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published in
New perspectives in Basin Analysis
1988

document version
Publisher's PDF, also known as Version of record

10 Intraplate Stresses: A New Element in Basin Analysis

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Abstract

Evidence reviewed in this paper indicates that intraplate stresses in the lithosphere are of substantial magnitude. Numerical modelling and observation of modern and paleo-stress fields demonstrate the existence of stress provinces of great areal extent in the interiors of the plates. The interaction of intraplate stresses with basin subsidence provides a new element in basin analysis. Fluctuations in intraplate stress fields influence basin stratigraphy and provide a tectonic explanation for short-term, relative sea-level variations inferred from the sedimentary record. Modelling shows that the incorporation of intraplate stresses in models of basin evolution can predict a succession of onlap and offlap patterns similar to those observed at basin flanks. Such a stratigraphy can be interpreted as the natural consequence of short-term changes in basin shape by moderate fluctuations in intraplate stresses, superimposed on long-term broadening of the basin with cooling since its formation. Basin stratigraphy could provide a new source of information for paleo-stress fields.

Introduction

During the last decade considerable progress has been made in the study of the stress field within lithospheric plates. Detailed analysis of earthquake focal mechanisms (e.g., Ahorner 1975; Bergman 1986), in-situ stress measurements (e.g., McGarr and Gay 1978; Illies *et al.* 1981), and analysis of break-out wells drilled for commercial purposes (Bell and Gough 1979; Blumling *et al.* 1983; Zoback *et al.* 1985) have demonstrated the existence of consistently oriented stress patterns in the lithosphere. Simultaneously, numerical modelling (Richardson *et al.* 1979; Wortel and Cloetingh 1981, 1983; Cloetingh and Wortel 1985, 1986) has resulted in better understanding of the causes of the observed variations in stress level and stress directions in the various lithospheric plates. Such studies have shown a causal relationship between the processes at plate boundaries and the deformation in the plates' interiors (e.g., Johnson and Bally 1986).

In models of the evolution of sedimentary basins located in the interiors of the plates (e.g., Beaumont 1978; Watts *et al.* 1982), however, the role of intraplate stresses has been largely ignored. These current geophysical models are not yet able to explain much of the observed evolution of the tectonic component of basin subsidence. Recently, the first steps have been taken towards exploring the consequences of the existence of intraplate stress fields for models of the formation and evolution of intracratonic basins (Lambeck 1983; de Rito *et al.* 1983). Intraplate stresses have also been demonstrated to be an important element in basin stratigraphy. Work by Cloetingh *et al.* (1985) and Cloetingh (1986) has shown that temporal fluctuations of the intraplate stress could provide a tectonic explanation for short-term apparent sea-level changes inferred from seismic stratigraphic analysis of the sedimentary record (Vail *et al.* 1977, 1984; Haq *et al.* 1987).

The present paper reviews evidence for the existence of intraplate stress fields in the lithosphere. This is followed by a discussion on some implications of intraplate stress for quantitative modelling of the evolution of sedimentary basins. Finally, the potential use of the sedimentary record to extract information on paleo-stress fields in the plates is explored.

Intraplate Stress Fields

The present stress field in the various plates has been studied in great detail by the application of a wide range of observational techniques. The results of a recent compilation of stress-direction data for the northwestern European Platform by Klein and Barr (1986) are displayed in Figure 10.1. The observed modern stress orientations show a remarkably consistent pattern, especially considering the heterogeneity in lithospheric structure in this area. These stress-orientation data indicate a propagation of stresses away from the Alpine collision front over large distances in the platform region. Observations of stress orientation in different continental and oceanic regions have demonstrated the existence of such large stress provinces with preferred stress directions to be a general characteristic of lithospheric plates (e.g., Bell and Gough 1979; Zoback and Zoback 1980; Lambeck *et al.* 1984; Bergman and Solomon 1985). These observations indicate that regional stress fields are dominated by the effect of plate-tectonic forces acting on the lithosphere.

For the analysis of paleo-stress fields, however, methods used to study the modern stress field cannot be applied. In this case, the information is derived from analysis of the geological record, such as stylolites, microstructures, or fault-orientation data (e.g., Letouzey 1986). As such, the inferred information on paleo-stress fields is less precise than the results of studies of modern stress indicators. The study of paleo-stress fields, however, adds geological time as a parameter crucial to understanding the temporal fluctuations of stress fields in the plates.

In this paper I concentrate on the regional stresses in the lithosphere induced by plate-tectonic forces. Other sources of stress, however, might dominate on a more local scale. Examples are stresses associated with topographic anomalies and crustal-thickness inhomogeneities at passive margins. Further, temperature variations may induce thermal stresses in cooling lithosphere, while membrane stresses are possibly induced by plate motions on a nonspherical earth. A number of these stress mechanisms have been reviewed by Turcotte and Oxburgh (1976). Finally, flexural stresses are generated due to vertical loads on the lithosphere, in particular by sedimentary sequences at passive margins (Cloetingh *et al.* 1982). Of the various locally induced stress sources the flexural stresses and thermal stresses stand out in magnitude (up to order of kbars), while most of the other mechanisms produce stresses with a characteristic level of the order of a few hundred bars (Turcotte and Oxburgh 1976; Cloetingh *et al.* 1982).

A complementary approach to collecting stress-indicator data is the study of the intraplate stress field using numerical modelling 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 (Solomon *et al.* 1975; Richardson *et al.* 1979). These and several other studies (Forsyth and Uyeda 1975; Chapple and Tullis 1977) resulted in the overall understanding that ridge push, which results from the elevation of the spreading ridge above the adjacent ocean floor and the thickening of the lithosphere with cooling, and slab pull, which acts on the downgoing slab in a subduction zone, are the two main driving forces. Since then, deeper understanding has been obtained of the age dependence of the forces acting on the lithosphere (Lister 1975; England and Wortel 1980). This development has benefitted from advances in the analysis of the subduction process (Vlaar and Wortel 1976; Wortel 1984). By implementing these new features and insights in stress modelling, Wortel and Cloetingh (1981, 1983) and Cloetingh and Wortel (1985, 1986) showed that the dynamic basis of their numerical modelling procedure enabled the resulting intraplate stress field to be used to analyze, explain, and even predict various deformational processes within the 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 the oceanic plates or lithospheric plates with major oceanic parts. However, the comparison between calculated stress fields in various primarily oceanic plates sheds new light on observations of modern and ancient stress fields in oceanic and continental lithosphere. Modelling of the stress field in the Indo-Australian Plate (Cloetingh and Wortel 1985, 1986) has shown that the joint occurrence in this single plate of an exceptionally high level of compressive deformation in the plate interior (McAdoo and Sandwell 1985; Bergman and Solomon 1985; Lambeck *et al.* 1984) and near-ridge parallel extensional deformation (Wiens and Stein 1984; Stein *et al.* 1987) is a transient feature unique to the present

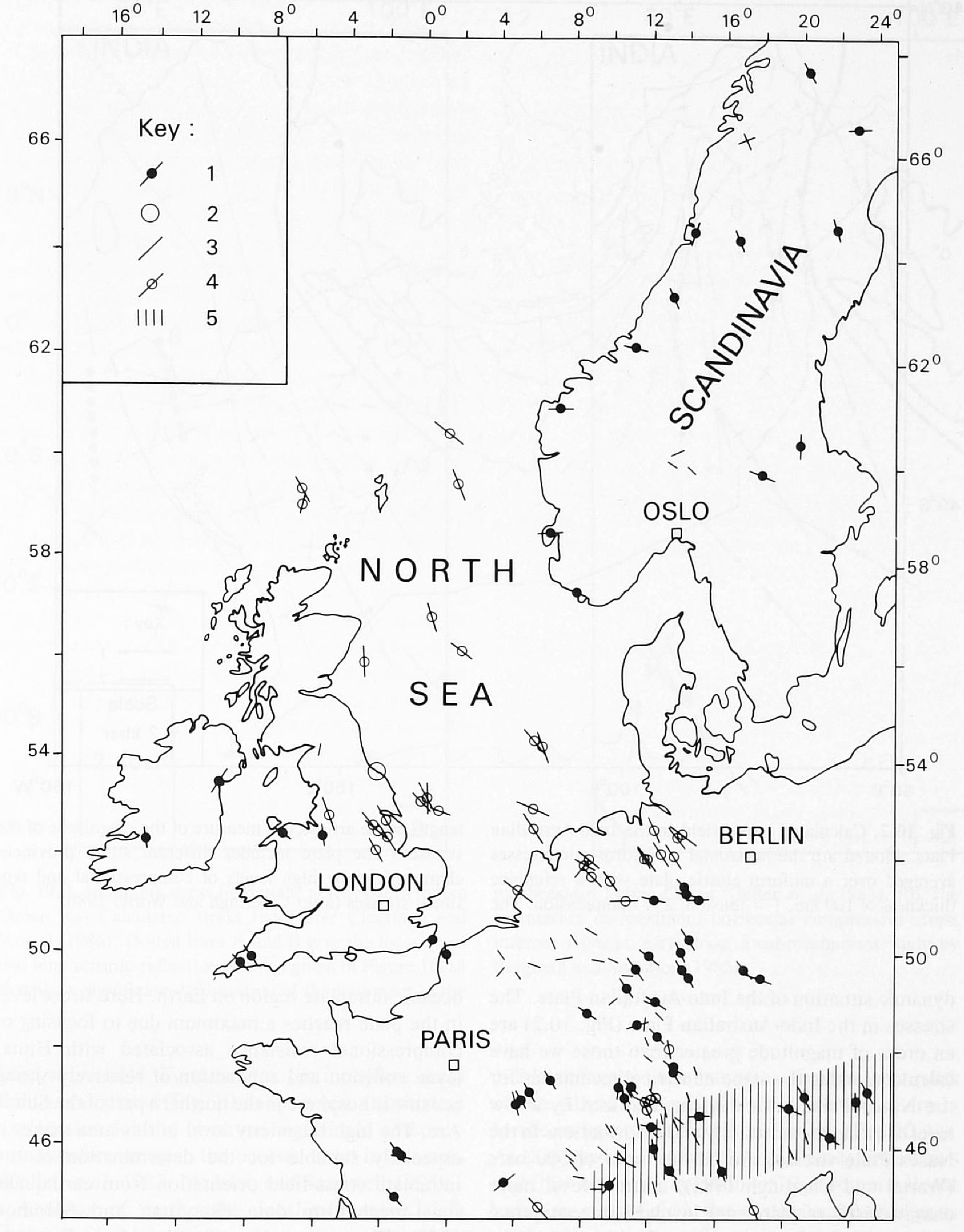


Fig. 10.1. Compilation of observed maximum horizontal present stress directions in the northwestern European Platform. 1 = the direction of maximum horizontal stress from *in-situ* measurements, 2 = a horizontal stress equal in all directions as found from *in-situ* stress measurements, 3 = the direction of maximum horizontal stress inferred from earthquake focal-mechanism studies, 4 = the direction of maximum horizontal stress inferred from break-out analysis, 5 = Alpine fold belt. The data indicate stress propagation away from the Alpine fold belt in the platform region (after Klein and Barr 1986).

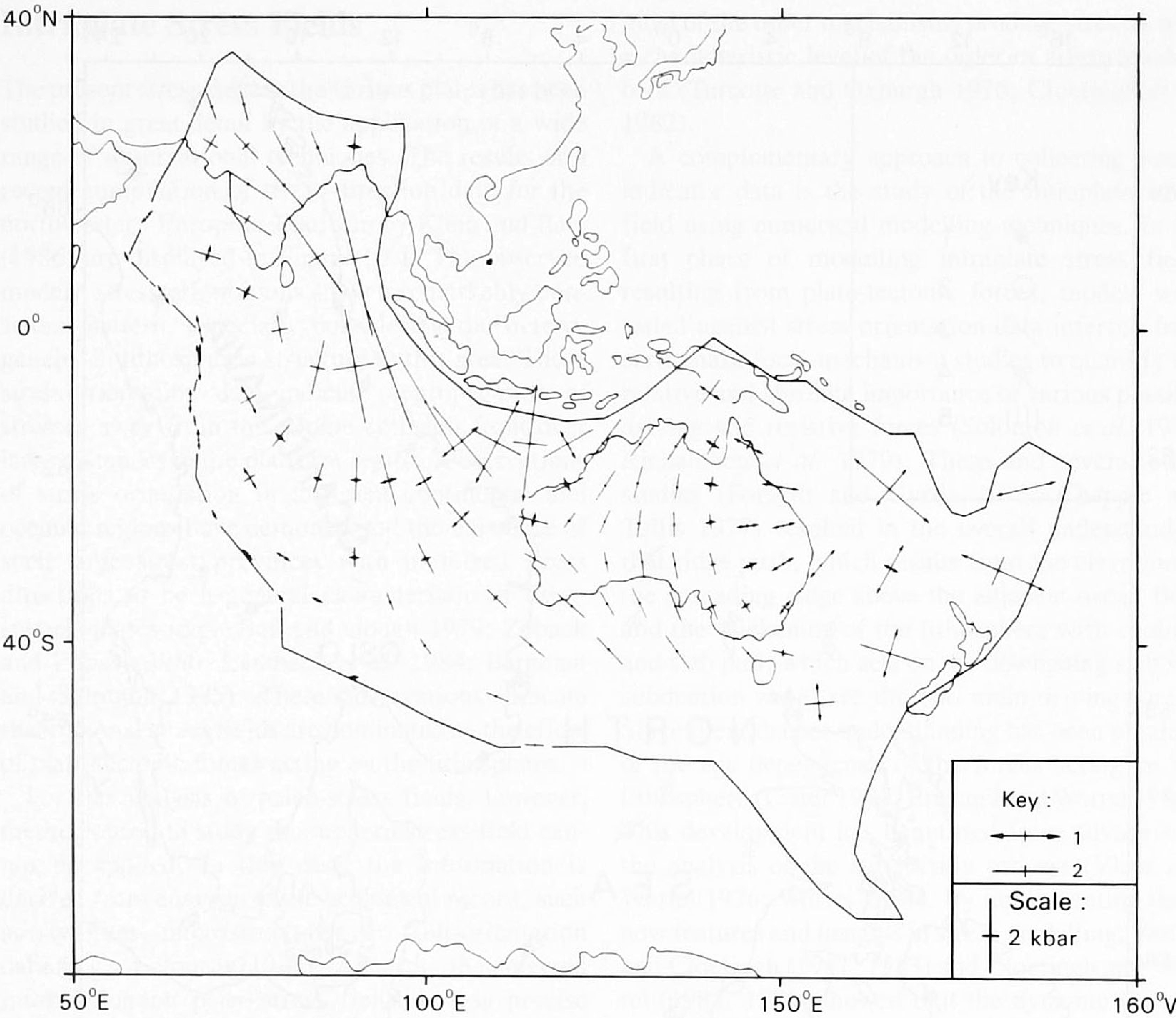


Fig. 10.2. Calculated stress field in the Indo-Australian Plate. Plotted are the horizontal nonhydrostatic stresses averaged over a uniform elastic plate with a reference thickness of 100 km. 1 = tension, 2 = compression. The length of the arrows is a measure of the magnitude of the stresses. The plate includes different stress provinces characterized by high levels of compressional and tensional stresses (after Cloetingh and Wortel 1986).

dynamic situation of the Indo-Australian Plate. The stresses in the Indo-Australian Plate (Fig. 10.2) are an order of magnitude greater than those we have calculated using the same numerical techniques for the Nazca Plate, which is characterized by a low level of intraplate seismicity and deformation. In the Nazca Plate stresses are on the order of 500 bars (Wortel and Cloetingh 1983), a stress level more characteristic of plates not involved in continental collision or rifting processes. The high level of intraplate deformation in the Indo-Australian Plate is manifested as an exceptionally high level of intraplate seismicity. This is particularly the case in the northeastern Indian Ocean sector of the plate, which is at present the most seismically active oceanic intraplate region on Earth. Here stress level in the plate reaches a maximum due to focusing of compressional resistance associated with Himalayan collision and subduction of relatively young oceanic lithosphere in the northern part of the Sunda Arc. The high seismicity level in this area makes it especially suitable for the determination of the intraplate stress-field orientation from earthquake focal-mechanism data (Bergman and Solomon 1985). The stress-orientation data from Bergman and Solomon (1985) given in Figure 10.3b 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 region, in agreement with the calcu-

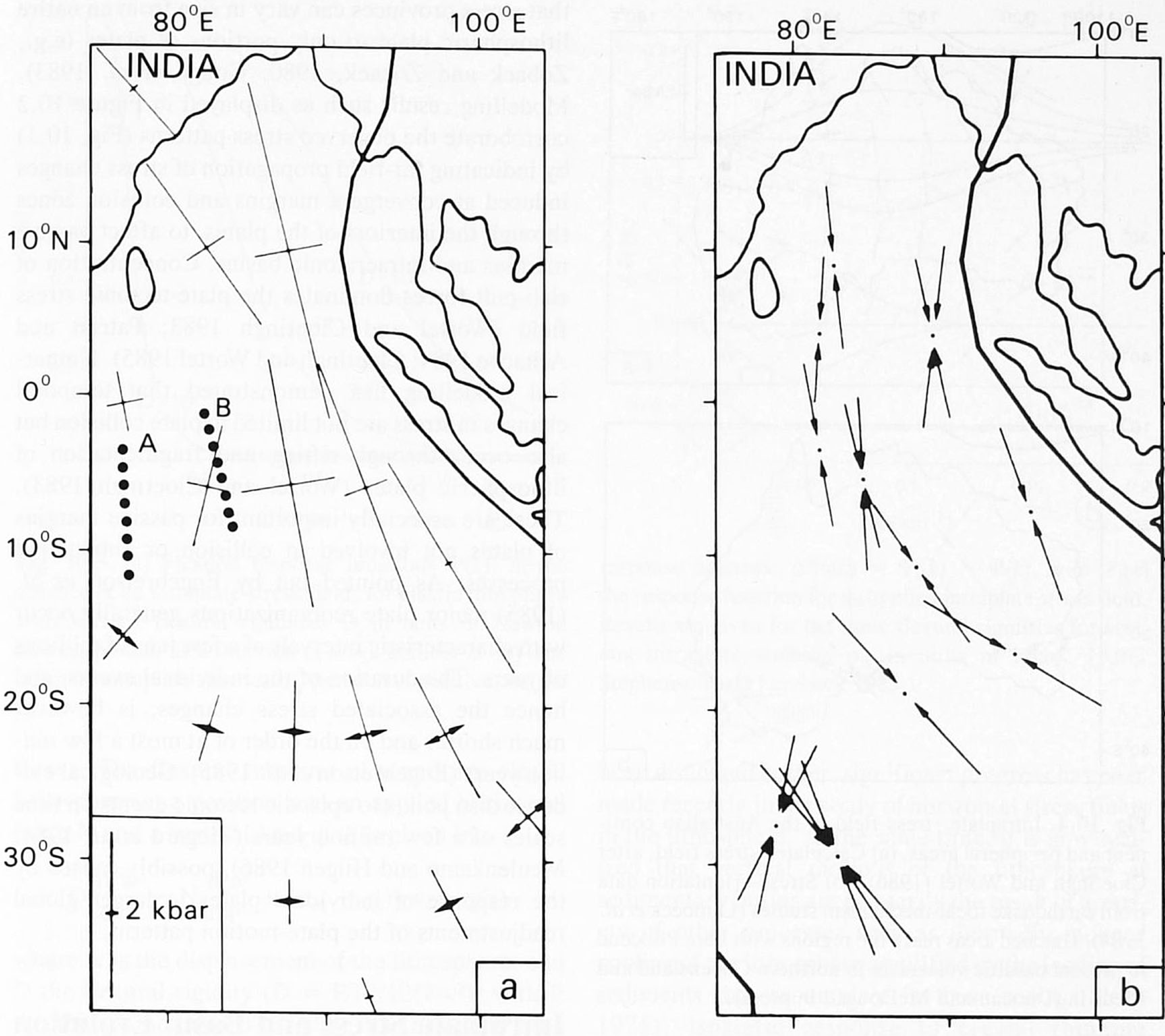


Fig. 10.3. Intraplate stress field in the northeastern Indian Ocean. (a) Calculated stress field after Cloetingh and Wortel (1986). Dotted lines A and B give the location of two long seismic-reflection profiles given in Figure 10.14 that show significant deformation of the basement in the northeastern Indian Ocean (Geller *et al.* 1983). (b) The orientation of maximum horizontal compressive stress inferred from an earthquake focal-mechanism study by Bergman and Solomon (1985).

lated stress field. Furthermore, the intraplate stress field as calculated (Fig. 10.3a) provides a consistent explanation for the observed significant compressional deformation in the oceanic crust in this area (Geller *et al.* 1983; McAdoo and Sandwell 1985).

Analysis of earthquake focal-mechanism data shows that large parts of the Australian continent are also in a state of significant horizontal compression (Lambeck *et al.* 1984). On the basis of observational evidence and modelling of gravity and topography, Lambeck *et al.* (1984) and Stephenson and Lambeck (1985) have suggested a magnitude of the order of 1–2 kbar for the intraplate stress field in large parts of the Australian continent (Fig. 10.4b), a stress level consistent with model predictions (Fig. 10.4a). As displayed in Figure 10.4, the modelling shows that the observed rotation of the compressional intraplate stress field in the western and central parts of the Australian continent is mainly the consequence of its geographic position relative to surrounding plate boundaries. Eastern Australia, an area characterized by recent volcanic activity, probably forms a separate stress province (Fig. 10.4a) with an intraplate tensional stress regime (Duncan

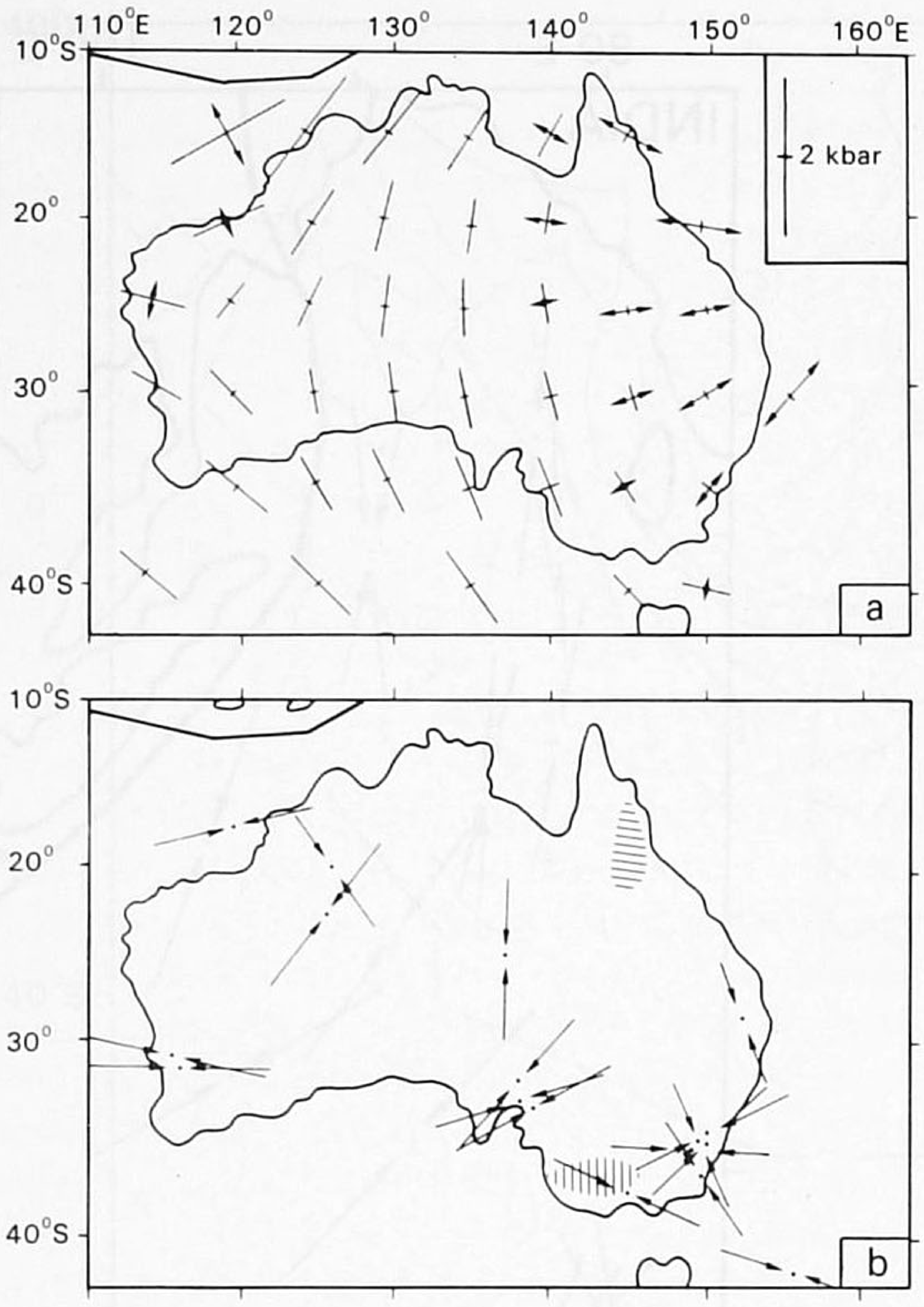


Fig. 10.4. Intraplate stress field in the Australian continent and peripheral areas. (a) Calculated stress field, after Cloetingh and Wortel (1986). (b) Stress-orientation data from earthquake focal-mechanism studies (Lambeck *et al.* 1984). Hatched areas mark the regions with Late Pliocene to present basaltic volcanism in northern Queensland and Victoria (Duncan and McDougall in press).

and McDougall in press), associated with transmission of tensional stresses from the adjacent plate boundaries eastward from the continent. Similarly, the stress province adjacent to the Southeast and Central India Ridges, which exhibits an exceptionally high level of intraplate tensional seismicity (Fig. 10.2), is causally related to far-field stress propagation of tensional stresses induced by the subduction of old oceanic lithosphere at the Java segment of the Sunda Arc (Stein *et al.* 1987).

These numerical models have shown that high-magnitude stresses can be concentrated in the plates' interiors. They have also shown, in agreement with observations (Fig. 10.1), that long-wavelength spatial variations in the stress field may occur, although such features do not necessarily exist in all plates. These models and observations make clear that stress provinces can vary in size from an entire lithospheric plate to only portions of plates (e.g., Zoback and Zoback, 1980; Gough *et al.* 1983). Modelling results such as displayed in Figure 10.2 corroborate the observed stress patterns (Fig. 10.1) by indicating far-field propagation of stress changes induced at convergent margins and collision zones through the interiors of the plates, to affect passive margins and intracratonic basins. Concentration of slab-pull forces dominates the plate-tectonic stress field (Wortel and Cloetingh 1983; Patriat and Achache 1984; Cloetingh and Wortel 1985). Numerical modelling has demonstrated that temporal changes in stress are not limited to plate collision but also occur through rifting and fragmentation of lithospheric plates (Wortel and Cloetingh 1983). These are especially important for passive margins of plates not involved in collision or subduction processes. As pointed out by Engebretson *et al.* (1985) major plate reorganizations generally occur with characteristic intervals of a few tens of millions of years. The duration of the individual events, and hence the associated stress changes, is however much shorter and on the order of at most a few million years (Engebretson *et al.* 1985). Geological evidence also points to episodic tectonic events on time scales of a few million years (Megard *et al.* 1984; Meulenkamp and Hilgen 1986), possibly caused by the response of individual plates to larger global readjustments of the plate-motion patterns.

Intraplate Stress and Basin Evolution

The evolution of sedimentary basins is in large part controlled by the response of the underlying lithosphere to the various tectonic loads. Lithospheric flexure forms an important element in determining this response (Beaumont 1981; Watts *et al.* 1982). We therefore begin this section with a brief discussion of the flexural response of the lithosphere to intraplate stresses.

The Effect of Intraplate Stress on the Deflection of the Lithosphere

In classical studies, Smoluchowski (1909), Vening Meinesz (see Heiskanen and Vening Meinesz 1958), and Gunn (1944) have investigated the flexural response of the lithosphere to applied horizontal

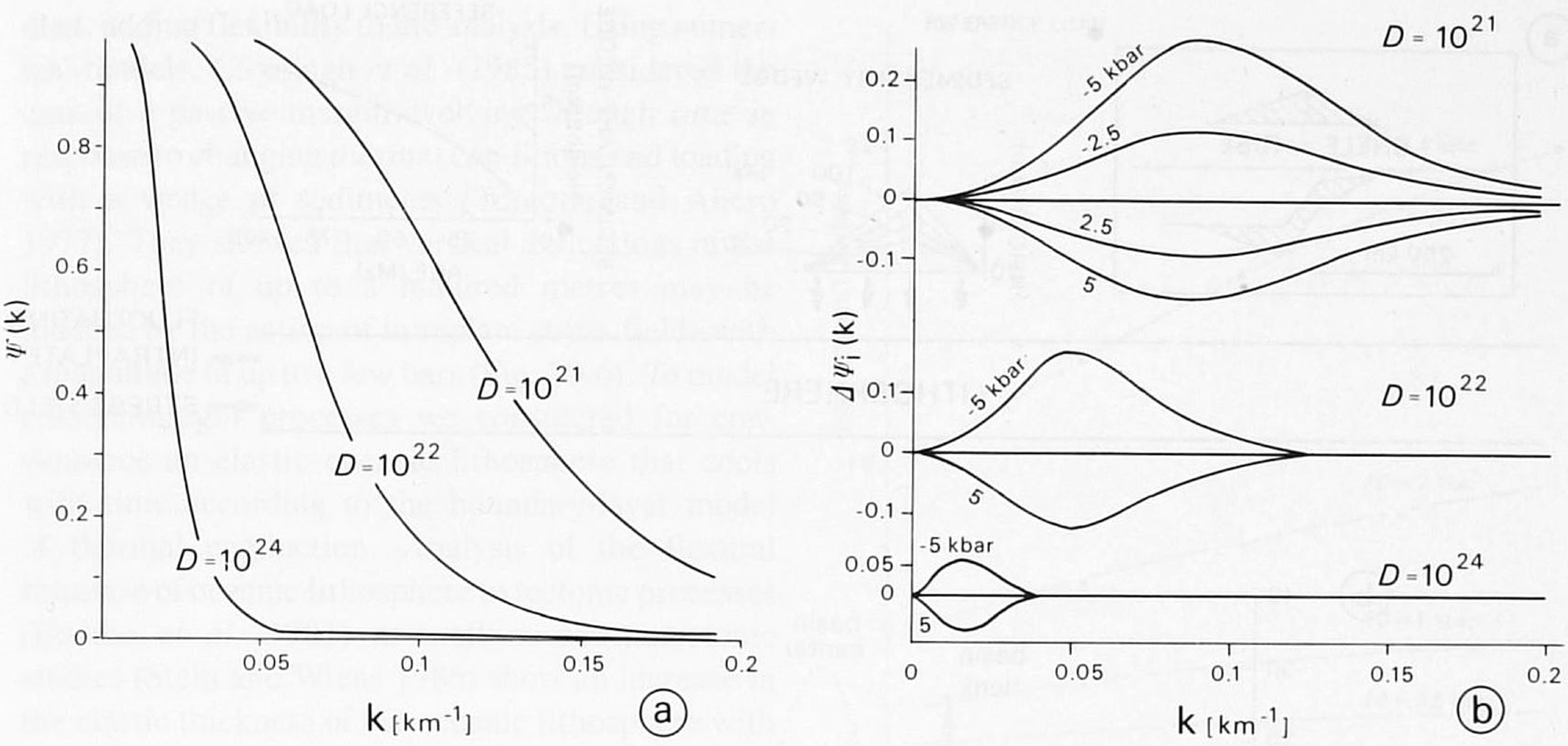


Fig. 10.5. (a) Flexural response functions $\Psi(k)$, in the absence of an intraplate stress field, for elastic thin plates with various flexural rigidities D in units of Newton metres, plotted as a function of wave number k. (b) The effect of intraplate stresses σ_N (tension is positive) on the response function. $\Delta\Psi_i(k) = \Psi_i(k) - \Psi(k)$, with $\Psi_i(k)$ the response function for an applied intraplate stress field. Results are given for the same flexural rigidities for various intraplate stresses σ_N in units of kbars. (After Stephenson and Lambeck 1985.)

forces. 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 the flexural rigidity ($D = ET^3/12(1-\nu^2)$), with E the Young's modulus, T the plate thickness, and ν the Poisson's ratio.

The axial load N is equivalent to the product of the intraplate stress σ_N and the plate thickness T. ρ_m and ρ_i are, respectively, the densities of mantle material and the infill of the lithospheric depression, usually water or sediment, and g is the gravitational acceleration. The solution to this classical equation is easily obtained for some simple loading cases (e.g., Turcotte and Schubert 1982). Assuming zero vertical load on the lithosphere, the early studies made a convincing case for neglecting horizontal forces in modelling the vertical motions of the lithosphere. They showed that for compressional forces below the buckling limit the induced vertical displacements of the lithosphere are negligible. This result, combined with lack of evidence for the existence of such horizontal forces, led for a long time to the withdrawal of attention from this topic. As has been discussed earlier, significant progress has been made recently in the study of horizontal stress fields in the lithosphere. At the same time, it is now realized that vertical motions of the lithosphere at sedimentary basins are primarily the result of a variety of other processes such as thermally induced cooling of the lithosphere amplified by the loading of sediments that accumulate in these basins (Sleep 1971), isostatic response to crustal thinning (McKenzie 1978), and flexural bending in response to vertical loading (Beaumont 1981). Hence, as pointed out by Cloetingh *et al.* (1985), 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 applied horizontal loads.

In analytical solutions of the equation describing the flexural behavior of thin elastic plates, the loading response of the plate is traditionally decomposed into its harmonic components by transforming the equation to the Fourier domain (Stephenson and Lambeck 1985). The flexural response function $\Psi_i(k)$ in the presence of a horizontal load N can be written as:

$$\Psi_i(k) = [1 + \frac{D\,(2\pi k)^4 - N\,(2\pi k)^2}{\rho_m g}]^{-1}$$

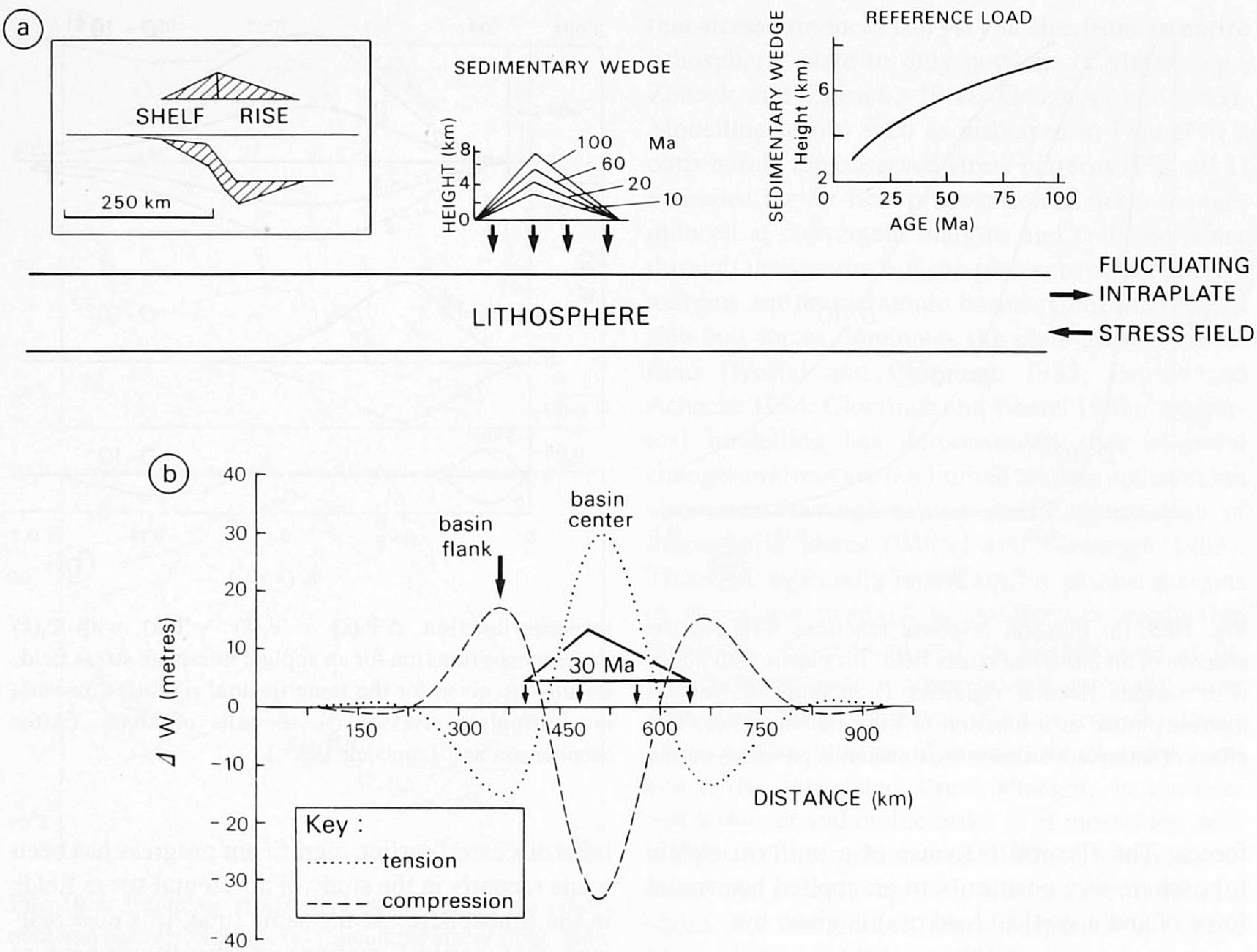


Fig. 10.6. (a) Model for apparent sea-level fluctuations resulting from variations in the intraplate stress field proposed by Cloetingh *et al.* (1985). The vertical displacement of the lithosphere at a passive margin evolves through time because of the thermal contraction and strengthening of the lithosphere and the loading with a wedge of sediments. Inset on the left shows position of this wedge on the outer shelf, slope, and rise. Sedimentation is assumed to be sufficiently rapid to equal approximately subsidence rate (Turcotte and Ahern 1977). The height and the width of the sedimentary wedge (see inset on the right for reference model of sediment loading) are given in kilometres at some selected time steps (specified in million years). The lithosphere is modelled as a uniform elastic layer. Differences between continental and oceanic lithosphere are neglected. (b) Effect of variations in intraplate stress fields on the deflection of a plate with a uniform elastic thickness (22 km) corresponding to a lithospheric age of 30 million years. Differential subsidence or uplift ΔW (metres) from the deflection in the absence of an intraplate stress field is given for intraplate stress fields of 1 kbar compression (dashed curve) and 1 kbar tension (dotted curve). Note the opposite effects at the flanks of the basin and at the basin center.

If N = 0, then $\Psi_i(k) = \Psi(k)$, the flexural response function of the plate in the absence of intraplate stress.

The wave number k at which the intraplate stresses most affect the flexural response of the lithosphere is almost completely determined by the plate's flexural rigidity (Stephenson and Lambeck 1985). These authors showed that the presence of intraplate stresses has a small but perceptible effect on this wave number, but exerts a controlling influence on the amplitude of the response for a given flexural rigidity. These features are illustrated in Figure 10.5, which shows the effect of intraplate stress fields of a magnitude of a few kbars on the flexural response of an elastic lithosphere.

The analytical formulation of specific simplified problems shows explicitly how the solution depends on various parameters. Numerical modelling techniques have the advantage of allowing more realistic geometries and variation in parameters to be handled, adding flexibility to the analysis. Using numerical models, Cloetingh *et al.* (1985) considered the case of a passive margin evolving through time in response to changing thermal conditions and loading with a wedge of sediments (Turcotte and Ahern 1977). They showed that vertical deflections of the lithosphere of up to a hundred metres may be induced by the action of intraplate stress fields with a magnitude of up to a few bars (Fig. 10.6). To model passive-margin processes we considered for convenience an elastic oceanic lithosphere that cools with time according to the boundary-layer model of thermal conduction. Analysis of the flexural response of oceanic lithosphere to tectonic processes (Bodine *et al.* 1981) as well as seismotectonic studies (Stein and Wiens 1986) show an increase in the elastic thickness of the oceanic lithosphere with age. Thus the response of the oceanic lithosphere to sediment load (Watts *et al.* 1982), and the intraplatehd stress field, is time dependent not only because the sediment load accumulates with time but also because of the changing mechanical properties of the lithosphere.

The vertical motions become more complex when the effective elastic thickness of the lithosphere is laterally variable (Artyushkov 1974; Cloetingh *et al.* 1985). For passive margins, this may result in an additional tilting of the lithosphere at the basin edge. This tilting is a consequence of the change in thickness of the layer that carries the intraplate stress, possibly due to changes in lithospheric thinning for rifted basins, and occurs even in the absence of sediment loading. Vertical motions at the basin edge due to intraplate stresses (Fig. 10.6) are amplified by this mechanism when the effective elastic thickness of continental lithosphere is less than that of oceanic lithosphere. As this effect is highly dependent on the rheological contrast between oceanic and continental lithosphere (e.g., Vink *et al.* 1984), the degree of induced amplification or reduction will vary for different basins.

Consequences for Basin Stratigraphy

In the previous section I pointed out that temporal fluctuations in stress are a natural consequence of the horizontal motions of the lithospheric plates. These findings have important consequences for short-term temporal variations in the basin shape through geological time (Cloetingh *et al.* 1985). This is illustrated by Figure 10.7, which displays the

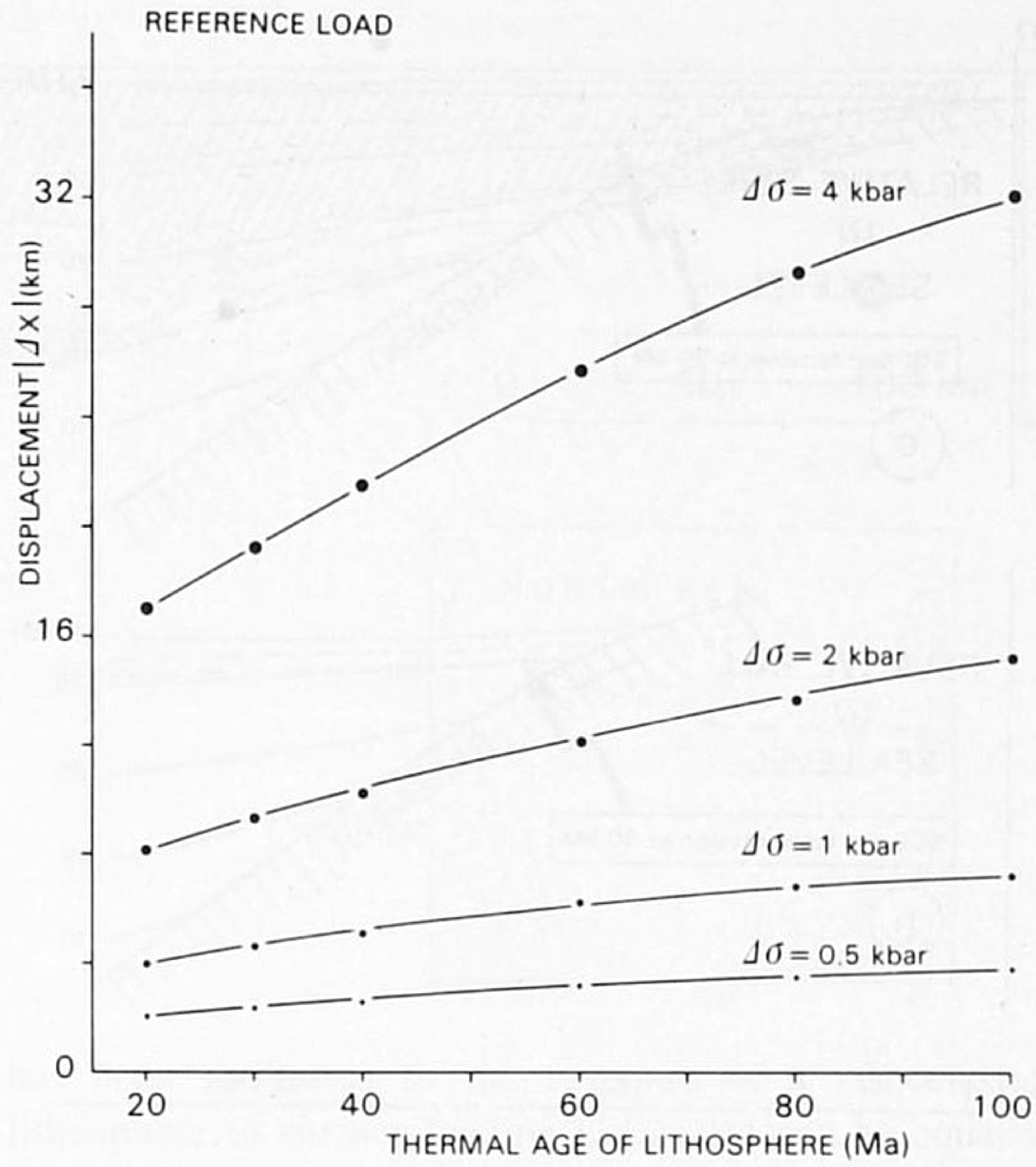


Fig. 10.7. Migration of the flexural node inferred from model calculations shown in Figure 10.6b. Apparent horizontal displacements $|\Delta(x)|$ (km) are plotted as a function of the age of the underlying lithosphere and sediment loading according to the reference model (after Turcotte and Ahern 1977) of Figure 10.6a. Curves give results for stress changes $\Delta\sigma$ of .5, 1, 2, and 4 kbar, respectively. Note that the horizontal expression of stress-induced short-term phases of coastal onlap and offlap increases with age.

horizontal migration of the flexural node under the influence of intraplate stresses. Curves show the horizontal displacement as a function of the age of the lithosphere for changes in stress of 0.5 kbar, 1 kbar, 2 kbar, and 4 kbar. For an intraplate compressional pulse the deflection of the lithosphere is abruptly narrowed and the flexural node migrates inward to the basin center. For an intraplate tensional pulse, rapid widening of the basin occurs and the flexural node migrates outward away from the center of the basin. These short-term widening and narrowing events are equivalent to rapid phases of coastal onlap and offlap, respectively. Inspection of Figure 10.7 shows that for a given stress change the horizontal extent of the basin area involved in a short-term phase of coastal onlap or offlap increases with lithospheric age. This feature is primarily the consequence of the increase in flexural stiffness of the lithosphere with age.

Sedimentation is assumed here to be sufficiently rapid to equal approximately the subsidence rate

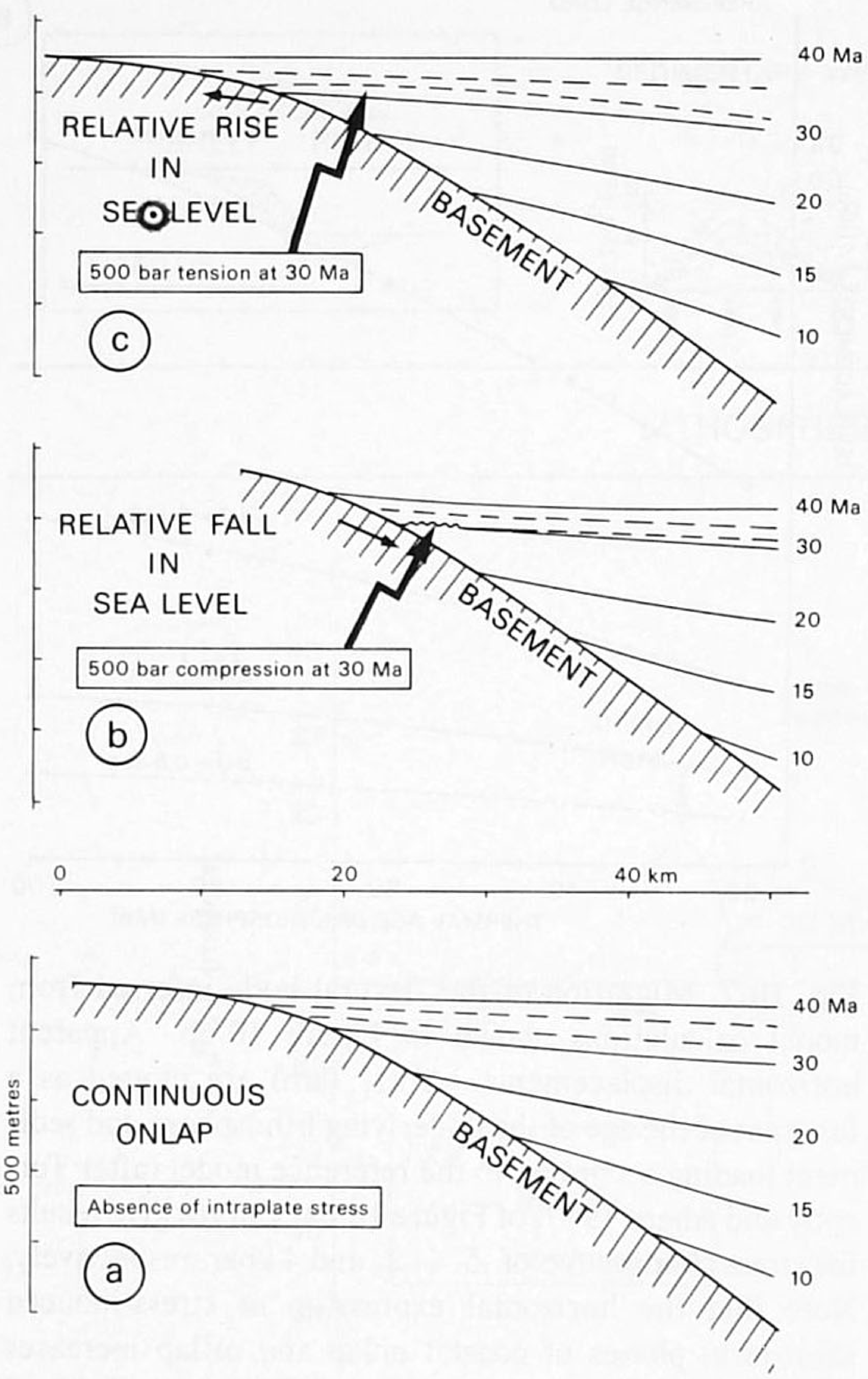


Fig. 10.8. Idealized stratigraphy at the edge of a basin underlain by 40 Ma lithosphere predicted on the basis of the calculations shown in Figure 10.6b. (a) Continuous onlap associated with long-term cooling of the lithosphere in the absence of an intraplate stress field. (b) A transition to 500-bar compression at 30 Ma induces a short-time phase of offlap at that time (indicated by position of short bold arrow) and an apparent fall in sea level superimposed on the thermal cooling subsidence curve. (c) A transition to 500-bar tension produces an additional short-term phase of onlap at 30 Ma (see position of short bold arrow) and an apparent rise in sea level.

(Turcotte and Ahern 1977). Although this reference model is certainly an oversimplification (see Pitman 1978 and Pitman and Golovchenko 1983 for a discussion), the assumption of sedimentation rate equalling subsidence rate allows one to equate onlap and offlap with rises and falls in sea level. The vertical motion of the lithosphere at the flank of the basin is of particular stratigraphic interest. Figure 10.8 schematically illustrates the relative movement between sea level and the lithosphere at the flank of the basin immediately landward of the principal sediment load as predicted on the basis of numerical calculations. When horizontal compression occurs, the peripheral bulge is magnified while simultaneously migrating in a seaward direction, uplift of the basement takes place, an offlap develops, and an apparent fall in sea level results, possibly exposing the sediments to produce an erosional or weathering horizon. Simultaneously, the basin center undergoes deepening (Fig. 10.6b), resulting in a steeper basin slope. For a horizontal tensional intraplate stress field, the flanks of the basin subside with its landward migration producing an apparent rise in sea level so that renewed deposition, with a corresponding facies change, is possible. In this case the center of the basin shallows (Fig. 10.6b), and the basin slope is reduced. The synthetic stratigraphy at the basin edge is schematically shown for the following three situations: long-term widening of the basin with cooling in the absence of an intraplate stress field (Fig. 10.8a), the same case with a superimposed transition to 500 bar compression at 30 Ma (Fig. 10.8b), and the case of a stress change to 500 bar tension at 30 Ma (Fig. 10.8c).

Beaumont (1978) and Watts *et al.* (1982) pointed out the differences in stratigraphy predicted on the basin edge for elastic and viscoelastic models in the absence of intraplate stress fields. These authors showed that for elastic models the width of the basin increases with time, inducing progressive onlap of sediments caused by the increase of the thickness of the elastic lithosphere since basin formation. They demonstrated that viscoelastic models for the lithosphere predict narrowing of the infill of the basin with younger sediments restricted to the basin center. However, care should be taken in rigidly classifying basins into elastic type or viscoelastic type strictly on the basis of this criterion. As pointed out by Allen *et al.* (1986), "from a sedimentologist's or stratigrapher's point of view there has been a largely futile but understandable search for basin geometries that support an elastic lithosphere at one extreme or a viscoelastic lithosphere at the other." The incorporation of intraplate stresses in elastic models of basin evolution can in principle predict a succession of onlaps and offlaps such as observed along the flanks of the Central North Sea Basin during Tertiary time (Fig. 10.9). Such a stratigraphy can be interpreted as a natural consequence of the mechanical widening and narrowing of basins by fluctuations in intraplate stress fields superimposed on the long-term broadening of the basin with cool-

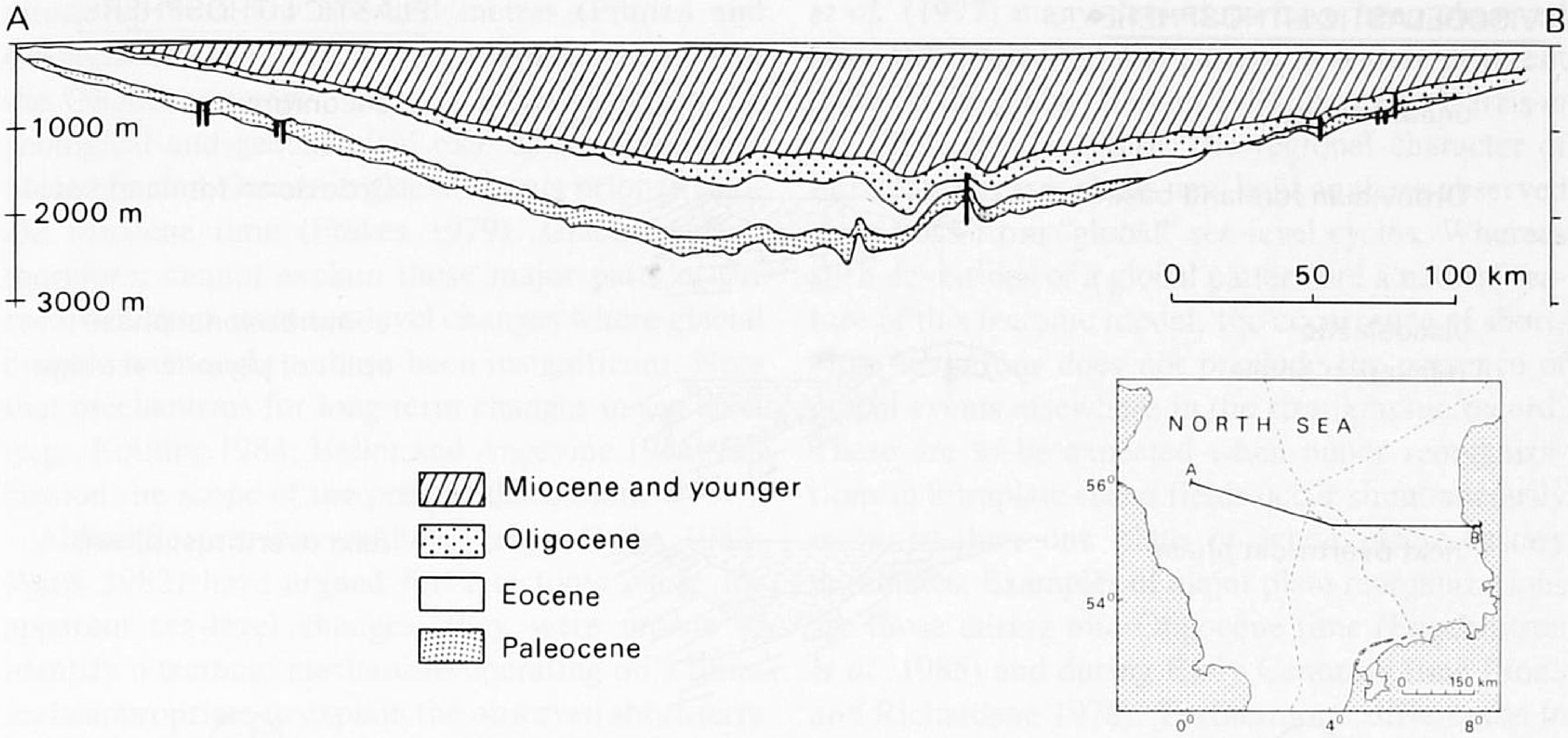


Fig. 10.9. Stratigraphic cross section through Cenozoic Central North Sea Basin (for location see inset). This type of stratigraphy with narrowing of the infilling of the basin with younger sediments restricted to the basin center often has been attributed to the response of a viscoelastic lithosphere to surface loading alone, but can be equally well explained in terms on an elastic plate under the influence of intraplate stresses (after Ziegler 1982).

ing since its formation. In particular, the modelling has shown that the stratigraphy predicted for a basin located on elastic lithosphere and under the influence of a compressional stress regime during deposition of younger strata will be very similar to the stratigraphy characteristically attributed (Watts *et al.* 1982) to the response of viscoelastic lithosphere to surface loading alone (see also Figs. 10.8, 10.9).

Although we have concentrated on the relationship between tectonics and stratigraphy at passive margins and intracratonic basins, the effect of intraplate stress fields is of importance to a wider range of sedimentary environments. Other settings where lithosphere is flexed downward under the influence of sedimentary loads occur in foreland basins (Beaumont 1981; Quinlan and Beaumont 1984). Despite its height of at most a few hundred meters the peripheral bulge flanking forelands basins is of particular stratigraphic interest (Quinlan and Beaumont 1984). These authors interpreted the development of unconformities in the Appalachian foreland basin in terms of uplift of the peripheral bulge caused by viscoelastic relaxation of the lithosphere (Fig. 10.10). However, the presence of intraplate stress of tensional or compressional type, of which the latter is more natural in this tectonic setting, can reduce or amplify the height of the peripheral bulge by an equivalent amount and thus greatly influence the stratigraphic record in foreland basins. In addition to lithospheric density variations (Stockmal *et al.* 1986), intraplate stresses might remove the need for the existence of hidden loads (Royden and Karner 1984) postulated to improve the fit between foreland basin deflection and the observed topographic load. As pointed out by Allen *et al.* (1986), syntectonic unconformities in basin-margin sequences of foreland basins demonstrate contemporaneous tectonic activity and sedimentation. It is interesting to note that the magnitude of the vertical motions caused by intraplate stresses is of the same order as those attributed to thermal blanketing effects superimposed on long-term foreland basin subsidence (Kominz and Bond 1986).

The dependence of the vertical motions at the edge of the basin on the age of the underlying lithosphere is demonstrated in Figure 10.11. The differential uplift, defined as the difference in deflection for a change in intraplate stress, is calculated for the flank of the basin as a function of the variation in the intraplate stress field. Curves illustrate the deflection for changes in stress of 0.5 kbar, 1 kbar, 2 kbar, and 4 kbar, with the same reference sediment load in each case. As pointed out by Cloetingh *et al.* (1985), stress changes of a few kbar occurring on time scales of a few million years and longer may furnish a tectonic explanation for short-term apparent sea-level

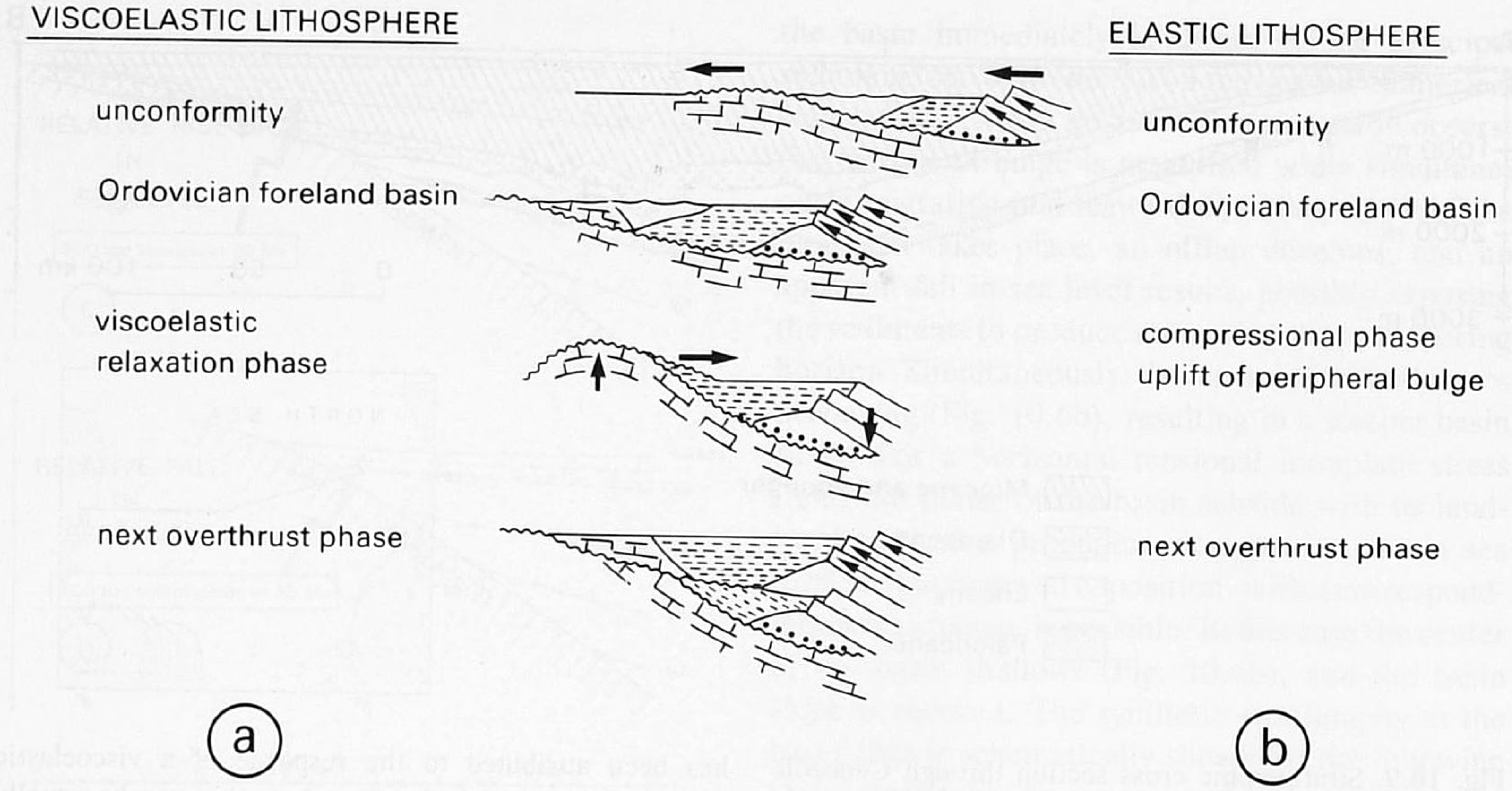


Fig. 10.10a. Diagram illustrating interpretation by Quinlan and Beaumont (1984) for the development of regional unconformities in a multistage foreland basin on viscoelastic lithosphere (column on the left). Bold arrows show overthrust and peripheral bulge migration. Fine arrows illustrate active overthrusting. The column on the right (b) shows interpretation of foreland-basin stratigraphy in terms of the action of intraplate stresses superimposed on flexure of an elastic lithosphere (modified from Quinlan and Beaumont 1984).

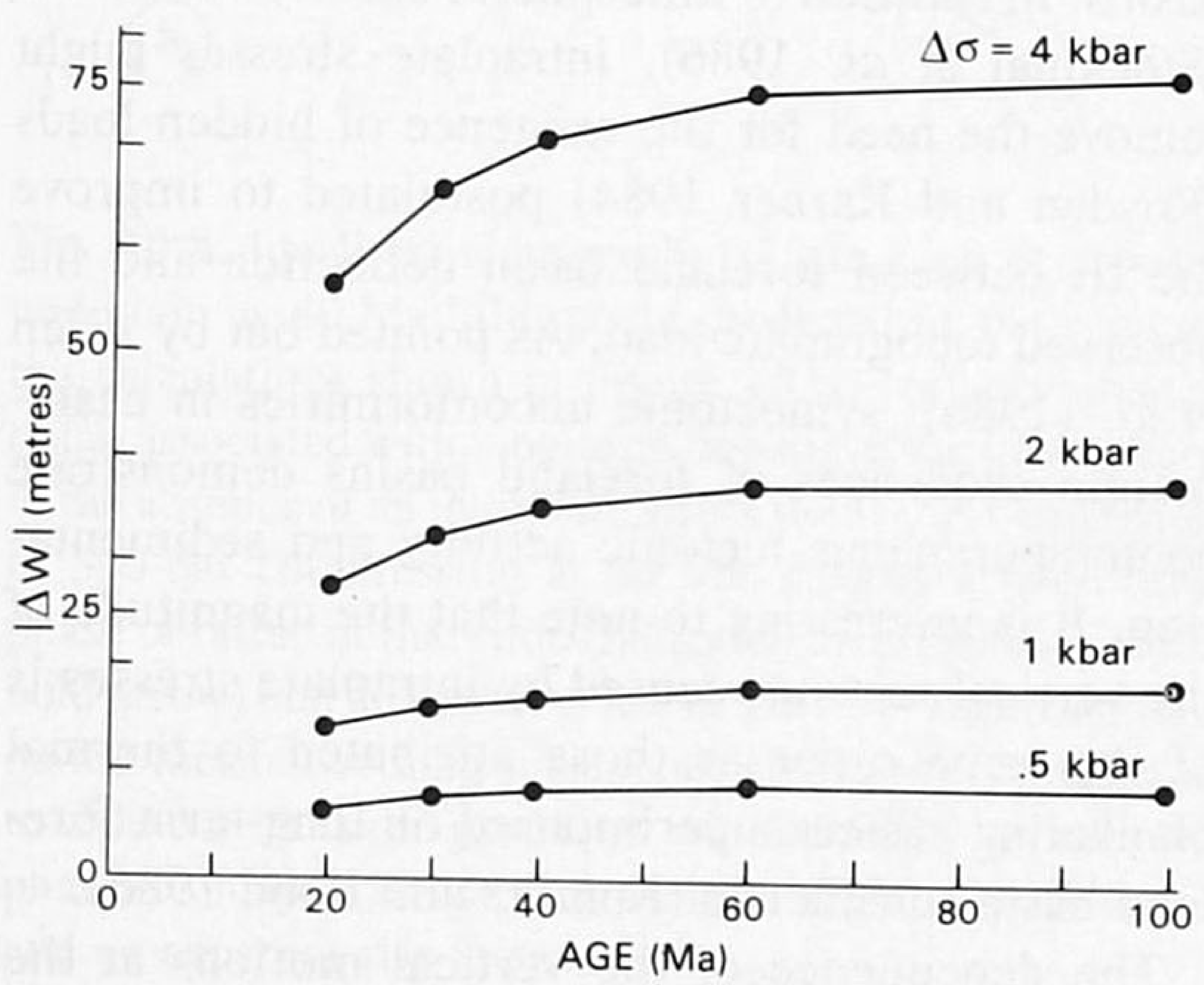


Fig. 10.11. Apparent sea-level fluctuations $|\Delta w|$ (*metres*) at basin edge (position marked by vertical arrow in Fig. 10.6b) due to superposition of variations in intraplate stress field on flexure caused by sediment loading. The data for $|\Delta w|$ are plotted as a function of the age of the underlying lithosphere and sediment loading according to the reference model (after Turcotte and Ahern 1977) of Figure 10.6a. Curves give results for stress changes ($\Delta\sigma$) of .5, 1, 2, and 4 kbar, respectively.

fluctuations with a magnitude of up to a hundred metres and a rate of 0.01–0.1 m/10^3 yr inferred from the seismic stratigraphic record. Note, however, that the modified Vail *et al.* curves (Vail *et al.* 1984; Haq *et al.* 1987) show a general reduction of the magnitude of the apparent sea-level fluctuations. Similarly, independent recent studies of the magnitude of the mid-Oligocene sea-level lowering point to a value much lower than previously thought. The magnitude of this fall in sea level, which is by far the largest shown in the Vail *et al.* curves, is now estimated to be between 50 metres (Miller and Fairbanks 1985; Watts and Thorne 1984) and 100 metres (Schlanger and Premoli-Silva 1986). Hence, a significant part of the short-term sea-level fluctuations inferred from seismic stratigraphy might have a characteristic magnitude of only a few tens of metres with a time interval of a few million years (Aubry 1985), which can be explained by relatively modest stress variations of a few hundred bars.

The short-term cycles in sea level have been traditionally interpreted in terms of glacio-eustatic controls. This situation has been caused partly by the inferred global character of the changes and partly by the absence of a tectonic mechanism to explain the inferred rapid lowerings in sea level with a magnitude up to a few hundred metres (Pitman and Golovchenko 1983). However, with the exception of the Oligocene events, there is no evidence in the geological and geochemical records for significant Mesozoic and Cenozoic glacial events prior to middle Miocene time (Frakes 1979). Glacio-eustasy, therefore, cannot explain those major parts of the record of short-term sea-level changes where glacial events are thought to have been insignificant. Note that mechanisms for long-term changes in sea level (e.g., Kominz 1984; Heller and Angevine 1985) fall beyond the scope of the present discussion.

Although previous authors (e.g., Bally 1982; Watts 1982) have argued for a tectonic cause for apparent sea-level changes, they were unable to identify a tectonic mechanism operating on a timescale appropriate to explain the observed short-term lowerings of the sea level. Changes in intraplate stress fields associated with tectonic reorganizations also explain the existence of a strong correlation (Bally 1982) in timing of plate reorganizations and rapid lowerings in sea level shown in the Vail *et al.* (1977) curves. Given current active research in sea-level fluctuation, the possibility that the effect of intraplate stresses is significant has been explored for various regions. Schlanger (1986) suggested that intraplate stress variations provide a possible explanation for the enigmatic high frequency (order of a few million years) variations in Cretaceous apparent sea levels. Hallam (in press) discussed the implications of such variations for the Jurassic sea levels of northwestern Europe. Meulenkamp and Hilgen (1986) explored a causal relationship between Neogene stratigraphy in the eastern Mediterranean and intraplate stress variations. Klein (1987) examined the stratigraphic record of the Paleozoic Appalachian foreland basin and concluded that local variation and local tectonic styles and associated changes in stress field might have masked the possible expression of global fluctuations in sea level in the basin.

Discrimination of Tectonics and Eustasy

It is generally agreed that unconformity-bound units caused by short-term changes in apparent sea level form the natural components of the sedimentary record (Van Hinte 1983). However, the observed boundaries could well reflect local or regional events and their supposed global nature remains to be proven (Parkinson and Summerhayes 1985). Since the publication of the onlap/offlap curves by Vail *et al.* (1977) many deviations have been observed from the inferred global pattern in widely different areas on the globe (Hallam 1984, in press; Harris *et al.* 1984; Jenkin 1984). The regional character of intraplate stresses sheds new light on these observed deviations from "global" sea-level cycles. Whereas such deviations of a global pattern are a natural feature of this tectonic model, the occurrence of short-term deviations does not preclude the presence of global events elsewhere in the stratigraphic record. These are to be expected when major reorganizations in intraplate stress fields occur simultaneously in more than one plate or when glacio-eustasy dominates. Examples of major plate reorganizations are those during mid-Oligocene time (Engebretson *et al.* 1985) and during Early Cenozoic time (Rona and Richardson 1978). Furthermore, differences in rheological structure of the lithosphere, which influence its response to applied intraplate stresses, might also explain differences in magnitude of the inferred sea levels such as observed between the North Sea region and the Gippsland Basin off southeastern Australia (Vail *et al.* 1977).

The discrimination of regional events from eustatic signals in the sea-level record of individual basins is usually a subtle matter, especially if biostratigraphic correlation is imprecise (Hallam 1984, in press). It is interesting to note that the vertical motions induced by the action of intraplate stresses in the center of the basin, although of the same magnitude as the displacements at the flanks, are small compared to the total subsidence in the basin center, which is on the order of several kilometres. The sign and magnitude of the apparent sea-level change will be a function of the sampling point (Figs. 10.6b, 10.8), which may provide a means of testing the effect of intraplate stresses or distinguishing this mechanism from eustatic contributions. In fact, Hallam (in press) has shown that a significant number of Jurassic unconformities are confined to the flanks of North Sea basins, consistent with the predictions of the tectonic mechanism of Figure 10.6.

The effect of intraplate stresses could in principle cause abrupt phases of rapid uplift and subsidence visible in many published subsidence curves of passive margins, intracratonic basins, and foreland basins. An example of such a curve for the Williston Basin of western North America is shown in Figure 10.12. As noted by Fowler and Nisbet (1985) and Sloss (in press), previously proposed subsidence mechanisms fail to explain the observed irregularities in the subsidence pattern. Similarly, changes in

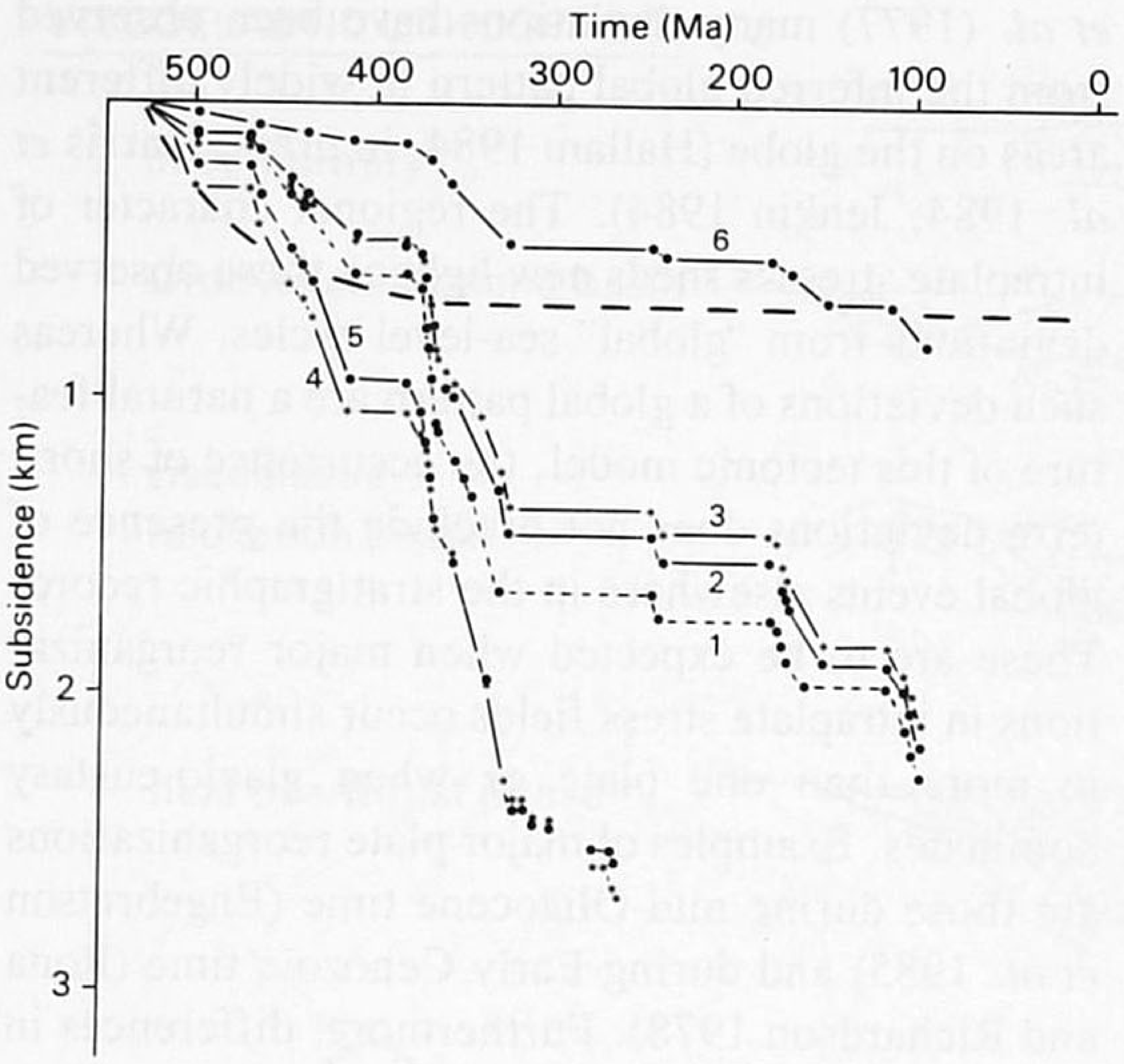


Fig. 10.12. Subsidence-time plots for five wells from the Williston Basin (after Fowler and Nisbet 1985). 1,2,3 = Saskatchewan wells; 4,5 = North Dakota wells; 6 = subsidence in well 2 after backstripping to show subsidence when basin is filled with water alone. Unconformities are shown as horizontal lines. Note that the curves have not been corrected for the erosion of sediments in the unconformities, which gives a distorted picture of the actual magnitude of the irregularities. Heavy dashed line is the subsidence curve calculated for a thermal time constant of 50 Ma. In curve 6, abrupt phases of rapid subsidence are visible, which cannot be explained by thermal processes, but which could possibly result from intraplate stresses.

stress field provide a simple explanation for the observed tilting (Ahern and Mrkvicka 1984) and associated temporal rotation of the source area of the sediments deposited in the Williston Basin. Flexural bulges have probably greatly influenced the extent of erosion that has occurred in the Williston Basin. Episodic subsidence patterns have been observed for many other intracratonic basins, such as the Illinois Basin (Heidlauf *et al.* 1986), passive margins including the United States Atlantic margin (Heller *et al.* 1982), and foreland basins such as the Swiss Molasse basins (Allen *et al.* 1986) and the Appalachian Basin (Klein 1987).

As noted earlier, the action of intraplate stresses causes different and opposite effects on the subsidence at the flanks and in the basin center. These spatial variations have a time-dependent expression due to the shift in the position of the flexural node by the long-term widening of the basin during its thermo-mechanical evolution. This results in varying expressions of applied stresses with time. The laterally varying effect of stresses applied on various time intervals on subsidence curves of the basin is schematically illustrated in Figure 10.13.

In the present paper I have adopted a reference model in which sedimentation approximately equalled subsidence. As pointed out previously, this assumption allowed me conveniently to equate onlap and offlap with rises and falls in sea level, respectively. Actually, the position of coastal onlap reflects the position where rate of subsidence equals rate of fall in sea level. During the application of stress 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 (Turcotte and Ahern 1977). Hence, as pointed out by Thorne and Watts (1984), to produce offlaps during late stages of passive-margin evolution requires much lower values of rate of change of sea level than those needed to produce offlaps during the earlier stages of basin evolution. If these offlaps are caused by fluctuations in intraplate stresses, this implies that the rates of changes of stress needed to create them diminish with age during the flexural evolution of the basin (see also Fig. 10.7). This is of particular relevance for an assessment of the relative contribution of tectonics and eustasy as a cause for Cenozoic unconformities. For example, Cenozoic unconformities developed at old passive margins in association with short-term narrowing of margin basins could be produced by relatively mild changes in intraplate stresses. In this context it is interesting to note that late-stage narrowing of Phanerozoic platform basins and Atlantic margin basins is frequently observed (e.g., Sleep and Snell 1976), without clear evidence for active tectonism. Proterozoic basins, which also show in many cases narrowing during later stages of their evolution, differ, however, by displaying a clear association of the narrowing phase with orogenic events (Grotzinger and Gall 1986; Grotzinger and McCormick 1988: this volume). This possibly reflects closure of the basin and destruction of the basin margins in a regime of strong compression.

Obviously, a need exists to do careful two-dimensional and three-dimensional facies analysis to test the general application of intraplate stresses in sedimentary basin models. It is equally important to obtain precise ages of the sequence boundaries and rate and magnitude of apparent sea-level change to establish their global or regional nature. In this context, critical stratigraphic examination of seismic sections across the inner edge of passive margins to the seaward part is of particular importance (e.g., Thorne and Watts 1984). New criteria will have to be developed to facilitate separation of eustatic effects and effects of vertical motions of the lithosphere on the sedimentary record (e.g., Klein 1982). Of special importance in this respect might be the stratigraphic consequence of spatial and temporal variations of the magnitude of the apparent sea-level change. Similarly, syndepositional deformation (see below) might prove to be a useful criterion to identify intraplate stresses.

Syndepositional Folding of Lithosphere, the Formation of Foreland Arches, and Inversion Tectonics

More data are needed to examine the actual response of the lithosphere under the influence of very large intraplate stresses. The region in the northeastern Indian Ocean south of the Bay of Bengal provides a relatively well understood example of intense deformation caused by intraplate stresses. Seismic reflections profiles (Fig. 10.14; Geller *et al.* 1983) show widespread deformation of originally horizontal sediments. The deformation occurs by broad basement undulations, with wavelengths of roughly 200 km and amplitudes up to 3 km, and numerous high-angle reverse faults. The strike of the undulations and reverse faults is approximately east-west, in agreement with the proposed north-south orientation of the stress field in the area. A prominent Upper Miocene unconformity separates deformed sediments from overlying strata (Geller *et al.* 1983). The absence of syn-sedimentary deformation below the unconformity provides evidence for a Late Miocene age for the timing of the onset of the deformation. The basement undulations coincide with undulations in gravity and geoid anomalies (McAdoo and Sandwell 1985). For elastic plate models, stresses of several tens of kbars are required to induce the observed folding of the basement. McAdoo and Sandwell (1985) pointed out in their study of the folding of the oceanic crust in the Bay of Bengal the importance of incorporating a more realistic rheology in models of the response of the lithosphere to large intraplate stresses. Their work showed that a depth-dependent rheology of the

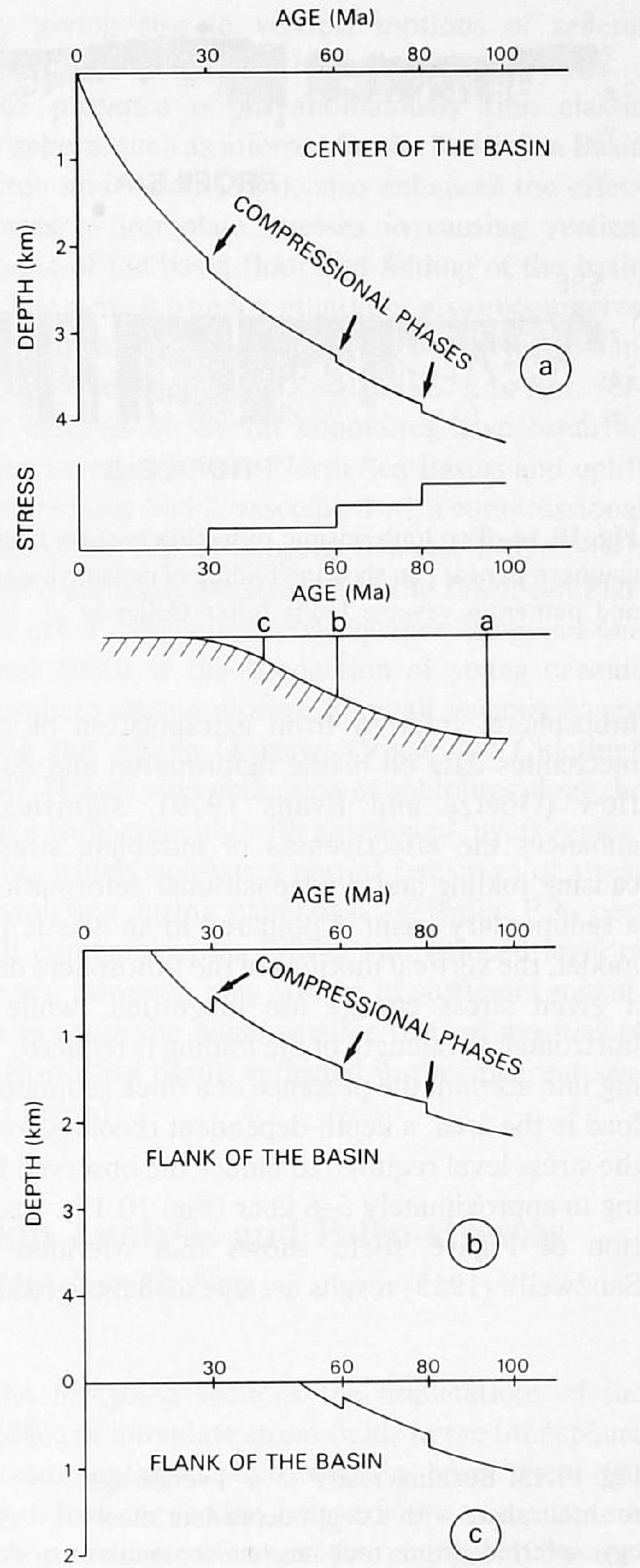


Fig. 10.13. Effect of intraplate stresses on subsidence curves at three different positions (a, b, and c) in the basin. (a) Effect of compression on subsidence predicted for a well in the basin center; (b) and (c) show the effect of compressional stress on subsidence for locations at the flanks of the basin, closer to the position of the flexural node. Note in these cases the different effects of compression at different time intervals, which are caused by shifts in the position of the flexural node caused by widening of the basin during its long-term thermal evolution.

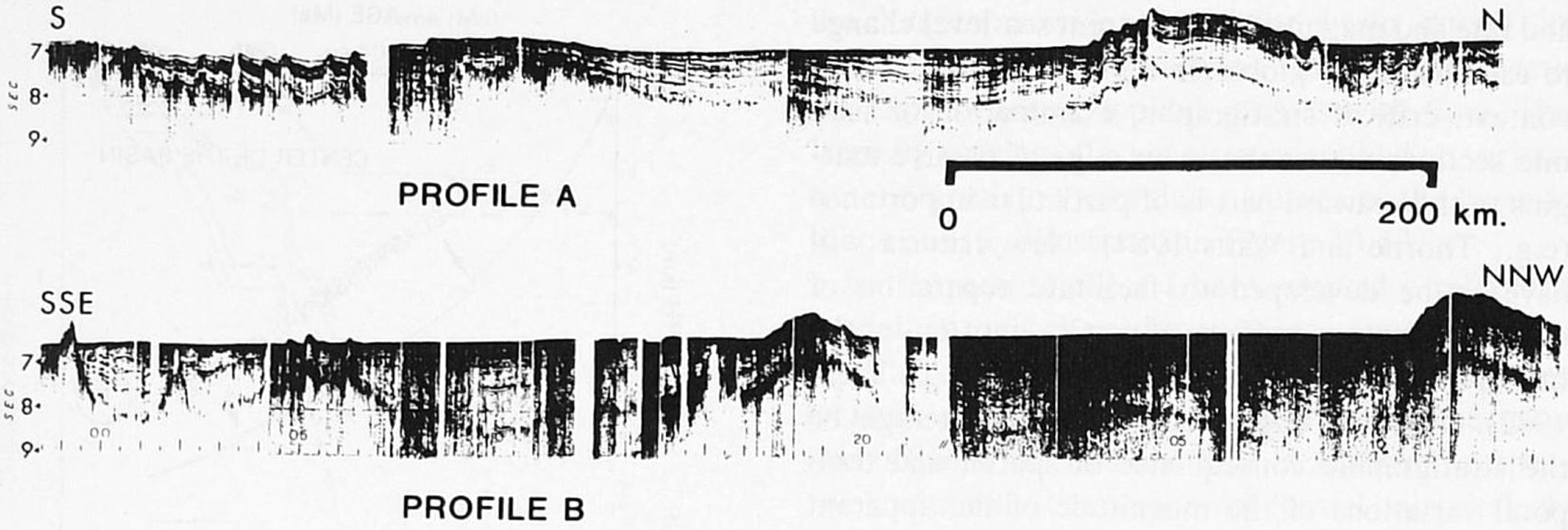
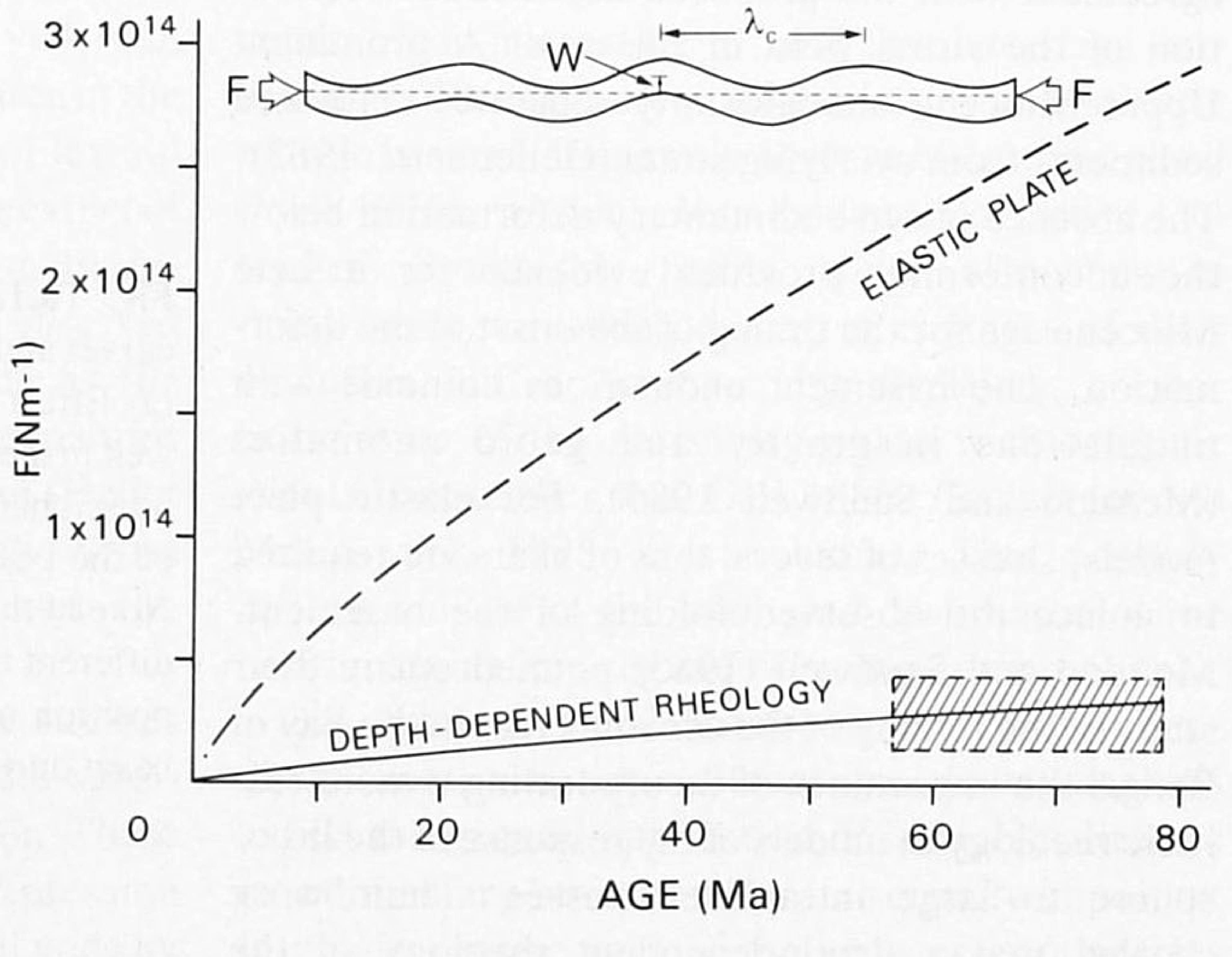


Fig. 10.14. Two long seismic reflection profiles from the southern Bengal Fan showing folding of oceanic basement and numerous reverse faults (after Geller *et al.* 1983). Time is given along the vertical scale in terms of seconds of two-way travel time. The location of the profiles is indicated in Figure 10.3a.

lithosphere, inferred from extrapolation of rock-mechanics data on brittle deformation and ductile flow (Goetze and Evans 1979), significantly enhances the effectiveness of intraplate stress in causing folding and syndepositional deformation in a sedimentary basin. Compared to an elastic plate model, the vertical motions of the lithosphere due to a given stress change are magnified, while the horizontal wavelength of the folding is reduced. Taking into account the presence of a thick sedimentary load in the area, a depth-dependent rheology lowers the stress level required to induce the observed folding to approximately 5–6 kbar (Fig. 10.15). Inspection of Figure 10.15 shows that McAdoo and Sandwell's (1985) results are in excellent agreement with independent estimates inferred from numerical modelling of intraplate stresses in the northeastern Indian Ocean (Cloetingh and Wortel 1985, 1986). Rheological weakening of the lithosphere due to flexural stresses induced by sediment loading (Cloetingh *et al.* 1982) also enhances the effectiveness of the action of intraplate stresses in a more general sense. However, for compressional deformation not reaching the buckling limit, the effect of incorporating more realistic rheologies in the models is less dramatic than in the northeastern Indian Ocean.

Syndepositional folding of oceanic lithosphere might have some interesting analoges in the continents (e.g., Porter *et al.* 1982). Rheological models

Fig. 10.15. Buckling load F (Nm^{-1}) versus age for lithosphere with a depth-dependent rheology inferred from rock-mechanics studies (Goetze and Evans 1979) given by the solid line and for fully elastic oceanic lithosphere given by the dashed line (after McAdoo and Sandwell 1985). Incorporation of depth-dependent rheology magnifies the vertical motions (W) and reduces the horizontal wavelength (λ_c) of stress-induced folding of the lithosphere. The box indicates stress levels calculated for the area in the northeastern Indian Ocean (Cloetingh and Wortel 1985) where folding of oceanic lithosphere under the influence of compressional stresses has been observed (McAdoo and Sandwell 1985).



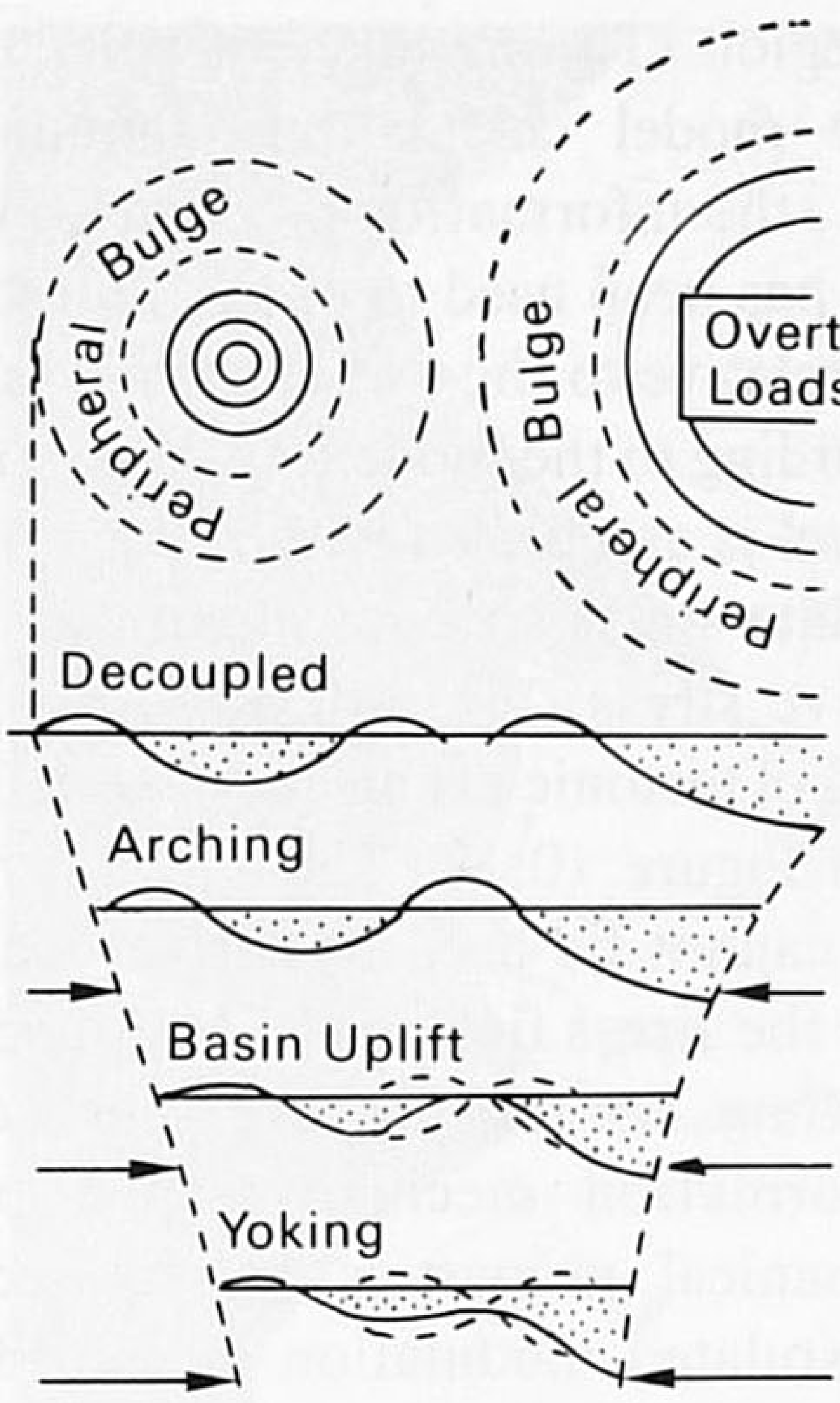


Fig. 10.16. Flexural interaction between foreland basin (right) and an intracratonic basin (left) causes the formation of arches. The upper part of the figure displays a plan view of the basins in the decoupled position, corresponding to the first cross section below. Subsequent cross sections show the nature of the interaction when the basins are closer together. The latter is indicated by an increasing length of horizontal arrows (after Quinlan and Beaumont 1984).

indicate that continental lithosphere is weaker than oceanic lithosphere (Vink *et al.* 1984; Cloetingh and Nieuwland 1984). This feature reflects primarily mineralogic differences caused by different stratifications within oceanic and continental lithosphere. Compared to oceanic lithosphere, the strength of which is primarily determined by olivine, the strength of continental lithosphere is reduced by the presence of relatively low strength crustal material, in particular quartz (Stein and Wiens 1986). Therefore, syndepositional folding in continental interiors is probably reached at a lower stress level than required for oceanic plates (Fig. 10.15). In general it is expected that a specific change in intraplate stress will induce larger vertical motions in continental lithosphere than in oceanic lithosphere. The presence of tensional or compressional intraplate stresses can reduce or amplify the height of the peripheral bulge at foreland basins (Fig. 10.10). Similarly, intraplate stresses may affect the postulated interaction (Quinlan and Beaumont 1984) between adjacent foreland arches (Fig. 10.16), possibly giving rise to vertical motions of several kilometres in the case of large intraplate stresses.

The presence of an anomalously thin elastic lithosphere, such as inferred for the North Sea Basin (Barton and Wood 1984), also enhances the effectiveness of intraplate stresses in causing vertical motions of the basin floor and folding of the basin fill. The same is true for situations, also encountered in the North Sea Basin, where subsidence is primarily fault-controlled (e.g., Ziegler 1987). In fact, vertical motions of several kilometres have occurred during inversion of the North Sea Basins and uplift of intervening highs, associated with compressional phases of the Alpine Orogeny (Ziegler 1982, 1987). These compressional stresses in the European Platform are a mechanical consequence (England and Wortel 1980) of the subduction of young oceanic lithosphere during closure of small oceanic basins during the Alpine Orogeny (Vlaar and Cloetingh 1984). Phases with obduction of ophiolites along the Alpine front coincide with absence of inversion tectonics, which suggests a partial reduction of stress transmission during such phases (Ziegler, P.A. personal communication 1987). The reduced level of stresses, however, may still be of sufficient magnitude to cause the much smaller vertical motions of the North Sea basins reflected in the apparent sea-level record.

Basin Analysis and Paleo-stresses in the North Sea

In the foregoing sections the implications of the presence of intraplate stress fields in the lithosphere for sedimentary-basin evolution have been discussed. In doing so I have argued for a tectonic cause for short-term apparent sea-level changes. If this tectonic mechanism is accepted, then the apparent sea-level curves can in principle be interpreted in terms of paleo-stress fields. In Cloetingh (1986) (see also Lambeck *et al.* in press for a more detailed discussion) the North Sea region was selected for this purpose. There were several reasons for this choice. Although the Vail *et al.* (1977) "global" coastal onlap/offlap curves are based on data from various basins around the world, they are heavily weighted in favor of North America, the Gulf coast, the northern and central Atlantic margins, and especially the North Sea. Therefore as noted by several authors (e.g., Miall 1986) the global cycles strongly reflect the seismic stratigraphic record of basins in a tectonic setting dominated by rifting events in the northern and central Atlantic. Moreover, in the North Sea area these rifting events have alternated with compressional phases induced by the Alpine Orogeny. Ziegler (1982, 1987) has pointed out the existence of a correlation between tectonic phases in the North Sea and changes in the history of the Atlantic and Tethyan Oceans. Changes in spreading rates in the Atlantic and Tethyan oceans are probably caused by or associated with changes in plate-tectonic forces. Paleostress-field orientation inferred from analysis of microstructures by Letouzey (1986) shows distinctly different stress patterns at different geological periods (Fig. 10.17). Tectonic lineation rosettes determined from spectral analysis of North Sea subsidence show a rotation from a prominent E-W trend in Triassic time to N-S in late Cretaceous time (Thorne 1985), a trend also reflected in the measured paleo-stress directions (Letouzey 1986). The influence of Alpine Orogeny is reflected in differences in orientation between Pliocene and other Cenozoic directions (Thorne 1985).

In the analysis it is assumed that apparent sea level is controlled by regional tectonics. This assumption is certainly an oversimplification, as parts of the sea-level record may be controlled by glacio-eustasy or may be the result of simultaneously acting tectonic or glacio-eustatic processes. However, stratigraphic studies favor an important role for tectonic processes in determining the sea-level record of the North Sea region. Hallam (in press), for instance, recognizes a number of significant regressive events that appear to be caused by regional tectonics rather than global sea level. This applies in particular to the exceptionally high frequency of occurrence of short-term sea-level fluctuations toward the end of Jurassic time, which occur simultaneously with a pronounced increase in fault-controlled tectonic activity (Hallam in press).

Lambeck *et al.* (in press) constructed a composite sea-level curve for northwestern Europe from regional sea-level curves published by Vail *et al.* (1977) for Tertiary time, Juignet (1980) and Hancock (1984) for Cretaceous time, and Hallam (1984) for Jurassic time. The composite sea-level curve was calibrated by adopting a magnitude of 50 metres for the mid-Oligocene fall of sea level based on recent estimates inferred from analysis of subsidence in the North Sea region (Thorne and Bell 1983). On the basis of the model calculations summarized in Figure 10.11, the information on changes in apparent sea level has been used to obtain an estimate of paleo-stress relative to the present stress level (Fig. 10.18). According to the modelling, apparent lowering of sea level is associated with relaxation of tensional intraplate stress or, equivalently, an increase in compressive stress (Figs. 10.6, 10.8). Ziegler's (1982) timing of tectonic events in western Europe is also given in Figure 10.18. The specific model of Figure 10.6 cannot be used to derive quantitative estimates for the stress field in the North Sea region in Mesozoic time. For this purpose, specification of the basin formation mechanism and pertinent thermo-mechanical properties is required. However, the postulated modulation of subsidence by changes in the magnitude of the regional intraplate stress field is relevant for this period. The occurrence of a long-term Mesozoic sea-level rise (and tensional intraplate stress) followed by a long-term Tertiary lowering is in agreement with the observed dominance of wrenching and rifting in Mesozoic time (Ziegler 1982). During Mesozoic time basin subsidence occurred through graben formation and crustal stretching (McKenzie 1978; Sclater and Christie 1980), which requires tensional stresses of the order of a few kbar (Cloetingh and Nieuwland 1984; Houseman and England 1986). At the Mesozoic/Cenozoic transition a reduction in tension or an equivalent change to overall compression could have induced a reversal of this long-term trend. Note, however, that this interpretation is not unique as long-term changes in sea level can equally be explained in terms of changes in long-term spreading rate and area/age distribution of the ocean floor caused by continental break-up (Heller and Angevine 1985).

Of prime interest here are the short-term fluctuations superimposed on these long-term changes in sea level. Figure 10.18 shows that in this case as well, the synthetic paleo-stress curve derived from the seismic stratigraphic record provides a mirror image of the tectonic evolution of northwestern Europe and the North Sea Basin: Rifting episodes correspond to relaxation of tensional paleo-stresses and Alpine orogenic phases correspond to episodes of increased compressive stresses. For reasons mentioned in the previous section, the values of the inferred fluctuations in intraplate stresses are certainly upper limits for the North Sea area.

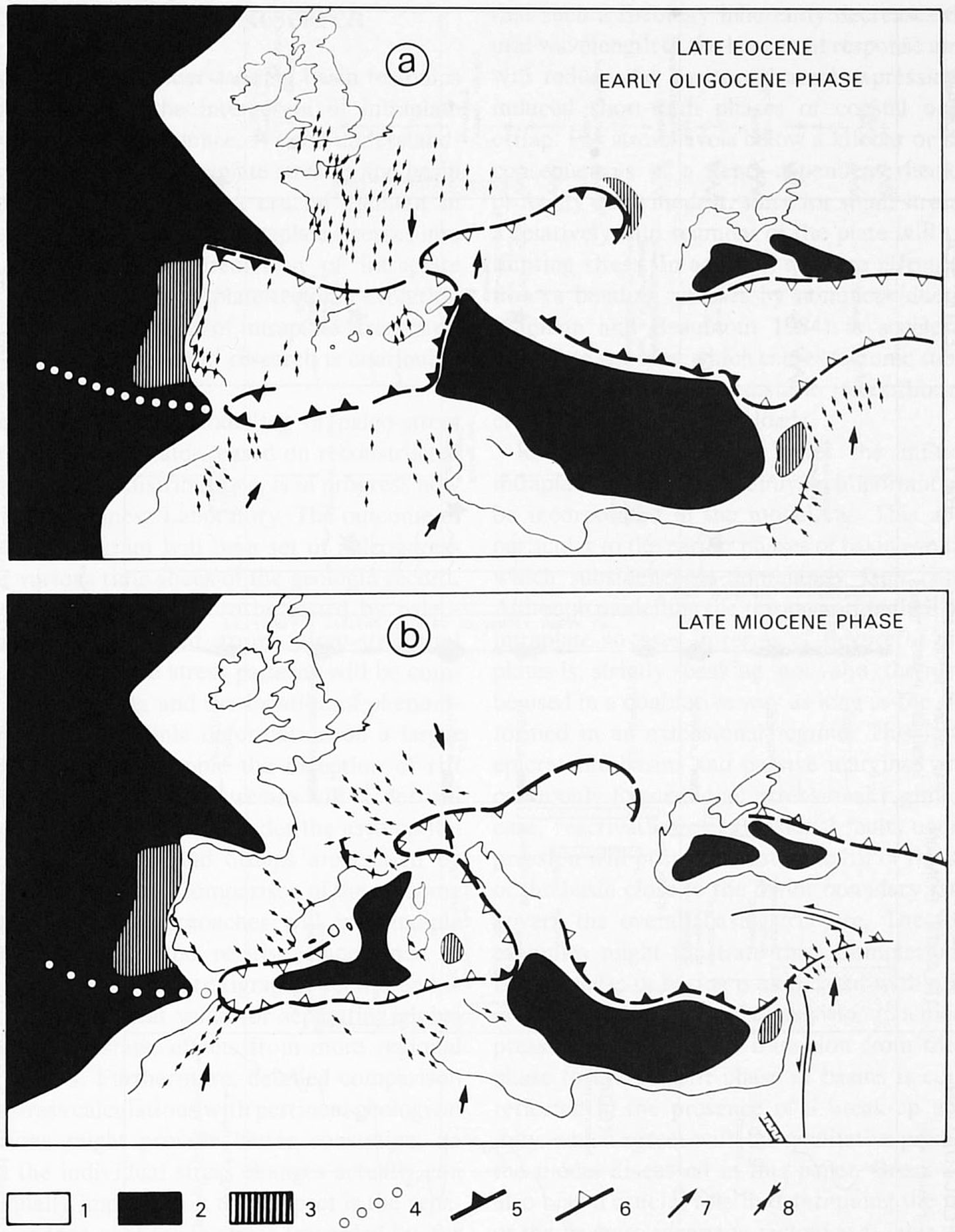


Fig. 10.17. Compressive paleo-stress field directions inferred from microtectonic studies of the northwestern European Platform and surrounding areas. 1 = continental crust and margins, 2 = oceanic crust, 3 = thin continental crust or oceanic crust, 4 = plate boundary, 5 = oceanic subduction, 6 = thrust and collision, 7 = relative Europe-Africa-Arabia azimuth vector of motion, 8 = maximum compressive paleo-stress axis orientation in fossil position relative to continental blocks (represented by present coast line). Position of the main Europe, Arabia, Africa continents is derived from kinematic reconstructions, with Europe arbitrarily fixed (after Letouzey 1986). (a) Late Eocene/Early Oligocene stress field orientations. (b) Late Miocene stress field. Comparison of (a) and (b) shows a rotation of the paleo-stress field in northwestern Europe during Tertiary time.

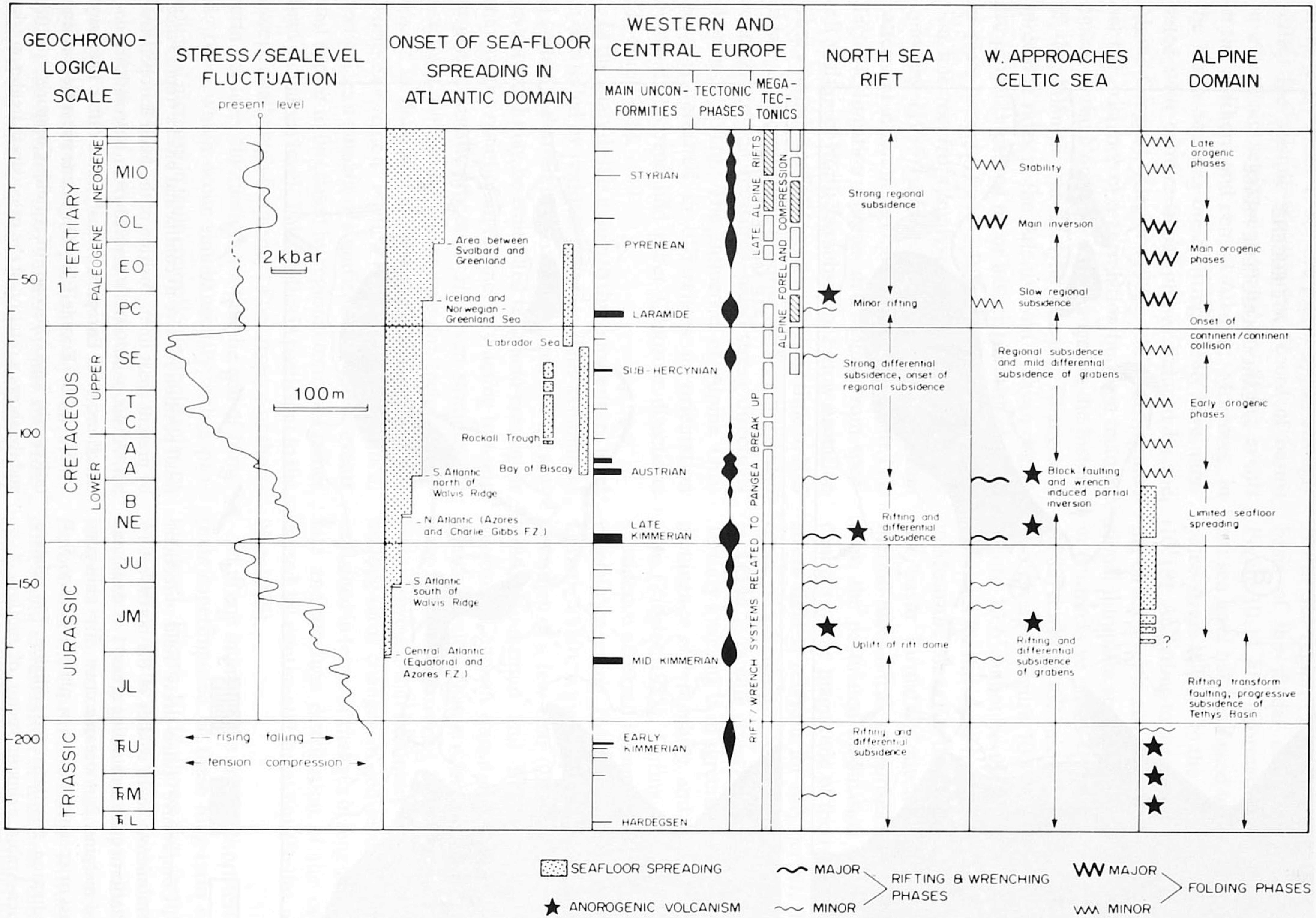


Fig. 10.18. Synthetic paleo-stress curve derived from a composite sea-level curve based on regional sea-level curves for northwestern Europe. Columns on right side show timing of tectonic events (after Ziegler 1982) in western Europe and North Sea Basin, and are given for comparison. Also indicated are timings of onset of sea-floor spreading in the Atlantic domain. Paleo-stresses are plotted relative to present-day stress level. Thin and heavy wavy lines denote minor and major rifting phases; thin and thick zig-zag lines represent minor and major folding phases; stars denote volcanic activity (after Lambeck *et al.* in press).

Directions for Future Research

The missing link for understanding basin tectonics may turn out to be the interaction of intraplate stresses with basin subsidence. A good understanding of the interaction of intraplate stresses and basin subsidence is thus possibly a crucial element in basin analysis. To incorporate intraplate stresses into basin analysis requires prediction of intraplate stresses from knowledge of plate-tectonic evolution and prediction of the effect of intraplate stresses on tectonic subsidence. Current research is continuing on these topics.

Systematic numerical modelling of paleo-stress fields in the various plates, based on reconstructed geometries and age distributions, is in progress now at the Vening Meinesz Laboratory. The outcome of this research program will be a set of paleo-stress maps for various time slices of the geologic record. The calculated stresses are to be tested by paleo-stress directions inferred from micro-structural analysis. Similarly, the stress patterns will be compared with the timing and the location of phenomena of intraplate tectonic deformation on a larger scale, including for example the inception of rift zones. Independently, paleo-stresses will be derived from the stratigraphic record under the assumption that short-term onlaps and offlaps are caused by fluctuating stress fields. Comparison of the outcome of the two different approaches will in principle allow us to test for the relative importance of intraplate stresses in the stratigraphic record and is expected to be of great value for separating global and primarily eustatic effects from more regional tectonic effects. Furthermore, detailed comparison of paleo-stress calculations with pertinent geological observations might provide better constraints on how fast the individual stress changes actually can occur. Equally important in this respect is the separation of sudden changes in stress (recorded by, for example, sudden depth changes) and stress changes continuous over tens of millions of years.

The motions of a flexural bulge are caused by an interplay of several factors, of which the level of intraplate stresses seems to be an important one. As noted previously, several effects not considered in the present analysis will influence the magnitude of the predicted short-term phases of subsidence or uplift. The effect of a depth-dependent rheology of the lithosphere will enhance the effectiveness of intraplate stresses as a cause of vertical motions, in particular for large values of stress. Note, however, that such a rheology inherently decreases the flexural wavelength of the basement response and hence will reduce the horizontal areal expression of the induced short-term phases of coastal onlap and offlap. For stress levels below a kilobar or so, these consequences of a depth-dependent rheology are probably quite modest, since for small stresses only a relatively thin segment of the plate will reach its limiting stress. In addition to these effects, relaxation of bending stresses by nonlinear ductile flow (Quinlan and Beaumont 1984) is accelerated by intraplate stresses, which causes tectonic subsidence by allowing further isostatic reequilibration of crustal and sedimentary loads.

Reactivation of faults under the influence of intraplate stresses is certainly an important aspect to be incorporated in the modelling. This applies in particular to the earlier phases of basin evolution, in which subsidence is dominantly fault controlled. Although modelling the response of faulted basins to intraplate stresses in terms of flexure of unbroken plates is, strictly speaking, not valid, the results can be used in a qualitative way as long as the faults are formed in an extensional regime. This is true for epicratonic basins and passive margins, which are commonly formed in an extensional regime. In this case, reactivation of extensional faults under compression will promote relative uplift of the segment of the basin close to the major boundary faults that govern the overall basin structure. The following examples might illustrate this. Completion of the rifting phase of basins is associated with relaxation of tensional stresses and a transition to a more compressional regime. The transition from the synrift phase to the postrift phase of basins is commonly reflected in the presence of a break-up unconformity, which agrees with the qualitative prediction of the model discussed in this paper. Break-up faults also play a crucial role in determining the response of the basin to inversion tectonics at later stages of its evolution (e.g., Ziegler 1987). It has been mentioned here that the simple model used by Cloetingh *et al.* (1985) provides only a very qualitative description which has to be refined in order to explore more quantitatively the dynamics of the actual inversion process. In principle, finite-element techniques such as used in the present analysis allow incorporation of pre-existing faults in the modelling. However, the success of more detailed modelling will be critically dependent on better knowledge of the actual strength of fault zones in the lithosphere under the action of intraplate stresses. Such insight is also crucial for relating sudden, probably episodic, displacements on basin faults to changes in stress of a possibly more continuous character.

Conclusions

The incorporation of intraplate stresses in quantitative models of basin evolution offers new perspectives in the analysis of sedimentary basins. Numerical modelling and observations of modern and ancient stress fields have demonstrated the existence of large stress provinces in the interiors of the plates. These studies have provided strong evidence for far-field propagation of intraplate stresses and established a causal relationship between processes at plate boundaries and deformation in the plate interiors, where stresses affect sedimentary basins both in passive-margin and intracratonic settings. Spatial and temporal fluctuations in stress fields apparently are recorded in the stratigraphic record of sedimentary basins. Numerical modelling has shown that the incorporation of intraplate stresses in models of basin evolution can in principle predict a succession of onlap and offlap patterns frequently observed along the flanks of passive margins and intracratonic basins. Such a stratigraphy can be interpreted as the natural consequence of the short-term mechanical widening and narrowing of basins by moderate fluctuations in intraplate stresses, superimposed on the long-term broadening of the basins with cooling since their formation.

Deformation of the lithosphere under the influence of large stresses generated by plate collision provides an explanation for observed syndepositional folding of oceanic lithosphere, the interaction of foreland peripheral bulges, and the inversion of basins as observed in the Alpine foreland. The seismic-stratigraphic record could provide a new source of information for paleo-stress fields. More detailed examination of the stratigraphic record of individual basins in connection with independent numerical modelling of paleo-stresses and analysis of paleo-stress data is required to exploit fully the role of intraplate stresses in basin evolution.

Acknowledgments. Kurt Lambeck and Herb McQueen made significant contributions to the modelling of the tectonic causes of sea-level changes. Rinus Wortel is thanked for stimulating cooperation in numerical modelling of intraplate stresses. Julian Thorne, Seth Stein, Gordon Shudofsky, Kerry Gallagher, and Peter Ziegler offered valuable suggestions and comments. George de Vries Klein, Anthony Hallam, Larry Sloss, Seymour Schlanger, and Peter Ziegler furnished preprints of papers. Gerard Bond, Julian Thorne, and R.H. Dott Jr. provided thoughtful reviews. The Netherlands Organization for the Advancement of Pure Research (ZWO) is acknowledged for partial support of this research.

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