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1 Tectonic subsidence and sea-level changes: a reappraisal

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Major advances in quantitative basin analysis have been made during the past decade. Lithospheric processes are not only crucial as a cause of long-term subsidence but are also of fundamental importance as a cause for short-term deviations from long-term patterns of thermal subsidence. Regional and global tectonics contribute significantly to the record of sea-level changes. Both rifting and compression can give rise to phases of rapid subsidence that were hitherto commonly attributed to renewed phases of crustal stretching. These findings allow a critical appraisal of some concepts in basin formation and basin evolution that are vital in modern analysis of tectonic subsidence and basin correlation.

INTRODUCTION

Two concepts that have shaped the past decade of quantitative analysis of the subsidence record of sedimentary basins are the McKenzie (1978) model for basin formation and Vail *et al.*'s (1977) framework of cyclic changes in sea level. The McKenzie (1978) stretching model provided a simple and elegant explanation for the succession of a rapid syn-rift phase of basin subsidence followed by a long-term phase of subsidence caused by subsequent cooling of the lithosphere. In earlier pioneering work, Sleep (1971) had already demonstrated the usefulness of models for lithospheric cooling to explain the phases of long-term subsidence documented at rifted margins. Simultaneously, the advent of quantitative stratigraphic techniques led to the construction of a set of charts of cyclic changes in sea level (Vail *et al.*, 1977). These cyclic short-term eustatic changes in sea level were offered as a mechanism to explain short-term deviations from long-term subsidence patterns (Watts and Steckler, 1979). Almost a decade has passed by since these key concepts were presented, and much work has been done in testing, evaluating and developing the basic ideas put forward by McKenzie and Vail. In the present chapter, we discuss the merits and shortcomings of these basic concepts in the light of recent theoretical advances in lithospheric dynamics.

SEA-LEVEL CHANGES AND BASIN STRATIGRAPHY

Vail *et al.* (1977) interpreted their short-term cycles of sea-level change in terms of a traditional glacio-eustatic

process. This view was partly based on the inferred global character of the inferred sea-level cycles and was partly due to the lack of a tectonic mechanism to explain both the rate and magnitude of their third-order (short-term) cycles. The issue of global synchronicity has attracted major debate (e.g. Parkinson and Summerhayes, 1985; Miall, 1986). 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 (see also Fig. 1). The issue of global synchronicity 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 and are therefore associated with a long thermal inertia of several tens of millions of years (Table I). Glacio-eustasy can easily 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) 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 (Fig. 2). This feature,

Fig. 1. Correlation of regional cycles of relative change in sea level from basins located on four continents used to construct global cycle charts (after Vail *et al.*, 1977).

Table I. Magnitudes and rates of sea-level changes by several mechanisms that were considered in the early 1980s as potential contributors to Vail *et al.*'s third-order cycles in sea level. Apart from glacio-eustasy, the proposed (tectonic) models failed to produce both the rate (1–10 cm/1000 yrs) and the magnitude (up to the order of 100 m). This feature is primarily caused by the long time constants of lithospheric thermal processes which grossly exceed the timescales (2–5 Ma) characteristic for short-term changes in relative sea-level (after Pitman and Golovchenko, 1983)

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

well known amongst modellers of postglacial rebound processes (e.g. Lambeck *et al.*, 1987; Fjeldskaar, 1989), is unfortunately not always fully appreciated by those who advocate glacio-eustasy as the key mechanism to explain global-synchronous changes in sea level of uniform magnitude.

More recently, 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 intraplate stresses, as well as being important in the formation of rifted basins, also play a critical role during their subsequent subsidence history. The ongoing World Stress Map Project of the International Lithosphere Program has convincingly established the existence of large-scale, consistently oriented stress patterns in the lithosphere (Zoback, 1989). Studies of, in particular, the French school of structural geology have provided compelling evidence for changes in the magnitudes and orientations of these stress fields on timescales of a few Ma (Letouzey, 1986; Bergerat, 1987; Philip, 1987) in association with

Fig. 2. Sea levels (in metres), relative to present-day sea level, in the North Sea 20 000 years ago, prior to the melting of the northern ice sheets. These results are based on a viscoelastic mantle of uniform viscosity and an elastic lithosphere of 50 km thickness (after Lambeck *et al.*, 1987).

collision and rifting processes in the lithosphere. These stresses are propagated away 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 that seems to be dominated by the effect of the Europe/Africa collision and the contribution of ridge push forces from ocean-floor spreading in the Atlantic (Klein and Barr, 1986; Zoback, 1989).

Intraplate stresses and basin stratigraphy

Intraplate stresses modulate the long-term basin deflection caused by thermal subsidence and induce rapid differential vertical motions of a sign and magnitude that depends on the position within the basin (Fig. 3). For example, intraplate compression causes relative uplift of

Fig. 3. Stress-induced subsidence/uplift at rifted basins. Above: a 60 Ma old rifted basin initiated by stretching. The wedge of sediments flexurally loads an elastic lithospheric plate. Below: differential subsidence or uplift (metres) induced by a change to 1 kbar compression (solid line) and 1 kbar tension (dashed line) (after Cloetingh *et al.*, 1985).

the basin flank, subsidence at the basin centre, and seaward migration of the shoreline. An offlap develops and an unconformity is produced. Increases in the level of tensional stress induce widening of the basin, lower the flanks and cause landward migration of the shoreline, producing a rapid onlap phase (Fig. 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, 1987). Similarly, pre-existing sub-basinal fractures/faults can be reactivated to cause intrabasinal 'noise' in the overall subsidence pattern (see Nemec, 1988).

Modelling of the stratigraphy of the U.S. Atlantic margin (Fig. 5; see also Cloetingh *et al.*, 1989a) and the North Sea (Lambeck *et al.*, 1987; Kooi and Cloetingh, 1989a) has shown that the punctuated stratigraphy can be successfully simulated by a stress field, the magnitude of which fluctuates through time, superimposed on the long-term thermal evolution. The inferred palaeostress was found to be largely consistent with independent data sets on the kinematic (Klitgord and Schouten, 1986) and tectonic evolution (Letouzey, 1986; Ziegler and Van Hoorn, 1989) of the northern/central Atlantic with a tensional stress field during Mesozoic times followed during the Tertiary by a compressional stress field, the magnitude of which increases with age (Fig. 6). 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. This author recognized that a significant part of the Mesozoic sequence and megasequence boundaries are associated with large-scale, fault-controlled rifting activity, while Cenozoic sequence boundaries in the Arctic were largely controlled by large-scale compressional activity and inversion tectonics (Fig. 7). Hallam (1989) recognized

Fig. 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 750 bar in-plane compression at 50 Ma induces uplift of the peripheral bulge, narrowing of the basin and a phase of rapid offlap, which is followed by a long-term phase of gradual onlap due to thermal subsidence. (c) A transition to 750 bar in-plane tension at 50 Ma induces downwarp of the peripheral bulge, widening of the basin and a phase of rapid basement onlap (after Cloetingh *et al.*, 1989a).

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.*, 1989b) strongly suggests that the majority of Mesozoic sequence boundaries reflect 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 seafloor spreading. On a large scale, unconformity-bounded 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

Fig. 5. Schematic model of the stratigraphy of the U.S. Atlantic margin at Cape Hatteras. (a, b) Modelled stratigraphy for elastic of rheology of the lithosphere in the absence of intraplate stresses. (c) Modelled stratigraphy showing effect of incorporating a fluctuating intraplate stress field in the analysis (after Cloetingh *et al.*, 1989a).

phases of accumulated tensional 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 the 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.

Discrimination of tectonic and eustatic effects

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 co-workers (Embry, 1989; Mørk *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: (1) sediment source areas that often vary greatly from one sequence to the next; (2) sedimentary regimes that change drastically and abruptly across sequence boundary; (3) faults that terminate at sequence boundaries; (4) significant changes in subsidence and uplift patterns that occur within basins across sequence boundaries; (5) sequence boundaries that are absent in parts of basins; (6) significant differences in the magnitude and extent of some of the sub-aerial unconformities and time-equivalents recognized by Vail *et al.* (1977) (see also Fig. 1). These criteria have been successfully applied to the Canadian Sverdrup Basin (Embry, 1989), as well as

Fig. 6. Paleostress curve inferred from stratigraphic modelling of U.S. Atlantic margin, and comparison with timing of large-scale tectonic events in the North/Central Atlantic Ocean. Tension is positive, and compression is negative (after Cloetingh *et al.*, 1989a).

Fig. 7. Timing and nature of megasequences and sequence boundaries of rifted basins around the Atlantic and in the Arctic (after Hubbard, 1988).

to the Svalbard basin and the Barents shelf (Mørk *et al.*, 1989).

The discrimination of tectonics and eustasy is a subtle matter, especially if biostratigraphic resolution is limited. The regional character of intraplate stresses can shed light on documented deviations from global sea-level charts. Whereas such deviations from global patterns are a natural feature of this tectonic model, the occurrence of short-term deviations does not preclude the presence of global events of tectonic origin elsewhere in the stratigraphic record. These are to be expected when major plate re-organizations in intraplate stress field occur simultaneously in more than one plate (e.g. Pollitz, 1988) or in time intervals prior to and shortly after the break-up of Pangea where continents and rift basins were in a largely uniform stress regime. The magnitude of the stress-induced phases of rapid uplift and subsidence varies with position within the basin, thus providing another important criterion to separate this contribution

from eustatic effects. Similarly, differences in the rheological structure of the lithosphere control the magnitude of the vertical motions. The presence of weak attenuated continental lithosphere enhances the ability of the stresses to cause substantial vertical motions and can explain differences in magnitude such as those observed between the Tertiary North Sea and the Gippsland Basin off south-east Australia (see Fig. 1).

TECTONIC SUBSIDENCE AND THE MECHANICS OF BASIN FORMATION

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

ripping 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; Houseman and England, 1986) 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 (e.g. Goetze and Evans, 1978).

Rapid phases of basin subsidence following the initial event of basin formation and with magnitudes 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 increases in the level of intraplate compression can equally well produce such deviations from the predictions of thermal models of basin subsidence. In Fig. 8 it is illustrated that 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 steep-head geometries of sedimentary basins (Kooi and Cloetingh, 1989b). The effect of such late-stage compressional phases is enhanced by brittle-ductile rheologies of the lithosphere, in particular where stress levels approach the lithospheric strength. Late-stage compression can explain the rapid phases of Late Neogene subsidence such as that encountered around the northern Atlantic (Cloetingh *et al.*, 1989c). 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 north-west Europe and the northern Atlantic is compressional in character (Klein and Barr, 1986; Zoback, 1989). A detailed study of the Gulf de Lions passive margin by the Institut Français du Pétrole has revealed a post-basin formation stage of strong excess subsidence in the centre of the basin, starting after 7 Ma (Fig. 9), while simultaneous rapid uplift has occurred at the basin flanks (Burrus *et al.*, 1987). As pointed out by Burrus *et al.* (1987), this rapid differential motion across the Gulf de Lions passive margin deviates from predictions of thermal models but coincides with a documented tectonic compressional phase. Stratigraphic modelling (Cloetingh *et al.*, 1989a) suggests that an increase in intraplate compression can adequately explain the observed subsidence pattern.

Stress-induced subsidence, therefore, can contribute substantially to the total tectonic subsidence. This is

Fig. 9. Subsidence of the Gulf de Lions passive margin. Curve 1 (solid curve): reconstructed tectonic subsidence, assuming local isostasy. Curves 2 and 3 (dashed and dotted, respectively): calculated tectonic subsidence from thermal models. Note the rapid excess subsidence since 7 Ma (after Burrus *et al.*, 1987).

Fig. 8. Schematic effect of late-stage compression on stratigraphy and subsidence of a 200 Ma old rifted basin incorporating a depth-dependent brittle-ductile rheological model for the lithosphere. Stretching factors are given in the upper panel. (a) Synthetic stratigraphy in the absence of intraplate stress displaying syn-rift sediments and stratigraphic horizons corresponding to 100 and 200 Ma after rifting. (b) Basement subsidence at both the flank and centre of the basin (locations indicated by vertical bars in Fig. 8 (a)). Figs 8(c) and (d) show similar results for an increase in compressional stress level from 100 Ma after basin formation onward. Note the dramatic narrowing and deepening of the basin, resulting in subsidence patterns deviating from the predictions of conventional thermal models (after Kooi and Cloetingh, 1989a).

Fig. 10. Different methods to obtain estimates for crustal extension. (a) The relation between predicted total subsidence for the McKenzie (1978) stretching model and the extension factor β . The rapid initial subsidence is followed by a long-term thermal subsidence phase (after Sawyer, 1985). (b) Estimates of crustal extension derived from structural analysis of displacements on faults.

important because it is common practice (Fig. 10) to use total tectonic subsidence as a measure for the determination of values of crustal extension (e.g. Sawyer, 1985; Sclater and Celerier, 1987). 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. Quantitative modelling of the post-Jurassic subsidence of the North Sea (Sclater and Christie, 1980; Barton and Wood, 1984) has yielded estimates of about 50–80 km, while Ziegler and Van Hoorn (1989) have pointed out that the total amount of extension on the Zechstein (Late Permian) level measured on high-angle normal faults is probably only 20–25 km. Late-stage compression can strongly contribute to the total tectonic subsidence. Therefore, current backstripping techniques, correcting only for vertical loading of the lithosphere, tend to overestimate values of crustal extension and therefore might result in overestimates of temperatures at depths corresponding to the hydrocarbon window (Kooi and Cloetingh, 1989b). 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 intraplate 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. 11). Similarly, effects of multiple stretching phases and increased levels of intraplate 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 vital 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 the effectiveness of stratigraphic correlation.

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Fig. 11. Two approaches to the analysis of tectonic subsidence. (a) Traditional interpretation: short-term deviations from long-term trends in basin subsidence are attributed to eustatic sea-level changes and renewed phases of crustal stretching. (b) Alternative interpretation: short-term deviations from long-term (thermal) subsidence are interpreted in terms of stress-induced vertical motions of the lithosphere.

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