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Section A: DOCUMENTARY EVIDENCE

1 Climate since A.D. 1500: Introduction

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1.1 The need for perspective

It is now common knowledge that today's climate is unlikely to prevail into the 21st century. An increase in "greenhouse gases", as a result of human activity, is likely to perturb the world's energy balance, leading to higher surface temperatures and a redistribution of precipitation patterns (National Research Council 1982; Bolin *et al.* 1986; I.P.C.C., 1990). The magnitude of any climatic change, and its distribution both geographically and seasonally, is estimated by the use of computer models of the general circulation. Simulations of equilibrium climatic conditions with early twentieth century CO_2 levels are generally compared with those resulting from doubling of CO_2 to obtain the differences that might be expected in the future (Schlesinger 1984). More recently, transient climate models have been developed to simulate the gradually changing climate as CO_2 and the other greenhouse gases increase from year to year. The focus of all these experiments is, of course, to isolate the impact of human activities on climate over the next century or so. However, whatever the anthropogenic climatic effects may be in the future, they will be superimposed on a climatic system which also responds to "natural" forcing factors. Unless we improve our understanding of what these factors are, and how the climate system has responded to them in the past, there is little prospect of interpreting, or anticipating, future climatic changes. Therefore, in order to understand how climate may vary in the future we must understand how and why it has varied in the past. With such knowledge we may be able to place our contemporary climate in a longer term perspective and identify any underlying trends or periodicities in climate upon which future climatic changes might be superimposed. With such knowledge we may be able to isolate the causes of past climatic fluctuations, causes which may continue to operate in the future and influence the course of forthcoming climatic events (Bradley 1990).

In this book we focus on the most recent period of climate history, the last 500 years. This is an important interval for a number of reasons. Firstly, we can construct a fairly comprehensive picture of climatic variations during this period, and we can also document variations in potentially important forcing factors. There is thus the opportunity to develop and test hypotheses about how the climate system responds to these factors. Secondly, climatic variability on a decadal to century time-scale is of most relevance to concerns about future climate and the extent to which "natural" variability will amplify or subdue anthropogenic effects. Thirdly, in the last 500 years, world population has increased by a factor of 12, at least; our society has changed from one which produced local, or perhaps regional environmental impacts, to one which now produces environmental impacts of global extent. Understanding how human activity may already have altered climate locally, regionally and perhaps globally is of great significance as we enter a new century facing even more rapid increases in world population and continuing environmental degradation.

1.2 Climates of the past

Although the focus of this book is the last 500 years, it is appropriate to consider briefly how this period fits in with the longer term changes that we know have affected the earth's climate system. Twenty thousand years ago the world was experiencing a period of major continental glaciation, the most recent of a series of such events which have occurred with some regularity over the last 2 million years (Imbrie and Imbrie 1979). These events were brought about by small changes in the position of the earth relative to the sun and the consequent redistribution of solar radiation across the earth (Berger 1980). Changes in atmospheric composition and of atmospheric aerosol loading may also have played a role in the evolution of climate from glacial to interglacial periods and back again to glacial periods (Barnola *et al.* 1987; Genthon *et al.* 1987; Chappellaz *et al.* 1990). The last continental ice sheets in the Northern Hemisphere had largely disappeared by 6000-7000 years B.P. (Before Present) and this induced dramatic changes in the distribution of plants and animals and of the world's coastlines as the sea returned to higher levels.

There is evidence that some parts of the world experienced quite warm summers around 5000 to 6000 years ago, a period sometimes referred to, rather loosely, as the mid-Holocene Optimum (cf. Webb and Wigley 1985). Whether this was a globally extensive warm period, or if it was warm in other seasons is not yet known. However, it is thought that glaciers in many parts of the world reached their post-glacial minima around this time, a condition made all the more significant by the fact that glaciers subsequently expanded over the course of the next few thousand years. Such periods of glacier expansion are known collectively as Neoglacial episodes, times of renewed glacier activity (Porter and Denton 1967). It has been argued that such periods have a certain regularity in time, suggesting some periodic forcing, but the evidence is quite weak and the causes of these glacier advances, whether regional or global in extent, remains obscure. For our purposes, the important point is that the most recent of these neoglaciations occurred during the last 500 years, and there is abundant evidence that this most recent episode was the most significant of all the periods of glacier expansion that have occurred since the last Ice Age. The period since A.D. 1500 is thus of extraordinary scientific interest.

Although there is voluminous indirect evidence that climatic conditions in the past 500 years were often quite different from our contemporary experience (e.g. von Rudloff 1967; Lamb 1982; Grove 1988) the precise nature of these differences, and what caused them, remains elusive. Certainly, there were a number of cold intervals which had dramatic environmental consequences. In almost every glacierised mountain region of the world, glaciers grew and advanced down-valley, often to positions as extensive as at any time since the last Ice Age, more than 10,000 years ago. These changes in alpine glaciers are so characteristic of the period that it is often referred to as "the Little Ice Age". However, among those who use this term there is little consensus on when it began, or when it ended.

Furthermore, the term suggests a period of uniformly cold conditions which obscures the fact that relatively warm intervals did occur (see Chapter 33). It also focuses attention on those parts of the world where snow and ice are common phenomena; how climate changed in tropical and sub-tropical regions during the last 500 years is far less well-documented. Only by careful paleoclimatic reconstruction can we hope to unravel the sequence of events which occurred during this "Little Ice Age" and to understand how extensive the changes really were geographically. With this information, it may then be possible to isolate the causes of such climatic variations.

1.3 The world in A.D. 1500

The beginning of the 16th century was somewhat of a watershed in the history of civilisation. Between 1492, when Columbus reached the Caribbean, and 1532, when Pisarro arrived in Peru, the map of the known world had irrevocably changed. Vasco de Gama rounded the southern tip of Africa in 1497 and reached as far as Calicut in India, opening up an entirely new trading route between Europe and the East. By 1519, Magellan's expedition had circumnavigated the globe. The stage was set for the emergence of colonial states, ruled from small but powerful countries, geographically isolated from their remote territories. Only Australasia and the extreme polar regions were to remain beyond the reach of explorers for a further 200 years or more.

Over the course of the next 500 years extraordinary changes in society took place. These occurred against a background of environmental changes, which may have played a critical role in some of the events which occurred. However, until we can document climatic variations of the last 500 years, the extent of such influences will remain controversial.

1.4 Sources of high resolution data for paleoclimatic reconstruction

Meteorological measurements (from which we can assess climatic conditions) are only available for relatively short periods (generally a century or less) from most parts of the world. Although observations have been maintained in a few locations for two centuries or more (see Jones and Bradley, Chapter 13) to obtain a broader picture of past climatic variations we must rely on additional non-instrumental records, from which climatic conditions can be deduced. Such records of climatically sensitive natural phenomena are surrogate or proxy measures of past climate; they contain climatic information which must be extracted and separated from the non-climatic matrix in which it is embedded. The analyst must isolate the climatic signal from the extraneous noise. As more detailed and geographically extensive records are built up, the possibility of identifying causes and mechanisms of climatic variation is increased and so the prospects of understanding future climate are enhanced.

Although there are numerous approaches to the reconstruction of past climates (Bradley 1985) only a few types of evidence have the potential of providing a record which can be resolved to the annual or seasonal level (Table 1.1). Of these, quantitative estimates of past

Table 1.1 Sources of high resolution paleoclimatic data.

Type of record	**Main distribution**	**Potential Information:**
Historical Documents	All continents	Almost all aspects of climate.
Tree rings	Continental areas*, excluding desert and tundra regions	Temperature precipitation pressure patterns, drought runoff.
Ice cores	Polar and high mountain regions	Temperature, precipitation atmospheric aerosols, atmospheric composition.
Varved sediments	Continents and some coastal basins	Temperature, precipitation solar radiation.
Corals	Tropical oceans	Sea surface temperatures, adjacent continental rainfall.

* Studies of tree growth in tropical areas have not yet provided useful paleoclimatic reconstructions

climate from studies of varved sediments, and corals, have not yet been widely carried out. These natural archives have great potential for paleoclimatology, but a number of problems have yet to be resolved. Two approaches to climatic reconstruction have the potential of providing wide geographic coverage: the analysis of documentary records (discussed in Section A) and of tree growth (dendroclimatic) indices (Section B). These can be supplemented in high latitude and high altitude regions by the analysis of ice cores (Section C). Figure 1.1 shows the distribution of records discussed in the various chapters of this book. By combining these different approaches, the aim is to construct a picture of past climatic variations on larger and larger spatial scales in which the whole is greater than the sum of its individual parts. At the same time, we need records of those phenomena which are thought to have played a role in causing climatic variations of the past. The most likely candidates for climate forcing on this time scale include explosive volcanic eruptions, solar activity variations and El Nino-Southern Oscillation (ENSO) events of varying magnitude. As with the paleoclimatic record itself, proxy records can be used to reconstruct the history of these climatic forcing factors. Such records are discussed in Section D.

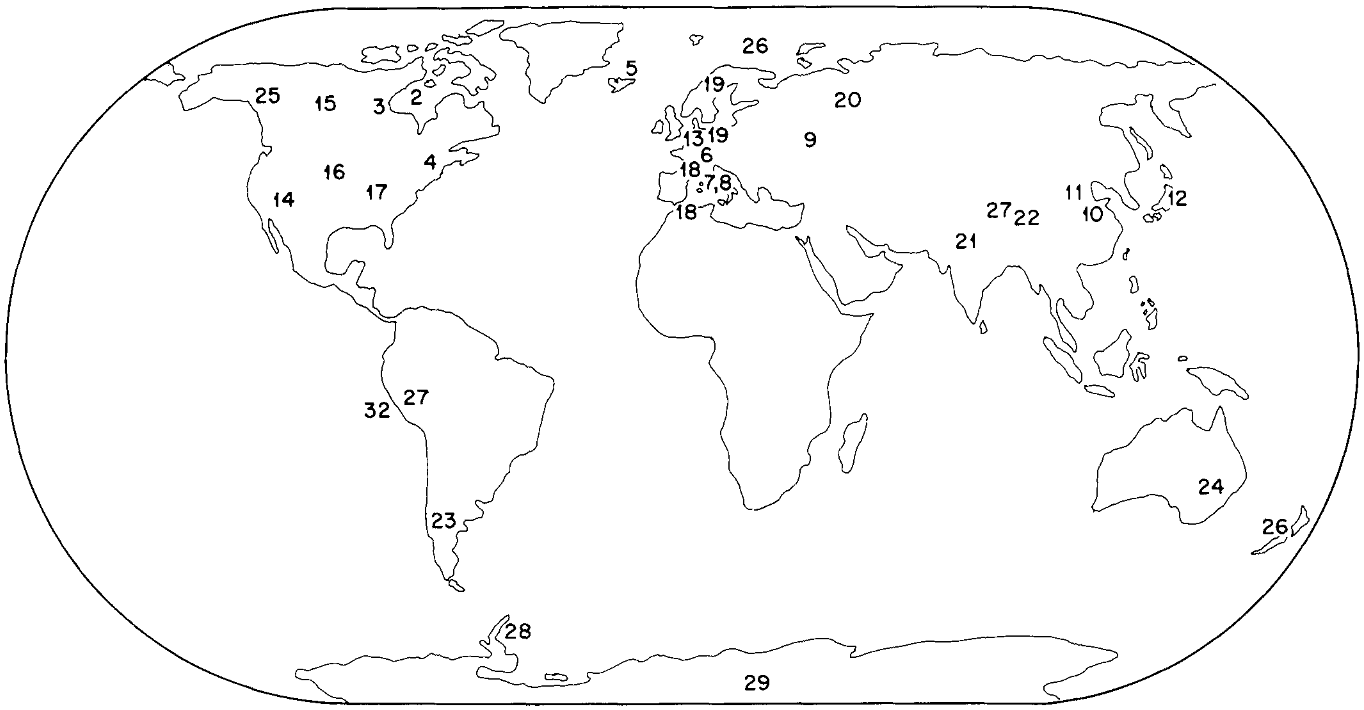

Figure 1.1 Approximate location of studies discussed in text. Numbers refer to chapters in this volume.

1.5 Methods of paleoclimatic reconstruction

1.5.1 Documentary records

Historical data can be grouped into three major categories. First, there are observations of weather phenomena *per se*, for example, the frequency and timing of frosts, or the occurrence of rainfall or snowfall recorded by early diarists. Secondly, there are records of weather-dependent natural phenomena (sometimes termed parameteorological phenomena) such as droughts, floods, lake or river freeze-up and break-up, etc. Thirdly, there are phenological records, which deal with the timing of recurrent weather-dependent biological phenomena, such as the dates of flowering of shrubs and trees, the timing of harvest (both fruit and grain) or the arrival of migrant birds in the spring. Within each of these categories there is a wide range of potential sources and an equally wide range of possible climate-related phenomena.

Potential sources of historical paleoclimatic information include: (a) ancient inscriptions, (b) annals, chronicles, etc., (c) government records, (d) private estate records, (e) maritime and commercial records, (f) personal papers, such as diaries or correspondence, (g) scientific or quasi-scientific writings, such as (non-instrumental) weather journals and (h) fragmented early instrumental records (Ingram *et al.* 1978). In all these sources, the historical climatologist is faced with the difficulty of ascertaining exactly what the qualitative description of the past is equivalent to, in terms of modern-day observations. What do the terms "drought," "frost," "frozen over," really mean? How can qualifying terms (e.g. "extreme" frost) be interpreted? Baker (1932) for example, notes that one 17th century diarist recorded three droughts of "unprecedented severity" in the space of only five years!

An approach to solving this problem has been to use content analysis (Baron 1982) to assess in quantitative terms, and as rigorously as possible, climatic information in the historical source. Historical sources are examined for the frequency with which key descriptive words were used (e.g. "snow," "frost," "blizzard," etc.) and the use which the writer may have made of modifying language (e.g. "severe frost," "devastating frost," "mild frost," etc.). In this way an assessment can be made of the order of increasing severity as perceived by the original writer. The ranked terms may then be given numerical values so that statistical analyses can be performed on the data. This may involve simple frequency counts of one variable (e.g. snow) or more complex calculations using combinations of variables. The original qualitative information may thus be transformed into more useful quantitative data on the climate of different periods in the past. Perhaps the most comprehensive work of this kind has been carried out for Switzerland by Pfister who discusses his approach in Chapter 6.

Non-climatic information is often required to interpret the climatic aspects of the source. Where did the event take place (was the event only locally important; was the diarist itinerant or sedentary?) and precisely when did it occur and for how long? This last question may involve difficulties connected with changing calendar conventions as well as trying to define what is meant by terms such as "summer" or "winter," and what time span might be represented by a phrase such as "the coldest winter in living memory." Not all of these problems may be soluble, but content analysis can help to isolate the most pertinent and unequivocal aspects of the historical source (Moody and Catchpole 1975).

Historical sources rarely give a complete picture of former climatic conditions. More commonly, they are discontinuous observations, often biased towards the recording of extreme events, and even these may pass unrecorded if they fail to impress the observer. Furthermore, long-term trends tend to go unnoticed since they are beyond the temporal perspective of one individual. In a sense, the human observer acts as a high-pass filter, recording short-term fluctuations about an ever-changing norm (Ingram *et al.* 1981).

It is worth noting that not all historical sources are equally reliable. It may be difficult in some cases to determine if the author is writing about events of which he has first-hand experience, or if events have been distorted by rumor, or the passage of time. Ideally, sources should be original documents rather than compilations; many erroneous conclusions about past climate have resulted from climatologists relying on poorly compiled secondary sources which have proved to be quite erroneous when traced back to the original data (Bell and Ogilvie 1978, Wigley 1978, Ingram *et al.* 1981). The problem of dealing with these types of fragmentary evidence is discussed for Italy by Camuffo and Enzi and Pavese *et al.* in Chapters 7 and 8, respectively, and for the Soviet Union by Borisenkov in Chapter 9.

As with all proxy data, historical observations need to be calibrated in some way, in order to make comparisons with recent data possible. This is commonly done by utilizing early instrumental data which may overlap with the proxy record, to develop an equation relating the two data sets. Thus, Bergthorsson (1969) was able to calibrate observations of sea-ice frequency on the north Icelandic coast with mean annual temperatures during the 19th century and then use this equation to assess long-term temperature fluctuations over the last 400 years from sea-ice observations. This work has been re-assessed by Ogilvie in Chapter 5 using historical information on both sea-ice extent and climatic conditions inland. Sea-ice observations (from the Hudson's Bay Company trading ships) are also used by Catchpole in Chapter 2 to compare sea-ice conditions during the late eighteenth and early nineteenth century with modern records.

Some observations may not need direct calibration if recent comparable observations are available. This applies to such things as rain/snow frequency, dates of first and last snowfall, river freeze/thaw dates etc., providing urban heat island effects, or technological changes (such as river canalization) have not resulted in a non-homogeneous record. For example, rain day counts are used to reconstruct monthly precipitation data for eastern China by Wang and Zhang in Chapter 11 and by Murata in Chapter 12 for Japan. The relationship between precipitation occurrence and temperature in Beijing during the eighteenth century is used by Wang *et al.* in Chapter 10 to estimate summer temperatures. Early weather diaries and instrumental measurements are used by Baron in Chapter 4 to estimate climate changes over New England during the seventeenth to nineteenth centuries. Observations of weather conditions by employees of the Hudson's Bay Company are discussed by Ball in Chapter 3.

1.5.2 Dendroclimatology

Variations in annual tree ring parameters from one year to the next have long been considered an important source of past climatic information. Climatic reconstructions have been made not only from measurements of a series of ring widths but more recently from measurements of maximum latewood density and from isotopic studies of the cellulose in individual tree rings (e.g. Briffa *et al.* 1988). Dendroclimatic reconstructions have been made in sub-tropical, temperate and high latitude regions of both hemispheres, using trees which can be shown to contain a primary climate-dependent signal. The technique is limited in tropical regions because most tree species there do not form distinct annual rings and growth is less susceptible to the inter-annual variability of climate.

There are three important steps in dendroclimatic reconstruction:

(1) Standardization of the tree-ring parameters to produce a site chronology.
(2) Calibration of the site chronology with instrumentally-recorded climatic data, and production of a climatic reconstruction based on the calibration equations.
(3) Verification of the reconstruction with data from an independent period not used in the initial calibration.

Here we briefly discuss these steps to provide some background for the chapters which follow in Section B.

Standardization Following the cross-dating of all series of tree-ring measurements from a site, it is generally necessary to standardize these in some way. The reason for this is that single series of tree-ring width measurements often exhibit systematic trends which are directly attributable to the aging of the tree. The best example of this is the wider widths of young rings compared to the thinner widths of older rings. In order for series from young and old trees to be amalgamated or compared it is necessary to remove the influence of this growth trend. This is generally done by fitting some form of mathematical function to the raw data and dividing or subtracting each measured value by the expected value according to the fitted curve. This procedure leads to a new series of dimensionless indices with a variance that is roughly equal through time (Fritts 1976).

A site chronology is produced by averaging the indexed series from each tree. Averaging the indices increases the expression of the common signal because variability which differs from tree to tree is lost in the averaging. Much of the common signal is climatically related. Ideally, a chronology should contain enough samples to ensure that the residual noise, left after averaging, is minimal. One way of gauging this is to calculate the Expressed Population Signal (EPS). This parameter estimates how well the site chronology approximates the population chronology. It is based on the average intersample correlation coefficient (r) (see Briffa 1984 and Wigley *et al.* 1984 for details). The lower the r value, the greater the replication required to achieve a certain level of confidence in the overall chronology.

The key point in standardization is the degree of fit between the functional form and the measured ring width series. If the fit is tight then the standardization procedure will take out more of the low-frequency variance, which may be related to climate, than desired. In some circumstances it may be necessary to trade-off the loss of low-frequency variance with the increase in statistical reliability of any reconstructions. For further information on this aspect of standardization the reader is referred to Cook (1985) and Briffa *et al.* (1987).

Dendroclimatic studies of trees from the arid southwestern United States tend to use modified exponential functions of the form $Yt = ae^{-bt} + k$, where a, b and k vary according to the tree-ring measurements (Fritts 1976). This approach is conservative in that it minimized the loss of potential climatic information on relatively long timescales. The use of such 'functions' is justified by the particular growth habitat, lack of competition and the relatively long-lived nature of semi-arid tree species.

The extension of dendroclimatology to many other parts of the world, most notably to Europe and eastern North America, involved the use of deciduous trees often growing in closed-canopy situations. Their growth curves are poorly modelled by the negative exponential function as they often contain periods of growth enhancement and suppression related to non-climatic factors such as competition, management, insect infestation etc. Several functional forms have been suggested to overcome these problems. Examples include orthogonal polynomials (Fritts 1976) Gaussian filters (Briffa 1984) and cubic splines (Cook and Peters 1981). Whichever functional form is used, it is important to decide how tight to fit the function to the ring width series. For the most complete discussion of this often neglected issue in dendroclimatology, the reader is referred to Cook and Briffa (1990).

Calibration Once a master chronology of standardized indices of some tree growth parameter (e.g. ring width, maximum latewood density) has been produced, the next step is to relate this to variations in climatic data. Mathematical and/or statistical procedures are used to derive an equation relating the tree indices to a climate variable. The equation is developed over a period known as the calibration period. If the equation adequately describes instrumental climatic variability over the calibration period, then it can be applied to past tree growth data to reconstruct the climate variable back as far as the beginning of the chronology series. As with all mathematical-statistical relationships of this kind, it is preferable to retain some climate data outside the calibration period. These data can then be used to assess the performance of the fitted equation with independent data from a verification period. The importance of this is discussed in the next section.

Calibration methods can be classified according to a hierarchical structure of complexity (see Table 1.2). The following is a brief outline of the statistical approaches illustrated with some recent examples from the dendroclimatic literature. More complete descriptions of some of the statistical procedures are given by Fritts (1976) Hughes *et al.* (1982) and Kairiukstis and Cook (1990).

Table 1.2 Calibration methods.

Level	**Number of variables of: Tree Growth**	**Climate**	**Statistical Technique**
1	1	1	Simple regression analysis
2a	n	1	Multiple linear regression analysis (MLR)
2b	nP	1	Principal component analysis (PCA)
3a	nP	nP	Orthogonal spatial regression (PCA + MLR)
3b	n	n	Canonical regression analysis

nP = number of variables after discarding unwanted ones from PCA

SIMPLE REGRESSION (LEVEL 1) This is the simplest level of climate reconstruction. Variations in growth indices at a single site are related to a single climatic parameter, such as mean summer temperature or total summer precipitation. In mathematical terms the equation might be:

$$t_i = c + bw_i$$

where t_i and w_i are the mean summer temperature and tree ring widths in year i, and c and b are the coefficients estimated by simple regression. An example of this approach is the work of Jacoby and Ulan (1982) where the date of the first complete freezing of the Churchill River estuary on Hudson Bay is reconstructed from a single chronology located near Churchill, Manitoba. Graybill and Shiyatov in Chapter 20 use this approach to reconstruct summer temperatures in the northern Ural mountains.

Increasing complexity might involve using additional predictors in the regression such as the previous year's growth, or using the average of a number of separate site chronologies. For example, Duvick and Blasing (1981) reconstructed annual (August-July) precipitation totals for Iowa from a single regional chronology which was an average of three site chronologies of the same species. Similarly, in Chapter 21, Hughes averages together ring width chronologies from eight sites and maximum latewood density chronologies from seven sites to reconstruct early spring and late summer temperatures and precipitation totals in the Kashmir region of the Himalayas.

MULTIPLE LINEAR REGRESSION (LEVEL 2) Most dendroclimatic reconstructions make use of multivariate statistics relating a single climate variable to an array of tree ring chronologies. An example is

$$t_i = c + b_1w_{i1} + \ldots, + b_nw_{in}$$

where there are *n* regression coefficients b_j for each of the *n* chronologies w_{ij}. This approach is simply an expansion of the level 1 technique, but bringing in more predictors generally accounts for progressively more climate variance. Incorporating too many terms (coefficients) into the equation is not to be recommended, even though it would be possible to explain all the variance in the climate parameter if enough terms were used. Adding too many coefficients simply widens the confidence limit about the reconstruction estimates.

Two procedures have been proposed for deciding how many of the possible n chronology predictors should be retained in the prediction equation. The first uses stepwise multiple regression to select the number of climate variables. In this technique, the chronology which accounts for most of the climate variables is selected, followed successively by the chronology that accounts for most of the remaining variance. Predictor chronologies are continually added until the point is reached where the addition of another chronology explains an amount of variance below some pre-determined threshold (e.g. 5%). An example of this approach is the reconstruction of drought in Southern California by Meko *et al.* (1980). Variations of the basic approach are used by D'Arrigo and Jacoby, Wu, and Boninsegna to reconstruct paleoclimatic conditions in northern North America, western China and southern South America, in Chapters 15, 22 and 23, respectively.

A major problem with stepwise regression is that intercorrelations between the tree ring predictors can lead to instability in the predictor equation. In statistical terms this is referred to as multi-collinearity. The second procedure attempts to overcome this by expressing the variance of the tree ring data in terms of principal components (PCs) (also called empirical orthogonal functions or EOFs) and using these in the regression procedure. Principal components analysis (PCA) is a statistical transformation of the original (intercorrelated) variables to produce a set of orthogonal (uncorrelated) principal components (for a review, see Richman 1986). The first PC is the mode of variation of the data that explains the most variance. The second is that which explains most of the remainder, but which is orthogonal to the first, and so on. Although there are as many PCs as there were original variables, the transformation means that most of the variance of the original data set is explained by a few PCs. A selection criteria will then be needed to decide how many PCs to retain. Various methods have been proposed (see e.g. Preisendorfer *et al* 1981).

Principal components analysis, apart from reducing the number of potential predictors, also considerably simplifies the multiple regression. It is not necessary to use the stepwise procedure because the new potential predictors are all orthogonal. An example of this kind of procedure is the reconstruction of July drought (Palmer Indices; Palmer 1965) in the Hudson Valley in New York by Cook and Jacoby (1979). Meko, Serre-Bachet *et al.* and Norton and Palmer (in Chapters 16, 18 and 24, respectively) use modifications of the principal component regression to reconstruct the paleoclimatic history of the U.S. Plains, southwestern Europe and northwest Africa and Australia and New Zealand.

SPATIAL REGRESSION (LEVEL 3) The procedures used in this level are the most complex of all reconstructions attempted in dendroclimatology. A spatial array of tree-growth data is used to reconstruct the spatial field of some climatic parameter. Details of this complex statistical procedure are given by Fritts (1976) and Blasing (1978). The earliest spatial dendroclimatic reconstruction used canonical regression analysis to establish the relationship between patterns of tree ring growth over western North America and mean sea-level pressure patterns over the North Pacific and adjacent continental regions of eastern Asia and North America (Fritts *et al.* 1971). This approach is extended in chapter 14 by Fritts and Shao by mapping temperature and precipitation patterns over the United States.

Related procedures have been used by other workers. Briffa *et al.* (1988) for example, use orthogonal spatial regression techniques to reconstruct April to September temperatures over Europe west of 30°E using densitometric information from conifers over Europe. In their procedure both the spatial array of temperature and the spatial array of densitometric data are first reduced to their principal components. Only significant components in each set are retained. Each retained PC of climate is then regressed in turn against the set of retained densitometric PCs. This procedure can be thought of as repeating the level 2 approach m times, where m is the number of retained climate PCs. Having found all the significant regression coefficients, the set of equations relating the climate PCs to the tree growth PCs are then transformed back to original variable space, resulting in an equation for each temperature location in terms of all the densitometric chronologies. In Chapter 19, Briffa and Schweingruber use this approach to present temperature reconstructions from northern Europe, and in Chapter 17 Cook *et al.* reconstruct Palmer (drought) Indices for the eastern United States using a spatial version of the level 2 principal components regression. In this technique the Indices are reconstructed point-by-point over the study region. It differs from the level 3 spatial regression approach by not reducing the drought indices to principal components.

Verification No matter how complex the calibration procedure is, it is essential that the empirically-derived equation(s) be tested with independent data from the verification period. The statistical procedures described in the previous section ensure that the maximum amount of climate variance is explained by the tree growth parameters over the calibration period. It is likely that this amount of explained variance will be reduced when the prediction estimates are tested against independent data. The performance of the equation over the verification period gives the best guide to the likely quality of the reconstruction for periods prior to instrumental measurement. All the authors in section 3 pay particular attention to this aspect of dendroclimatic reconstruction.

There is really no substitute for independent data to assess the reconstruction performance. At least one-third to one-half of climate data should be retained for verification. Many reconstructions split the climate data into two halves calibrating on one half and verifying on the other. The process may then be repeated, reversing the calibration and verification periods.

A number of statistical comparisons between reconstructed and independent climate data have been proposed (see, for example, Fritts 1976). These include comparisons of the mean, standard deviation, correlation coefficient, reduction of error and the Durbin-Watson test for series correlation on the residuals. For further discussion of verification procedures the reader is referred to Gordon (1982) where there is discussion of the many sub-sample replication techniques proposed by Mosteller and Tukey (1977).

1.5.3 Ice core records

The accumulation of past snowfall in the remote polar and alpine ice caps and ice sheets of the world provides an extraordinarily valuable record of paleoclimatic and paleoenvironmental conditions. These conditions are generally studied by detailed physical and chemical analyses of ice and firn (snow which has survived the summer melt season) in cores recovered from the highest elevations on the ice surface, where snow melt and sublimation are essentially zero. On ice caps where melting may accasionally occur, the re-freezing of meltwater within the underlying snowpack produces an identifiable crystal fabric which can be studied as an indicator of warm summer conditions in the past (Koerner 1977). In Chapter 26, Tarussov discusses such a record from Spitsbergen and the Soviet Arctic Islands.

Snowfall provides a unique record, not only of accumulation amounts *per se*, but also of air temperature at the time the snow formed in the atmosphere, atmospheric composition (including gaseous composition and aerosol content) and the occurrence of explosive volcanic eruptions. Although many ice cores have now been recovered, the detailed record of the last 500 years has been studied in relatively few of them. In Section 4 of this volume, details of recent research on high elevation ice cores from the Alaska-Yukon border, from Peru and from Tibet are discussed, together with records from Antarctica, the Soviet Arctic Islands and Spitsbergen.

A primary source of paleoclimatic information in ice cores is the variation of oxygen isotopes in the water molecules which made up the original snow crystals. These variations arise because the vapor pressure of $H_2{}^{16}O$ is 1% higher than that of $H_2{}^{18}O$. Evaporation from a water body thus produces a vapor which is relatively depleted in the heavier isotope, ^{18}O, than the original water. However, when condensation occurs, the lower vapor pressure of $H_2{}^{18}O$ results in this heavier molecule passing from the vapor to the liquid state more readily than the lighter $H_2{}^{16}O$ molecules. As condensation proceeds, this fractionation causes the remaining vapor to become increasingly depleted in the heavier isotope, and so precipitation from the remaining vapor contains less and less $H_2{}^{18}O$. The greater the fall in temperature, the more condensation will occur and the lower will be the heavy isotope concentration in the precipitation, compared to that in the original water source. Similarly, falls in temperature lead to lower proportions of deuterium (^2H) relative to hydrogen (^1H) in precipitation. Isotopic concentration in precipitation can therefore be considered as a function of the temperature at which the condensation occurs. Some of the problems of establishing the precise nature of this relationship are discussed by Peel in Chapter 28. The relative proportions of ^{16}O and ^{18}O in a sample are expressed in terms of departures ($\delta^{18}O$) from a standard, known as SMOW (Standard Mean Ocean Water); a $\delta^{18}O$ value of -10 indicates that a sample has an $^{18}O/^{16}O$ ratio 1% (10 per mil) less than SMOW (Bradley 1985).

Because of the relationship of oxygen (and hydrogen) isotopes in snowfall to temperature, there is a strong seasonal variation of $\delta^{18}O$ (and δD) which enables seasonal layers in annual accumulation to be detected in ice cores. Such variations are a primary means of identifying annual layers in the ice core record, enabling a chronology to be established in that part of the core where distinct layering can be detected (see Mosley-Thompson, Chapter 29). At greater depths, the accumulation of overlying snow and molecular diffusion between layers causes the seasonal signal to become less distinct and other means of dating have to be employed. However, in most of the sites discussed in Section 4, such problems are only encountered at significantly greater depths than those occupied by the record of the last 500 years. When confusion arises as to whether a clear sequence of seasonal layering occurs, other signals may be used to help establish the correct chronology; for example, in many ice core records, a seasonal variation in aerosols is found, commonly peaking in spring or early summer months and reaching a minimum in winter months (see, for example, records from the Quelccaya Ice Cap, discussed by Thompson in Chapter 27). Cross-checking between the isotopic variations and the aerosols generally resolves any uncertainties in the chronology (cf. Hammer *et al.* 1978 and Mosley-Thompson, Chapter 29). As an additional check, distinct horizons with elevated acidity content are often detectable by measuring the electrolytic conductivity of the ice. These layers represent times of explosive volcanic eruptions, many of which are known from historical sources, often to the day of the eruption. Holdsworth *et al.*, (Chapter 25) demonstrate the value of such chronostratigraphic markers in their study of an ice core from Mount Logan, Alaska. The overall record of explosive volcanism is discussed in more detail in Chapter 31 by Bradley and Jones.

Once a reliable stratigraphy has been established, it may be possible to determine variations in accumulation rate over time, assuming losses due to ablation and deformation are minimal. Such information provides important insights into atmospheric circulation, as illustrated by studies of the Quelccaya Ice Cap ice cores (Chapter 27) and of the Mount Logan ice core (Chapter 25).

1.6 Causes of Climatic Variations

Ice cores provide a unique opportunity to study some of those factors which may have caused (forced) climatic variations (such as explosive volcanic eruptions) in direct comparison with a proxy climatic record which may also record the *impact* of the forcing factor. This enables the rate of climate change to be established, at least locally, without any uncertainties about dating and comparisons of separate, independent records. However, not all important forcing factors are recorded in ice cores and past variations in climate may still require additional records of forcing to understand the paleoclimatic record. Section D provides details of three important factors which may be responsible for large scale climatic variations in the past – solar activity variations (Stuiver and Braziunas, Chapter 30) explosive volcanic eruptions (Bradley and Jones, Chapter 31) and El Nino-Southern Oscillation (ENSO) events (Quinn and Neale, Chapter 32).

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