Chapter 7

Stratigraphy

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7.1 Introduction and Concepts

Stratigraphy is the study of the succession and time-related architecture of rock strata. Stratification is not limited to sedimentary deposit, but is also found in igneous rocks, particularly volcanic developments, and in certain plutonic rocks. All bedded rocks can be treated stratigraphically, i.e. to establish age relations between beds. However, the term "stratigraphy" is used primarily in the context of sedimentary research.

Stratigraphy involves the study, subdivision and documentation of sedimentary successions, and on this basis to interpret the geological history they represent. In order to reconstruct an environment or special events in the geological past, regionally or even globally, it is necessary to characterise and correlate sedimentary horizons from different areas. It is important to establish which sedimentary units were deposited at the same time or by the same or similar sedimentological or biological processes, even if they are not quite contemporary. Correlation is generally a question of what is possible to achieve with the available amount and quality of data.

We usually have no possibility of determining definitely which beds were deposited at the same time, but attempt to use all the information available in the rocks. This information falls into five main categories: (1) Rock composition and structures resulting from sedimentological processes. (2) Fossil content, which is a result of biological, environmental and ecological evolution throughout geological history. (3) Content of radioactive fission products in minerals or rocks which may be used for age dating. (4) Magnetic properties of rock strata. (5) Geochemical features of sediments. (6) Geophysical expression. These correlation methods are so different that it has been found useful to work with three forms of stratigraphy which can be used in parallel:

1. *Lithostratigraphy:* Classification of sedimentary rock types on the basis of their composition, appearance and sedimentary structures.
2. *Biostratigraphy:* Classification of sedimentary rocks according to their fossil content.
3. *Chronostratigraphy:* Classification of rocks on the basis of geological time. *Geochronology* is the actual subdivision of geological time.

The first two are thus based on rock relationships which can be described, and are often collectively referred to as rock stratigraphy.

Despite radiometric dating, geological time is not absolute. Even this dating method gives different ages, depending on which half-life is used for calculating radioactive decay, and is encumbered with several other uncertainty factors. Chronostratigraphy is therefore a theoretical and abstract concept which describes a time scale we cannot measure exactly. Rock stratigraphy, on the other hand, takes the rocks themselves as its starting point and is based on boundaries which are identified as potentially suitable for correlation across greater or lesser areas. For rules for stratigraphic nomenclature, see stratigraphy.org. The different types of stratigraphic units are listed below in hierarchic arrangement. See also Stratigraphy.com.

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Lithostratigraphic	Biostratigraphic	Chronostratigraphic	Geochronological
Supergroup	Assemblage zone	Eonothem	Eon
Group	Range zone	Erathem	Era
Subgroup	Acme zone	System	Period
Formation	Interval zone	Series	Epoch
Member	Biozone	Stage	Age
Bed		Chronozone	Chron

7.2 Lithostratigraphy

The *formation* is the fundamental lithostratigraphic unit defined in a bedded succession. An important property of a formation is that it should be easily recognised in the field or a borehole due to its composition (lithology). The formation is thus a mappable unit, which can be visualised on an ordinary geological map (scale e.g. 1:50,000) or recognised in the description of a bedded succession. There is in principle no limit to the thickness of a formation, but it usually varies from a few tens to several hundred metres. A 50–300 m thick sandstone, over- and underlain by quite different rocks such as shale or limestone, would constitute a natural formation. However, a formation will seldom be homogeneous, and for more detailed mapping it will be useful to divide the formation into smaller units. Parts of a sandstone formation containing shale or conglomerate beds, for example, can be defined as *members* of the formation.

The smallest unit in the lithostratigraphic classification is a *bed* which is assumed to have been deposited by a single depositional process without a break in sedimentation, and is distinguishable from the beds above and below. When sections are logged, beds and other lithological entities are often recorded as "units". Units which are <1 cm thick are called *laminae*, but this is a purely size-descriptive term. Most beds/units contain, or may even be entirely comprised of, laminae. Only when a lamina lies isolated between significantly different sediments will it be named, e.g. if it is a thin tephra (ash bed) or bentonite in a shale succession.

For some purposes it is expedient to group several formations together into a larger unit called a *group*. A group normally consists of three to six formations and can be divided into two or more *subgroups*. The largest lithostratigraphic unit is a *supergroup*, which consists of two or more groups. This unit is used when a common name is needed to encompass a thick sedimentary package.

In sedimentary basins which are partly or entirely buried in the subsurface, for example the North Sea basin, lithostratigraphic units like formations and groups are defined on the basis of records from wells, such as well logs. In offshore areas where it is difficult to find enough geographical names to give the stratigraphic units, other names are then also used, including historical names, names of animals etc.

7.2.1 Lithostratigraphic Terminology

As research into sedimentary successions expanded, it required a standardised lithostratigraphic nomenclature that could be applied anywhere. Therefore, a set of international rules was established for naming lithostratigraphic units and are summarised below:

1. A stratigraphic unit should preferably be defined with reference to a type section (stratotype) present in a good exposure, or in a well where the unit is adequately represented.
2. Each stratigraphic unit ought to be named after a geographical site, located preferably near the type section if exposed on land. Offshore stratigraphic units may be named after marine features (e.g. fishing banks), although in the North Sea basin other names have also been used.
3. The same name ought not to be used for more than one stratigraphic unit. The unit which is first defined has priority.
4. Stratigraphic names may consist of a unique name, i.e. from a locality, and a stratigraphic unit hierarchy designation, e.g. Kimmeridge Formation or Kimmeridge Clay. In the US many stratigraphic names are well established even if they do not strictly follow the international rules for stratigraphic nomenclature.

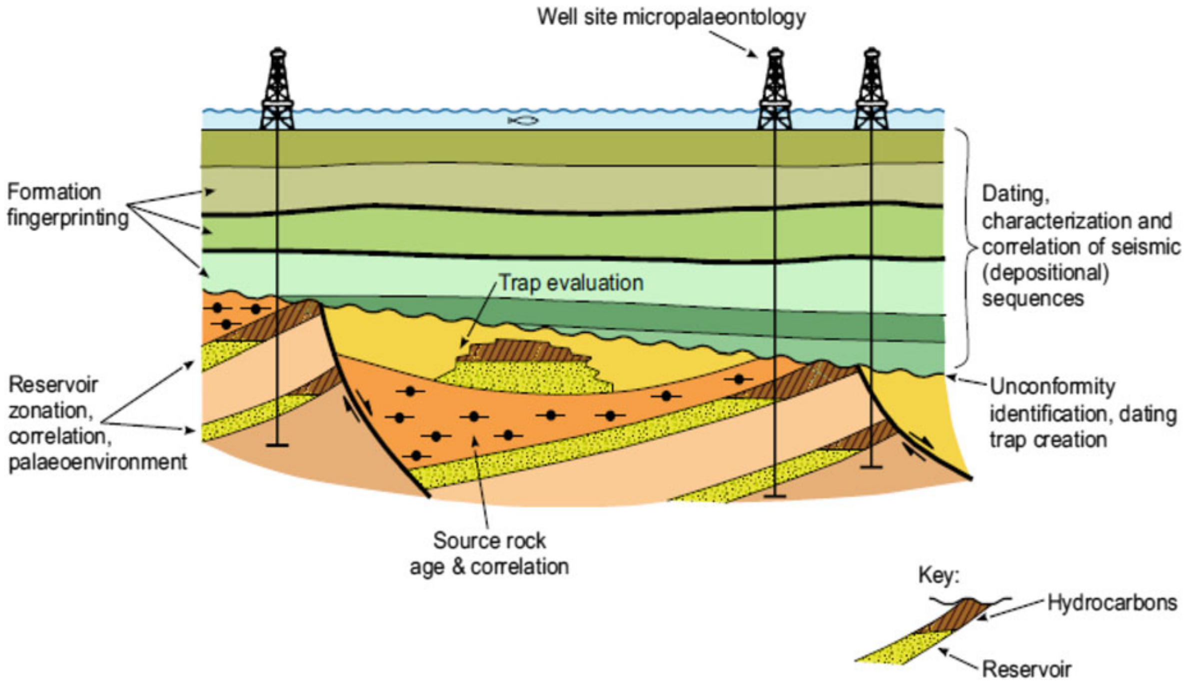

Fig. 7.1 Application fields of micropalaeontological methods in petroleum exploration and appraisal activities (based on Copestake 1993)

7.3 Biostratigraphy

Biostratigraphy is based on the fossils in sedimentary rocks. Its tasks are to group strata into units characterized by fossil content, and apply these to date and correlate sedimentary successions. Biostratigraphy has made it possible to correlate sedimentary rocks on a global scale, and forms the foundation for global stratigraphic classification of sedimentary successions in combination with radiometric and other age dating methods. The biostratigraphic application of fossils is based on the fact that during the course of geological time, biological evolution took place whereby some species died out and new ones appeared. Any particular species will therefore be represented only in sediments deposited within a limited time span, and can be used to recognise this time span or part of it.

7.3.1 Nature of the Fossil Record

7.3.1.1 Fields of Application

Industrial application of biostratigraphy is mainly based on microfossils, owing to the fact that these microscopic remains commonly occur in large amounts, and can be extracted easily both from drill cuttings, sidewall cores as well as conventional cores. The rapid evolution of many microfossil groups makes them valuable tools in subsurface stratigraphic work (Fig. 7.1). In addition to the general tasks of age determination and correlation, their applications include: unconformity identification, characterisation and correlation of seismic (depositional) sequences, fingerprinting of formations, reservoir zonation and palaeoenvironmental modelling. These applications are crucial in petroleum exploration, contributing significantly to assessments of reservoir distribution, source rock evaluation, trap evaluation, estimation of reserves, field development studies and impacts on drilling problems. The fast evaluation potential of microfossil samples is also an important factor in any subsurface work.

7.3.1.2 Distribution of Microfossils

Practically all rocks of sedimentary origin contain microfossils, although their abundance, diversity and state of preservation are highly variable. These features are strongly influenced by the age, depositional environment, composition and diagenetic

Rocks \ Microfossils	Spores, pollen	Dinoflagellates	Acritarchs	Foraminifera	Conodonts	Ostracods	Prasynophyceans	Calpionellids	Chitinozoans	Botryococcaceans	Coccolithophores	Radiolarians	Silicoflagellates	Diatoms
Clays and shales	●	●	●	●	⧗	●	⧗	⧗	●	⊗	●	●	⊗	⊗
Limestones and marls	⧗	⧗	⧗	●	●	●	⊗	●	⧗		●	⊗	⊗	
Flints and cherts	⊗	⧗	⊗	⧗	⊗	⊗					⊗	●	⧗	⧗
Coal, lignite, peat	●	⧗	⧗				○			⧗				
Sands and sandstones	⧗	⊗	⊗	⧗	⊗	○	○		⊗					
Dolomites, ankerites	⊗	⊗	⊗	⧗	⧗	○	○	○						
Evaporites: gypsum, halite	⊗	○	○											
Metamorphic rocks: slates, phyllites, marbles	○	○	○	○	○									

● Abundant
⧗ Common
⊗ Rare
○ Sporadic

Fig. 7.2 Distribution trends and importance of major microfossil groups in different types of rocks

history of the sediments. Marine mudstones, marls and limestones usually have a particularly rich microfossil content of high diversity (Fig. 7.2). Well sorted sandstones have a generally low microfossil content, as most microfossils are both lighter and smaller than the average sand grains. Dolomites are also poor in microfossils because most of the fossil organisms with primary calcite have been dissolved during dolomitisation, or because the beds represent an evaporite environment. Coals and related organic sediments are rich in terrestrial microfossils (spores and pollen). Evaporites are characterised by small quantities of microfossils, mainly of terrestrial origin.

Microfossils occur in Precambrian rocks but rapid development of many central groups started in the Cambrian (Fig. 7.3). Every part of the stratigraphic column from Cambrian to recent time contains one or several microfossil groups which are potentially useful for biostratigraphical or palaeoecological purposes. According to skeletal composition, microfossils can be arranged into five groups, which comprise several taxonomic divisions:

(1) Calcareous microfossils: calcitic and aragonitic foraminifera, single-celled heterotrophs; ostracods, microscopic crustaceans; calpionellids, uncertain origin; carophytes, algal reproductive organs; calcareous nannoplankton, single-celled algae.
(2) Arenaceous microfossils: agglutinated foraminifera, single-celled heterotrophs.
(3) Siliceous microfossils: radiolarians, single-celled heterotrophs; diatoms, single-celled algae; silicoflagellates, single-celled algae; ebridians, single-celled heterotrophs.
(4) Phosphatic microfossils: conodonts, primitive vertebrates.
(5) Organic-walled microfossils: acritarchs, uncertain origin; prasinophytes, cysts of green algae; chitinozoa, uncertain origin; chlorophytes, freshwater algae, algal colonies; dinoflagellates, single-celled algae.

7.3.2 Factors Controlling Stratigraphic Application

There are several factors influencing the applicability of microfossils to stratigraphic purposes. The most important ones having a positive effect on the application potential are: (1) Presence of hard parts with high preservation potential. (2) High evolution rate of taxa. (3) Extensive regional distribution, or high potential for dispersal. (4) High degree of facies independence. (5) Small redepositional potential.

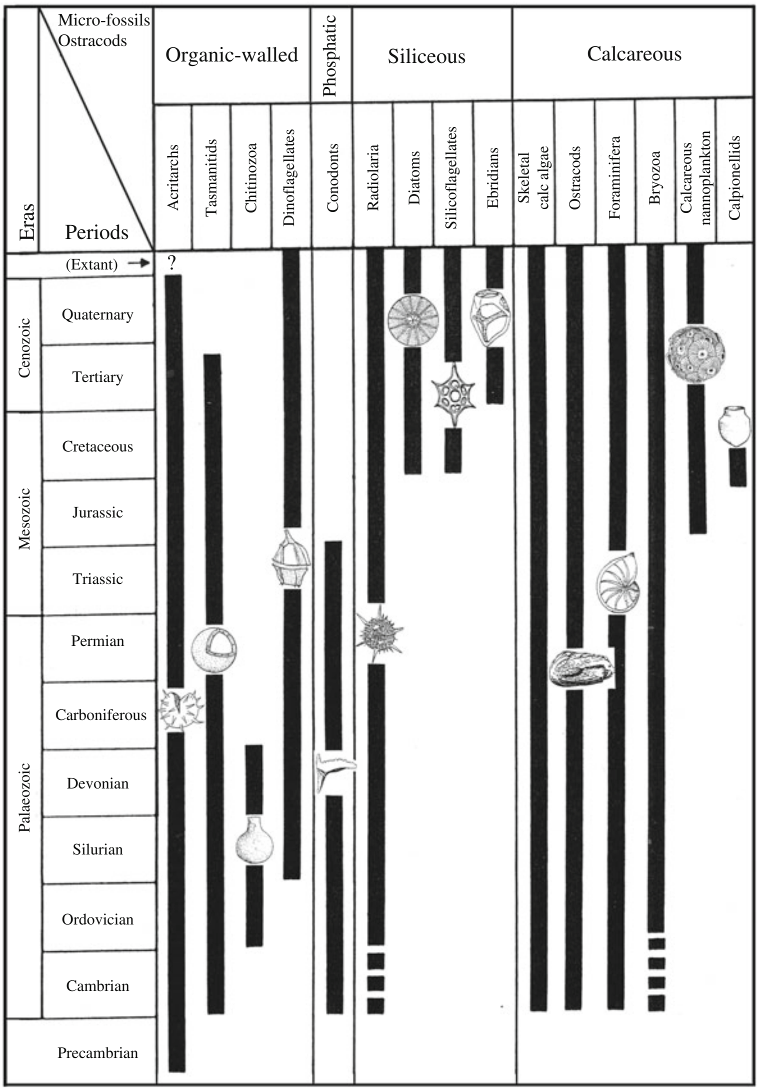

Fig. 7.3 Outline of the stratigraphic distribution of microfossil groups from Precambrian to present time, arranged according to skeletal composition (modified after Haq and Boersma 1978)

(6) Presence of many distinct morphological features, increasing the precision level of taxon identification.

7.3.2.1 Mode of Life

The mode of life (*ethology*) of organisms is crucial for the regional distribution of their fossil remains. Living and fossil biota can be arranged into two ethological groups (Fig. 7.4). *Pelagic* organisms live freely in the water column and comprise the *planktonic* forms (passively floating) and *nektonic* forms (actively swimming). The *benthos* comprises bottom dwelling organisms belonging to *epibenthos* (living at the sediment/water interface), and *endobenthos* (living buried in the sediment). Epibenthic forms are freely moving or are attached to firm surfaces. Foraminifera as a group reveal the widest environmental range by being represented in all the ethological categories except nektonic. Ostracods and diatoms also have a relatively wide range as they occur both in the benthic and pelagic habitats. Conodonts are regarded to have a nektonic origin. Acritarchs, dinoflagellates, calcareous nannoplankton, radiolarians and several minor groups are exclusively planktonic. Spores and pollen occupy a special position because they are the reproductive

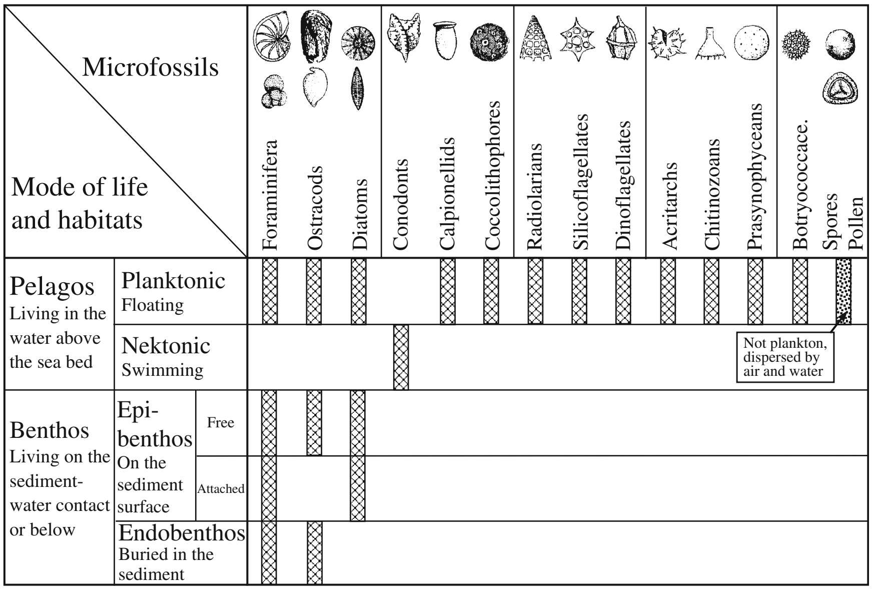

Fig. 7.4 Mode of life and habitats of major microfossil groups in aquatic environments

organs of terrestrial plants, but are dispersed principally as plankton (by air or water).

It is obvious that the pelagic group contains microfossils with particularly high stratigraphic applicability, owing to their relative independence of facies, e.g. planktonic foraminifera, calcareous nannoplankton, dinoflagellates and the nektonic conodonts. Benthic biota usually have less stratigraphic potential, although they are also commonly used for this purpose, particularly in marginal marine deposits where planktonic biota are rare. On the other hand, benthic forms are commonly excellent facies indicators, e.g. benthic foraminifera and ostracods.

7.3.2.2 Environmental Distribution

The life habitat of biota is of prime importance both for stratigraphical purposes and biofacies analysis. In non-marine subaquatic environments important fossil groups include diatoms and ostracods (Fig. 7.4). In marginal marine brackish environments low diversity assemblages of foraminifera, ostracods, and diatoms are dominant. Shallow shelf to bathyal environments are typified by their high diversity assemblages of benthic and pelagic biota. In abyssal environments, below the compensation depth of calcium carbonate, microfossils are exclusively siliceous, arenaceous or organic-walled.

The members of the ubiquitous group of organic-walled microfossils are produced in a wide range of contrasting habitats (Fig. 7.4). Pollen and spores are produced by land and freshwater flora. They are transported far out into the marine realm and show decreasing frequency with increasing distance from source. They are commonly used in the stratigraphy of terrestrial and marginal marine deposits in the absence of other fossils, although they provide a low age resolution. Chlorophyte freshwater algae (as Botryococcus) are also prone to be transported to marine areas. Prasionophyte marine algae and acritarchs are common in brackish marginal marine and shallow shelf settings. Dinoflagellates are most common in marine shelf and oceanic waters, and are widely used for stratigraphical purposes.

Foraminifera occur in a wide variety of environments ranging from open ocean to estuarine waters, and are extensively used in stratigraphic and facies analyses. Marginal marine environments (such as tidal marshes, estuaries and lagoons) are typified by low diversity benthic assemblages composed of agglutinated and mixed calcareous-agglutinated components. The low species diversity restricted nature of these assemblages reduces their stratigraphical applicability. Normal marine shelf areas are typified by high diversity calcareous faunas and

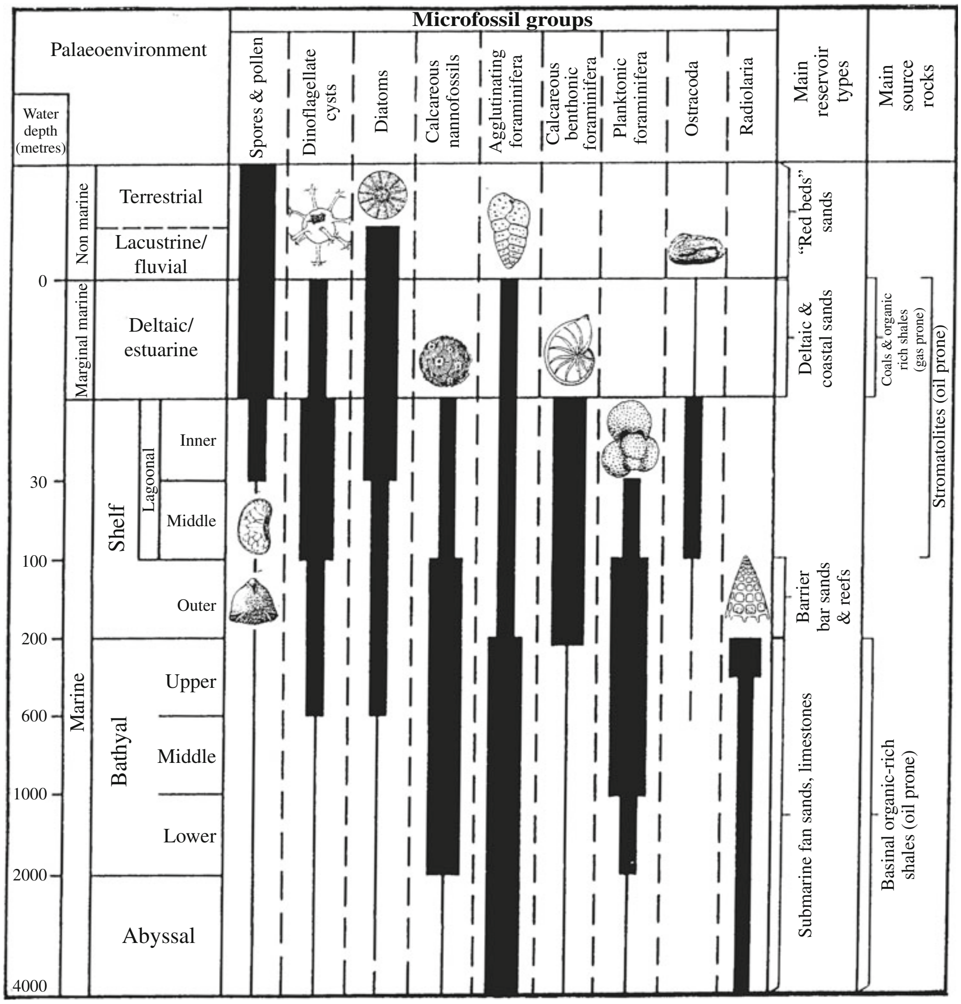

Fig. 7.5 Depth-related palaeoenvironmental distribution of major microfossil groups in the North Sea basin (Triassic to Tertiary), and position of the main petroleum reservoir and source rock facies (from Copestake 1993)

increasing frequency of planktonic species toward the shelf break.

The foraminiferal planktonic component reaches its maximum in bathyal areas. These shelf and bathyal assemblages form an excellent basis for age determination and stratigraphic correlation. The amount of calcareous benthic and planktonic foraminifera decreases downslope through the lysocline and they disappear below the compensation depth of calcium carbonate. Below this level the sea bed is populated by agglutinated species, usually of a cosmopolitan nature, which provide an adequate basis for stratigraphic and environmental application.

The environmental distribution of the various microfossil groups in the North Sea basin Mesozoic and Cenozoic succession is well known owing to the prevailing oil and gas exploration (Fig. 7.5). The microfossil distribution in this area reflects the general bathymetric distribution pattern of the various groups. Abyssal deposits below the carbonate compensation depth are dominated by radiolarians and cosmopolitan agglutinated foraminifera. Bathyal and outer shelf sediments are characterised by high dominance of planktonic foraminifera and calcareous nannoplankton. Mid-shelf sediments contain abundant calcareous benthic foraminifera and dinoflagellates. Deltaic deposits are typified by spores, polen and agglutinated foraminifera. In non-marine deposits the most common fossils are spores and pollen.

7.3.2.3 Redeposition

Erosion of fossil-bearing sediments with subsequent transport and redeposition of microfossils can lead to serious problems for dating and correlation. Palynomorphs are particularly liable to redeposition owing to their small size and weight. Conodonts also have a high redeposition potential because their calcium phosphate skeletons are strongly resistant to corrosion by weathering and transport. Redeposition is suspected if microfossils appear at higher stratigraphic levels than expected. Reworked microfossils can be distinguished by changes in colour and traces of wear.

7.3.2.4 Facies Barriers and Biostratigraphy

Sediment cores from deep oceanic settings usually provide undisturbed sections containing a continuous fossil record, commonly of high stratigraphic resolution. Based on this type of material, biostratigraphy combined with magnetostratigraphy and chemostratigraphy contributed heavily to development of a global biochronozonal framework. Although with marked reductions, this framework has been extended to formations deposited in marine shelf, deltaic and coastal areas, where it forms an important basis for dating and correlation.

The basic unit of biostratigraphic correlation is the *taxon chronozone*, also called *biochronozone*. As defined below, a taxon chronozone represents all sediments deposited between the evolutionary inception and extinction (total range) of its marker species (Fig. 7.8). In contrast, the *biozone* represents only those sediments at a given place that contain one or several marker species. In the case of a single marker, the biozone can be defined by the local appearance and disappearance (local range) of the species. The time interval represented by the biozone may differ from place to place.

In deep oceanic regimes, the latitudinal distribution of each species is generally defined by the water mass temperature and chemistry. Therefore, the regional (lateral) extent of biozones and taxon chronozones is of restricted nature. Outside its optimal stratigraphic distribution area, the range of the marker species can vary considerably owing to environmental factors. At such sites the chronozone cannot be defined and the term biozone is used. Correlation between tropical and temperate oceanic regions is difficult, because of the low number of biota that include reliable zonal indicator species common to both areas. Interfingering of biozones (local ranges) and integration with magnetostratigraphy and chemostratigraphy is a useful approach to this type of correlation.

7.3.3 *Biozones*

The basic unit of biostratigraphy is the biozone which is a body of strata defined by its fossil content. During the history of biostratigraphy, a variety of names and definitions of units have been proposed, but there are only three basic types of zones: interval zones, assemblage zones and abundance zones. Zones defined in sedimentary successions are usually arranged into a zonal scheme, which can be developed for an extensive region and even for global application.

7.3.3.1 Interval Zones

These are the most commonly used type of zones, with boundaries defined by particular fossil events, specifically the first and last occurrence of taxa (usually species). Thus, the interval zone comprises the strata deposited between these two events. There are five types of interval zones (Fig. 7.6).

1. *Taxon range zone* is defined by the first and last occurrence of a particular taxon.
2. *Concurrent range zone* is the interval between the first occurrence of a taxon and the last occurrence of another taxon.
3. *Partial range zone* is the interval between the last occurrence of a taxon and the first occurrence of another taxon partitioning the range of a third taxon.
4. *Successive first appearance zone* represents the interval between the successive first appearance of two taxa.
5. *Successive last appearance zone* is the interval between the successive last appearance of two taxa.

The successive last appearance zone is the type most commonly used in industrial biostratigraphy, because the samples are mainly derived from well drilling cuttings. In such sample sets, the actual first appearances are unreliable owing to downhole (caving) contamination. Precise positions of first appearances are required in order to recognise the first four interval zone types. Taxon range zone and successive first appearance zone are the most applied in zonation of onshore sections.

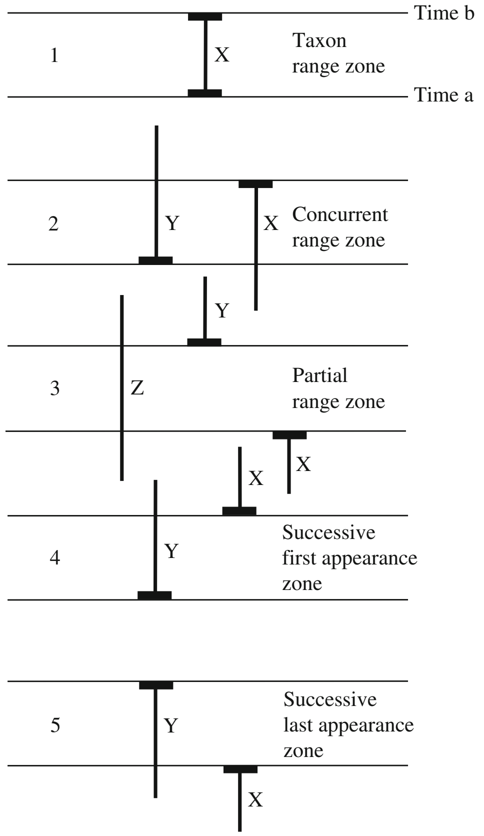

Fig. 7.6 Outline of the five types of biostratigraphic interval zones shown by occurrences of microfossils X, Y, and Z. *Vertical lines* are taxon ranges, *horizontal bars* are first and last occurrences

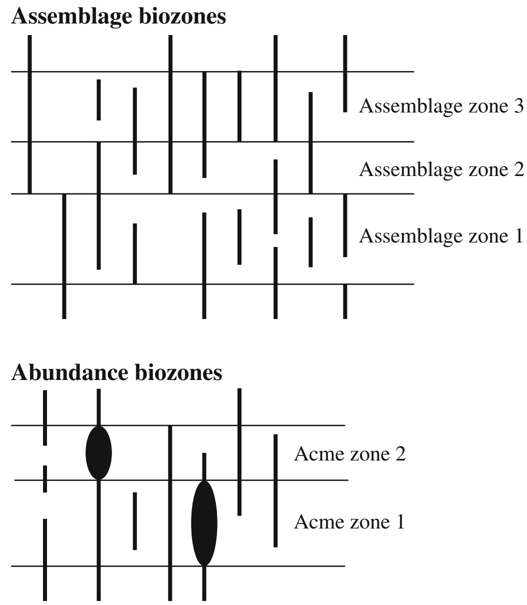

Fig. 7.7 Outline of the biostratigraphic assemblage and abundance zones

7.3.3.2 Other Biozones

1. *Assemblage zone* is a biozone characterised by the association of three or more taxa (Fig. 7.7). This zone is recognised without regard to range limits of taxa, and is usually of regional importance.
2. *Acme zone* is based on the numerical maximum of one or several taxa, and is also termed *abundance zone*. Taxon maxima are strongly influenced by local environmental conditions, and therefore this zone is usually of local stratigraphic importance.
3. *Taxon chronozone* represents the interval between two unique bioevents: the first evolutionary inception of a taxon, and its extinction. Thus, this zone comprises the total range of the taxon, and is of global nature (Fig. 7.8). The distribution of taxa, however, is to a varying degree facies-controlled, which leads to the development of geographically restricted taxon range zones. One or several of such local taxon range zones can be included in the total range of the taxon.

7.3.4 Biostratigraphic Correlation

Biostratigraphic correlation is based on the fact that during the course of geological time, biological evolution has taken place, whereby some species and higher taxa have died out and new ones have appeared (Fig. 7.3). Any particular species will therefore be represented only in sediments deposited within a limited time span. However, this method also has many limitations. Some fossil groups developed rapidly, with individual species only existing for a short period. Other groups evolved slowly and persisted throughout long periods of geological time. Some fossil organisms are found only in rocks deposited in particular environments or facies. Many were dependant on a certain salinity, oxygen content and water depth. Bottom conditions played a major role for epifaunal (attached) and infaunal (burrowing) animals.

Biozones of sedimentary successions are arranged into zonal schemes, which are the main tools of fossil-based correlation. The ideal index fossil for defining a

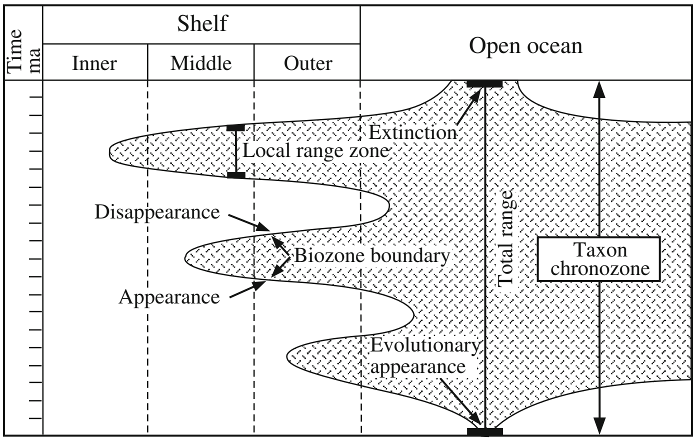

Fig. 7.8 Facies-related distribution of an imaginary fossil taxon exemplifying the difference between a taxon chronozone (global) and a biozone (local) taxon range zone

biozone ought to belong to a fossil group which underwent rapid biological evolution and had a global distribution. It should not have had special environmental requirements, and should have been cosmopolitan with respect to habitat. One such fossil group is the graptolites, which are common in Cambro-Silurian shales. They were planktonic and drifted in the surface seawater. However, even this type of plankton may show a distribution pattern which is influenced by ocean currents and temperatures etc. At the present, microfossils are being used more and more for biostratigraphic correlation particularly by the petroleum industry. Several animal and plant groups (e.g. conodonts, planktonic foraminifera, radiolarians and spores and pollen) have wide distributions and have the advantage that even small samples, e.g. from boreholes, provide adequate material for statistical treatment.

Although fossils are always helpful in stratigraphic correlations, we now avoid defining important geological boundaries, for example that between the Ordovician and the Silurian, by means of fossil occurrences. Such a boundary would have to be moved whenever one found a new occurrence of a certain fossil. The geological periods are at the present defined by international committees which select a type section with continuous sedimentation and preferably fossiliferous record, and the boundary is physically marked as a fixed point in the section. The boundary is then unambiguously defined. All available means can then be used, including fossils, to correlate the boundary with other areas. This is the principle of the arbitrary boundary. In reality it is not entirely arbitrary. We try to put it on a section with optimal potential for correlation with other areas and the global time scale (Fig. 7.9).

Geologists used to define a stratigraphic boundary at a break (hiatus) in the sedimentary record. However, sediments later found elsewhere that had been deposited in the period represented by the hiatus, could not be assigned to either unit. For example, the boundary between the Tertiary and Cretaceous was defined in England where a hiatus is developed between these systems. As a result it was difficult to reach agreement on whether sediments which were deposited for example in Denmark during this period (Danian), should belong to the Cretaceous or the Tertiary.

7.4 Time Stratigraphy

Chronostratigraphy is an attempt to correlate rocks deposited at the same time, across larger areas. The accuracy achievable with chronostratigraphic correlation depends on whether the sediments contain evidence of well-defined geological events which were simultaneous across the region. These events may be biological (e.g. appearance of particular species), sedimentological (e.g. deposition of ash layers) or geophysical (e.g. reversals of the Earth's magnetic field). Stratigraphical research has a long tradition of correlating synchronous events in geological history. Such correlations are independent of an absolute time scale. Only after the development of radiometric dating methods did it become possible to set up a series of

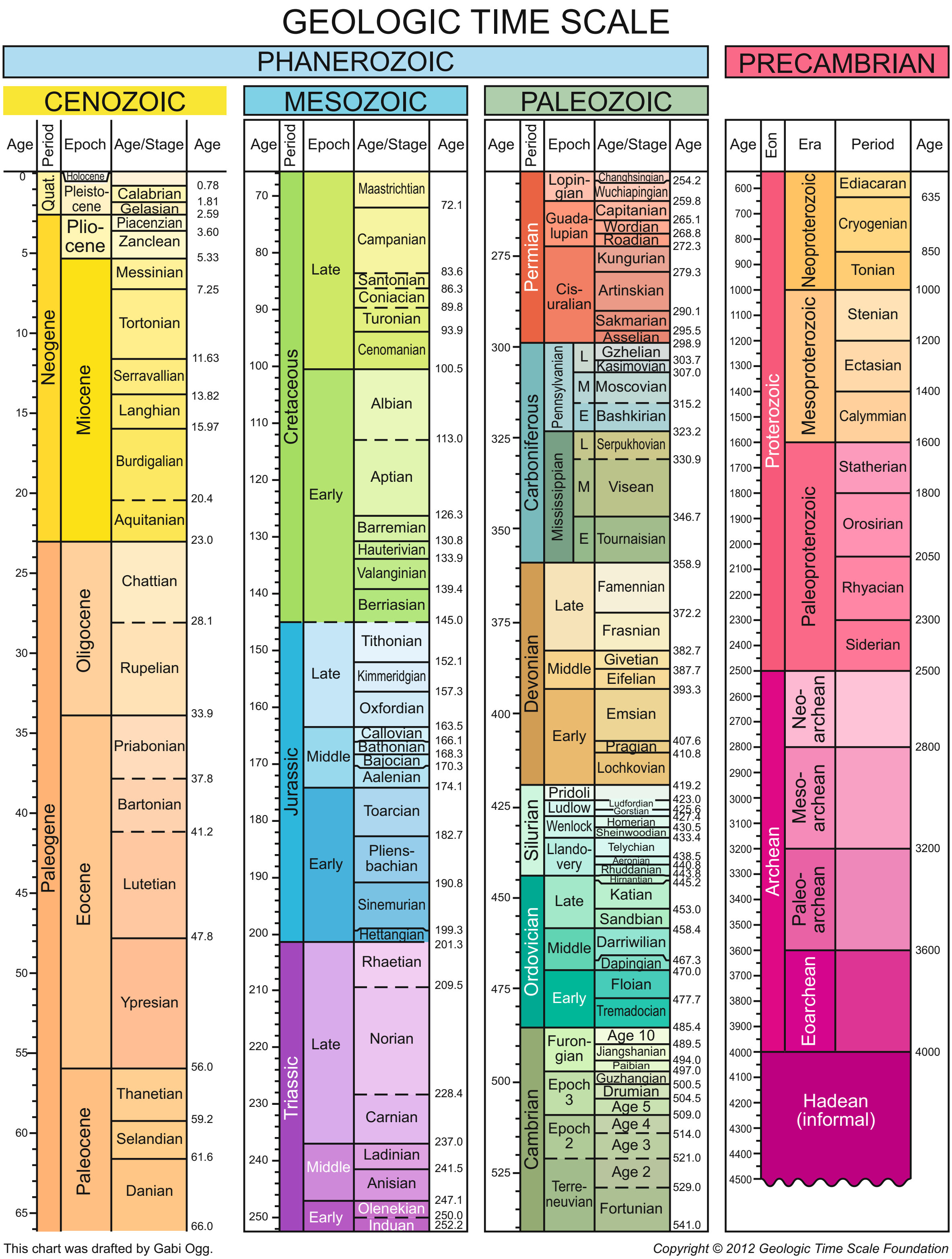

Fig. 7.9 Geological time scale for the Phanerozioc Aeon (Geological Time Scale Foundation 2012)

datings approaching an absolute time scale, but radiometric datings are not absolute with respect to time.

We thus distinguish between two types of time stratigraphy: (1) Geochronology, which is the subdivision of the Earth's history into finite time units. A geochronological unit is a specific interval of geological time. (2) Chronostratigraphy or time-rock stratigraphy is the subdivision of sedimentary successions, and their correlation, on the basis of time. Chronostratigraphic units are by definition synchronous.

A geochronological unit represents a specific interval in the geological time scale. For example, the Jurassic *period* is a geochronological unit. A geochronological unit may define the time span between two specific geological events. We can say, for example, that some of the rocks in the North Sea were deposited during the Jurassic period. The corresponding chronostratigraphic unit (system) signifies the rocks which were formed during the same period. We therefore say that some of the rocks in the North Sea belong to the Jurassic *system*.

The basic chronostratigraphic unit is a *chronozone*. A chronozone includes all the deposits formed during a particular, relatively short, geologic time interval and which are defined by a geological phenomenon or by a particular interval of a rock succession. In most cases a chronozone is a *taxon chronozone*, which is defined as the period between the first appearance and last occurrence of a particular fossil taxon (Fig. 7.8). Whereas a biozone can only be defined where the fossil is present, a chronozone represents all the rocks that were formed during this period, regardless of whether they contain fossils. A *chron* is the interval of time during which the rock in a chronozone was formed. It is thus the geochronological equivalent of a chronozone.

A chronozone may be named after a biostratigraphic unit, e.g. a *Didymograptus extensus* chronozone, or after a lithostratigraphic unit, which can be recognised over large areas. Such characteristic strata are often called *marker beds, key beds*, or *datum beds*. Chronostratigraphic horizons are important for correlation within sedimentary basins, and form the framework for all facies reconstructions. If we have two *chronohorizons* or datum beds, we can measure the variation in thickness and composition of the sediments which were deposited during a particular time period. This may make it possible to map variations in sedimentation rate and the ratio of sandstone to shale. Maps showing the sediment thickness between two marker beds are called isopach maps. This is very useful for reconstructing sedimentary facies on the basis of borehole data, a standard method in connection with oil prospecting.

The best chronostratigraphic horizons (marker beds) are bentonite (ash) layers; on a smaller regional scale also coal beds, phosphate beds, thin limestone or sandstone beds, or particular fossil horizons. When analysing stratigraphic records from wells (logs) one tends to use beds which produce a distinctive log pattern. Seismic reflectors can also in certain instances be used as datum beds (horizons).

A *stage* is a chronostratigraphic unit which includes one or more chronozones but which nevertheless covers a limited period of time, usually 3–10 million years (Fig. 7.9). This is the smallest unit in the chronostratigraphic hierarchy which is used for correlation all over the world. A stage is defined in a type section and usually designated by a geographical name near the type profile. For example, the Kimmeridgian stage is well exposed on the Dorset coast at Kimmeridge. The correlation of a stage is usually based on biostratigraphy. An *age* is the period of time (geochronological unit) which corresponds to a stage.

A *series* is a chronostratigraphic unit larger than a stage. For example, the Late Jurassic is a series constituting part of the Jurassic system. The geochronological unit which corresponds to a series is an *epoch*. We can say that a certain limestone was deposited during the Late Jurassic epoch.

A geochronological *period* varies in duration from about 20–30 million years (Silurian) to about 60–70 million years (Cretaceous). The Quaternary period, however, is much shorter, only about 2.5 million years. The rocks formed during a period constitute a *system*. An *era* is comprised of two or more periods. The Palaeozoic era had a duration of about 300 million years, but the Cenozoic era did not last longer than the longest Palaeozoic periods (65 million years).

The largest units in the chronostratigraphic scale, *erathem* and *enothem*, are not used much, since it is seldom relevant to group rocks which were deposited over such long periods of time. However, when we discuss geological history in relation to the biological development on the Earth, for example, it can be useful to speak about the Phanerozoic *aeon* which covers the Palaeozoic, Mesozoic and Cenozoic eras. The Precambrian is divided into two aeons: the Proterozoic (542–2,500 million years ago) and the Archean (2,500–4,000 million years ago).

7.5 The Relation Between Lithostratigraphy, Biostratigraphy and Chronostratigraphy

These three types of stratigraphy are based on different criteria, and geological experience has shown that it is useful to maintain this tripartite division. Stratigraphical boundaries defined in these three ways may sometimes nearly coincide, but usually show considerable divergences particularly over larger distances. This can be demonstrated schematically by the occurrence of a sandstone formation between two shale units (Fig. 7.10). Lithostratigraphically, this sandstone formation is unambiguously defined by the boundaries between shale and sandstone. However, if we find good index fossils, X and Y, in both the shale and the sandstone, it may turn out that the top of the sandstone in area B corresponds chronostratigraphically to the bottom of the sandstone in area A. Sedimentation of sand has thus moved from B to A during the course of a measurable period of geological time. Sandstone sedimentation along coasts and on deltas will tend to shift over long periods of time, and the sandstones deposited will therefore have an upper and a lower boundary which are not parallel to a theoretical time plane, i.e. they are time-transgressive.

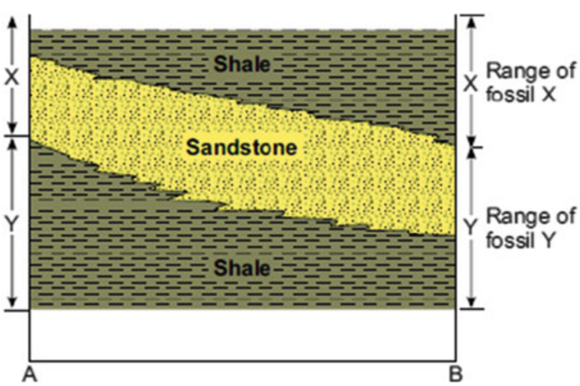

Fig. 7.10 Occurrence of a time-trasgressive sandstone formation between shale units. X and Y are good index fossils which are little affected by facies. Consequently, they indicate that the sandstone is younger in area A than in B, and has been deposited by progradation from B to A. This could be a shallow marine sandstone or a sandstone deposited by a prograding delta.

In practical geological work (mapping, well and borehole studies, etc.) one is primarily interested in correlating rock types, in other words lithostratigraphy. To construct a facies map that gives an overview of the distribution of sediments at a particular time, it is necessary to establish a time correlation. Good index fossils or marker beds will provide an important framework for portraying depositional conditions, for example during regressive or transgressive phases. If we study a modern coastal area, we find different fauna in different environments. We can easily see that the distribution of biotas within a limited geological time period is not controlled by stratigraphic time, but by facies. Many animal groups provide good indices for sedimentary facies (Figs. 7.4 and 7.5).

7.5.1 Correlation

Rocks from the Phanerozoic Aeon were originally classified on the basis of fossils. Later, committees for each period have been set up, responsible for selecting type sections where the lower and upper boundaries are physically defined by a bolt ("golden spike"). Correlation with other areas can then be carried out with the aid of fossils or other age indicators, anywhere in the world.

Although biostratigraphic correlation is still the most important method, magnetostratigraphy and radiometric dating methods are gaining importance. Within the confines of a sedimentary basin, however, certain types of lithostratigraphic correlation can turn out to be the most accurate. The various types of well logs provide good opportunities for lithostratigraphic correlation. Seismic profiles have perhaps to an even greater degree made it possible to correlate lithostratigraphy over great distances as a basis for sequence stratigraphy (see next chapter).

It was realised relatively early that the lateral changes in facies corresponded to the vertical changes in sedimentary basins if there are no breaks in sedimentation (= unconformity). Walthers law states that, "In a conformable succession the only facies that can occur together in vertical successions are those that can occur side by side in nature".

7.6 Radiometric Dating Methods

7.6.1 Concept of Dating

The geological time scale is mainly founded on age determinations based on radioactive processes. There are a number of other indirect methods of measuring geological time, but only radiometric datings give satisfactory quantitative results. The measurements are based on the fact that radioactive nuclides undergo decay at a certain rate. The rate of fission can be determined with some accuracy, and there is no reason to believe that it has varied through geological time, even though this is difficult to prove. The rate at which radioactive decay proceeds is proportional to the assumed number of fissionable atoms present. The number of atoms remaining which can undergo fission decreases according to the following formula, however:

$$dN/dt = -\lambda N - \text{which is the rate of decay at any time } t.$$

$$N = N_o e^{\lambda t}$$

where
N_o = number of atoms at time T_o
N = number of atoms at time T
λ = decay (disintegration) constant, which is the fraction of the total number of atoms which will decay in a given time.

A convenient measure of the rate of fission is the half-life ($T_{1/2}$) i.e. the time it takes for half of the original atoms to decay. That is to say

$$N/N_o = 1/2 = e^{-\lambda T}_{1/2}$$

The half-life is then:

$$T_{1/2} = \ln 2/\lambda = 0.693/\lambda$$

Radioactive decay processes result in new nuclides called daughter nuclides. In order to date geological material we measure the ratio between the fissionable nuclide (parent nuclide) and the daughter nuclide.

7.6.2 Potassium-Argon Method

One of the most commonly used dating methods is the potassium-argon method. ^{40}K is an unstable nuclide which decays mainly to ^{40}Ca with the emission of beta particles, but about 12% is transformed into argon-40 through capture of electrons and emission of X-rays (gamma-rays). Any mineral which contains potassium, for example biotite, will also contain some ^{40}K which decays to ^{40}Ar. The amount of argon in the mineral is thus an expression of the mineral's age, which can be calculated if we know the half-life ($T^{1/2}$) or the decay rate (λ) (half-life is 1.31×10^9 years).

The amount of argon gas in the minerals is analysed with a sensitive mass spectrometer. Argon is a very volatile gas, however, and when minerals containing argon are heated up or deformed, for example during metamorphism or folding, the argon will escape and the radioactive "clock" will indicate an age which corresponds more to the time of metamorphism than to the formation of the mineral. Therefore, a radioactive age determination does not necessarily give a figure for the age of the rock, but will tell us something about the geological processes which have affected the rock and thus help us to reconstruct its geological history.

7.6.3 Rubidium-Strontium Method

Another commonly used dating method is the rubidium-strontium rato, ^{87}Rb to ^{87}Sr. Rubidium has a radioactive isotope, ^{87}Rb, which decays to ^{87}Sr with a half life of 4.89×10^{10} years. The amount of ^{87}Sr is then a function of the time since the formation of the mineral or rock. Instead of measuring absolute amounts of ^{87}Sr, it is simpler to measure the $^{87}Sr/^{86}Sr$ ratio on the mass spectrometer. Then the $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ ratios are plotted as axes on a graph. Analyses of minerals or rocks which have the same age will then form points on a straight line (isochron). The slope of this line will be an expression of the age of the rock.

Using this method we can analyse both minerals and whole rocks, since we assume that the ^{87}Sr formed does not escape from the rock as easily as argon (in the K-Ar method). When dating sediments one should choose the finest-grained clay sediments with low permeability. It can then be assumed that after deposition the sediment was homogeneous with respect to strontium isotopes, and will provide an isochron which dates any diagenesis occurring a relatively short time after deposition. Larger clastic fragments, however, will contain a strontium isotope ratio which corresponds to the age of the source rock, and the obtained date will be intermediate between the age of the source rock and the time of deposition of the sediments.

7.6.4 Carbon 14 Method

The carbon 14 method is the one most commonly used for dating the youngest sediments, from about 50,000 years old up to the present. ^{14}C is formed in the atmosphere by cosmic rays when a ^{14}N atom absorbs a neutron and gives off a proton. ^{14}C is unstable, and decays to ^{14}N. The production of ^{14}C thus occurs only in the atmosphere, at altitudes over 10,000 m. The ^{14}C then mixes with the lower air layers and the seawater.

The ^{14}C method is based on the assumption (discussed below) that the ^{14}C content of the atmosphere has been constant for a long time, due to an equilibrium between the ^{14}C added from the atmosphere and the ^{14}C which decays to ^{14}N. The half-life of ^{14}C is 5,730 years. This means that after 5,730 years half of the ^{14}C atoms will have changed to ^{14}N. ^{14}C enters the carbon dioxide (CO_2) in the air and is taken up by plants through photosynthesis. If we assume that the carbon dioxide in the air in the past had as much ^{14}C as now, the ^{14}C content of older plants or plant remains is an expression of their age. This age can be determined analytically with relatively great accuracy, for example 10,200 ± 100 years, depending on the nature of the sample. When 50–60,000 years has elapsed since plant material formed (i.e. since it ceased to take up CO_2), the ^{14}C content will be so small that we will be approaching the limit of detection. This is then the upper limit to the age of material we can analyse. We now know that the concentration of ^{14}C has varied in the last 10,000 years and more due to variation in cosmic radiation.

^{14}C from the atmosphere also becomes mixed with sea water and freshwater. Carbonate-secreting organisms which live in the sea and in lakes will take up CO_2 from the water, and the amount of ^{14}C in their $CaCO_3$ skeletons can be measured. Molluscan and foraminiferal shells are particularly suitable for age dating. Chemically precipitated carbonates can also be dated in this way. Living organisms have a tendency to fractionate the lightest isotopes from the heaviest. The ratio between two stable carbon isotopes, ^{12}C and ^{13}C, can therefore be used to correct ^{14}C determinations.

7.6.5 Other Radiometric Methods

There are numerous other radiometric dating methods, of which the uranium-lead method and the lead-lead method (relationship between different lead isotopes) are the most important.

The methods based on fission processes mentioned above, involve long half-lives, 10^9–10^{10} years, so for young sediments the quantity of decay products available to be analysed will be small and the accuracy poor. An exception is the protactinium method $\left(^{231}Pa/^{230}Th\right)$, which has given age determinations on younger sediments showing good agreement with the ^{14}C method.

The fission-track method is especially well suited for younger rocks. It is applied to glass or minerals which contain a sufficient amount of uranium 238. The material will show tracks from fission products which are observed as deformations. By counting the number of such tracks, an expression of the age is obtained. In apatite the tracks disappear (aneal) at about 100°C, which is called the blocking temperature. The frequency of tracks is therefore an expression of the time elapsed since rocks have been uplifted and cooled to below the blocking temperature.

7.7 Chemostratigraphy

7.7.1 The $^{87}Sr/^{86}Sr$ Method

The $^{87}Sr/^{86}Sr$ ratio is particularly useful for carbonates and other minerals precipitated in the ocean. In the ocean water the $^{87}Sr/^{86}Sr$ ratio has varied greatly over geologic time (Fig. 3.8). By analysing this ratio in minerals precipitated in marine environments the ages can be constrained. During Tertiary time, the $^{87}Sr/^{86}Sr$ ratio increased and analyses of calcareous organisms will give a unique age, in some cases with a resolution of 1–2 million years. This method can be applied also to calcareous microfossils such as foraminifera and coccoliths. In carbonate the Sr/Ca ratio may be useful also in detecting depositional facies, because the data will reflect the primary aragonite/calcite ratio.

7.7.2 Bulk Chemical Composition Analyses

Chemical analyses of bulk samples of cuttings and cores may be useful in correlation between wells, because the chemical composition reflects changes in the composition of sediments supplied to the basin. Particularly good results are expected in correlation of reservoirs. For this purpose major elements, trace elements and isotopic composition can be used. The samples can be analysed by XRF, by energy dispersive systems in a SEM, or in mass-spectrographs. Analyses can be performed for bulk chemical composition.

The bulk chemical composition varies as a function of grain size but the ratio between major elements such as Na/Al and K/Al may be characteristic for the provenance rocks of sediments. Both sandstones and shales may have very different ratios between Na-feldspar and K-feldspar. Trace elements may be useful also in provenance studies.

7.8 Magnetostratigraphy

7.8.1 Palaeomagnetism

When clastic sediments are deposited or when volcanic rocks solidify, minerals orientate themselves in the prevailing magnetic field. Magnetic minerals in clastic sediments can orient themselves according to the magnetic field during deposition, while diagenetic minerals will become oriented in the magnetic field during diagenesis. The magnetic minerals in rocks thus define a magnetic vector, which indicates the direction and strength of the magnetic field during formation. By compensating for later magnetic effects, the orientation of this *remanent magnetism* can be measured. The data will reveal the position of the geographic pole, and indicate the palaeogeographical latitude and longitude.

Palaeomagnetic measurements are of great help in reconstructing the positions of the continents during the geological past, and are important for plate tectonic and palaeoclimatic reconstructions. From the Palaeozoic onwards there is quite good agreement between palaeomagnetic determinations of latitude and palaeoclimatic indications like biogeography, glacial deposits and evaporites, but this is not the case with Precambrian deposits, which are more difficult.

7.8.2 Magnetic Field Polarity Changes

It has become evident that the polarity of the Earth's magnetic field has been reversed for numerous time periods of varying duration. A number of measurements of magnetism in rocks of known age have provided us with a time scale based on periods of normal and reversed magnetic field. We thus find that we can divide geological time into periods with dominantly normal or reversed polarity (Fig. 7.11). Within these we also find several shorter intervals when the magnetic field switched between the two polarities. Since we must assume that the switching between normal and reversed magnetism has taken place simultaneously and suddenly all over the world, such physical changes offer an ideal basis for correlation.

There are often major practical problems, however, since many periods of the Earth's history are characterised by a predominance of successions with rapidly changing reversals. Where we have many measurements and continuous profiles, e.g. in deep-sea cores, we will be able to correlate with relative certainty on the basis of the longer periods of magnetic field stability. Volcanic rocks and sediments deposited in fluvial or shallow-water environments, however, will have numerous hiatuses between beds, hampering registration of continuous variations in the residual magnetism. During the last 700,000 years we have had apparently normal polarity, possibly with the exception of a short period about 200,000–300,000 years ago. If we find sediments or volcanic rocks

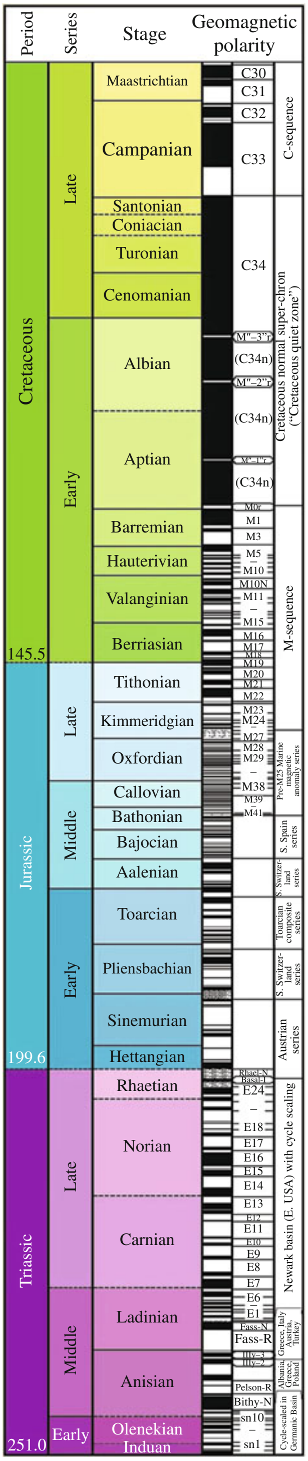

Fig. 7.11 Reversals of the Earth's magnetic field from Triassic to Cretaceous time. Normal polarity: black bands, reversed polarity: white bands (From Gabi Ogg 2010, http://stratigraphy.science.purdue.edu/charts/educational.html)

with reversed magnetism, we know that they are highly probably more than 700,000 years old.

7.9 Sequence Stratigraphy

7.9.1 General Aspects

Sequence stratigraphy is defined as the subdivision of sedimentary basin fills into genetically related stratal packages bounded by unconformities and their correlative conformities. Thus, the concept of sequence stratigraphy is closely related to that of allostratigraphy. Allostratigraphic analysis applies bounding discontinuities, e.g. erosional surfaces and marine flooding surfaces, for recognition of sedimentary entities independent of any genetic model. Sequence stratigraphy applies allostratigraphic features to interpret the depositional origin of sedimentary packages by analysing these in a framework of transgressive-regressive developments producing base level changes.

The infilling of a basin is controlled by the interaction of tectonics and eustasy, producing base level changes which define the accommodation space for sediments to be deposited. This is the space or height between the seafloor and the sea level. The evolution of basin architecture is controlled by the balance between accommodation space and sediment supply. If sediment supply exceeds the accommodation space available, *progradational geometries* will be the result. If accommodation space and sediment supply are roughly balanced, *aggradational geometries* result. When sediment supply is less than the creation of accommodation space, *retrogradational geometries* are formed. Tectonics and climate are the most important factors determining the production and supply of sediments.

The major bounding and subdividing surfaces of an Exxon-type depositional sequence are commonly represented by: the lower and upper sequence boundaries, transgressive surfaces and maximum flooding surface (Fig. 7.12). These surfaces principally define three sediment bodies: lowstand systems tract, transgressive systems tract and regressive systems tract. The building blocks of the systems tracts are parasequences, forming parasequence sets.

The boundary of a depositional sequence is defined as a subaerial unconformity and its correlative

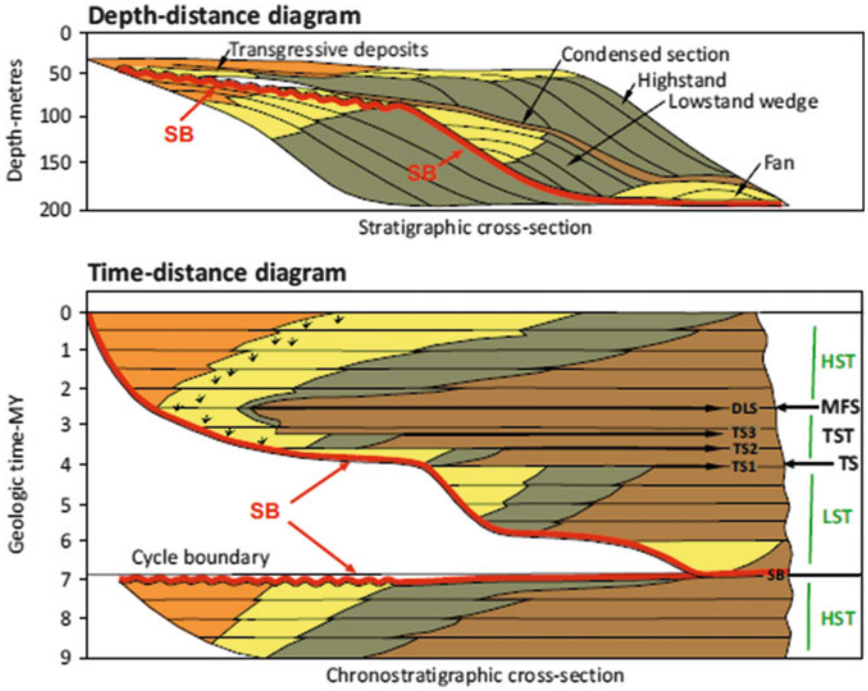

Fig. 7.12 Depositional sequence portrayed in a depth-distance and a time-distance diagram. It shows the regional extent of the condensed section from deep to shallow water, with its time span strongly expanding from coastal to basinal areas. The time span of the sequence boundary unconformity decreases from marginal to basinal areas. Abbreviations of sequence elements: SB = sequence boundary, LST = low stand systems tract, TS = transgressive surface, MFS = maximum flooding surface, HST = high stand systems tract (modified after Loutit et al. 1988)

conformity. This can be recognised as a change in lithology and facies. The sediments between two unconformities or their correlative maximum regressive surface have been called a transgressive-regressive (T-R) sequence.

7.9.2 Accommodation Space

Local accommodation space for sediments represents simply the water depth in the depositional area. A global factor influencing the accommodation space is the eustatic sea level stand which is defined by a combination of several factors. One of these is the volume of the ocean basins, which is believed to be mainly controlled by the rate of seafloor spreading. Young lithosphere is warm and buoyant, therefore it will rise and decrease the volume of the world ocean. On the contrary, old lithosphere is colder and denser, and sinks. Another factor contributing to ocean volume changes is the sedimentation rate.

The volume of water in the ocean basins is also a primary factor influencing eustatic sea level changes. Continental and mountain glaciations and deglaciations decrease and increase, respectively, the volume of ocean water. These developments are exemplified by the well-documented regressions and transgressions accompanying Quaternary glacial and interglacial periods (Fig. 7.13). Other factors influencing the volume of oceanic waters are the expansion of water with increasing temperature and the amount of continental groundwater and surface water.

Tectonic movements result in transgressions and regressions of local to regional nature. These movements are driven by factors that affect how continental lithosphere floats on the asthenosphere, which is controlled by three main mechanisms: (1) Stretching of continental lithosphere resulting in replacement of continental lithosphere by denser and thinner astenosphere which sinks. (2) During stretching, continental lithosphere is heated, becomes less dense and

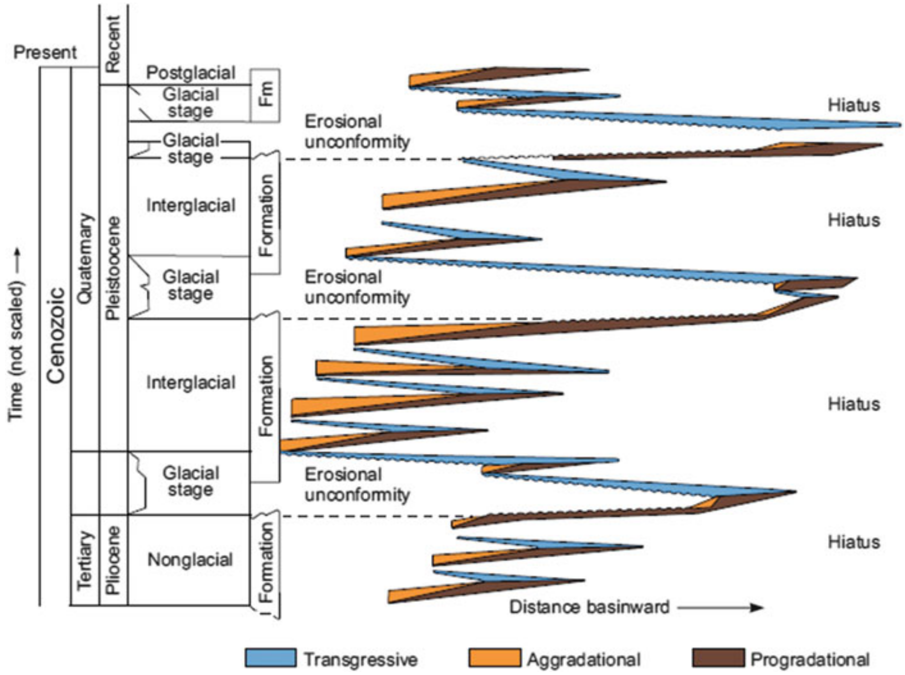

Fig. 7.13 Schematic illustration of transgressive-regressive cycles in the Quarternary, resulting from glacially-induced changes of the volume of ocean water. During a complete cycle, sedimentation takes place only during short episodes (note the three types of depositional architecture). During most of the time there is non-deposition or erosion (hiatus)

tends to uplift. Subsequent cooling results in subsidence. (3) Weight of tectonic load added to lithosphere can produce subsidence e.g. wedges of fold and thrust belts pushed into foreland basins.

When analysing sedimentary successions, it is usually extremely difficulty to distinguish between the effects of eustatic sea level changes and tectonic movements. Therefore, in sequence stratigraphic studies, the phrase "relative sea level" is used to express the local sum of global sea level and tectonic movements.

7.9.3 Parasequences

Parasequences are sediment packages of genetically related beds representing a single minor transgression event followed by sediment progradation. In the marine realm parasequences are bounded by flooding surfaces. Most of the marine parasequences are formed in offshore shelf to shoreface environments, and show an upwards-coarsening development (Fig. 7.14). In coastal areas upwards-fining parasequences are formed during transgressions, encroaching over the supratidal zone, and contain shallowing-up subtidal, intertidal to supratidal strata (Fig. 7.15).

Stacked parasequences form parasequence sets, which show stacking patterns according to their position in the sequence stratigraphic architecture (Fig. 7.16). If sediment supply exceeds the available accommodation space, progradational stacking geometry will be the result. If accommodation space and sediment supply are roughly balanced, aggradational geometries occur. When sediment supply is less than the creation of accommodation space, retrogradational stacking geometries are formed.

7.9.4 System Tracts

The lowstand systems tract is bounded by the sequence base and the first transgressive surface (Fig. 7.12). It was deposited when the sea level was located at or below the shelf break. The shelf was exposed to erosion and crossed by rivers eroding incised valleys and transporting sediments to the shelf margin. The lowermost part of the lowstand

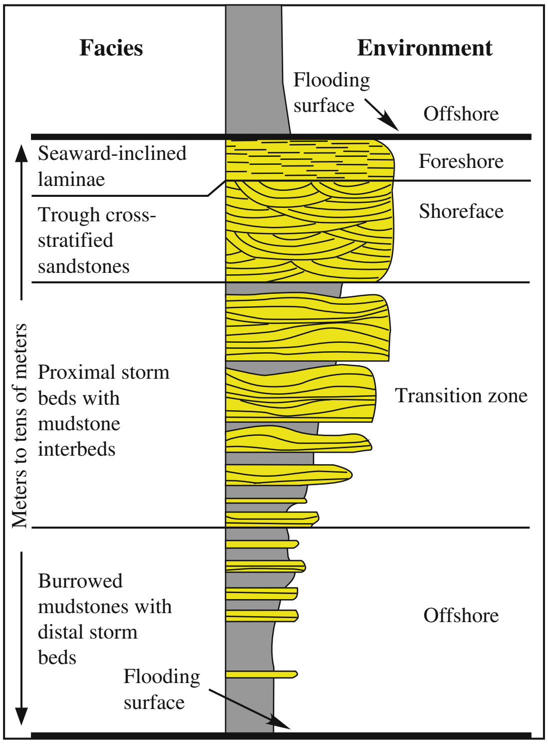

Fig. 7.14 Example of marine parasequence showing coarsening- and shallowing-upwards from offshore shales to foreshore sandstones (from University of Georgia web site, 2009)

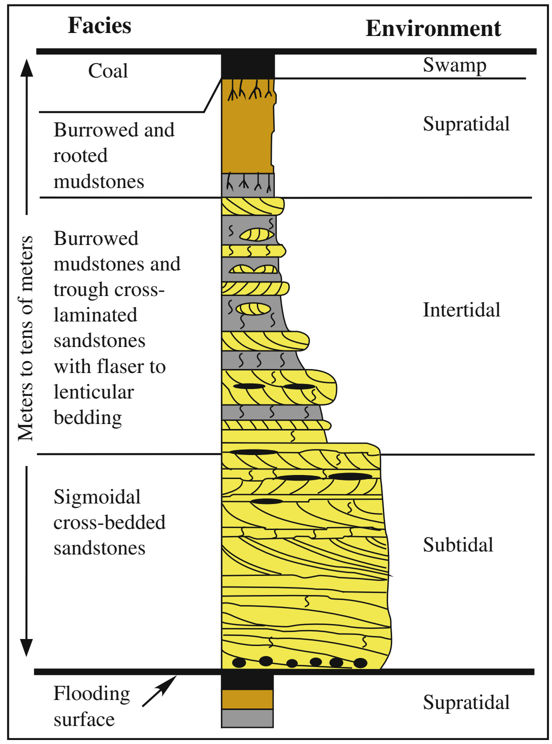

Fig. 7.15 Example of coastal parasequence showing fining- and shallowing-upwards from subtidal sandstones to supratidal swamp deposits (from University of Georgia web site, 2009)

systems tract is the basin floor fan, dominated by gravity-driven deposition of turbidite lobes and in feeder channels. The upper part of the lowstand systems tract is the lowstand wedge, which consists of progradational parasequences. Uppermost, estuarine sediments are deposited in drowned incised valleys (Fig. 7.17).

The transgressive systems tract is bounded by the transgressive surface and the maximum flooding surface. It is composed of retrogradational parasequence sets showing a generally upward-fining and upward-deepening development. This systems tract is generally thin because much sediment is trapped in estuaries, and the advancing transgression successively creates new accommodation space.

The highstand systems tract is developed between the maximum flooding surface and the upper sequence boundary. This systems tract represents a relatively thick succession of progradational parasequences showing an upward-coarsening and upward-shallowing development (Fig. 7.17). The parasequences downlap to the maximum flooding surface. The estuaries are now filled up by sediments, therefore the systems tract is characterised by outbuilding of deltas and shorelines.

7.9.5 Condensed Sections

The transgressive surface and maximum flooding surface were formed during periods of transgression characterised by sediment starvation, which leads to development of condensed intervals forming important marker horizons in the sequence stratigraphic architecture of a basin (Fig. 7.12). Condensed sections are thin marine horizons commonly showing several of the following features: pelagic to hemipelagic sedimentation; presence of apparent hiatuses; borrowed horizons and hardgrounds; presence of authigenic minerals such as glauconite, phosphate, siderite; high faunal abundance (if not anoxic); high faunal diversity (if oxic), low diversity (if hypoxic). Deposition of organic-rich black shales with high gamma readings is typical of many condensed sections.

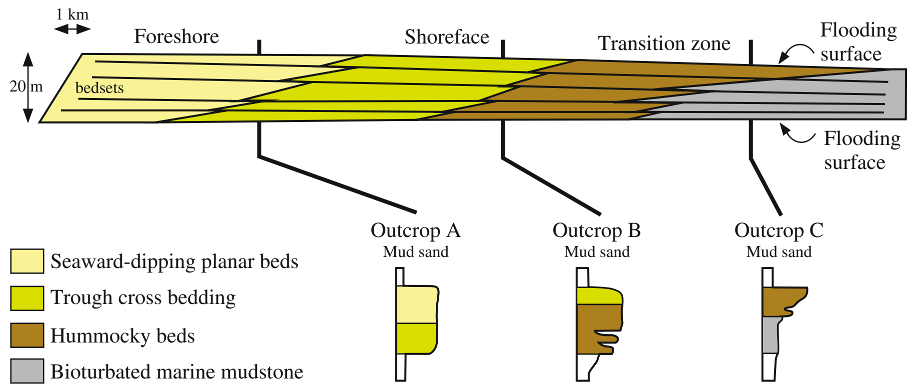

Fig. 7.16 Lateral and vertical relationships between parasequences forming a parasequence set, which progrades from foreshore to offshore shelf (from University of Georgia web site, 2009)

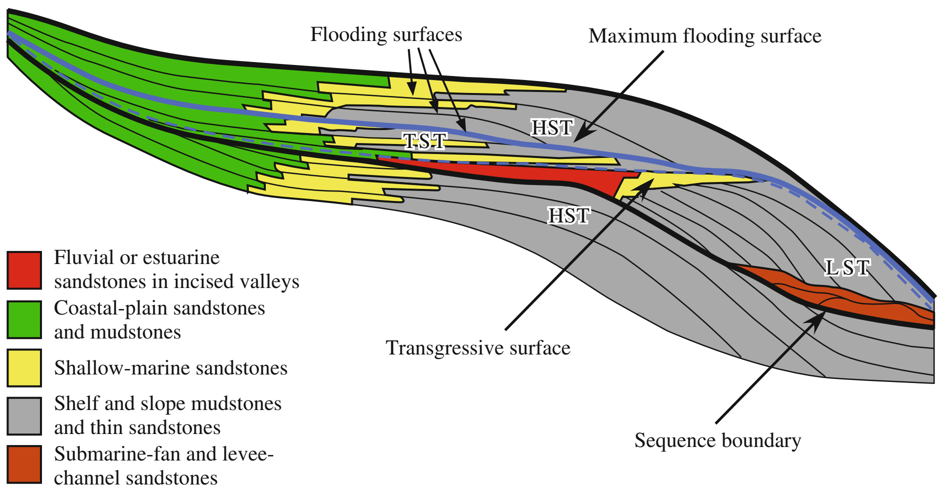

Fig. 7.17 Depositional sequence (type 1) portrayed as a down-dip section across an idealised passive continental margin with shelf break. The diagram shows sequence boundaries, systems tracts and parasequences. LST = low stand systems tract, TST = transgressive systems tract, HST = high stand systems tract (from University of Georgia web site, 2009)

Condensed sections have a large regional extent, and show an expanding time span from coastal to basinal areas (Fig. 7.12). They are particularly well developed at the maximum flooding surface and help to identify this horizon. Thus, condensed sections represent important levels of correlation within sedimentary basins. They are also applicable to recognition of global transgressive-regressive cycles.

7.9.6 Sequence Types

During development of the sequence boundary the relative sea level fall is of varying magnitude. The sea level fall can continue to a position below the shelf-slope break, or stop on the shelf. According to these developments, two types of sequences are distinguished.

Type 1 sequence: This is formed if the sea level falls below the shelf break at the sequence boundary (Fig. 7.17). The shelf is exposed for erosion and a low stand systems tract is formed. The boundary reveals a sea level fall below the previous shoreline. A type 1 sequence can be formed in two basinal developments: by deposition in a basin having a shelf-slope break; or by deposition in a basin that lacks a shelf-slope break but has an evenly sloping ramp-like margin.

Type 2 sequence: The sea level position above the sequence boundary corresponds to the position of the previous shoreline below the boundary. Consequently, the boundary reveals an aggradational development. During the sea level low stand a shelf margin systems tract is deposited.

7.9.7 Carbonate Sequence Stratigraphy

Sequence stratigraphy was originally developed for siliciclastic systems, but the same principles can be applied to carbonate systems as well. However, the special features of carbonates make the appearance of stratigraphic elements somewhat different in the two system types.

Carbonate production takes place mainly *in situ*, and is particularly high during moderate sea level rise, which typically results in extremely thick transgressive systems tracts. Parasequences formed in peritidal environments are particularly thick. The highstand systems tracts in carbonate successions are much thinner than in siliciclastic sequences. During transgression, extremely rapid sea level rise stops *in situ* carbonate production, which leads to the formation of condensed sections. Carbonates are prone to dissolution, leading to sequence boundaries commonly marked by solution surfaces, karst development and palaeosols.

Further Reading

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