PALEOCLIMATOLOGY

Reconstructing Climates of the Quaternary

Second Edition

Raymond S. Bradley

University of Massachusetts
Amherst, Massachusetts

HARCOURT ACADEMIC PRESS

San Diego London Boston
New York Sydney Tokyo Toronto

10 DENDROCLIMATOLOGY

10.1 INTRODUCTION

Variations in tree-ring widths from one year to the next have long been recognized as an important source of chronological and climatic information. In Europe, studies of tree rings as a potential source of paleoclimatic information go back to the early eighteenth century when several authors commented on the narrowness of tree rings (some with frost damage) dating from the severe winter of 1708–1709. In North America, Twining (1833) first drew attention to the great potential of tree rings as a paleoclimatic index (for historical reviews, see Studhalter, 1955; Robinson *et al.*, 1990; Schweingruber, 1996, p. 537). However, in the English-speaking world, the "father of tree-ring studies" is generally considered to be A.E. Douglass, an astronomer who was interested in the relationship between sunspot activity and rainfall. To test the idea of a sunspot-climate link, Douglass needed long climatic records and he recognized that ring-width variations in trees of the arid southwestern United States might provide a long, proxy record of rainfall variation (Douglass, 1914, 1919). His efforts to build long-term records of tree growth were facilitated by the availability of wood from archeological sites, as well as from modern trees (Robinson, 1976). Douglass' early work was crucial for the development of dendrochronology (the use of tree rings for dating) and for dendroclimatology (the use of tree rings as a proxy indicator of climate).

10.2 FUNDAMENTALS OF DENDROCLIMATOLOGY

A cross section of most temperate forest trees will show an alternation of lighter and darker bands, each of which is usually continuous around the tree circumference. These are seasonal growth increments produced by meristematic tissues in the tree's cambium. When viewed in detail (Fig. 10.1) it is clear that they are made up of sequences of large, thin-walled cells (earlywood) and more densely packed, thick-walled cells (latewood). Collectively, each couplet of earlywood and latewood comprises an annual growth increment, more commonly called a tree ring. The mean width of a ring in any one tree is a function of many variables, including the tree species, tree age, availability of stored food within the tree and of important nutrients in the soil, and a whole complex of climatic factors (sunshine, precipitation,

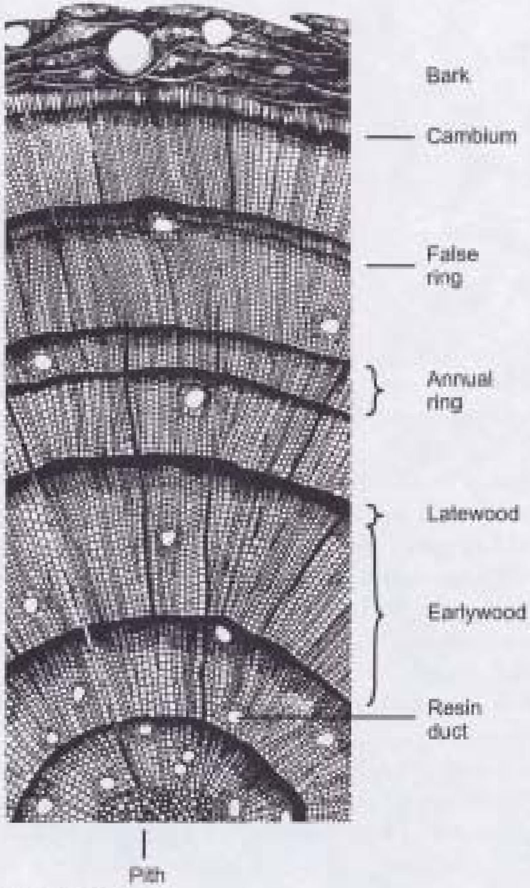

FIGURE 10.1 Drawing of cell structure along a cross section of a young stem of a conifer. The earlywood is made up of large and relatively thin-walled cells (tracheids); latewood is made up of small, thick-walled tracheids. Variations in tracheid thickness may produce false rings in either earlywood or latewood (Fritts, 1976).

temperature, wind speed, humidity, and their distribution throughout the year). The problem facing dendroclimatologists is to extract whatever climatic signal is available in the tree ring data and to distinguish this signal from the background noise. Furthermore, the dendroclimatologist must know precisely the age of each tree ring if the climatic signal is to be chronologically useful. From the point of view of paleoclimatology, it is perhaps useful to consider the tree as a filter or transducer which, through various physiological processes, converts a given climatic input signal into a certain ring width output that is stored and can be studied in detail, even thousands of years later (Fritts, 1976; Schweingruber, 1988, 1996).

Climatic information has most often been gleaned from interannual variations in ring width, but there has also been a great deal of work carried out on the use of density variations, both inter- and intra-annually (densitometric dendroclimatology). Wood density is an integrated measure of several properties, including cell wall thickness, lumen diameter, size and density of vessels or ducts, proportion of fibers, etc. (Polge, 1970). Tree rings are made up of both earlywood and latewood, which vary markedly in average density and these density variations can be used, like ring-width measurements, to identify annual growth increments and to cross-date samples (Parker, 1971). It has also been shown empirically that density variations contain a strong climatic signal and can be used to estimate long-term climatic variations over wide areas (Schweingruber *et al.*, 1979, 1993). Density variations are measured on x-ray negatives of prepared core sections (Fig. 10.2) and the optical density of the negatives is inversely proportional to wood density (Schweingruber *et al.*, 1978).

Density variations are particularly valuable in dendroclimatology because they have a relatively simple growth function (often close to linear with age). Hence standardization of density data may allow more low-frequency climatic information to be retained than is the case with standardized ring-width data (see Section 10.2.3). Generally, two values are measured in each growth ring: minimum density and maximum density (representing locations within the earlywood and latewood layers, respectively), although maximum density values seem to be a better climatic indicator than minimum density values. For example, Schweingruber *et al.* (1993) showed that maximum density values were strongly correlated with April-August mean temperature in trees across the entire boreal forest, from Alaska to Labrador, whereas minimum and mean density values and ring widths had a much less consistent relationship with summer temperature at the sites sampled (D'Arrigo *et al.*, 1992). Maximum latewood density values are calibrated in the same way as with the ring-width data using the statistical procedures described in Section 10.2.4. However, optimum climatic reconstructions may be achieved by using both ring widths and densitometric data to maximize the climatic signal in each sample (Briffa *et al.*, 1995).

Isotopic variations in wood have been studied as a possible proxy of temperature variations through time, but the complexities of fractionation both within the hydrological system, and in the trees themselves, make simple interpretations very difficult (see Section 10.4). Ring-width and densitometric and isotopic approaches to paleoclimatic reconstruction are complementary and, in some situations, could be used independently to check paleoclimatic reconstructions based on only one of

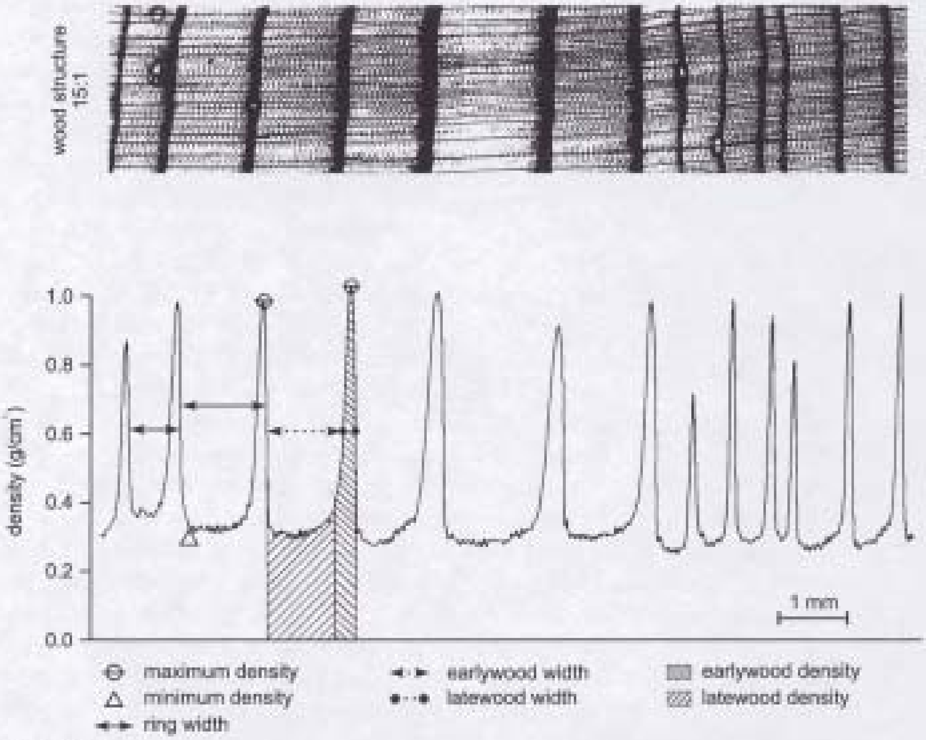

FIGURE 10.2 Example of a tree-ring density plot based on an x-ray negative of a section of wood (top of figure). Minimum and maximum densities in each annual ring are clearly seen, enabling the annual ring width to be measured as well as the width of both the earlywood and latewood (courtesy of F. Schweingruber).

the methods, or collectively to provide more accurate reconstructions (Briffa *et al.*, 1992a).

10.2.1 Sample Selection

In conventional dendroclimatological studies, where ring-width variations are the source of climatic information, trees are sampled in sites where they are under stress; commonly, this involves selection of trees that are growing close to their extreme ecological range. In such situations, climatic variations will greatly influence annual growth increments and the trees are said to be sensitive. In more beneficent situations, perhaps nearer the middle of a species range, or in a site where the tree has access to abundant groundwater, tree growth may not be noticeably influenced by climate, and this will be reflected in the low interannual variability of ring widths (Fig. 10.3). Such tree rings are said to be complacent. There is thus a spectrum of possible sampling situations, ranging from those where trees are extremely sensitive to climate to those where trees are virtually unaffected by interannual climatic variations. Clearly, for

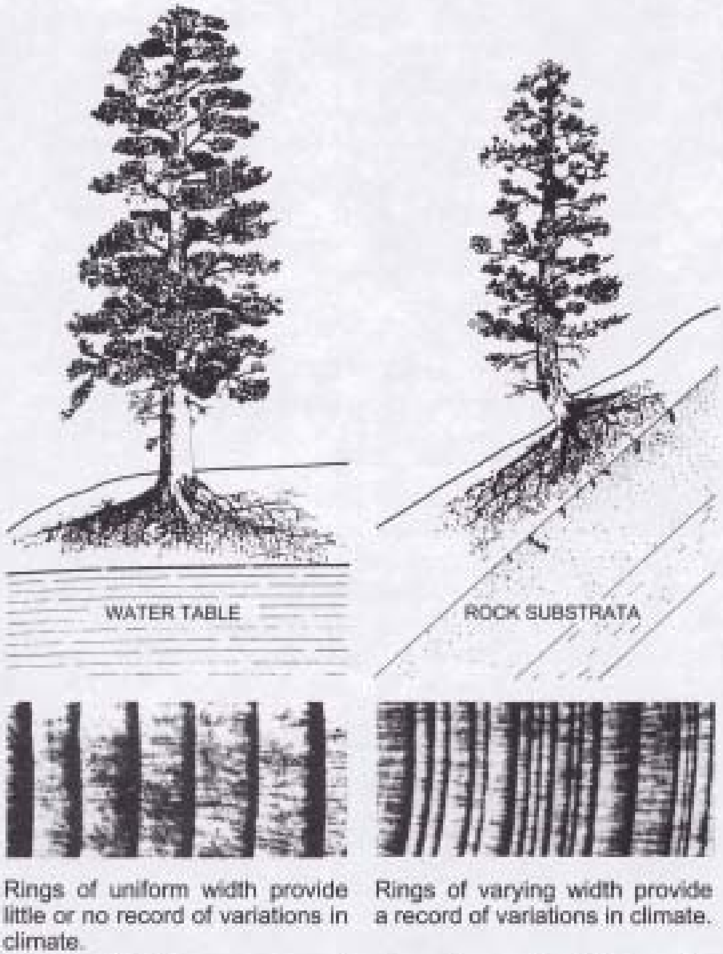

FIGURE 10.3 Trees growing on sites where climate seldom limits growth processes produce rings that are uniformly wide **(left)**. Such rings provide little or no record of variations in climate and are termed complacent. **(right)**: Trees growing on sites where climatic factors are frequently limiting produce rings that vary in width from year to year depending on how severely limiting climate has been to growth. These are termed sensitive (Fritts, 1971).

useful dendroclimatic reconstructions, samples close to the sensitive end of the spectrum are favored as these would contain the strongest climatic signal. Often, therefore, tree-ring studies at the range limit of trees are favored (e.g., alpine or arctic treeline sites). However, climatic information may also be obtained from trees that are not under such obvious climatic stress, providing the climatic signal common to all the samples can be successfully isolated (LaMarche, 1982). For example, ring widths of bald cypress trees from swamps in the southeastern United States have been used to reconstruct the drought and precipitation history of the area over the last 1000 years or more (Stahle *et al.*, 1988; Stahle and Cleaveland, 1992).

Paleoclimatic reconstructions have also been achieved using teak from equatorial forests in Indonesia (D'Arrigo *et al.*, 1994) as well as from mesic forest trees in Tasmania (Cook *et al.*, 1992a). For isotope dendroclimatic studies (Section 10.4), the sensitivity requirement does not seem to be as critical and it may, in fact, be preferable to use complacent tree rings for analysis (Gray and Thompson, 1978). Sensitivity is also less significant in densitometric studies and good relationships between maximum latewood density and temperature have been found in "normal" trees, which are growing on both moist and well-drained sites (Schweingruber *et al.*, 1991, 1993).

In marginal environments, two types of climatic stress are commonly recognized, moisture stress and temperature stress. Trees growing in semiarid areas are frequently limited by the availability of water, and ring-width variations primarily reflect this variable. Trees growing near to the latitudinal or altitudinal treeline are mainly under growth limitations imposed by temperature and hence ring-width variations in such trees contain a strong temperature signal. However, other climatic factors may be indirectly involved. Biological processes within the tree are extremely complex (Fig. 10.4) and similar growth increments may result from quite different combinations of climatic conditions. Furthermore, climatic conditions prior to the growth period may "precondition" physiological processes within the tree and hence strongly influence subsequent growth (Fig. 10.5). For the same reason, tree growth and food production in one year may influence growth in the following year, and lead to a strong serial correlation or autocorrelation in the tree-ring record. Tree growth in marginal environments is thus commonly corre-

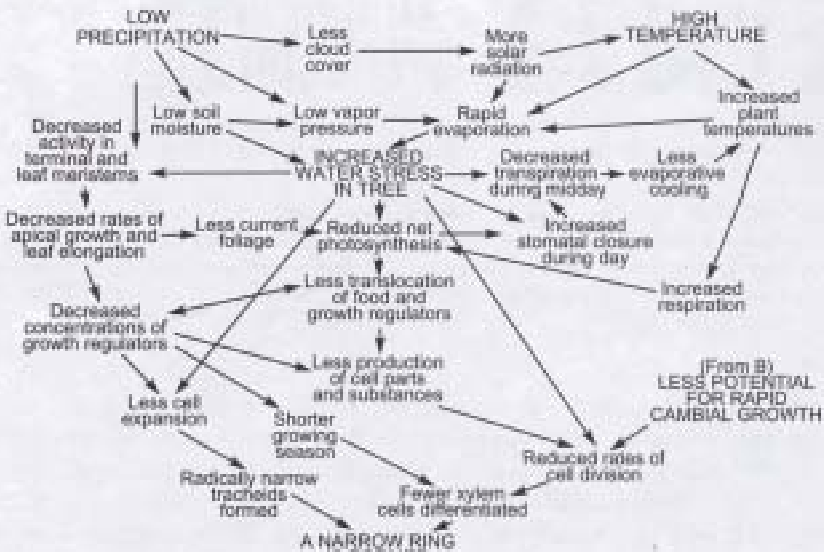

FIGURE 10.4 A schematic diagram showing how low precipitation and high temperature during the growing season may lead to the formation of a narrow tree ring in arid-site trees. Arrows indicate the net effects and include various processes and their interactions. It is implied that the effects of high precipitation and low temperature are the opposite and may lead to an increase in ring widths (Fritts, 1971).

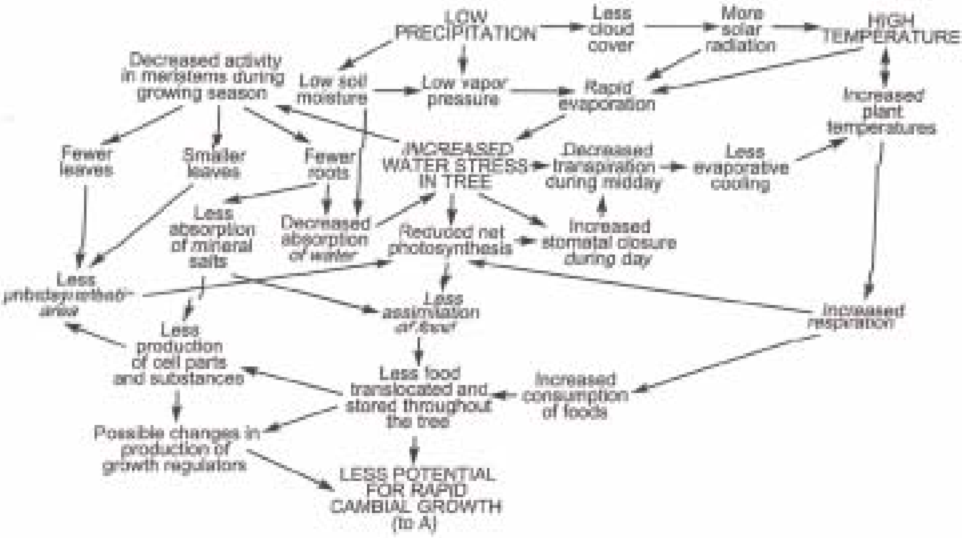

FIGURE 10.5 A schematic diagram showing how low precipitation and high temperature *before* the growing season may lead to a narrow tree ring in arid-site trees (Fritts, 1971).

lated with a number of different climatic factors in both the growth season (year t_0) and in the preceding months, as well as with the record of prior growth itself (generally in the preceding growth years, t_{-1} and t_{-2}). Indeed in some dendroclimatic reconstructions, tree growth in subsequent years (t_{+1}, t_{+2}, etc.) may also be included as they also contain climatic information about year t_0.

Trees are sampled radially using an increment borer, which removes a core of wood (generally 4–5 mm in diameter) and leaves the tree unharmed. It is important to realize that dendroclimatic studies are unreliable unless an adequate number of samples are recovered; two or three cores should be taken from each tree and at least 20 trees should be sampled at an individual site, though this is not always possible. Eventually, as will be discussed, all of the cores are used to compile a master chronology of ring-width variation for the site and it is this that is used to derive climatic information.

10.2.2 Cross-Dating

For tree-ring data to be used for paleoclimatic studies, it is essential that the age of each ring be precisely known. This is necessary in constructing the master chronology from a *site* where ring widths from modern trees of similar age are being compared, and equally necessary when matching up sequences of overlapping records from modern and archeological specimens to extend the *chronology back in time* (Stokes and Smiley, 1968). Great care is needed because occasionally trees will produce *false rings* or intra-annual growth bands, which may be confused with the actual earlywood/latewood transition (Fig. 10.6). Furthermore, in extreme years some

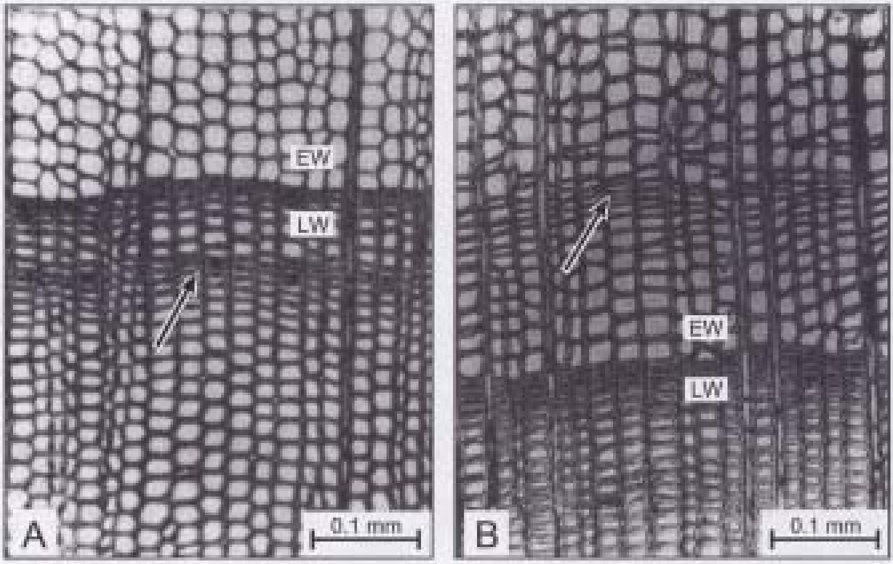

FIGURE 10.6 Annual growth increments or rings are formed because the wood cells produced early in the growing season (earlywood, EW) are large, thin-walled, and less dense, while the cells formed at the end of the season (latewood, LW) are smaller, thick-walled, and more dense. An abrupt change in cell size between the last-formed cells of one ring (LW) and the first-formed cells of the next (EW) marks the boundary between annual rings. Sometimes growing conditions temporarily become severe before the end of the growing season and may lead to the production of thick-walled cells within an annual growth layer (arrows). This may make it difficult to distinguish where the actual growth increment ends, which could lead to errors in dating. Usually these intra-annual bands or false rings can be identified, but where they cannot the problem must be resolved by cross-dating (Fritts, 1976).

trees may not produce an annual growth layer at all, or it may be discontinuous around the tree or so thin as to be indistinguishable from adjacent latewood (i.e., a partial or missing ring) (Fig. 10.7). Clearly, such circumstances would create havoc with climatic data correlation and reconstruction, so careful cross-dating of tree ring series is necessary. This involves comparing ring width sequences from each core so that characteristic patterns of ring-width variation (ring-width "signatures") are correctly matched (Fig. 10.8). If a false ring is present, or if a ring is missing, it will thus be immediately apparent (Holmes, 1983). The same procedure can be used with archeological material; the earliest records from living trees are matched or cross-dated with archeological material of the same age, which may, in turn, be matched with older material. This procedure is repeated many times to establish a thoroughly reliable chronology. In the southwestern United States, the ubiquity of beams or logs of wood used in Indian pueblos has enabled chronologies of up to 2000 years to be constructed in this way. In fact, accomplished dendrochronologists can quickly pinpoint the age of a dwelling by comparing the tree-ring sequence in supporting timbers with master chronologies for the area (Robinson, 1976). Similarly, archeologically important chronologies have been established in western Eu-

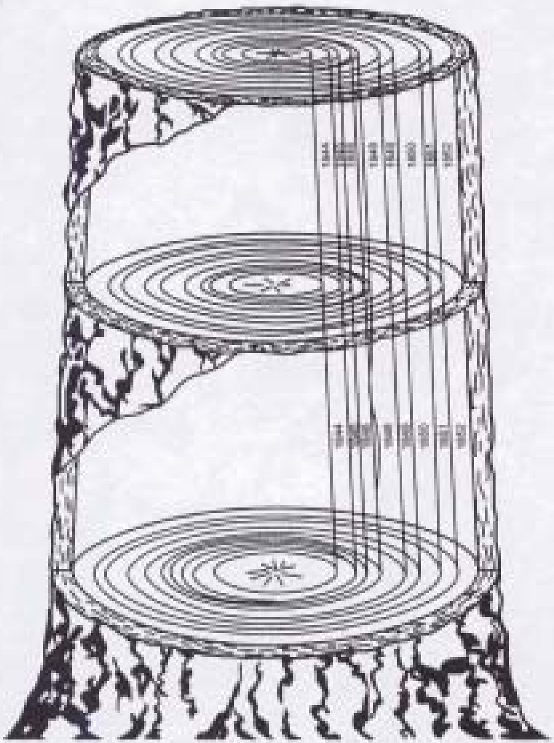

FIGURE 10.7 Schematic diagram illustrating the potential difficulty presented by the formation of a partial ring (in 1847). In the lowest two sections the ring might not be sampled by an increment borer, which removes only a narrow wood sample. In the upper section the ring is thin, but present all around the tree circumference. Such missing or partially absent rings are identified by careful cross-dating of multiple samples (Glock, 1937).

rope. Hoffsummer (1996) for example, has used beams of wood from buildings in southeastern Belgium to establish an oak chronology extending back to A.D. 672, and in several regions of France chronologies of over 1000 yr in length have been assembled from construction timbers (Lambert *et al.*, 1996). Dendrochronological studies have also been used in studies of important works of art, for example in dating wooden panels used for paintings, furniture, and even the coverboards of early books (Eckstein *et al.*, 1986; Lavier and Lambert, 1996). Finally, tree stumps recovered from alluvial sediments and bogs have been cross-dated to form composite chronologies extending back through the entire Holocene. Long tree-ring series such as these are so accurate that they are used to calibrate the radiocarbon timescale (see Section 3.2.1.5). Tree rings are unique among paleoclimatic proxies in that, through cross-dating of multiple cores, the absolute age of a sample can be established. This attribute distinguishes tree rings from other high resolution proxies (ice cores, varved sediments, corals, banded speleothems, etc.) because in such proxies, comparable replication of records and cross-dating of many samples across multiple sites is rarely possible. Consequently, with the exception of specific marker

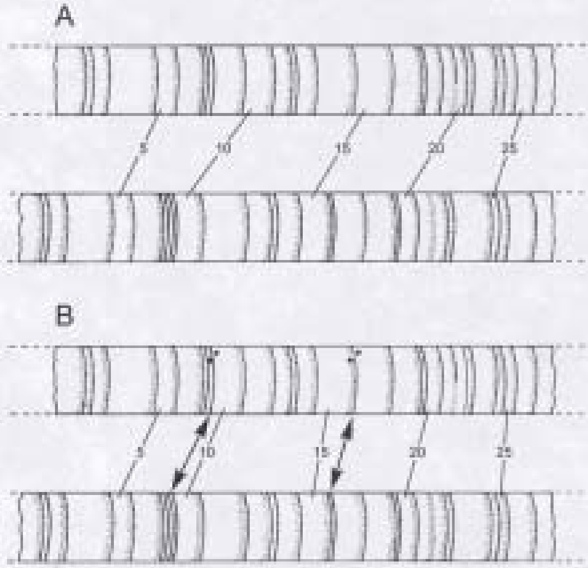

FIGURE 10.8 Cross dating of tree rings. Comparison of tree-ring widths makes it possible to identify false rings or where rings are locally absent. For example, in **(A)** strict counting shows a clear lack of synchrony in the patterns. In the lower specimen of **(A)**, rings 9 and 16 can be seen as very narrow, and they do not appear at all in the upper specimen. Also, rings 21 (lower) and 20 (upper) show intra-annual growth bands. In **(B)**, the positions of inferred absence are designated by dots (upper section), the intra-annual band in ring 20 is recognized, and the patterns in all ring widths are synchronously matched (Fritts, 1976).

horizons (e.g., those associated with a volcanic event of known age), dating of such proxies is always more uncertain than in dendroclimatic studies (Stahle, 1996).

Once the samples have been cross-dated and a reliable chronology has been established there are three important steps to produce a 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; and
3. verification of the reconstruction with data from an independent period not used in the initial calibration.

In the next three sections, each of these steps is discussed in some detail.

10.2.3 Standardization of Ring-Width Data

Once the chronology for each core has been established, individual ring widths are measured and plotted to establish the general form of the data (Fig. 10.9). It is common for time series of ring widths to contain a low frequency component resulting entirely from the tree growth itself, with wider rings generally produced during the early life of the tree. In order that ring-width variations from different cores can be

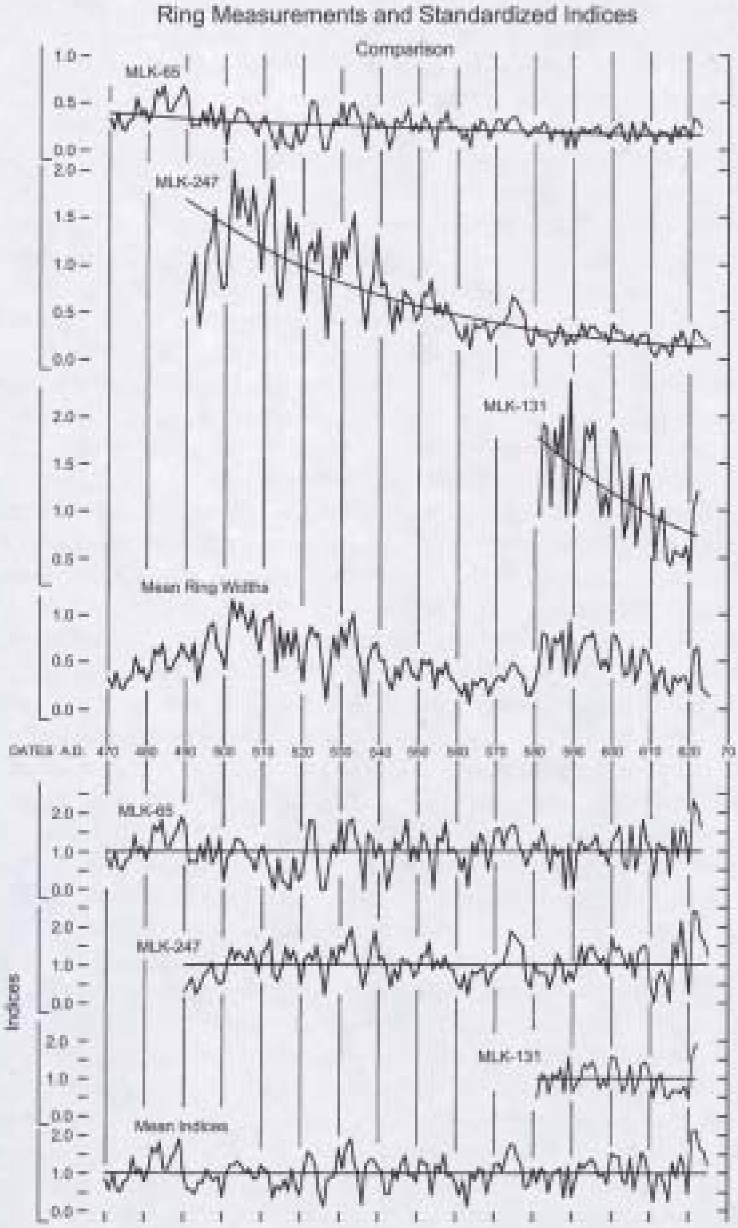

FIGURE 10.9 Standardization of ring-width measurements is necessary to remove the decrease in size associated with increasing age of the tree. If the ring widths for the three specimens shown in the upper figure are simply averaged by year, without removing the effect of the tree's age, the mean ring-width chronology shown below them exhibits intervals of high and low growth, associated with the varying age of the samples. This age variability is generally removed by fitting a curve to each ring-width series, and dividing each ring width by the corresponding value of the curve. The resulting values, shown in the lower half of the figure, are referred to as indices, and may be averaged among specimens differing in age to produce a mean chronology for a site (lowermost record) (Fritts, 1971).

compared, it is first necessary to remove the growth function peculiar to that particular tree. Only then can a master chronology be constructed from multiple cores. Growth functions are removed by fitting a curve to the data and dividing each measured ring-width value by the "expected" value on the growth curve (Fig. 10.9). Commonly, a negative exponential function, or a lowpass digital filter is applied to the data. Cook *et al.* (1990) recommend that a cubic-smoothing spline be used, in which the 50% frequency response equals ~75% of the record length (n). This means that low frequency variations in the data (with a period $>0.75\ n$) are largely removed from the standardized data, so the analyst then has an explicit understanding of the frequency domain that the resulting series represents (Cook and Peters, 1981).

The standardization procedure leads to a new time series of ring-width indices, with a mean of one and a variance that is fairly constant through time (Fritts, 1971). Ring-width indices from individual cores are then averaged, year by year, to produce a master chronology for the sample site, independent of growth function and differing sample age (Fig. 10.9, lowest graph). Averaging the standardized indices also increases the (climatic) signal-to-noise ratio (S/N). This is because climatically related variance, common to all records, is not lost by averaging, whereas non-climatic "noise," which varies from tree to tree, will be partially cancelled in the averaging process. It is thus important that a large enough number of cores be obtained initially to help enhance the climatic signal common to all the samples (Cook *et al.*, 1990).

Standardization is an essential prerequisite to the use of ring-width data in dendroclimatic reconstruction but it poses significant methodological problems. Consider, for example, the ring-width chronologies shown in Fig. 10.10. Drought-sensitive conifers from the southwestern United States characteristically show ring-width variations like those in Fig. 10.10a. For most of the chronology a negative exponential function, of the form $y = ae^{-bt} + k$, fits the data well. However, this is not the case for the early section of the record, which must be either discarded or fitted by a different mathematical function. Obviously, the precise functions selected will have an important influence on the resulting values of the ring-width indices. In the case of trees growing in a closed canopy forest, the growth curve is often quite variable and unlike the negative exponential values characteristic of arid-site conifers. Periods of growth enhancement or suppression related to non-climatic factors such as competition, management, insect infestation, etc., are often apparent in the records. In such cases (Fig. 10.10b) some other function might be fitted to the data and individual ring widths would then be divided by the local value of this curve to produce a series of ring-width indices. Care must be exercised not to select a function (such as a complex polynomial) that describes the raw data too precisely, or *all* of the (low frequency) climatic information may be removed; most analysts select the simplest function possible to avoid this problem but inevitably the procedure selected is somewhat arbitrary. Further problems arise when complex growth functions are observed, such as those in Fig. 10.10c. In this case it would be difficult to decide on the use of a polynomial function (dashed line) or a negative exponential function (solid line) and in either case the first few observations should perhaps be discarded. Such difficulties are less significant in densitometric or isotope dendroclimatic studies because there is generally less of a growth trend in density

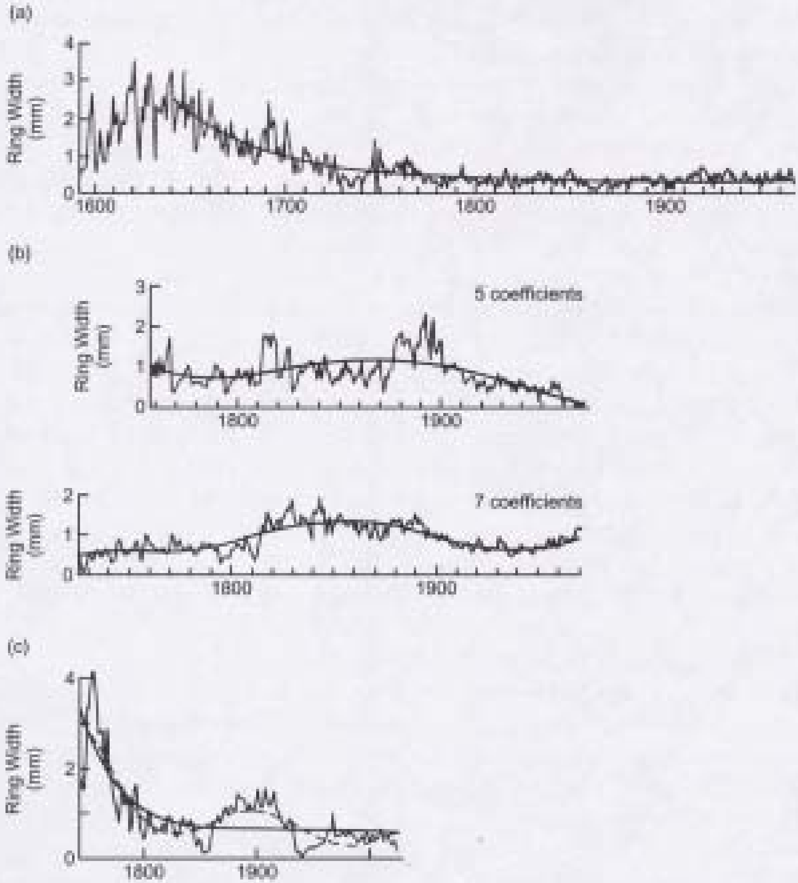

FIGURE 10.10 Some problems in standardization of ring widths. In **(a)** most of the tree-ring series can be fitted by the exponential function shown. However, the early part of the record must be discarded. In **(b)** the two ring-width series required higher-order polynomials to fit the lower frequency variations of each record (the greater the number of coefficients for each equation, the greater the degree of complexity in the shape of the curve). In **(c)** the series could be standardized using either a polynomial (dashed) or exponential function (solid line). Depending on the function selected and its complexity, low-frequency climatic information may be eliminated. The final ring-width indices depend very much on the standardization procedure employed (examples selected from Fritts, 1976).

and isotope data; hence these approaches may yield more low-frequency climatic information than is possible in the measurement of ring widths alone (Schweingruber and Briffa, 1996).

It is clear that standardization procedures are not easy to apply and may actually remove important low-frequency climatic information. It is not possible, *a priori*, to decide if part of the long-term change in ring width is due to a coincident climatic trend. The problem is exacerbated if one is attempting to construct a long-term dendrochronological record, when only tree fragments or historical timbers spanning limited time intervals are available, and the corresponding growth function may not be apparent.

The consequences of different approaches to standardization are well illustrated by the studies of long tree-ring series (Scots pine, *Pinus sylvestris*) from northern Fennoscandia by Briffa *et al.* (1990, 1992a). In order to produce a long dendroclimatic reconstruction extending over 1500 yr, Briffa *et al.* (1990) constructed a composite chronology made up of many overlapping cores which varied in their individual length, from less than 100 to more than 200 yr (Figs. 10.11 and 10.13). In the shorter segments, the growth function is significant over the entire segment length, but in longer segments the growth factor becomes less significant (see Fig. 10.9, upper panel). In Briffa *et al.* (1990) each segment was standardized individually (the procedure used in almost all dendroclimatic studies), in this case by the use of a cubic spline function that retains variance at periods less than ~2/3 of the record length. Thus, in a 100 yr segment, variance at periods >66 yr would be removed, whereas in a 300 yr segment, variance at periods up to 200 yr would be retained. All standardized cores were then averaged together, producing the record shown in Fig. 10.12c. This shows considerable interannual to decadal scale variability, but little long-term low frequency variability. In fact, as the mean segment length varies over time (Fig. 10.11) so too will the low frequency variance represented in the composite series.

In Briffa *et al.* (1992a) the standardization procedure was revised by first aligning all core segments by their relative age, then averaging them (i.e., all values of the first year in each segment (t_1) were averaged, then all values of t_2 etc. . . . to t_n). This assumes that in each segment used, t_1 was at, or very close to, the center of the tree and that there is a tree-growth function (dependent only on biological age) that is

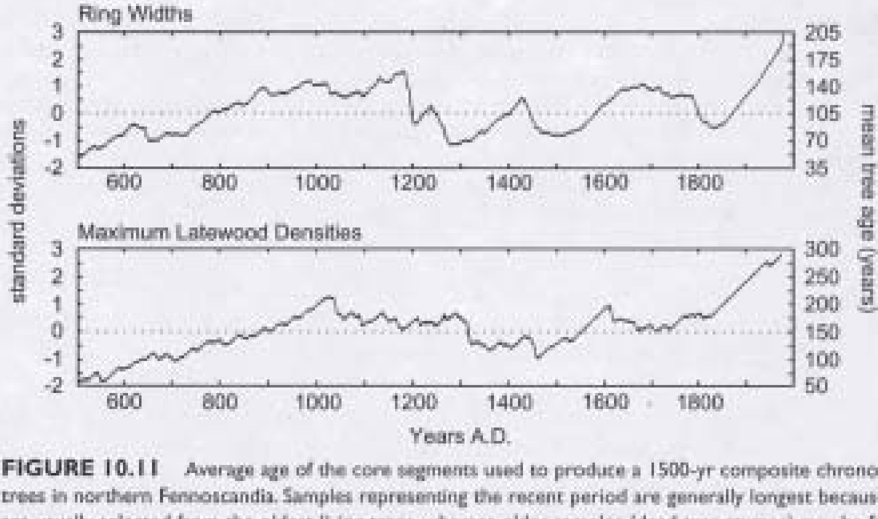

FIGURE 10.11 Average age of the core segments used to produce a 1500-yr composite chronology from trees in northern Fennoscandia. Samples representing the recent period are generally longest because samples are usually selected from the oldest living trees, whereas older samples (dead tree stumps) may be from trees of any age (Briffa et al., 1992a).

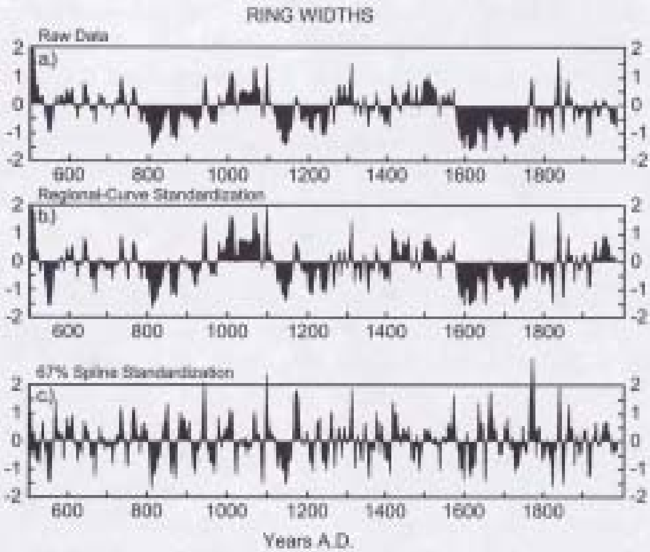

FIGURE 10.12 Ring-width data from trees in northern Fennoscandia showing **(a)** the mean indices without any standardization, **(b)** indices derived from standardization using the regional curve standardization, or **(c)** indices derived from standardization using a cubic-smoothing spline function (see text for discussion) (Briffa et al., 1992a).

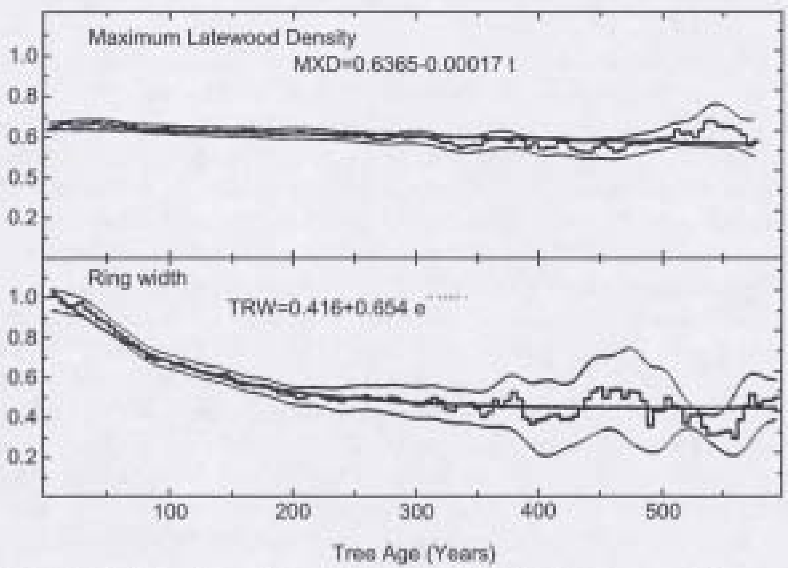

FIGURE 10.13 Regional standardization curve (RCS) of tree-ring samples from northern Fennoscandia based on a least squares fit to the mean values of all series, after they were aligned according to their biological age (Briffa et al., 1992a). Ring-width data commonly has a more pronounced growth function than maximum latewood density data.

common to all samples. The resulting "regional curve" provided a target for deriving a *mean* growth function, which could be applied to all of the individual core segments regardless of length (Fig. 10.13). Averaging together the core segments, standardized in this way by the regional curve, produced the record shown in Fig. 10.12b. This has far more low frequency information than the record produced from individually standardized cores (Fig. 10.12c) and retains many of the characteristics seen in the original data (Fig. 10.12a). From this series, low growth from the late 1500s to the early 1800s is clearly seen, corresponding to other European records that record a "Little Ice Age" during this interval. Also seen is a period of enhanced growth from A.D. ~950–1100, during a period that Lamb (1965) characterized as the "Medieval Warm Epoch." It is apparent from a comparison of Figs. 10.12b and 10.12c that any conclusions drawn about which were the warmest or coldest years and decades of the past can be greatly altered by the standardization procedure employed. All of the high frequency variance of Fig. 10.12c is still represented in the record produced by regional curve standardization but potentially important climatic information at lower frequencies is also retained. The problem of extracting low frequency climatic information from long composite records made up of many individual short segments is addressed explicitly by Briffa *et al.* (1996) and Cook *et al.* (1995) who refer to this as the "segment length curse"! Although it is of particular concern in dendroclimatology, it is in fact an important problem in all long-term paleoclimatic reconstructions that utilize limited duration records to build up a longer composite series (e.g., historical data).

10.2.4 Calibration of Tree-Ring Data

Once a master chronology of standardized ring-width indices has been obtained, the next step is to develop a model relating variations in these indices to variations in climatic data. This process is known as calibration, whereby a statistical procedure is used to find the optimum solution for converting growth measurements into climatic estimates. If an equation can be developed that accurately describes instrumentally observed climatic variability in terms of tree growth over the same interval, then paleoclimatic reconstructions can be made using only the tree-ring data. In this section, a brief summary of the methods used in tree-ring calibration is given. For a more exhaustive treatment of the statistics involved, with examples of how they have been used, the reader is referred to Hughes *et al.* (1982) and Fritts *et al.* (1990).

The first step in calibration is selection of the climatic parameters that primarily control tree growth. This procedure, known as response function analysis, involves regression of tree-ring data (the predictand) against monthly climatic data (the predictors, usually temperature and precipitation) to identify which months, or combination of months, are most highly correlated with tree growth. Usually months during and prior to the growing season are selected but the relationship between tree growth in year t_0, t_{-1} may also be examined as tree growth in year t_0 is influenced by conditions in the preceding year. If a sufficiently long set of climatic data is available, the analysis may be repeated for two separate intervals to determine if the relationships are similar in both periods (Fig. 10.14). This then leads to selection of the

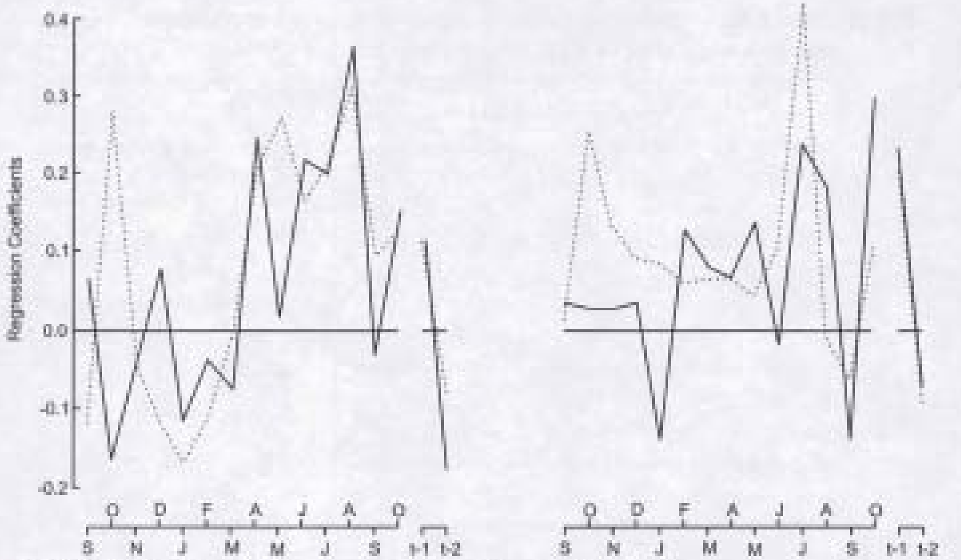

FIGURE 10.14 Results of a response function analysis to determine the pattern of growth response in Scots pine (from northern Fennoscandia) to temperature in the months September (in year t_{-1}) to October in year t_0. Values plotted are coefficients from multiple regression analysis of tree-ring maximum latewood density (left) and ring widths (right) in relation to instrumentally recorded temperatures in the region. Two periods were used for the analysis to examine the stability of the relationships: 1876–1925 (dotted lines) and 1926–1975 (solid line). In addition to the monthly climate variables, the ring width values from the two preceding years (t_{-1} and t_{-2}) are also included to assess the importance of biological preconditioning of growth in one summer by conditions in preceding years. The analysis shows that maximum latewood density is increased by warm conditions from April–August of year t_0, and a similar, but less strong signal is found in tree-ring widths. Growth in the previous year is also important (Briffa et al., 1990).

month or months on which the tree-ring records are dependent, and which the tree rings can therefore be expected to usefully reconstruct. For example, by this approach Jacoby and D'Arrigo (1989) found that the ring widths of white spruce (*Picea glauca*) at the northern treeline in North America are strongly related to mean annual temperature, whereas Briffa *et al.* (1988, 1990) showed that summer temperature (April/May to August/September) is the major control on ring widths and maximum latewood density in Scots pine (*Pinus sylvestris*) in northern Fennoscandia.

Once the climatic parameters that influence tree rings have been identified, tree-ring data can be used as predictors of these conditions. Various levels of complexity may be involved in the reconstructions (Table 10.1). The basic level uses simple linear regression in which variations in growth indices at a single site are related to a single climatic parameter, such as mean summer temperature or total summer precipitation. 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 was reconstructed from a single chronology located near Churchill, Manitoba. Similarly, Cleaveland and Duvick (1992) reconstructed July hydrological drought indices for Iowa from a single regional chronology which was an average

TABLE 10.1 Different Levels of Complexity in the Methods Used to Determine Relationship Between Tree Ring Parameter and Climate

Level	Number of variables of: Tree growth	Climate	Main statistical procedures used
I	1	1	Simple linear regression analysis
IIa	*n*	1	Multiple linear regression (MLR)
IIb	*n*P	1	Principal components analysis (PCA)
IIIa	*n*P	*n*P	Orthogonal spatial regression (PCA and MLR)
IIIb	*n*P	*n*P	Canonical regression analysis (with PCA)

1 = a temporal array of data.

n = a spatial and temporal array of data.

*n*P = number of variables after discarding unwanted ones from PCA.

From Bradley and Jones (1995).

of 17 site chronologies of the same species. More commonly, multivariate regression is used to define the relationship between the selected climate variables and a set of tree-ring chronologies within a geographical area where there is a common climate signal. The tree ring data may include both ring widths and density values. Equations that relate tree rings (the predictors) to climate (the predictand) are termed transfer functions with a basic form (assuming linear relationships) of:

$$y_t = a_1 x_{1t} + a_2 x_{2t} + a_3 x_{3t} \ldots + a_m x_{mt} + b + e_t$$

where y_t is the climate parameter of interest (for year t); $x_{1t}, \ldots, x_{mt}$ are tree-ring variables (e.g., from different sites) in year t; $a_1, \ldots, a_m$ are weights or regression coefficients assigned to each tree-ring variable, b is a constant, and e_t is the error or residual. In effect, the equation is simply an expansion of the linear equation, $y_t = ax_t + b + e_t$, to incorporate a larger number of terms, each additional variable accounting for more of the variance in the climate data (Ferguson, 1977). Theoretically, it would be possible to construct an equation to predict the value of y_t precisely. However, adding too many coefficients simply widens the confidence limits about the reconstruction estimates so that eventually the uncertainty become so large that the reconstruction is virtually worthless. What is needed is an equation that uses the minimum number of tree ring variables to account for the maximum amount of variance in the climate record. Commonly, the procedure of stepwise multiple regression is used to achieve this aim (Fritts, 1962, 1965). From a matrix of potentially influential predictor variables, the one that accounts for most of the climate variance is selected; next, the predictor that accounts for the largest proportion of the remaining climatic variance is identified and added to the equation, and so on in a stepwise manner. Tests of statistical significance, as each variable is selected, enable the procedure to be terminated when a further increase in the number of variables in the equation results in an insignificant increase in variance explanation. In this way, only the most important variables are selected, objectively, from the large array of potential predictors. An example of this approach is the reconstruction of drought in Southern California by Meko *et al.* (1980).

A major problem with stepwise regression is that intercorrelations between the tree-ring predictors can lead to instability in the prediction equation. In statistical terms this is referred to as multicollinearity. To overcome this, a common procedure is to transform the predictor variables into their principal modes (or empirical orthogonal functions, EOFs) and use them as predictors in the regression procedure. Principal components analysis involves mathematical transformations of the original data set to produce a set of orthogonal (i.e., uncorrelated) eigenvectors that describe the main modes of variance in the multiple parameters making up the data set (Grimmer, 1963; Stidd, 1967; Daultrey, 1976). Each eigenvector is a variable that expresses part of the overall variance in the data set. Although there are as many eigenvectors as original variables, most of the original variance will be accounted for by only a few of the eigenvectors. The first eigenvector represents the primary mode of distribution of the data set and accounts for the largest percentage of its variance (Mitchell *et al.*, 1966). Subsequent eigenvectors account for lesser and lesser amounts of the remaining variance (Fig. 10.15). Usually only the first few eigenvectors are considered, as they will have captured most of the total variance. The value or amplitude of each eigenvector varies from year to year, being highest in the year when that particular combination of conditions represented by the eigenvector is most apparent. Conversely, it will be lowest in the year when the inverse of this combination is most apparent in the data. Eigenvector amplitudes can then be used as orthogonal predictor variables in the regression procedure, generally accounting for a higher proportion of the dependent data variance with fewer variables than would be possible using the "raw" data themselves. A time series of eigenvector values (amplitudes) is referred to as a principal component (PC), the dominant eigenvector being PC1, the next most common pattern PC2, etc.

Apart from reducing the number of potential predictors, principal components analysis also simplifies multiple regression considerably. It is not necessary to use the stepwise procedure because the new potential predictors are all orthogonal. This approach was used in the reconstruction of July drought in New York's Hudson Valley by Cook and Jacoby (1979). They selected series of ring-width indices from six different sites, and calculated eigenvectors of their principal characteristics. These were then used as predictors in a multiple regression analysis with Palmer Drought Severity Indices (Palmer, 1965)[12] as the dependent variable. The resulting equation, based on climatic data for the period 1931–1970, was then used to reconstruct Palmer indices back to 1694 when the tree-ring records began (Fig. 10.16). This reconstruction showed that the drought of the early 1960s, which affected the entire northeastern United States, was the most severe the area has experienced in the last three centuries.

[12] Palmer indices are measures of the relative intensity of precipitation abundance or deficit and take into account soil-moisture storage and evapotranspiration as well as prior precipitation history. Thus they provide, in one variable, an integrated measure of many complex climatic factors. They are scaled from +4 or more (extreme wetness) to −4 or less (extreme drought) and are widely used by agronomists in the United States as a guide to climatic conditions relevant to crop production.

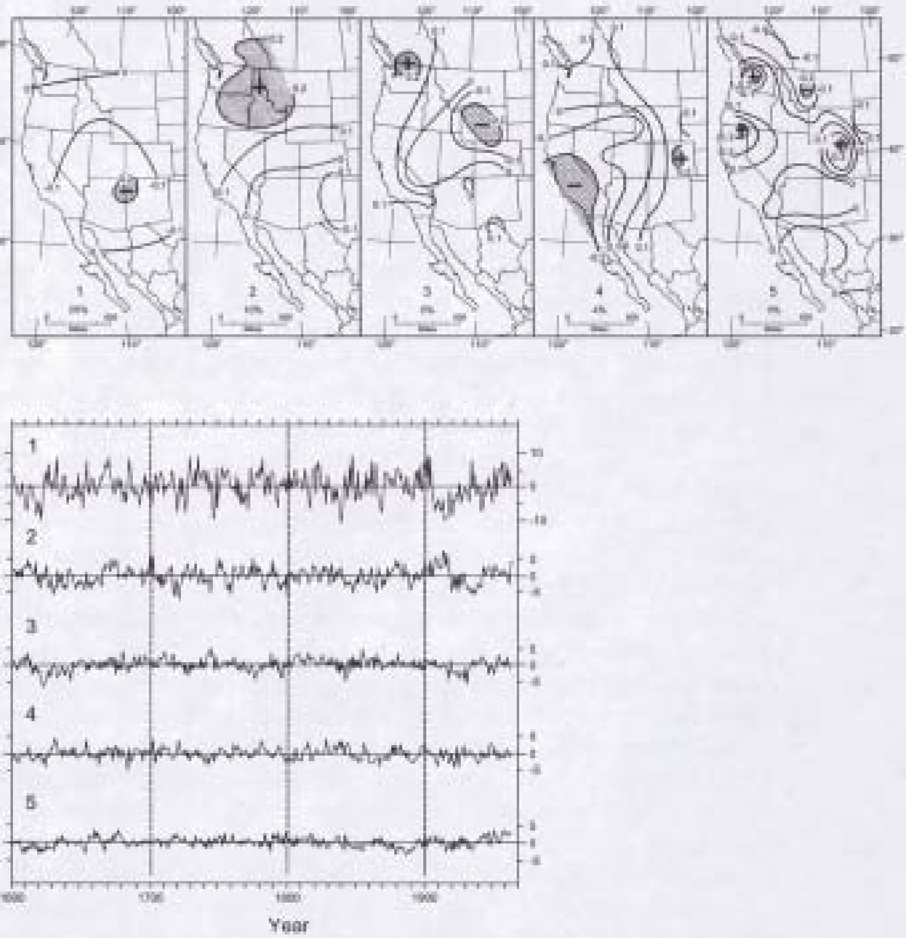

FIGURE 10.15 The first five eigenvectors of tree-ring widths based on a network of 65 chronologies distributed across the western United States, northern Mexico, and southwestern Canada. These represent the major patterns of growth anomalies in the region. Eigenvector 1 accounts for 25% of the overall variance; each subsequent eigenvector accounts for progressively less. The lower diagram shows the relative amplitudes of these five eigenvectors since A.D. 1600 (i.e., the principal components, [PCs] 1–5). These and other PCs were used in canonical regressions with gridded temperature, precipitation, and pressure data, first to calibrate the tree-ring data and then to reconstruct maps of each climatic parameter back to A.D. 1600 (Fritts, 1991).

Simple univariate transfer functions express the relationship between one climatic variable and multiple tree-ring variables. A more complex step is to relate the variance in multiple growth records to that in a multiple array of climatic variables (e.g., summer temperature over a large geographical region) (Table 10.1). To do so, each matrix of data (representing variations in both time and space) is converted into its principal modes or eigenvectors; these are then related using canonical regression or orthogonal spatial regression techniques (Clark, 1975; Cook *et al.*, 1994). These

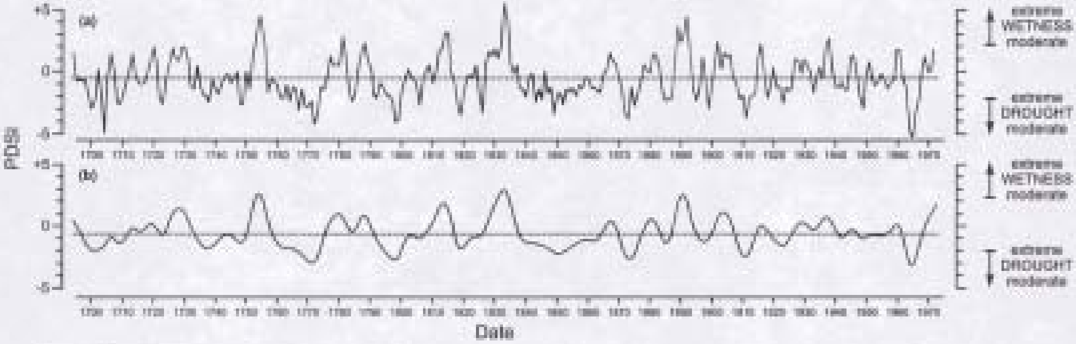

FIGURE 10.16 July Palmer drought indices for the Hudson Valley, New York, from 1694 to 1972 reconstructed from tree rings. **(a)** Unsmoothed estimates; **(b)** a lowpass filtered version of the unsmoothed series that emphasizes periods of ≥10 yr (Cook and Jacoby, 1979).

techniques involve identifying the variance that is common to individual eigenvectors of the two different data sets and defining the relationship between them. The techniques are important in that they allow spatial arrays (maps) of tree-ring indices to be used to reconstruct maps of climatic variation through time (Fritts *et al.*, 1971; Fritts, 1991; Fritts and Shao, 1992; Briffa *et al.*, 1992b).

The most comprehensive work of this sort is that of Fritts (1991). Ring-width indices from 65 sites across western North America (i.e., 65 variables) were transformed into eigenvectors, where each one represented a spatial pattern of growth covariance among the sites (Fig. 10.15). The first ten eigenvectors accounted for 58% of the joint space-time variance of growth anomaly over the site network. Eigenvectors were also derived for seasonal pressure data at grid points over an area extending from the eastern Pacific (100° E) to the eastern U.S. (80° W) and from 20 to 70° N; the first three eigenvectors of pressure accounted for ~56% of variance in the data. Using amplitudes of all these eigenvectors for the years common to both data sets (1901–1962), canonical weights were computed for the growth eigenvectors to give maximum correlations with pressure anomalies. Amplitudes of these weighted eigenvectors were then used as predictors of normalized pressure departures at each point in the pressure grid network, by applying the canonical weights to the standardized ring-width data (the level IIIb approach in Table 10.1). This resulted in estimates of pressure anomaly values at each point in the grid network, for each season, for each year of the ring-width network. Maps of mean pressure anomaly could thus be produced for any interval by simply averaging the individual anomaly values (Fritts and Shao, 1992). The same procedure was used for a network of gridded temperature and precipitation data across the United States. Figure 10.17 gives the mean pressure, temperature and precipitation anomalies for 1602–1900 for each gridded region compared to twentieth century means. This ~300 yr interval spans much of what is commonly referred to as the "Little Ice Age" and it is interesting that the reconstruction suggests there was a stronger ridge over western North America, with higher pressure over Alaska during this time. This was associated with

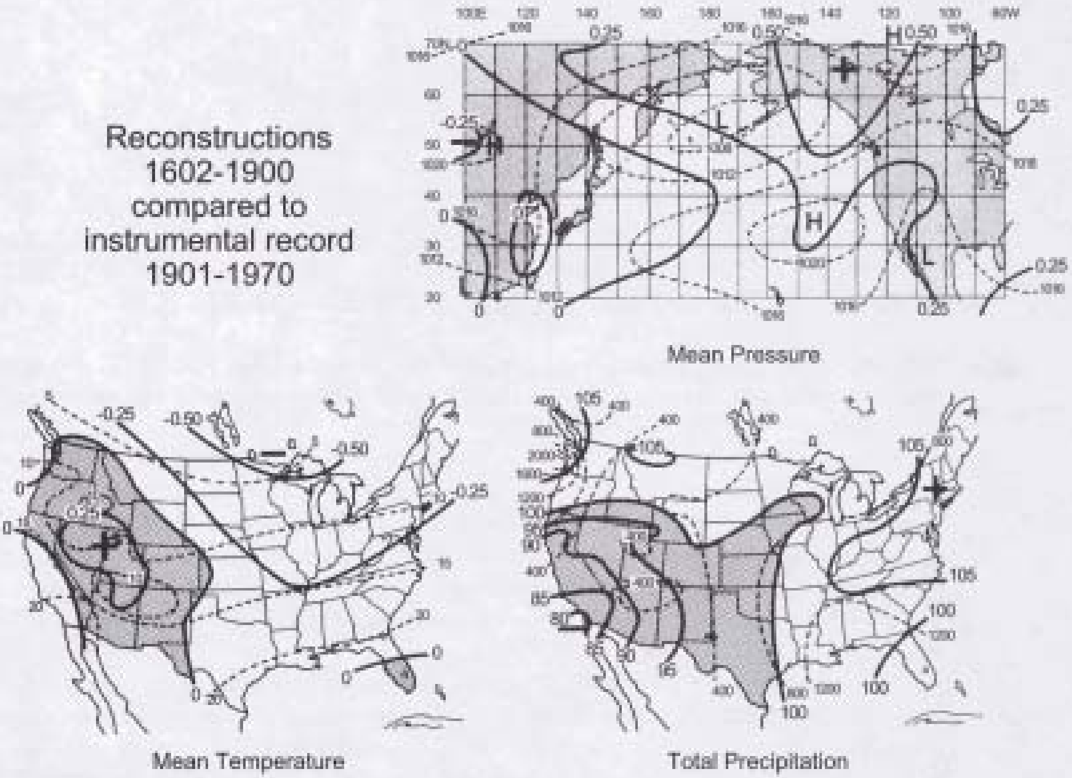

FIGURE 10.17 Anomalies (from 1901–1970 mean values) of mean sea-level pressure, temperature, and precipitation (in % of mean) for the period 1602–1900. Values were reconstructed from a 65-chronology network of tree-ring width data (see Fig. 10.15). Shaded areas on the temperature and precipitation maps are warm or dry anomalies, respectively (Fritts, 1991).

higher temperatures and lower precipitation over much of the western and southwestern United States, but cooler and wetter conditions in the east and northeast. Precipitation was also higher in the Pacific Northwest, presumably reflecting more frequent depressions affecting this area. In effect, the reconstruction points to an amplification of the Rossby wave pattern over North America, with an increase in cold airflow from central Canada into the central and eastern U.S. (Fritts, 1991; Fritts and Shao, 1992). Reconciling these reconstructions with evidence for extensive glacier advances in the Rockies and other mountain ranges of western North America during this period is clearly very difficult (Luckman, 1996).

Related procedures have been used by other workers. Briffa *et al.* (1988) for example, used orthogonal spatial regression to reconstruct April–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 were first reduced to their principal components. Only significant components in each set were retained. Each retained PC of climate was then regressed in turn against the set of retained densitometric PCs. This procedure can be thought of as repeating the level IIb approach (Table 10.1) *m* times, where *m* is the number of retained climate PCs. Having found all of the significant regression coefficients, the set of equations relating the climate PCs to the tree-growth PCs were then transformed back to original variable space, resulting in an equation for each temperature location in terms of all the densitometric chronologies. A similar approach was used by Schweingruber *et al.* (1991) and Briffa and Schweingruber (1992) to derive temperature reconstructions for Europe, and by Briffa *et al.* (1992b), who derived temperature anomaly maps for the western United States from a network of tree-ring density data.

Fig. 10.18 shows some examples of these reconstructions in comparison with instrumental data for the same years, providing a qualitative impression of how good the pre-instrumental reconstructions might be. In fact, verification statistics over an independent period are generally good, providing a more quantitative assessment that earlier reconstructions are likely to be reliable. Of particular interest is the reconstruction of temperatures in the early nineteenth century, around the time of the eruption of Tambora (Fig. 10.19). Tambora (8° S, 118° E) exploded in April 1815; it is considered to have been the largest eruption in the last thousand years if not the entire Holocene (Rampino and Self, 1982; Stothers, 1984). Contemporary accounts

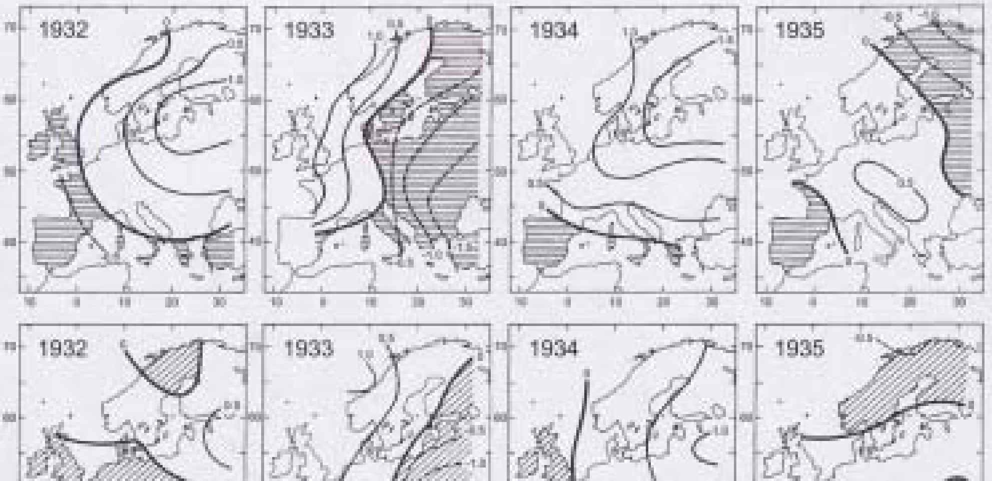

FIGURE 10.18 Observed April–September mean temperature anomalies in the summers of the early 1930s (expressed as departures from the 1951–1970 mean) compared with (bottom panel) the corresponding reconstruction based on a tree-ring density network made up of 37 chronologies distributed across the region. (Schweingruber *et al.*, 1991).

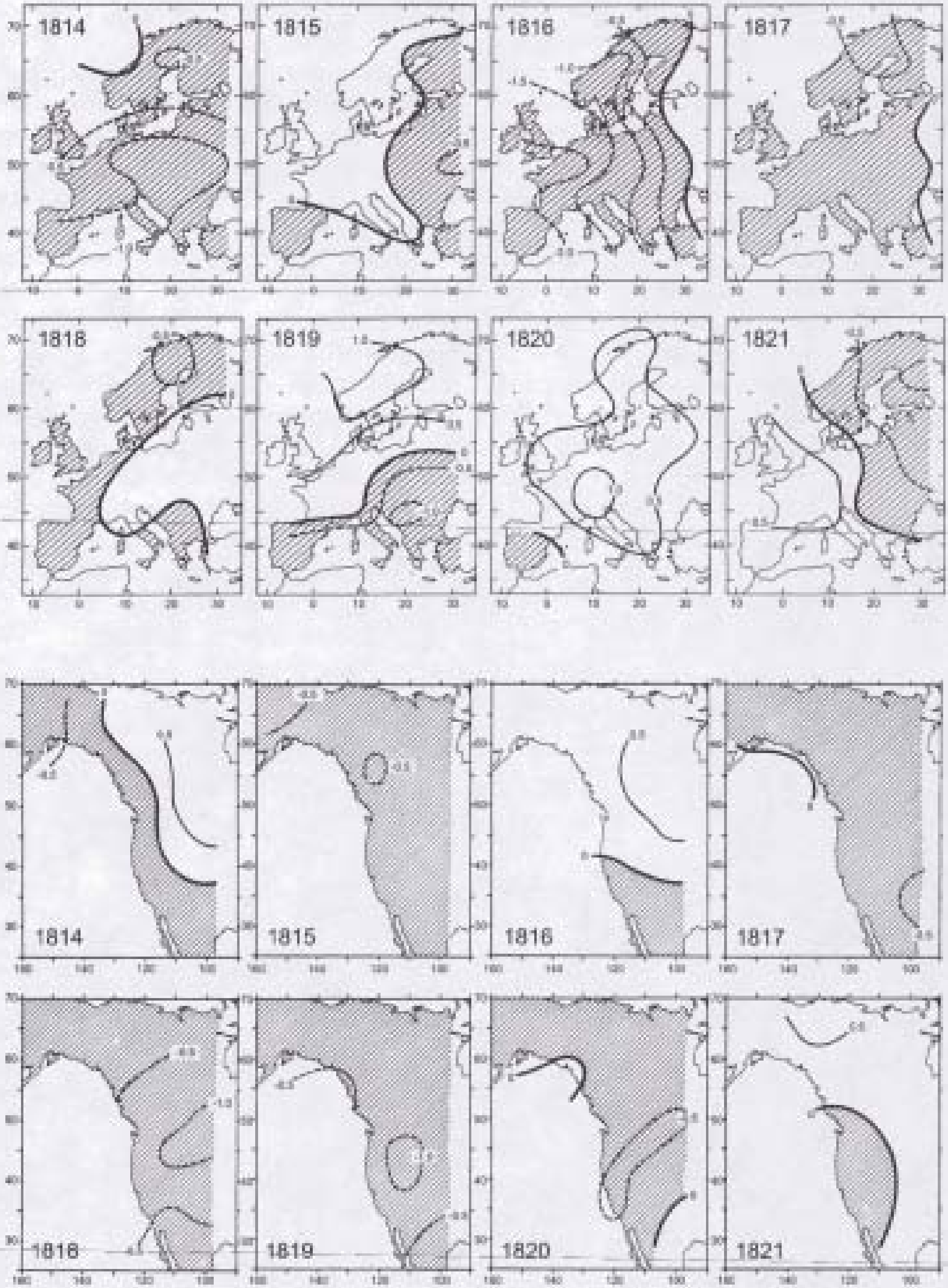

FIGURE 10.19 Reconstructed summer (April–September) temperature anomalies (from 1951–1970 means) for western Europe and western North America for the early nineteenth century based on a 37 site tree-ring density network in Europe and a 53 site tree-ring density network in North America. Conditions were cold in both regions in the years following the eruption of Tambora, though 1816 itself was mostly warm in western North America (Schweingruber *et al.*, 1991).

document the severe climatic conditions that were experienced in western Europe and in the eastern United States in the following year, which became known as "the year without a summer" (Harington, 1992). The dendroclimatic reconstruction for summer temperatures (April–September) in western Europe indeed show extremely cold conditions in 1816 with cold summers also in 1817 and 1818 over most of western Europe (and northwestern Russia; Shiyatov, 1996) continuing into 1819 in southern Europe. Over most of the western United States and Alaska, 1816 was not cold, but the following four summers were uniformly cooler than the 1951–1970 reference period. The coldest summer in the last few centuries in the western United States (1601) was also associated with an eruption, probably Huaynaputina in Peru (Briffa *et al.*, 1992b, 1994). Temperatures across the region averaged 2.2 °C below the 1881–1982 mean as a result of that event.

Before concluding this section on calibration, it is worth noting that tree-ring indices need not be calibrated only with climatic data. The ring-width variations contain a climatic signal and this may also be true of other natural phenomena that are in some way dependent on climate. It is thus possible to calibrate such data directly with tree rings and to use the long tree-ring records to reconstruct the other climate-related series. In this way, dendroclimatic analysis has been used to reconstruct runoff records (Stockton, 1975; Stockton and Boggess, 1980) and lake-level variations (Stockton and Fritts, 1973). Some of these applications are discussed in more detail in Section 10.3.3.

10.2.5 Verification of Climatic Reconstructions

An essential step in dendroclimatic analysis (indeed in all paleoclimatic studies) is to test or verify the paleoclimatic reconstruction in some way. The purpose of verification is to test if the transfer function model (derived from data in the calibration period) is stable over time, usually by comparing part of the reconstruction with independent data from a different period. Inevitably, when the prediction estimates are tested against an independent data set the amount of explained variance will almost always be less than in the calibration period. To quantify how good the reconstruction is, in comparison with independent data, various statistical tests are generally performed (Gordon, 1982; Fritts *et al.*, 1990; Fritts, 1991, Appendix 1). These statistics then provide some level of confidence in the rest of the reconstruction; the performance of the transfer function over the verification period is the best guide to the likely quality of the reconstruction for periods when there are simply no instrumental data.

Two approaches to verification are generally adopted. First, when calibrating the tree-ring data, very long instrumental records for the area are sought. Only part of these records are then used in the calibration, leaving the remaining early instrumental data as an independent check on the dendroclimatic reconstruction. If the reconstruction is in the form of a map, several records from different areas may be used to verify the reconstruction, perhaps indicating geographical regions where the reconstructions appear to be most accurate (Briffa *et al.*, 1992b). This approach is difficult in some areas where tree-ring studies have been carried out (e.g., the western United States and northern treelines) because these are areas with very few early instrumental records (Bradley, 1976). Dendroclimatic studies in western Europe (Serre-Bachet *et al.*, 1992; Briffa and Schweingruber, 1992) can be more exhaustively tested because of the much longer instrumental records in that area. Indeed, it is sometimes possible to conduct two calibrations, with both tree-ring and climatic data from different time periods and to compare the resulting dendroclimatic reconstructions for earlier periods derived independently from the two data sets (Briffa *et al.*, 1988). This provides a vivid illustration of the stability of the derived paleoclimatic reconstructions (Fig. 10.20).

A second approach is to use other proxy data as a means of verification. This may involve comparisons with historical records or with other climate-dependent phenomena such as glacier advances (LaMarche and Fritts, 1971b) or pollen variations in varved lake sediments (Fritts *et al.*, 1979) etc. It may even be possible to use an independent tree-ring data set to compare observed growth anomalies with those expected from paleoclimatic reconstructions. Blasing and Fritts (1975), for exam-

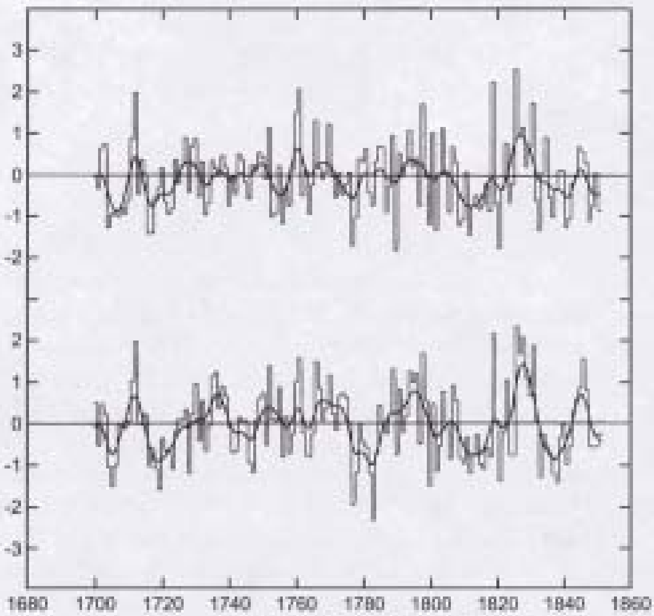

FIGURE 10.20 Two reconstructions of northern Fennoscandinavian temperatures for July–August using the same transfer function model, but with calibration based on 1852–1925 in the upper diagram, and 1891–1964 in the lower diagram. Ordinate axis is in standard deviation units with one unit approximately equal to 1 °C. The two calibrations accounted for 69% and 56% of the variance of instrumental temperature over the same (calibration) interval, respectively. Both gave statistically significant statistics when verified against data from the other "independent" interval. Thus, both reconstructions could be viewed as statistically reliable. Although the two series are strongly correlated (r = 0.87) there are important differences between them, which warns against overinterpreting the reconstructions for individual years. For example, the lower diagram shows an extreme in 1783 that does not appear in the upper diagram, and even the lower frequency variations show important differences requiring careful examination of the regression weights used for each reconstruction (Briffa et al., 1988).

ple, used a network of trees from an area between northern Mexico and southern British Columbia to reconstruct maps of sea-level pressure anomalies over the eastern Pacific and western North America. A separate temperature-sensitive data set from Alaska and the Northwest Territories of Canada was then used to test the reconstructions. Periods of anomalously low growth in the northern trees were associated with increased northerly airflow as predicted by the pressure reconstructions.

In all verification tests, one is inevitably faced with two questions:

(a) If the verification is poor, does the fault lie with the dendroclimatic reconstruction (and hence the model from which it was derived) or with the proxy or instrumental data used as a test (which may itself be of poor quality and subject to different interpretations)? In such cases, re-evaluation of the tree-ring data, the model, and the test data must be made before a definitive conclusion can be reached.

(b) Is the dendroclimatic reconstruction for the period when no independent checks are possible as reliable as for the period when verification checks can be made? This might seem an insoluble problem but it is particularly important when one considers the standardizing procedure employed in the derivation of tree-growth indices (Section 10.2.3). Errors are most likely to occur in the earliest part of the record, whereas tests using instrumental data are generally made near the end of the tree-growth record (where replication is generally highest and the slope of the standardization function is generally lowest) and least likely to involve the incorporation of large error. The optimum solution is for both instrumental and proxy data checks to be made on reconstructions at intervals throughout the record, thereby increasing confidence in the overall paleoclimatic estimates.

Dendroclimatologists have set the pace for other paleoclimatologists by developing methods for rigorously testing their reconstructions of climate. Many other fields would benefit by adopting similar procedures.

10.3 DENDROCLIMATIC RECONSTRUCTIONS

The following sections provide selected examples of how tree rings have been used to reconstruct climatic parameters in different regions of the world. This is not by any means an exhaustive review and for further information the reader should examine the sections on dendroclimatology in Bradley and Jones (1995), Jones *et al.* (1996), and Dean *et al.* (1996).

10.3.1 Temperature Reconstruction from Trees at the Northern Treeline

Many studies have shown that tree growth at the northern treeline is limited by temperature; consequently, both tree-ring widths and density provide a record of past variations in temperature. A 1500-yr long summer temperature reconstruction using Scots pine (*Pinus sylvestris*) from northern Scandinavia has already been discussed (see Fig. 10.12). Even longer records may be possible because in this region logs of Holocene age have been dredged from lakes and bogs at or beyond the present treeline. By cross-dating these samples, it should eventually be possible to construct a well-replicated chronology extending back over most of the Holocene (Zetterberg *et al.*, 1995, 1996). Similarly, in the northern Ural mountains of Russia, the dendroclimatic record of living trees has been extended back over 1000 yr by overlapping cores from dead larch trees (*Larix sibirica*) found close to or above present treeline (Graybill and Shiyatov, 1992; Briffa *et al.*, 1995) (see Section 8.2.1). Ring-width and density studies of these samples show warm conditions in the fourteenth and fifteenth centuries, declining to uniformly low summer temperatures in the sixteenth and seventeenth centuries. There is little similarity between Fennoscandian and northern Urals data prior to ~1600, after which time both records show a gradual increase in temperatures through the 20th century. In fact, in the northern Urals, the 20th century (1901–1990) is the warmest period since (at least) 914 AD, although this does not appear to be the case in northern Scandinavia. Care is needed in the interpretation of these apparent long-term changes because of the number of cores and the individual core segment lengths contributing to the overall record vary over time (Briffa *et al.*, 1996). In the northern Urals data set, the replication (sample depth) and mean core length are at a minimum from ~1400–1650 and ~1500–1700, respectively. For similar reasons, reliability of the mean chronology is poor before ~1100. The overall summer temperature reconstruction is therefore very sensitive to the standardization procedure used to correct for growth effects in each core segment; quite different temperature reconstructions can be produced depending on whether cores are individually standardized (by a cubic spline function) or if a regional curve standardization method is employed (Fig. 10.21). This again points to the dangers implicit in long-term paleotemperature reconstructions built up from multiple short cores when sample replication and mean record length are limited.

Shorter records, derived only from living trees have been recovered all along the Eurasian treeline by Schweingruber and colleagues (Schweingruber and Briffa,

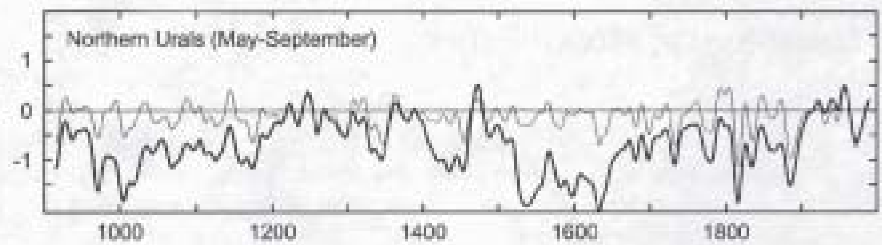

FIGURE 10.21 May–August mean temperature reconstruction for the northern Ural Mountains, Russia, based on tree-ring widths standardized by a cubic-smoothing spline (thin line) and maximum latewood density data standardized by deriving a regional curve for all samples (see Fig. 10.13) (thick line). Data are shown in °C anomalies from the 1951–1970 mean and are smoothed with a 25-yr lowpass filter. The large differences between the two reconstructions indicate that low frequency information is lost in the reconstruction in which only ring widths are used (Briffa *et al.*, 1996). Note that both approaches give comparable reconstructions in the calibration and verification periods, thus providing no *a priori* warning that there might be a problem with the spline-standardized data set.

1996; Vaganov *et al.*, 1996). Both ring-width and density studies of these samples enable a picture of regional variations to be built-up, from Karelia in the west to Chukotka in the east. This shows that the temperature signal is not uniform across the region; for example, from ~80–150° E the early 19th century was uniformly cold, but this cool period was far less pronounced or as persistent farther west, from ~50–80° E. Such variations point to possible shifts in the upper level Rossby wave circulation, which may have been amplified or displaced during this interval. Building up networks of tree-ring data in this way will eventually lead to continental-scale maps of Eurasian temperature anomalies through time (see Fig. 10.19; Schweingruber *et al.*, 1991).

Ring widths in spruce (mainly *Picea glauca*) from the northern treeline of North America contain a strong record of mean annual temperature (Jacoby and D'Arrigo, 1989; D'Arrigo and Jacoby, 1992). Indices averaged over many sites along the treeline, from Alaska to Labrador, reveal a strong correlation with both North American and northern hemisphere temperatures over the last century (Jacoby and D'Arrigo, 1993). Based on these calibrations, long-term variations in temperature have been reconstructed (Fig. 10.22). This indicates low temperatures throughout the seventeenth century with a particularly cold episode in the early 1700s. The eighteenth century was relatively warm, followed by very cold conditions (the coldest of the last 400 years) in the early to mid-1800s. Temperatures then increased to the 1950s, but have since declined. This record is broadly similar to that derived from trees in the northern Ural mountains of Russia, but has less in common with the Fennoscandian summer temperature reconstruction discussed earlier.

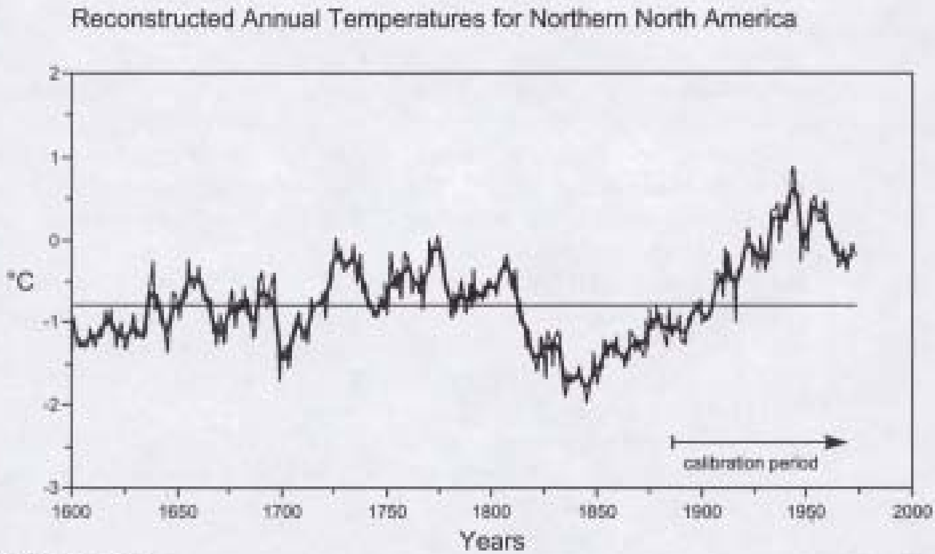

FIGURE 10.22 Reconstruction of annual temperature across northern North America for 1601–1974, based on a set of 7 tree-ring width chronologies from Alaska to Quebec; 5-yr smoothed values also shown by darker line (D'Arrigo and Jacoby, 1992).

Using a different network of trees sampled for maximum latewood density, Schweingruber *et al.* (1993) and Briffa *et al.* (1994) identify distinct regional signals along the North American treeline; by grouping records with common signals they reconstructed summer (April–September) temperatures back to A.D. 1670 for the Alaska/Yukon, Mackenzie Valley, and Quebec/Labrador regions. These series do not capture low frequency variability very well, but do show significant anomalies associated with major volcanic eruptions known from historical records (Jones *et al.*, 1995). However, not all regions are affected in the same way; for example, 1783 (the year that Laki, a major Icelandic volcano erupted) was a very cold summer in Alaska (also noted by Jacoby and D'Arrigo, 1995) but it was relatively warm in the Mackenzie valley. By contrast, 1816 (following the eruption of the Indonesian volcano Tambora in 1815) was exceptionally cold in Quebec and Labrador, but was warm in Alaska/Yukon. Such variations indicate that volcanic aerosols do not always produce uniform climatic responses across the globe and may, in fact, produce strong meridional temperature gradients (Groisman, 1992).

10.3.2 Drought Reconstruction from Mid-latitude Trees

Drought frequency is of critical importance to agriculture in the central and eastern United States. Several dendroclimatic studies have attempted to place the limited observational record in a longer-term perspective by reconstructing precipitation or Palmer Drought Severity Indices (PDSI) for the last few centuries. Cook *et al.* (1992b) used a network of over 150 tree-ring chronologies from across the eastern United States to reconstruct summer PDSI for each of 24 separate regions. On average, 46% of the summer PDSI variance was explained by the tree-ring data. The entire set of reconstructed PDSI for all 24 regions was then subjected to principal components analysis to identify the principal patterns of drought distribution across the entire eastern United States. Six main patterns were identified, with each one corresponding to a different subregion affected by drought (Fig. 10.23). By examining the time series of each principal component, the drought history of each region could be assessed (Fig. 10.24). This revealed that while drought is common in some areas, it is quite rare for drought to extend over many different regions simultaneously. One exception was the period 1814–1822, when severe drought devastated the entire northeastern quadrant of the United States, from New England to Georgia and from Missouri to Wisconsin. This may have been related to circulation anomalies resulting from the major eruption of Tambora in 1815 and/or to four moderate-to-strong El Niño events in quick succession (1814, 1817, 1819, and 1821) (Quinn and Neal, 1992).

Further analysis of drought history in Texas (see Factor 4 in Fig. 10.24) using ring-width chronologies from post oak trees (*Quercus stellata*) has enabled the PDSI to be reconstructed separately for northern and southern Texas (Stahle and Cleaveland, 1988). From these series, the recurrence interval of different levels of drought severity can be calculated (Fig. 10.25). This shows that while the probability of moderate drought (a PDSI of –2 to –3) is high every decade in both north and south Texas, there is a >50% chance of severe drought (PDSI<-4) once a decade in south Texas; in north Texas, however, such droughts are less likely (once in 15 yr).

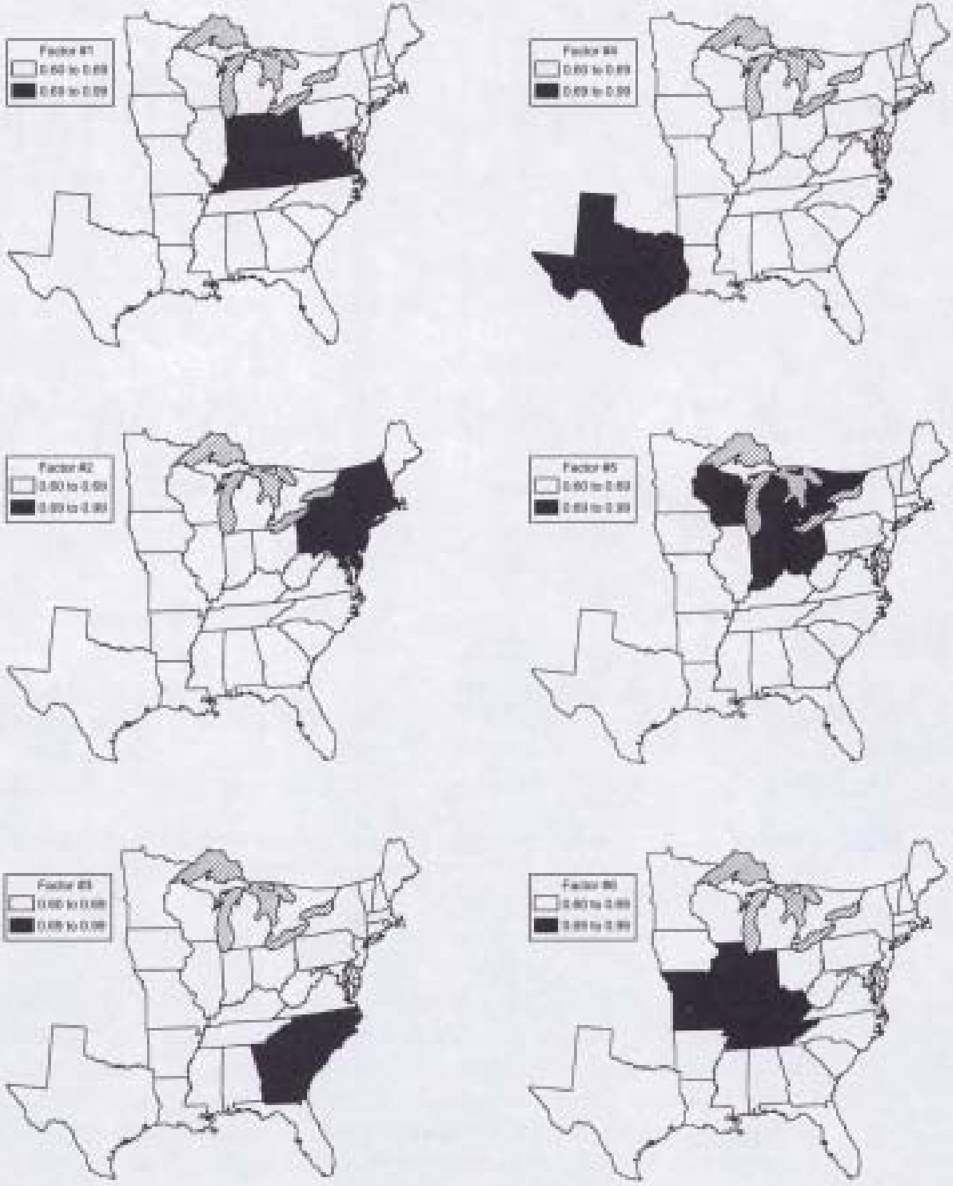

FIGURE 10.23 Major drought patterns across the eastern United States, based on principal components analysis. Shaded areas are where 49% or more of the variance of each drought factor (1 to 6) is explained. The temporal history of pattern 2, for example, will primarily reflect drought in New England, whereas pattern 4 will reflect conditions in Texas (Cook et al., 1992b).

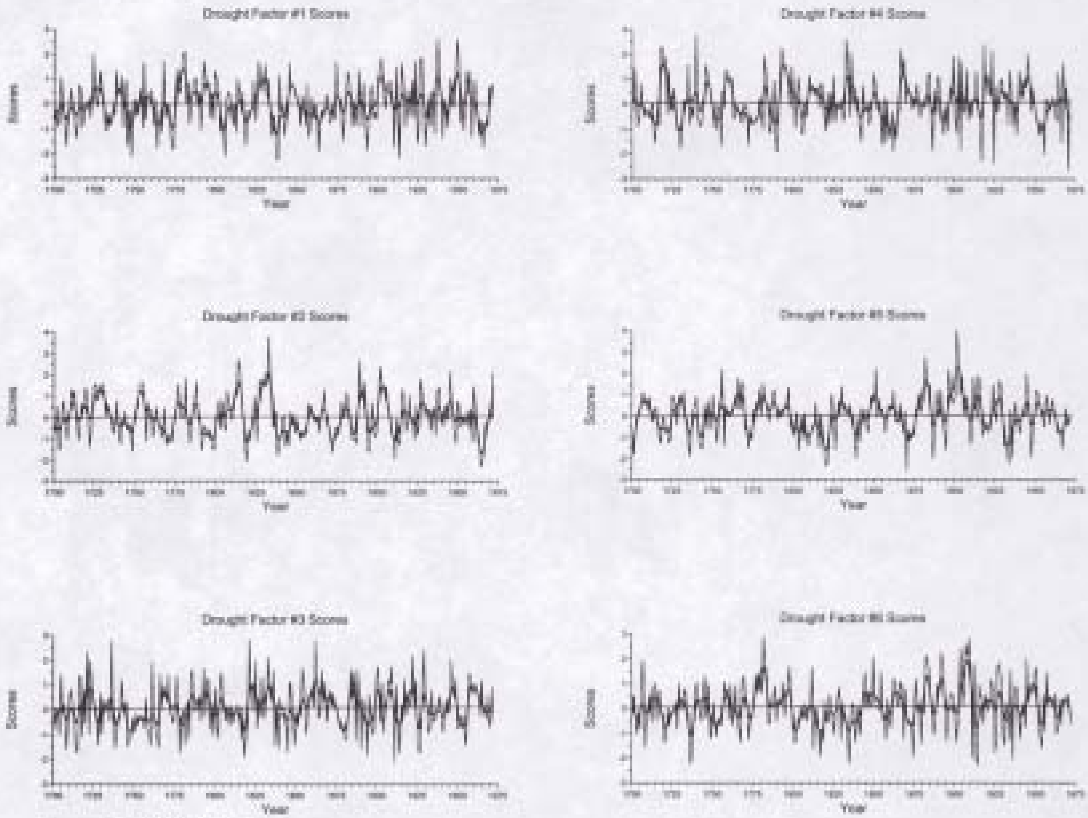

FIGURE 10.24 Reconstructed drought history in the eastern United States. Each series corresponds to a drought pattern shown in Fig. 10.23. The drought in New England in the 1960s (drought factor 2) was unprecedented in the last 270 yr (Cook *et al.*, 1992b).

In the southeastern United States, bald cypress trees (*Taxodium distichum*) growing in swamps have provided a surprisingly good record of precipitation and drought history extending back over 1000 years in some places (Stahle *et al.*, 1985; 1988; Stahle and Cleaveland, 1992). Apparently, tree growth is affected by the water level and water quality in the swamps (oxygenation levels, pH etc.) so tree-ring records show excellent correlation with spring and early summer rainfall in the region. In fact, individual tree-ring chronologies in Georgia and the Carolinas can explain almost as much of the variance in state-wide rainfall averages as a similar network of individual rainfall records (Stahle and Cleaveland, 1992). Long-term reconstructions (>1000 yr) show that non-periodic, multidecadal fluctuations from predominantly wet to dry conditions have characterized the southeastern United States throughout the last millennium (Fig. 10.26). However, since 1650, rainfall in North Carolina appears to have systematically increased, with the period 1956–1984 being one of the wettest periods in the last 370 yr. This trend was

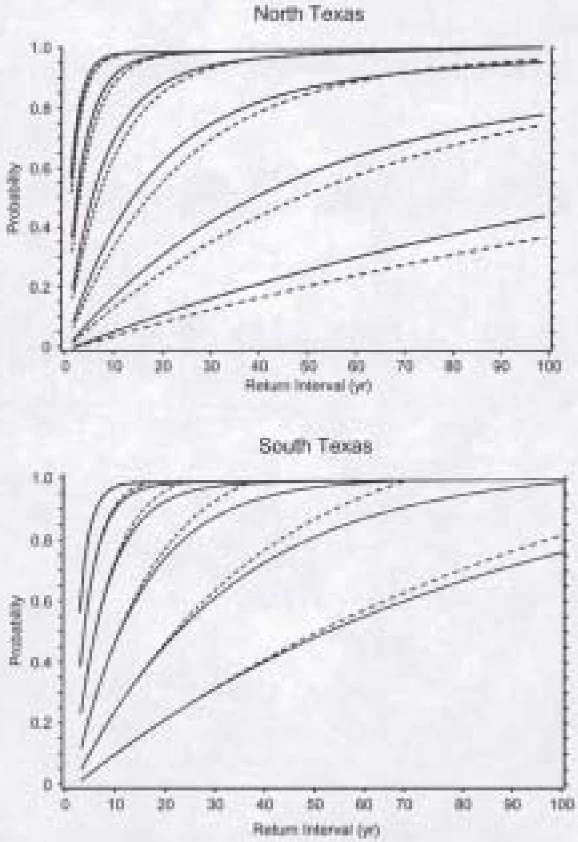

FIGURE 10.25 Return intervals for various levels of drought (solid line) and wetness (dashed line) in north and south Texas. Values shown are for June Palmer Drought Severity Indices ranging from ±1 to ±6, from top left to bottom right. Moderate droughts (second line from the left) have a return interval of a decade or less in both regions, whereas extreme drought (4th line from the left) can be expected with a 50% probability once every 15 years in north Texas, and once every 10 years in south Texas (Stahle and Cleaveland, 1988).

abruptly ended by two remarkable, consecutive drought years (1986 and 1987); only five other comparable two-year droughts have been registered by the bald cypress trees since A.D. 372 (Stahle *et al.*, 1988).

Many meteorological studies have noted the relationship between El Niño/ Southern Oscillation (ENSO) events in the Pacific (or their cold event equivalents, La Niñas) and anomalous rainfall patterns in different parts of the world. These teleconnections result from large-scale displacements of major pressure systems and consequent disruptions of precipitation-bearing storm systems (Ropelewski and Halpert, 1987, 1989; Diaz and Kiladis, 1992). Several attempts have been made to identify an ENSO signal in tree-ring data in order to reconstruct a long-term ENSO index (Lough and Fritts, 1985; Lough, 1992; Meko, 1992; Cleaveland *et al.*, 1992; D'Arrigo *et al.*, 1994). The strongest regional signal in North America is in the

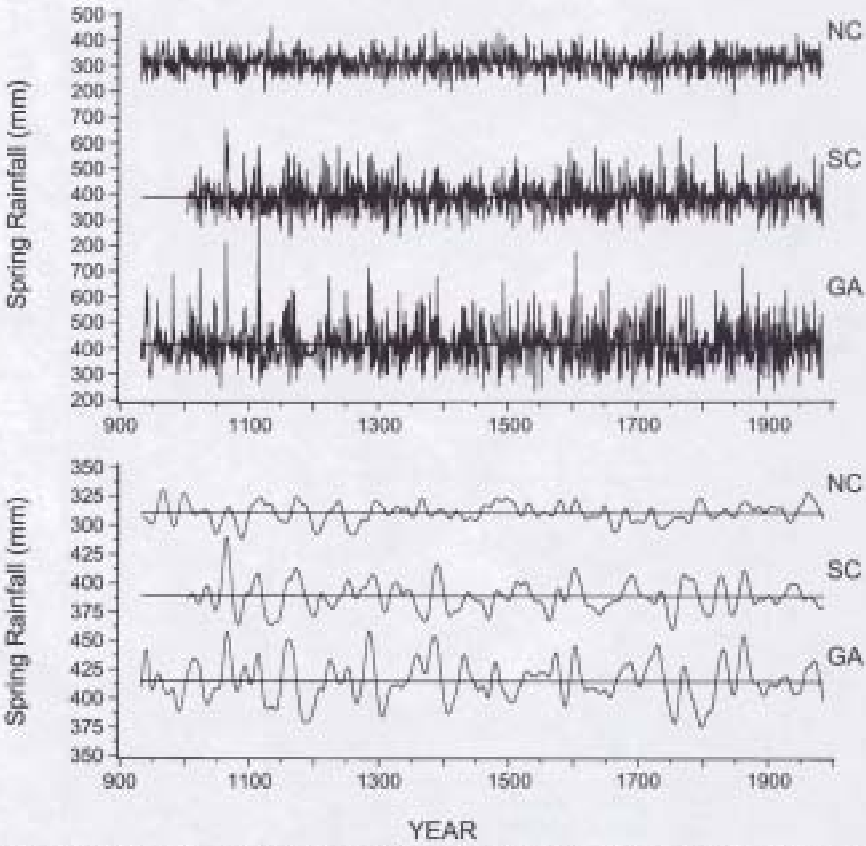

FIGURE 10.26 Reconstructed state-wide rainfall for North Carolina (NC) in April–June and South Carolina and Georgia (SC and GA) in March–June. The upper diagram shows the annual values reconstructed; the lower diagram shows the same data smoothed to emphasize low frequency variations more clearly (with periods >30 yr) (Stahle and Cleaveland, 1992).

southwestern U.S. and northern Mexico, where warm events tend to be associated with higher winter and spring rainfall, which leads to increased tree growth (Fig. 10.27). This led Stahle and Cleaveland (1993) to focus on trees from that area to reconstruct a long-term South Oscillation Index (SOI) back to 1699. Although their results showed considerable skill in identifying many major ENSO events in the past (in comparison with those known from historical records) they estimate that only half of the total number of extremes were clearly defined over the last 300 yr. Similar problems were encountered by Lough and Fritts (1985) using a network of arid-site trees from throughout the western United States. The principal SOI extremes they identify for the last few centuries are not the same as those selected by Stahle and Cleaveland's analysis. This points to the problem of characterizing ENSOs from teleconnection patterns, which are spatially quite variable in relation to both positive and negative extremes of the Southern Oscillation. Consequently, although

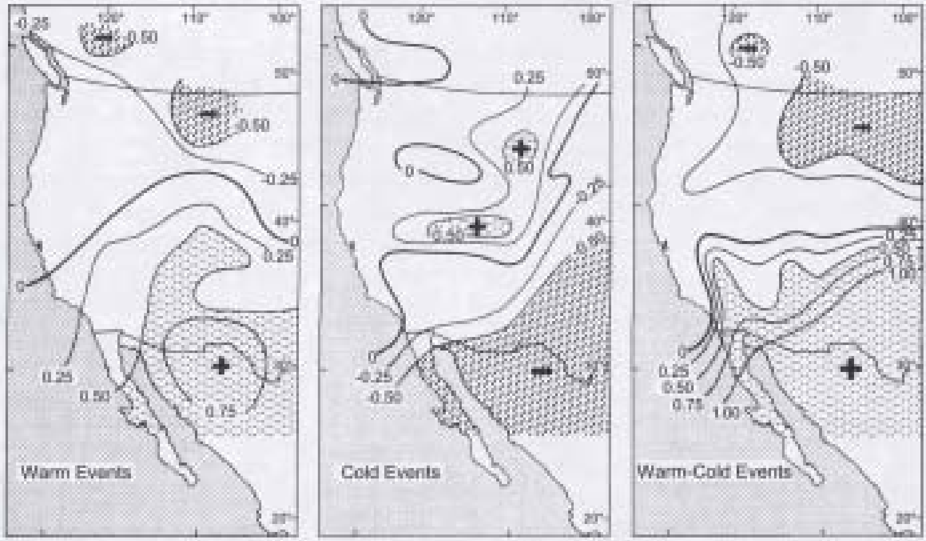

FIGURE 10.27 Tree-ring anomalies across western North America associated with the high and low phases of the Southern Oscillation. The pattern on the left is associated with El Niños; teleconnections lead to heavier winter and spring rainfall in northern Mexico and the U.S. Southwest. By contrast, cold events in the Pacific (La Niñas) are associated with drier conditions and reduced tree growth in the same region (Lough, 1992).

there may be an *overall* signal related to ENSO in one region, precipitation patterns vary enough from event to event that precise, yearly ENSO reconstructions are very difficult (Lough, 1992; D'Arrigo *et al.*, 1994).

One additional factor related to drought is the frequency of fire in some areas (Swetnam, 1993). Fire scars damage the cambium of trees and are clearly visible in tree sections. By building up a chronology of fire history from non-contiguous regions, the frequency of large-scale fires affecting wide areas (related to regional drought episodes) can be identified. In California, fires affecting widely separated groves of giant Sequoia are associated with significant negative winter/spring precipitation anomalies. Over the last 1500 yr, fire frequency was low from A.D. 500–800, reached a maximum ~A.D. 1000–1300, then generally declined. Interestingly, the relationship with temperature in the region is not significant on an annual basis, but over the long-term fire frequency and temperature are positively correlated. Swetnam (1993) attributes this to long-term temperature fluctuations controlling vegetation changes on decadal to century timescales, whereas fire activity from year to year is more related to fuel moisture levels, which are highly correlated with recent precipitation amounts. In the southwestern United States, dry springs and extensive fires are associated with "cold events" in the Pacific (La Niñas) as a result of related circulation anomalies that block moisture-bearing winds from entering the region (Swetnam and Betancourt, 1990). Thus, tree-ring widths and fire occurrence are manifestations of large-scale teleconnections linking sea-surface temperatures in the tropical Pacific to rainfall deficits in the arid Southwest.

10.3.3 Paleohydrology from Tree Rings

Tree rings can be used to reconstruct climate-related phenomena that in some way integrate the effects of the climate fluctuations affecting tree growth. In particular, much work has been devoted to paleohydrological reconstructions involving streamflow. Stockton (1975) was interested in reconstructing long-term variations in runoff from the Colorado River Basin, where runoff records date back only to 1896. As runoff, like tree growth, is a function of precipitation, temperature, and evapotranspiration, both during the summer and in the preceding months, it was thought that direct calibration of tree-ring widths in terms of runoff might be possible. Using 17 tree-ring chronologies from throughout the watershed, eigenvectors of ring-width variation were computed. Stepwise multiple regression analysis was then used to relate runoff over the period 1896–1960 to eigenvector amplitudes over the same interval. Optimum prediction was obtained using eigenvectors of ring width in the growth year (t_0) and also in years t_{-2}, t_{-1}, and t_{+1}, each of which contained climatic information related to runoff in year t_0. In this way an equation accounting for 82% of variance in the dependent data set was obtained; the reconstructed and measured runoff values are thus very similar for the calibration period (Fig. 10.28). The equation was then used to reconstruct runoff back to 1564, using the eigenvector amplitudes of ring widths over this period (Fig. 10.29). The reconstruction indicates that the long-term average runoff for 1564–1961 was ~13 million acre-feet (~16 × 10^9 m^3) over 2 million acre-feet (~2.47 × 10^9 m^3) less than during the period of instrumental measurements. Furthermore, it would appear that droughts were more common in this earlier period than during the last century, and the relatively long period of above average runoff from 1905 to 1930 has only one comparable period (1601–1621) in the last 400 yr. Stockton argues that these estimates, based on a longer time period than the instrumental observations, should be seriously considered in river management plans, particularly in regulating flow through Lake Powell, a large reservoir constructed on the Colorado River. In this

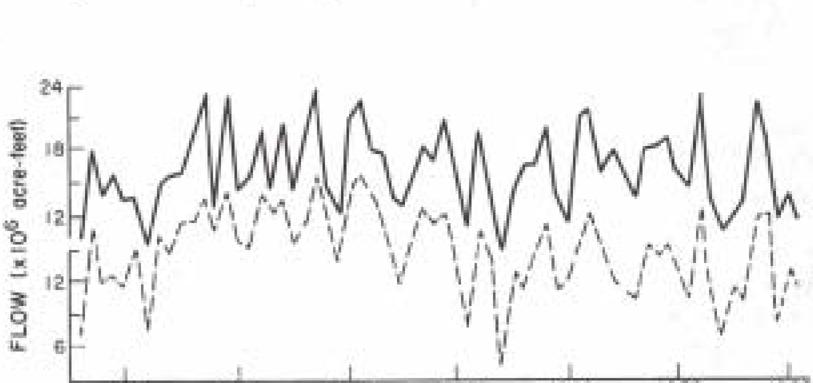

FIGURE 10.28 Runoff in the Upper Colorado River Basin. Reconstructed values (-----) are based on tree-ring width variations in trees on 17 sites in the basin. Actual data, measured at Lee Ferry, Arizona, are shown for comparison (——). Based on this calibration period, an equation relating the two days sets was developed and used to reconstruct the flow of the river back to 1564 (Fig. 10.29) (Stockton, 1975).

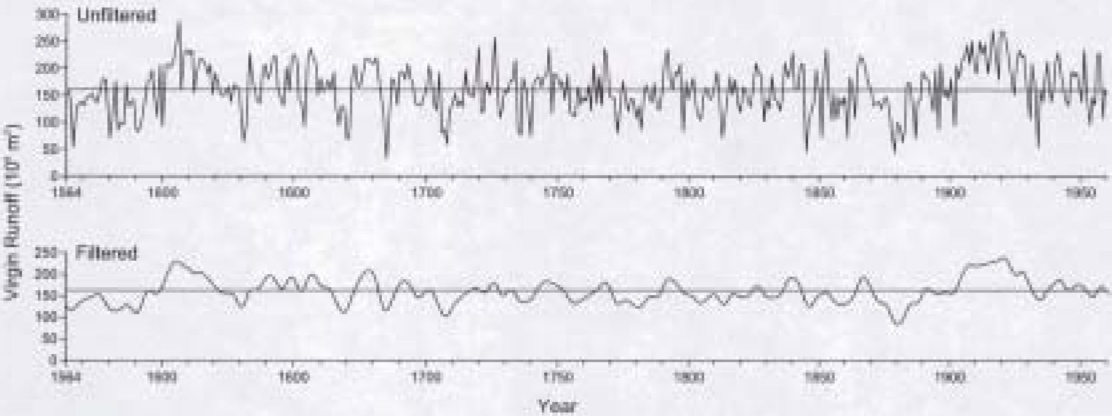

FIGURE 10.29 Annual virgin runoff of the Colorado River at Lee Ferry, as reconstructed using ring-width index variation, calibrated as shown in Fig. 10.28. Growth for each year, and the three following years, was used to estimate water flow statistically. Smooth curve (below) represents essentially a 10-yr running mean. Runoff in the period ~1905–1925 was exceptional when viewed in the context of the last 400 yr (Stockton, 1975).

case, dendrohydrological analysis provided a valuable long-term perspective on the relatively short instrumental record. Similar work has been accomplished by Stockton and Fritts (1973), who used tree-ring eigenvectors calibrated against lake-level data to reconstruct former levels of Lake Athabasca, Alberta, back to 1810 (Figure 10.30). Their reconstruction indicated that although the long-term average lake level is similar to that recorded over the last 40 yr, the long-term variability of lake levels is far greater than could be expected from the short instrumental record. To preserve this pattern of periodic flooding, essential to the ecology of the region, the area is now artificially flooded at intervals that the dendroclimatic analysis suggests have been typical of the last 160 yr.

A few studies have attempted to reconstruct streamflow in humid environments, but they have been less successful than dendroclimatic reconstructions in more arid areas (Jones *et al.*, 1984). In humid regions, periods of low flow are generally more reliably reconstructed than high flows, as trees are more likely to register prolonged drought than heavy precipitation events that may lead to high streamflow and flooding. Cook and Jacoby (1983) reconstructed summer discharge of the Potomac River near Washington, DC, back to 1730 using a set of 5 tree-ring width chronologies; their results indicate that there was a shift in the character of runoff around 1820. Before this time, short-term oscillations about the median flow were common, generally lasting only a few years. After 1820 more persistent, larger amplitude anomalies became common. For example, runoff was persistently below the long-term median from 1850–1873. If such an event were to be repeated in the future with all of the modern demands on water from the Potomac River, the consequences would be extremely severe. Such a perspective on natural variability of the hydrological system is thus invaluable for water supply management and planning.

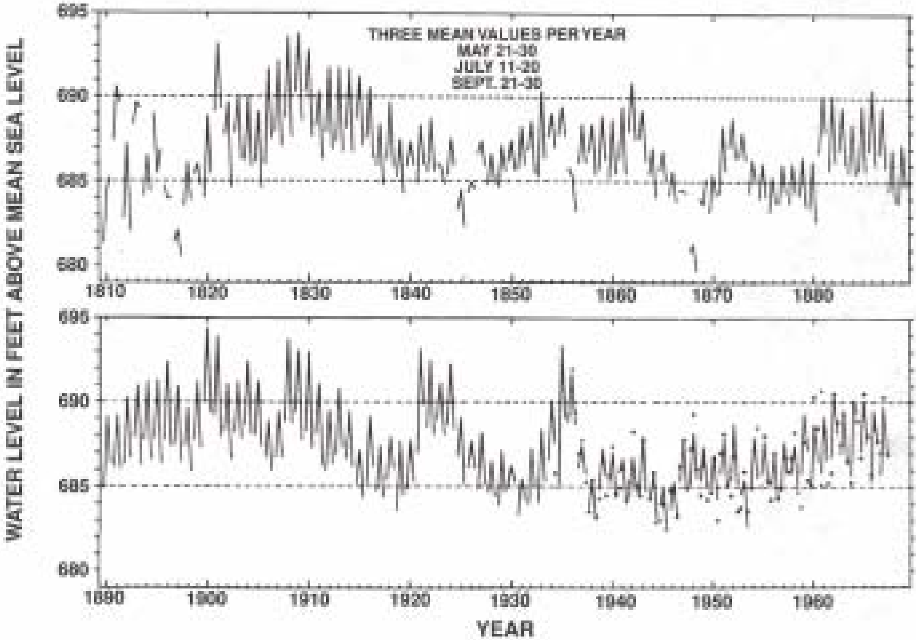

FIGURE 10.30 Levels of Lake Athabasca, Alberta, Canada, as reconstructed from tree-ring data. Tree rings indicate that prior to 1935 there was greater variability in lake levels during May and July, but there was less variability in lake levels for September than during the recent calibration period. Dots indicate actual lake levels used for calibration. Lines connect the three estimates from tree rings, representing mean lake level for May 21–30, July 11–20, and September 21–30. Points are not connected over the winter season, as calibrations of levels for the frozen lake could not be made (Stockton and Fritts, 1973).

10.4 ISOTOPE DENDROCLIMATOLOGY

Many studies have demonstrated empirically that variations in the isotopic content of tree rings ($\delta^{13}C$, $\delta^{18}O$, and $\delta^{2}H$) are in some way related to climate.[33] Early studies used whole wood samples but some studies now suggest that latewood may need to be isolated from earlywood to get a clear signal of climatic conditions in the growth year (Switsur *et al.*, 1995; Robertson *et al.*, 1995). In mid- and high latitudes, there is a positive relationship (*sensu lato*) between the oxygen and hydrogen isotopic composition of rainfall and temperature (Rozanski *et al.*, 1993) so it is reasonable to expect that the isotopic content of wood in trees might preserve a record of past temperature variations in such regions (Epstein *et al.*, 1976). The problem is

[33] To avoid the difficulties of chemical heterogeneity in wood samples, a single component, α-cellulose (polymerized glucose), is extracted for isotopic analysis; α-cellulose contains both carbon-bound and oxygen-bound (hydroxyl) hydrogen atoms, but the latter exchange readily within the plant. It is therefore necessary to remove all hydroxyl hydrogen atoms (by producing nitrated cellulose) to avoid problems of isotopic exchange after the initial period of biosynthesis. For a discussion of isotopes, deuterium/hydrogen (D/H) and $^{18}O/^{16}O$ ratios, see Sections 5.2.1 and 5.2.2.

that additional isotopic fractionation (of hydrogen, oxygen, and carbon atoms) occurs within trees during the synthesis of woody material and these biological fractionations are themselves dependent on many factors, including temperature, relative humidity, and wind speed (evapotranspiration effects) (Burk and Stuiver, 1981; Edwards, 1993). Nevertheless, in spite of such complications, a number of studies have found very strong positive correlations between $\delta^{13}C$, $\delta^{18}O$, and $\delta^{2}H$ and temperature, and negative correlations with relative humidity. The exact temperature relationship varies, but in several studies of $\delta^{13}C$ and $\delta^{18}O$, it is commonly in the range of 0.3–0.4‰ per °C (Burk and Stuiver, 1981; Yapp and Epstein, 1982; Stuiver and Braziunas, 1987; Lipp *et al.*, 1991; Switsur *et al.*, 1995). Most studies have focused on only the last few decades, but a few have attempted longer paleoclimatic reconstructions. For example, based on the strong correlation observed in recent data between August temperature and $\delta^{13}C$ in fir (*Abies alba*) from the Black Forest of southern Germany, Lipp *et al.* (1991) reconstructed temperature back to A.D. 1000 (Fig. 10.31). This record suggests that there was a steady decline in temperature from the early fourteenth century to ~1850, with an earlier warm episode centered on A.D. 1130. Other long-term temperature reconstructions include a 2000 yr $\delta^{13}C$-based record from the western United States (Stuiver and Braziunas, 1987) and a 1500 yr $\delta^{2}H$-based record from California (Epstein and Yapp, 1976). There is little similarity between these two records, possibly reflecting real temperature differences, but perhaps also highlighting the many complications that may confound any simple interpretation. Certainly, the processes involved are complex. For example, Lawrence and White (1984) found that δD in trees from the northeastern United States does not contain a strong temperature signal, as might

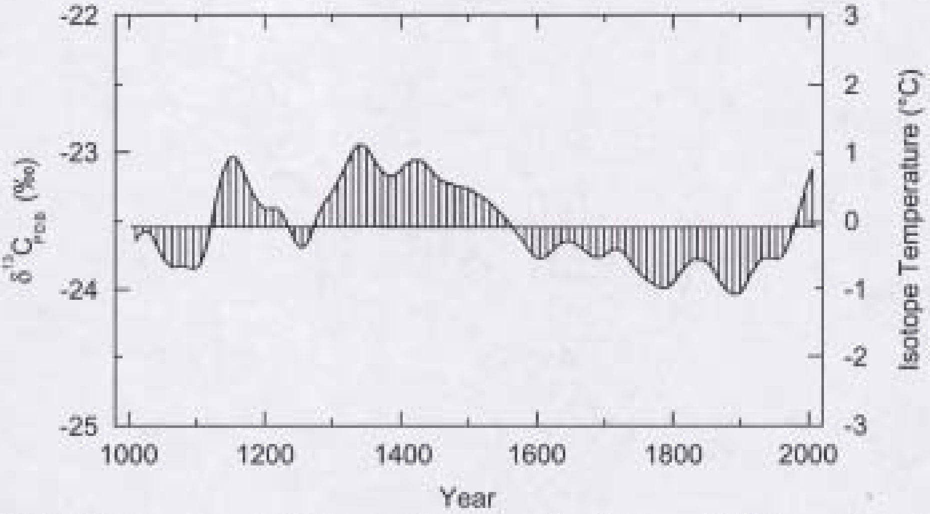

FIGURE 10.31 Average $\delta^{13}C$ values of cellulose extracted from the latewood of 19 fir trees (*Abies alba*) in the Black Forest, from A.D. 1004–1980. Values shown are smoothed by a 100-yr Gaussian lowpass filter; values from 1850–1980 are corrected for the effects of contamination by fossil fuel CO_2. Based on calibration over recent decades, the scale of August mean temperature is shown on the right, relative to the long-term mean (Trimborn *et al.*, 1995).

be expected, but is inversely correlated with summer rainfall amount. By contrast, Ramesh *et al.* (1989) found δD in teak from western India was positively correlated with rainfall. In each case it is possible to construct an explanation for the observed relationship, but one feels that this approach is a little too *ad hoc*. There is a real need to build on the empirical approach by constructing comprehensive models to help in understanding all the processes involved (Edwards and Fritz, 1986; Clague *et al.*, 1992; Edwards, 1993). Isotopic dendroclimatology has much potential, but significant efforts are still needed to develop reliable paleoclimatic records to complement ring-width and densitometric studies.

10.4.1 Isotopic Studies of Subfossil Wood

Yapp and Epstein (1977) showed that δD in wood was strongly correlated with δD in associated environmental waters, which were consistently 20–22‰ *higher* than the nitrated cellulose (Epstein *et al.*, 1976; Epstein and Yapp, 1977). As δD of annual precipitation is correlated (geographically) with mean annual temperature (Yapp and Epstein, 1982) long-term records of δD from trees may be a useful proxy for temperature variations. This idea is supported by mapping δD values from modern plants and comparing them with measured δD values of meteoric waters (Fig. 10.32). It is clear

FIGURE 10.32 Isolines of δD based on modern meteoric waters (i.e. in precipitation) compared to δD values inferred from cellulose C-H hydrogen (cellulose nitrate) in modern plants (underlined values). With the exception of only two Sierra Nevada samples, the inferred values differ by an average of ~ 4‰ from the values measured in precipitation samples (Yapp and Epstein, 1977).

that the δD in plants provides a good proxy measure of spatial variations in δD of precipitation. Assuming that this relationship has not changed over time, it is possible to reconstruct former δD values of meteoric water by the analysis of radiocarbon-dated subfossil wood samples (Yapp and Epstein, 1977). Figure 10.33 shows such a reconstruction, for "glacial age" wood (dated at 22,000–14,000 yr B.P.). Surprisingly, the ancient δD values at all sites are consistently *higher* than modern values (an average of +19‰). The higher δD of meteoric waters implies that temperatures over the ice-free area of North America were warmer in late Wisconsin times than today. However, there are many other factors that could account for the high δD values observed. In particular, δD values in the wood may reflect rainfall in warm growing seasons and so would be isotopically heavier than the annual values mapped in Fig. 10.32. This would be especially true if more of the extracted cellulose came from latewood rather than earlywood.

Other factors which could help to explain this surprising set of data include: (a) a reduction in the temperature gradient between the ocean surface and the adjacent precipitation site on land; (b) a change in δD of the ocean waters as a result of ice growth on land (probably corresponding to an increase in oceanic δD of 4–9‰); (c) a change in the ratio of summer to winter precipitation; and (d) a positive shift in the average δD value of oceanic water vapor, which at present is not generally evaporating in isotopic equilibrium with the oceans.

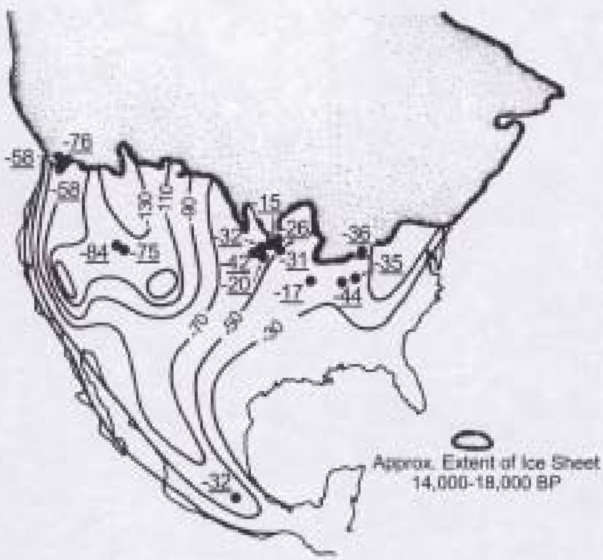

FIGURE 10.33 Distribution of 16 δD values of meteoric waters from 15 different sites during the late Wisconsin glacial maximum, as inferred from δD values of tree cellulose C-H hydrogen (underlined). The approximate position of the southern margin of the ice sheet is shown at its maximum extent (hatched line). The "glacial age" meteoric waters of the 15 sites have, on average, δD values which are 19‰ more positive than the corresponding modern meteoric waters at those sites as deduced from the data shown in Fig. 10.32. The "glacial age" distribution pattern of δD values is similar to the modern pattern, but is systematically shifted by the positive bias of the ancient waters. The North American coastline shown is that of today and does not take into account the lower sea level at the time of glacial maximum (Yapp and Epstein, 1977).

Whatever the reason, the isotopic composition of glacial age precipitation has important implications for the isotopic composition of the Laurentide Ice Sheet, which is generally assumed to have been composed of ice that was very depleted in ^{18}O and deuterium. Using δD values from trees growing along the shores of Glacial Lakes, Agassiz and Whittlesey, Yapp and Epstein (1977) calculated that the $\delta^{18}O$ value of waters draining from the former Laurentide ice sheet probably averaged around −12 to −15‰, far higher than would be expected by analogy with glacial age ice in cores from major ice sheets. Measured δD values from glacial age ice in cores from Greenland and Antarctica (Chapter 5) average 80‰ *below* those of modern precipitation in the same area. These results are thus somewhat enigmatic and it would be extremely valuable to extend this work to other formerly glaciated areas (particularly Scandinavia) to study "glacial age" δD values in more detail because they have an important bearing on the interpretation of other paleoclimatic records, particularly ice and ocean cores. On the other hand, there may have been other factors operating to alter the δD/temperature relationship observed today, or to bias the wood isotopic signal in some way, illustrating once again the difficulty of interpreting isotopic signal in tree rings.