Chapter 13

Tectonics and Sea-level Changes: a Controversy?

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ABSTRACT

Numerical modelling and observations of modern and paleo-stress fields demonstrate the existence of stress provinces of great areal extent in the interiors of plates. These tectonic stresses are of fundamental importance in basin analysis as a cause for short-term deviations from long-term patterns of thermal subsidence. Fluctuations in tectonic stress levels in the lithosphere 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 tectonic stresses in quantitative models of basin stratigraphy can predict a succession of offlap and onlap patterns similar to those observed in seismic-stratigraphic studies. Such a punctuated stratigraphy can be interpreted as the natural consequence of short-term changes in basin shape by moderate fluctuations in tectonic stresses, superimposed on long-term broadening of the basin with cooling since its formation. Hence, regional and global tectonics might contribute significantly to the record of sea-level changes. The “controversy” is not to choose between tectonics and eustatics as a cause for short-term sea-level changes but rather the challenge is to quantify the changing roles and interplay of these mechanisms through geological time.

INTRODUCTION

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 not so much in terms of the development of new modelling techniques and insights into the rheological makeup of the lithosphere but rather in the processing of new, high-quality data sets from previously unexplored areas of the globe. Virtually all modelling carried out so far has been in terms of lithospheric displacements, thus refraining from a full examination of the dynamic controls exerted by tectonic stress. This is because stresses are very sensitive to adopted lithospheric rheologies and these rheologies have been by convention unrealistically simple. This is true for models of both extensional and compressional sedimentary basins. 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. Moreover, 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 (Cloetingh et al., 1985; Cloetingh, 1986). Similarly, models of basin development in compressional environments have been conventionally related to flexure profiles, again not invoking the dynamic control of the compressional stresses intrinsic to this particular tectonic setting. Another reason that the relationship between lithospheric stress and displacement in tectonic modelling has not received full attention is because 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 carried out by the International Lithosphere Program (Zoback and World Stress Map Team, 1989). Further, the application of structural techniques to establish the temporal evolution in paleo-stress patterns has begun in a number of sedimentary basins (e.g. Letouzey, 1986). Simultaneously, numerical modelling (e.g. Cloetingh and Wortel, 1985, 1986) has resulted in a better understanding of the causes of the observed variation in stress level in the various lithospheric plates. As a result of these efforts considerable progress has been made in the study of the stress field within lithospheric plates.

The stretching model for basin formation (McKeezie, 1978) provides a simple and elegant explanation for the succession of a rapid syn-rift phase of basin subsidence by a long-term phase of subsidence caused by subsequent cooling of the lithosphere. As discussed by Watts et al. (1982) models for lithosphere cooling are useful to explain the long-term phases of subsidence documented at rifted margins. The advent of quantitative stratigraphic techniques has led to the construction of a set of charts of cyclic changes in sea level by Vail and coworkers (Vail et al., 1977; Haq et al., 1987, 1988). Almost a decade has passed by since these key concepts were presented in the Vail et al. (1977) paper and much work has been done in testing, evaluating and developing these ideas. In the present chapter, we discuss these basic concepts in sedimentary basin analysis in the light of recent theoretical advances in lithospheric dynamics. Studies of the tectonic stress field in the plates have shown a causal relationship between processes at plate boundaries and deformation in the plate's interiors. In models of the evolution of sedimentary basins located in the interiors of the plates, however, the role of these tectonic stresses has been largely ignored. Only recently the first steps have been made in exploring the consequences of the existence of tectonic stress fields in the lithosphere for models of the formation and evolution of sedimentary basins, shedding new light on the controversy on the causes of sea-level change (Cloetingh, 1988).

The question whether globally synchronous sea-level changes really do exist has been one of the most exciting scientific events of the past decade. The recognition of the effects of sea-level change has revolutionized the interpretation of marine seismic data from sedimentary environments, whereas the fundamental and widespread interaction of sea-level change with other marine and atmospheric phenomena is unique (Watkins, 1989). As discussed by Watkins (1989), sea-level changes have been shown to correlate with changes in CCD, oxygen and carbon isotope composition, faunal productivity and distribution, silica diagenesis, climate, deep oceanic circulation, seismic reflectors in carbonate sediments and preservation of organic carbon. It is often not clear whether these processes are the direct result of sea-level changes or are caused by observed phenomena, like tectonics, that also affect sea level, but it is clear that the topic of sea-level change is an important scientific problem in modern geology (Watkins, 1989). The problem of the causes of sea-level change is at the same time also the subject of a major controversy between advocates of a eustatic mechanism coupled to waning and waxing of ice sheets (e.g. Vail et al.,

Vail et al. (1977) interpreted short-term cycles in sea-level changes in terms of a traditional glacio-eustatic explanation. This view was partly based on the inferred global character of the sea-level cycles and was partly due to the lack of a tectonic mechanism to explain both the rate and magnitude of the third-order cycles. The issue of global synchronity has attracted major debate (e.g. Miall, 1986; Hubbard, 1988). Several authors have noted that Vail's cycles, although based on data from different basins around the world, are heavily weighted in favour of the North Sea and the northern/central Atlantic margins. The issue of global synchronity is important as it obviously strongly influences present discussions on the causes of short-term changes in sea level. As noted by Pitman and Golovchenko (1983), 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 (Kominz, 1984) and are therefore associated with a long thermal inertia of several tens of millions years (Table 13.1). 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 to suggest low altitude glaciation. This presents a fundamental problem of explaining eustatic sea-level changes in the Vail et al. (1977) and Haq et al. (1987, 1988) charts at times prior to the Late Cenozoic. The second problem is the inability of glacio-eustatics to cause uniform lowerings and rises of sea level. The sign and magnitude of the induced sea-level change is dependent on the distance to the location of the ice cap. This feature, well known in circles of modellers of post-glacial rebound processes (e.g. Lambeck et al., 1987), is unfortunately not always fully appreciated by those who advocate glacio-eustasy as the key mechanism to explain global synchronous changes of uniform magnitude in sea level.

It has been shown that short-term changes in relative sea level can equally well be caused by rapid, stress-induced vertical motions of the lithosphere within sedimentary basins (Cloetingh et al., 1985; Cloetingh, 1986). This work showed that tectonic

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stresses, apart from being important in the formation of rifted basins, also play a critical role during their subsequent subsidence history. We first review evidence for the existence of tectonic stress fields in the lithosphere. This is followed by a discussion of some implications of tectonic stress for quantitative modelling of sedimentary basins. Finally the potential to separate the tectonic contribution from the eustatic signal in the record of relative sea-level changes is explored. The "controversy" is not to choose between a tectonic mechanism and a eustatic one in explaining the causes of short-term sea-level changes but rather the challenge is to discriminate their relative contributions.

TEC TONIC STRESSES IN THE PLATES

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 observed modern stress orientations show a remarkably consistent pattern, especially considering the heterogeneity in intraplate lithospheric structure. 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 of the International Lithosphere Program has convincingly established the existence of these large-scale, consistently oriented stress patterns as a general characteristic of lithospheric plates (Zoback and World Stress Map Team, 1989). These observations indicate that regional stress fields are dominated by the effect of plate-tectonic forces acting on the lithosphere. Figure 13.1 shows a compilation of the World Stress Map.

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 (Letouzey, 1986; Bergerat, 1987; Philip, 1987; De Ruig, 1990). The results of a recent compilation of stress direction data for the European platform are displayed in Fig. 13.2. In this case, the orientations and evolution of the principal paleo-stress axes are inferred from microstructure measurements in sedimentary rocks, such as pressure solution surfaces (stylolites), veins with secondary mineralizations, or small faults with a clear indication of the sense of displacement (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. These stresses are propagated from the plate boundaries into the interiors of the plates where they affect the vertical motions within sedimentary basins (Cloetingh, 1988; Nemec, 1988). An example of this is provided by the stress field in the northwestern European platform where studies of borehole elongations from oil wells and of earthquake mechanisms have revealed an orientation of the stress field (Fig. 13.2) that seems to be dominated by the effect of Europe/Africa collision and the contribution of ridge push forces from ocean-floor spreading in the Atlantic (Le Pichon et al., 1988; Zoback and World Stress Map Team, 1989).

TEC TONIC STRESSES AND STRATIGRAPHY OF RIFTED CONTINENTAL MARGINS

Tectonic stresses modulate the long-term basin deflection caused by thermal subsidence and induce rapid differential vertical motions of a sign and magnitude
Figure 13.2  Compilation of observed maximum horizontal stress directions in the European platform. (a) Present-day stress field. (b) Paleo-stress field during Late Eocene times. The data indicate stress propagation away from the Alpine fold belt in the platform region (after LePichon et al., 1988).
that depends on the position within the basin (Fig. 13.3). For example, a compressional stress state 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, lowering of the flanks, and cause landward migration of the shoreline, producing a rapid onlap phase (Fig. 13.4). 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 (e.g. Galloway, 1989). Similarly, pre-existing sub-basinal fractures/faults can be reactivated to cause intrabasinal "noise" in overall subsidence pattern (see Nemec, 1988).

Modelling of the stratigraphy of the US Atlantic margin (Fig. 13.5; see also Cloetingh et al., 1989) and the North Sea (Lambeck et al., 1987; Kooi and Cloetingh,
Tectonics and Sea-level Changes: a Controversy?

1990) has shown that the punctuated stratigraphy can be successfully simulated by a model in which a stress field whose magnitude fluctuates through time, is super-imposed on the long-term thermal evolution. The inferred paleo-stress was found to be largely consistent with independent data sets on the kinematic (Kitgord and Schouten, 1986) and tectonic evolution (Letouzey, 1986; Ziegler, 1989a; Ziegler and Van Hoorn, 1990) of the northern/central Atlantic with a tensional stress field during Mesozoic times followed during the Tertiary by a compressional stress field whose magnitude increases with time. These findings are largely in agreement with the results of stratigraphic work by Hubbard (1988) on the timing and nature of megasequence and sequence boundaries of rifted basins around the Atlantic and the Arctic. Hubbard recognized that a significant part of the Mesozoic sequence and megasequence boundaries is associated with large-scale, fault-controlled rifting activity, while Cenozoic sequence boundaries in the Arctic were mainly controlled by large-scale compressional activity and inversion tectonics (Fig. 13.6). Hallam (1988) also recognized the apparently 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 recorded in the Lower Jurassic of the North Sea.

Similarly, detailed analysis of the Canadian Atlantic margin (Cloetingh et al., 1990b) strongly suggests that the bulk of Mesozoic sequence boundaries reflects the adjustment to stress changes that occur during rifting in the north-central Atlantic. The tectonic and structural evolution of the eastern Canadian continental margin suggests tectonic linkage between continental crust and oceanic crust. This linkage is reflected in the history of extensional subsidence of the Mesozoic basins underlying the Grand Banks and the timing and patterns of sea-floor spreading. On a large scale, unconformity-bound sequences match major basin-forming stages representing syn-rift and post-rift subsidence. On a smaller scale, local unconformities and regional limestone markers reflect adjustment to intermittent subsidence. Intermittent phases of accumulated tectonic stress, associated with rift episodes in the northern and central Atlantic, and subsequent rapid relaxation of these stresses explain the asymmetry in Vail’s onlap–offlap charts. Both the timing and nature of Vail’s second-order and third-order cycles appear to a large extent to be controlled by plate tectonic evolution and the associated changes in stress regimes of the northern and central Atlantic.

**TECTONIC STRESSES AND FORELAND AND INTRACRATONIC BASIN EVOLUTION**

In the previous section we have concentrated on the relationship between tectonics and stratigraphy at rifted margins. The effect of tectonic stress is also quite important in compressional settings. Of primary interest in this respect are foreland basins,

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**Figure 13.4** 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 sedimentary package bounded by isochrons of 50 Ma and 52 Ma after basin formation. (a) Continuous onlap associated with long-term cooling of the lithosphere in the absence of stress fields. (b) A transition to 500 bar in-plane compression at 50 Ma induces uplift of the peripheral bulge, narrowing of the basin and a phase of rapid onlap, which is followed by a long-term phase of gradual onlap due to thermal subsidence. (c) A transition to 500 bar in-plane tension at 50 Ma induces downwarp of the peripheral bulge, widening of the basin and a phase of rapid basement onlap.
where lithosphere is flexed down under the influence of thrust sheet and sedimentary loads (Quinlan and Beaumont, 1984; Zoetemeijer et al., 1990). Despite its height of at most a few hundred metres the peripheral bulge flanking foreland basins is of particular interest (Quinlan and Beaumont, 1984; Tankard, 1986). These authors interpreted the development of unconformities such as in the Appalachian foreland basin in terms of uplift of a peripheral bulge caused by viscoelastic relaxation of the lithosphere (Fig. 13.7). However, the presence of tectonic stresses can greatly reduce or amplify the height of the peripheral bulge and thus greatly influence the stratigraphic record at foreland basins. For example, incorporation of tectonic stresses into elastic models of foreland and cratonic basin evolution explains key features of the Late Carboniferous tectonic evolution of North America, which were hitherto interpreted in terms of more complex viscoelastic models (Quinlan and Beaumont, 1984). With increasing horizontal compression, flexural bulges are amplified, reaching high elevation when lithospheric stresses approach lithospheric strength (Cloetingh et al., 1989). Conversely, during reduction of stress levels in the lithosphere by thrusting events, or on a larger scale by changes at convergent zones, the magnitudes of flexural bulges are diminished (Cloetingh et al., 1989). Temporal changes in level of E-W oriented compression generated by plate collision along the Appalachian front successfully explain the spacing and time-varying patterns of flexural arches during Late Carboniferous time (McKee and Crosby, 1975; Ziegler, 1989b). During Ouachita orogen activity, tectonic stress with a more N-S oriented direction of compression became a more crucial component in the control of large-scale tectonic evolution. When basin geometry is characterized by a N-S spacing close to the flexural wavelength of the lithosphere, the response of the basin to an increasing level of N-S compression is a deepening of the basin centre and peripheral uplift of the basin flanks with development of intervening basement arches. Occurrence of strong phases of tectonic subsidence deviating from thermal subsidence patterns within platforms and cratonic basins at great distances from convergent plate boundaries in the interior of north America is time-equivalent with changes in tectonic regimes along the collision margin (Klein and Cloetingh, 1989; Bond and Komín, 1990). These features suggest a strong coupling of tectonic processes at plate boundaries and interior basins and platforms. Analysis of fracture orientations in Paleozoic carbonate rocks of east-central North America (Cradock and van der Pluijm, 1989) also provides strong evidence for far-field propagation of tectonic stress, as well as the extensive analysis of platform basins in NW Europe where the role of far-field Alpine tectonics has been abundantly demonstrated (Ziegler, 1987). A characteristic feature of the effect of stresses on basin evolution is the predicted variation in the differential uplift and subsidence patterns with respect to position within the basin. There is some evidence in support of this feature of the model. For example, late-stage subsidence is characterized by acceleration near the basin centres, whereas at basin flanks, subsidence is observed to slow down (Némec, 1988). Observations from subsidence patterns associated with Late Carboniferous cyclothem are also in agreement with the predictions of the model (Klein and Cloetingh, 1989).

Figure 13.5 Schematic model of the stratigraphy of the US Atlantic margin at Cape Hateras. (a) Modelled stratigraphy for elastic rheology of the lithosphere in the absence of tectonic stresses. (b) Modelled stratigraphy in the absence of tectonic stresses but adopting long-term changes in sea level after Komín (1984). (c) Modelled stratigraphy showing the combined effect of long-term changes in sea level and a fluctuating tectonic stress in the analysis (after Cloetingh et al., 1989).
Figure 13.6 Timing and nature of megasequences and sequence boundaries of rifted basins around the Atlantic and in the Arctic (after Hubbard, 1988).
Short-term deviations from long-term patterns of tectonic subsidence observed on backstripped subsidence curves are a natural consequence of changes in stress levels within tectonic plates (Cloetingh, 1988) and were observed, for instance in the Illinois basin (Watso and Klein, 1989). Perturbations in the long-term patterns of tectonic subsidence reflect the temporal variation in, for example, the processes of repeated thrusting in a foreland basin, or repeated faulting during thermal subsidence in a cratonic basin. The scaling of the magnitude of the inferred variation in paleo-stress levels is, however, dependent on the rheology of the lithosphere (Kooi and Cloetingh, 1989; Cloetingh et al., 1989). The order of magnitude of the stress perturbations inferred from analysis of subsidence patterns associated with cyclothem deposition (Klein and Cloetingh, 1989) is compatible with estimates of stress drops of earthquakes (Aki and Richards, 1980) and stress changes associated with the mechanics of fault motion (De Bremaecker, 1987). The analogy of accumulating stresses interrupted by discrete phases of stress relaxation (Fig. 13.8) compares well with the stress patterns associated with earthquake generation where the level of stress drops (being the change in stress) is significantly lower than the level of lithospheric stress itself. Hence, the temporal and spatial evolution of the basin and its stress history inferred from subsidence analysis provide an internally consistent explanation of the coupling between for example foreland basin stratigraphy and the mechanics of thrust motion.

**TECTONIC STRESSES AND HIGH-AMPLITUDE COMPRESSIONAL DEFORMATION**

Compared to an elastic plate, a plate with a depth-dependent brittle-ductile rheology will by its finite strength, and lower flexural rigidity, be more sensitive to tectonic
ELASTIC MODEL WITH INTRAPLATE STRESSES AND FORELAND BULGE EVOLUTION

Figure 13.8  Accumulation of compressional stresses in foreland basin interrupted by episodic phases of stress relaxation associated with discrete thrusting. Compressional stresses build up to a threshold at which movement of thrust sheets is initiated, causing migration of the flexural bulge, changes in sea level and deposition of transgressive/regressive cycles.

stresses. This feature is illustrated in Fig. 13.9, which shows the magnification of the vertical deflection resulting from the incorporation of a depth-dependent rheology in the modelling. Inspection of Fig. 13.9 demonstrates that the effect of tectonic stresses on the height of the flexural bulge is substantially larger in the case of a depth-dependent rheology than in the case of an elastic rheology of the lithosphere. Figure 13.10 displays the dependence of the vertical motions of the peripheral flexural bulge on the thermal age of the lithosphere. The magnitudes of the stress-induced vertical motions are calculated for rheological models based on the extrapolation of rock mechanics data for both oceanic (Fig. 13.10a) and continental lithosphere (Fig. 13.10b,c). It is obvious from inspection of these figures that the stress-induced deflection rapidly increases for stresses approaching lithospheric strength. As such the stress-induced vertical motions cover a wide spectrum of magnitudes, from values of the order of a few tens of metres to a hundred metres or so (characteristic for variations in relative sea level) to kilometres (characteristic for folding of the lithosphere which occurs as a result of accumulation of stress levels close to the strength of the lithosphere). Lithospheric folding will occur in cases of extremely high compressional stresses induced in the lithosphere in connection
with a unique dynamic situation. An example of folding due to collision processes is presently occurring in the northeastern Indian Ocean.

The northeastern Indian Ocean is probably the most active region of present-day intraplate deformation (Stein et al., 1989, 1990). In the central Indian Ocean, sediments and acoustic basement are deformed into long-wavelength (~200 km) undulations associated with large gravity and geoid anomalies (Fig. 13.11). Superimposed on the folds are closely spaced (~5–20 km) faults (Fig. 13.12). The orientation of the fold axes changes from east–west in the Central Indian basin to ENE-WSW on the Ninetyeast Ridge, to NE-SW in the Warton Basin (Fig. 13.11). The orientation and spatial distribution of the folds and the directions of maximum and minimum compression from earthquake focal mechanism solutions are in good agreement with the signs and directions of the stresses calculated for the Indo-Australian plate (Cloetingh and Wortel, 1985, 1986). The stress model by Cloetingh and Wortel predicts large (on the order of several kilobars) compressional stresses in the northeastern Indian Ocean. However, it is more difficult to determine the magnitude of the stresses required to produce the observed deformations. The same order of magnitude for stress is required when modelling the wavelengths of folding (Stephenson and Cloetingh, 1990, see also Fig. 13.10). Similar stresses can also be inferred from the depth of seismicity, which often occurs at depths where the lithosphere should have considerable strength (Stein et al., 1990). If faulting requires that the ambient stress equals or exceeds that expected on rheological grounds, the focal depths imply stresses in good agreement with those predicted (Govers et al., 1990).

**TECTONICS AND SEDIMENTATION: UNCONFORMITY GENERATION IN THE DEEP SEA**

Stress-induced vertical motions within sedimentary basins also have important consequences for the dynamics of erosion and sedimentation. For example, the
compression-induced uplift at the basin flank is accompanied by enhanced erosion of the uplifted area and increased sedimentation in the basin centre (Fig. 13.13). On the other hand, a decrease in sedimentation might also occur when the uplift area induces significant changes in the drainage pattern of the hinterland. Stress-induced uplift at the basin flank changes the geometry of sediment sources and sinks. There has been much discussion on the linkage and correlation between margin unconformities and deep-sea unconformities (e.g. Winterer, this volume). As has been pointed out by several authors, the existence of a correlation between these unconformities is not proof of a causal link. Also, because of the large number of deep-sea unconformities, it has been pointed out that a correlation of a subset of these with margin unconformities is in principle always possible (Winterer, this volume). Traditionally, deep-sea unconformities have been interpreted primarily in terms of eustatic. However, as noted by Pickering et al. (1989), tectonic stress might also generate deep-sea unconformities. For example, stress-induced changes in shape of the margin will possibly trigger erosion and deposition by turbidity currents, perhaps in association with canyon cutting. (Miller et al., 1987), producing unconformities that should be traceable from the margin into the deeper parts of the

Figure 13.10 Stress-induced uplift at basin flank. (a) Deflections adopting a depth-dependent oceanic rheology of the lithosphere (Goetze and Evans, 1979). (b), (c) Deflections adopting depth-dependent continental rheology of the lithosphere for a quartz/diabase/olivine and a quartz/diorite/olivine petrological layering of the lithosphere (Carter and Tsenn, 1987) (after Stephenson and Cloetingh, 1990).
Figure 13.11 Intraplate earthquakes, gravity anomalies, and present-day stresses predicted by finite element modelling. Gravity highs corresponding to the folded oceanic lithosphere generally trend normal to the predicted compressive stress (after Stein et al., 1989).

basin. Similarly, significant distortion of the basin shape at the flanks of passive margins during plate reorganizations could also give rise to changes in water circulation such as that documented by Dillon and Popenoe (1988) for the US Atlantic margin with possible consequences for climatic changes.

On seismic reflection records, folding of lithosphere is most clearly seen in the Central Indian Ocean Basin because of the thick accumulation of turbidite sediments in the Bay of Bengal originating from material eroded from the Himalayan mountains and transported to the Bay of Bengal by river systems. Within the region of folding in the Central Indian Ocean Basin the sediment thickness ranges from greater than ~3 km in the north to less than ~0.5 km in the south (Cochran et al., 1989; Stein et al., 1990). A widespread unconformity marks a surface separating folded strata below from relatively flatlying, but faulted, sediment above. This has been dated at 7.5 Ma (ODP Leg 116 sites, Cochran et al., 1989). This unconformity can be traced throughout much of the region of the observed folding (Fig. 13.11). Since all the layers within the folds appear to have been folded about equally (Fig. 13.12b), the folding probably occurred relatively rapidly, without time for large-scale syntectonic deposition. Had major folding occurred over a relatively long time interval then the stratigraphy would be quite different. Only in a few locations are the sediments above the unconformity upturned, related either to a mild subsequent folding episode or uplift to local faulting (Stein et al., 1990). This suggests that the sediments above the unconformity are deposited after the major folding phase ended. Hence it is likely that the unconformity does not present the initiation of deformation but
Figure 13.12 (a) Long seismic reflection profile from the southern Bengal Fan showing folding of oceanic lithosphere and numerous reverse faults. Time is given along the vertical scale in terms of seconds of two-way travel time (after Stein et al., 1990). (b) Seismic reflection profile at ODP Leg 116 showing the fine structure of the compressional deformation and reverse faulting together with the presence of the unconformities (after Cochran et al., 1989).

rather the end of deformation by folding and presumably since then compressional deformation has occurred mainly by reverse faulting.

However, as spectacular as the folds are, they represent only a very small amount of shortening and most of the compression and deformation has been

Figure 13.13 Effect of tectonic compression on basin subsidence/uplift and associated sedimentation and erosion.
accommodated by faulting (Stein et al., 1990). An examination of the seismic reflection profile records (C. A. Stein, 1989 pers. comm.) suggests that the sediments in the Central Indian Basin are not continuing to be folded. Since the sediments on top of the acoustic basement often have faults extending to the surface, faulting must have occurred recently.

The period of time during which major folding occurred is uncertain, but the stratigraphy of the oceanic sediments suggests that folding occurred over a relatively short time interval, probably not significantly earlier than 7.5–8 Ma. Alternatively, if the folded region, especially the lows, did not receive much sedimentation after the initiation of the folding, then the phase of folding may have lasted longer. It has been noted that once the lithosphere was folded the sediment patterns were locally changed (Stein et al., 1990). Some of the elevated regions presumably blocked the flow of turbidites from the north. In some regions, the younger sediment has since completely covered the folds. In others, the fold highs have not yet been covered and have received little sediment since uplift. Only at ODP Leg 116 is there a stratigraphic record through the unconformity. Here, pre-and post-unconformity sediment rates and sediment types appear to be similar (Cochran et al., 1989). Hence it is likely that major parts of the floor of the central Indian Ocean continued to receive sufficient sediment supply after the initiation of the folding, so if folding occurred over a very long time period, this should be prominently reflected in the stratigraphy. It should be noted that the unconformity between the folded (below) and mainly flat-lying faulted sediments (above) is only clearly seen due to the large sedimentation rates before and after the end of folding (Stein et al., 1990). In particular, after the end of the major folding episode, with sedimentation rates closer to typical deep-sea ones, we would not be able to clearly identify this unconformity. Cochran et al. (1989) have identified some major unconformities, both of a structural nature and sedimentation type, in the sedimentary record in the last 17 Ma. The main controls on the sedimentation appear to be the uplift and erosion history of the Himalayas, the position of sea level relative to the shelf edge, paleoceanography and the complications due to fan growing processes. The unconformity 8–7.5 Ma in the Central Indian Basin separating folded from relatively flat-lying sediments is related to the large compressive stresses within the Indo-Australian plate (Fig. 13.11), but is not accompanied by a change in sediment type. The next unconformity 4.6–5.4 Ma appears related to a rise in relative sea level with a change in sediment type. The next unconformity at 0.8 Ma appears to be related to the Plio–Pleistocene glaciation resulting in a significant lowering of sea level with an exposure of the Indian continental shelf.

Apparent sea-level changes along the Indian margin and northwestern Australian margins should have occurred, given the high stress levels predicted. Unfortunately, little information has been published in a format useful for analysis and the Indian sea-level studies are complicated by the Himalayan uplift and fan deposition. Pandey (1986) suggests that the eastern Indian shelf experienced major uplift and regression at about 12 Ma ago. New ODP data for the Northwestern Australian shelf hold the promise of shedding light on this issue and the key-question of global synchronity of the sea-level record. Given that the stress required to produce relative sea-level variations at rifted margins is an order of magnitude less than that to fold the Indian ocean lithosphere, and if the stress slowly increased with time, the continental margins might record the stress changes prior to the 7.5 Ma (Cochran et al., 1989) onset of the deformation.

Linking the other unconformities dated at Leg 116 to tectonic events or even to the stratigraphic unconformities of the ocean’s margins is a difficult problem. It is often quite difficult to correlate in a meaningful sense deep-sea conformities with those at
the continental margin. Traditionally, the marine unconformities have been related to eustatic sea-level changes, although the correlation is not very good. The marine sedimentation is quite sensitive (Pickering et al., 1989) to changes in shape of the margin that effect erosion/deposition (e.g. turbidity currents/canyon cutting) as demonstrated by Miller et al. (1987) for the Atlantic Ocean.

Also, changes in the shape of the margin will affect paleoceanography and hence the sedimentation patterns (Cloetingh et al., 1990a). Until recently the effect of changing the tectonic stress field has not been considered a major factor affecting the continental margin stratigraphy and hence perhaps also the deep-sea record.

**DISCRIMINATION OF TECTONIC AND EUSTATIC SIGNALS**

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 critical, unbiased stratigraphic analysis is on the spatial and temporal differences in the expressions of tectonic processes and eustatics. The development of stratigraphic criteria to differentiate between tectonics and eustasy is therefore vital. This should occur both on an interbasinal and an intrabasinal scale (see also Nemec, 1988). An important step in that direction has been made by Embry and coworkers (Embry, 1990; Mork et al., 1989). These authors recognized a number of features in the stratigraphic record of rifted basins that were hard to explain in terms of the eustatic framework.

Amongst these, the following are of particular interest:

- sediment source areas that often vary greatly from one sequence to the next;
- sedimentary regimes that change drastically and abruptly across a sequence boundary;
- faults that terminate at sequence boundaries;
- significant changes in subsidence and uplift patterns that occur within basins across sequence boundaries;
- sequence boundaries that are absent in parts of basins; and
- significant differences in the magnitude and extent of some of the subaereal unconformities and time-equivalents recognized by Vail et al. (1977)

These criteria have been successfully applied to the Canadian Sverdrup Basin (Embry, 1990) as well as to the Svalbard basin and the Barents shelf (Mork et al., 1989). The discrimination of tectonics and eustasy is a subtle matter, especially if biostratigraphic resolution is limited. The regional character of tectonic 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 reorganizations in tectonic stress fields occur simultaneously in more than one plate (e.g. Pollitz, 1988) or in time intervals prior to and shortly after break-up of Pangaea where continents and rift basins were in a largely uniform stress regime (Dewey, 1988). The magnitude of the stress-induced phases of rapid uplift and subsidence varies with position within the basin, thus providing another important criterion to separate this contribution from eustatic effects. Similarly, differences in rheological structure of the lithosphere control the magnitude of the vertical motions. The presence of weak attenuated continental lithosphere enhances the effectivity of the stresses to cause substantial vertical motions and could explain
differences in magnitude such as those observed (Vail et al., 1977) between the Tertiary North Sea and the Gippsland Basin off southeast Australia.

**EXTRACTION OF SEA-LEVEL CHANGES FROM THE SUBSIDENCE RECORD**

The McKenzie (1978) 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 are assumed to relax. An essential assumption made in the 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 (Cloetingh and Nieuwland, 1984; Sawyer, 1985a) has demonstrated that tensional stresses of the order of several kilobars are required to stretch continental lithosphere. Important in this respect have been the theoretical advances in lithospheric rheology based on extrapolation of rock mechanics data (Brace and Kohlstedt, 1980).

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 could be explained in terms of multiple stretching phases (Greenlee et al., 1988). However, care should be taken with this interpretation as an increase in the level of tectonic compression can equally well produce this type of deviations from predictions of the thermal models 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 (Kooi and Cloetingh, 1989). 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. As demonstrated by Cloetingh et al. (1990a), late-stage compression can explain the rapid phases of Late Neogene subsidence such as those encountered around the northern Atlantic (Fig. 13.14). A prominent reorganization of spreading directions and rates occurred at 2.5 Ma along the entire Atlantic spreading centre (Klitgord and Schouten, 1986). Important tectonic phases during late Neogene times also occurred on the western side of the Atlantic Ocean. A climax in compressional tectonics in the Arctic of northern Alaska and northern Canada has occurred at about 6 Ma, possibly connected with the formation of an incipient convergent plate boundary (Hubbard et al., 1987). Similarly, the termination of extension in the Basin and Range province during Pliocene times coincides with important changes in the basin evolution in the Gulf of Mexico (Galloway, 1989). Episodicity in plate motions and associated changes in the Pacific and Atlantic plates in the Upper Miocene (9 Ma) and Pliocene (4 Ma) have been documented in great detail recently by Pollitz (1988). His analysis provides strong evidence of a mechanical coupling between the plates by showing, for example, that the causes of the Pacific and northern American changes in plate motion are related to plate driving forces originating in the northwestern Pacific subduction zones.

Recent work (Cloetingh et al., 1990a) strongly suggests that the sediments of the rifted basins around the Atlantic record a phase of intensive global compressional tectonics, associated with an important late Neogene plate reorganization of possibly global nature. These findings are also interesting in view of the partly overlapping occurrence of glaciation in Quaternary times. The onset of the observed acceleration
in basin subsidence occurs well before the first occurrence of glaciation (1.6–1.9 Ma). This observation and the differential character of the uplift and subsidence at different positions within the rifted basins around the northern Atlantic rules out glaciation as the main cause for the late Neogene subsidence phases. On the other hand, it is well known that periods of increased elevations promote the development of glaciation (e.g. Powell and Vevers, 1987). Although uplift in its own is only part of the dynamics of glaciation and changes in the air circulation patterns (Ruddiman and Raymo, 1988), the intimacy of glaciations and uplift seems to be more than casual. An interesting consequence of stress-induced downbending in the central North Sea could be an uplift of Norway and the British isles. Therefore, a causal link (Fig. 13.15) could exist between the accumulation of compression during late Neogene time inferred from the modelling of subsidence and stratigraphy of rifted margins around the northern Atlantic, and the onset of glaciation during Quaternary time. Thus the interesting perspective arises that uplift-induced glaciation by Himalayan and Rocky Mountain plateau uplift is augmented by uplift-induced
glaciation along the northern Atlantic itself. Such an explanation for the rapid Late Neogene accelerations in basin subsidence (Heller et al., 1982; Nilsen et al., 1986) is more likely than a phase of renewed stretching, as the present-day stress field in NW Europe and the northern Atlantic is compressional in character (Zoback and World Stress Map Team, 1989).

Stress-induced subsidence, therefore, can contribute substantially to the total tectonic subsidence. That is important because it is common practice to use total tectonic subsidence as a measure for the determination of values of crustal extension (e.g. Sawyer, 1985b). Another independent estimate of extension can be obtained by measuring horizontal displacements of the normal faults active during extension. In general, a discrepancy exists between these two estimates of crustal extension, as has been discussed for the North Sea basin (Sclater and Christie, 1980; Ziegler and Van Hoorn, 1990). Late-stage compression can strongly contribute to the total tectonic subsidence. Therefore, current back-stripping 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 (Kooi and Cloetingh, 1989). Stress-induced phases of short-term basin subsidence might intrinsically explain at least part of the discrepancies between the various estimates of crustal extension obtained from subsidence analysis and structural interpretations.

**DISCUSSION AND CONCLUSIONS**

Sedimentary basins form and evolve in the interiors of plates that are subject to episodic changes in tectonic regime. Recent work on intraplate tectonics has established the strong mechanical coupling between geodynamic processes at plate boundaries and the deformation 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 (Fig. 13.16). Similarly, effects of multiple stretching phases and increased levels of tectonic compression both create rapid vertical motions of a magnitude well beyond the range covered by apparent sea-level changes. Integrated analysis of the structural and stratigraphic evolution of sedimentary basins, together with the ongoing development of dynamic models for basin formation and evolution, are of great interest. Quantification of the relative contributions of the different mechanisms that are offered to explain the long-term and short-term frequency ranges of the complex spectrum of vertical motions in sedimentary basins should contribute to resolve the ongoing controversy on the causes of sea-level change.

Undoubtedly both the waxing and waning of ice sheets and tectonics are operating on time scales that sometimes interfere. The simple fact that one of these mechanisms operates during a certain time slice does not prove that the contribution of the other does not exist. Similarly, a coupling might exist between glaciation and tectonics. Plate tectonics can operate on a global scale (plate reorganizations) and on a more regional scale, which is quite important for putting the discussion on global synchronicity into a meaningful perspective. Discovering the fine structure of the mechanisms responsible for the non-glacial second and third order sea-level changes is considered by many investigators one of the most important issues in the area of sequence stratigraphy (Watkins, 1989). The ongoing development of tectonic models with predictive capabilities of global and regional amplitudes and durations of sea-level changes could provide a solid foundation for fruitful future discussions on the relative contributions of tectonics and eustasy to the record of sea-level change. With the recent advent of a tectonic mechanism operating on a time scale hitherto solely addressed by the eustatic mechanism, the discussion should evolve from a controversy between advocates of either eustasy and tectonics into a quantitative search for the changing roles and interplay of these mechanisms through geological time.

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Tectonics and Sea-level Changes: a Controversy?


Sea-level History and Sedimentation


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