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Dating Methods and the Quaternary

Whatever withdraws us from the power of our senses; whatever makes the past, the distant or the future, predominate over the present, advances us in the dignity of thinking beings.

Samuel Johnson

1.1 Introduction

The Quaternary is the most recent period of the geological record. Spanning the last 2.5 million years or so of geological time[1] and including the **Pleistocene** and **Holocene** epochs,[2] it is often considered to be synonymous with the 'Ice Age'. Indeed, for much of the Quaternary, the earth's land surface has been covered by greatly expanded ice sheets and glaciers, and temperatures during these **glacial** periods were significantly lower than those of the present. But the Quaternary has also seen episodes, albeit much shorter in duration, of markedly warmer conditions, and in these **interglacials** the temperatures in the mid- and high-latitude regions may have exceeded those of the present day. Indeed, rather than being a period of unremitting cold, the hallmark of the Quaternary is the repeated oscillation of the earth's global climate system between glacial and interglacial states.

Establishing the timing of these climatic changes, and of their effects on the earth's environment, is a key element in Quaternary research. Whether it is to date a particular climatic episode, to estimate the rate of operation of past geological or geomorphological processes, or to determine the age of an artefact or cultural assemblage, we need to be able to establish a chronology of events. The aim of this book is to describe, evaluate and exemplify the different dating techniques that are applicable within the field of

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Quaternary science. It is not, however, a dating manual. Rather, it is a book that is written from the perspective of the user community as opposed to that of the laboratory expert. It is, above all, a book that lays emphasis on the practical side of Quaternary dating, for the principal focus is on examples or case studies. To paraphrase the words of the actor John Cleese, it is intended to show just what Quaternary dating can do for us!

In this chapter, we examine the development of ideas relating to geological time and, in particular, to Quaternary dating. We then move on to consider the ways in which the quality of a date can be evaluated, and to discuss some basic principles of radioactive decay as these apply to Quaternary dating. Finally, we return to the Quaternary with a brief overview of the Quaternary stratigraphic record, and of Quaternary nomenclature and terminology. These sections provide important background information, and both a chronological and stratigraphic context for the remainder of the book.

1.2 The Development of Quaternary Dating

Early approaches to dating the past were closely associated with attempts to establish the age of the earth. Some of the oldest writings on this topic are to be found in the classical literature where the *leitmotif* of much of the Greek writings is the concept of an infinite time, equivalent in many ways to modern day requirements for steady-state theories of the universe (Tinkler, 1985). This position contrasts markedly with that in post-Renaissance Europe where biblical thinking placed the creation of the world around 6000 years ago, and when the end of the universe was predicted within a few hundred years. This restricted chronology for earth history derives from the biblical researches of James Ussher, Archbishop of Armagh, who in 1654 published his considered conclusion, based on Old Testament genealogical sources, that the earth was created on Sunday 23 October 4004 BC, with ‘man and other living creatures’ appearing on the following Friday. Another momentous event in the Old Testament, the ‘great flood’, occurred 1656 years after the creation, between 2349 and 2348 BC.

In his magisterial review of the history of earth science, Davies (1969) has observed that although modern researchers have tended to scoff at Ussher’s chronology he was, in fact, no fanatical fundamentalist but rather a brilliant and highly respected scholar of his day. It is perhaps for this reason that his chronology had such a pervasive influence on scientific thought, although it is perhaps less clear to modern geologists why it still forms a cornerstone of contemporary creationist ‘science’! During the eighteenth and nineteenth centuries, however, with the development of uniformitarianist thinking in geology,[3] the pendulum began to swing once more towards longer timescales for the formation of the earth and for the longevity of operation of geological processes, a view encapsulated by James Hutton’s famous observation in his *Theory of the Earth* (1788) that ‘. . . we find no vestige of a beginning, no prospect of an end’.

The difficulty was, of course, that pre-twentieth-century scientists had no bases for determining the passage of geological time. One of the earliest attempts to tackle the problem was William McClay’s work in 1790 on the retreat of the Niagara Falls escarpment, which led him to propose an age of 55 440 years for the earth (Tinkler, 1985). Others tried a different tack. The nineteenth-century scientist John Joly, for example, calculated the quantity of sodium salt in the world's oceans, as well as the amount added every year from rock erosion, and arrived at a figure of 100 million years for the age of the earth. Increasingly, however, came an awareness that even this extended time frame was simply not long enough to account for the entire history of the earth and, moreover, for organic evolution, a view that was underscored by the publication of Darwin's seminal work *Origin of Species* in 1859. Further challenges to the Ussher timescale and to its successors came from the field of archaeology, with noted antiquarians such as John Evans (and his geological colleague Joseph Prestwich) arguing, on the basis of finds of ancient handaxes, for a protracted period of human occupation extending into a period of antiquity '.... remote beyond any of which we have hitherto found traces' (Renfrew, 1973).

It was into this atmosphere of chronological uncertainty that Louis Agassiz introduced his revolutionary idea of a 'Great Ice Period', which arguably marks the birth of modern Quaternary science. This notion, first propounded in 1837, was initially received with a degree of scepticism by the geological establishment, but the idea not only of a single glaciation but, indeed, of multiple glaciations rapidly gained ground. By the beginning of the twentieth century, most geologists were subscribing to the view that four major glacial episodes had affected the landscapes of both Europe and North America, although the basis for dating these events remained uncertain. An early attempt at establishing a glacial–interglacial chronology was made by the German geologist Albrecht Penck, using the depth of weathering and 'intensity of erosion' in the northern Alpine region of Europe to estimate the duration of interglacial periods. On this basis, an age of 60 000 years was assigned to the Last Interglacial and 240 000 years to the Penultimate Interglacial, the duration of the Quaternary being estimated at 600 000 years (Penck and Bruckner, 1909). An alternative approach using the astronomical timescale based on observed variations in the earth's orbit and axis[4] again arrived at a similar figure, although if older glaciations recorded in the Alpine region were included, the time span of the Quaternary was extended to around 1 million years (Zeuner, 1959). This figure has since been widely quoted and, for the first half of the twentieth century at least, was generally regarded as the best estimate of age for the Quaternary.

At about the time that the Quaternary glacial chronology was being worked out for the European Alps, the first attempts were being made to develop a timescale for the last deglaciation, using laminated or layered sediment sequences which were interpreted as reflecting annual sedimentation cycles. These are known as **varves**, and are still employed as a basis for Quaternary chronology at the present day (section 5.3). Some of the earliest studies were made on the sediments in Swiss lakes and produced estimates of between 16 000 and 20 000 years since the last glacial maximum (Zeuner, 1959), results that are not markedly different from those derived from more recent dating programmes. The seminal work on varved sequences, however, was carried out in Scandinavia where Gerard de Geer (Figure 1.1) developed the world's first high-resolution deglacial chronology in relation to the wasting Fennoscandian ice sheet (section 5.3.3.1). This approach was subsequently applied in North America to date glacial retreat along parts of the southern margin of the last (Laurentide) ice sheet (Antevs, 1931).

The early years of the twentieth century saw the development of another dating technique which is still widely used in Quaternary science, namely **dendrochronology** or **tree-ring dating** (section 5.2). Research on tree rings has a long history, and the relationship between tree rings and climate (a field of study known as **dendroclimatology**) has

Figure 1.1 *Gerard de Geer measuring varves at Beckomberga, Stockholm, in 1931. Varve chronology was the first dating technique to provide a realistic estimate of Quaternary time (photo: Ebba Hult de Geer, courtesy of Lars Brunnberg and Stefan Wastegård)*

intrigued scientists since the Middle Ages. Indeed, some of the earliest writings on this subject can be found in the papers of Leonardo da Vinci (Stallings, 1937). The basics of modern dendrochronology, however, were formulated by the American astronomer Andrew Douglass, who was the first to link simple dendrochronological principles to historical research and to climatology (Schweingruber, 1988). Together with Edmund Schulmann, he founded the world-famous Laboratory for Tree-Ring Research at the University of Arizona in 1937. In Europe, it was not until the end of the 1930s that dendrochronology began to gain a foothold, largely through the work of the German botanist, Bruno Huber. His research laid the foundation for the modern school of German dendrochronology which has remained at the forefront of tree-ring research in Europe to the present day.

The most significant advance in Quaternary chronology, however, came during and immediately after the Second World War, with the discovery that the decay of certain radioactive elements could form a basis for dating. Although measurements had been made more than 30 years earlier on radioactive minerals of supposedly Pleistocene age (Holmes, 1915), it was the pioneering work of Willard Libby and his colleagues that led to the development of radiocarbon dating, and to the establishment of the world's first radiocarbon dating laboratory at the University of Chicago in 1948. During the 1950s and 1960s, other **radiometric methods** were developed that built on technological advances (increasingly sophisticated instrumentation) and an increasing understanding of the nuclear decay process. These included uranium-series and potassium–argon dating (Chapter 3), while a growing appreciation of the effects on minerals and other materials of exposure to radiation led to the development of another family of techniques which includes thermoluminescence, fission track and electron spin resonance dating (Chapter 4). In the late 1960s and 1970s, advances in molecular biology enabled post-mortem changes in protein structures to be used as a basis for dating (amino acid geochronology), while remarkable developments in coring technology led to the recovery of long-core sequences from ocean sediments and from polar ice sheets, out of which came the first marine and ice-core chronologies. The last two decades of the twentieth century have been characterised by a series of technological innovations that led not only to a further expansion in the range of Quaternary dating techniques, but also to significant improvements in analytical precision. A major advance was the development of accelerator mass spectrometry (AMS), which not only revolutionised radiocarbon dating (Chapter 2), but also made possible the technique of cosmogenic nuclide dating (section 3.4). The last decade has also witnessed the creation of the high-resolution chronologies from the GRIP and GISP2 Greenland ice cores, and from the Vostok and EPICA cores in Antarctica (section 5.5).

These various developments and innovations mean that Quaternary scientists now have at their disposal a portfolio of dating methods that could not have been dreamed of only a generation ago, and which are capable of dating events on timescales ranging from single years to millions of years. The year 2004 sees the 350th anniversary of the publication of the second edition of Ussher's ground-breaking volume on the age of the earth. How he would have reconciled the recent advances in Quaternary dating technology with his 6000-year estimate for the age of the earth is difficult to imagine!

1.3 Precision and Accuracy in Dating

Before going further, it is important to say something about how we can judge the quality of an age determination. Two principal criteria reflect the quality of a date, namely **accuracy** and **precision**, and these apply not only to dates on Quaternary events, but to all age determinations made within the earth, environmental and archaeological sciences. For dating practitioners and for interpreting dates, it is important to understand the meaning and significance of these terms. **Accuracy** refers to the *degree of correspondence* between the true age of a sample and that obtained by the dating process. In other words, it refers to the degree of *bias* in an age measurement. **Precision** relates to the *statistical uncertainty* that is associated with any physical or chemical analysis that is used as a basis for determining age. As we shall see, all dating methods have their own distinctive set of problems, and hence each age measurement will have an element of uncertainty associated with it. These uncertainties tend to be expressed in statistical terms and provide us with an indication of the level of precision of each age determination (Chapter 2).

An example of the distinction between accuracy and precision in the context of a dated sequence is shown in Figure 1.2. In sample A, there is close agreement in terms of mean age between the four dated samples, and the standard errors (indicated by the range bars) are small; however, the dates are 2000–2500 years younger than the 'true age'. These dates are therefore precise, but inaccurate. In sample B, the reverse obtains; the dates cluster around the true age but have wide error bars. Hence they are accurate but imprecise. In sample C, however, the dates are of similar age and have narrow error bars. These age determinations are both accurate and precise, which is the optimal situation in dating.

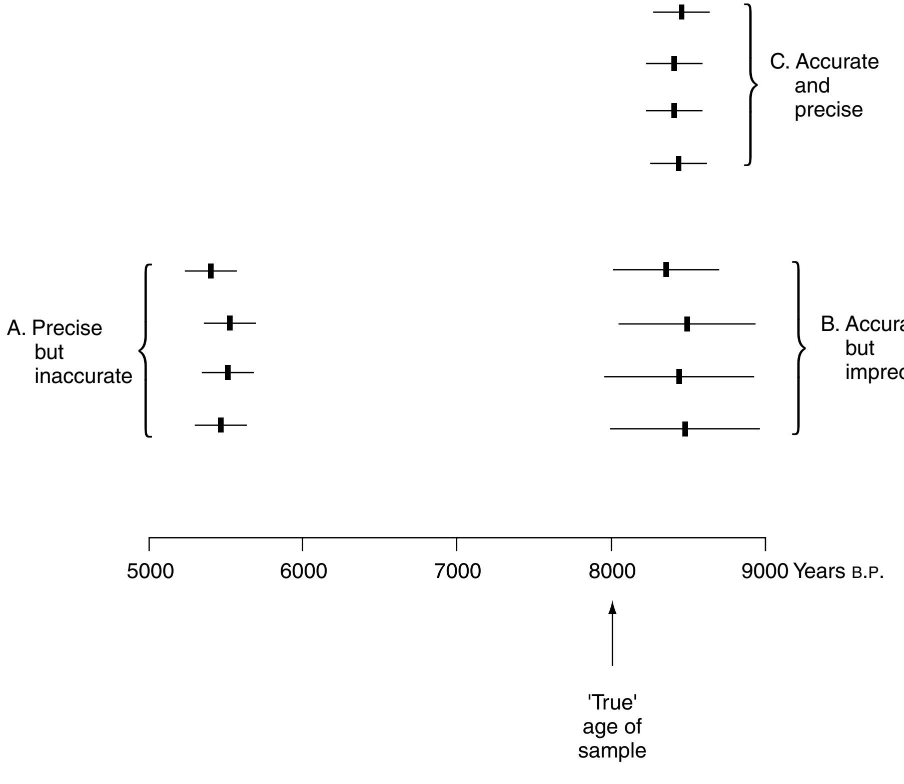

Figure 1.2 *Accuracy and precision in a dated sequence (modified after Lowe and Walker, 1997); see text for details*

1.4 Atomic Structure, Radioactivity and Radiometric Dating

Radiometric dating methods form a significant component of the Quaternary scientist's dating portfolio. Indeed, half of the chapters in this book that deal specifically with dating methods are concerned with radiometric dating. All radiometric techniques are based on the fact that certain naturally occurring elements are unstable and undergo spontaneous changes in their structure and organisation in order to achieve more stable atomic forms. This process, known as **radioactive decay**, is time-dependent, and if the rate of decay for a given element can be determined, then the ages of the host rocks and fossils can be established.

In order to understand the basics of radiometric dating, it is necessary to know something about atomic structure and the radioactive process. Matter is composed of minute particles known as **atoms**, the nuclei of which contain positively charged particles (**protons**), and particles with no electrical charge (**neutrons**), which together make up most of the mass of an atom. Third elements are **electrons**, which are tiny particles of negative charge and negligible mass that spin around the nucleus. Collectively, protons, neutrons and electrons are referred to as **elementary** or **sub-atomic particles**, and for many years were considered to be the fundamental building blocks of matter. With the development of large particle accelerators however, machines that are capable of accelerating samples to such high speeds that matter breaks down into its constituent parts, dozens of new sub-atomic particles have been discovered and current research suggests that atomic matter is made up of elementary particles from two families, **quarks** and **leptons**. Our understanding of electrons (which are members of the lepton family of particles) and their behaviour has also changed. At one time it was believed that electrons orbited the nucleus in shells (or orbitals), similar to the way in which the planets orbit the sun, and that in each of these orbits they had certain energy. However, the situation now appears to be more complex, as modern physics has shown that it is not possible to determine both the location *and* the velocity of a sub-atomic particle.[5] More recent work on atomic structure therefore envisages electrons with a particular energy existing in *volumes of space* around the nucleus, even though their exact location cannot be established. These volumes are known as **atomic orbitals**. The build-up of electrons in atomic orbitals allows scientists to explain many of the physical and chemical properties of elements, and lies at the heart of our modern understanding of chemistry.

When an atom gains or loses electrons, it acquires a net electrical charge, and such atoms are known as **ions**. The electrical charge can be positive or negative; a positive ion is referred to as a **cation** and a negative ion as an **anion**. The nature of the electrical charge is based on the number of protons minus the number of electrons, and is often referred to as the **valence**. Hence, an element with eight protons and eight electrons has a net electrical charge of 0. If it gains two electrons, it has a negative electrical charge (valence $= 2^-$) and it becomes an anion. If it loses two electrons, it develops a positive charge (valence $= 2^+$) and becomes a cation. **Ionisation**, which is the process whereby electrons are removed (usually) or added (occasionally) to atoms, is an important element in radiation (see below).

The atoms of each chemical element have a specific **atomic number** and **atomic mass number**. The former refers to the number of protons contained in the nucleus of an atom, while the latter is the number of protons plus neutrons. In other words, the mass number is the total number of particles (**nucleons**) in the nucleus. The atomic number is usually written in subscript on the left-hand side of the symbol for the chemical element (e.g. oxygen – $_{8}O$; uranium – $_{92}U$), while the atomic mass number is shown in superscript (e.g. ^{16}O; ^{238}U). In some elements, although the number of protons in the nucleus remains the same, the number of neutrons may vary. Elements that possess the same number of protons but different numbers of neutrons are referred to as **isotopes**. The number of electrons is constant for isotopes of each element, and hence they have the same chemical properties, but the isotopes differ in mass, and this will be reflected in change in the atomic mass number. Examples include carbon (^{12}C, ^{13}C, ^{14}C) and oxygen (^{16}O, ^{17}O, ^{18}O). Individual isotopes of an element are referred to as **nuclides**. Most of these are stable; in other words the binding forces created by the electrical charges are sufficient to keep the atomic particles together. In some cases, however, where there are too many or too few neutrons in the nucleus, for example, the nuclides are unstable and this results in a spontaneous emission of particles or energy to achieve a stable state. This is the process of **radiation** (or **radioactive 'decay'**), and such isotopes are known as **radioactive nuclides**.

Unstable nuclei can rid themselves of excess energy in a variety of ways, but the three most common forms are **alpha**, **beta** and **gamma decay**. In **alpha (α) decay**, a nucleus emits an *alpha particle* consisting of two protons and two neutrons, which is a nucleus of helium. Nuclides that emit alpha particles lose both mass and positive charge. The atomic mass number changes to reflect this, and the result is that one chemical element can be created by the decay of others. In **beta (β) decay**, a different kind of particle is ejected – an electron. The emission of a negatively charged electron does not alter mass, hence there is no change in atomic mass number. There is, however, a change in atomic number because the reason for the ejection of the electron is that as the nucleus decays, a neutron transmutes into a proton, and the nucleus must rid itself of some energy and increase its electrical charge. The emission of an electron, with its negative charge and small amount of excess energy, enables this to be achieved. The third common form of radioactivity is **gamma (γ) decay**. Here, the nucleus does not emit a particle, but rather a highly energetic form of electromagnetic radiation. Gamma radiation does not change the number of protons and neutrons in the nucleus, but it does reduce the energy of the nucleus. Gamma rays are not important in most forms of radiometric dating (with the exception of some short-lived isotopes: Chapter 3), but they do contribute to the build-up of luminescent properties in minerals (Chapter 4). In addition, the cosmic rays from deep space that constantly bombard the earth's upper atmosphere, and which initiate the chemical reaction that leads to the formation of radiocarbon (Chapter 2) and other cosmogenic isotopes (Chapter 3), are largely composed of gamma radiation.

An atom that undergoes radioactive decay is termed a **parent nuclide** and the decay product is often referred to as a **daughter nuclide**. Some parent–daughter transformations are accomplished in a single stage, a process known as **simple decay**. Others involve a more complex reaction in which the nuclide with the highest atomic number decays to a stable form through the production of a series of intermediate nuclides, each of which is unstable. This is known as **chain decay** and occurs, for example, in uranium series (section 3.3). The intermediate nuclides that are formed during the course of decay are therefore both the products (or daughters) of previous nuclear transformations and the parents in subsequent radioactive decay. Such nuclides are referred to as **supported**. Where the decay process involves a nuclide that has not, in itself, been created by the decay process, or where that nuclide has been separated from earlier nuclides in the chain through the operation of physical, chemical or biological processes, this is known as **unsupported decay**. The distinction between supported and unsupported decay is considered further in the context of ^{210}Pb (lead-210) dating (section 3.5.1).

Radioactive decay processes are governed by atomic constants. The number of transformations per unit of time is proportional to the number of atoms present in the sample and for each decay pathway there is a **decay constant**. This represents the probability that an atom will decay in a given period of time. Although the radioactive decay of an individual atom is an irregular (stochastic) process, in a large sample of atoms it is possible to establish, within certain statistical limits, the rate at which overall distintegration proceeds. In all radioactive nuclides, the decay is not linear but exponential (e.g. Figure 2.1) and is usually considered in terms of the **half-life**, i.e. the length of time that is required to reduce a given quantity of a parent nuclide to one half. For example, if 1 gm of a parent nuclide is left to decay, after $t_{1/2}$ only 0.5 gm of that parent will remain. It will then take the same period of time to reduce that 0.5 gm to 0.25 gm, and to reduce the 0.25 gm to 0.125 gm, and so on. The half-life concept is fundamental to all forms of radiometric dating.

1.5 The Quaternary: Stratigraphic Framework and Terminology

As we saw above, the Quaternary is conventionally subdivided into **glacial** (cold) and **interglacial** (temperate) stages, with further subdivisions into **stadial** (cool) and **interstadial** (warm) episodes. The distinction between glacials and stadials on the one hand, and interglacials and interstadials on the other, is often blurred, but glacials are generally considered to be cold periods of extended duration (spanning tens of thousands of years) during which temperatures in the mid- and high-latitude regions were low enough to promote extensive glaciation. Stadials are cold episodes of lesser duration (perhaps 10 000 years or less) when cold conditions obtained and when short-lived glacial readvances occurred. Interglacials, on the other hand, were warm periods when temperatures in the mid- and high latitudes were comparable with, or may even have exceeded, those of the present, and whose duration may have been 10 000 years or more. Interstadials, by contrast, were short-lived (typically less than 5000 years) warmer episodes within a glacial stage, during which temperatures did not reach those of the present day. This type of categorisation, which is based on inferred climatic characteristics, is known as **climatostratigraphy** (Lowe and Walker, 1997).

Evidence for former glacial and interglacial conditions (as well as stadial and interstadial environments) has long been recognised in the terrestrial stratigraphic record. Former cold episodes are represented by glacial deposits, by periglacial sediments and structures, and by biological evidence (such as pollen or vertebrate remains) which are indicative of a cold-climate régime. Interglacial and interstadial phases are reflected primarily in the fossil record (pollen, plant macrofossils, fossil insect remains, etc.), or in biogenic sediments that have accumulated in lakes or ponds during a period of warmer climatic conditions. However, because of the effects of erosion, especially glacial erosion, the Quaternary terrestrial stratigraphic record is highly fragmented and, apart from some unusual contexts such as deep lakes in areas that have escaped the direct effects of glaciation, long and continuous sediment records are rarely preserved. During the later twentieth century, therefore, Quaternary scientists turned to the deep oceans of the world, where sedimentation has been taking place continuously over hundreds of thousands of years. Indeed, many ocean sediment records extend in an uninterrupted fashion back through the Quaternary and into the preceding Tertiary period. One of the great technological breakthroughs of the twentieth century was the development of coring equipment mounted on specially designed ships (Figure 1.3) which enabled complete sediment cores to be obtained from the deep ocean floor, sometimes from water depths in excess of 3 km!

What these cores revealed was a remarkable long-term record of oceanographic and, by implication, climatic change. This is reflected in the **oxygen isotope 'signal'** (or trace) in marine microfossils contained within the ocean floor sediments. The variations in the ratio between two isotopes of oxygen, the more common and 'lighter' oxygen-16 (^{16}O) and the rarer 'heavier' oxygen-18 (^{18}O), are indications of the changing isotopic composition of ocean waters between glacial and interglacial stages. As the balance between the two oxygen isotopes in sea water is largely controlled by fluctuations in land ice volume,[6] downcore variations in the oxygen isotope ratio ($\delta^{18}O$) can be read as a record of glacial/interglacial climatic oscillations, working on the principle that ice sheets and glaciers would have been greatly expanded during glacial times but much less extensive during interglacials (Shackleton and Opdyke, 1973). The sequence can therefore be divided into a series of **isotopic stages** (marine oxygen isotope or **MOI stages**) and these are numbered from the top down, interglacial (temperate) stages being assigned odd numbers, while even numbers denote glacial (cold) stages. The record shows that over the course of the

Figure 1.3 *The* Joides Resolution*, a specially commissioned ocean-going drilling ship for coring deep-sea sediments (photo Bill Austin)*

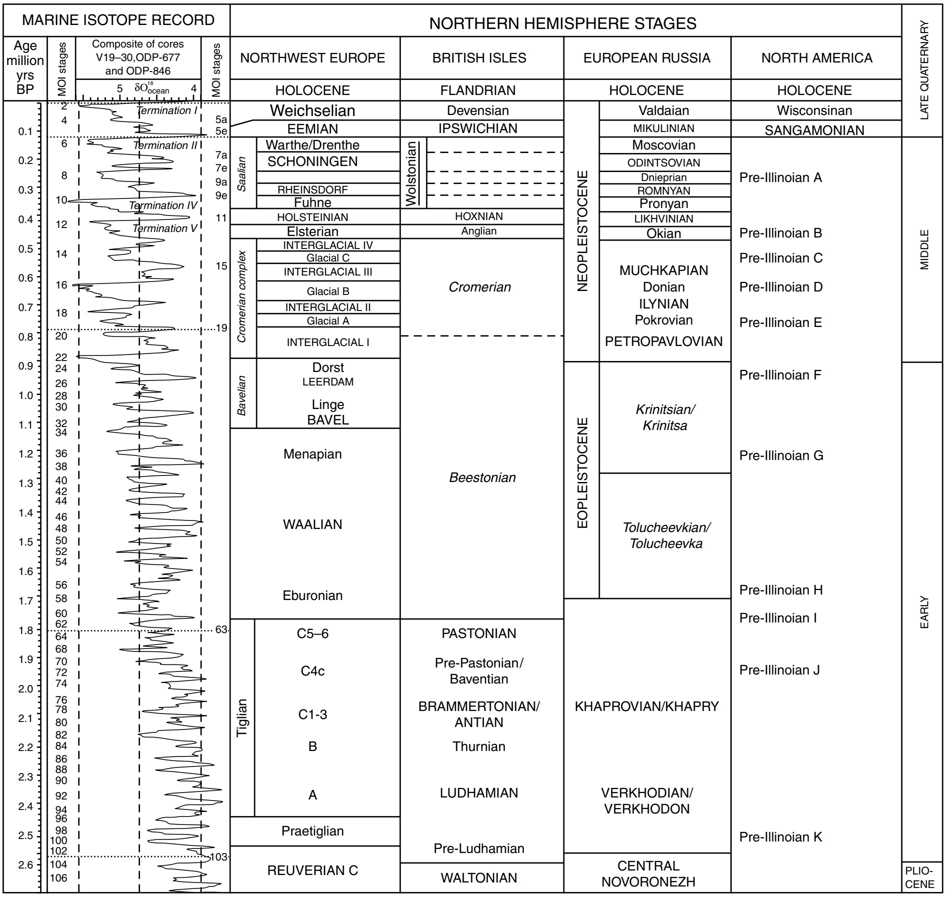

Figure 1.4 *The MOI record based on a composite of deep-ocean cores (V19–30; ODP-677 and ODP-846) (left) and the Quaternary stratigraphy of the northern hemisphere set against this record (right). The marine isotope signal shows the oxygen isotope stages back to 2.6 million years* BP*. In the correlation table, temperate (interglacial) stages are shown in upper case, while cold (glacial) stages are shown in lower case. Complexes which include both temperate and cold stages are in italics (based on Gibbard* et al.*, 2004)*

past 800 000 years or so, there have been around ten interglacial and ten glacial stages, while over the course of the entire Quaternary, back to 2.5 million years or so, more than 100 isotopic stages have been identified (Shackleton *et al.*, 1990; Figure 1.4, left). This is many more temperate and cold stages than has been recognised in terrestrial sequences, and hence the deep-ocean isotope signal provides a unique proxy record[7] of global climate change. It also constitutes an independent climatostratigraphic scheme against which terrestrial sequences can be compared. This approach is exemplified in a number of case studies discussed in the following pages, while the use of oxygen isotope stratigraphy as a basis for dating Quaternary events is considered in Chapter 7.

The Quaternary terrestrial stratigraphic sequence in different areas of the northern hemisphere, and possible correlatives with the MOI record, is shown on the right-hand side of Figure 1.4. Broadly speaking, the Quaternary can be divided into Early, Middle and Late periods. The **Late Quaternary**, which includes the present interglacial, last cold stage and last interglacial (ca. 0–125 000 years ago), is readily correlated between the various regions, and this warm–cold–warm sequence can be equated with the MOI stratigraphy (MOI stages 1–5). Prior to that, however, the various regional records are less easily correlated. During the **Middle Quaternary**, which encompasses the period from ca. 125 000 to 780 000 years ago, a number of glacial and interglacial episodes are reflected in the various terrestrial stratigraphic records, but several of these have no formal designation. Moreover, some designated warm and cold periods appear to contain both warm and cold stages (often several), while there are clearly gaps or hiatuses in the stratigraphic sequences. As a result, correlation not only between each of the regional sequences but also between these and the MOI 'template' becomes increasingly uncertain. These problems are even more acute during the **Early Quaternary** (prior to ca. 780 000 years ago) where the number of designated stages is even fewer, and both regional correlations and links with the MOI sequence become increasingly speculative. In many ways, Figure 1.4 exemplifies one of the principal difficulties in Quaternary science, namely the lack of a universal dating technique that is applicable to the entire Quaternary time range and to all stratigraphic contexts. The figure is, nevertheless, a useful aide-memoir, and the reader will find it helpful to refer back to it when working through some of the case studies later in the book.

One part of the Quaternary record where there is a broad measure of agreement and where, moreover, there is also closer dating control is the climatic oscillation that occurred at the end of the Last Cold Stage (ca. 15 000 and 11 500 years ago) and which is most clearly reflected in proxy climate records from around the North Atlantic region. This episode is referred to as the **Weichselian Lateglacial** in northern Europe and the **Devensian Lateglacial** in Britain (Figure 1.5, right). It is characterised by rapid warming around 14 800 years ago (the **Bølling-Allerød Interstadial** in Europe; **Lateglacial** or **Windermere Interstadial** in Britain), a significant cooling (**Younger Dryas** or **Loch Lomond Stadial**) around 12 900 years ago, and finally an abrupt climatic amelioration at the onset of the present (**Holocene**) interglacial at ca. 11 500 years ago. In the Greenland (GRIP) ice core (Chapter 5), this climatic oscillation is reflected in a series of clearly defined 'events' in the oxygen isotope record (GS-2; GI-1; GS-1: Figure 1.5, left). Greenland Interstadial 1 is further divided into a series of sub-events, with GI-1a, GI-1c and GI-1e representing warmer intervals, and GI-1b and GI-1d reflecting cooler episodes (Björck *et al.*, 1998; Walker *et al.*, 1999). Whereas the timescale for terrestrial sequences from Britain and northern Europe is based on calibrated radiocarbon years (section 2.6), the Greenland (GRIP) record is in ice-core years (section 5.5). Again, the reader may find it useful to cross-reference some of the later case studies with Figure 1.5.

1.6 The Scope and Content of the Book

In the following chapters, the various dating techniques that are available to Quaternary scientists (Figure 1.6) are introduced, explained and evaluated. The last element is

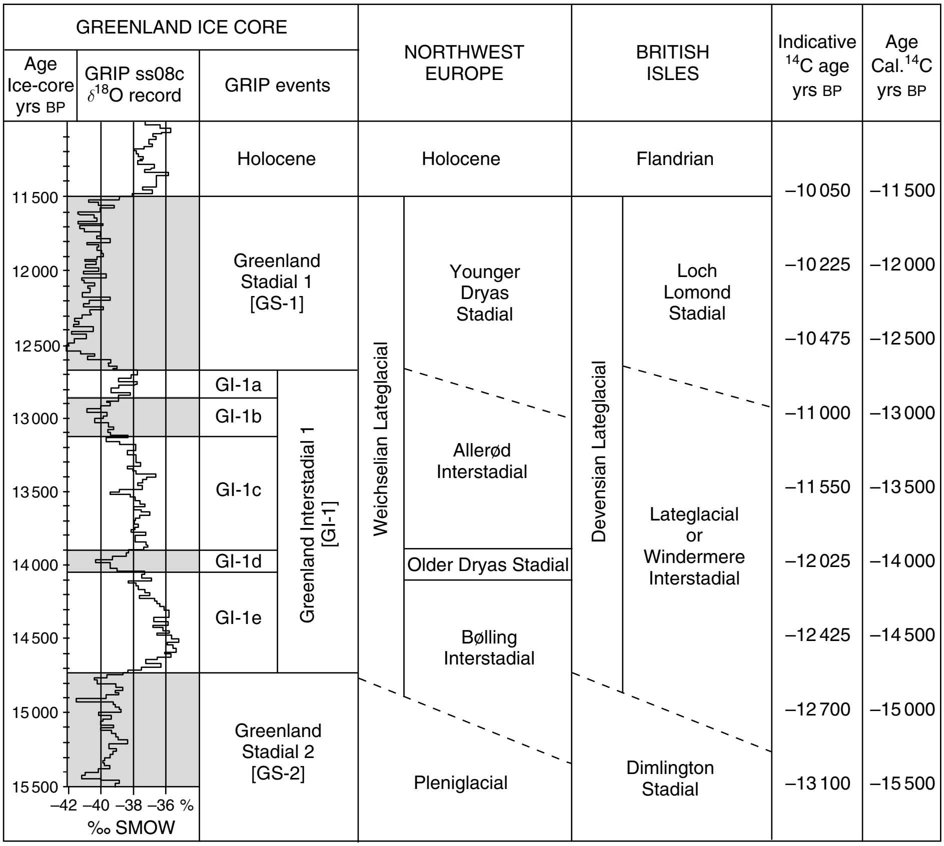

Figure 1.5 *The $\delta^{18}O$ record from the GRIP Greenland ice core showing the Lateglacial event stratigraphy (left), and the stratigraphic subdivision of the Lateglacial in northwest Europe and the British Isles. The isotopic record is based on the GRIP ss08c chronology, and the colder stadial episodes are indicated by dark shading. The radiocarbon ages should be regarded as indicative ages only (partly after Lowe* et al.*, 2001)*

especially important because it is important to understand not only how each method works, but also where and why errors are likely to occur. Some of these may arise from the nature of the sample; others from analytical limitations. Whatever the cause, these will impact on the resultant age determinations. Each section concludes with a number of examples or case studies. These have been carefully selected to show how the different techniques can be employed in Quaternary science and to give an indication of the range of applications of each method.

Chapters 2–5 deal with techniques that enable ages to be determined in years before the present (years BP). In other words, they allow **estimates of age** to be obtained. Chapters 2–4 describe **radiometric dating techniques**, where age is determined from measurements either of radioactive decay of some unstable chemical elements (Chapters 2 and 3), or of the effects of radioactive decay on the crystal structure of certain minerals or fossils (Chapter 4). Chapter 5 reviews a group of methods based on the regular accumulation of

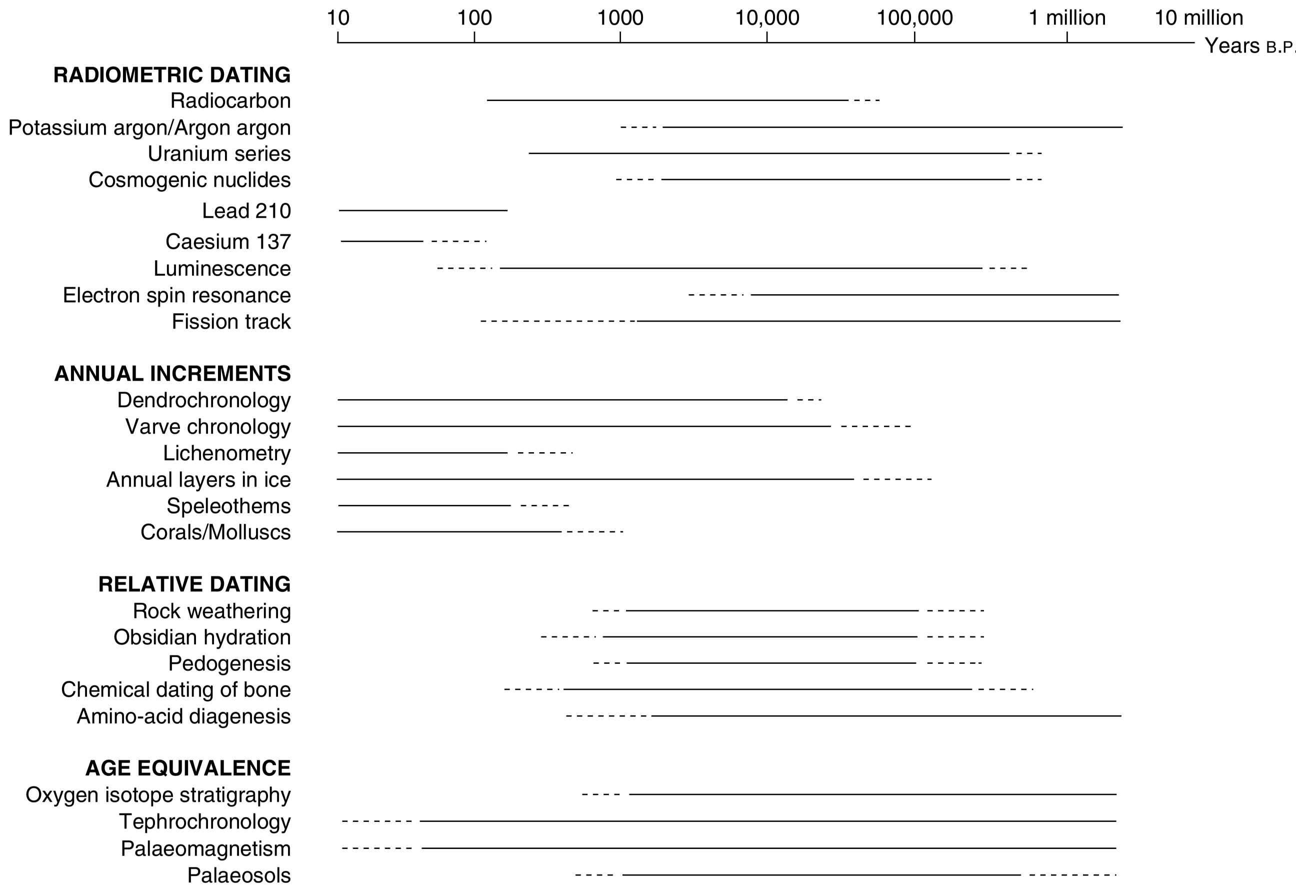

Figure 1.6 *The effective dating ranges of the different techniques discussed in this book*

sediment or biological material through time, and which form **annually banded records**. All of the techniques that enable estimates of age to be made have sometimes been referred to as *absolute dating methods*. This term has not been used here; indeed it has been deliberately avoided because it implies a level of accuracy and precision that can seldom, if ever, be achieved in reality. As we saw above, and as will be amplified in the following discussions, where age estimates are being obtained, errors are unavoidable and hence there will inevitably be an element of uncertainty associated with each age determination. There is, therefore, nothing 'absolute' about a date, and it should not be referred to as such.

In Chapters 6 and 7, two further groups of dating techniques are considered. The first involves the grouping of fossils or sedimentary units which are then ranked in relative order of antiquity; hence, these are known as **relative dating methods**. Some are based on the principles of stratigraphy where relative age can be determined by the position of stratigraphic units in a geological sequence; others use the degree of degradation or chemical alteration (both of which may be time-dependent) on rock surfaces, in soils or in fossils, to establish relative order of age. Chapter 7 considers methods that enable **age equivalence** to be determined, based on the presence of contemporaneous horizons in separate and often quite different stratigraphic sequences. With respect to both relative dating and age-equivalence techniques, where the various stratigraphic units or fossil materials can be dated by one of the age-estimate methods described in Chapters 2–5, it may prove possible to fix the relative or age-equivalent chronologies in time. In other words, they can be calibrated to an independently derived timescale.

Notes

1. The duration of the Quaternary is still a matter for debate, with some authorities arguing for a 'shorter' timescale of around 1.6–1.8 million years, while others subscribe to the view that a longer timescale of 2.5–2.6 million years is more appropriate. There is, perhaps, a majority in favour of the longer chronology and this interpretation has been followed here.
2. The **Pleistocene** epoch ended around 11 500 years ago and was succeeded by the **Holocene**, the warm period in which we live. As the present temperate period is simply the most recent of a number of temperate episodes that form part of a long-term climatic cycle, the last 11 500 years can be seen as part of the Pleistocene (West, 1977). Hence, the terms 'Quaternary' and 'Pleistocene' are often used interchangeably.
3. **Uniformitarian reasoning**, as initially developed by the Scottish geologist James Hutton in the later eighteenth century, emphasises the continuity of geological processes through time. Hence contemporary processes (**modern analogues**) can be used as a basis for interpreting past events. Uniformitarianism is often described by the dictum 'the present is the key to the past'.
4. The **Astronomical Theory of Climate Change** is based on the assumption that surface temperatures of the earth vary in response to regular and predictable changes in the earth's orbit and axis. The three principal components are the **precession of the equinoxes** (apparent movement of the seasons around the sun) with a periodicity of ca. 21 000 years, the **obliquity of the ecliptic** (variations in the tilt of the earth's axis) with a periodicity of ca. 41 000 years, and the **eccentricity of the orbit** (changes in the shape of the earth's orbit) with a periodicity of ca. 96 000 years. Collectively these govern the amount of heat received by the earth and the distribution of this heat around the globe. First developed in its modern form by the Scottish scientist James Croll in

the nineteenth century, the theory was subsequently elaborated by the Serbian geophysicist Milutin Milankovitch in the 1930s. The radiation balance curves that he produced can be calibrated to the orbital parameters and used to provide an **astronomical timescale** for glacial–interglacial cycles (Chapter 5). Further explanation of the astronomical theory can be found in standard Quaternary texts, such as those by Lowe and Walker (1997), Roberts (1998), Williams *et al.* (1998), Wilson *et al.* (2000) and Bell and Walker (2005).

5. A major discovery in physics during the early twentieth century was that sub-atomic particles sometimes behave as if they are waves, a concept that lies at the heart of the science of quantum physics. One consequence is that it is impossible to measure both the *position* of a sub-atomic particle and its *velocity*, an idea that was first proposed by the German physicist Werner Heisenberg in his **Uncertainty Principle**. This indeterminacy in the sub-atomic world can be seen very clearly whenever a single atomic event can be observed, such as in radioactivity. Although quantum physics is a highly complex field, there are a number of accessible texts that deal with this subject or that include sections on it. Ones that I have found particularly informative (and enjoyable!) are by Gribbin (1984), Close *et al.* (1987), Barrow (1988), Gribbin (1995), Penrose (1999), Rees (2000) and, of course, Bryson (2003).
6. During evaporation from the free ocean surface, a **fractionation** (or separation) occurs so that more of the lighter oxygen isotope, ^{16}O, is drawn into the atmosphere than the heavier isotope, ^{18}O. In the cold stages of Quaternary, therefore, large amounts of the lighter isotope would have been transported poleward by moisture-bearing winds and locked into the greatly expanded ice sheets. As a consequence, ocean water would have been relatively 'enriched' in the heavier isotope ^{18}O. The reverse would obtain during interglacial stages for, with reduced land ice cover, more ^{16}O would have been returned to the oceans where water would have become relatively 'depleted' in ^{18}O. Accordingly, the $\delta^{18}O$ trace provides a record of changing volumes of land ice, and hence of glacial/interglacial climatic fluctuations.
7. A '**proxy climatic record**' is one based on an indirect measure of climate. In other words it is based on *inferential* evidence (pollen, plant macrofossils, etc.), as opposed to *direct evidence* obtained using a thermometer or rain gauge.