**Intraplate stress and sedimentary basin evolution**

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CHAPTER

# 21

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# INTRAPLATE STRESS AND SEDIMENTARY BASIN EVOLUTION

## Introduction

The origin of sedimentary basins is a key element of the geological evolution of the continental lithosphere. During the last decade substantial progress has been made in understanding the thermomechanical aspects of sedimentary basin evolution and the isostatic response of the lithosphere to surface loads such as basins. Most of this progress has been made by the development of new modelling techniques, insights into the mechanical properties of the lithosphere (the **rheology**), and in the processing of new, high-quality data sets from previously unexplored areas of the globe.

The fundamental quantities of rheology are stress and strain respectively. **Stress** can be defined as the force per unit area acting on a surface within a solid. Once the dynamic quantity (stress) is specified, the kinematic quantity (deformation or **strain**) can be derived. Almost all basin modelling carried out so far has been in terms of lithospheric displacements, thus refraining from a full examination of dynamic controls exerted by the lithospheric stress. This is because the calculation of stress is very sensitive to the adopted mechanical properties of the lithosphere. These lithosphere rheologies have been by definition unrealistically simple, especially in view of recent advances in rock mechanics studies. This is true of models for both extensional and compressional sedimentary basins as discussed in Chapter 15. For example, most models for extensional basin formation are keyed to lithospheric strain due to an unknown and unspecified stress field rather than to the strain response of the lithosphere to a known and/or realistic stress state. However, changes in plate-tectonic regimes and associated stress fields have been shown to be quite important in controlling the subsidence record and stratigraphic architecture of extensional basins. Similarly, models of basin development in compressional environments have been conventionally related to lithospheric flexure profiles, again not involving the dynamic control of the compressional stresses intrinsic to this particular tectonic setting. A major reason that the relationship between lithospheric stresses and displacements in tectonic modelling has been largely neglected is that little has been known about the actual stress state in the lithosphere. This situation has recently changed drastically as the result of the World Stress Map Project of the International Lithosphere Project. This project has revealed the existence of regionally consistent patterns of tectonic stress in the lithosphere. Moreover, structural measurements to establish the temporal evolution of palaeo-stress have begun in a number of sedimentary basins. Simultaneously, numerical modelling has resulted in a better understanding of the causes of the observed present-day stress levels and stress directions in the various lithospheric plates. Such studies have shown a causal relation-

## BOX 21.1

# Stress

The state of stress of a cube of rock is usually described in terms of nine stress components of which only six are independent if the body is in equilibrium. The stress on each face of the cube (Figure 21.1a) can be separated into a stress component **normal** to the face ($\sigma$), and a **shear stress** component ($\tau$) along the face. The shear stress itself can be resolved into two components parallel to the directions of the two other axes of the coordinate system. In general, it is possible to calculate the state of stress once the stresses on three mutually perpendicular planes are known. Three planes, called principal planes of stress, exist on which the shear stresses are zero. Stresses in the mechanically strong upper part of the lithosphere are compressive at depths larger than a few tens of metres. The principal stresses (denoted by $\sigma_1$, $\sigma_2$, and $\sigma_3$ for the largest, intermediate and minimal principal stress respectively) are in the lithosphere located in approximately horizontal and vertical planes (Figure 21.1b).

The **vertical stress** $\sigma_v$ is approximately equal to the product of density of the overlying mass, gravitational acceleration and depth. The maximum horizontal stress and the minimum horizontal stress are indicated by $\sigma_{Hmax}$ and $\sigma_{Hmin}$ respectively. The style of deformation is determined by the relative stress magnitudes. For example, **normal faulting** occurs when $\sigma_v > \sigma_{Hmax} > \sigma_{Hmin}$, while **strike-slip faulting** occurs when $\sigma_{Hmax} > \sigma_v > \sigma_{Hmin}$, and **thrust** or **reverse faulting** occurs when $\sigma_{Hmax} > \sigma_{Hmin} > \sigma_v$. When the principal stresses are equal, shear stresses are zero and the stress state is hydrostatic. When the principal stresses are different, shear stresses $\tau$ exist. For example, as shown in the left-hand side panel of Figure 21.1b, the maximum shear stress on the plane bisecting the $\sigma_{Hmax}$ and $\sigma_v$ directions and coinciding with the $\sigma_{Hmin}$ direction equals $(\sigma_{Hmax}-\sigma_v)/2$.

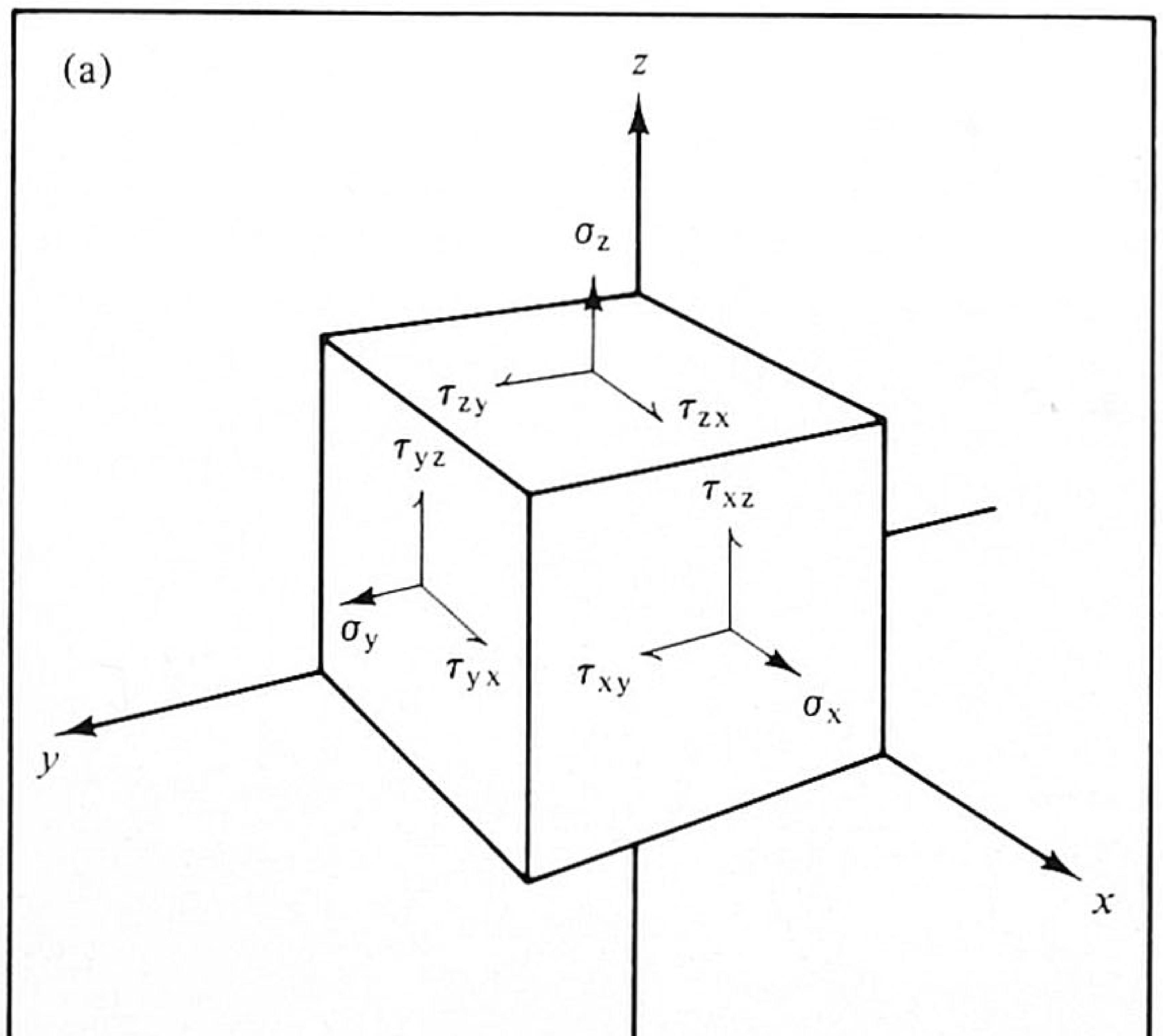
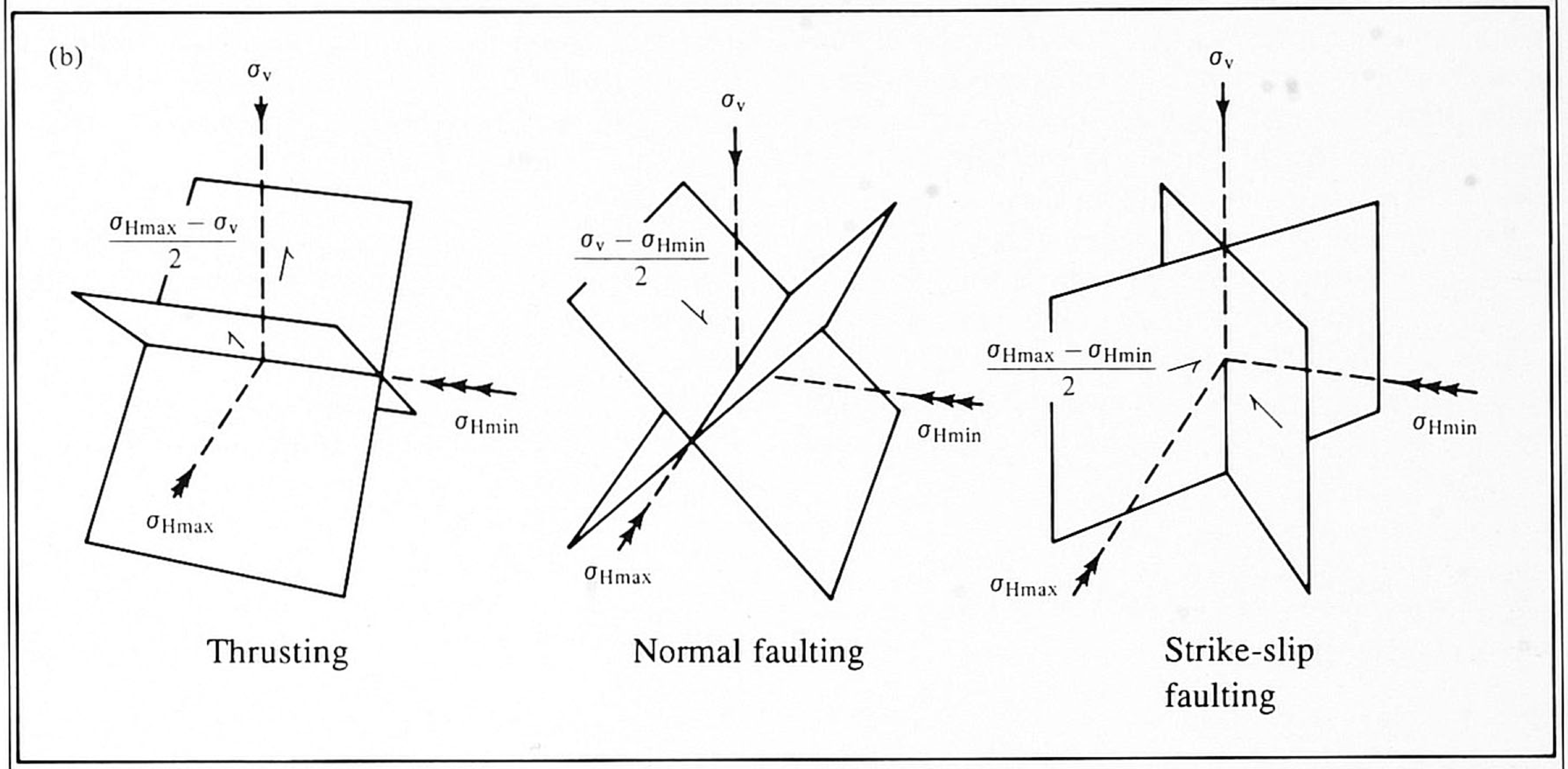
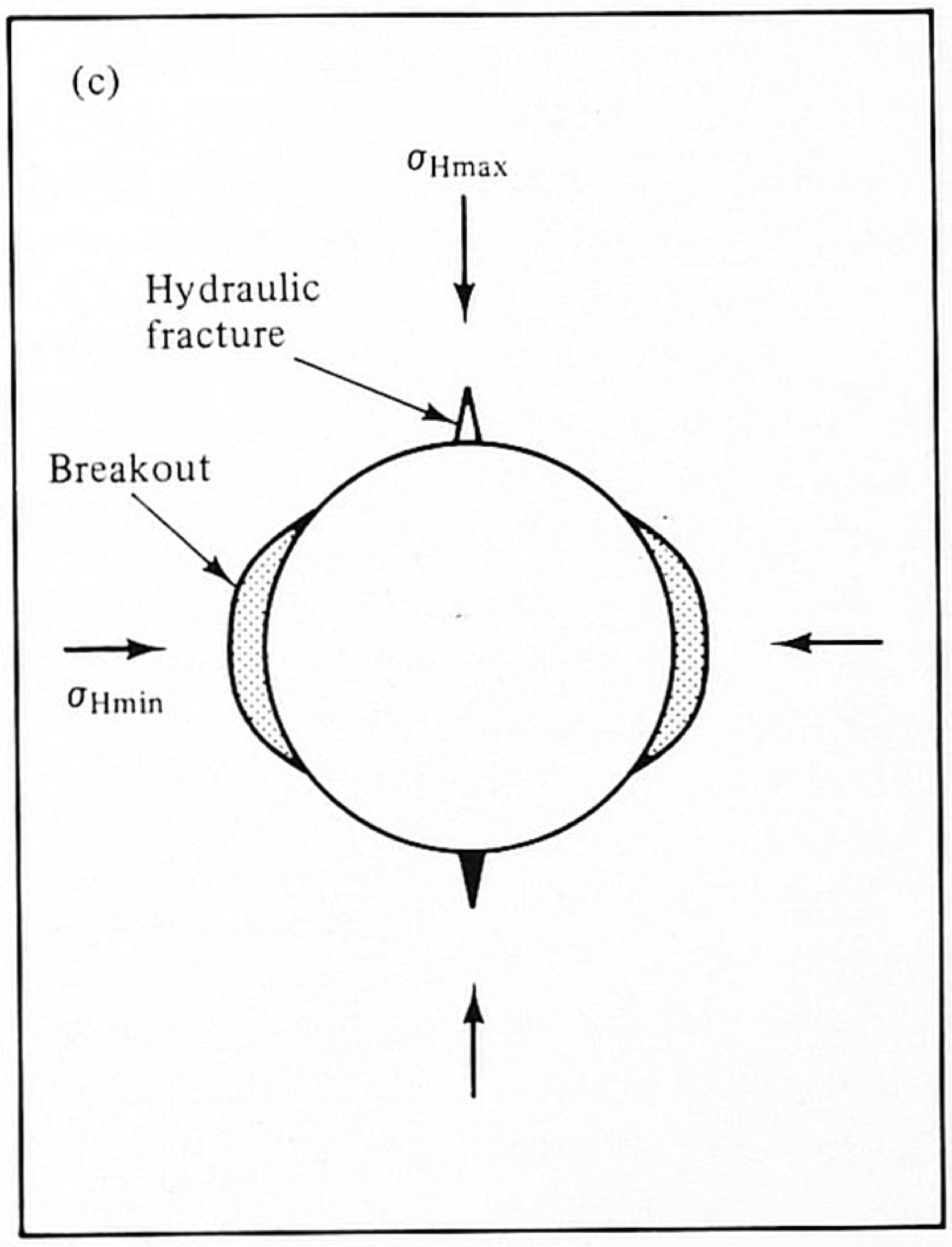



***Figure 21.1*** Stress conventions and orientations. (a) Stress components on three faces of an infinitesimal cube of rock volume. A stress acting on a plane can be separated into a component normal to the plane ($\sigma_x$, $\sigma_y$, or $\sigma_z$ respectively) and two shear stress components acting within the plane ($\tau_{xy}$, $\tau_{xz}$; $\tau_{yx}$, $\tau_{yz}$; $\tau_{zx}$, $\tau_{zy}$). When in equilibrium ($\tau_{zy} = \tau_{yz}$; $\tau_{zx} = \tau_{xz}$; $\tau_{xy} = \tau_{yx}$) the stress field can be described in terms of six independent components. (b) Principal stresses in the lithosphere and planes of maximum shear stress. The principal stresses are in approximately horizontal and vertical planes. The vertical stress $\sigma_v$ is approximately equal to the product of density, gravitational acceleration and depth. $\sigma_{Hmax}$ and $\sigma_{Hmin}$ denote maximum and minimum horizontal tectonic stress respectively. The style of deformation is determined by the relative stress magnitudes. Thrusting or reverse faulting occurs when $\sigma_{Hmax} > \sigma_{Hmin} > \sigma_v$, normal faulting occurs when $\sigma_v > \sigma_{Hmax} > \sigma_{Hmin}$, and strike-slip faulting occurs when $\sigma_{Hmax} > \sigma_v > \sigma_{Hmin}$. (c) Deformation of a horizontal circular cross-section of a bore-hole is caused by horizontal intraplate stresses $\sigma_{Hmax}$ and $\sigma_{Hmin}$ leading to the formation of breakouts (parts of the bore-hole wall that collapse) and hydraulic fractures. The maximum compressive stress induced around the bore-hole (and hence the breakout) is centred on the azimuth of the least horizontal tectonic stress. In a hydraulic fracturing experiment a portion of the well-bore is isolated and pressurised by injected fluids until a tensile fracture develops in the direction of the maximum horizontal stress. Down-hole measurements have confirmed that breakouts and hydraulic fractures form perpendicular to each other. □

**21.1 continued**




ship between the processes at plate boundaries and deformations in the plates' interiors. In models of the evolution of sedimentary basins located in the interiors of the plates, however, the role of these lithospheric stresses had been largely ignored. Recently, the first steps have been taken towards exploring the consequences of the existence of intraplate stress fields in the lithosphere for models of the formation and evolution of sedimentary basins.

Rates of tectonic subsidence following extensional rifting are dependent on stresses affecting the basin as well as on crustal cooling. This is important, as cyclic short-term changes in eustatic sea level have been regarded by many as the sole explanation for these short-term deviations from long-term subsidence. In particular, the development of quantitative stratigraphic techniques has led to the widespread use of a set of charts of cyclic changes in sea level for stratigraphic correlation. Since these key-concepts were presented by Vail and colleagues (see Chapter 20) much work has been done in testing, evaluating and developing them. In the present chapter, we discuss these basic concepts in sedimentary basin analysis in the light of recent theoretical advances in lithospheric dynamics.

We first review evidence for the existence of intraplate stress fields in the lithosphere. This is followed by a discussion of some implications of intraplate stress for quantitative modelling of sedimentary basins. Finally, the potential to separate tectonic contributions from eustatic contributions to the apparent sea-level record is explored.

## Intraplate stresses in the lithosphere

The intraplate stress field in the lithosphere is the superposition of regional components and local sources of stress. The regional stresses in the lithosphere are induced by plate-tectonic forces and can be traced over large distances. On a more local scale, however, other contributors to the stress field might dominate. Examples are stresses associated with topographic anomalies and crustal thickness inhomogeneities at passive continental margins. Further, as a result of temperature variations rocks expand or contract inducing large thermal stresses in cooling lithosphere. The Earth's surface has the shape of a spheroid with polar flattening and an equatorial bulge and, consequently, plates deform as they change latitude. This deformation generates **membrane stresses** in the lithosphere. Finally, **flexural stresses** are induced due to vertical loads on the lithosphere, in particular by sedimentary sequences at passive continental margins. Stress components are described in Box 21.1. Stress magnitudes are expressed in the SI unit for stress which is the **Pascal** (Pa), equivalent to one Newton per square metre ($Nm^{-2}$), although it is more convenient in stress studies of the lithosphere to use the **megaPascal** (1 MPa = $10^6$ Pa). At the same time, multiples of the CGS unit ($dyncm^{-2}$) are quite frequently used, in particular the **bar** (1 MPa = 10 bar) and the **kilobar**. Of the various locally induced stress sources the flexural stresses and thermal stresses stand out in magnitude (up to the order of several hundred MPa) while, as shown in Table 21.1, most of the other mechanisms produce stresses with a characteristic level of the order of a few tens of MPa.

The present stress field in the various plates has been studied in great detail by the application of a wide range of observational techniques. These include detailed analysis of earthquake focal mechanisms, *in situ* stress measurements, including hydraulic fracturing and stress-relief measurements, and analysis of stress-induced elliptical bore-hole deformations of wells or 'breakouts' (see Box 21.2).

The observed modern stress orientations show a remarkably consistent pattern, especially considering the heterogeneity in lithospheric structure in the plates. These stress-orientation data indicate a propagation of stresses away from the plate boundaries over large distances into the plate interiors. The World Stress Map Project has convincingly established the existence of these large-scale, consistently oriented stress patterns to be a general characteristic of lithospheric plates. These observations, compiled in Figure 21.2, prove that regional stress fields are usually dominated by the effect of plate-tectonic forces acting on the lithosphere.

Strong evidence exists for changes in the magnitudes and orientations of these stress fields on time scales of a few million years in association with collision and rifting processes in the lithosphere. The results of a recent compilation of stress-direction data for the European platform are displayed in Figure 21.3. In this case, the orientation and evolution of the principal **palaeo-stress** axes is inferred from **microstructure meas-**

Table 21.1 *Stress mechanisms and stress levels in the lithosphere (1 kbar = 100 MPa). Apart from regional stresses in the lithosphere induced by plate-tectonic forces, other sources might dominate on a more local scale. Examples are stresses associated with topographic anomalies and crustal thickness inhomogeneities at passive margins. As a result of temperature variations, rocks expand or contract which can induce large thermal stresses in cooling lithosphere. Deformation of the plates as they change latitude leads to membrane stresses in the lithosphere because the Earth is not a perfect sphere. Flexural stresses are generated due to vertical loads on the lithosphere, in particular by the accumulation of sedimentary sequences at passive continental margins.*

| Mechanism | Stress | Order of magnitude |
|---|---|---|
| Ridge push | $\sigma_{RP}$ | 10 MPa |
| Drag | $\sigma_{D}$ | 1 MPa |
| Slab pull | $\sigma_{SP}$ | 10–100 MPa |
| Topography | $\sigma_{TO}$ | 10–100 MPa |
| Crustal thickness contrast | $\sigma_{CT}$ | 1–10 MPa |
| Thermal | $\sigma_{TH}$ | 100 MPa |
| Membrane | $\sigma_{ME}$ | 10 MPa |
| Flexure | $\sigma_{FL}$ | 100 MPa |
| Overburden | $\sigma_{OB}$ | 10 MPa |

BOX 21.2

## Techniques to measure the direction of the modern stress field

**Earthquake focal mechanisms** are obtained from radiation patterns of seismic waves. These define a set of two perpendicular fault planes and slip vectors, with P and T axes corresponding to the directions of maximum shortening and extensional strain for these shear faults. It is very difficult to determine the stress levels at depths of the earthquake source. However, it is generally possible to constrain relative stress magnitudes from the style of active tectonic faulting. ***In situ* stress-relief measurements** determine the strain relaxation when a rock sample is separated from the volume of surrounding rock. The change of stress after the sample is removed is then calculated from the strain relief. These measurements are only possible near the Earth's surface or on an excavated surface in mines. **Well-bore breakouts** occur spontaneously as a result of stress relaxation around the bore-hole. As predicted by simple elastic theory, the induced maximum compressive stress around a vertical bore-hole (and hence the breakout) is centred on the azimuth of the least horizontal stress (Figure 21.1c). In a **hydraulic fracturing experiment** a portion of the well-bore is isolated and pressurised by injected fluids until a tensile fracture develops in the direction of the maximum horizontal stress. Breakouts and hydraulic fractures are formed perpendicular to each other as has been shown by down-hole experiments in many wells in different tectonic settings. Breakout data are a very valuable and flexible tool in the study of the lithospheric stress field. They allow, for example, multiple determinations in a single well as well as the possibility to investigate regional consistency by the use of a large number of wells. Recently the Ocean Drilling Program has started to incorporate stress measurements in its drilling operation scheme. This is particularly important for our knowledge of the state of stress in those oceanic plates where earthquake focal mechanism studies are hampered by a low level of intraplate seismicity. The World Stress Map Project has demonstrated that the various techniques show, in general, excellent agreement.

**urements** in sedimentary rocks, such as pressure solution surfaces (stylolites), veins with secondary mineralisations, or small faults with a clear indication of the sense of displacement (Figure 21.4). Box 21.3 discusses in more detail techniques used to measure palaeostress.

As such, the inferred information on palaeo-stress fields is less precise than the results of studies of modern stress indicators. The study of palaeo-stress fields, however, adds geological time as a parameter crucial to understanding the temporal fluctuations of stress fields in the plates.

A complementary approach to collecting stress indicator data is the study of intraplate stress fields using numerical techniques. In the first phase of modelling intraplate stress fields resulting from plate-tectonic forces, models were tested against stress orientation data inferred from earthquake focal mechanism studies to quantify the relative and absolute importance of various possible driving and resistive forces. These studies resulted in the overall understanding that ridge push and slab pull (see Chapter 9) are the two main driving forces. Since then, a better understanding has been obtained of the age-dependence of the forces acting on the lithosphere. This development has benefited from advances in the analysis of the subduction process. Subsequent implementation of these new features and insights in stress modelling led to the successful prediction of various deformation processes within lithospheric plates.

Because of better constraints on the thermo-mechanical and tectonic evolution of the oceanic lithosphere, which is relatively well understood, these models have concentrated on oceanic plates or lithospheric plates with major oceanic parts. An example is given in Figure 21.5a which shows a comparison of the predicted and observed stress fields in the northeastern Indian Ocean. This area is at present the most seismically active oceanic intraplate region on Earth. Here stresses reach high levels due to focusing of compressional resistance associated with collision of the Indian and the Eurasian plate and subduction of relatively young oceanic lithosphere in the northern part of the Java–Sumatra Arc. Stress-orientation data

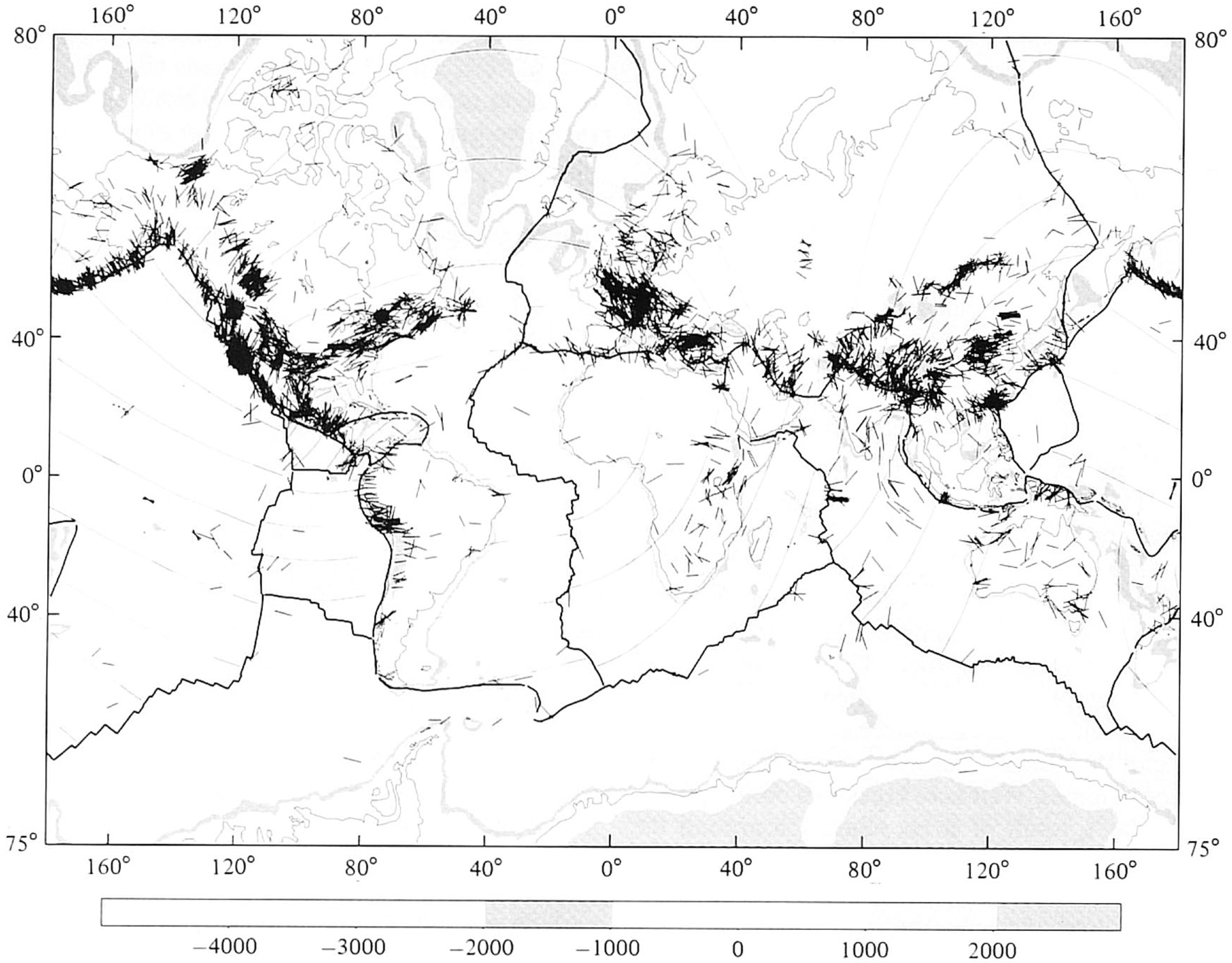



*Figure 21.2* The World Stress Map. Axes give the orientation of the maximum horizontal stress field $\sigma_{Hmax}$. The line length of the data is proportional to the quality of the data, and therefore our confidence in the data. Earthquake focal mechanisms are obtained from the radiation pattern of seismic waves and lead to the definition of a set of two orthogonal fault planes and slip vectors, with P and T axes that represent the directions of maximum shortening and extension. The World Stress Map Project has demonstrated that the various techniques show, in general, excellent agreement, revealing the existence of regionally consistent patterns of tectonic stress in the lithosphere. The observed stress directions also agree quite well with the directions of absolute plate motions indicated by the dashed lines in the figure. The Indo-Australian plate, where the stress field is probably dominated by age-dependent slab pull forces, forms an exception to this pattern. Note the uneven distribution of the data as an artefact of the closer sampling of certain areas such as California and the North Sea in association with intensive studies of seismic risk and drilling for oil exploration and production respectively. Future stress measurements by the Ocean Drilling Program will forward our knowledge of the state of stress in oceanic plates such as the Pacific plate, where a low level of intraplate seismicity limits the determination of stress directions by earthquake focal mechanism studies. Shading relates to topography with top of Everest at 0 m. □

demonstrate a rotation of the observed stress from N–S oriented compression in the north to a more NW–SE directed compression in the southeastern part of the Bay of Bengal area, in agreement with the calculated stress field. The intraplate stress field as calculated also explains folding of the entire lithosphere observed in this area from seismic profiles and satellite gravity data (Figure 21.5b).

These numerical models have shown that high-magnitude stresses can be concentrated in the plates' interiors. They have also shown, in agreement with observations such as shown in Figure 21.5, that spatial variations and large-scale rotations in orientation in the stress fields may occur. These models and observations make clear that stress provinces can vary in size from an entire lithospheric plate to only portions of plates. Modelling results such as displayed in Figure 21.5 corroborate the observed stress patterns by

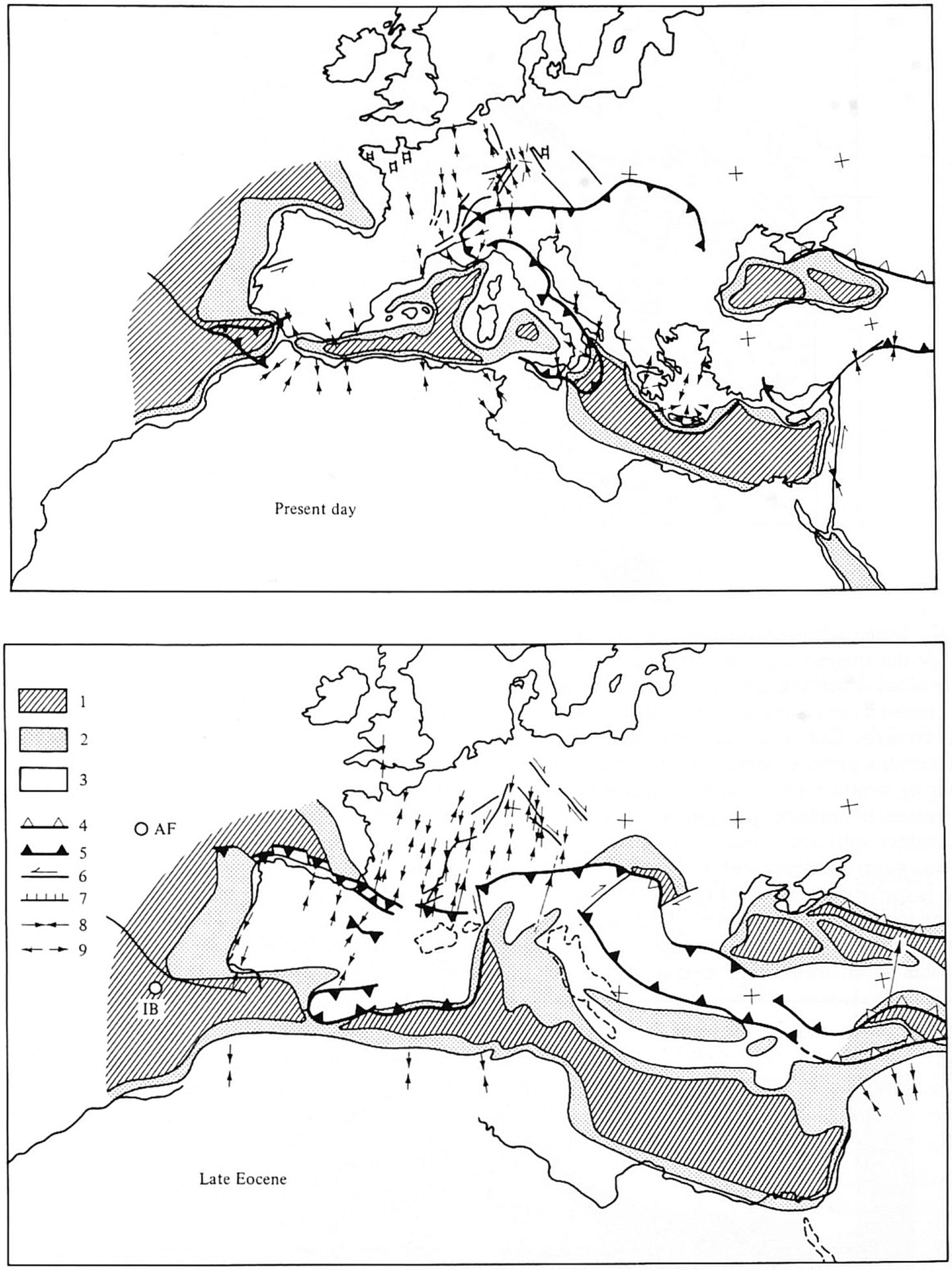



***Figure 21.3*** Compilation of observed maximum horizontal stress directions in the European platform. Key: 1, oceanic crust; 2, thinned continental crust; 3, continental crust; 4, subduction; 5, overthrust; 6, strike-slip fault; 7, normal fault; 8, azimuth of maximum principal stress $\sigma_1$; 9, azimuth of minimum principal stress $\sigma_3$. (a) Present-day stress field from *in situ* measurements, focal mechanism studies and geologic stress indicators. (b) Palaeo-stress field during Late Eocene times and reconstructed geodynamic evolution in a framework of Cenozoic Africa/Eurasia collision. Eurasia is fixed and AF and IB denote the Africa/Eurasia and Iberia/Eurasia rotation poles. The rotation poles and motion vectors are given for a time slice between 54 and 35 Ma. The stress data for the European platform given in a framework of Cenozoic Africa/Eurasia collision demonstrate stress propagation away from the Alpine fold belt in the platform region. □

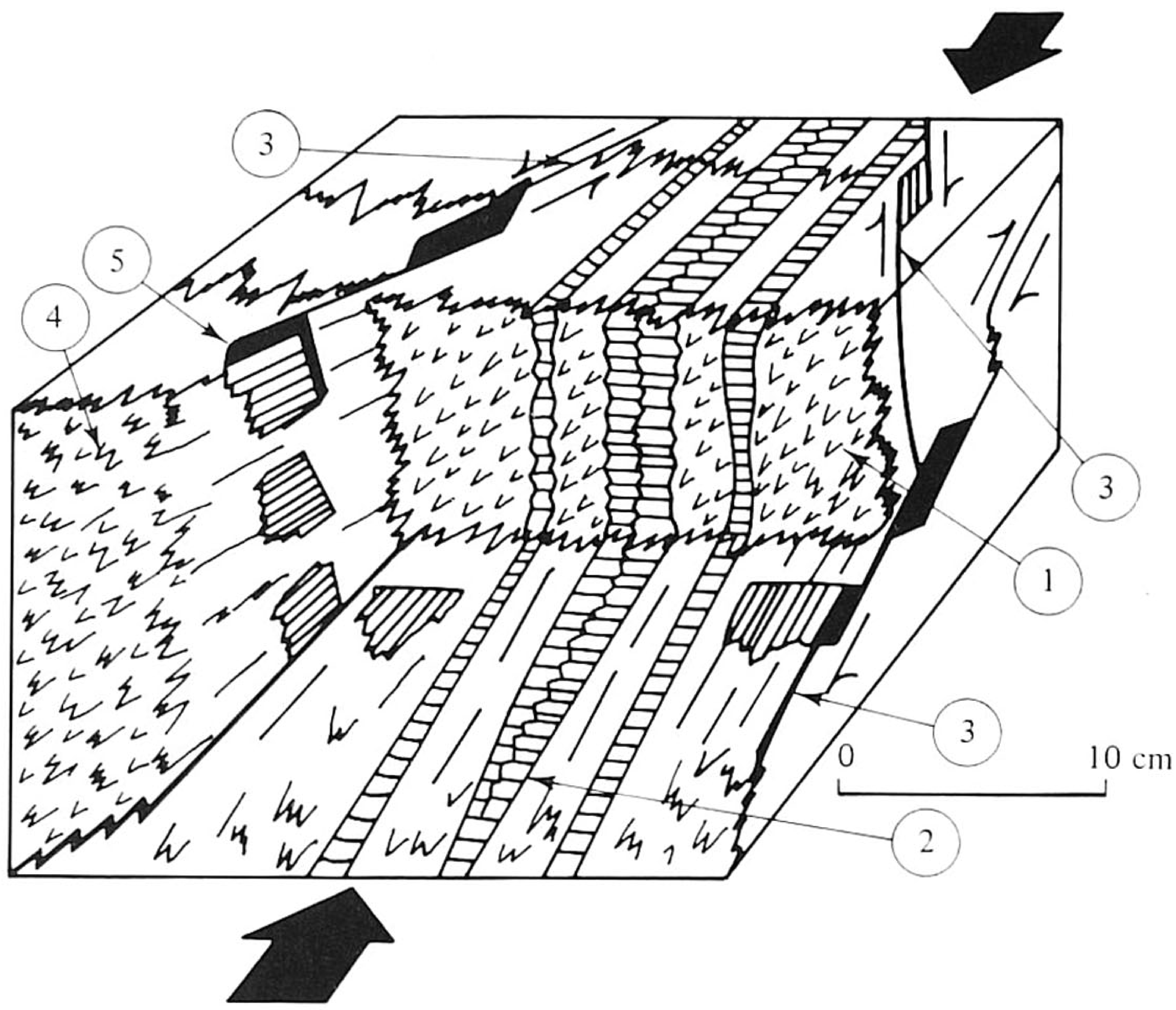
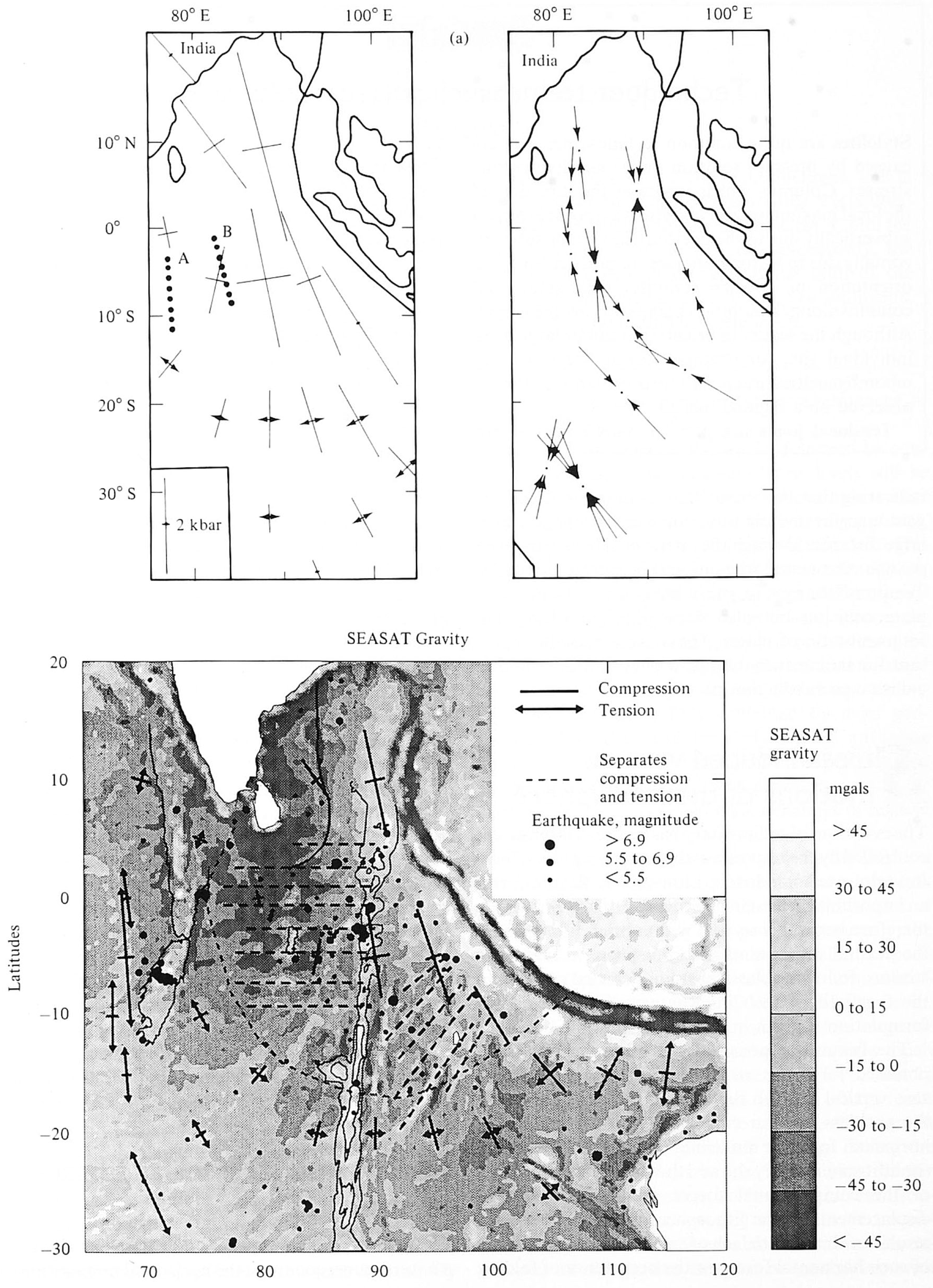



*Figure 21.4* Schematic presentation of microstructures measured for the analysis of palaeo-stress patterns. 1, tectonic stylolites which are common in limestones and which are caused by pressure solution under non-hydrostatic stresses. Columns represent the direction of the local maximum principal stress axis $\sigma_1$. Stylolites form subvertically by overburden or subhorizontally due to tectonic stresses. In practice, only the orientation of tectonic stylolites with subnormal columns along the solution plane seam are measured. Although the scatter in orientation is sometimes large in each site, due to fractures and inhomogeneities, the method can provide quite consistent stress patterns on a regional scale. 2, tensional joints which are parallel veins of calcite (or quartz), with crystals growing normal to the joints. These structures are commonly associated with tectonic stylolites and form parallel to the stylolitic columns. The local minimum principal stress axis ($\sigma_3$) is inferred from the average normals to these joints. 3, measurements on a small fault plane. 4, measurements on a pressure solution surface. Reliable measurements on a small fault plane and on a pressure solution surface require the determination of the sense of displacement. 5, asymmetric steps of fibrous calcite or quartz (accretionary growth of crystal fibres during slip movement), or mechanical striation on the rocks also allow the determination of the sense of displacement. The black arrows indicate the maximum compressive stress axis. □

*Figure 21.5* (a) Regional stress field in the northeastern Indian Ocean. Left: predicted stress field induced by plate-tectonic forces acting on the lithosphere. Dotted lines give the location of long seismic reflection profiles that show folding of the entire lithosphere caused by the high stress levels in the area. Right: the orientation of maximum horizontal compressive stress inferred from earthquake focal mechanism studies. Note the good agreement between theoretical modelling of the stress field and stress observation data. (b) Gravity anomalies and stresses predicted by numerical modelling. Gravity highs from satellite data corresponding to the folded oceanic lithosphere generally trend normal to the predicted stress field. Note the large-scale rotation of the fold axes(indicated by dashed lines) from an E–W trend in the area close to Sri Lanka to a NE–SW orientation in the ocean region close to the Sumatra trench. The line that marks the transition between tensional and compressional stresses in the northeastern Indian Ocean also coincides roughly with the boundary of the area where folding occurs. □

(a)



80° E
100° E
India
10° N
0°
10° S
20° S
30° S
A
B
2 kbar
SEASAT Gravity
Compression
Tension
Separates compression and tension
Earthquake, magnitude
> 6.9
5.5 to 6.9
< 5.5
SEASAT gravity
mgals
> 45
30 to 45
15 to 30
0 to 15
−15 to 0
−30 to −15
−45 to −30
< −45
Latitudes
20
10
0
−10
−20
−30
70
80
90
100
110
120


(b)

## BOX 21.3

### Techniques to measure palaeo-stress fields

**Stylolites** are quite common in limestones and are caused by pressure solution under non-hydrostatic stresses. Columns which represent the direction of the local maximum principal stress axis $\sigma_1$ are formed subvertically due to the overburden load or subhorizontally due to tectonic stresses. In practice, only the orientation of tectonic stylolites with subnormal columns along the solution plane seam are measured. Although the scatter in orientation can be large in an individual site, for example due to fractures and inhomogeneities, usually a consistent stress pattern is observed on a regional scale.

**Tensional joints** are parallel veins of calcite (or quartz), with crystals growing normal to the joints. These structures are frequently associated with tectonic stylolites and are parallel to the stylolitic columns. The orientation of the local minimum principal stress axis ($\sigma_3$) is given by the average normals to the joints. Measurements on a small fault plane or on a pressure solution surface require an indication of the sense of displacement. These follow usually from tectonic stylolitic columns oblique to the surface containing them, asymmetric steps of fibrous calcite or quartz (accretionary growth of crystal fibres during slip movement), as well as from mechanical striation on fault planes.

indicating that the stress changes induced at convergent margins and collision zones can propagate over large distances through the interiors of plates, to affect passive continental margins and intracratonic basins. Temporal changes in stress levels are not limited to plate collision but also occur through rifting and fragmentation of plates. These are especially important for sedimentary basins in plates not involved in collision or subduction processes.

## Stress-induced vertical motions of the lithosphere

The evolution of sedimentary basins is to a large extent controlled by the response of the underlying lithosphere to various tectonic loads. Lithospheric flexure forms an important element in determining this response. We therefore begin this section with a brief discussion of the flexural response of the lithosphere to intraplate stresses, following classical studies carried out during the first half of the 20th century. A mathematical formulation is given in Box 21.4.

The flexural response of the lithosphere is easily obtained for some simple loading cases. Assuming zero vertical load on the lithosphere, early studies of the problem made a convincing case for neglecting horizontal forces in modelling the vertical motions of the lithosphere. They showed that for reasonable levels of the compressional forces the induced vertical displacements of the lithosphere are negligible. This result, combined with lack of evidence for the existence of such horizontal forces or the occurrence of folding of the entire lithosphere such as is now known from studies of Indian Ocean tectonics led, for a long time, to the withdrawal of attention from this topic. As mentioned earlier, significant progress has been made recently in the study of horizontal stress fields in the lithosphere. Furthermore, it is now realised that vertical motions of the lithosphere at sedimentary basins are primarily the result of a variety of other tectonic processes. These include thermally induced cooling of the lithosphere amplified by the loading of sediments that accumulate in basins, isostatic response to crustal thinning, and flexural bending in response to concentrated vertical loads. The flexural response function is described in Box 21.5. Hence, it is essential to account for the presence of already existing vertical loads on the lithosphere when solving for the response of the lithosphere to horizontal stress fields, as the magnitudes of these stresses are in general far below the stresses that are required to produce folding of the entire lithosphere.

The analytical formulation of specific simple problems shows explicitly how the solution depends on various parameters. Numerical modelling techniques have the advantage of allowing more realistic geometries and variations in parameters to be handled, adding flexibility to the analysis.

## Intraplate stresses and basin stratigraphy

We have seen that temporal fluctuations in stress are a natural consequence of the horizontal motions of the

BOX 21.4

## Mathematical formulation of flexure

The flexural response of a uniform elastic lithosphere at a position $x$ to an applied horizontal force $N$ and a vertical load $q(x)$ is given by:

$$D\frac{d^4w}{dx^4} - N\frac{d^2w}{dx^2} + (\rho_m - \rho_i)gw = q(x)$$

where $w$ is the displacement of the lithosphere, and $D$ is the flexural rigidity ($D = E\,T^3/12(1-\varpi^2)$), with $E$ the Young's modulus, $T$ the plate thickness, and $\varpi$ the Poisson's ratio. The horizontal force $N$ is equivalent to the product of the intraplate stress $\sigma_N$ and the plate thickness $T$. $\rho_m$ and $\rho_i$ are, respectively, the densities of mantle material and the infill of the lithospheric depression, usually water or sediments, and $g$ is the gravitational acceleration.

lithosphere plates. These findings have important consequences for short-term temporal variations in the basin shape throughout geological time. Intraplate stresses, apart from being important in the formation of rifted basins, also play a critical role during their subsequent subsidence history. Intraplate stresses modulate the long-term basin deflection caused by thermal subsidence and induce rapid differential vertical motions of a sign and magnitude that depends on the position within the basin. Figure 21.6 schematically illustrates the relative movement between sea level and lithosphere at the flank of a flexural basin landward of the principal sediment load as predicted by a numerical model. The synthetic stratigraphy at the basin edge is schematically shown for three situations in Figure 21.7. In one, long-term flexure under the basin results from cooling in the absence of an intraplate stress field. Also shown is the same situation with a superimposed transition to 750 bars compression or a superimposed transition to 750 bar tension after 50 Ma. As described in Chapter 15, the thermally induced flexural widening of the basin provides an elegant explanation for long-term phases of coastal onlap. However, by their long-term nature, changes in basin shape by thermal cooling or heating, fail to produce the punctuated character of the stratigraphy of sedimentary basins. As shown by Figure 21.7, intraplate compression causes relative uplift of the basin flank, subsidence at the basin centre, and seaward migration of the shoreline. An offlap develops and an unconformity is produced. Increases in the level of tensional stress induce widening of the basin, lower the flanks and cause landward migration of the shoreline, producing a rapid onlap phase. Stress-induced vertical motions of the crust can also drastically influence sedimentation rates. Flank uplift due to an increased level of compression, for example, can significantly enhance sedimentation rates and modify the infilling pattern, promoting the development of unconformities.

During rifting phases, eventually followed by continental break-up, the tensional stress levels will be reduced. Rifting in the Atlantic, for example, rather than instantaneously, occurred at discrete rifting phases, with stepwise relaxation of tensional stresses. This process might explain sea-level fluctuations that do not correlate with accelerations in plate-spreading rates or increases in ridge lengths. Simultaneously, the correlation of short-term sea-level changes at both sides of the Atlantic may reflect rifting-related accumulation and relaxation of tensional stresses. Break-up unconformities, not successfully explained by most geodynamical models, can be explained by intraplate stresses. Similarly, the break-up of the supercontinent Pangea probably occurred with major changes in the Earth's stress field and associated changes in relative sea level.

The position of coastal onlap reflects the position where rate of subsidence equals rate of sea-level fall. During changes in the intraplate stress field the rate of subsidence is temporarily changed. Consequently, the equilibrium point of coastal onlap is shifted in position. The thermally induced rate of long-term subsidence strongly decreases with the age of the basin. Hence, the production of offlaps during late stages of passive continental margin evolution requires much lower rates of change in apparent sea level than in earlier stages of basin evolution. If these offlaps result from fluctuations in intraplate stresses, the rates of stress change required diminish with age during the post-rift evolution of the basins. This is of particular relevance for an assessment of the relative contributions of tectonics and eustasy to Cenozoic unconformities. For example, Cenozoic unconformities developed at old passive margins in association with

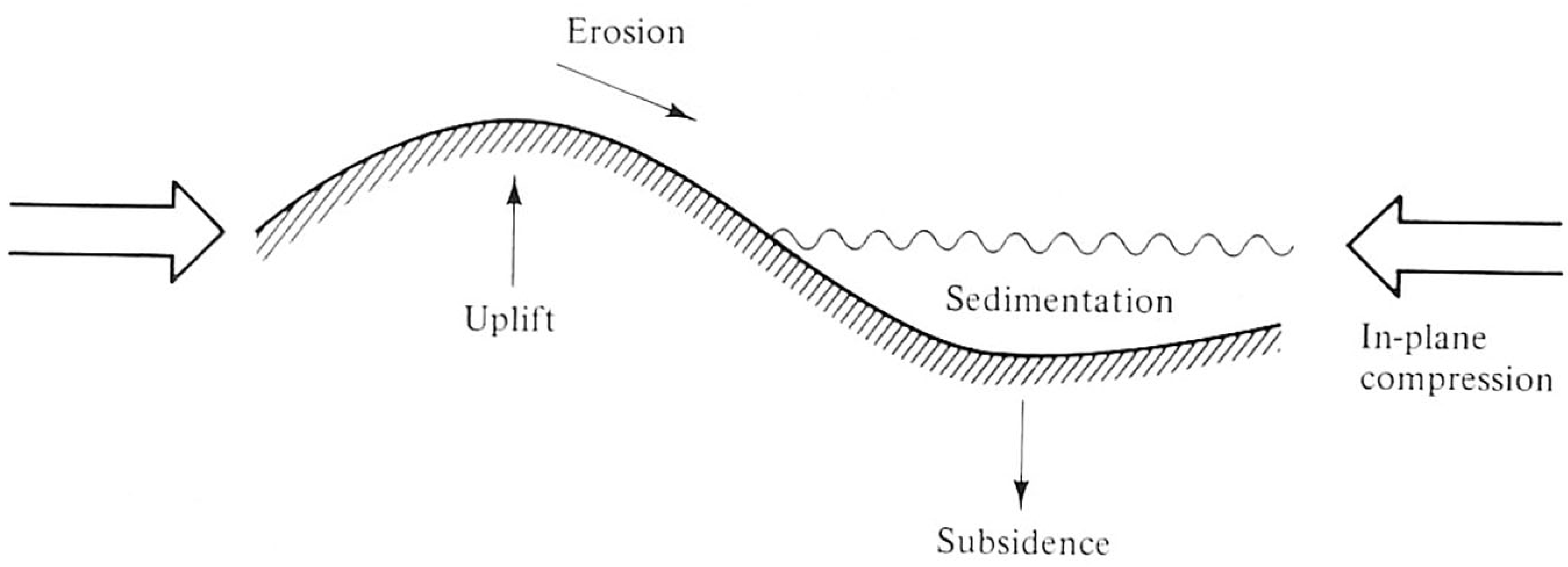



**Figure 21.6** Relative movement induced by intraplate compression at the flank of a sedimentary basin landward of the principal sedimentary load. The stress-induced change in the shape of the basin affects the dynamics of sedimentation and erosion and possibly the oceanic circulation patterns. Changes in basin geometry can trigger erosion and deposition by turbidity currents perhaps in association with canyon cuttings. This could lead to the formation of unconformities traceable from the margin into the deeper part of the basin. □

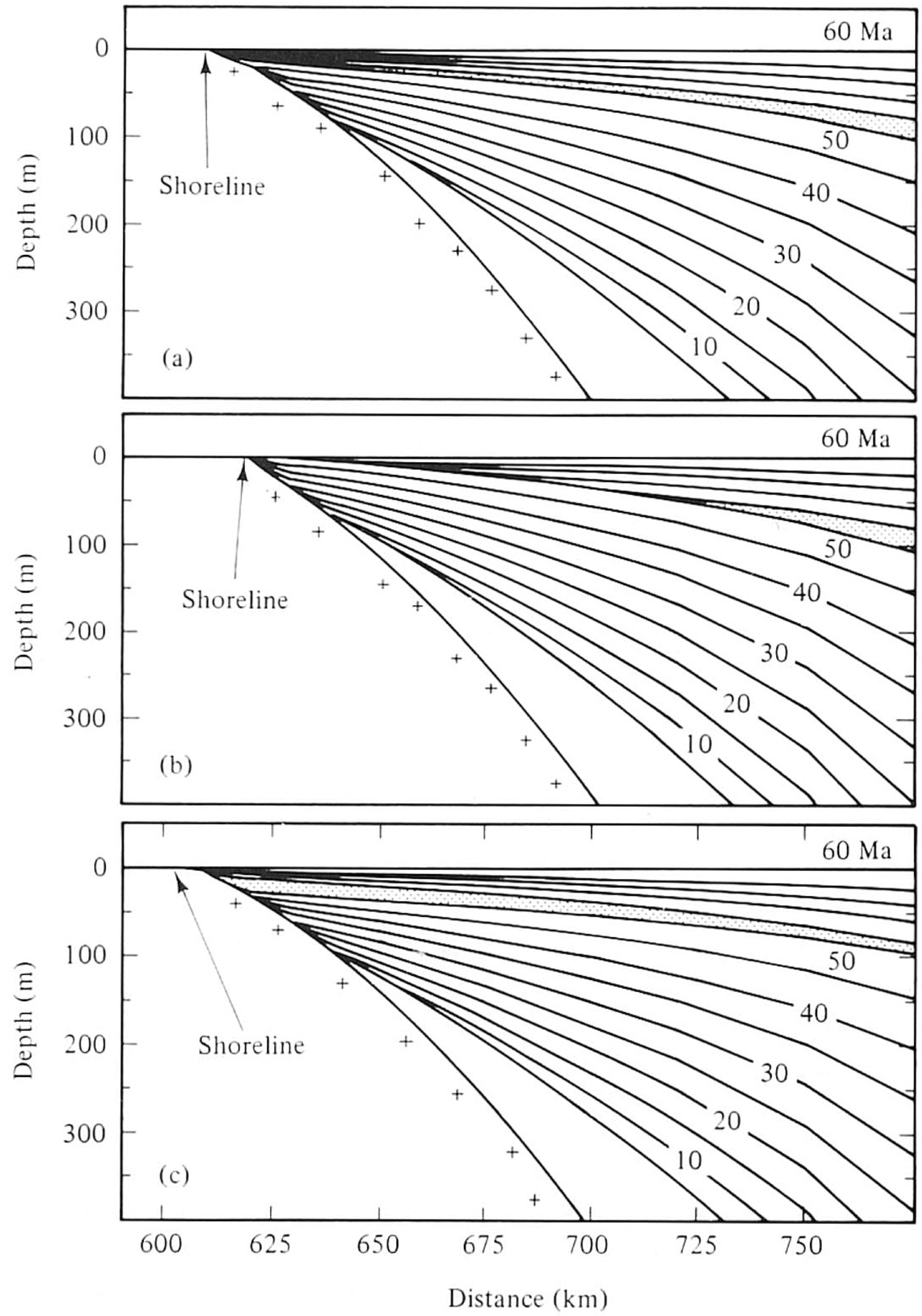
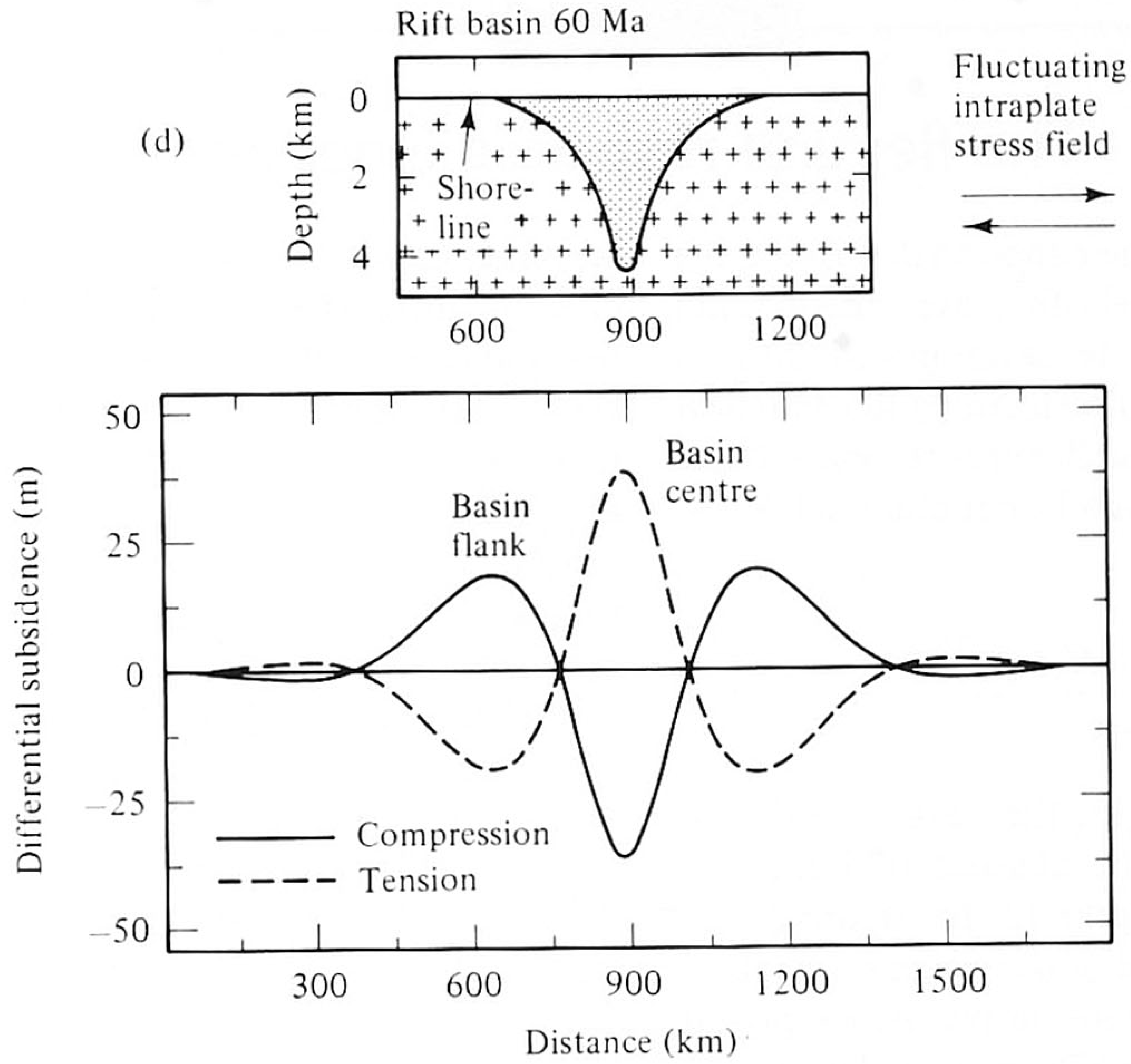



**Figure 21.7** Synthetic stratigraphy for a 60-Ma-old passive margin, which is initiated by lithospheric stretching followed by thermal subsidence and flexural infilling of the resulting depression. Shading indicates the position of the




plane compression at 50 Ma induces uplift of the peripheral bulge, narrowing of the basin and a phase of rapid offlap, which is followed by a long-term phase of gradual onlap due to thermal subsidence. (c) A transition to 750 bar in-plane tension at 50 Ma induces downwarp of the peripheral bulge, widening of the basin and a phase of rapid basement onlap. (d) The differential subsidence or uplift (metres) induced by a change to 1 kbar compression (solid line) and 1 kbar tension (dashed line). □

short-term basin narrowing could be produced by relatively mild changes in intraplate stress levels. Renarrowing and later erosion of Phanerozoic platform basins and passive margins is frequently observed, without clear evidence for active tectonism.

Figure 21.9 demonstrates that the incorporation of intraplate stresses in elastic models of basin evolution can, in principle, predict a succession of alternating rapid onlaps and offlaps observed along the flanks of basins such as the US Atlantic margin. In the figure, a two-layered stretching model for basin initiation is incorporated, as well as the effects of finite and multiple stretching phases and intraplate stresses. Inspection of Figure 21.9 shows the well-known failure of the standard elastic model of basin evolution to predict narrowing of the basin with younger sediments restricted to the basin centre. The narrowing of the basin during its late-stage evolution has often been interpreted as reflecting either the response of the basin to a phase of visco-elastic relaxation or to a long-term eustatic sea-level fall. The stratigraphic model demonstrates that although the incorporation of a long-term change in sea level enhances the Cenozoic narrowing of the basin margin, a long-term post Cretaceous decline in sea level alone cannot cause both the documented basin narrowing and the total thickness of sediments accumulated at this time. Therefore, it could be that much of the observed non-depositional or erosional character of the shelf surface is caused by stress-induced uplift of the basin flank.

Thus, modelling of the stratigraphy of the US Atlantic margin (Figure 21.9) has shown that the punctuated stratigraphy can be successfully simulated by a stress field, whose magnitude fluctuates through time, superimposed on the long-term thermal evolution of the basin. The inferred palaeo-stress is largely consistent with independent data sets on the kinematic and tectonic evolution of the northern/central Atlantic. These show a mainly tensional stress regime during Mesozoic times followed during the Tertiary by a more compressional stress field whose magnitude increases with age. A significant part of the Mesozoic unconformities of rifted basins around the Atlantic and the Arctic might be associated with large-scale, fault-controlled rifting activity, while Cenozoic sequence boundaries in the Arctic could be largely controlled by

BOX 21.5

### The flexural response function

In analytical solutions of the equation describing the flexural behaviour of thin elastic plates, the loading response of the plate can be decomposed into its harmonic components by transforming the equation to the Fourier domain. The flexural response function $\Phi_e(k)$ in the presence of a horizontal load $N$ can be written as:

$$\Phi_e(k) = \left[1 + \frac{D(2\pi k)^4 - N(2\pi k)^2}{\rho_m g}\right]^{-1}.$$

If $N = 0$, then $\Phi_e(k) = \Phi(k)$, the flexural response function of the plate in the absence of intraplate stress, as discussed in Chapter 15. In Figure 21.8a, $\Phi(k)$ is plotted as a function of the wavenumber k of the surface load, which is here simply the reciprocal of the load's wavelength (in kilometres). The curves show the relative flexural response of the elastic plate to a spectrum of surface loads of equal amplitude but different wavelengths. A value of 1 means that the flexural response is maximised (or, equivalently, isostatic compensation is 'local') and a value of 0 means that there is no flexural response (or no isostatic compensation). Figure 21.8b shows the effect of intraplate stress fields of a magnitude of a few hundred MPa on the flexural response of an elastic lithosphere. The plotted curves $\Delta\Phi_e(k) = \Phi_e(k) - \Phi(k)$ (with $\Phi_e(k)$ the response function when an intraplate stress field is applied) show the incremental changes to the flexural response functions in Figure 21.8a, in equivalent units, resulting from the application of various magnitudes of intraplate stresses to the thin elastic plate. Positive increments result from the application of compressional intraplate stresses and mean that the flexural response of the plate to a given surface load is enhanced by the presence of these stresses. The degree of the enhancement is seen to depend on the load wavelength and the flexural rigidity of the plate. The wavenumber k at which the intraplate stresses most affect the flexural response of the lithosphere is almost completely determined by the plate's flexural rigidity. The presence of intraplate stresses has a small but perceptible effect on this wavenumber, but exerts a controlling influence on the amplitude of the response for a given rigidity.

large-scale compressional activity and inversion tectonics. Similarly, simultaneous occurrence of a high frequency of faulting activity in the North Sea area and an increased intensity in the occurrence of sea-level lowerings has been reported in the Upper Jurassic of the North Sea. Intermittent phases of accumulated tensional stress, associated with rifting episodes in the northern and central Atlantic, and subsequent rapid relaxation of these stresses could explain the asymmetry in Vail's onlap–offlap charts. Both the timing and nature of Vail's second-order and third-order cycles might be to a large extent controlled by the plate-tectonic evolution and the associated changes in stress regimes of the northern and central Atlantic.

Although we have concentrated in this chapter on the relationship between tectonics and stratigraphy for rifted basins, the effect of intraplate stresses is of importance to a wider range of sedimentary environments. Another setting where the lithosphere is flexed downward under the influence of sedimentary loads is in foreland basins. Several studies have interpreted the development of unconformities in foreland basins in terms of uplift of the peripheral bulges flanking these basins. The presence of tensional or compressional intraplate stresses, the latter being more natural in this tectonic setting, can amplify or reduce the height of the peripheral bulge by an equivalent amount and thus greatly influence the stratigraphic record.

## Discussion

Vail and coworkers initially interpreted their short-term cycles of sea-level change in terms of waning and waxing of ice sheets. This view was partly based on the inferred global character of the apparent sea-level cycles, partly based on the repetitious character of the cycles, and was partly due to the lack of a tectonic mechanism to explain both the rate and magnitude of the third-order cycles. Crucial in this respect has also been the basic assumption that tectonic subsidence should be slow, requiring absolute changes in sea level to explain irregularities in the subsidence record. The issue of global synchroneity has attracted major debate. Several authors have noted that Vail's cycles, although based on data from different basins around the world, are heavily weighted

### 21.5 continued

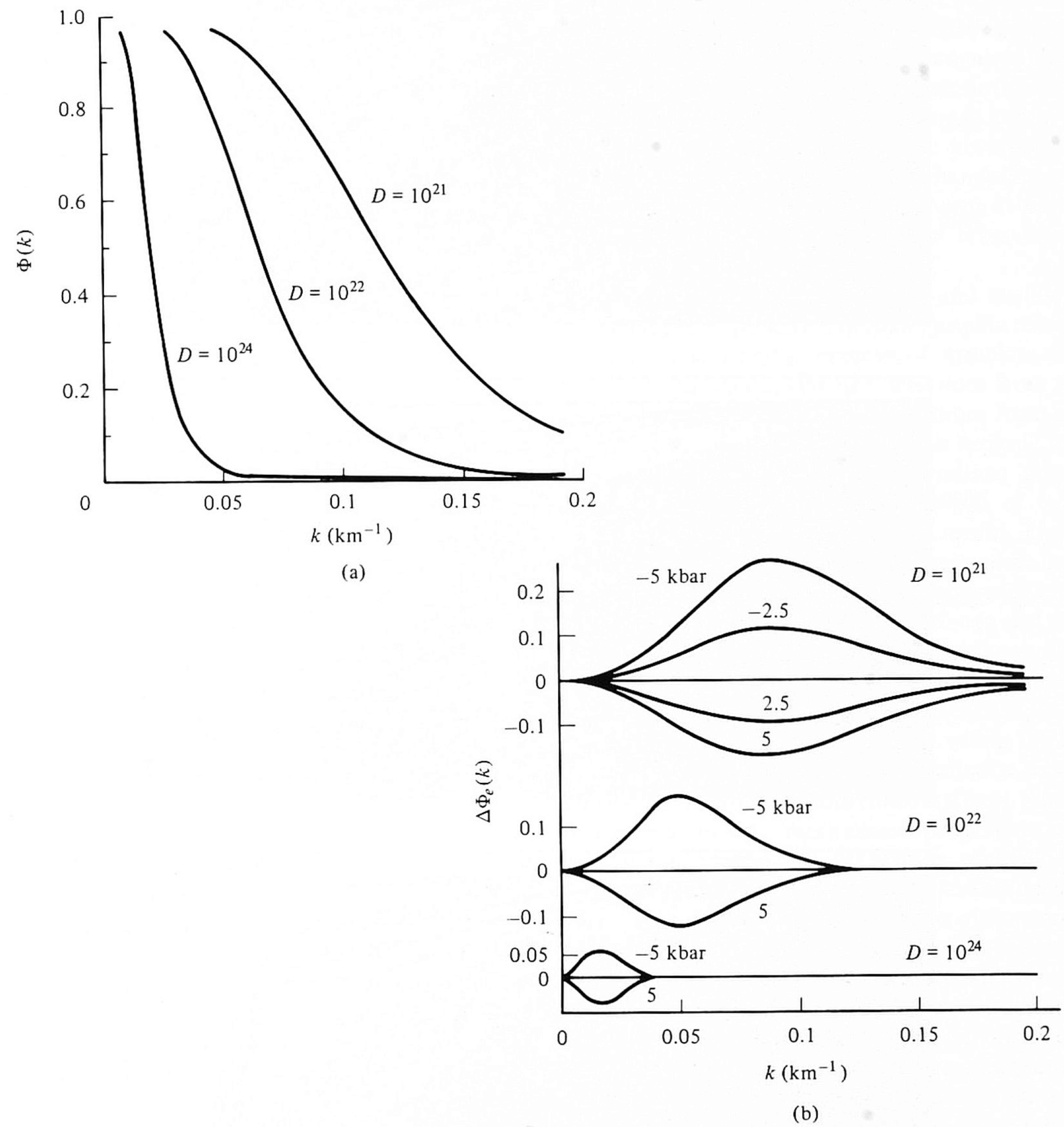



**Figure 21.8** (a) The flexural response $\Phi(k)$ of an elastic thin plate, with various flexural rigidities $D$ (in units of newton metres), to a surface load when there are no intraplate stresses. $\Phi(k)$ is plotted as a function of the wavenumber $k$ of the surface load, which is here simply the reciprocal of the load's wavelength (in km). Each curve shows the relative flexural response of the elastic plate to a spectrum of surface loads of equal amplitude but different wavelengths. A value of 1 means that the flexural response is maximised (or, equivalently, isostatic compensation is 'local') and a value of 0 means that there is no flexural response (or no isostatic compensation). (b) The effect of intraplate stresses $\sigma_N$ (in units of kbars, 1 kbar = 100 MPa; tension is positive) on the flexural response functions shown in (a). The plotted curves $\Delta\Phi_e(k) = \Phi_e(k) - \Phi(k)$ (with $\Phi_e(k)$ the response function for an applied intraplate stress field) show the incremental changes to the flexural response functions in (a), in equivalent units, resulting from the application of various magnitudes of intraplate stresses to the thin elastic plate. Positive increments result from the application of compressional intraplate stresses and mean that the flexural response of the plate to a given surface load is enhanced by the presence of these stresses. The degree of the enhancement is seen to depend on the load wavelength and the flexural rigidity of the plate. □

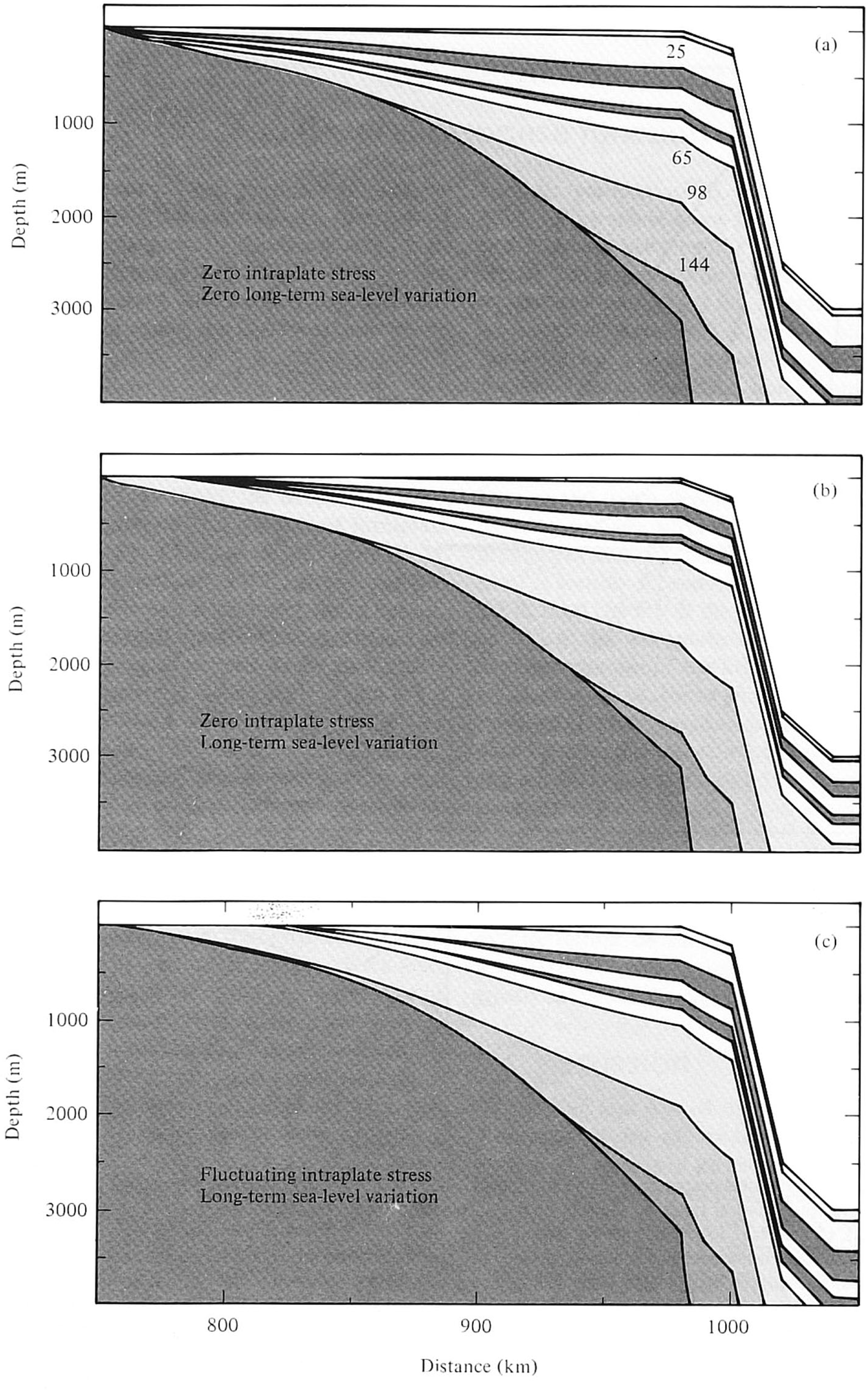



***Figure 21.9*** Schematic model of the stratigraphy of the US Atlantic margin at Cape Hateras. The numbers in the strata refer to stratigraphic ages in Ma. (a, b) Modelled stratigraphy for elastic rheology of the lithosphere in the absence of intraplate stresses. (c) Modelled stratigraphy showing the effect of incorporating a fluctuating intraplate stress field in the analysis. □

in favour of the North Sea and the northern/central Atlantic margins. The issue of global synchroneity is important as it obviously strongly influences present discussions on the causes of short-term changes in sea level. Tectonic mechanisms such as variations in spreading rates, hot spot activity and orogeny fail to produce changes at the rate of third-order cycles. This is because such explanations are derivatives of the thermal evolution of the lithosphere and are therefore associated with a long thermal inertia of several tens of millions of years (Table 21.2). Changes in water volume by the waning and waxing of ice sheets (**glacio-eustasy**) easily can induce both the rate and magnitude of the inferred sea-level changes but raises two basic problems. The first problem is the occurrence of third-order sea-level cycles during time intervals where there is no geological evidence in support of low-altitude glaciation. This presents a fundamental problem of explaining sea-level changes by this mechanism at times prior to the Late Cenozoic. The second problem is the inability of glacio-eustasy to cause uniform global lowerings and rises of sea level. Studies of post-glacial rebound have shown that the concept of eustasy is not valid in the case of sea-level changes resulting from substantial lateral transfer of mass, as between ice sheets and oceans. The sea level coincides with an equipotential surface and is partly controlled by the gravitational attraction of the ice sheets. For this reason, the response of sea level to the melting of an ice sheet varies strongly between sites near the ice sheets and sites farther away. Furthermore, the response of the crust to the removal of the ice load and to the loading by the meltwater causes differential motion between sea level and land. As a result, the sign and magnitude of the induced sea-level change is dependent on the distance to the location of the ice cap. This is quite important for scenarios in which glacio-eustasy is the key mechanism to explain global synchronous changes of uniform magnitude in sea level.

We have shown in the previous section that short-term changes in relative sea level can equally well be caused by rapid, stress-induced vertical motions of the lithosphere. Hence intraplate stresses, apart from being important in the formation of rifted basins, probably also play a critical role during their subsequent subsidence history. Undoubtedly, both eustasy and tectonics have contributed to the record of short-term changes in sea level. The relative contributions are, by their nature, of variable magnitude. The key question to be answered from stratigraphic analysis is related to the spatial and temporal differences in the expressions of tectonic processes and eustasy. The development of stratigraphic criteria to differentiate between tectonics and eustasy is therefore vital, and is needed both on an interbasinal and an intrabasinal scale. Recently, a number of features in the stratigraphic record of rifted basins have been recognised that are difficult to explain in terms of the eustatic framework. Among these are sediment source areas and/or sedimentary regimes that often change abruptly across sequence boundaries, intrabasinal changes in subsidence and uplift patterns that also occur at sequence boundaries, and faults that do not cross sequence boundaries.

The discrimination of tectonics and eustasy is a subtle matter, especially if biostratigraphic resolution is limited. The regional character of intraplate stresses can shed light on documented deviations from global sea-level charts. Whereas such deviations from global patterns are a natural feature of this tectonic model, the occurrence of short-term deviations does not preclude the presence of global events of tectonic origin elsewhere in the stratigraphic record. These are to be expected when major plate reorganisations in intraplate stress fields occur simultaneously in more than one plate or in time intervals prior to and shortly after the break-up of Pangea where continents and rift basins were in a largely uniform stress regime. The magnitude of the stress-induced phases of rapid uplift and subsidence varies with position within the basin, thus providing another important criterion to distinguish this contribution from eustatic effects. Similarly, differences in the mechanical properties of the lithosphere control the magnitude of the vertical motions. The presence of weak, attenuated continental lithosphere enhances the effectiveness of the stresses to cause substantial vertical motions and can explain differences in magnitude of the apparent sea-level changes such as those observed between the Tertiary North Sea and the Gippsland Basin off southeast Australia. Only the larger short-term fluctuations in sea level, with magnitudes in excess of 50 metres, require stress changes large enough to be related to major plate boundary reorganisations. This observation could explain the frequently observed correlation between the timing of plate reorganisations and rapid lowerings in the sea level. Furthermore, glacio-eustatic events sometimes occur simultaneously with a major tectonic reorganisation which could, for example, have contributed to the large magnitude of the Oligocene sea-level lowering. It appears that the repetitious character of sea-level cycles and plate-wide correlation, which are often considered to be diagnostic for a eustatic cause of the sea-level change, can equally well be explained by episodic accumulation and re-

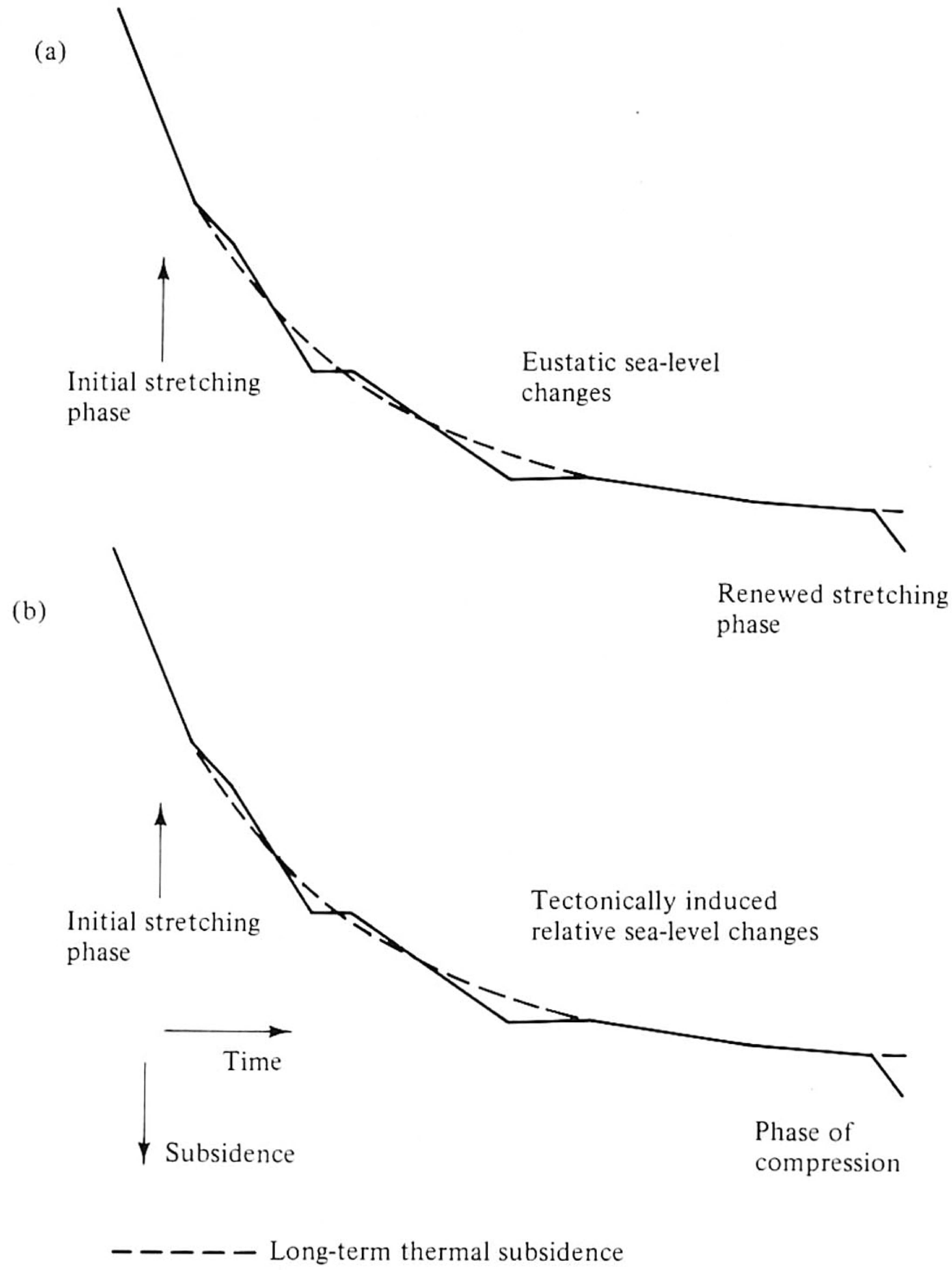



***Figure 21.10*** Two approaches to the analysis of tectonic subsidence. (a) Traditional interpretation: short-term deviations from long-term trends in basin subsidence are attributed to eustatic sea-level changes and renewed phases of crustal stretching. (b) Alternative interpretation: short-term deviations from long-term (thermal) subsidence are interpreted in terms of stress-induced vertical motions of the lithosphere. □

laxation of stress in the lithosphere. Similarly, as stress-induced basement subsidence and uplift phases can be of a rate and magnitude corresponding to Vail's third-order cycles (Figure 21.10), care must be taken when interpreting the record of sea-level change solely in terms of waxing and waning of ice sheets.

The stretching model of basin formation predicts a rapid phase of initial subsidence followed by long-term subsidence associated with cooling of the lithosphere. Lithospheric stretching occurs due to passive rifting of the lithosphere, after which stresses relax. An essential assumption made in this model is, therefore, that stresses are zero after basin formation. The original stretching formulation was a strictly kinematic one. More recently, work on the dynamics of stretching has demonstrated that tensional stresses of the order of several hundred MPa are required to form a sedimentary basin by this mechanism. Important in this respect have been the recent theoretical advances in lithospheric rheology based on extrapolation of rock mechanics data.

Future work has to address more fully the dynamic element of lithosphere deformation. Furthermore, present quantitative models of the origin of basins are incapable of solving problems related to subsequent structural development that may be intrinsically coupled with basin formation. For example, late-stage compression during the post-rift evolution of extensional basins

Table 21.2. *Magnitudes and rates of sea-level changes by several mechanisms that were considered in the early 1980s as potential contributors to short-term changes in relative sea level (Vail et al's third-order cycles). Apart from glacio-eustasy, the proposed (tectonic) models failed to produce both the rate (1–10 cm/1000 yrs) and the magnitude (up to the order of a 100 metres). This is primarily because of the long time constants of lithospheric thermal processes which grossly exceed the time scales (2–5 Ma) characteristic for short-term changes in relative sea level.*

| | Probable maximum | | Maximum maximum | | |
|---|---|---|---|---|---|
| Mechanism | Magnitude (m) | Rate (cm/1000 yr) | Magnitude (m) | Rate (cm/1000 yr) | Time interval (Ma) |
| Glaciation | 150 | 1000 | 250 | 1000 | 0.1 |
| Ridge volume | 350 | 0.75 | 500 | 1.2 | 70 |
| Orogeny | 70 | 0.10 | 150 | 0.20 | 70 |
| Sediment | 60 | 0.11 | 85 | 0.25 | 70 |
| Hot spots | 50 | 0.08 | 100 | 0.14 | 70 |
| Flooding of ocean basins | | | | instantaneous | |

can largely explain current discrepancies between estimates of crustal thinning derived from structural analysis (obtained by measuring horizontal displacements of the normal faults active during extension) and subsidence data (where the total tectonic subsidence is used as a measure of crustal thinning). Rapid phases of basin subsidence after the initial event of basin formation and with a magnitude too large to attribute to changes in sea level are usually explained in terms of multiple stretching phases (Figure 21.10). However, care should be taken with this interpretation as an increase in the level of intraplate compression can equally well produce this type of deviation from the thermal model predictions of basin subsidence. Phases of lithospheric compression during the post-rift evolution of rifted basins can give rise to substantial deepening of the basin centre, accompanied by uplift at basin flanks, promoting the development of steershead geometries of sedimentary basins. The effect of such late-stage compressional phases is enhanced by brittle–ductile rheologies of the lithosphere, in particular where stress levels approach the lithospheric strength. Late-stage compression could, for example, explain the rapid phases of Late Neogene subsidence such as that encountered around the northern Atlantic. Therefore, current backstripping techniques, correcting only for vertical loading of the lithosphere, tend to overestimate values of crustal extension and, therefore, might result in overestimates of temperatures at depths corresponding to the hydrocarbon window.

## Conclusions

Sedimentary basins form and evolve in the interiors of the plates that are subject to episodic changes in tectonic regime. Recent work on intraplate tectonics has established that there is a strong mechanical coupling between geodynamic processes at plate boundaries and deformations in the plate interiors. In fact, the record of vertical motions in sedimentary basins holds the potential for unravelling the full complexity of the interplay between the processes of basin dynamics and basin fill. Careful analysis is required to separate effects of eustatic sea-level changes from stress-induced short-term motions of the lithosphere, as both mechanisms produce rather similar short-term distortions from long-term patterns of thermal subsidence. Plate tectonics can operate on a global scale (plate reorganisations) and on a more regional scale, which is important for the discussion on global synchroneity of apparent sea-level change. Integrated studies of the structural and stratigraphic evolution of sedimentary basins, together with further development of dynamic models for basin formation and evolution, will contribute to the success of basin analysis.

## Further reading

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*Authoritative and extremely well-documented collection of papers on the stratigraphy and basin evolution of extensional basins in the northern Atlantic. Very useful for the wealth of high-quality seismic sections of sedimentary basins.*

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*A unique and very-well-illustrated overview of the tectonic evolution of the Arctic/Northern Atlantic and western Tethys based on a compilation of industry and non-industry data.*

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