Betty for putting me on the right road,
Sylvie for keeping me there.

The Holocene

An Environmental History

SECOND EDITION

Neil Roberts

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CHAPTER TWO

Reconstructing Holocene Environments

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Dating the Past

Dating techniques are fundamental to an understanding of the natural and cultural changes which have taken place during the Holocene. Without them events such as the Neolithic (agricultural) revolution would float aimlessly in time. Perhaps worse, without independent dating methods we would be forced to depend on environmental and archaeological evidence to provide chronologies, thus robbing this evidence of much of its potential meaning (Vita-Finzi, 1973).

Curiously, the most obvious reason for wanting to date the past may be the least important; that is, the desire to know how old something is simply for its own sake. It may stagger the imagination to think that there are trees living today in California's Sierra Nevada which were saplings before Stonehenge was completed, but the significance of neither henge monuments nor the bristlecone pine is better explained as a result. A much more significant role for independent dating lies in the testing of hypotheses (Deevey, 1969). Take, for example, the mid-Holocene elm decline widely recorded in pollen diagrams from northwest Europe. If, as has been hypothesized, this was a consequence of prehistoric agriculture, then the fall in elm pollen values should coincide with the time of arrival of the first farming communities. Dating early Neolithic sites, on the one hand, and the appropriate section of pollen diagrams, on the other, allows the anthropogenic hypothesis to be tested. In particular, if the elm decline is found to pre-date the start of the Neolithic, then this hypothesis would be shown to be false, and our agricultural ancestors exonerated from blame (see chapter 5 for further discussion of the elm decline).

Another reason for wanting to have precise estimates for the age of Holocene events is in order to calculate past rates of change. Some events, such as **eustatically** controlled sea-level rise, would have occurred at the same time everywhere across the ocean; that is, the change was a **synchronous** one. More usually, however, events begin earlier at some places than at others. The spread of a disease or the movement of a glacier snout are of this type, and they are termed **time-transgressive**. The rate of disease diffusion or of glacier retreat will vary between cases and precise dating techniques make it possible to establish whether rates were fast or slow, constant or variable. Dating is also essential to a number of the other techniques discussed later in this chapter. Past influx of pollen or of sediment into a lake, or of algal productivity within it, can only be obtained if a sound and detailed chronology exists for the sequence of lake sediments.

One simple form of dating is provided by the fact that in undisturbed sediments younger layers overlie older ones. This law of superposition indicates which layer was deposited first, but it fails to provide the actual age of either. Preferably, the layers should be placed not only in a relative sequence such as this but also firmly in time, and for this it is necessary to assign them absolute ages in years. Four principal approaches to absolute dating exist: (1) those based on historical records; (2) those employing **radiometric dating** techniques; (3) those utilizing incremental dating methods; and (4) those based on palaeomagnetism. Each will be discussed here, but most attention will be paid to the second and third approaches, which will be illustrated in detail by reference to the specific techniques of radiocarbon dating and **dendrochronology**.

Historical dating

This form of dating is most obviously associated with documentary-based studies of historical ecology or climate. For instance, medieval manuscripts recording the extent and condition of England's royal forests almost invariably have dates attached to them, as do Icelandic chronicles referring to drift ice – and hence sea temperature – around that island. We may go further and suggest that without a date, documentary records such as these are of little use in helping to piece together past environments.

Historical or archaeological evidence can also be important in dating non-documentary records such as pollen diagrams. In the case of one core from southwest Syria, the presence of an exotic pollen type, that of maize (*Zea mays*), helped show that the upper part of the core dated to recent centuries and not to the early Holocene as had previously been proposed (Bottema, 1977). Maize is native to the Americas and was introduced into the Old World as a crop only after the Spanish conquest of Mexico in the sixteenth century.

The remnants of past human activity can provide other important clues for dating Holocene environmental changes. A Roman tile drain buried beneath 5 m of alluvium, or a classical port now silted up and many km from the sea, both testify to active sediment transport and deposition by rivers during the last 2000 years. Artifacts, including pottery, stone tools and coins, can all be assigned ages with greater or lesser precision (see plate 2.1), as can less obvious cultural evidence such as hemp fibre from retting that has been incorporated in lake sediments. If discovered in a stratigraphic sequence, artifacts provide maximum ages for the layer in which they were found; maximum because they may have been re-worked since being deposited initially. A bicycle frame pulled from supposedly

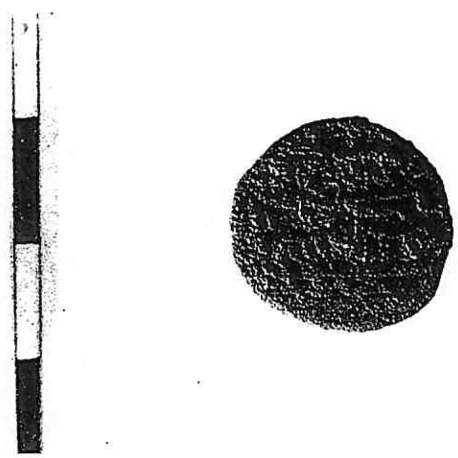


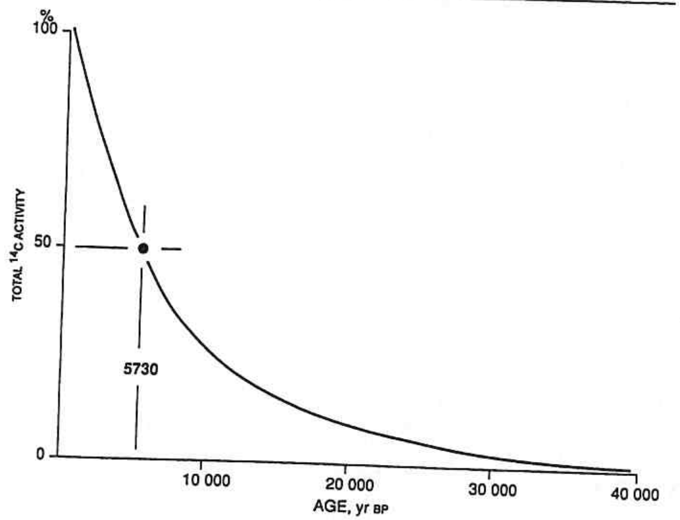
Plate 2.1 This Arab coin from Crete, made between 847 and 861 AD, helped date the alluvial fill in which it was found

mid-Holocene coastal dunes provides a – revised – maximum age for the dunes, for even if the bicycle were manufactured 50 years ago it could have been dumped there as recently as last year. On the other hand, structures such as a former irrigation channel or a shell midden will not have been redeposited in the same way as artifacts and they may therefore offer tighter dating control. In fact, built structures will often provide minimum rather than maximum ages, say for a land surface on which they are found. However, the use of human artifacts for dating purposes can be fraught with dangers, as it necessarily involves time-transgressive phenomena. The stone age may have ended 5000 years ago in eastern Europe, but it arguably lives on with certain isolated hunter-gatherer groups in the tropics.

Radiometric dating methods

Radiometric dating techniques have been discovered and applied only during the second half of the twentieth century. They involve the radioactive properties of different materials which contain within them a natural time signal, most often involving the principle of isotopic decay. Most natural elements are a mixture of several **isotopes**, which have the same chemical properties and atomic numbers but different numbers of neutrons and hence different atomic masses. One isotope is always dominant for each element; for instance, in the case of carbon the dominant isotope is carbon-12 (or ^{12}C). Carbon isotopes with atomic masses different from this are ^{13}C and ^{14}C, of which the latter is by far the least abundant.

Figure 2.1 Decay curve for radiocarbon



Carbon-14 is also different from the other two in that it is isotopically unstable; in other words, it decays to form stable ^{14}N over time. Most importantly of all, unstable isotopes such as ^{14}C decay at a fixed rate. The rate of radioactive decay is not a straight line but occurs rapidly to begin with and slows down progressively over time (see figure 2.1). Because decay rates are measurable, unstable isotopes represent natural archaeological or geological clocks. The clock is set to zero when isotopic decay starts, and the time elapsed can be gauged from how far the decay process has proceeded since then, as indicated by the amount of radioactivity left in the element. The decay rate for a particular isotope is most usually recorded by its half-life, or the time taken for its radioactivity to be reduced by half. Half-lives of elements vary from the very short (e.g. radon-222, 3.8 days) to the very long (e.g. potassium-40, 1300 million years). In general the useful dating range of individual isotopic methods is about ten times their half-life. Beyond this, radioactive emissions are difficult to distinguish from normal background levels of radioactivity. Consequently, different isotopes provide dating methods for very different time periods, and only a few of these will be useful for the Holocene.

The most important of the radiometric dating methods for the Holocene is based on the isotope carbon-14. RADIOCARBON DATING was pioneered by Willard Libby at Chicago in the late 1940s, work which later earned him the Nobel Prize for Chemistry (Burleigh, 1981). ^{14}C is formed in the upper atmosphere by cosmic ray bombardment of nitrogen (N) atoms. The resulting carbon isotope is rapidly oxidized to form carbon dioxide (CO_2), and is then mixed uniformally and rapidly through the atmosphere. Photosynthesis leads to ^{14}C being taken up by plants, which in turn is passed on to higher organisms including humans. Consequently, ^{14}C is present in small but significant quantities in all living organisms. When organisms die their ^{14}C content is no longer replenished and the isotopic clock is set in motion. Libby calculated the half-life of ^{14}C experimentally as 5560 ± 30 years, a value close to the present best estimate of 5730 ± 40 years. Libby cross-checked samples which he had dated by the ^{14}C method against materials of known historical age, most of which came from ancient Egypt and were up to 5000 years old. The close level of agreement led him to believe that his radiocarbon dating method was a valid one and that it could be applied universally.

Since its discovery about 100 000 ^{14}C determinations have been carried out by over 100 different laboratories. About half of all dates have been for archaeological investigations, with most of the rest having been for studies of past environments, and with the vast majority of radiocarbon dates being of Holocene age. The results of radiocarbon dating are usually presented as ages in years before present, normally written using the notation yr BP. In practice, the present is taken as AD 1950 in order to prevent dates appearing to be older simply because they were analysed more recently. Archaeological chronologies use the BC/AD system more often than the yr BP one, but the former can be turned into the latter simply by adding 1950 or – as a quick approximation – 2000 years to BC dates.

Also listed along with each individual ^{14}C date is a laboratory code number and an error function representing one standard deviation about the mean. Thus 8160 ± 110 yr BP (BM – 1666) was dated at the British Museum radiocarbon laboratory and indicates that there is a 68 per cent probability that the ^{14}C age of this sample lies between 7940 and 8380 yr BP. It should be borne in mind that radiometrically determined dates are statistical estimates, and they should therefore be treated as such. ^{14}C dating can be used to determine the age of any material containing carbon, including wood, charcoal, peat, seeds, bone, carbonate, shell, iron, cloth, rope, groundwater and soil. Among the more exotic materials dated have been ostrich eggshell, lime mortar and human brain!

The ^{14}C method cannot, however, be applied to samples of very recent age. The burning of fossil fuels since the Industrial Revolution has caused an injection of geologically old carbon into the atmosphere which has lowered its ^{14}C content and made recent samples seem older than they actually are – the so-called Suess effect. More recently still, this has been overcompensated and further complicated by the testing of atomic weapons. In short, ^{14}C dating is of little use for samples younger than about 150 years.

Many other forms of sample contamination are possible, and although ash from a careless investigator's cigarette has doubtless accounted for more than one erroneous ^{14}C date, contamination is usually of natural rather than of human origin. Carbon both older and younger than the date of the death of the sample may be involved. The latter is a particular problem in areas of limestone or coal-bearing bedrock. This problem is especially difficult to assess where bicarbonate-rich lake waters have been taken up by aquatic plants which subsequently formed part of the lake sediment. The dilution of ^{14}C levels causes ages to appear older than they actually are, and this is known as the hard-water effect (Deevey et al., 1954). Contamination by young carbon, by contrast, makes dates appear too recent. There are countless such sources, including rootlets penetrating into underlying stratigraphic layers, recalcification of mollusc shells, and animal burrows in archaeological sites. However, many problems of contamination by younger material can be avoided by careful selection of samples in the field and by pre-treatment during dating. Young carbon is, in any case, a much more serious problem for samples of Pleistocene than of Holocene age. Improved dating precision is also being achieved by a method developed in the 1980s of measuring ^{14}C directly using a mass spectrometer. This **AMS** (or **Accelerator Mass Spectrometer) dating** method allows the dating of small samples weighing as little as 5 mg. It allows scientists to pick out and date individual seed remains, especially of land plants which should be free of any hard-water error, or charcoal fragments. More controversially, it allowed the age of threads from the Turin Shroud to be determined, which could otherwise have been dated only by destroying the shroud itself. The result showed the Shroud to be a – very clever – medieval forgery, fabricated between AD 1260 and 1390.

Some materials are often considered to be more prone to contamination than others – for example, bone, shell and soil – but ^{14}C dating will be more reliable if samples of these materials are adequately prepared before age determination. In the case of bone, only the protein collagen should be dated (Gillespie, 1984, p. 13), while mollusc shell should be tested by x-ray diffraction and acid leached or mechanically cleaned if necessary (Vita-Finzi and Roberts, 1984). Soil, of course, takes longer to form than either of these two, but the consequent broad time range associated with ^{14}C dates on buried soils is as much a stratigraphic as a dating problem (it also applies, for instance, to the interpretation of soil pollen) (Matthews, 1985). ^{14}C age determination of soil should ideally involve dating selected organic fractions; but even without

this, Rothlisberger and Schneebeli (1979) obtained consistent and meaningful results in their study of Holocene soil and moraine stratigraphy in the Swiss Alps. All of these contamination problems are local to sites under study. By contrast, the one major revision to the ^{14}C timescale since Libby's initial discovery involves variations that were global in scale.

It was initially assumed by Libby that there had been no significant changes in atmospheric levels of ^{14}C during recent millennia. Libby had indeed carried out ^{14}C determinations on historically dated materials to check this, but after initial enthusiasm he subsequently found that older dates systematically underestimated the true age of samples. This was confirmed in 1965 when Hans Suess presented results comparing ^{14}C dates with those from another dating method – dendrochronology. These results are discussed in detail below, but suffice it to say that ^{14}C dates older than 2500 years significantly underestimate actual age. But because these deviations from true age are systematic and world wide, it is possible to apply correction, or calibration, factors to ^{14}C determinations to turn them into true, or calendar dates. The need for ^{14}C calibration has not, therefore, invalidated the ^{14}C method, which remains a remarkably robust and successful one. The radiocarbon method and its applications are discussed in more detail by Gillespie (1984) and Olsson (1986), and the AMS method by Hedges (1991).

Two other radiometric techniques of increasing importance are luminescence and uranium-thorium dating. LUMINESCENCE, or Optical, dating was initially applied to archaeological materials such as pottery, burnt clay and flint (Aitken, 1990). It was then discovered that 'bleaching' of the sample, which resets the radiometric clock, could be accomplished by exposure to sunlight as well as by firing. Thermo-luminescence or TL dating subsequently came to be successfully applied to the quartz or feldspar grains of wind-blown sediments such as sand dunes and loess (Wintle and Huntley, 1982), assuming that they were not deposited in night-time darkness! The technique has since added other variants, such as Optically Stimulated Luminescence (OSL), which require shorter 'bleaching' times, and are capable of dating other types of sediment such as river alluvium (Aitken, 1994; Duller, 1996). URANIUM-THORIUM, or U-Th, dating is based on a more complex decay chain than ^{14}C, with a sequence of 'daughter' isotopes being produced from the original 'parent' (Smart, 1991; Ivanovich and Harmon, 1995). On the other hand, the half-lives involved are longer than with ^{14}C, so that the technique can be used to date older materials. Unstable uranium or thorium isotopes are taken out of water by corals, or precipitated in cave speleothems or in lake sediments. U-Th, especially with the use of a high-precision mass spectrometer, has proved valuable in calibrating the radiocarbon method beyond the range of dendrochronology (Bard et al., 1990).

Luminescence, uranium-thorium and ^{14}C all cover timespans of thousands of years or longer, and none is useful for dating changes during the last century. Partly because of this, palaeoecological studies tended to stop precisely at that point in time when human impact has often become most apparent. However, a number of dating techniques with short half-lives have been developed which allow the timing of recent human impact on ecosystems to be determined. Probably the most useful of these new radiometric dating techniques are LEAD-210 and CAESIUM-137 (see Technical Box X, pp. 236–8).

Dendrochronology and radiocarbon calibration

Incremental techniques, which are based on natural seasonal rhythms or annual growth rates, have been utilized since the early years of the twentieth century. However, it would be wrong to imagine that they were made redundant by the advent of radiometric dating techniques. As will be made clear from the discussion that follows, they remain as valuable today as they ever were.

The principles of this type of dating can be illustrated by what is probably the best-known incremental dating method: tree-ring dating or DENDROCHRONOLOGY. This technique was largely developed through the efforts of A.E. Douglass working in Arizona between 1910 and 1940. Although it had been known since at least the time of Aristotle that trees produced annual growth rings, Douglass was the first to systematize this into a dating method. At its simplest the technique involves counting the number of growth rings present between the bark and the centre of a living tree. As the outermost ring can be assumed to represent the current year's growth, the age of the tree can be calculated by counting the number of rings present. This need not require lumberjack skills, however, for a tree-ring sequence can be obtained without harming the tree by using an increment borer, which provides pencil-sized tree cores. This basic form of tree-ring dating can be used to provide a chronology for valley glacier retreat or alluvial fan activity over recent decades and centuries, although the ages it provides in these cases are only minimum ones because of the time lag between deglaciation or channel change and subsequent colonization by trees.

Potentially of greater value are those approaches which exploit the characteristic tree-ring signatures produced as a result of differential growth from one year to the next. Favourable environmental conditions lead to the formation of a broad growth ring, while adverse conditions have the opposite effect. The result is a unique sequence of wide and narrow rings not unlike the bar-codes used for store price tags. The main environmental factors which cause ring-width variations are climatic ones, but the critical stress factor will not be the same everywhere. In the Alps a narrow tree-ring is usually the product of a cold summer, while in the American Southwest low rainfall is more likely to be responsible for stunted growth. Tree-rings can therefore provide proxy climatic data, the study of which is known as dendroclimatology (Schweingruber, 1988; Fritz, 1991).

Tree-ring signatures also fulfil another function, that of allowing cross-dating or correlation between tree-ring sequences. Such cross-dating is not restricted to live wood, so that sequences from dead trees or wood used in buildings can be matched up with those from living trees (see figure 2.2). The former would otherwise be free-floating chronologically, as their outer edges do not date to the present day and consequently lack a firm reference point in time. The building up of a chronology from separate tree-ring sequences amounts to no less than a jigsaw puzzle with time. 'The pieces are scattered round as living trees, stumps, timbers in buildings and buried, either as archaeological material or as naturally preserved timbers, in bogs, rivers or lake beds' (Baillie, 1995). Douglass, in Arizona, compiled both a master chronology stretching from living yellow (ponderosa) pines back to the thirteenth century AD, and a floating one based on pine timbers from prehistoric buildings. In 1929, after much searching, he found the missing piece in his jigsaw – an excavated wooden beam from Pueblo Bonito dating to between AD 1237 and 1380 – and the two sequences were joined.

Huber subsequently applied Douglass's approach to German tree-rings and by 1963 had completed a 1000-year chronology based on oak rather than pine timbers. This oak chronology has since been extended and provisionally joined to a 1605-year floating pine sequence which together go back right to the beginning of the Holocene (Kromer and Becker, 1993). A similar 7000+ year chronology has been built up from Irish bog oaks by a research team at Queen's University, Belfast, using the same cross-dating approach (Pilcher et al., 1984). The most famous dendrochronological record, however, comes from the bristlecone pine (*Pinus longaeva*) (plate 2.2), which grows at altitudes of 3000 m in the mountains of eastern California and Nevada. This is the world's longest-living tree and, incredibly, living bristlecone pines have been recorded up to 4600 years old (Ferguson, 1968). Cross-dating of sub-fossil trunks with living trees has produced a continuous bristlecone pine tree-ring chronology 8200 years long. The construction of tree-ring

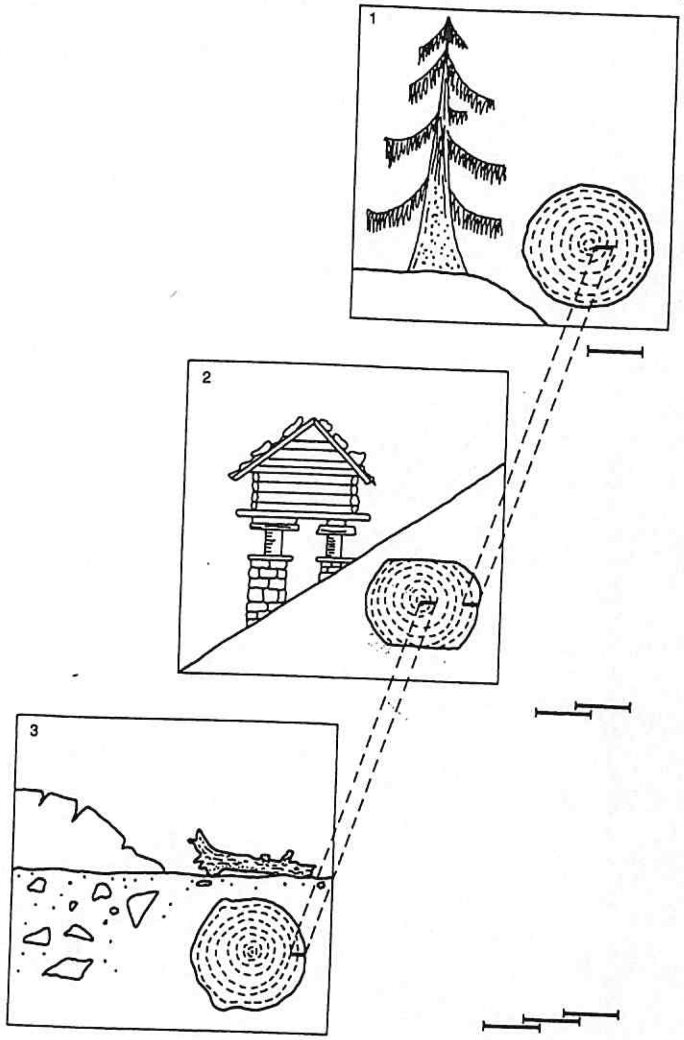


Figure 2.2 Cross-dating alpine timbers from overlapping tree-ring signatures

chronologies such as these involves the science of computer matching and x-ray densitometry as well as the art of matching tree-ring series by eye (Baillie, 1995).

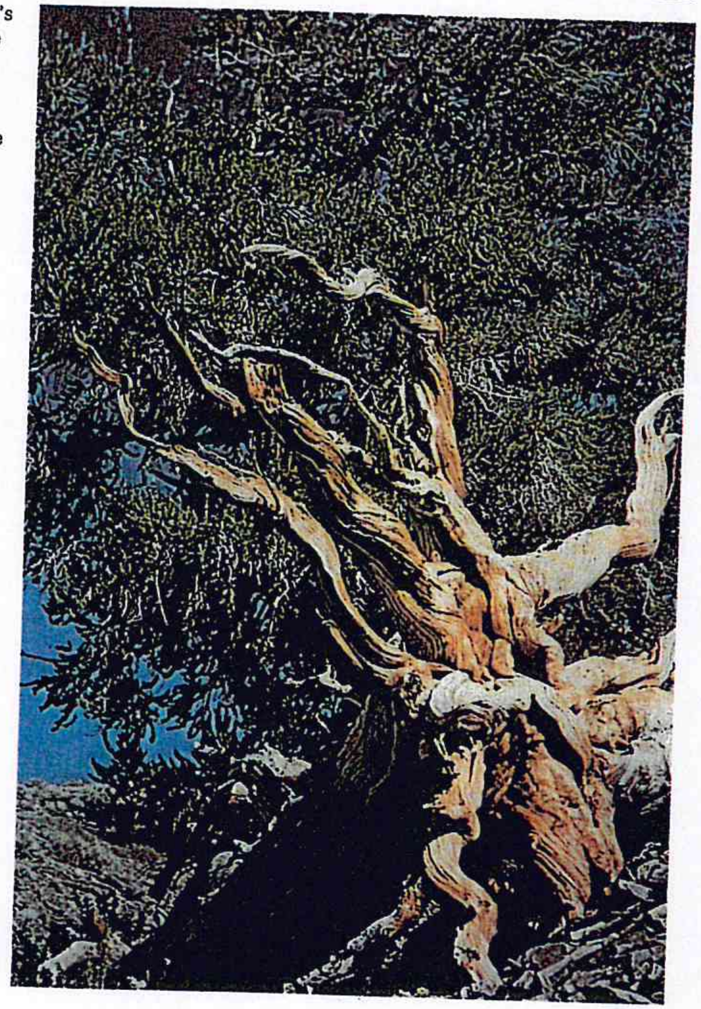
Possibly the most important feature of dendrochronology is the fact that the wood used for tree-ring counts can also be dated by radiocarbon, and this has allowed a cross-check to be made between the two methods. At first sight it may seem surprising that it is possible to get 'old' dates from 'living'

wood, but most of the woody cells in trees are empty, being used only for transporting water from the roots to the branches. The cells laid down in each growth ring are the only part to retain a new ^{14}C imprint. In the 1960s, Hans Suess dated samples of bristlecone pine of known tree-ring age by radiocarbon. His results revealed a systematic divergence between ^{14}C and tree-ring ages in samples more than about 2500 years old (see figure 2.3). Mid-Holocene ^{14}C dates appear to be too young, with a discrepancy amounting to 600 years (or 12 per cent) for material 5000 years old. Tree-ring dates are more accurate than their ^{14}C equivalents, although this does not invalidate the ^{14}C method, but rather requires that ^{14}C dates be calibrated if they are to be expressed as true, or calendar years. Bristlecone pine calibration caused important revisions to be made to archaeological chronologies based on ^{14}C, and in consequence the theories of culture change in European Prehistory (Renfrew, 1973). Palaeoecologists and other environmental scientists have been more reluctant than archaeologists to convert their ^{14}C dates into calendar years. This is partly because until recently, calibration curves covered only part, not all, of the Holocene. Now, however, tree-ring calibration has been taken further back in time and has been joined by other dating methods which extend to the period of the last glaciation. Two of these, tree-rings and U-Th ages from corals, have been combined in a widely used ^{14}C age calibration computer program, CALIB 3.0 (Stuiver and Reimer, 1993).

There now seems every reason, therefore, to adopt a calendar rather than a conventional, uncalibrated ^{14}C timescale for an environmental history of the Holocene. In consequence, **the ages in this book will use the convention of quoting dates calibrated, expressed as Cal. yr BP** (Calendar years before present). One advantage of this is that it removes the discrepancy between radiometric and historical chronologies during the period from about 5000 to 2500 years ago. Calibration has been based on CALIB 3.0 and OxCal (Ramsay, 1995), with minor modification at the Pleistocene–Holocene boundary (see Technical Box I). A summary conversion table from ^{14}C to calendar ages (and vice versa) can be found in the appendix on p. 253. Calibration has been applied to individual ^{14}C dates as well as to overall chronologies. In most published papers, ^{14}C determinations are quoted only as raw, uncalibrated dates, and they would need to be converted to calendar ages in order to make a comparison with the timescale used here.

It is important to recognize that there are some potential pitfalls in making a conversion from ^{14}C to calendar years. First, the tree-ring calibration curve is not a smooth line but contains a number of wiggles which represent short-term variations in atmospheric ^{14}C content (figure 2.3). This is significant

Plate 2.2 The world's longest-lived tree, the bristlecone pine (*Pinus longaeva*), which lives in the high mountains of the Sierra Nevada in western North America



in that some ^{14}C dates have more than one calendar age; a ^{14}C date of 4500 yr BP for instance has calibrated ages between 5056 and 5246 years. Calibration can therefore have an important effect on studies where high dating resolution is required because an apparently synchronous event may be turned into one that is potentially time-transgressive. This is a particular problem for dating the onset of the Holocene (Technical Box I). In high-resolution studies, on the other hand, it may be possible to obtain sufficient ^{14}C dates to use the 'wiggles' of the calibration curve to match the series being dated. A second

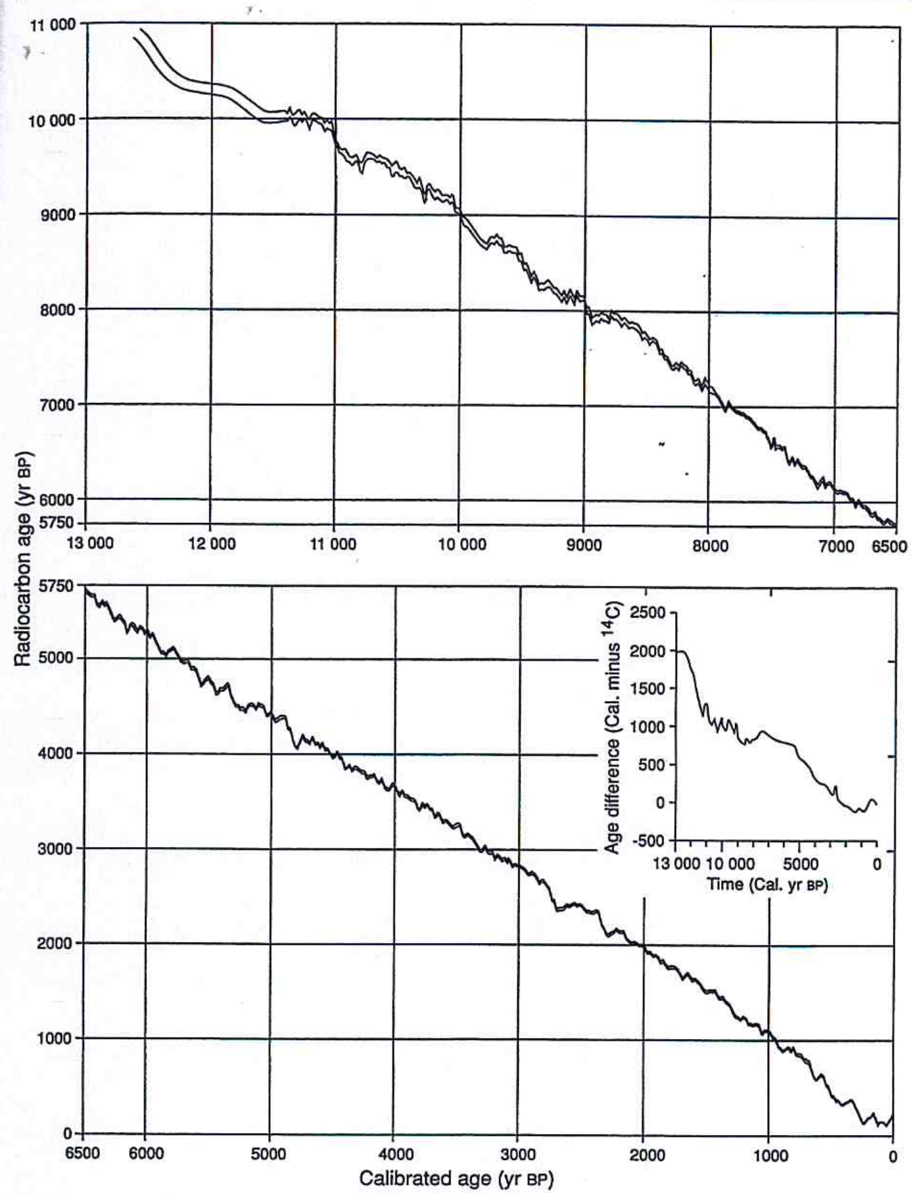


Figure 2.3 Tree-ring radiocarbon calibration curve for the Holocene (inset: departures from calendar age with time) (from OxCal. 14L program; Younger Dryas based on Goslar et al., 1995)

TECHNICAL BOX

Technical Box I: Dating the onset of the Holocene

There are different schools of thought about how the Holocene should be formally defined. Because the Holocene still 'lives', debate has focused on defining its onset, namely the Pleistocene–Holocene boundary. The first school of thought (e.g. Watson and Wright, 1980) believes that the beginning of the Holocene is easily recognized in individual records of deglaciation or biotic change (e.g. in pollen diagrams), but that because these changes are not exactly the same age everywhere, the Pleistocene–Holocene boundary should therefore be time-transgressive. A second school prefers the 'type-section' approach commonly used in geology to define stratigraphic boundaries. Mörner (1976), for example, proposed that a sequence in southern Sweden might be the standard reference point for the Pleistocene–Holocene boundary. The third view, and the one which has been most widely adopted, is that the Holocene is simply defined as beginning 10 000 ^{14}C years ago.

However, there is a difficulty in knowing what 10 000 ^{14}C yr BP actually means in terms of real calendar ages. ^{14}C-dated tree-ring sequences have shown a systematic divergence between the two timescales which amounts to 1000 years or more by the beginning of the Holocene. Moreover, ^{14}C time is elastic; that is, it has been stretched or compressed at different periods in the past. And perversely – if not coincidentally – 10 000 ^{14}C yr BP happens to coincide with a major ^{14}C 'plateau', when the ^{14}C 'clock' stood still, while at least 400 years of real time elapsed (Taylor et al., 1996) (figure 2.3). This ^{14}C plateau was probably caused by abrupt changes in atmospheric CO_2 and oceanic circulation at the end of the last glaciation (Goslar et al., 1995), and it has meant that 10 000 ^{14}C yr BP represents not one moment in time, but several! It would clearly be preferable to define the onset of the Holocene in calendar and not in ^{14}C years. But which date should be chosen? There is evidence of rapid climatic change around this time in a number of independently dated natural archives, and the different calculated ages for the change are shown in table 2.1. These different techniques do not provide identical results, but most of them lie between 11 000 and 11 800 Cal. yr BP. The CALIB 3.0 program gives *c.*11 200 Cal. yr BP as the calibrated equivalent of 10 000 ^{14}C yr BP, although this is a little younger than counts of annual snow layers in Greenland ice cores (Alley et al., 1993). Following Gulliksen et al. (1998), an age of 11 500 Cal. yr BP will be used for the start of the Holocene, although it is recognized that this age could be in error by up to 300 years either way.

Table 2.1 Calendar age estimates for the beginning of the Holocene

Age (Cal. yr BP)	*Method*
11 200	Tree-ring calibration of ^{14}C on German pines
11 360–11 920	Counting of annual layers in GISP2 Greenland ice core
11 440–11 580	Counting of annual layers in GRIP Greenland ice core
11 200–11 700	U-Th calibration of ^{14}C in tropical corals
11 395–11 540	'Wiggle matched' ^{14}C lake sediment record, Kråkenes Lake, Norway
11 360–11 600	Varve calibration of ^{14}C, Lake Gosciaz, Poland
*c.*11 490	Varve calibration of ^{14}C, Holzmaar, Germany
10 960–11 100	Varve calibration of ^{14}C, Soppensee, Switzerland
11 440	Varve counting, Sweden

Sources: Alley et al. (1993), Björck et al. (1996), Gulliksen et al. (1998)

problem is that ^{14}C calibration beyond about 11 000 Cal. yr BP is far from universally agreed. Different dating techniques have produced significantly different age estimates for the start of the **Late-glacial** period, for instance. None the less, all agree that ^{14}C and calendar ages continue to diverge back to the last glacial maximum. For our purposes, they cross-match sufficiently well to make it possible to use a calibrated timescale for all of the events and changes covered in this book, even if the precision achieved for ^{14}C calibration is much reduced before about 10 000 years ago.

Other incremental dating methods

Dendrochronology is only one of a number of incremental dating methods. One potentially precise technique involves counting the annual layers of snow that build up to form ice sheets. If the accumulation significantly exceeds the summer melting, and if it is not deformed by lateral movement (as it is in a valley glacier), then the snow layers will be preserved even after later compaction turns them to ice (see plate 3.1, p. 64). After a certain point, these layers become so squashed by the overlying weight of ice that they are no longer visible.

None the less, in the GISP2 Greenland ice-sheet core it was possible to count every annual layer back to 17 400 Cal. yr BP with an estimated precision of ±3 per cent (Alley et al., 1993).

In similar fashion, lake sediments may be finely layered, or laminated, representing one year's accumulation. Annually laminated lake sediments are known as **varves**, and can form in a number of different ways. The varves studied by the Swedish scientist Baron de Geer at the end of the nineteenth century had formed in lakes adjacent to the former Scandinavian ice sheet, and comprised couplets of alternating coarse and fine-grained layers (Sturm, 1979). These clastic varves provided one of the first quantitative estimates for the duration of the Holocene. On the other hand, laminated lake sediments can form wherever there is a strong seasonal variation in sediment supply or where circulation of lake water is absent or incomplete for at least part of the year (O'Sullivan, 1983). One of the commonest situations in which non-glacial varves form is in lakes that freeze over in winter, as is the case in much of North America and Scandinavia. Organic matter accumulates in the still-water conditions beneath the ice cover, often followed in spring by a deposition of **allochthonous** sediment washed into the lake during snow-melt. In the summer months photosynthesis by algae may lead to a layer of **diatoms** settling out on the lake bed. Varves can also be chemical in origin if the lake water is either strongly acid or alkaline, or if it becomes saline. Alkaline lakes such as Zürich See in Switzerland are rich in carbonate, which summer photosynthesis causes to be precipitated out as a $CaCO_3$ white-out (Kelts and Hsü, 1978). Varves have proved useful even when they are not continuous to the present day and are therefore free-floating in time, for example, in calculating the length of the cold Younger Dryas **stadial** at the end of the last ice age (Lotter, 1991).

Using LICHENS for dating achieves a very different degree of precision. Lichens represent a symbiotic relationship between fungi and algae. The algae photosynthesize to provide nutrition for the partnership and the resulting compound organism behaves as if it were a single biological unit. Lichens are able to exploit a wide range of habitats and grow on substrates ranging from tree trunks to bare rocks. Because of their ability to survive in unfavourable environments they often take the role of primary colonizers in plant succession. Their utilization for dating purposes exploits the fact that most lichen species grow at rates that are both fixed and measurable. The larger the lichen, the greater its age. Empirical work by R.E. Beschel in arctic and alpine environments during the 1950s and 1960s showed there to be a clear relationship between lichen size – as measured by diameter – and age. Beschel went on to formalize this as a dating technique which has subsequently been applied to many field studies, notably of Holocene glacier fluctuations (Innes, 1985). Lichenometry is especially valuable in glacial studies because of the difficulty of applying other dating techniques to such problems as establishing a date for exposure of a bare rock surface following deglaciation (Matthews, 1992).

Field techniques for the measurement and sampling of lichens for dating are relatively straightforward (Lock et al., 1979), but doubts have been raised about whether the assumptions behind them have been adequately tested (McCarroll, 1994). In particular, it remains uncertain how far lichen growth rates are affected by variables other than time, including competition for space, light and nutrients. In consequence, lichenometry is best considered as a dating method that can provide relative rather than absolute ages.

Palaeomagnetic dating

The Earth acts as a giant magnet with its own magnetic field, and this field changes constantly. Perhaps the most dramatic example of these changes is the periodic reversals in the Earth's magnetic polarity. However, the last time the compass needle slipped from the south to the north pole was over 700 000 years ago, and although claims have been made for more recent short-lived magnetic excursions, polarity reversals occur too infrequently to help date the Holocene. On the other hand, so-called secular magnetic variations can be important in helping to provide a timescale over hundreds or thousands of years. It has long been recognized that magnetic north does not coincide with geographic or true north, and that the position of the former has wandered through time. Measurements of the geomagnetic field have been taken at London for over 400 years and show that magnetic north migrated westwards between AD 1580 and 1820, when it lay 20°W of true north. Since then it has shifted eastwards once more so that today it lies at about 11°W.

The angle between true and magnetic north, or declination, is only one of three components of the magnetic field. The compass needle will not only move from side to side but, if freely suspended, dip down at an angle towards the Earth. This dip, or inclination, varies from the equator to the poles, as does strength or intensity of the Earth's magnetic field. All three of these components vary through time and in combination produce a characteristic palaeomagnetic signature. Their magnetic alignments can be incorporated and preserved in baked materials or in sediment particles which settle out in standing water. The study of archaeomagnetic samples such as hearths has extended the record of secular magnetic variations

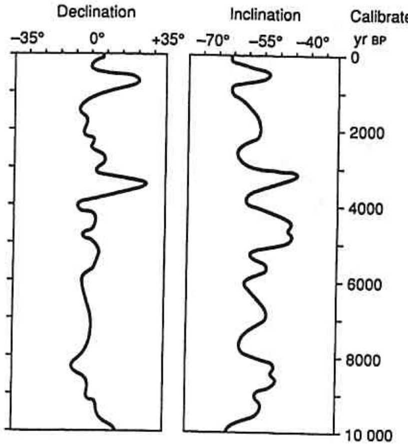


Figure 2.4 Palaeomagnetic dating curves for South Australia (data from Barton and McEllhinny, 1982)

back to before AD 1000 (Aitken, 1990). For earlier periods magnetic measurements have been applied to wet lake sediment cores with considerable success (Thompson and Oldfield, 1986). Secular magnetic variations do not, on their own, provide age estimates. However, once a master geomagnetic curve has been dated by another method such as ^{14}C, it can be applied to sites elsewhere. Holocene master curves, which are regional rather than global in extent, now exist for six areas of the world (Thompson, 1984). Those for South Australia come from the crater lakes Bullenmeri and Keilambete and are shown in figure 2.4. Measurement by magnetometer is rapid and, in the case of declination, non-destructive. On the other hand, not all lakes are equally well suited for the technique, which is effectively restricted to lake sediments with homogeneous particle size and organic content down core.

Conclusion

Possibly the most notable feature of dating techniques is the great diversity of ways in which age estimates can be obtained for the Holocene. Diversity is the key in other respects as well; dating techniques have different age ranges over which they can operate; they cannot all be applied to the same sample materials; some destroy the sample during dating, while others do not; and so forth (see table 2.2). Overall dating reliability is consequently improved by the use of several techniques in combination. Some of the most precise dates, other than historical ones, come from tree-rings. These are accurate as well as precise, but can be relatively time consuming to obtain. Lake varves and ice layers can achieve a similar precision of one

Table 2.2 Dating methods applicable to the Holocene

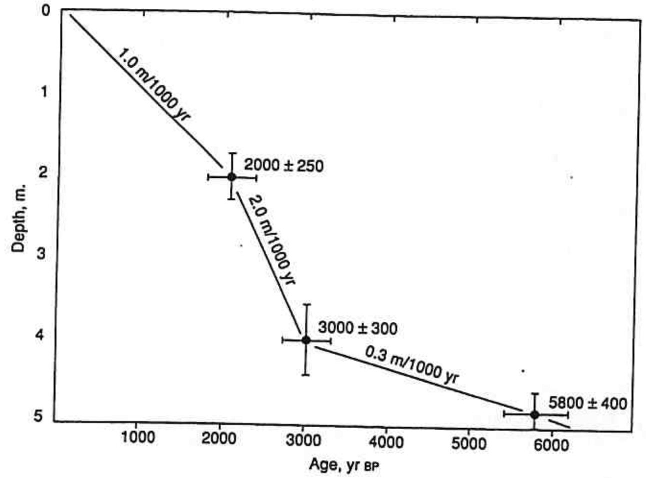
Dating method	*Time range Cal. yr BP*	*Precision yr*	*Applications*
Historical			
archival	1–5000	1–50	varied; e.g. glacier histories
archaeological	50–>40 000	1–1000	varied; e.g. alluvial histories
Radiometric			
radiocarbon	200–40 000	20–1000	organic materials
luminescence	50–>40 000	50–5000	burnt clay and pottery, most minerogenic sediments (e.g. loess)
uranium series	1000–>40 000	20–5000	carbonates (e.g. coral), volcanics
lead-210	10–200	1–10	lake and coastal sediments, peat
caesium-137	1–35	1–5	lake sediments
Incremental			
tree-ring	1–11 000	1–5	wood timber
varves	1–20 000	1–10	lake and some marine basin sediments
annual snow/ice layers	1–>20 000	1–350	ice-sheet cores
lichenometry	10–5000	10–1000	glacial histories
Palaeomagnetism			
secular variations	100–15 000	10–100	lake sediments

year or less, and have the additional advantage of dating deposits which are prime sources of palaeoenvironmental information. On the other hand, none of these is as widely applicable as ^{14}C dating.

Dating is not only a matter of precision and accuracy, but also of attribution and interpretation. Dated samples almost always come from a stratigraphic context of one kind or another, and after age determination they must be related back to the site stratigraphy. Because ^{14}C and ^{210}Pb, in particular, provide spot dates rather than a continuous chronology, the age of intervening stratigraphic layers needs to be assessed by interpolation between or extrapolation beyond dates obtained by these methods. This is normally achieved by use of an age-depth curve, in which dates are related to their stratigraphic position in a sediment core or profile (see figure 2.5). In sediments which are still accumulating, the uppermost layers can be assumed to date to the present; otherwise at least two dates are necessary for an age-depth curve to be drawn up.

If change through time at a single sequence is investigated via stratigraphy, the matching of different sequences over space is achieved via **correlation**. Features such as the European elm decline are best demonstrated to be of the same age if they are dated independently at every site. Correlation using pollen or other biostratigraphic evidence is likely to be imprecise and

Figure 2.5 Age-depth curve and accumulation rates for a lake sediment core with three ^{14}C dates



may lead to circular reasoning, but other means of correlation can be a reliable alternative to independent dating. Prominent among these is the use of volcanic ash layers, or **tephrochronology**, as marker horizons. Fossil soils and landforms can also sometimes be traced over considerable distances and are normally age-equivalent rather than time-transgressive.

Palaeoecological Techniques

Palaeoecology is the study of fossil organisms in order to reconstruct past environments. Accurate palaeoenvironmental reconstruction also depends on modern ecological data, and is consequently only as good as our knowledge of present-day ecosystems. A fundamental assumption made here is that relationships which can be observed at the present time also held good in the past. This principle is known as **uniformitarianism**. Take the case of *Hippopotamus* bones found in the middle of the arid Saharan desert and dated to between 10 000 and 6000 years ago. Assuming that the bones have not been moved significantly since death, then two possible explanations might be offered. The first and more obvious would be to employ our knowledge of modern hippo ecology to infer that the Sahara was a land of lakes and rivers during the early Holocene. The alternative explanation would be that hippos have changed their habits and habitats, formerly being happy scaling sand dunes and only recently taking to wallowing in mud.

In this example the case for the present not being the key to the past seems somewhat absurd. But for organisms with

Table 2.3 Uniformitarian* assumptions in palaeoecology

1	We understand the environmental factors governing present-day plant and animal distributions.
2	Their ecological affinities have not changed through time.
3	Present (and past) distributions are (and were) in equilibrium.
4	Modern analogues exist.
5	The origin (taphonomy) of a fossil assemblage can be established, and that . . .
6	. . . this assemblage is not biased by contamination or differential preservation.
7	Fossils can be identified to a meaningful level of taxonomic resolution.

* *sensu* actualism

short life cycles, such as beetles (Coleoptera), the validity of the principle of uniformitarianism may be called into question. Up to 20 000 generations of beetles have lived and died since the last glacial maximum, perhaps sufficient for evolutionary changes to have taken place. In fact, detailed comparisons of modern and ice-age beetle skeletons have revealed no discernible changes in their anatomy, indicating evolutionary stability (Coope, 1970). Even so, uniformitarianism remains assumed not proven, and the further back in time we go, the more uncertain this assumption becomes. Uniformitarianism remains an important issue in other ways too; for example, in the problem of interpreting an assemblage of fossils which has no modern equivalent, or analogue (see table 2.3).

The process by which a group of living plants or animals is transformed into an assemblage of fossils is by no means always a straightforward one. This transformation and the subsequent interpretation of the fossil assemblage is illustrated here with reference to pollen analysis. Pollen analysis, or palynology, is only one of many branches of palaeoecology, but it is broadly representative of the wider subject in terms of the methods it employs.

Principles of pollen analysis

Palynology is the single most important branch of palaeoecology for the late Pleistocene and Holocene. It has been extremely widely applied since it was first applied in stratigraphic studies by the Swede Lennart von Post at the start of the twentieth century (Faegri and Iversen, 1989). Its basis is that pollen grains and spores produced as part of plant reproduction may be incorporated and preserved in lake muds, peat bogs or other sediments which can later be analysed to reconstruct the vegetation of an area.

Table 2.4 Relative pollen productivity of selected temperate trees

× 1	lime (*Tilia*), ash (*Fraxinus excelsior*), maple (*Acer*)
× 2	beech (*Fagus*)
× 4	elm (*Ulmus*), spruce (*Picea*)
× 6	hornbeam (*Carpinus*)
× 8	alder (*Alnus*), birch (*Betula*), hazel (*Corylus*), pine (*Pinus*), oak (*Quercus*)

Source: Andersen (1973)

Pollen production starts with a life assemblage, such as mixed temperate woodland. Within the woodland ecosystem, pollen is produced more abundantly by some plants than by others (see table 2.4); in particular, productivity is much greater for wind-pollinated than for insect-pollinated taxa. Of the many pollen grains produced, only a few are used for reproductive purposes, with the remainder being dispersed through the environment. Pollen dispersal involves several pathways, notably via high- and low-level air-borne transport and via a water-borne route in runoff and streams (Tauber, 1965). Each pathway introduces its own selective bias into the dispersal process. High-level wind dispersal, for example, leads to a well-mixed pollen 'rain' with the pollen of some taxa being carried over long distances, notably the conifers (gymnosperms), whose pollen grains have sacs to help keep them air-borne.

The processes of pollen production, dispersal and deposition thus generate a death assemblage which is significantly different in its composition from the initial life assemblage. This is further altered by sediment diagenesis and by differential fossil preservation and destruction (Birks and Birks, 1980, ch. 1). Pollen is best preserved under **anaerobic**, typically acid conditions, such as are encountered in blanket peat bogs. The by-now fossil assemblage can be investigated through field and laboratory work, typically involving coring at a field site (see plate 2.3) followed by laboratory preparation of samples. In order that only the sporopollenin remains intact, inorganic components of the sediment such as carbonate and silica are removed by reagents like hydrochloric and hydrofluoric acids. After mounting on a glass slide, pollen grains can be examined under the microscope, usually at magnifications between ×400 and ×1000. Individual pollen taxa have different shapes, sizes, number of apertures and other features which allow them to be differentiated (Moore et al., 1991). Identification is normally to the level of genus for trees and shrubs, and only to family for many herbs and grasses: species-level identification is the exception rather than the rule in palynology.

We can illustrate the sequence described above by means of the example shown in figure 2.6. In this hypothetical case only four plant taxa are present; pine (*Pinus*) and deciduous

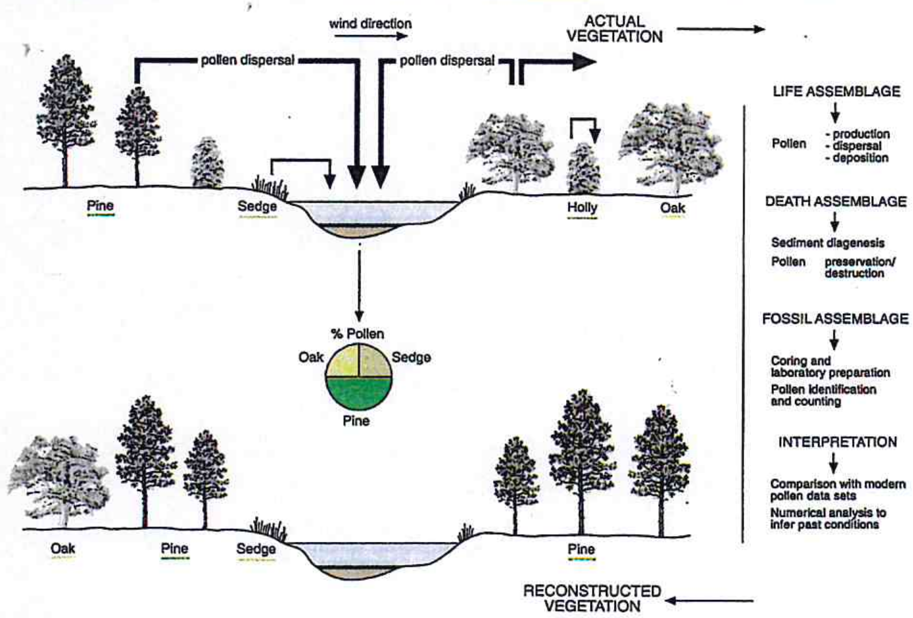


Figure 2.6 Vegetation reconstruction based on pollen analysis (see text for explanation)

oak (*Quercus*) form separate woodland stands either side of a lake, holly (*Ilex aquifolium*) is an understorey shrub, and there is a fringing mat of sedges (Cyperaceae) around the lake. Although pine and oak produce abundant pollen, the prevailing wind direction causes pine to be more strongly represented in the pollen rain reaching the lake. In contrast, holly is hardly represented at all, because it is insect-pollinated. The pattern of vegetation reconstructed from fossil pollen includes aspects of the original flora, but these have been modified and are incomplete. Not only does pine appear more abundant than oak, but it is difficult to know if the two types of tree formed separate or mixed woodland stands. The sedge pollen could also be interpreted in more than one way, for while some species form sedge swamps, other species occupy very different habitats such as tundra. The rarity of holly pollen grains might cause it to be overlooked altogether in the reconstruction. It is precisely in order to overcome some of the biases illustrated here that palaeoecologists carry out studies of modern pollen deposition and preservation as an integral part of their work.

The reconstruction shown in figure 2.6 is based on the relative proportions of different pollen taxa. To obtain representative proportions it is normally necessary to count at least 200 pollen grains in any sample. However, even if a much larger

Table 2.5 Absolute and relative pollen-counting statistics

	Absolute numbers			Relative proportions		
	pine	*oak*	*sedge*	*pine*	*oak*	*sedge*
time 1	10	5	5	50%	25%	25%
time 2	10	5	35	20%	10%	70%

number than this were to be counted, the 'relative' approach has certain inherent limitations. Imagine that the lake in our example was progressively infilled with sediment, and that as it became shallower so sedge swamp largely replaced open water. Tree pollen would be outnumbered by a superabundance of locally produced sedge pollen (see table 2.5). In a pollen count based on relative proportions, pine and oak would appear to decline in importance between times 1 and 2 even though the surrounding woodland vegetation was in fact unchanged. In an attempt to overcome this difficulty palynologists often exclude some local, wetland pollen types from the total pollen sum on which percentages are based. On the other hand, it may be difficult to assign particular pollen types to a local wetland as opposed to a dryland origin with certainty. An alternative approach is to count 'absolute' numbers of microfossils to obtain pollen concentration per cm^3 of sediment. This is most often done by adding a known quantity of exotic markers, such as *Lycopodium* spores or polystyrene microspheres, into the sample during preparation. If a timescale is available for the site under study, concentration values can be translated into rates of pollen influx per year.

Most pollen analyses are carried out at sites – such as lakes – which have experienced more or less continuous sediment accumulation. A series of pollen samples from different depths in the sediment profile can then be presented as a pollen diagram, which conventionally shows depth or age on the vertical axis and frequency on the horizontal axes. Pollen and other microfossil diagrams are used extensively in this book, in some cases based on absolute counts, in others based on proportional ones. The diagrams sometimes show only selected taxa, or major pollen groups such as arboreal (tree) versus non-arboreal pollen (AP *vs* NAP). Figure 2.8A is a percentage pollen diagram from a site in southern France, which shows changes in selected important tree types against core depth. As the calibrated ^{14}C dates indicate, this section of the core covers the transitional period from the Late-glacial stage through to the middle of the Holocene. Like most pollen diagrams, it is subdivided into zones within which samples (or spectra) are similar in their essential characteristics. Before the advent of

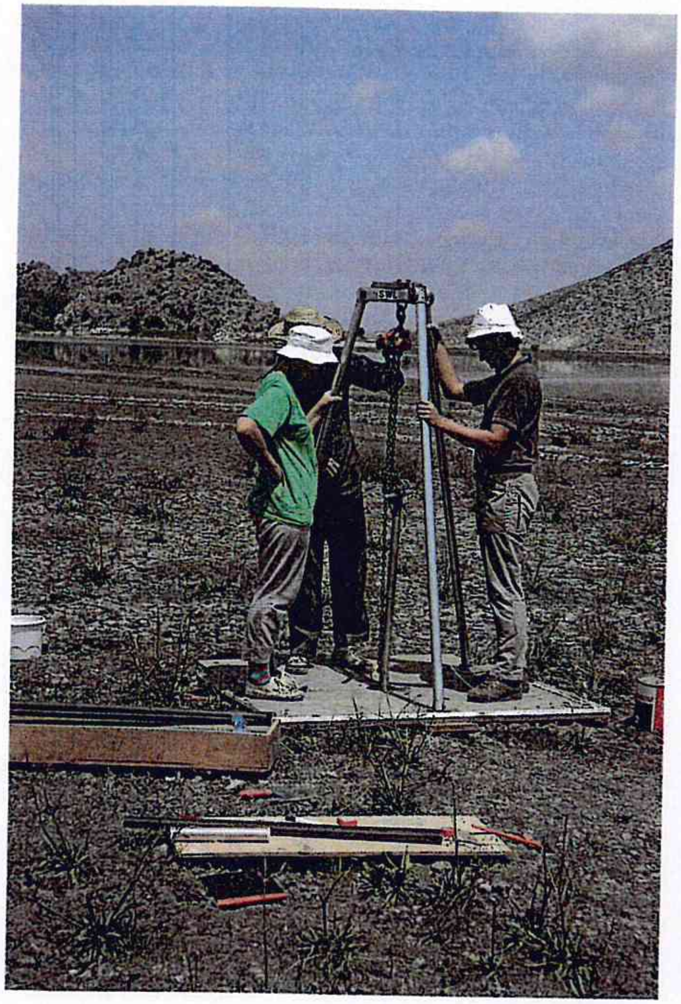
radiocarbon dating, these zones were assumed to be synchronous over whole regions. In the British Isles, for instance, Sir Harry Godwin proposed a pollen zonation scheme that began in the Late-glacial (zones I–III) and continued through the Holocene to the present day (IV–VIII). Regional pollen zones are now known to be time-transgressive and are therefore used less widely than previously, but local pollen assemblage zones remain an essential part of individual site investigations. Increasingly, these local pollen zones are designated with the help of statistical methods such as cluster analysis (Birks, 1986b).

Pollen, climate and human impact

The initial objective of a pollen analytical study is normally to reconstruct past vegetation, but there is often a second objective – that of establishing the factors which determined the former flora. The most important of these controlling factors are CLIMATE and HUMAN ACTIVITIES. Assuming little human disturbance, it is widely believed that vegetation will eventually achieve a state of equilibrium with the prevailing climate (Webb, 1986; Edwards and MacDonald, 1991a; Ritchie, 1995). Modern data on climate–plant relationships may therefore be applied to pollen-based vegetation histories to produce estimates of past temperatures or rainfall levels (Birks, 1981). In West Africa, for example, the montane forest tree *Olea hochstetteri* is today restricted to land above 1100 m. Its pollen is prominent in the bottom part of a diagram from the lowland site of Lake Bosumtwi, and this has been used to infer that prior to 8500 years ago temperatures here were at least 2–3°C colder than at present (Maley and Livingstone, 1983).

Comparable northern European examples are the arctic/alpine species, dwarf birch (*Betula nana*) and mountain avens (*Dryas octopetala*), whose pollen or macrofossils are common in Late-glacial assemblages. In the British Isles, dwarf birch is today virtually restricted to the Scottish Highlands, where mean summer temperatures are below 22°C (Connolly and Dahl, 1970). But before accepting this as the temperature of a typical July day in southern England 11 000 years ago, it should be remembered that species distribution is a function of more than one climatic variable. In an attempt to overcome this problem, Johannes Iversen (1944) combined two climatic variables and several different indicator species. He showed, for example, that holly can tolerate cool summers but not cold winters, whereas mistletoe (*Viscum album*) is the reverse – requiring summer warmth but being relatively hardy in winter. Ivy (*Hedera helix*) lies somewhere between the other two

Plate 2.3 Coring at the edge of lake Gölhisar in southern Turkey. The resulting 8.3 m-long core produced a pollen record spanning the whole of the Holocene



in its climatic tolerances (see figure 2.7). Using the overlapping ranges of the three shrubs in combination it is possible to provide more precise temperature reconstructions than if each were used on its own, and to provide an indication of past climatic continentality.

To extend this approach to a 'multivariate' analysis of vegetation and climate requires the use of more complex numerical methods, possible only with the help of a computer. Tom Webb and colleagues (Webb, 1985; Bartlein et al., 1986) developed numerical, multivariate techniques for reconstructing

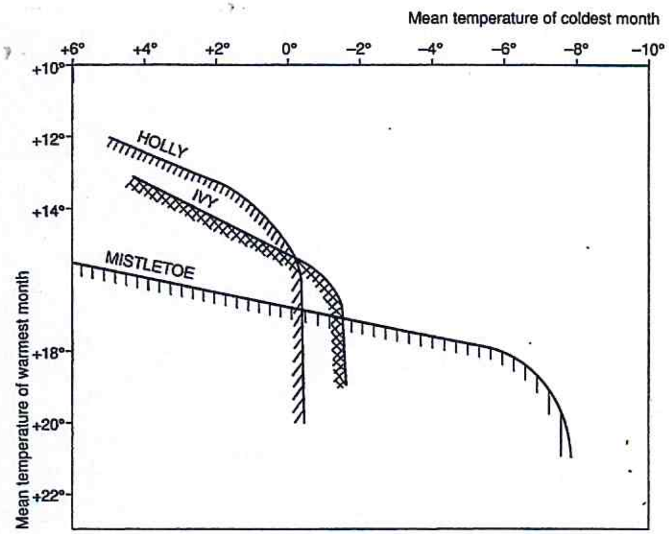


Figure 2.7 The thermal limits of holly, ivy and mistletoe (after Iversen, 1944)

climate directly from pollen data. This collapses the normal two-step reconstruction of pollen to vegetation to climate into a one-step process, involving the creation of so-called pollen–climate 'response surfaces'. Climatic calibration of modern and fossil pollen data has allowed maps to be produced showing temperature, precipitation or air mass distribution for different times in the past over eastern North America (Webb et al., 1993). This procedure cannot be applied so easily in Europe because human impact on vegetation has a much longer antiquity here, and mapping work has concentrated more on pollen or vegetation than on climate (Huntley and Birks, 1983; Prentice et al., 1996).

The interpretation of pollen records in terms of past human impact is, if anything, more difficult than it is for climate (Edwards and MacDonald, 1991b). Many former land uses, such as hunting-fishing-foraging in Mesolithic Europe, have no modern analogue for comparison. Furthermore, it is often the indirect consequences of human impact that are most obvious palynologically (Edwards, 1979, 1982). The pollen of wheat, barley and other cereal crops is poorly dispersed and only weakly represented in European pollen diagrams until historical times. More prominent evidence of prehistoric agriculture instead comes from the pollen of ruderals (weedy plants) like docks (*Rumex* sp.) and nettles (*Urtica* sp.) (Turner, 1964; Behre, 1981, 1986). These species take advantage of disturbed ground regardless of cause, and are therefore not diagnostic indicators of anthropogenic disturbance. Many, for example, were relatively abundant during the Late-glacial before a stable forest cover had been established (Godwin, 1975; Huntley and Birks, 1983, p. 518ff.).

The recognition of human disturbance in pollen diagrams usually involves a combination of evidence, including changes in the overall composition of vegetation as well as indicator species (Maguire, 1983; Birks et al., 1988). Amongst the latter would be grassland indicators such as ribwort plantain (*Plantago lanceolata*), whilst the former would include changes in the AP–NAP ratio. Where several lines of evidence converge – as in the pollen diagram shown in figure 6.11 (see p. 190) – interpretation is likely to be straightforward. In other cases a former vegetation change could be attributable to more than one cause: the mid-Holocene European elm decline, for example, has been variously attributed to prehistoric agriculture, disease, climate change, or a combination of more than one of these factors. Equifinality is the term used when a similar end-result could have been produced by more than one different cause. It is one of the tasks of palaeoecology to test between competing causal hypotheses, for instance in explaining what has caused many Scandinavian lakes to be acidified (see chapter 7).

Plant macrofossils and charcoal

Notwithstanding the enormous success of palynology in reconstructing former vegetation, climate and land use, pollen research alone cannot hope to answer all our questions about past ecologies. In some situations other sources of proxy data may prove more informative, and the potential range of biological indicators is very wide – from microscopic algae to mammoths. On calcareous rocks where pollen sites are almost absent, such as English chalk, land snails (Mollusca) have proved of great value in the investigation of land use and vegetation history (see Technical Box VII, pp. 196–7). Beetles (Coleoptera) are another important type of organism which can respond rapidly and sensitively to changes in climate (see Technical Box III, pp. 69–70). In **palaeolimnological** studies aquatic indicators such as diatoms (see Technical Box IX, pp. 230–1) are especially valuable. The remains of many types of micro-organisms can furnish valuable palaeoecological information in different contexts, ranging from Protozoa such as Foraminifera (calcareous marine zooplankton) and testate amoebae (Warner, 1988), through ostracods (Holmes, 1992) to the pigments left by phytoplankton (Leavitt, 1993). The value of alternative, complementary sources of information is well illustrated by the example of plant macrofossils, that is, plant remains like seeds, leaves and wood that are visible with the naked eye.

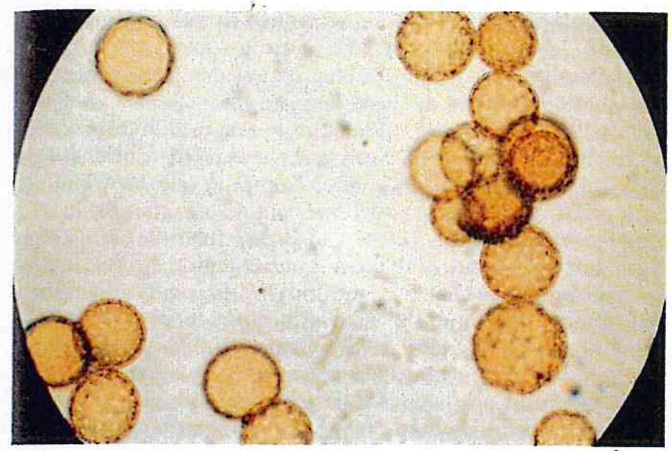


Plate 2.4 Pollen grains of *Chenopodium album*, an indicator of open-ground conditions, *c.*40 microns (0.04 mm) in diameter

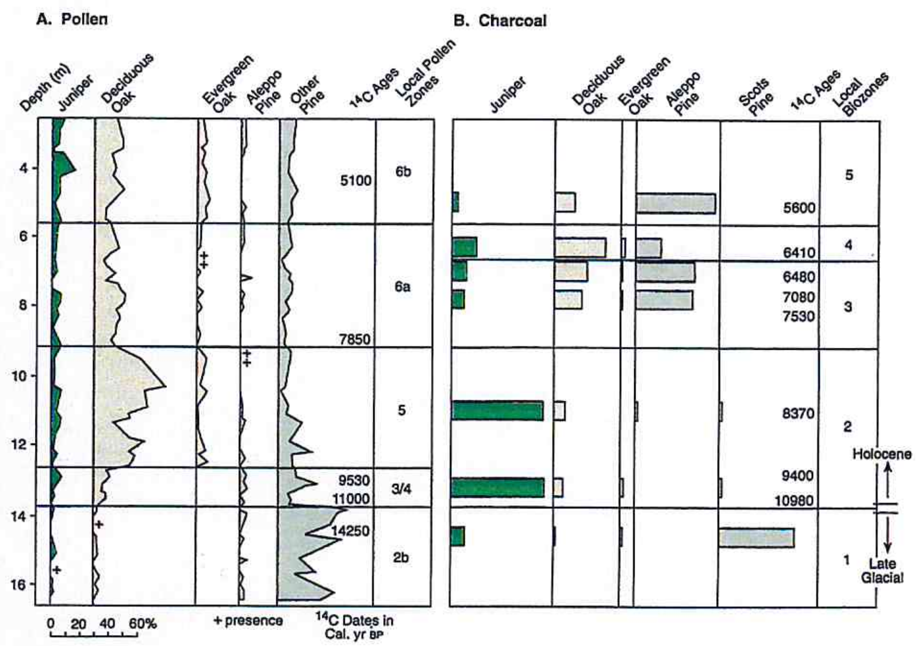


Figure 2.8 Comparative records of Late-glacial and early Holocene changes in selected trees from southern France; **A** pollen record from a valley peat bog at Tourves (after Nicol-Pichard, 1987); **B** wood charcoal record from prehistoric rockshelter site of Fontbrégoua (data from Thiébault, 1997)

Most of the same principles examined in relation to pollen – life and death assemblages, and so forth – apply also to plant macrofossils, and indeed to other organisms. However, bigger sample volumes are required because macrofossils are larger than pollen, and counting statistics are more difficult because of the wider variety of parts preserved. (This applies even more to bone remains of mammal, reptile and bird faunas.) Compared with pollen, however, plant macrofossils have a number of advantages (Birks and Birks, 1980, p. 66ff.). They form part of the once-living plant rather than the reproductive stage of the life cycle. As a result, they are usually identifiable to species level and the ecological inferences that can be drawn from them are relatively precise. Unlike pollen grains, macrofossils are not transported far from their point of origin and their local presence can be assumed with greater confidence. They can also help record plants with poor pollen production or dispersal, such as the arctic flower, mountain avens (*Dryas octopetala*).

On archaeological sites macrofossils are often found in the form of charred seeds or wood charcoal, which offers a direct indication of the plant resources exploited by former human communities (Helbaek, 1969; Vernet and Thiébault, 1987). The study of wood charcoals – or anthracology – can offer useful insights into localized vegetation changes which complement those provided by pollen analysis (Chabal, 1992). Figure 2.8 compares the Late-glacial to mid-Holocene woodland history from a charcoal record in an excavated prehistoric rockshelter in southern France with a pollen core from a valley mire nearby. The latter represents the more continuous record of regional vegetation change, but there are some problems with interpretation of pollen data which the charcoal can help to resolve. For instance, the Late-glacial period before *c.*11 500 Cal. yr BP has high pine pollen percentages at Tourves, but does this really mean that pine trees were common in the vicinity? Pine is notoriously well dispersed by the wind and often makes up a significant proportion of the total pollen found in open, exposed landscapes such as tundra, even though it may not be growing locally (Ritchie, 1987). Fortunately, we know in this case that pine trees really were present – and Scots pine at that – because the Epi-Palaeolithic occupants of the rockshelter collected it at the time (figure 2.8B). On the other hand, changes in the charcoal record can also have been caused by cultural preferences, say, because one type of wood burns better than another. The switch from juniper to oak and pine charcoal just before 7500 Cal. yr BP, for example, coincides with the appearance of the first (Neolithic) farmers in this area. This apparent vegetation change may reflect the fact that oak and pine woods were being cut down to make way for cultivated crops, rather than because juniper – which had been used previously – became less common in the regional vegetation. Charcoal particles are also found in lake sediments and peat bogs (Patterson et al., 1987), but as microscopic, rather than macroscopic, fragments. While they are normally too small to permit identification to tree type, these tiny charcoals do offer very important information about fire histories (e.g. Bennett et al., 1990).

Plant macrofossil analysis has been applied to a range of palaeoecological problems. In western Norway, as in Provence, there were some striking differences between the vegetation records indicated by pollen and plant macrofossils, in this case from lake sediment cores (Birks, 1993). In particular, birch (*Betula*) was prominent in the pollen diagrams, but absent in the macrofossil record from the same lakes during a Late-glacial stadial. As the latter normally provides the more reliable index of local conditions, we can assume that birch was not, in fact, growing in the vicinity of the lake, and that its pollen must have been blown in from further afield. One task for which plant macrofossils are well suited is reconstructing the history of wetlands as they progress through a **hydrosere** succession from open water through sedge swamp and fen carr to peat bog (Walker, 1970). Because peat deposits are almost entirely made up of plant macrofossils, studies of peat stratigraphy and of macrofossils often amount to the same thing. The relative importance of the remains of *Sphagnum* moss and heathland plants (e.g. *Calluna*, *Eriophorum*) in peat bog sequences provides an indicator of surface wetness and hence changes in hydro-climatic conditions (e.g. Barber, 1981).

A very different application of plant macrofossil analysis has been developed in the arid lands of the American Southwest. Here, the desert packrat (*Neotoma* spp.) collects plant material from within a 30 m radius to form midden deposits (Wells, 1976; Betancourt et al., 1990) rather as the beaver (*Castor*) does in the boreal forest. These middens represent a sample of the local vegetation, which is subsequently preserved by being soaked and cemented by packrat urine! The packrat's somewhat unhygienic habits have given us an excellent record of vegetational and climatic history. The plant macrofossils that make up the packrat middens are not only identifiable to the level of species, but can also be dated directly by radiocarbon. They have been used, for instance, to show that during the late Pleistocene the Sonoran and southern Mohave deserts supported woodland of pinyon pine (*Pinus monophylla*) and other pygmy conifers, not the creosote-bush scrub existing today (Spaulding et al., 1983). This implies at least 400 m lowering in the altitudinal range of coniferous trees prior to

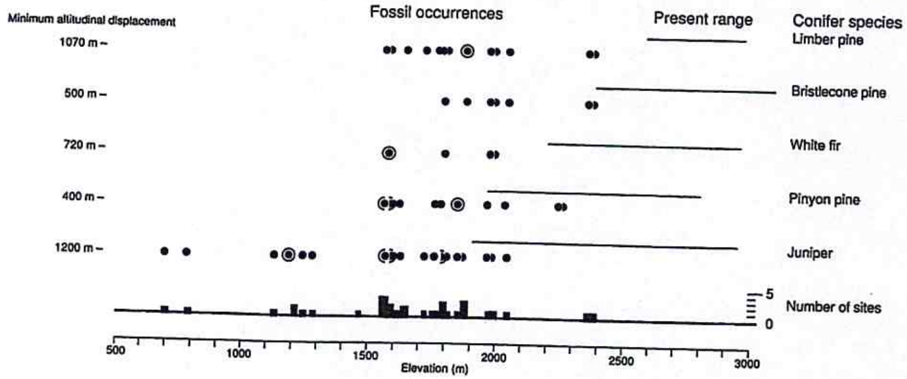


Figure 2.9 Present-day and late Pleistocene altitudinal ranges of coniferous trees in the Mohave desert based on packrat middens (after Spaulding et al., 1983). Circled dots indicate an occurrence at more than one site at that elevation

13 000 Cal. yr BP (see figure 2.9). Middens formed by hyrax (Procaviidae) and similar 'desert rats' are found in Africa, south-west Asia and Australia, opening the possibility of applying more widely this remarkable source of palaeoecological data (Scott, 1990; Fall, 1990; Nelson et al., 1990).

Geological Techniques

Recent Earth history can be understood by examining sediment sequences and landforms at the Earth's surface. These can be classified into a series of different geomorphological regimes, each with its own characteristic set of depositional environments (see table 2.6). As in palaeoecology, the reconstruction of past geological environments is based on uniformitarian principles; in other words, our knowledge of the present is used as the key to understanding the past. Modern data about Earth surface processes are applied to fossil landforms and sediments, first, to assign them to a particular geomorphological regime,

Table 2.6 Characteristics of different geomorphic regimes

Geomorphic regime	*Sedimentation*	*Sediments*	*Landforms*
Lacustrine	continuous (lake bed)	lake marl	shoreline terrace
Fluvial	semi-continuous	river gravel	palaeo-channel
Coastal marine	semi-continuous	beach sand	raised beach
Aeolian	semi-continuous	dune sand	fossil dune
Glacial	discontinuous	till	end-moraine
Mass movement	discontinuous	'head'	solifluction lobe

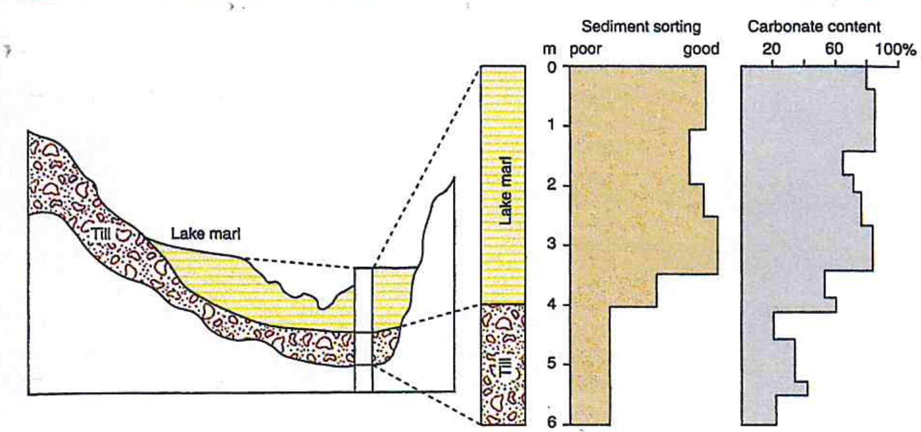


Figure 2.10 Environmental reconstruction based on landform and sedimentary evidence

and, second, to interpret their origin in terms of past processes and climates. No-analogue situations are encountered less often than in palaeoecology, although they do sometimes occur, notably for high-magnitude events. The so-called 'Channelled Scabland' of Washington, for instance, was for many years explained in terms of observable modern fluvial processes (Baker and Bunker, 1985). It is now realized that it was in fact created by catastrophic, short-lived flooding when ice-dammed Lake Missoula burst at the end of the last glaciation.

LANDFORMS may be produced by erosion, deposition, or a combination of both types of process. Terrace features, in particular, are formed by a phase of sedimentation – say, by an aggrading river – followed by a period of erosional downcutting. Terraces provide valuable information on previous river, lake and sea levels that can be recorded by survey and mapping. However, it is possible to misinterpret the origin of terraces and other landforms unless data on surface morphology are supplemented by an analysis of the underlying sediments. Earth histories can therefore be studied most effectively by combining landform-based techniques with those based on sediment stratigraphies (Goudie, 1994), as illustrated by the sequence shown in figure 2.10. Here a glacial moraine is overlain by lake sediments, now being dissected by fluvial erosion. Of these three geomorphic regimes, only the first two have left sedimentary records, and their contrasting physical and chemical characteristics are recorded on one side of the diagram.

Minerogenic or clastic sediments, such as river gravels, comprise rock particles of allochthonous origin – that is, they were brought to the point of deposition from elsewhere. Analysis of their physical characteristics involves examining particle shape, size distribution and arrangement in space. One standard

Table 2.7 Comparative properties of glacial and aeolian sediments

	Glacial till	*Dune sand*
particle shape	angular	spherical
grain surface	fractured	smooth
size distribution	mix of fine and coarse particles	homogenous sand
particle sorting	very poor	good
sediment structures	weak or absent	obvious cross-bedding
particle alignment	may be parallel to direction of ice flow	not evident

measurement technique is particle size analysis, carried out by sieving, pipette analysis and other methods, from which size and sorting indices may be derived (Folk, 1974). The characteristics of two contrasting clastic sediment types are listed in table 2.7. Till and dune sand are easily distinguished from each other, but differentiation may be harder in other cases. Soliflucted 'head' has similar properties to some tills, for instance, while desert dune and beach sand could be confused on the basis of sediment characteristics alone. Minerogenic deposits can also be analysed for their clay and 'heavy' mineral compositions, and for their mineral magnetic properties.

PRECIPITATES, such as lake marl, are sediments chemically deposited from aqueous solution and are therefore **autochthonous** in origin. Most commonly, deposition takes place in standing-water conditions such as are encountered in lakes and the sea, and involves one of two main mechanisms. The first of these is biochemical 'fixing' of solutes by aquatic or marine organisms, like the polyps responsible for building up coral reefs. The second is direct chemical precipitation from salinealkaline water where solutes are highly concentrated, for example in a playa lake or coastal lagoon (see plate 2.5). Different types of salt are deposited from different solutions, depending on the chemical composition and concentration of the water (Eugster and Kelts, 1983). Another type of chemical precipitate is speleothem carbonates found in caves, including stalagmites and flowstones. Like other precipitates, speleothems can be analysed isotopically (see below).

Biogenic sediments comprise primarily organic carbon and are most often autochthonous, for example in a raised peat bog. More rarely, organic matter may be allochthonous, as when soils are eroded and washed into streams and lakes. Biogenic sediments typically contain abundant micro- and macrofossils and they are therefore extensively used in pollen and other palaeoecological investigations.

In practice, most sediments comprise a mixture of clastic, chemical and biogenic elements. One way in which the differing proportions of elements making up a sediment sample can be assessed is by using a system of sediment classification, such as that developed by the Danish palaeoecologist J. Troels-Smith (Birks and Birks, 1980, p. 38ff.; Aaby and Berglund, 1986). Of Troels-Smith's six main sediment categories, three are biogenic, a further two are minerogenic, and a further one is either chemical or biogenic (see table 2.8). Although this system of sediment classification is widely applied, its usage is restricted to freshwater lake and mire sediments in humid temperate regions. Lakes and mires certainly provide some of the most accessible and best-studied sediment 'archives' recording Holocene environmental change, with the former being especially informative when the sediments are annually laminated, or varved. Sediment in small temperate-zone lakes and mires typically builds up at about 1 m per thousand years, so that a full Holocene record is likely to be represented by around 10 m of accumulated sediment (Webb and Webb, 1988).

Table 2.8 Elements of the Troels-Smith sediment classification

(a)	*Substantia humosa*: undifferentiated disintegrated organic matter
(b)	*Turfa*: macroscopic plant remains of below-ground origin and mosses; e.g. roots
(c)	*Detritus*: macroscopic plant remains of above-ground origin; e.g. leaves
(d)	*Limus*: lake mud of biogenic or chemical origin; e.g. marl
(e)	*Argilla*: fine-grained minerogenic particles; e.g. clay
(f)	*Grana*: medium or coarse-grained minerogenic particles; e.g. sand

Lakes and mires are, however, but two among many types of stratigraphic record. The deep ocean bed, for example, is a depositional environment which has proved invaluable for understanding the Earth's history during the Pleistocene ice age. Deep-sea sediment cores contain a record of long-term fluctuations between cold glacial and warmer interglacial climates (Lowe and Walker, 1997). The slow rate of sedimentation is ideal for studying such long spans of time, although it means that the uppermost sediments formed during the Holocene are typically no more than 50 cm thick. This does not allow the kind of detailed analysis of change through time that is possible in lake sequences, although in enclosed seas like the Mediterranean and in some other basins, sediment accumulation rates have often been faster than in most of the open ocean. Here, pollen and other analyses can help link deep-sea core records to those on land (Rossignol-Strick, 1995). At the opposite extreme, the stratigraphic record preserved in ice sheets can be thousands of metres thick. In the Summit cores from Greenland the Holocene is represented by no less than

Plate 2.5 Chemical sediments precipitated from a hyper-saline lake

1600 m vertical thickness of ice, and this permits annual time resolution – although coring to this depth does present some formidable logistical challenges (see Technical Box II, pp. 64–5).

Stable isotope analysis

As noted earlier in this chapter, most chemical elements have rare forms with the 'wrong' number of neutrons. Some of them, like ^{14}C, are isotopically unstable and undergo radioactive decay with time, but others are stable and provide information instead on past environmental conditions. Among the most important are the isotopes of carbon and oxygen.

The ratio between the heavy isotope oxygen-18 (^{18}O) and the lighter, and much more common, oxygen-16 (^{16}O) varies with temperature, and this fact led its discoverer – Harold Urey – to claim he had a 'geological thermometer' in his hands. Cesare Emiliani measured this $^{18}O/^{16}O$ (or $\partial^{18}O$) ratio in the fossil shells of tiny marine planktonic foraminifera to try to reconstruct changes in Pleistocene temperatures. The oxygen-isotope stages that he devised now form the way that the **Quaternary** ice age is classified into episodes of warmer and colder climate. It has since been established that a more important direct factor than temperature in influencing the marine isotopic record was the growth and decay of ice sheets.

These lock up isotopically 'light' water as ice, and leave the oceans enriched in the heavier isotope ^{18}O. In other words, oxygen-isotope analysis of deep-sea cores provides an index of past glaciation, rather than of temperature. Temperature has been a more important control over the $\partial^{18}O$ record in some other systems, such as freshwater lake sediments, ice sheets and speleothems. Oxygen isotopes can indicate seasonal as well as annual variations in temperature or rainfall, for example in mollusc shells as they accrete (Abell et al., 1995). In salt lakes, the stable oxygen-isotope ratio varies primarily according to the intensity of evaporation from the water surface, and here they reflect past salinity changes (Talbot, 1990).

Stable carbon-isotope analysis provides a way to determine past changes in the type of plants and in the global carbon cycle, for example as the atmospheric concentration of greenhouse gases has fluctuated. Some plants, such as tropical grasses and sedges, incorporate CO_2 as a four-carbon molecule during photosynthesis; the so-called C4 pathway. Other grasses, along with trees and shrubs, use a three-carbon (C3) pathway. The relative importance of C4 plants can be established by measuring the ratio of two stable isotopes of carbon, ^{13}C and ^{12}C. That ratio, expressed as $\partial^{13}C$, has been analysed in plant macrofossils, soils, ostracods, land snails, timbers and peats among others (Goodfriend, 1992; Street-Perrott et al., 1997; Heaton et al., 1995). In lake sediments, the stable carbon-isotope ratio will have been affected by the productivity of aquatic algae, as well as the type of vegetation in the lake catchment. Carbon isotopes have also been used to help determine past changes in human (and animal) diet. Maize, along with some other grain crops, incorporates CO_2 via the C4 pathway (Chisholm, 1989). The relative dietary importance of C4 plants can be established by measuring the $\partial^{13}C$ ratio in human bone. That ratio is high in the case of diets with a large proportion of C4 plants, such as maize, or of animals fed off C4 plants. The $\partial^{13}C$ values are low when C3 plants, including wild foodstuffs, dominate the food chain.

Geomorphology and climate

Climatic conditions exert a powerful influence over geomorphic processes and landforms. Geomorphologists such as Jean Tricart (1972) and Julius Büdel (1982) have argued that at a global scale distinct morphoclimatic regions can be recognized, within which a common set of processes are dominant. Thus ice-wedge polygons are exclusively found in the periglacial zone of winter freezing and summer thaw, where the average annual temperature is below −6°C (Black, 1976; Harris, 1986). Geomorphic processes such as freeze-thaw are so closely linked

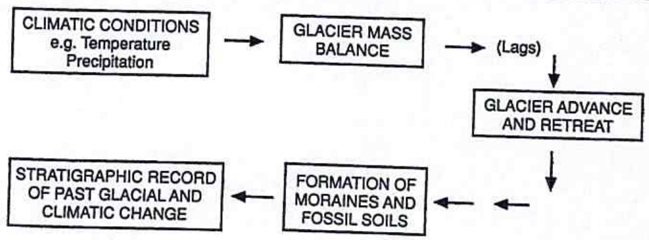


Figure 2.11 The glacier–climate linkage

to climate that it is possible to calibrate them against temperature or precipitation. However, not all landforms are harmoniously adjusted to the prevailing climate because some were produced in the past when climatic conditions were different. These relict landforms may furnish indications about past or palaeoclimates.

One geomorphic regime intimately associated with specific climatic conditions is the GLACIAL environment. Valley glaciers form under cold, moist climates, specifically in areas where the accumulation of winter snow exceeds summer melting or ablation. The surplus snow builds up, becomes compressed to form ice, and moves downslope by basal sliding or flowage. In due course the glacier enters a zone of warmer air temperatures in which there is a net deficit between accumulation and ablation, and eventually the glacier terminates. The mass balance of a land-based glacier is therefore a function of winter snowfall on the one hand, and summer temperature and cloudiness on the other (Drewry, 1985). Under colder or more snowy conditions the glacier snout will advance downvalley, under warmer or drier conditions it will retreat. On the other hand, glacier response to a change in climate is rarely instantaneous, mainly because it takes time for an increase or decrease in ice volume in the accumulation zone to be felt at the snout (see figure 2.11). Even though changes are transmitted downvalley four times faster than the actual rate of ice movement, there will still be a time lag in glacial response amounting to a few decades for an alpine glacier, and a few millennia for a major ice sheet. In any case, portions of the major ice sheets – past and present – have had margins ending in the sea rather than on land, and in these cases sea level has been a more important control over ice-melt than air temperature.

Geologic evidence of former ice extent comes from moraines and till sequences which can be dated by ^{14}C measurement, dendrochronology and lichenometry. The existence of moraines several km downvalley from modern alpine glacier snouts represents convincing evidence of Holocene fluctuations in climate (Grove, 1979; Porter, 1981). The most recent of these moraines is little more than a century old and has only been

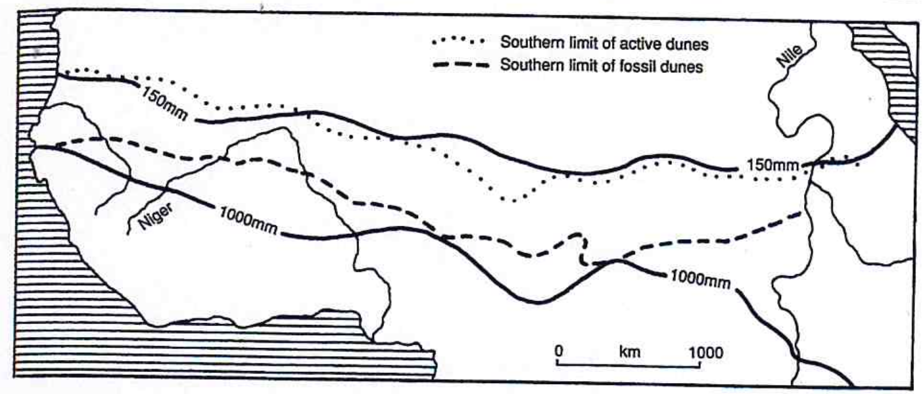


Figure 2.12 Modern and fossil dune limits on the south side of the Sahara (based on Goudie, 1983)

partially recolonized by vegetation (see plate 2.6). In contrast to those found in alpine valleys, the moraines created by Pleistocene ice sheets now lie far removed from any existing glaciers and are now covered by well-developed soils and vegetation. Some ice sheets have survived right through the Holocene, notably in Antarctica and Greenland, and cores drilled into them have provided high-resolution records of late Quaternary climate change (see Technical Box II, pp. 64–5).

Geomorphology provides important records of Holocene environmental change in the tropics and sub-tropics too, although in this case the primary climatic variations have been from wet to dry (Douglas and Spencer, 1985). Two of the most climatically sensitive geomorphic regimes are lakes and desert dunes. AEOLIAN processes are most effective where sediment is not bound together by surface soil or vegetation and which has a poor internal cohesiveness when dry. Wind-blown silt, or loess, originated in some cases from deflation of glacial sediments left behind by retreating ice sheets, and in others from high-altitude rock weathering in mountains. However, it is in arid lands where ideal conditions for aeolian processes are most often met, with climatically induced moisture deficiency causing vegetation to be sparse or absent. Because of the close relationship between aeolian processes and climatic aridity, fossil dunes help to provide an indication of the changing extent and location of the world's arid zone (Sarnthein, 1978; Lancaster, 1990).

On the southern (Sahelian) margin of the Sahara, active sand dunes are today restricted to land with less than 150 mm mean annual rainfall (see figure 2.12). Stable, vegetated dune ridges are none the less found up to 450 km south of this limit, in areas presently receiving up to 1000 mm rainfall (Grove and Warren, 1968). These mark a former extension of the Sahara southwards, when there was reduction of effective

Plate 2.6 Oblique air photograph of the Arolla (left), Pièce (centre) and Tsidjioure Nouve (right) glaciers in Switzerland. The moraines and trim-lines visible downvalley of the modern glaciers were created at times during the Holocene when the ice was more extensive and the climate cooler and snowier. The most recent advance occurred during the 'Little Ice Age', which ended in the nineteenth century

precipitation to only a quarter of its present value. Traces of fossil sand dunes have even been traced beneath the tropical rainforests of Zaire and Amazonia (Tricart, 1974). Aeolian landforms not only indicate times of climatic aridity but also reveal former wind strengths and directions (e.g. Wells, 1983; Bowler and Wasson, 1984; Thomas, 1984). Because dunes are aligned relative to the dominant wind vector, they may be used to establish palaeo-wind directions, and differences from the present can then be computed.

If dunes indicate periods of climatic aridity, high-level LAKES usually reflect wet phases or **pluvials**. In hydrologically closed or non-outlet lakes, water is lost only through evaporation, assuming there is negligible groundwater outflow (Street, 1980). Lake area and water depth adjust so as to increase or reduce evaporation losses, in the same way as a glacier expands or contracts in response to changes in snowfall and melting. In essence, closed lakes can behave like giant rain gauges: more rainfall (or less evaporation) and the water level rises, less rainfall (or more evaporation) and the water level falls. Direct geomorphological evidence of former high levels comes from

shoreline terraces left 'high and dry' above present-day lakes; indirect evidence comes from proxy records of changing lake salinity (Street-Perrott and Harrison, 1985). When a lake moves from being open to closed, it also changes from freshwater to saline (Langbein, 1961). Instead of being washed away down the outflow stream, solutes are retained within the lake water and are progressively concentrated by evaporation. Some closed lakes have become so concentrated that they are now hyper-saline; the Dead Sea, for example, now has a salinity ten times that of ordinary sea water.

Palaeosalinities are recorded in the chemical composition of lake sediments, because different salts are precipitated as lake salinity increases; for instance, in a series from freshwater carbonates through sulphates to chlorides (Bowler, 1981; Teller and Last, 1990). Equally useful in salinity reconstruction are the surviving hard parts of aquatic organisms such as diatoms, ostracods and molluscs (de Deckker, 1981; Gasse, 1987; Holmes, 1992). Numerical estimates for past salinity and ionic composition can be obtained by means of statistical transfer functions which relate modern species assemblages to water chemistry in a 'training set' which spans a range of different environments. This approach has proved especially successful for diatoms (Fritz et al., 1991; Cumming and Smol, 1993). Ostracods can also be analysed chemically for trace elements such as strontium, calcium and magnesium, whose ratios are salinity- and temperature-dependent, and for stable isotopes (Chivas et al., 1986; Engstrom and Nelson, 1990).

Climatic calibration of lake-level data has been attempted using a number of different approaches. The first involves the calculation of past precipitation, runoff and evaporation for a lake and its catchment based on a model of the contemporary water balance (Street, 1980). The major assumption that this approach makes is that past temperatures – the primary control over evaporation rates – are known. An alternative approach developed by John Kutzbach (1980) utilizes a combined energy- and water-balance model. Both have been used to provide estimates of early Holocene precipitation in Africa and elsewhere in the tropics (Kutzbach, 1983). Some lakes are mainly fed from below rather than above, and act as groundwater 'windows' rather than as rain gauges. In such cases the whole groundwater system has to be modelled, as was done for the lakes of the Parker's Prairie Sandplain in the Mid-West United States (Almendinger, 1993).

Geo-archaeology

The interface between earth science and past human activity has been variously termed geo-archaeology (Davidson and

Shackley, 1976) and archaeological geology (Rapp and Gifford, 1982). It includes the study of sediments from archaeological sites but also extends off-site to investigate, for example, palaeogeographic reconstructions of coastlines of historical importance (e.g. Kraft et al., 1977) and sites buried by river alluvium (Brown, 1997). One of the ways archaeology and geology have helped each other has been in the study of CAVE deposits. Most caves are ancient landforms, and during the Quaternary they have served as the residence of many creatures, including owls, bears and our own human ancestors. Indeed, it is popularly believed that all early hominids were 'cave men'. As caves have been infilled with sediment, so they have also incorporated in their stratigraphies materials brought in by their former occupants: bone remains, stone tools, charcoal from hearths and food refuse. These provide archaeologists with rich pickings for fieldwork, and cave sequences such as at Franchthi in Greece (Jacobsen and Farrand, 1987) and Cresswell Crags in England (Jenkinson, 1984) have been meticulously excavated in recent years in collaboration with environmental scientists, often employing techniques such as micromorphological analysis (Courty et al., 1989).

Interdisciplinary research of the geo-archaeological type has many potential advantages (Vita-Finzi, 1978; Stein and Farrand, 1985). Archaeological and palaeoenvironmental data can be matched and dated from the same stratigraphic context without the need for correlation between sites. In the case of Lake Mungo in Australia, it was survey work on lunette dunes by geomorphologist Jim Bowler that led to discovery of an archaeological site in the first place (Bowler et al., 1972). The human remains found at Mungo pushed back the antiquity of aboriginal peoples in Australia to over 25 000 Cal. yr BP, and also helped to provide a timescale for the geological and climatic changes recorded in the site's sediment stratigraphy (see figure 2.13). In particular the Mungo skeleton provided an age for the beach gravels with which it was associated, and hence for a phase of high lake levels.

Over the course of the Holocene, human impact on geomorphic processes has increased progressively, especially on the rate of SOIL LOSS from slopes. Soil is eroded and lost as part of the natural geological cycle of denudation, but under dense grassland or woodland that loss is minimized because rainfall is intercepted and runoff reduced (Thornes, 1987). Removal of the vegetation, by human or natural agencies, causes increased potential for erosion by raindrop impact, surface sheetwash, rilling, gullying, leaching and wind action (Limbrey, 1975; Boardman et al., 1990). Nor is impact restricted to the site of degradation. Eroded soils are washed downslope and downstream to push up the sediment loads carried by rivers.

In fact, soil erosion is an important indicator of the degree of disturbance in an ecosystem.

Geomorphic environments which record soil erosion histories may be broadly divided into those of net erosion and those of net deposition (Bell, 1983; Bell and Boardman, 1992; Dearing, 1994). The former ought to provide the more accurate soil loss data, but on their own, degraded hillslopes are an unreliable guide to erosion history, their mere existence providing no clue to the age or origin of erosion. However, where archaeological sites lie *in situ*, the former soil surface may be preserved beneath them. Examination of soils beneath prehistoric burial mounds has shown that on the English chalk downlands, later Holocene erosion has almost completely removed the original cover of loess (Catt, 1978).

Most studies of past erosion rates have focused on the point of deposition rather than the point of erosion. Suitable environments include valley fills, floodplain and estuarine deposits, and lake sediments. In all cases sediment yield is used as a surrogate for soil loss, which tends to underestimate true erosion rates. In using sedimentary records it is best to use relatively small, 'closed' catchments such as drainage basins with a high trap efficiency (e.g. Brown and Barber, 1985), often containing a lake or reservoir (e.g. Foster et al., 1985). Mineral magnetic properties such as **magnetic susceptibility** and magnetic remanence allow different types of iron minerals to be distinguished, and this often helps pinpoint the source area from which sediments were derived (Thompson and Oldfield, 1986). Soils, for example, have a different magnetic 'signature' from the parent bedrock because of the formation of secondary ferrimagnetic oxides such as magnetite, and this is enhanced if the soil has been subject to burning.

Models of Environmental Reconstruction

Up to now, palaeoenvironmental techniques have been dealt with individually. The question therefore remains of how they and individual site investigations should be fitted together to provide a comprehensive picture of landscape change. One conceptual framework frequently used is the environmental system model, in which different physical, chemical and biological components of the environment are seen to be interconnected over a definable area of space. A good example is the LAKE-CATCHMENT ECOSYSTEM, in which a natural and easily defined system boundary is provided by the drainage basin watershed (Oldfield, 1977; O'Sullivan, 1979). Nutrients and other materials are washed down slopes into streams, and eventually end up in the lake. Not only does the lake therefore

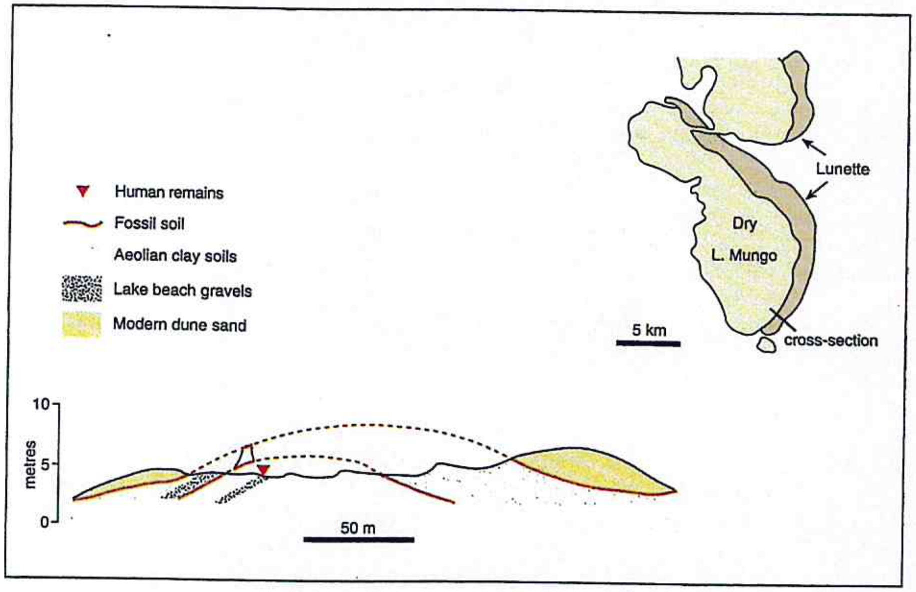


Figure 2.13 Sediment and archaeological sequence at Lake Mungo, Australia (after Bowler et al., 1972)

serve to integrate changes over the whole catchment – that is, across space – it also records the history of such changes through time. Whereas in a normal drainage basin materials would be lost from the environmental system along with stream discharge, in a lake catchment they are trapped and incorporated in bottom sediments, so preserving a record of past environmental processes (Binford et al., 1983; Smol, 1992). Furthermore, different techniques of dating and analysis may be used in combination on the same lake sediment cores; for example, to reconstruct both catchment erosion rates via sediment influx calculations, and a history of land use via the pollen record.

So although palaeoenvironmental data typically derive from specific field localities, the material they contain does not originate solely from that one place. In other words, each individual site locus actually represents an aggregated record over a wider catchment area. The size and shape of each site catchment will vary according to the type of site and other factors (Vita-Finzi, 1974). Whereas a lake's catchment is effectively the same as its drainage basin, that of a pollen core on a peat bog is defined atmospherically, being the area upwind during the pollen production season (see figure 2.14). Birds, animals and human populations may be thought of as having catchments too. The faunal remains found in a cave reflect the area exploited by its occupants, such as the hunting territories of

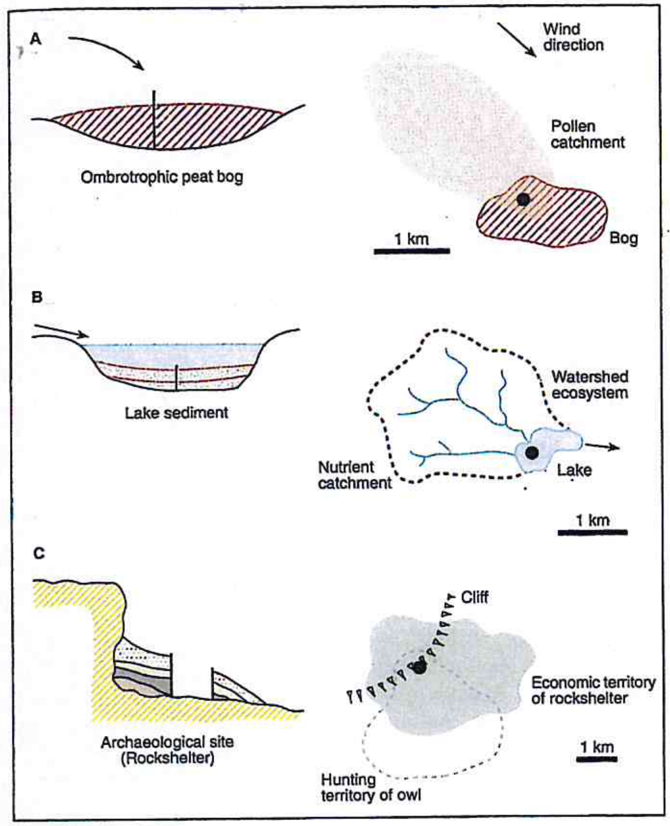


Figure 2.14 Palaeoenvironmental sites and their catchments: **A** ombrotrophic peat bog; **B** lake watershed ecosystem; **C** rockshelter

an owl or a Mesolithic hunting band. The nature of the territory around an archaeological site helps to explain the on-site record of economic activities preserved in seed and bone remains, or in historical archives for individual estates or monasteries (Higgs and Vita-Finzi, 1972; Butlin and Roberts, 1995).

Of course, catchment areas may not be so easily defined as the examples shown in figure 2.14. Even so, they serve to emphasize that environmental reconstruction takes place principally at the landscape scale (Birks et al., 1988). In order to study environmental processes operating at a smaller scale – say, within a stand of trees – it is necessary to select those sites which only reflect local changes (Bradshaw, 1988). Whereas the pollen entering a large lake normally reflects regional-scale vegetation (*c.*100 km^2), that found in a small woodland hollow will have travelled a short distance and have been determined instead by the nearby flora (*c.*0.1 ha). At the opposite extreme, large-scale changes involving vegetation formations or climatic zones (*c.*1 million km²) can only be derived by combining data from many individual sites. For example, pollen records from over 400 core localities are held in the North American Pollen Database, and 650 in its equivalent for Europe, both used to reconstruct continent-wide vegetation changes since the last glacial maximum.

Just as there is a spatial resolution to the reconstruction of past environments, so there is also a temporal resolution. Palaeoecological and geological evidence is resolvable down to one year in the case of varves and tree-rings, but more usually the time interval between sample points is 10 to 100 years (see table 2.2). While this inevitably means the 'loss' of information about year-to-year environmental variations, it is compensated by the 'gain' of a smoothed record of underlying trends over centuries and millennia. These are timescales of change which are too long to be observed directly, and for which proxy records are uniquely well suited.

Tying the various complementary sources of palaeoenvironmental data together needs good liaison between different specialists. For this reason many projects on environmental history involve a team of researchers working over a number of years at a particular site or in a particular area, whether this be on a long core through an ice cap, or a study of cultural transformation of a regional landscape. Integrating information from archaeological excavations, historical records, off-site pollen cores and alluvial sequences, and dating them all, can be a major challenge (not to say headache), but at its best the combined results can reveal much more than the sum of their component parts. For example, in the Ystad project a coordinated team of archaeologists, palaeoecologists, landscape historians and others worked together over a number of years to uncover the history of one area of southern Sweden (Berglund, 1991). There are limitations to any one technique, but together they are able to provide us with an insight into a past that would otherwise remain inaccessible.

Only selected techniques of dating and environmental reconstruction have been reviewed in this chapter or included in subsequent technical boxes. Fortunately, several texts provide further detail on these and other palaeoenvironmental methods, for example, by Birks and Birks (1980), Bradley (1984), Berglund (1986), and Lowe and Walker (1997).