7 Stratigraphy: The Modern Synthesis

Contents

7.1 Introduction

Stratigraphy is defined as the study of layered rocks. In the context of sedimentary geology in general, and of this book in particular, stratigraphy is the discipline that pulls everything together. In Chaps. 2–5 of this book we deal with increasingly large and complex sedimentological concepts, and in Chap. 6 we discuss mapping methods, which are essentially the methods for extending our interpretations beyond our immediate data points by interpolation and extrapolation. In this chapter we add the elements of chronostratigraphic dating and correlation and demonstrate the interdisciplinary nature of modern stratigraphic methods.

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Interpretations that focus on local to regional facies studies or sequence stratigraphy rely on the principles of stratigraphy to create and confirm local and regional correlations. Rock units and the depositional environments in which they formed may be related to each other in a way that enables additional correlations to be made to, for example, regional and global climatic and tectonic events. Stratigraphic methods are required to construct such correlations, and a dependence on these methods increases with the scale of a project, from the local to the regional to the continental to the global.

Formal stratigraphic practices, including the definition of formations and stages, had their origins in nineteenth-century field geology, beginning with William Smith (Conkin and Conkin 1984; Miall 2004), and have evolved into a set of carefully defined procedures for naming and correlating the various kinds of stratigraphic unit (Salvador 1994; NACSN 2005). They have also, for many years, been applied successfully to subsurface well data. These methods are mainly based on detailed lithostratigraphic and biostratigraphic information, the analysis of which is discussed later in this chapter, along with other important aids to correlation, including radioisotopic dating and magnetic reversal stratigraphy. Lithostratigraphic classification of the sedimentary record remains the basic descriptive process for stratigraphic documentation, simply because of the large body of historical documentation based on lithostratigraphy; but the ultimate goal is now to develop stratigraphic frameworks based on sequence stratigraphy (Catuneanu 2006; Catuneanu et al. 2011; see Sect. 7.7).

A broader, more regional approach to correlation generally is taken by those dealing with reflection seismic data (Veeken 2007). Commercial seismic work in frontier regions and the deep reflection profiles produced by groups such as the Consortium for Continental Reflection Profiling (COCORP) in the United States, Canada's Lithoprobe Project, and many other international projects, provide sweeping regional cross-sections, within which correlations at the detailed level may be far from clear. This work may be supplemented by detailed three-dimensional reconstructions created using 3-D seismic methods applied to prospective rock volumes (Brown 2011).

With outcrop and well data, the problem may be to establish the regional framework from a mass of local detail, whereas with seismic data it is the detail that may be hard to see (depending on depth, on which seismic resolution depends). The ideal combination is, of course, a basin with a network of key exploration holes tied to regional seismic lines, plus local 3-D seismic volumes. Most advanced petroleum exploration projects now achieve this level of detail. Examples are provided in Chap. 6.

These differences in data type and scale have led to two different approaches to stratigraphic correlation. In industry exploration work in frontier regions, particularly the great offshore basins, a rather informal, pragmatic approach may be taken to such topics as biostratigraphic zonation and the naming of formations, at least in the early stages of basin development. The broad picture can be derived from seismic cross-sections, and the details gradually resolve themselves as more well data become available. Application of various basin mapping methods (Chap. 6) and use of the genetic, depositional systems approaches and sequence stratigraphy concepts (Chap. 5) are of particular value here. Some examples of detailed mapping and correlation and the kinds of research questions these projects raise are discussed in Chap. 8.

In the absence of seismic data, it is necessary to construct the forest from the individual trees (the second approach). The stratigraphic framework in most well-explored (mature) basins was built up this way using lithostratigraphic methods, and in the past the work has usually been accompanied by considerable controversy, as local specialists have argued about the relationships between the successions in different parts of a basin or between outcrop and subsurface units. Sequence concepts, because of their predictive power, can now make this task easier, although the work is usually not without its problems.

Whether a basin analysis exercise starts from seismic sections or outcrop work, it is desirable, eventually, to document the fine detail of the stratigraphy by establishing a sequence-stratigraphic framework and, ideally, to tie this to the global time scale (formal sequence-stratigraphic methods have yet to be finalized; see Sect. 7.7). Every local biostratigraphic, radioisotopic and magnetostratigraphic study can potentially contribute to the long-term effort to perfect a global chronostratigraphic (time) scale, which goes by the formal name of the **Geologic Time Scale (GTS)**. This last step is beyond the needs of most exploration companies and is an area of research commonly taken over by state geological surveys and individuals in academic institutions, although the data base and expertise built up in industry may form an essential component. This is one area in which the government core and data repositories can prove their usefulness.

Several commissions and subcommissions and numerous working groups of the *International Union of Geological Sciences* (IUGS), mostly under the auspices of the *International Commission on Stratigraphy* (ICS), have been carrying this work for many years, like many national groups. The *North American Commission on Stratigraphic Nomenclature* has been particularly influential. Some of the results are reported in this chapter, and the interested reader may wish to examine the IUGS journal *Episodes*, which reports on the activities of these various groups and announces important publications. Other important publications include *Newsletters on Stratigraphy* (published by E. Schweizerbart, founded in 1965), and the journal *Stratigraphy* (published by

Micropaleontology Press, founded by William Berggren and John A. Van Couvering in 2004). They both publish original research articles and reviews and may also include background information on the ongoing work of ICS and the *International Subcommission on Stratigraphic Classification* (ISSC). The North American Commission published an updated, comprehensive guide to stratigraphic procedure in 2005. Recent synthesis publications describing the geological time scale have brought this topic to a high level of sophistication (Gradstein et al. 2004a, 2012, 2020). We discuss this later in this chapter (Sect. 7.8). The website https://stratigraphy.org is the official website of the ICS.

This chapter is intended to provide an introduction to practical working methods. Chronostratigraphic (including biostratigraphic) research must form an integral part of any ongoing basin analysis project unless it is strictly local in scope. The work is usually performed by specialists. Correlation methods based on lithostratigraphy and sequence stratigraphy are also compared and contrasted here, although the procedures for erecting formal, named units (included here for consistency and completeness) can be left to advanced stages of the analysis (and do not yet include procedures relating to sequence stratigraphy). Such naming is best carried out by individuals with sedimentological training so that the depositional systems and sequence concepts described in Chaps. 3–5 can be incorporated into the work. Research trends, emerging problems and some new developments are discussed in Chap. 8.

As discussed in detail in Chap. 8, the sedimentary record is "more gap than record." For 100 years, since Barrell (1917) published his ground-breaking work on the rates of sedimentary processes, the fragmentary nature of the record has largely been ignored by practicing stratigraphers and sedimentologists. Typically, in many stratigraphic sections, only about 10% of elapsed time is represented by the preserved rock at a 10^6-year time scale. All the developments in dating and correlation methods and the emergence of the powerful new descriptive-interpretive methods of sequence stratigraphy have all been accomplished despite this fact. Stratigraphy continues to "work" as a practical method of basin mapping and resource exploration. How can this be? It is because there is a limited suite of natural processes that deposit and preserve sediment, and these predominate in the development of the stratigraphic record, including both the accumulated sedimentary successions and the hiatuses that separate them (Miall 2015). These are widespread and are now well known. For example, modern work on sequence stratigraphy has identified a limited range of allogenic processes, all characterized by a specific range of time scales, that are now known to generate sequences (Miall 1995; see Sect. 7.6). These typically extend through a sedimentary basin and may correlate to other successions regionally, or even globally. The orders-of-magnitude range of time scales over which these processes operate was the basis for the original hierarchical "order" classification of sequences (Vail et al. 1977). The range of natural processes that build the sedimentary record, from the burst-and-sweep turbulence of traction current to the accumulation of a basin-fill over millions of years, constitutes a crudely fractal distribution of rates and time scales, but these processes are genetically unrelated (Miall 2015). This is important because it means that the distribution of relevant time scales is not mathematically precise, and that, therefore, quantification of processes, for example on the basis of fractal theory, may not be particularly useful, illuminating or, indeed, relevant. Also, as noted elsewhere, interpretations of sedimentary processes that imply continuity, such as sediment-transport mass-balance calculations and the reconstruction of shoreline trajectories through transgressions and regressions, need to take into account the interruptions in the record represented by hiatuses, the frequency and duration of which have consistently been ignored.

A formal approach to the issue of unconformities is discussed in Sect. 7.6. We address rate and time scale issues in more detail in Chap. 8.

7.2 Types of Stratigraphic Unit

Rocks may be described in terms of any of their physical, chemical, organic or other properties, including lithology, fossil content, geochemistry, petrology, mineralogy, electrical resistivity, seismic velocity, density (gravity), magnetic polarity or age. Theoretically, any of these properties may be used for description and correlation, and most are so used for various purposes. In practice, lithology is the most important criterion; fossil content is also crucial for rocks of Phanerozoic age. Magnetic polarity has become useful as a correlation tool, particularly for the younger Mesozoic and the Cenozoic, and radioisotopic ages are used to assign numerical ("absolute") ages to biostratigraphic, magnetic and other chronostratigraphic units. A standard oxygen isotope scale is available for the late Cenozoic (Pliocene to present), and increasing use is being made of other isotopic signatures, particularly carbon and strontium. Later sections of this chapter deal with all these techniques in more detail. Other geophysical properties are used in the early reconnaissance stages of exploration of a sedimentary basin. Not all these properties will necessarily give rise to the same correlations of a given rock body; for example, it is commonly difficult to relate geophysical properties precisely to lithology. Therefore, no single type of stratigraphic unit can be used to define all the variability present in nature.

Reflection-seismic data have been widely used in the exploration of sedimentary basins since the 1970s, making use of the concepts of seismic stratigraphy (Sect. 6.3), and with particular application to the interpretive field of sequence stratigraphy (Chap. 5). Sheriff (1976) pointed out that for reflecting events to be distinguished on seismic data they must each represent a clear velocity contrast and must be at least the equivalent of a quarter wavelength apart. At shallow depths, velocities are in the range of 1.5–2.5 km/s and reflections are of relatively high frequency, about 5–100 Hz, so that a quarter wavelength is on the order of 5–12 m. At greater depths, typical reflection wavelengths increase considerably. Therefore, stratigraphic resolution is fairly coarse (Fig. 7.1). Early work in seismic sequence stratigraphy had portrayed sequence boundaries as distinctive reflection surfaces correlatable over wide distances (Mitchum et al. 1977), but improved acquisition and processing techniques have demonstrated that the broad, through-going correlations on which this early work was based may in many cases be suspect. Cartwright et al. (1993) demonstrated that many of the critical features of sequence architecture, including the sequence boundaries, break down upon detailed examination into complex reflection patterns representing local facies variability, and major through-going surfaces may not be present or easy to trace. Seismic data are essential for studying large-scale stratigraphic features such as depositional systems and regional (or global) sequences, but may be of less use in the development of the refined stratigraphic subdivisions that are the subject of this chapter. The reader interested in the application of seismic methods to basin analysis may wish to consult standard textbooks in this area, such as Catuneanu (2006) or Veeken (2007).

Two basic categories of stratigraphic information are essential for the complete documentation of the stratigraphic record: (1) descriptive lithic units and (2) geochronologic information, dealing with correlation and age of the strata (Harland 1993).

The most important types of stratigraphic units are:

Lithostratigraphic units: These are strictly empirical, based on observable lithologic features including composition and grain size, and possibly also including certain basic sedimentological information, such as types of sedimentary structures and cyclic successions.

Biostratigraphic units: These are based on fossil content. Life forms evolve with time, permitting subdivision into **biozones** on the basis of changes in the fauna or flora. The first and last appearance of particular species or variants may also serve as useful time markers. When used on their own, biostratigraphic units provide relative ages. Numerical ("absolute") ages are derived by cross-correlation with chronostratigraphic data, as noted below.

Unconformity-bounded units: These are units bounded above and below by unconformities. They may consist of any kinds of rocks, igneous, metamorphic or sedimentary. Unconformity-bounded sedimentary successions may be formalized using the empirical, descriptive classification procedures of **allostratigraphy** (see NACSN 2005), but increasingly geologists now employ the interpretive procedures and models of **sequence stratigraphy** as the main basis for subdivision and mapping of the basin fill.

Chronostratigraphic units: These comprise an interpretive stratigraphy, in contrast to lithostratigraphic and biostratigraphic units, which are strictly descriptive. Chronostratigraphy concerns itself with correlation and the age of the strata in years ("absolute" age), which may be determined by a variety of means, of which the most important are fossil content, radioisotopic dating, magnetic polarity (for the post Middle Jurassic), and the isotopic record of oxygen, carbon and strontium (for different parts of the Phanerozoic record). The principal chronostratigraphic units, which form the main

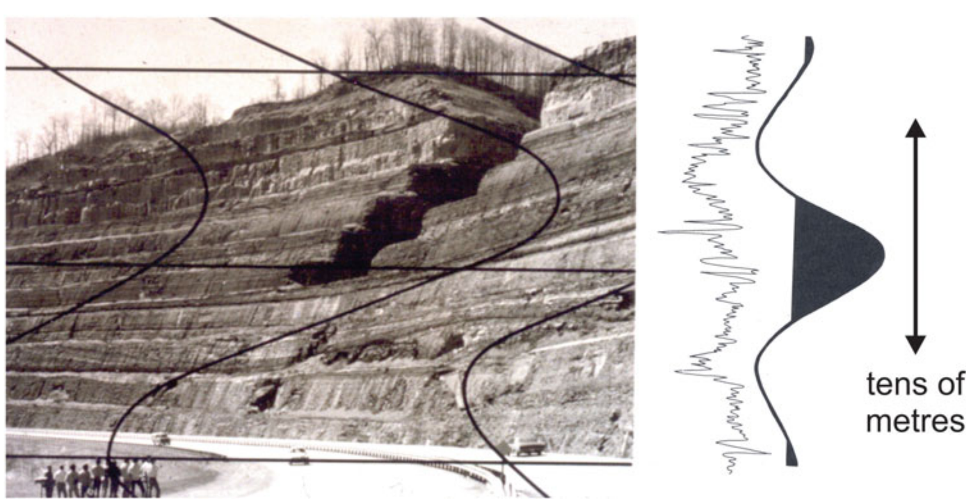


Fig. 7.1 Scale of a typical seismic wave form as compared to an outcrop (left), and compared to a wireline log (right). The frequencies characteristically used in petroleum exploration seismic work (10–60 Hz) have long wavelengths. Seismic resolution is therefore limited to large-scale stratigraphic features (based on an idea from A. E. Pallister and A. E. Wren)

foundation of the **Global Time Scale** (or **Geologic Time Scale**) are **stages**. Increasing use is being made of the cyclostratigraphy of selected stratigraphic sections to build a highly accurate **astrochronologic time scale**.

Both lithostratigraphic and biostratigraphic units may be local in extent. Lithologic character depends on the depositional environment, sediment supply, climate, rate of subsidence etc., all of which can vary over short distances. Lithostratigraphic units are diachronous to a greater or lesser degree, that is, they represent a different time range in different places, reflecting gradual shifts in the environment, for example, during transgression or regression. The limits of a stratigraphic unit are either its erosional truncation at the surface or beneath an unconformity or a facies change into a contemporaneous unit of different lithology. A special type of lithostratigraphic unit is that formed by short-lived **stratigraphic event**s, which are those that have widespread depositional effects within very short time spans (Ager 1981). Examples of such events are volcanic ash falls, the deposits of violent storms and tsunamis, and certain regional or global sea-level changes. Such event deposits may prove very useful as local correlation tools, as discussed in Sect. 7.8.4.

Biostratigraphic units are based on fauna or flora, the distribution of which is ecologically controlled. Also, contemporaneous faunas located in ecological niches that are similar but geographically isolated may show subtly different evolutionary patterns, making comparisons or correlations between the areas difficult (this is the subject of **biogeography**). All life forms evolve with time so that faunas and floras show both spatial and temporal limits on their distribution.

The unconformities that demarcate unconformity-bounded units are caused by subaerial erosion during times of low stands of sea level, by erosion following tectonic uplift, submarine erosion or sudden environmental change (Sect. 7.6). These events are typically widespread. Sea-level change may be caused by tectonic elevation of the basement or by eustatic changes in sea level. In either case, the unconformity surfaces define units of considerable lateral extent. Rapid changes in the depositional environment may generate what Schlager (1989) termed **drowning unconformities**. Some unconformities may be of global significance, although this is typically difficult to demonstrate (Miall 2010, Chap. 14) Unconformities provide an excellent basis for the regional subdivision of basin fill, and their interpretation may throw considerable light on regional tectonic evolution. **Sequence stratigraphy** has become the method of choice for subsurface mapping by the petroleum industry precisely because of its practical utility in focusing on and documenting these unconformity-bounded successions (Chap. 5). Unconformities and unconformity-bounded units are discussed in Sect. 7.6.

Chronostratigraphy attempts to resolve the difficulties in regional and global correlation by establishing a global, time-based reference frame. A standard **Geological Time Scale (GTS)** has gradually evolved since the discovery of radioisotopic dating early in the twentieth century. However, the accuracy of chronostratigraphic correlation is only as good as that of the time-diagnostic criteria on which it is based. Imprecision and error remain (Sects. 7.8 and 8.10).

The evolution of these four types of units has had a long and complex history (Hancock 1977; Conkin and Conkin 1984; Miall 2004) and there has been controversy about definitions. Hedberg (1976), Hancock (1977) and Harland (1978, 1993) discussed some of the early practical and philosophical problems. The nineteenth-century geological practice did not distinguish lithology from age, causing severe correlation problems, wherever a facies change or a diachronous boundary occurred. More recently, there has been controversy over whether the rocks (lithologic units) or interpreted age range should form the primary basis of chronostratigraphy (e.g., Zalasiewicz et al. 2004). The discussions are likely to seem somewhat academic and theoretical to the average basin analyst and will not be discussed at length here. A historical summary of the methods that gradually evolved for the construction of the geological time scale is provided by Gradstein and Ogg (in Gradstein et al. 2020, Chap. 2).

7.3 The Six Steps Involved in Dating and Correlation

Six main "steps" are involved in the dating and correlation of stratigraphic events (Miall 1994). Figure 7.2 summarizes these steps and provides generalized estimates of the magnitude of the uncertainty associated with each aspect of the correlation and dating of the stratigraphic record. Some of these errors may be cumulative, as discussed in the subsequent sections. The assignment of ages and correlations with global frameworks is an iterative process that, in some areas, has been underway for many years. There is much feedback and cross-checking from one step to another. What follows should be viewed, therefore, as an attempt to break down the practical business of dating and correlation into more readily understandable pieces, all of which may be employed at one time or another in the unravelling of regional and global stratigraphies. The main steps are as follows:

1. Identification of the units or stratigraphic events to be correlated, and development of regional correlation frameworks, including the mapping of hiatuses,

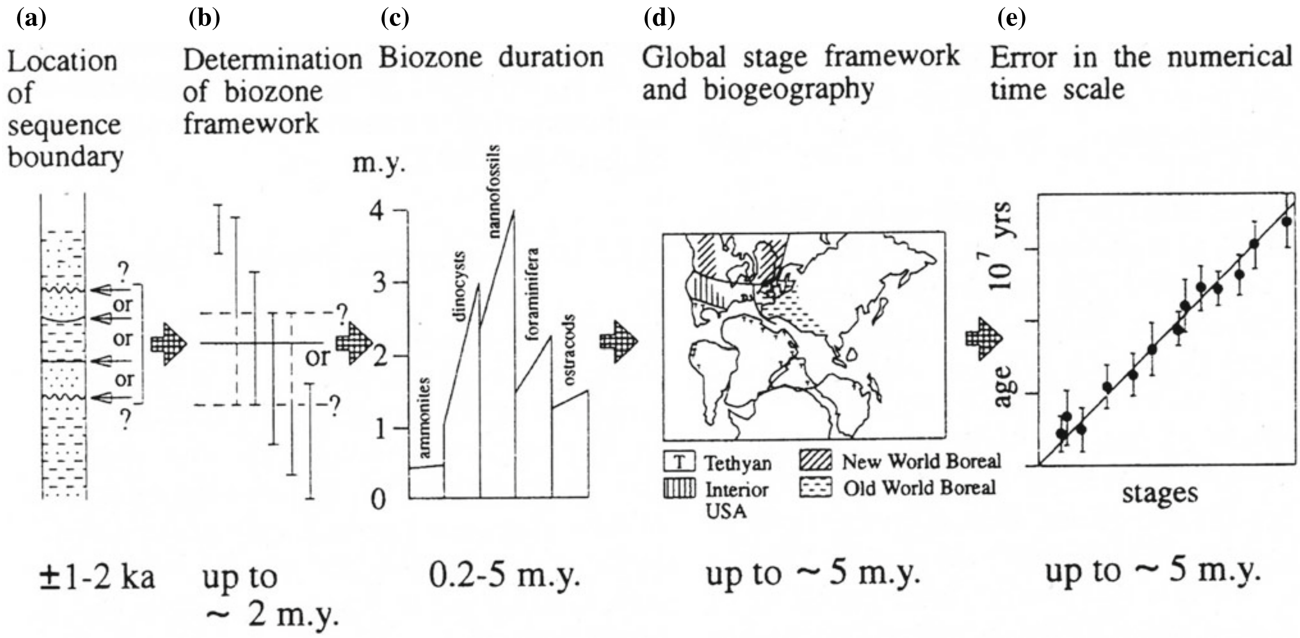


Fig. 7.2 Steps in the correlation and dating of stratigraphic events. e = typical range of error associated with each step. **a** In the case of the sequence framework, the location of sequence boundaries (step 1) may not be a simple matter but depends on the interpretation of the rock record using sequence principles. **b** Assignment of the boundary event to the biozone framework (step 2). An incomplete record of preserved taxa (almost always the case) may lead to ambiguity in the placement of biozone boundaries. **c** The precision of biozone correlation depends on biozone duration (step 3). Shown here is a simplification of Cox's (1990) summary of the duration of zones in Jurassic sediments of the North Sea Basin. **d** The building of a global stage framework (step 4) is fundamental to the development of a global time scale (step 5). However, global correlation is hampered by faunal provincialism. Shown here is a simplification of the faunal provinces of Cretaceous ammonites, shown on a mid-Cretaceous plate-tectonic reconstruction. Based on Kennedy and Cobban (1977) and Kauffman (1984). **e.** The assignment of numerical ages to stage boundaries and other stratigraphic events (step 6) contains an inherent experimental error and also the error involved in the original correlation of the datable horizon(s) to the stratigraphic event in question. Diagrams of this type are a standard feature of any discussion of the global time scale (e.g., Haq et al. 1988; Harland et al. 1990). The establishment of a global biostratigraphically-based sequence framework involves the accumulation of uncertainty from step (**a**) through (**d**). Potential error may be reduced by the application of radioisotopic, magnetostratigraphic or chemostratigraphic techniques which, nonetheless, contain their own inherent uncertainties (step 6)

unconformities and other key surfaces. Local correlations may be based on lithostratigraphy, but sequence stratigraphic concepts and methods are now practically universal. Correlations may be guided or constrained by supplementary data, such as biostratigraphic zonation. Determining the position of events such as sequence boundaries may or may not be a straightforward procedure, and requiring the application of facies mapping and sequence mapping techniques (Chaps. 3, 4, 5).

2. Determining the extent and chronostratigraphic significance of unconformities. Unconformities, including sequence boundaries, represent finite time spans which vary in duration from place to place. In any given location this time span could encompass the time span represented by several different sedimentary breaks at other locations. Resolving such problems may require that some of the other steps be completed, particularly step 3.
3. Determination of the biostratigraphic framework. One or more fossil groups is used to assign the selected event to a biozone framework, and zones are defined and correlated from section to section. Error and uncertainty may be introduced because of the incompleteness of the fossil record. Graphic correlation or other quantitative techniques may be employed (Sect. 7.5).
4. Assessment of relative biostratigraphic precision. The length of time represented by biozones depends on such factors as faunal diversity and rates of evolution. Durations of biozones vary considerably through geological time and between different fossil groups. Steps 3 and 4 may be aided by the availability of numerical ages obtained from the radioisotopic dating of igneous material, such as lava flows and ash beds. Increasing use is now being made of chemostratigraphic data to cross-reference with and calibrate biostratigraphic data.
5. Correlation of biozones with the global stage framework (Sect. 7.8.3). Much of the existing stage framework was initially, with notable exceptions, built from the study of macrofossils in European-type sections, although

microfossils have become increasingly important for subsurface work and global studies (McGowran 2005). Correlation with this framework raises questions of environmental limitations on biozone extent, our ability to inter-relate zonal schemes built from different fossil groups, and problems of global faunal and floral provinciality and diachroneity.

6. Assignment of numerical ("absolute") ages (Sect. 7.8.2). The use of radioisotopic and magnetostratigraphic dating methods, plus the increasing use of chemostratigraphy (oxygen, carbon and strontium isotope concentrations) permits the assignment of numerical ages in years to the biostratigraphic framework (in addition to the possible direct dating of stratigraphic units, as noted in point 4, above). Such techniques also constitute methods of correlation in their own right, especially where fossils are sparse. The geological time scale (GTS) has become an instrument of considerable geological importance and practical utility in recent years, contributing to the emergence of what Miall (2013) termed "Sophisticated Stratigraphy" (Sect. 7.8; see Fig. 7.33).

7.4 Lithostratigraphy

Until the 1980s it was standard practice to describe and map stratigraphic successions on the basis of lithostratigraphic principles (Fig. 7.3). In the field, particularly in arid regions where the rocks are well exposed, it is still the historically established formations that are the basis for field location and identification. Such is the case, for example, in the Grand Canyon and Canyonlands areas of the United States, and the Front Ranges of the Rocky Mountains in Alberta. Some lithostratigraphic names have long been part of geological language and are unlikely to be deposed for a long time (e.g., in the UK: Carboniferous Limestone, Old Red Sandstone; in the United States Austin Chalk, Mancos Shale; in Canada: Leduc Formation, Rundle Group). It is necessary, therefore, to be able to read older publications and maps and understand what type of information they convey.

Among the problems with lithostratigraphy as a method of description is that the defined units carry no meaning regarding the origins or age of the units. Formations are commonly diachronous, and many stratigraphic names were established many decades ago, long before the advent of modern facies and sequence analysis. Older literature may therefore be replete with the names of local, poorly defined units, with a given body of rocks defined and named differently in different parts of a basin. Procedures are available (e.g., see NACSN 2005) for the revision and redefinition of units as new information becomes available from surface mapping or subsurface exploration.

7.4.1 Types of Lithostratigraphic Units and Their Definition

A hierarchy of units has been developed based on the **formation**, which is the primary lithostratigraphic unit (NACSN 2005).

Group
Formation
Member
Tongue or lentil
Bed

The formation. An important convention has long since been established that all sedimentary rocks should be subdivided (when sufficient data have been collected) into formations. No other types of lithostratigraphic subdivisions

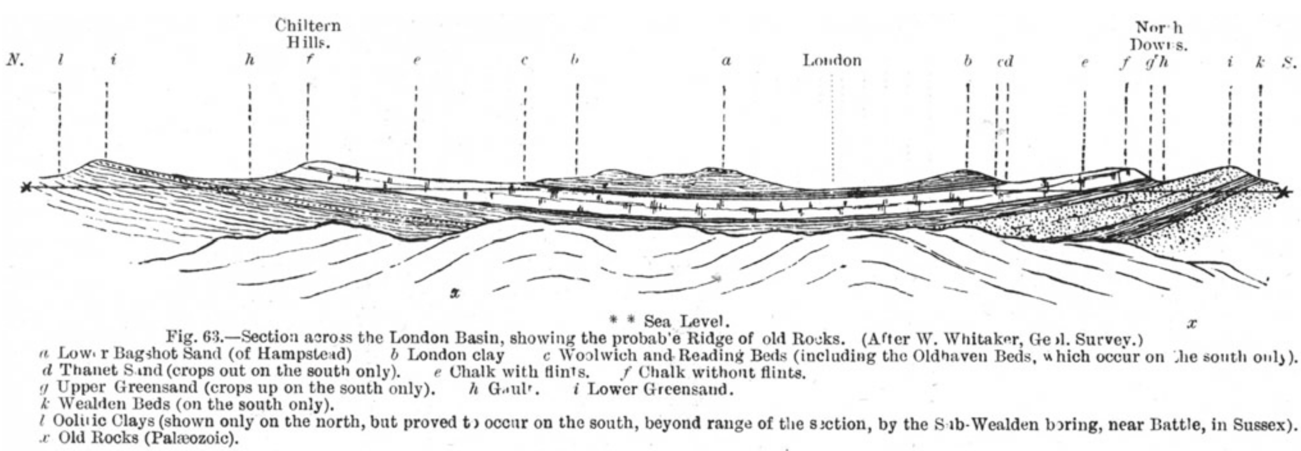


Fig. 7.3 A cross-section of the London Basin, England, showing the development of descriptive terminology for stratigraphic units. From "Geology of the counties of England and Wales", by Jerome Harrison, 1882

need to be used, although convenience of description may require them.

What is a formation? There are no fixed definitions that deal with the scale or variability of what should constitute a formation, although the procedures for establishing limits (contacts) and names are well established (e.g., see NACSN 2005). Figure 7.4 provides a good example of the way in which stratigraphic successions are subdivided on the basis of lithology. The lithologies, colors and weathering characteristics of the rocks suggest a fourfold subdivision of the exposure. Comparison with other exposures nearby and the presence of distinctive fossils permit three of the subdivisions to be assigned to previously existing formations, while the fourth (oldest) unit is different from the local succession, and has yet to be given a name. This outcrop is large enough that the angular unconformity between two of the units (the Nansen and the Barrow formations) can clearly be seen.

The degree of lithologic variability required to distinguish a separate formation tends to reflect the level of information available to the stratigrapher. Formations may be only a few meters or several thousands of meters in thickness; they may be traceable for only a few kilometers or for thousands of kilometers. Formations in frontier basins usually are completely different in physical magnitude from those in populated, well-explored basins, such as much of western Europe and the United States. As an exploration in frontier basins proceeds, some of the larger formations first defined on a reconnaissance basis may subsequently be subdivided into smaller units and the ranking of the names changed. NACSN (2005) provides the procedures for making these kinds of revisions.

The most important criteria for establishing a formation are its usefulness in subdividing stratigraphic cross-sections and its "mappability." For reconnaissance mapping, a thin unit that cannot accurately be depicted at a scale of, for example, 1:250,000, may be of little use, although the definition and mapping of thin but widespread marker units may be of considerable utility. For more detailed work, mappability at a scale of 1:50,000 or even 1:10,000 may be a more useful criterion. Problems of consistency may arise when detailed work is conducted around a mine site within what is otherwise a poorly explored frontier basin.

Formations should not contain major unconformities, although minor disconformities may be acceptable (indeed, as we now recognize, they are all but unavoidable). The contacts of the formation should be established at obvious lithologic changes. These may be sharp or gradational. An unconformity is a logical choice for a formation contact. Where lithologies change gradually, either vertically or laterally, it may be difficult to choose a logical place to draw the contact. For example, a mudstone may pass up into a sandstone through a transitional succession with sandstone beds becoming thicker and more abundant upward. The mudstone–sandstone formation contact could be drawn at the oldest thick, coarse sandstone (with thickness and coarseness carefully spelled out), at the level where sandstone and mudstone each constitute 50% of the section, or at the youngest extensive mudstone bed. The choice is arbitrary, and it is immaterial which method is selected as long as the same method is used as consistently as possible throughout the extent of the formation.

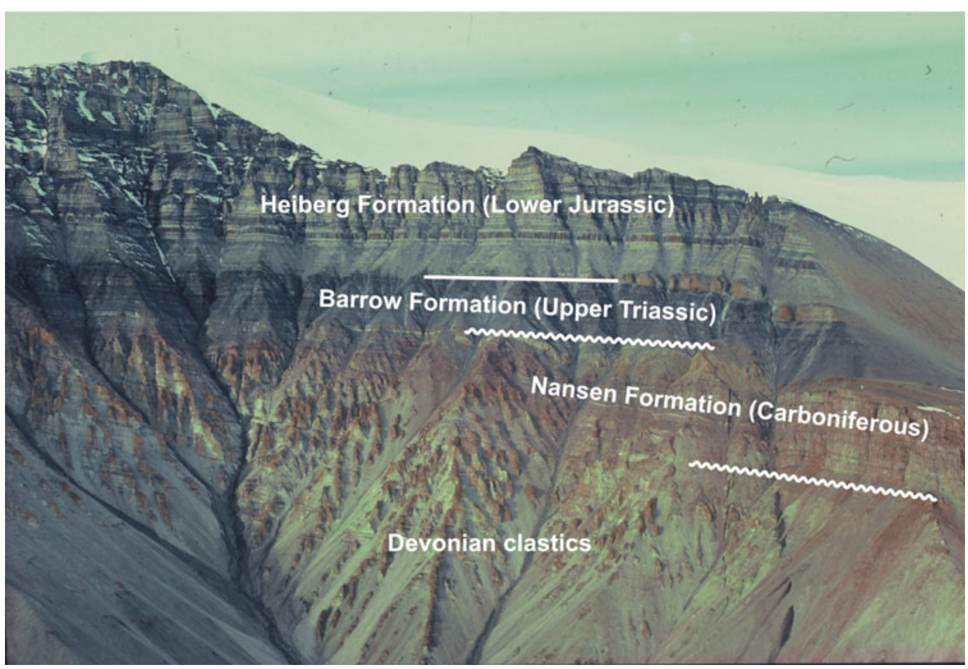


Fig. 7.4 An example of a lithostratigraphic subdivision of a rock succession. Stratigraphic units exposed in the mountains of northern Ellesmere Island, Arctic Canada (photo: A. F. Embry)

Other problems of definition arise where there are lateral lithologic changes, requiring a definition of a new formation. A simple diachronous contact is not a problem, but where the two units intertongue with one another, it may be virtually impossible to draw a simple formation contact. One solution is to give each tongue the same name as the parent formation. A section passing through the transition region may then show the two formations succeeding each other several times. The only problem this causes is if formation contact and thickness data are stored in a data bank and used in automated contouring programs. Without additional input from the operator, a computer program might not be able to handle this type of data. Other alternatives are to define the whole transitional rock volume in terms of one of the parent units, to separate the transitional lithologies as a separate lithostratigraphic entity or to give separate tongues their own bed, tongue or member names. Published stratigraphic codes (e.g., Salvador 1994; NACSN 2005) provide procedures and practical solutions but do not specify any rigid rules for the resolution of such problems. The main criteria should be practicality, convenience and consistency.

The sometimes arbitrary nomenclatural issues raised by lithostratigraphic methods may be clarified or avoided by the use of modern sequence methods, but it should not be forgotten that sequence stratigraphy is an interpretive approach to the rocks, and the inductive, empirical nature of lithostratigraphy will likely remain as an essential underpinning of basinal stratigraphic frameworks for some time to come.

A range of other terms is used to group or subdivide stratigraphic successions on the basis of lithostratigraphy (Fig. 7.5). Vertical and lateral contacts between units may be defined on the basis of clear lithologic change, but are commonly somewhat arbitrary, and as noted above, such subdivisions contain no useful information about the depositional relationships of the strata. We address these issues later, in the discussion on sequence stratigraphy (Sect. 7.6), where it is demonstrated how this genetic approach to stratigraphy can lead to much more meaningful reconstructions and interpretations.

The group. All other stratigraphic units are based on the formation. A group consists of two or more formations related lithologically. In the past, named groups have been established for thick and varied successions without first defining the constituent formations. This is not recommended practice. In contrast, formations defined during reconnaissance exploration may be subdivided into constituent formations and the original name retained and elevated to group status if detailed mapping subsequently provides appropriate data. Groups should not contain major unconformities.

The component formations of a group may not be the same everywhere. Lateral facies changes requiring the definition of different formations can occur within a single group. In contrast, a component formation may extend laterally from one group to another. Groups are normally defined for regions of complex stratigraphy. Toward the basin margin or basin center, the component formations may lose their individuality, in which case the group may be "demoted" to a formation, while still retaining the same name.

The terms supergroup and subgroup are occasionally used to provide an additional hierarchy of subdivisions. Usually, there are historical reasons for this; some of the higher ranking names may have started out as member, formation or group names, with reclassification and promotion being required as additional work demonstrated the need for further subdivision.

The member. This is the next ranking unit below the formation. Not all formations need to be divided into members, and formal names need to be used for only a few, one or none of the constituent members, depending on the convenience or the level of information available.

There are no standards for the thickness or extent of members, and commonly it is difficult to decide whether to define a given lithostratigraphic unit as a member or a

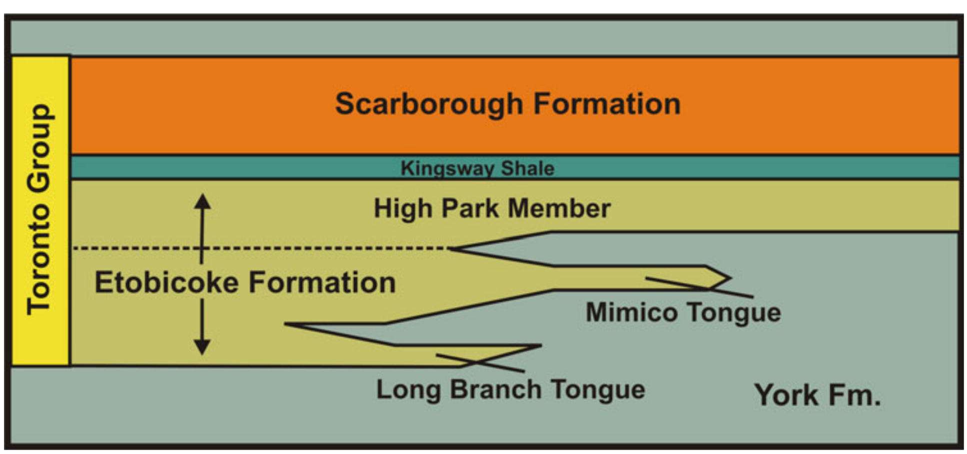


Fig. 7.5 A correlation table for a hypothetical basin fill, drawn to illustrate the various ways in which lithostratigraphic terminology may be adapted to best capture existing stratigraphic variability

formation. However, the recommended practice is that all parts of a succession be subdivided into formations, and so this is the best level at which to start. A member cannot be defined without its parent formation.

For mapping and other purposes, it is commonly convenient to establish informal units, such as the lower sandstone member, which do not require formal names.

Tongue or lentil. These are similar to members. Because of their geometric connotations, the terms are useful for parts of formations where they interfinger with each other. Formal names may be established for one, several or all such units, depending on convenience and practicality.

Bed. This is the smallest formal, named unit in the hierarchy of lithostratigraphic units. Normally, only a few parts of a stratigraphic succession will be subdivided into named beds. Coal seams in mine areas, prominent volcanic tuff horizons and other marker beds are typical examples. Certain stratabound ore-bearing beds, such as placer units, may also be named.

7.4.2 The Names of Lithostratigraphic Units

When establishing a named unit, it is standard practice to give it a geographical name, chosen to suggest the location or areal extent of the unit. This may be a river, lake, bay, headland, hill, mountain, town, village, etc. Permanent names are preferable. Subsurface work in frontier basins, particularly in offshore areas, may rapidly use up all the available names, in which case the name of the well chosen as the type section may be used. Failing this, names may have to be invented.

In most cases, the geographical name will be followed by the rank designation, for example, Wilcox Group, Pocono Formation. For beds, this is commonly not done, particularly in the case of coal seams, for which a complex mine terminology may have evolved. Many older stratigraphic units use a lithologic term instead of a rank term, for example, Gault Clay, Dakota Sandstone and Austin Chalk (e.g., see Fig. 7.3), but this is not recommended because the rank of the unit is not clear from the name alone (NACSN 2005).

Workers should beware of using a geographic name that has already been employed in a different context or renaming units without justification. Geological survey organizations commonly retain a file of current and obsolete stratigraphic names that the worker may wish to consult. Formal naming of units requires that the name be published in a recognized publication, such as a national or international journal. Information required to establish a name includes a designated type section or **stratotype**, with a detailed description of the succession and information about the distribution of the unit and its relationship to overlying, underlying or age-equivalent units in adjacent locations. Further details on the establishment of stratotypes are provided in Sect. 7.8: Chronostratigraphy.

Lithostratigraphic units may be changed in rank as the level of knowledge improves. For example, the Cornwallis Group in the Canadian Arctic Islands started as the Cornwallis Formation and was raised to group rank when it was realized that it contained three mappable units of formation rank. Conversely, the Eureka Sound Group was named for a thick and varied clastic succession, but it was never subdivided into named constituent formations by the original author and was reduced to formation status. The unit has now been subdivided and has been formally redefined as a group once again.

When a unit is raised in rank, the original name should not be used for any of the subdivisions but is best retained for the higher ranking unit or abandoned altogether.

7.5 Biostratigraphy

Biostratigraphy is the study of the relative arrangement of strata based on their fossil content. Descriptive or empirical biostratigraphy is used in erecting zones for local or regional stratigraphic correlation and forms the basis for a global system of chronostratigraphic subdivision. Gradstein (in Gradstein et al. 2020, Chap. 3) provided a succinct description of the major fossil groups used in biostratigraphic studies.

Fossil content varies through a stratigraphic succession for two main reasons: evolutionary changes and ecological differences, such as changes in climate or depositional environment. Biostratigraphy should be based only on evolutionary changes, but it is always difficult to distinguish these from changes that take place in a biostratigraphic assemblage as a result of ecological modifications, and this problem is a cause of continuing controversy for many fossil groups.

Biostratigraphy obviously can only be studied, and a classification erected where fossils are present. This rules out all of the Precambrian, except for the concluding subdivision of the Proterozoic—the Ediacaran (Knoll et al. 2006). Even in the Phanerozoic, there are many rock units for which the fossil record is very sparse, and biostratigraphic subdivision is correspondingly crude. This is particularly the case in nonmarine strata or those (particularly carbonates) in which fossil remains have been destroyed by diagenesis.

Biostratigraphy is a study for specialists. Refined work requires intimate knowledge of the phylogeny of a large number of fossil groups and their regional or global distribution. To accumulate this knowledge may take half a lifetime, and the subject is an excellent example of science in which the practitioner seems to spend inordinate amounts of time “learning more and more about less and less.” Some of the leading authorities in a particular fossil group may be able to discuss the cutting edge of their research with only half a dozen other colleagues around the world. This gives them considerable value if one happens to find their kind of fossil, but it may somewhat restrict their scientific scope. Geologists engaged in basin analysis of the Phanerozoic are very rarely such specialists. Biostratigraphers are therefore employed by many organizations to provide these specialized service skills, or they function independently as consultants. They may be engaged much of the time in pursuing paleontological research but are able to provide biostratigraphic diagnoses for selected fossil types over a specified age range.

Professional biostratigraphic work may take a great deal of field and laboratory time. Sections that a sedimentologist may dismiss as sparsely fossiliferous may yield hundreds or even thousands of specimens to the careful collector. Laboratory extraction of microfossils or palynomorphs may yield similar numbers. It is this kind of work that is necessary for modern, refined biostratigraphic studies. Much of the submitted material, particularly that from frontier exploration wells, may itself provide the basis for new biostratigraphic zoning schemes.

Basin analysts should understand what they are getting when they submit their own material to a specialist for identification. Commonly, they are interested in two items of information: (1) age of the enclosing rock, the information that can be used for correlation purposes, and (2) information regarding the ecological environment of the fossils, which can aid in the interpretation of depositional environments. Age is a chronostratigraphic interpretation based on taxonomic descriptions, but commonly there are problems of fossil identification or interpretation, particularly where the fossil record is sparse, or the material is from a new, poorly studied area. It is particularly important that fossil collections (or the rock containing microfossils or palynomorphs for laboratory extraction) be located as precisely as possible in the outcrop section or well samples from which they were taken. The types of increased stratigraphic precision that are now required for advanced sequence-stratigraphic analysis require this (this is further discussed in Chap. 8). The purpose of this section, therefore, is to discuss some of the problems of the biostratigraphic record and to describe the methods that biostratigraphers use in plying their trade.

7.5.1 The Nature of the Biostratigraphic Record

Biofacies and Biogeography: The geographical distribution of taxa reflects the restriction of ranges due to ecological variations and the geographical isolation of populations. Two topics are included under this heading, the facies control of faunas and floras and the problem of faunal and floral provincialism.

Some taxa are adapted to a benthonic (bottom-dwelling) mode of life and others to a nektonic (swimming) or planktonic (floating) habit. In principal, nektonic or planktonic forms should be preferred for biostratigraphic purposes because of the likelihood of being more widely distributed and therefore more broadly useful. Benthonic forms tend to be more facies dependent because of their need for certain water conditions or sediment types for feeding and dwelling behavior. However, in practice, benthonic forms are widely used by professional biostratigraphers. Even such static forms as corals, burrowing mollusks, and anchored brachiopods have been found to be invaluable for zoning the deposits of the continental shelves. Many benthonic taxa have a planktonic larval stage that ensures wide distribution via marine circulation. Conversely, many planktonic forms, such as the graptolites, are too fragile to survive in agitated, shallow-water environments and are therefore just as facies-bound as their benthonic contemporaries. In practice, virtually every taxonomic group has some biostratigraphic utility, although considerable problems may arise in attempts to determine the relationships between the various facies-bound faunas, unless environmental fluctuations cause lithofacies of different types, with their accompanying faunas or floras, to become interbedded (Fig. 7.6). Where well exposed, such mixed successions are of great value in establishing a global chronostratigraphic framework. Figure 7.6 illustrates such a scenario schematically, where marine foraminiferal zones can be correlated to nonmarine palynomorphs because of the interfingering of these facies zones. The zig-zag “shazam” interfingering configuration used to illustrate facies relationships in this diagram is an overly simplistic representation of the progradation and retrogradation that occurs as a result of changes in relative sea level but helps to make the point that the vertical range of a given assemblage may be **diachronous**, changing in age laterally as a result of shifts through time of facies belts.

Classic examples of facies-bound faunas widely used by biostratigraphers are the **shelly** and **graptolitic** faunas of the lower Paleozoic. The shelly fauna actually includes two more or less distinct subfaunas, one in the inner, shallower shelf dominated by brachiopods and the other on the outer shelf, characterized by trilobites. The term appears to have arisen with Russian work that defined a “small shelly fauna,” consisting of small exoskeletons of many different types (Matthews and Missarzhevsky 1975). The graptolitic fauna is confined mainly to low-energy deposits of the continental slope, rise and abyssal plain (Berry 1977). In Newfoundland, on what was the ancient eastern continental margin of North America during the Cambrian and Ordovician, these two facies interfinger. Carbonate turbidites and debris flows,

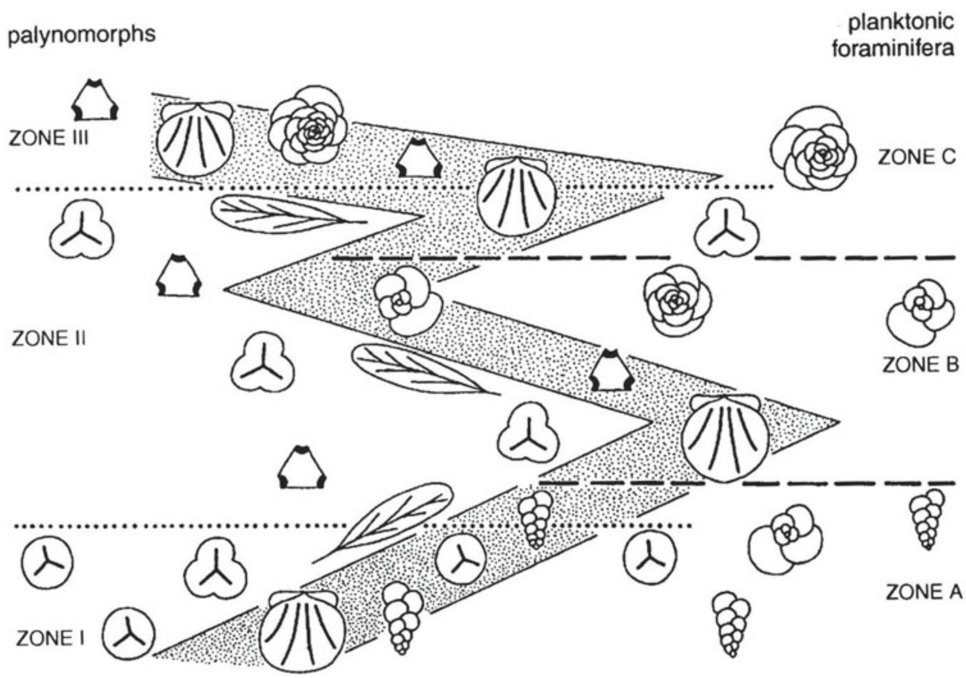


Fig. 7.6 The interbedding of three biofacies resulting from shifting of environments. Typically, such interbedding is caused by a sea-level change (McGowran 2005, Fig. 1.4, p. 8)

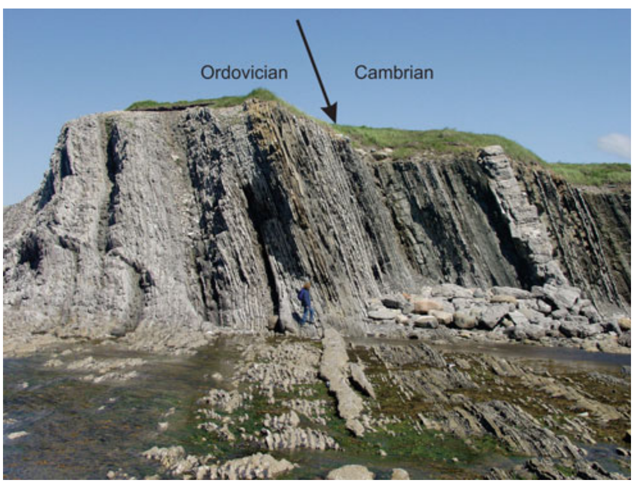


Fig. 7.7 Interbedding of a shelly fauna in limestone sediment-gravity flows and a graptolitic fauna in interbedded shales, Cambrian–Ordovician continental-margin deposits, Green Point, Newfoundland. The beds here are overturned. This location has been designated as a stratotype for the Cambrian–Ordovician boundary

derived from the collapse of the continental margin, are interbedded with graptolitic shales at the base of the continental slope (Fig. 7.7). At Green Point, on the west coast, this assemblage straddles the Cambrian–Ordovician boundary, and the location has been established as the stratotype for this important chronostratigraphic boundary (the GSSP for the base of the Tremadocian Series: Cooper et al. 2001; see Sect. 7.8.1 for a discussion of the GSSP).

A good example of facies control of what might appear at first sight to be a recurrent, biostratigraphically controlled fauna is provided by the brachiopod communities of the Upper Ordovician to Middle Silurian of the Welsh Borderlands. Ziegler et al. (1968) showed that at many localities there is a sequence of assemblages containing, in upward stratigraphic order, *Lingula*, *Eocoelia*, *Pentamerus*, *Stricklandia* and *Clorinda*, followed by a graptolitic fauna. Careful correlation of these sections using graptolites showed that the brachiopod sequence is markedly diachronous. This is shown in Fig. 7.8, in which the graptolite zones are shown by horizontal correlation lines labeled A1 to C7. One might ask, why are the graptolites trusted more than the brachiopods for the purposes of chronostratigraphic correlation? The answer is that brachiopods are benthonic organisms known to be prone to facies control, whereas graptolites are tried and tested biostratigraphic indicators. The correlations shown in Fig. 7.8 are supported by additional work on two of the brachiopod genera, *Eocoelia* and *Stricklandia*. When examined in detail, evolutionary trends can be detected within the populations of these genera as they are traced from west to east across the line of the section shown in Fig. 7.8 (and other sections not shown). The overall interpretation of these faunal data is that the sections reveal a gradual eastward marine transgression and deepening of the water such that successive brachiopod communities represent ecological adjustments to increased depths (*Lingula* inhabited brackish waters, and in part, the subsequent succession represents changing shell thicknesses in response to wave and tidal energy). There is no

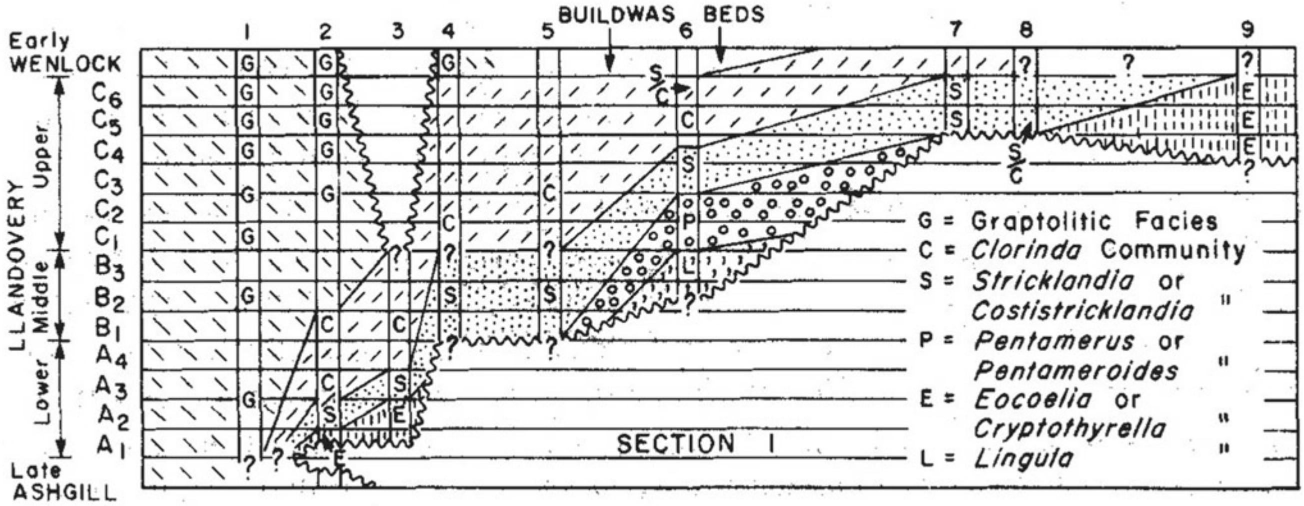


Fig. 7.8 An example of facies-bound faunas. The succession of brachiopod communities in each of these sections is the same (L, E, P, S, C), but the use of graptolite biostratigraphy (biozones A1 to C7) shows that they are markedly diachronous and therefore facies-controlled. Silurian, Welsh Borderland (Ziegler et al. 1968; McKerrow 1971)

obvious relationship between brachiopod assemblage and sediment type in this case.

Biogeography and evolution: William Smith recognized the significance of the succession of faunas in the stratigraphic record long before Darwin's theory of evolution was established. The point is that it is not necessary to understand evolutionary relationships of fossil groups in order to make use of them as tools in the inductive establishment of stratigraphic order and correlation. However, it can certainly help to explain the nature of taxonomic change and the distribution of distinct groups through time and space.

Molecular biology shows that the underlying process of evolution is genetic drift, the gradual accumulation of random incremental change in gene variants. Natural selection favors some mutations over others, which generates steady change. Isolation of populations will also tend to increase genetic divergence, eventually to the point that populations will be unable to interbreed and then constitute distinct species (see the review of current ideas by Kelley et al. 2013).

Three styles of evolution were described by Eldridge and Gould (1977), and are illustrated in Fig. 7.9. The first, termed **phyletic gradualism** or **transformational evolution** (McGowran 2005, p. 382), refers to long-term evolutionary change, typically in response to geographical, climatic or other environmental pressures (Fig. 7.9a). Certain varieties of a species may be favored by these changes so that there is a gradual adjustment in the stock until a distinctive new species appears.

Kauffman (1977) described two examples of phyletic gradualism in Cretaceous pelecypods. Figure 7.10 illustrates a series of histograms of height–width ratios of the *Inoceramus pictus* lineage, derived from populations collected at about 40 cm intervals (lower graph), and the number of growth ridges in the first 25 mm of shells of *Mytiloides labiatus* (upper graph). The data permit subdivision of the population into species (S), subspecies (SS) and morphological zones (MZ), as indicated in the adjacent columns.

Another example of gradual evolution is provided by the foraminifera *Globigerina* and *Orbulina* (Fig. 7.11). The species listed in the top are an evolutionary series, three of which are illustrated (1: *Globigerina quadrilobatus*; 2: *G. bisphericus*; and 3: *Orbulina universa*). The gradual variation between these types has permitted the erection of seven zones, as indicated by the horizontal lines at the left.

Two types of environmental adaptation can occur, that which is accompanied by permanent genetic change and that which can, to some extent (never precisely), reverse itself to recreate the same variety or race of a species more than once, whenever the same environmental conditions are repeated (homeomorphy). Clearly, the first type is the only one of use to biostratigraphers, but the literature is replete with ambiguous biostratigraphic determinations that may be falsely based on diachronous environmental change. For example, this has been a serious problem with the ammonites (Kennedy and Cobban 1977), one of the best biostratigraphic indicators.

Taxa that evolve by phyletic gradualism have the most potential for refined biostratigraphic zonation, but they require specialist study to recognize the very subtle changes between the varieties. This type of work is beyond the abilities of the generalist basin analyst.

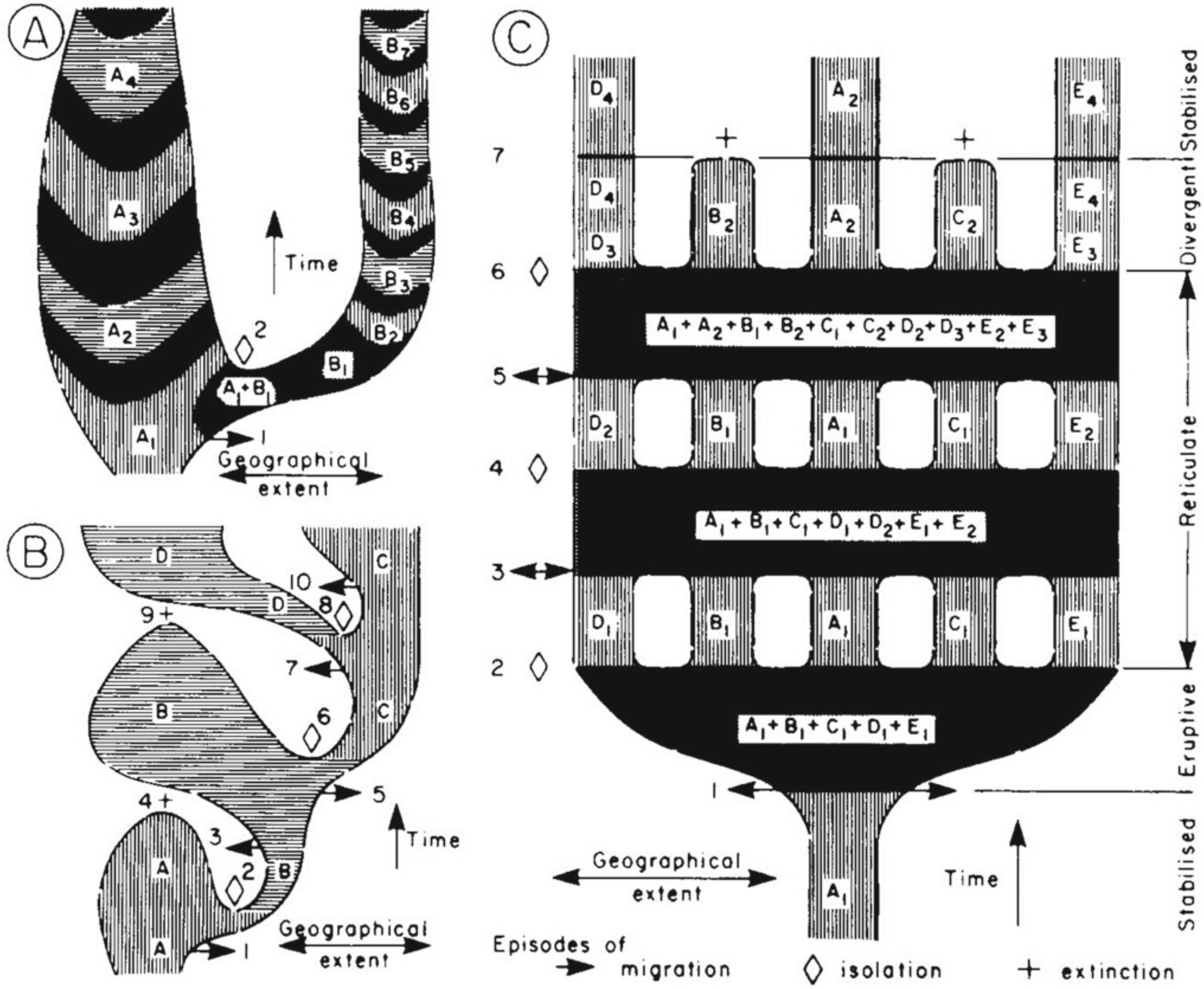


Fig. 7.9 Three models of evolution: A phyletic gradualism; B punctuated equilibrium; and C. reticulate speciation. A, B, etc., refer to successive varieties or species; 1, 2, etc., refer to the chronology of events. In B, the same area at the left of the diagram is successively occupied by three species A, B, and C, which evolve elsewhere and migrate in (Sylvester-Bradley, 1977)

The second style of evolution was named **punctuated equilibrium** by Eldridge and Gould (1972, 1977). The concept was adapted from an earlier term, **allopatric speciation**, and is based on the premise that in a successful, widely distributed taxon the population is genetically conservative. Evolution is thought to occur only where extreme variants are selected by environmental pressures on the fringes of the species range. Rather than a gradual adaptation to an ecological niche or a broadening of a species range by extending slowly into subtly different niches, as in phyletic gradualism, the hypothesis of punctuated equilibrium proposes the spasmodic occurrence of bursts of relatively rapid evolutionary change. Extreme variants of a species can only evolve into a new species if they become isolated by changes in the environment, climate or geography, as through the rifting and drifting apart of continental plates. Also, the catastrophic extinction of organisms by bolide impacts or other catastrophes empties out many ecological niches and permits rapid adaptive radiation and the explosive development of many new taxa in the period immediately following these extinctions (Fig. 7.9b).

Sylvester-Bradley (1977) proposed a third style of evolution, which he termed **reticulate speciation** (Fig. 7.9c). This combines, on a small scale, the mechanisms of both the other two evolutionary styles. Gene transfer may take place by several processes, including symbiosis, lateral transfer and hybridization. Sylvester-Bradley offered the modern common vole as an example of a taxon that has evolved in this way. The vole is distributed virtually globally and comprises numerous races reflecting adaptation to local

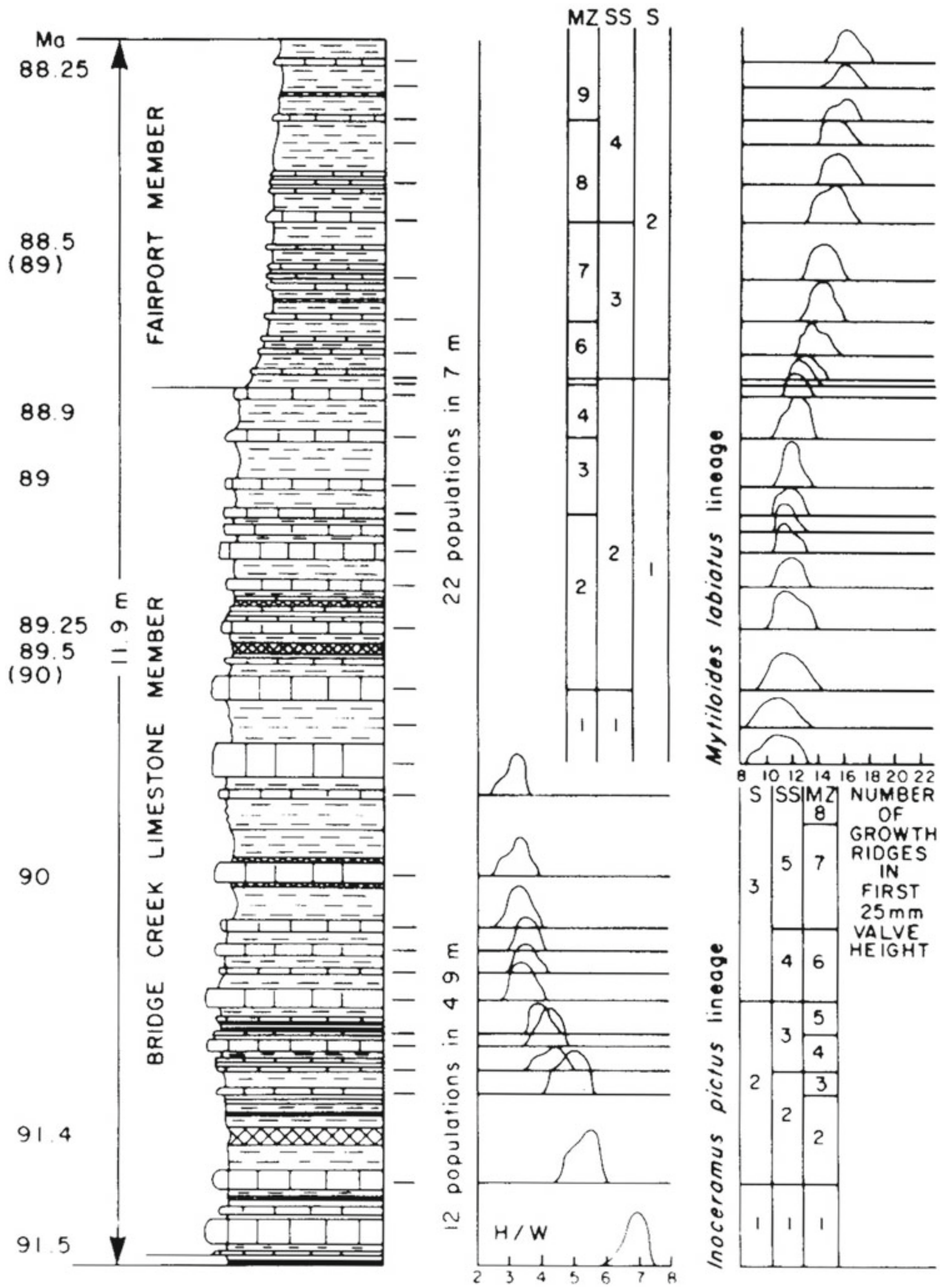


Fig. 7.10 Evolutionary trends in two pelecypod species, Cretaceous of Western Interior. Symbol X in the lithologic column indicates bentonite beds used for radioisotopic age dating. Suggested systematics and zonal subdivisions are shown in the numbered columns: MZ, morphological biozone; SS, subspecies; S, species (Kauffman 1977)

variations in climate, vegetation, altitude, isolation on islands etc. These varieties have evolved in response to rapid global changes following the Pleistocene ice age, and they demonstrate the rapidity with which geographical and ecological changes may bring about evolution. Grant and Grant (2008) similarly observed rapid reticulate speciation in Darwin's finches on the Galapagos Islands. The apparent stability of the species of many taxonomic groups for several million years or more at times during the geological past contrasts with the rapid adaptability of the modern vole and the finch. To what extent reticulate speciation will be recognized for fossil groups remains to be determined. However, to recognize this style of evolution would seem to require an immense bank of detailed descriptive data, and therefore it is very much a subject of study for specialists.

A review of modern concepts in evolution as applied to the fossil record is provided by Kelley et al. (2013).

In the geological record, many distinct populations have been recognized, based on geographical distributions, and are defined as **faunal provinces**. These are much discussed by biostratigraphers. To the nonspecialist such concepts as the Malvinokaffric Province or the Tethyan or Boreal

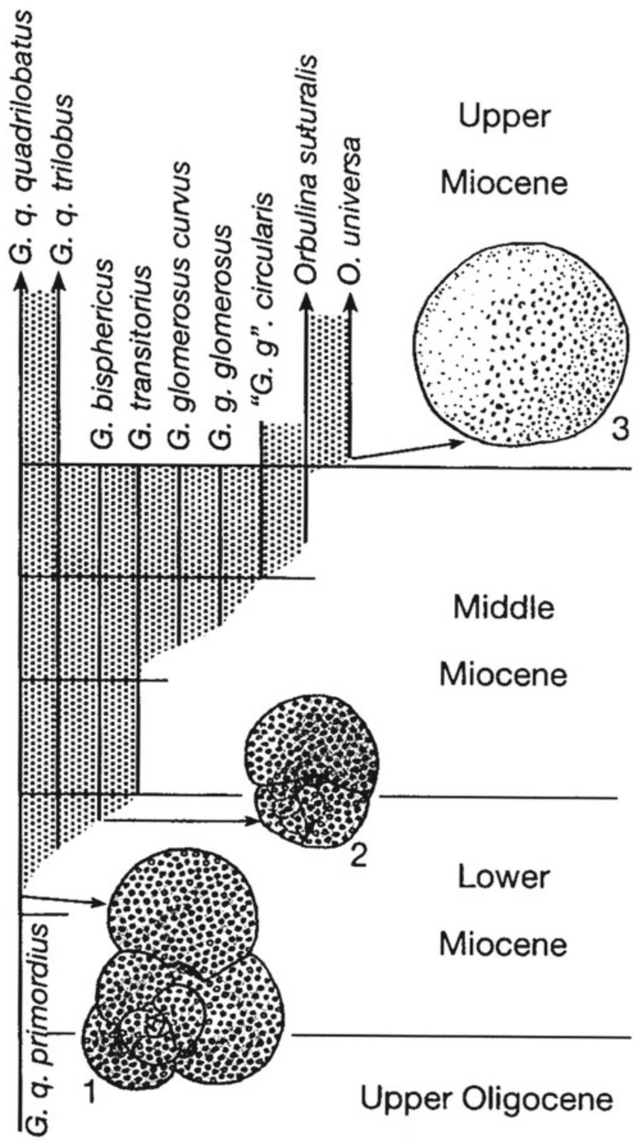


Fig. 7.11 Evolution of the foraminifera *Globigerina*, culminating in the different genus *Orbulina*, in southern Australia. The stippling pattern indicates continuous variation between the various morphotypes and their ranges. Seven zones are defined by the horizontal lines at left (from McGowran 2005, Fig. 1.4, p. 95)

Realms are sometimes difficult to understand. The definition of what constitutes a given faunal province requires a great deal of specialist knowledge, but even the specialists had difficulty before the advent of plate tectonics in comprehending why many of these provinces existed. Until the 1970s there was much discussion of the appearance and disappearance of strange, narrow "land bridges" to explain the merging and divergence of provincial variations. But in fact, faunal provincialism provides some of the most convincing geological lines of evidence for plate tectonics (Tarling 1982).

One of the most famous of these is the example of the trilobite faunas flanking Iapetus, the proto-Atlantic Ocean that developed between Laurentia (ancient North America) and Baltica and Africa in the earliest Paleozoic. Specialists noted significant differences in Cambrian trilobites between those found in England and those in Scotland, and between those from western and eastern occurrences in Newfoundland, whereas Scottish trilobites are similar to those in western Newfoundland, and English trilobites are similar to those in eastern Newfoundland (Fig. 7.12). Given the current geographic distributions of the fossils, these similarities and differences make no biogeographic sense. However, Wilson (1966) cited these distributions as one of several lines of evidence in his proposal—now universally accepted—that the line demarking the two distinct trilobite faunal provinces constitutes an ancient continental suture that formed when a former ocean closed as a result of subduction and continental collision. That ocean, now called Iapetus, occupied approximately the position of the present Atlantic Ocean, but many significant continental fragments changed margins when the present ocean developed; that is, the continental rift occurred along a somewhat different line. Landing et al. (2013) reviewed in detail the faunal provincialism that developed in the Cambrian following the breakup of Rodinia.

The ammonites provide an excellent example of the various biogeographic styles that can occur in organisms, some offering considerable advantages to the biostratigrapher, others a severe hindrance. Many ammonites underwent a planktonic larval stage that may have lasted from hours to weeks. Where this occurred, it would have been of some importance to the distribution of the species. Not all ammonites showed this. Distribution patterns and varying degrees of facies independence show that some ammonites were benthonic in adult life habitat, some were nektonic and some may have been planktonic. Their facies distribution and provincial tendencies thus varied considerably. Some ammonites may have drifted long distances after death. The modern Nautilus, the only living relative of the ammonites, has a buoyant shell after death, and observations in modern oceans suggest that the shell may drift for hundreds, if not thousands of kilometers. Geologically this could be of great importance, but Kennedy and Cobban (1977) suggested that many ammonites in fact became rapidly waterlogged after death and did not float appreciable distances.

Kennedy and Cobban (1977) summarized much of the data for Cretaceous ammonite distribution and concluded that there were five types of faunal provinces. Some genera have a virtually worldwide, or **pandemic**, distribution. Pandemic taxa would seem to offer the best possibilities for global correlation. They are relatively facies independent,

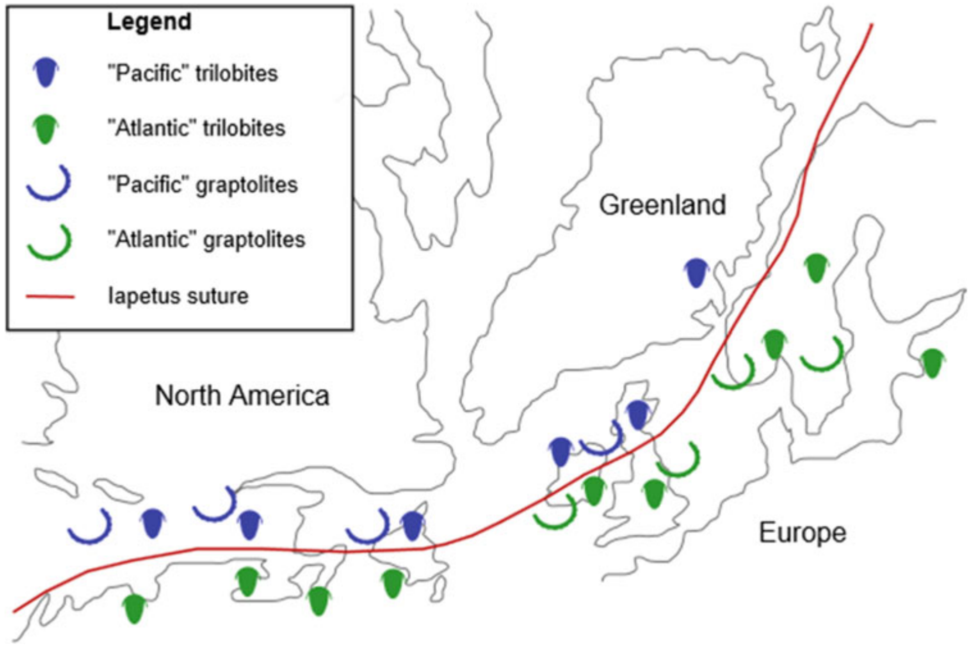


Fig. 7.12 The Cambrian trilobite and graptolite faunas of the Iapetus margins. Based on Wilson (1966)

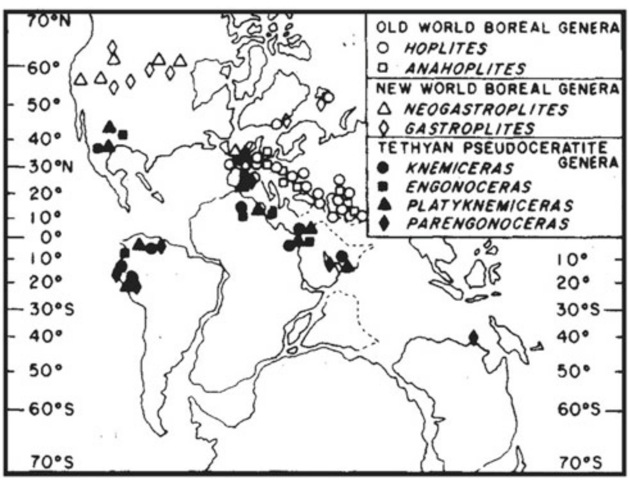


Fig. 7.13 The biogeography of selected Cretaceous ammonites, plotted on a Cretaceous plate-tectonic reconstruction of the continents. Climatic tolerances underlie the different geographic spread of Boreal versus Tethyan forms (an example of latitudinally restricted distributions), while the gradually widening Atlantic Ocean caused the gradual isolation and separate evolutionary development of ammonite faunas on different sides of the ocean (longitudinally restricted distributions). These faunas occasionally mixed in the Western Interior Seaway of North America because of shifts in climate and changes in sea level that caused local faunal migrations along a north–south axis (Kennedy and Cobban 1977)

but it turns out that many are long-ranging forms and thus of limited biostratigraphic usefulness.

Some ammonites have **latitudinally restricted** distributions, reflecting their preference for waters of a certain temperature or salinity and their tolerance of seasonal fluctuations. Some examples are shown in Fig. 7.13. They define two provinces, the northern, colder water Boreal province (open symbols) and the more tropical Tethyan province (closed symbols). In many parts of Europe and North America, faunal fluctuations through stratigraphic successions between Tethyan and Boreal (and other) faunas have been cited as evidence of the existence of connecting seaways and transgressions across otherwise barren areas.

Longitudinal restrictions on distribution, such as the presence of land masses or large ocean basins, are a cause of further provincialism. These are added to latitudinal restrictions in the generation of the third type of faunal province: **endemic** distributions. Note that in Fig. 7.13 the Tethyan genera show no longitudinal restriction, whereas the four Boreal genera are typical endemic taxa, restricted to either Eurasia or North America (these are examples chosen to illustrate a point and should not be taken to define a universal difference between the Boreal and Tethyan provinces). Endemic ammonites have been shown to have evolved rapidly and thus are of prime biostratigraphic importance, although their provincialism has hindered intercontinental correlation.

Disjunct distributions are those of scattered but nevertheless widely distributed taxa. The distributions are not thought to represent inadequate data or severe facies control but probably reflect very low population densities.

As noted, some ammonite taxa may drift in oceanic currents after death. In extreme cases, where an endemic form is involved, such drifted or **necrotic** distributions may prove invaluable for long-distance correlation.

In general, the provinciality of taxa increases the difficulty by which they may be related to the global time scale, because provincialism reduces the variety of forms that may be used to establish relative ages.

7.5.2 Biochronology: Zones and Datums

By the mid-nineteenth century, the work of the early stratigraphers, following Smith, had clearly established the value of fossil assemblages for the establishment of stratigraphic order and for the purposes of comparison between stratigraphic sections. The similarity of the succession of faunas or floras between sections in different basins, even different continents—termed **homotaxis**—was well established. The main elements of the Phanerozoic geological time scale had been defined, including all the names of the periods (Berry 1987), and the concepts of the biozone and the stage were already developed (Hancock 1977). However, there remained the issue of age. In the absence of a clear understanding of how faunas and floras changed with time, and without the tools to establish numerical age, it remained a legitimate question whether homotaxis could be equated with **synchrony**. Darwin's *The Origin of Species* was published in 1859, yet in 1862 T. H. Huxley stated "for anything that geology or palaeontology are able to show to the contrary, a Devonian fauna or flora in the British Isles may have been contemporaneous with Silurian life in North America, and with a Carboniferous fauna and flora in Africa."

Eventually, this philosophical dilemma was resolved by developments in the understanding of the processes of evolution, coupled with the establishment of even more detailed systems of zonation and correlation, which left little room for doubt regarding the reality of the principle of relative age and time correlation based on fossil content. Some of the steps in the evolution of thought are described by Hancock (1977), Miall (2004) and McGowran (2005, pp. 54–65).

It is now universally accepted that formally established biozones represent specific intervals of time, subject to two important caveats: (1) the changes in taxonomy that lead to the definition of discrete datum planes of change, or zones of similarity in the fossil record, are not globally instantaneous. The appearance of a successful evolutionary step requires a discrete period of time for it to spread throughout its full range. This time period may be short, in geological terms, but it is not instantaneous and may be measurable. We return to this problem in Sect. 7.5.3. (2) The geographic range of a biostratigraphic datum or zone is limited by ecological factors. A biozone may not, therefore, have exactly the same age range everywhere.

There are various methods available for making the most efficient use of fossil occurrences. Particularly distinctive and/or abundant forms may serve to represent a specific span of time. Such forms are called **index fossils**. The first or last appearance of particular, distinctive species is commonly employed as biomarkers. These horizons are termed the **first appearance datum (FAD)** and the **last appearance datum (LAD).** Suites of fossils may be used to define **biozones**. This may be done in several different ways (Fig. 7.14). Figure 7.15 illustrates one of the more common methods of defining a biozone, which takes advantage of the fact that the ranges of different species typically overlap. In this diagram a **concurrent range zone** is defined as that interval of the rocks within which all three of the fossils A, B and C are

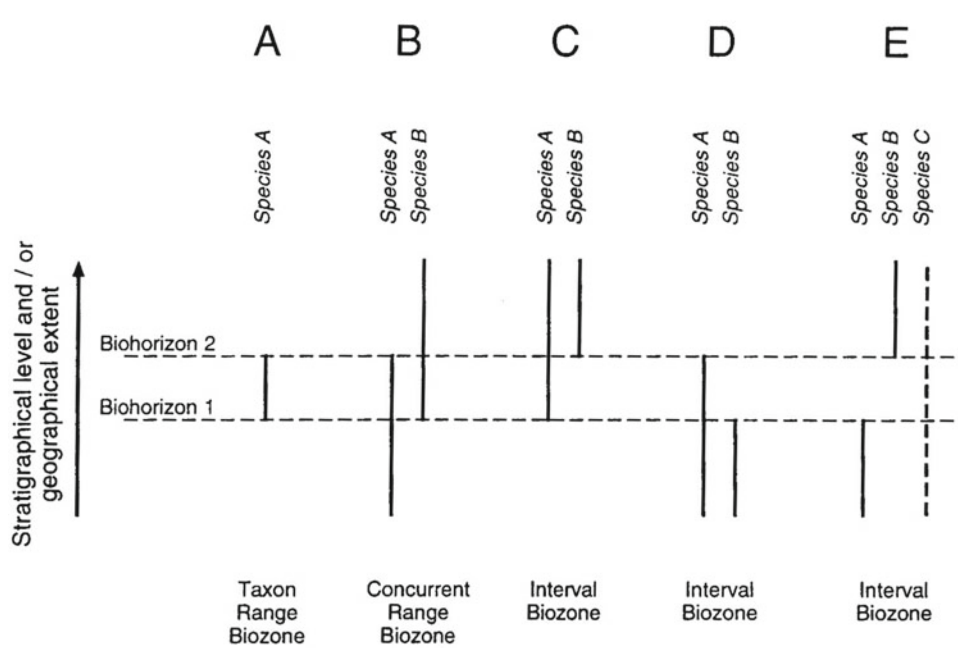


Fig. 7.14 Types of biozone, as defined in the International Stratigraphic Guide (Salvador 1994. Diagram from Pearson (1998), Fig. 5.2, p. 126)

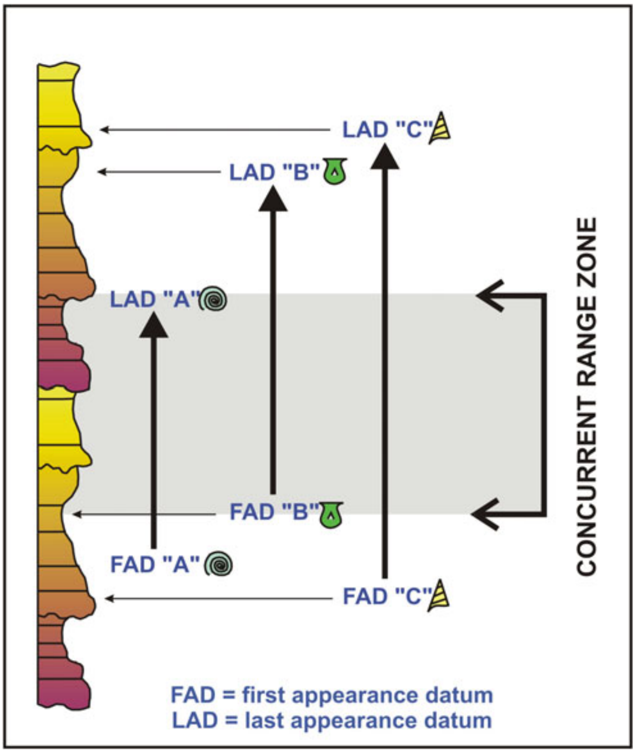


Fig. 7.15 The use of first and last appearance datums to define a concurrent range biozone. The area colored in grey is the only part of the section where all three species, A, B and C are expected to be present

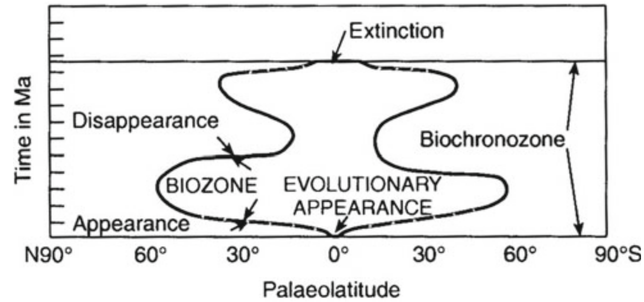


Fig. 7.16 The diachroneity of a biozone. A taxon first appears near the equator and takes a discrete amount of time (up to 2 m.y.) to migrate into higher latitudes north and south of the equator. The range does not then remain constant because of ecological factors, in this case, climate change, which leads to contraction and then renewed expansion of the range, until the taxon becomes extinct. The biochronozone (or biochron) of this taxon (the time span it represents) is global in extent but, in practice, the time span is that which is indicated by the actual presence of the taxon, which varies from place to place and, in this illustration, is absent altogether in high latitudes (McGowran 2005, Fig. 2.11, p. 38; after Loutit et al. 1988)

present, and falls between the FAD of species B and the LAD of species A.

It is important to be aware that the time range indicated by a biozone is not necessarily the same everywhere. Figure 7.16 illustrates a hypothetical example where the diachronous spread of a fossil taxon and subsequent variations in its range owing to climatic factors has led to significant variability in the local range of the biozone. Diachroneity is discussed further in the next section. There are sophisticated methods for managing these issues and, as discussed in Sect. 7.8, biostratigraphy is still the major foundation of the geological time scale for the Phanerozoic.

In addition to ecological factors, there are considerations of preservation and potential sampling bias. Diagenetic destruction of fossils is common, and sampling bias might simply reflect bad luck in the choice of sampling site, or a bias induced by poor collection practices (Fig. 7.17). For example, whereas a field geologist undertaking a reconnaissance mapping exercise might be satisfied with a cursory examination of an outcrop for fossil content, a professional biostratigrapher is likely to carry out more thorough investigations. For example, soft sands and clays may be run through a sieve or water-washed on site to isolate macrofossils, or large lithologic samples may be collected at routine stratigraphic intervals for water or acid treatment to extract microfossils or palynomorphs. In the subsurface, drilling disturbance may constitute a major problem. The drilling process penetrates layers from the top down, so the stratigraphically last appearance of a fossil (the LAD) will be the first encounter with a given taxon, and the level of this horizon is one that can be trusted. However, the tendency for holes to cave can lead to fossils (and rock cuttings) being fed into the mud stream after the drill bit has passed on down their point of origin (Fig. 2.28). For this reason, the FAD of a fossil taxon in the subsurface needs to be treated with caution.

As Sadler et al. (2014, p. 4) stated: "Signals and noise mingle among these contradictions. Records of real ecological patch dynamics, biogeographical habitat shifts and evolutionary turnover are confounded by incomplete preservation and collection."

Figure 7.18 is an example of palynological zonation of two wells through a Cretaceous succession in Delaware (from Doyle 1977). The wells are represented by their gamma ray logs with sample collection depths given in feet. The ranges of the principal angiosperm pollen types are shown by vertical bars and are shown dashed where identification is uncertain. Concurrent range zones are delimited by dashed lines perpendicular to the depth scale and are numbered I to IV next to the Series, Stage and Formation designations. It was found that the zones could be most easily defined on the basis of the first (oldest) appearance of a taxon, partly because extinct species tended to be reworked, and partly because taxa were found to die out slowly at the upper limit of their range. Work of this type required the counting and documentation of several

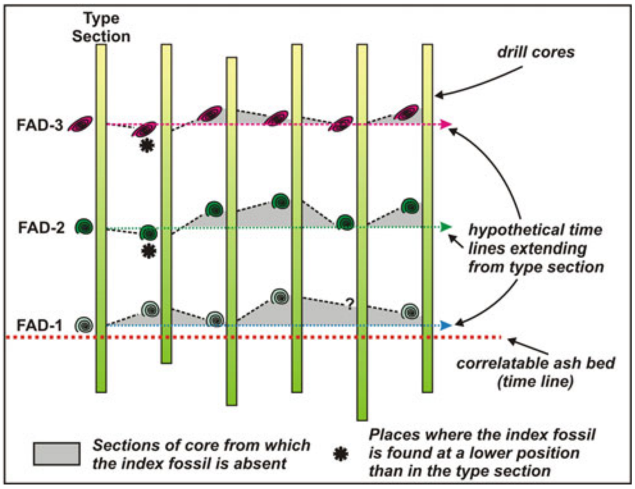


Fig. 7.17 First and last appearance data define ideal timelines in the rocks. However, in this example, many first occurrences are higher (younger) than the hypothetical timelines would predict and a few are lower (older). The reasons for this are discussed in the text

hundreds of individual pollen grains in each sample. Several or many complete sections through the succession of interest may be required before the data are adequate for the definition of the biozones. Range charts, such as that illustrated here, must be prepared for each and carefully compared.

Another way to define biozones focuses on the gradual change in the anatomy of a particular evolutionary lineage. This is called a **lineage zone**. McGowran (2005) used an example from the late Cenozoic mammalian evolution to illustrate this concept (Fig. 7.19). The evolution of dental morphology provided much of the information on which this zonal scheme was based.

7.5.3 Diachroneity of the Biostratigraphic Record

Another common item of conventional wisdom is that evolutionary changes in faunal assemblages are dispersed so rapidly that, on geological time scales, they can be essentially regarded as instantaneous. This argument is used, in particular, to justify the interpretation of FADs as time-stratigraphic events (setting aside the problems of preservation discussed above). However, this is not always the case. Some examples of detailed work have demonstrated considerable diachroneity in important pelagic fossil groups. Landing et al. (2013) provided an overview of the problem, focusing in particular on the Cambrian. Cramer et al. (2015) pointed out that many "events" that are assumed to be instantaneous on a geological time scale, may, in fact, be diachronous on a finer time scale. With our increasing ability to provide chronostratigraphic control on events in deep time to a $\pm 0.1\%$ level of accuracy (Sects. 7.8.2, 8.10.3), quantifying diachroneity may become increasingly important.

MacLeod and Keller (1991) explored the completeness of the stratigraphic sections that span the Cretaceous-Tertiary boundary, as a basis for an examination of the various hypotheses that have been proposed to explain the dramatic global extinction occurring at that time. They used graphic correlation methods and were able to demonstrate that many foraminiferal FADs and LADs are diachronous. Maximum diachroneity at this time is indicated by the species *Subbotina pseudobulloides,* the FAD of which may vary by up to 250 ka between Texas and North Africa. However, it is not clear how much of this apparent diachroneity is due to preservational factors.

An even more startling example of diachroneity is that reported by Jenkins and Gamson (1993). The FAD of the Neogene foraminifera *Globorotalia truncatulinoides* differs by 600 ka between the southeast Pacific Ocean and the North Atlantic Ocean, based on analysis of much DSDP material. This is interpreted as indicating the time taken for the organism to migrate northward from the South Pacific following its first evolutionary appearance there. As Jenkins and Gamson (1993) concluded:

> The implications are that some of the well documented evolutionary lineages in the Cenozoic may show similar patterns of evolution being limited to discrete ocean water masses followed by later migration into other oceans ... If this is true, then some of these so-called 'datum planes' are diachronous.

Cody et al. (2008) provided another example of diachroneity. They reported on the distribution of diatoms in 32 Neogene cores from the Southern Ocean, which allowed an estimate to be made of the differences between the levels of the observed local first occurrence and last occurrence events and the projected levels of the global FADs and LADs for the same species. Around 50% of local event levels do not accurately record the timing of the global event: a few are off by 4 Ma or more from the total global FAD and LAD, due mostly to a small set of individually incomplete local ranges.

These conclusions are of considerable importance, because the results were derived from excellent data, and can, therefore, be regarded as highly reliable. They relate to some of the most universally preferred fossil groups for Mesozoic–Cenozoic biostratigraphic purposes, diatoms and foraminifera. It would appear to suggest a limit of up to about one-half million years on the precision that can be expected of any biostratigraphic event. Figure 7.16 illustrates the general problem.

The cases reported here may or may not be a fair representation of the magnitude of diachroneity in general,

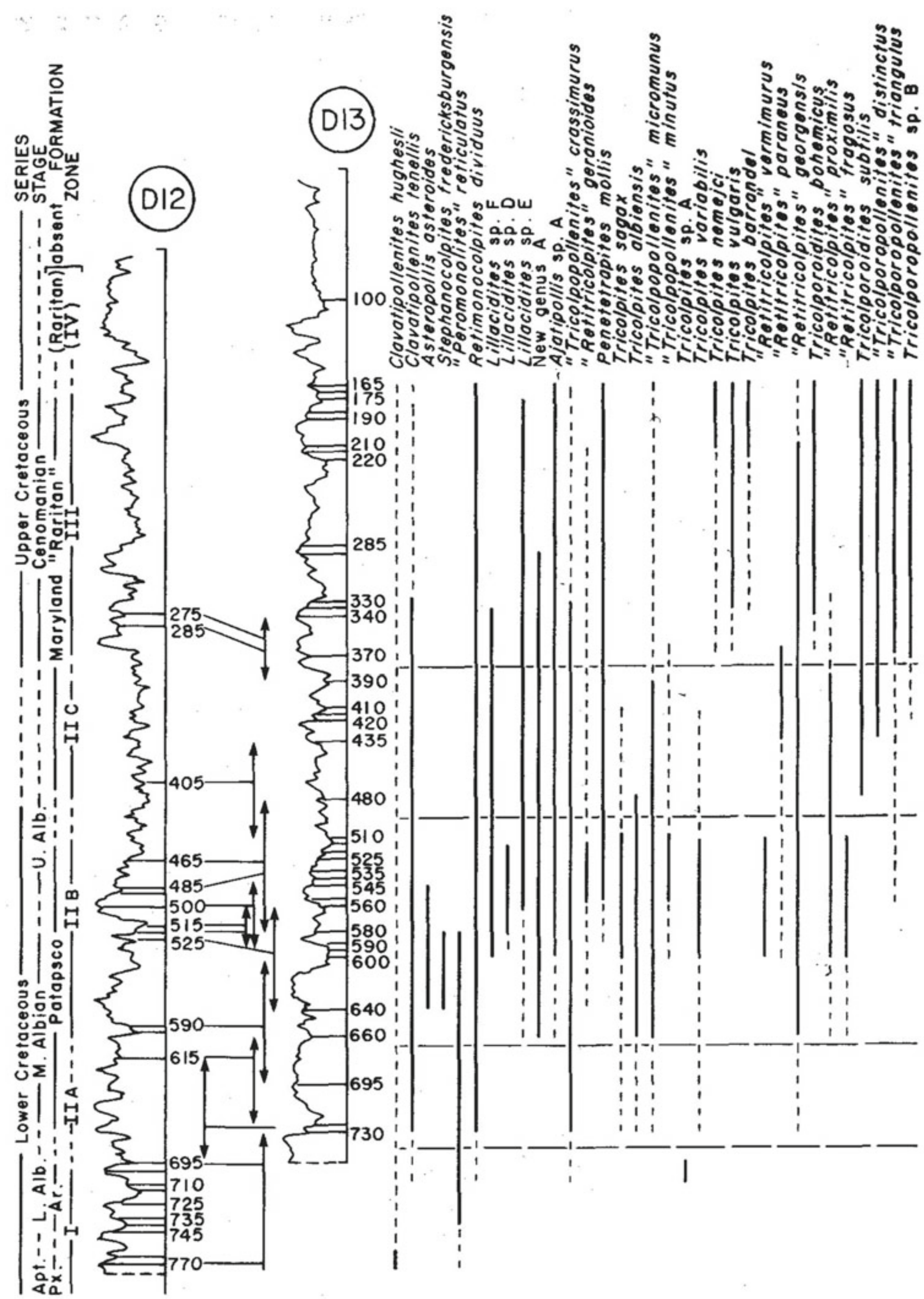


Fig. 7.18 Use of palynological concurrent range biozones to correlate two subsurface wells (as shown by gamma ray logs). Correlation brackets (double-headed arrows) terminate just above and below samples in the other well that bracket the age of the indicated sample (Doyle 1977)

although this is recognized as a general problem (Smith et al. 2015). After a great deal of study, experienced biostratigraphers commonly determine that some species are more reliable or consistent in their occurrence than others. Such forms may be termed **index fossils**, and receive a prominence reflecting their usefulness in stratigraphic studies. Studies may indicate that some groups are more reliable than others as biostratigraphic indicators. For example, Ziegler et al. (1968) demonstrated that brachiopod successions in the Welsh Paleozoic record were facies controlled and markedly diachronous, based on the use of the zonal scheme provided by graptolites as the primary indicator of relative time (Fig. 7.8). Armentrout (1981) used diatom zones to demonstrate that molluscan stages are time transgressive in the Cenozoic rocks of the northwest United States. Wignall (1991) demonstrated the diachroneity of Jurassic ostracod zones. Landing et al. (2013, p. 136) offered this general caution.

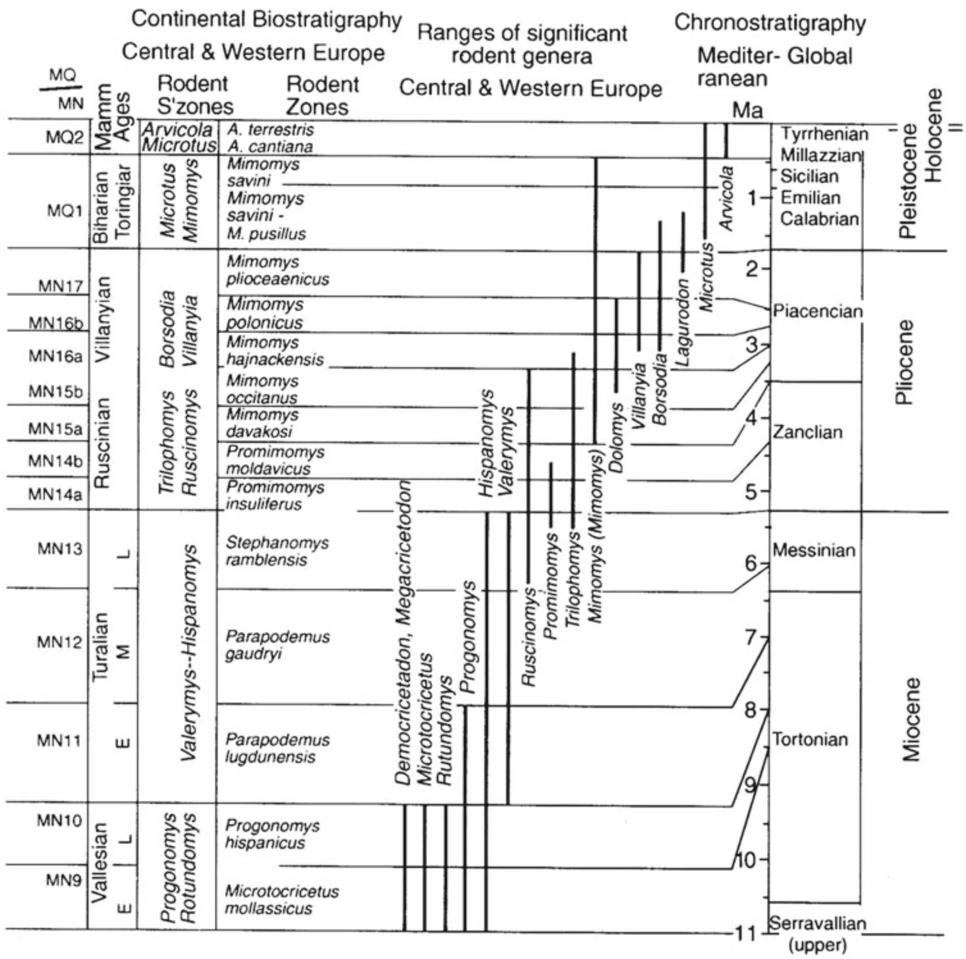


Fig. 7.19 Lineage zones defined by rodent evolution (McGowran 2005, Fig. 4.31, p. 150, based on Fejfar and Heinrich 1989)

Use of the local FADs of a fossil for correlation between sections without rigorous supplementary information will lead to errors in correlation or poorly defined chronostratigraphic units because significant time intervals likely will separate the local FADs. Each FAD must mirror biological phenomena ranging from evolutionary origination, to dispersal, successful local colonisation, and appearance of facies that allow a species' fossilisation

7.5.4 Quantitative Methods in Biochronology

The graphic correlation technique. This method was first described in a landmark book by Shaw (1964). Useful explanations of the technique were given by Miller (1977) and Edwards (1984, 1985). Mann and Lane (1995) edited a research collection devoted to the application of this topic to practical problems in basin analysis. Gradstein et al. (2004b) discussed the use of the method in the construction of the geological time scale. The example used herein has been borrowed from Miller (1977).

As with conventional biostratigraphy, the graphic method relies on the careful field or laboratory recording of occurrence data. However, only two items of data are noted for each taxon, the first (oldest) and last (youngest) occurrence (the FAD and LAD). These define a local range for each taxon. The objective is to define the local ranges for many taxa in at least three complete sections through the succession of interest. The more sections that are used, the more nearly these ranges will correspond to the total (true) ranges of the taxa. To compare the sections, a simple graphical method is used.

One particularly complete and well-sampled section is chosen as a **standard reference section**. Eventually, data from several other good sections are amalgamated with it to produce a **composite standard reference section**. A particularly thorough paleontologic study should be carried out on the standard reference section, as this enables later sections, for example, those produced by exploration drilling, to be correlated with it rapidly and accurately.

The graphic technique, which will be now described, is used both to amalgamate data for the production of the composite standard and for correlating the standard with new sections. Figure 7.20 shows a two-dimensional graph in which the thicknesses of two sections X and Y have been

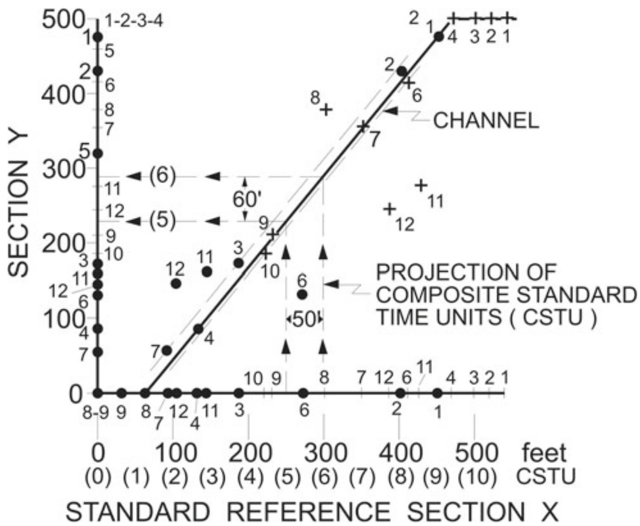


Fig. 7.20 This and the next three figures illustrate Shaw's (1964) graphic correlation method as discussed by Miller (1977). This plot shows the distribution of first occurrences (open circles) and last occurrences (crosses) in two sections and the positioning of the line of correlation. The channel is the zone on either side of the line of correlation encompassing observation error

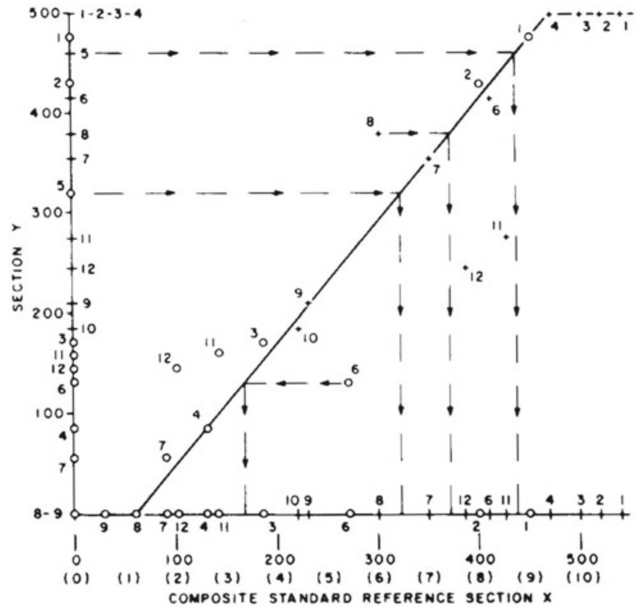


Fig. 7.21 The method used to compile a composite standard reference section. Data from new sections may be used to extend the range of occurrence of taxa that do not show their full range in the standard section (lowest occurrence of species 6, highest occurrence of species 8) and may also be used to transfer data on to the standard section, such as the range of species 5, which does not occur in the latter (Miller 1977)

marked of on the corresponding axes. The first occurrence of each taxon is marked by a circle on each section and the last occurrence by a cross. If the fossil taxon occurs in both sections, points can be drawn within the graph corresponding to the first and last occurrences by tracing lines perpendicular to the X and Y axes until they intersect. For example, the plot for the top of fossil 7 is the coincidence of points X = 350 and Y = 355.

If all the taxa occur over their total range in both sections and if sedimentation rates are constant (but not necessarily the same) in both sections, the points on the graph fall on a straight line, called the line of correlation. In most cases, however, there will be a scatter of points. The X section is chosen as the standard reference section, and ranges will presumably be more complete there. The line of correlation is then drawn so that it falls below most of the first occurrence points and above most of the last occurrence points. The first occurrence points to the left of the line indicate the late first appearance of the taxon in section Y. Those to the right of the line indicate the late first appearance in section X. If X is the composite standard, it can be corrected by using the occurrence in section Y to determine where the taxon should have first appeared in the standard. The procedure is shown in Fig. 7.21. Arrows from the first occurrence of fossil 6 show that in section X the corrected first appearance should be at 165 ft. The same arguments apply to the points for last appearances. Corrections of the kind carried out for fossil 6 (and also the last occurrence of fossil 8) in Fig. 7.21 enable refinements to be made to the reference section. Combining several sections in this manner is the method by which the composite standard is produced. Data points can also be introduced for fossils that do not occur in the reference section. In Fig. 7.21, arrows from the first and last occurrences of fossil 5 in section Y show that it should have occurred between 320 and 437 ft in section X.

If the average, long-term rate of sedimentation changes in one or other of the sections, the line of correlation will bend. If there is a hiatus (or a fault) in the new, untested sections (sections Y), the line will show a horizontal terrace. Obviously, the standard reference section should be chosen so as to avoid these problems as far as possible. Harper and Crowley (1985) pointed out that sedimentation rates are in fact never constant and that stratigraphic sections are full of gaps of varying lengths (we discuss this problem in Chap. 8). For this reason, they questioned the value of the graphic correlation method. However, Edwards (1985) responded that when due regard is paid to the scale of intraformational stratigraphic gaps versus the (usually) much coarser scale of biostratigraphic correlation, the presence of gaps is not of critical importance. Longer gaps of the scale that can be detected in biostratigraphic data (e.g., missing biozones) will give rise to obvious bends in the line of correlation, as noted previously.

The advantage of the graphic method is that once a reliable composite standard reference section has been drawn up it enables chronostratigraphic correlation to be determined between any point within it and the correct point on any comparison section. Correlation points may simply be read of the line off correlation. The range of error arising from such a correlation depends on the accuracy with which the line of correlation can be drawn. Hay and Southam (1978) recommended using linear regression techniques to determine the correlation line, but this approach assigns equal weight to all data points instead of using one standard section as a basis for a continuing process of improvement. But as Edwards (1984) noted, all data points do not necessarily have equal value; the judgment and experience of the biostratigraphers are essential in evaluating the input data. For this reason, statistical treatment of the data is inappropriate.

Figures 7.22 and 7.23 illustrate an example of the use of the graphic method in correlating an Upper Cretaceous succession in the Green River Basin, Wyoming, using palynological data (from Miller 1977). The composite standard reference section has been converted from thickness into **composite standard time units**, by dividing it up arbitrarily into units of equal thickness. As long as the rate of sedimentation in the reference section is constant, these time units will be of constant duration, although we cannot determine by this method alone what their duration is in years. Figure 7.22 shows the method for determining the position of selected time lines on each test section, and in Fig. 7.23 the time units are used as the basis for drawing correlation lines between four such sections. Note the unconformity in each illustration and the variation in sedimentation rates in Fig. 7.23.

The value of the graphic method for correlating sections with highly variable lithofacies and no marker beds is obvious, and it is perhaps surprising that the method is not more widely used. An important difference between this method and conventional zoning schemes is that zoning methods provide little more than an ordinal level of correlation (biozones, as expressed in the rock record, have a finite thickness which commonly cannot be further subdivided), whereas the graphic method provides interval data (the ability to make graduated subdivisions of relative time). Given appropriate ties to the global time frame the composite standard time units can be correlated to absolute ages in years and used to make precise interpolations of the age of any given horizon (such as a sequence boundary) between fossil occurrences and tie points (Gradstein et al. 2004b). The precision of these estimates is limited solely by the accuracy and precision obtainable during the correlation to the global standard. MacLeod and Keller (1991) provided excellent examples of this procedure, and their results suggest an obtainable precision of less than ±100 ka.

Constrained optimization (CONOP): Gradstein et al. (2004b) pointed out several disadvantages of the basic graphic correlation method. It relies on only a few sections, placing particular importance on a single section that becomes the basis for the composite standard. A superior, automated correlation method, called constrained optimization, has been used in the construction of several parts of the Phanerozoic time scale. This method automates graphic correlation so that multiple sections are compared and correlated simultaneously. In this way, gaps and changes in sedimentation rate in the initial standard section do not influence the outcome. The method is described by Kemple et al. (1995) and Sadler (1999) and critically evaluated by Smith et al. (2015).

Several of the time scales for the Phanerozoic periods described in detail in Gradstein et al. (2004a) were developed using these techniques. For example, the Ordovician scale in Gradstein et al. (2004b) made use of 669 graptolite taxa in 119 sections. In the recent study of Ordovician–Silurian graptolite biostratigraphy by Sadler et al. (2009),

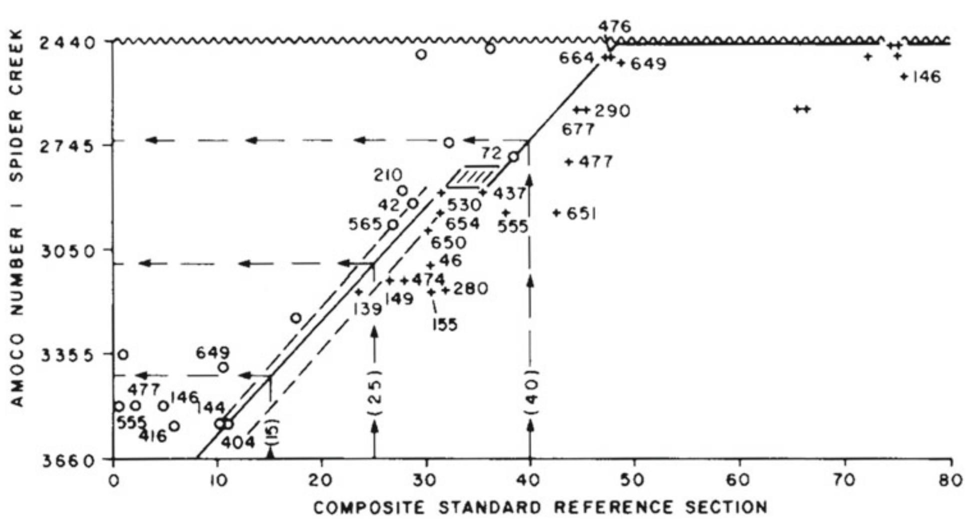


Fig. 7.22 An example of the use of the graphic method, showing a plot of data from one well against the composite standard. *Break* and *shaded areas* on the line of correlation are interpreted as an unconformity. The reference section has been divided into arbitrary thickness units (composite standard time units) (Miller 1977)

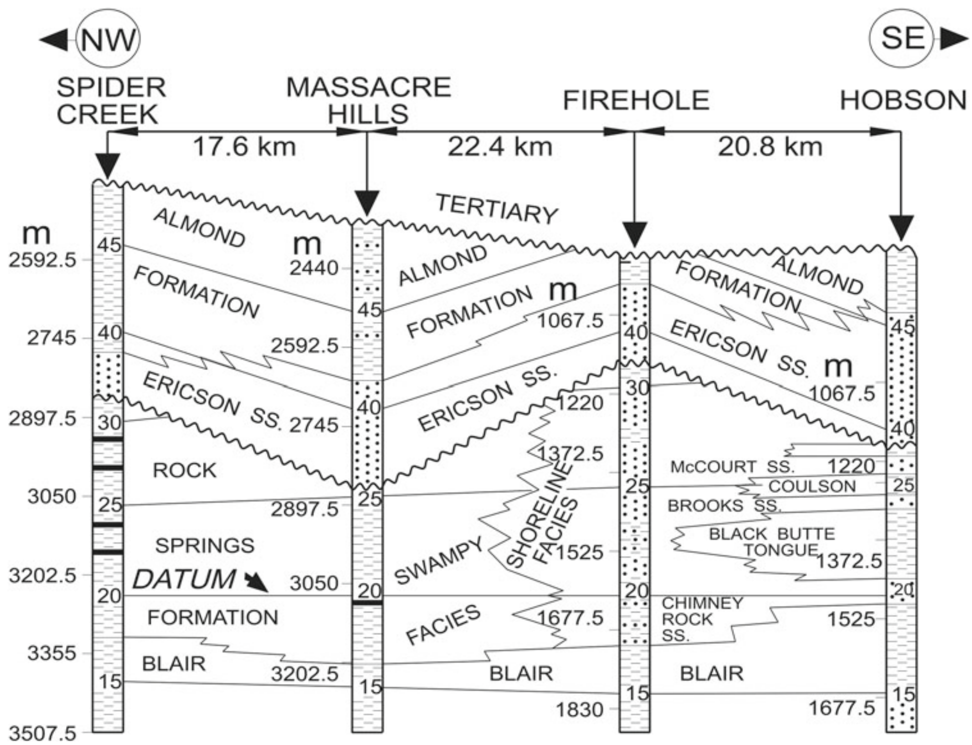


Fig. 7.23 Correlation of four wells in an area of marked lateral facies change using the composite standard time units from Fig. 7.22 (Miller 1977)

they noted (p. 887) that "The Graptolite zones vary widely in duration from as short as 0.1 m.y. to nearly 5.0 m.y. The mean duration of zones or zonal groupings calibrated here is 1.44 m.y. in the Ordovician and 0.91 m.y. in the Silurian." Current developments in the biostratigraphic basis of chronostratigraphy are discussed further in Sects. 8.9 and 8.10.

Calibration against the chemostratigraphic record. The global development of chemostratigraphic methods (Sect. 7.8.2) has provided an entirely new method for the quantitative calibration of biostratigraphic data. It is now a common practice among stratigraphers developing detailed time scales for specific basins and those working on the geological time scale to carry out chemostratigraphic sampling through the sections from which biostratigraphic data have been obtained. The increasing reliability of the oxygen, carbon and strontium isotopic global time scales has provided the basis for the numerical dating of biozones, which can be then used on their own, if necessary, as independent indicators of numerical age. For example, Cramer et al. (2010) employed $\delta^{13}C$ isotope chemostratigraphy to calibrate conodont and graptolite biozones across the Llandovery–Wenlock (early-middle Silurian) boundary in several sections in the Baltic area, with cross-referencing to sections in Britain and the Niagara area of New York state. They determined that the seven conodont and four graptolite zones investigated spanned about one million years and that the biozones each represented less than 500,000 years.

7.6 Unconformities and Unconformity-Bounded Units

The stratigraphic record is replete with sedimentary breaks ranging from the localized scours swept out by migrating bedforms, to the major angular unconformities that record significant orogenic events. As discussed in Chap. 5, unconformities have served as the primary basis for the subdivision of stratigraphic successions into sequences. In many basins there are hierarchies of sequences nested within each other and separated by sedimentary breaks of varying time significance (e.g., Figs. 5.30 and 5.31). Various terms have been erected in attempts to reflect the significance of different sedimentary breaks, in terms of the sedimentologic or structural discordance represented by the break (e.g., diastem, paraconformity etc.), but none of these have been rigorously defined, and their use is not recommended. Likewise, the expression "relatively conformable" has been used to define the nature of the successions that characterize the sequence contained between the unconformities, but as discussed in Sect. 5.4, this term, also, has not been defined and, as we now know from several decades worth of

Table 7.1 A classification of unconformities (Miall 2016)

SRS	Time scale (yrs)	Inst Sed Rate (m/ka)	Process	Description of break	Field characteristics of sedimentary break and/or of beds above and below
1–4	10^{-6}–10^{-1}	10^{4}–10^{6}	Bedform migration; diurnal to normal meteorological changes in runoff; tidal cycles	Local channel scours	Nesting of channels, macroforms and bedforms within a structure of minor bounding surfaces (ranks 1–5 of Miall, 1996)
5–7	10^{0}–10^{3}	10^{0}–10^{3}	Subtle tectonism, including in-plane stress	Migration and switching of depositional systems	Minor cut-and-fill erosion, early cementation, “unconformity paleosols” in nonmarine settings
			Autogenic seasonal to long-term geomorphic processes	Minor erosion, mature paleosols	Superimposition of delta and shelf-margin clinoform lobes separated by transgressive ravinement surfaces, rare preservation of falling-stage incised distributary channels, incised valleys;
			Rare extreme weather events	Marked facies change, minor regional erosion	Facies blanket, regional marker horizon
7-9	10^{4}–10^{5}	10^{-2}–10^{0}	High-frequency tectonism	Syndepositional unconformities	Strong but very localized angularity, coarse clastic wedges (“growth strata”)
			Regional response to flexural loading/unloading	Basin-wide low-angle unconformities	Low- to very-low-angle clinoform sets Evidence of fluvial or marine erosion, transgressive lag deposits at breaks
			Far-field intraplate stress changes	Tilting and warping of sequences and sequence sets	Widespread shifts in paleocurrent patterns, shoreline trends
			Orbital forcing	Continental (potentially global)—scale, nonangular break	Cyclothemic facies changes, potentially deep erosion of unconformity surface, coastal and shelf-margin clinoform onlap-offlap cycles
9–12	10^{6}–10^{7}	10^{-3}–10^{-1}	Orogenic tectonism	Regional angular unconformity	May be associated with deep erosional relief, clastic wedges
			Dynamic unconformities associated with basin formation	Onlap and offlap caused by basin subsidence	Onlap of extensional margins during flexural subsidence Onlap/offlap during motion of foreland-basin forebulge
			Dynamic topography	Sub-continental unconformity	Low-angularity (units above and below have similar dip). Commonly little field evidence of major time break
			Global eustasy	Global unconformity	Similar to above
			Long-term environmental change	Regional disconformities	Eolian supersurfaces, drowning unconformities (carbonates)

SRS = Sedimentation rate scale (from Miall 2015)

research on sequence stratigraphy, is misleading to the extent that all stratigraphic successions contain sedimentary breaks of varying temporal magnitude and physical extent (Table 7.1).

Table 7.1 provides a classification of sedimentary breaks based on their temporal significance and the mechanisms that generated them (from Miall 2016). They are conveniently classified into four broad classes on the basis of their temporal significance. Firstly, there are the minor breaks formed by bedform turbulence, representing minutes to seconds during normal bedform migration, and days to weeks as channels migrate laterally. These need careful examination, because subtle indicators, such as evidence of organic activity may indicate the passage of significant time (Davies and Shillito 2018). The second group represents breaks of up to a few hundred years duration caused by seasonal to long-term geomorphic processes, such as the migration and switching of depositional systems. The third group comprises the products of high-frequency tectonism and those caused by orbitally forced climate change, including glacioeustatic fluctuations in base level. These represent 10^4–10^5 years. Those caused by tectonism may be angular; those generated by glacioeustasy will be structurally conformable. Lastly, there are the angular unconformities formed by orogenesis, basin development and dynamic topography, and those formed by long-term eustasy, with time significance of 10^6–10^7 years.

Unconformities, typically appear in the rock record as apparently simple time planes, such as a stratal surface marked by contrasting lithologies having a significant difference in weathering characteristics. However, a great deal of information may be gained by a regional study of such surfaces (Miall 2016). The following are examples of "group 1" unconformities in Miall's (2016) classification, that is, those caused by regional tectonism on a 10^6–10^7-year time scale (Table 7.1). The continent-wide unconformity that defines the contact between the Canadian Shield and the Phanerozoic sedimentary record across North America is an ancient paleo-landscape surface, with remnants of fluvial drainage valleys and shorelines, remnant sea stacks and boulder beaches of late Precambrian age. Over much of western Australia, there is evidence of a regional Cretaceous surface. Erosional valleys with preserved regolith grade down into valleys occupied by lavas of Eocene age; Precambrian granite domes emerge as hills from beneath a remnant cover of Jurassic rocks that onlap their lower slopes (Twidale 1997). In the Rocky Mountain states a series of regional unconformities in the Triassic and Jurassic succession was recognized and named by Pipiringos and O'Sullivan (1978). These contain a great deal of information about the tectonic and paleogeographic evolution of the Mesozoic Western Interior Seaway. Zuchuat et al. (2019) carried out a detailed study of one of these surfaces, the J-3 Unconformity, separating the Middle Jurassic Entrada Sandstone from the Upper Jurassic Curtis Formation (and laterally equivalent units) in east-central Utah (USA). This is a laterally variable surface, generated by either erosion-related processes, such as eolian deflation, and water-induced erosion, or by deformational processes. The J-3 Unconformity is a composite surface formed by numerous processes that interacted and overlapped spatially and temporally.

Unconformities have commonly served as convenient boundaries for various types of stratigraphic units, particularly those based on lithostratigraphy. In many areas, regional unconformities have long been used to define natural subdivisions of the stratigraphic record (Blackwelder 1909). As originally defined by Sloss et al. (1949), sequences were defined as operational units separated by "marked discontinuities in the stratal record of the craton which may be traced and correlated for great distances on the objective bases of lithologic and faunal 'breaks' …." The use of unconformities as boundaries is now avoided in the definition of chronostratigraphic units, for reasons explained in Sect. 7.7. However, it is increasingly, being recognized that the stratigraphic record is subdivided into unconformity-bounded units of regional and possibly even global extent, caused by widespread changes in sea level or by regional tectonic or climatic events (Vail et al. 1977; Emery and Myers 1996; Miall 1995, 2010). As noted by Miall (2015), sequence stratigraphy "works" because there is a limited number of processes that operate at the 10^4–10^8-year time scale to create the sequence record and the unconformities that serve to subdivide it into convenient packages for description and interpretation. The methods of sequence stratigraphy have now become virtually universal (Catuneanu 2006).

Unconformity-bounded units include other types of units within them, such as biostratigraphic and lithostratigraphic units. They are not the same as these and are not necessarily equivalent to chronostratigraphic units, because the ages of the bounding unconformities may change from place to place. However, unconformity-bounded units have a certain chronostratigraphic significance because, with certain unusual exceptions, all the rocks below an unconformity are older than all of those above, and time lines do not cross unconformity surfaces. Some exceptions to these rules include (1) where a disconformity surface is caused by submarine erosion by a deep oceanic current that changes position with time, as a result of changing the configuration of the ocean basin (Christie-Blick et al. 1990), but this is not a problem that is likely to be encountered very frequently; (2) Ravinement surfaces are time-transgressive (Nummedal and Swift 1987); (3) Large-scale-incised valleys, caused by base-level fall at a coastline, may be significantly time-transgressive (Strong and Paola 2008); (4) The subaerial erosion surfaces that form the basis for most sequence subdivisions are typically highly complex and diachronous surfaces, as demonstrated by Holbrook and Bhattacharya (2012; see Fig. 5.8).

Proposals for the formal definition of unconformity-bounded units were given by the International Subcommission on Stratigraphic Classification (1987; see also Salvador 1994) and are also contained in the North American Commission on Stratigraphic Nomenclature (1983, 2005). The term **sequence**, as originally used by Sloss et al. (1949), was not recommended by these authorities, because this term had come to be used in a number of slightly but significantly different ways. In common geological parlance, the term *sequence* has commonly been used as a synonym for *succession,* a practice that needs to be discouraged because of the value now associated with the sequence method. The ISSC (1987) recommended the formal term **synthem**, a proposal followed by Salvador (1994), whereas the North American Stratigraphic Commission proposed the definition of **allostratigraphic units**, including alloformation, allogroup and allomember (NACSN 1983). The term synthem has not been accepted by the stratigraphic community. **Allostratigraphy**, as a formal system of definition and naming, has had modest success. The intent was that major unconformity-bounded units would be termed alloformations, with minor units nested within them, such as parasequences, labeled allomembers (NACSN 2005). The intent of this approach was to be purely descriptive, with no genetic connotations built into the terminology or methods.

Several groups of workers have made use of allostratigraphic methodology. For example, Autin (1992) subdivided the terraces and associated sediments in a Holocene fluvial floodplain succession into alloformations. R. G. Walker and his coworkers employed allostratigraphic terminology in their study of the sequence stratigraphy of part of the Alberta Basin, Canada. Their first definition of unconformity-bounded units is described in Plint et al. (1986), where the defining concepts were referred to as **event stratigraphy**, following the developments of ideas in this area by Einsele and Seilacher (1982). With the increasing realization that sequences and their bounding surfaces may be markedly diachronous, we no longer refer to sequences and sequence boundaries as "events." Walker (1990) discussed some of the practical problems in making use of sequence and allostratigraphic concepts. Explicit use of allostratigraphic terms appears in later papers by this group (e.g., Plint 1990). A text on facies analysis that built extensively on the work of this group recommends the use of allostratigraphic methods and terminology as a general approach to the study of stratigraphic sequences (Walker 1992). Martinsen et al. (1993) compared lithostratigraphic, allostratigraphic and sequence concepts as applied to a stratigraphic succession in Wyoming. As they were able to demonstrate, each method has its local advantages and disadvantages. Sequence stratigraphy is characterized by powerful, genetic concepts and interpretive methods, which provide it with a major advantage where appropriate. However, some workers, notably A. G. Plint, have found that the basic mapping of unconformities, and flooding surfaces is practical and efficient, and he continues to use allostratigraphic terminology in his work (see, for example, Shank and Plint 2013, and Fig. 6.4, in which correlations are focused on the mapping of surfaces of erosion and transgression; what they term E/T surfaces).

7.7 The Development of Formal Definitions for Sequence Stratigraphy

Allostratigraphy represented one of the first attempts to incorporate unconformity-bounded units into the formal framework of stratigraphy (see Sect. 1.2.9 for earlier ideas on this topic). Acceptance of its successor, sequence stratigraphy, by the "official" international community—the *International Commission on Stratigraphy*—has nearly been achieved. Controversies about how to define sequences have hindered the development of formal procedures.

One of the commonest complaints about sequence stratigraphy was that it is "model-driven." Catuneanu (2006, pp. 6–9) summarized the various approaches that have been taken to defining sequences and argued the case that the differences between the various models are not important, so long as sequences are described properly with reference to a selected standard model, with correct and appropriate recognition of systems tracts and bounding surfaces. His diagram comparing the various approaches is reproduced here as Fig. 7.24, and the suite of important surfaces that are used in sequence and systems-tract definition is shown in Fig. 7.25. The major difference between the sequence models is where different workers have chosen to place the sequence boundary. It should be noted that in each of the sequence definitions shown in Fig. 7.26, a similar set of systems tracts is shown in much the same relationship to each other. Exceptions include the T-R sequence, which makes use of a simplified definition of systems tracts, and such differences as that between the "late highstand" of depositional sequence III and the "falling stage" of depositional sequence IV.

There has been extensive discussion between the original proponents of the modern sequence models (P. R. Vail, H. Posamentier, J. Van Wagoner and their colleagues at Exxon) and others, regarding sequence definitions, centered on such characteristics as the facies shifts that take place within sequences and their significance with regard to the base-level cycle. This discussion has led to a number of different ways of defining sequences (Figs. 7.24, 7.25 and 7.26). In a masterly synthesis of the controversies, Catuneanu (2006) showed how to resolve these differences, and in three major collaborative publications (Catuneanu et al. 2009, 2010, 2011) discussed proposals for formal definitions that could be accepted by all workers (Fig. 7.26). As he demonstrated, "all approaches are correct under the specific circumstances for which they were proposed" (Catuneanu et al. 2011, pp. 232–233). Additional details, including descriptions and illustrations of the key bounding surfaces that develop on continental shelves, are described by Zecchin and Catuneanu (2013).

In the original Exxon model (Vail et al. 1977) the sequence boundary (commonly abbreviated as SB on diagrams) was drawn at the subaerial unconformity surface, following the precedent set by Sloss (1963), an approach that readily permits the sequence framework to be incorporated into an allostratigraphic terminology, at least for coastal deposits, where the subaerial erosion surface is readily mapped. Offshore may be a different story. The first sequence model (Vail et al. 1977) did not recognize the falling-stage systems tract. The highstand of one sequence was followed directly by the lowstand of the next sequence, with the sequence boundary falling

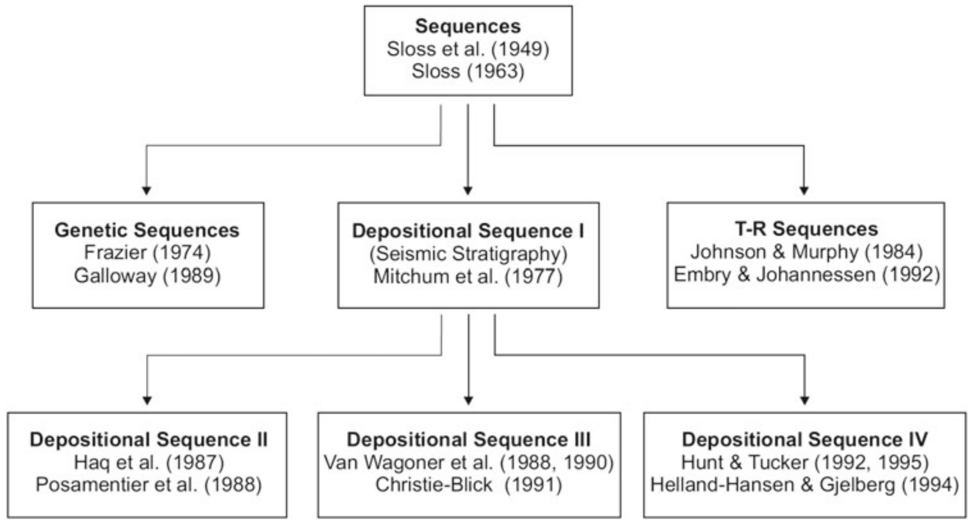


Fig. 7.24 The evolution of sequence definitions. T-R = transgressive–regressive. From Catuneanu et al. (2011)

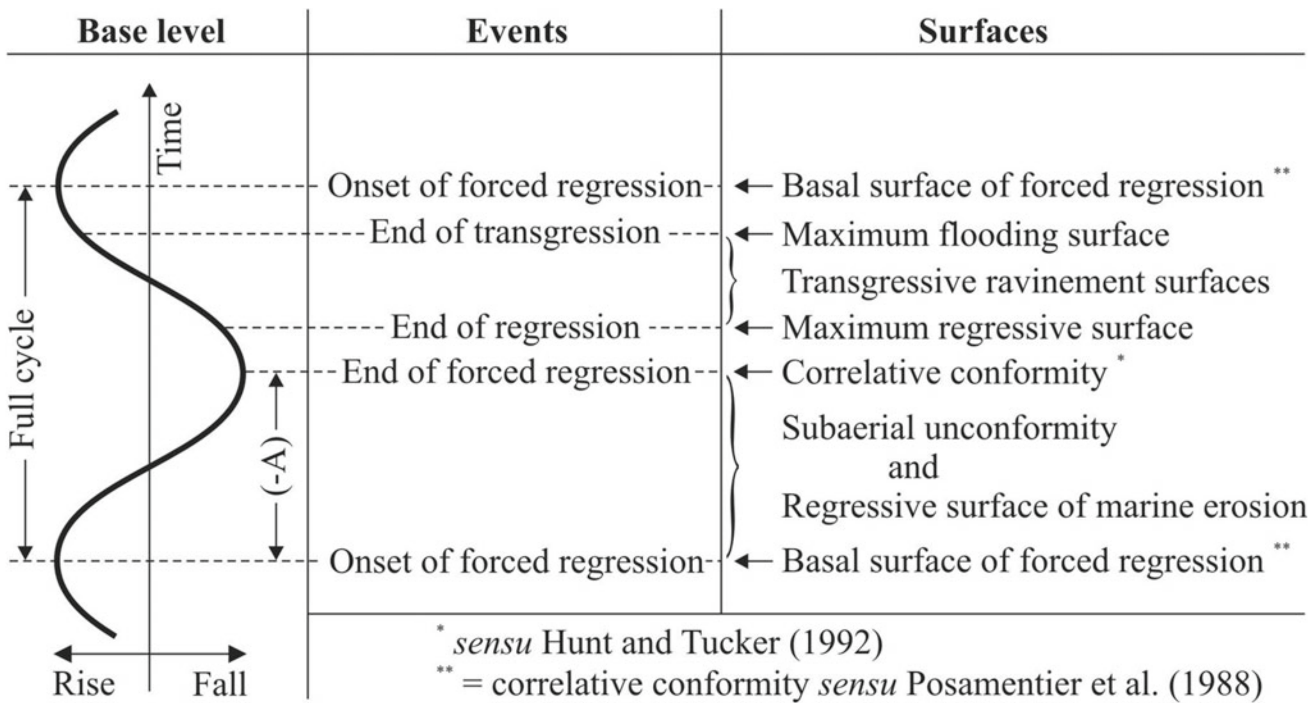


Fig. 7.25 Stratigraphic surfaces used in the definition of sequences and systems tracts, and their timing, relative to the cycle of base-level change (Catuneanu 2006, Fig. 4.7)

between the two systems tracts. Plint (1988) was the first to recognize the importance of the process of **forced regression** that generated by a fall in base level, and that included the formation of "sharp-based sandstone bodies" and the **regressive surface of marine erosion** (Fig. 5.14). The forced regressive deposits could be assigned either to a "late highstand" or an "early lowstand." Based on assumptions about the changing rate of sea-level fall, the sequence boundary—the coastal equivalent of the subaerial erosion surface—was initially placed at the **basal surface of forced regression** (assumed commencement of forced regression). This placement of the sequence boundary is the basis for what Catuneanu (2006) refers to as "depositional sequence II" (Fig. 7.26). The problem with this definition is addressed below. In addition, the early Exxon work defined several different types of sequence-bounding unconformity. Vail and Todd (1981) recognized three types, but later work (e.g., Van Wagoner et al. 1987) simplified this into two, termed type-1 and type-2 unconformities, based on assumptions about the

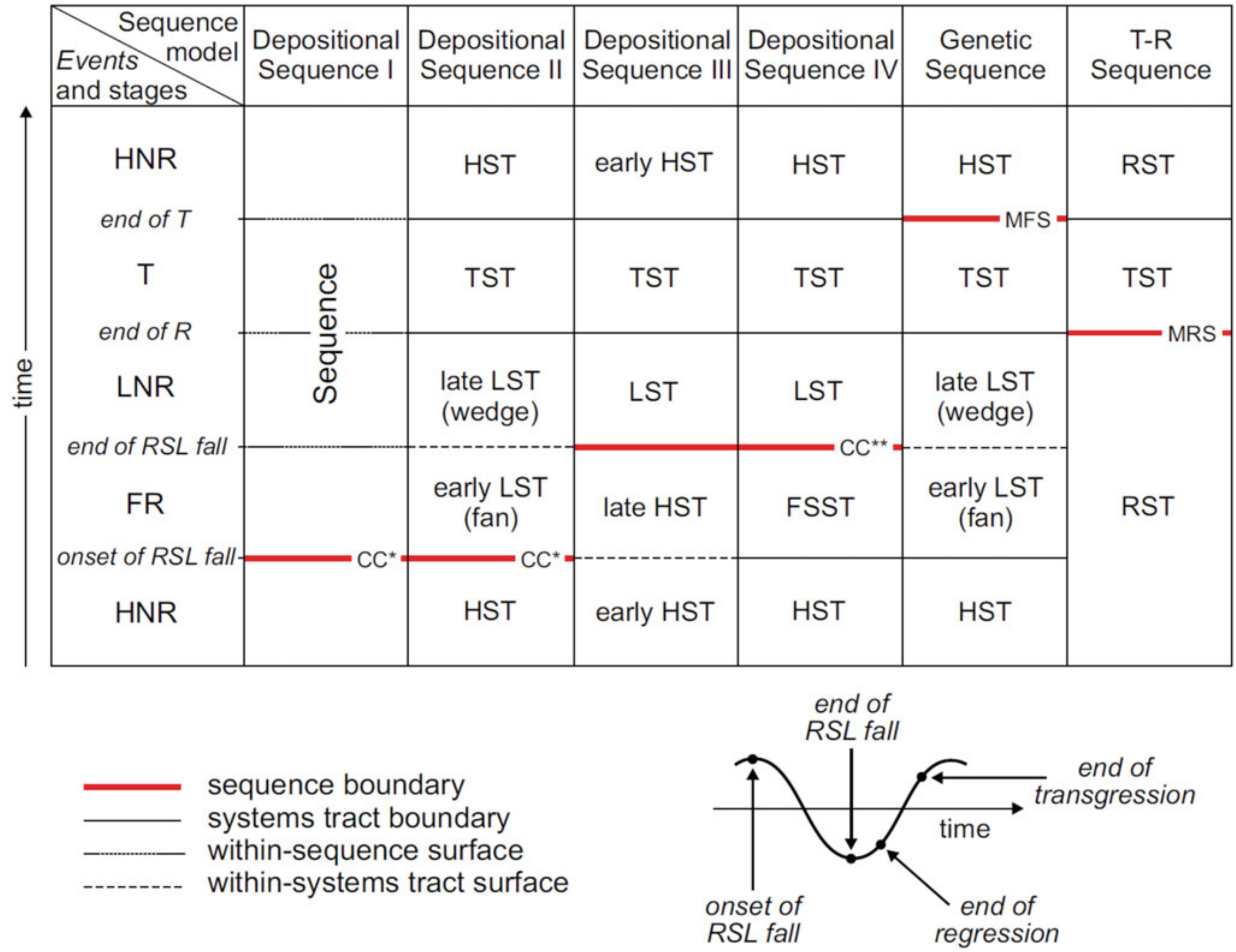


Fig. 7.26 Nomenclature of systems tracts, and timing of sequence boundaries for the various sequence stratigraphic approaches. Abbreviations: RSL—relative sea level; T—transgression; R—regression; FR—forced regression; LNR—lowstand normal regression; HNR—highstand normal regression; LST—lowstand systems tract; TST—transgressive systems tract; HST—highstand systems tract; FSST—falling-stage systems tract; RST—regressive systems tract; T-R—transgressive–regressive; CC*—correlative conformity in the sense of Posamentier and Allen (1999); CC**—correlative conformity in the sense of Hunt and Tucker (1992); MFS—maximum flooding surface; MRS—maximum regressive surface (from Catuneanu et al. 2011)

rate of change of sea level and how this was reflected in the sequence architecture. In the rock record, they would be differentiated on the basis of the extent of subaerial erosion and the amount of seaward shift of facies belts. However, Catuneanu (2006, p. 167) pointed out the long-standing confusions associated with these definitions and recommended that they be abandoned. These "types" are not discussed further in this book. Schlager (2005, p. 121) recommended the separate recognition of the third type of sequence boundary, the **drowning unconformity**, which forms "when sea level rises faster than the system can aggrade, such that a transgressive systems tract directly overlies the preceding highstand tract often with a significant marine hiatus. … Marine erosion frequently accentuates this sequence boundary, particularly on drowned carbonate platforms."

Hunt and Tucker (1992) were among the first (since Barrell!) to point out that during sea-level fall, subaerial erosion continues until the time of sea-level lowstand, with the continuing transfer of sediment through clastic delivery systems to the shelf, slope and basin, and with continuing downcutting of the **subaerial erosion surface** throughout this phase. The age of the subaerial erosion surface, therefore, spans the time up to the end of the phase of sea-level fall, a time substantially later than the time of initiation of forced regression. The use of the basal surface of forced regression as a sequence boundary, as in "depositional sequence II" is, therefore, not an ideal surface at which to define the sequence boundary, although, as Catuneanu (2006) reports, it is commonly a prominent surface on seismic-reflection lines. In fact, as Embry (1995) pointed out (see Fig. 7.29), there is no through-going surface associated with forced regression that can be used to extend the subaerial erosion surface offshore for the purpose of defining a sequence boundary. He argued that "from my experience I have found that the most suitable stratigraphic surface for the conformable expression of a sequence boundary is the transgressive surface" (Embry, 1995, p. 4). This meets his criterion—one with which all stratigraphers would agree—that "one of the main purposes of sequence definition [is] a coherent genetic unit without significant internal breaks" (Embry 1995, p. 2). His preferred definition of sequences, the T-R sequence, places the sequence boundary at the TR surface, at the end of the phase of regression and the time of initial transgression (Figs. 7.28, 7.29). There is, of course, a delay in time between the end of downcutting of the subaerial erosion surface during the falling stage, and the flooding of the same surface during transgression. The results of the two processes may coincide in the rocks, which is why this surface may provide a good stratigraphic marker, but it is important to remember that the surface is not a time marker and represents a time gap, with the gap decreasing in duration basinward.

Highlighting the timing of development of the subaerial erosion surface by Hunt and Tucker (1992) also served to highlight the inconsistency of assigning the main succession of submarine fan deposits on the basin floor to the lowstand systems tract, as shown by the Vail et al. (1977) and Posamentier et al. (1988) models, in which these deposits are shown resting on the sequence boundary. Notwithstanding the discussion of the Hunt and Tucker (1992) paper by Kolla et al. (1995), who defended the original Exxon models, this is an inconsistency that required a redefinition of the standard sequence boundary. It is now recognized that in the deep offshore, within submarine-fan deposits formed from sediment delivered to the basin floor during a falling stage (which is often the most active interval of sediment delivery to the continental margin; see Fig. 5.15), there may be no sharp definition of the end of the falling stage nor of the turn-around and subsequent beginning of the next cycle of sea-level rise and, therefore, no distinct surface at which to draw the sequence boundary. The boundary here would be a **correlative conformity**, and may be very difficult to define in practice.

An alternative sequence model, termed the **genetic stratigraphic sequence** (Figs. 7.26, 7.28), was defined by Galloway (1989), building on the work of Frazier (1974). Although Galloway stressed supposed philosophical differences between his model and the Exxon model, in practice, the difference between them is simply one of where to define the sequence boundaries. The Exxon model places emphasis on subaerial unconformities, but Galloway (1989) pointed out that under some circumstances, unconformities may be poorly defined or absent and, in any case, are not always easy to recognize and map. For example, in all but the largest outcrops a fluvial channel scour surface may look exactly like a regional subaerial erosion surface. Galloway's (1989) preference was to draw the sequence boundaries at the **maximum flooding surface**, which corresponds to the highstand downlap surfaces. He claimed that these surfaces are more prominent in the stratigraphic record, and therefore more readily mappable.

Galloway's proposal has not met with general acceptance. For example, Walker (1992) disputed one of Galloway's main contentions, that "because shelf deposits are derived from reworked transgressed or contemporary retrogradational deposits, their distribution commonly reflects the paleogeography of the precursor depositional episode." Galloway (1989) went on to state that "these deposits are best included in and mapped as a facies element of the underlying genetic stratigraphic sequence." However, as Walker (1992) pointed out, most sedimentological parameters, including depth of water, waves, tides, basin geometry, salinity, rates of sediment supply and grain size, change when an unconformity or a maximum flooding surface is crossed. From the point of view of genetic linkage, therefore, the only sedimentologically related packages lie (1) between a subaerial unconformity and a maximum flooding surface, (2) between a maximum flooding surface and the next younger unconformity or (3) between a subaerial erosion surface and the overlying unconformity (an incised valley fill)(Walker 1992, p. 11).

However, some workers have found Galloway's use of the maximum flooding surface much more convenient for sequence mapping, for practical reasons. For example, it may yield a prominent gamma ray spike in wireline logs (Underhill and Partington 1993), or it may correspond to widespread and distinctive goniatite bands (Martinsen 1993), or it may provide a more readily traceable marker, in contrast to the surface at the base of the lowstand systems tract, which may have irregular topography and may be hard to distinguish from other channel-scour surfaces (Gibling and Bird 1994). In nonmarine sections it may be hard to find the paleosol on interfluves that correlates with the sequence-bounding channel-scour surface (Martinesen 1993). In some studies (e.g., Plint et al. 1986; Bhattacharya 1993) it has been found that ravinement erosion during transgression has removed the transgressive systems tract so that the marine flooding surface coincides with the sequence boundary.

Catuneanu et al. (2011, p. 183) pointed out that however the sequences are defined, there may be unconformities contained within them. For example, Galloway's (1989) genetic stratigraphic sequences are explicitly defined with the sequence boundary at the maximum flooding surface, not the subaerial erosion surface. Therefore, sequences can no longer be defined as "unconformity-bounded." Given the need to encompass all types of sequence model, Catuneanu et al. (2009, p. 19) proposed redefining a sequence as "a succession of strata deposited during a full cycle of change in accommodation or sediment supply." As discussed in Sect. 5.4, sequences commonly occur in nested hierarchies, with thinner sequences of shorter duration nested within larger sequences that represent longer time spans. The boundaries between all of these sequences are unconformities of varying time significance (Figs. 5.29–5.31).

As described in Chaps. 3 and 4, many stratigraphic successions contain small-scale cycles nested within a sequence. Van Wagoner et al. (1988, 1990) erected the term **parasequence** to encompass the shoaling-upward successions, capped by flooding surfaces, that are common in coastal clastic successions. The term was proposed originally as part of a hierarchy of terms, the bed, bedset, parasequence, parasequence set and sequence. Prograding delta lobes, regressing clastic shorelines and peritidal carbonate cycles are examples of parasequences that are particularly common in the geological record. Catuneanu et al. (2009, p. 19) noted that the term has also been used for cyclic deposits in some fluvial and deep-marine deposits, where the concept of "flooding surface" is irrelevant. A particular source of confusion comes from the incorporation of the word "sequence" within the term parasequence. The nomenclature problem was not improved by the usage employed by Mitchum and Van Wagoner (1991), who equated parasequences with "4th-order paracycles." Sequences are allogenic products of regional controls, whereas parasequences may be a product of autogenic processes, such as delta lobe switching (Fig. 7.27). This has been demonstrated to be the case in the example of the Dunvegan delta illustrated in Fig. 5.11 (Bhattacharya 1991). The shingles and their bounding flooding surfaces are therefore local in distribution, and their development has little, if anything, to do with the allogenic mechanisms that generate sequences. However, to apply to these successions a term that contains the word "sequence" in it is inevitably to introduce the implication that they are allogenic in origin and constitute regionally correlatable units. The correct interpretation clearly depends on good mapping to determine the extent and correlatability of each shingle, and it would seem advisable not to use a term in a descriptive sense that carries genetic implications. Yet Mitchum and Van Wagoner (1991) illustrated parasequences/paracycles with a strike-oriented cross-section 400 km long, which claims to show correlations between several separate depositional systems. This, and the implication that parasequences are the same thing as minor sequences, added to the confusion. Furthermore, some high-frequency sequences formed by Milankovitch processes are comparable in thickness and outcrop appearance. It would be this author's preference that the term parasequence be abandoned entirely, but Catuneanu et al. (2009, 2011) (papers to which this author contributed!) recommended that the term may continue to be used if it is restricted to its original definition. An additional discussion of the term parasequence is provided in Sect. 5.4.

The selection of which sequence definition or model to use in any given stratigraphic setting is a matter of choice. The original Sloss/Vail **depositional sequence** model (Sloss 1963; Vail et al. 1977) and the **T-R model** of Embry and Johannessen (1992) use the subaerial erosion surface as the sequence boundary (Figs. 7.28, 7.29). The sequence then represents a full cycle of increasing and decreasing accommodation preserved between two erosion surfaces. The Galloway (1989) **genetic stratigraphic sequence** model uses the maximum flooding surface (MFS) as the sequence boundary (Figs. 7.28, 7.29). In many stratigraphic settings, recognizing and correlating a maximum flooding surface (MFS) is much more easily accomplished than mapping the subaerial erosion surface. The MFS is commonly represented by a marine shale or a condensed section which, because it is deposited at a time of maximum transgression, is typically widespread and forms a distinctive marker bed between packages of coarser clastic or carbonate/evaporite facies. Subaerial erosion surfaces may be characterized by significant erosional relief, which may make them difficult to trace within suites of wireline logs. Furthermore, where this surface occurs within successions of nonmarine strata it can be very difficult to distinguish the sequence boundary from local fluvial channel scour surfaces (see Fig. 5.8). However, these differences in mappability do not need to be reflected in the choice of sequence model.

The TR sequence model uses the subaerial erosion surface as the sequence boundary for the nonmarine and near-shore portions of a sequence but differs from the depositional sequence where sequences are traced into the offshore.

In a nonmarine to coastal to marine section, the maximum seaward extent of the subaerial erosion surface depends on the amplitude of relative sea-level change. Traced far enough seaward there will be a point beyond which water depths are such that exposure and erosion do not take place during base-level fall. Correlating sequence boundaries and systems tracts into the offshore may be difficult. The erosion surface generated by forced regression (Fig. 5.14; RSE in Fig. 7.29) defines the base of the falling-stage systems tract but, as demonstrated by Plint (1988), this surface is not at a

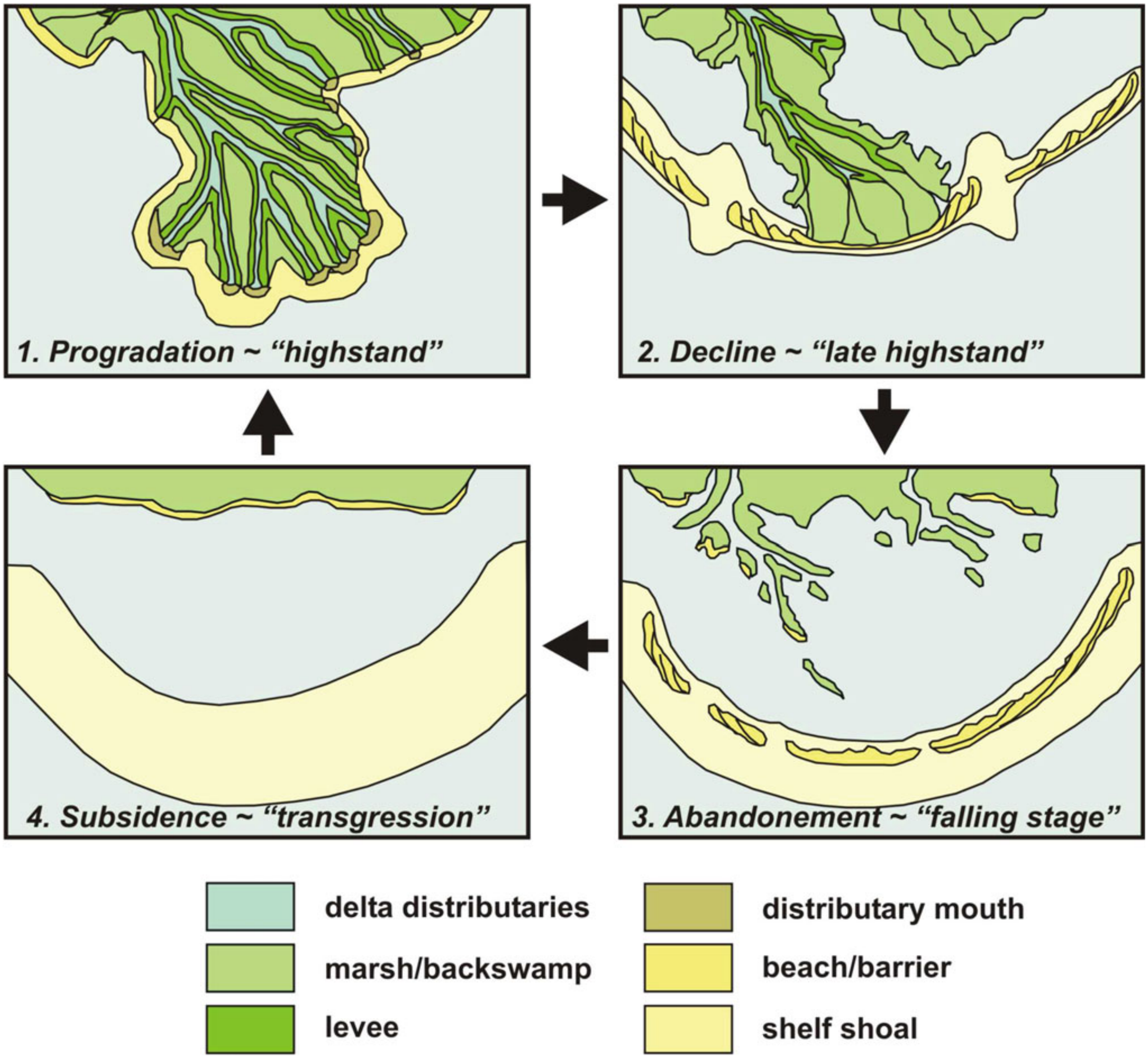


Fig. 7.27 The development and abandonment of delta lobes in a river-dominated, Mississippi-type delta (e.g., see Fig. 4.21), based on a detailed analysis of the Mississippi delta system by Boyd and Penland (1988). In stage 1, progradation develops an upward-shoaling deltaic succession. Abandonment, followed by subsidence (resulting from compaction) cause the upper layers of the succession to be reworked (stage 2), resulting in the development of an extensive barrier island system (stage 3). Finally, the deposit undergoes transgression and is covered by marine shale (stage 4). Repetition of this succession of events when a new delta lobe progrades back over the older deposit results in shoaling upward successions bounded by transgressive flooding surfaces, that is, parasequences. In this case, however, they are clearly of autogenic origin. Systems-tract designations for each of the four stages are indicated in parenthesis

stratigraphically consistent level, but may recur at stratigraphically higher positions in the offshore direction as base level continues to fall. The offshore limit of the RSE records the end of the phase of base-level fall, and this point is used to define the sequence boundary in **depositional sequences III** and **IV** (Fig. 7.29). Regressive sedimentation is likely to continue until the rate of rising accommodation during the subsequent cycle of base-level rise equals the rate of sediment supply. At this time, regression ends, transgression begins, and this point is used by Embry (1995) and Embry and Johannessen (1992) to define the sequence boundary for **TR cycles**. The transgressive surface which then develops marks the beginning of the next cycle of base-level rise (TS in Fig. 7.29). In many shelf settings, the TS is marked by lag deposits or a condensed section and is readily recognizable in cores and on wireline logs. It passes landward into the

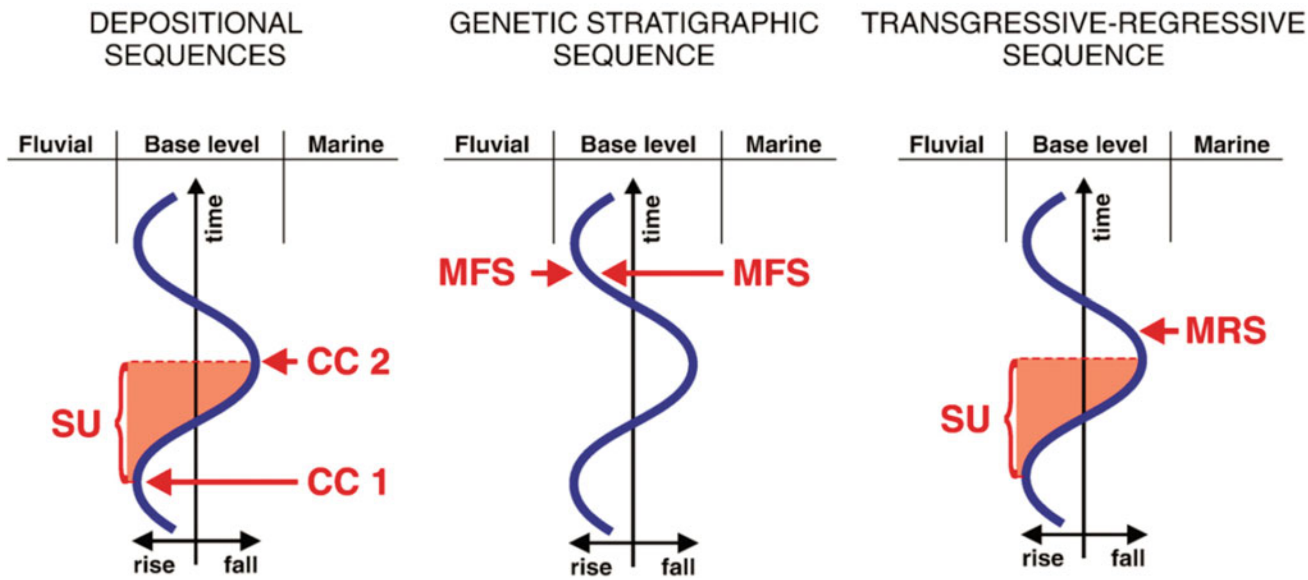


Fig. 7.28 Selection of sequence boundaries according to the "depositional," "genetic stratigraphic" and "transgressive–regressive" sequence models. The choice of sequence boundary is less important than the correct identification of all sequence stratigraphic surfaces in a succession (Fig. 7.10). Abbreviations: SU—subaerial unconformity; CC 1—correlative conformity *sensu* Posamentier and Allen (1999); CC 2—correlative conformity *sensu* Hunt and Tucker (1992); MFS—maximum flooding surface; MRS—maximum regressive surface. The subaerial unconformity is a stage-significant surface, whereas all other surfaces shown in this diagram are event-significant (Catuneanu et al. 2009, Fig. 24, p. 17)

ravinement surface (R in Fig. 7.29), a diachronous surface formed by wave erosion during transgression (Fig. 5.17).

In earlier definitions of the depositional sequence (**depositional sequences I** and **II** in Fig. 7.26), the seaward correlation of the subaerial erosion surface was equated with the beginning of base-level fall and this surface, and its seaward extension as correlative conformity (CC1 in Fig. 7.28; this surface is also shown in Fig. 7.29) was used to demarcate the sequence boundary. The inconsistencies in this approach were pointed out by Hunt and Tucker (1992). Subaerial erosion continues throughout the falling stage of the base-level cycle, and so the erosion surface that is subsequently preserved by being transgressed and buried during the succeeding base-level rise will correspond in age to the end of the falling stage, not the beginning. The first regressive surface marking the beginning of the falling stage (e.g., as in Fig. 5.14) may be followed by others, further seaward and stratigraphically younger, as the falling stage proceeds. Another of the inconsistencies of the early sequence definitions was the assignment of offshore submarine fan deposits primarily to the "lowstand wedge" formed (according to this earlier interpretation) during the beginning of the next cycle of base-level rise. Given that much of the sediment supply for submarine fans come from subaerial erosion during falling base level (Fig. 5.15), this does not seem plausible.

Embry and Johannessen (1992) claimed that the TR sequence definition is the only one that adheres to objective criteria, meaning the designation of the sequence boundary at a mappable surface. This is correct in that, as shown in Fig. 7.29, the surfaces marking the beginning and end of a cycle of base-level change are not necessarily associated with any recognizable facies change in the rocks, meaning that their recognition and mapping in surface outcrops or in subsurface cores, logs or seismic data, requires interpolation and extrapolation. The surface marking the end of regression is a lithologically mappable surface, but it will be directly mappable only so far seaward as regressive deposits extend. Beyond that point, a correlation may be possible by the tracing of reflections in seismic data, but recognition of the surface of maximum regression in cores or logs through deep marine deposits is not likely to be possible, so this particular feature of the TR definition is not significantly better than that of the depositional sequence.

A generalized sequence model is shown in Fig. 7.30 (see also Fig. 5.11). In this model the base-of-slope submarine fan and other deep-water deposits are assigned to the falling stage systems tract, contrary to the first sequence definitions (depositional sequences I and II). The sequence boundary (timeline 4) includes an incised valley, which is filled by the lowstand systems tract and during transgression (timelines 5–7). Commonly, incised valleys are filled toward the end of

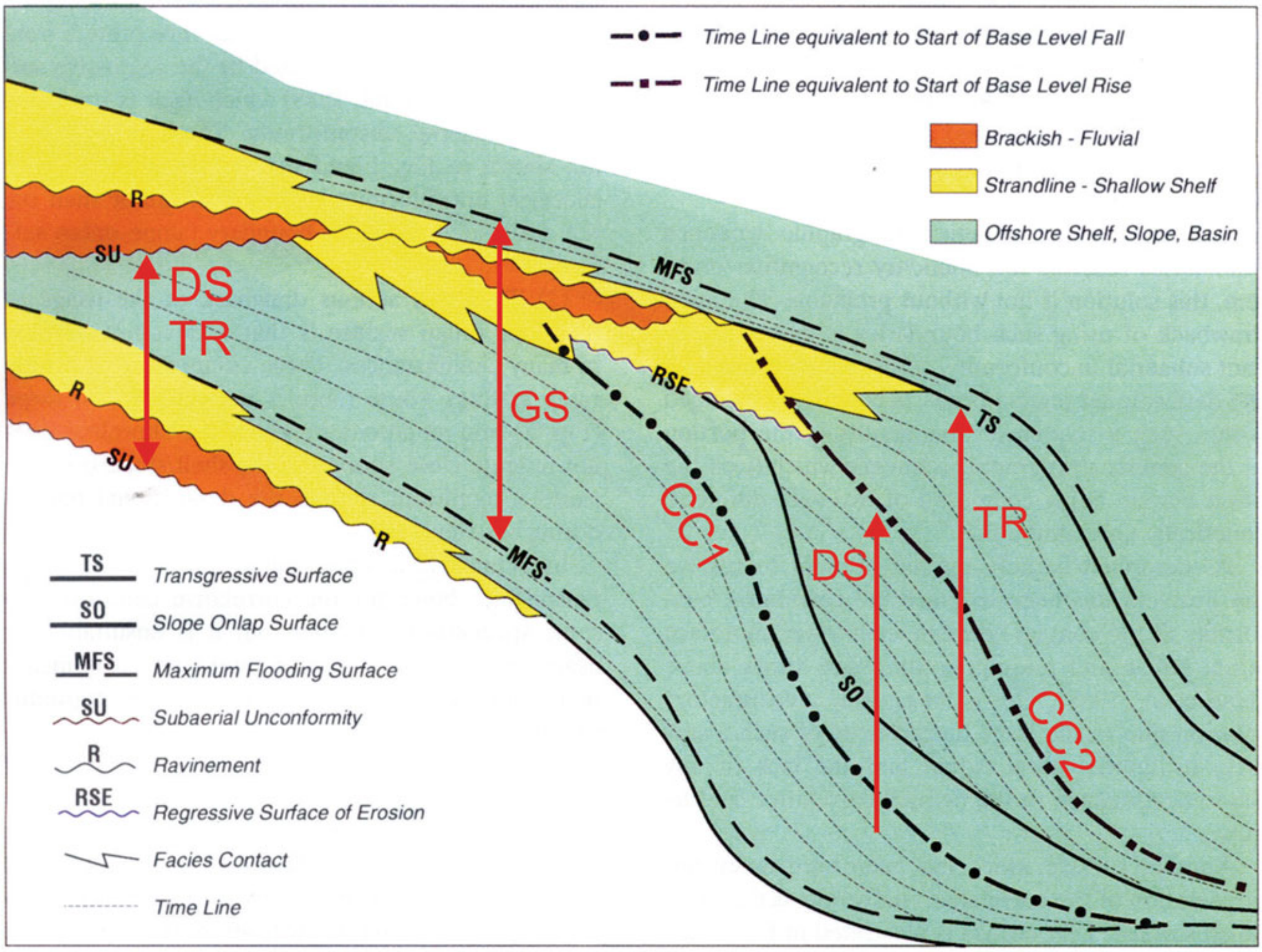


Fig. 7.29 Schematic cross-section through an ideal continental-margin sequence, showing the relationships between the major surfaces. Red arrows and letters denote three alternative ways to define sequences. In coastal and nearshore deposits, the sequence boundary for depositional sequences (DS) and TR sequences (TR) is defined by the subaerial erosion surface (SU). Correlating this surface offshore may be difficult in the absence of high-quality seismic-reflection data. Genetic stratigraphic sequences (GS) are defined by the maximum flooding surfaces. Depositional sequences III and IV use the correlative conformity corresponding to the end of base-level fall as the sequence boundary (CC** of Fig. 7.26, CC2 of Fig. 7.28 and this figure). This may be indicated by a change from erosional to aggradational deposition in the forced-regressive deposits. The time of initiation of base-level fall corresponds to correlative conformity CC* of Fig. 7.26 and CC1 of Fig. 7.28 and this figure, and defines the contact between early and late highstand in depositional sequence III. The top of a DS is indicated here by the arrow labeled DS that terminates at the surface marking the beginning of base-level rise. TR sequences are defined offshore by the end of regression. In this diagram the wedge of regressive deposits is truncated by the transgressive surface, which here marks the top of a TR sequence. In depositional sequence terminology, this latest wedge of regressive deposits is classified as the lowstand systems tract (adapted from Embry 1995, Fig. 1)

the transgressive phase, with deposition extending beyond the margins of the valley. Elsewhere on the continental margin transgressive deposits are characteristically thin and may consist of condensed deposits or a coarse lag. Hardgrounds and extensive bioturbation are common.

A significant flaw of the TR model is that the surface of maximum regression may not correlate with the subaerial erosion surface. As Embry and Johannessen (1992) noted, regression may continue in time beyond the end of the cycle of base-level fall, which corresponds to the age of the subaerial erosion surface. The deposits that form after this time are assigned to the lowstand systems tract and may comprise a substantial thickness of regressive deposits, accumulated on the shelf, slope and deep basin during the beginning of the next cycle of transgression. Two seismic lines are shown here which contain this feature (Figs. 7.31, 7.32). The lowstand normal regressive deposits (LNR) in Fig. 7.31 are the deposits that form after the end of base-level fall—

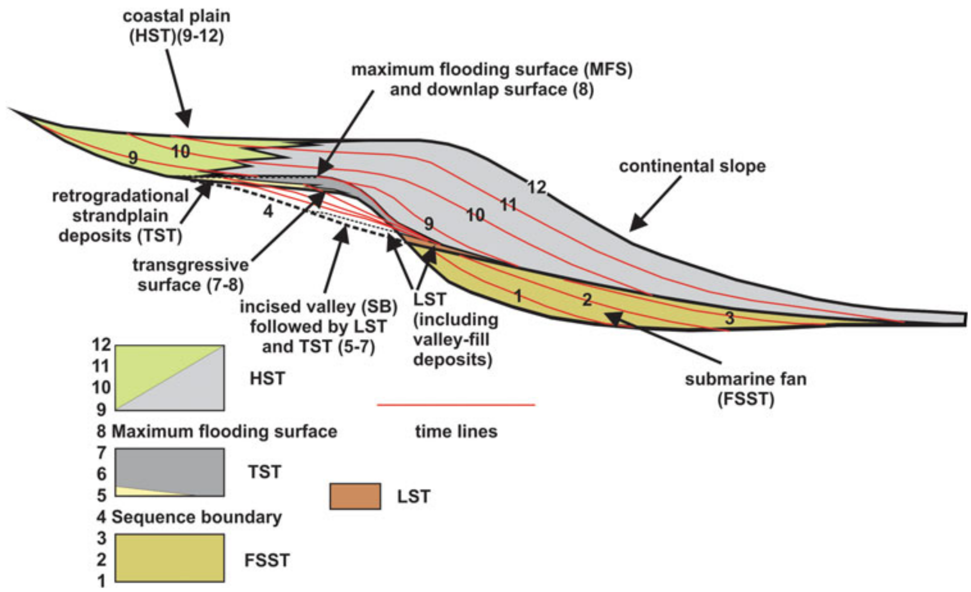


Fig. 7.30 A generalized sequence model. Twelve time lines reveal the internal architecture and evolution of the sequence and its component systems tracts

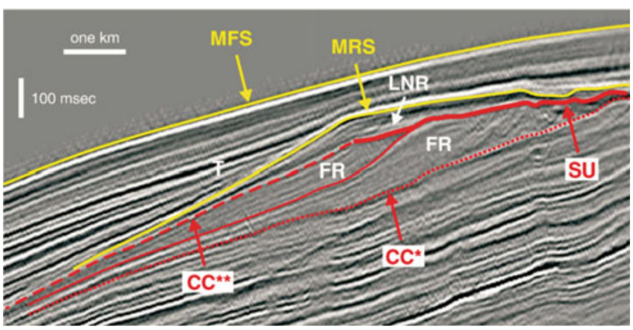


Fig. 7.31 Seismic line in the Gulf of Mexico showing different genetic types of deposits (forced regressive, normal regressive, transgressive) and stratigraphic surfaces that may serve as sequence boundaries according to different sequence stratigraphic models (modified from Posamentier and Kolla, 2003). Abbreviations: FR—forced regressive; LNR—lowstand normal regressive; T—transgressive; SU—subaerial unconformity; CC*—correlative conformity sensu Posamentier and Allen, 1999 (= basal surface of forced regression); CC**—correlative conformity sensu Hunt and Tucker 1992; MRS—maximum regressive surface; MFS—maximum flooding surface. The line displays the typical stacking patterns and stratal terminations associated with forced regression (offlap, downlap, toplap, truncation), normal regression (downlap, topset), and transgression (onlap) (Catuneanu et al. 2009, Fig. 7, p. 6)

indicated by the correlative conformity (CC**) extending basinward from the subaerial erosion surface—and are capped by the **maximum regressive surface (MRS)**. A particularly thick succession of shelf and slope deposits is assigned to the LNR in Fig. 7.32. This section illustrates the migration through time of the shelf margin and the shoreline. In this example a high sediment supply was maintained throughout transgression, resulting in a thick transgressive shelf section. This is commonly not the case, with transgression commonly recorded as a lag deposit or a condensed section.

7.8 Chronostratigraphy and Geochronometry

7.8.1 The Emergence of Modern Methods

Geochronometry is the study of the continuum of geologic time. Geologic events and rock units may be fixed within this time frame by a variety of methods, of which radioisotopic dating is the most direct.

Chronostratigraphy is the study of the standard stratigraphic scale, comprising the familiar eras, periods and ages (e.g., Paleozoic, Triassic, Campanian) (Table 7.2).

For the Phanerozoic, biostratigraphy is the main basis of the chronostratigraphic method, but radioisotopic methods

Table 7.2 The conventional hierarchy of formal chronostratigraphic and geochronologic terms

Chronostratigraphic	Geochronologic
Eonothem	Eon
Erathem	Era
System	period
Series	Epoch
Stage	Age
Substage	Subage

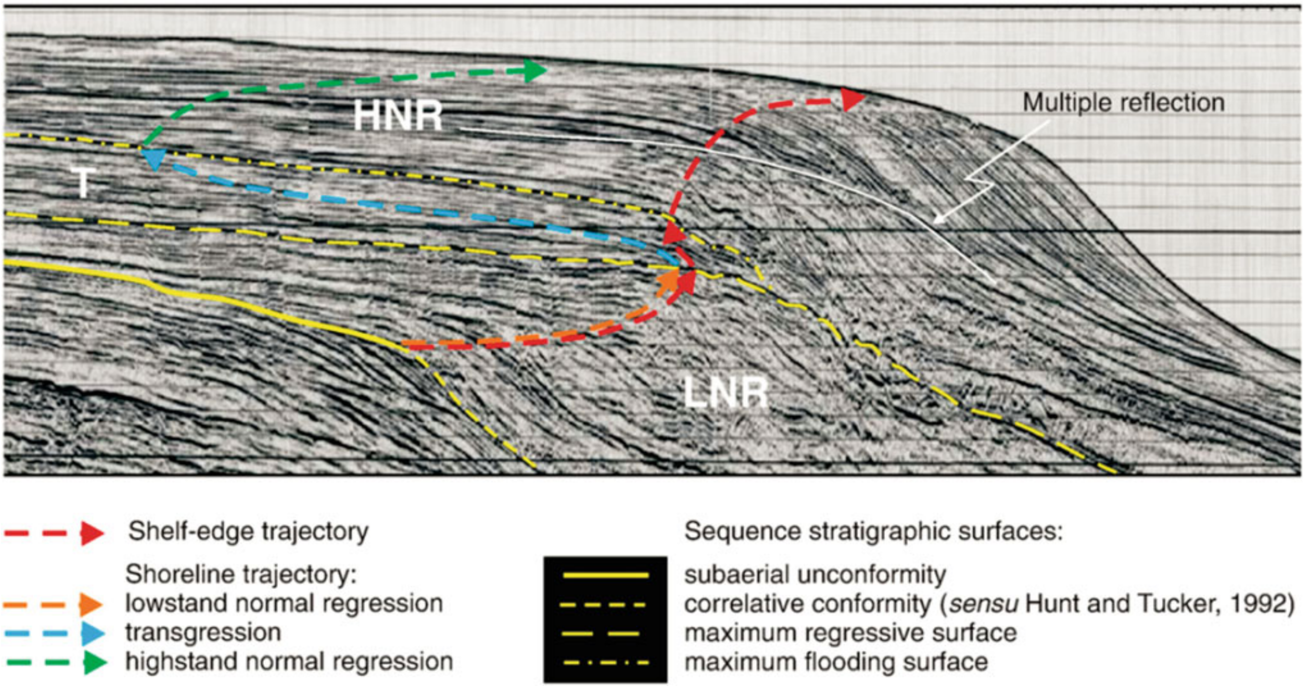


Fig. 7.32 Dip-oriented regional seismic profile from the Pelotas Basin, southern Brazil (modified from Abreu 1998), showing large-scale (high-rank) lowstand normal regressive (LNR), transgressive (T) and highstand normal regressive (HNR) systems tracts. Lower rank sequences are nested within these higher rank systems tracts. The transgressive systems tract thickens landward, which reflects the direction of shift of the depocenter. Individual backstepping parasequences are difficult to observe within the transgressive systems tract due to the limitation imposed by vertical seismic resolution. The shoreline trajectory and the shelf-edge trajectory may coincide during lowstand normal regression but are separate during transgression and highstand normal regression. The change in depositional trends from dominantly progradational to dominantly aggradational is typical for lowstand normal regressions. Conversely, the change in depositional trends from dominantly aggradational to dominantly progradational is typical for highstand normal regressions. Horizontal scale: approximately 50 km. Vertical scale: 2 s two-way travel time (Catuneanu et al. 2009, Fig. 19, p. 14)

and magnetic reversal stratigraphy are essential for providing numerical ("absolute") ages, and chemostratigraphy is becoming increasingly important. Radioisotopic dating is the only chronostratigraphic tool available for the Precambrian, and is essential for providing a calibration scale of biostratigraphic subdivisions in the Phanerozoic (Harland et al. 1964, 1990; Berggren et al. 1995). Magnetic reversals are now of great importance for studying Upper Cretaceous and Cenozoic strata but are difficult to use in older rocks owing to the difficulty of obtaining complete reversal sequences and the problems of post-depositional modification (Kennett 1980).

Early speculations about the age of the Earth, and of rates of geological processes, began to be situated within a modern quantitative framework with the development of radioisotopic dating by Ernest Rutherford in 1905. The English geologist Arthur Holmes made the development of a radioisotopically calibrated geologic time scale a central part of his life's work, and it could be argued that the first modern era in the quantification of geologic time ended with his death in 1965. Two major events occurred at about this time. The second edition of his great book "*Principles of Physical Geology*" appeared (Holmes 1965), containing, among other things, a lengthy treatment of radioisotopic dating (plus what were then his very advanced ideas about continental drift), and the Geological Society of London published the first comprehensive compilation of data in support of a modern geological times scale (Harland and Francis 1964, 1971).

It is important to distinguish between the two quite different, but interrelated concepts of time, geochronometry, the measurement of time, in standard units such as the year and the second, and chronostratigraphy, the compilation of standard rock units. A distinction between "time" and "rocks" is essential, given the incompleteness of the stratigraphic record, and the need to continually revise, expand and update the means by which we relate the chronostratigraphic to the geochronometric scale. Harland (1978) and Harland et al. (1990, Chaps. 1–3) provided detailed explanations of the theory and terminology surrounding these terms. Because of the very fragmentary nature of the sedimentary record, there is still no chronostratigraphic standard for the Precambrian, which is subdivided mainly on a geochronometric basis, except for the Ediacaran System (635–542 Ma), the youngest part of the Precambrian, which is the only exception (Knoll et al. 2006).

The subdivision of the stratigraphic record into even smaller units based on detailed studies of their fossil content had reached a remarkably sophisticated level by the early twentieth century, based on the specialized study of some unique units in which the fossils are abundant and contain readily measurable indicators of rapid evolution. The ammonites of the Jurassic in southern Britain figure prominently in this history, as described by Callomon (1995, 2001). It can be demonstrated that local subdivisions of relative stratal time representing time spans in the range of 10^5 years are possible based on such work (this is discussed further in Chap. 8). But the question of how to incorporate this information systematically in the construction of a chronostratigraphic scale that could be used worldwide continued to be controversial and problematic until the 1970s (this early history is summarized by Miall 2004, p. 6–11). There are three obvious reasons: the record is everywhere locally incomplete, fossils are facies bound, and with very rare exceptions it is not possible to assign specific numerical ages to fossil horizons without interpolations and extrapolations that might incorporate the other two problems. Some boundaries, including many established early in the history of the science, had been very poorly defined. For example, the Silurian–Devonian boundary, first recognized in Britain, occurs at a major angular unconformity, or within a marine to nonmarine transition, within which biostratigraphic correlation was problematic, raising (but only much later, when time began to be quantified) the question of how to define and categorize the time undocumented at the break.

Solutions to these problems began to emerge in the 1960s and might be said to constitute the beginning of the second phase of the modern era of modern stratigraphy. The key development was the evolution of the **Global Stratigraphic Sections and Points (GSSP)** concept. The ideas appear to be primarily British in origin (e.g., Ager 1964; Bassett 1985; Cowie 1986; Holland 1986). They encompass two important concepts: the idea of designating key marker boundaries at specific type locations within continuous sections, and the idea that multiple criteria—biostratigraphy and radioisotopic dating, and more recently magnetostratigraphy and chemostratigraphy, wherever applicable, should be used to nail down the precise age of the boundary "by all available means," to quote Torrens (2002, p. 256). McLaren (1970, p. 802) explained the desirability of defining boundaries within continuous sections in this way:

> There is another approach to boundaries, however, which maintains that they should be defined wherever possible in an area where "nothing happened." The International Subcommission on Stratigraphic Classification, of which Hollis Hedberg is Chairman, has recommended in its Circular No. 25 of July, 1969, that "Boundary-stratotypes should always be chosen within sequences of continuous sedimentation. The boundary of a chronostratigraphic unit should never be placed at an unconformity. Abrupt and drastic changes in lithology or fossil content should be looked at with suspicion as possibly indicating gaps in the sequence which would impair the value of the boundary as a chronostratigraphic marker and should be used only if there is adequate evidence of essential continuity of deposition. The marker for a boundary-stratotype may often best be placed *within* a certain bed to minimize the possibility that it may fall at a time gap." This marker is becoming known as "the Golden Spike."

By "nothing happened" it meant a stratigraphic succession that is apparently continuous. The choice of boundary is then purely arbitrary and depends simply on our ability to select a horizon that can be the most efficiently and most completely documented and defined (just as there is nothing about time itself that distinguishes between, say, February and March, but to define a boundary between them is useful for purposes of communication and record). This is the epitome of an empirical approach to stratigraphy. Choosing to place a boundary where "nothing happened" is to deliberately avoid having to deal with some "event" that would require interpretation (see Miall 2004 for a discussion of the importance of this point). This recommendation was accepted in the first International Stratigraphic Guide (Hedberg 1976, pp. 84–85). The concept also includes the proviso that only the base of a unit is so defined at the chosen location, not the top of the underlying unit, lest future work determines that at the type section the boundary is marked by a hiatus—hence the term **topless stage**. The "missing time" so identified is then assigned to the underlying unit, permitting continuous revision without the need for new boundary definitions.

The concept of the Global Stratigraphic Section and Point (GSSP), informally called the **Golden Spike** concept, was rapidly accepted, and has led to an explosion of specialized work under the auspices of the *International Commission on Stratigraphy* (ICS), a division of the *International Union of Geological Sciences* (IUGS), to correlate key sections worldwide in order to develop internationally recognized markers for epochs and stages that could then become part of the standard chronostratigraphic scale (e.g., Remane 2000a). This work is regularly reported in the IUGS journal *Episodes* and, nowadays, on the website https://stratigraphy.org/ maintained by the ICS. The criteria for the selection and ratification of a GSSP are as follows (https://stratigraphy.org/gssps/):

- A GSSP has to define the lower boundary of a geologic stage.
- The lower boundary has to be defined using a primary marker (usually the first appearance datum of a fossil species).
- There should also be secondary markers (other fossils, chemical, geomagnetic reversal).
- The horizon in which the marker appears should have minerals that can be radioisotopically dated.
- The marker has to have a regional and global correlation in outcrops of the same age
- The marker should be independent of facies
- The outcrop has to have an adequate thickness
- Sedimentation has to be continuous without any changes in facies
- The outcrop should be unaffected by tectonic and sedimentary movements, and metamorphism
- The outcrop has to be accessible to research and free to access
- This includes that the outcrop has to be located where it can be visited quickly (International airport and good roads) and has to be kept in good condition (ideally a national reserve), in accessible terrain, extensive enough to allow repeated sampling and open to researchers of all nationalities.

A discussion of selected GSSPs is presented in Sect. 7.8.3. As of 2021, 102 Phanerozoic boundaries had been established by the definition of a GSSP. Of these, only 72 had been agreed upon and ratified by the Commission. The first to be ratified, in 1972, was that for the base of the Lochkovian Stage, which defines the base of the Devonian. The GSSP was defined at Klonk, in the Czech Republic (Fig. 7.42). All the GSSPs for the Silurian and Devonian have been ratified, and those for the Silurian were completed in 1984. Only 4 of the 12 Cretaceous GSSPs have been ratified, a list that does not include that for the base of the Berriasian, which defines the base of the Cretaceous. This is surprising, considering the wealth of biostratigraphic and chemostratigraphic data now available for the Cretaceous and younger intervals in the Phanerozoic.

The importance of using "all available means" to identify, calibrate and date GSSPs cannot be over-emphasized. As noted in Sect. 7.5.3, the use of single criteria, such as first appearance data (FAD) and last appearance data (LAD) can lead to errors associated with biostratigraphic diachroneity. Quantitative methods, particularly graphic correlation and its advanced version, constrained optimization, make use of multiple criteria. As Smith et al. (2015) noted, despite the best efforts of stratigraphers, it has emerged that some GSSPs have been defined at sedimentary breaks, and others still over-rely on few chronostratigraphic criteria, such as the range of a single key taxon. As of 2015, some 74 of currently defined GSSPs actually only make use of biostratigraphic criteria. Smith et al. (2015) also noted that most GSSPs have been defined at outcrop sections of shallow-marine rocks (where the probability of multiple hiatuses and much missing section is greatest: see Chap. 8), with little use being made of subsurface drill-core sections, and none have been established in deep-marine sediments, where the preserved record is likely to contain fewer hiatuses.

It is instructive to review the development of the modern geologic time scale from 1964 to the present. Among the first to appear were the scales for the Jurassic and Cretaceous developed by Van Hinte (1976a, b). Useful discussions of stratigraphic concepts were compiled by Cohee et al. (1978). Numerous attempts to synthesize existing data for the Phanerozoic have been made, notably by British (Harland et al. 1982, 1990), French (Odin 1982) and American (Berggren et al. 1995) groups. An important review of modern Chinese work was provided by Ogg (2019). Each group drew on its own data base and made different interpolations and extrapolations, with the result that there are significant differences in the assigned ages of many of the important chronostratigraphic boundaries.

A quantum leap forward was achieved by the ICS with the publication in 2004 of its updated geologic time scale (GTS2004: Gradstein et al. 2004a). Gradstein et al. (2012) subsequently published their own updated version (although this is not an official product of the International Stratigraphic Commission), and a further major revision appeared in 2020 (Gradstein et al. 2020). The 2004 version incorporated numerous new data points, documented with the use of quantitative biostratigraphy, much-improved radioisotopic dating methods, chemostratigraphy and (for the Neogene) cyclostratigraphy (Fig. 7.33). The new scale (now updated to GTS2020) presents us with unprecedented opportunities for the comparison and calibration of detailed local and regional studies of rates and processes. All Cambrian to Paleogene ages are given to the nearest 100,000 years, although for much of the Cambrian–Devonian scale, potential errors of > 1 m.y. remain. Post-Paleogene ages are given to within 0.01 m.y. This scale, like all before it, incorporates numerous revisions of assigned ages. Almost all major chronostratigraphic boundaries in the Mesozoic and Paleozoic have been revised by several million years relative to earlier scales, such as that of Berggren et al. (1995), reflecting new data or changing interpretations of earlier data. There is no sign, yet, that the time scale has finally stabilized, although the incremental changes from one scale to the next do appear to be getting smaller. We return to this point later.

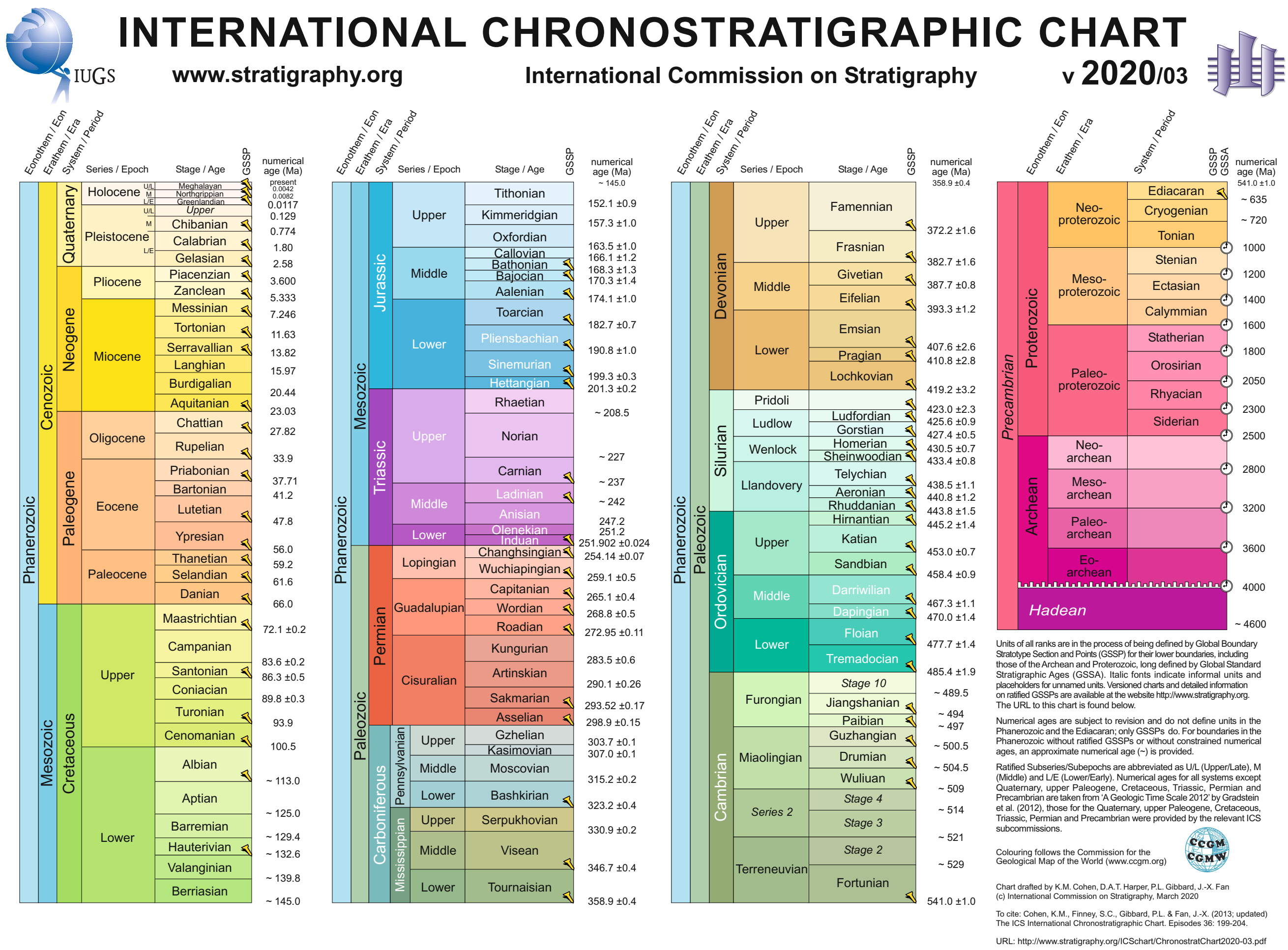


Fig. 7.33 The geological time scale prepared by the International Commission on Stratigraphy. Internationally agreed GSSPs are marked with a golden-spike icon. This figure shows the version current as of March 2021 (https://stratigraphy.org)

The time scale now undergoes continuous revision, under the auspices of the ICS. For example, a 2012 version was published in time for the 34th international Geological Congress at Brisbane (Cohen et al. 2012). (The latest version available at the time of going to press is provided in Fig. 7.33.) An example of the detail involved in the construction of the scale is shown in Fig. 7.34.

7.8.2 Determining the Numerical ("Absolute") Age of a Stratigraphic Horizon

Direct dating of sedimentary rocks by radioisotopic dating can only be carried out on a few potassium-bearing minerals, such as glauconite, because potassium is the only constituent of authigenic sedimentary minerals that contains naturally occurring radiogenic isotopes. Potassium–argon and argon–argon methods are the most common ones employed. However, the ages determined by the use of this method may relate to diagenetic age rather than depositional age, and there are problems associated with the loss of the daughter product, argon.

More commonly, radioisotopic ages are determined for interbedded volcanic horizons, especially ash beds or bentonites, and the stratigraphic age of the rocks of interest are then determined by interpolation, as shown in Fig. 7.35. In order to make use of this method, two major assumptions have to be invoked, firstly, that sedimentation was continuous and, secondly, that sedimentation was at a constant rate. In most cases, neither of these assumptions can be assumed to be fulfilled. This does not mean that the technique is useless, only that it must be employed with care. Ideally, many radioisotopic determinations should be made on successive volcanic units, and then interpolations and extrapolations can be fine-tuned by methods of averaging and by making local corrections. We come back to this issue in Chap. 8, where we examine further some deeper questions regarding the interpretation of time as preserved in the rock record.

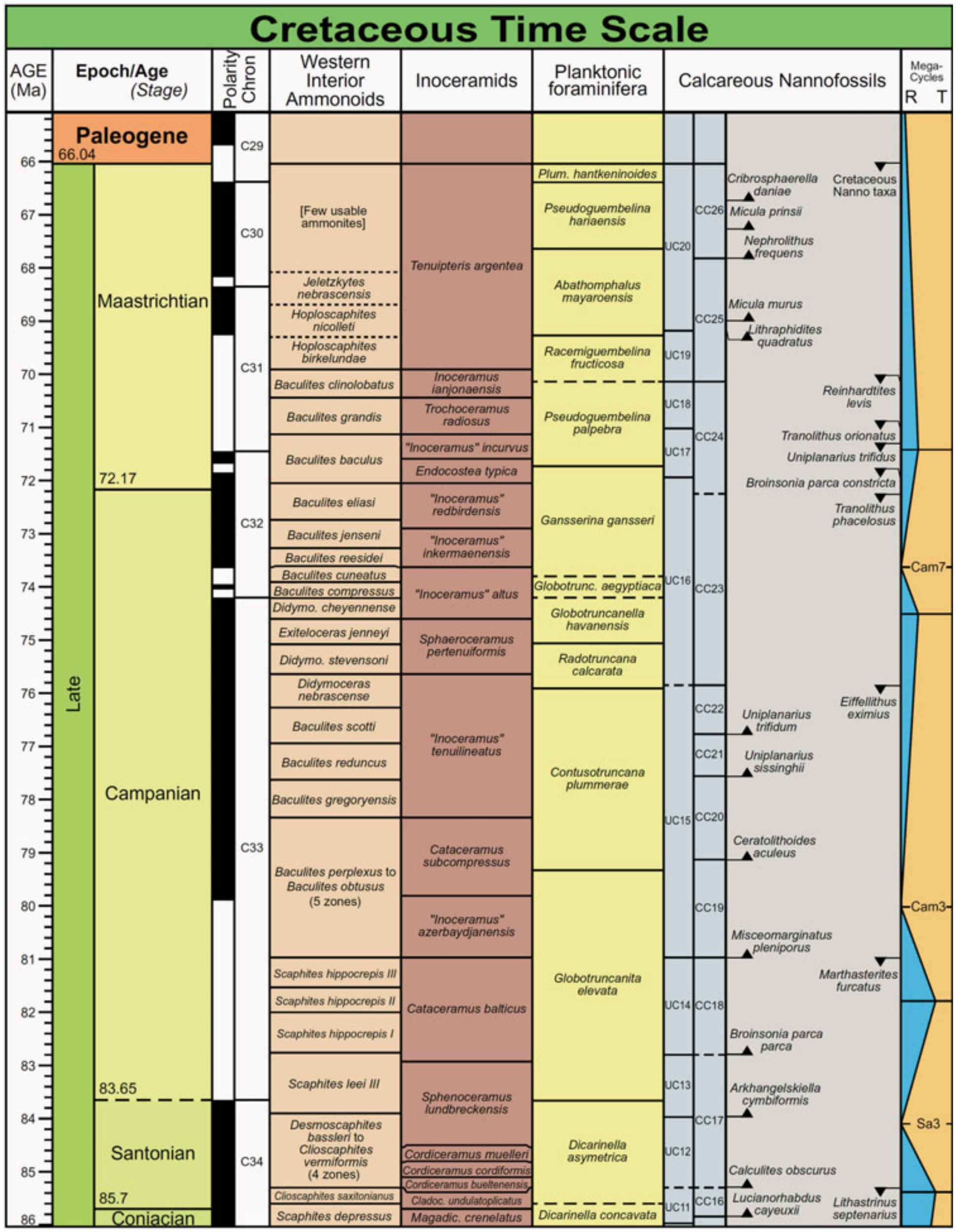


Fig. 7.34 Part of the GTS2020 scale for the Late Cretaceous, showing the magnetic polarity chrons and selected biozone schemes. Arrows in the microfossil columns indicate FADs (pointing upward) and LADs (pointing downward) (Gale et al., in Gradstein et al. 2020, Fig. 27.9, p. 1044)

Modern radioisotopic dating methods. Not surprisingly, the methods of radioisotopic dating and the data base have undergone orders-of-magnitude transformations over the last half-century (Mattinson 2013). Holmes (1960) provided a list of 63 ages, mostly obtained by the K/Ar method. Of these, 19 were used to construct a linear time scale, adjusting the space allotted to each system according to cumulative stratal thickness, it being assumed at the time that this constituted an indication of the duration of the period. By the time of the 1964 Geological Society of London scale, the data base consisted of some 337 measurements. The compilation of Harland et al. (1990, Table 4.2) lists some 750 entries. Figure 7.36 illustrates the increased accuracy and precision of the dating of the Permian–Triassic boundary over the 22-year period from 1991 to 2011. The current ICS date for this boundary is 252.17 ± 0.06 Ma.

Among the improvements in methods has been the establishment of the chronogram method by Harland et al. (1982, pp. 3–4) for plotting and visualization of measurement error, a standardization of decay constants, much enhanced analytical methods and a better appreciation of the systematics of the various isotopic decay paths. The K/Ar

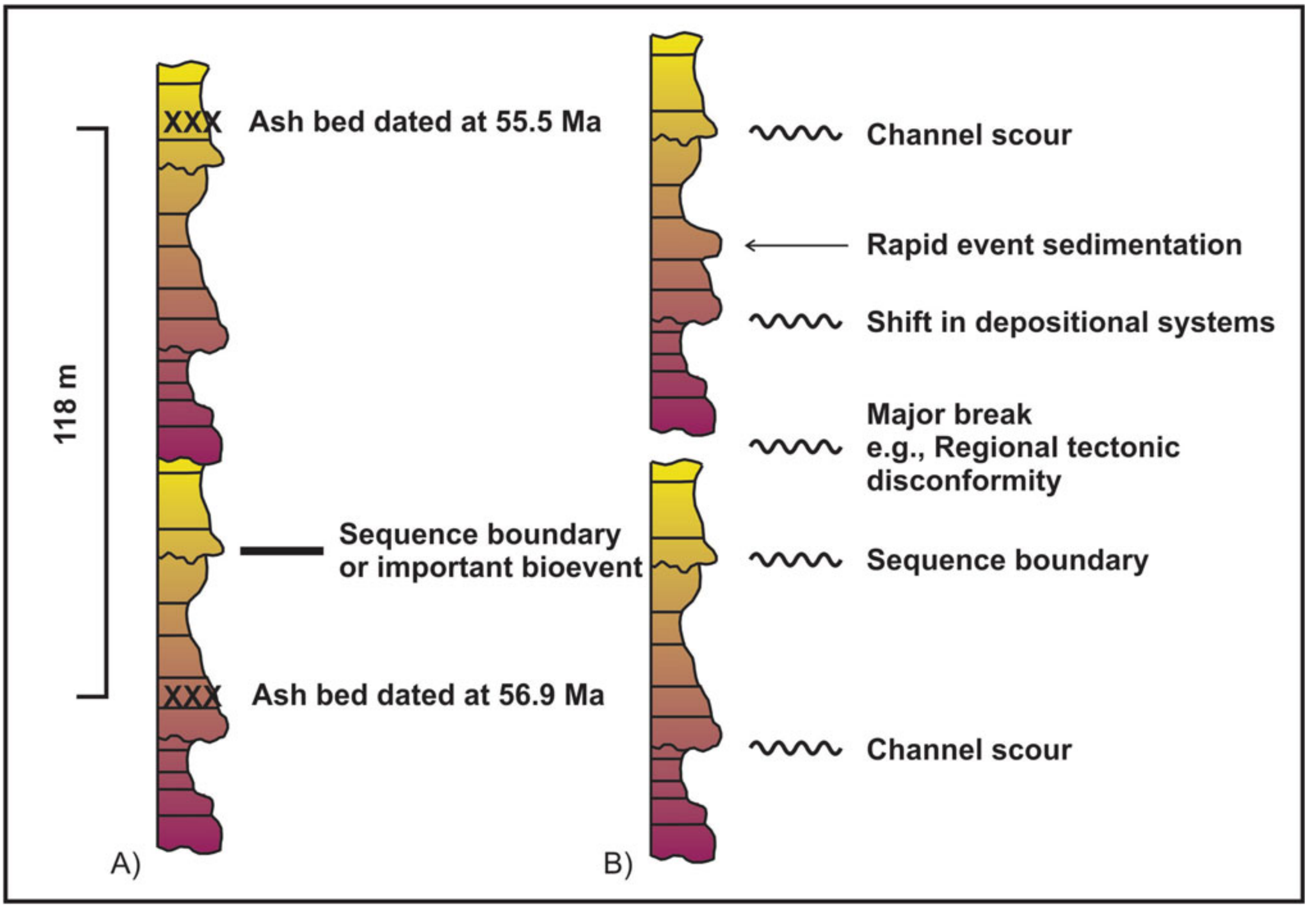


Fig. 7.35 The determination of "absolute" ages of stratigraphic events by using the method of "bracketing" the event by a pair of radioisotopic ages; The principle of the method (left) and the reality in practice (right). A) On the left, 118 m of beds are shown accumulating in 56.9–55.5 m.y., = 1.4 m.y. The average sedimentation rate is therefore 118/1.4 m/m.y., = 84 m/m.y. The sequence boundary of interest is 30.3 m above the lower ash bed. Assuming a constant sedimentation rate and no hiatuses, the 30.3 m of beds accumulated in 30.3/84 = 0.36 m.y. Therefore the age of this boundary is 56.9–0.36 = 56.54 Ma. However, as shown in B), at right, real sections are full of sedimentary breaks, which represent missing time, and variations in sedimentation rate, all of which render the concept of the average sedimentation rate suspect

and Rb/Sr methods have been found to be less reliable than those obtained by the $^{40}Ar/^{39}Ar$ and U–Pb methods (Villeneuve 2004; Mattinson 2013). The use of ages derived from glauconites, popular in the 1980s (Odin 1982), has decreased, because of the increasing realization of the unreliability of the method. It has been demonstrated repeatedly that daughter products, Sr or Ar, may be lost from the mineral grains, yielding ages that are too young.

Radioisotopic age determinations are characterized by a normal experimental error. The current practice may achieve a precision of ± 0.1% or better (Mattinson 2013, p. 310; Smith et al. 2015; Cramer et al. 2015; Schmitz et al., in Gradstein et al. 2020, Chap. 6), e.g., ± 100,000 years at 100 Ma. However, the attainment of such accuracy and precision in the stratigraphic record depends on the availability of appropriate datable material at the rights places in the rock record, and the accuracy of the existing time scale varies from stage to stage because of this. Refinements in decay constants and inter-laboratory calibration become even more important. As Sageman et al. (2014) demonstrated in their discussion of U–Pb and $^{40}Ar/^{39}Ar$ data sets from an Upper Cretaceous interval in the Western Interior Basin, dating accuracy and precision are now such that differences between the two methods may relate to internal differences in the processes that set the final isotopic ratios—differences in Ar closure temperature between different minerals, and the length of the cooling phases of igneous bodies from which such minerals as zircon are ultimately derived. There may be differences of several hundred thousand years between the results from different methods, which then need to be reconciled by calibration against other data sets, including biochronology, chemostratigraphy and magnetostratigraphy. An important multi-authored review of the $^{40}Ar/^{39}Ar$ dating method was provided by Schaen et al. (2021).

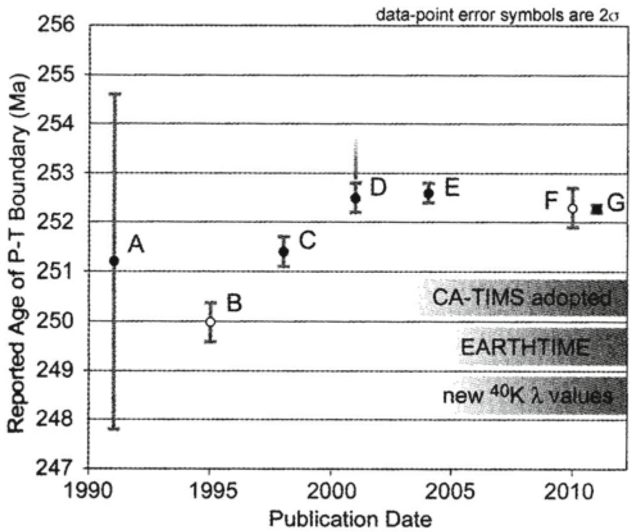


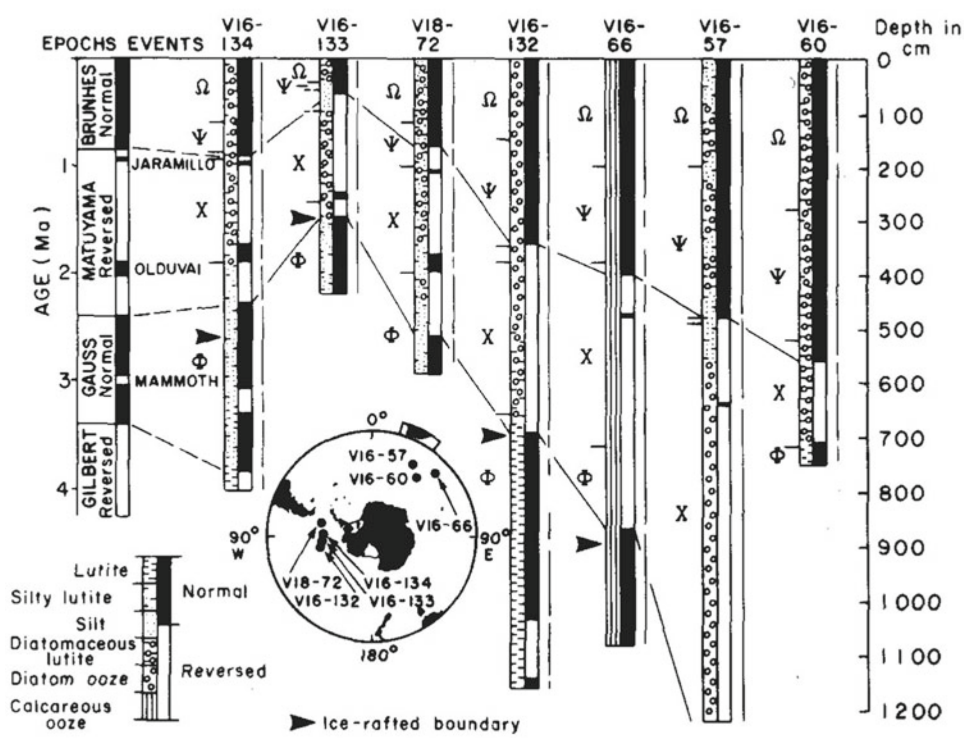
Fig. 7.36 The improvements in accuracy and precision of the dating of the Permian–Triassic boundary from 1991 to 2011. Solid circles are dates based on U–Pb zircon analysis; open circles are dates based on Ar–Ar analysis of sanidine (Mattinson 2013, Fig. 11, p. 315)

Even given a very precisely dated GSSP for a key chronostratigraphic boundary, it is then quite another matter to exploit this information to date and calibrate new sections elsewhere. This exercise requires a biostratigraphic, radioisotopic or chemostratigraphic record capable of yielding correlations of equivalent accuracy and precision.

Magnetostratigraphy: The first major development, the erection of a satisfactory geomagnetic polarity scale for the last 4.5 Ma, was published by Cox (1969), based primarily on the sampling of successions of lava flows. The standard scale now in use was developed primarily from the study of cores through undisturbed deep-sea sediments. The first use of such cores predated the *Deep Sea Drilling Project* (DSDP), and the cores were very short (see review by Kennett, 1980). Harrison and Funnell (1964) were the first to combine biostratigraphy and chronostratigraphy in an attempt to correlate a reversal event. Later, Opdyke et al. (1966) studied some longer cores, up to 12 m in length, and were able to correlate reversal events with the land-based lava sequence. Radiolarian zone boundaries closely parallel the reversal correlations (Fig. 7.37). The DSDP started in 1968 and began to have an important effect on chronostratigraphy. However, difficulties were encountered in establishing magnetic stratigraphy directly from the cores because of drilling disturbance and bioturbation. The practice developed of calibrating polarity and biostratigraphic data by correlating deep marine with exposed on-land sections and dating the latter radioisotopically.

Since the 1960s, magnetostratigraphy has become almost equal in importance to biostratigraphy, chemostratigraphy and radioisotopic dating in establishing a time scale for the last 160 Ma of Earth's history. The establishment of a reliable reversal sequence requires close sampling of an assumed, undisturbed, continuous section, and relies on the ability to recognize a kind of "bar-code" pattern of normal and reversed intervals. However, undisturbed sections for the pre-Late Jurassic are rare—there is no undisturbed sea

Fig. 7.37 Correlation of magnetic stratigraphy and radiolarian zones in seven cores from the Antarctic Ocean (Opdyke et al. 1966)



floor of greater age—and it is therefore much more difficult to standardize the scale for these older rocks.

The technique is most useful for time spans characterized by frequent reversals. During the Cenozoic, reversals occurred two or three times each million years, providing a very distinctive pattern that has enabled the establishment of 29 reversal intervals, named **chrons** for that era (Ogg and Smith 2004). The availability of age information on a 10^4–10^5-year time scale has provided powerful new tools for exploring the rates of sedimentary and stratigraphic processes in the geological record. Reversal sequences for older parts of the Mesozoic and the Paleozoic have been assembled from partial sections (Langereis et al. 2010), but are not yet reliable enough to become part of the standard geological time scale. Figure 7.34 shows part of the GTS2020 time scale for the Cretaceous, including the numbered polarity chrons.

Magnetostratigraphy may also be used to correlate local sections with each other without regard to the global scale. However, because reversal events are not unique, it is not possible to correlate them by matching sequences from different stratigraphic sections unless they contain particularly distinctive long or short polarity intervals. Some supplementary criteria may be required to assist in matching, such as marker beds or biostratigraphic zonation. Picard (1964) and Irving (1966) were among the first Western workers to use paleomagnetic correlation for sediments on the continents. The technique has become widely used for nonmarine sediments because of the scarcity of other means of precise correlation.

A single example of the use of magnetostratigraphy in a practical field problem is described here briefly. For many years, a team has been exploring the nonmarine Siwalik Group (Oligocene-Quaternary) of Pakistan, in part because of its rich vertebrate fauna and in part because of the information the sediments yield about the tectonics of the Himalayas, from which the sediments were derived. Magnetostratigraphy, coupled with radioisotopic dating of several ash beds, provided a useful means of local correlation between sections that show marked lateral facies changes (Keller et al. 1977; Barndt et al. 1978; Johnson et al. 1979). It was also possible to propose a correlation with the global scale. Figure 7.38 illustrates the correlation of three closely spaced sections in the Pabbi Hills area, near Jhelum. From three to five oriented rock specimens were collected from each of 113 sites within the sections. These were subjected to laboratory tests to determine the stability of the field and the absence of magnetic overprinting, and the pole positions obtained corrected for structural dip. The reversal zones were correlated with the standard scale of Opdyke (1972) using the following argument. The oldest remains of *Equus* (horse) were found at the 400 m level in the composite section. The oldest occurrence of *Equus* in North America is dated 3.5 Ma and in Asia 2.5 Ma. It is considered unlikely that the Pabbi Hills fossils are older than 3.5 Ma. Note that the fossil locality is in a short normal polarity sequence within a long reversed interval. Only two such dominantly reversed intervals are present in the magnetostratigraphic scale, the Matuyama and the Gilbert zones. The Gilbert zone extended from 5.1 to 3.3 Ma, and the Matuyama from 2.41 to 0.70 Ma (revised ages of Opdyke 1972; modified from Cox 1969). The evidence of *Equus* suggests that this is the Matuyama zone. The two short normal events then correlate with the Olduvai and Jaramillo subzones. The Gilsa event (subzone) of Fig. 7.38 is not universally recognized and does not appear on Opdyke's (1972) chart.

Four composite sections have been correlated using the presence of two tuff horizons (Visser and Johnson 1978) and the polarity zones, as shown in Fig. 7.38.

The importance of this work is that it permits precise local and global correlation of vertebrate localities, permits accurate calculations of sedimentation rates and provides accurate control for studying sedimentological characteristics, basin architecture and tectonic events. Some of these aspects have been explored in later papers by this research group and other workers in the area. For example, Johnson et al. (1985) were able to determine the various time scales represented by fluvial cycles and to explore the implications for rates of channel wandering and the nature of tectonic and climatic controls on sedimentation. Behrensmeyer (1987) developed detailed two-dimensional reconstructions of the stratigraphic architecture as a basis for an examination of the taphonomy of the vertebrate remains.

Chemostratigraphy: **Oxygen isotope stratigraphy** has made an enormous contribution to the development of stratigraphy. The method depends on measurements of the $^{16}O/^{18}O$ ratio. Because ^{16}O is the lighter of the two isotopes, water molecules containing this light oxygen are preferentially evaporated from seawater. During times of ice-free global climate, they are recycled to the oceans and the isotopic ratio remains in a stable balance corresponding to the natural proportions of the two isotopes in the hydrosphere. However, glacial ice is composed of condensed ^{16}O-enriched water so that continental ice buildups are preferentially enriched in ^{16}O, with the result that the $\delta^{18}O$ content of the oceans is increased. Sediments preserve the isotopic ratios of the oxygen that existed in the hydrosphere at the time the sediments were formed. Measurements are made on the carbonate comprising foraminiferal tests. Emiliani (1955) was the first to demonstrate that the oxygen isotope record is cyclic, and was among the first to argue that the fluctuations should reflect the high-frequency oscillations between glacial and interglacial stages that have characterized the Cenozoic record.

We now know that the $^{16}O/^{18}O$ ratio is a highly sensitive indicator of global ocean temperatures and ice cover, and

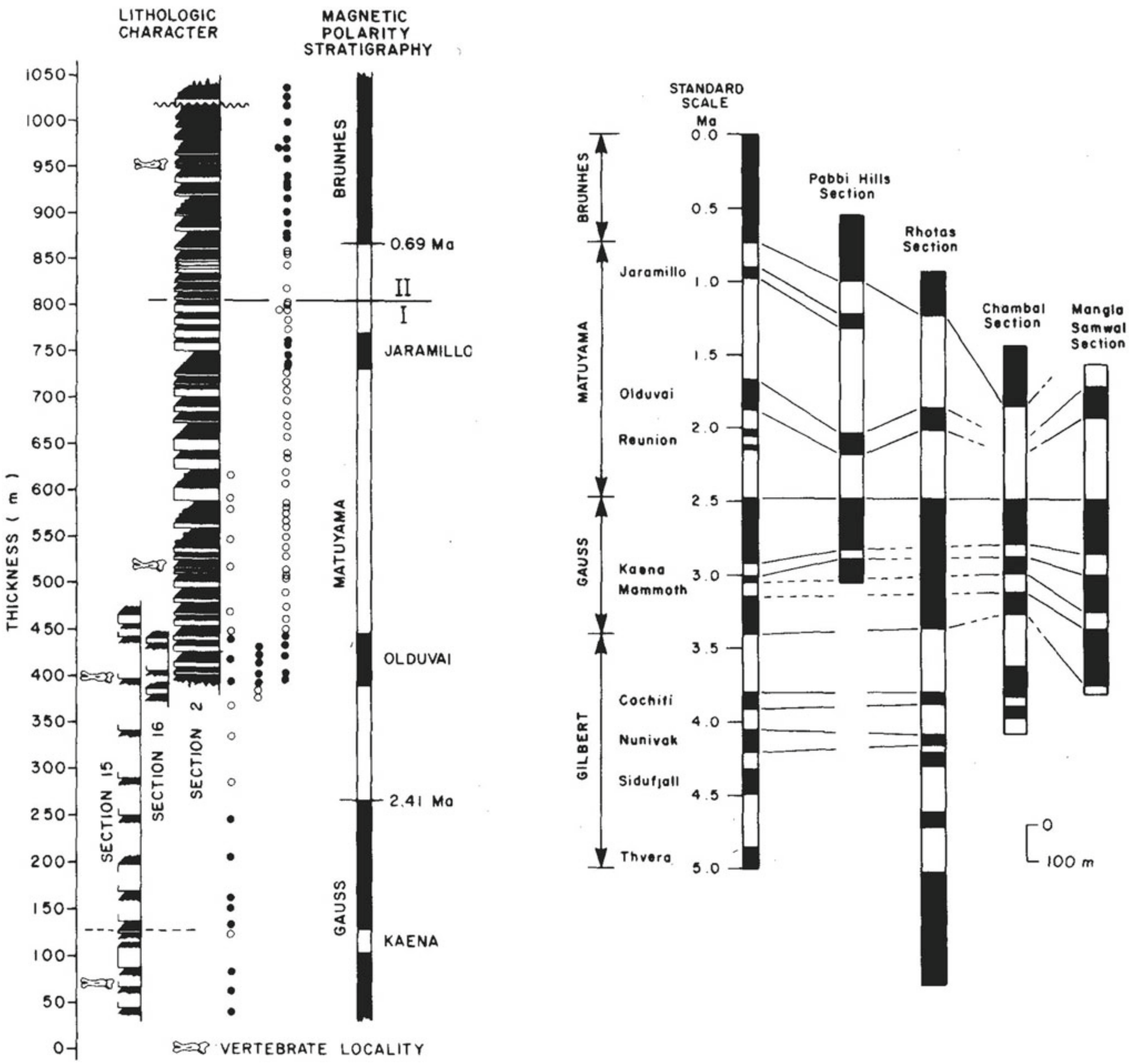


Fig. 7.38 The practical use of magnetostratigraphy to correlate sections of nonmarine strata in Pakistan. Three partial sections were compiled at the left to provide a single composite section for the Pabbi Hills region. Key vertebrate locations are shown, and field samples showing normal and reversed magnetic polarities are indicated by the vertically arranged suites of black and white circles. Correlation of the Pabbi Hills section with the standard scale and with other nearby stratigraphic sections is shown at right. Note the presence of missing reversals in some of the sections, indicating either incomplete sampling or reversals that are actually missing because of local erosion (redrawn from Johnson et al. 1979)

can therefore be used as an analog recorder of ice volumes (Hays et al. 1976; Shackleton and Opdyke 1976; Matthews 1984, 1988). Matthews (1984) suggested a calibration value of δ^{18}O variation of about 0.011‰ per meter of sea-level change. Miller et al. (2005) compiled the available data for the Late Cretaceous to present and used this compilation to develop a scenario for the growth and variation in the extent of the ice cover of Antarctica (see Sect. 7.9 and Fig. 7.55). The appearance of large ice caps in the early Oligocene, and the beginning of the northern hemisphere glaciation at about 5 Ma are clearly indicated by large stepwise increases in δ^{18}O, but also of interest are the fluctuations in the isotopic data that would suggest the development of temporary, small ice caps in the Antarctic in the Late Cretaceous. This provides important independent evidence to support the growing body of sequence stratigraphic work that suggests the likelihood of orbital forcing of glacioeustasy in the Cretaceous. The mapping of suites of high-frequency sequences in Alberta (Plint 1991), New Mexico (Lin et al. 2021; see Sect. 8.12) and elsewhere, and the development of a chemostratigraphic record of orbital climate variations in Colorado (Sageman et al. 2014) are all consistent with the revelation that, contrary to long-held ideas, the Cretaceous was not a period of uniform "greenhouse" climate.

A modern oxygen isotope scale developed from the early systematic work of the SPECMAP (SPEctral MAPping) project (Imbrie et al. 1984; Imbrie 1985). Following, in particular, the pioneering work of Hays et al. (1976) the SPECMAP project established a detailed record for the late Pleistocene (the last 780 ka) which calibrated the scale against an insolation energy index calculated from the integration of the three major orbital cycles, obliquity, precession and eccentricity. Since that time it has become standard procedure to record the oxygen isotope signal in deep-sea cores, and numerous studies have contributed to the refinement of the scale. A complete modern treatment of the subject is provided by Grossman and Joachimski (in Gradstein et al. 2020, Chap. 10).

The current scale used in GTS2020 includes a systematic series of marine cycles back to 5.5 Ma (Gradstein et al. 2020, Chap. 10; Fig. 7.39 of this book). As discussed in Sects. 7.8.7 and 8.11, the calibration of this scale against the record of orbital variations through the last few million years of Earth's history is now providing the main basis for a new, astrochronological time scale. A more generalized plot of $\delta^{18}O$ variations since the mid-Cretaceous is provided in Fig. 7.55.

Strontium isotope stratigraphy is a relatively new topic. It was not mentioned at all by Harland et al. (1990). In the late 1970s, it began to be recognized that the ratio $^{87}Sr/^{86}Sr$ varied systematically through time, as recorded in marine sediments (Burke et al. 1982). Veizer (1989) provided a detailed review of the origins of strontium in seawater and the processes that affected the preservation of the signal, including the effects of diagenesis. He argued that the rate of mixing of ocean waters and the long-residence time of strontium in seawater could potentially yield a reliable signal. McArthur (1994, 1998; McArthur and Howarth 2004) took the lead in the development and application of the tool to age determination and standardization of the method and its integration into the process of refining the time scale.

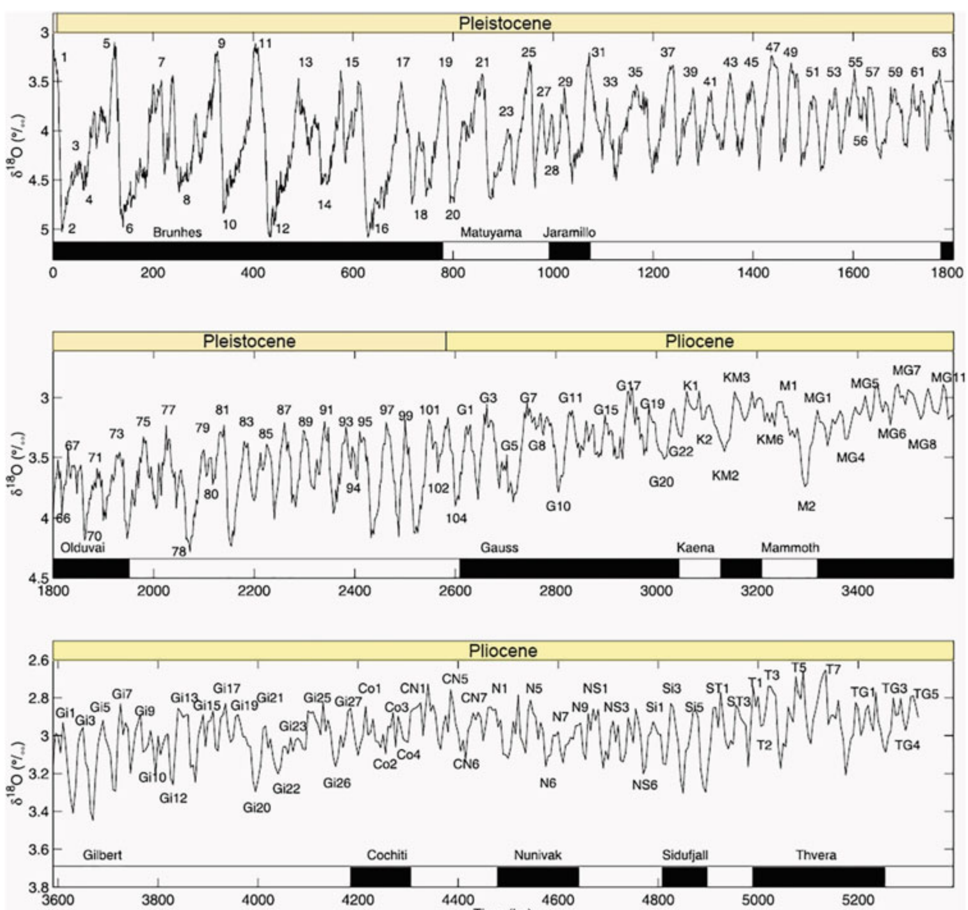


Fig. 7.39 The GTS2020 time scale for the last 5.3 Ma, showing the magnetic polarity chrons and the marine oxygen isotope record (Gradstein et al. 2020, Fig. 10.1, p. 284)

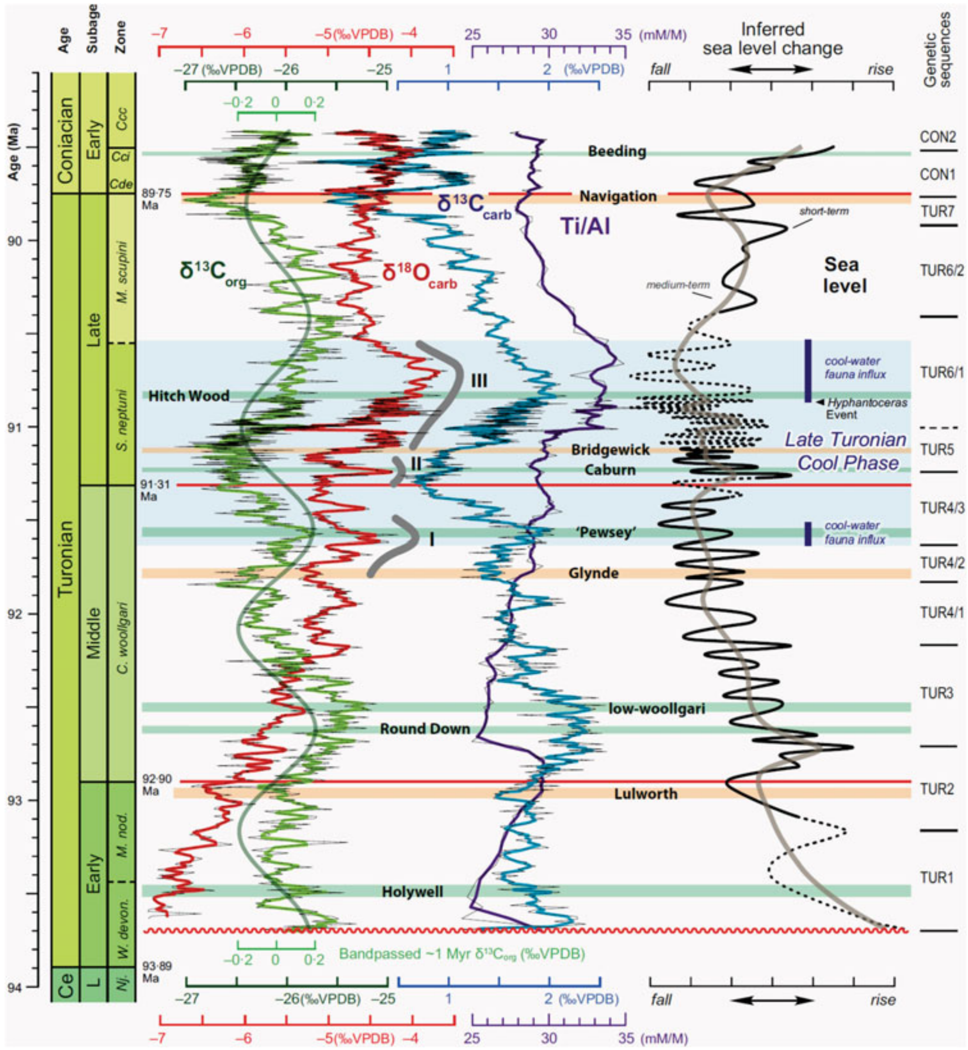


Fig. 7.40 Stable isotope profiles for the Bohemian Cretaceous Basin, showing the main named carbon excursion events (Jarvis et al. 2015, Fig. 7)

The $^{87}Sr/^{86}Sr$ ratio varies with age in the carbonate of various marine shells, mainly because of fluctuations in the rates and types of continental weathering, and the measurement of this value enables a sample to be situated on a graph showing the variation in composition with time. The graph for the Phanerozoic, as currently in use (McArthur and Howarth 2004, Fig. 7.1) is remarkably similar to that compiled by Veizer (1989), which indicates that the technique rapidly reached maturity. However, El Meknassi et al. (2018) warned that water mixing from continental or submarine groundwater sources may introduce inaccuracies in the dating method.

The values of the ratio are not unique with respect to time, because it has risen and fallen in a crudely cyclic manner within the range 0.7070–0.7090 during the last 450 m.y. so that some intermediate values correspond to several different ages on the curve. It is, therefore, necessary to know, within a few million years, which part of the graph to use for reading off the age against the calculated ratio.

Another chemostratigraphic tool of increasing importance is the **carbon isotope record**, $\delta^{13}C$. Research on this topic began with the work of Scholle and Arthur (1980) on certain Cretaceous intervals. The $\delta^{13}C$ record was calibrated against a rich biostratigraphic record in order to ensure chronostratigraphic precision. It was found that many carbon excursions could be correlated globally, and this was interpreted as a result of "oceanographic changes driving biotic turnover in the marine fossil record" (Jarvis et al. 2006, p. 565). Jarvis et al. (2006) expanded the work to define 39 **carbon excursion events** spanning the Cenomanian to Santonian interval, from which they developed a global reference curve, and it was speculated that the events were generated by Milankovitch-band orbital forcing of climate change. Additional research on the Turonian interval, with an analysis of pCO_2 changes and climate variability, was reported by Jarvis et al. (2015). An example of this work is shown here (Fig. 7.40) to indicate the potential for the tool to identify and correlate carbon and oxygen isotope values globally. Note a large number of named carbon excursion events. Cramer and Jarvis (2020) provided an overview of the topic for the GTS2020 volume, in which they provided scatter plots for all the data assembled to date (nearly 60,000 measurements) for the entire Phanerozoic.

7.8.3 Stages and Boundaries

The evolution of the **stage** concept was a confused one. D'Orbigny and Oppel in the mid-nineteenth century were the first to use the term with basically its present meaning (Hancock 1977; McGowran 2005).

Stages are effectively convenient groupings of biozones. Stage boundaries may be drawn at the top or base of a particularly well or widely developed biozone or a prominent faunal change. Many of our modern stage names were rather loosely defined, perhaps on only a handful of taxa when originally established, but have subsequently been refined by a more detailed study using many different life forms. Biozones are now usually established using only the members of a single phylum (except, perhaps, Oppel zones), whereas many stages have now been defined in many different ways. For example, Devonian stages are based mainly on brachiopods, corals, trilobites, fish, conodonts and palynomorphs; Cretaceous stages are based mainly on ammonites, pelecypods, brachiopods, foraminifera, nannofossils and palynomorphs.

What has happened over the years is that stage terms became so useful that geologists began to define the same stage in different ways. They added descriptions of suites of different kinds of fossils, which helped to define parallel successions of biozones and to transcend biogeographic problems, and then when radioisotopic and magnetostratigraphic data became available, this information was added in as well. Chemostratigraphic data are built-in wherever they are available. Gradually the rock-term stage became a more broadly based term referring globally to a particular interval of time. The modern use of stage concepts is well illustrated in current treatments of the geologic time scale, e.g., in Gradstein et al. (2004a, 2012, 2020). The stage is now regarded as "the basic working unit of chronostratigraphy (Hedberg 1976; Salvador 1994; Smith et al. 2015; Gradstein et al. 2020; Chap. 2). A total of 102 stages is now used to subdivide the Phanerozoic. Their average duration is 5.3 m.y.

In any given sedimentary basin, the recognition of a stage and its component chronozones depends on the nature of the fossil and other evidence that can be compiled for analysis. Surface studies may be able to benefit from the very detailed studies of macrofossils that have been carried out over the last 150 years. For example, paleontologists have found that ammonites evolved so rapidly during the Mesozoic that they have been able to erect scores of biozones. Modern dating methods have shown that for some parts of the Mesozoic these zones may represent time intervals as small as 90,000 years, which permits an astonishing precision in dating where the record is complete enough (Callomon 1995). An extreme example is shown in Fig. 7.41 (we return

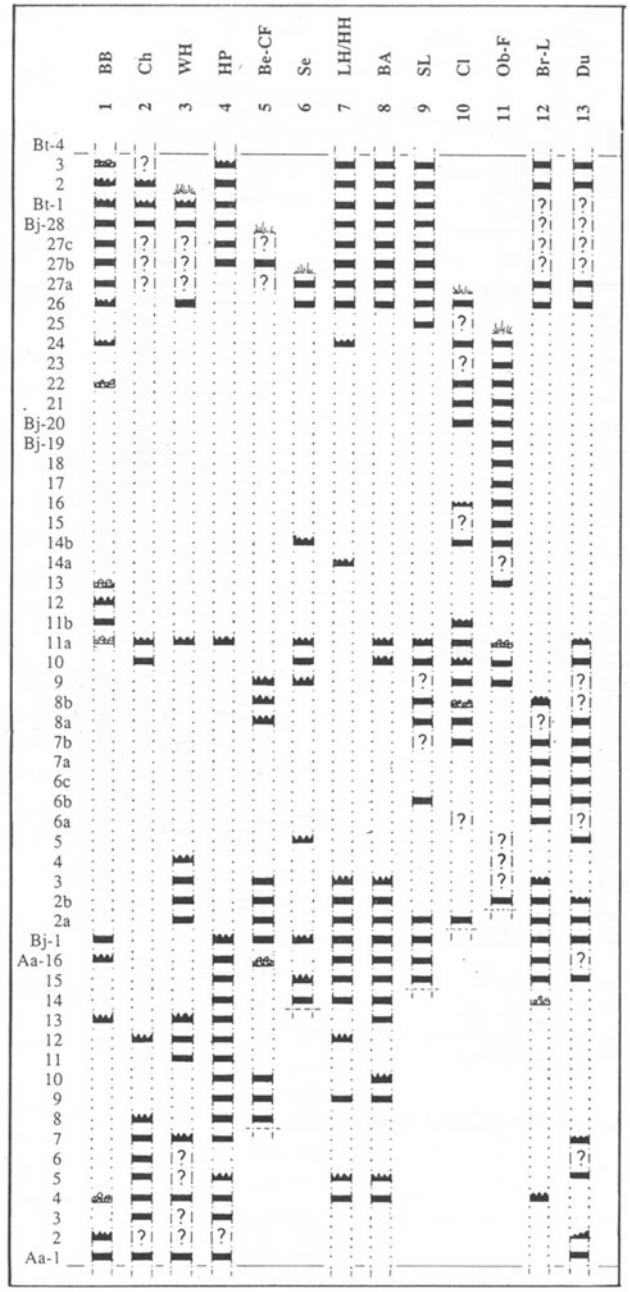


Fig. 7.41 The 56 ammonite faunal horizons recognized in 13 sections of the Inferior Oolite of Dorset and Somerset, England. The sections average 5 m in thickness are spaced out over a total distance of about 80 km, and span about 5 Ma of the Middle Jurassic; therefore each faunal horizon represents an average of about 90,000 years. Aa = Aalenian, Bj = Bajocian, Bt = Bathonian stages (Callomon 1995, Fig. 5, p. 143)

to a discussion of this particular example in Chap. 8). For parts of the early Paleozoic, trilobites and graptolites offer similar biostratigraphic precision (see Sect. 7.5.4), with graptolites now providing some of the most tightly defined and numerically dated zonal schemes in the Phanerozoic (Sadler et al. 2009).

Subsurface studies must rely on microfossil or palynological evidence (McGowran 2005). Foraminifera evolved rapidly during the Cenozoic, but for most of the geologic time microfossils, though commonly occurring in great abundance, seemed to have evolved more slowly, and so individual biozones necessarily represent longer intervals of time.

Most fossil groups tend to be facies bound, and the key to successful stage correlation is to locate sections where more than one useful fossil group is present or where repeated facies variations cause ecologically incompatible taxa to occur close together by interdigitation. This is largely a matter of chance. Much depends on lucky exposure of the right section, and the literature is replete with obscure geographic localities that have attained a specialized kind of fame because of the excellence of the biostratigraphic work carried out there. Examples are such small Welsh towns as Llanvirn and Llandeilo (Ordovician), the Eifel district, Belgium (Devonian), and the Barrandian area of the Czech Republic (Silurian-Devonian boundary). Geologists worldwide knew about Maastricht long before the average European politician had heard about this minor city. It seems like a happy accident of onomatopoeia that the hammering in of the first golden spike to locate a GSSP (the base Lochkovian 1972) would be at a place called Klonk! (Fig. 7.42). Most stages are named after such places. Many of the classic stages that were erected in the nineteenth or early twentieth century in Europe and North America have been replaced during the international work to develop the GTS as it was determined that the defining faunas were endemic or too facies-bound. Plate-tectonic movements cause faunal provincialism to vary in many taxonomic groups simultaneously, and so the perfection of this correlation varies from place to place and time to time. For example, Berry (1977) reported that the correlation of early and middle Ordovician graptolitic faunas between Europe and North America has been fraught with controversy because of faunal provincialism. At that time, the proto-Atlantic Ocean (Iapetus) was at its widest. We earlier referred to the comparable problem of trilobite correlation across Iapetus (Fig. 7.12). The end result of this extended effort is that most stages can now be recognized globally, with varying degrees of confidence.

As an example of modern work in establishing stages by means of detailed faunal work on several animal groups, a brief discussion is presented here of the Pridolian and Lochkovian stages, as defined in the Barrandian area of the Czech Republic. Chlupác (1972) published a detailed description of the faunas and revised the earlier biostratigraphic subdivisions of these rocks. His faunal list for the two stages includes over 300 species, of which graptolites, conodonts and trilobites constitute the most important biostratigraphic indicators. Other groups providing subsidiary control include eurypterids, phyllocarids, ostracodes, echinoids, cephalopods, gastropods, pelecypods and brachiopods. Figure 7.42 illustrates one of several short but critical sections measured through the Pridolian-Lochkovian boundary near Klonk. Limestone beds are numbered 1 to 53. This section constitutes the very first internationally agreed GSSP, having been ratified by the International Union of Geological Sciences in 1972. It constitutes the stratotype for the base of the Devonian epoch. A vital characteristic of the Barrandian area is that here the section straddling the Silurian–Devonian boundary is basinal-marine in origin, and therefore preserves a more continuous marine section relative to the marginal-marine, nonmarine or unconformable contact in the British type areas of the Silurian-Devonian contact.

The Pridolian–Lochkovian succession in this area consists mainly of thinly interbedded, grayish-black, calcareous mudstone and grayish-black to dark gray, fine-grained (micritic), skeletal, platy, weathering limestone, with subordinate beds of pale limestone and coarser, detrital limestone (calcarenite to calcirudite). Graptolites are abundant in the mudstones; trilobites and conodonts occur sparsely in the limestones, particularly in the paler, less muddy units, and become more abundant north of Klonk, where the rocks undergo a facies change into a predominantly pure carbonate succession. This interbedding of different facies with their contrasting faunas is one of the most important features of these central European sections from the point of view of biostratigraphic stratotype definition.

Two graptolite biozones have been recognized in the Pridolian and form the main basis for the definition of the unit. They are the *Monograptus ultimus* zone below and the *M. transgrediens* zone above. The latter does not reach the top of the Pridolian but is followed by a graptolite interregnum containing only sparse, nondiagnostic forms (Fig. 7.42). The base of the Lochkovian is defined by the sudden widespread appearance and abundance of *Monograptus uniformis* in bed 20. Other species of *Monograptus* and *Linograptus* appear in the upper part of the lower Lochkovian, while in the upper Lochkovian, *M. hercynicus* is typical. Some of these species occur in the pure carbonate facies, permitting close correlation with the shelly fauna. It is interesting to note that in spite of the effort biostratigraphers have made to formalize their biozone types, Chlupác's work is typical of many in that no attempt is made to state what kind of biozone is in use. The older of these graptolite biozones appear to be single-taxon range biozones, with concurrent-range biozones for the upper part of the lower Lochkovian and the upper Lochkovian.

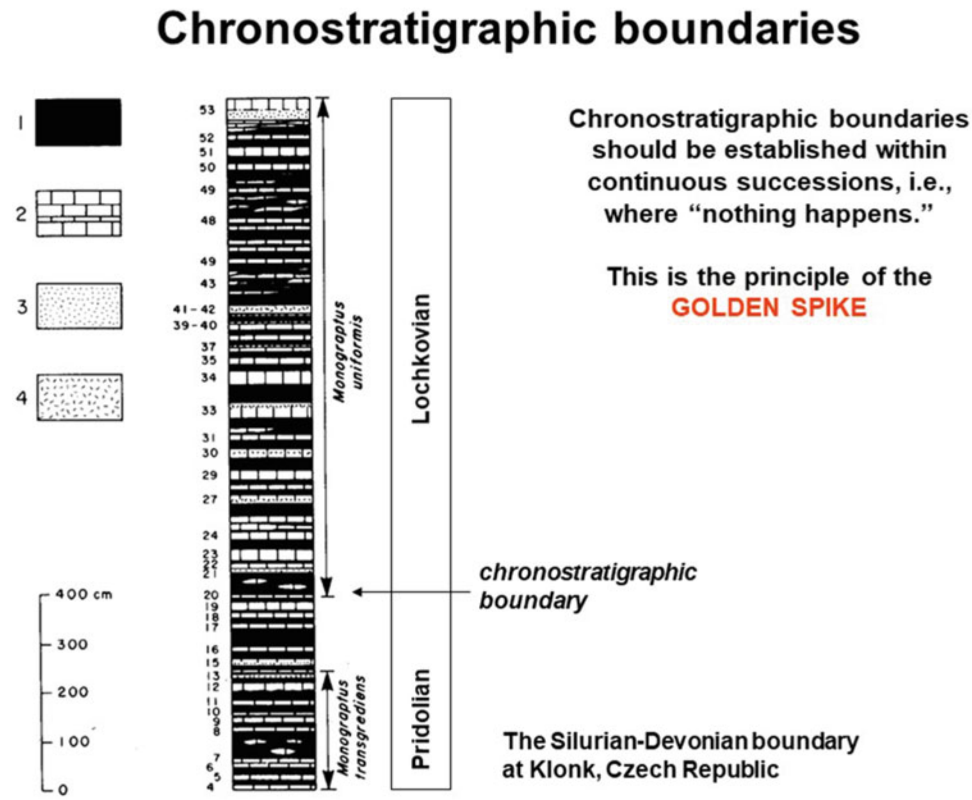


Fig. 7.42 The principle of the Golden Spike. The boundary between the Silurian and the Devonian was defined here in 1968 at a bed a certain distance below the first occurrence (FAD) of the graptolite *Monograptus uniformis*. This definition also serves to define the boundary between the Pridolian and the Lochkovian stages. Adapted from Chlupác (1972). This became the first internationally recognized GSSP

The trilobite *Warburgella (Podolites) rugulosa rugosa* is of primary importance in delineating the lower boundary of the Lochkovian. In the Klonk section, it appears in limestone bed 21, immediately above the first appearance of *M. uniformis* in the upper part of bed 20 (Fig. 7.42).

The conodont *Icriodus woschmidti* defines a range biozone corresponding approximately with the lower part of the Lochkovian, although in the Barrandian area, it ranges down through the graptolite interregnum into the top of the *M. transgrediens* biozone. Conodonts are not common in the somewhat argillaceous facies at Klonk.

This discussion could be extended considerably into a consideration of other faunal groups and some of the subsidiary species that define concurrent range zones. It is to be hoped, however, that by this time the reader can discern the main threads of a procedure that has now been followed innumerable times by many different workers.

Barnes et al. (1976) discussed the correlation between graptolites, conodonts, trilobites and brachiopods for the Ordovician rocks of Canada. This is an interesting case in that the benthonic fauna was used to establish a North American stage nomenclature in the craton (the Richmond, Maysville, etc.), whereas graptolites occur mainly in deeper water deposits and were correlated with the classic British stages (Ashgill, Caradoc, etc.), in spite of the difficulties of faunal provincialism across the Iapetus Ocean referred to earlier. To be of any regional use, these biostratigraphic schemes had to be integrated with each other, and this depended on finding locations, as at Klonk, where the different biofacies are interbedded. In order to cover the entire Ordovician system, it was necessary to study partial sections in the Canadian Rocky Mountains, Nevada, Texas, Newfoundland, the St. Lawrence Lowlands, and parts of the Canadian Arctic Islands. One of the key sections was the Lower to Middle Ordovician Cow Head Group of Newfoundland. At Green Point, where the beds are overturned, the stratotype for the Cambrian–Ordovician boundary has been established (Fig. 7.7; Cooper et al. 2001). Here, trilobites and conodonts occur in limestone boulders slumped from the shelf into the deep water basin, where graptolite-bearing shales buried them. Integration of the biozone schemes had to allow for the fact that the boulders were probably slightly younger than the enclosing shales. Elsewhere, transgressions and regressions caused the intermingling of faunas from different facies and different faunal provinces.

Another example of GSSP is that for the base of the Turonian at Pueblo, Colorado, the description of which was provided by Kennedy et al. (2005). The succession here consists of alternating limestones and shales (Figs. 7.43,

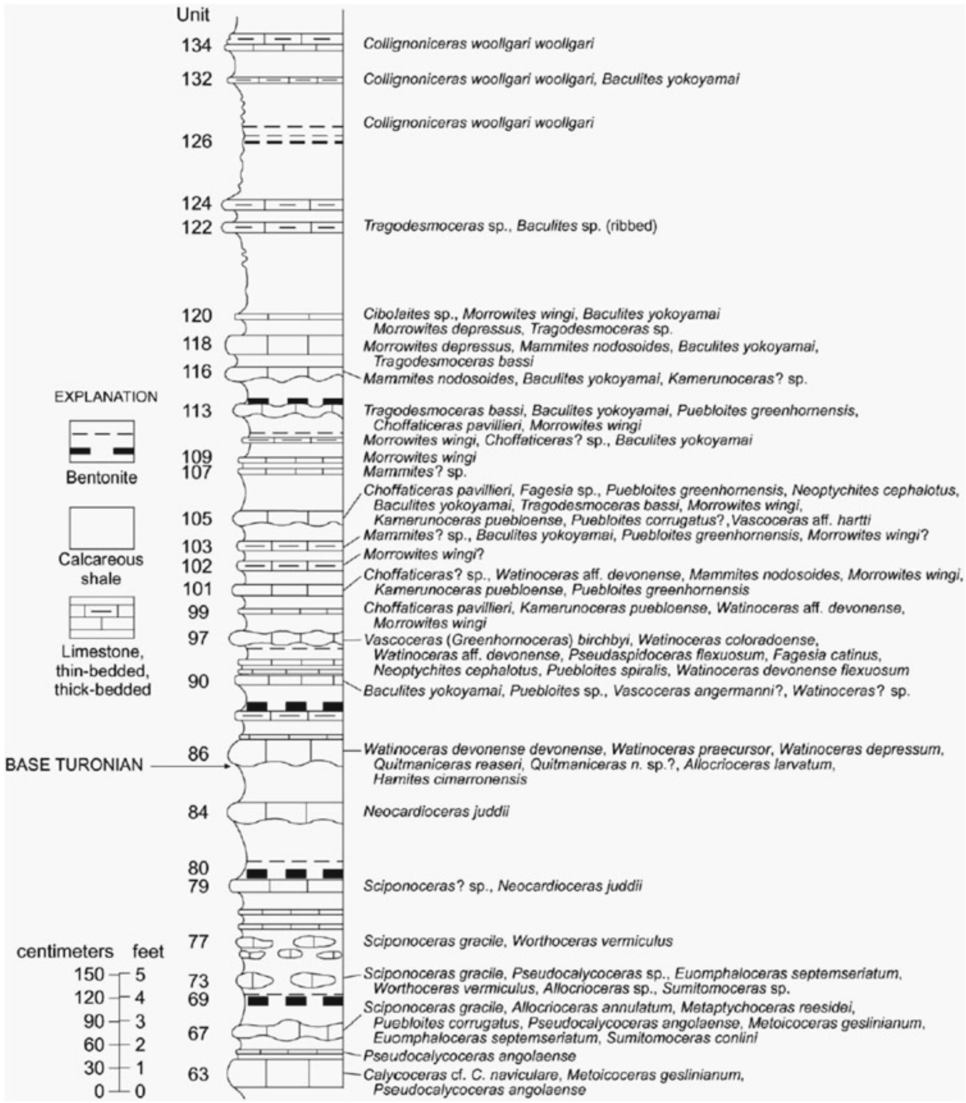


Fig. 7.43 The type section of the base of the Turonian. The lithological succession is that of the Bridge Creek Member of the Greenhorn Limestone on the north side of the Pueblo Reservoir State Recreation area, Colorado. The GSSP for the base of the Turonian is placed at the base of bed 86 (Kennedy et al. 2005, Fig. 8, p. 101)

7.44). The limestones are fossiliferous biomicrites. Much of the succession is bioturbated. The alternation of the two major lithologies is attributed to orbitally forced climatic cyclicity. The base of the Turonian has been placed at the base of bed 86, which corresponds to the first occurrence of the ammonite *Watinoceras devonense* (Fig. 7.43). There are many secondary biostratigraphic indicators of the boundary, including bivalves, dinoflagellates and other ammonites. Figure 7.43 shows the ranges of the inoceramids in this section, and the inoceramid and ammonite zones. The boundary is also indicated by a carbon-isotope excursion which can be correlated worldwide, and corresponds to an Oceanic Anoxic Event. The numerical age of the boundary has been derived from radioisotopic dating of several bentonite beds that be traced throughout much of the Western Interior Basin. The bentonites at the Pueblo location are too weathered to be dated, but Kennedy et al. (2005) provide dates determined from six other samples collected from correlative units in other parts of the basin.

7.8.4 Event Stratigraphy

The term **event stratigraphy** has been attributed to Ager (1973), although, as noted by Torrens (2002, p. 258), geologists have been aware of the importance of sudden events for some time. Typical geological "events" include sudden sedimentary events, such as storms and sediment gravity flows; volcanic events, generating widespread ash beds; earthquakes; biologic events such as first- and last-appearances of taxa, and mass extinction events; and chemostratigraphic events, such as "carbon excursions." Comprehensive treatments of this topic have been provided by Einsele and Seilacher (1982) and Kauffman (1988).

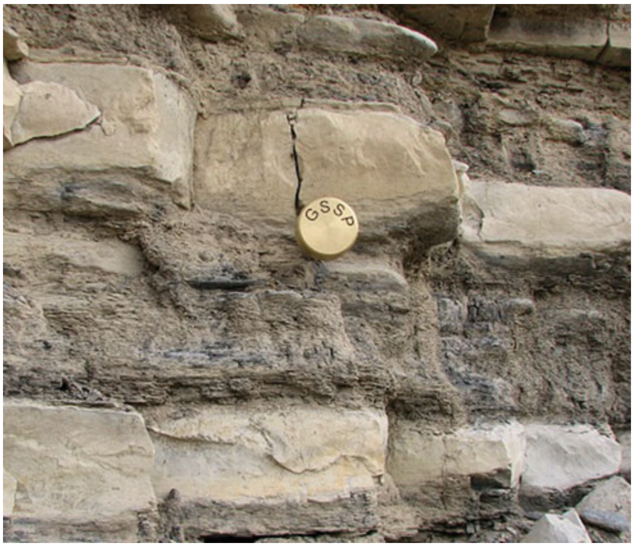


Fig. 7.44 The GSSP for the base of the Turonian. See Fig. 7.42 for stratigraphic location (photo by Brad Sageman)

Where the products of specific events can be recognized reliably, they provide invaluable markers for local and regional correlation. Bentonite layers, because they can be characterized by their petrology and radioisotopic age, have been used for this purpose for many years. Some events appear to be truly unique, such as the global K-T boundary event as the product of a meteorite impact, as first proposed by Alvarez et al. (1980); a hypothesis now almost universally accepted. However, there are many other types of events that are repeatable (e.g., major sediment-gravity flows; liquefaction events attributable to earthquakes), which means that their reliability as markers may be limited unless correlation can be substantiated by secondary means.

Kauffman (1988) documented in detail the methods of what he termed **high-resolution event stratigraphy,** in which he made use of bentonites, biomarkers and other events to construct regional stratigraphic correlations using the methods of graphic correlation, with an estimated accuracy of $\pm$ 100 ka. In Sect. 7.8.6 we discuss a debate regarding the use of event stratigraphy in the construction of the standard suite of GSSPs. We return to this topic in Sect. 8.12 as part of a brief introduction to modern research to develop the astrochronological time scale.

7.8.5 Absolute Ages: Their Accuracy and Precision

Time scales like the one in Fig. 7.33 look finished. It would appear that if fossils can be assigned to one or other of the stages shown in this diagram, it should be possible to determine the age of a section almost anywhere within the Phanerozoic to within a fraction of a million years. Often this is, in fact, the case, but the precision of the scale is deceptive. It represents geologists' best guesses, given all the available information at the time of compilation. It does not show the amount of interpolation and extrapolation that has gone into the construction of the scale. To get an idea about this, look at Fig. 7.45. This is a table that compares the assigned ages of the Mesozoic stages between eleven different published compilations. Each of these syntheses represents the research of the highest international standard.

Why are there so many variations in assigned age from one scale to the next? For example, the age of the Jurassic–Cretaceous boundary has moved up and down by 15.3 Ma years since 1982. How can this be?

The answer is that many of the assigned ages depend on extrapolation or interpolation from a limited number of well-known fixed points. It is rare for an important biomarker, such as a key FAD, to occur stratigraphically next to a datable ash bed or a magnetic reversal event. Many event ages are determined using the kinds of calculations shown in Fig. 7.35. A key event is bracketed by two or more dated horizons, and the age is worked out by assuming continuous sedimentation at a constant rate. Both these assumptions are commonly incorrect. Making the same kind of calculations at several locations where the same kinds of assumptions have to be made may result in a range of ages for the same event. The geologist then makes a best guess, picking the one that seems to be the most reliable, or averaging them in some way. Opinions may differ about which particular field locality or which dating exercise offers the most reliable tie point for the scale. In the Introduction to GTS2020 it is noted that "30% of Phanerozoic stage boundary ages have a change of their lower boundary by more than 0.5 Myr [relative to GTS2012], and in some cases much more" (Gradstein et al. 2020, p. 5). For example, some of the stage boundaries in the Early Cretaceous and Late Jurassic have shifted by more than 2 m.y. since the 2004 version of the GTS, and are characterized by potential errors of more than 0.5 m.y. Gradstein et al. (2020, p. 11) stated that "the new philosophy, which was started with GTS2004 and GTS2012, is to select analytically precise radioisotopic dates with high stratigraphic resolution. More than 330 radioisotopic dates were thus selected for their reliability and stratigraphic importance to calibrate the geologic record in linear time." Further, "Ages and durations of Cenozoic stages derived from orbital tuning are considered to be accurate to within a precession cycle ($\sim$20 ka) assuming that all cycles are correctly identified, and that the theoretical astronomical tuning for progressively older deposits is precise."

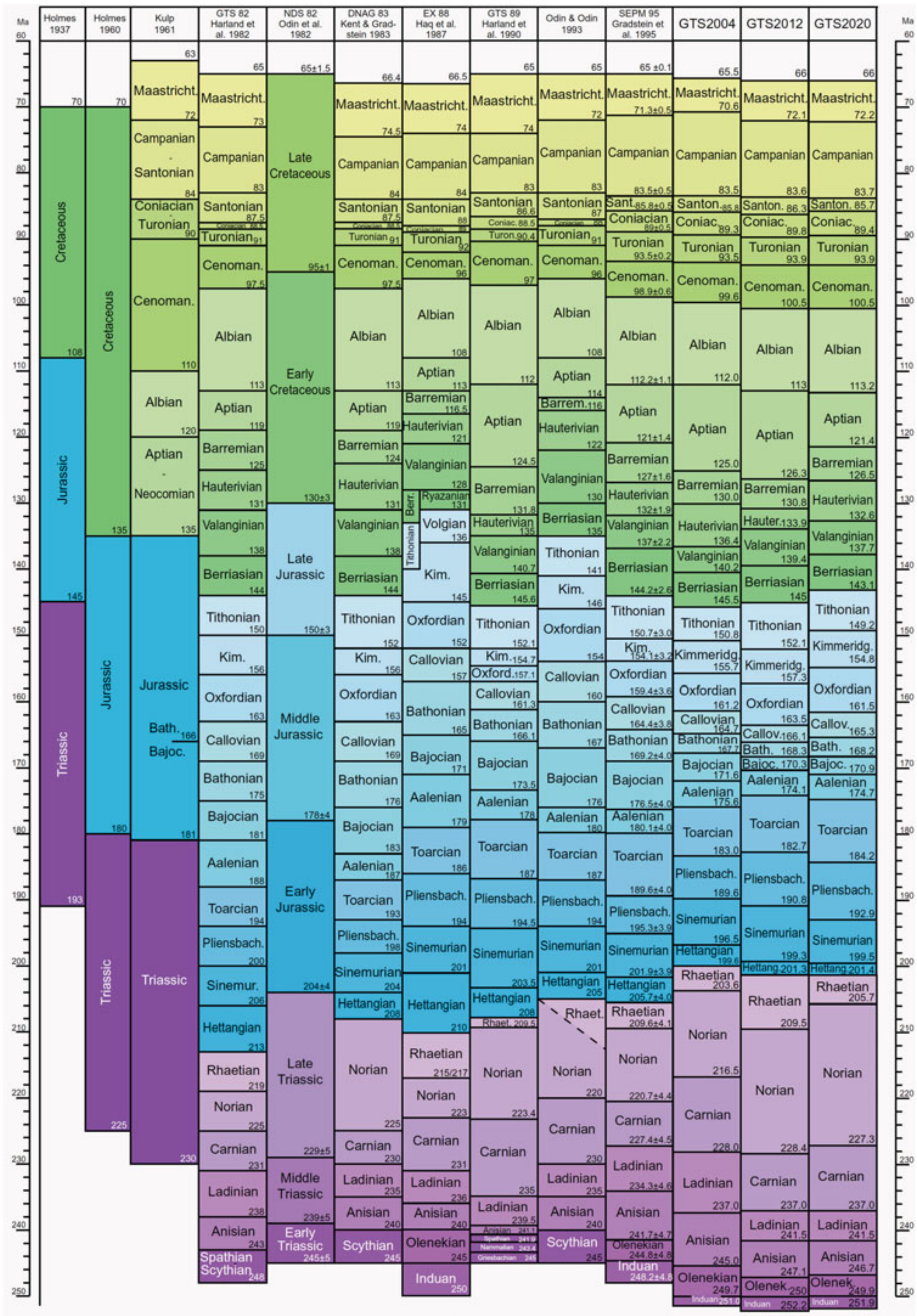
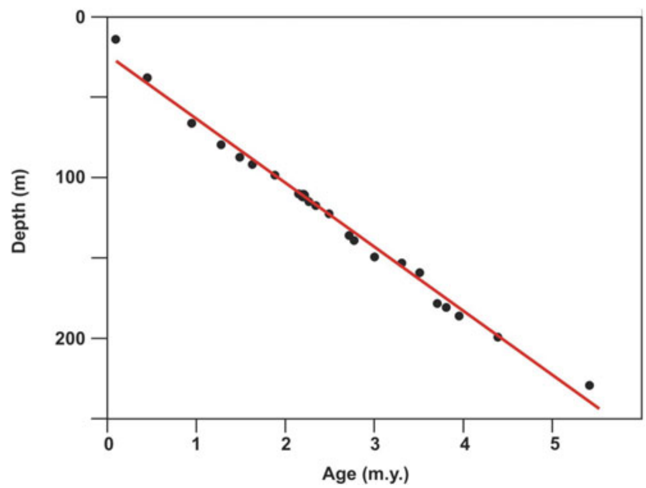


Fig. 7.45 Comparison of Mesozoic time scales (Gradstein et al. 2020, Fig. 1.4)

In the case of the Jurassic–Cretaceous boundary, Gale et al. (in Gradstein et al. 2020, p. 1024) stated:

> The Cretaceous is the only Phanerozoic system that does not yet have an accepted global boundary definition, despite over a dozen international conferences and working group meetings dedicated to the issue since the 1970s …. Difficulties in assigning a global Jurassic/Cretaceous boundary are the product of historical usage, the lack of any major faunal change between the latest Jurassic and earliest Cretaceous, a pronounced provincialism of marine fauna and flora and a concentration of previous studies on often endemic ammonites. Another problem is the occurrence of widespread hiatuses or condensations in many European and Russian epicontinental successions caused by the long-term "Purbeckian regression."



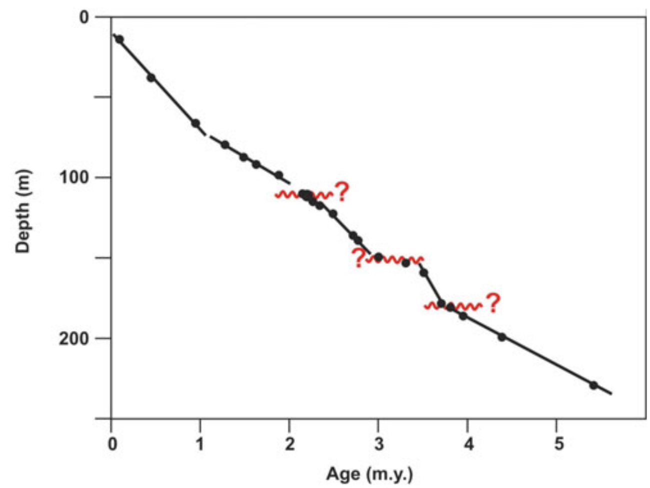


Fig. 7.46 The problem of calibration. The data points in both graphs represent the same set of real data from a well in the Gulf Coast of Florida (from Roof et al. 1991). Each point is a biomarker that has been dated against the available time scale. The question then arises, does this data represent continuous sedimentation at a constant rate? The straight line of correlation in the left-hand figure suggests that this is the case, making allowance for small errors in the dating of the fossils. But it is possible to look at the data a different way, as seen in the right-hand figure. Sedimentation rate may vary, and there may be several minor unconformities. Careful examination of the section is necessary to determine the nature of facies changes, the presence of breaks in sedimentation, and so on, in order to test the validity of the interpretations suggested in the right-hand diagram

This is a particularly severe example of the types of problem that have affected all the projects to develop an international agreement on the designation of GSSPs for stage boundaries.

In the absence of sufficient datable field locations, all kinds of assumptions have had to be made. For example, assumptions that the species of a given group of animals evolve at a constant rate so that their zones can all be assumed to be of the same duration. The first work to date the reversal events in the magnetostratigraphic scale, was carried out on the magnetized oceanic crust, and made use of an assumption that sea-floor spreading rates were constant so that the width of the reversal zone could be transformed into a value for its duration. This particular assumption has since been proved to be completely wrong—sea-floor spreading rates vary almost continuously.

Figure 7.46 illustrates the problem, using some real data from a well of the Florida coast. Which of the two interpretations is correct? Both will yield approximate ages for samples lying between the dated points, but the differences, though small, could be important. Where the slope between two points in this diagram is low, this means that the thickness of strata between the points is untypically small, given the indicated age span. A simple explanation would be that there is a local disconformity. An unusually steep slope could mean the occurrence of a rare event, such as the passage of a turbidity current, which locally thickens the stratal record. Careful reexamination of the original rock record could help to resolve this problem. This is a good example of why the **field context** of the rocks is important, as noted at the beginning of this book (Sect. 1.1).

Further developments in chronostratigraphic accuracy and precision are discussed in Sects. 8.10 and 8.11.

7.8.6 The Current State of the Global Stratigraphic Sections and Points (GSSP) Concept and Standardization of the Chronostratigraphic Scale

Despite the apparent inductive simplicity of the approach described here to the refinement of the time scale, the completion of the necessary suite of GSSPs has been slow, in part because of the inability of some working groups to arrive at an agreement (Vai 2001). In addition, two contrasting approaches to the definition of chronostratigraphic units and unit boundaries have now evolved, each emphasizing different characteristics of the rock record and the accumulated data that describes it (Smith et al. 2015). Castradori (2002) provided an excellent summary of what has become a lively controversy within the International Commission on Stratigraphy. The first approach, which Castradori described as the **historical and conceptual approach**, emphasizes the historical continuity of the erection and definition of units and their boundaries, the data base for which has continued to grow since the nineteenth century by a process of inductive accretion. Aubry et al. (1999, 2000) expanded upon and defended this approach. As noted by

Smith et al. (2015), in some cases, precedence and historical continuity have had to be set aside in favor of choosing new stratotypes that provide a greater data base for correlation purposes.

The alternative method, which Castradori terms the **hyper-pragmatic approach** (a very misleading label, in this writer's opinion), focuses on the search for and recognition of significant "events" as providing the most suitable basis for rock-time markers, from which correlation and unit definition can then proceed. The followers of this methodology (see the response by Remane 2000b, to the discussion by Aubry et al. 2000) suggest that in some instances historical definitions of units and their boundaries should be modified or set aside in favor of globally recognizable event markers, such as a prominent biomarker, a magnetic reversal event, an isotopic excursion, or, eventually, events based on cyclostratigraphy. This approach explicitly sets aside McLaren's (1970) recommendation (cited above) that boundaries be defined in places where "nothing happened," although it is in accord with suggestions in the first stratigraphic guide that "natural breaks" in the stratigraphy could be used or boundaries defined "at or near markers favorable for long-distance time-correlation" (Hedberg 1976, pp. 71, 84). The virtue of this method is that where appropriately applied it may make field recognition of the boundary easier. The potential disadvantage is that it places prime emphasis on a single criterion for definition, and relies on assumptions about the superior time-significance of the selected boundary event. The deductive flavor of the hypothesis is therefore added to the methodology. In this sense the method is not strictly empirical (as discussed below, assumptions about global synchroneity of stratigraphic events may in some cases be misguided. See Miall and Miall 2001). Very few "events" are likely to be global in scope, which means that where they are absent, boundary determination has to revert to the historical, inductive approach.

Smith et al. (2015) noted that even where the traditional approach has been used to define a GSSP, there has tended to be an overreliance on single biostratigraphic criteria for the definition, which may limit their usefulness and flexibility. To the key defining criterion of (for example) the first or last appearance datum of a chosen taxon could be added the stratigraphic distance above or below the FAD or LAD of other taxa. They also noted the importance of supplementary criteria.

The hyper-pragmatic approach builds assumptions into what has otherwise been an inductive methodology free of all but the most basic of hypotheses about the time-significance of the rock record. The strength of the historical and conceptual approach is that it emphasizes multiple criteria, and makes use of long-established practices for reconciling different data bases, and for carrying correlations into areas where any given criterion may not be recognizable. For this reason, this writer is not in favor of the proposal by Zalasiewicz et al. (2004) to eliminate the distinction between time-rock units (chronostratigraphy) and the measurement of geologic time (geochronology). Their proposal hinges on the supposed supremacy of the global stratotype boundary points. History has repeatedly demonstrated the difficulties that have arisen from the reliance on single criteria for stratigraphic definitions, and the incompleteness of the rock record, which is why "time" and the "rocks" are so rarely synonymous in practice (see also Aubry 2007, on this point; and Heckert and Lucas 2004, for other comments on the Zalasiewicz et al. proposal). Some of the current controversies surrounding the placement of GSSPs in the Cenozoic are discussed by Berggren (2007) and Walsh (2004). The latter paper also contains a lengthy discussion regarding the controversies surrounding the definitions and usages of the key terms, including stage, boundary stratotypes, GSSP etc., most of which are beyond the concerns of the practicing stratigrapher.

A different debate has arisen since the power of astrochronological calibration of the time scale became evident, a topic we take up in the next section, and again in Sect. 8.13. Astrochronology is based on the assertion that, in certain, carefully selected sections, a complete record of orbital forcing is preserved by cyclic variations in facies and thicknesses in the sedimentary record, and that by counting the cycles and correlating them to numerical ages derived from radioisotopic dating of some other means, a precise time scale with accuracy and precision in the 10^4-year range may be established.

As noted above, the traditional (at least since the 1970s) method of defining the time scale has been by the erection of GSSPs and topless stages, the purpose of the latter feature being to automatically allow for the presence of hidden hiatuses in the succession at the stratotypes. However, the relevance of this issue has been called into question when the method of selecting and dating a GSSP involves the necessary assumption of stratigraphic completeness, either in the actual stratotype or in the composite section upon which the stratotype is based. Such is the case with astrochronology. As noted by Hilgen et al. (2006, p. 117):

> … all late Neogene GSSPs are by now defined in land-based deep marine sections. All these sections have an integrated high-resolution stratigraphy, uniting detailed cyclo-, magneto- and biostratigraphies and have been astronomically tuned. Moreover, they cover the entire interval of the stage in a demonstrable continuous succession. As such, the sections perfectly embody the concept of a stage and may serve as unit stratotype for that stage in addition to accommodating its GSSP.

These authors argued for the extension of the concept to the remainder of the Cenozoic and also to the Mesozoic, as the astrochronological data base becomes more complete (Hilgen et al. 2015, 2020; see Sect. 8.11). They also pointed out that if the GTS is based on complete sections then the distinction between "rock" and "time" becomes unnecessary. However, the determination of "completeness" is always difficult, and as we now know, all sections contain some sedimentary breaks, although in deep-marine sections they may be of minor importance. I remain convinced that we need to be careful about abandoning the long-held cautions that support this dual terminology. The preservation of sections that are complete at the 10^4–10^5-year time scale is likely to be very unusual, requiring especially undisturbed basinal conditions, and the identification of the sections that are complete enough to be used in the establishment of an astrochronological time scale should be considered a rare event (see next section). The issue of the incompleteness of the stratigraphic record is discussed further in Chap. 8.

7.8.7 Cyclostratigraphy and Astrochronology

Cyclostratigraphy: the subdiscipline of stratigraphy that deals with the identification, characterization, correlation and interpretation of cyclic variations in the stratigraphic record.

Astrochronology: The dating of sedimentary units by calibration of the cyclostratigraphic record with astronomically tuned time scales.

Tuning: Adjusting the frequencies, including harmonics, of a complex record preserved in natural succession to best-fit a predicted astronomical signal.

Croll (1864) and Gilbert (1895) were the first to realize that variations in the Earth's orbital behavior may affect the amount and distribution of solar radiation received at the Earth's surface, by latitude and by season, and could be the cause of major climate variations. Several classic studies were undertaken to search for orbital frequencies in the rock record, and theoretical work on the distribution of insolation was carried out by the Serbian mathematician Milankovitch (1930, 1941), who showed how orbital oscillations could affect the distribution of solar radiation over the Earth's surface (the mathematical work of Milankovitch was so advanced and important for its time that an image of Milankovitch is now used on one of the Serbian currency bills). However, it was not for some years that the necessary data from the sedimentary record was obtained to support his model. Emiliani (1955) was the first to discover periodicities in the Pleistocene marine isotopic record, and the work by Hays et al. (1976) is regarded by many (e.g., de Boer and Smith 1994) as the definitive study that marked the beginning of a more widespread acceptance of orbital forcing, the so-called **Milankovitch processes**, as a major cause of stratigraphic cyclicity on a 10^4–10^5-year frequency—what is now termed the **Milankovitch band**. The model is now firmly established, particularly since accurate chronostratigraphic dating of marine sediments has led to the documentation of the record of faunal variations and temperature changes in numerous upper Cenozoic sections (an early summary was provided by Miall 2010, Sects. 7.2, 11.3. Later reports were given by Gradstein et al. 2004a; Hilgen et al. 2015). These show remarkably close agreement with the predictions made from astronomical observations. Many high-frequency sequence records are now interpreted in terms of the orbital-forcing model (summaries and reviews in Miall 2010, Chap. 11; Hilgen et al. 2015), and there is increasing evidence from detailed stratigraphic studies of orbital forcing and glacioeustasy in the Cretaceous record. We return to this work in Sect. 8.11.

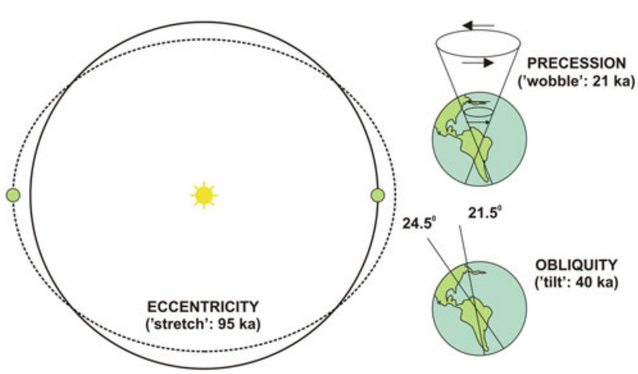


Fig. 7.47 Perturbations in the orbital behavior of the earth, showing the causes of Milankovitch cyclicity. Adapted from Imbrie and Imbrie (1979)

There are several separate components of orbital variation (Fig. 7.47). The present orbital behavior of the Earth includes the following cyclic changes (Schwarzacher 1993).

1. Variations in orbital **eccentricity** (the shape of the Earth's orbit around the sun). Several "wobbles," which have periods of 2035.4, 412.8, 128.2, 99.5, 94.9 and 54 ka. The major periods are those at around 405 and 100 ka.
2. Changes of up to 3° in the **obliquity** of the ecliptic, with a major period of 41 ka, and minor periods of 53.6 and 39.7 ka.
3. **Precession** of the equinoxes. The Earth's orbit rotates like a spinning top, with a major period of 23.7 ka. This affects the timing of the perihelion (the position of the closest approach of the Earth to the sun on an elliptical orbit), which changes with a period of 19 ka.

Imbrie (1985, p. 423) explained the effects of these variables as follows:

> Variations in obliquity alter the income side of the radiation budget in two fundamental ways: they modulate the intensity of the seasonal cycle, and they alter the annually integrated pole-to-equator insolation gradient on which the intensity of the atmospheric and oceanic circulations largely depend. (Low values of obliquity correspond to lower seasonality and steeper

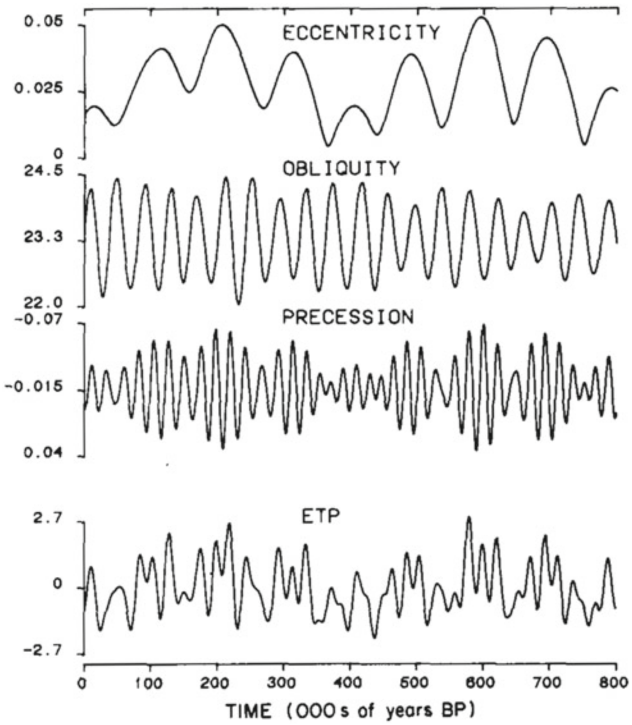


Fig. 7.48 The three major orbital-forcing parameters, showing their combined effect in the eccentricity-tilt-precession (ETP) curve at the bottom. Absolute eccentricity values are shown. Obliquity is measured in degrees. Precession is shown by a precession index. The ETP scale is in standard deviation units (Imbrie 1985)

> insolation gradients.) Variations in precession, on the other hand, alter the structure of the seasonal cycle by moving the perihelion point along the orbit. The effect of this motion is to change the earth-sun distance at every season, and thereby change the intensity of incoming radiation at every season.

Each of these components is capable of causing significant climatic fluctuations given an adequate degree of global sensitivity to climate forcing. For example, when obliquity is low (rotation axis nearly normal to the ecliptic), more energy is delivered to the equator and less to the poles, giving rise to a steeper latitudinal temperature gradient and lower seasonality. Variations in precession alter the structure of the seasonal cycle, by moving the perihelion point along the orbit. This changes the Earth–Sun distance at every season, thus changing the intensity of insolation at each season. “For a given latitude and season typical departures from modern values are on the order of ~5%” (Imbrie, 1985, p. 423). Because the forcing effects have different periods they go in and out of phase (Fig. 7.48). One of the major contributions of Milankovitch was to demonstrate these phase relationships on the basis of laborious time-series calculations. These can now, of course, be readily carried out by computer. The success of modern stratigraphic work has been to demonstrate the existence of curves of change in temperature, redox state, carbonate content, organic productivity and other variables in the Cenozoic record that can be correlated directly with the curves of Fig. 7.48. For this purpose, sophisticated time-series spectral analysis is performed on various measured parameters, such as oxygen-isotope content or cycle thickness. This approach has led to the development of a special type of quantitative analysis termed **cyclostratigraphy** (House 1985).

The Earth became highly sensitized to orbital variations during the cool climates of the Late Cenozoic, possibly as a result of a northern hemisphere cooling of air masses by the uplift of the Tibetan plateau (Ruddiman 1997), and it is now generally accepted, following the work of Hays et al. (1976) that the fluctuations in glaciation that characterized the Neogene were driven by orbital forcing, a process that gave us the oxygen isotope time scale (Emiliani 1955; Imbrie and Imbrie 1979; Imbrie 1985; Imbrie et al. 1984). The last ice age was ended by a phase of increasing solar insolation, which peaked about 10,000 years ago. A period of climatic warming, called the Holocene Optimum, existed from about 9000 to 6000 years ago. Since then there has been a very slow, long-term cooling trend (Fig. 7.56; possible anthropogenic influences are not discussed here).

To develop a time scale from the orbital record in sedimentary successions requires several important assumptions, given that, unlike, for example, bioevents or datable ash beds, cyclostratigraphy consists of a succession of identical, or near-identical, cyclic fluctuations in some primary depositional characteristic, such as carbon or calcium carbonate content, redox state, bed thickness or, more generally, facies. These assumptions include the following:

1. The section is continuous, or
2. (alternate): Discontinuities in the section can be recognized and accounted for in the subsequent analysis.
3. Sedimentation rate was constant, or event beds (such as turbidites) can all be recognized and discounted.
4. Orbital frequencies may be reconstructed for the distant past based on astronomical calculations of planetary motions.
5. Thickness can be converted to time using a simple sedimentation rate transformation.
6. The variabilities in stratigraphic preservation (facies changes, hiatuses) can be effectively managed by pattern-matching techniques.
7. Orbital frequencies can be reconstructed from the rock record of the distant geological past, based on independent age-bracketing of the section.

In general, these assumptions are more likely to be met in deep-marine and lacustrine settings, where it may be

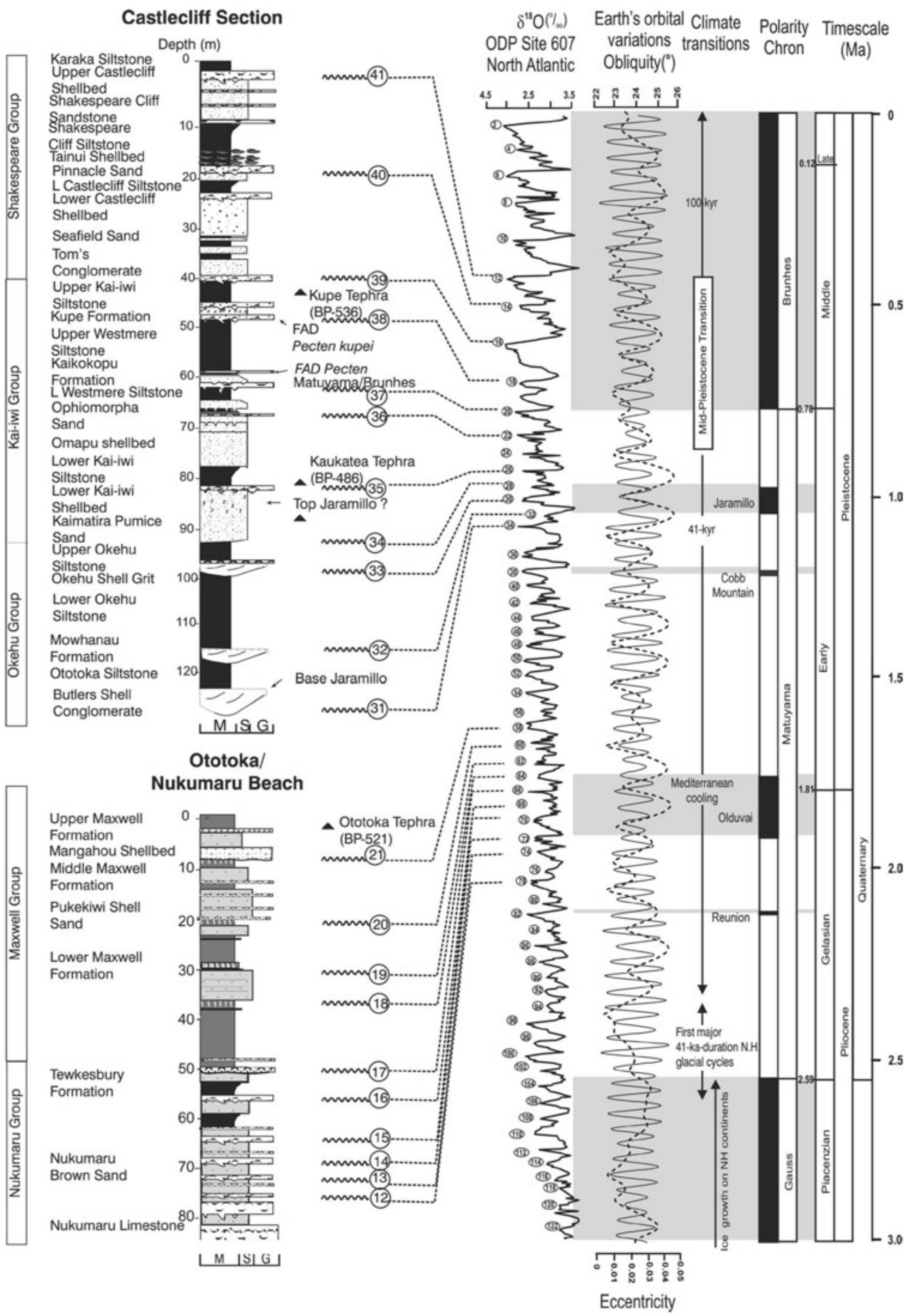


Fig. 7.49 Composite stratigraphic columns for the Nukumaru and Castlecliff coastal sections, Wanganui Basin, New Zealand, showing lithostratigraphy, sequence stratigraphy, and correlations with the oxygen isotope timescale (Naish et al. 2005). This particularly complete Pliocene to Pleistocene section serves as a regional standard for geological time (Naish et al. 2005)

expected that the allogenic forcing of sedimentary processes by orbital mechanisms might be expected to overwhelm local autogenic influences. However, one of the more convincing studies of a high-frequency sequence stratigraphic record being used as the basis for a cyclostratigraphic time scale is that of the shallow-marine Plio-Pleistocene cycles of the Wanganui Basin in North Island, New Zealand (Fig. 7.49). Also, Hilgen et al. (2015) cite several studies of fluvial systems in which it would appear that orbital forcing is the primary determinant of stratigraphic cyclicity.

The first major modern study of cyclostratigraphy (Hays et al. 1976) demonstrated that the last 800 ka of Earth time was dominated by a 100-ka cyclicity, as indicated by the cyclic climatic fluctuations from glacial to interglacial

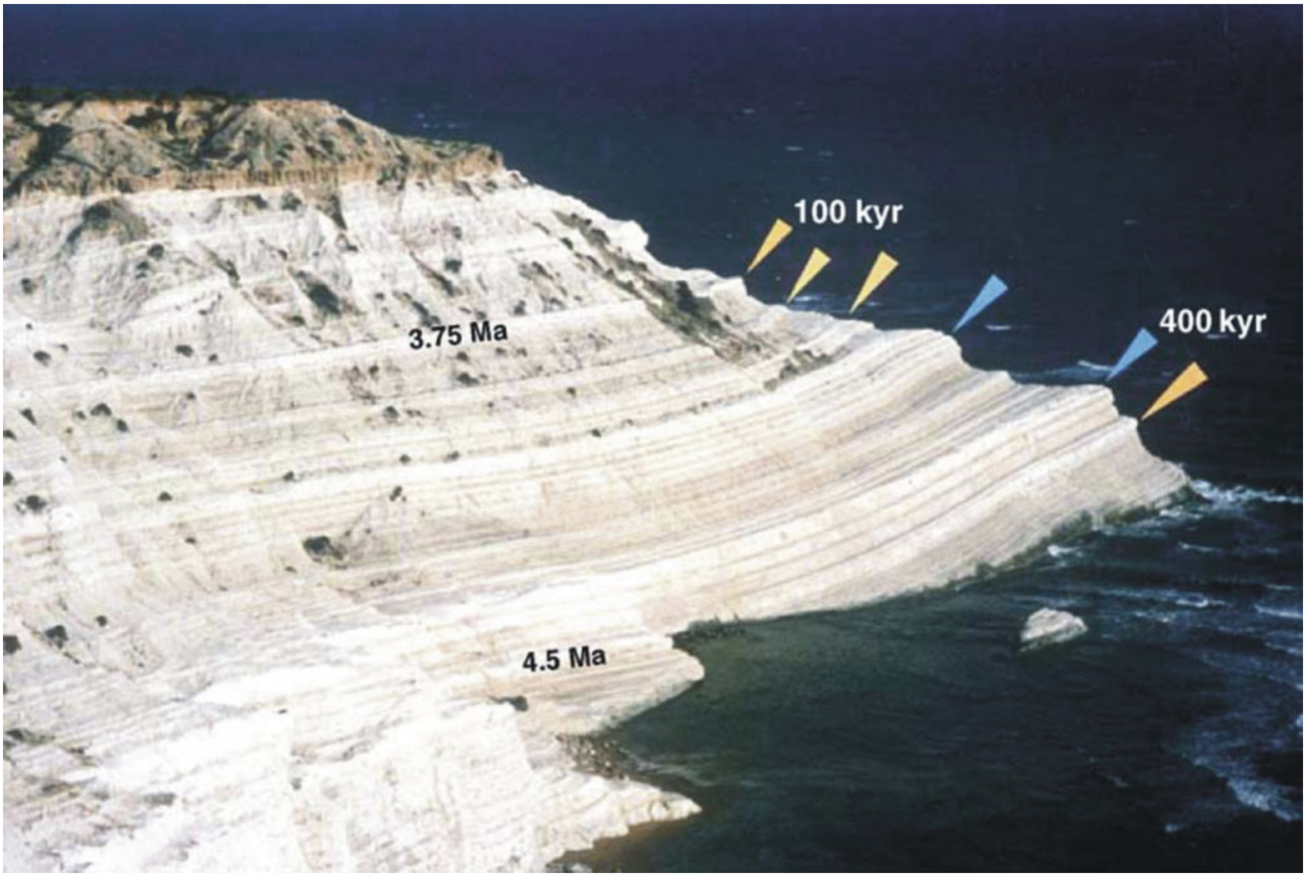
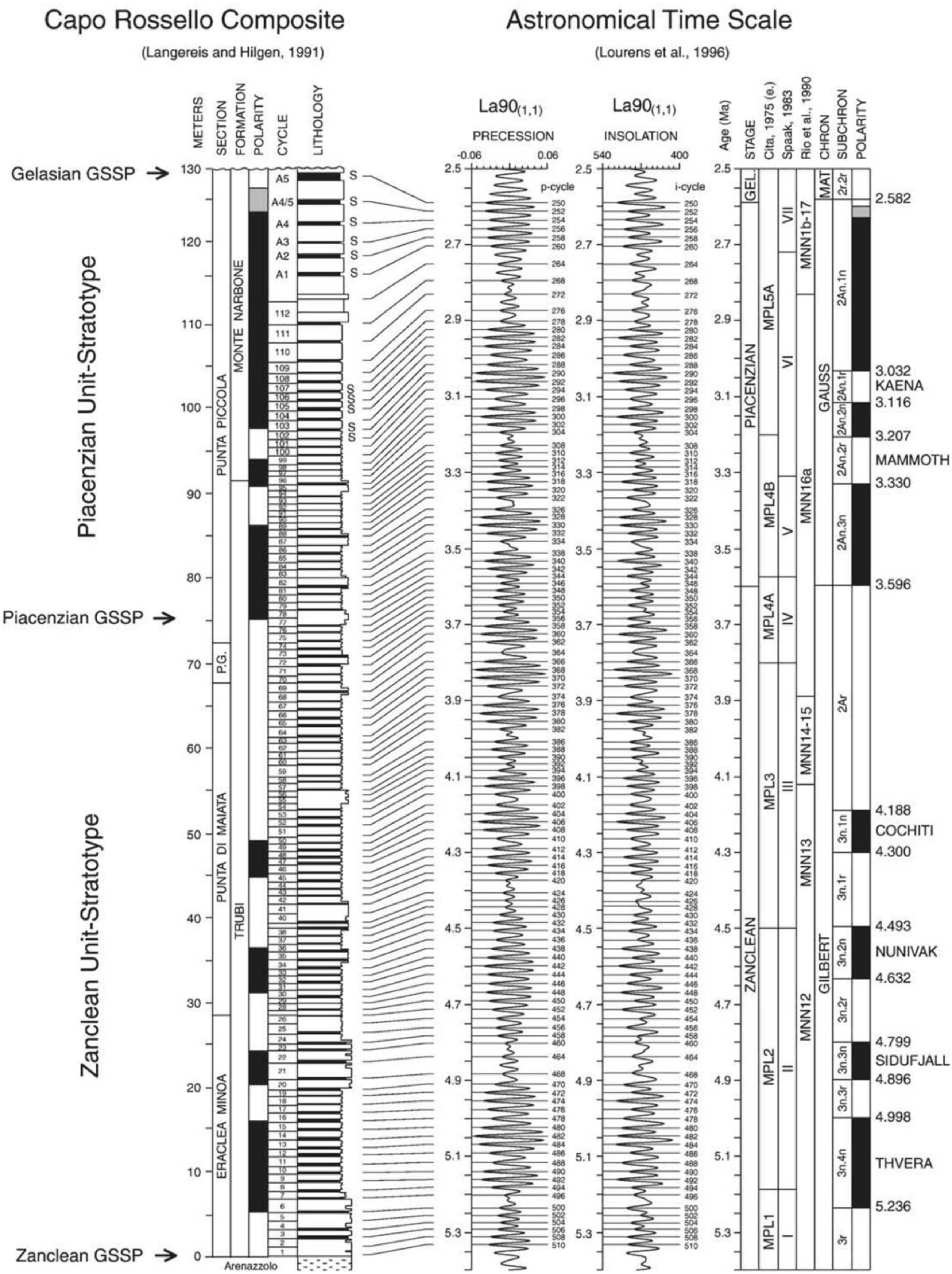


Fig. 7.50 The Punta di Maiata section on Sicily. Punta di Maiata is the middle partial section of the Rossello Composite and part of the Zanclean unit stratotypes, which defines the base of the Pliocene (Van Couvering et al. 2000; Hilgen et al. 2006). Larger-scale eccentricity-related cycles are clearly visible in the weathering profile of the cape. Small-scale quadripartite cycles are precession-related; precession-obliquity interference patterns are present in particular in the older 400-kyr carbonate maximum indicated in blue. All cycles have been tuned in detail and the section has excellent magnetostratigraphy, calcareous plankton biostratigraphy and stable isotope stratigraphy (Hilgen et al. 2006)

stages. These are recorded in fine detail by the oxygen-isotope record, particularly as this is measured in ODP cores (Fig. 7.39). In the longer term, extending back into the Paleogene and earlier, the 405-ka eccentricity cycle seems to be the most stable and the most likely to provide the basis for astrochronology.

There were several early focused attempts to examine the use of the orbital "pacemaker" as the basis for a high-precision time scale (Herbert et al. 1995; House and Gale, 1995; Shackleton et al. 1999), and several important regional studies were carried out that began to make a substantial contribution to the growth of this field of research. The stratigraphic and sedimentologic basis for this research was summarized by Miall (2010, Chap. 11). A range of indicators may be used to examine for orbital cyclicity, beyond the physical "cyclic" appearance of the rocks themselves. These include bed thickness, oxygen isotope ratios, weight-percent calcium carbonate or organic carbon, grayscale pixel data (from core scans), magnetic susceptibility, and resistivity data from a microimaging scanner.

Pioneering studies to establish an astrochronological time scale have been led by Fritz Hilgen. A reliable cyclostratigraphic (astrochronologic) time scale was first established for the youngest Cenozoic strata, back to about 5 Ma (Hilgen 1991; Berggren et al. 1995; Hilgen et al. 2006; see Figs. 7.50, 7.51). Over the succeeding decade, astronomically calibrated sections were used to extend the astrochronological time scale back to 14.84 Ma, the base of the Serravallian stage, in the mid-Miocene (https://stratigraphy.org), and research is proceeding to extend the time scale not only to the base of the Cenozoic, but through at least the Mesozoic (Hinnov and Ogg 2007; Hilgen et al. 2006, 2015). Figure 7.52 illustrates the correlation of Upper Miocene sections in the Mediterranean basin to the orbital scale. Westphal et al. (2008) offered a sharply critical review


Capo Rossello Composite
(Langereis and Hilgen, 1991)
Astronomical Time Scale
(Lourens et al., 1996)
METERS
SECTION
FORMATION
POLARITY
CYCLE
LITHOLOGY
La90(1,1)
PRECESSION
La90(1,1)
INSOLATION
p-cycle
i-cycle
Age (Ma)
STAGE
Cita, 1975 (e.)
Spaak, 1983
Rio et al., 1990
CHRON
SUBCHRON
POLARITY
Gelasian GSSP
Piacenzian GSSP
Zanclean GSSP
Piacenzian Unit-Stratotype
Zanclean Unit-Stratotype
PUNTA PICCOLA
MONTE NARBONE
P.G.
PUNTA DI MAIATA
TRUBI
ERACLEA MINOA
Arenazzolo
GEL.
PIACENZIAN
ZANCLEAN
MPL5A
MPL4B
MPL4A
MPL3
MPL2
MPL1
MNN1b-17
MNN16a
MNN14-15
MNN13
MNN12
MAT
GAUSS
GILBERT
2r.2r
2An.1n
2An.1r
2An.2n
2An.2r
2An.3n
2Ar
3n.1n
3n.1r
3n.2n
3n.2r
3n.3n
3n.3r
3n.4n
3r
2.582
3.032
KAENA
3.116
3.207
MAMMOTH
3.330
3.596
4.188
COCHITI
4.300
4.493
NUNIVAK
4.632
4.799
SIDUFJALL
4.896
4.998
THVERA
5.236

◀ **Fig. 7.51** The Rossello composite section (RCS, Sicily, Italy,) the unit stratotypes for series spanning the base of the Pliocene, incorporating the orbital tuning of the basic precession-controlled sedimentary cycles and the resulting astronomical time scale with accurate and precise astronomical ages for sedimentary cycles, calcareous plankton events and magnetic reversal boundaries. The Zanclean and Piacenzian GSSPs are formally defined in the RCS while the level that time-stratigraphically correlates with the Gelasian GSSP is found in the topmost part of the section. The well-tuned RCS lies at the base of the early–middle Pliocene part of the astrochronological time scale and the Global Standard Chronostratigraphic Scale and as such could serve as unit stratotype for both the Zanclean and Piacenzian Stage (Hilgen et al. 2006)

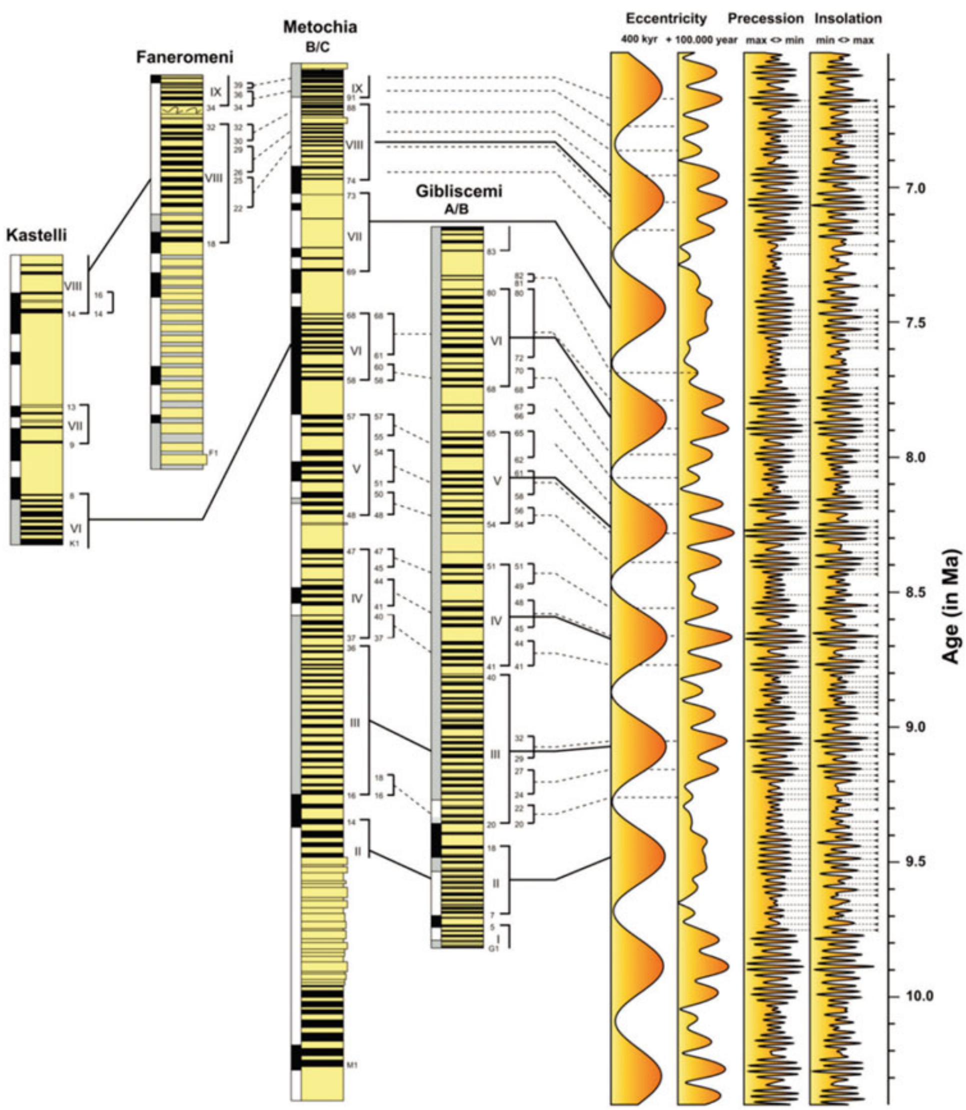


Fig. 7.52 Astronomical tuning of sapropels and associated grey marls in land-based deep marine sections in the Mediterranean for the interval between 10 and 7 Ma (upper Miocene). Colors in the lithological columns indicate sapropels (black), associated grey marls (grey) and homogeneous marls (yellow). Colors in the magnetostratigraphic columns indicate normal polarities (black), reversed polarities (white) and uncertain polarities (grey). Sapropels and associated grey marls have been numbered per section and lumped into large-scale groups (roman numerals) and small-scale groups. The initial age model is based on magnetobiostratigraphy. Phase relations between sapropel cycles and the orbital parameters/insolation used for the tuning are based on the comparison of the sapropel chronology for the last 0.5 myr with astronomical target curves (Hilgen et al. 2015, Fig. 2)

of the field data base on which part of the astrochronological time scale is based, pointing to problems of diagenesis and differential compaction of contrasting lithologies, that render direct one-for-one correlations between the critical field sections problematic. A careful comparison of the correlations between two critical field sections shows that the correlations of the astronomical cycles are not always supported by the correlations of bioevents. In some cases, there are different numbers of cycles between the occurrences of key bioevents. Multiple cross-checks are required to evaluate and, if possible, correct for such discrepancies.

Astronomical studies suggest that because of the long-term chaotic nature of the Earth's response to the gravitational influence of the planets, it is not possible to extend the present orbital frequencies back beyond about 60 Ma, with the exception of the eccentricity harmonic of

405 ka, which appears to have been stable through the Phanerozoic (Laskar et al. 2004). A modern treatment of the astronomical solutions is provided by Laskar (2020).

For the older part of the geological record (particularly the Mesozoic and Paleozoic), several studies have now established convincing "floating" scales for specific stratigraphic intervals; that is, scales that exhibit reliable orbital frequencies, once tuned, but that cannot be precisely correlated to the numerical time scale because of residual

Table 7.3 Major changes in Earth's history as revealed in the stratigraphy

#	Age	Process	Result	References
1	Late Archean	Generation of cratons	Appearance of shallow-marine and nonmarine environments, microbial life, stromatolites	Hawkesworth et al. (2017), Beall et al. (2018), Eriksson et al. (1998, 2013)
2	~2.4–2.0 Ga	Great Oxygenation Event	Increase in atmospheric oxygen led to oxidized environments, appearance of much red Fe^{3+} mineralization	Cloud (1968), Eriksson et al. (1998, 2013)
3	780–630 Ma	Snowball Earth?	Hypothesis of frozen Earth. Disputed on sedimentological and chronostratigraphic grounds	Hoffman and Schrag (2002), Allen and Etienne (2008), Eyles (1993, 2008), Eyles and Januszczak (2004), Le Heron et al. (2019)
4	555–500 Ma	Cambrian Explosion	Apparent sudden appearance of numerous new life forms.	Cloud (1948), Smith and Harper (2013)
5	~500–400 Ma	Great cratonic seas	Continental-scale cratonic seas caused by high sea levels. Widespread shallow-water carbonate sediments	Worsley et al. (1984), Pratt and Holmden (2008), Derby et al. (2012)
6	~400–250 Ma	Creation of Pangea (Caledonian orogenies)	Global changes in geology. Great unconformities and mismatch of biogeographic provinces. Old Red Sandstone.	Wilson (1966), Scotese (2001), Miall and Blakey (2019), Friend and Williams (2000)
7	~360–300 Ma	Gondwana Glaciation	Marine glacial deposits in Gondwana; cyclothems in the Northern hemisphere	Crowell (1978), Heckel (1986), Eyles (2008)
8	~350–300 Ma	Carboniferous System	Near-global extent of great forests. Source of much of the world's coal	Stanley (2005)
9	252 Ma	End Permian extinction	Dramatic change in global faunas	Esmeray-Senlet (in Gradstein et al. 2020, Chap. 3L)
10	~300–200 Ma	Breakup of Pangea	Nonmarine rift basins throughout Europe and eastern North America. New Red Sandstone	Ziegler (1988), Scotese (2001), Stanley (2005), Withjack et al. (1998)
11	Jurassic and Cretaceous	Oceanic anoxic events	Preservation of mudrocks rich in organic carbon	Schlanger and Jenkyns (1976), Schlanger et al. (1987), Cramer and Jarvis (in Gradstein et al. 2020)
12	~100-65 Ma	Late Cretaceous high sea levels	Continental flooding. Widespread chalk in southern US and NW Europe	Stanley (2005)
13	~100-65 Ma	Orbital forcing of climate, glacioeustasy	Cyclothemic stratigraphy	Elder et al. (1994), Plint and Kreitner (2007), Sageman et al. (2006), Shank and Plint (2013)
14	65 Ma	End Cretaceous impact and extinction	Iridium clay, tsunami deposits, shocked quartz	Alvarez et al. (1980), Hildebrand et al. (1991), Esmeray-Senlet (in Gradstein et al. 2020, Chap. 3L)
15	~40-5 Ma	Alpine and Himalayan orogeny	Creation of Alpine-Himalayan ranges, Tibetan Plateau, global cooling	Kennett (1977), Raymo and Ruddiman (1992), Ruddiman (2008), Summerhayes (2015)
16	2.5-0 012 Ma	Northern hemisphere Cenozoic glaciation	Multiple glacial episodes	Ruddiman (2008), Summerhayes (2015)
17	0.12-0.00 Ma	The Holocene	Variable post-glacial climates	Bradley (1999), Anderson et al. (2013), Plimer (2009), Carter (2010)
18	Post-WW2	The Anthropocene	Increasing dominance of human influence on global processes	

imprecisions in numerical dating methods (Hilgen et al. 2015).

Gradstein and Ogg (in Gradstein et al. 2020, p. 24) stated:

> Now, recent developments in integrated high-resolution stratigraphy and astronomical tuning of continuous deep marine successions combine potential unit stratotypes and boundary stratotypes for global stages as basic building blocks of the standard Global Chronostratigraphic Scale (GCS) (Hilgen et al. 2006, 2020). For the late Neogene with its outstanding orbitally tuned stratal record, some of the Global Stratotype Section and Point (GSSP) sections may also serve as unit stratotypes.

As this quote indicates, the pressure to recognized "unit stratotypes" and remove the distinction between "rock" and "time" is building. However, recent work on the generally fragmentary nature of stratigraphic preservation, which we detail in Chap. 8, means that this debate continues. A discussion of some current research in the area of astrochronology is provided in Sect. 8.11.

7.9 Stratigraphy Reflects Changing Earth Environments

The Earth's record of deep time is contained in its stratigraphy. The changing thermal structure of the planet, which led to the initiation of plate tectonics at about 3 Ga, the development of cratonic crust, the growth of continents by accretion and their migration, collisions and separations to form and disaggregate supercontinents, and Earth's ever-changing climate are all recorded in the deep structural geology and sedimentary cover of Earth. As sea-floor spreading moved continents through different climatic regimes, and continents collided, forming mountain belts that were uplifted, eroded and shed huge volumes of clastic debris, all was recorded more or less faithfully in the stratigraphic record. The evolution of plants led to dramatic changes in Earth's atmosphere and this, in turn, helped to fuel the appearance and diversification of animal life. At times, it seems, whole continents were characterized by particular types of sedimentary records and their contained fossils, reflecting the particular paleogeography and regional paleoclimate of the period. Some of the unique environmental conditions displayed by rocks from the distant past are discussed in Sect. 3.4.3. In this section we follow, very briefly, some of these changes that helped to create some particularly distinctive regional stratigraphies. Eighteen specific events and periods are discussed in this section, as summarized in Table 7.3

The evidence from the Earth's crust and modeling of mantle thermal behavior suggest that the distinctive form of mantle and crustal evolution that we group under the heading of plate tectonics began around 3 Ga (Hawkesworth et al. 2017; Beall et al. 2018). This was when the first cratons

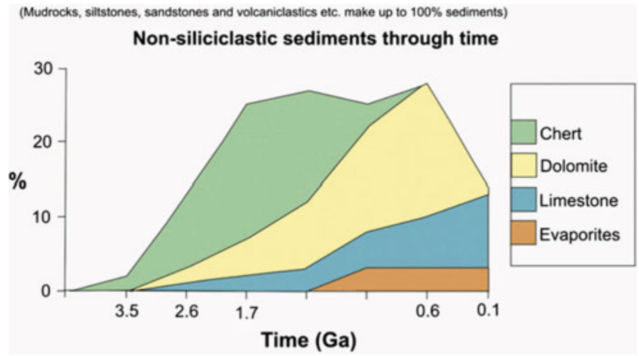


Fig. 7.53 The relative abundance of non-siliciclastic sediments through time, showing dolomites to be more abundant than limestones during much of the Proterozoic when microbial ecosystems dominated the biosphere (Eriksson et al. 2013, Fig. 7)

began to form; that is, the skins of stable sialic crust that rest isostatically on the mantle, just above sea level. We can recognize their appearance from the first evidence of shallow water, even nonmarine rocks in their sedimentary cover, where this has not been deformed out of recognition by subsequent tectonism or metamorphism. Eriksson and Mazumder (2020) provided an overview of modern work on Archean Earth processes based on recent research around the globe on preserved Archean structural belts that preserve protocontinental fragments. For example, the 3.55–3.2 Ga Barberton greenstone belt of southern Africa is characterized by "carbonaceous cherts with filamentous, spheroidal, and lenticular microstructures; traces of hydrothermal biofilms; pseudocolumnar stromatolites; large spheroidal microfossils; and apparently photosynthetic microbial mats" preserved in sediments of intertidal, supratidal and fluvial origin (Eriksson and Mazmunder 2020, p. 2).

Eriksson et al. (1998) provided an extensive overview of Precambrian sedimentation systems. The atmospheric composition changed significantly through the Precambrian. The Archean atmosphere consisted primarily of carbon dioxide, nitrogen and methane, with minor amounts of water, hydrogen, carbon monoxide and reduced sulfur gases. CO_2 content was much higher than at any time during the Phanerozoic, likely several thousand parts per million. Atmospheric oxygen began to be generated by photosynthesis when cyanobacteria evolved, somewhere around 3 Ga, but for hundreds of millions of years the quantities were modest, and the atmosphere continued to be dominated by the volcanic outgassing of CO_2, and gases of nitrogen and sulfur (Eriksson et al. 1998, 2013). Banded iron formation, generated in a reducing atmosphere, was an important sedimentary product during this phase (Archean to Paleoproterozoic; ~3700–2450 Ga). Photosynthesis increased in importance through the Paleoproterozoic, and about 2.4 billion years ago the atmosphere became essentially an oxidizing environment, with the increased formation and preservation of sedimentary ferric iron minerals (Cloud 1968). Red, yellow and brown-colored sandstones became common for the first time. This is called the Great Oxygenation Event, although the transition lasted for many millions of years, so this was not strictly an “event.”

Figure 7.53 provides estimates of the changing composition of the non-siliciclastic constituents of the sedimentary record through the Proterozoic and Phanerozoic. As can be seen, cherts and dolomite were abundant through the Proterozoic, with limestone and evaporite increasing in importance from the Neoproterozoic.

Regional glaciations occurred several times through the Precambrian, as evidenced by deposition of thick glaciomarine sediment-gravity-flow and ice-rafted deposits. Hoffman and Schrag (2002) focused on a series of widespread glaciations that took place during the Neoproterozoic, and developed the hypothesis of “Snowball Earth,” the central idea of which is that the Earth descended into a deep freeze that also partly froze the oceans and essentially shut down hydrological systems and biological evolution. This idea has been challenged by several groups of workers. Allen and Etienne (2008) focused on sedimentological evidence for the continuation of active marine sedimentation through the supposed snowball period, indicating the continuing presence of open oceans and the aqueous transport and deposition of siliciclastic sediments. Many supposed glacial deposits, on close sedimentological examination, prove to have little or no connection to glaciation (e.g., Kennedy et al. 2019). Eyles (1993, 2008) and Eyles and Januszczak (2004) argued that the Earth needed to be conditioned to widespread glaciation by regional uplift, which cooled the atmosphere. The evidence from the location and timing of major glacial episodes indicates that the uplift that occurs during continental separation and extension is often associated with regional glacial episodes. Chronostratigraphic data indicate that glaciation during the Neoproterozoic, while widespread, consisted of regional episodes occurring over an extended period of tens of millions of years, with no single intervals when the Earth could be said to be largely frozen (Le Heron et al. 2020).

At the beginning of the Phanerozoic, the atmosphere still contained an order of magnitude more CO_2 than at the present day, at nearly 7000 ppm (Fig. 7.54). The CO_2 level gradually dropped through the Paleozoic, largely because of the rise of land plants, which increased the extraction of this gas from the air for use as a nutrient source by photosynthesis (Summerhayes, 2015, p. 160). It has been demonstrated that, over the Phanerozoic, there has been no clear relationship between CO_2 levels and global temperature (Berner and Kothavala 2001; Davis 2017). The stratigraphic record contains qualitative paleoclimate indicators, such as certain plants and animals, which have an expected latitudinal range; for example, corals, which grow best between 30° north and south. Such sediments as glacial deposits, eolian sands and evaporites also have paleoclimatic significance if they are regionally widespread. When regional paleoclimates reconstructed using this evidence are related to their paleolatitude, based on reconstructions of the plate-tectonic history, the familiar global climate belts appear (Summerhayes 2015, Chap. 6). Global climates have

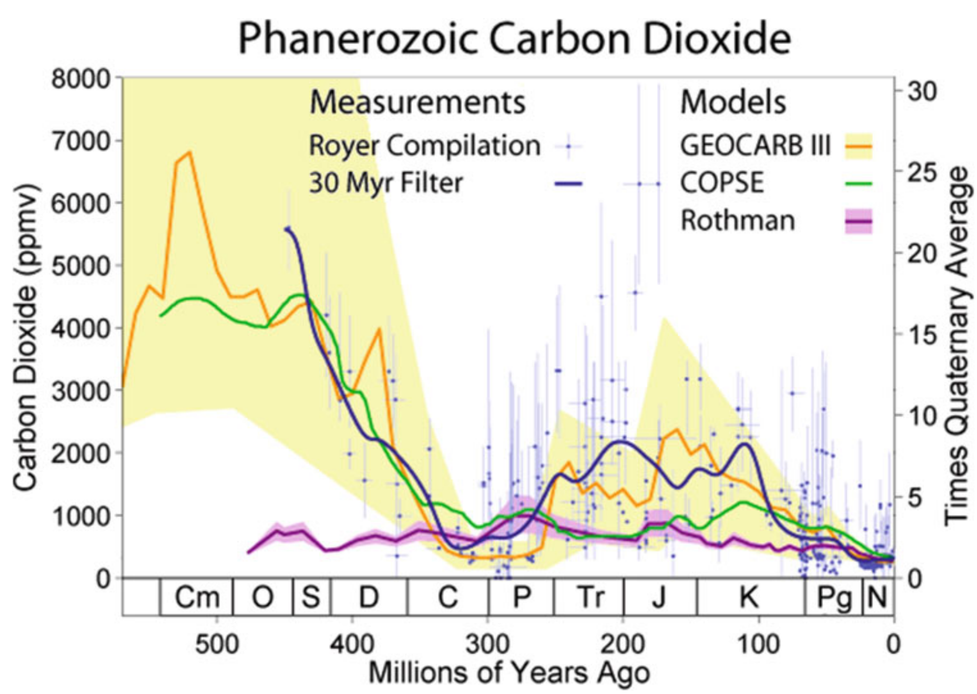


Fig. 7.54 Changes in atmospheric carbon dioxide composition of the atmosphere through the Phanerozoic. Compiled by Rhode (2019), https://commons.wikimedia.org/wiki/File:Phanerozoic_Carbon_Dioxide.png. GEOCARBIII is from Berner and Kothavala (2001)

experienced extremes of heat and cold, such as the glacial to interglacial fluctuations of the late Cenozoic. The terms "greenhouse climate" and "icehouse climate" were proposed for long-term Phanerozoic global climate states by Fischer (1984), and he related these to cycles of "high" and "low" CO_2, a model generally supported by Summerhayes (2015, Chap. 10). However, in detail, these models become hard to support, in part because modern data on the ancient atmospheric composition (Fig. 7.54) indicate a more complex history than was available to Fischer, and in part, modern stratigraphic data likewise reveal substantial short- to medium-term variability in climate through the Phanerozoic. We discuss some of these variations below.

One of the most distinctive features of the Phanerozoic sedimentary record that distinguishes it from the Precambrian is the abundance of fossil invertebrates. This led to the coining of the term "Cambrian Explosion" for the supposed burst of evolution that led to the faunal abundance (Cloud 1948). However, a detailed examination of the fossil record and the date of the first appearance of the various phyla and classes in the rock record reveals that the so-called explosion took tens of millions of years. For example, Smith and Harper (2013) summarized the enormous growth in diversity between the appearance of the Ediacaran fauna (645 Ma) and the completion of the first biodiversity expansion at the end of the Ordovician (443 Ma), a period of some 192 m.y. The first mollusks, phyoliths, brachiopods, archeocyatha and trilobites appeared over a span of some 15 million years in the Early Cambrian; hardly an explosion. Nonetheless, some explanations seem necessary for the eventual abundance of life forms. Smith and Harper (2013, p. 1356) suggested that a range of

> interacting processes generated an evolutionary cascade that led to the rapid rise in diversity. The initiating event is likely to have been the early Cambrian sea-level rise that led to inundation of continental margins and interiors and the rapid input of erosional by-products. This sea-level rise would also have generated a very large increase in habitable area lying between the base of wave turbulence and the depth to which light penetrates, providing a further driver for large increases in diversity.

Clearly, the processes of evolution had to have reached a point where spontaneous mutations led to advantages for organisms that were already poised for diversification. One of these advantages was the development of biomineralization, which provided both defensive and predatory hard tissue nearly simultaneously and, as a byproduct, greater preservability for exoskeletons in the fossil record.

Sea levels underwent a major long-term eustatic rise commencing at the end of the Neoproterozoic and continuing into the Ordovician, remaining at high levels until the Carboniferous (Vail et al. 1977; Fischer 1981). It is widely accepted that this was the result of the breakup of the supercontinent Rodinia, with accelerated sea-floor spreading that led to the production of large areas of young oceanic crust at isostatically shallow levels, and the resultant displacement of ocean waters in the form of a long-continued global transgression (Worsley et al. 1984). The early Paleozoic was characterized by some of the most extensive cratonic epeiric seas that the Earth has experienced. As described by Pratt and Holmden (2008) and Derby et al. (2012), among the stratigraphic products was thick and areally extensive shallow-marine carbonate deposits. In the United States the term "Great American Bank" has been used to describe this stratigraphy (Derby et al. 2012).

The construction of Pangea began with regional orogenies, such as the early Taconic orogeny in eastern North America in the late Cambrian. A long series of regional to continental orogenic episodes, collectively referred to as the Caledonian Orogeny, marked the suturing of continental fragments during the Cambrian to Carboniferous period. The major episode of continental accretion was the suturing of Laurentia with Baltica, the Scandian phase, which lasted from mid-Silurian to early Devonian (Scotese 2001; Miall and Blakey 2019). The closure of large oceans, such as the Iapetus and Rheic oceans brought long-separated and distinctive biogeographic provinces together, creating proximity between mismatched faunas and floras. One of the first of these mismatches to be recognized and mapped was that of early Paleozoic trilobite and graptolite provinces, which constituted part of the evidence used by Wilson (1966) to postulate the closing and reopening of the Atlantic Ocean, as discussed in Sect. 7.5.1 (Fig. 7.12). Stratigraphically, among the most distinctive products of this era was the deposition of the largely nonmarine Devonian Old Red Sandstone in a wide variety of syn- to post-orogenic tectonic settings within or adjacent to the Caledonian orogen, including Maritime and Arctic Canada, Svalbard, Greenland, the British Isles and Norway (Friend and Williams 2000). One of the most well known of these occurrences is at Hutton's famous unconformity at Siccar Point in southeast Scotland, where the sandstone rests on deep-marine Silurian arenites, close to the Laurentian-Baltic suture.

During the Paleozoic, the Gondwana continent, comprising Africa, South America, India, Australia and Antarctica, drifted across the south pole (Crowell 1978; Scotese 2001). Widespread continental glaciation occurred on these continental areas, commencing in Africa and South America in the Late Devonian, extending to southern Africa in the Mississippian and to India, Australia and Antarctica in the Pennsylvanian and Permian. Thick glaciomarine deposits are widespread in these continents (Eyles 2008). The near-polar locations of the continents constituted an important precondition for glaciation, but as Eyles (2008) emphasized, tectonism, leading to broad areas of regional uplift was also important in generating long-term cooling.

Stratigraphic research in Britain, Germany and North America led to the recognition of the cyclic nature of Carboniferous stratigraphy, and Wanless and Weller (1932) coined the term "cyclothem" for these deposits in the mid-continental United States (see Sect. 1.2.2). Shepard and Wanless (1935) attributed the cyclicity to glacioeustasy driven by the growth and decay of Gondwana ice caps, based on comparisons to the high-frequency glacial-interglacial climatic fluctuations documented for the late Cenozoic glaciation. This interpretation has been confirmed and extended by subsequent research. Heckel (1986) developed a sea-level curve for the midcontinent deposits based on the cyclic repetition and areal extent of key lithofacies, such as conodont-bearing shales, open-marine limestones, prograding fluvial-deltaic deposits and unconformities, many marked by paleovalleys indicating short-term base-level fall. The base-level fluctuations reconstructed for this curve suggest high-frequency sea-level cycles with periodicities comparable to modern orbital frequencies (Heckel 1986).

Coal of Carboniferous age occurs throughout northern Europe, Asia, and midwestern and eastern North America (Stanley 2005). In fact, the term "Carboniferous" comes from England, in reference to the rich deposits of coal that occur there. The evolution of land plants led to the first appearance of woody tissue and bark, which facilitated the rapid development and spread of widespread forests. Most of the coal is of Pennsylvanian age, the term now used for the Upper Carboniferous. It was this coal that was the basis for the Industrial Revolution, first in Britain toward the end of the eighteenth century, and then in continental Europe and North America.

There were five great biological extinctions during the Phanerozoic: (1) Late Ordovician, (2) late Devonian, (3) end-Permian, (4) end-Triassic, and (5) end-Cretaceous (Esmeray-Senlet, in Gradstein et al. 2020). Of these, the end-Permian extinction is considered the most severe, with 80% of marine genera and 75% of terrestrial genera extinguished. Among the major groups that were terminated at this time were eurypterid arthropods, trilobites, acanthodians and blastoid echinoderms. Other groups that were rendered nearly extinct, but which underwent a recovery during the Triassic include the ammonoids, brachiopods, corals, bryozoans, anthozoans, crinoids, gastropods, foraminifera and radiolaria. Over two-thirds of terrestrial labyrinthodont amphibians, sauropsid reptiles and therapsid proto-mammals also became extinct. Several kill mechanisms have been proposed for the end-Permian mass extinction, including carbon cycle disruption, ocean anoxia, ocean acidification, global warming, acid rain, ozone destruction and toxic metal poisoning. Many of these kill mechanisms were linked to the eruption and emplacement of the Siberian Trap Large Igneous Province, the largest volume of preserved continental basaltic magmatism generated in the Phanerozoic (Esmeray-Senlet, in Gradstein et al. 2020, p. 129). None of the proposed extinction mechanisms has widespread support at this time.

The breakup of Pangea commenced with a diachronous Late Permian–Triassic rift along the northern margin of Gondwana (Arabia–India–Australia), with the separation of a Cimmerian continent (Turkey–Iran–Tibet) and the opening of the Neotethyan ocean (Scotese 2001). Rifting extended to the site of the future north Atlantic Ocean in the Triassic, with the development of an enormous series of faulted rift basins extending from Florida to New England and Atlantic Canada, Greenland, North Africa and most of western Europe (Ziegler 1988). These were mostly located in low-latitude settings, and are characterized by nonmarine redbed deposits (fluvial, lacustrine and eolian deposits, some with evaporites). Some old stratigraphic terms, including the New Red Sandstone (Britain), and the Keuper and Buntsandstein (Germany), became familiar names for these deposits. Many of these basins subsided as flexural subsidence accompanied the opening of the Atlantic Ocean, and are now deeply buried beneath the resulting extensional-margin sedimentary wedge. See, for example, the transects across the Atlantic margins of North America prepared by Withjack et al. (1998).

As noted by Schlanger et al. (1987, p. 372): "One of the prime results of the Deep Sea Drilling Project during the 1970s was the discovery that the major ocean basins, during Cretaceous time, were the sites of deposition of sediments anomalously rich in organic carbon in comparison to the average organic-carbon content of Phanerozoic sediments." Schlanger and Jenkyns (1976) were among the first to note the "anomalous stratigraphic concentration of carbonaceous sediments, loosely described as 'black shales', and [they] came to the general conclusion that whatever the mechanism involved, the Cretaceous oceans, from roughly Hauterivian through Santonian time, were locally or regionally oxygen deficient" (Schlanger et al. 1987, p. 372). They termed these periods Oceanic Anoxic Events. At first, two broad time envelopes were recognized within which these events occurred, late Barremian through Albian and late Cenomanian through early Turonian time. Modern work (Cramer and Jarvis, in Gradstein et al. 2020) now recognizes a single Jurassic event (Toarcian) and five Cretaceous events. These "events" are marked by high $\delta^{13}C$ values in the ocean waters, caused by the enhanced preservation of ^{12}C in biomass deposited on the ocean floor. The explanation for these events is interpreted as the spread of oxygen-deficient waters in the world ocean, due to global high temperatures and low latitudinal temperature gradients which, coupled with the decreased solubility of oxygen in these warm waters,

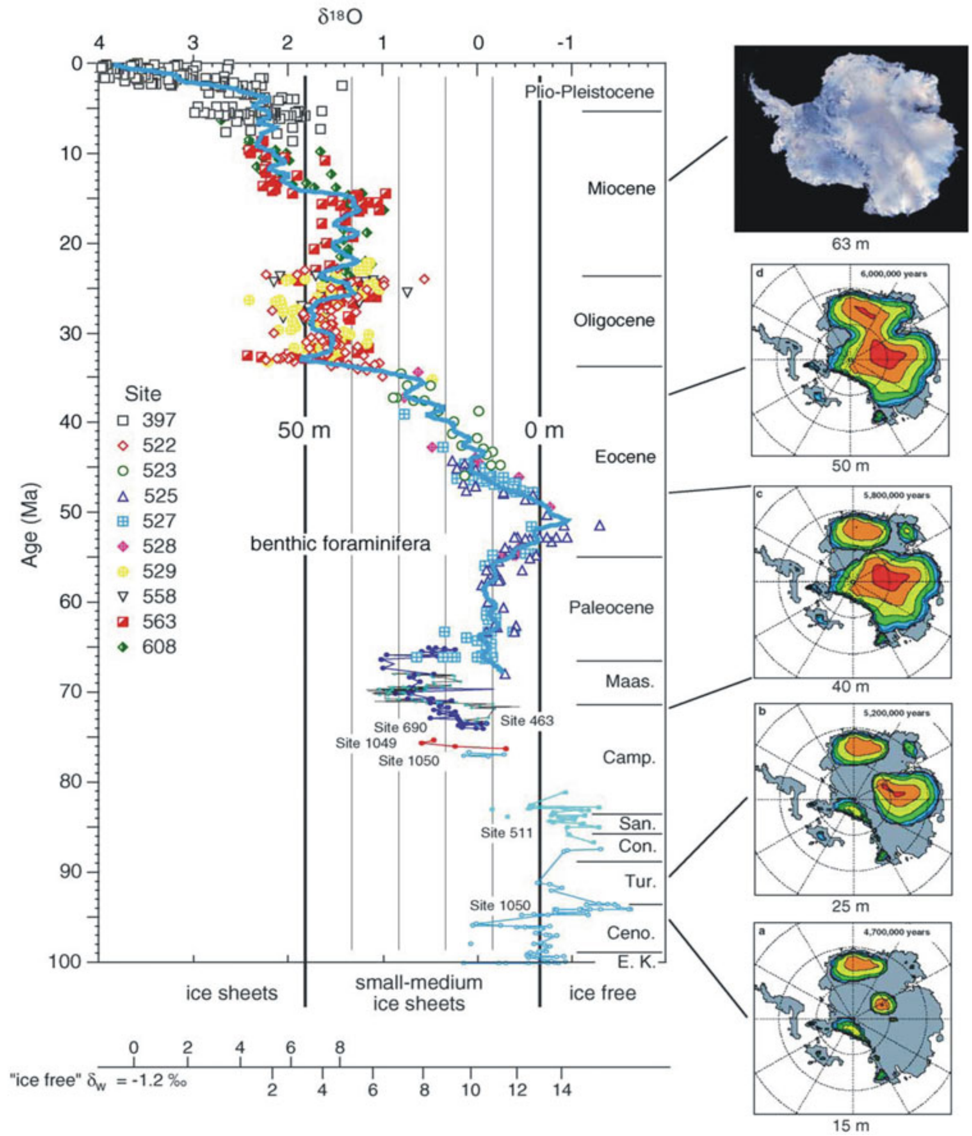


Fig. 7.55 The long-term trend in $\delta^{18}O$ values from the mid-Cretaceous to the present, as measured in deep-sea cores. The computed relationship between $\delta^{18}O$, ocean temperature and continental ice volume provides the basis for the estimates of changing long-term Antarctic ice cover. Not shown are the high-frequency glacial to interglacial fluctuations generated by orbital forcing (from Miller et al. 2005)

decreased the rate of reoxygenation of bottom waters (Fischer and Arthur 1977). A link with the great Cretaceous transgressions has also been suggested (Schlanger and Jenkyns 1976). It is now realized that these event beds are among the most important source beds for the world's petroleum.

The Cretaceous transgressions constituted the second long Phanerozoic period of high global sea level, and are attributed to the active sea-floor spreading that led to the breakup of Pangea (Worsley et al. 1984). Once again, as during the early Paleozoic, the world's cratons were flooded, leading to the development of widespread epeiric seas. It has also long been thought that much of the Cretaceous period was characterized by a greenhouse climate, that is, one that is significantly warmer than today, and globally more equable (Stanley 2005). Much of western Europe was covered by epeiric seas warmed by equatorial currents flowing westward from the Tethys Ocean that lay between Africa and Eurasia. One of the most distinctive sedimentary products of this period is the Upper Cretaceous Chalk, famous for forming the white cliffs of "Albion" (from the Latin word for white) along the south coast of England (Stanley 2005). This unit underlies most of southeast England and also occurs through much of western Europe, including France, Germany and Denmark. Its formation is attributed to the flourishing of microscopic organisms, notably the coccoliths, in warm-temperate marine waters. A comparable unit is the Austin Chalk of southern Texas, also of the Late Cretaceous age.

The long-term persistence of greenhouse climates began to be questioned when oxygen isotope measurements suggested cool periods during the Cenomanian and the

Maastrichtian. High $\delta^{18}O$ values were compiled by Miller et al. (2005) and led to the suggestion that there may have been small, short-lived ice caps on Antarctica at several times during the Cretaceous (Fig. 7.55).

Stratigraphic evidence for cyclic successions of cyclothem type, with durations and periodicities suggesting orbital frequencies have been mapped in several areas. Elder et al. (1994) correlated basinal marl-shale cycles of Turonian age in Kansas with prograding shoreface clastic successions in Utah. Plint and Kreitner (2007) traced thin sequences bounded by marine flooding surfaces of Cenomanian age without changes in thickness or facies across syndepositional structural elements in parts of the Alberta Basin, and likewise suggested orbital control. In both cases, glacioeustasy is suspected. In a quite different setting, Sageman et al (2006) defined an orbital time scale and a new C-isotope record for a Cenomanian–Turonian boundary stratotype in Colorado (see Sect. 7.8.3 and Figs. 7.43 and 7.44). Shank and Plint (2013) constructed a regional framework for the Turonian Cardium Formation across southern Alberta and began the work of linking the high-frequency allostratigraphy to the Colorado sections. Lin et al. (2021) were able to estimate the amplitude of glacial sea-level change at up to 50 m from backstripped cross-sections of the Upper Cretaceous Gallup Sandstone in New Mexico (see Sect. 8.12). The influence of glacioeustasy in the Late Cretaceous no longer seems in doubt.

The second of the five great Phanerozoic extinction events was that which brought the Cretaceous to a close. On land, land dinosaurs and pterosaurs disappeared; many bird, lizard, snake, insect and plant groups underwent drastic changes. In the oceans, $\sim$75% of species and $\sim$40% of genera became extinct (Esmeray-Senlet, in Gradstein et al. 2020, Chap. 3L). The discovery of anomalously high abundances of the rare earth iridium and other platinum-group elements at the boundary between the Cretaceous and the Cenozoic led to the hypothesis that a large bolide had impacted the Earth, creating a major environmental catastrophe, including tsunamis, wildfires and a lengthy "nuclear winter" (Alvarez et al. 1980). The discovery of a major crater located offshore from the coast of Yucatan, Mexico suggested a possible site for the impact (Hildebrand et al. 1991), and subsequent studies have confirmed the likelihood of this interpretation. Other interpretations, including environmental perturbations caused by large volcanic eruptions, have not withstood detailed examination. The boundary event was initially referred to as the K-T boundary, after the abbreviations for the Cretaceous and Tertiary, but with the abandonment of the term Tertiary for reasons of chronostratigraphic consistency, the boundary is now abbreviated as the K-Pg event (Pg = Paleogene).

The Earth underwent a short but pronounced warming episode during the late Paleogene and early Eocene, the causes of which are obscure. Following this, at about 50 Ma, as indicated by $\delta^{18}O$ values, the Earth commenced a dramatic long-term cooling, which initially led to the development of major ice caps on Antarctica, and then, around the beginning of the Oligocene, to ice coverage of almost all of the continent, comparable to the ice coverage today (Fig. 7.55). At around 5 Ma a further sharp increase in $\delta^{18}O$ values occurs, and this is associated with the commencement of the major northern hemisphere glaciation, characterized by repeated glacial to interglacial fluctuations on orbital time scales, up to the interglacial of the present day (Fig. 7.39). A number of processes have been suggested for this long-term cooling trend, but there is no general agreement on the causes (Stanley 2005; Ruddiman 2008; Summerhayes 2015). Kennett (1977) suggested that the separation of South America from Antarctica, and then the further separation from Australia during the breakup of Pangea, would have gradually isolated the Antarctic continent over the south pole, cut off the flow of warm ocean waters from low latitudes, and created the vigorous circum-Antarctic air and ocean currents that keep the continent cold. Plate tectonic reconstructions indicate that complete separation between the Antarctic and Australia occurred around 30 Ma. The pronounced cooling trend suggested in Fig. 7.55 began at about 34 Ma, the lag between these dates perhaps indicating the gradual buildup of the circum-Antarctic currents once rifting began.

Different explanations have been offered for the further cooling that commenced at about 5 Ma. Raymo and Ruddiman (1992) based their proposal on the supposed importance of the atmospheric concentration of carbon dioxide in controlling global temperature variations. They suggested that the uplift of the Tibetan plateau and the Himalayan ranges following the India-Asia plate-tectonic collision would have exposed an unusual area of the Earth's crust to vigorous weathering, a process that takes carbon dioxide out of the atmosphere and therefore, according to this hypothesis, could have led to major cooling. The emergence of these elevated areas in the midst of the global air circulation patterns would, by itself, have tended to cool the atmosphere, so it is possible that multiple processes were at work.

As recently as the 1960s mainstream interpretations still followed the hypothesis of Penck and Bruckner (1909), that there were four major glaciations in Europe, named the Gunz, Mindel, Riss and Wurm phases. However, since the development of oxygen isotope stratigraphy, and particularly since the recovery of continuous Late Cenozoic sections from deep-sea cores, it has been realized that there have been multiple glaciations. The current $\delta^{18}O$ chronostratigraphic chart contains 104 Quaternary cycles and 105 for the Pleistocene (Fig. 7.39). The sparse fourfold chronology initially developed in Europe in part reflects the very fragmentary nature of the record of continental glaciation, a

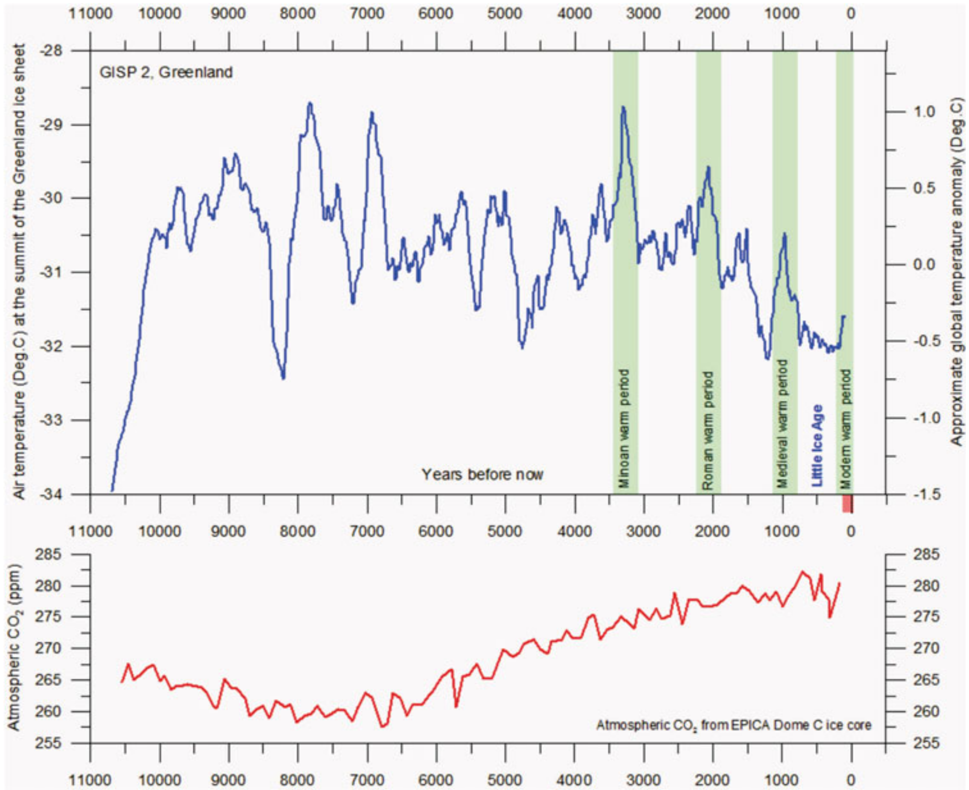


Fig. 7.56 Top: Holocene temperature variations over central Greenland, based on the GISP2 ice core. Bottom: atmospheric CO_2 measurements, from the EPICA Dome C ice core. Diagram from http://www.climate4you.com/ (downloaded April 9, 2021). Ice core data ends in the mid-nineteenth century. Dashed red line in the temperature graph shows the beginning of the modern warm period

process that is predominantly erosional, and in part reflects the lack of accurate dating methods until the advent of the oxygen isotope method coupled with radioisotopic dating.

Until about 1 Ma the $\delta^{18}O$ values fluctuated on an approximately 41 ka cycle, corresponding to the obliquity period. At around 1 Ma the amplitude of the cyclicity became much more pronounced and the periodicity shifted to a 100-ka cycle, modulated by 400-ka periodicity, corresponding to the eccentricity cycle and its prominent harmonic. The reasons for this are still not well understood (Ruddiman 2008; Summerhayes 2015) and are, in any case, beyond the scope of this book.

The Holocene epoch should be of particular interest to Earth scientists, because this is the era, since the end of the last ice age, about 12,000 years ago, when human society underwent its rapid evolution to the present complex industrial era in which we live. In particular, this era offers the opportunity to compare past natural climate variations with the climate of the present day, under long-term geological conditions, such as plate-tectonic setting, the elevations of mountains and continents, ocean currents, weather systems, etc., that have barely changed in 12,000 years. H. H. Lamb was the pioneer in this field, and his many books and articles helped to establish such terms as the Medieval Warm Period and the Little Ice Age by the 1970s (e.g., Lamb 1972). Yet the approach of the Earth science community to this question since Lamb has been curiously ambiguous. Some paleoclimatologists have published very detailed treatises on the Holocene (e.g., Bradley 1999; Anderson et al. 2013), yet the crucial question: what can we learn from Holocene climate change to help us to understand present-day variability, is typically not directly addressed.

Figure 7.56, which is from the website of Ole Humlum (http://www.climate4you.com), is an exception, and a very similar graph was published by Lewis and Maslin (2015, Fig. 2). It compares two post-glacial data sets derived from ice core data, temperature and atmospheric CO_2 composition. The temperature data are from Greenland and the CO_2 are from Antarctica. It is not to be expected that data sets from opposite sides of the Earth should be directly correlated because atmospheric and oceanic processes are commonly hemispheric and out of phase; yet the long-term (>10,000 year) trends that this illustration reveals are in direct opposition to those which the anthropogenic global warming model would predict. Atmospheric CO_2 shows a slow trend of modest decrease from 10,500 to 7,500 years ago, followed by a slow but steady increase. The Greenland temperature trend shows the opposite: a dramatic post-glacial increase of 5° from 11,000 to 8,000 years ago, followed by a slow decrease to the very recent. The long-term temperature trend illustrated in Fig. 7.56 is attributed to orbitally forced changes in solar insolation.

There are many oscillations into short warm and cool episodes, many of which can be correlated to historical trends in human development There is an ongoing debate about how "regional" versus "global" these episodes are, but it is generally acknowledged that during the Holocene Optimum that peaked about 8000 years ago regional continental ice caps were much less extensive than at the present day. In the Canadian Rocky Mountains, ice caps and glaciers did not reappear until the "Neoglacial" period, about 3000 years ago (Rutter et al. 2006).

These Holocene temperature variations are in direct contradiction to a long-lasting impression among some geologists and the wider public, that Holocene climates were stable. Raikes (1967) is quoted by Anderson et al. (2013, p. 155) as stating that "From at least 7000 BC, and possibly earlier, the worldwide climate has been essentially the same as today." This is the perception held by the public today. For example, Thomas Friedman, a very influential columnist with the New York Times, based one of his regular columns, on October 7 2015, on the work of Johan Rockström (Director of the Stockholm Resilience Center) and Mattias Klum. Quoting from their new book "*Big World, Small Planet*," Friedman stated:

> It's only been in the last 10,000 years that we have enjoyed the stable climate conditions allowing civilizations to develop based on agriculture that could support towns and cities. This period, known as the Holocene, was an 'almost miraculously stable and warm interglacial equilibrium, which is the only state of the planet we know for sure can support the modern world as we know it.' It finally gave us "a stable equilibrium of forests, savannahs, coral reefs, grasslands, fish, mammals, bacteria, air quality, ice cover, temperature, fresh water availability and productive soils. 'It is our Eden,' Rockström added, and now 'we are threatening to push earth out of this sweet spot.'

Anderson et al. (2013, p. 155) strongly disagreed. They state that, based on all the paleoclimatological evidence, which is described in detail in their book, "clearly … the concept of a stable Holocene environment is quite untenable." It is my contention that were Holocene climates fully discussed and understood by the general public, the current obsession with the urgency of climate change would never have developed, because there is nothing about present-day climates, including temperature extremes and severe weather events, that has not been preceded at some time during the Holocene. It is likely that the current global temperature regime is no warmer than that which prevailed during the Medieval warm period, at around 1000 AD. For a more balanced view of Holocene climates, see Plimer (2009) and Carter (2010).

The concept of the "Anthropocene" has received a great deal of attention in recent years, defining a period in Earth history when human influence on surface processes became global in scope. Such issues as pollution, deforestation, loss of habitat and biodiversity, climate change, extreme weather events, and possibly accelerated rise in global sea level owing to the melting of continental ice caps, have led some specialists to predict impending catastrophe. Mungall and McLaren (1990) referred to a "planet under stress." Crutzen (2002) proposed the term **Anthropocene**, to define a new geological era, and some Earth scientists have proposed that the term be formally defined and added to the geologic time scale (Zalasiewicz et al. 2008). There is no question that human influence has been profound (e.g., Ruddiman 2013; Gibling 2018; Koster 2020), and the term is a useful one, in the same sense that the "Renaissance" identifies a period of political and artistic liberation in European history. But to incorporate the term into the geological time scale is, in this author's opinion, an unnecessary step that would be a misuse and misunderstanding of the purpose of the GTS. Finney and Edwards (2016) discussed the proposal from the perspective of the International Commission on Stratigraphy, and suggested that the ICS should not be involved in an activity that could be perceived as purely political. As this chapter has attempted to make clear, the purpose of the GTS and the procedures involved in its construction, such as the selection of GSSPs for selected boundaries, is to facilitate the accurate and precise correlation of the sedimentary record of geological processes and events worldwide. There is no need for a formally defined Anthropocene to facilitate the procedure of using geological methods to correlate processes that are largely within historical memory.

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