CHAPTER

15

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THE FORMATION OF SEDIMENTARY BASINS

Introduction

Sedimentary basins are a characteristic feature of the Earth's crust and lithosphere and range in age from Archaean to the present day. While sedimentary basins have long been of interest to the exploration geologist, it is only in the last decade or so that their mode of formation has received much attention from Earth scientists in academia.

At the present day, the world's largest basins are found at the margins of the continents near the mouths of river systems. For example, up to 15 km of sediments have accumulated at the continental margin off the Amazon, Niger and Mississippi rivers. Other margins with large sediment thicknesses occur off the Ganges river in the Indian Ocean and the Colorado river in the Pacific Ocean. The largest sediment thicknesses in the geological record are also believed to have occurred at the margins of the continents. For example, several kilometres of sediments are known to have accumulated at the margins that flanked the Tethys Ocean prior to the collision of the African and Eurasian plates and formation of the Betic-Alps mountain ranges during the Tertiary.

Sedimentary basins are dominated during their evolution by *vertical movements*. They are therefore difficult to explain in terms of plate tectonics, since this theory emphasises the horizontal motions of the plates. On a large scale, an undeformed basin has a generally synclinal form that reflects the tectonic movements that modified it while it was being infilled by sediment. Basins may form close to plate boundaries, or in intra-plate settings. The continental margin basins described above are related to extension of the lithosphere that often leads to continental break-up and ocean opening. On the smaller scales of geological field mapping, the internal reflector geometry of a basin reflects not only tectonics, but the modifying effects of erosion, compaction, sea-level changes and sediment supply.

The past few years has seen a rapid increase in our understanding of sedimentary basin formation due in large part to the acquisition of high-resolution multichannel seismic reflection and refraction profile data and advances in our understanding of the thermal and mechanical properties of the Earth's lithospheric plates. In addition, sedimentary basins are the world's largest repository of oil and gas deposits (Figure 15.1) and so have been subject to intense study by the petroleum industry.

The different types of basins

If we define basins as depressions of the Earth's crust that contain more than one kilometre of sediment, then approximately 70 per cent of the Earth's surface is underlain by basins of one type or another. Such a definition excludes basins whose sedimentary infill is now incorporated in fold belts, but includes those in the stable continental interiors and flanking regions that have escaped the destructive effects of plate subduction and rifting.

Sedimentary basins are found in a variety of tectonic settings. Some of the world's largest basins occur within or on the stable continental interiors, such as the central USA (e.g. Michigan/Illinois), the Russian Platform, and Congo (e.g. Zaire) and are termed **intracratonic basins**. They are also associated with the flanks of orogenic belts as **foreland basins**, such as the Appalachians (e.g. Allegheny, eastern USA), Rockies (e.g. Denver, western USA), Alps (e.g. Molasse, Switzerland), Betic (e.g. Guadalquivir, Spain) and Pyrenees (e.g. Ebro, Spain) ranges. Other types of basins are associated with **passive continental margins** (e.g. Atlantic-type), **back-arc** (e.g. western Pacific) and **fore-arc** (e.g. Cook Inlet, Alaska) regions, and **strike-**

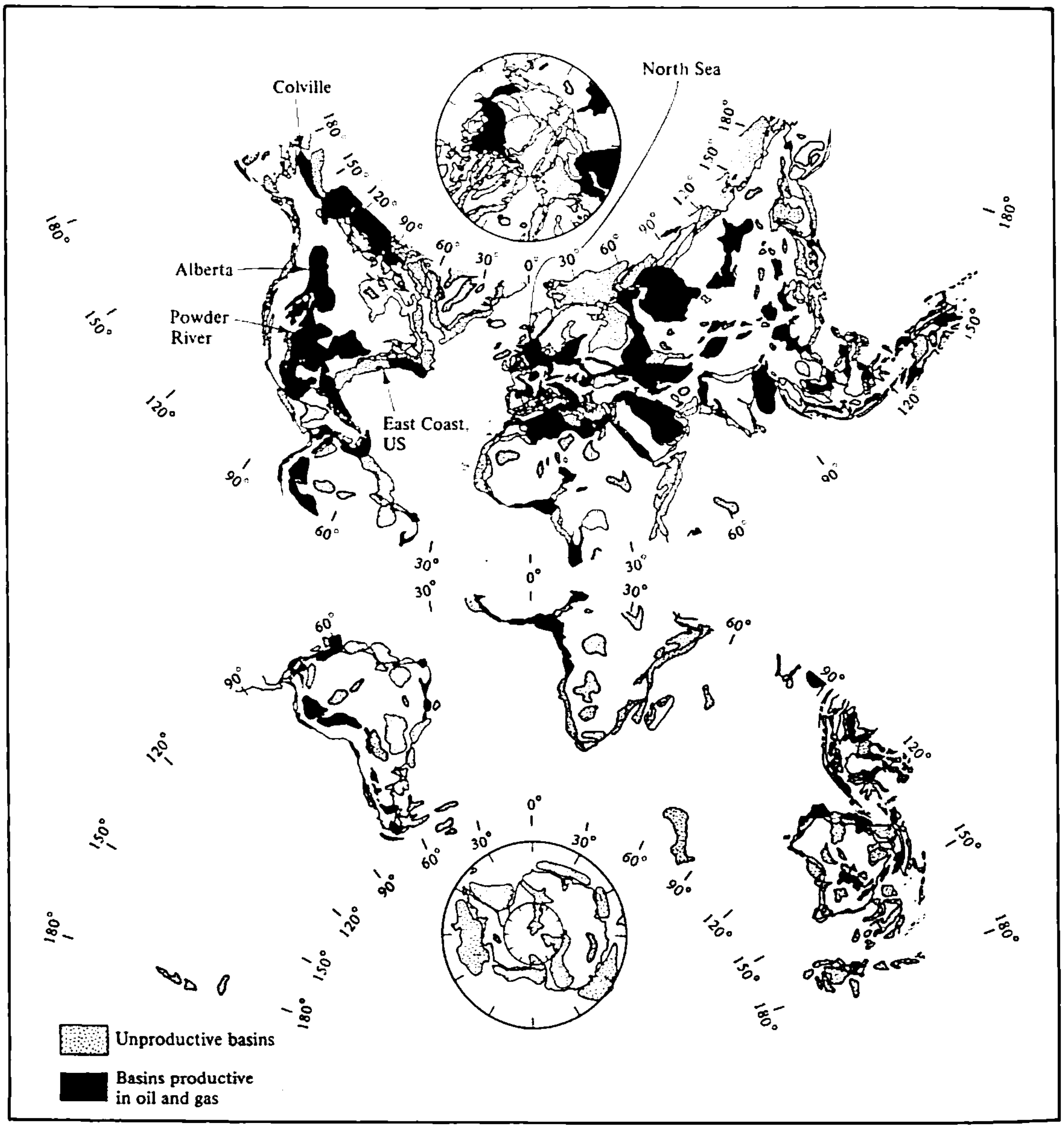

Figure 15.1 The global distribution of sedimentary basins. Heavily shaded regions correspond to basins that are productive in oil and gas. The basins referred to in the text are labelled.

slip basins associated with faults showing lateral movement (e.g. Ventura, California).

Even though basins show considerable tectonic diversity, they have some common features that depend on their setting. For example, most passive continental margin and intra-plate basins are associated with rifting (i.e. extension of the crust along normal faults) of rigid continental lithosphere. Foreland basins are associated with compressional plate boundaries. By way of contrast, smaller basins of the fore-arc, back-arc and strike-slip type develop in response to an extensional, compressional or strike-slip stress field along a plate collision zone.

In some cases, a basin may change tectonic setting during its evolution. For example, it is quite common in the geological record to find a foreland basin located on or near a previous passive margin (e.g. Colville Trough of the North Slope, Alaska). During convergence, the steep plunge in basement at the edge of passive margin basins serves to localise thrusts and may cause them to preferentially 'telescope' onto the more gently dipping flanking coastal plain. Other examples of rift-type basins within compressional-type orogenic belts occur (e.g. Alboran Sea between the Betic and Rif mountains of southern Spain and north Morocco). The extension may, in this case, be the result of collapse following thickening of the crust and lithosphere during mountain building.

The deep structure of sedimentary basins

The shallow stratigraphy of many basins is now quite well known, but the underlying crystalline basement is often too deep to be penetrated by drilling or to be imaged by seismic reflection profiling techniques. Moreover, there have only been a few seismic refraction studies carried out to determine the velocity structure and depth to the Moho (see Chapter 3) beneath basins.

One example where both stratigraphic and deep seismic data are available is the North Sea basin (Figure 15.1). The Palaeozoic to Recent sediments that fill this basin reach maximum thicknesses of up to 9 km. There is good evidence in the basin of rifting in the form of graben formation, volcanism and flank uplift. Seismic refraction data indicate that the depth to the Moho (i.e. the seismic definition of the base of the crust) is about 18 to 25 km beneath the basin, and about 25 to 35 km beneath adjacent areas of Britain and Norway. The Moho therefore shallows locally beneath the North Sea basin.

While a shallow Moho depth is a feature of many basins, some notable exceptions exist. For example, deep seismic reflection profiling studies show that the Fastnet and Celtic basins in the continental shelf off Britain and France have a flat underlying Moho (see Chapter 13). Other examples appear to exist on deep seismic profiles of the Canadian margin.

An interesting feature of the deep structure of many basins is the presence of sub-horizontal reflectors in the lower part of the crust. The origin of these reflectors is not clear, but they have been attributed to either mechanical transformations of the crust during orogeny, or to some form of ductile flow in the crust during rifting. The deep crustal reflectors are quite common throughout Europe, but have only been clearly seen beneath the Basin and Range province in the USA.

Mechanisms of subsidence and uplift

A sedimentary basin will not form unless there is an initial depression for the sediments to fill in. Rift-type, compressional-type and strike-slip basins are often characterised by thick sequences of continental and shallow-water sediments and therefore require substantial tectonic driving forces in order to explain them.

Perhaps the best known of the basin-forming mechanisms is thermal contraction of the oceanic lithosphere as it cools away from a mid-ocean ridge crest. At a ridge crest, hot mantle material is accreted to the lithosphere which, as it cools, increases density and subsides (see Chapter 9). Observations of sea-floor depth show that the subsidence is exponential in form (Figure 15.2), increasing from depths of about 2500 m at the ridge crest (i.e. zero age) to greater than 6000 m for sea floor older than 150 million years (Ma).

At some continental margins, a considerable thickness of continental-derived material has built out onto the oceanic lithosphere. In the central Atlantic, for example, turbidity deposits have caused the Mesozoic age oceanic crust to be bent down at least 1 km more than would be expected on the basis of its age. Locally, at the mouth of the Mississippi and Niger rivers, continental-derived sediments have depressed the oceanic crust by as much as 4 to 5 km below expected depths.

As a rule, sediment thicknesses much in excess of 2 km rarely accumulate on the oceanic crust. The greatest thicknesses occur instead on the continental crust – at or near the present-day shelf break in slope. Off the East Coast US margin, for example, seismic reflection profiles reveal up to 10 to 15 km of Mesozoic–Tertiary sediments that overlie a crystalline basement of Palaeozoic or greater age.

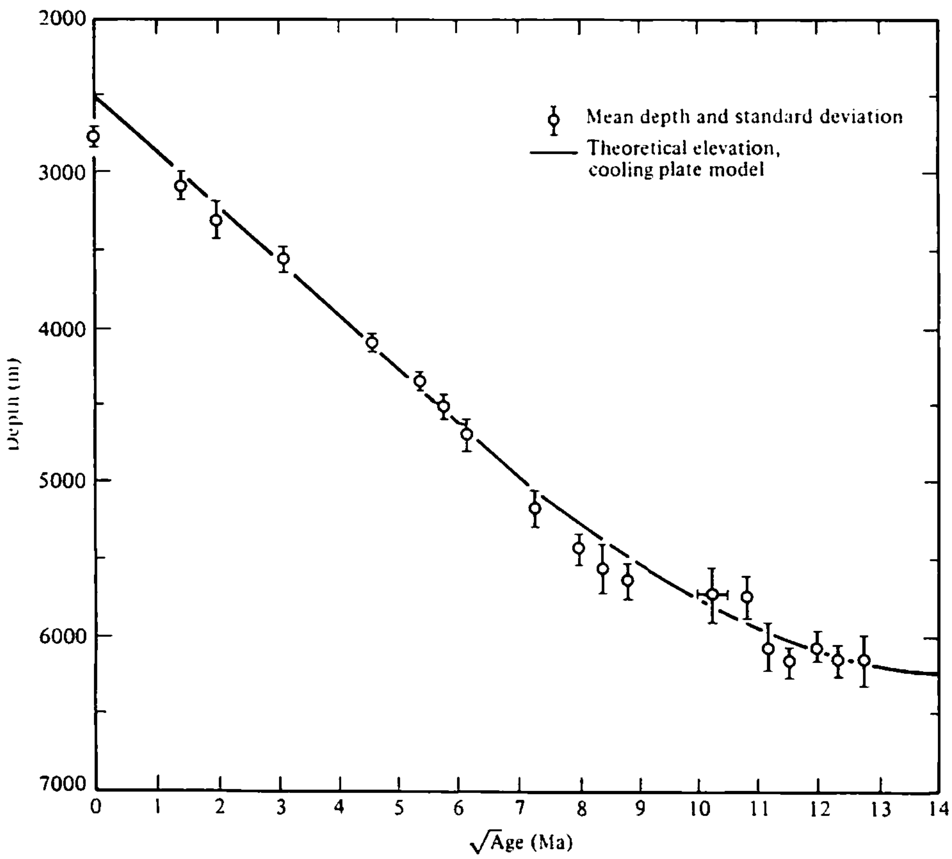

Figure 15.2 Subsidence of the North Pacific Ocean basin based on observations of the sea-floor depth and theoretical models based on a cooling plate that show an exponential decline in subsidence rates. □

In the late 1970s, Dan McKenzie suggested that during rifting continental lithosphere is stretched, cools and subsides with age in a manner similar to that of the oceanic lithosphere. According to this model, **rift basins** are characterised by a relatively rapid initial subsidence, due to crustal and lithospheric thinning, followed by a slow thermal subsidence, as the crust and the lithosphere cool. This behaviour is commonly referred to as the **McKenzie model of basin formation.**

A **stretching model** is supported by the observation that rift structures are generally associated with thinner than usual crust. Off northwest Spain, for example, seismic reflection profiling (Figure. 15.3) and deep drilling data reveal a series of tilted fault blocks in the upper part of the continental crust. The crust in this region is 20 km thinner than it is beneath the adjacent shelf, and the geometry of the thinning suggests uniform extension. A gentle undulating reflector separates the faulted upper crust from a highly reflective lower crust below, and this reflector has been interpreted by some workers as a **'detachment'**, which during extension separated a region of brittle deformation above from one of pervasive ductile flow below.

The rift structures illustrated in Figure 15.3 are unusual in that they have only a thin sediment cover, despite their location in water depths of up to 4000 m. In most basins, the rift structures are obscured by substantial thicknesses of post-rift sediments. Since sediments displace water they represent a load on the crust and lithosphere which should sag under their weight. The extent of sediment loading depends, however, on the water depth that is available for sedimentation, the density of the sediments and mantle materials, and the strength of the crust and lithosphere.

Let us assume that due to tectonic or some other means a water depth, W_d, is made available for sedimentation. According to the **principle of isostasy** there will be some depth in the Earth, known as the **depth of compensation,** where the pressure (or force/unit area) on a loaded and unloaded column are equal. The pressure on an unloaded column is given by

$$W_d \rho_w g + T_c \rho_c g + r \rho_m g, \qquad (1)$$

where W_d = water depth of deposition, T_c = mean thickness of the crust, r = distance from the base of the crust to the depth of compensation, ρ_w = density of water, ρ_c = density of crust, ρ_m = density of mantle, g = average gravity, and

$$r = d - T_c - W_d,$$

where d = depth of compensation. The pressure on the base of a loaded column is given by

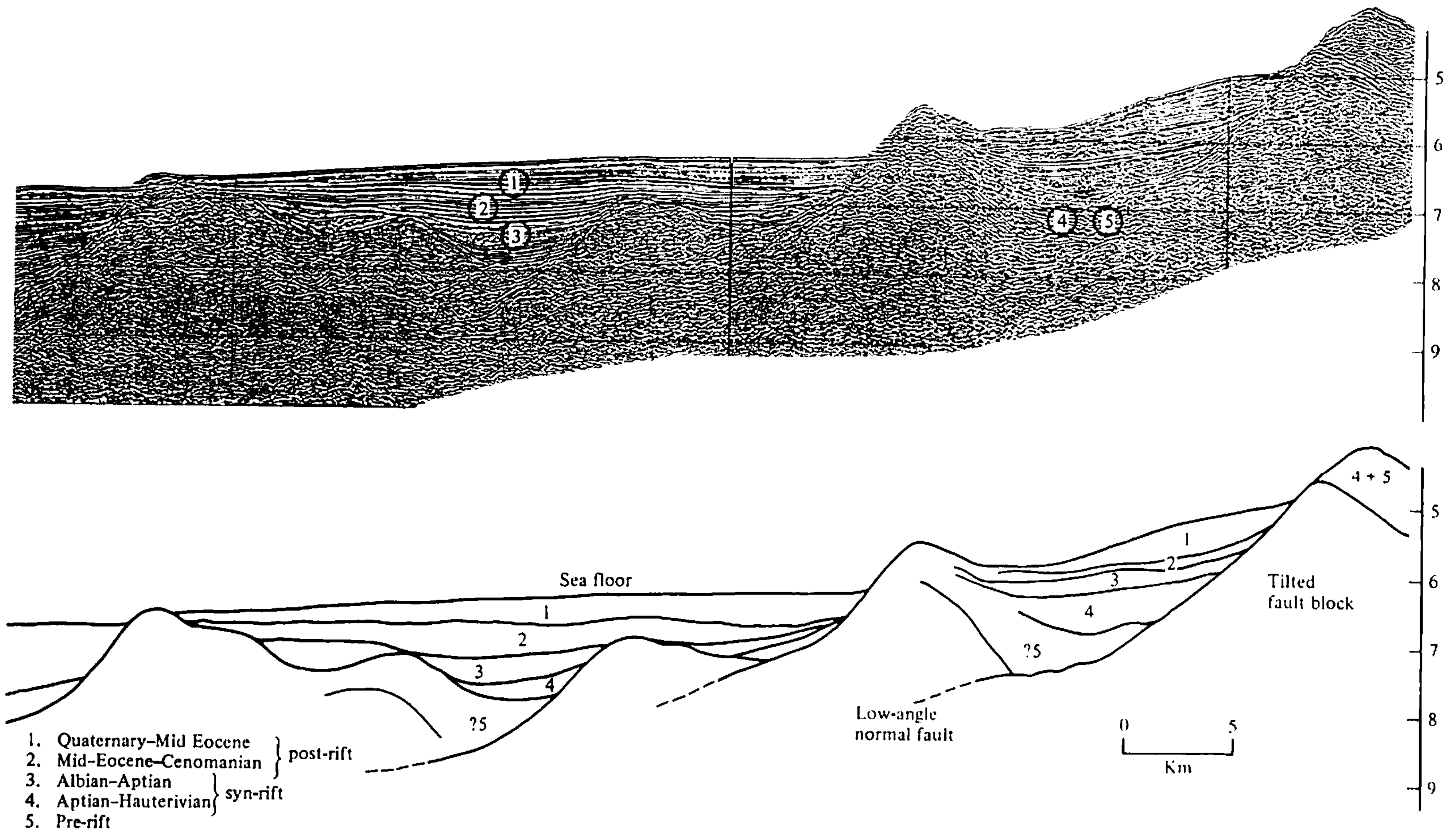

Figure 15.3 Seismic reflection profile of the Galicia Bank in the continental margin off northwest Spain. The profile, which was obtained by the Institut Français du Petrole (Paris, France), shows the tilted fault block structure associated with the initial rifting between Europe and North America. □

$$\rho_s Sg + T_c \rho_c g, \tag{2}$$

where S = thickness of sediments and ρ_s = density of sediment. If the depth of compensation is at the base of the loaded crust then

$$d = S + T_c$$

or

$$r = S - W_d. \tag{3}$$

Combining equations 1–3 we then get (neglecting effects of sea-level and water depth changes):

$$S = W_d \frac{(\rho_m - \rho_w)}{(\rho_m - \rho_s)}. \tag{4}$$

With ρ_s = 2500 kg m^{-3}, ρ_m = 3300 kg m^{-3} and ρ_w = 1030 kg m^{-3}, equation (4) shows that only a thickness of up to about 2.5 times the water depth can form as a result of sediment loading.

In the case of a land-locked basin, the water is eventually replaced by sediment and no further deposition can occur. The situation is different in basins where there is a limited supply of sediment such that the basin is unable to fill completely. At a continental margin, for example, sediments may be able to build out, but they can rarely fill a growing ocean basin.

Let us consider the case of a 'hypothetical' margin in which sediment loads progressively build up and out from some initial shelf break in slope (Figure 15.4a). Oceanic flexure studies reveal that when the crust and lithosphere are loaded they respond by bending over a broad region. The pattern of deformation is similar to that which would be expected for an elastic plate overlying a weak fluid substratum (see Box 15.1 for a mathematical treatment of how an elastic plate deforms when loaded). Figure 15.4b shows that the flexure due to a wedge-shaped sediment load is characterised by a broad depression, a thinning of the stratigraphic unit in a landward and seaward direction, and a flanking region of uplift or bulge (often termed a **flexural bulge**).

As a margin builds out, the stratigraphy due to an initial load and its flexure is modified by subsequent loading. For example, a previous load and flexure that is located in the flexural depression of a new load will experience subsidence. Conversely, a previous load and flexure located on the bulge will be subject to uplift. The cumulative stratal geometry (Figure 15.4c,d) shows **onlap** (see Figure 20.4a, p. 394) in the direction of load migration and a basinward shift in the pattern of onlap, or **offlap**. Furthermore, each depositional surface is flexed to a sigmoidal shape. The calculated stratal patterns are similar to structures observed in river delta systems, especially those which are referred to as clino-forms, (see Figure 20.4b and c, p. 394) observed on seismic reflection profiles.

The calculated stratigraphy in Figure 15.4 is based on a model which assumes that during sediment loading there is little or no stress relaxation of the lithosphere. Oceanic flexure studies, however, show that the elastic thickness of the lithosphere (which is determined by the flexural rigidity) is some two to three times thinner than the seismic thickness, suggesting that on loading there must be some form of relaxation of the mechanically supportive part of the lithosphere from its short-term (? seismic) thickness to its long-term elastic thickness. Since oceanic island loads take at least about 1 to 2 Ma to form, an elastic model is probably a good approximation for the response of the lithosphere to sedimentary packages that are a few million years or longer in duration. Unfortunately, it is not currently known whether continental lithosphere behaves in a similar way to the oceanic lithosphere or, whether, because of compositional differences, it takes more or less time to complete its relaxation.

Irrespective of the relaxation time, Figure 15.4 shows that sediment loading is capable of forming basins of considerable thickness. The infilling of a water-filled depression 5 km deep (the maximum that could be produced by a thermal cooling model) by sediment would produce a basin up to 12.5 km thick. The sediment loading effect is large enough that it complicates the use of stratigraphic data to isolate the effects of those processes, such as thermal contraction, that are believed to be responsible for basin subsidence. The **'backstripping'** technique is one approach that has been adopted to attempt to correct the stratigraphic data for the effects of sediment loading, and is described in Box 15.2.

Backstripping of sediments from deep commercial wells shows that the post-rift subsidence of many basins is characterised by a simple exponential decrease – in accord with the predictions of the McKenzie model. At some basins, the agreement is remarkably close suggesting that thermal contraction may be a major contributor to basement tectonics. Others show irregularities on the smooth exponential decrease due to superimposed tectonic events. Off Sable Island, Nova Scotia, for example, an abrupt step in the tectonic subsidence at about 140 Ma has been attributed to piercement structures caused by the mobilisation of an underlying salt layer.

At some wells, the McKenzie model is unable to explain the tectonic subsidence deduced by backstripping. For example, wells in the East Coast US coastal plain show uplift, rather than subsidence, during the first few tens of million years of the post-

BOX 15.1

Loading of thin elastic plates

A useful way to calculate the flexure of the crust and lithosphere due to arbitrary shaped sediment loads is based on the response function technique. Consider, for example, the case of a load, $h(x)$, on the Earth's surface of the form:

$$h(x) = h \cos(kx)\, g\, (\rho_s - \rho_w), \qquad \text{(A.1)}$$

where h = the maximum height of the load, g = average gravity, ρ_s = density of the sediment, ρ_w = density of water, k = wavenumber ($2\pi/\lambda$, where λ is the wavelength), and x = horizontal distance. The flexure of the crust and lithosphere, $z(x)$, is obtained by solving the fourth-order differential equation for the bending of a thin elastic plate overlying a weak fluid substratum and is given by:

$$z(x) = \frac{(\rho_s - \rho_w)\, h\, g \cos(kx)}{((\rho_m - \rho_s)\, g + Dk^4)}, \qquad \text{(A.2)}$$

where ρ_m = density of the mantle and D is the flexural rigidity of the plate which is determined from the elastic thickness, T_e by:

$$D = \frac{E T_e^3}{12(1 - \sigma^2)}, \qquad \text{(A.3)}$$

where E = Young's modulus and σ = Poisson's ratio. We see from equation (A.2) that when $k \to 0$ (i.e. $\lambda \to \infty$) then

$$z(x) \to \frac{(\rho_s - \rho_w)\, h}{(\rho_m - \rho_s)},$$

which if $\rho_s = \rho_c$ where ρ_c = density of the crust gives:

$$z(x) \to \frac{(\rho_c - \rho_w)\, h}{(\rho_m - \rho_c)}.$$

This is the well known 'Airy root' response to loading. If, on the other hand, $k \to \infty$ (i.e. $\lambda \to 0$) then

$$z(x) \to 0,$$

and there is no deflection. This can be considered as the Bouguer response to loading. We see from equation (A.2) that the crust and lithosphere is behaving as a sort of filter in the way that it responds to sediment loads. The term filter is used in the usual sense to denote a system with an input and output. In the application discussed here, the input is the load $h(x)$ and the output is the flexural response $z(x)$. The filter characteristics can be determined by defining a certain wavenumber parameter given by:

$$\phi_e = \frac{\text{input}}{\text{output}}.$$

For convenience, we will consider the Airy-type response as input and the flexural response as output. We are then able to write:

$$\text{Input} = \frac{(\rho_s - \rho_w)\, g\, h \cos(kx)}{(\rho_m - \rho_s)\, g},$$

$$\text{Output} = \frac{(\rho_s - \rho_w)\, g\, h \cos(kx)}{[(\rho_m - \rho_s)\, g + Dk^4]},$$

and

$$\phi_e = \left[\frac{D k^4}{(\rho_m - \rho_s)\, g} + 1\right]^{-1}. \qquad \text{(A.4)}$$

This is the function that modifies the input (i.e. the Airy-type response) to produce the output (i.e. the flexure). Equation (A.4) shows that for long-wavelength loads, the crust and lithosphere do not modify the Airy-type response, and these loads are supported mainly by buoyancy of the substratum. For short-wavelength loads the crust and lithosphere greatly modify the Airy-type response. These loads are mainly supported by the rigidity of the plate.

The response function approach outlined here can be used to rapidly compute the flexure due to any shape of sediment load. Let us replace the load $h \cos(kx)$ by a function $\delta(x)$ such that $\delta(x) = 0$ when $x \neq 0$ and

$$\delta(x)\, \partial x = 1, \qquad x = 0.$$

Now, if $H(k)$ and $Z(k)$ are the discrete Fourier transforms of $h(x)$ and $z(x)$ respectively, then equation (A.2) reduces to

$$Z(k) = \phi_e(k)\, H(k)\, \frac{(\rho_s - \rho_w)}{(\rho_m - \rho_s)}.$$

To calculate the flexure, $z(x)$, due to an arbitrary shaped sediment load we simply take its transform, multiply by the wavenumber parameter, ϕ_e, and a density factor, and inverse transform the result.

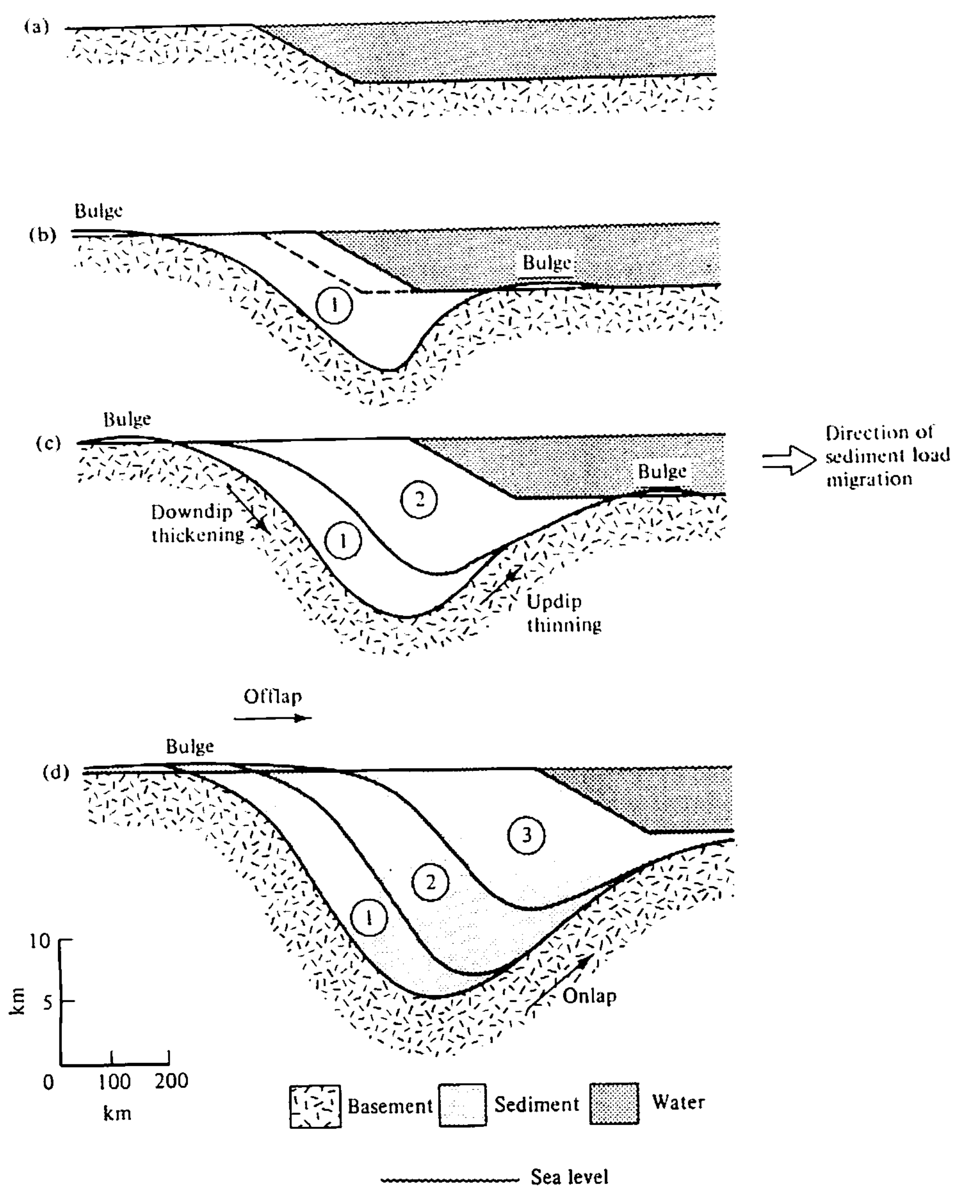

Figure 15.4 Simple model for the effects of sedimentary loading at a passive continental margin. (a) The initial state of margin with a slope and abyssal sea-floor depths of 5 km; (b) flexure due to a single wedge-shaped load; (c) the flexure after two loads have been applied; (d) the margin configuration after three loads. □

rift period. This uplift cannot be explained by a thinner than usual crust at the time of rifting (an initial crustal thickness of 18 km or less will cause uplift in the McKenzie model) since the edge of the margin eventually did subside forming a gently dipping coastal plain on to which Late Cretaceous to Recent sediments were deposited.

One way to explain the well data is by a **'two-layer' model** where the crust and mantle are extended by different amounts: the uplift being the result of mantle thinning beneath a region of little or no crustal thinning. The total amount of extension in the crust and mantle must be the same, of course, otherwise there would be a space problem. McKenzie and co-workers used simple Gaussian curves for the distribution of the extension in the crust and mantle. A model (e.g. Figure 15.5) in which the mantle extension occurs over a broader region than the crustal extension but the total amount of mantle and crustal extension are the same explains the absence of the Jurassic beneath

BOX 15.2

Backstripping

Let us consider a so-called 'loaded' column that represents the sedimentary unit that accumulated during a certain interval of geological time and an 'unloaded' column that represents the position of the underlying 'basement' without the effects of the sediments (see figure). The pressure at the base of a loaded column is given by:

$$W_d \rho_w g + S^* \rho_s g + T_c \rho_c g, \tag{B.1}$$

where W_d = water depth of deposition, T_c = mean thickness of the crust, S^* = the sediment thickness corrected for compaction (see discussion in later section), g = average gravity and ρ_w, ρ_s and ρ_c are the densities of the water, sediment and crust respectively. The pressure at the base of the unloaded column is given by:

$$Y\rho_w g + T_c \rho_c g + r\rho_m g, \tag{B.2}$$

where Y is the corrected or tectonic subsidence, ρ_m is the density of the mantle, and r (see figure) is the distance from the base of the unloaded crust to the depth of compensation (assumed to be at the base of the loaded crust) and is given by:

$$r = S^* + W_d - \Delta_{SL} - Y, \tag{B.3}$$

where Δ_{SL} = sea-level change as it would be viewed on the continent or 'freeboard'.

Combining equations (B.1, B.2 and B.3) we then get:

$$Y = S^* \frac{(\rho_m - \rho_s)}{(\rho_m - \rho_w)} + W_d - \Delta_{SL} \frac{\rho_m}{(\rho_m - \rho_w)}. \tag{B.4}$$

Equation (B.4) corrects the observed stratigraphic record for the effects of sediment and water loading and changes in water depth – a technique that has come to be known as '**backstripping**'.

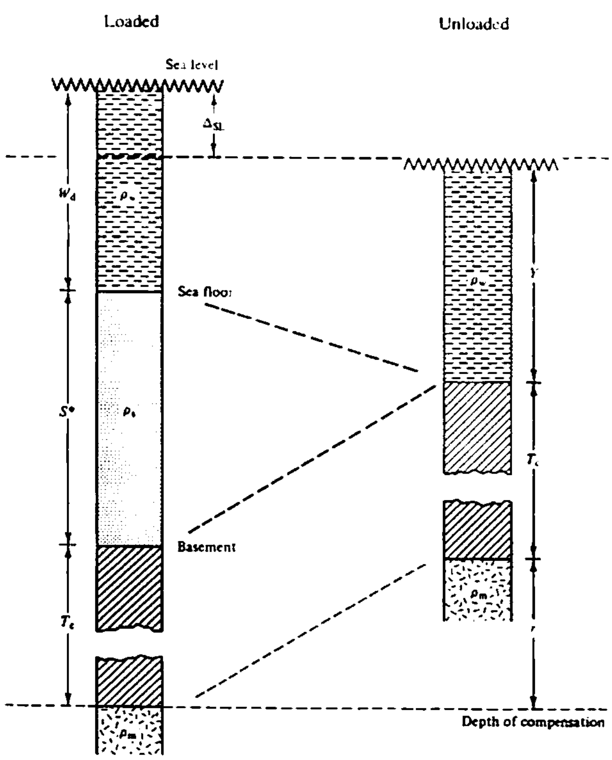

Figure B.1 Schematic diagram illustrating the backstripping technique.

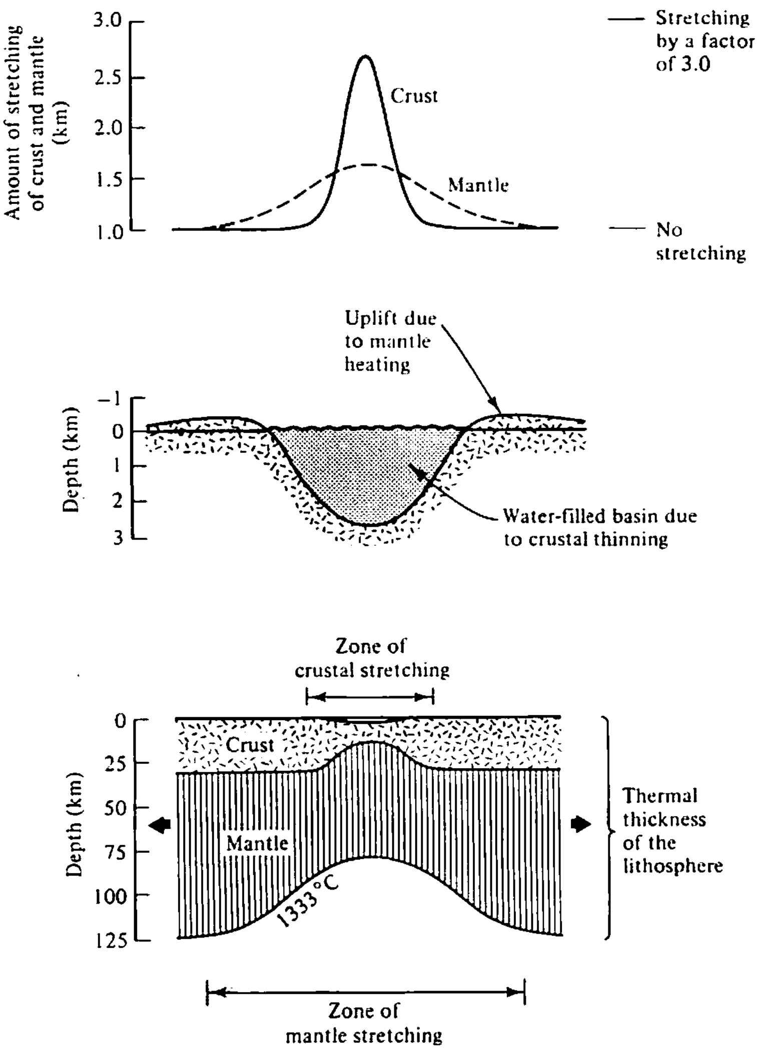

Figure 15.5 The initial water-filled subsidence for a rift-type basin formed by stretching of the crust and lithosphere. The thermal bulges are the result of a transfer of heat from the hot stretched region to the unstretched crust and lithosphere. The model assumes that the stretching is distributed unevenly with depth but that the total amount of extension in the crust and lithosphere is equal. □

the East Coast US coastal plain, even though the rift/drift transition (i.e. the change from rifting to sea-floor spreading) at this margin is believed to be of Late Triassic/Early Jurassic age.

The distribution of extension with depth will influence the shape of a water-filled basin (Figure 15.6). If the extension is accomplished by a simple necking of the crust and lithosphere (i.e. **pure shear**) then the maximum crustal and mantle extension will coincide and there will be an overall symmetry to the pattern of basin uplift and subsidence. If, on the other hand, the extension is by **simple shear** (as some recent geologic studies in the Basin and Range province, western USA suggest) then the crustal and mantle extension may be offset such that the syn- and pre-rift are displaced from the post-rift sediments and there is an overall asymmetry to the basin shape.

At present, it is not known whether the extension in rift-type basins occurs by pure or simple shear or by some combination of these mechanisms. A critical test is seismic reflection and refraction profile data since these have the potential to image deep layers in the crust as well as determine its detailed velocity structure. During the past decade several countries (e.g. USA, UK, France and Canada) have developed programmes to apply seismic techniques to the study of the deep structure of the continental crust (see Chapter 13).

A compilation of a deep seismic LITHOPROBE (Canada) line across the Canadian margin with a BIRPS and ECORS (UK, France) profile of the conjugate margin (i.e. one that was opposite to another prior to continental separation) off southwest Britain is shown in Figure 15.7. This profile shows a highly layered lower continental crust and several strong reflectors that may represent detachment surfaces within the crust. The most significant result of the seismic profiles is that they reveal that, while the

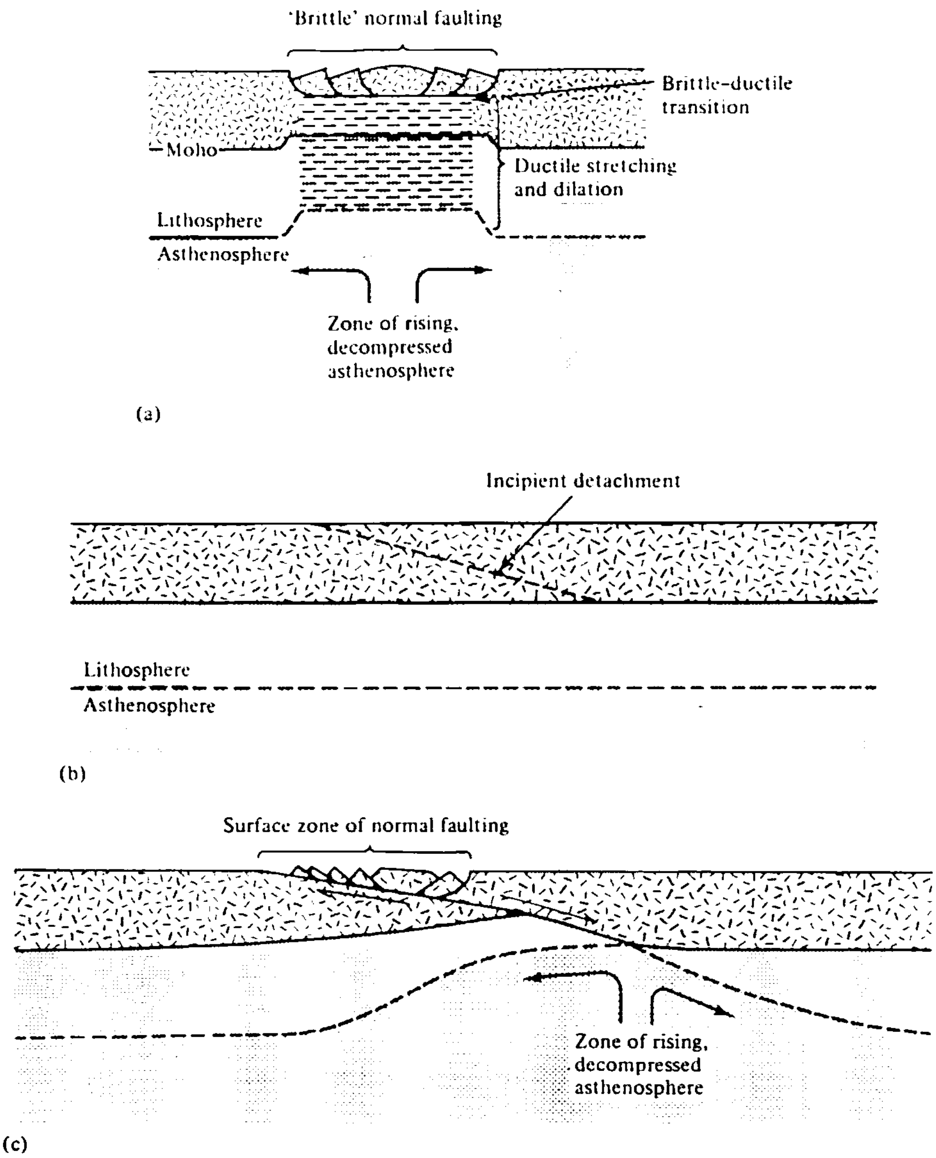

Figure 15.6 Illustration of the two end-member models for lithospheric extension and basin formation. (a) Shows the pure shear model; deformation in both brittle and ductile layers is uniform. (b) and (c) are successive stages in the simple shear model, where brittle fault blocks are carried along a shear plane that propagates through the entire lithosphere; deformation in both brittle and ductile layers is asymmetrical. □

geometry of faulting in the upper crust is asymmetric, the lower crust is more symmetric providing general support to the stretching model.

A number of basins fail to show any evidence of the exponential decrease of subsidence that is so characteristic of extensional basins formed by rifting and that is predicted by the McKenzie model. Foremost among these are the foreland basins that form in compressional settings in front of advancing thrust/fold belts. Although exceptions exist, many of these types of basins are characterised by an initially slow subsidence which progressively *increases* with time.

It is generally agreed that foreland basins are mechanically, rather than thermally, driven. Stratigraphic studies based on deep wells, seismic reflection profiles and gravity modelling show that basement beneath the Colville (Alaska), Powder River (western USA) and Alberta (western Canada) basins, for example, dips gently toward the mountain belt. Furthermore, these basins are bounded on their landward

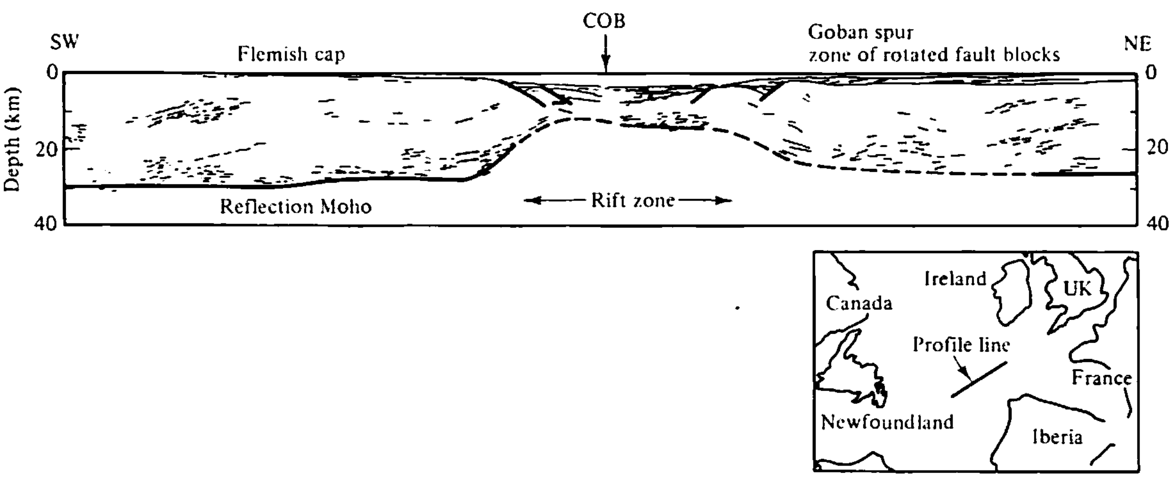

Figure 15.7 A compilation of deep seismic reflection profile data obtained over the eastern margin of the Canadian shelf and across the conjugate margin off southwest Britain. The compilation shows that, whereas the fault pattern in the upper crust is asymmetric, the thinning of the lower crust is symmetric, suggesting stretching occurred by pure shear (COB: continent–ocean boundary). □

side by a broad flexural arch which, in some cases, plays an important role in the supply of sediment to the trough. One of the most well-studied foreland basins is the Colville Trough in the North Slope, Alaska (Figure 15.8). The trough is bounded to the south by the Brooks Ranges and to the north by the Barrow Arch, which is one of the world's largest petroleum plays. Figure 15.8 shows that the basement underlying the Colville Trough has a flexural form and previous studies have shown that it can be explained by an elastic plate model which is loaded at one end.

If mountain belts comprise a series of thrust/fold loads then a flexure model predicts that as each load advances, the flexural depression and arch migrates away from the mountain belt. The tectonic movement at a particular locality in a basin would depend, however, on its location on the flexure profile. For example, a locality that is near the flexural node (i.e. a point of no subsidence or uplift) will not experience subsidence until the flexural depression of a new load reaches it. The tectonic subsidence, as obtained say by backstripping, would be expected to be initially slow and then progressively increase.

A difficulty with the modelling of foreland basins is that the sources of the loads responsible for the tectonic subsidence are not well known. In the case of the Himalaya, gravity anomaly studies suggest that the present-day topography is sufficiently large to account for the subsidence of the Australian–Indian plate beneath the Ganges basin. However, the Himalaya consists of a 'stack' of individual thrust/fold loads so that if we are to predict the stratigraphy of the flanking foreland, the geometry of each load needs to be reconstructed first. Unfortunately, this requires estimates of the amount of shortening, the height of the series of thrusts, and the extent of erosion, and these are difficult to access. Gravity anomaly studies of the Alps and Appalachians suggest, on the other hand, that the present-day topography is insufficient to cause flexure of the Swiss Molasse and Appalachian foreland. These mountain belts require 'buried' loads within the crust. Again, buried loads pose a difficulty in stratigraphic modelling since their distribution with time is not known.

The observation in Precambrian shield areas of foreland basins *without* their accompanying mountain belt raises the possibility that buried rather than surface loads are the primary mechanism by which forelands are preserved in the geological record. A mountain belt is subject to erosion and, if it constitutes the main load, eventually the adjacent foreland basin would be removed as the mountain belt and its flank respond to the denudation by uplift. Buried loads, however, are 'protected' from erosion and could remain in the crust for long periods of geological time. Currently, there is considerable controversy regarding the origin of these loads. In the Alps and Appalachians, buried loads are associated with large-amplitude gravity anomalies and, in some cases, outcrops of ultra-basic rocks. They could be wedges of oceanic crust (and mantle?) material that were thrust into the thickened continental crust during convergence. Alternately, they may represent areas of formerly thin crust, formed in passive margin settings, the mantle 'anti-

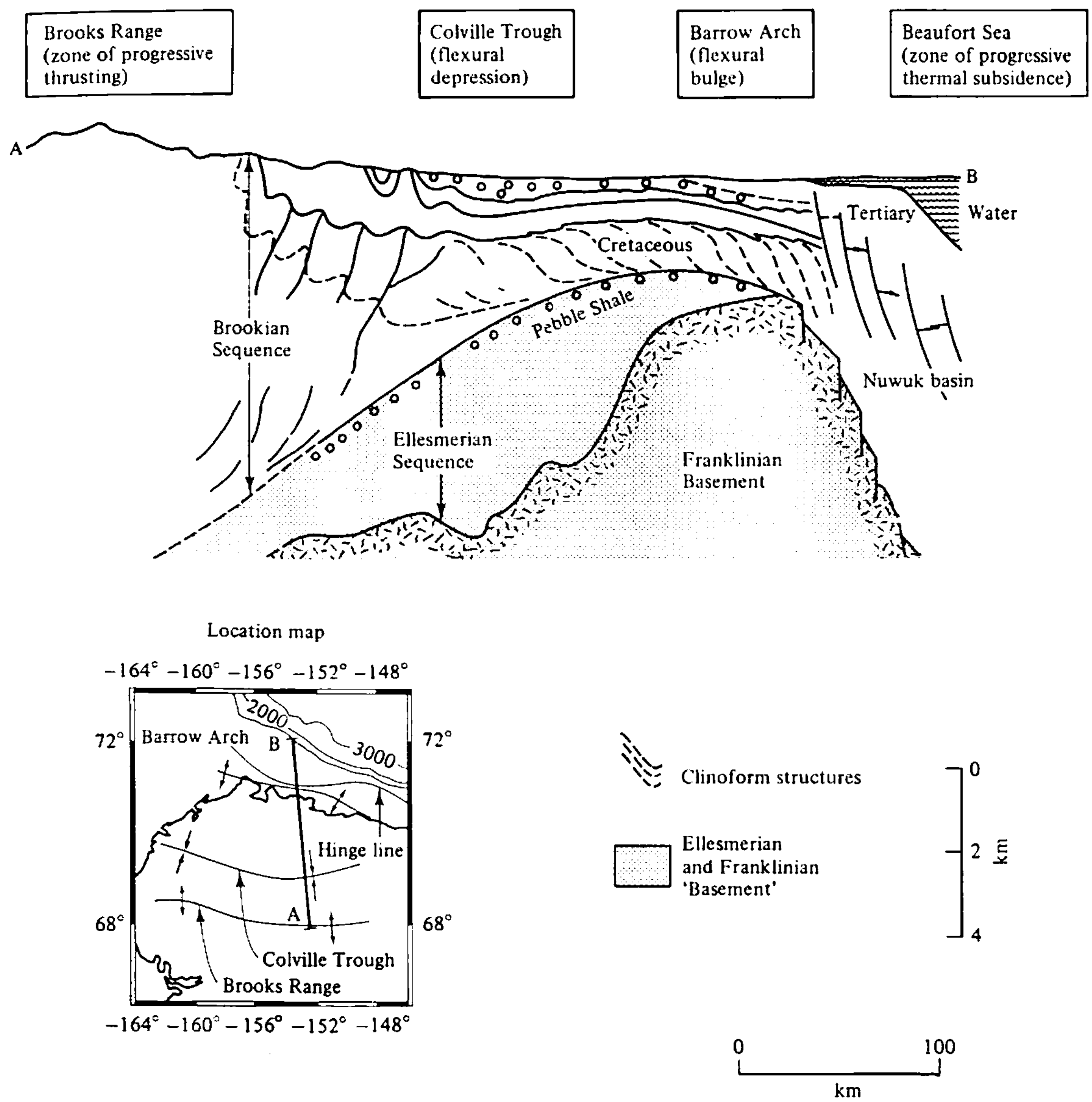

Figure 15.8 Schematic cross-section of the Brooks Range – Colville Trough – Barrow Arch system. The Pebble Shale marks the base of the fore-deep sequence and was probably flat prior to loading in the mountain belt. □

roots' of which later become loads on the base of the crust.

The mechanical response of the crust (and lithosphere) to loads also appears to play a role in the development of the small basins that develop in regions of localised extension and compression. These basins are distinguished from their rift-type and foreland counterparts by their narrowness, rapid subsidence, variable facies, and; in the case of marine basins, the inability of sedimentation rates to keep pace with subsidence. The stratigraphy of these basins suggests that they alternately experience compression and extension on very short time scales due to rotations of adjacent crustal blocks.

One of the best examples of the mechanically driven small basins are the **'pull-apart' basins** (e.g. Ventura, California) that form between overlapping strike-slip faults such as occur along the San Andreas fault system. Although, a McKenzie-type stretching model has been applied to explain subsidence in the Ventura basin, other basins in the fault system (e.g. Santa Barbara) show marked interruptions in their subsidence due apparently to compressional events and their associated flexural effects. Other examples of small basins occur in compressional and extensional settings. In compressional settings it is quite common to find 'piggy-back' basins perched above the main thrust front (e.g. Rocky Mountains). These basins occupy the sites of large pre-existing inter-montane depressions that probably formed in response to compression. Often these basins are characterised by sub-aerial, lacustrine or fluviatile sediments. The major control on the subsidence of these basins appears to be sediment loading. In extensional settings, small basins are found within rift systems, especially on the hanging walls of normal faults (i.e. the side of the fault that moved relatively downwards). Again, sediment loading is thought to be a major cause of subsidence in these basins.

Basin stratigraphy

The previous discussion suggests that it should be possible to construct synthetic stratigraphic cross-sections of basins simply by combining sediment loading with mechanisms such as thermal contraction and thrust/fold emplacement. This is difficult to do in practice because (a) the calculation of sediment load depends on a knowledge of the water depth through time and (b) there are certain feedback effects between sediment loading and thermal contraction and thrust/fold loading which need to be taken into account.

The problem is illustrated schematically in Figure 15.9. Let us assume a basin margin characterised by a wedge of sediment that formed in some depth of water. Within the wedge there will be a surface which will be the depth the underlying 'basement' would be without the sediment. This surface is the backstripped 'basement' depth, referred to here as the base. The concept of base is important since it allows the sediment thickness to be partitioned into two parts: a region above which defines the load and a region below which defines the flexure. Since we now have models for the movement of base (e.g. by thermal contraction) and the flexural response of the crust and lithosphere, stratigraphic modelling reduces to estimating the sediment load as increments through time. Unfortunately, estimation of the load depends also on the water depth of deposition and, for most basins, this is a poorly known parameter.

The water depth is a product of the combined processes of erosion and sedimentation, each of which will vary spatially and temporally. Coastal studies show that shorelines are continually being modified in such a way that they tend to acquire just the right slope to ensure that the incoming supplies of sediment can be removed at about the same rate as they are received. An adjusted profile of this type is referred to by geologists as an **'equilibrium profile'**. Unfortunately, because of the complexities of tides, storms, surges and climate variations it is unlikely that a long-lived approach to equilibrium can ever be achieved. The concept is a useful one, however, and most basin models now assume that sediments infill to either a horizontal, pre-deformation surface or to some constant sloping palaeo-bathymetric surface.

A further potential difficulty in stratigraphic modelling is that the mechanical and thermal properties of the crust and lithosphere are linked in such a way that the flexural strength depends on their thermal structure. For example, in a rift-type basin the lithosphere

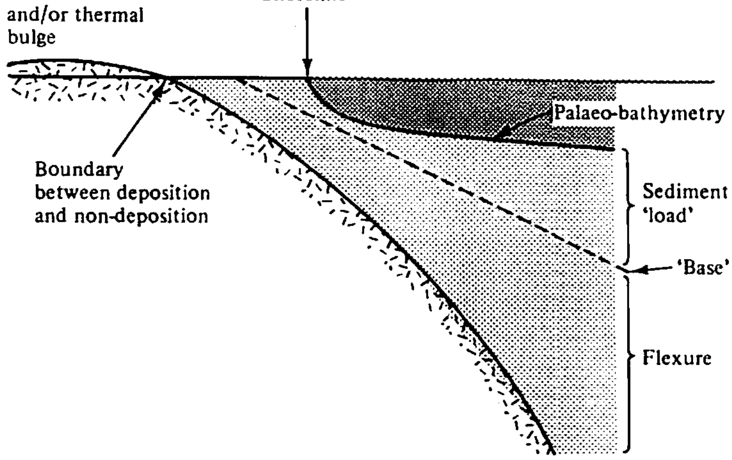

Figure 15.9 Schematic illustration of the 'beach problem' in basin modelling. For explanation, see text.

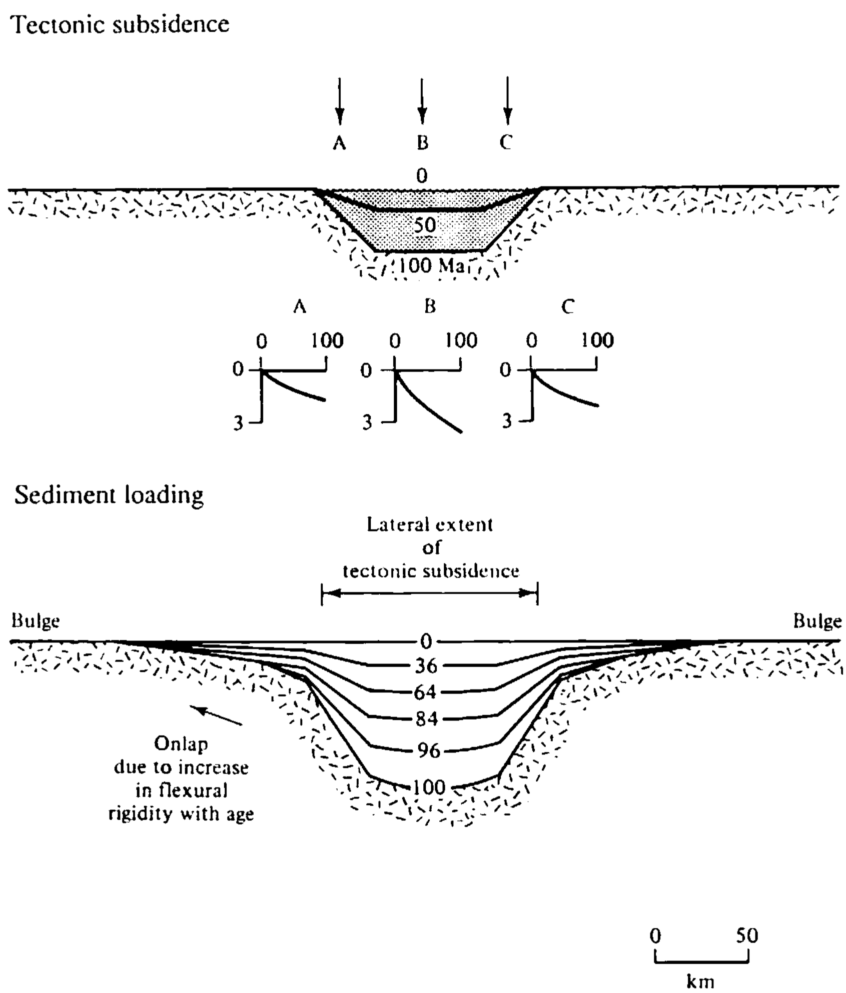

Figure 15.10 Simple model for the stratigraphic patterns that develop in a thermally subsiding basin formed by extension at the time of rifting. The onlap at the basin margin is produced by an assumption that as the lithosphere cools following rifting it becomes stronger, thereby spreading out the deformation due to individual sediment loads. □

is likely to be hot and weak early in basin evolution while later on it is likely to be cold and strong. This change in strength during basin evolution could greatly affect the basin architecture (i.e. the stratal patterns developed within the basin).

Oceanic flexure studies have shown that as the lithosphere cools, its flexural rigidity (which is a measure of its resistance to deformation) increases. The elastic thickness of the lithosphere, T_e, is determined by the flexural rigidity and is given approximately by:

$$T_e = Z_{450\,°C}, \qquad (5)$$

where Z is the depth to the 450 °C oceanic isotherm. If the flexural rigidity is also coupled to the temperature as it is in the case of rift-type basins, then we would expect that T_e would increase from small (zero?) values for sediments deposited at the time of rifting to about 30 km for sediments formed 100 Ma after the basin-forming event.

Figure 15.10 shows a model for the stratigraphy of a rift-type basin based on the concept of an equilibrium profile and a T_e which increases with age as the crust and lithosphere cool. The tectonic subsidence is assumed to be exponential in form, and to increase from the basin edge to its centre. The predicted stratigraphy shows a progressive overstep, or onlap, of the basement at the basin edge. This onlap is the result of our assumption that the rigidity of the underlying lithosphere, like that of oceanic lithosphere, increases with age. The result after a number of stratigraphic units have been deposited is to produce a flexural basin with a characteristic sag shape, which is referred to by some as the 'steers head'.

A number of rift-type basins show onlap at their edges especially during the thermal subsidence phase of basin development. Figure 15.11 shows an example which developed following the rifting and separation of Africa and North America during the Early Jurassic.

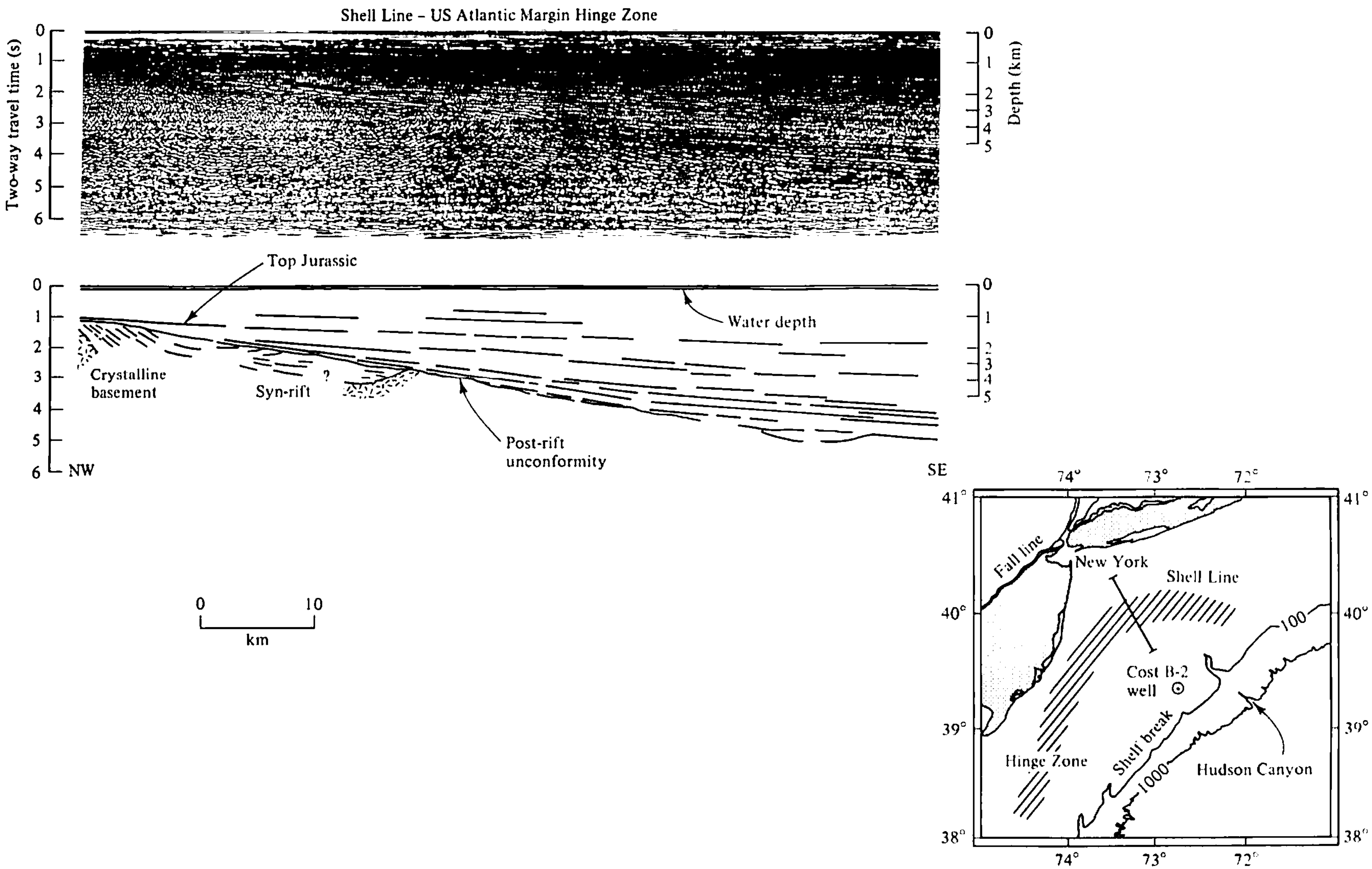

Figure 15.11 Seismic reflection profile obtained by the Shell Oil Company across a portion of the hinge zone of the Atlantic-type continental margin off New York. □

A flexure model with a T_e that increases with age predicts that a basin should widen with time and that young sediments progressively overstep basement rocks at the basin edge. The fact that some basins (e.g. Michigan, central USA) show the youngest sediments in the centre is therefore problematical. Initially, it was thought that the young sediments were the result of some form of viscous relaxation in the lithosphere. Inspection of Figure 15.10 shows, however, that an elastic plate model with a T_e that increases with age could explain the occurrence of young sediments at the basin centre, provided that its edges were subject at some time during their history to a widespread erosional event.

As pointed out earlier, the assumption that T_e depends on age is based mainly on oceanic flexure observations and not on any direct evidence for the behaviour of the continental lithosphere. Some researchers have recently suggested that the continental lithosphere may not follow such a simple T_e versus age dependence and that in extension it may be intrinsically weak. If this is the case can onlap patterns, such as observed in Figure 15.11, still be explained by a tectonic mechanism?

Several workers have pointed out that an Airy-type model (i.e. $T_e = 0$, see Box 15.1) could also explain the post-rift onlap, provided it is combined with a thermal model in which extension varies with depth. The thermal model required is one in which the basin flanks are characterised by mantle heating with little or no crustal stretching. The mantle heating causes the flanks to rise while the lack of crustal stretching prevents subsidence. Eventually, the mantle cools, the basin flanks subside below sea level and sediments progressively overstep the basement.

Foreland basins apparently do not have a thermally induced driving component, although both the sediments (and basement) may heat up as subsidence proceeds. The movement of base (Figure 15.9) in this case is produced by flexure due either to thrust and fold loads emplaced on the surface of the crust or to 'buried loads' within the adjacent orogenic belt.

Figure 15.12 shows the stratigraphy that would be expected for a foreland basin formed by four separate thrust/fold loading events. Each load was emplaced on the plate assuming that sediment infilled the basin to a constant horizontal pre-deformation surface. As each load advances, the sedimentary infill progressively oversteps older units and the flexural bulge migrates outward of the thrust/fold front. The result

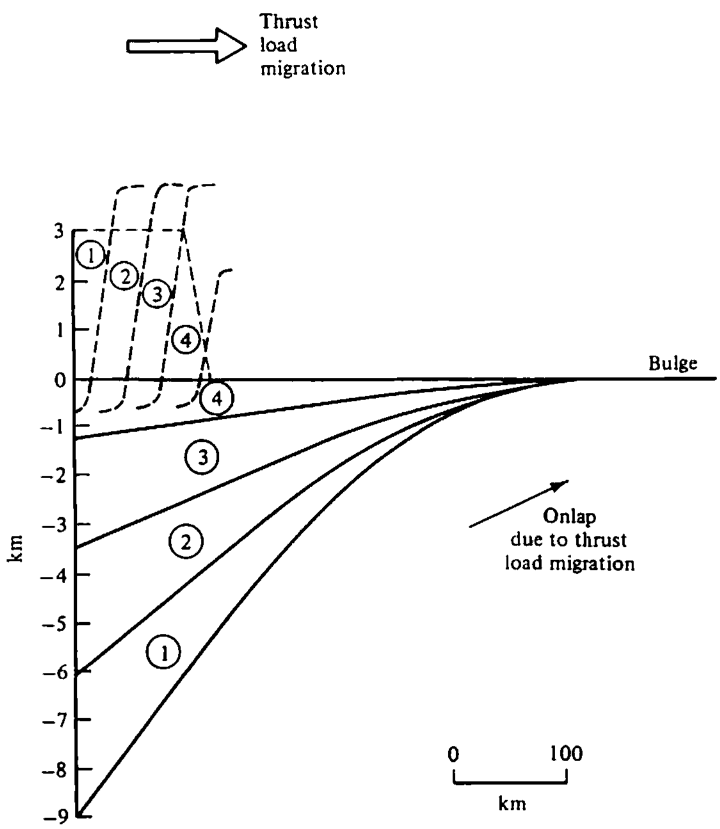

Figure 15.12 **Simple model for the stratigraphy of a foreland basin formed by flexure due to successive advancing thrust/fold loads.**

after four loads is to produce a flexural sag basin that is dominated by onlap at its edge. Onlap is a feature of all models: the widest pattern of onlap corresponding to large values and the narrowest pattern to small values of the flexural rigidity.

Foreland basins range in width from 20 to 300 km, suggesting that the rigidity of the underlying basement ranges from weak (e.g. Appenine, Ebro, Guadalquivir) to strong (e.g. Allegheny, Ganges). The cause of such large strength variations in the continental lithosphere is not clear, but it may be due to the different tectonic settings of these basins. For example, the Appenine, Ebro and Guadalquivir basins are located on lithosphere that was quite close to the site of passive margin development at the borders of the Tethys Ocean, while the Allegheny and Ganges basins developed on the cold, rigid, interiors of the continents.

Modulation of the stratigraphic record

The basin models discussed so far are based on only the simplest of thermal and mechanical concepts. They ignore the disturbing effects of other sedimentary factors such as erosion, variations in the supply of sediment, compaction and changes in sea level. While it is believed these factors do not contribute significantly to the overall geometry of basins, they nevertheless may interact with the primary controls in such a way as to exert a major control on the smaller-scale packaging of stratigraphic units.

Erosion

Basins formed by thermal contraction and flexure are associated with peripheral bulges which may be subject to erosion. Thermal bulges are a feature of the initial phases of rift-type basin evolution and can be produced by either the lateral flow of heat from the hot stretched region to the relatively cold unstretched continental lithosphere, or a region of mantle heating that extends over a wider region than the crustal stretching. Flexural bulges, on the other hand, are a consequence of the mechanical loading of an elastic plate overlying a weak fluid substratum and they can develop in strike-slip, extensional and compressional settings. The removal of material by erosion has two effects: one is to unload the crust and cause uplift and the other is to thin it and cause subsidence. At the edge of a basin, the interaction of these effects will have important consequences for the position of the shoreline.

Let us consider a hypothetical mountain range that initially is in isostatic equilibrium such that its elevation, h, is balanced by a deep crustal root, r. Erosion disturbs this balance and an uplift, w, results. The pressure at the base of a column in the mountain range is given by:

$$(h + r + T_c)\rho_c g,$$

where ρ_c = density of the crust, T_c = mean thickness of the crust and g = average gravity. The pressure at the base of the eroded column is:

$$(T_c + r)\rho_c g + \rho_m wg,$$

where ρ_m = density of the mantle. Assuming the columns before and after erosion are in isostatic equilibrium we then get

$$w = h\,\frac{\rho_c}{\rho_m}. \qquad (6)$$

Using typical values of ρ_m = 3330 kg m^{-3} and ρ_c = 2700 kg m^{-3} we see that an elevated region can be reduced to about 70 per cent of its height by erosion.

If the erosion is a prolonged event then this uplift will in turn be subject to erosion. Equilibrium is achieved when the uplift is reduced to zero which occurs as soon as the total amount of material that is removed, S_c, reaches $h + r$. The Airy root r is given by

$$r = \frac{h\rho_c}{(\rho_m - \rho_c)},$$

so the total amount that can be removed is given by

$$S_c = h\left[1 + \frac{\rho_c}{(\rho_m - \rho_c)}\right]. \qquad (7)$$

Using the same values of ρ_m and ρ_c as before, equation (7) shows that up to about 5.5 times the initial topography could be removed by erosion.

The estimates of the amount of erosion discussed so far are maximum likely ones since they ignore that the uplift may be limited by the flexural rigidity of the plates. Moreover, erosion may be inhibited by other factors such as lithology, vegetation and climate. Unfortunately the dependence of erosion rate on these variables is not well understood.

Studies of the catchment areas of large river systems have revealed that the rate of erosion, due to the combined effects of lithology, vegetation and climate, is approximately proportional to the average elevation. We can then write that

$$\frac{\partial h}{\partial t} = K\,w_{av}, \qquad (8)$$

where w_{av} = average elevation and K = the constant of proportionality. The rate of uplift (assuming an Airy-type crust) is obtained by differentiating equation (6) with respect to time:

$$\frac{\partial w}{\partial t} = \frac{\partial h}{\partial t}\frac{\rho_c}{\rho_m} .$$

The rate of change in the average elevation is given by the difference between the rate of uplift and the rate of erosion

$$\frac{\partial w_{av}}{\partial t} = \frac{\partial w}{\partial t} - \frac{\partial h}{\partial t} ,$$

which gives:

$$\frac{\partial w_{av}}{\partial t} = \frac{\partial h}{\partial t}\left[\frac{\rho_c}{\rho_m} - 1\right]$$

or

$$\frac{\partial w_{av}}{\partial t} = -K\left[1 - \frac{\rho_c}{\rho_m}\right] w_{av} .$$

Hence,

$$w_{av(t)} = w_{av(t=0)}\left|e^{-(1-\frac{\rho_c}{\rho_m})Kt}\right|, \quad (9)$$

where $w_{av(t=0)}$ is the initial elevation and $w_{av(t)}$ is the average elevation at subsequent times. Equation (9) shows that an initial uplift will decay exponentially with time. The parameter $1/K$ is defined as the **erosion time constant**.

The question for basin modelling is what are likely values for the erosion time constant? Modern estimates for the rate of erosion range from 50 cm per 1000 years for the Himalaya to 5 cm per 1000 years for the Appalachians. The average elevation of these features are 5 and 0.5 km respectively, so equation (8) suggests a time constant of about 10 Ma. At this rate, it would take about 60 Ma to reduce the Himalaya and Appalachian mountains to about a third of their present-day elevation.

Several workers have pointed out that since erosion continually modifies the landscape it is important to take into account the conservation of mass. One approach is to assume that erosion is a linear diffusion process and that a balance exists between the rate of erosion and the spatial variation in flux of material. The change in topography can then be written

$$w_{(t)} = w_{(t=0)}\left[e^{-(1-\frac{\rho_c}{\rho_m}\eta)Kt}\right], \quad (10)$$

where η is the slope of the topography and $1/K$ is the **diffusion time constant**.

A diffusion model is important to basin modelling since it allows the transport of sediment from source to sink to be specified, even in basins where the supply of sediment is not known. For example, the difference between the slope after diffusion and the original slope gives the amount of material removed by erosion and the amount that is added by sedimentation. Diffusion causes the area of the material that is removed and added to broaden with time. This effect has consequences for the development of stratigraphic patterns at the basin edge – causing offlap in the region of erosion and onlap in the more distal region of sedimentation.

Variations in the supply of sediment

It is generally assumed in basin modelling that the supply of sediment to a basin is sufficient to maintain a constant bathymetry profile through time. In fact, the amount of sediment supply is highly variable. Studies of bio-stratigraphic data from deep wells in the East Coast US outer continental shelf, for example, show that in Jurassic to Late Cretaceous time this margin was prograding and water depths were in middle shelf environments (i.e. water depths of about 20–100 m) but, by the Eocene, water depths had increased to slope environments (200–2000 m). In the Miocene there was a renewed period of deltaic progradation and a return to shallow-water depths of deposition.

One way to evaluate the effects of sediment supply is to use the backstripping equation discussed earlier. However, rather than consider the 'static' case, we will examine the relationships that exist between changes in the variables with time. Equation (B.4) can then be written

$$\dot{Y} = \dot{S}*\frac{(\rho_m - \rho_s)}{(\rho_m - \rho_w)} + \dot{W}_d - \dot{\Delta}_{SL}\frac{\rho_m}{(\rho_m - \rho_w)} , \quad (11)$$

where the superscript dot refers to the rate of change of the variable. If we assume no sea-level change then $\dot{\Delta}_{SL} \rightarrow 0$ and

$$\dot{Y} \rightarrow \dot{S}*\frac{(\rho_m - \rho_s)}{(\rho_m - \rho_w)} + \dot{W}_d . \quad (12)$$

Equation (12) shows that changes in water depth are determined by the relative difference between the rate of tectonic subsidence and sediment supply. If the sediment supply rate exceeds the tectonic subsidence, $\dot{W}_d < 0$ and water depths decrease. If, on the other hand, tectonic subsidence exceeds the sediment supply, $\dot{W}_d > 0$ and water depths increase. These considerations suggest that the sediment supply has the capacity to control the water depth or, as some workers have referred to it, the **accommodation space** (see Chapter 21) in a basin.

Usually, in the early evolution of a basin (e.g. during the initial subsidence phase) $\dot{Y}$ is at a maximum so there is an increased likelihood that $\dot{W}_d > 0$, and that water depths will increase. This may explain the gradual landward shift of transgressive sequences and onlap that dominates the early evolution of many rift-type basins. As a rift-type basin cools (e.g. during the thermal subsidence phase), $\dot{Y}$ becomes less significant and there is an increased likelihood that $\dot{W}_d < 0$ and water depths will decrease. This may result in a gradual seaward shift of regressive sequences and could explain the occurrence of offlap that tends to dominate the later stages in the evolution of many rift-type basins.

The problem is that the rate of supply of sediment to a basin is a poorly known quantity. This is because the supply of sediments from a source to a basin depocentre is frequently interrupted. Even if continental-derived sediments eventually make their way to a basin depocentre, erosion due to bottom counter currents could remove them and deposit them elsewhere as re-worked material.

As Figure 15.4 shows, the progradation of sediments in river delta systems is probably the simplest form of infill that can be taken into account in basin modelling. The geometry of progradation can be seen as 'clinoformal' structures on seismic reflection profiles. Other margins, however, during the same time period have mainly built up (i.e. have aggraded rather than prograded, see Chapter 20) and show only a limited development of such structures.

Some of the best-documented examples of prograding sediment wedges occur within a foreland basin setting – as clastics move out across a basin following deepening of a basin by thrust/fold loading. Figure 15.13 shows a model in which a prograding clastic wedge migrates out across a foreland basin from the thrust/fold tip towards the flexural bulge. The model includes a tectonic subsidence beyond the flexural bulge that causes a shift in the main depocentre and the migrating wedge to build out over the bulge, thereby limiting the amount of sediment that is deposited in the basin. Eventually, as the second thrust/fold load is emplaced, the main depocentre returns to its original position and the clastic wedge once again migrates out toward the bulge.

Compaction

One of the most important post-depositional processes that may considerably alter the 'architecture' of a sedimentary basin is compaction. The compaction may involve either a mechanical change that closes the pore spaces in a rock or a chemical alteration of the rock. Usually, sedimentary geologists restrict their definition of compaction to the physical removal of pore space, and ignore cementation. However, the latter process needs to be taken into account when modelling the burial history of sediments. Box 15.3 gives a mathematical treatment of mechanical compaction.

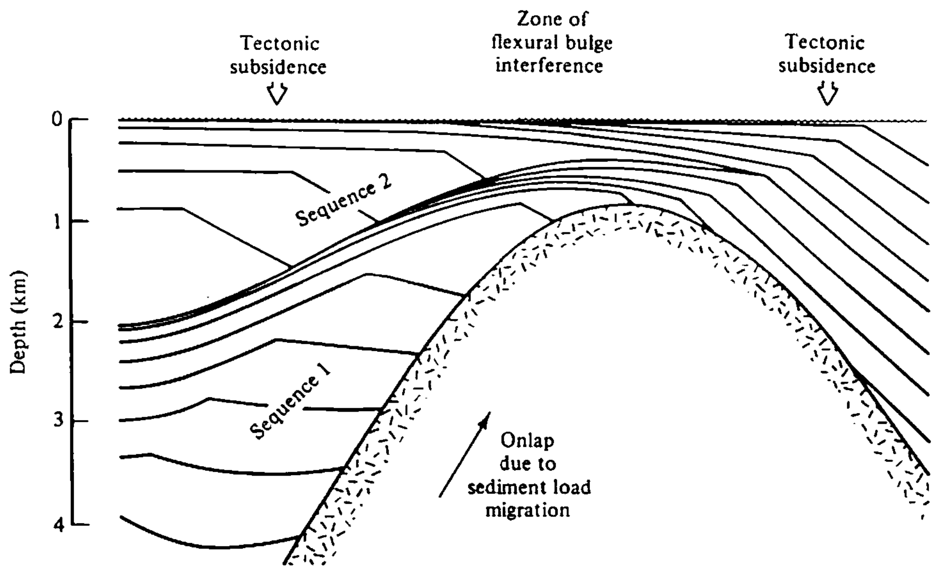

Figure 15.13 Stratigraphic model of a foreland basin assuming that it results from the superposition of two loads: one due to thrust/fold loading and the other due to thermal contraction of the lithosphere following rifting. The centres of the two loads are displaced by about 400 km, which causes uplift in the intervening region due to interference of two flexural bulges. The model assumes that in the foreland basin sediments prograde across the basin from the thrust/fold tip towards the bulge while in the rift-type basin sediments build up to a constant bathymetry profile. □

BOX 15.3

Compaction of sediments

The easiest process to take into account in modelling is a mechanical compaction. Let us consider a cylinder of rock such that its porosity (ϕ) is given by:

$$\phi = \frac{h_w}{h_t}, \tag{C.1}$$

where h_w is the height of the voids and h_t is the total height. Now

$$h_t = h_w + h_g, \tag{C.2}$$

where h_g is the height of the grains. So,

$$h_g = h_t (1 - \phi). \tag{C.3}$$

We can then write for a layer at shallow depth (say one that has not been very compacted) that

$$h_g = h_{ts} (1 - \phi_s),$$

where h_{ts} and ϕ_s are the thickness and porosity respectively at shallow depth. At great depth we have

$$h_g = h_{td} (1 - \phi_d),$$

where h_{td} and ϕ_d are the thickness and porosity respectively at great depth. If we assume that the grains are incompressible then h_g will be constant with depth and,

$$h_{ts} = \frac{h_{td}(1 - \phi_d)}{(1 - \phi_s)}. \tag{C.4}$$

Equation (C.4) can be used to evaluate the thickness any layer would originally have prior to it being buried from its present thickness, provided some information is available on the relationship between porosity and depth. For many normally pressured sediments, the variation of porosity with depth is thought to follow an exponential path.

In basins, there is usually a variation in lithology both horizontally and vertically that reflects changes in the environment of deposition. From the basin edge to its centre the sediment type may change from well-cemented quartz-rich sandstones, grade to siltstones and eventually shales. Since each rock type has a different porosity–depth curve, then the effects of compaction will vary greatly across a basin.

Consider, for example, a 100-m thick stratigraphic unit which changes in lithology across a basin. Towards the basin depocentre, the unit is a sandstone which has a porosity of 20 per cent at a depth of 3000 m. The same rock type at the surface would have a porosity of 56 per cent. Substituting these values in equation (C.4) gives an original thickness of 182 m, nearly twice the preserved thickness. Towards the basin margin, the unit is a well-cemented quartzite with a porosity of 10 per cent. The same rock has a porosity of 20 per cent at the surface. In this case, the original thickness is 120 m which is similar to the preserved thickness.

Simple considerations such as these show that compaction effects may significantly 'distort' the geometry of a basin with time. Since the porosity–depth profile can be expected to vary laterally, in sympathy with changes in sediment type, the distortion will not only be vertical, but also lateral in extent.

Sea-level changes

Sea-level changes modify the stratigraphic record in two different ways: by loading or unloading the basement and by causing a shift in the location of different types of sediment and limits of deposition. If the magnitudes of sea-level changes are known, then equation (B.4) can be used to incorporate their loading effects. Also, since the past few thousands of years were times of known sea-level changes due to waxing and waning of continental ice sheets, the sedimentary response and shoreline shifts that result from a sea-level change can be tested empirically using the geological record.

The main controversy at present is focussed not so much on the stratigraphic response to sea-level changes, but on questions concerning the causes, magnitude and period of these changes during the geological past. There are only a few generally accepted mechanisms for sea-level changes and these produce substantially different types of sea-level curve. One of the mechanisms that has already been mentioned is a change in ocean water volume due to the waxing and waning of continental glaciations (or **glacio-eustasy**). This mechanism tends to produce oscillatory sea-level changes that are of relatively short period with magnitudes of up to 100 m or more.

Another is the change in volume of the ocean basins due to variations in length and/or spreading rate along the world's ocean ridge system (**tectono-eustasy**). This mechanism produces 'smooth' sea-level changes that are of relatively long period. Although the individual contributions to sea-level changes of glaciation, changes in the volume of oceanic ridges and other plate tectonic mechanisms, such as mid-plate swells, volcanism or mountain building, can be estimated, it is not known to what extent these mechanisms contribute to intermediate-period sea-level changes in the past.

Most researchers are now of the view that long-term changes in sea-level (i.e. changes over periods of a few million years to a few hundred million years) are caused by changes in the volume of ocean ridge flanks caused by spreading rate changes. At times of fast spreading (e.g. during the Late Cretaceous) there is a relative increase in the amount of water displaced onto the continental edges, whereas at slow spreading ridges there is a decrease. For example, a ridge system 10^4 km in length that spread at 2 cm/year for 70 Ma and was then subject to a three-fold spreading rate increase of 6 cm/year would cause an increase in sea-level of about 127 m.

W. Pitman carried out a study of the cumulative contribution to sea level of spreading rate changes along the entire length of the world's ocean ridges as well as some ridges (e.g. Pacific/Kula) which have long since been consumed at the deep-sea trenches. The sea-level curve obtained by Pitman shows a maximum value in Late Cretaceous time of 350 m above present levels and then a steady decline to the present day.

The Pitman curve may be tested using stratigraphic data since times of sea-level rise should have left a record of transgressive and highstand deposits in the continental interiors. As sea level fell, these deposits would have been stranded as elevated outcrops providing, of course, that erosion did not completely remove them first. One group of rocks that apparently escaped significant erosion is the Late Cretaceous Coleraine formation of Minnesota in the central USA. This formation is at a present-day elevation of 390 m which, when corrected for local glacial loading/unloading effects, is close to the predicted value using the Pitman curve.

One problem with using continental flooding estimates as an indicator of past sea-level changes is that the continents themselves may have undergone large-scale tectonic movements subsequent to a sea-level change. There is therefore considerable debate at the present time about the suitability of certain continents to act as **stable reference frames**, although some workers have argued that the central USA and, perhaps, the Russian platform, have probably both been relatively stable since the Late Cretaceous.

Because their tectonic history can now be reasonably well described by thermal models, rift-type basins have become the main focus to test the effects of sea-level changes on the stratigraphic record. In equation (B.4), for example, if y is known and well data can be used to estimate W_d and S^* then Δ_{SL} could be isolated by backstripping. Several attempts have been made in the past to estimate Δ_{SL} from bio-stratigraphic data but, while the resulting curves agree with the timing of the Late Cretaceous maximum in Pitman's curve, they are about a factor of two smaller in magnitude. The reason for this difference is not clear, but it could be due to an inadequate global data set for the backstripped well data, errors in the calculation of ridge crest volume changes, or the competing effects of other tectonic factors.

Another test of sea-level changes is to examine the stratigraphic record itself. For example, if sea level fell by either 350 m or 150 m there should be an observable difference in the sediment accumulation, especially in the later stages of rift-type basin development when the tectonic subsidence is small. At the East Coast US margin, for example, the change in the thermal subsidence since 85 Ma is about 320 m. If on the shelf sea level fell by 350 m, then little sediment should have accumulated. If, on the other hand, sea level fell by 150 m, then as much as 375 m could accumulate. The post-Cretaceous sediment thickness at the East Coast US margin is 500 m or greater, suggesting the use of a low rather than a high sea-level curve.

To date, the most exhaustive study using the stratigraphic record to deduce sea level was the one carried out by P. R. Vail and colleagues at Exxon in the late 1970s. By correlating unconformity bounded stratigraphic sequences within and between basins, the Exxon group were able to document that certain unconformities are widespread in extent and therefore are most likely caused by sea-level changes. The resulting **'Vail sea-level curve'**, as it has come to be known, was an 'oscillatory' curve, with periods that vary from several hundreds of thousand years to a few tens of million years and amplitudes of up to 300 m, that was superimposed on the 'smooth' Pitman curve.

The method used by Vail and colleagues to estimate the magnitude of sea-level change was based primarily on the geometry of stratal relationships observed on seismic reflection profiles. Unfortunately, their procedure may have over-corrected for tectonic effects. The resulting sea-level curve was characterised by gentle sea-level rises, but very rapid falls. Later studies by the Vail group did apparently take into account tectonics, since the sea-level falls were less abrupt and, in some cases, as smooth as the sea-level rises.

Of particular interest in basin modelling is the stratigraphic response to sea-level changes – whether they be smooth of the type proposed by Pitman or oscillatory as suggested by Vail. It might be expected that a sea-level rise would cause a landward shift of sediment belts and onlap, while a fall would cause a seaward shift and offlap. The stratigraphic response to sea-level changes, however, is complex and depends on the relative rates of tectonic subsidence, sediment supply and water depth changes.

There has been considerable progress in the past few years in understanding the stratigraphic response to simple periodic changes in sea level. However, these studies are limited by the fact that we still do not know precisely enough the actual magnitude and period of sea levels in the past. The smooth Pitman curve is presently in dispute because of uncertainties in both the lengths and spreading rates of the ocean ridge systems and the geological time scales. According to one recent analysis, the amplitude of the Late Cretaceous sea-level highstand could range from -5 to 450 m. Similarly, the Vail curve is disputed because a sea-level rise and fall may give rise to quite similar depositional environments and onlap/offlap patterns, depending on the relative rates of tectonic subsidence and sediment supply. Thus, it may be difficult to use the stratigraphic record to infer sea-level changes, unless precise knowledge is obtained of the geometry (onlap and offlap) and depositional environment of individual sequences. One way to obtain this information for an offshore basin is by a combination of seismic reflection profiling and deep drilling data. While such a data set was apparently available to Vail and his colleagues for many of the world's sedimentary basins, most workers do not have access to it and it has proved difficult to verify the details of their sea-level curve.

As a result of these difficulties, some workers have taken a different approach, the most promising of which have been the derivation of sea-level curves from studies of the ^{18}O record in deep-sea sediments and basin modelling that incorporates tectonics and simple periodic waveforms for sea-level changes.

Conclusions

The various sedimentary factors discussed in the previous sections complicate our ability to construct more complex basin models than ones constructed on the basis of simple thermal and mechanical concepts alone. The problem is that while we understand how erosion, changes in the sediment supply, compaction and sea-level changes may influence the stratigraphy individually, it is not known how they combine with tectonics, or with each other, to control the architecture of basins. A single dynamical model for the formation of sedimentary basins is therefore still beyond reach at the present time.

A basin model which incorporates tectonics together with the sedimentary processes discussed in this chapter would have great predictive power. Of particular interest would be the ability to construct synthetic stratigraphic profiles for basins of different ages in different tectonic settings. By incorporating the effects of erosion, changes in sediment supply, and compaction it should be possible, for example, to quantitatively evaluate the effects of sea-level changes in the past. It has been suggested, on the basis of existing models, that sea-level changes may only be an important control on the stratigraphy during the *later* stages of basin formation when tectonic effects are subdued. During the *early* stages, tectonics may dominate. If subsequent modelling studies are able to verify a tectonic control, then this would have profound implications for correlative stratigraphy and our understanding of the nature of the control of the stratigraphic record.

One aspect of basin modelling that has not been discussed here is its application to oil and gas exploration. The thermal and mechanical parameters that best fit observed stratigraphic data can be used to compute the temperature history of the sediments and the maturation of any source beds that are thought to be present (see Chapter 18). Although most industry geologists consider actual measurements of maturity to be far more reliable, a predicted maturity may be useful in a frontier basin where there has been little or no drilling. Basin models may also be used to compute how the stratal relationships within a basin may vary with time and position. This is important for models which consider how fluids such as oil may flow from low to high points in a basin.

In addition to its importance to the oil and gas industry, basin modelling represents an unparalleled challenge to the academic community. The reason why individual basins have the shape they do is still enigmatic and remains one of the major outstanding questions to be addressed by the next generation of students in the Earth sciences.

Further reading

Allen, P. A. & Allen, J. A. (1990) *Basin Analysis: Principles and Applications*, Blackwell Scientific Publications, 451 pp.

A book that summarises the modern approach to basin

analysis. The book provides further reading on many of the concepts developed here, particularly the sedimentary factors of erosion, sediment supply, compaction, and sea-level changes.

Bally, A. W. (1981) *Geology of Passive Continental Margins: History, Structure and Sedimentologic Record* (with special emphasis on the Atlantic margin), American Association of Petroleum Geologists, Education Course Note Series No. 19, 350 pp.
An introductory text devoted to the structure, subsidence history and sedimentological evolution of passive continental margins. Emphasis is on the East Coast US Atlantic margin, but discussions of the Gulf Coast and Tethys margins are included.

Bally, A. W. (ed.) (1984) *Seismic Expressions of Structural Styles*, American Association of Petroleum Geologists, Studies in Geology Series No. 15.
An atlas showing the structure of sedimentary basins as revealed by multichannel seismic reflection profiling.

Biddle, K. T. & Christie-Blick, N. (eds.) (1985) *Strike-Slip Deformation, Basin Formation, and Sedimentation*, Society of Economic Paleontologists and Mineralogists, Tulsa, Oklahoma, 396 pp.
A collection of papers describing the origin of the small basins that form in strike-slip settings.

Kent, P., Bott, M. H. P., McKenzie, D. P. & Williams, C. A. (eds.) (1982) *The Evolution of Sedimentary Basins*, Philosophical Transactions of the Royal Society of London, Series 305A, 338 pp.
An authoritative account that summarises the state of knowledge in the early 1980s of the principal basin-forming mechanisms. Emphasis is on the rift-type and foreland basins.

St John, B. (1980) *Sedimentary Basins of the World*, American Association of Petroleum Geologists, Map and chart series at 1:40 000 000.
A useful map showing the distribution of the world's sedimentary basins.