Chapter 8

Seismic Stratigraphy, Sequence Stratigraphy and Basin Analysis

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8.1 Seismic Stratigraphy

Seismic records are based on measurements of the time sound waves (seismic waves) take to travel through rock. The sound or signal is produced by explosives or compressed air (air guns). Rock is an elastic medium and the velocity of sound conveys a lot of information about the properties of the rock. Normal sound waves (P-waves) travel through both the solid phase, which for the most part consists of minerals or rock fragments, and the liquid or gas in the pores. Shear waves (S-waves) on the other hand can only go through the solid phase.

The most important parameters influencing the velocity of sound are: porosity, mineral composition, and the degree of cementation. These factors determine the stiffness of the rock (Bulk modulus, see Chap. 11). The velocity of sound waves in water is about 1,500 m/s, but depends on temperature and salt concentration. Sound passes through unconsolidated sediments at velocities which are only slightly higher than the velocity in water (1,500–2,000 m/s, and sometimes even lower) because they have high water content and because the framework on which the sediment grains are based does not offer any real strength (stiffness) as a medium for the seismic waves.

Cementation of sand with carbonate or siliceous cement will bind the grains together in a framework which will increase the stiffness and velocity considerably even if the porosity is relatively high. Compaction due to overlying sediments which causes water to be expelled will also give higher velocities, not only because the water content decreases, but because more numerous and larger contacts are formed between the clastic grains. Velocities in moderately consolidated sediments, such as the Tertiary sediments of the North Sea, are 2–3 km/s. In more consolidated (compacted and cemented) sedimentary rocks which have not been subjected to metamorphosis, velocities are mostly between 3 and 5 km/s. This is the case for many of the Mesozoic sediments in the North Sea. Metamorphic and eruptive rocks will have velocities of about 5–6 km/s. Limestones will often have higher velocities than sandstones at the same depth because they often are more cemented and because carbonate cement has a high degree of rigidity and low compressibility. Carbonate reefs may be strongly cemented and have high velocities at shallow depth. Sandstone in turn provides a more rigid medium for sound waves than shale at the same depth, because of its grain-supported structure.

If the rocks do not contain oil and gas we can assume that their porosity is identical with the water content in the rock. Velocity will then be a function of porosity (φ), and if we know the velocity of sound in the rock matrix, we can calculate the porosity using Wyllie's equation:

$$1/V_{\mathrm{r}} = (1-\varphi)/V_{\mathrm{m}} + \varphi/V_{\mathrm{f}}$$

where
V_{r} = velocity in rock when saturated with liquid, i.e. the measured velocity
V_{f} = velocity in the fluid
V_{m} = velocity in the rock matrix.

The inverse values of the velocities are expressions of the time the signals take to travel through a layer of certain thickness.

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If the pores are filled with gas instead of liquid (water or oil), the velocity reduction will be even greater because the sound travels much more slowly through gas than through a liquid. Whyllie's equation is however not an accurate description of the relation between porosity and velocity.

V_r thus approaches V_m at zero porosity, and for sandstone V_m is about 5.5–6 m/s.

If we know the velocity of the rock matrix and the fluid we should be able to calculate the porosity as a function of velocity, but the Wyllie equation is very much a simplification. Rocks with the same porosities can have rather different velocities depending on the type of grain contacts and the distribution of cement.

When sound waves move between sedimentary beds with different velocities, they will be refracted according to Snell's law (see Fig. 8.1):

$$\sin x_1 / \sin x_2 = v_1/v_2$$

Here x_1 is the angle of incidence of the waves where they meet the boundary plane between two strata, and x_2 is the angle of refraction of the emergent waves. v_1 and v_2 are the velocities through the respective rock strata.

If the two beds have different velocities, they will as a rule also have different densities, and part of the acoustic energy will not be refracted, but reflected. How much of the energy is reflected depends on the difference in the *acoustic impedance*, which is the product of velocity and density (Fig. 8.2).

The coefficient for reflection (R) is then:

$$R = (\rho_2 \cdot v_2 - \rho_1 \cdot v_1)/(\rho_2 \cdot v_2 + \rho_1 \cdot v_1)$$

where ρ_1 and ρ_2 are the densities of the two rocks, and v_1 and v_2 their respective velocities (Fig. 8.2). We see that the greater the difference in density and velocity, the greater the amount of energy which will be reflected. Sandstone will often have significantly different acoustic impedance from shale, and a considerable amount of sound energy will be reflected from the boundary between a sandstone bed and a shale bed.

This is however not always true and the contrast depends on the type of clays and their clay mineral composition. Limestones will tend to have both high velocities and high densities. The result will be even greater contrast in acoustic impedance between limestones and, for example, shales. However, this contrast will always depend on the porosity of the limestone in question.

On a seismic section, beds which have greatly contrasting acoustic impedances stand out as strong reflectors. This makes it possible to map characteristic rock boundaries, e.g. the top of a limestone or the boundary between shales and sandstones, using seismic sections (Fig. 8.3).

As we have seen, the critical parameters determining the reflection coefficient are the velocities and densities of the different lithological units. Using a well, we may measure a velocity log (sonic log) and a density (ρ) log which record how these properties change through the sequence (see Fig. 8.3a). The product of velocity and density may then be computed and presented as an acoustic impendance ($p{\cdot}v$) log.

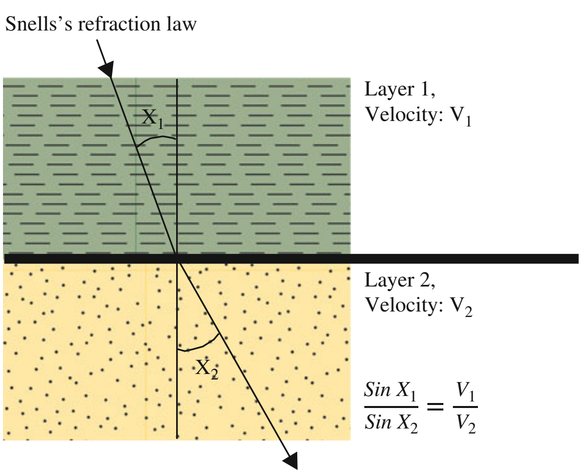

Fig. 8.1 Snell's law for the refraction of sound waves

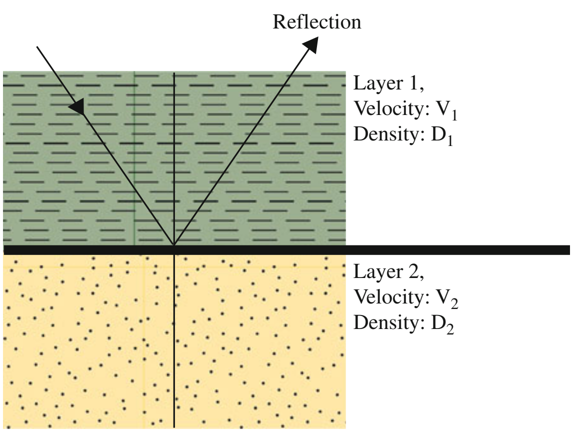

Fig. 8.2 Diagram for the reflection of waves in a layered sedimentary sequence. The amount of energy reflected is a function of acoustic impedance, which is the product of the density of the beds (ρ) and the velocity of the sound waves (v)

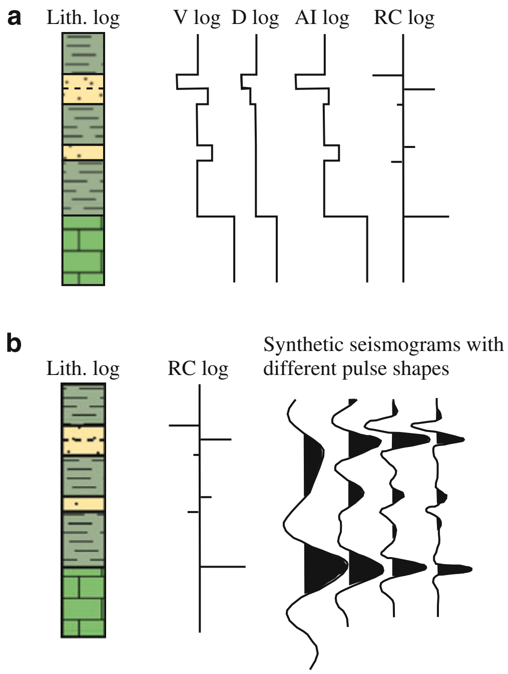

Fig. 8.3 (**a**) Illustration of a reflection coefficient log based on velocity (v) and density (ρ) (from Anstey 1982). (**b**) Synthetic seismogram of the different reflection coefficients (RC log) in a bedded sequence. The resolution of the seismogram varies with the width of the seismic pulse (from Anstey 1982). Normally, sediments have to exceed 20–30 m in thickness to be distinctly recorded on a seismic section, depending on the wavelength of the seismic signal. The upper sand contains some gas in the upper part, causing lower velocities

Note that at the boundary between limestones and shales there is a very significant drop in both velocity and density, resulting in a marked change on the acoustic impedance log. Sandstones usually have higher velocities than shales, but they may not differ much in density, so the difference in acoustic impedance will be small. The reflection coefficient, which is an expression of differences in acoustic impedance, is a synthetic seismic trace such as we would have seen on a seismic cross-section through the sequence.

If we have gas instead of water in a rock, the velocity will be considerably reduced. The velocity of sound in gas is much lower than it is in liquid, depending on composition, temperature and pressure. The boundary between gas-bearing and water-bearing rocks may produce a strong reflection because there is a large difference in impedance between the two layers. For this reason the boundary between gas and oil is often revealed as a strong reflector because it is horizontal and does not always follow the other rock strata. This is called a “flat spot” and exemplifies direct indication of hydrocarbons (usually gas) through seismic methods. At greater depth and higher pressure the contrast between gas and oil and also oil and water will be lower.

Reflections which are multiples of a relatively flat sea-bottom reflection are also near-horizontal and may be confused with “flat spots”, but these may be removed by filtering the data during processing. Temperature-dependent diagenetic reactions may also produce horizontal reflections, e.g. the transformation of amorpheous silica (opal A) and opal CT to quartz which produces a strong increase in velocity and density. If the geothermal gradients are rather uniform (horizontal isotherms) this diagenetic transition will form horizontal reflections which may crosscut the bedding.

It may also be possible using AVO (see Chap. 17) to detect contacts between oil and water. Oil has a lower velocity than water and if the oil has a high gas content, the difference is even greater.

Changes in the oil/water contact during production can be monitored by shooting seismic in an oilfield at intervals of several years (see 4D Seismics). In deeper reservoirs with lower porosities it is more difficult to detect fluid contacts.

High pore-pressure causing reduced effective stress and stiffness (elasticity) results in lower seismic velocities in sandstones and clays, particularly at depths less than 3 km. Limestones can have relatively high velocities even at shallow depth.

At greater depths chemical compaction is the most important factor, and in clastic sediments it is generally a function of temperature and less dependent on the effective stresses. Nevertheless, seismic velocities are often observed to fall in overpressured rocks even if the porosity is not significantly reduced. This may be related to reduced stress at grain contacts.

As the quality of seismic data has improved, one has been able to use seismic profiles for detailed interpretation of stratigraphic relations and even depositional environments. The basis for this is that seismic reflections usually follow time lines in a sedimentary sequence. In other words, seismic reflections follow surfaces which constituted the seafloor surface at the time when the sediments were deposited.

Seismic reflections can be followed from a sandy facies into a clay/siltstone shale facies. We can, for example, follow seismic reflections from the fluvial part of a delta out into the pro-delta muds. Relatively thin transgressive sandstones may be deposited on the delta top, and are useful for correlation. This is because the relief on land may be very low so that a few metres sea level rise causes a large transgression. Transgressions will then result in a wide shelf with little clastic supply, causing deposition of carbonates. Thin calcareous sediments and limestones may therefore also be relatively close to time lines. Limestones typically have rather high velocities and so we find strong reflectors. On the delta slope, however, fluvial sediments and marine delta front sand may prograde into a marine basin over considerable geological time, depending on the depth of the basin and the sediment supply. The seismic reflections will follow the surface from delta front sand to delta slope, where we have sand and mud (shale) beds lying parallel with the slope. Even though we can often follow the reflector a little further into the basin, it will be less marked there because there is less contrast in lithology and hence in acoustic impedance. Different types of clay may however also produce differences in seismic response and particularly smectitic clays are characterised by low density and velocity.

The shifting of sedimentation input from one part of the delta to another through channel switching (distributary abandonment as part of delta-lobe shifting), also contributes to the formation of lithological contrasts on the delta slope. Progradation of new delta lobes results in deposition of sheets of sand over mud near time-stratigraphic boundaries. The inactive delta lobes will be compacted and often develop a thin carbonate or transgressive sandstone layer, while sedimentation takes place in the active lobe. The small unconformities produced in this way also tend to produce lithological contrasts which may be recorded on the seismic record.

Seismic sections through prograding deltas provide information about the water depth, the rate of sediment input and the wave energy in the basin.

8.2 Different Types of Seismic Signatures

A stratigraphic unit which is composed of a conformable bedding series, genetically linked together at the top and bottom by unconformities, is called a *depositional sequence*. A depositional sequence is thus a package of sediments deposited during a definite period of time, defined by unconformities above and below.

The unconformities may be due to a break in sedimentation due to relative changes in sea level or other causes such as changes in sediment supply.

A seismic profile through a sedimentary succession has reflectors that show layers of contemporary deposits. Terminolgy has been established to describe the geometry of seismic reflections (Fig. 8.4).

Baselap is the term for gradual deposition above the lower boundary of a depositional series and represents a small unconformity.

If a sequence progrades out across an unconformity, depositing successively younger beds basinwards, we call this type of contact *downlap*. The building of beds out into the basin like this is called *offlap*. We are thus dealing with a bed which has a primary depositional slope with respect to the unconformity surface.

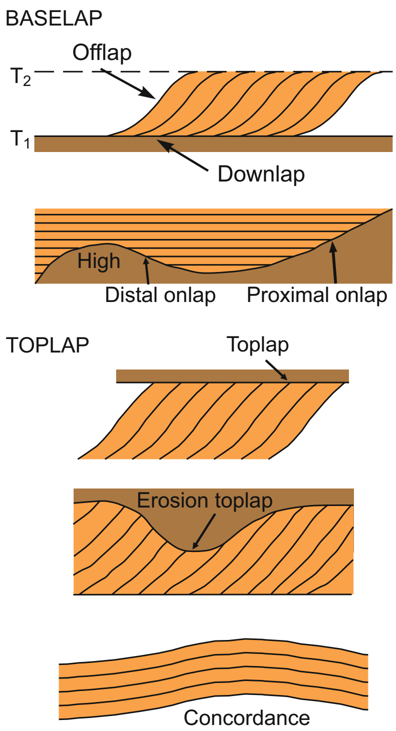

Fig. 8.4 Different types of seismic stratigraphic relations. The seismic reflectors represent time lines as a rule, i.e. rocks deposited at the same time

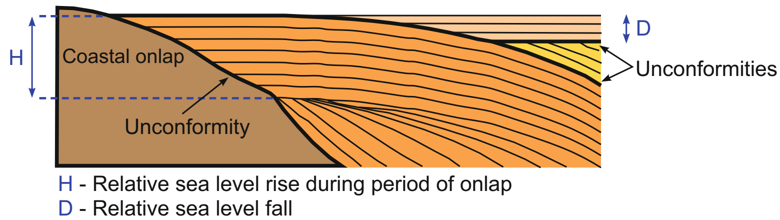

Fig. 8.5 Coastal onlap followed by a sea level drop and a renewed onlap. The coastal onlap indicates a relative sea level rise of H metres but this may be due to local tectonic movements and loading by water and sediments

Onlap is the term for primarily almost horizontal beds against a sloping unconformity which may be a submarine or a subarial slope. It occurs most commonly as a result of sedimentation gradually burying an unconformity during a transgression onto a land surface that provides the unconformity (Fig. 8.5). We call this *proximal onlap* or *coastal onlap*. If sedimentation covers a positive relief structure in the basin, for example a horst or a salt dome, we get *distal onlap* (Fig. 8.4).

Toplap is the contact between the seismic reflectors and an upper unconformity. An erosion surface will truncate the reflections sharply, forming an erosional truncation (*erosional toplap*).

Transgression may form a coastal onlap across an unconformity (Fig. 8.5). Later the relative sea level fell, and a prograding offlap sequence formed. Finally the sea level rose again and an onlap sequence formed. H and D are not to be regarded as absolute values for sea level changes. They must be adjusted for isostatic responses to loading and unloading and to tectonic uplift or subsidence. H is the relative rise in sea level during the period with onlap. D is the relative fall in sea level which led to a downward shift in the prograding downlap.

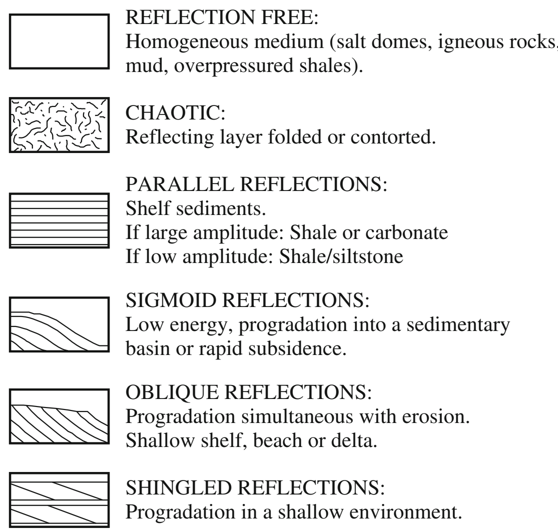

Fig. 8.6 Classification of internal structures in seismic units. Layering in the sedimentary sequences causes changes in the acoustic impedance

8.3 Interpretation of Lithology and Sedimentary Facies by Means of Seismic Profiles

In addition to structural data, a seismic profile provides us with information about the properties of sedimentary rocks. The internal properties of seismic units provide important information about the lithology (Fig. 8.6).

Because seismic reflections mainly represent time lines, i.e. sedimentary beds which were deposited contemporaneously, it is also possible to a certain extent to interpret the depositional environment. The most important parameters we use are:

1. *Reflection amplitudes* – the strength of the reflections. As we saw above, the proportion of the energy reflected at the boundary between two beds is a function of the difference in the acoustic impedances (velocity multiplied by density). If we have an alternating series of different beds, the distance between the bed boundaries in relation to the wavelength of the transmitted seismic waves will play a major part (Fig. 8.3).
2. *Reflector frequency*. The distance between the reflectors will indicate the thickness of the bed, but there will be a lower limit to the thickness that can be detected, which should correspond to about the half wavelength of the seismic waves.

3. *Internal velocity in the beds*. The internal velocity of the bed can provide information about lithology and porosity.
4. *Reflector continuity*. The continuity of reflectors will be a function of how continuous the sediment beds are, information which is essential for reconstructing the environment.
5. *Reflector configuration*. If we take the compaction effect into account, the shape of the reflecting beds gives us a picture of the sedimentation surface as it was during deposition. The slope of the reflectors, for example, represents the slope of prograding beds in a delta sequence with later differential compaction and tilting superimposed. Erosion boundaries with unconformities will in the same way show the palaeo-topography during erosion.

Seismic profiles provide information about the filling of the basin in response to the type of subsidence and sedimentary facies.

When interpreting lithology and depositional environment from seismic profiles, it is important to use sedimentological models as aids. If there are well data on lithologies, these must also be integrated. The information we gain from seismic profiles is often not sufficient by itself for an unambiguous interpretation. There may be several lithological compositions and environments which could give similar seismic signatures. Only by looking at the whole basin in a sedimentological context do we have a good basis for selecting the interpretation which seems most reasonable.

8.4 Seismic Interpretation of Sedimentary Basins

Seismic profiles present a picture of the way the basin has been filled in. This is a result of an interaction between the rate of subsidence, rate of deposition and the energy of the depositional environment. The seismic signatures can be used to interpret facies and basin infilling.

8.5 Faults and Tectonic Boundaries

It must be remembered that the principle we use for calculating the depth to a reflecting boundary assumes that the layering is relatively horizontal.

Primary seismic reflections will be deformed through tectonic deformation so that they become tilted or folded. Folded beds will only be realistically depicted if the folds are sufficiently gentle that the beds have a low angle of dip.

We can distinguish faults where good reflections suddenly stop, suggesting an abrupt lateral change in lithology. Faults are generally too steep to reflect the sound wave straight back again, and the fault plane itself will not appear as a reflector on the seismic profile. Because of the special "edge" effects near faults, the ends of the reflecting layers which should define faults will not be quite correctly located on the seismic profile, and it may therefore be difficult to trace the fault entirely accurately. The termination of beds against faults may produce diffraction from a point source, giving a curved alignment. Special treatment of seismic data (migration) will rectify a fair number of these errors and give a more correct picture. In recent years seismic lines have been shot with smaller and smaller grid spacings to obtain a better map of the reservoir structure. A three-dimensional seismic data set is then produced and seismic sections can be constructed at any angle relative to the grid. This method also allows us to construct horizontal time-slices through the structure. This is almost like a topographic or geological map which is a horizontal projection of the geology. It is also a very powerful method of delineating faults and other important structural elements.

Another relatively new development is *borehole seismics*, particularly vertical seismic profiles (VSPs). This method involves firing shots near the seafloor close to a well and recording signals at regular depth intervals in the well. The main advantage of VSPs is that they produce a very good profile of the seismic velocity as a function of depth, better than a synthetic seismic log.

8.6 Changes in Sea Level

It has been clear for a long time that there are unconformities in sedimentary sequences which can be correlated over long distances, and that there were periods in geological history with a high sea level and others when it was low.

Proximal onlaps are due to sedimentation moving landwards over an unconformity surface. If we are dealing with a coastal deposit, a proximal onlap will mean that the sea level has risen in relation to the land surface which forms the top of the unconformity. On seismic profiles we can see onlaps onto the land, measure the height range between the lowest and uppermost onlaps, calculate the difference in seismic time and convert this into approximate thickness.

However, we must remember that the thickness of the sediments deposited is due not only to a rise in eustatic sea level, but also to local subsidence of this part of the basin. The weight due to increased water depth will cause further subsidence, and sedimentation will increase the load, resulting in further subsidence to attain isostatic equilibrium. Local tectonic subsidence may produce a relative change in coastal onlap in a seismic profile. Regressions are defined as the boundary between land and sea being displaced out into the basin. They may be caused by a fall in sea level which will shift the coastline to further out on the shelf or to the edge of the continental slope. Here unloading of some of the water plus erosion of sediments leads to isostatic uplift of the area landward of the coast line, so that the measured regression is greater than the real lowering of the sea level.

The definition of a transgression is that the sea encroaches over what was previously land, while a regression is a situation where the boundary between sea and land ("shoreline") moves seaward so that seabed becomes land.

Transgressions and regressions are not always directly related to sea level changes. When a delta builds out into the sea, there is a local regression on the delta even if the sea level has not fallen. If sedimentation is sufficiently rapid, we can have a local regression on a delta even with a rising sea level.

Along a coastline we can have transgressions in some areas and regressions in others at the same time, depending on the rate of sedimentation or erosion with respect to sea level change.

The term "forced regression" is used to indicate that there is a primary lowering of the sea level.

Changes in sea level can be due to:

a. Local tectonic movements, for example uplift of a horst or subsidence of a graben structure.
b. Plate-tectonic movements which can be of great extent, but are not global.
c. Sea level changes. These are global and are called eustatic sea level changes.

During the Quaternary, cyclic changes in sea level of up to 120 m accompanied the growth and decay of continental ice sheets. These changes were rapid and of large magnitude and can be traced throughout much of the world. Nevertheless it is often the local conditions which count most. In the areas which had supported ice sheets, such as Scandinavia, the melting of the ice led to uplift due to unloading, and this exceeded the rise in sea level so that there was a regression. A 2,000 m thick ice sheet will cause about 600–700 m of isostatic depression of the crust under the ice because the density of ice (0.9 g/cm^3) is about one third of that of rocks (2.7 g/cm^3).

When seismic stratigraphy was established (Vail et al. 1976), it was assumed that most of the variations in sea level that could be interpreted from seismic records, were attributable to eustatic changes and thus could be employed for correlation across great distances. There was a problem in that one did not know of any other processes than the accumulation of ice on the continents capable of producing large and rapid global sea level changes. However we only know of such ice ages from the Quaternary and late Tertiary, the Carboniferous-Permian, late Ordovician and the end of the Precambrian.

The general consensus now is that the rapid sea level changes outside the ice ages were mainly caused by tectonic activity. Nevertheless, transgressions and regressions can be correlated over greater or smaller distances, depending on the type of tectonic movements. In particular large-scale plate tectonic movements involving changes in seafloor spreading rates and subduction rates cause global sea level changes.

In order to prove that a transgression is eustatic, we need rather accurate age determinations of the transgressive deposits from many parts of the world, but it is difficult to get high resolution datings. However seismic reflectors correlated with well data can be used to obtain good stratigraphic control and a picture of onlap and offlap can be interpreted in terms of sea level variation. They may represent relative sea level changes for a part of a sedimentary basin, or eustatic (global) transgressions or regressions.

Although one might perhaps expect that the sedimentation within a basin would primarily be characterised by local tectonic conditions, drainage, depositional conditions etc., studies of thousands of seismic profiles and wells from many sedimentary basins indicate that there have been simultaneous transgressions and regressions in completely different parts of the world. This has been documented by correlating seismic profiles with oil wells in the same area where the age of seismic unconformities and depositional sequences can be dated by means of biostratigraphy. It turns out that characteristic seismic reflectors which represent falls in sea level are of approximately the same age, for example in the North Sea, South China Sea, Mexican Gulf and Alaska. It has become clear that many areas, especially continental margins, have a rather similar tectonic history which is ascribable to global seafloor spreading.

Much of geological time is not represented by deposits, though; there are usually greater or lesser breaks in deposition (hiatuses). This becomes very clear when one looks closely at the continental sequences. Ocean floor sequences are more continuous, but also there one finds clear breaks in deposition. The time represented by breaks (hiatuses) varies greatly because there will always be some sedimentation somewhere, and correlation with sequences with continuous sedimentation entails trying to find signs of rapid changes in facies and deposition depth. Microfossils can be helpful in indicating water depth, even though they are not always reliable.

Mapping unconformities in the field can be difficult in the absence of abundant exposures. Many of the best examples have therefore been found from desert areas in the USA and elsewhere.

8.7 Changes in the Volume of the Ocean Basins

As we have seen, large-scale ocean floor topography is a function of the age of the seafloor, i.e. how long it has had to cool down since it was formed, and of the overlying sediment thickness. Periods with rapid seafloor spreading will result in relatively broad spreading ridges which cause the volume of the ocean basins to decrease, and seawater then will spread further onto continent margins. If all seafloor spreading ceased, the spreading ridges would slowly sink and within about 100 million years would have disappeared almost completely, The volume of the oceans basins would then be greater, as there would be a sea level drop corresponding to the volume of the ocean ridges. This mechanism can explain the great fluctuations in sea level through geological time. Note too that periods with major transgressions can be correlated with periods of rapid seafloor spreading, for example in the Cretaceous and Carboniferous periods. The deep channels formed in connection with subduction are, in fact, small in relation to the width of the spreading ridge. In Permian and Triassic times we had one big supercontinent and little seafloor spreading. This was a regressive period with a large land area.

Drying out of cut-off ocean basins may also lead to eustatic changes in sea level. There is much to indicate that the Mediterranean Sea was cut off from the Atlantic and dried up in Upper Miocene (Messinian) times. This reduced the world's total volume of ocean basins, increasing the sea level by about 5–6 m.

Crustal thickening and thinning. An increase in the depth to the *Moho* (seismic discontinuity separating the Earth's crust and mantle) will lead to elevation of the land. The greatest land elevation results from continental collision, when the thickness of the continental crust may be doubled (to 70–80 km), as in the Himalayas.

Stretching and thinning of the continental crust will move heavy mantle rocks upwards and increase the average density of the rocks down to a compensation depth of 100 km. This will lead to subsidence and more low density sediments can be accumulated. We see this at the transition between continental and oceanic crust, and where we have rift formation the continental crust thins below the rift, causing graben formation.

Variations in the temperature gradient affect the density of the rocks and thereby the isostatic equilibrium. Rifting causes elevation of the areas along the margin of the rift where the crust is not thinned (e.g. East Africa) and subsidence of the whole area when the rifting ceases and the crust cools (e.g. the North Sea).

The major transgressions in the Cambrian, Ordovician and Cretaceous, can be explained fairly satisfactorily by means of plate tectonic models. A relatively low sea level at the end of the Palaeozoic (Carboniferous-Permian) and the beginning of the Mesozoic (Triassic) can be explained as being due to limited seafloor spreading, for example along the Atlantic Ocean. With fewer and smaller spreading ridges the oceans could accommodate more water and consequently less seawater would flood the continents. On the continents there was active rifting without seafloor spreading and this increased the geothermal gradient and elevated areas which were previously covered by shallow epicontinental seas, to above sea level.

The short-term variations in sea level are more difficult to explain, however. For geological periods which experienced major continental glaciations (Quaternary-Upper Miocene, Permian-Carboniferous, Upper Ordovician and late Precambrian) we can resort to glacioeustatic changes in sea level (changes in sea level due to glaciation), which are very rapid in geological terms. Widespread sea level changes in the Lower Tertiary and Mesozoic must be due to plate tectonic factors.

Subsidence of the seafloor and continental crust of various thicknesses is a function of time after the rifting stage. The spreading ridges with hot basalt (oceanic crust) often lie at about 2 km water depth or above sea level (island). Seafloor basalt gradually subsides to almost 6 km due to cooling over 100 ma without sediment loading (Fig. 8.7a). Sedimentation on top of the sea floor basalt will gradually reduce the average density of the crust down to the compensation depth (about 100 km) and the water depth will then be reduced. Great water depth is found on old continental crust where there is little sediments (Fig. 8.7b).

8.8 Sedimentation and Isostatic Equilibrium

8.8.1 Why Do Sedimentary Basins Subside?

Sedimentary basins can be assumed to be in isostatic equilibrium in relation to the Earth's crust. However, the crust has a certain rigidity, and it takes time before equilibrium is attained after loading. Studies of uplift curves, for example from Scandinavia, show that during the course of 10,000 years the crust has undergone major adjustments in the form of uplift to compensate for the unloading of ice after the last glaciation. This is geologically a very fast response and we can assume that major sedimentary basins are in approximate isostatic equilibrium with regard to most geological processes. Nevertheless, even the filling of water reservoirs for hydroelectric plants causes local subsidence.

For the most part, then, we can apply the classical Airy isostasy model to sedimentary basins, which enables us to draw a number of interesting conclusions.

Uplift and subsidence are also linked to variations in geothermal gradients and heat flow in sedimentary basins.

Movement of rocks in relation to the surface affects the geothermal gradients. Erosion removes the uppermost, colder strata so that warmer strata come closer to the surface, and the geothermal gradient increases. As a result of subsidence of a sediment basin and sedimentation, heat flow upward will be partly offset by rock subsidence, giving lower geothermal gradients. Sedimentary basins with high rates of sedimentation are therefore often characterised as "cold basins". Tectonic elevation and erosion will increase geothermal gradients.

The geothermal gradient in seafloor rocks is consistently greater than it is over the continents, and on the continents it is highest in areas of volcanic activity.

Areas with a high geothermal gradient due to volcanism will slowly cool down to normal gradients when volcanic activity ceases. It may take about 100 million years before a normal geothermal gradient is re-established, due to cooling and contraction in accordance with the crust's coefficient of expansion and isostatic subsidence due to increased density. Seafloor basalt is hot and flows at relatively shallow depths in the Earth's crust, and the spreading oceanic ridges are only about 2.2 km below the surface of the sea. After about 180 million years of cooling, the water depth at isostatic equilibrium is about 5.7 km without sediment loading. With sediment loading the oceanic crust may subside to about 17 km (Fig. 8.7a). This is the theoretical maximum thickness of a sedimentary sequence overlying oceanic crust. Sediment basins on the continental crust have a sedimentary thickness which is a function of the thickness of the crust and the density of the sediments and the basement. The thinner the continental crust beneath a sedimentary basin, the more sediments can accumulate while maintaining isostatic equilibrium (Fig. 8.8). Along continental margins sedimentary basins have formed on thin (stretched) continental and oceanic crust.

The formation of sedimentary basins clearly requires that the density of the rocks below the basin is greater than that of the rocks which surround it. The sediment and water which fill the basin are lighter and

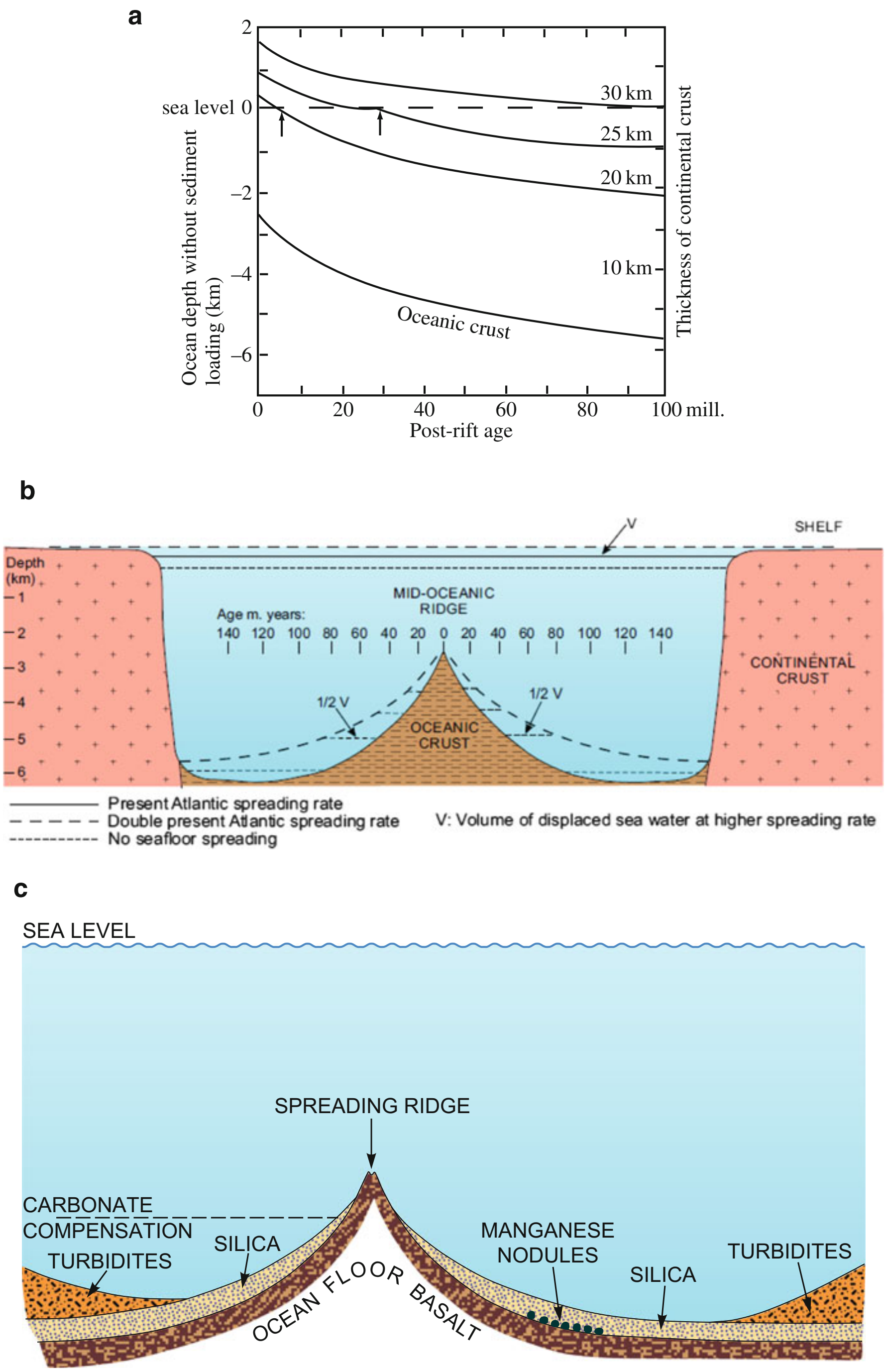

Fig. 8.7 (**a**) Subsidence of the sea cooling in the bed and continental crust of various thicknesses as a function of post-rifting age (after Kinsman 1975). Basaltic spreading ridges (oceanic crust) are often at 2 km depth, then the seafloor sinks to almost 6 km depth after 100 million years. (**b**) The rate of seafloor spreading control the average water depth which is a function of cooling of the oceanic crust. The basalt shrinks and becomes heavier with time. High spreading rates are therefore associated with transgressions. (**c**) Schematic presentation of an Atlantic type spreading ridge. The shallow part may be above the carbonate compensation depth while the deeper part will be without carbonate sediments and contain little clastic sediments, being far away from land. The sea floor with then be sediment-starved and concentrate silica from the volcanism and manganese nodules

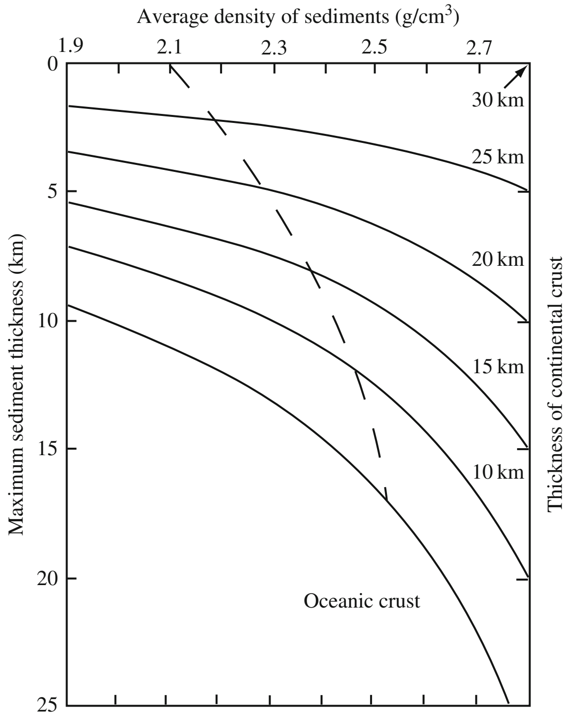

Fig. 8.8 Potential loading capacity to isostatic equilibrium on oceanic crust and of various thicknesses of continental crust as a function of sediment density (after Kinsman 1975). If the average sediment density is 2.5 g/cm^3, the maximum thickness of sediments can be 17 km on top of the oceanic crust. If the sediments have lower density (higher porosity) the sediment thickness must be lower to main isostatic balance with respect to the crust and the sediments down to about 100 km depth. If there is 20 km of continental crust and the sediment density is 2.3 g/cm^3, there is room for 4 km of sediments

have to be compensated for with heavier rocks below the basin. We can assume a depth of compensation of about 100 km. This means that the weight of a column of rock plus any water present down to a depth of 100 km must be the same everywhere. Subsidence of a sedimentary basin may be due to several crustal processes:

1. Stretching and thinning of the continental crust. Heavier mantle rocks then make up a greater percentage of the rock column down to a compensation depth of about 100 km.
2. Cooling, i.e. lower geothermal gradients in the crust, lead to contraction and a higher rock density (thermal contraction), and this will result in subsidence.
3. Increased loading by water, sediments or other rocks will cause subsidence. Water loading could be due to a transgression increasing the weight of the water column. Sediment loading occurs with basin infilling.
4. Subsidence along subduction zones. This results in lower geothermal gradients in the downward-deflected crust, so that the density also increases.
5. Tectonic loading. Thrusting of tectonic plates leads to increased loading on the part of the Earth's crust in question and we have subsidence, especially in front of nappes (foreland basins).

8.8.2 Changes in Sea Level and Sedimentation and Isostatic Compensation

Variations in sea level due to eustatic transgressions or tectonic subsidence will represent loading or unloading of the crust which will reinforce the primary change in sea level. If the sea rises 100 m, for example due to ice sheet melting, this will increase the isostatic loading on the seafloor. We can calculate that there will be a further 43.5 m of subsidence, so that the total increase in the depth of water at equilibrium will be 143.5 m.

If a sedimentary basin with this depth of water is filled with sediment, there will be further isostatic subsidence because of the weight of sediments, providing accommodation for deposition totalling 250–300 m depending on the density of the sediments. A 100 m rise in sea level will thus lead to deposition of almost 300 m of sediment. In the same way, primary tectonic subsidence due to cooling of the ocean floor will lead to further subsidence due to increased water and sediment loading.

Using stratigraphic data in the form of measured profiles or oil wells as our starting point, we can calculate backwards to the primary tectonic or eustatic changes in sea level. This method is called "backstripping".

$$Z = Y((\rho_m - \rho_s)/t(\rho_m - \rho_w))\Delta H\rho_w/(\rho_m - \rho_w) + (H - \Delta H)$$

where Z is the primary tectonic subsidence, F is a factor which is a function of the rigidity of the Earth's crust, ΔH is the change in sea level and H is the water depth (Watts 1983). Y is the sediment thickness compensated for compaction, i.e. the sediment thickness (and density ρ), shortly after deposition (Stickler and Watts 1978b, Bally et al. 1981). If we know the variations in sea level and the depth of the water from environmental interpretations, for example, these can be inserted into the equation so that the primary tectonic movement can be worked out.

The backstripping technique which is used to reconstruct the subsidence history of different parts of a sedimentary basin lends itself very well to computer modelling. The data obtained on the depth and temperature history of the source rock in particular has allowed much better assessments of kerogen maturity and the times of oil expulsion and migration. One parameter which is crucial but difficult to estimate is the variation of the geothermal gradient as a function of geological time. It is also often difficult to estimate the palaeodepth during the deposition of different sedimentary formations. Whether a formation was deposited at a depth of 200 or 1,000 m will make a very significant difference, and it is often difficult to make accurate palaeoenvironmental estimates of the palaeobathymetry.

To form a sedimentary basin with a thick infill of sediments requires a crustal depression large enough to provide the *accommodation space* for the deposits. This may be a result of large-scale crustal movements like seafloor spreading and crustal thinning (extension). The supply of sediments is also critical and clastic sediments have to be supplied from adjacent land areas which are being uplifted and eroded.

Chemical and biogenic sediments are formed locally from seawater and thus are not reliant on sediment supply from land areas. This typically applies to limestones, which accumulate where there is little clastic sedimentation. Sedimentary basins that are more or less isolated from the sea in arid regions can be filled up with evaporites at a rather high sedimentation rate.

8.9 Continental Rifting

Stretching and thinning of the continental crust takes place by tensional tectonics in connection with rifting. Crustal stretching may be a result of tensional forces and uplift due to the high geothermal gradients associated with rifting. This leads to a thinning of the continental crust, causing isostatic subsidence.

Fracture zones in the continental crust (rift valleys) produce subsidence which often is located in a sedimentary basin because the continental crust is thin and because heavy rocks from the mantle push their way up, thus increasing the average density of the rocks. Those parts of the rift valley system which have much volcanism will have less space for the sediments. Along the margins of a rift system, where the continental crust is not stretched, uplift occurs due to the higher geothermal gradient. This causes the basement rock at the surface to slope away from the rift valley as the geothermal gradients are reduced. This will to some extent be compensated for by the formation of erosional valleys which cut backwards into the raised shoulders on the sides of the rift valley.

Because of the high relief around these basins and the short transport distance for the sediment eroded from the bedrock, these basins will be characterised by mineralogically very immature sediments, largely arkoses and conglomerates deposited in fan deltas along the active faults. In the central and deeper parts of the basins we find deposition of finer-grained sandstones and clayey sediments. Rift valley basins may be continental lacustrine basins as in East Africa, or marine as those offshore East Africa and the Jurassic basins of the North Sea. Horsts, which are unstretched (thick) pieces of continent crust, may become topgraphically very high due to high heat flow. The Ruwenzori Mountains of East Africa, reaching more than 5,000 m, are one example. Both marine and lacustrine rift basins will tend to have reducing conditions in the deeper part due to limited circulation of oxygenated water.

Rift basins formed in areas with wet climates will be occupied by large lakes. Lacustrine basins often have an even better potential for producing source rocks than marine rift basins, because water stratification (density stratification) is usually more marked in lakes. We will therefore often find black, organic-rich shales in these basins.

Rift basins formed in arid zones are characterised by evaporite deposits. Block faulting will readily lead to isolated basins or horsts which cut off contact with the open sea. Evaporites are typical of rift deposits today, e.g. in East Africa, and were widespread in Europe and North America during the Permo-Triassic, before the ocean-floor spreading which created the Atlantic Ocean started in the Mid-Jurassic. The Zechstein salt deposits in Germany and the North Sea are typical examples. Jurassic and early Cretaceous rifting during the early phase of the opening of the South Atlantic Ocean, evaporite basins were located in the arid regions of the time. They can form almost perfect cap rocks that are not likely to leak. If they are thick enough they may form salt domes which produce structural traps in overlying rocks, as at Ekofisk and in the Gulf Coast basin. Rising salt domes also strongly influence the clastic sediment distribution in a basin.

Salt has high conductivity causing the temperatures above the salt to be higher than normal and the sediments below the salt to be cooler than normal. This must be taken into account when modelling the maturation of source rocks associated with the salt. These temperature anomalies may also be important when modelling diagenesis and reservoir properties. Large subsalt discoveries have been made in recent years both in the Gulf Coast and offshore Brazil.

The subsidence due to rifting renders the adjacent rocks unstable and promotes gravitational sliding of blocks in the crust, in towards the rift structure. There is a tendency for listric faults to form, i.e. parallel, curved fault planes which start as normal faults and curve round with depth until they are almost horizontal. The blocks then become rotated so that they tilt over and slope away from the rift. During the initial part of the spreading phase, basins with limited circulation will be formed so that evaporites and carbonates are often deposited. Upper Jurassic and Lower Cretaceous deposits of this sort are found extensively along the margins of the Atlantic Ocean.

8.10 Subsidence Along Passive Margins

Passive margins develop from a rift phase to a spreading phase. The transition between continental crust and oceanic crust therefore consists of a thinned continental crust with listric faults and horsts formed during the rifting phase (Fig. 8.9). As ocean-floor spreading progresses, the geothermal gradients in this part of the continental shelf will decline, resulting in cooling and thermal subsidence of the continental margins. The oceanic crust will also experience thermal subsidence, as a function of the age of the seafloor. Subsidence flexure will develop where the subsidence is most rapid, on the outer parts of the continental shelf and slope nearest the oceanic crust. Further in from the passive margin the subsidence rate will be slower. Eventually sedimentation fills the prism between the oceanic and continental crust. With cooling, the rigidity of the crust increases, so that the bending of the continental crust near the continental margin broadens and there is overall subsidence of the continental shelf, and in consequence transgression and onlap. It will subside isostatically and make accommodation space. During high sea level the clastic supply is pushed back onto the continent and during low stand a progradational sequence is formed (Fig. 8.10).

In other places the drainage discharge from land focuses the sediment in large delta areas. This is the situation with the Mississippi delta in the Mexican Gulf, which has been receiving sediment from large expanses of the North American continent since Mesozoic times.

The sediment supply is controlled by the drainage system on the continents, which may follow old rift systems because most of the sediment is produced on the continents and has plenty of space to be deposited along the margin of the deep ocean.

The Niger delta represents a similar focusing of sedimentation on the African side of the Atlantic. Sediment basins along passive margins are deposited above old rift basins which have cooled. If basin subsidence is rapid, this will also contribute to a low geothermal gradient because the sediments have to be heated during burial (20–30°C/km). A relatively great thickness of sediment (4–5 km) must therefore be deposited for the underlying source rocks to achieve maturation. That is to say, attain 100–150°C, depending on the subsidence rate and hence the time available for the heating. This kind of sedimentary basin situated along a passive margin is called a “cold basin”.

Foreland basins form in front of mountain chains when they are uplifted and eroded. The advancing nappes help to depress the continental crust and provide accommodation space for the sediments shed from the mountains. Foreland basins tend to be broad and gently folded into giant structures in the distal parts. The Middle East is a large foreland basin in front of the alpine mountain chain running along Iran and Turkey. The Persian Gulf is a continuation of this basin collecting sediments coming from the mountains of Iran. The Jurassic and Cretaceous sequences in Iraq, Saudi Arabia, Kuwait and The Emirates contain some of the largest oil fields in the world.

On the east side of the Rocky Mountains we have similar foreland basins from the Denver Basin all the

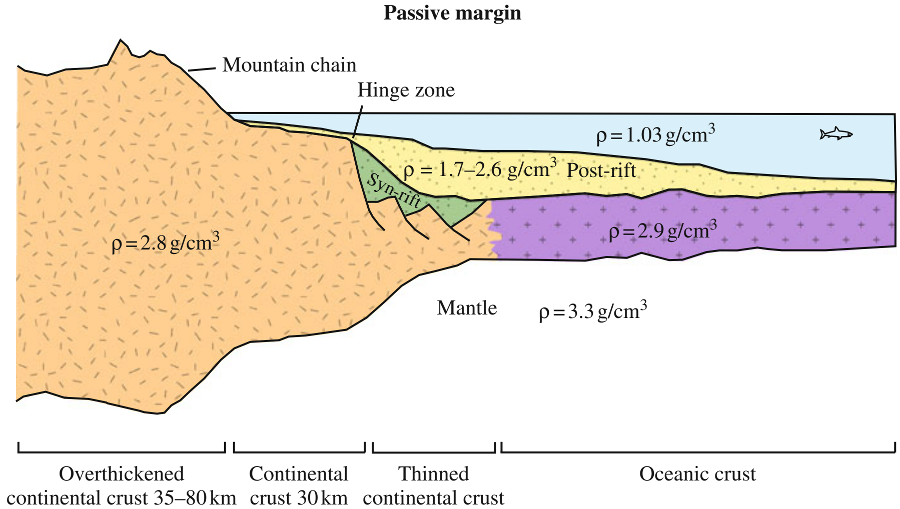

Fig. 8.9 Simplified cross-section of a passive continental margin. The sediments that were deposited during the initial phase of rifting lie beneath the younger sequence which was deposited along the passive margin. The thinner the crust is, the more sediments can accumulate and the maximum thickness is reached when the progradation reaches cold oceanic crust

way to Alberta, Canada, where there are very large deposits of heavy oil and tar sand. The heavy oil in Venezuela (and Columbia) is located in a similar tectonic position in front of the Andes.

8.11 Strike-Slip Faults and Pull-Apart Basins

We find "strike slip" faults in both the oceanic and the continental crust. Transform faults in the oceanic crust result in ridges of younger basaltic material which can help to limit the extent of sedimentary basins, particularly in the early phases of the opening (see Chap. 6).

Strike-slip faults in the continental crust can lead to both compression, i.e. thickening, forming small mountains, and stretching of the continental crust forming deep sedimentary basins. Calculations show that relatively modest stretching of the continental crust will cause considerable thinning and subsidence. Bends in the strike-slip fault system like the San Andreas Fault open up deep holes in the continental crust which can be filled in with very thick sequences of sediments like in the Ventura and Los Angeles Basins. The sedimentation rate may be very high because of the large uplifted land areas and the relatively small basins.

Here, too, relief which has developed through stretching of the continental crust will be made more pronounced by sediment loading and thermal subsidence. When two plates move parallel to one another, we have *strike-slip* faults of the San Andreas type in California, but there is no new plate formation, nor any subduction. We distinguish between "right-lateral" or "dextral" faults, where the opposite side of the fault has moved to the right, and "left-lateral" or "sinistral" faults. The San Andreas is a dextral fault, and the western part of California (Salina block) has moved northward in relation to the North American continent. If the fault plane follows a completely straight structure in the rock parallel with the direction of movement there will be neither tension nor compression along the fault plane. However, faults usually follow older structures in the basement rocks which may be curved and strike slip movements may then produce both compression and tension along the fault plane. Compression will lead to folding and the formation of small mountain ranges, while tension will lead to the opening of deep sedimentary basins.

We find the most typical examples of this in California. When a fault branches we may also see both compression and tension, with elevation and subsidence respectively of blocks, depending on their orientation. If the fault shifts to another parallel fault plane, we have crustal tension in the area in between, and often also basalt flows. A so-called "pull-apart" basin forms, which is a "hole" in the continental crust formed by the strike slip movement. The Salten

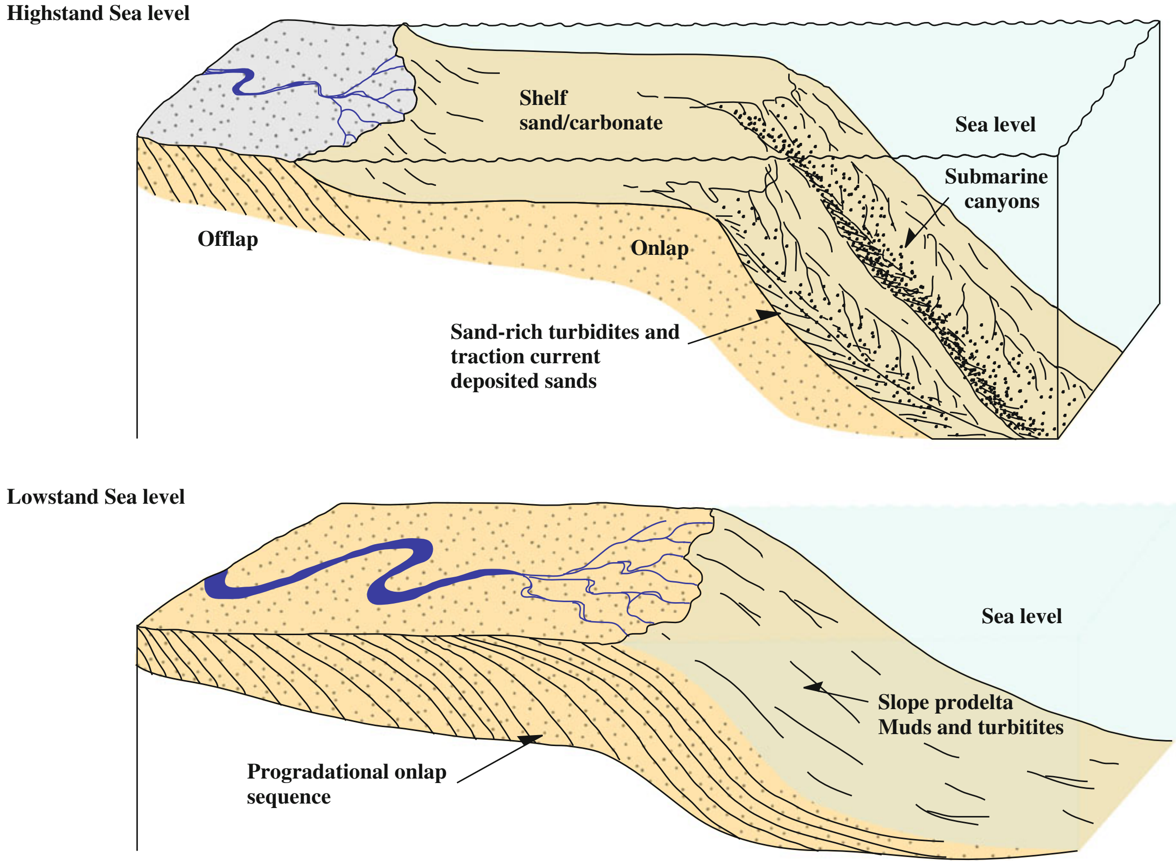

Fig. 8.10 A modal for transgression with sediment starvation and renewed progradation across the shelf onlapping (downlapping) onto an unconformity

Trough in Southern California is a good example and much of the Los Angeles and Ventura basins originated in this way during the Pliocene.

The basins are very deep holes in the continental crust, and are sometimes floored with volcanic rocks. Because of the limited size of the basins and the continued uplift of the areas surrounding them, we get a very high rate of sedimentation. In the Miocene-Pliocene basins in California, 5–8 km or more of sediments have been deposited in roughly as many million years. This corresponds to an average of 1 mm/year, which is a very high figure in this context. The sediments consist largely of proximal turbidites along the edge of the basin and distal turbidites and fine-grained clay sediments in the more central, deeper parts. In the Ventura Basin which continues into the Santa Barbara Channel, there are about 15 km of young sediments (<4–5 million years) which have been folded and which contain large amounts of oil.

Basins of this type provide almost ideal conditions for oil accumulation and are amongst the richest oil regions in the USA. Very few basins can compete with the California strike-slip related basins in terms of barrels of oil per square kilometre. This is because these deep basins often have reducing conditions in their bottom water due to limited circulation of oxidised water. We therefore get organic-rich sediments deposited. Turbidite sandstones, particularly the more proximal parts, often have sufficiently high porosity to become reservoir rocks. Even if the porosity is not especially high, turbidite deposits often form thick sequences that extend over large areas. The reservoir quality can also be controlled by turbidite lobes or channels with levees.

Sediments deposited in basins of this type tend to undergo folding, which will result in good structural traps. Geothermal gradients will be relatively high, and allow thorough maturation.

8.12 Converging Plate Boundaries

When plates collide (converge), there can be boundaries between:

- Oceanic crust and continental crust, such as along the west side of South America (Andes type).
- Continental crust against continental crust (Alpine or Himalayan type mountain chain).
- Oceanic crust against oceanic crust, with formation of volcanic islands.

The oceanic plate has a higher density than the continental crust, so in a subduction zone the oceanic plate will be carried downwards under the continental one. The oceanic plate is relatively cold and its geothermal gradient is low because since it is descending, the upward heat flux is reduced as a function of the rate of subduction. Where there is active subduction, ocean trenches are formed, such as the 10–11 km deep ones outside the Philippines and Japan. This is because the underlying rocks are cold and therefore dense.

In subduction zones the downward-moving part of the oceanic crust is characterised by extremely low geothermal gradients. This is because the heat flow must move counter to the direction of movement of the plate undergoing subduction which is therefore colder and denser than the average old ocean floor crust. The subduction zones form the deepest parts of the ocean (up to 10–11 km deep) where there is not a great deal of sedimentation. Ordinary, cold, 100–150 million year old ocean floor is in isostatic equilibrium at a depth of 5–7 km without sediment loading.

Where we have subduction of oceanic crust beneath continents we find the following types of basin:

- Trench basin (oceanic sedimentation within the trench).
- Accretionary prism.
- Fore-arc basin (relatively shallow basin in front of island arcs).
- Inter-arc basin (sedimentary basin between island arcs of continental crust).
- Back-arc basin (sedimentary basin on the oceanic crust).
- Retro-arc basin (formed if continental blocks are upthrust and faulted, creating a secondary basin in the continental crust).

Trench basins seldom contain large amounts of sediment if separated from the continent by an island arc preventing sediment supply from the continents. One of the reasons why they are the deepest oceanic areas is because they do not fill up with sediment. This is because the ocean floor beneath trench basins is part of a dynamic system and is being consumed through subduction as sedimentation occurs. The downward movement of the oceanic crust helps to maintain a low geothermal gradient in the trench since heat must flow up through rocks that are descending. Because of the continual lateral displacement of the rock floor under the trenches, the sediment thickness will be limited. The island arcs are the main source of sediment supply to the trench basins.

Most of the sediment coming from the continents will be trapped in back-arc basins. The sediments in trench basins consist of deep-water conglomerates, turbidites and pelagic sediments. Trench basins lie mostly below the carbonate compensation depth and therefore contain little carbonate and are also usually starved with respect to clastic sediment supply. Sedimentary transport of very fine-grained pelagic sediments tends to occur along the axis of the trench. The inner slope up to the island arcs is steeper than the outer one, and the subduction zone (Benioff zone) is at the foot of this slope (Fig. 2.48).

When the oceanic plate descends into the Benioff zone, sediments deposited on the sea bed may be scraped off, forming wedges of sediment. This is called an *accretionary prism* and consists of a complicated pattern of folded and brecciated sedimentary rocks exhibiting overthrusts and underthrusts (Fig. 2.49). The sedimentary sequence is the right way up within each thrusted unit, but the sequence of thrusted units is inverted. The youngest sediment wedges are at the bottom because they are the last sediments to have been scraped up from the seafloor.

On the inside of the trench, slope basins are formed with sediment supplied by the island arcs. These deposits are folded, deformed and reworked by the subduction processes.

Although sediments in trench basins may be rich in organic material, their oil potential is poor due to a lack of reservoir rocks and the intense tectonic deformation. In addition the geothermal gradient will also be lower than in other basins, at least at the time of deposition.

Fore-arc basins form between the actual island arcs with volcanoes, and the subduction trench. They are not subjected to the intense tectonic deformation which occurs in trench basins. Fore-arc basins often transgress onto the island arcs as the belt of volcanic activity gradually withdraws towards the continent. These basins will contain fluvial and deltaic sediments closest to the island arcs, then shallow marine continental shelf sediments. In the outer parts sediments are deposited in deeper water towards the oceanic slope. There is good potential here for organic-rich sediments to accumulate, but the geothermal gradient may be low despite the proximity to a volcanic island chain, and maturation consequently slow.

Large parts of California in front of the Rocky Mountains (Sierra Nevada) are a fore-arc basin (San Joaquin Basin) and there are numerous oil fields here, not least in the area around Bakersfield.

Intra-arc basins are a result of tensional tectonics in the island arcs within an area otherwise characterised by convergent plate movement. Here a system of fault-controlled basins (graben) is formed. Sediments deposited in these basins will for the most part be immature, and volcanic material will be common. Volcanic sandstones will often lose their primary porosity rapidly due to unstable minerals and diagenetic transformation. The geothermal gradient is high, but conditions for formation of source and reservoir rocks are not very good.

Intra-arc basin sediments will frequently be subjected to intense tectonic deformation which may deform potential reservoirs.

Back-arc basins develop on the oceanic crust due to seafloor spreading behind the island arcs. There is likely to be an abundant supply of sediment from the continent, and back-arc basins may also fill with deltaic and shallow marine sediments. Because of later tectonic deformation, back-arc sedimentary basins are not the most promising oil prospects.

The Jurassic and Cretaceous seaway through North America east of the Rocky Mountains was a back-arc basin during subduction of the Pacific plate. The area is a good oil region but Tertiary uplift resulted in a very strong meteoric water drive which flushed out many of the oil fields.

8.13 Conclusions

Stratigraphy is a response to changes (movements) in the Earth's crust and sediment supply. Changes in the environments and the biological production of sediments are also important. The accommodation space for sediments is also controlled by the rate of compaction of the underlying sediments.

Geophysical methods, mainly seismics, are used to map out sedimentary basins on a regional scale and also on a small scale for detailed exploration and production. It is also important to reconstruct the evolution of sedimentary basins though time and the changes with respect to the sediment supply and the environment.

This must to a large extent be based on reconstruction of plate tectonic movements and basin modelling.

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