Plain of Idaho, studied by O. K. Davis (1984b) and Bright (1966). PS8607-160

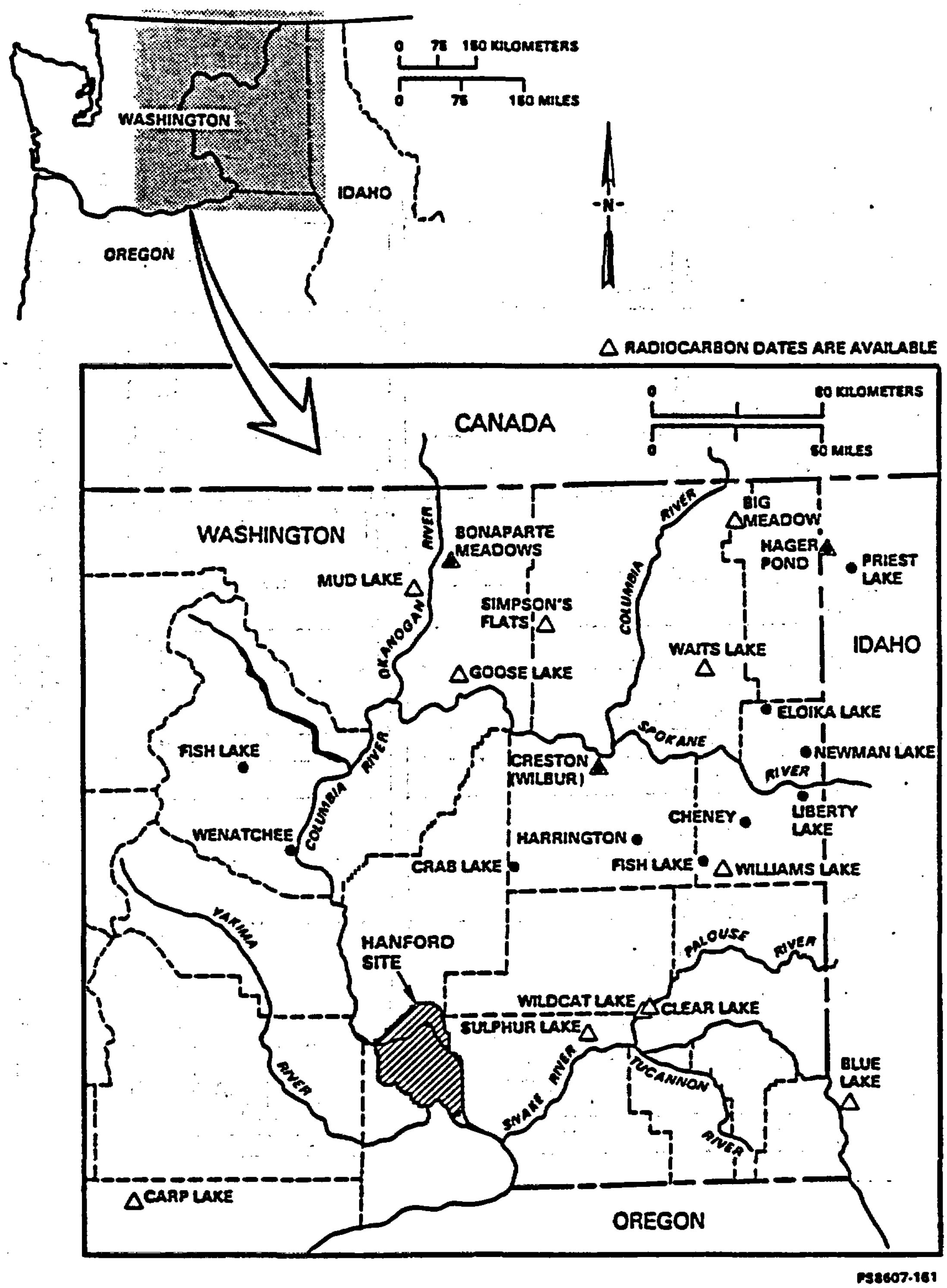


Figure 5.2-8. Locations within the Columbia Basin of Washington at which palynological samples have been obtained for climatic reconstructions. These include the six sites discussed here (Mehringer, 1985b, Fig. 16, p. 36). PS8607-161

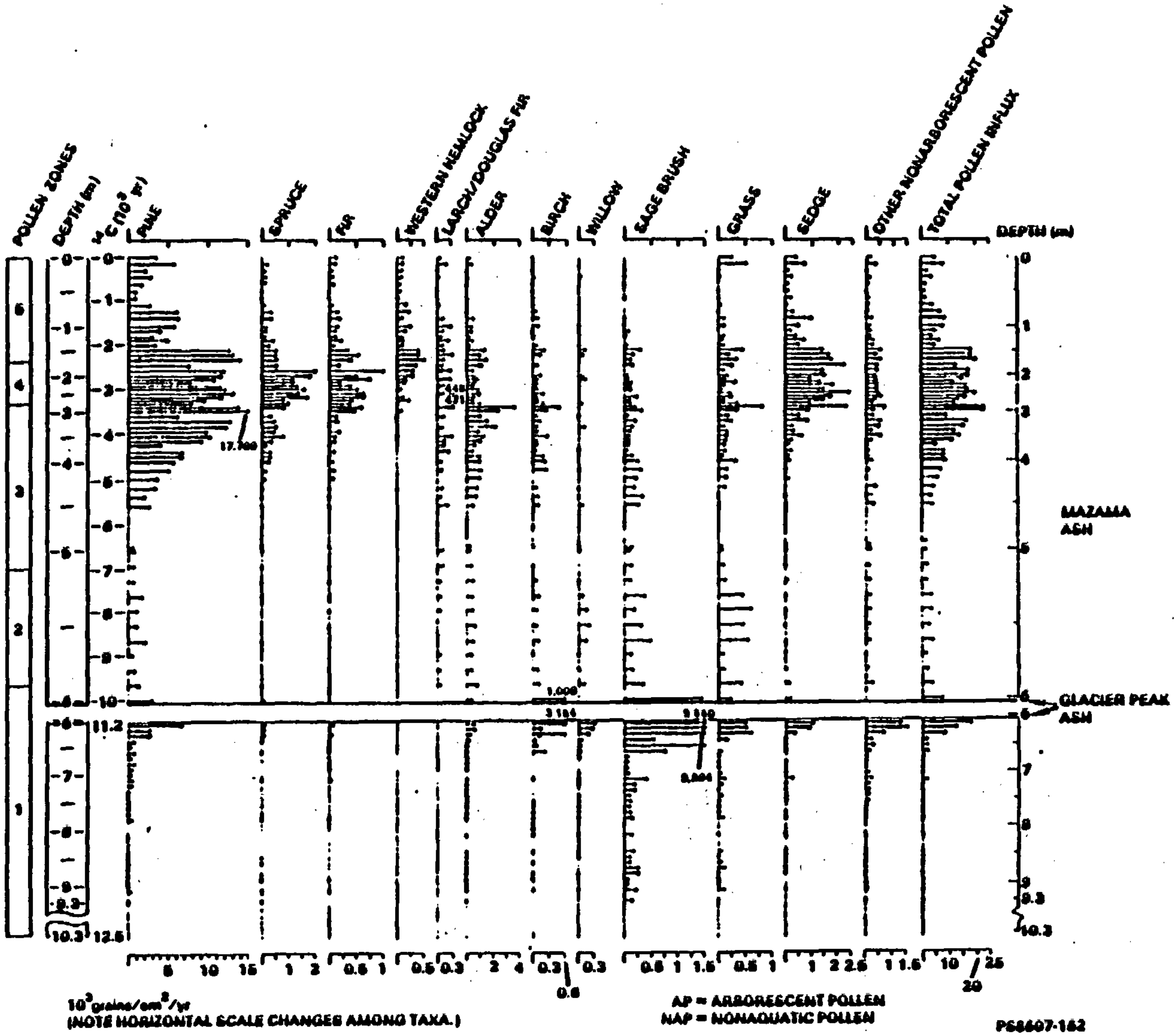


Figure 5.2-9. Pollen influx diagram from Big Meadow Fen, Washington (Mack et al., 1978a). PS8607-162

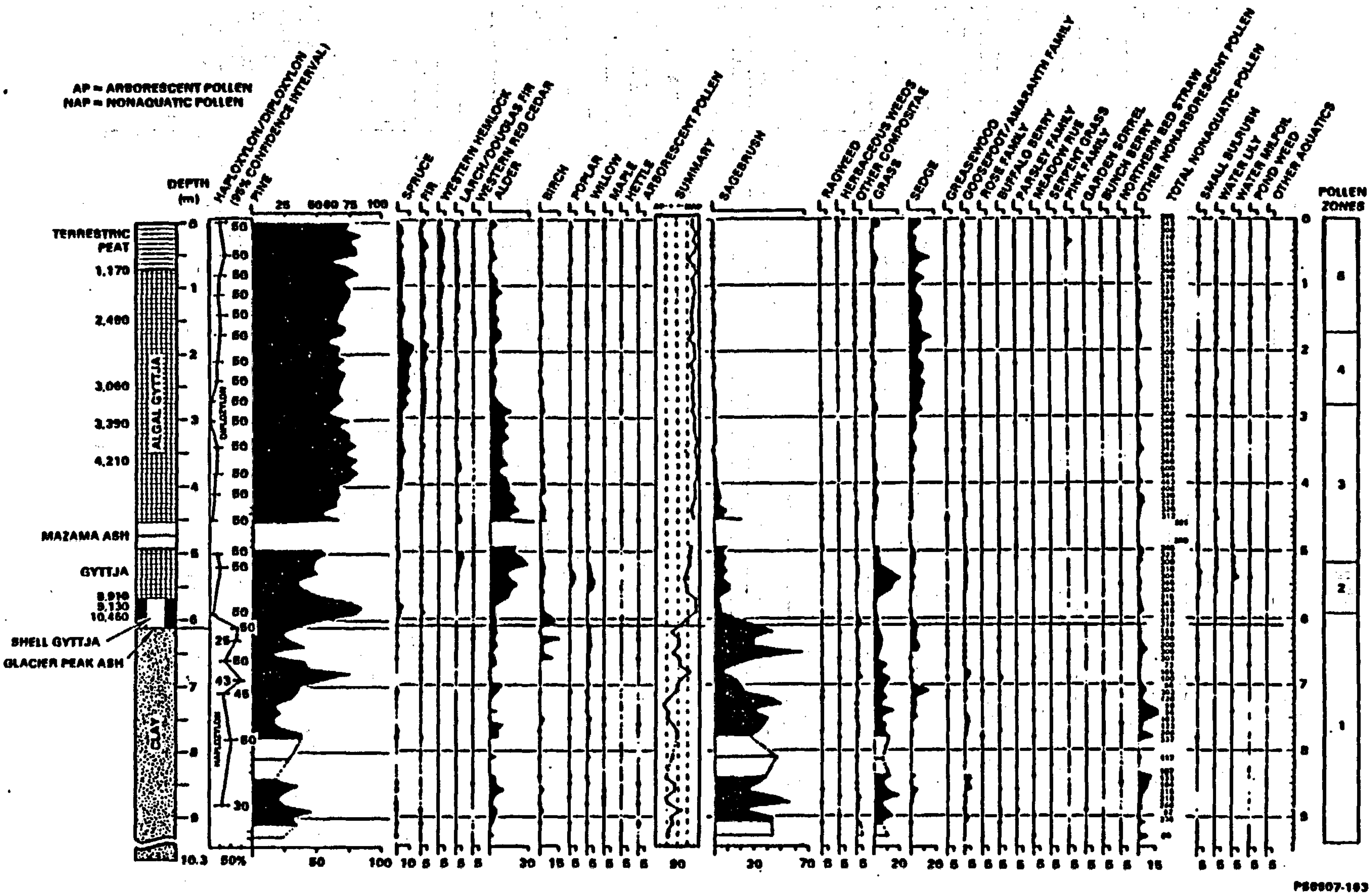


Figure 5.2-10. Pollen percentage diagram from Big Meadow Fen (Mack et al., 1978a). Shading is added to improve clarity of the figure. PS8607-163

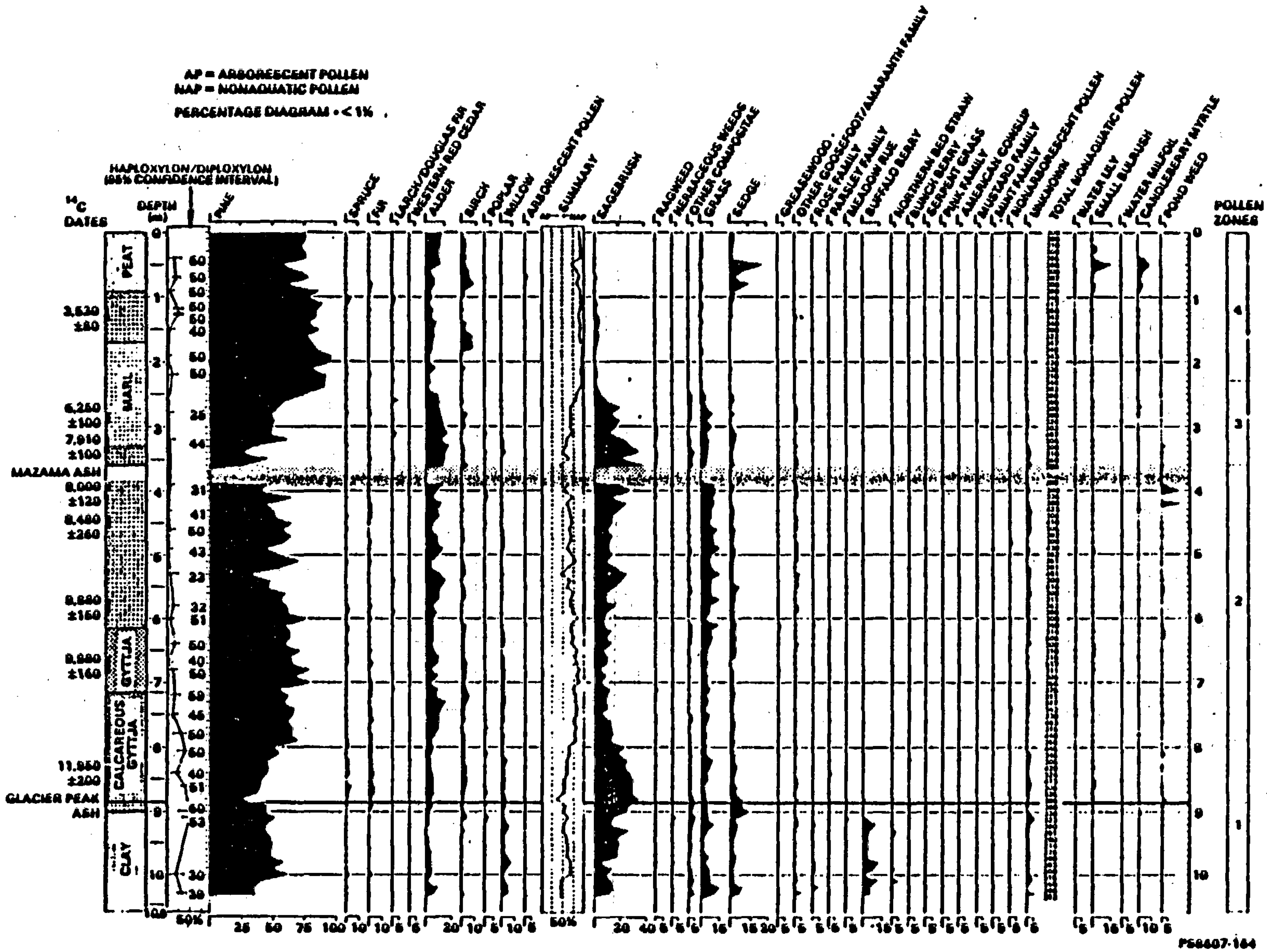


Figure 5.2-11. Pollen percentage diagram from Waits Lake. Radiocarbon dates bracket the Mazama tephra at 6,900 years before present (Mack et al., 1978d). PS8607-164

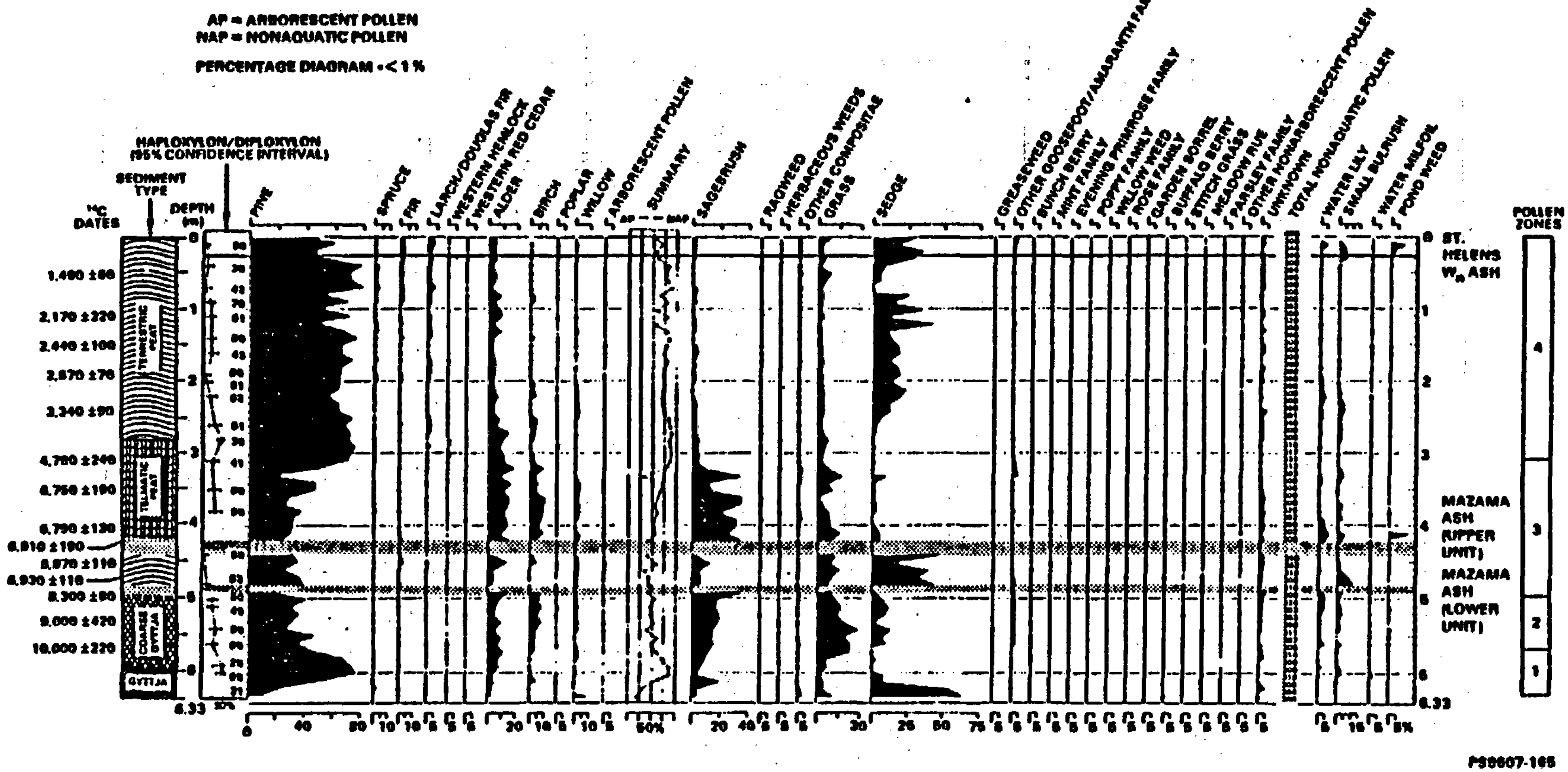


Figure 5.2-12. Pollen percentage diagram from Bonaparte Meadows (Mack et al., 1979). PS8607-165

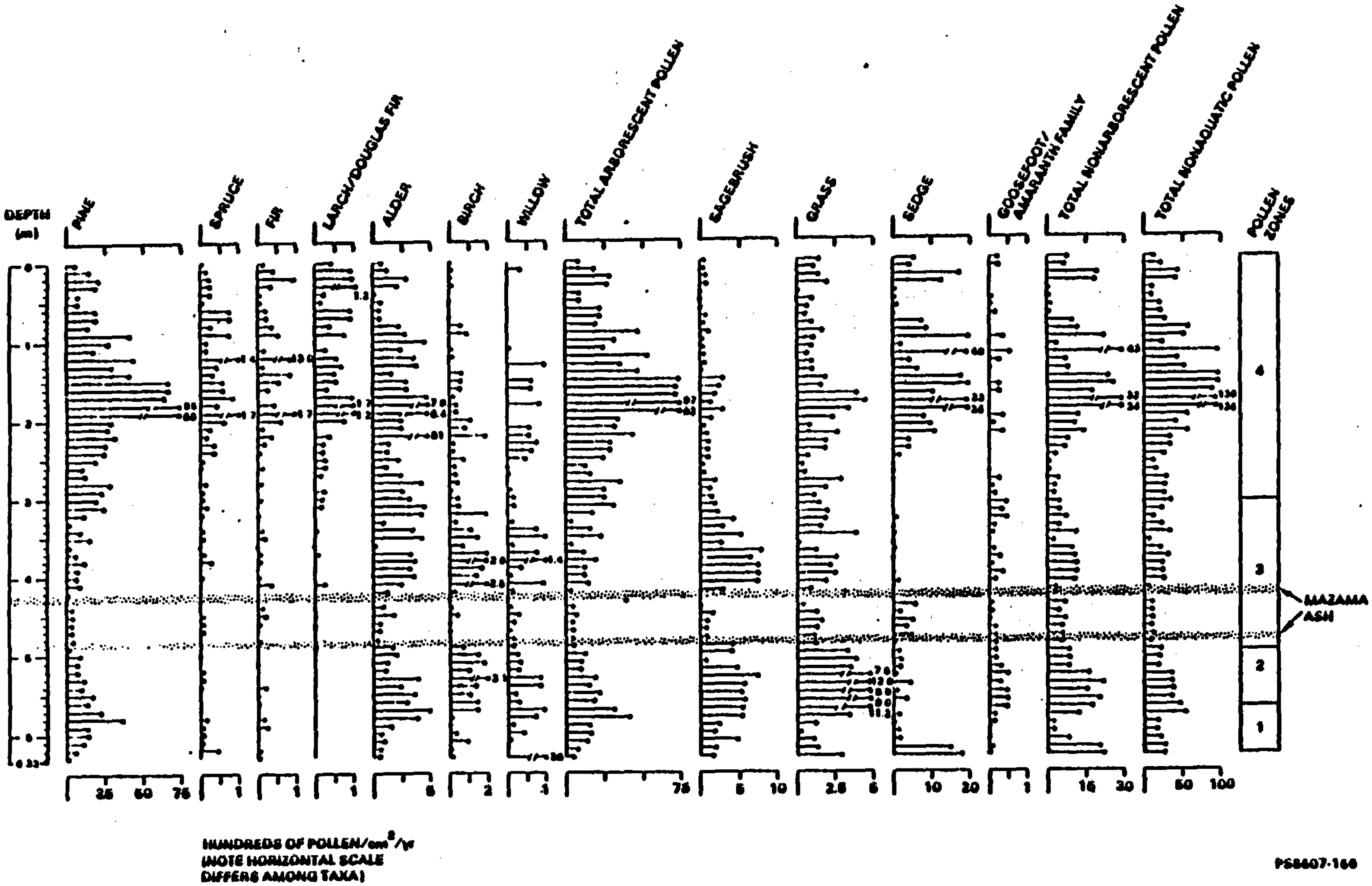


Figure 5.2-13. Pollen influx diagram from Bonaparte Meadows (Mack et al., 1979). PS8607-166

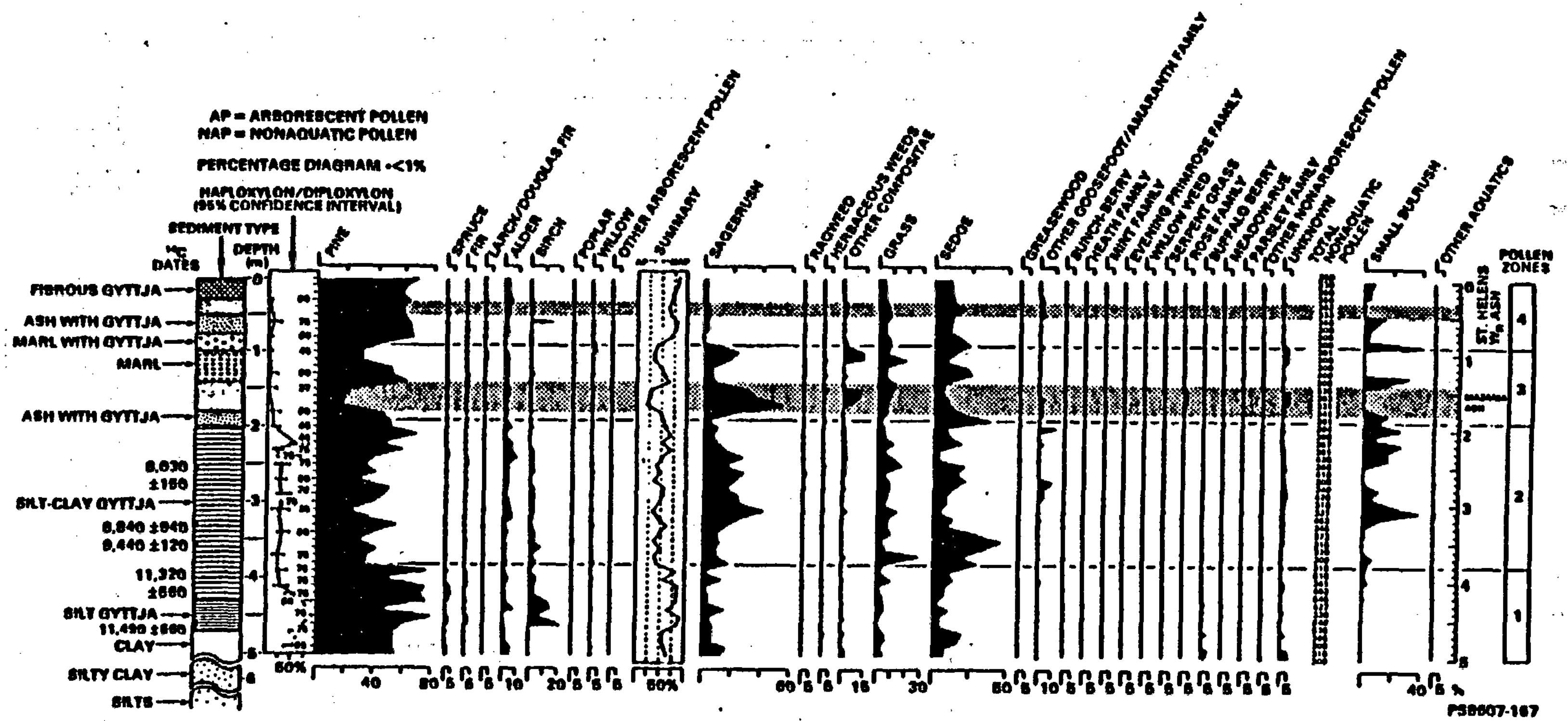


Figure 5.2-14. Pollen percentage diagram from Mud Lake (Mack et al., 1979). PS8607-167

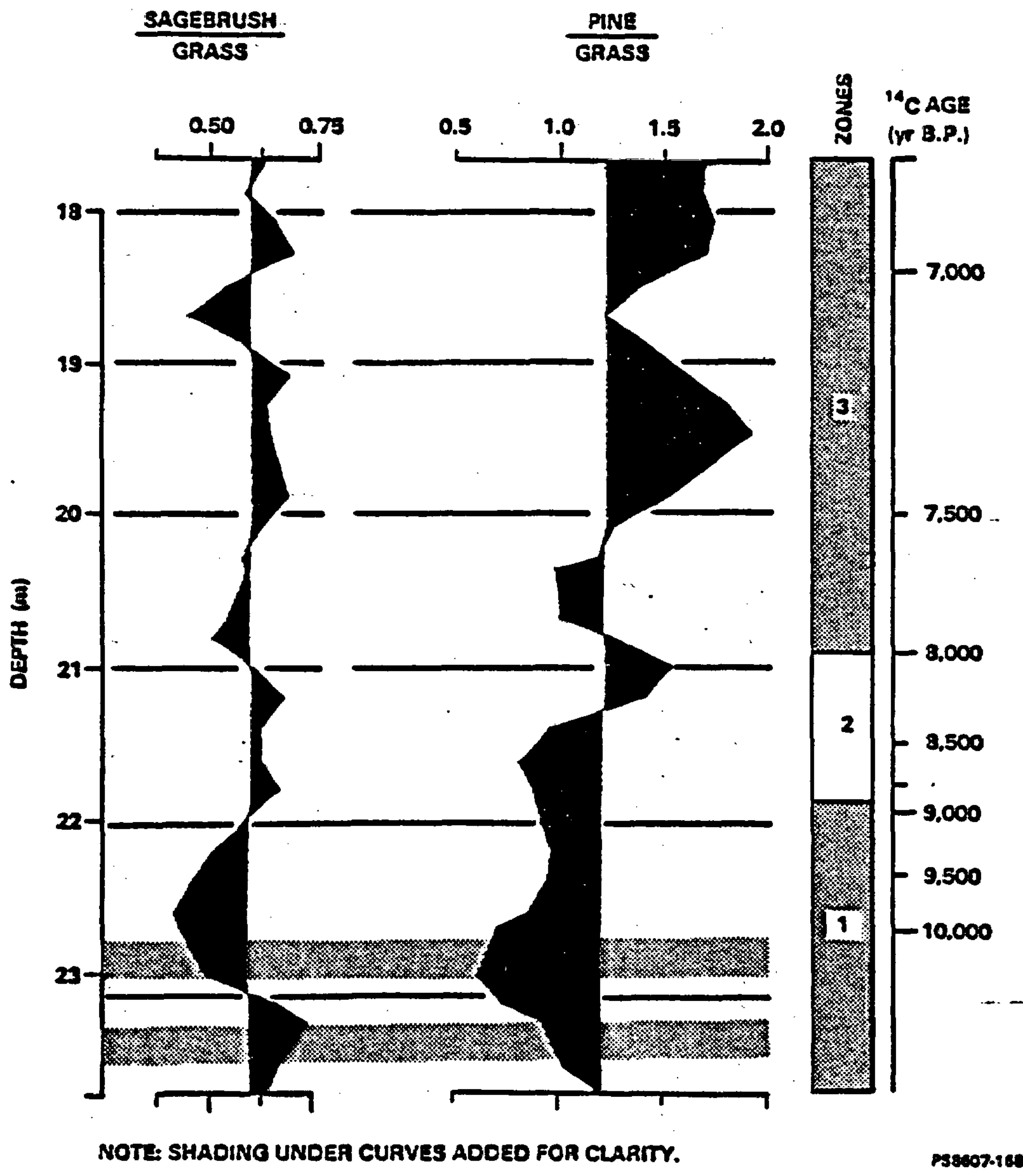


Figure 5.2-15. Sagebrush/grass and pine/grass ratios for Wildcat Lake pollen zones below Mazama ash (Mehringer, 1985b, Fig. 12, p. 25). PS8607-168

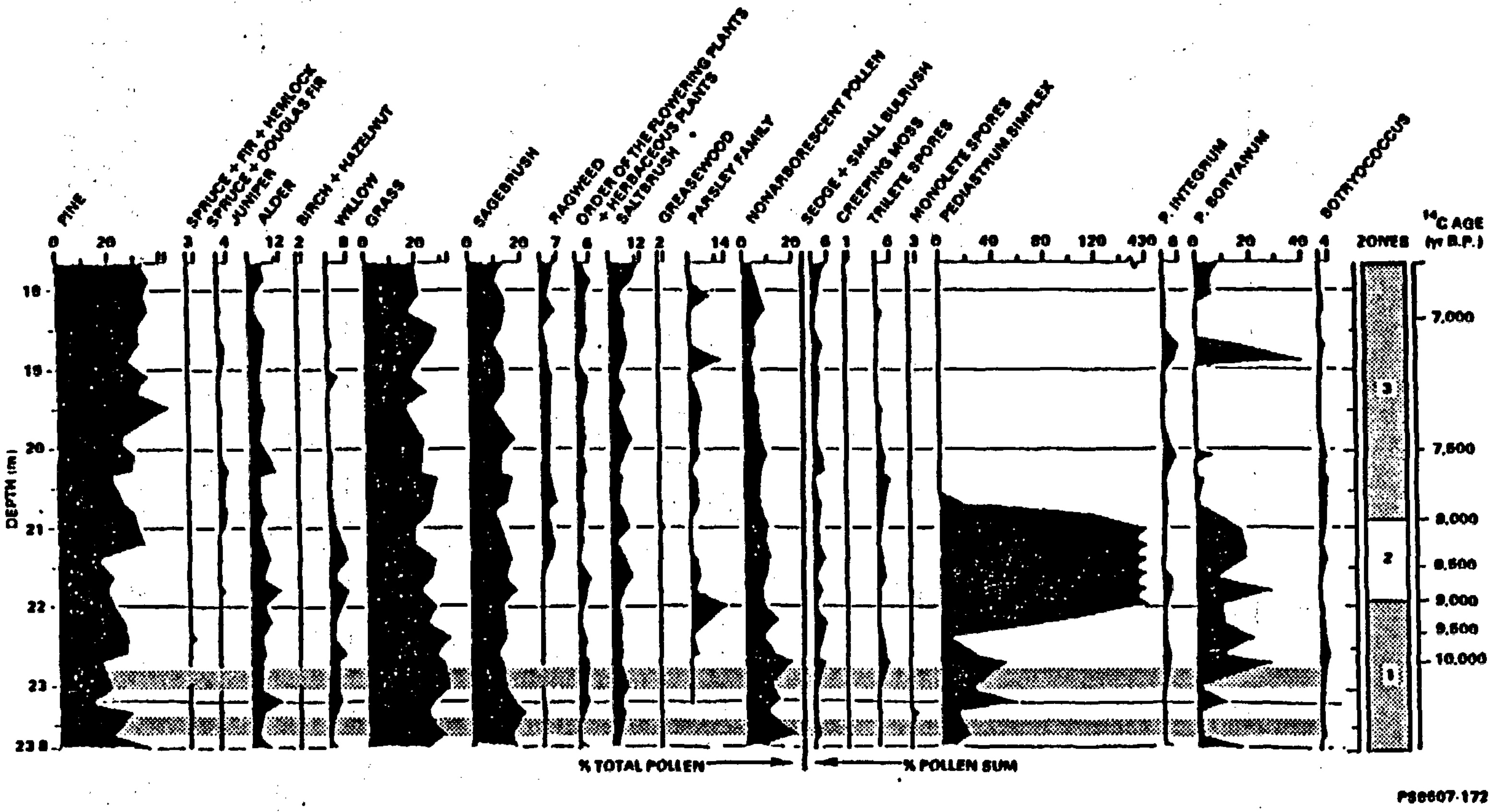


Figure 5.2-16. Pollen percentage diagram for Wildcat Lake pollen zones below Mazama ash (Mehringer, 1985b, Fig. 10, p. 22). PS8607-172

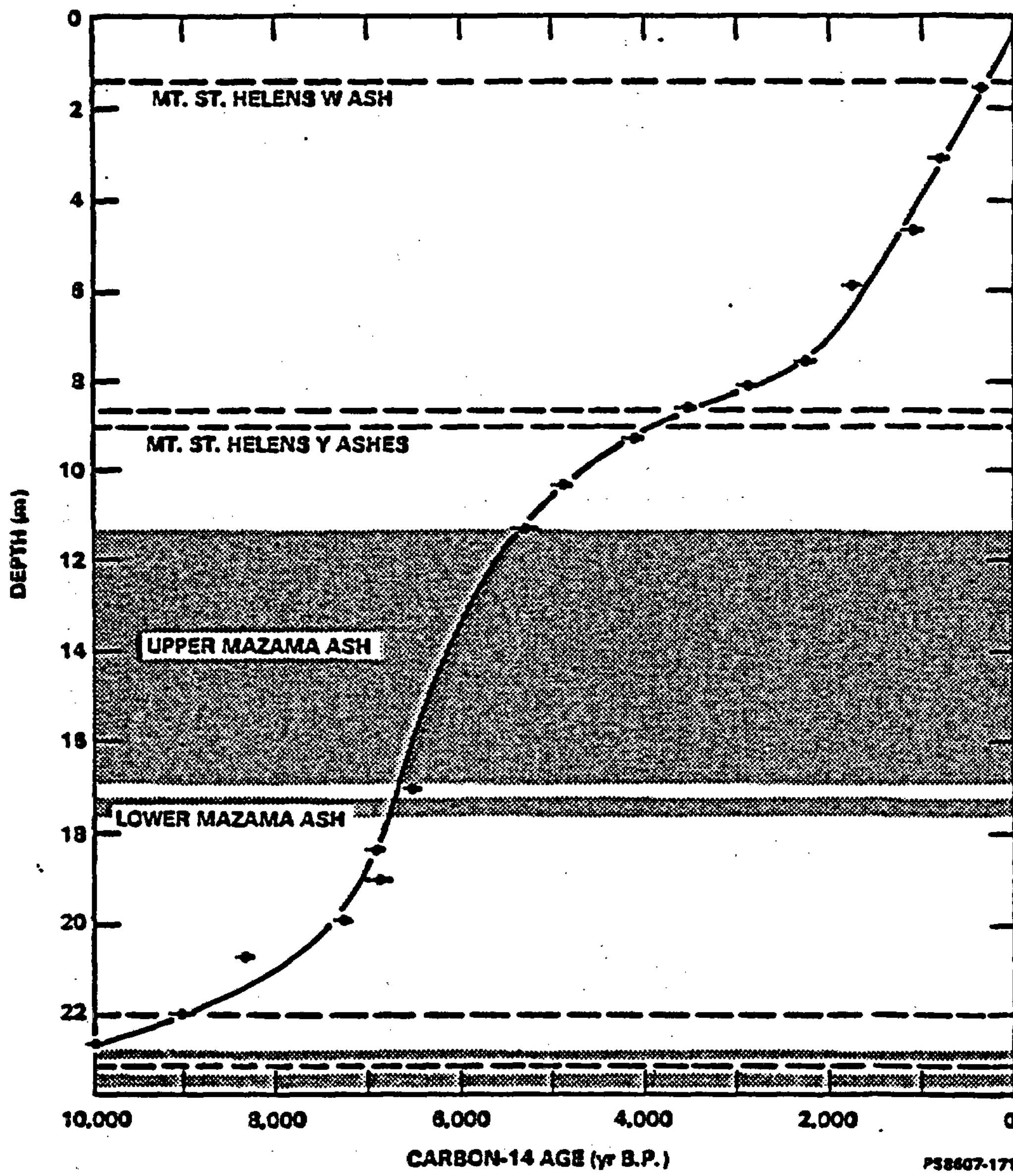


Figure 5.2-17. Radiocarbon dates and sediment accumulation curve for sediments from Wildcat Lake (Mehringer, 1985b, Fig. 2, p. 9). Lines through points represent the error in the carbon-14 dates. PS8607-171

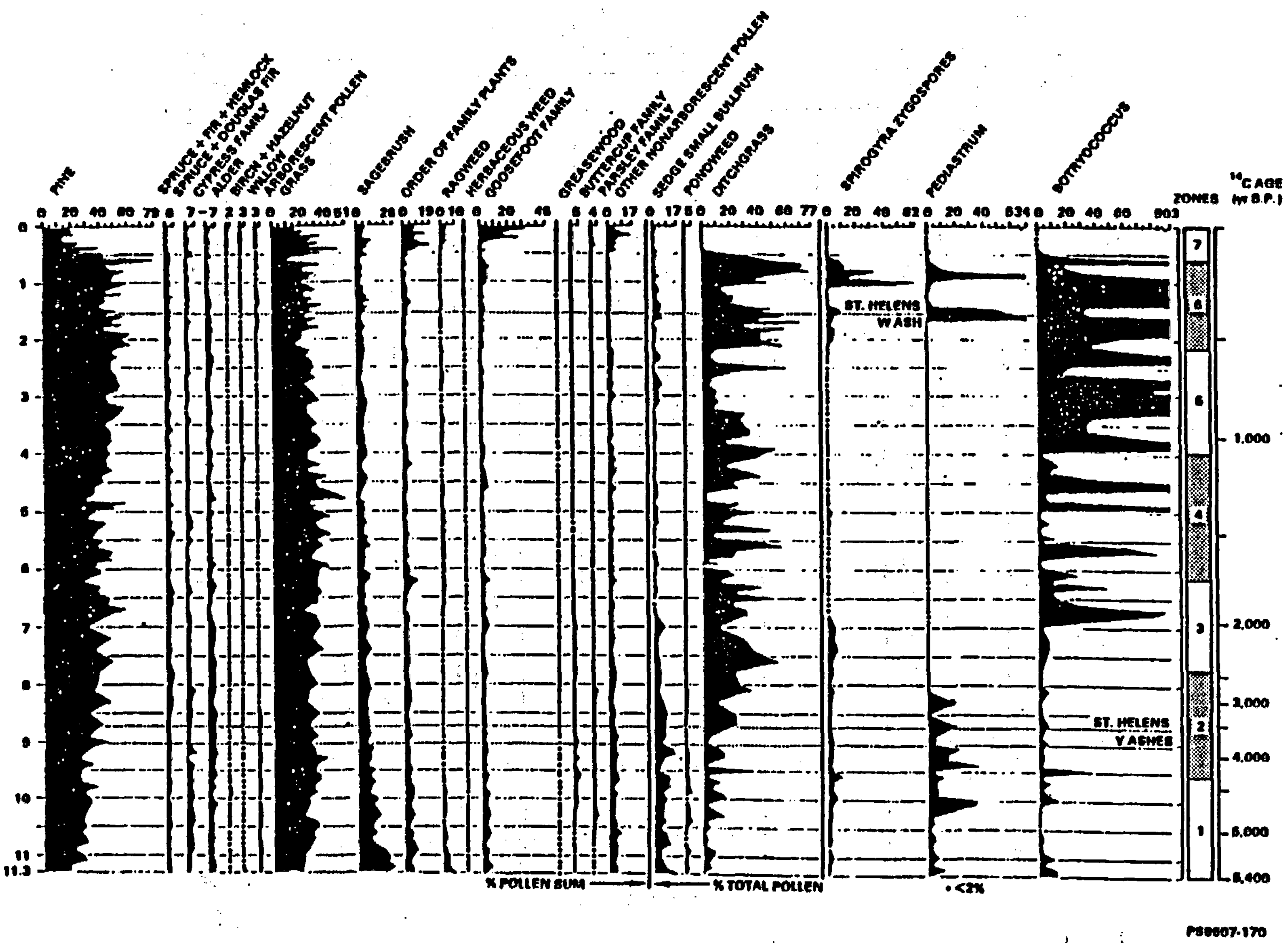


Figure 5.2-18. Pollen percentage diagram for Wildcat Lake pollen zones above Mazama ash (Mehringer, 1985b, Fig. 9, p. 21). PS8607-170

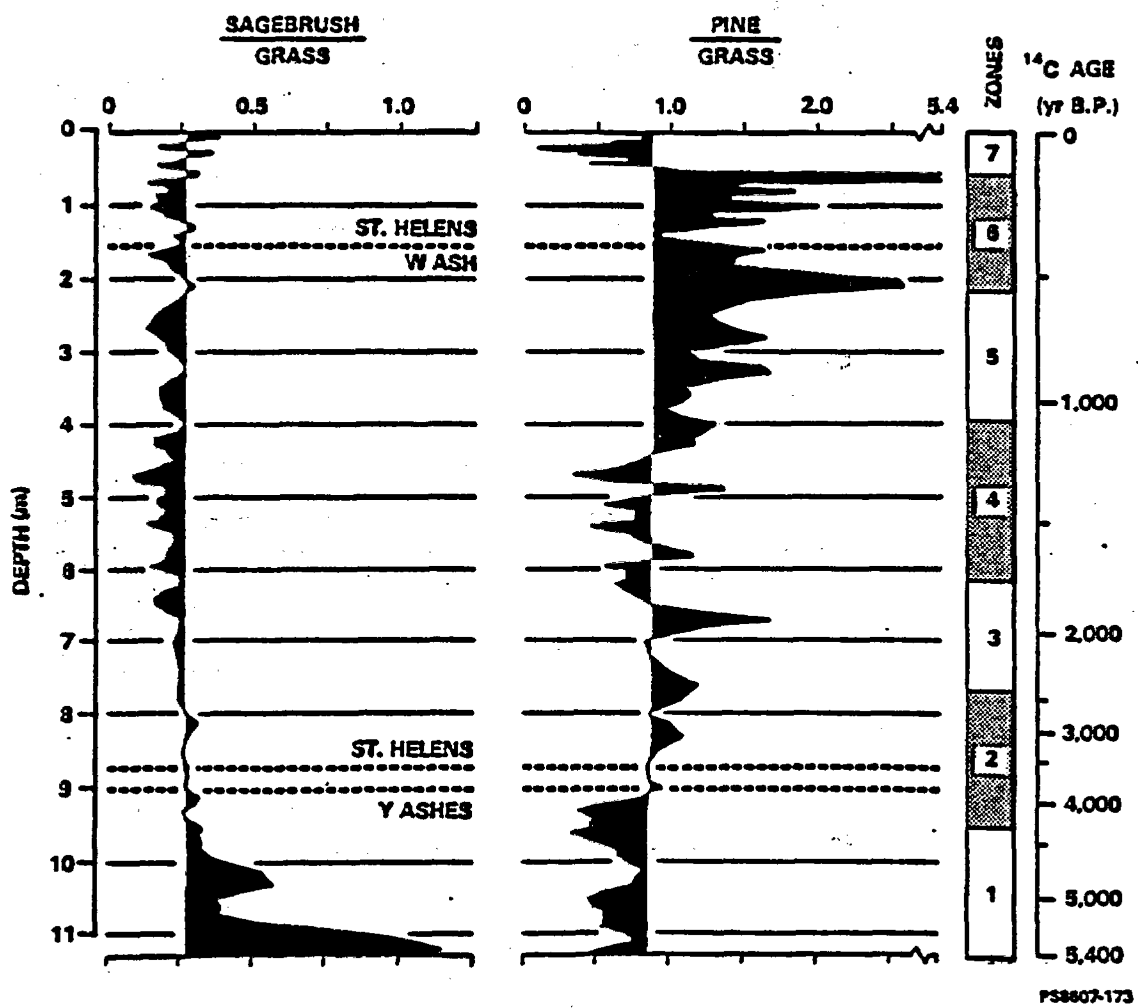


Figure 5.2-19. Sagebrush/grass and pine/grass pollen ratios for Wildcat Lake pollen zones above Mazama ash (Mehringer, 1985b, Fig. 13, p. 27). PS8607-173

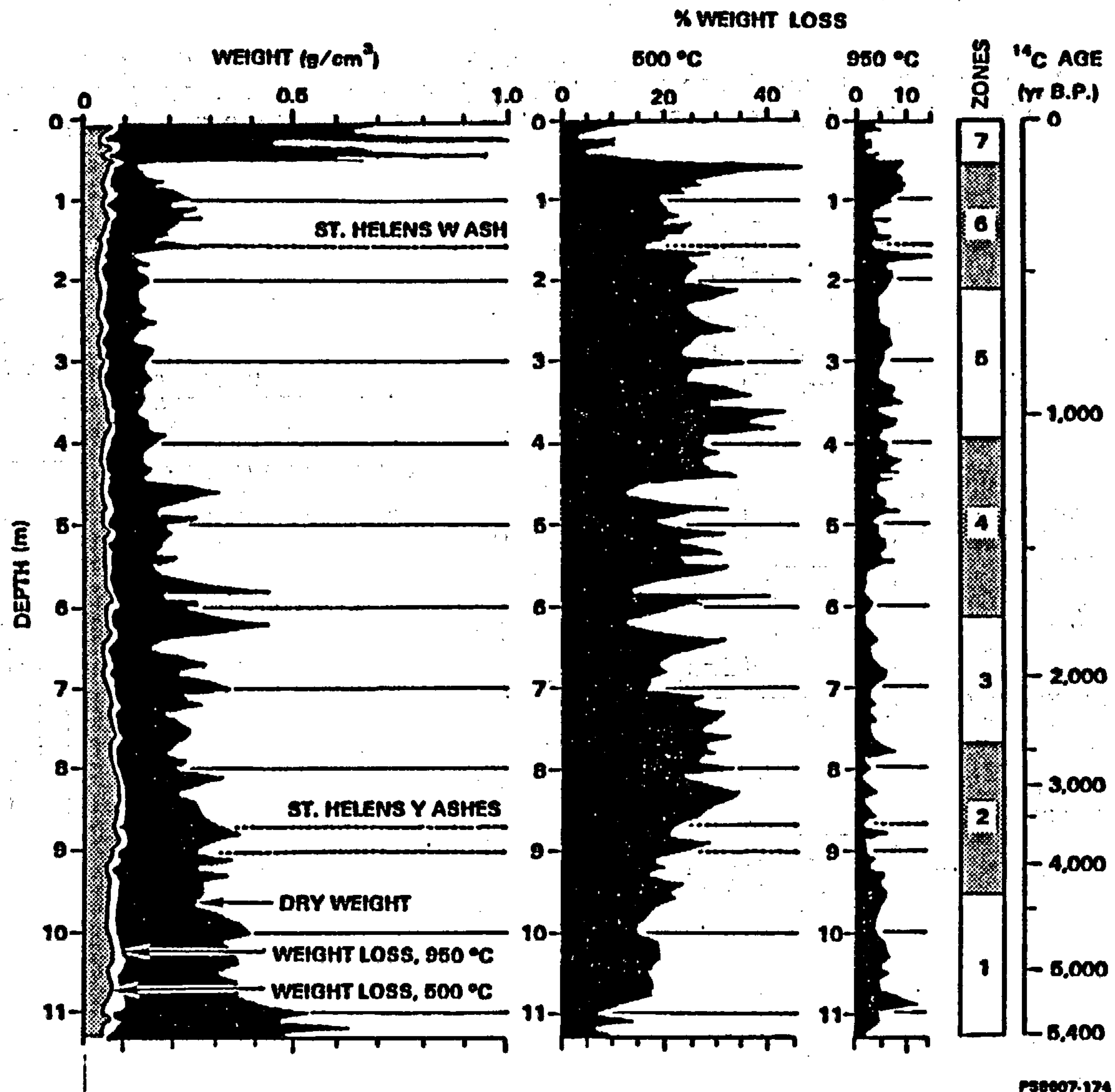


Figure 5.2-20. Dry weight and percent weight loss for Wildcat Lake sediments above Mazama ash (Mehringer, 1985b, Fig. 4, p. 12). PS8607-174

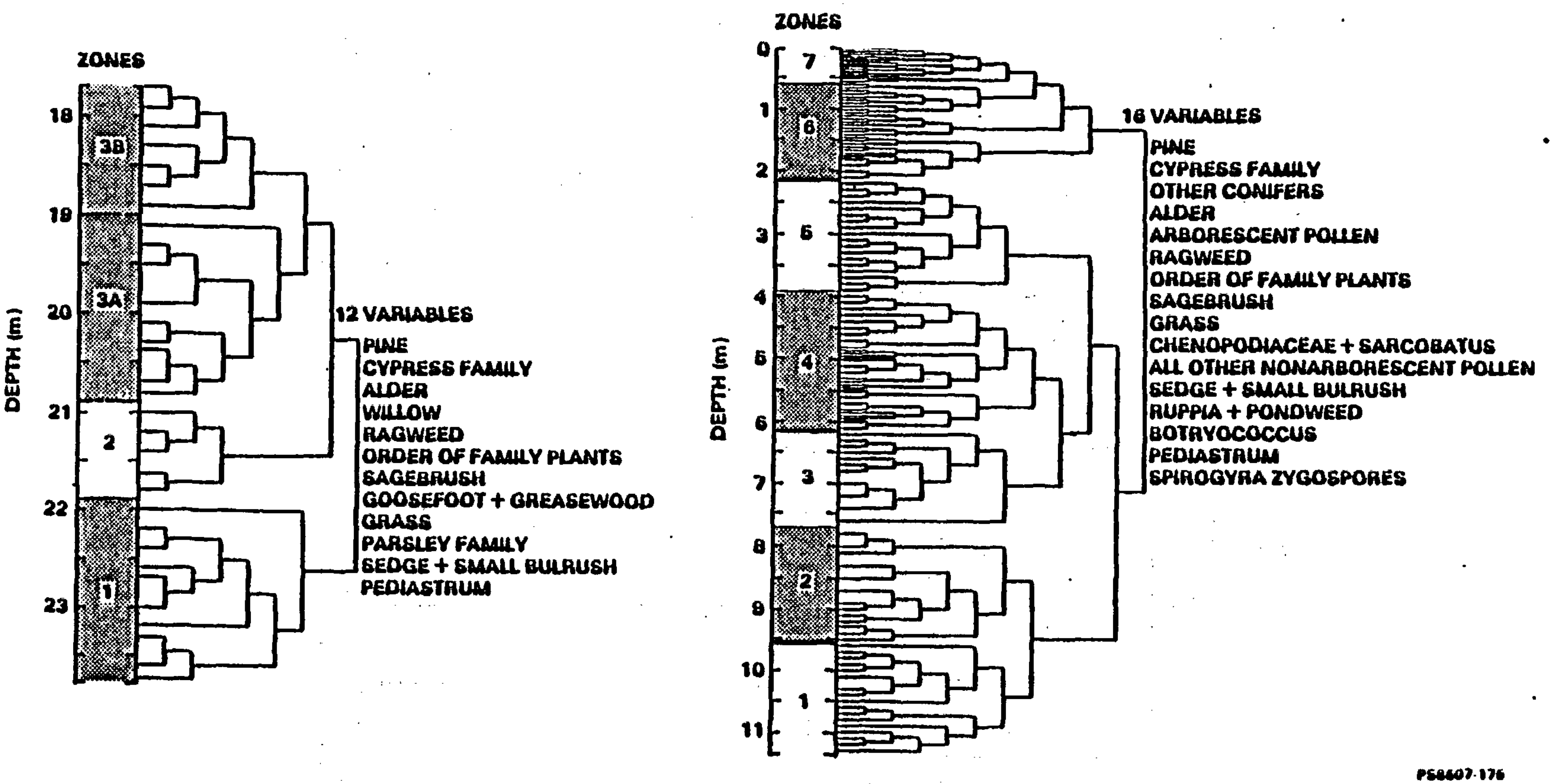


Figure 5.2-21. Cluster analysis and discrimination of Wildcat Lake pollen zones below Mazama ash (left), and above Mazama ash (right) (Mehringer, 1985b, Fig. 6 and 7, pp. 17 and 18). PS8607-175

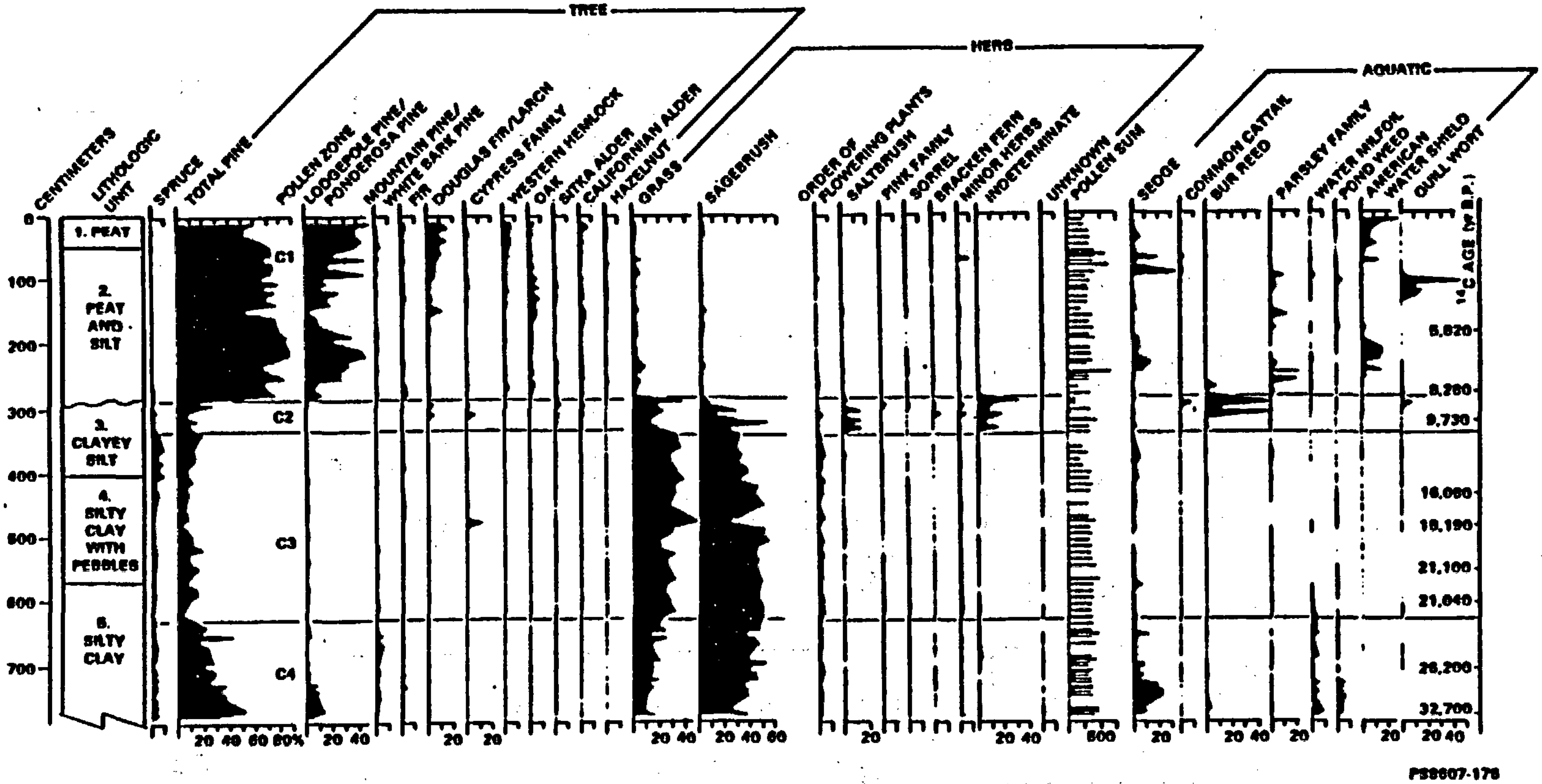


Figure 5.2-22. Pollen percentage diagram for Carp Lake (Barnosky, 1983; Barnosky, 1985). PS8607-176

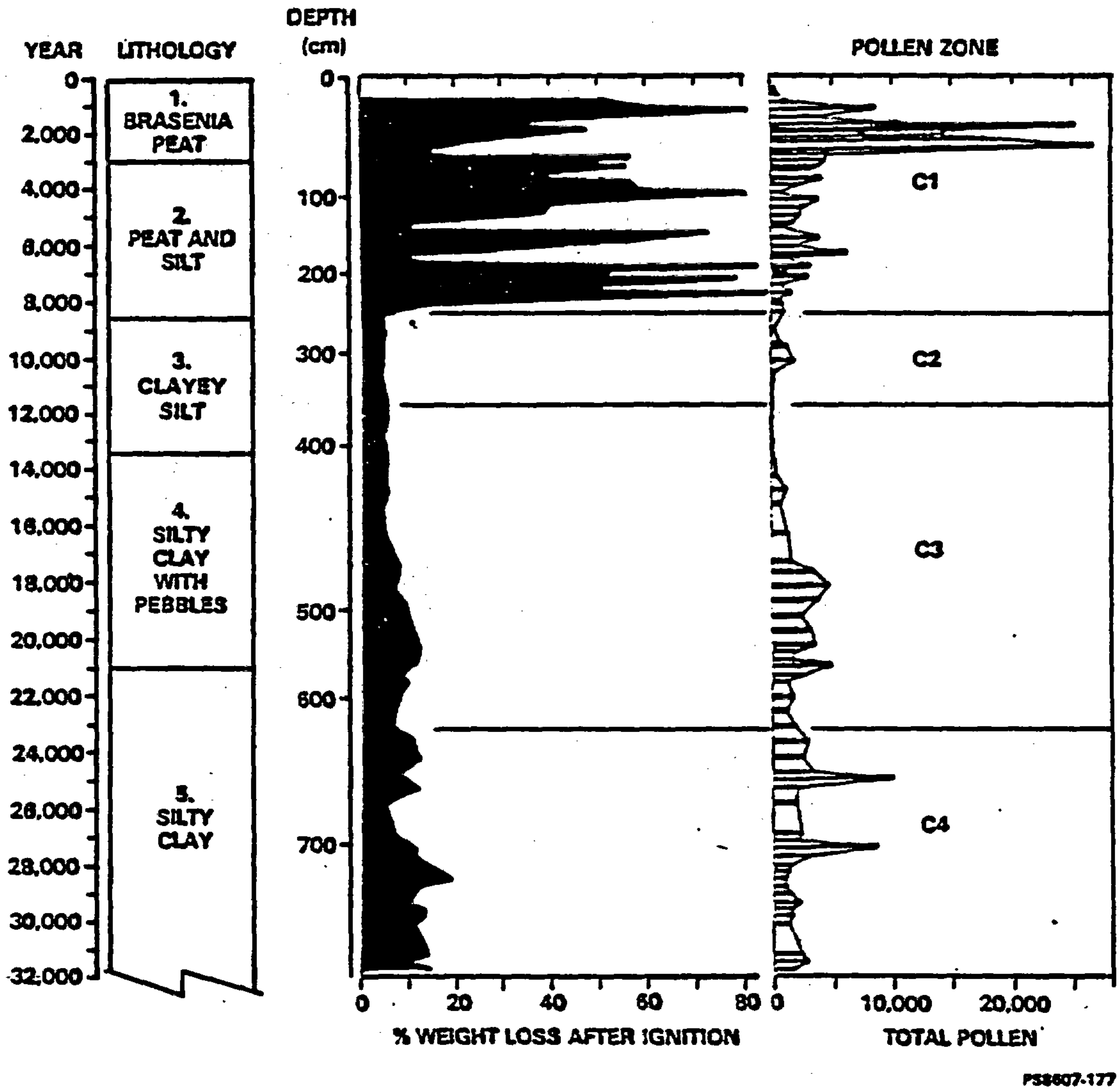


Figure 5.2-23. Percent weight loss and pollen concentration for samples from Carp Lake (Barnosky, 1983; Barnosky, 1985). PS8607-177

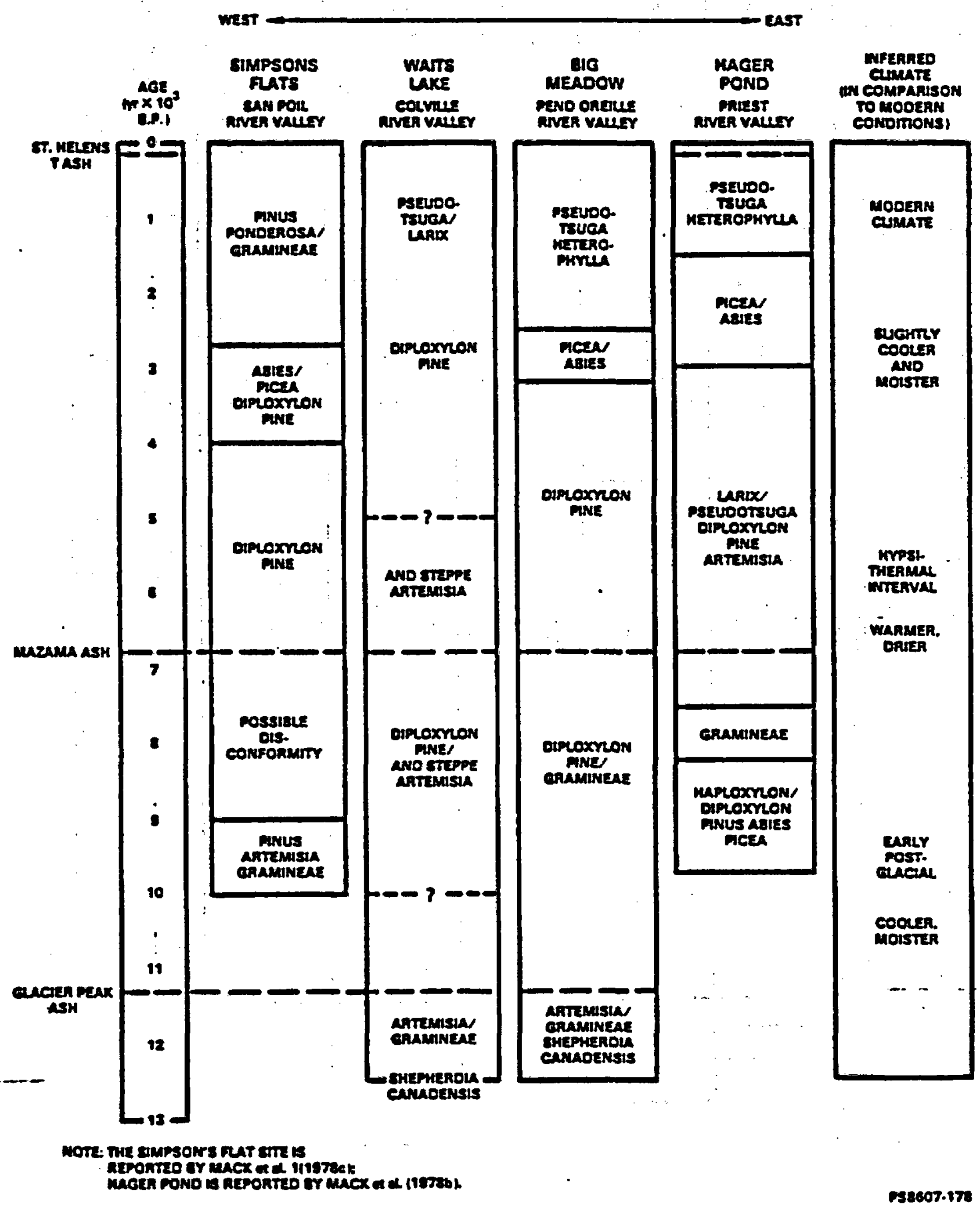


Figure 5.2-24. Correlation and inferred timing of vegetation and climatic change at sites from northeastern Washington and adjacent Idaho (Mehringer, 1985a, Fig. 7, p. 177). PS8607-178

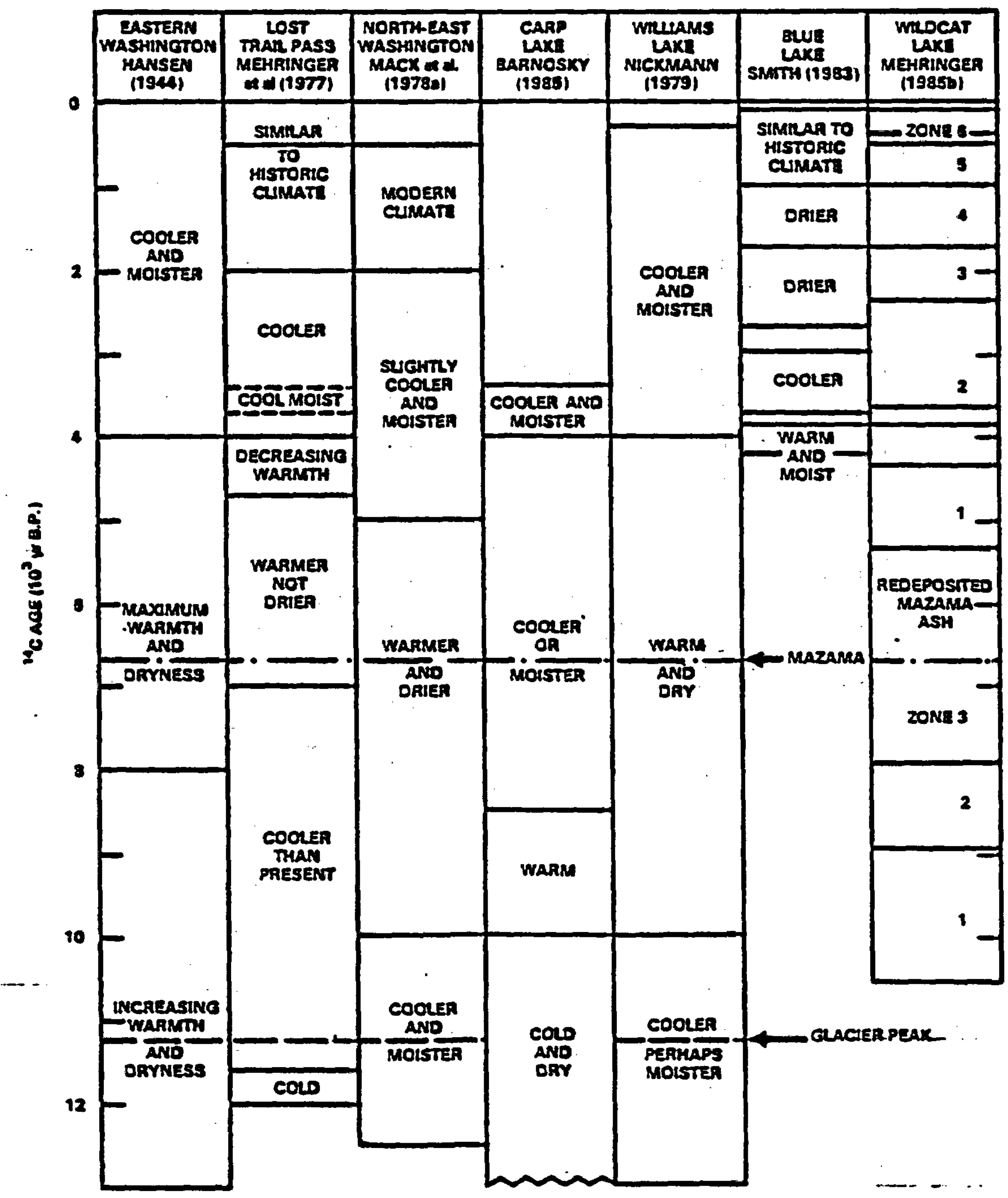


Figure 5.2-25. Correlation and inferred vegetation and climatic change at sites from eastern Washington, adjacent Idaho, and Montana (Mehringer, 1985b, Fig. 17, p. 37). PS8607-179

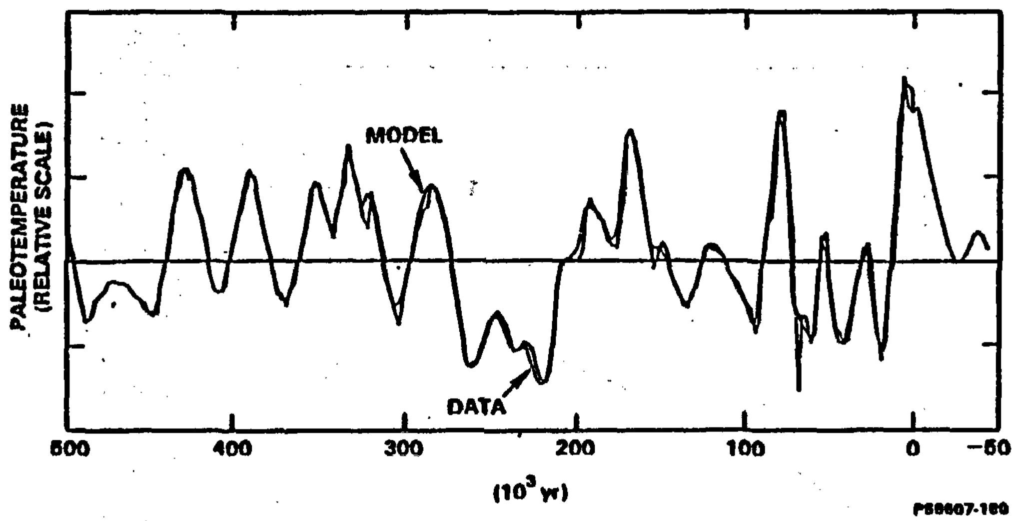


Figure 5.2-26. Paleoclimate record and model estimates from Lake Biwa, Japan, for the past 500,000 years (Kanari et al., 1984, Fig. 2, p. 408). PS8607-180

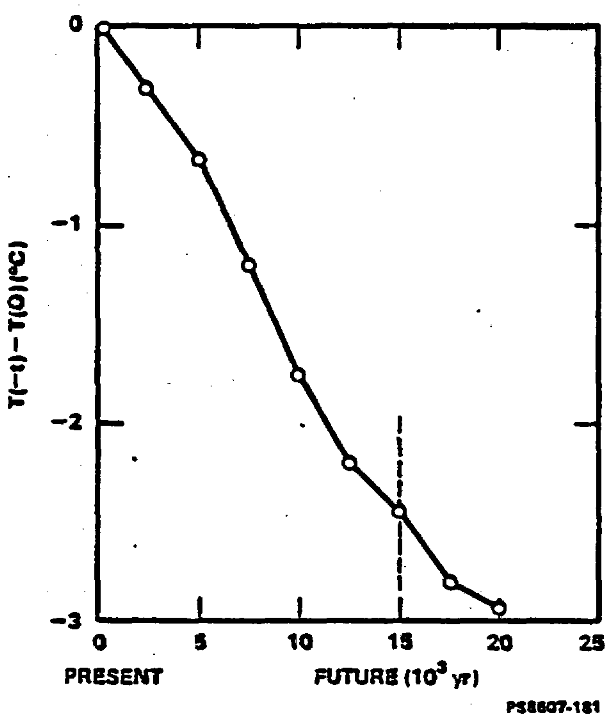


Figure 5.2-27. Relative temperature decline forecast illustrated in Figure 5.2-26. Uncertainty becomes substantial after 15,000 year (dashed line) (Kanari et al., 1984, Fig. 4, p. 412). PS8607-181

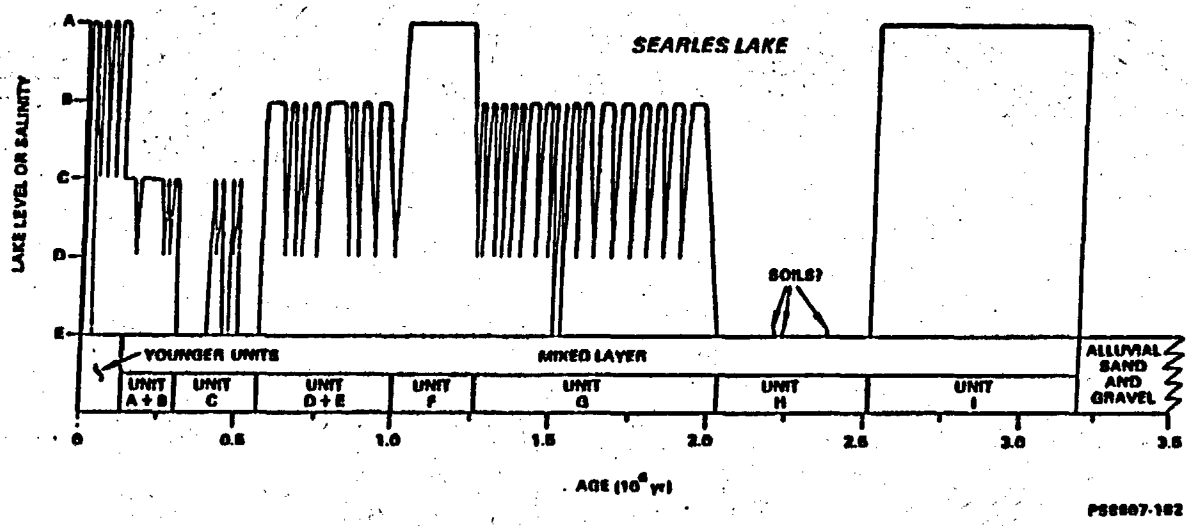


Figure 5.2-28. Reconstructed history of Searles Lake, based primarily on core KM-3. Brief possible fluctuations during deposition of Units F, H, and I are not plotted (Smith et al., 1983, Fig. 6, p. 21). PS8607-182

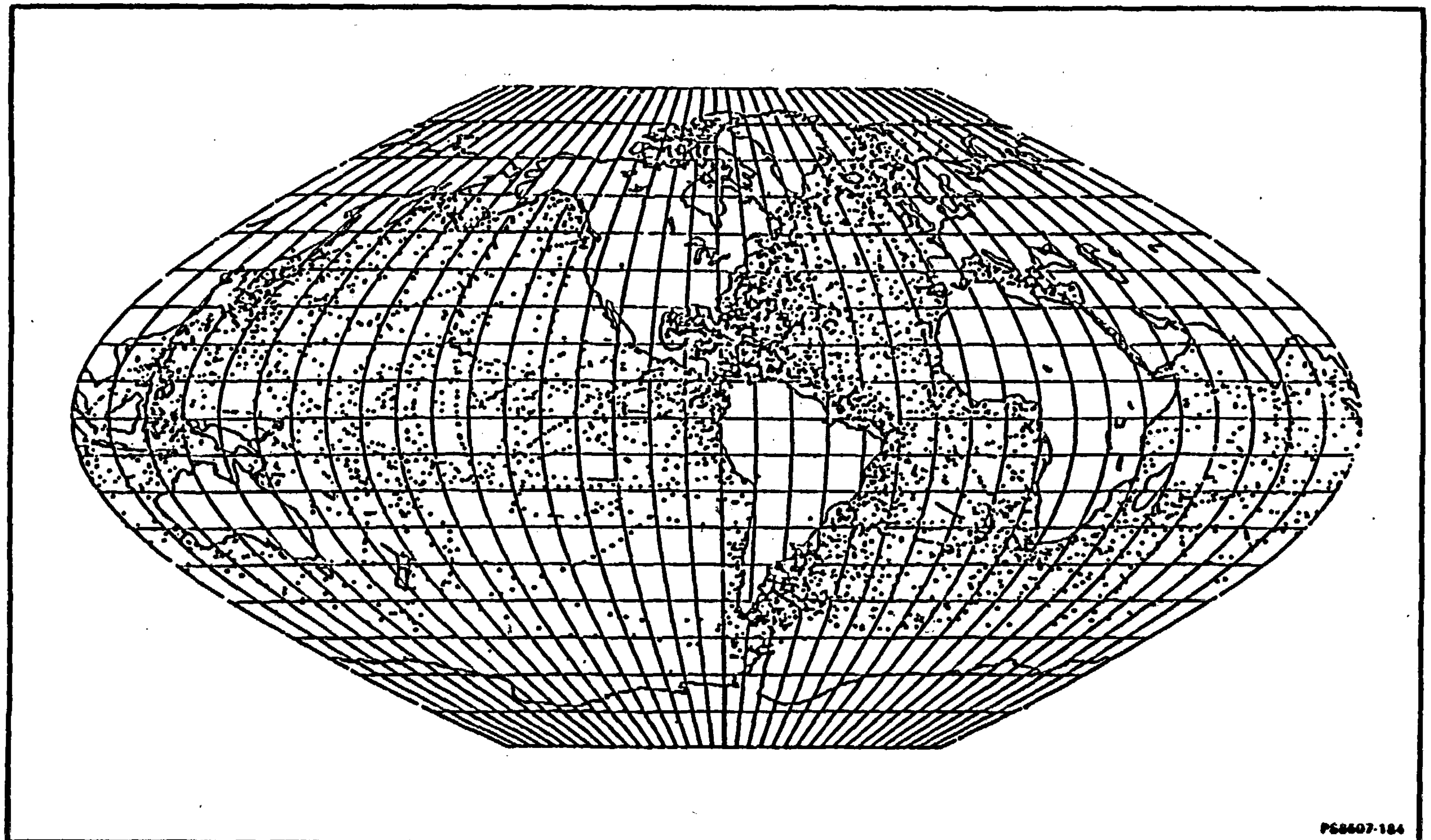


Figure 5.2-29. Distribution of sites worldwide at which cores of deep-sea sediments have been obtained. Over 16,000 cores are now available (Ruddiman, 1985, Fig. 1, p. 199). PS8607-184

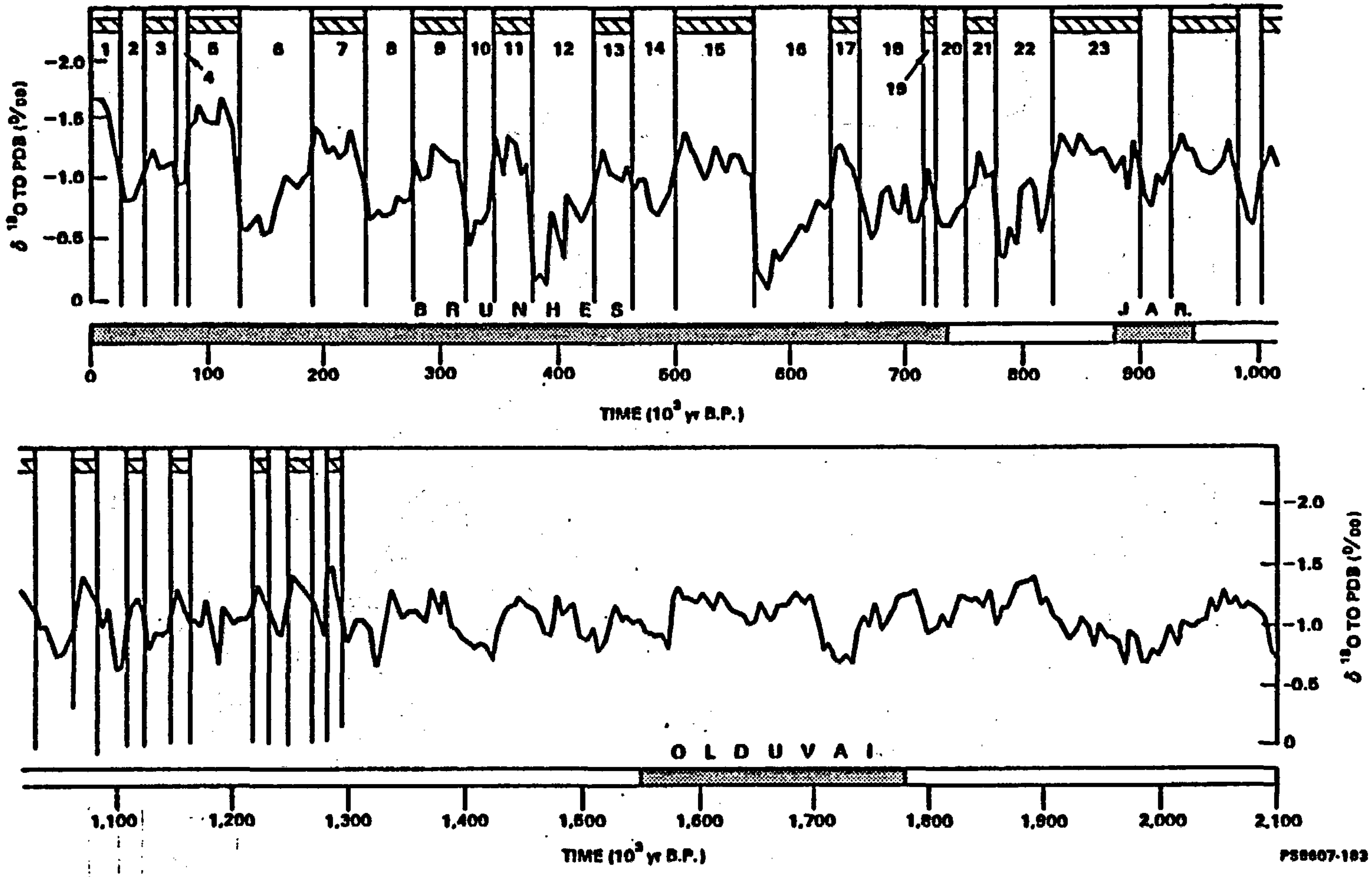


Figure 5.2-30. Oxygen isotope record obtained from core V28-239 in the equatorial Pacific Ocean. Numbered oxygen isotopic stages are shown as are the major polarity episodes of the paleomagnetic record (Shackleton and Opdyke, 1976, Fig. 1, pp. 450-451). PS8607-183

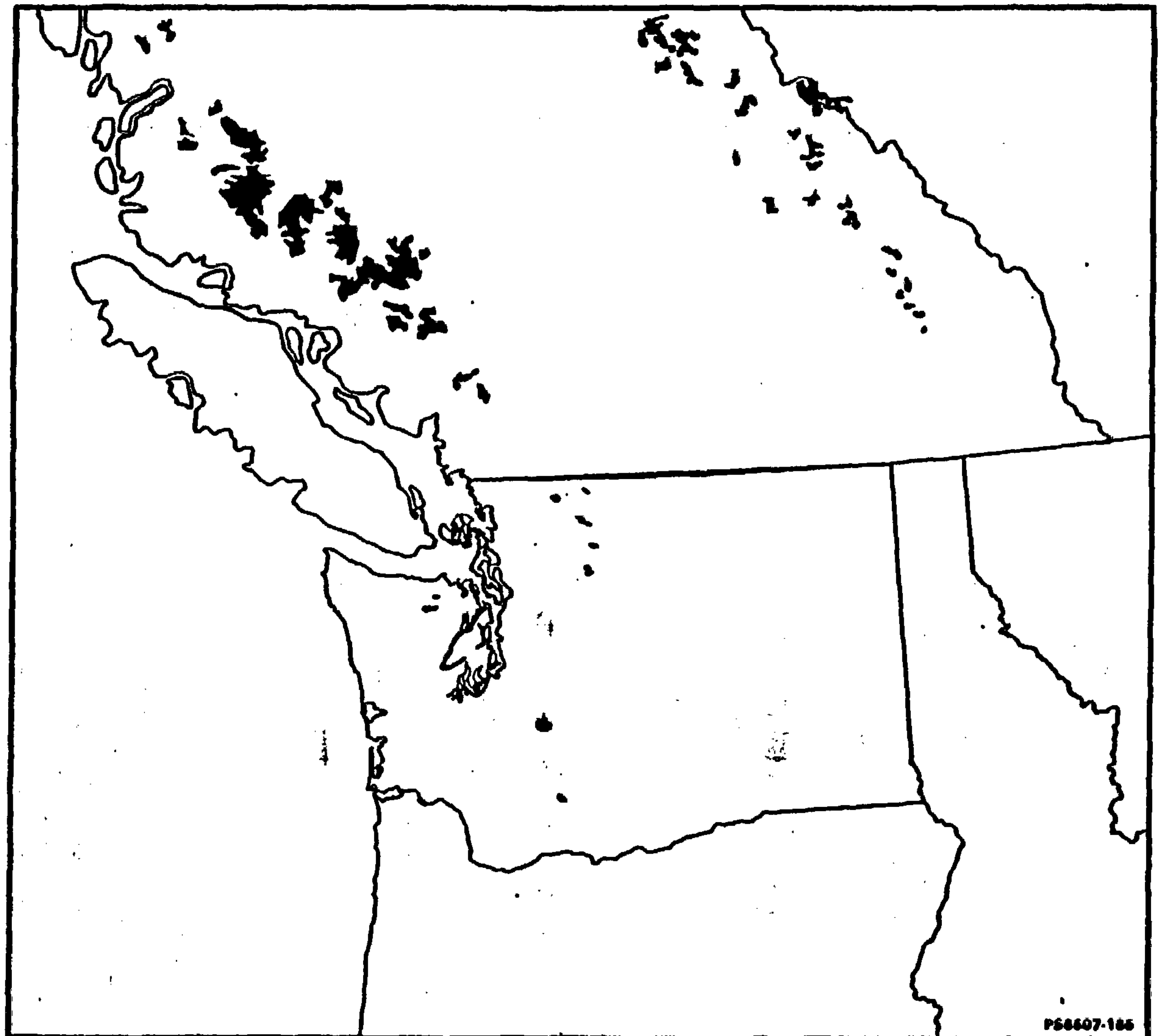


Figure 5.2-31. Locations of modern glaciers in the Pacific Northwest (USGS, 1980, p. 5).

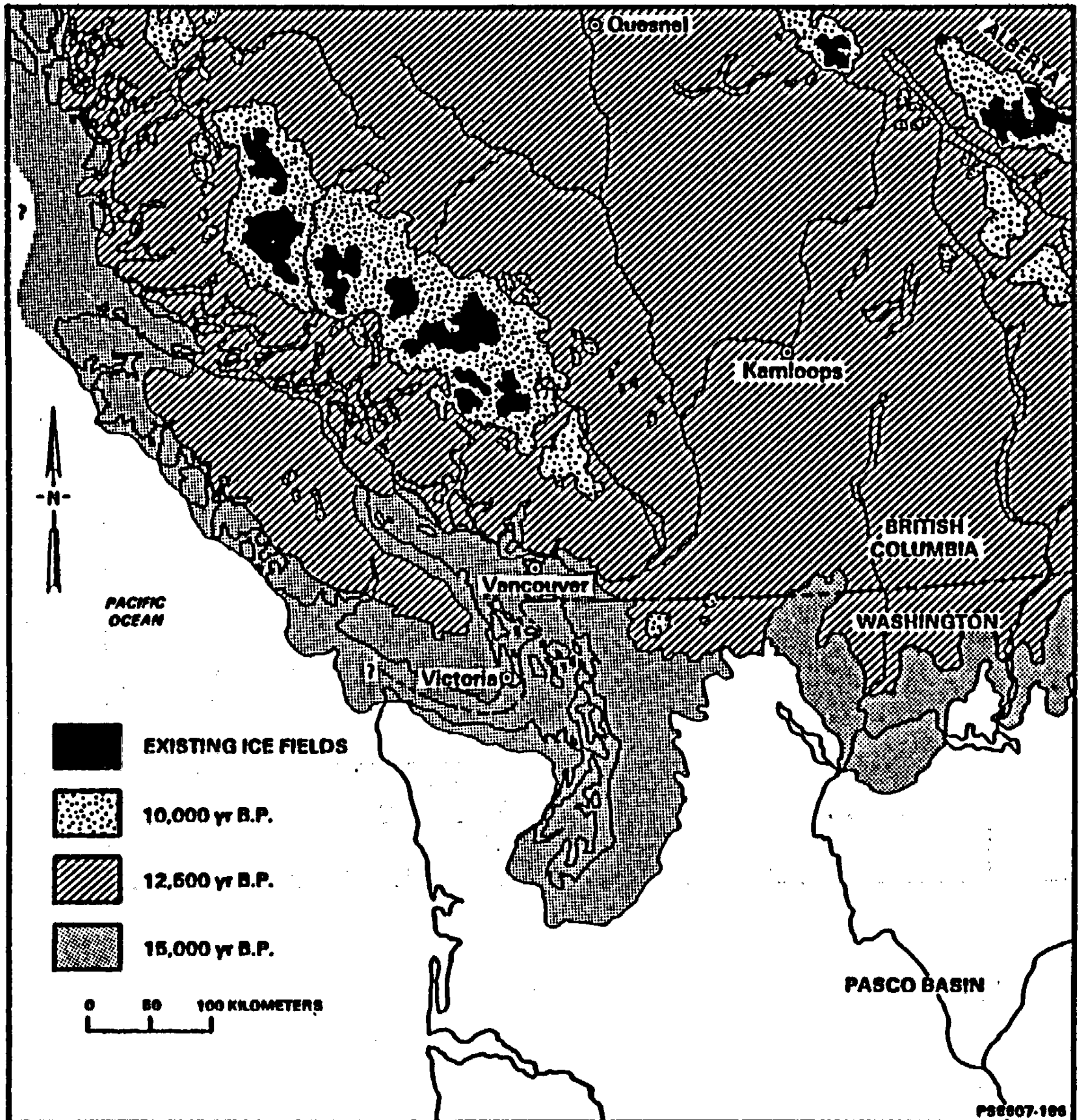


Figure 5.2-32. Stages in the retreat of the Cordilleran ice sheet from 15,000 to 10,000 years before present (Clague, 1981, Fig. 4, p. 14). PS8607-186

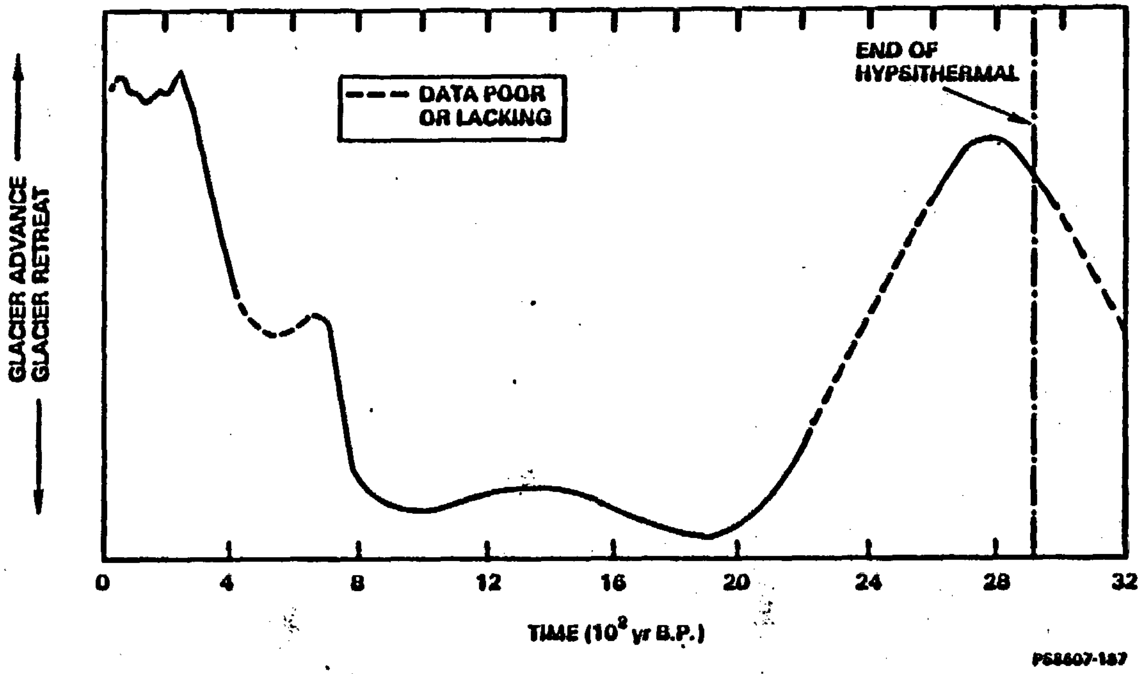
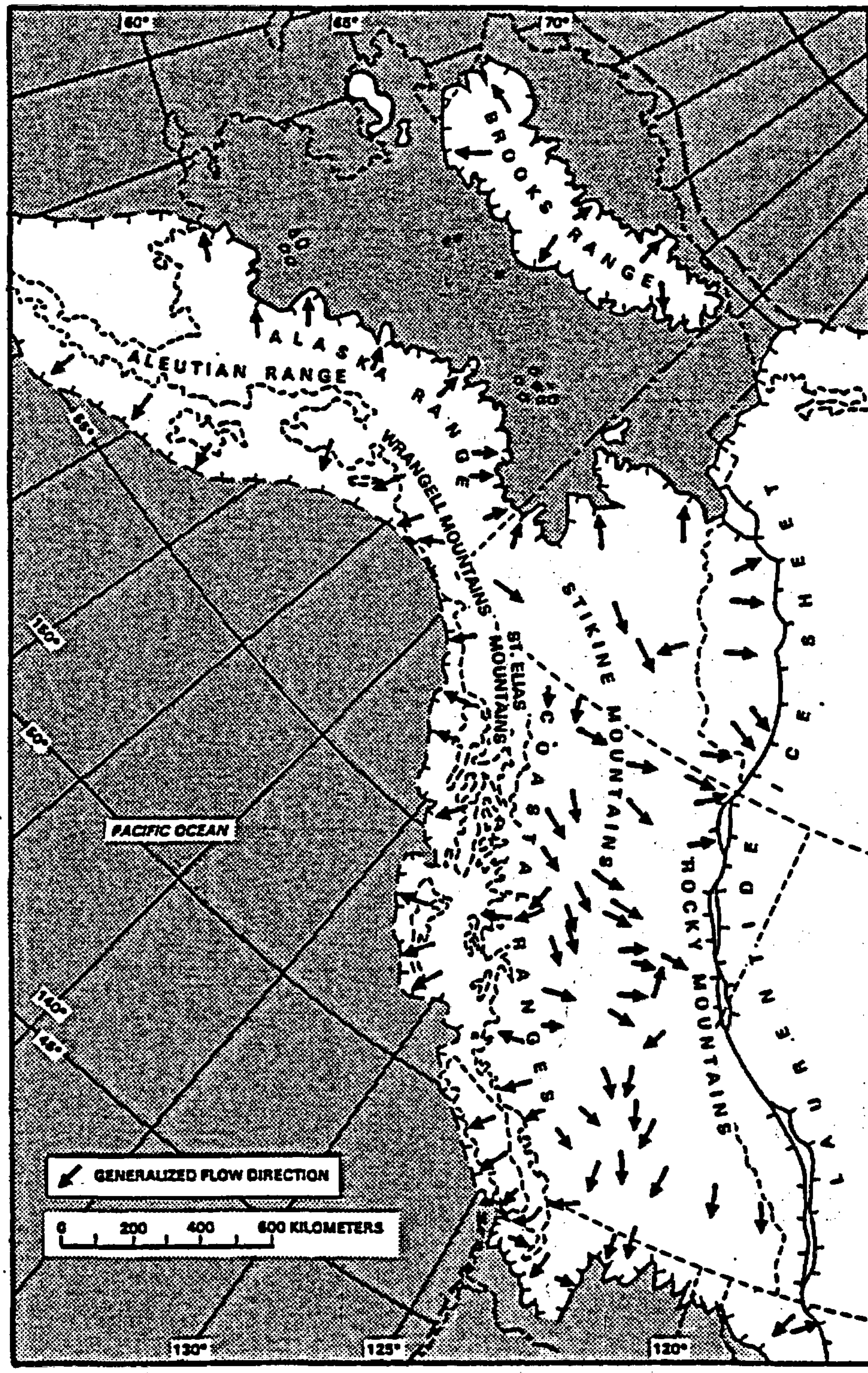


Figure 5.2-33. General trend of glacier advances in the western United States during the Holocene (Porter and Denton, 1967, p. 201). PS8607-187



PS8607-188

Figure 5.2-34. Configuration of the Cordilleran ice sheet at the time of maximum glaciation in the late Pleistocene. (Compiled from Prest et al. (1968) and maps in Wright and Frey (1965) (Flint, 1971).) PS8607-188

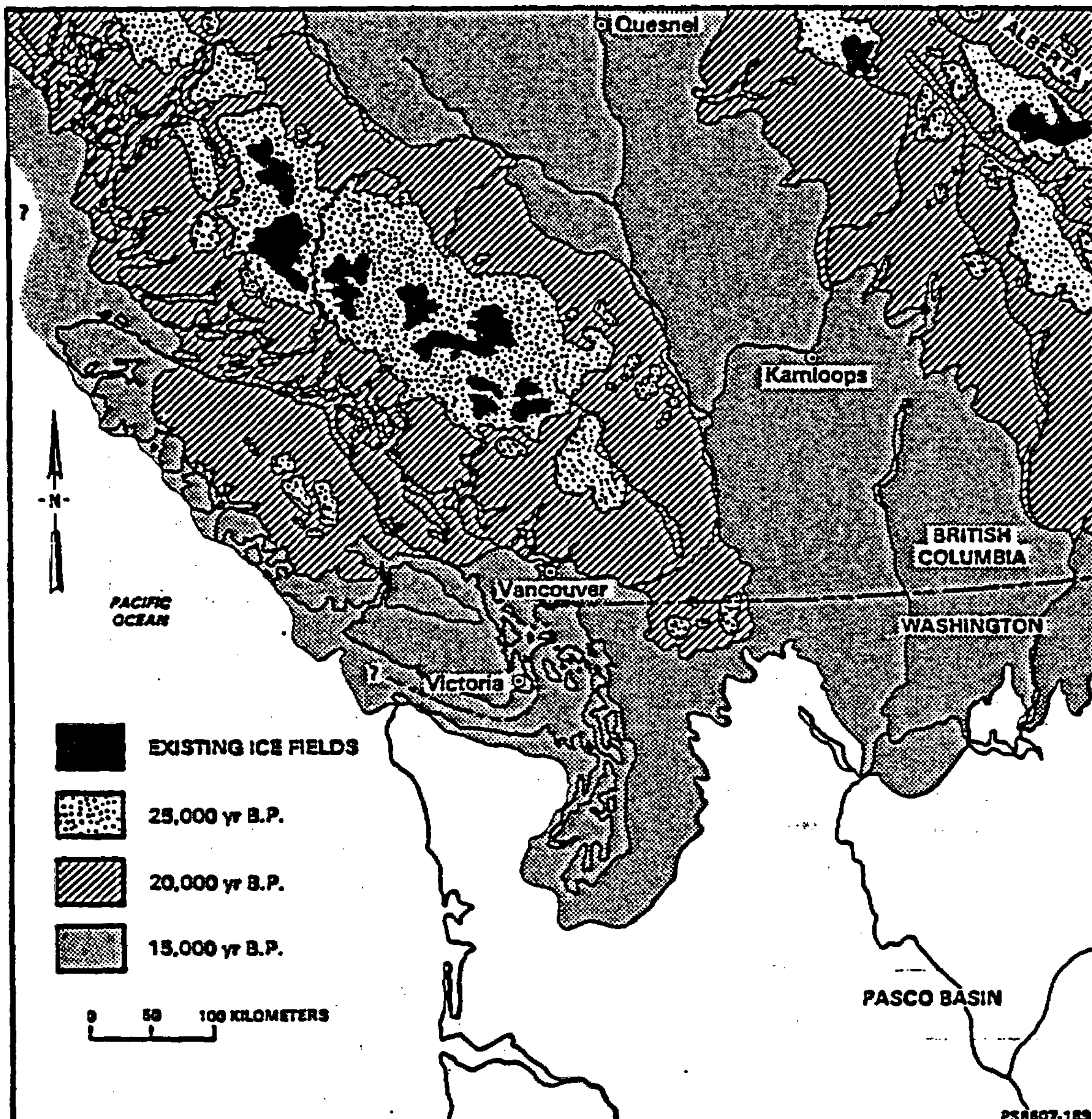


Figure 5.2-35. Stages in the advance of the Cordilleran ice sheet during the last glaciation of the late Pleistocene (Clague, 1981, Fig. 3, p. 10). PS8607-189

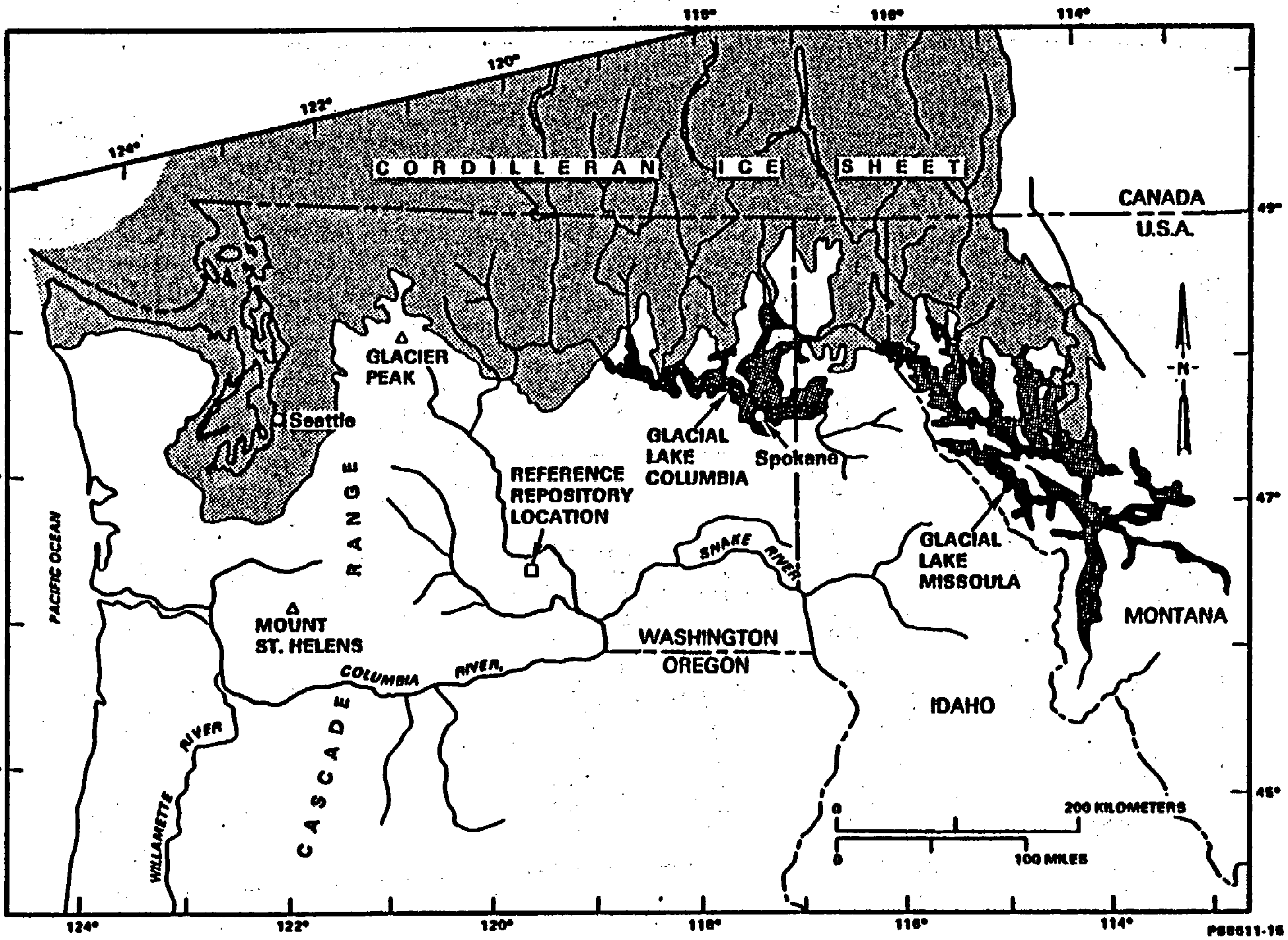


Figure 5.2-36. Configuration of the southern margin of the Cordilleran ice sheet at the time of maximum advance during the late Pleistocene (Waitt, 1985, Fig. 1, p. 1272). PS8611-15

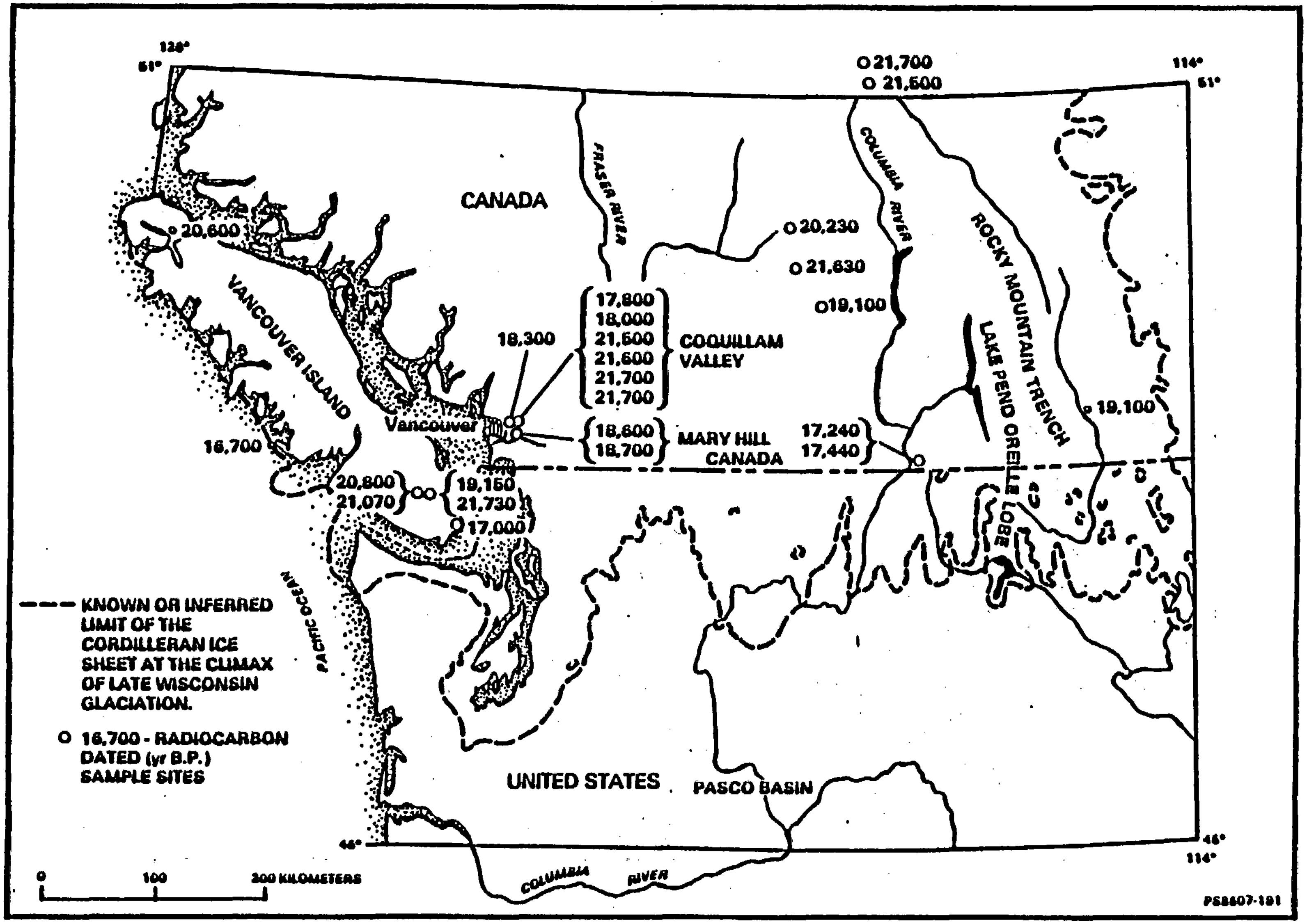


Figure 5.2-37. Some published radiocarbon-dated sample sites in the Pacific Northwest that bracket the age of the last advance of the Cordilleran ice sheet into the United States (Clague et al., 1980, p. 323, Fig. 1). PS8607-191

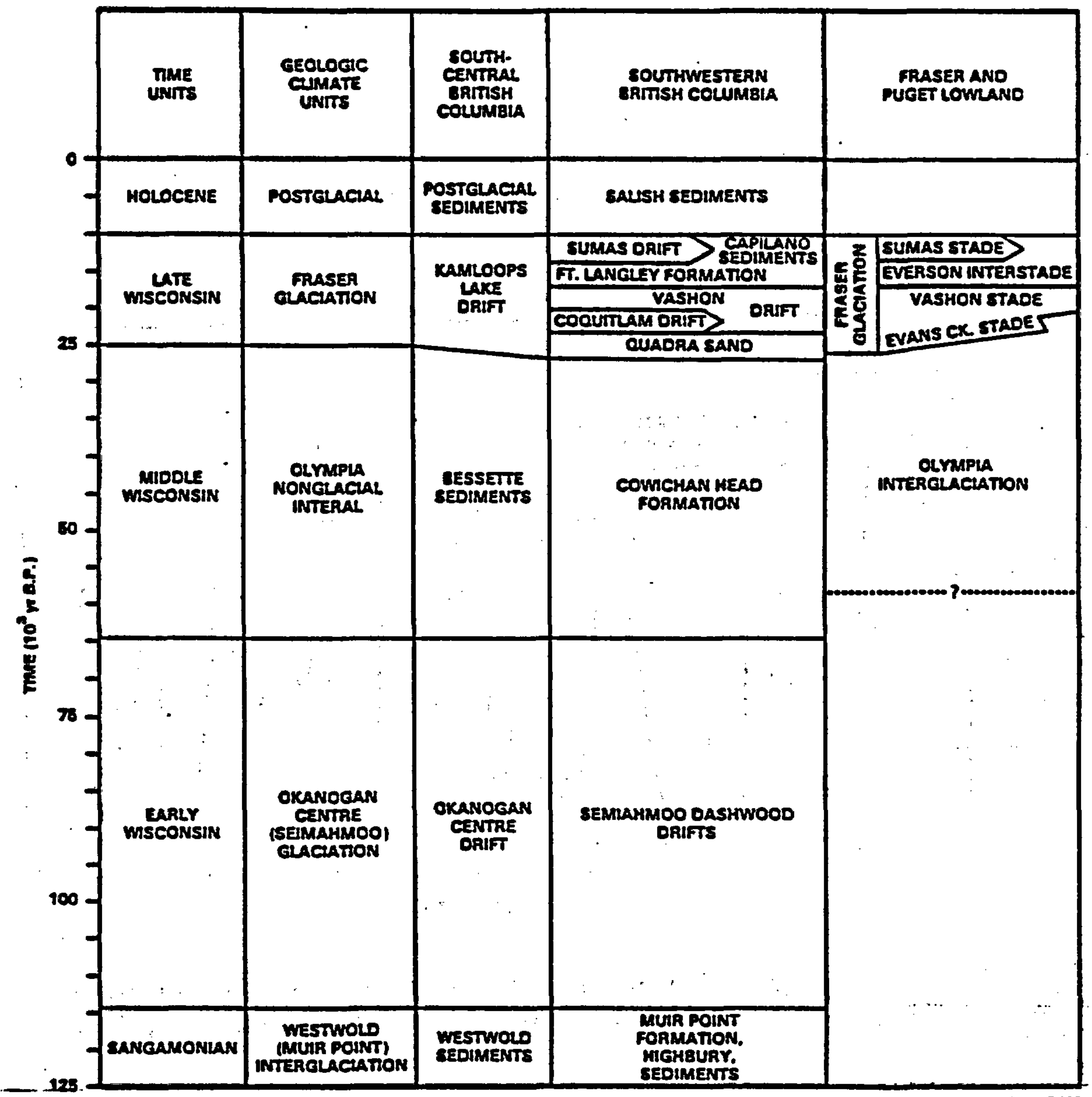


Figure 5.2-38. Stratigraphic relations observed at sites in southern British Columbia in sediments of the Fraser glaciation (Clague, 1978, Fig. 16.1, p. 96; Clague, 1981, Fig. 2, p. 6). (The relationships between lithostratigraphic units and the climatic events that influenced their deposition are discussed in Fulton and Smith, 1978; Armstrong, 1981; and Clague, 1981.) PS8607-192

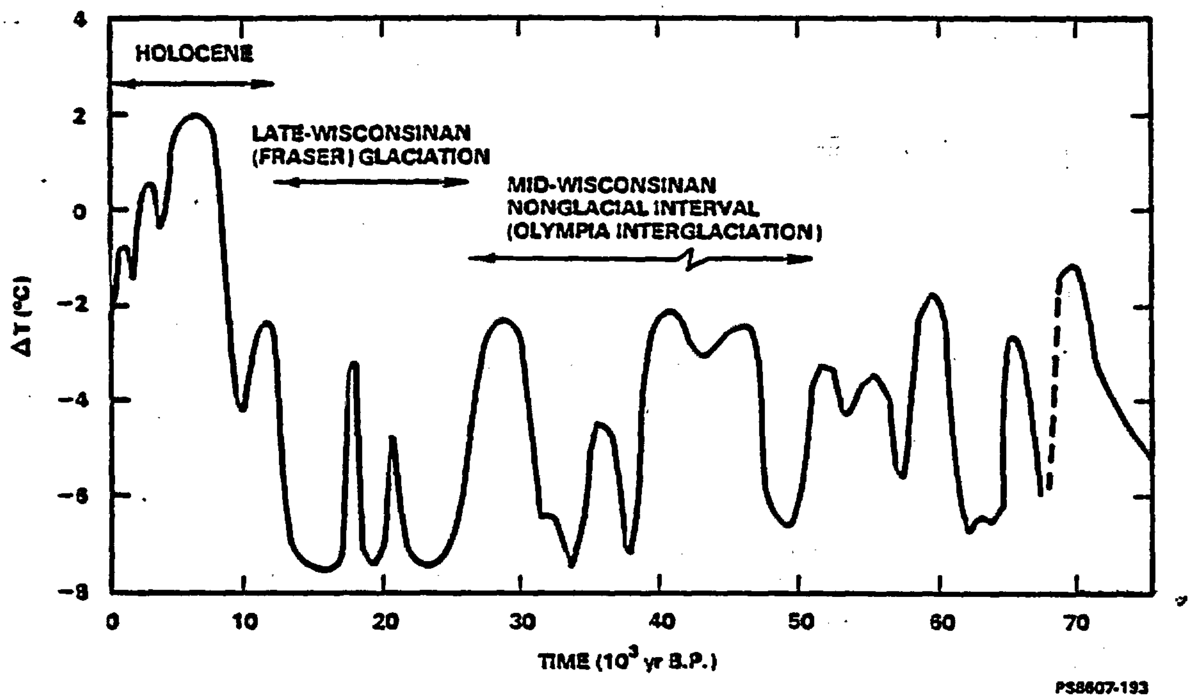


Figure 5.2-39. Pattern of temperature fluctuations reconstructed for the Pacific Northwest throughout the past 70,000 years (Heusser, 1977, Fig. 16). PS8607-193

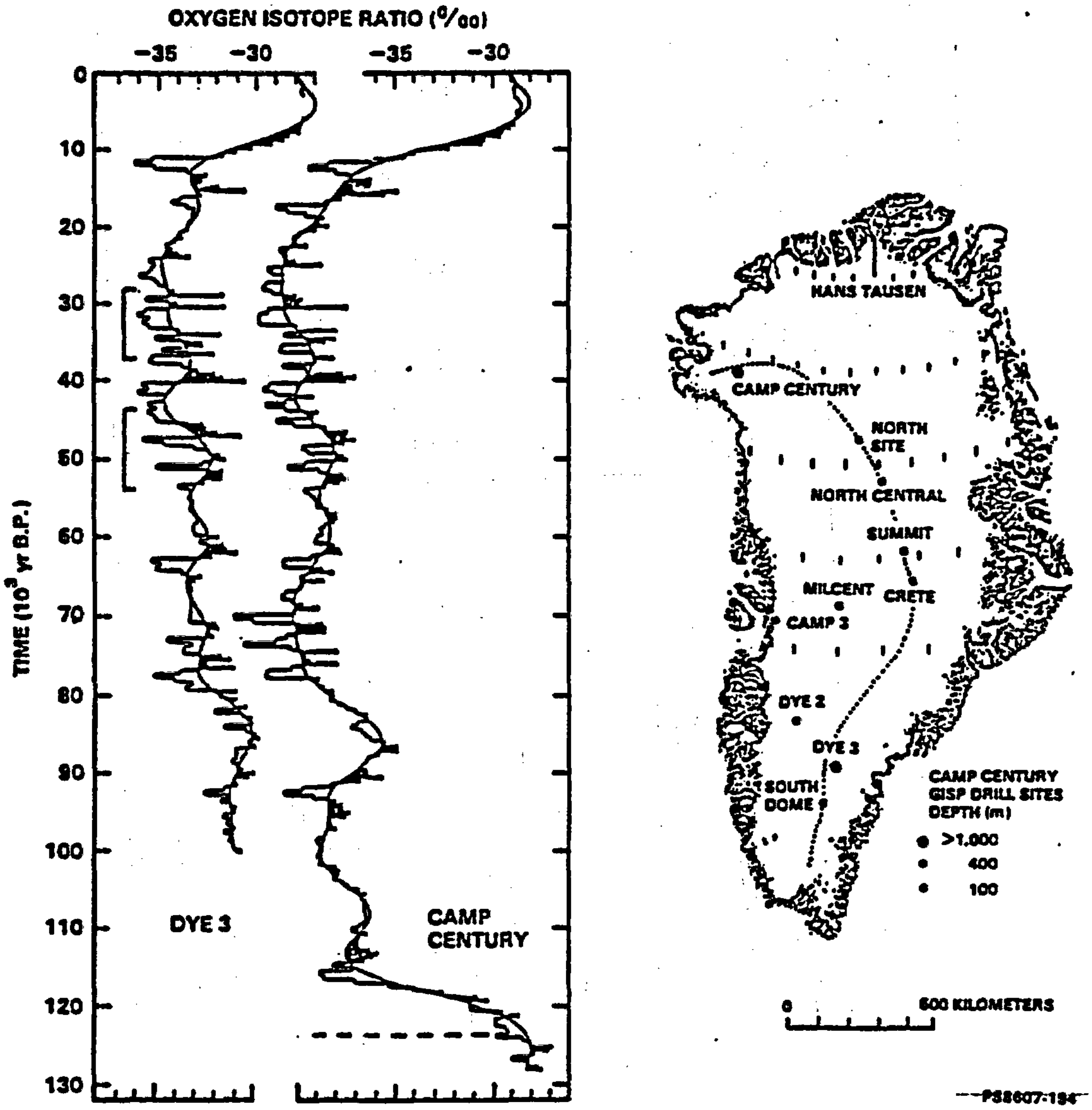


Figure 5.2-40. The oxygen isotopic ratios observed in ice cores obtained from two drilling sites in Greenland. Locations of these and other available cores are shown on index map at right (Dansgaard et al., 1984, Fig. 1 and 3, pp. 289-90). PS8607-194

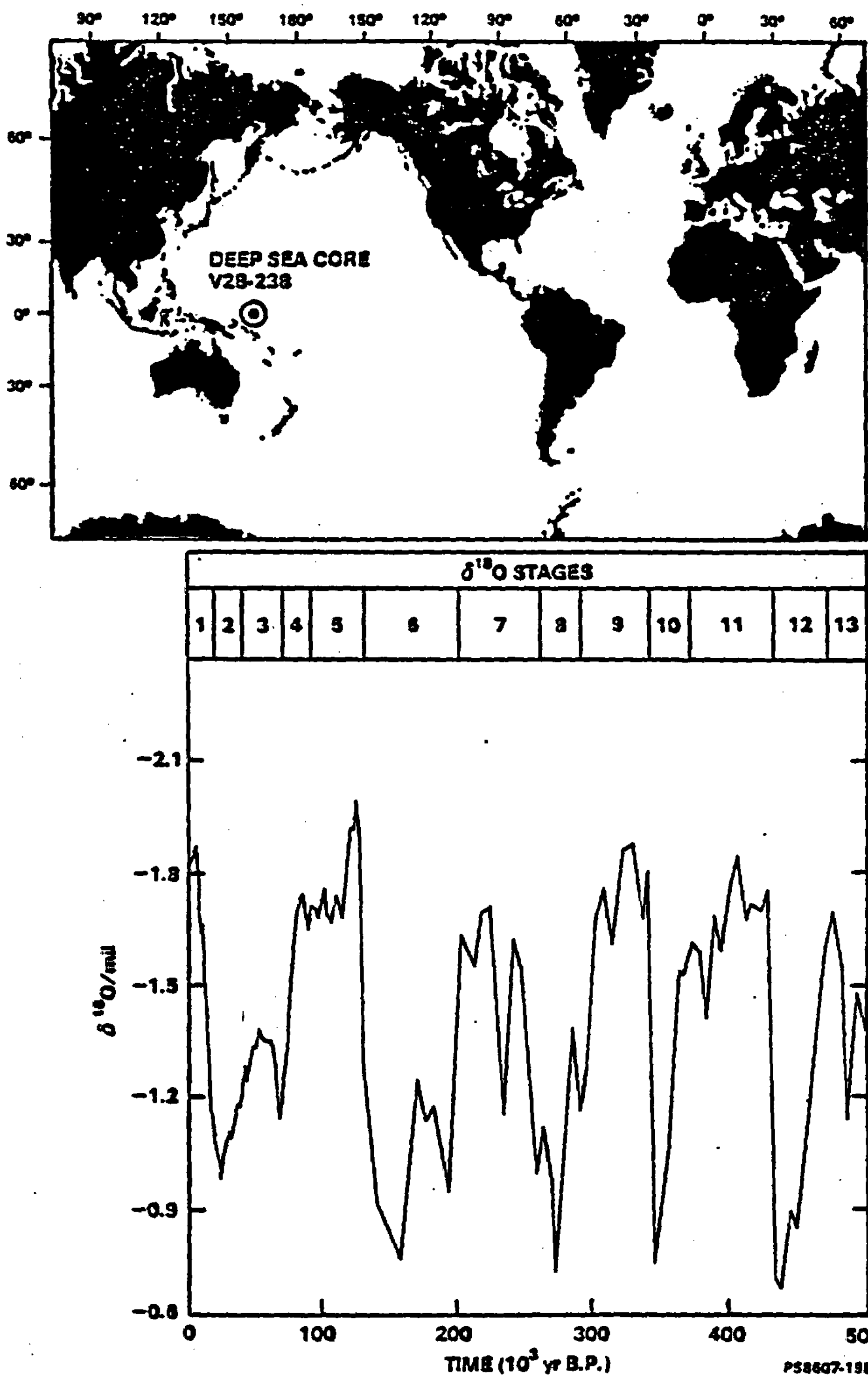


Figure 5.2-41. The oxygen isotopic record observed in deep-sea core v28-238 from the equatorial Pacific (Imbrie et al., 1984, Fig. 1, p. 271; Imbrie and Imbrie, 1980). PS8607-195

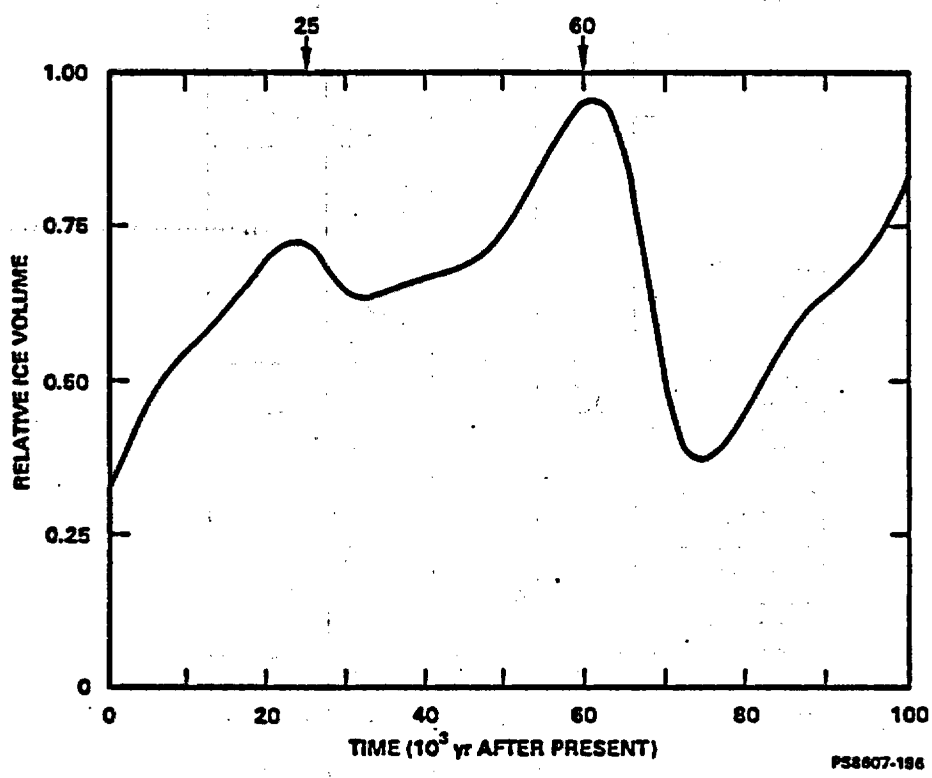


Figure 5.2-42. Estimated relative volume of ice on the surface of the Earth over the next 100,000 years based on the assumption that the future behavior is predictable from the record of the past 780,000 years (Imbrie and Imbrie, 1980). PS8607-196

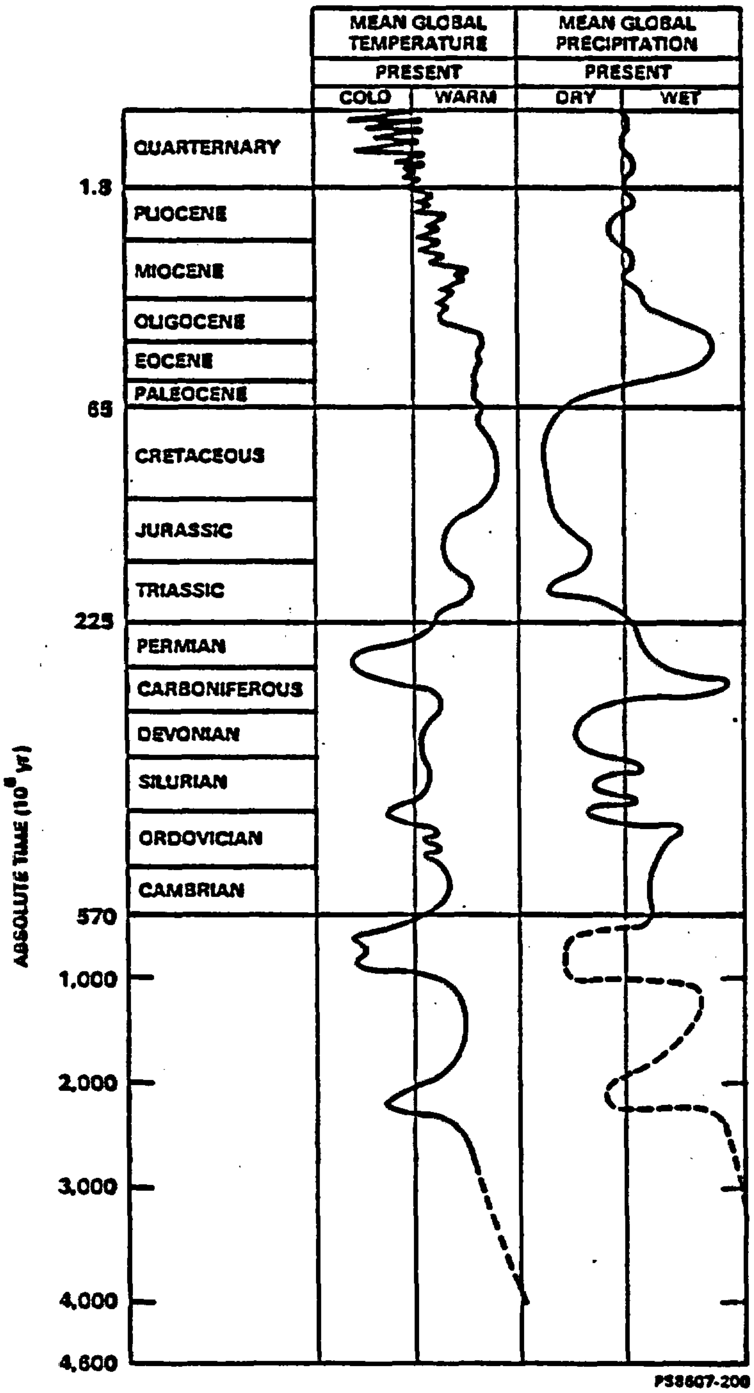


Figure 5.2-43. Schematic reconstruction of variations of temperature and precipitation throughout the Phanerozoic. Trends are dashed where data are very sparse (Frakes, 1979, p. 261, Fig. 9-1). PS8607-200

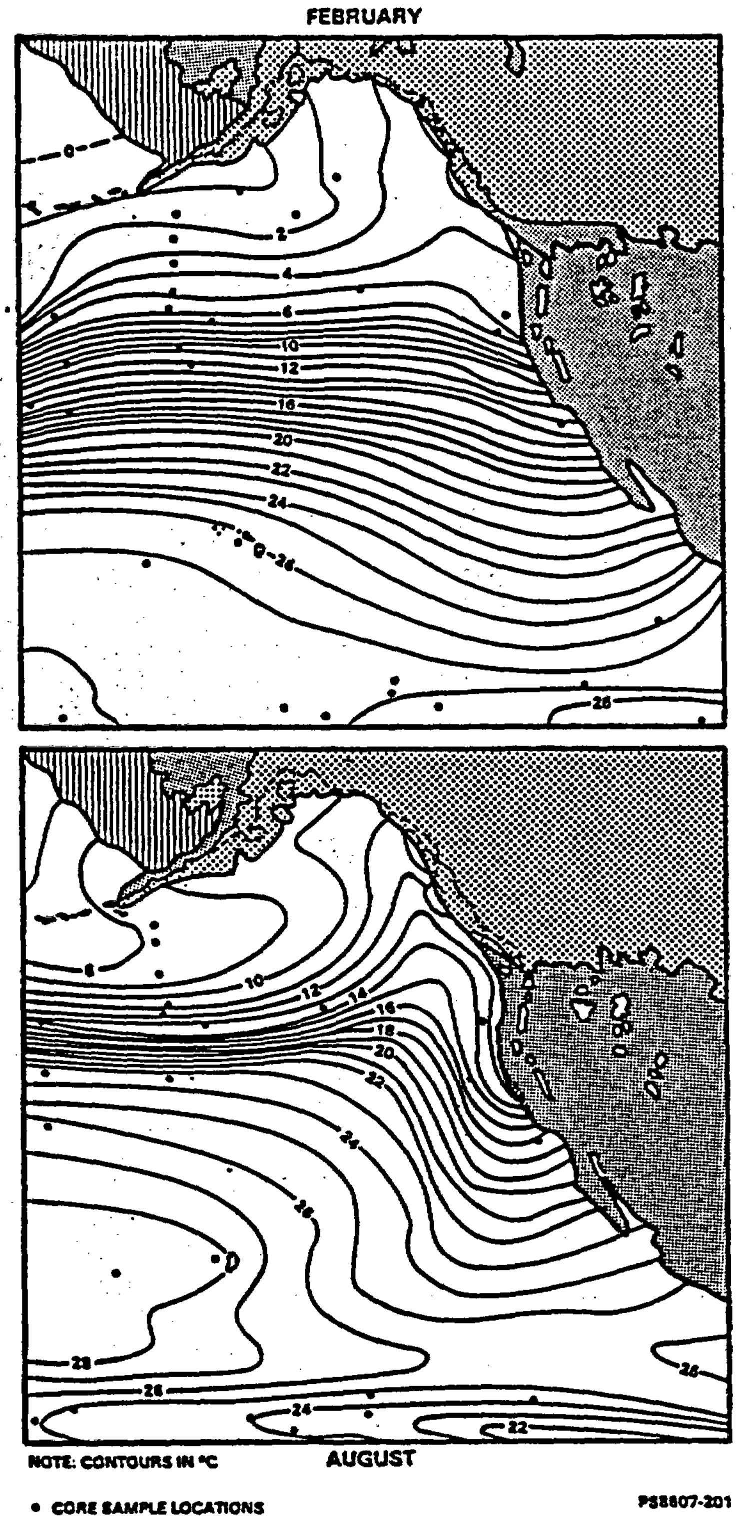


Figure 5.2-44. Sea surface temperature pattern reconstructed at the last glacial maximum (CLIMAP, 1981). PS8607-201

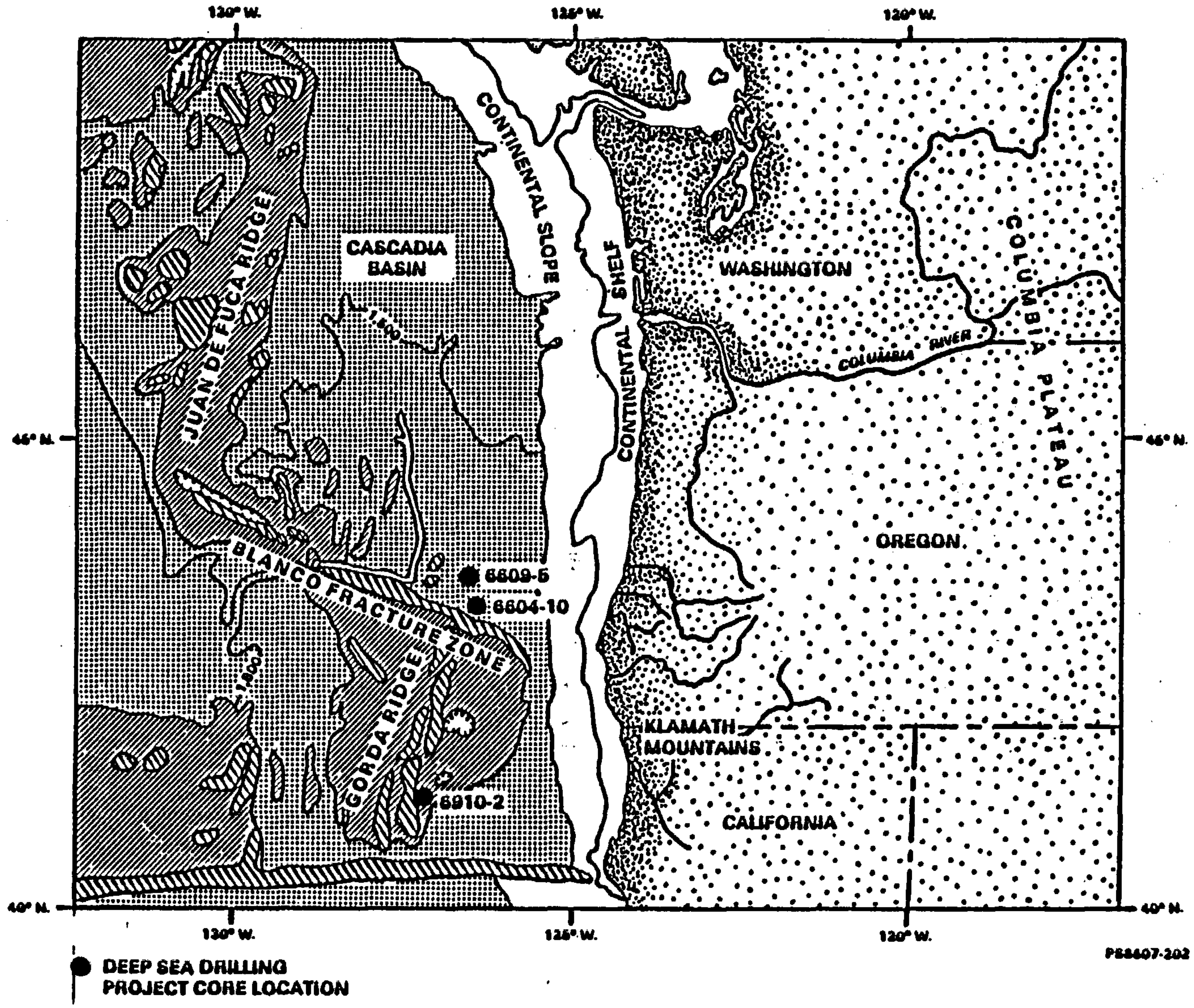


Figure 5.2-45. Locations of piston cores obtained within the Cascadia Basin used for Late Pleistocene paleotemperature studies (Heath et al., 1976, Fig. 1, p. 395). PS8607-202

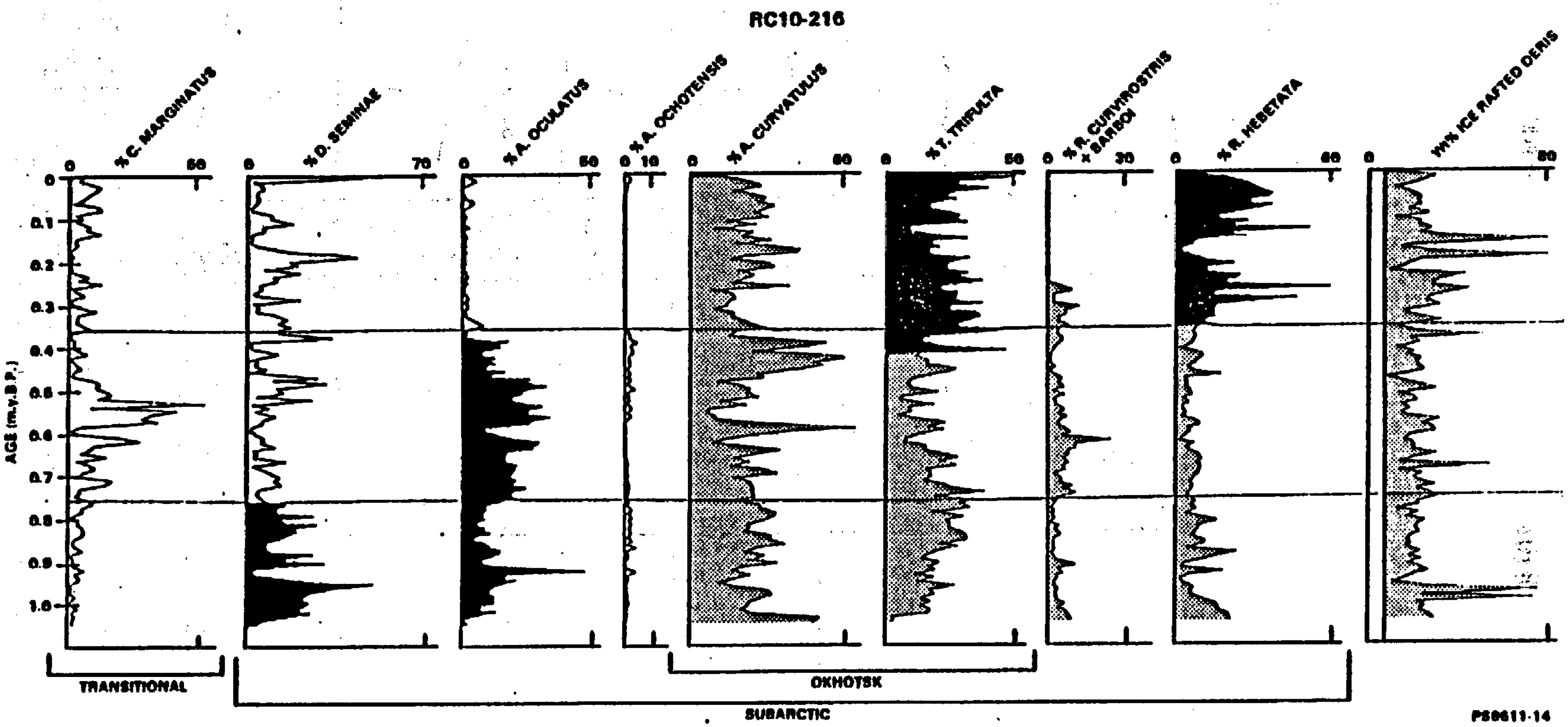


Figure 5.2-46. Record of climatic variability of the past million years preserved in core RC10-216 from the northeast Pacific Ocean. PS8611-14

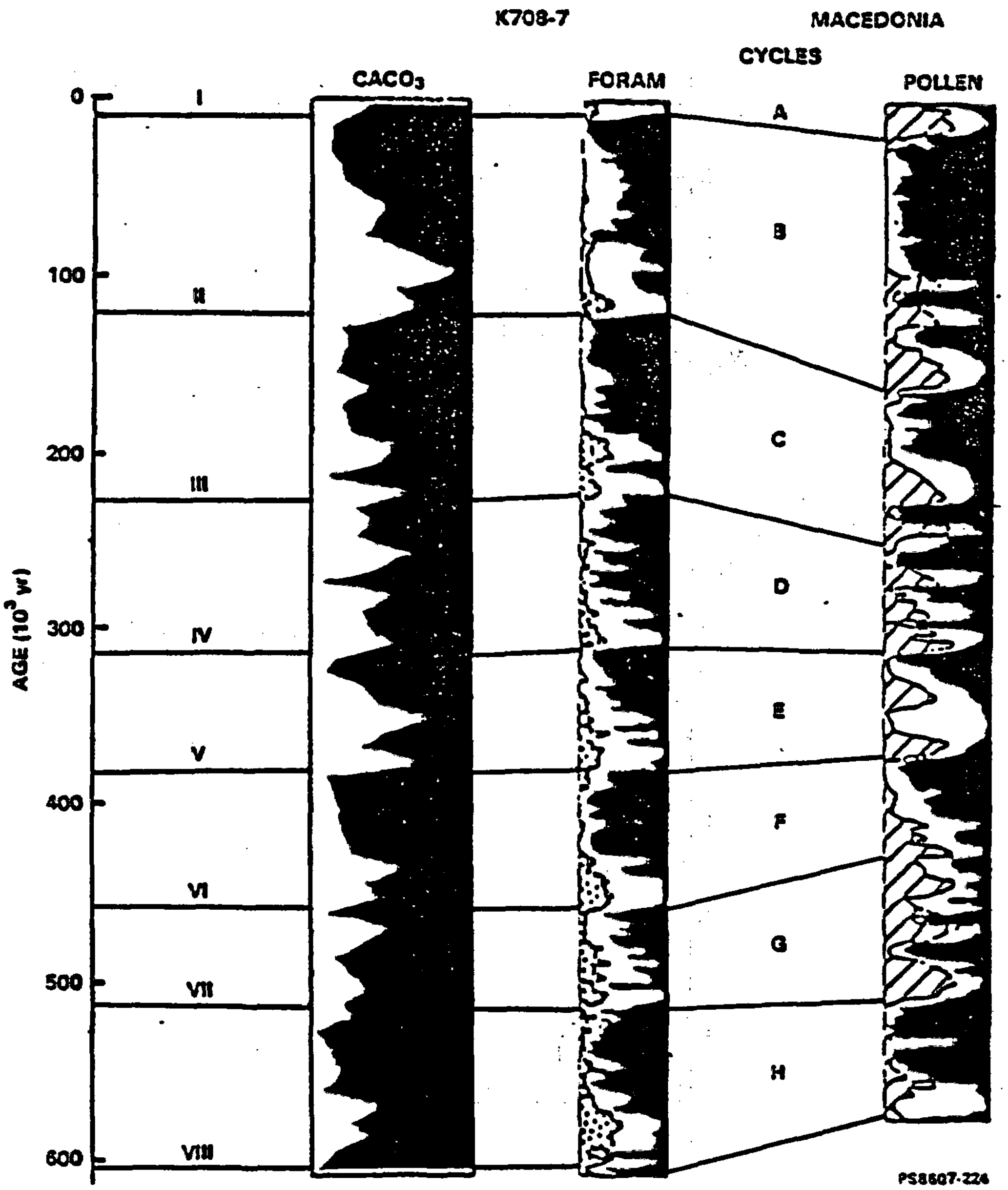


Figure 5.2-47. Correlation of a pollen core from Macedonia, Greece, with deep-sea core indicators of paleo sea surface temperatures (Ruddiman and McIntyre, 1976, Fig. 14, p. 138). Dates of oxygen isotopic stage terminations (Roman numerals) are given in Table 5.2-6. PS8607-224

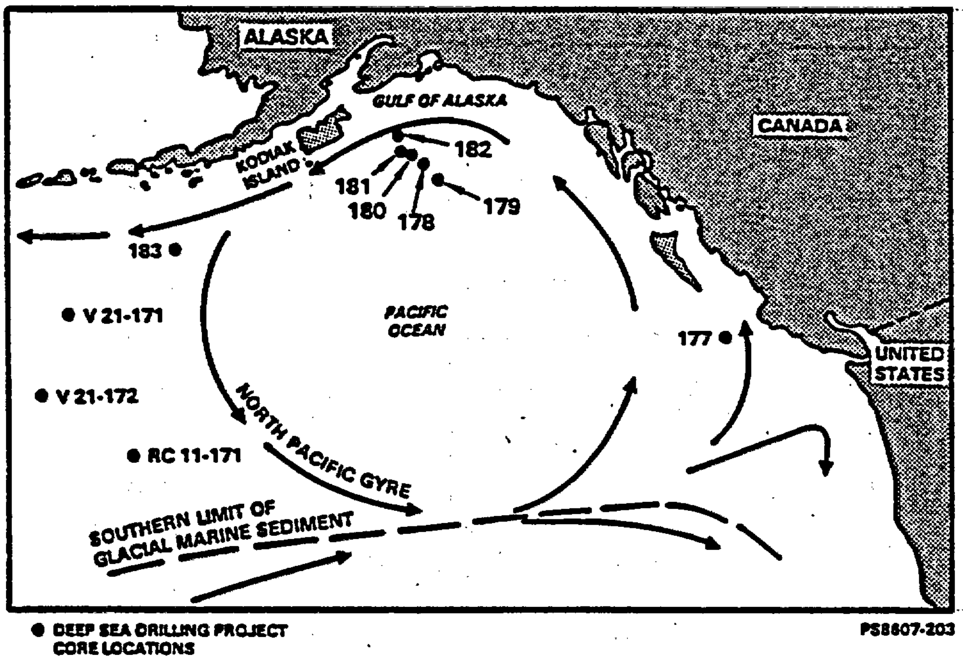


Figure 5.2-48. Locations of Deep-Sea Drilling Project cores obtained from the North Pacific and Gulf of Alaska (von Huene et al., 1976, Fig. 1, p. 412; Kent et al., 1971). PS8607-203

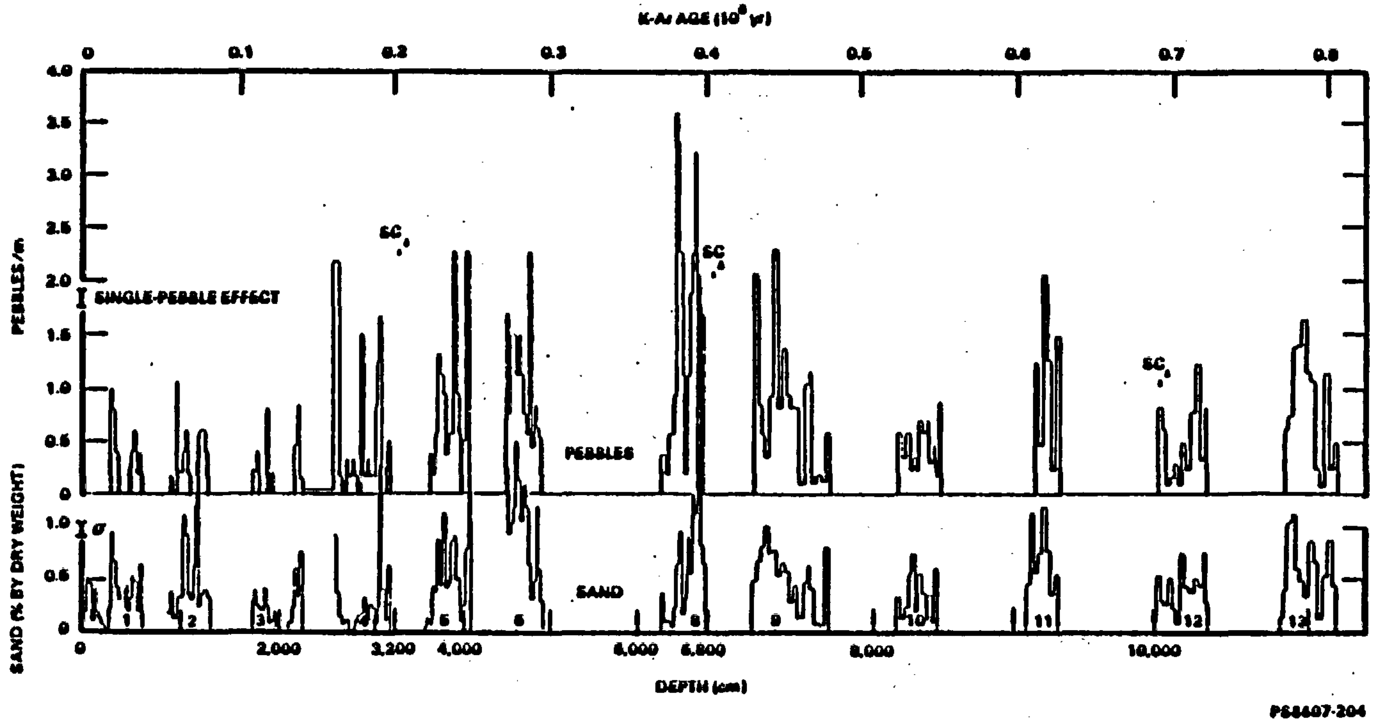


Figure 5.2-49. Variations in the percentage of sand and pebble count measured in core DSDP-178 from the North Pacific Ocean. The location of this core is shown in Figure 5.2-48 (von Hune et al., 1976, Fig. 2, p. 414). PS8607-204

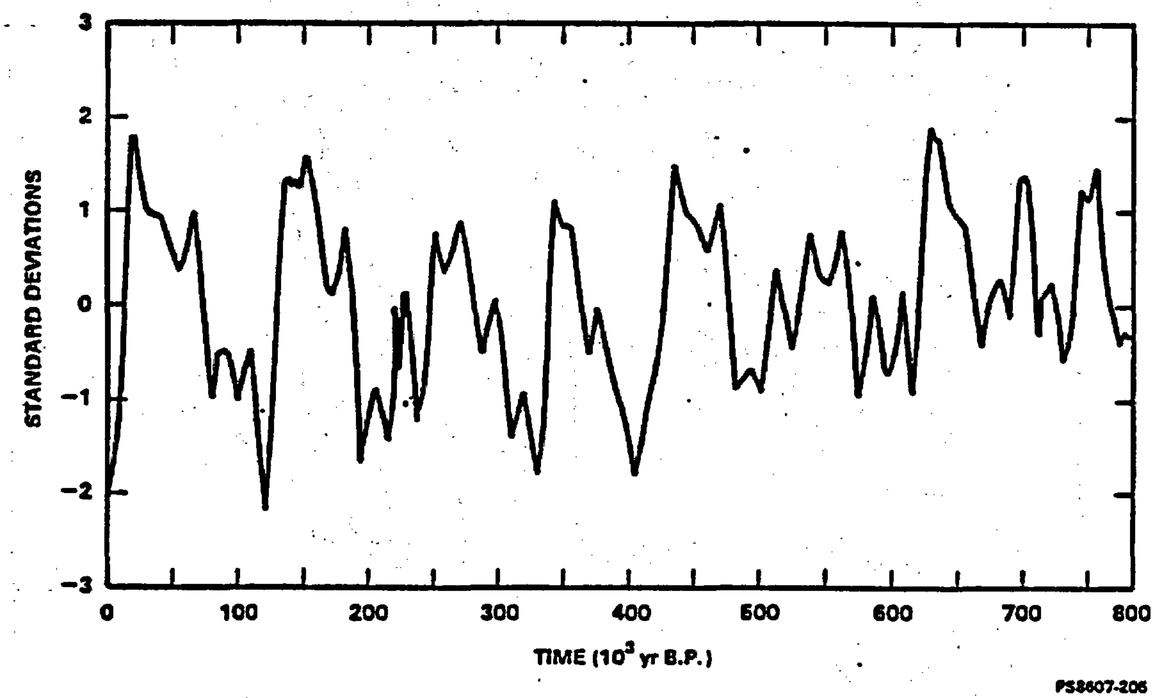


Figure 5.2-50. Oxygen-isotopic variations reconstructed by the Spectral Mapping Project (plotted from data given by Imbrie et al., 1984, Table 7, pp. 291-293). PS8607-205

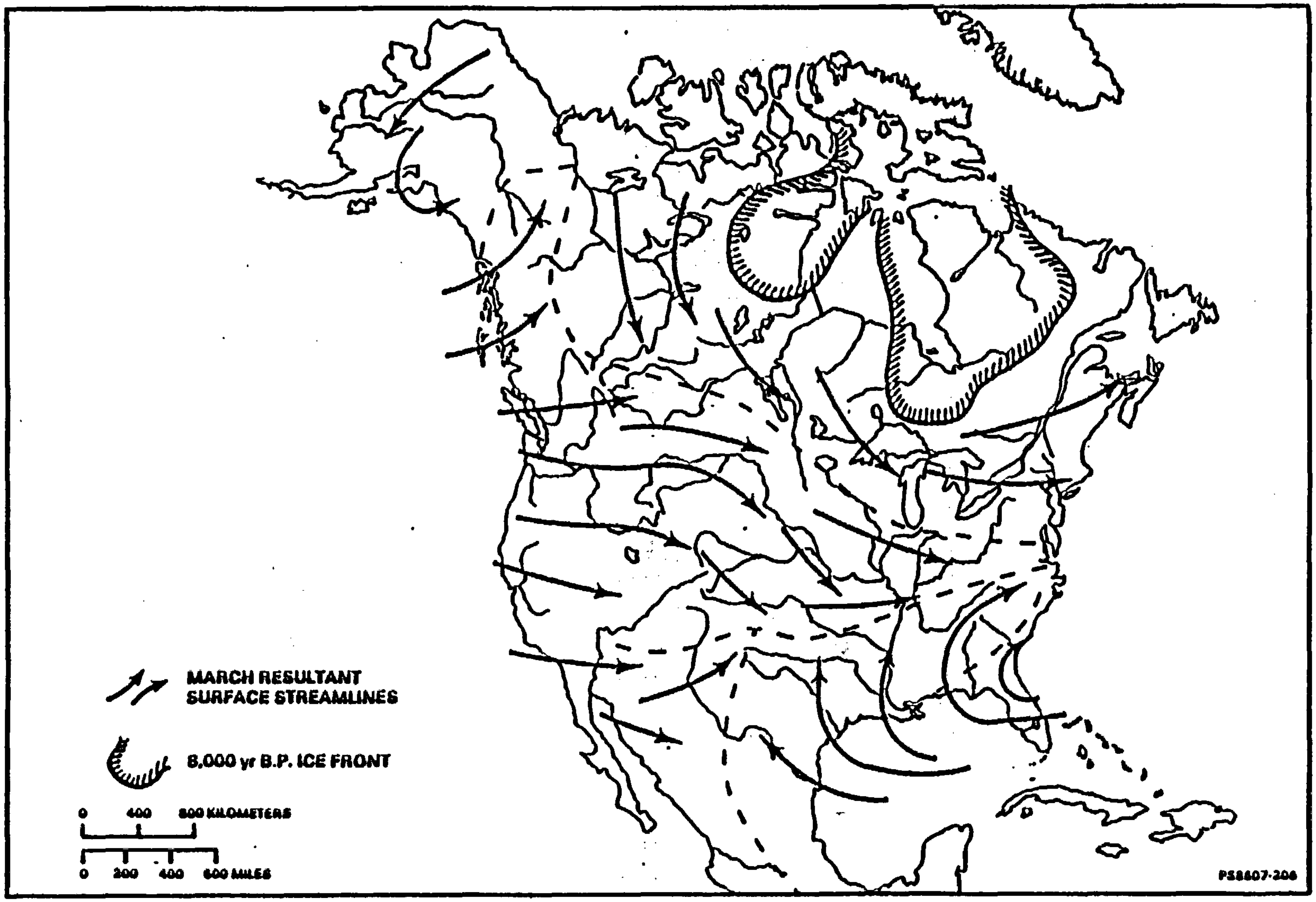


Figure 5.2-51. Pattern of atmospheric circulation reconstructed for 8,000 years before present (Knox, 1983, Fig. 3-6, p. 31). PS8607-206

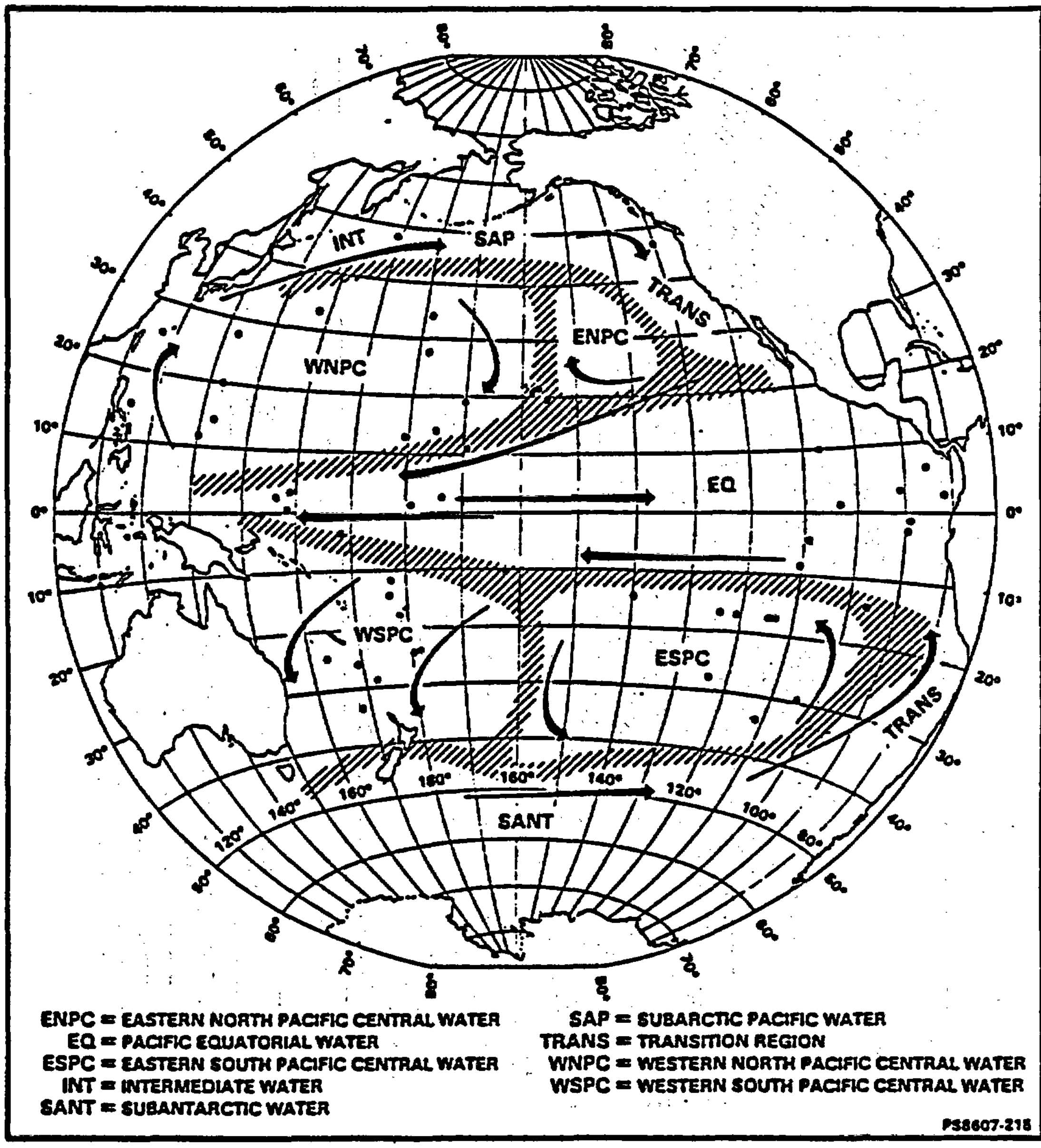


Figure 5.2-52. Circulation patterns in the Pacific Ocean (Geitzenauer et al., 1976, Fig. 2, p. 429; Sverdrup et al., 1942, p. 740, Fig. 209 A). PS8607-215

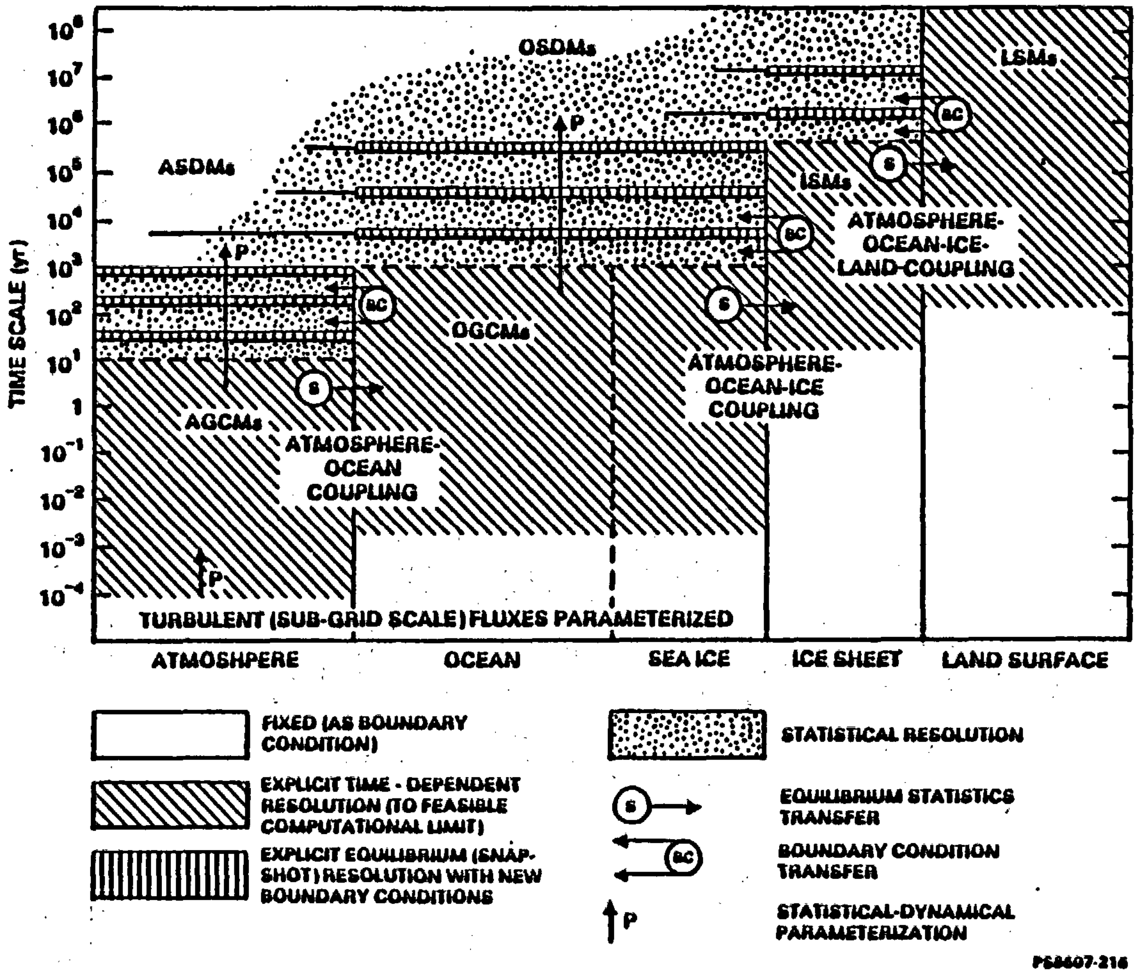


Figure 5.2-53. Schematic summary of a snapshot strategy for integrating the components of the coupled climatic system (Gates, 1982, Fig. 2.12, p. 40). PS8607-216

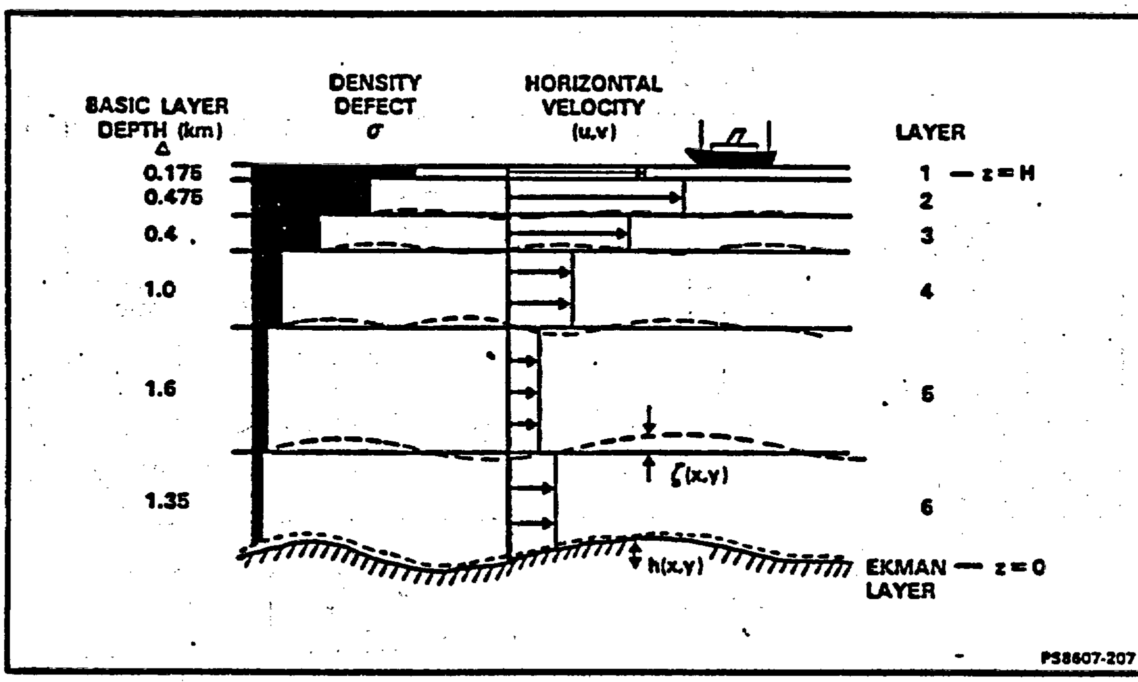


Figure 5.2-54. Example of layers included in a mixed-layer ocean model (Bretherton and Karweit, 1975, p. 238). PS8607-207

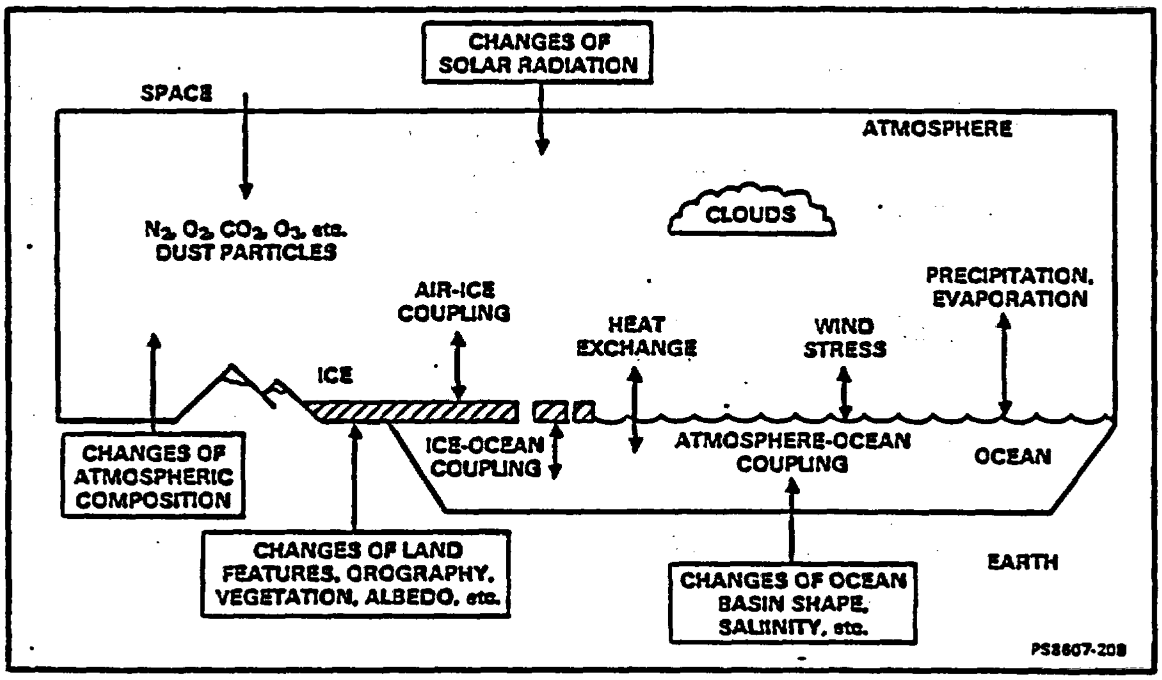


Figure 5.2-55. The global cryosphere-atmosphere-ocean system including the major sources of variation at geologic time scales and the interactions between these sources (EOS, 1981). PS8607-208

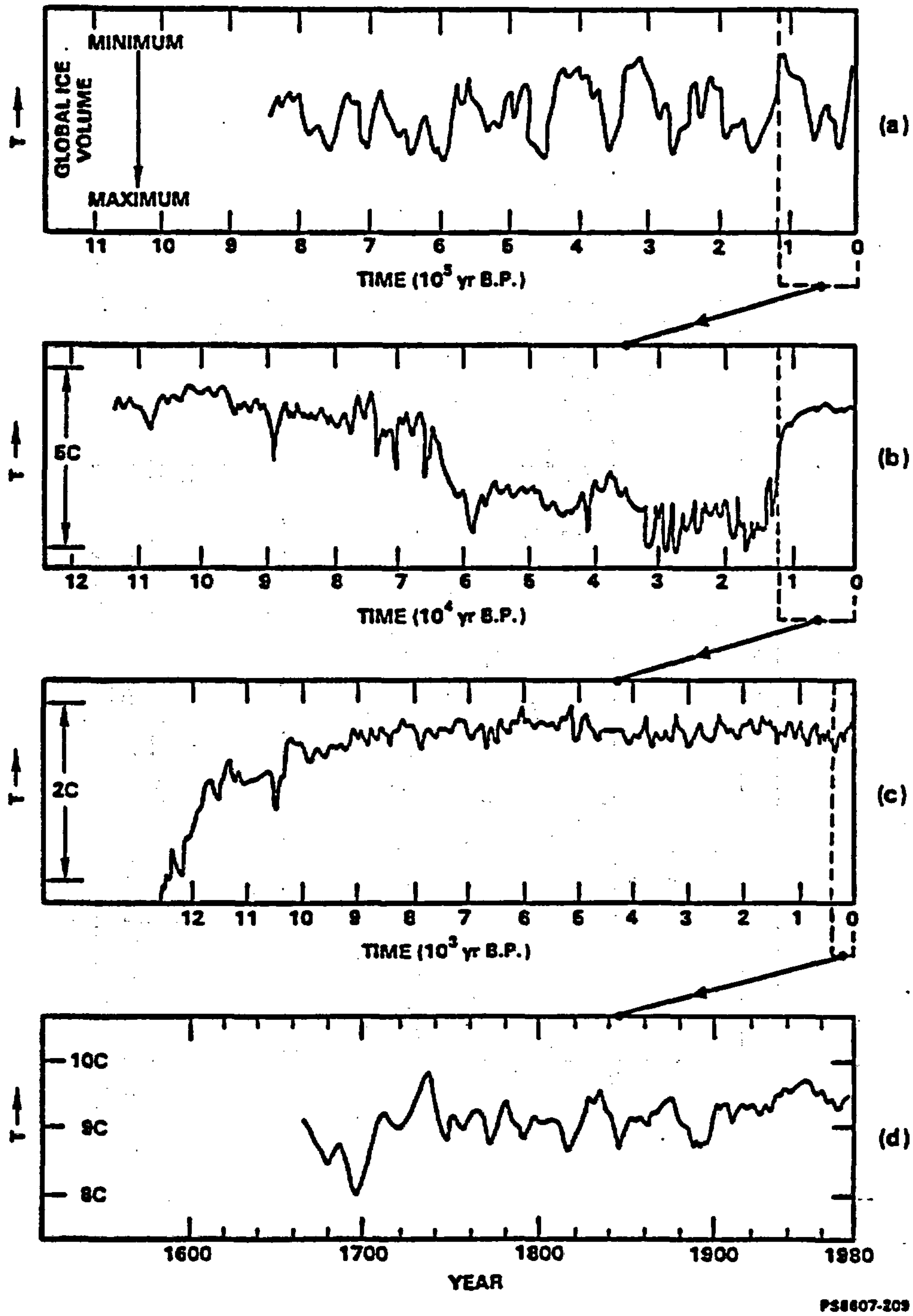


Figure 5.2-56. Variations in the global ice budget of the Earth over the arious time scales of interest in nuclear waste disposal (Saltzman, 1985, Fig. 1, p. 342). PS8607-209

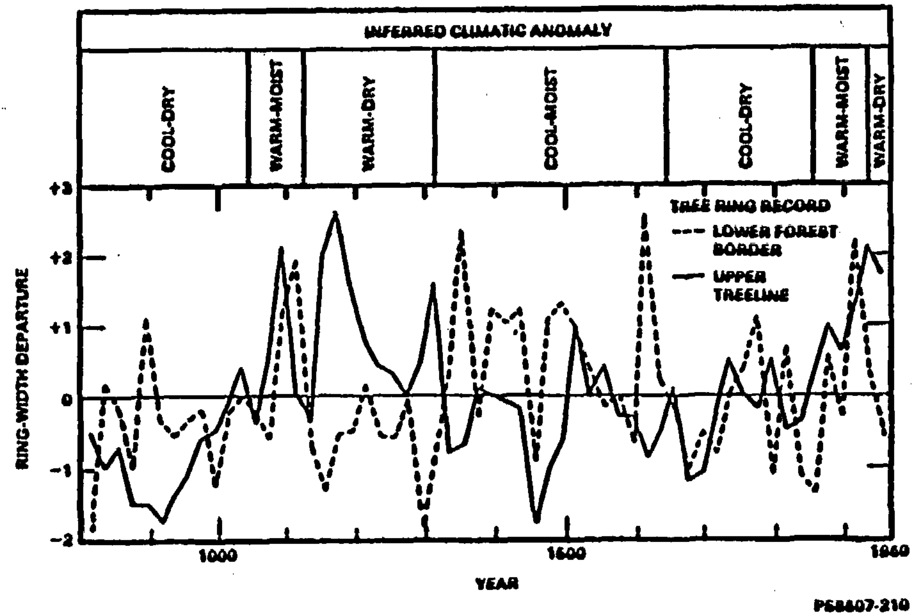


Figure 5.2-57. Alternation of periods of warmth and cold and (or) moist and dry reconstructed by dendroclimatic studies of Bristlecone Pines from the Sierra Nevada, California (LaMarche, 1974, Fig. 6, p. 1047). PS8607-210

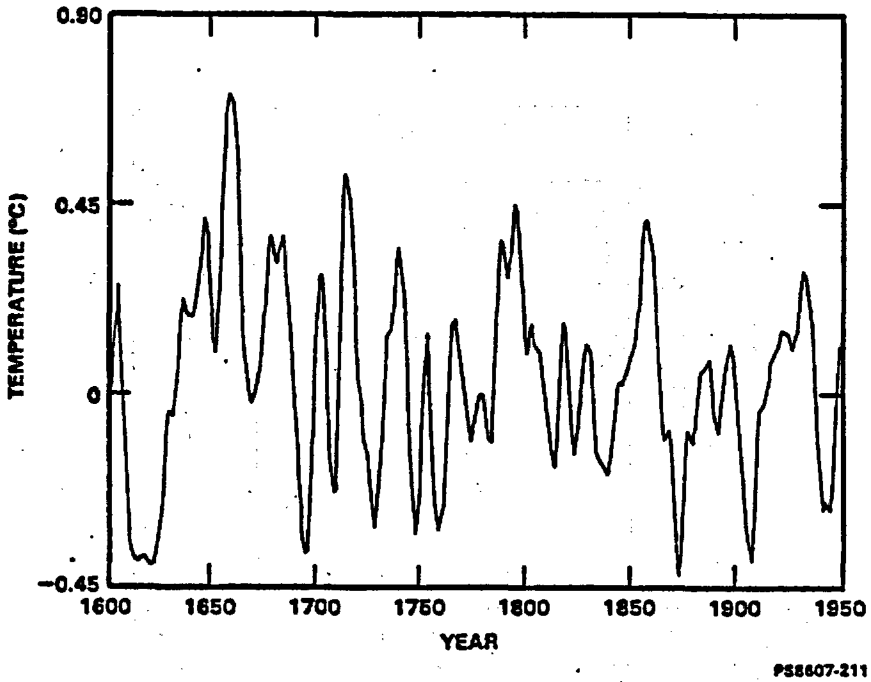


Figure 5.2-58. Reconstructed variations of temperature in the Pacific Northwest from the year 1600 to 1950 (Cropper and Fritts, 1986, annual temperature table, App. 3). PS8607-211

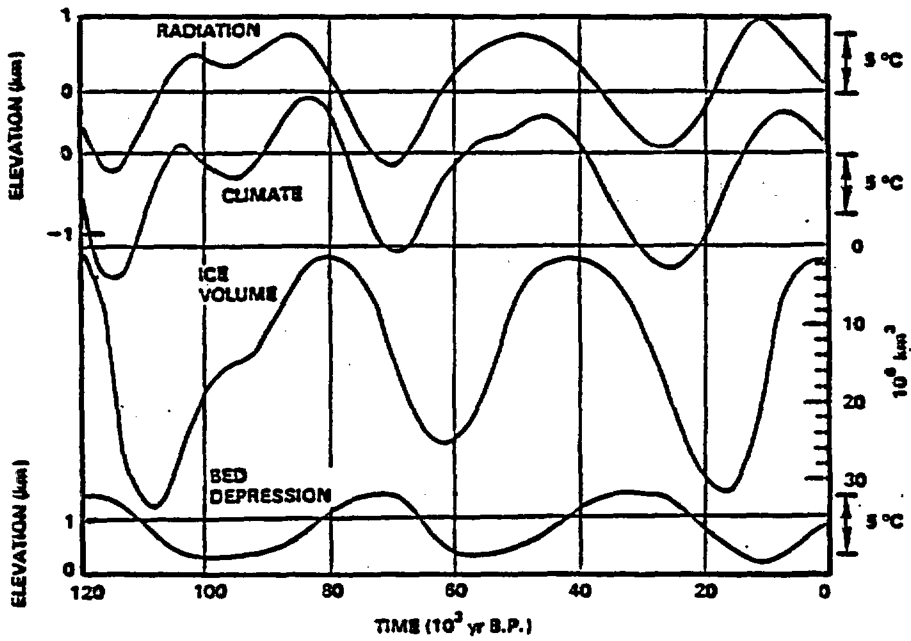


Figure 5.2-59. Model results illustrating the behavior of several components of the climatic system which controlled the fluctuations of the North American ice sheet over the past 120,000 years (Budd and Smith, 1981, pp. 369-409). PS8607-212

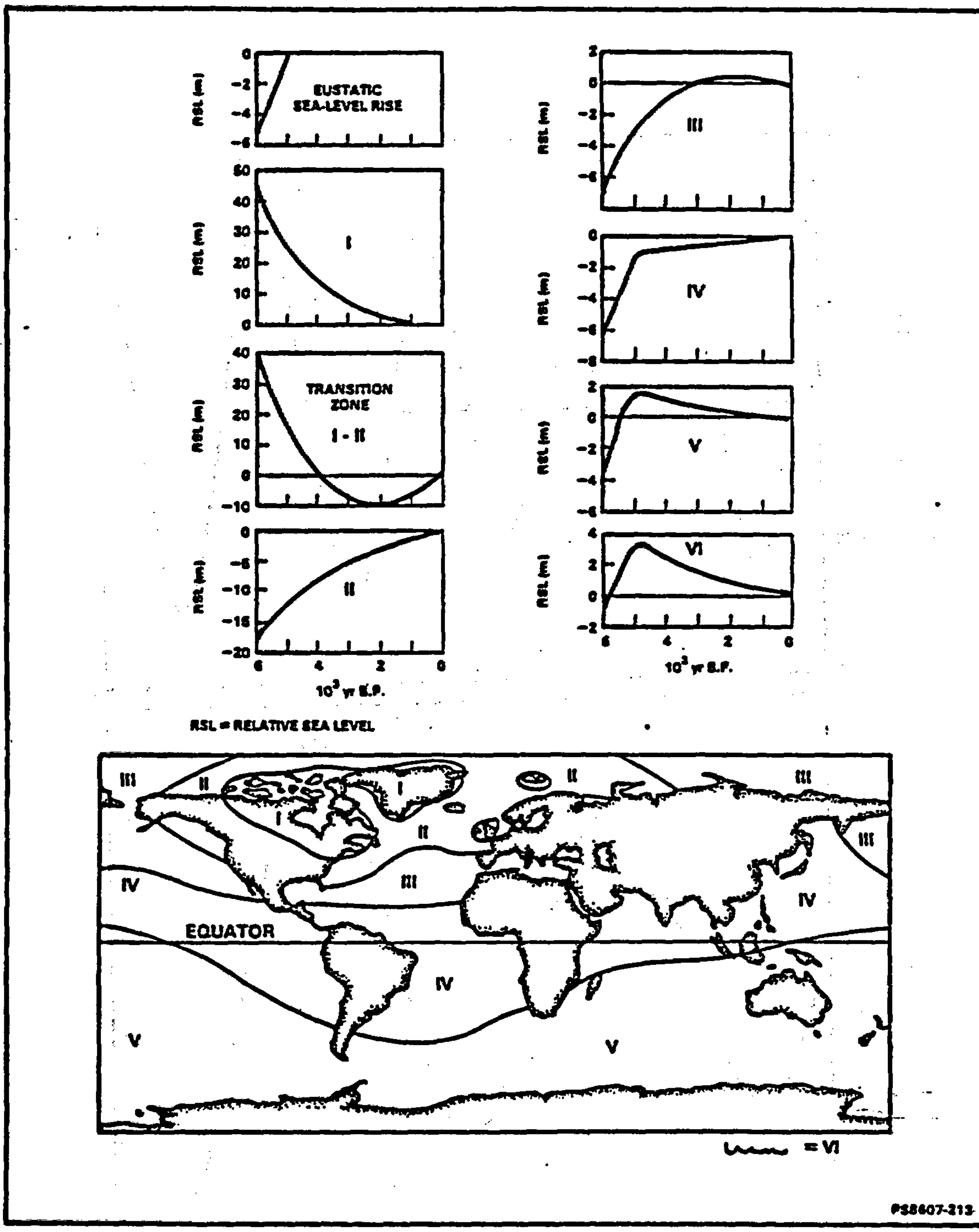


Figure 5.2-60. Relative sea-level curves for the Holocene deglaciation. Based on an assumed sea-level rise of 75.6 meters between 17,000 and 5,000 years before present (Clark and Lingle, 1979, Fig. 1). PS8607-213

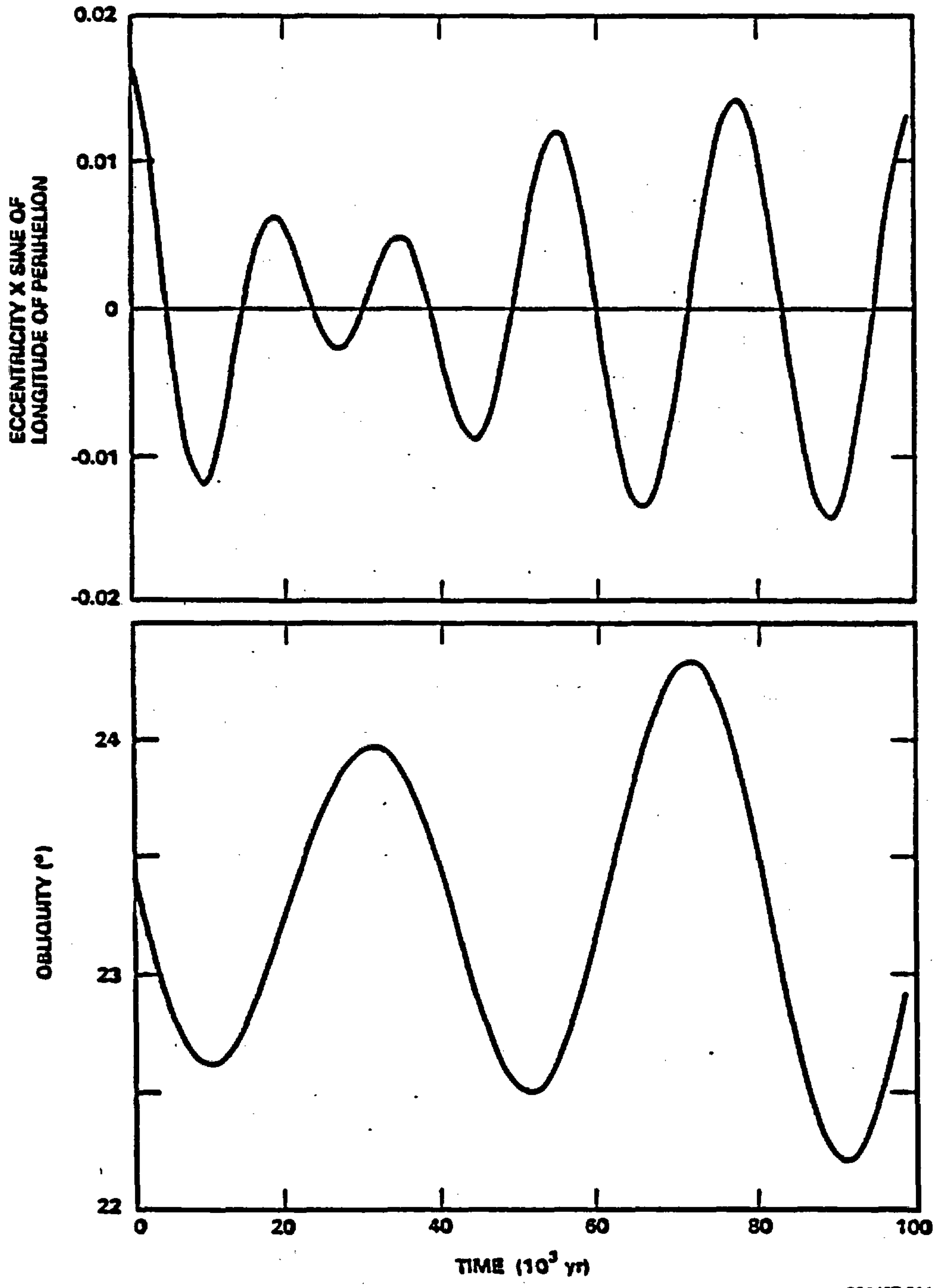


Figure 5.2-61. Precession (top) and variations in obliquity (bottom) of the Earth's orbit over the next 100,000 years. PS8607-214

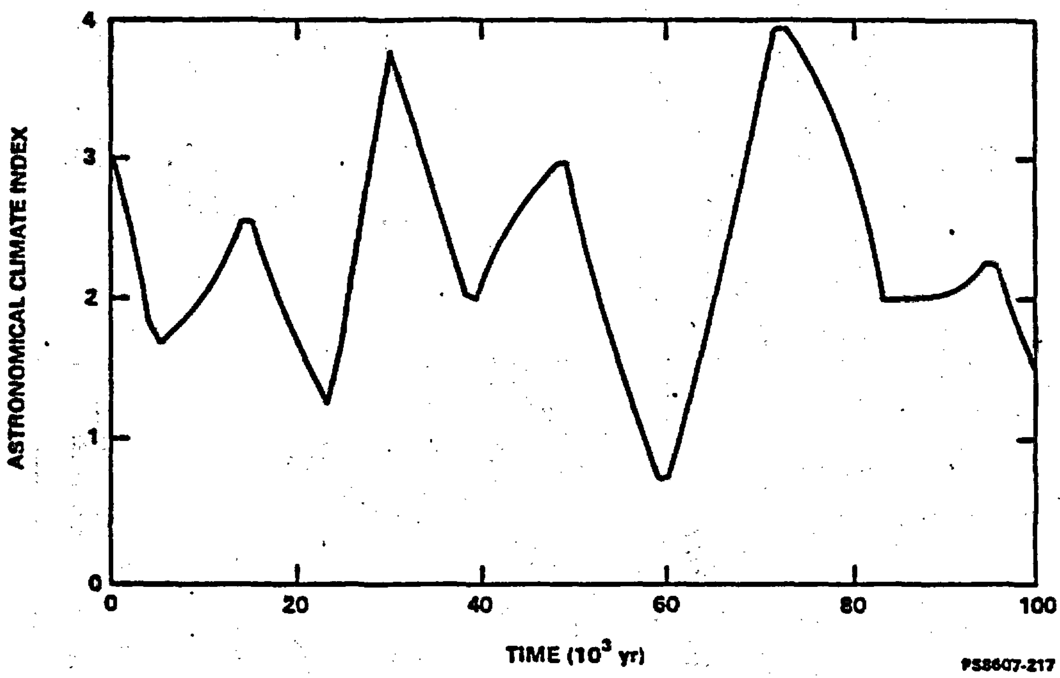


Figure 5.2-62. Variations in the Astronomical Climate Index (ACLIN) over the next 100,000 years (Kukla et al., 1981). PS8607-217

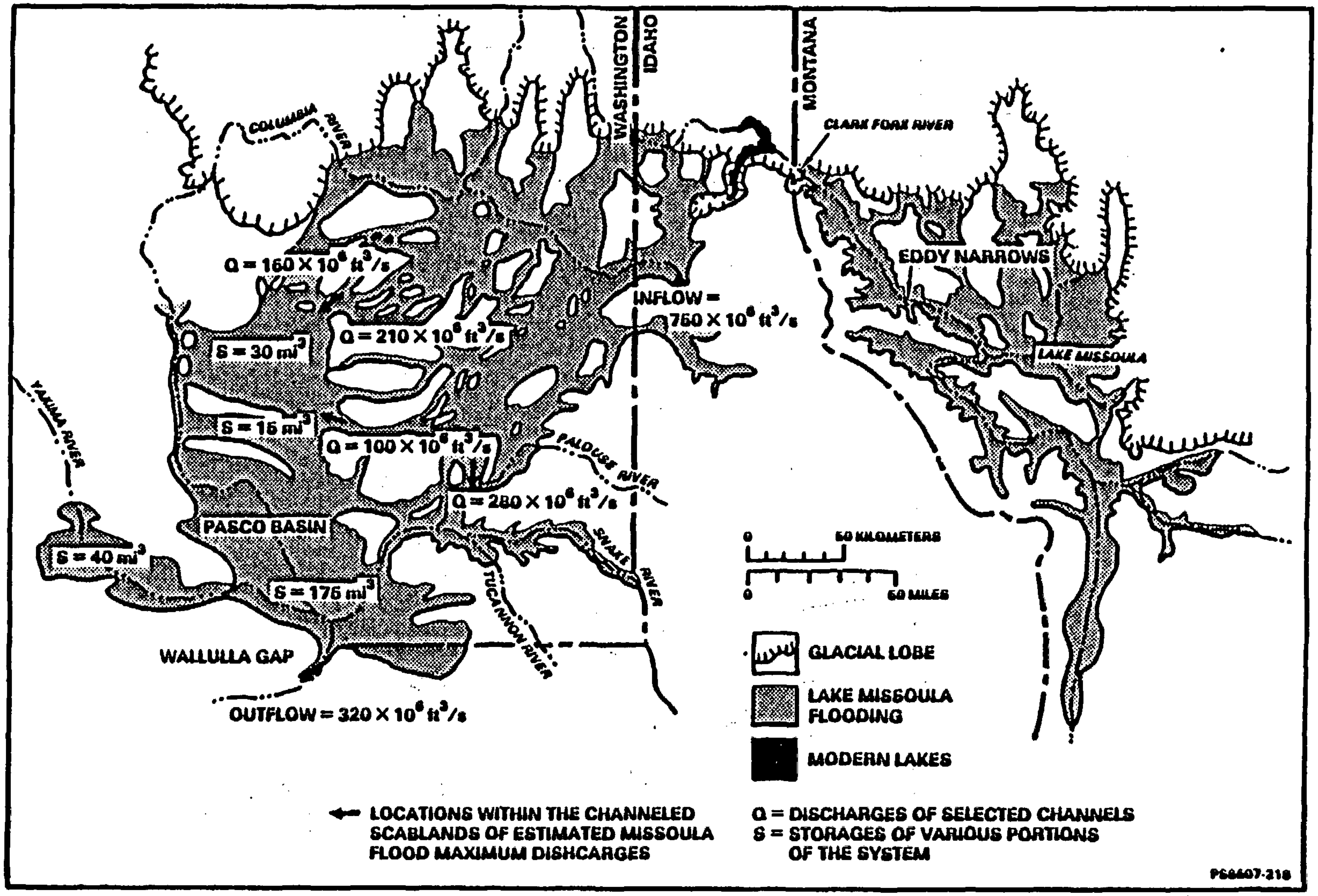


Figure 5.2-63. Regional relationship of glacial features of concern in this study (V. R. Baker, 1973, p. 21, Fig. 12). PS8607-218

Table 5.1-1. Monthly and annual prevailing wind directions, average speeds, and peak gusts at the Hanford Meteorology Station

Month	Wind (m/s) (mi/h in parentheses)								
	1945 – 1980 Averages						Peak gusts		
	Prevailing direction	Mean monthly speeds	Highest		Lowest				
			Monthly	Year	Monthly	Year	Speed	Direction	Year
January	NW	2.9 (6.4)	4.6 (10.3)	1972	1.4 (3.1)	1955	36 (80)	SW	1972
February	NW	3.2 (7.1)	4.5 (10.8)	1976	2.1 (4.6)	1963	29 (65)	SW	1971
March	WNW	3.8 (8.5)	4.8 (10.7)	1977[a]	2.6 (5.9)	1958	31 (70)	SW	1956
April	WNW	4.0 (9.0)	5.0 (11.1)	1972[a]	3.3 (7.4)	1958	33 (73)	SSW	1972
May	WNW	4.0 (8.9)	4.7 (10.5)	1965[a]	2.6 (5.8)	1957	32 (71)	SSW	1948
June	WNW	4.1 (9.2)	4.8 (10.7)	1949	3.0 (7.7)	1950[a]	32 (72)	SW	1957
July	WNW	3.9 (8.7)	4.3 (9.6)	1963	3.0 (6.8)	1955	31 (69)	WSW	1979
August	WNW	3.6 (8.0)	4.1 (9.1)	1946	2.7 (6.0)	1956	30 (66)	SW	1961
September	WNW	3.4 (7.5)	4.1 (9.2)	1961	2.4 (5.4)	1957	29 (65)	SSW	1953
October	WNW	3.0 (6.6)	4.1 (9.1)	1946	2.0 (4.4)	1952	28 (63)	SSW	1950
November	NW	2.7 (6.1)	3.7 (8.2)	1977	1.3 (2.9)	1956	29 (64)	SSW	1949
December	NW	2.7 (6.1)	3.7 (8.3)	1968	1.7 (3.0)	1963[a]	32 (71)	SW	1955
Year	WNW	3.4 (7.7)	3.7 (11.1)	April 1972[a]	2.8 (2.9)	November 1956	36 (80)	SW	January 1972

NOTE: Values presented are based on measurements made at 15.2 m (50 ft) above the ground (1945 to 1980).

[a]Also occurred in an earlier year.

PST87-2005-5.8-1

Table 5.1-2. Average wind speed, peak gust, and 100-year peak gust for selected levels from the Hanford Meteorology Station

Level	Average wind speed	Peak gust	100-year peak gust
15.2 m (50 ft)	3.5 m/s (7.8 mi/h)	36 m/s (80 mi/h)	38 m/s (86 mi/h)
61 m (200 ft)	4.7 m/s (10.5 mi/h)	40 m/s (89 mi/h)	42 m/s (94 mi/h)
122 m (400 ft)	5.1 m/s (11.4 mi/h)	42 m/s (93 mi/h)	46 m/s (103 mi/h)

Table 5.1-3. Averages and extremes of temperature at the Hanford Meteorology Station, 1912 to 1980

Month	Temperature (°C) (°F in parentheses)														
	1912 – 1980 Averages							1912 – 1980 Extremes							
								Daily maximum				Daily minimum			
	Daily maximum	Daily minimum	Mean monthly	Highest		Lowest		Highest		Lowest		Highest		Lowest	
				Monthly	Year	Monthly	Year	Recorded	Year	Recorded	Year	Recorded	Year	Recorded	Year
January	2.7 (36.9)	−5.5 (22.1)	−1.4 (29.5)	5.8 (42.5)	1953	−11.1 (12.1)	1950	22.2 (72)	1971	−18.9 (−2)	1950	11.7 (53)	1971	−30.6 (−23)	1934
February	7.7 (45.8)	−2.6 (27.3)	2.6 (36.6)	6.9 (44.5)	1958	−5.9 (21.4)	1929	21.7 (71)	1924	−19.4 (−3)	1950	12.8 (55)	1932	−30.6 (−23)	1950
March	13.5 (56.3)	0.8 (33.4)	7.1 (44.8)	9.9 (49.8)	1926	4.1 (39.4)	1955	28.3 (83)	1960	−4.4 (24)	1960	12.2 (54)	1942	−14.4 (6)	1955
April	18.8 (65.9)	4.2 (39.6)	11.6 (52.8)	15.3 (59.6)	1934	8.6 (47.5)	1955	35.00 (95)	1934	5.0 (41)	1945	15.6 (60)	1956	−11.1 (12)	1935
May	24.2 (75.6)	8.8 (47.9)	16.6 (61.8)	20.4 (68.8)	1947	14.8 (56.6)	1933	39.4 (103)	1936[a]	9.4 (49)	1918	21.1 (70)	1956	−2.2 (28)	1954
June	28.4 (83.2)	12.9 (55.3)	20.7 (69.2)	24.1 (75.4)	1922	16.2 (63.0)	1953	43.3 (110)	1912	12.8 (55)	1966	27.2 (81)	1924	0.6 (33)	1933
July	33.3 (91.9)	16.2 (61.1)	24.7 (76.5)	27.7 (81.8)	1960	22.4 (72.4)	1963	46.1 (115)	1939	15.0 (59)	1966	27.8 (82)	1925	3.9 (39)	1979
August	31.9 (89.5)	15.2 (59.4)	23.6 (74.4)	27.5 (81.5)	1967	21.0 (69.8)	1964	45.0 (113)	1961	17.2 (63)	1920	27.2 (81)	1961[a]	4.4 (40)	1918
September	26.6 (79.9)	10.6 (51.1)	18.6 (65.5)	22.1 (71.7)	1967	14.9 (58.8)	1926	38.9 (102)	1976[a]	11.1 (52)	1934	22.2 (72)	1955	−3.9 (25)	1926
October	18.5 (65.3)	4.8 (40.6)	11.7 (53.0)	15.0 (59.0)	1952	9.3 (48.8)	1930	32.2 (90)	1933	−0.6 (31)	1935	15.6 (60)	1945[a]	14.4 (6)	1935
November	9.1 (48.4)	−0.3 (31.4)	4.4 (39.9)	7.8 (46.0)	1954	−0.4 (31.3)	1955	23.9 (75)	1975	−10.0 (14)	1955	11.1 (52)	1959[a]	−18.3 (−1)	1955
December	4.2 (39.5)	−3.3 (26.1)	0.4 (32.8)	3.6 (38.5)	1957	−7.5 (18.5)	1919	20.6 (69)	1980	−19.4 (−3)	1919	13.3 (56)	1975	−32.8 (−27)	1919
	Mean daily maximum	Mean daily minimum	Mean annual												
Year	18.2 (64.8)	5.2 (41.3)	11.7 (53.1)												

[a]Also occurred in an earlier year.

PST87-2005-5.8-2

Table 5.1-4. Facility design and planning engineering weather data for the Hanford Meteorology Station. Temperatures are presented using the Fahrenheit scale that is traditionally used to present these values (Latitude: 46° 34'N; Longitude: 119° 36'W; Elevation: 223 m (733 ft) above sea level)

Three coldest months (December, January, and February)				
Temperature °F (°C)		Prevailing wind direction	Average speed (m/h)	Annual heating degree days
99%[a]	97.5%[b]			
1 (-17)	8 (-13)	West-northwest	6.5	5,299

Four warmest months (June, July, August and September).								
Temperature (dry bulb) and mean coincident wet bulb (MCWB)								
1%[c]	MCWB	2.5%[d]	MCWB	Mean daily range	Prevailing wind direction	5%[e]	MCWB	
101 (38)	68 (20)	98 (37)	67 (19)	30 (18)	Northwest	95 (35)	66 (19)	

Wet bulb °F (°C)			Dry bulb temperature °F (°C)		Wet bulb temperature °F (°C)	
1%[a]	2.5%[b]	5%[e]	Hours ≥93 (34)	Hours ≥80 (27)	Hours ≥73 (23)	Hours ≥67 (19)
69 (21)	69 (21)	66 (19)	189	813	813	813

[a]Exceeded 99% of the time.
[b]Exceeded 97.5% of the time.
[c]Exceeded 1% of the time.
[d]Exceeded 2.5% of the time.
[e]Exceeded 5% of the time.

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Table 5.1-5. Averages and extremes of precipitation at the Hanford Meteorology Station, 1912 to 1980

Month	Precipitation (mm) (in. in parentheses)													
	Liquid water content of all precipitation[a]							Snow, ice pellets (sleet and hail)[b]						
	Mean monthly	Maximum		Minimum		Maximum		Mean monthly	Maximum		Maximum		Maximum	
		Monthly	Year	Monthy	Year	In 24 h	Year		Monthly	Year	In 24 h	Year	Depth	Year
January	23.4 (0.92)	62.7 (2.47)	1970	2.0 (0.08)	1977	27.4 (1.08)	1948	132 (5.2)	594 (23.4)	1950	180 (7.1)	1954	305 (12.0)	1969
February	15.2 (0.60)	78.2 (3.08)	1940	c	1967	31.5 (1.24)	1916	58 (2.3)	660 (26.0)	1916	457 (18.0)	1916	622 (24.5)	1916
March	9.4 (0.37)	47.2 (1.86)	1957	0	1942[b]	15.0 (0.59)	1949	8 (0.3)	107 (4.2)	1951	56 (2.2)	1957	58 (2.3)	1957
April	9.9 (0.39)	31.0 (1.22)	1969	0	1933[b]	14.7 (0.58)	1980	c	c	1968[b]	c	1968[d]	0	–
May	12.2 (0.48)	51.6 (2.03)	1972	0	1931	35.3 (1.39)	1972	c	c	1960	c	1960	0	–
June	13.7 (0.54)	74.2 (2.92)	1950	0	1919	38.1 (1.50)	1934	0	0	–	0	–	0	–
July	3.8 (0.15)	20.6 (0.81)	1966	0	1939[b]	31.7 (1.25)	1942	0	0	–	0	–	0	–
August	6.1 (0.24)	34.5 (1.36)	1977	0	1955[b]	22.6 (0.89)	1977	0	0	–	0	–	0	–
September	7.9 (0.31)	34.0 (1.34)	1947	c	1976[b]	20.8 (0.82)	1947	0	0	–	0	–	0	–
October	14.5 (0.57)	69.1 (2.72)	1957	0	1917[b]	48.5 (1.91)	1957	c	38 (1.5)	1973	38 (1.5)	1973	38 (1.5)	1973
November	21.8 (0.86)	77.5 (3.05)	1926	c	1976[b]	19.8 (0.78)	1966	36 (1.4)	323 (12.7)	1955	211 (8.3)	1978	231 (9.1)	1978
December	22.9 (0.90)	64.3 (2.53)	1931	2.8 (0.11)	1976[b]	25.4 (1.00)	1958	102 (4.0)	485.1 (19.1)	1964	137 (5.4)	1965	307 (12.1)	1964

[a]Average total annual preciptitation--160.8 mm (6.33 in.).
[b]Average total annual snowfall--33.6 mm (13.2 in.).
[c]Trace amount.
[d]Also occurred in an earlier year.

PST87-2005-5.0-3

Table 5.1-6. Frequency of occurrence of mixing heights at the Hanford Site within various ranges

Mixing height (m)	Spring	Summer	Fall	Winter
<50	0.03	0.01	0.07	0.08
50 - 100	0.06	0.05	0.08	0.11
100 - 200	0.17	0.15	0.23	0.38
200 - 400	0.25	0.19	0.28	0.29
400 - 600	0.08	0.09	0.10	0.05
600 - 1,000	0.18	0:25	0.15	0.03
>1,000	0.24	0.26	0.10	0.05

NOTE: Values presented are based on 4 yr of hourly mixing depth observations (March 1982 - March 1986). Hourly values are estimated by the duty meteorologist at the Hanford Meteorology Station based on measurements from the station meteorological tower and acoustic sounder, pibal observations, and regional radiosonde data.

Table 5.1-7. Annual joint wind speed, direction, and stability frequencies for the Hanford Meteorology Station (sheet 1 of 2)

Stability category	Wind speed (m/s)[a]	Wind direction (%)																		
		N	NNE	NE	ENE	E	ESE	SE	SSE	S	SSW	SW	WSW	W	WNW	NW	NNW	VAR	Calm	Total
Extremely unstable	Calm	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	0.01	0.01
	0.5–1.5	0.15	0.19	0.19	0.09	0.10	0.08	0.09	0.04	0.04	0.02	0.04	0.03	0.03	0.03	0.07	0.09	0.15		1.43
	2.0–3.0	0.47	0.47	0.40	0.24	0.23	0.21	0.24	0.15	0.18	0.16	0.17	0.11	0.11	0.14	0.35	0.38	0.14		4.15
	3.5–5.5	0.18	0.22	0.16	0.05	0.03	0.03	0.05	0.04	0.05	0.10	0.19	0.14	0.07	0.17	0.43	0.17	0.00		2.08
	6.0–8.0	0.06	0.12	0.07	0.01	0.01	0.00	0.00	0.01	0.01	0.04	0.17	0.18	0.07	0.21	0.35	0.04	0.00		1.36
	8.5–10.5	0.01	0.03	0.02	0.01	0.00	0.00	0.00	0.00	0.00	0.03	0.10	0.12	0.04	0.13	0.30	0.01	0.00		0.80
	11.0–up	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.12	0.10	0.03	0.08	0.22	0.00	0.00		0.59
	Total	0.87	1.03	0.84	0.40	0.37	0.32	0.39	0.24	0.29	0.37	0.79	0.68	0.36	0.76	1.72	0.69	0.29	0.01	10.42
Moderately unstable	Calm	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	0.00	0.00
	0.5–1.5	0.05	0.06	0.08	0.03	0.05	0.04	0.05	0.02	0.02	0.02	0.02	0.01	0.01	0.01	0.03	0.03	0.05		0.58
	2.0–3.0	0.12	0.09	0.11	0.08	0.07	0.07	0.08	0.04	0.06	0.05	0.07	0.04	0.05	0.05	0.12	0.11	0.03		1.24
	3.5–5.5	0.06	0.08	0.04	0.02	0.02	0.01	0.03	0.01	0.02	0.04	0.06	0.07	0.04	0.06	0.15	0.06	0.00		0.76
	6.0–8.0	0.01	0.03	0.02	0.01	0.00	0.00	0.00	0.01	0.01	0.02	0.07	0.08	0.03	0.08	0.12	0.01	0.00		0.51
	8.5–10.5	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.04	0.05	0.01	0.05	0.10	0.00	0.00		0.30
	11.0–up	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.05	0.04	0.01	0.02	0.07	0.00	0.00		0.22
	Total	0.24	0.28	0.26	0.14	0.14	0.13	0.16	0.09	0.10	0.16	0.29	0.29	0.16	0.28	0.59	0.22	0.08	0.00	3.61
Slightly unstable	Calm	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	0.01	0.01
	0.5–1.5	0.07	0.05	0.08	0.05	0.07	0.05	0.05	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.03	0.03	0.05		0.68
	2.0–3.0	0.12	0.11	0.10	0.06	0.08	0.07	0.09	0.05	0.05	0.06	0.06	0.05	0.05	0.06	0.11	0.13	0.03		1.26
	3.5–5.5	0.06	0.05	0.05	0.02	0.02	0.01	0.02	0.02	0.02	0.05	0.08	0.08	0.05	0.07	0.17	0.06	0.00		0.83
	6.0–8.0	0.02	0.03	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.03	0.07	0.10	0.04	0.10	0.13	0.01	0.00		0.59
	8.5–10.5	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.05	0.06	0.01	0.05	0.08	0.00	0.00		0.29
	11.0–up	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.07	0.05	0.01	0.03	0.08	0.00	0.00		0.29
	Total	0.28	0.25	0.24	0.14	0.16	0.14	0.17	0.11	0.10	0.20	0.35	0.36	0.19	0.33	0.61	0.24	0.08	0.01	3.96
Neutral	Calm	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	0.66	0.66
	0.5–1.5	0.71	0.49	0.65	0.46	0.55	0.60	0.60	0.28	0.26	0.19	0.21	0.16	0.29	0.31	0.52	0.64	0.34		7.25
	2.0–3.0	0.67	0.43	0.47	0.30	0.37	0.42	0.56	0.31	0.30	0.27	0.36	0.31	0.41	0.60	1.14	0.82	0.10		7.82
	3.5–5.5	0.24	0.24	0.16	0.08	0.08	0.10	0.16	0.14	0.14	0.27	0.45	0.49	0.48	0.88	1.48	0.40	0.00		5.77
	6.0–8.0	0.12	0.17	0.07	0.02	0.01	0.01	0.05	0.07	0.12	0.25	0.53	0.71	0.52	1.06	1.12	0.11	0.00		4.95
	8.5–10.5	0.04	0.05	0.05	0.01	0.00	0.00	0.01	0.03	0.07	0.20	0.47	0.49	0.21	0.79	0.81	0.02	0.00		3.26
	11.0–up	0.01	0.02	0.02	0.00	0.00	0.00	0.00	0.01	0.06	0.24	0.54	0.30	0.08	0.52	0.76	0.00	0.00		2.56
	Total	1.79	1.40	1.41	0.89	1.02	1.13	1.38	0.83	0.95	1.42	2.56	2.45	1.98	4.15	5.83	2.00	0.44	0.66	32.28

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Table 5.1-7. Annual joint wind speed, direction, and stability frequencies for the Hanford Meteorology Station (sheet 2 of 2)

Stability category	Wind speed (m/s)[a]	Wind direction (%)																		
		N	NNE	NE	ENE	E	ESE	SE	SSE	S	SSW	SW	WSW	W	WNW	NW	NNW	VAR	Calm	Total
Slightly stable	Calm	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	0.55	0.55
	0.5–1.5	0.36	0.23	0.24	0.20	0.25	0.33	0.43	0.25	0.23	0.16	0.19	0.18	0.32	0.31	0.43	0.36	0.18	|	4.67
	2.0–3.0	0.44	0.23	0.18	0.15	0.18	0.25	0.38	0.25	0.23	0.21	0.28	0.38	0.73	0.85	0.98	0.68	0.00	|	6.40
	3.5–5.5	0.23	0.15	0.13	0.09	0.10	0.12	0.21	0.22	0.13	0.18	0.33	0.65	1.24	2.18	1.74	0.49	0.00	|	8.19
	6.0–8.0	0.10	0.09	0.05	0.02	0.04	0.02	0.07	0.12	0.10	0.17	0.42	0.73	1.06	2.89	1.85	0.18	0.00	|	7.91
	8.5–10.5	0.01	0.02	0.02	0.01	0.00	0.00	0.01	0.04	0.07	0.13	0.35	0.36	0.20	1.07	1.03	0.02	0.00	|	3.37
	11.0–up	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.05	0.18	0.26	0.13	0.04	0.28	0.38	0.00	0.00	|	1.33
	Total	1.14	0.73	0.63	0.48	0.56	0.72	1.10	0.90	0.81	1.03	1.83	2.44	3.58	7.58	6.41	1.72	0.21	0.55	32.42
Moderately stable	Calm	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	0.20	0.20
	0.5–1.5	0.20	0.10	0.10	0.08	0.09	0.13	0.21	0.11	0.12	0.10	0.11	0.10	0.17	0.16	0.20	0.19	0.11	|	2.28
	2.0–3.0	0.18	0.09	0.07	0.06	0.06	0.09	0.18	0.14	0.14	0.13	0.17	0.22	0.44	0.51	0.54	0.38	0.00	|	3.39
	3.5–5.5	0.07	0.04	0.03	0.01	0.01	0.02	0.09	0.11	0.06	0.06	0.11	0.30	0.62	0.93	1.17	0.34	0.00	|	3.95
	6.0–8.0	0.02	0.03	0.01	0.01	0.01	0.00	0.04	0.08	0.02	0.02	0.05	0.20	0.32	0.86	1.25	0.11	0.00	|	3.03
	8.5–10.5	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.04	0.03	0.09	0.21	0.00	0.00	|	0.44
	11.0–up	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.01	0.00	0.00	|	0.05
	Total	0.47	0.26	0.21	0.15	0.17	0.24	0.52	0.46	0.36	0.32	0.47	0.86	1.59	2.55	3.38	1.02	0.12	0.20	13.35
Extremely stable	Calm	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	0.06	0.06
	0.5–1.5	0.09	0.06	0.11	0.03	0.07	0.07	0.12	0.07	0.08	0.07	0.07	0.07	0.10	0.05	0.11	0.13	0.11	|	0.41
	2.0–3.0	0.11	0.07	0.07	0.05	0.06	0.05	0.08	0.09	0.10	0.06	0.08	0.12	0.18	0.16	0.16	0.18	0.03	|	1.64
	3.5–5.5	0.02	0.01	0.01	0.01	0.01	0.01	0.03	0.04	0.03	0.04	0.04	0.07	0.13	0.16	0.31	0.19	0.00	|	1.10
	6.0–8.0	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.02	0.02	0.01	0.03	0.07	0.14	0.32	0.05	0.00	|	0.67
	8.5–10.5	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.03	0.00	0.00	|	0.08
	11.0–up	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	|	0.01
	Total	0.22	0.14	0.19	0.09	0.14	0.13	0.25	0.21	0.24	0.20	0.22	0.29	0.50	0.52	0.92	0.54	0.14	0.06	4.99

NOTE: Values are based on 15 yr of wind measurements (1955 to 1970) at 61 m (200 ft) above the ground and Pasquill-Gifford delta-temperature stability categories.

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[a]To convert m/s to mi/h, multiply by 2.24.

Table 5.2-1. The climate stations used in the dendroclimatic analyses

Map symbol*	Site	Used in final model
a	Aberdeen, Washington	
b	Baker, Oregon	T, P
c	Kalispell, Montana	
d	Kamloops, British Columbia	
e	Roseburg, Oregon	
f	Spokane, Washington	T
g	Vancouver, British Columbia	
h	Walla Walla, Washington	T
i	Colville, Washington	P
j	Lewiston, Idaho	P
k	Port Angeles, Washington	
l	Porthill, Washington	
m	Waterville, Washington	

Source: Cropper and Fritts, 1986.

NOTE: All stations listed were used in preliminary models. Stations used in final models are indicated by a T if they were used for temperature reconstruction or a P if they they were used for precipitation reconstruction.

*See Figure 5.2-2.

Table 5.2-2. Statistical comparison of the 20th century record of climate in the Pasco Basin with the reconstructed record for 1602 through 1900

Variable	Instrumental record 1901-1970	Reconstructed record 1602-1900	Instrumental minus reconstructed record
Mean temperature, °C			
Annual	9.71	9.81	-0.10
Winter	-0.33	0.01	-0.34
Spring	11.23	11.41	-0.18
Summer	21.28	21.25	0.03
Autumn	10.04	9.86	0.18
Mean precipitation, cm			
Annual	34.87	35.69	-0.82
Winter	10.53	10.70	-0.17
Spring	12.68	12.30	0.38
Summer	2.99	2.71	0.29
Autumn	8.67	9.98	-1.31
Standard deviation temperature, °C			
Annual	0.41	0.43	0.53
Winter	1.06	0.98	0.60
Spring	0.65	0.30	0.45
Summer	0.58	0.67	0.48
Autumn	0.24	0.33	0.40
Standard deviation precipitation, cm			
Annual	2.95	3.80	1.97
Winter	0.91	1.13	2.05
Spring	2.08	2.24	2.35
Summer	0.83	0.76	2.77
Autumn	1.63	2.63	1.58

NOTE: Right column in lower half of table gives the standard deviations of the differences in the means of the two time periods rather than the differences in the standard deviations. That is, the differences between each pair of values is computed, then the means and standard deviations of the difference data are computed.

Table 5.2-3. Cross-referenced list of scientific and common names of taxa referred to in text, with annotations (sheet 1 of 3)

Common and scientific names[a]		Notes
alder	Alnus	Most species of this genus are disturbance adapted, likely to have been important on recently deglaciated landscape, or when forest fire frequency is high.
Alnus	alder	
Artemisia	sagebrush/wormwood	
Betula	birch	
birch	Betula	In this region there are two: Betula occidentalis (paper birch) and Betula papyrifera (paper birch).
bracken fern	Pteridium aquilinum	A plant typically abundant in pollen spectra of the early Holocene of the maritime Pacific Northwest.
Brasenia	water-shield	
buffalo-berry	Shepherdia canadensis-type	Believed to have been a seral species on nutrient poor, recently-deglaciated soils.
Bunchgrass		In this region, cheifly the genus Agropyron.
cedar		One of a number of tree genera in the family Cupressaceae. Their pollen is normally indistinguishable, but is occasionally identified to species on ecologic grounds.
Chenopodiineae	saltbush	
Cyperaceae	Sedge	The sedge family.
diploxylon	Pinus	Subgen. Diploxylon. Yellow pine. In this region, the most widespread yellow pines are the ponderosa pine and the lodgepole pine.
ditch-grass	Ruppia	The genus Ruppia; an aquatic taxon.
Douglas-fir	Pseudotsuga menziesii	A conifer intermediate in its moisture requirements between mesophytic species such as western red cedar and western hemlock, and xerophytes such as ponderosa pine and western juniper.
Douglas-fir/larch		An arboreal pollen taxon, either Douglas-fir or larch.
Engelmann spruce	Picea engelmannii	
fescue	Festuca	A perennial or annual grass.
fir	Abies	Any of the true firs. Trees normally grow in habitats with greater average annual precipitation than those in which Douglas-fir grow.
goosefoot		See saltbush.
grand fir	Abies grandis	
Gramineae		The grass family.
grass	Gramineae (or Poaceae)	
haploxylon pine	Pinus	Subgen. Haploxylon. White pine. In this region, the most widespread white pines are western white pine and whitebark pine.

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Table 5.2-3. Cross-referenced list of scientific and common names of taxa referred to in text, with annotations (sheet 2 of 3)

Common and scientific names[a]		Notes
hawkweed	Hieracium	
juniper	Juniperus	In this region, there are two species likely to have been present, Juniperus occidentalis, a tree of the dry hills and sand dunes of the area, and J. communis, a shrub of montane and subalpine habitats.
Latrix	larch	
lodgepole pine	Pinus contorta	A disturbance adapted, seral species; likely to have been important on the recently deglaciated landscape and when forest fire frequency was high.
mallow ninebark	Physocarpus malvaceus	A shrub.
mountain lover	Pachistima myrsenites	A shrub.
Nuphar	pond lily	
oak	Quercus	In the vicinity of the Columbia Basin, Oregon white oak (Quercus garryana) is the only species. It is a relative thermophile compared to almost all other arboreal taxa.
ocean spray	Holodiscus discolor	A shrub.
Pinus	pine	
pond lily	Nuphar	An aquatic macrophyte.
ponderosa pine	Pinus ponderosa	The most thermophilous of the pine species in the region.
Pseudotsuga	Douglas-fir	
Pteridium	bracken ferm	
Quercus	oak	
Ruppia	ditch-grass	
sagebrush	Artemisia	There exist both shrub species of the genus Artemisia and herbaceous species, the former (sagebrushes) typify steppe regions, while the latter are tundra or subalpine taxa.
Salix	willow	
saltbush	Chenopodiineae	Chenopodiineae pollen attributed primarily to xerophytic shrubs such as Artiplex and Grayia, unless otherwise specified.
saltbush/goosefoot	Chenopodiineae	Chenopodiineae pollen attributable to either steppe shrubs such as Atriplex, Ceratoides, and Grayia, but also herbaceous, disturbance-adapted plants such as non-woody species of Atriplex, Chenopodium, and the tumble-weed Salsola.
sedge	Cyperacea	Graminoids; the sedges normally thrive in swampy situations.
Shepherdia canadensis-type	buffalo-berry	

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Table 5.2-3. Cross-referenced list of scientific and common names of taxa referred to in text, with annotations (sheet 3 of 3)

Common and scientific names[a]		Notes
snowberry	Symphoricarpos albus	A shrub.
spruce	Picea	In this region there are two interior species likely to have occupied the glacial-age landscape, engelmann spruce (Picea engelmannii) or white spruce (P. glauca).
Tsuga heterophylla	western hemlock	
water shield	Brasenia schrebneri	An aquatic macrophyte.
western hemlock	Tsuga heterophylla	
western larch	Larix occidentalis	
western red-cedar	Thuja plicata	More typical of maritime coastal forests,but also occurring in the mountains of northern Washington.
western white pine	Pinus monticola	A pine common in the coniferous forests of the northern and eastern periphery of the Columbia Basin.
white bark pine	Pinus albicaulis	A subalpine species, adapted to extreme cold and effectively arid soils.
willow	Salix	In this region, commonly a shrub, and commonly a riparian species, growing on moist to wet soils, or close to water courses.
wormwood	Artemisia	Herbaceous species.

[a]Nomenclature follows Hitchcock and Cronquist (1973).

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Table 5.2-4. Eastern Washington tephra layers found in sediment cores

Tephra unit	Approximate age (yr)
Mount St. Helens Set W	500
Mount St. Helens Set Y	3,350 to 3,510
Mount Mazama Layer O	6,700 to 7,000
Glacier Peak Layer B	11,200
Glacier Peak Layer G	11,750

Table 5.2-5. Other paleoenvironmetal records from the interior Pacific Northwest (sheet 1 of 2)

Author	Title	Reason these records were not used
Bartholomew (1982)	Pollen and Sediment Analysis of Clear Lake, Whitman County, Washington: The Last 600 Years	a
Bright and Davis (1982)	Quaternary Paleoecology of Idaho National Engineering Laboratory, Snake River Plain, Idaho	b
Davis (1984a)	Multiple Thermal Maxima During the Holocene	b
Davis et al. (1977)	Pollen Analysis of Wildcat Lake, Whitman County, Washington: The Last 1,000 Years	a
Freeman (1929)	Evidence of Prolonged Droughts on the Columbia Plateau Prior to White Settlement	a, c
Hansen (1939a)	Paleoecology of a Central Washington Bog	c
Hansen (1939b)	Pollen Analysis of a Bog near Spokane, Washington	c
Hansen (1939c)	Pollen Analysis of a Bog in Northern Idaho	c
Hansen (1941a)	A Pollen Study of Post-Pleistocene Lake Sediments in the Upper Sonoran Life Zone of Washington	c
Hansen (1941b)	Paleoecology of a Montane Peat Deposit near Wenatchee, Washington	c
Hansen (1943a)	Paleoecology of a Peat Deposit in East Central Washington	c
Hansen (1943b)	A Pollen Study of a Subalpine Bog in the Blue Mountains of Northeastern Oregon	c
Hansen (1944)	Postglacial Vegetation of Eastern Washington	c
Hansen (1946)	Postglacial Forest Succession and Climate in the Oregon Cascades	b, c

Table 5.2-5. Other paleoenvironmetal records from the interior Pacific Northwest (sheet 2 of 2)

Author	Title	Reason these records were not used
Hansen (1947)	Postglacial Forest Succession, Climate, and Chronology in the Pacific Northwest	c
Henry (1984)	Holocene Paleoecology of the Western Snake River Plain	b
Mack and Bryant (1974)	Modern Pollen Spectra from the Columbia Basin, Washington	--
Mack et al. (1976)	Pollen Sequence from the Columbia Basin, Washington: Reappraisal of Postglacial Vegetation	a
Mack et al. (1978b)	Reexamination of Postglacial Vegetation History in Northern Idaho, Hager Pond, Bonner County	b
Mack et al. (1978c)	Late Quaternary Pollen Record from the Sanpoil River Valley, Washington	a
Mack et al. (1983)	Holocene Vegetational History of the Kootenai River Valley, Montana	b
Mehringer et al. (1977)	Postglacial History of Lost Trail Pass Bog, Bitterroot Mountains, Montana	b
Nickman (1979)	The Palynology of Williams Lake Fen, Spokane County, Washington	a
Nickman and Leopold (1984)	A Postglacial Pollen Record from Goose Lake, Okanogan County, Washington: Evidence for an Early Holocene Cooling	a
Smith (1983)	A 4,300-Year History of Vegetation, Climate, and Fire from Blue Lake, Nez Perce County, Idaho	a

a = fossil record less than 10,000 yr long or accompanied by a hiatus that exceeds 2,000 yr.
b = the site is outside the limits of the Columbia Basin.
c = no or inadequate time control of the stratigraphic section.

Table 5.2-6. Pollen-stratigraphic records from the Columbia Basin and vicinity

Area	Site	Duration of record (yr)	References
Selkirk Range, northeastern Washingon	Big Meadow	~12,500	Mack et al., 1978a
	Waits Lake	~13,500	Mack et al., 1978d
Okanogan Valley, central-northern Washington	Bonaparte Meadows	~10,500	Mack et al., 1979
	Mud Lake	~11,500	Mack et al., 1979
Eastern Columbia Basin	Creston	~13,000	Mack et al., 1976
	Williams Lake	~13,000	Nickmann, 1979
	Wildcat Lake	~10,600	Mehringer, 1985b
Southwestern Columbia Basin	Carp Lake	~33,000	Barnosky, 1983, 1984, 1985

Table 5.2-7. Depth and extrapolated ages of major climatic cycles and their terminations in core K708-7 from the northeast Atlantic Ocean

Major climatic cycles (m)	Termination depths	Termination numbers and extrapolated ages (yr B.P.)	Alternate termination ages[a]
A	45	(I) 13,500	11,000
B(1-2)[b]	310	(II) 127,000	127,000
C(3-5)[b]	560	(III) 225,000	225,000
D(6-7)[b]	740	(IV) 311,000	300,000
E	920	(V) 384,000	380,000
F	1,090	(VI) 457,000	--
G	1,220	(VII) 513,000	--
H	1,430	(VIII) 603,000	--

[a]From Broecker and van Donk (1971).
[b]Numbers in parentheses are the equivalent carbonate minimums from McIntyre et al. (1972).

Table 5.2-8. Characteristics and values of the constants of integration for the obliquity, precession, and eccentricity of the earth's orbit, calculated for seven different solutions

	Accuracy[a]	Epoch	Obliquity (°)	Precession ("/yr)	Precession constant (°)	Maximum eccentricity
M	1 1 2	1800	23.306263	50.471698	4.105013	0.063865
B1	1 1 2	1850	23.318130	50.446713	3.448476	0.063845
B2	1 3 2	1850	23.316059	50.444577	3.472760	0.07446
SH	2 1 2	1950	23.317083	50.450194	1.964146	0.067414
B3	2 3 2	1850	23.320548	50.439287	3.392528	0.072864
B4	2.3.2 R	1850	23.319756	50.440672	3.334857	0.073917
B5[b]	2.3.2.R	1850	23.319670	50.440822	3.328585	0.073917

Sources: Taken from Berger, 1984, Table 1, p. 14.
M - LeVerrier (1855)
B1,B4,B5 - Bretagnon (1984)
B2,B3 - Berger (1978b)
SH - Sharaf and Budnikora (1967)

NOTE: The labeling of the solutions corresponds to the authors from which the developments originate. All further computations are following the model developed in Berger (1978c).

[a]The three numbers given to characterize the accuracy are related respectively to the order to the masses, the degree with respect to e's and i's, and the degree with respect to eccentricity of the earth's orbit in the expansion of obliquity and precession. R means that some relativity terms are included.

[b]Same as B4 but where the mean rates have been changed according to expected accuracy from the Bretagon (1984) and Berger (1978b) studies.

Table 5.2-9. Listing of active prognostic mesoscale models

Model formalism	Reference	Computer	Grid size	Time
Hydrostatic PE, 3-D[a]	Jones, 1980	CDC Cyber 750	2700 x 2700 dx = 10 km	24-h 12 CPU-h
Hydrostatic incompressible PE, 1-, 2- or 3-D[b]	Brown et al., 1982	NCAR CRAY-1	12 x 13 x 29	48-h 24 CPU-m dt = 6
Hydrostatic incompressible PE, 2- and 3-D[c].	Pielke, 1974	NCAR CRAY	30 x 36 x 13 dx = 10 km	12-h 24 CPU-m dt = 90s
Hydrostatic incompressible PE, dry 2-D (x and sigma)[d]	Keyser and Anthes, 1982	NCR CRAY	--	36-h 2-m
Hydrostatic incompressible PE, 1-, 2-, and 3-D[e]	Mellor and Yamada, 1974; Yamada, 1978, 1981	CDC 7600	--	.004 CPU-s /point /dt
Hydrostatic PE, 2- and 3-D, nested, 2-way interactive grids[f]	Mathur, 1983	--	Tropical storm David	48-h 2 CPU-h
Hydrostatic PE, 2- and 3-D[g]	Anthes and Warner, 1978	CRAY 1A	50 x 50 x 10 dx = dy = 100 km dt = 200s	24-h 7-m
Hydrostatic PE, 3-D[h]	Boudra, 1981	CRAY-1	--	36-h 50 m
Hydrostatic: incompressible PE, compressible 3-D, lake breeze[i]	Ballentine, 1980, 1982	--	49 x 41 x 11 dx ~ 10 km leap-frog time diff. Dry Lake Michigan	12-h 27-h

Source: Pielke, 1984, Appendix B. PST87-2005-5.8-6

[a]Atlantic Oceanographic and Meteorological Laboratory/Hurricane Research Division.
[b]Center for the Environment and Man.
[c]Colorado State Mesoscale Model.
[d]Frontogenesis Modeling within the Mesoscale Analysis and Modeling Group of the Severe Storm Branch of the Goddard Laboratory for Atmosphere Sciences, NASA/Goddard Space Flight Center.
[e]Los Alamos National Laboratory.
[f]National Meteorological Center.
[g]NCAR - Penn State.
[h]University of Miami.
[i]University of Wisconsin, Milwaukee.