Gradstein, F. M., Ludden, J. N., et al., 1992
Proceedings of the Ocean Drilling Program, Scientific Results, Vol. 123

37. CONTINENTAL MARGIN STRATIGRAPHY, DEFORMATION, AND INTRAPLATE STRESSES FOR THE INDO-AUSTRALIAN REGION[1]

Sierd Cloetingh,[2] Carol Stein,[2] Paul Reemst,[3] Felix Gradstein,[4] Paul Williamson,[5] Neville Exon,[5] and Ulrich von Rad[6]

ABSTRACT

The stratigraphic record along the northwestern Australian continental margin provides constraints on the tectonic evolution of the Indo-Australian region, which is characterized by an extremely high level of intraplate deformation, presumably reflecting high levels of regional stress. The patterns of folding and faulting in the northeastern Indian Ocean (shown by focal mechanisms, gravity, and seismic-reflection data) are consistent with predictions of a stress model based on age-dependent driving forces on the lithosphere. Separate compressional and extensional stress provinces are found, each characterized by regionally consistent stress orientations. For example, folding occurs only within the region of predicted deviatoric compression. The evolution of the stress field can be inferred from the stratigraphic data of ODP Leg 116. An unconformity dated at 7.5 Ma marks the end of the recent major phase of folding since which deformation has occurred primarily through faulting. Here, we examine the stratigraphic record of the northwestern margin of Australia using data from Legs 122 and 123 to investigate whether the continental margin stratigraphy records the spatial and temporal variations in stress inferred from the folding in the northeastern Indian Ocean. With the exception of the Pliocene unconformities, the major unconformities inferred from Leg 122 coincide with lowstands in sea level on regional sea-level curves proposed for the northwestern Australian margin, while the major unconformities found during Leg 116 show a lack of correlation with lowstands in these regional curves. Although the correlation of the unconformities from Legs 116 and 122 with lowstands in the Haq et al. (1987) eustatic curve is sometimes good, the major unconformities drilled during Legs 116 and 122 also coincide with plate-tectonic changes of a global character. The Neogene stratigraphic record of the Northwest Shelf can be explained by a paleostress field characterized by an overall tensional stress regime during middle Miocene time to a regime with an increasing level of intraplate compression from late Miocene time onward. The paleostress field inferred from the stratigraphic modeling of the Northwest Shelf is consistent with observed tectonic activity, including middle Miocene fault reactivation in the Browse and Carnarvon basins and Pliocene tilting of the Wombat Plateau. The inferred increase in the level of compression for the northwestern margin occurs at a time corresponding with the onset of compressional folding in the northeastern Indian Ocean, which was documented during Leg 116. The tectono-stratigraphic evolution of the northwestern Australian margin seems to reflect both the changes of intraplate tectonics associated with the unique dynamic situation of the Indo-Australian Plate, as well as the interplay between a major global plate reorganization in Neogene time with eustatic contributions to the sea-level record.

INTRODUCTION

Since the publication of Vail et al.'s short-term cycles (1977) in relative sea level, major debate has been going on about the inferred global synchroneity of the third-order cycles, both in support of this hypothesis (e.g., Haq et al., 1987) and in opposition (e.g., Parkinson and Summerhayes, 1985; Miall, 1986; Hallam, 1988). Several authors have noted that Vail et al.'s cycles, although based on data from different basins around the world, are heavily weighted in favor of the North Sea and the northern/central Atlantic margins. The issue of global synchroneity is important, as it obviously strongly influences present discussions about 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 thus are associated with a long thermal inertia of several tens of millions of years. Glacio-eustasy can easily produce both the rate and magnitude of the inferred sea-level changes, but raises two basic problems. First, the occurrence of third-order cycles in sea level during time intervals where no geological evidence can be found suggests low-altitude glaciation. This presents a fundamental problem of explaining changes in eustatic sea level in the Vail et al. (1977) and Haq et al. (1987) charts prior to the late Cenozoic. Second, we see the inability of glacio-eustatics to cause uniform decreases and increases in sea level. The sign and magnitude of the induced sea-level changes depend upon the distance to the location of the ice cap. This feature, well-known among modelers of post-glacial rebound processes (e.g., Lambeck et al., 1987), is not always fully appreciated by those who advocate glacio-eustasy as the key mechanism for explaining global-synchronous changes in sea level of uniform magnitude.

More recently, it has been shown that short-term changes in relative sea level can be caused equally well by rapid, stress-induced, vertical motions of the lithosphere in sedimentary basins (Cloetingh et al., 1985; Cloetingh, 1986; Karner, 1986). This research showed that intraplate stresses, as well as being important in the formation of rifted basins (e.g., Houseman and England, 1986), also play a critical part during their subsequent subsidence history. The World Stress Map Project of the Interna-

[1] Gradstein, F. M., Ludden, J. N., et al., 1992. *Proc. ODP, Sci. Results,* 123: College Station, TX (Ocean Drilling Program).

[2] Institute of Earth Sciences, Vrije Universiteit De Boelelaan, 1085, 1081 HV Amsterdam, The Netherlands.

[3] Department of Geological Sciences, University of Illinois at Chicago, P.O. Box 4348, Chicago, IL 60680, U.S.A.

[4] Bedford Institute of Oceanography, Atlantic Geoscience Centre, Dartmouth, Nova Scotia B2Y 4A2, Canada.

[5] Bureau of Mineral Resources, P.O. Box 378, Canberra, ACT 2601, Australia.

[6] Bundesanstalt für Geowissenschaften und Rohstoffe, P.O. Box 51054, D-3000 Hannover 51, F.R.G.

tional Lithosphere Program has convincingly established the existence of large-scale, consistently oriented stress patterns in the lithosphere (Zoback et al., 1989). In particular, studies of the French school of structural geology provided compelling evidence for changes in the magnitude and orientations of these stress fields on time scales of a few million years (Letouzey, 1986; Bergerat, 1987; Philip, 1987) in association with collision and rifting processes in the lithosphere. These stresses have been propagated away from the plate boundaries into the interiors of the plates, where they affect the vertical motions within sedimentary basins (Cloetingh, 1988; Ziegler, 1987, 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 the ridge push forces from oceanfloor spreading in the Atlantic (Zoback et al., 1989).

Recent studies of the tectono-stratigraphic evolution of the northern-central Atlantic suggest that the inferred record of changes in sea level reflects, to a large extent, the changes in seafloor spreading of the northern Atlantic (Tankard and Balkwill, 1989; Cloetingh et al., 1989a, 1989b). Similarly, as shown by a number of recent studies (Hubbard, 1988; Hallam, 1988; Tankard and Balkwill, 1989), third-order variations in relative sea level correlate with tectonic events. It seems that the spatial and temporal synchroneity of sequence boundaries in different basins surrounding the northern Atlantic occurs simultaneously with tectonic phases (Fig. 1). The preference for global eustasy causing third-order cycles has been partly based on their inferred global character. For this reason, it is essential to consider the stratigraphic record of sedimentary basins outside of the Atlantic/North Sea area to establish the relative contributions of tectonics and eustasy to the sea-level record. Since the publication of Vail et al.'s study (1977), scientists have noted, for example, that the Gippsland Basin of southeastern Australia (Steele, 1976) does not

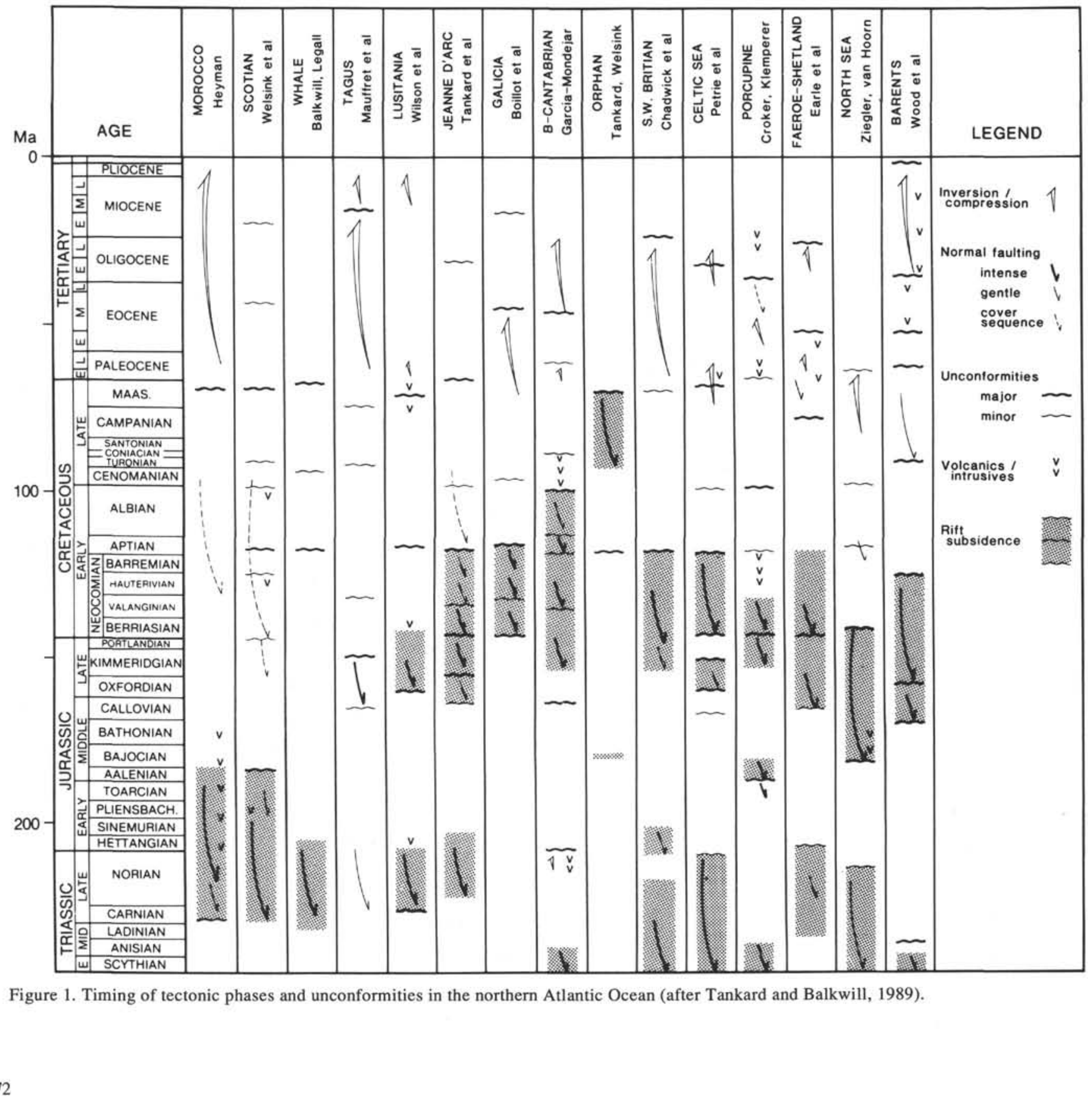


Figure 1. Timing of tectonic phases and unconformities in the northern Atlantic Ocean (after Tankard and Balkwill, 1989).

conform to the predictions of the global sequence charts. Similarly, the record at other basins of the Australian margin shows evidence for deviations from the expected third-order cycles (Chaproniere, 1984; Carter, 1985).

With the advent of new data from Legs 122 and 123 (Ludden, Gradstein, et al., 1990; Haq, von Rad, O'Connell, et al., 1990) for the northwestern Australian margin and Argo Abyssal Plain (Fig. 2) and with enhanced insight in the intraplate tectonic evolution of the central Indian Ocean in the last 10 yr from plate kinematic models (Wiens et al., 1985), marine geophysical studies (Geller et al., 1983; Stein et al., 1989a, 1990), stress modeling studies (Cloetingh and Wortel, 1985; 1986), earthquake studies (Wiens and Stein, 1984; Bergman and Solomon, 1985; Petroy and Wiens, 1989), and data from Leg 116 (Cochran, Stow, et al., 1989), a quantitative framework is now available to investigate the possible link between the sea-level record and intraplate evolution of the Indo-Australian region. The existence of spatially coherent stress patterns in the Indo-Australian Plate (Fig. 3A), predicted by stress modeling (Cloetingh and Wortel, 1985; 1986) and confirmed by data from marine geophysics (Fig. 3B) and drilling by the petroleum industry, offers one the prospect of studying the temporal and spatial variations in stress-induced vertical motions in different parts of the plate.

In the following, we first briefly review the response of the lithosphere to horizontal stresses. Subsequently, we discuss the implications of the existence of different stress provinces in the Indian Ocean on the stratigraphy of the seafloor and rifted continental margins.

TECTONIC STRESS AND APPARENT CHANGE IN SEA LEVEL

Intraplate stresses modulate the long-term basin deflection caused by thermal subsidence (Fig. 4A) and induce rapid differential vertical motions (Fig. 4B) of a sign and magnitude that depend on the position within the basin (Cloetingh et al., 1985). For example, for an elastic model of the lithosphere, intraplate compression causes relative uplift of the basin flank, subsidence at the basin center, and seaward migration of the shoreline (Fig. 4C). An offlap develops, and an unconformity is produced. Increases in the level of tensional stress induce widening of the basin, lower the flanks, and cause landward migration of the shoreline, producing a rapid onlap phase. Stress-induced vertical motions of the crust can also drastically influence sedimentation rates. Flank uplift due to an increased level of compression, for example, can significantly enhance sedimentation rates and modify the infilling pattern, promoting the development of unconformities (e.g., Galloway, 1987). Similarly, preexisting sub-basinal fractures/faults can be reactivated to cause intrabasinal "noise" in the overall subsidence pattern (see Nemec, 1988).

A more realistic model of the mechanical properties of the lithosphere is based on the extrapolation of rock mechanics data (Goetze and Evans, 1979). These scientists developed a depth-dependent rheology of the lithosphere that combined Byerlee's law (Byerlee, 1978) in the brittle domain with temperature-dependent constitutive equations describing the deformation in the ductile regime. Although extrapolated over several orders of magnitudes,

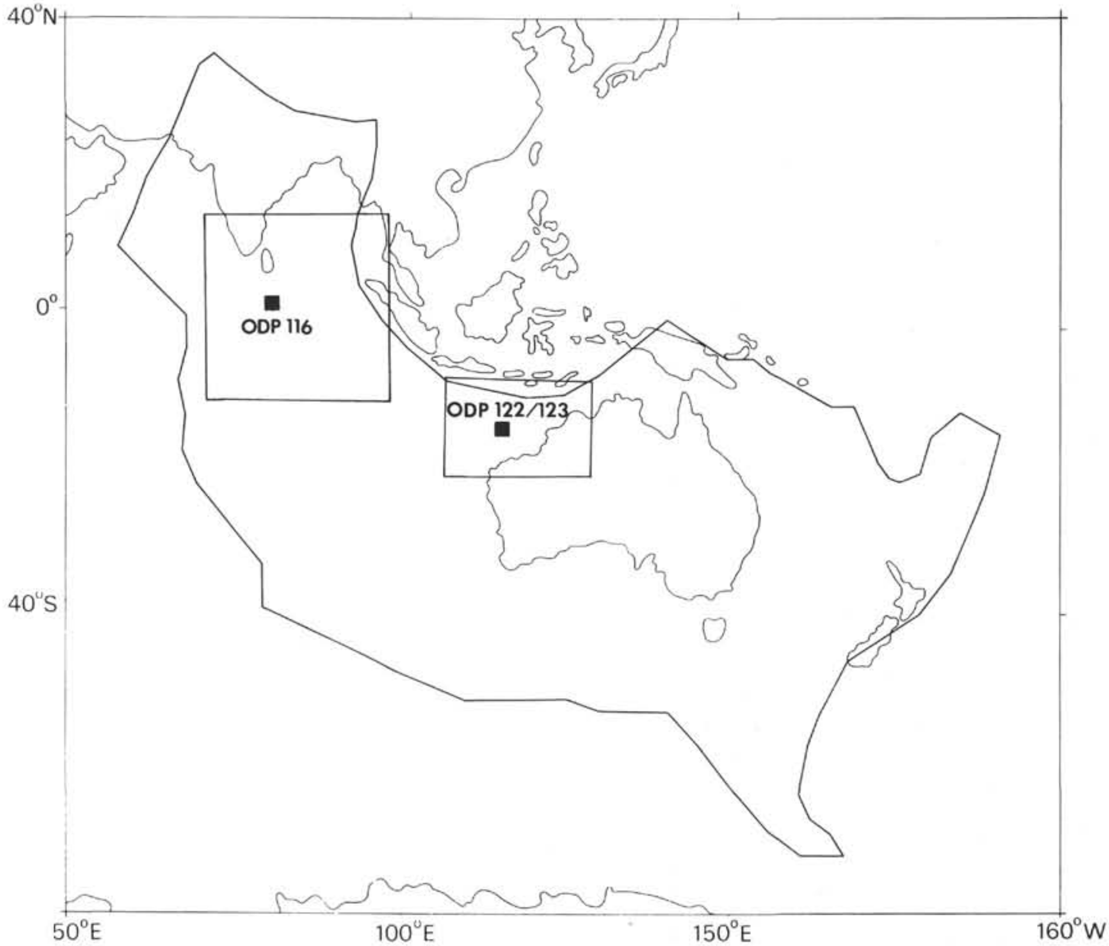


Figure 2. Location map showing areas in the Indo-Australian Plate studied during Legs 116, 122, and 123.

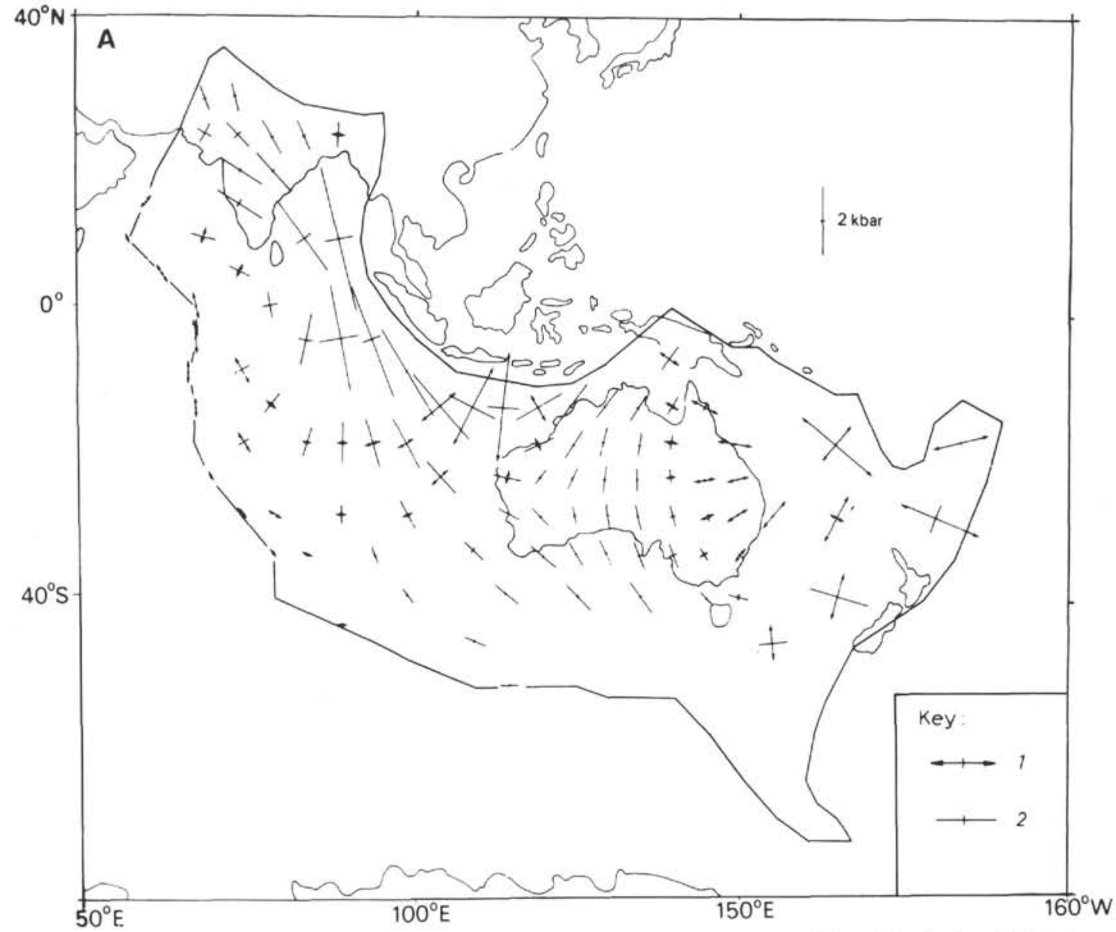


Figure 3. Stress field in the Indo-Australian Plate. **A.** Predicted from finite-element modeling (Cloetingh and Wortel, 1986). Plotted are the horizontal nonlithostatic stresses in the plate. Arrows pointing outward denote deviatoric tensional stresses. **B.** Stress directions inferred from analysis of earthquake focal mechanism data labeled by numbers (see Cloetingh and Wortel [1986] for a full compilation of the data).

the resulting strength envelopes have been demonstrated to be consistent with the outcome of studies of intraplate seismicity (Wiens and Stein, 1983) and bending of the lithosphere (McAdoo et al., 1985). The rheological studies suggest that continental lithosphere in general is weaker than oceanic lithosphere (Kirby, 1985; Vink et al., 1984). One would expect rifted continental margins to be characterized by lateral variations in lithospheric strength, corresponding to an oceanward increase in the equivalent elastic thickness of the lithosphere (Fig. 5A). These lateral variations introduce an additional tilting effect not present in the models, based on a uniform elastic plate (Cloetingh et al., 1985; see Fig. 5B).

Compared with an elastic plate, a plate having a depth-dependent rheology will by its finite strength, and hence lower flexural rigidity, be more sensitive to applied horizontal loads. This feature is illustrated in Figure 6A, which shows the magnification of the vertical deflection in the basin that resulted from the incorporation of a depth-dependent rheology in the modeling. At the same time, and also due to its finite strength, a plate having a depth-dependent rheology will begin to fail upon the application of tectonic loads in the lithosphere (Fig. 6B). This failure process will begin at the uppermost and lowermost boundaries of the mechanically strong part of the lithosphere, where lithospheric strengths are less. As a result, the mechanically strong part of the lithosphere will be thinned to a level that ultimately depends on the ratio of the integrated strength of the lithosphere and the applied stress (Fig. 6B). This behavior results in a reduction of the flexural wavelength of the plate once a tectonic load has been applied. Note that the latter effect applies to stresses of both compressional and tensional character. This feature adds a complexity not present in the case of an elastic rheology, where compression and tension have an opposite effect on the flexural shape of the basin. Rheological weakening of the lithosphere will also occur because of flexural stresses induced by the sediment loading itself (Cloetingh et al., 1982). Superimposed on this bending stress are the in-plane stresses. An important but unsolved question is how much of the bending stress is relaxed, as discussed by Stein et al. (1989b), for the Canadian margin.

The net effect of an in-plane compressional force on the sign of the induced differential vertical motion at the position of the peripheral bulge flanking the basin is similar to the predictions of the elastic model. A difference with the predictions of the elastic model occurs for tensional stresses (Fig. 6A). Because of rheological weakening, both compression and tension will induce a basinward migration of the flexural node of the basin. As the bulge is commonly located below the shelf of passive margins (Watts and Thorne, 1984), its vertical displacement from varying stress levels will dominate the stratigraphic response and apparent sea-level record induced by intraplate stresses. Inspection of Figure 6A demonstrates that the stress-induced deflections of the litho-

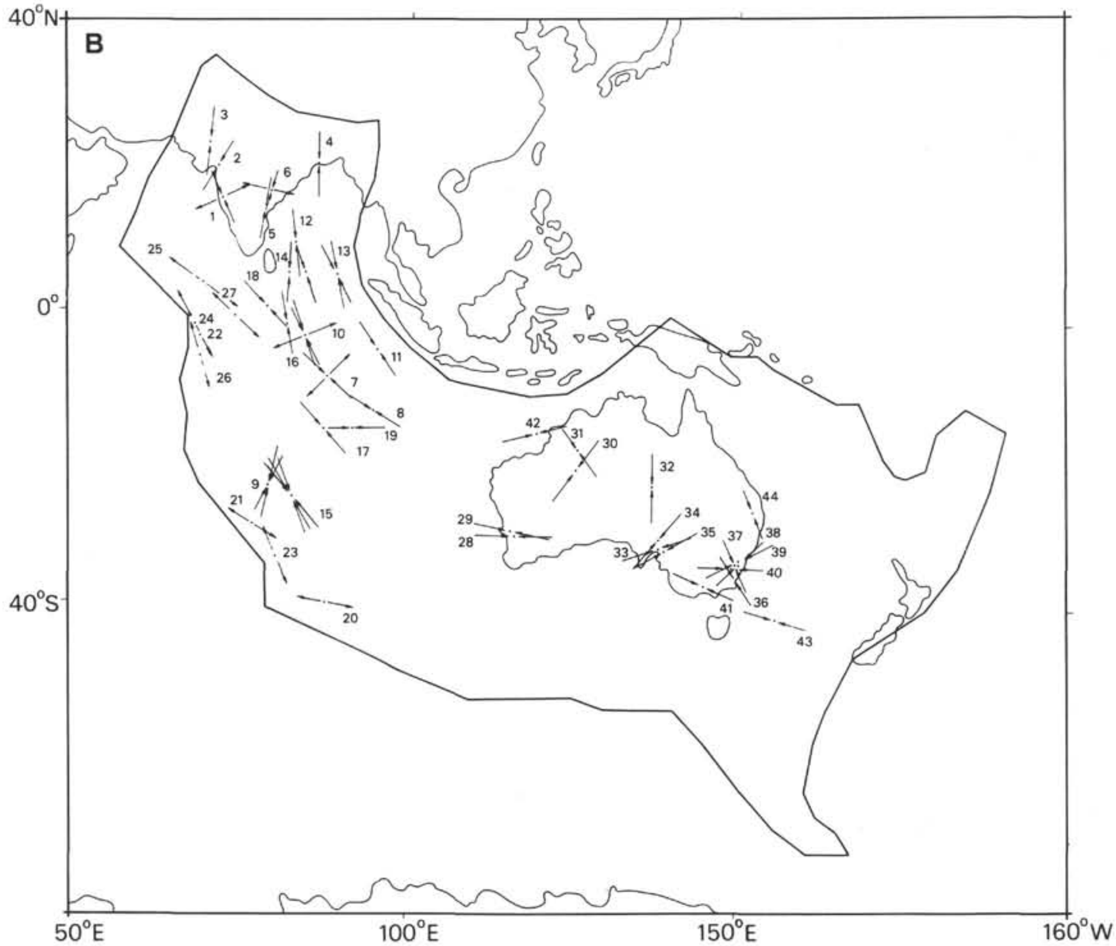


Figure 3 (continued).

sphere are substantially larger than in the case of an elastic rheology of the lithosphere. Figure 7 shows the magnitudes of the vertical motions at the basin flanks induced by the action of tectonic stress for rheological models that are based on extrapolation of rock mechanics data for oceanic and continental lithosphere. The differential uplift or subsidence, defined as the difference in deflection for a change in-plane force, is calculated for the flank of the basin as a function of the variation in-plane force, with the same reference sediment load in each case. Two models for continental rheology (Carter and Tsen, 1987) are adopted using different petrological models for the lithosphere. The first model adopts a quartz-diabase-olivine layering, while the second model is based on the quartz-diorite-olivine stratification. The stress-induced deflection rapidly increases for stresses approaching lithospheric strength. Figure 7 displays the transition between vertical motions of the order of about 100 m that are reflected in the apparent sea-level record, and the more dramatic vertical motions with magnitudes of the order of a few kilometers that occur as a result of the accumulation of stress to levels near the strength of the lithosphere. Lithospheric folding will occur if high compressional-stress levels are induced in the lithosphere in response to a unique dynamic situation of a lithospheric plate. An example of folding caused by such a process (collision) is presently occurring in the northeastern Indian Ocean and will be discussed next. 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 the tectonic model presented above, 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 occur simultaneously in more than one plate (e.g., Pollitz, 1988) or in time intervals such as prior to and shortly after the breakup of Pangea, when 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 for separating this contribution from eustatic effects (Embry, 1990; Embry and Podruski, 1988). Similarly, differences in the rheological structure of the lithosphere control the magnitude of the vertical motions (Cloetingh et al., 1989a). The presence of weak, attenuated continental lithosphere enhances the ability of the stresses to cause substantial vertical motions and may explain differences in magnitude, such as those observed between the Tertiary North Sea and the Gippsland Basin off southeast Australia (Vail et al., 1977).

Thus, it is the interplay among variations in stress level and orientation with rheological variations that controls the record of vertical motions. For example, compared to the predictions of the elastic model of the lithosphere, incorporation of depth-dependent rheological properties significantly enhances the magnitude of stress-induced vertical motions (Cloetingh et al., 1989a).

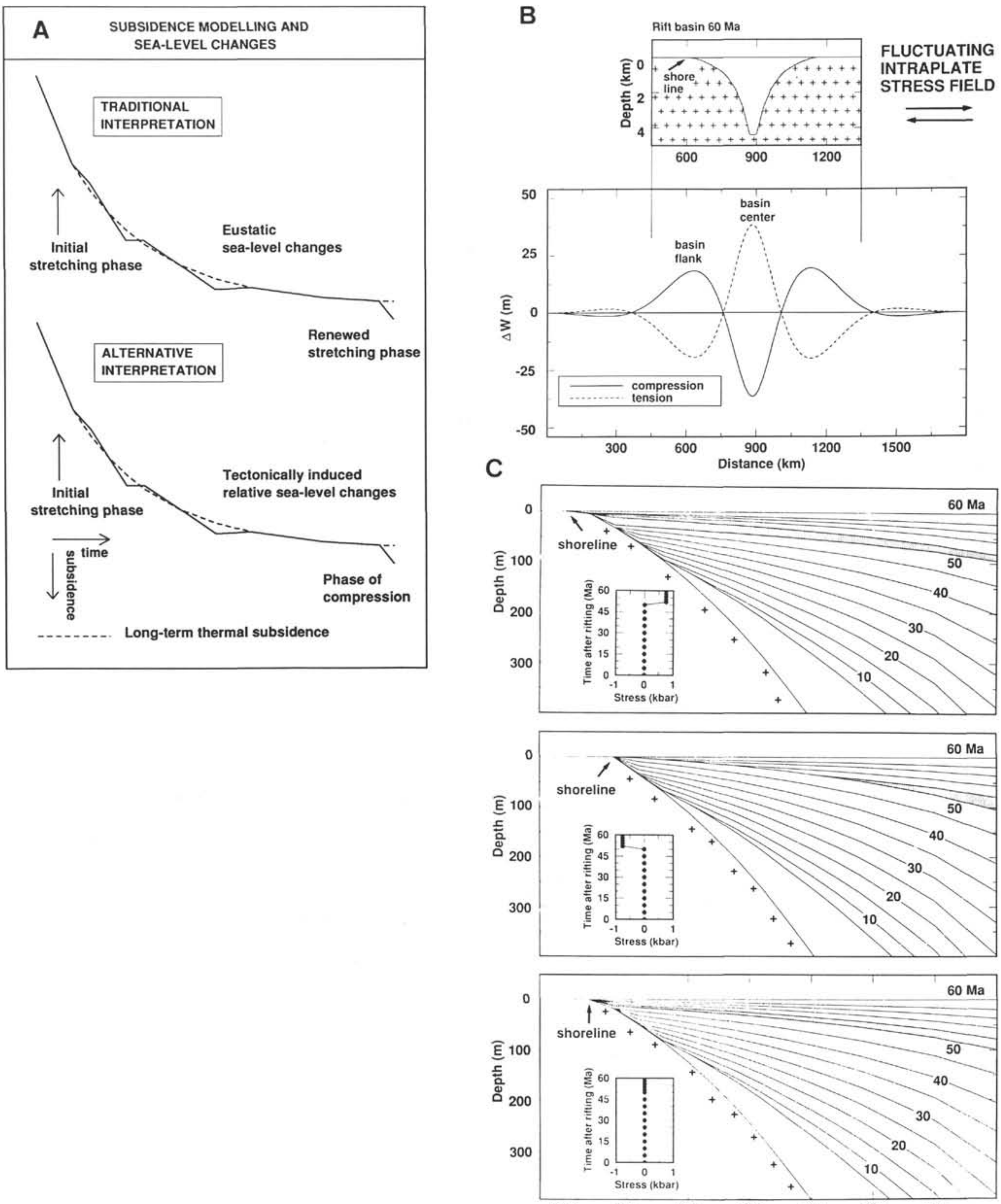


Figure 4. Effects of intraplate stress on the record of short-term vertical motions in sedimentary basins. **A.** Two approaches to the analysis of tectonic subsidence. Traditional interpretation (top): short-term deviations from long-term trends in basin subsidence are attributed to changes in eustatic sea level and to renewed phases of crustal stretching. Alternative interpretation (bottom): short-term deviations from long-term (thermal) subsidence are interpreted in terms of stress-induced vertical motions of the lithosphere. **B.** Stress-induced vertical motions in the lithosphere in a uniform elastic plate model. **C.** Effect of changes in intraplate stress on basin stratigraphy for a uniform elastic plate model for the mechanical properties of the lithosphere. Note the opposite effects at the flanks of the basin and at the basin center.

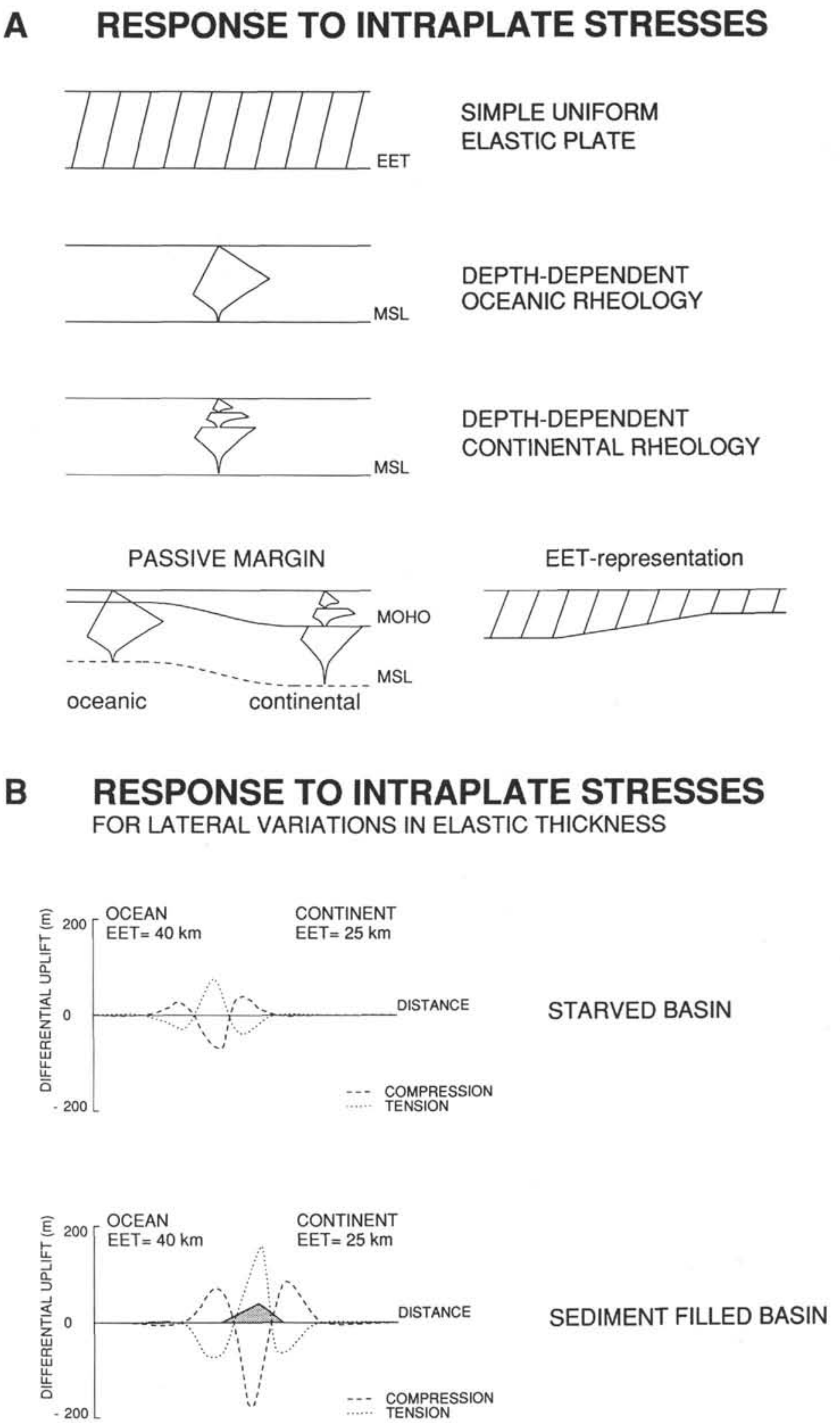
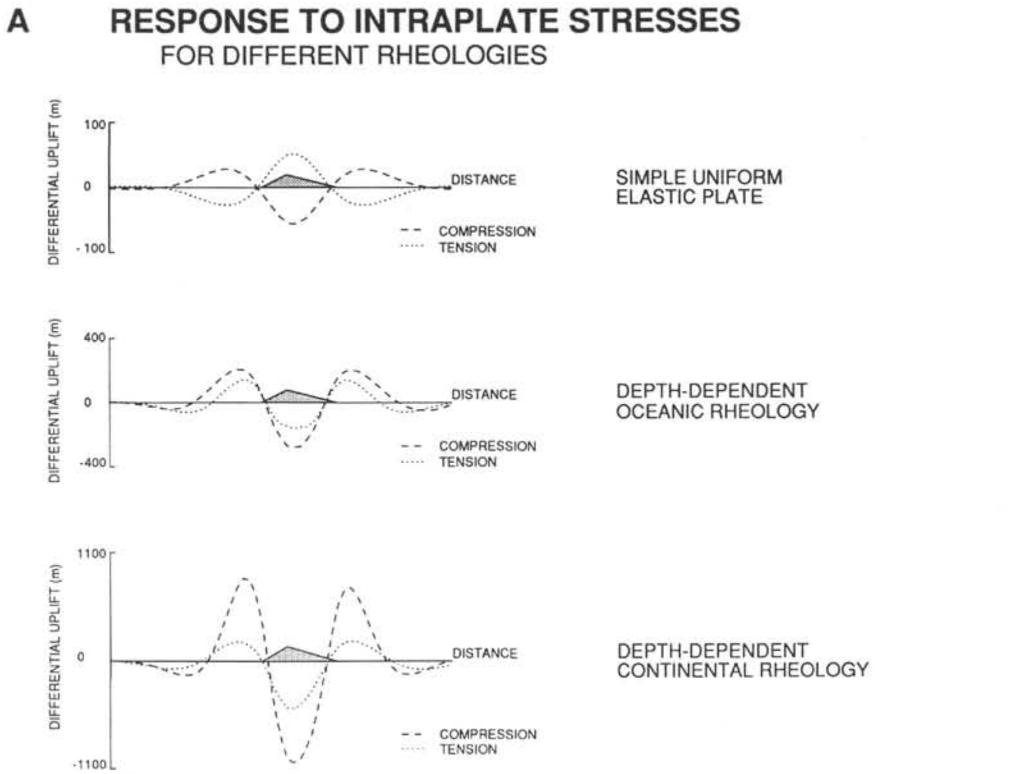


Figure 5. **A.** Different rheological models of the lithosphere considered when studying the response of the lithosphere to intraplate stress. The elastic plate model is characterized by an equivalent elastic thickness of the lithosphere (EET). The models (based on extrapolation of rock mechanics data) are characterized by thickness of the mechanically strong part of the lithosphere (MSL), defined by the depth at which (as a result of increasing temperature) lithospheric strength levels approach zero. **B.** Response to intraplate stresses for a lithosphere having lateral variations in elastic thickness and for a starved and sediment-filled basin, respectively. Shaded triangle indicates location of sedimentary wedge.



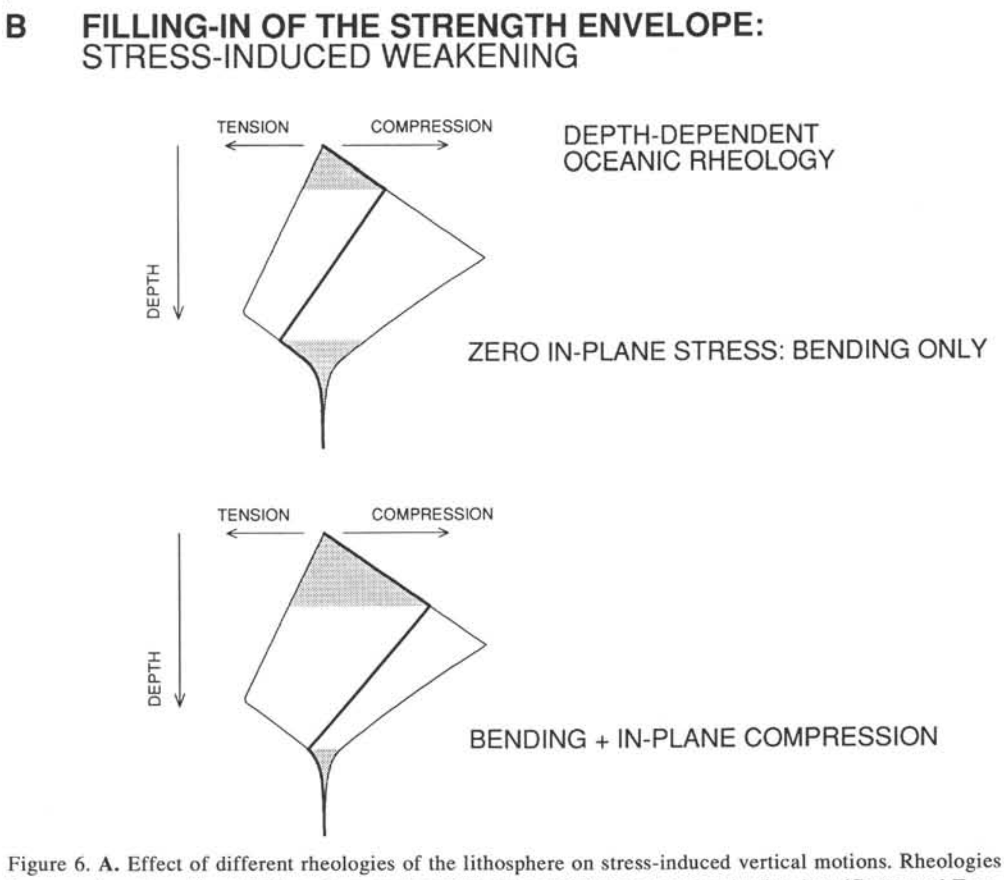


Figure 6. **A.** Effect of different rheologies of the lithosphere on stress-induced vertical motions. Rheologies for oceanic and continental lithosphere are based on extrapolation of rock mechanics data (Carter and Tsen, 1987) for olivine and for quartz/diorite/olivine models, respectively. Stress-induced vertical motions are given for the two depth-dependent rheologies and, for comparison, the predictions from the simple elastic model. Shaded triangle indicates location of sedimentary wedge. **B.** Effect of stress-induced rheological weakening of the lithosphere in response to intraplate compressive stress.

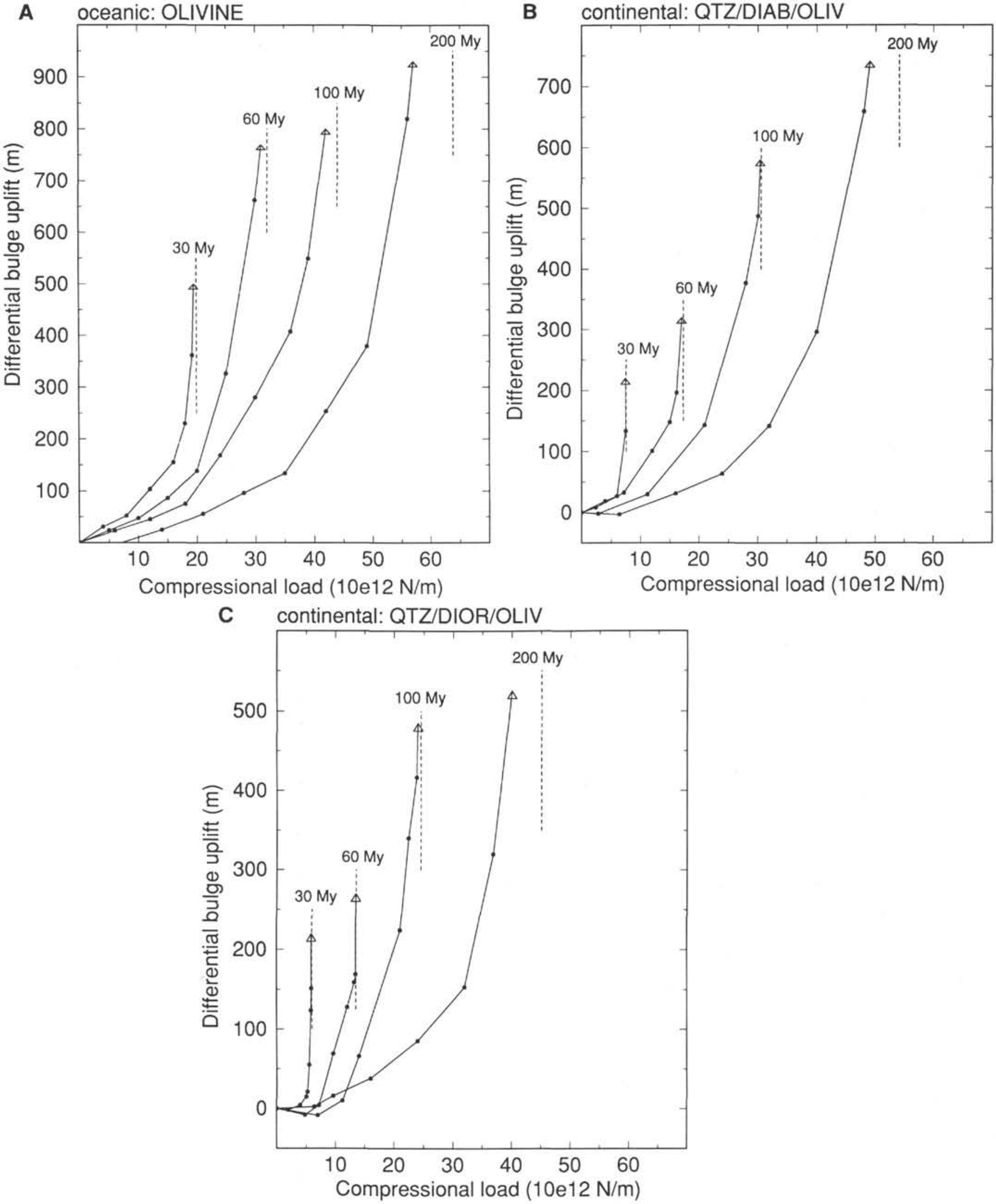


Figure 7. Stress-induced differential uplift (meters) at basin flank resulting from superposition of variations in in-plane force on flexure caused by sediment loading that has been induced by intraplate stresses up to the buckling load as a function of lithospheric age. Vertical dashed lines give limits of integrated lithospheric strength derived from rock mechanics data for different thermo-tectonic ages of the lithosphere. **A.** Deflection for depth-dependent oceanic olivine rheology. **B.** Deflection for depth-dependent quartz-diabase-olivine continental rheology. **C.** Deflection for depth-dependent quartz-diorite-olivine continental rheology.

STRESS-INDUCED DEFORMATION IN THE CENTRAL INDIAN OCEAN

Stress Field in the Central Indian Ocean

Until recently, the northeastern Indian Ocean was considered the most active region of oceanic intraplate seismicity (Gutenberg and Richter, 1954; Sykes, 1970; Stein and Okal, 1978; Bergman and Solomon, 1985). North of ~10°S in the central Indian Basin, sediments and acoustic basement have been deformed into long-wavelength (~200 km) undulations (Fig. 8A), associated with large free-air gravity (Weissel et al., 1980; Geller et al., 1983; Neprochnov et al., 1988) and geoid anomalies (McAdoo and Sandwell, 1985). Superimposed on the folds (Fig. 8B) are closely spaced (~5–20 km) faults (Geller et al., 1983; Cochran, Stow, et al., 1989). The Wharton Basin to the east of the Central Indian Basin is characterized by high seismicity, faulting, and folding north of ~15°S (Stein et al., 1989a; Shcherbakov et al., 1989). The Ninetyeast Ridge (between the Central Indian and Wharton basins) has also been affected by the regional deformation. North of ~10°S, the complex blocky structure of the ridge appears to have been formed by folding (Petroy and Wiens, 1989; Stein et al., 1989a), with wavelengths and amplitudes similiar to those in the Central Indian and Wharton basins. The spatial distribution of the folds can be traced using SEASAT-derived free-air gravity data. The axes of the folds change from east–west in the Central Indian Basin, to east-northeast–west-southwest on the Ninetyeast Ridge, to northeast–southwest in the Wharton Basin (Fig. 9A).

This seismicity and folding and faulting may be interpreted as manifestations of a diffuse plate boundary that separates distinct Indian and Australian plates (Wiens et al., 1985). The orientation and spatial distribution of the folds (Fig. 9A) and the directions of maximum and minimum compression from earthquake focal mechanism solutions (Fig. 9B) are in good agreement with signs and directions of the stresses calculated for the Indo-Australian Plate (Cloetingh and Wortel; 1985;1986). Cloetingh and Wortel's

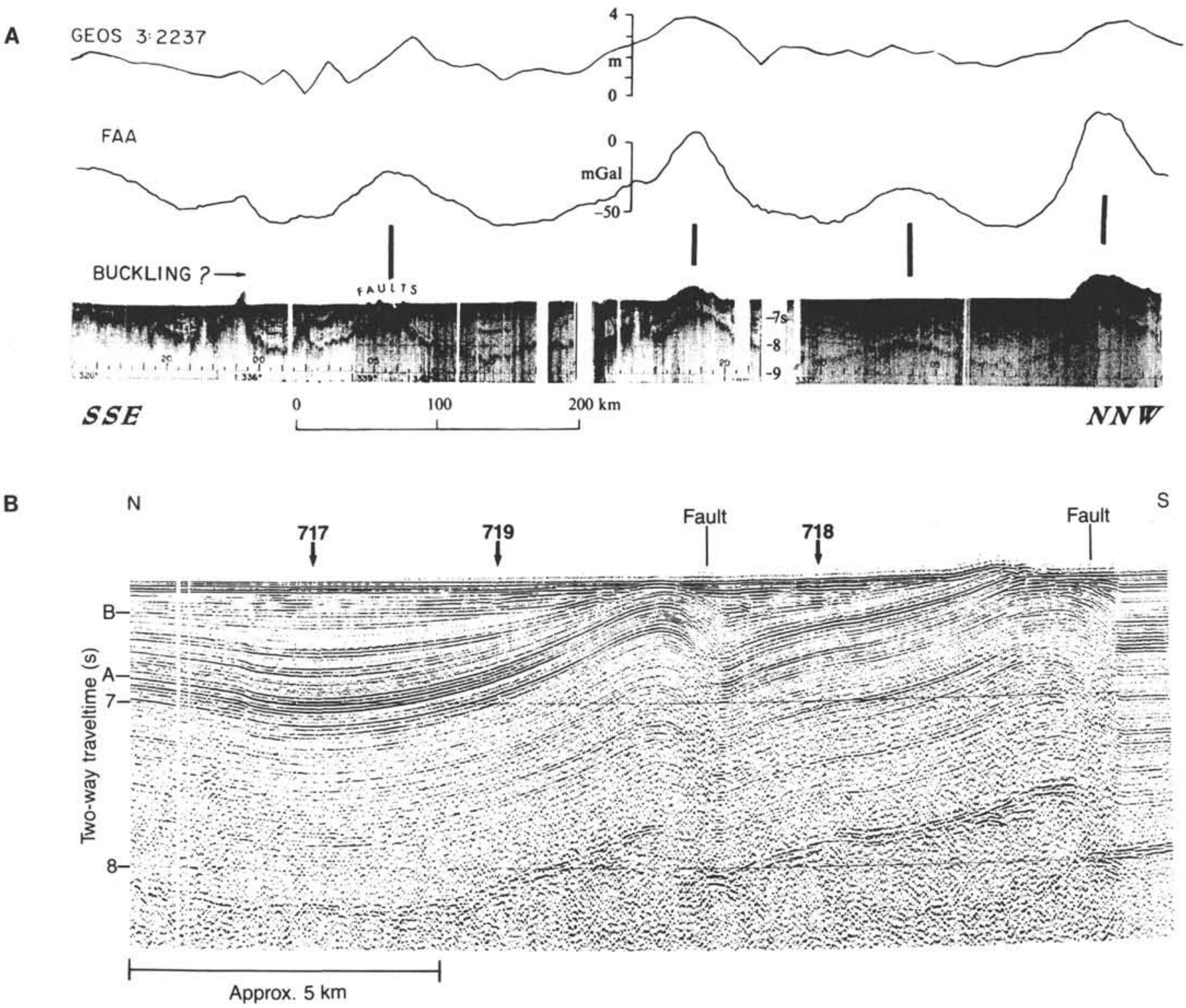


Figure 8. **A.** Seismic-reflection profile in the area of intense intraplate deformation, showing long wavelength folding of the lithosphere associated with undulations in gravity (FAA) and geoid height (after Weissel et al., 1980). **B.** Seismic-reflection profile showing faulted blocks with spacing of 5 to 10 km at the sites of Leg 116 (After Shipboard Scientific Party, 1989b).

stress model predicts large (on the order of hundreds of megapascals) compressional stresses in the northeastern Indian Ocean.

The folds were observed principally only within the region of predicted compression for both principal horizontal nonlithostatic stresses. These stresses are commonly not observed south of the transition to the extensional stress region.

However, it is more difficult to determine the magnitude of the stresses required to produce the observed deformation. The same order of magnitude of stress is required when modeling the wavelengths of the folding (McAdoo and Sandwell, 1985; Zuber, 1987; Fig. 10). Similar stresses may also be inferred from the depth of seismicity, which often occurs at depths where the lithosphere should have considerable strength (Wiens and Stein, 1983; Stein, 1984). 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., 1991).

Not all of the northeastern Indian Ocean is in compression. The region of the Chagos Bank area, in the southern Chagos-Laccadive Ridge, is a site of large normal-fault earthquakes (Stein, 1978; Bergman and Solomon, 1984; Wiens, 1986). The direction of the minimum and maximum compressional axes from these focal mechanisms agrees well with the location and direction of maximum extensional stresses, calculated by Cloetingh and Wortel (1985; 1986). This region, part of the diffuse plate boundary between the Indian and Australian plates, is predicted to be in extension from the NUVEL-1 plate boundary motion model (De Mets et al., 1990). In addition, the recent rapid subsidence and drowning reef complex of Chagos Bank (Darwin, 1842; Schlanger and Stein, 1987) may result from the regional extensional stresses. The Chagos Bank is an excellent location for studying the effect of intraplate stresses and changes in sea level. Carbonate platforms are sensitive recorders of changes in relative sea level (Schlager, 1988; Schlager and Gamber, 1986). Detailed analysis of the subsidence history provides a powerful tool for separating the contributions of tectonics and eustasy to the sea-level record. Of particular interest are the effects of earthquakes, subsidence, uplift, and tilt on platform anatomy. The fragmentation of platforms often is controlled not only by local effects, such as sediment supply and currents but also by tectonic events. In mid-Cretaceous time, a period of major plate reorganization, many carbonate platforms within the Tethyan region died (Eberli, 1991). For further discussion see Stein et al. (1987, 1989a, 1990).

Unconformity Generation in the Deep Sea and Timing of the Onset of Deformation

In seismic reflection-records (Fig. 8B), folding is more clearly seen in the Central Indian Basin than in the Wharton Basin because of the thick accumulation of turbidite sediments in the Bengal Fan that were derived from the Ganges-Brahmaputra delta system. Within the region of observed folding in the Central Indian Basin the sediment thickness ranges from greater than 3 km in the north to less than 0.3 km in the south. However spectacular the folds appear on the seismic-reflection records, they can only account for a few kilometers of shortening, suggesting that faulting must now be the major mechanism for convergence between the Indian and Australian plates (Gordon et al., 1990). A widespread unconformity marks a surface separating folded strata below from relatively flat lying, but faulted, sediment from above (Fig. 8B). The surface has been dated as 7.5 to 8 Ma at sites of Leg 116 (Cochran, 1990). This unconformity can be traced throughout much of the region of observed folding in the Central Indian Basin. From single-channel seismic reflection records, it appears that the layers below the unconformity may have been folded about equally.

The duration of folding associated with this unconformity depends on the sedimentation history. The most simplistic interpretation is that the folding occurred relatively rapidly, without time for large-scale syndeposition, given that only in a few locations are the sediments above the unconformity upturned to either a mild subsequent folding episode (Neprochnov et al., 1988) or uplifted due to local faulting. An alternative interpretation is that the folding occurred over a relatively long period of time. Hence, the lack of apparent growth structures in the fold could be due to slow sedimentation rates after the initiations of folding. The local sedimentary sedimentation deposition patterns have been altered subsequent to folding. The elevated seafloor may have blocked the flow of turbidites from the north, as observed that since folding higher parts of the folds have received mainly pelagic sediment at low (a few meters per million years) sedimentation rates (Weissel et al., 1980; Geller et al., 1983). However, based on the only location where there is a stratigraphic record through the unconformity (Leg 116, 1°S, 81.4°E) the pre- and post- sedimentation rates and types appear to be similar (Cochran, 1990). Thus, we conclude that folding probably was a relatively short-lived event and ended about 7.5 to 8 Ma.

Hence, we suggest that the unconformity does not represent the initiation of deformation, as previously suggested (Weissel et al., 1980; Geller et al., 1983), but rather the end of deformation by folding, and presumably, since then compressional deformation has occurred mainly by reverse faulting. We can speculate about the causes for the change from mainly folding deformation to faulting. Perhaps some change in the stress field and/or mechanical behavior of the lithosphere will be required to explain the mode of deformation. In theory, given unfractured material, more stress is required to produce faulting than folding. Hence, perhaps with time, the stresses have increased. Initially, the stress was sufficient to produce folding, but not faulting. As the magnitude of stress increased, the mode of deformation changed from folding to faulting. Alternatively, the folding sufficiently weakened the lithosphere, which permitted faulting to occur without a significant change of the stress. Similar changes in the mode of deformation are seen in the continental lithosphere underlying the northeastern Canadian Arctic Islands (Stephenson et al., 1990; Stephenson and Cloetingh, 1991). For the eastern Sunda Shelf region (off Indonesia), folding has been observed prior to faulting (Letouzey et al., 1990). The spacing of the faults (~5–20 km; Shipboard Scientific Party, 1989a) suggests that faults formed when the oceanic crust was near the ridge crest may have been reactivated by the Neogene compressional tectonics (Geller et al., 1983; Shipboard Scientific Party, 1989b). The unconformities younger than 7.5 Ma at Leg 116 sites may be related to controls on the sedimentation, such as uplift and erosion of the Himalayas, position of sea level relative to shelf edge, paleoceanography, and complications caused by fan growth processes. For example, the unconformity at 4.6 to 5.4 Ma may relate to an increase in sea level, with a change in sediment type from gray silt–mud turbidites to finer muddy, black, organic-rich turbidites. The next unconformity (at 0.8 Ma) may be related to the Pliocene–Pleistocene glaciation that resulted in a significant lowering of sea level, with an exposure of the Indian continental shelf, and that is marked by a change to more coarse, silty, turbidites that accumulated at a faster sedimentation rate. However, Cochran (1990) concluded that these variations in sedimentation were not related to the third-order variations in sea level.

Linking other unconformities dated at Leg 116 sites to tectonic events, or even the stratigraphic unconformities of the ocean's margins, is difficult. Scientists have had difficulty correlating (in a meaningful sense) deep-sea unconformities with those at the continental margins (Winterer, 1990). Traditionally, marine un-

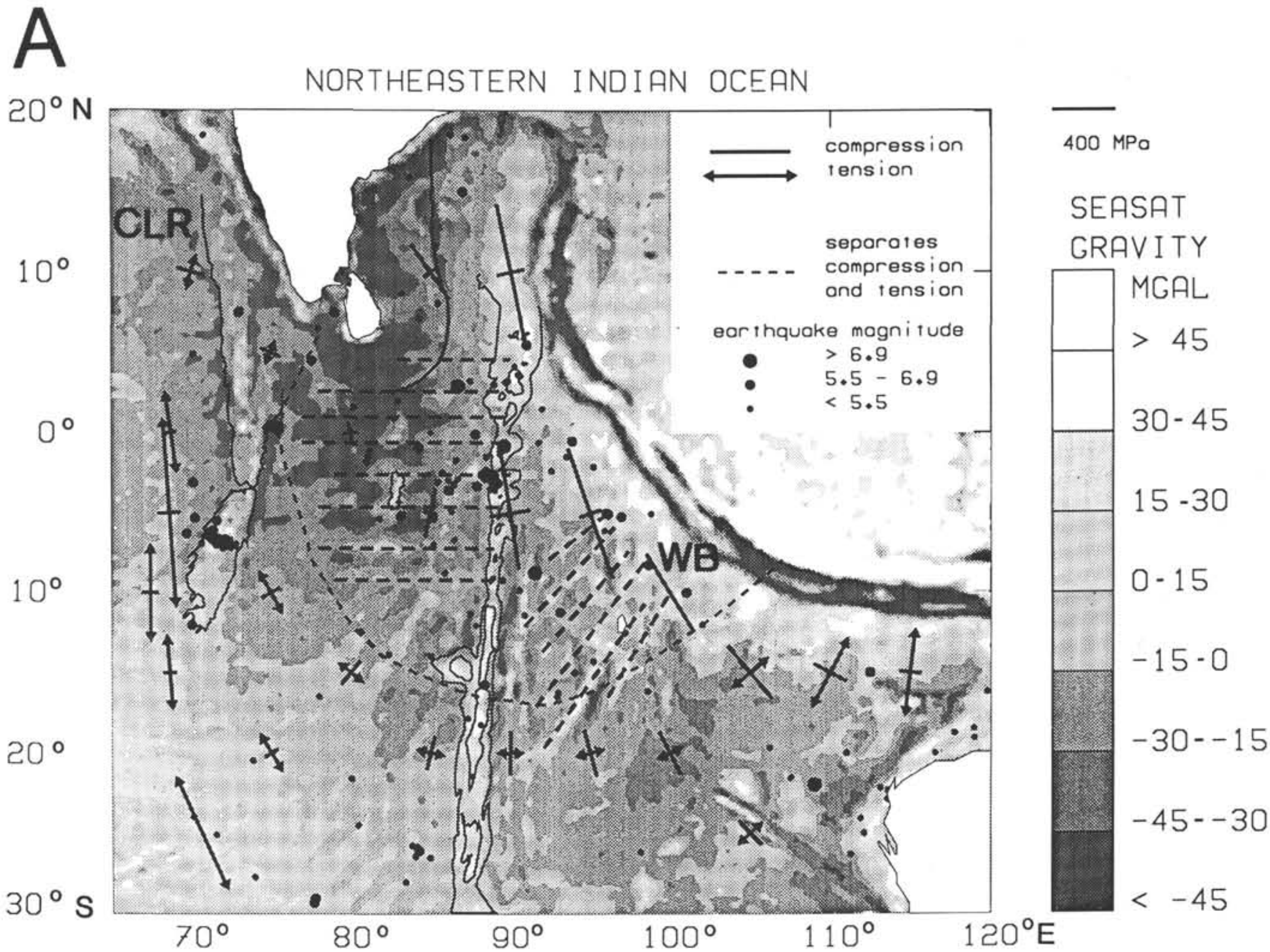


Figure 9. **A.** Fold axes from analysis of SEASAT data and predicted stresses from finite-element modeling (after Stein et al., 1989a). WB = Wharton Basin, CLR = Chagos Laccadive Ridge. **B.** Comparison of predicted stress field (left; (Cloetingh and Wortel, 1986) and stress directions inferred from earthquake focal mechanism data (right; Bergman (Bergman and Solomon, 1985).

conformities have been related to eustatic changes in sea level, although correlation is not good. Marine sedimentation is sensitive (Pickering et al., 1989) to changes in margin shape that affect erosion/deposition (e.g., turbidity currents/canyon cuttings), as demonstrated by Miller et al. (1987) for the Atlantic Ocean.

In addition, stress-induced changes in the shape of the margin will affect paleoceanography and, hence, sedimentation patterns (see Cloetingh et al., 1990). Until recently, the effect of changing the intraplate stress has not been considered a major factor affecting the continental margin stratigraphy and, hence, perhaps also the deep-sea record (Fig. 12). Hence, an examination of the stratigraphy of the margins of the Indian Ocean may give additional insight to these problems. Recently, such data has been acquired for the Northwest Australian margin, including information from Legs 122 and 123.

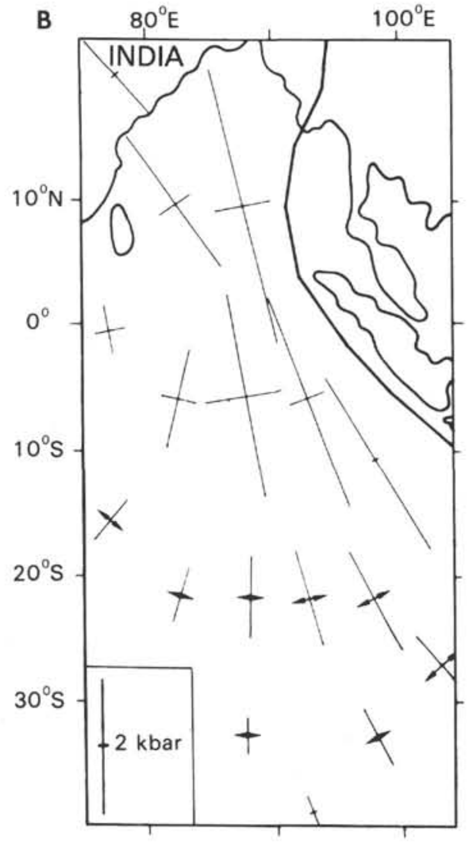
STRATIGRAPHIC RECORD OF THE NORTHWESTERN AUSTRALIAN MARGIN

The final configuration of the northwestern Australian margin was pre-determined by rifting within Gondwanaland during late Triassic–Late Jurassic time (Ludden, Gradstein, et al., 1990). Different estimates have been given for the timing of the onset of seafloor spreading of Greater India away from Australia. Using biostratigraphic data from Leg 123, this event was dated by Ludden, Gradstein, et al. (1990) at 140 Ma, while recent K/Ar dating of a celadonite sample from Site 765 has yielded an age of 155 Ma (J. Ludden, pers. comm., 1991). The latter age compares well with earlier estimates from DSDP data (Falvey, 1974; Exon et al., 1982; Veenstra, 1985) that provided an age of 155 Ma (Callovian) for the onset of spreading north of the Exmouth Plateau.

Stress modeling (Fig. 13) predicts tension to the north in the Argo Abyssal Plain and a change from a compressional to a tensional environment along the margin from the northeast to the southwest. We examined the geophysical and geological data from the northwestern Australian margin and recent drilling results from Legs 122 and 123 to determine the effect of a stress regime on stratigraphy.

Tectonic Setting

The rifting of the margin led to the formation of a horst-and-graben system (e.g., Barrow-Dampier Sub-basin, Malita Graben, Londonderry High, see Fig. 14; von Stackelberg et al., 1980), associated with volcanism (Exon et al., 1982). Seafloor spreading started in the north and started later during the Valangian (135 Ma) in the west and south.



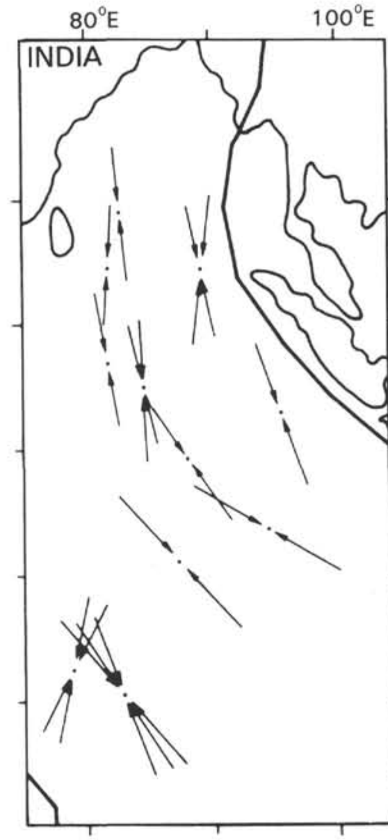


Figure 9 (continued).

Thus, during the Neocomian, the Argo, Gascoyne, and Cuvier abyssal plains developed (Powell and Luyendijk, 1982; Ludden, Gradstein, et al., 1990). Break-up and subsequent cooling of the lithosphere caused thermal subsidence of the rifted margin, which led to increasing water depths, as reflected in the change from clastic to carbonate deposition during Cretaceous time (Bradshaw et al., 1988). Carbonate deposition continued throughout the Tertiary. This depositional change was followed by a change in tectonic regime. Northwestern Australia, which occupied a position at a divergent plate boundary during most of the Phanerozoic, in Tertiary time, started to converge with the Eurasian Plate along the Banda Arc. In this collision zone, Cretaceous and Tertiary open-marine basin sediments were being incorporated into what has been described as either an imbricate subduction wedge (Karig et al., 1987) or an internal fold belt with a related foreland basin (Audley-Charles, 1986). For further details about the structural evolution of the margin, we refer the reader to a recent compilation by P. G. and R. R. Purcell (1988) of the Northwest Shelf of Australia.

In Figure 15, we summarize the structural framework and seafloor spreading events of the study area.

Stratigraphy

Figure 16A depicts a cross section through the northwestern margin of Australia in the region of the Exmouth Plateau that is based on an extensive data set from seismic surveys and well analysis. Figure 16B shows the stratigraphic sequences present on the Exmouth Plateau and the major reflectors that separate them. This stratigraphic framework was first established by Willcox and Exon (1976). More recent studies have been published by von Stackelberg et al. (1980), Colwell and von Stackelberg (1981), and von Rad and Exon (1983). Simultaneously, biostratigraphic studies were conducted by a number of authors, including Quilty (1980, 1981), von Rad et al. (1990), and Zachariasse (1991). Reflection seismic control consists of 12,000 km of data from the 1972 Bureau of Mineral Resources (BMR) continental margin survey, 9,300 km of GSI seismic data collected in 1976 and 1977 (Wright and Wheatly, 1979), and 30,000 km of subsequent petroleum industry seismic data. Most recently, seismic data have been collected by *Rig Seismic* research Cruises 7 and 8 (Exon, Williamson, et al., 1988).

The oldest reflector commonly seen is the latest Triassic to Lower Jurassic F-unconformity, interpreted by some scientists as a rift onset unconformity in the sense of Falvey (1974). The underlying Triassic sequence consists of early Triassic shallow marine muds and silts of the Locker Shale Formation and grades up into middle and late Triassic fluvio-deltaic sediments of the Mungaroo Formation. The whole sequence has a thickness of 2000 to 3000 m. The E-reflector is an unconformity previously thought to represent the break-up unconformity (Veevers, Heirtzler, et al., 1974), and to be of Callovian age. The underlying Early and Middle Jurassic sequences have a variable thickness that ranges from a few tens of meters on the Exmouth Plateau arch to more than 1000 m among the tilted fault blocks of the Barrow-Dampier sub-basin. These sequences consist largely of shallow-marine and fluvio-deltaic siltstones (Dingo Claystone, Brigadier Beds) and include coal measures. The Neocomian D-reflector represents the Neocomian break-up unconformity of the Exmouth Plateau and also represents the top of a deltaic sequence called the Barrow Group. The lowermost part of this sequence consists of Jurassic silts and shales of the upper Dingo Claystone, which vary greating in thickness. The overlying Barrow Group is a deltaic formation. It consists of prograding sands and siltstones in the south and prodelta muds north and northwest of Barrow

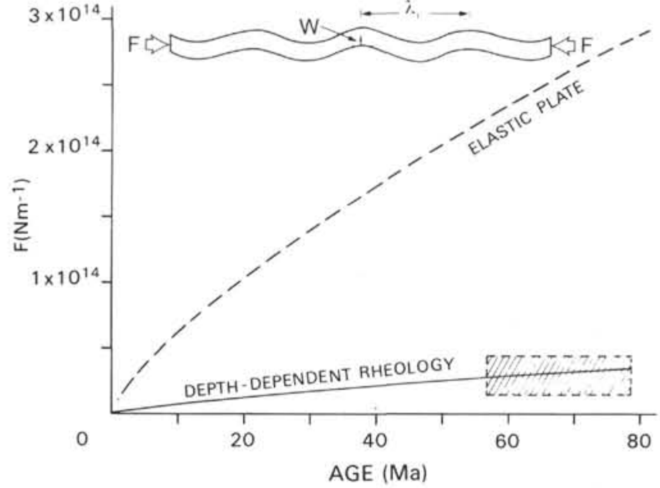


Figure 10. Buckling load, F (Nm^{-1}) vs. age for lithosphere, with a depth-dependent rheology inferred from rock-mechanics studies (solid line; Goetze and Evans, 1979) and for fully elastic oceanic lithosphere (dashed line; after McAdoo and Sandwell, 1985). Incorporation of depth-dependent rheology magnifies the amplitude, 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; Stein et al., 1989a).

Island. Maximum thickness varies from 1250 to 1750 m in places where the total sequence is present and thins rapidly northward (Barber, 1988). The mid-Cretaceous C-reflector is easy recognizable everywhere along the Northwest Shelf. It is a gentle unconformity in places. The underlying sequence consists of transgressive mid-Cretaceous mudstones (Gearle Siltstone) having a maximum thickness of 1500 m northwest of Barrow Island. The C-reflector marks a change in sedimentation type from terrigenous sedimentation to carbonate deposition of the overlying sequence. The B-reflector on the Cretaceous/Tertiary boundary is a mild unconformity of Paleocene age (Exon and Willcox, 1980). It overlies a well-bedded sequence of marls and calcilutites (the Gearle Siltstone and Haykock Marl) and underlies the Toolonga Calcilutite. The C–B sequence is thin on the Exmouth Plateau (approximately 100 m), but is much thicker west of the Swan Graben (approximately 300 m; Exon et al., 1982).

The A-reflector (or A_1 (Exon and Willcox, 1978]) is an unconformity formed during the Oligocene (von Stackelberg et al., 1980). The underlying Paleogene sequence is a thick, monotonous, prograding, eupelagic carbonate sequence of marls and calcilutites deposited in bathyal water depths. In some areas, an A_2-reflector of early Eocene age also has been described that is between the B- and A_1-reflector (Exon and Willcox, 1978). The total thickness between the A_1- and B-reflector varies between 100 and 600 m. The Oligocene to Holocene sequence between the seafloor and the A_1-reflector is gently unconformable in places. It consists of Miocene-to-Holocene eupelagic deposits of foraminifer nannofossil ooze (Haq, von Rad, O'Connell, et al., 1990) in a bathyal environment (Cape Range Group), with thicknesses ranging from 200 to 400 m.

Subsidence Analysis

We calculated water-loaded tectonic subsidence through time using methods discussed, for example, by Sclater and Christie (1980) and Bond and Kominz (1984). Here, we adopted Harland et al.'s (1982) time scale. Lithological effects, such as compaction, have been corrected for. Each stratigraphic unit between two chronostratigraphic horizons has been assigned sand, silt, shale, carbonate, anhydrite, and halite percentages, with each lithology responding according to its own depth-porosity curve: $\phi(z) = \phi_0$ e-cz. Minimum and maximum limits of lithological effects have been tested and showed that associated uncertainties in tectonic subsidence are on the order of up to several tens of meters. The parameters used for our compaction calculations are shown in Table 1. These compaction data are representative estimates covering a range of values commonly observed in quantitative subsidence analysis (Bond and Kominz, 1984). Our analysis of compaction parameters in the backstripping procedure taught us that sedimentary units of strongly different compactional behavior may (under conditions that are unlikely to occur on the Northwest Australian margin) lead to accelerations in tectonic and basement subsidence. For example, a sudden release of extreme overpressure can cause deviations from a smooth subsidence pattern of the same order of magnitude as deviations induced by tectonics. However, such drastic variations in compaction behavior can only produce accelerations in subsidence and do not explain the rapid phases of uplift that one can see in our subsidence curves. Paleowater depth data (inferred from faunal assemblages) were used for all the wells analyzed here. Uncertainties in water depth vary from a few tens of meters at the shelf to a few hundred meters in the plateau regions (see Fig. 17). We adopted the local Airy isostatic compensation for correcting the lithospheric response to loading and ignored the effect of finite strength of the lithosphere. However, this assumption does not significantly affect the shape of the inferred subsidence curves (Watts et al., 1982; Hegarty et al., 1988), particularly during the rift stage, when the lithosphere is mechanically weak (e.g., Hegarty et al., 1988). This also implies that flexural effects of intraplate stresses have been retained in these curves. Similarly, we ignored the reduced basement cooling that resulted from the blanketing effect of sediments (Turcotte and Ahern, 1977; Lucazeau and Le Douaran, 1985). To incorporate this effect in our analyses would require detailed knowledge of the thermal structure of the lithosphere throughout the basin's evolution. Furthermore, because of the long time scales on which lithospheric cooling operates (order of tens of millions of years), a correction for the effect of blanketing (order of magnitude up to 100 m) will not alter significantly the slope of the tectonic subsidence curves (Lucazeau and Le Douaran, 1985). Some of these wells were analyzed in earlier studies by Bradshaw et al. (1988) and Swift et al. (1988).

Figure 18 displays the results of the tectonic subsidence analysis for wells on the Northwest Shelf. These wells have been

Table 1. Maximum (upper part) and minimum (lower part) compaction parameters for different lithologies used in the backstripping procedure.

Lithology	ϕ_1	c_1	ϕ_0	c_0	zp	ρ_{gr}
Sand	0.29	0.216	0.4	0.51	1.0	2.65
Silt	0.42	0.375	0.6	1.0	0.5	2.68
Shale	0.50	0.475	0.7	1.1	0.5	2.72
Carbonate	0.52	0.442	0.78	1.33	0.5	2.71
Halite	0.00	0.10	0.0	0.10	0.0	2.03
Anhydrite	0.00	0.10	0.0	0.10	0.0	2.95
Sand	0.20	0.48	0.2	0.48	0.0	2.65
Silt	0.25	0.325	0.25	0.325	0.0	2.68
Shale	0.37	0.47	0.53	1.05	0.5	2.72
Carbonate	0.20	0.58	0.2	0.58	0.0	2.71
Halite	0.00	0.10	0.0	0.10	0.0	2.03
Anhydrite	0.00	0.10	0.0	0.10	0.0	2.95

After Bond and Kominz (1984).

grouped in accordance with the structural domains, as given in Figure 15. Dashed lines in the tectonic subsidence curves indicate a truncation in subsidence and correspond to stratigraphic unconformities. These truncations, each of which represents a hiatus, may be the result of complex basement movements or increased deep-water currents (Zachariasse, 1991).

Exmouth Plateau

On the Exmouth Plateau, a rapid subsidence (representing rifting during the Triassic) was followed by a period of slow subsidence or uplift during the Jurassic. In most wells, this is represented by a hiatus in the stratigraphic column (Fig. 18A). This is in agreement with the reported uplift of the Exmouth Plateau (Barber, 1988). After this period of slow subsidence or uplift, a phase of accelerated subsidence was again observed. This period of renewed subsidence spans a time slice between the Late Jurassic and the Late Cretaceous. Subsidence occurs in both a linear pattern (e.g., Mercury 1) and stepwise (e.g., Vinck 1/Site 763). The Eendracht 1/Site 762 curve exhibits a particularly prolonged period of no subsidence, followed by a phase of rapid subsidence. The Latest Jurassic to Early Cretaceous part of the generalized subsidence record coincides with the Argo/Gascoyne ocean spreading (Ludden, Gradstein, et al., 1990) between Greater India and Australia. The subsidence curves confirm that the process of breakup and crustal separation occurred in several (at least two) periods and that the onset of this process differs for individual wells within the Exmouth Plateau. The Vinck 1/Site 763 subsidence curve displays this two-stage rifting process well, while the Eendracht 1/Site 762 curve reflects only the second period of accelerated rifting. The other holes show a continuous transition between the two periods of rifting. From Late Cretaceous time onward, slow ongoing subsidence occurred, representing mainly the thermal cooling of the lithosphere. The Eendracht 1/Site 762 and Vinck 1/Site 763 drill holes exhibit a minor increase in tectonic subsidence during the late Miocene and Pliocene. The other holes lack information over this period because of poor recovery over this interval.

Kangaroo Syncline

To a large extent, the holes of the Kangaroo Syncline depict the same pattern as that observed for the Exmouth Plateau, with a first stage of rapid subsidence during the Triassic, followed by a truncation represented by a hiatus in the stratigraphic column. Subsequently, a long period of accelerated subsidence occurred, starting in the Late Jurassic until Late Cretaceous time, followed by a phase of more gradual subsidence. In most of the holes, no data were recorded for the Neogene. The tectonic subsidence curves of Zeewulf 1 and Saturn 1, however, do provide evidence for the occurrence of a phase of increased subsidence during early Pliocene time.

Rankin Trend

Wells on the Rankin Trend (Fig. 18C) record a stage of initial rapid subsidence, followed by a truncation, a pattern that is similar to the subsidence characteristics of the holes discussed above. The period of renewed rifting, however, seems to start somewhat later in this area. The profile (see Fig. 16) shows that the Rankin Trend was placed on a higher block, compared with adjacent areas. Therefore, the associated uplift may have retarded the subsequent subsidence phase. As observed in the other wells, rifting phases were followed by a roughly linear subsidence pattern and continued until Late Cretaceous time. In contrast, however, with the Exmouth Plateau and the Kangaroo Syncline, the Tertiary record is characterized by an overall mild uplift phase on which short-term phases of rapid uplift have been superimposed. A phase of uplift occurs from the Eocene onward, interrupted by a period of subsidence during the early Miocene, followed by uplift until Pliocene time, and a phase of renewed subsidence from late Pliocene time onward (Fig. 18C).

Brigadier Trend

Wells on the Brigadier Trend, which is located at the northeastern continuation of the Rankin Trend, show a phase of late Triassic subsidence and a subsequent decrease in subsidence rate toward Middle Jurassic time. This early phase of subsidence was followed by a truncation until Neocomian time. Inspection of the subsidence record of, for example, the Delambre 1 well, shows that in comparison to subsidence patterns observed for wells located on the Rankin Trend, subsidence in the Early and Middle Jurassic is more pronounced. During Late Jurassic time, truncation of the subsidence record occurred, followed by a linear pattern of subsidence during Cretaceous time and a pattern of alternating uplift and subsidence during the Cenozoic.

Barrow-Dampier Sub-basin

After a phase of rapid subsidence during the late Triassic, subsidence continued, with a lower but substantial rate during the Early and Middle Jurassic in the Barrow-Dampier Sub-basin. This is in contrast to the other areas, where this phase is characterized by a truncation. Accelerated subsidence occurred again during Late Jurassic and Neocomian time, followed by decreasing subsidence in the remaining part of the Cretaceous.

The incomplete subsidence record during Tertiary time shows a close similarity to patterns observed for the other regions on the flank of the northwestern margin in the area of the Exmouth Plateau.

Wombat Plateau

A tectonic subsidence curve (constructed from a combined data set of Sites 759, 760, 761, and 764) is presented in Figure 18F. A phase of rapid subsidence occurred during the late Triassic and Early Jurassic, followed by a phase of uplift and decreasing water depths, until the Early Cretaceous (Aptian).

Renewed rapid subsidence occurred from the mid-Cretaceous through the early Tertiary (Eocene), followed by slow subsidence until the Pliocene. From Pliocene time onward until the present, a phase of increased subsidence was recorded. The Wombat Plateau is made up of different blocks along both northeast-trending faults like those in the Montebello Trough and west-trending faults like those in the Wombat Halfgraben (Fig. 15). Deviations of the tectonic subsidence curve of the ODP sites on the Wombat Plateau from those subsidence patterns observed for the other areas are an expression of the highly faulted structural pattern of this region.

Abyssal Plains

Subsidence curves of the Argo and the Gascoyne abyssal plains (Fig. 18G) depict the expected exponential curve of oceanic lithosphere (Parsons and Sclater, 1977).

The different times of onset of rapid subsidence of 155 Ma at DSDP Site 261 in the northern Argo Abyssal Plain and 145 Ma at ODP Site 765 in the southern Argo Abyssal Plain suggest that sediments, deposited during a period of 10 m.y., are missing because the basement has been dated at 155 Ma (J. Ludden, pers. comm., 1991). The onset of rapid subsidence at 134 Ma at Site 766 in the Gascoyne Abyssal Plain supports the inferred existence (Ludden, Gradstein, et al., 1990) of a southward-propagating spreading center between Australia and Greater India.

In Figure 19, we summarize the timing of major tectonic events in the Indo-Australian region that occurred during the evolution of the northwestern Australian margin. Also given are the timing of the local tectonic events and the observed unconformities from

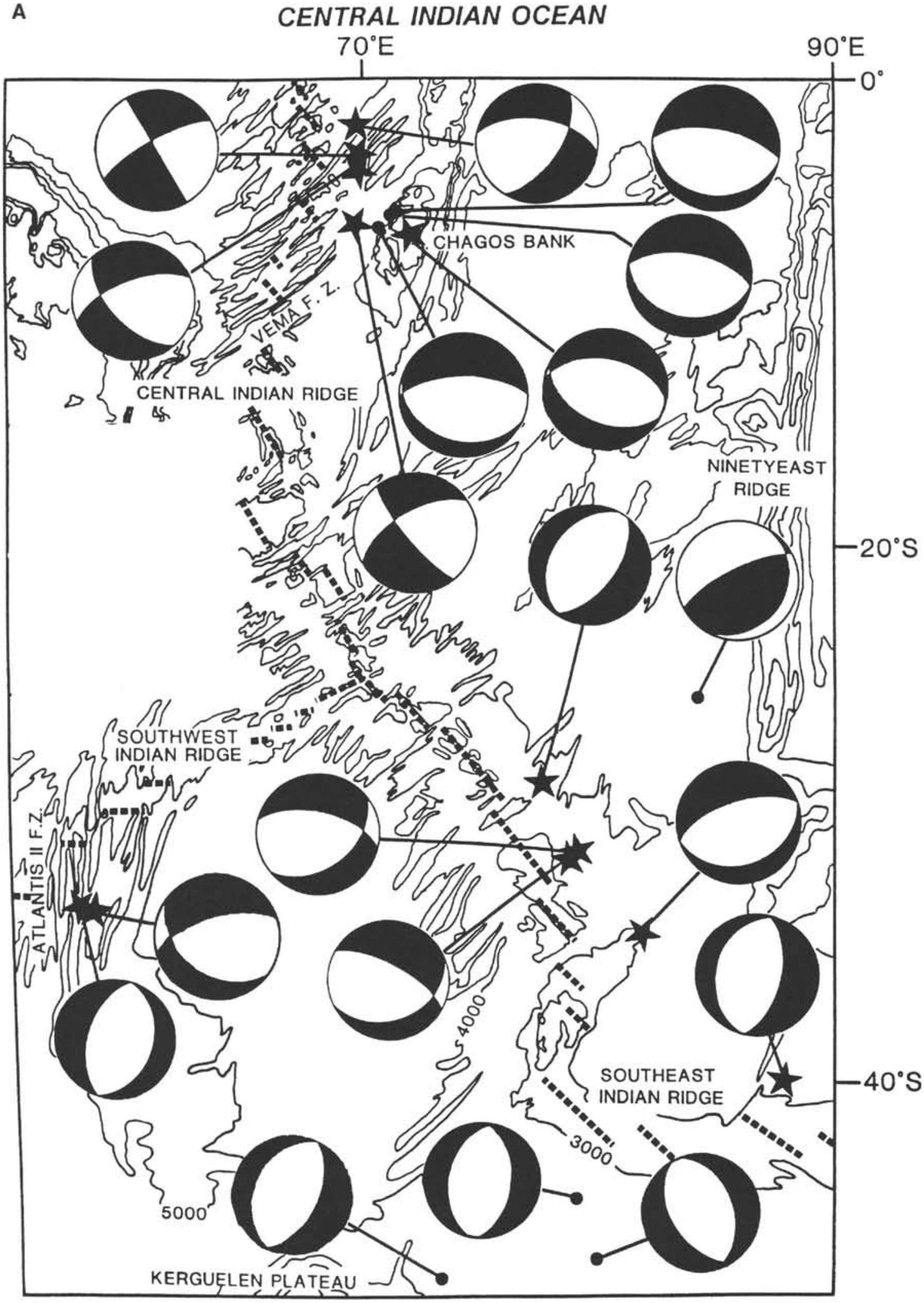


Figure 11. **A.** Extensional seismicity given by stars and dots in Chagos Bank and surrounding area (after Stein et al., 1987; Schlanger and Stein, 1987). **B.** Comparison of predicted stress (left) with stress directions from analysis of earthquake focal mechanism data (right) in Central Indian Ridge/ Chagos Bank area.

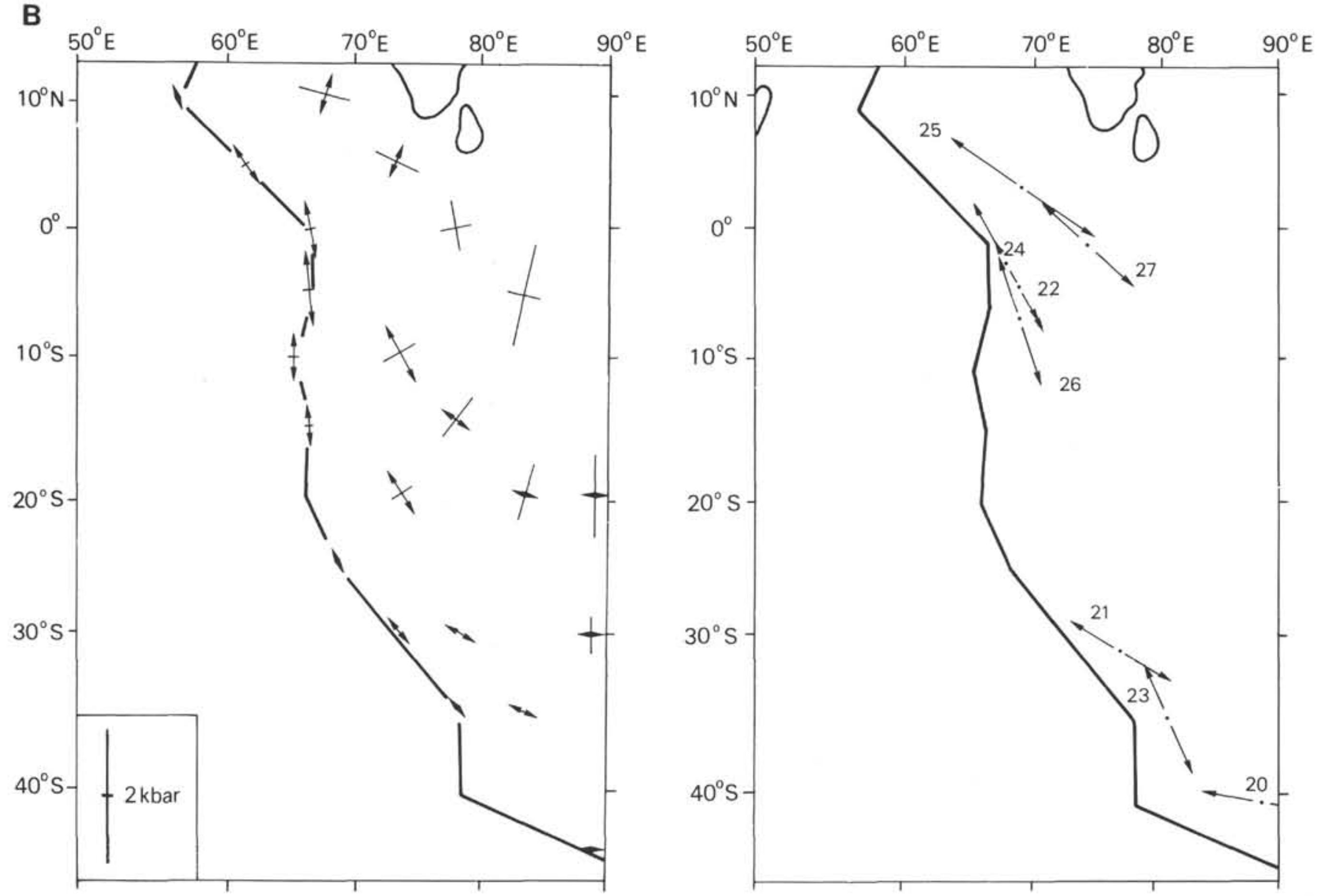


Figure 11 (continued).

well data. The major characteristic of the subsidence patterns for the basins on the Northwest Australian Shelf is an initial phase of rapid subsidence during the middle and late Triassic. This was followed by truncation during Early and Middle Jurassic time, a phase of rapid subsidence from Late Jurassic on, and slowing down from Late Cretaceous time onward. The latter phase of slow subsidence often occurs simultaneously with uplift phases during the late stage of the margin evolution. Thus, it appears that these subsidence patterns exhibit distinct deviations from the simple subsidence curves predicted by thermal models of basin evolution. These are to be expected (Kooi and Cloetingh, 1989a) in a region that has been subject to major tectonic phases throughout its evolution.

Stratigraphic Modeling

Modeling

Stratigraphy and synthetic subsidence curves are modeled along an east-west profile across the Exmouth Plateau and the Barrow-Dampier Sub-basin. The modeling approach used here is based on the stretching assumption of McKenzie (1978) and Royden and Keen (1980).

Depth-dependent stretching has been incorporated by employing crustal (δ) and subcrustal (β) stretching factors. Thermal calculations were performed using the finite-difference approach of Verwer (1977). This enabled us to incorporate finite- and multiple-stretching phases (Jarvis and McKenzie, 1980) and the effects of lateral heat flow (Watts et al., 1982; Cochran, 1983) into the model. The parameters used in the stratigraphic modeling are given in Table 2. The timing of the stretching phases is based on the results of the subsidence analysis and is presented in Figure 19. In our model, the lithosphere is made up of blocks 25 km wide, with each block characterized by a variable stretching factor. The cumulative crustal and subcrustal stretching factors for each box, δ and β, are given in the panel below the modeled stratigraphic section. These values were chosen to obtain an optimal fit to the stratigraphy and subsidence patterns. The temperature within each of the extended lithospheric blocks is primarily a function of the amount of stretching, duration of rifting, amount of stretching of neighboring blocks, and heat production (Kooi and Cloetingh, 1989b). We used an elastic plate analogy described by the 333°C isotherm for the mechanical properties of the lithosphere. The equivalent elastic thickness (EET) is reduced during each

Table 2. Parameters used in the forward modeling of basin stratigraphy.

Parameter	Symbol	Value
Initial lithospheric thickness	L	125 km
Initial crustal thickness	C	31.3 km
Coefficient of thermal expansion	α	3.4×10^{-5} °C^{-1}
Asthenospheric temperature	T^m	1333°C
Thermal diffusivity	κ	0.008 cm^{2}s^{-1}
Mantle density	ρ_m	3.33 gcm^{-3}
Crustal density	ρ_c	2.8 gcm^{-3}
Isotherm describing EET	T_e	333°C
Water density	ρ_w	1.03 gcm^{-3}
Surface porosity of sediment infill	$\phi(0)$	0.55
Characteristic depth constant	c	0.55 km^{-1}

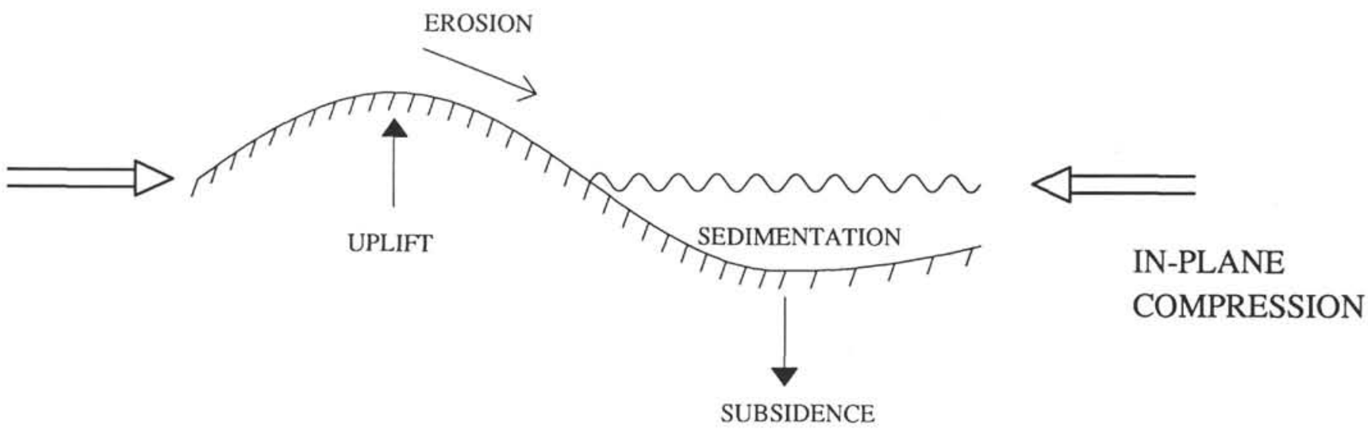


Figure 12. Compression-induced changes in sedimentation and erosion patterns at the center and flanks of depressions on flexed lithosphere.

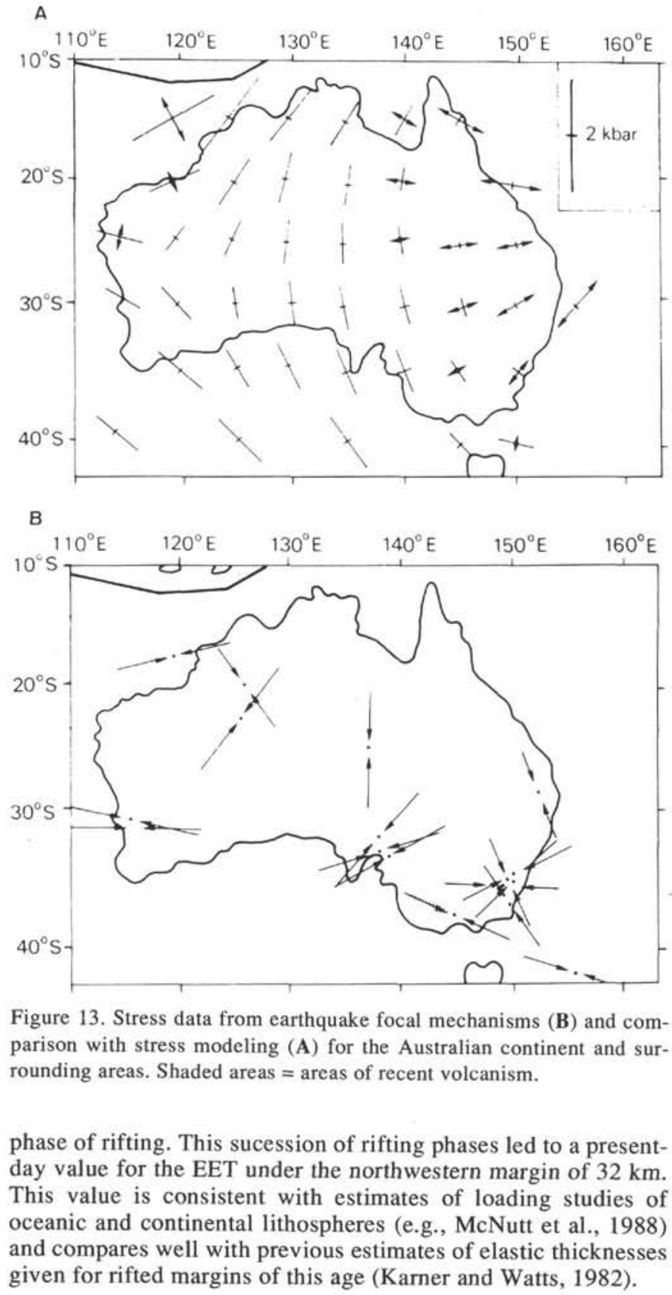


Figure 13. Stress data from earthquake focal mechanisms (**B**) and comparison with stress modeling (**A**) for the Australian continent and surrounding areas. Shaded areas = areas of recent volcanism.

phase of rifting. This sucession of rifting phases led to a present-day value for the EET under the northwestern margin of 32 km. This value is consistent with estimates of loading studies of oceanic and continental lithospheres (e.g., McNutt et al., 1988) and compares well with previous estimates of elastic thicknesses given for rifted margins of this age (Karner and Watts, 1982).

A significantly lower value for the EET of the lithosphere underlying the Exmouth Plateau area has been inferred from a recent study of satellite altimetry by Fowler and McKenzie (1989). However, in their study, these authors ignored the loading by the synrift sediments that locally form more than the one-half the total sedimentary pile. Consequently, they grossly underestimated the mechanical thickness of the lithosphere (Kooi et al., in press). Compaction of sediments is incorporated in the forward modeling of the stratigraphy using the standard porosity-depth relation of $\delta(z) = 0.55\ e^{-0.55}z$ for all sediments at the margin (see Table 2).

Results

Figure 20A indicates the generalized documented stratigraphy of the Exmouth Plateau region. Figure 20B shows the predicted stratigraphy for a model adopting uniform stretching and two finite-stretching phases at Late Triassic (238–210 Ma) and Late Jurassic (150–140 Ma) time, respectively. For the first stretching phase, a maximum β_1 of 1.8 was adopted for the Barrow-Dampier Sub-basin. For the second stretching phase, a maximum β_2 of 1.5 was used at this location. Also shown are synthetic subsidence curves for two wells (West Tryal Rocks 1 and Saturn 1) along the profile. Comparison with the observed stratigraphy and subsidence patterns (Figs. 16 and 18) demonstrates that this model successfully predicts the Triassic and Early Jurassic stratigraphy. However, the model fails to simulate the post-Middle Jurassic stratigraphy. In particular, the model does not produce the observed unconformity during Late Jurassic time. Also, the predicted thicknesses of the synrift sediments associated with the second stretching phase exceed the observed thicknesses. Similarly, inspection of the subsidence curves shows that the two-phase stretching model underestimates the thickness of the "post-rift" post-Neocomian sediments. These findings support the existence of a third stretching phase.

In a comparison of observed with modeled subsidence curves, the largest errors were introduced by uncertanties in paleobathymetry (see Fig. 17). Uncertainties are also intrinsically linked to age datings, which accuracy is on the order of a few million years, while a higher precision in age dating can be reached for the Cenozoic part of the subsidence curves.

Figure 20C gives the predicted stratigraphy for three stretching phases and incorporates a third phase of crustal thinning during Neocomian time (119–98 Ma), with a maximum β_3 of 1.2, a maximum β_2 of 1.5 and a maximum β_1 of 1.8. Also shown are synthetic subsidence curves for two wells (West Tryal Rocks 1 and Saturn 1) along the profile. Inspection of Figure 20B demonstrates that the model results in a much better fit to the overall stratigraphy. However, modeling fails to simulate some of the

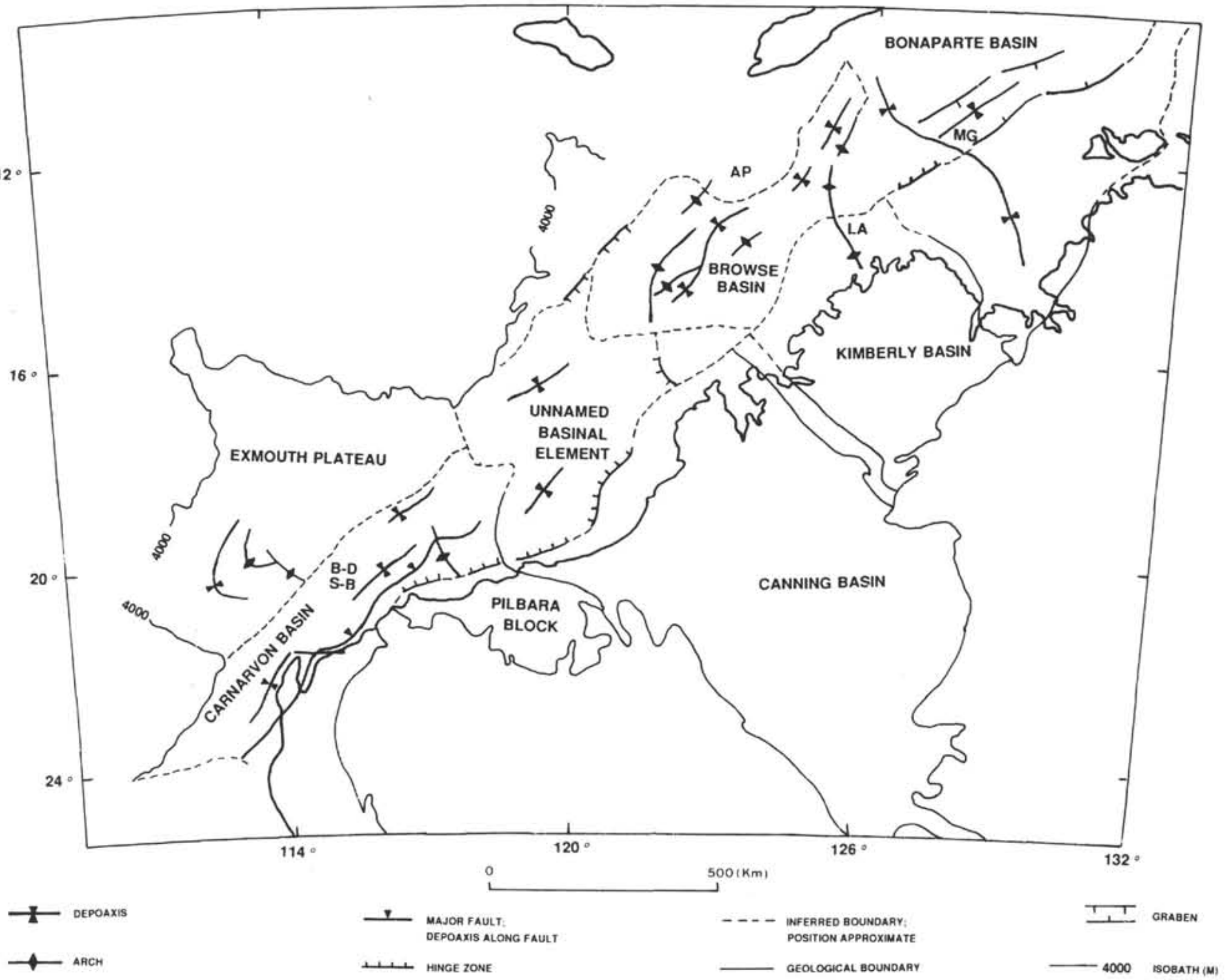


Figure 14. Location map and structural setting of the Northwest Shelf of Australia (modified after Barber, 1982). LA = Londonderry Arch, MG = Malita Graben, AP = Ashmore Platform, B-D S-B = Barrow-Dampier Sub-basin.

observed lateral variations in subsidence patterns that occurred at the positions of structural highs, overestimating the thickness of Late Jurassic sedimentary sequences at the wells located at the Rankin Trend and the Exmouth Plateau. Such features can be explained adequately by depth-dependent stretching in the lithosphere underlying the basins (e.g., Sclater and Celerier, 1987).

Figure 20D depicts the effect of incorporating depth-dependent stretching at the Exmouth Plateau in our analysis. We adopted crustal δ and subcrustal β stretching factors, with a δ_1 of 1.35 and a β_1 of 1.05 at the location of Eendracht 1 and Site 762, a $\delta_2 = 1.1$ and $\beta_2 = 1.35$ at the location of the Mercury 1 well, and $\delta_2 = \beta_2 = 1.5$ at the Barrow-Dampier Sub-basin. Also shown are synthetic subsidence curves for three wells (Tryal Rocks, West Tryal Rocks 1, and Saturn 1) along the profile. Compared to the predictions of the uniform stretching models, the incorporation of a two-layered stretching model results in a much better fit with the observed lateral variations in Late Jurassic stratigraphy. This better fit is the result of the effect of incorporating higher δ factors during the second stretching phase, inducing thermal uplift and leading to a decrease in the predicted thickness of the Late Jurassic section. Our modeling also successfully predicts the overall characteristics of the long-term patterns of tectonic subsidence during Cretaceous and Tertiary time. However, the observed subsidence patterns show short-term deviations from the predicted subsidence curves that cannot be attributed to thermal processes associated with the basin formation process. As discussed previously, these short-term deviations in the subsidence record have been frequently interpreted in terms of eustatic changes in short-term sea level, but can also be explained by short-term changes in tectonic stress. Here, we explore the latter possibility, focusing on the most pronounced short-term deviations in subsidence during late Oligocene (25 Ma), early Miocene (16 Ma), and late Miocene time (10 and 6 Ma).

Figure 20E illustrates the effects of changes in intraplate stress during Neogene time on the predicted stratigraphy. Also shown are synthetic subsidence curves for two wells (Tryal Rocks and West Tryal Rocks 1) along the profile. The paleostress field inferred from stratigraphic modeling is characterized by an increasing tensional stress field during 40 to 25 Ma, followed by a compressional stress regime from 25 to 16 Ma, a change to tension at 16 to 10 Ma, again a change into a stress regime of compression from 10 to 6 Ma, and finally, continuing compression from 6 Ma to the present (Fig. 21A). The inferred existence of a compressional stress field during late Miocene to Holocene time corresponds to the timing of the onset and occurrence of large-scale folding and faulting at the location of Leg 116 in the northeastern Indian Ocean. The magnitude of the inferred paleostress field should be considered as providing an upper limit based on an elastic rheology of the lithosphere, implying infinite strength. Modeling of the response of the lithosphere to intraplate stresses,

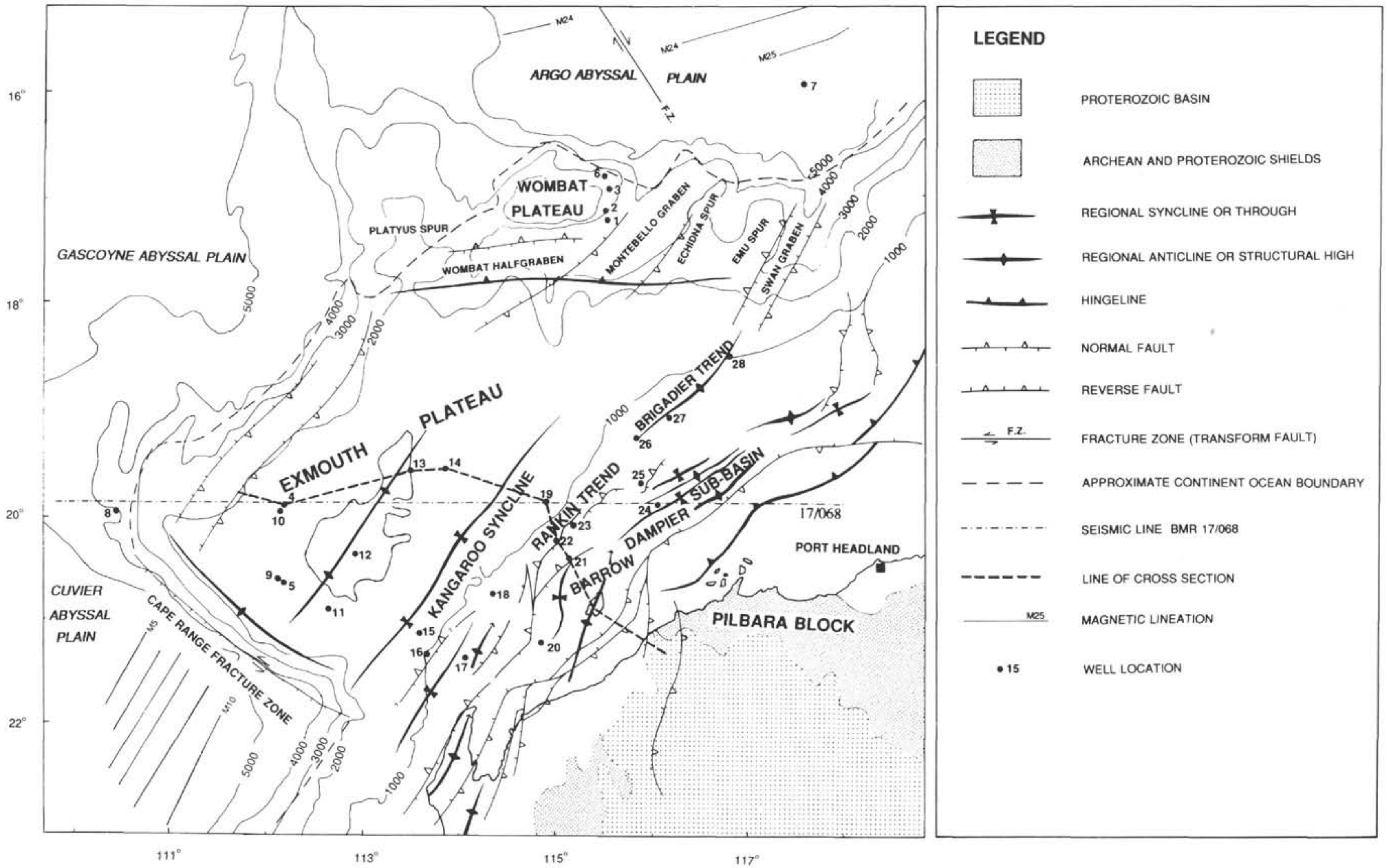


Figure 15. Location map and structural setting of the Exmouth Plateau region (modified after Barber, 1988; Woodside Offshore Petroleum, 1988). Numbers indicate well and site locations. ODP sites: 1 = 759, 2 = 760, 3 = 761, 4 = 762, 5 = 762, 6 = 764, 7 = 765, 8 = 766. Industry wells: 9 = Vinck 1, 10 = Eendracht 1, 11 = Sirius 1, 12 = Investigator 1, 13 = Jupiter 1, 14 = Mercury 1,15 = Zeewulf 1, 16 = Resolution 1, 17 = Novara 1, 18 = Zeepaard 1, 19 = Saturn 1, 20 = Rosily 1, 21 = Tryal Rocks 1, 22 = West Tryal Rocks 1, 23 = Sultan 1, 24 = Dampier 1, 25 = Goodwyn 1, 26 = Gandara 1, 27 = Brigadier 1, and 28 = Delambre 1.

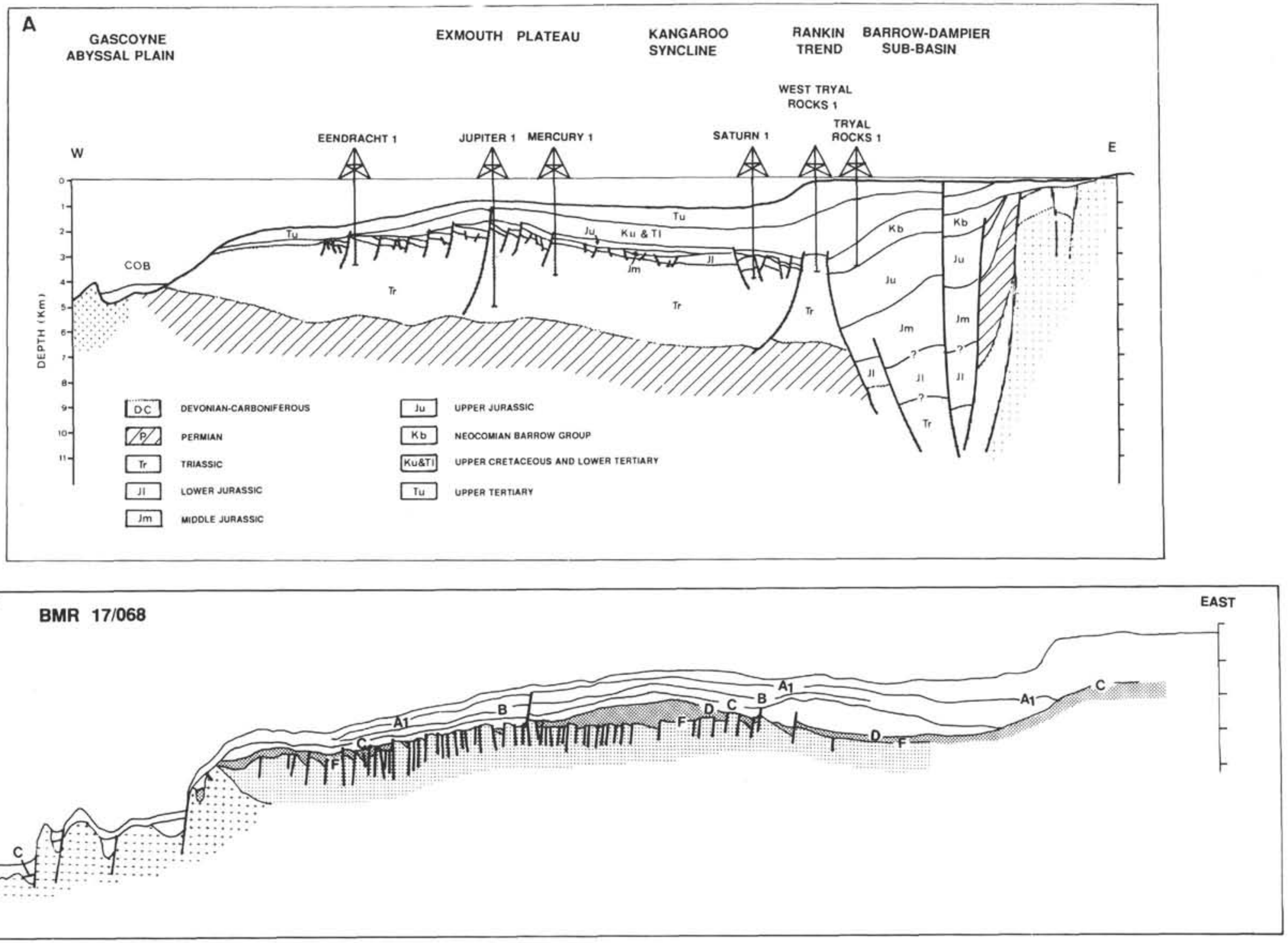
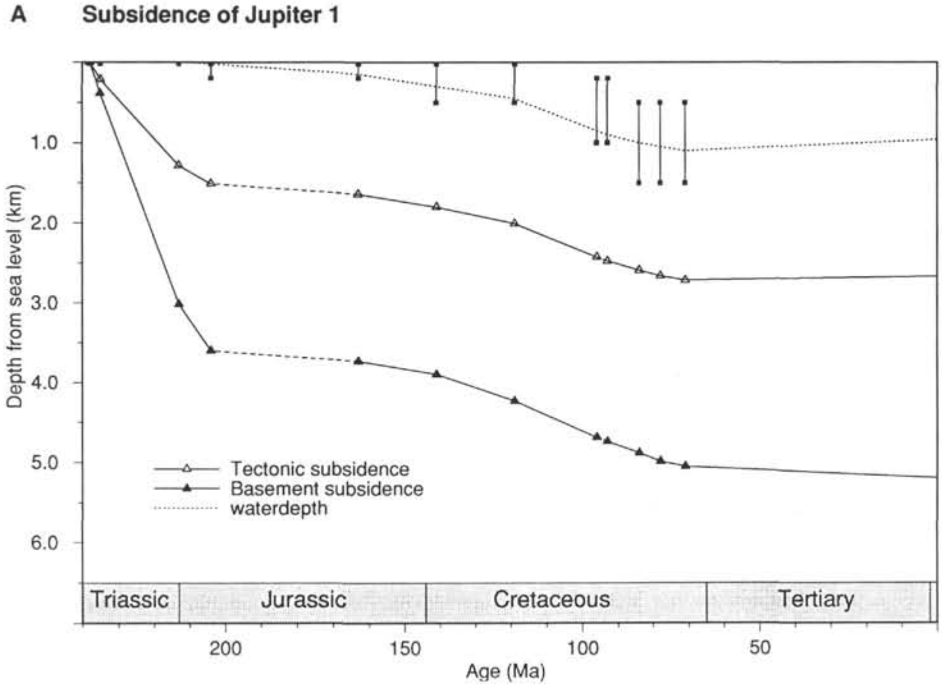


Figure 16. **A.** Stratigraphic cross section along profile given in Figure 15 through the Exmouth Plateau region, based on seismic data and well analysis (after Barber, 1988). **B.** Seismic Line 17/068 (location given in Fig. 15) across the Exmouth Plateau showing the positions of reflectors and their approximate ages (after Exon and Willcox, 1978). A_1 = Oligocene, B = Cretaceous/Tertiary boundary, C = Turonian, D = Early Cretaceous (Valanginian), F = late Triassic.



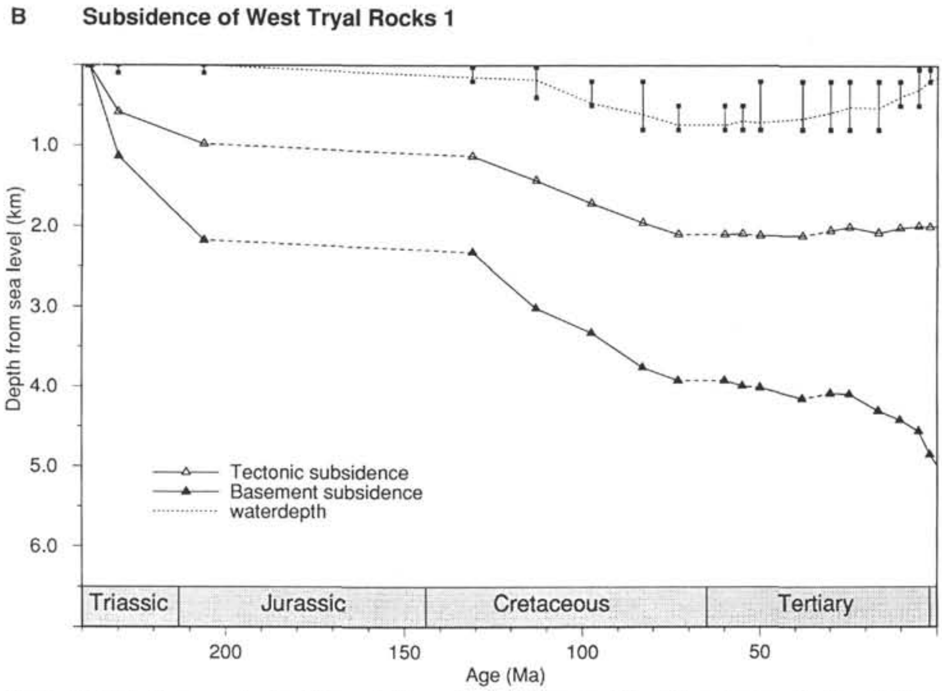


Figure 17. Subsidence analysis of three of the wells used here (position of wells given in Fig. 16A). **A.** Jupiter 1. **B.** West Tryal Rocks 1. **C.** Tryal Rocks 1. Dotted lines indicate the effect of correcting for paleobathymetry.

adopting a more realistic depth-dependent, brittle-ductile rheology based on extrapolation of rock-mechanics data (see Figs. 5, 6, 7, and 10), has shown that such a non-elastic rheology significantly lowers the magnitude of the inferred stresses. Considering the intrinsic uncertainties in the fine structure of the rheology of continental lithosphere, we concentrate here on the temporal changes in the patterns of the paleostress field that were derived from the stratigraphic modeling, rather than on their absolute magnitudes.

Changes in the intraplate stress field, which may have occurred as a result of plate reorganization (as at the start of subduction along the Banda Arc or during plate fragmentation and rifting [Wortel and Cloetingh, 1983]) are thought especially to influence the sedimentary record. However, this effect is dependent on position on the margin. For example, in the case of a position on a flexural node, tectonic movement would be relatively independent of intraplate stresses. On a position where the lithosphere showed a substantial deflection, we might expect stress-dependent differences in subsidence or uplift. In the Exmouth Plateau region on the Northwest Shelf, the largest sediment load is situated underneath the Barrow-Dampier Sub-basin, where the synrift sediments account for a substantial load and will cause the largest

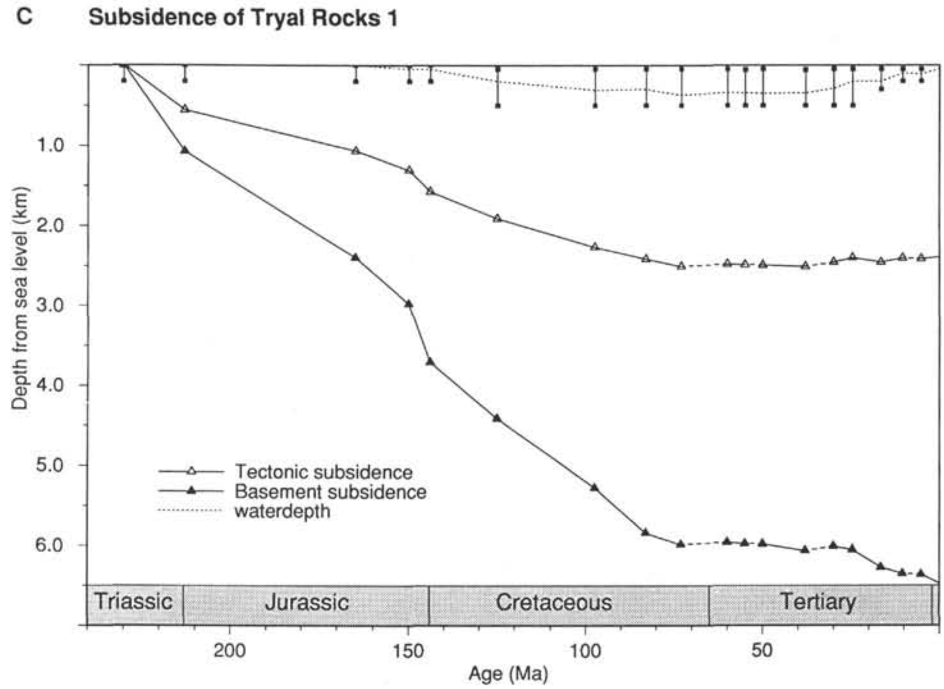


Figure 17 (continued).

deflection of the lithosphere. Intraplate stresses in the Indo-Australian Plate are expected to have their greatest influence on vertical movements in places such as this, where deflection is significant and the crust is weakened because of rifting.

Correlation of Changes in Sea Level with Local and Regional Tectonic Events

An interesting question is whether changes in depositional regime and stratigraphy observed on the Northwest Shelf reflect stress changes of regional extent in the neighborhood of the Northwest Shelf itself, changes in stress on a larger scale throughout the whole of the Indo-Australian Plate, or alternatively, eustatic changes in sea level.

Evidence exists to support a largely tectonic origin for the stratigraphic cycles on the Northwest Shelf. After studying the tectonic elements on the Northwest Shelf, Bradshaw et al. (1988) identified the existence of large-scale regional changes in the geometry of the structural elements of the margin through time. Bradshaw et al. (1988) also drew attention to the existence of a regional northwest tilt (flexure) of the Northwest Shelf that began during the Cretaceous and continued throughout the Cenozoic. Preexisting faults to the west of the Londonderry Arch (Fig. 14) were reactivated during the Miocene, a process that continues to the present. Widespread uplift in the Bonaparte Basin also occurred during the early Miocene. Farther south in the Carnarvon Basin, Miocene-to-Holocene strike-slip faulting disrupted the gradual subsidence patterns, forming several regional *en-echelon* anticlines (Bradshaw et al., 1988). This tectonic activity may reflect the Miocene–Pliocene continent-arc collision along the northern margin of the Australian continent (see Fig. 21B). As pointed out by Apthorpe (1988) time breaks of 2 to 4 m.y. duration remain between the cycles originally defined by Quilty (1980) in the early Eocene, at the end of the early Oligocene, and late in the middle Miocene. Bostratigraphic evidence exists of tectonism and downwarping immediately prior to the mid-Oligocene (Apthorpe, 1988). This period has been interpreted as marking the change from a passive margin to an incipiently active one. Coincident with this change is the beginning of shelf progradation on a massive scale. Some anticlines were present for all of the Cenozoic and much of the Cretaceous (Hocking, 1988). Woods (1988) and Wormald (1988) provided evidence for reactivation of normal faults in the Ashmore Platform and the Browse Basin at the northernmost edge of the Northwest Shelf during Miocene time, the result of the collision of northwestern Australia with the Southeast Asian Plate.

Inspection of Figure 21A shows an interesting correlation between the paleostress curve inferred from stratigraphic modeling and the timing and nature of local tectonic events on the Northwest Australian Shelf. For example, the reactivation of faults and faulting-induced folding in the Browse and Carnarvon basins during middle Miocene time occurred at a time when the inferred paleostress field was characterized by a high level of tensional stress. The tilting phases of the Wombat Plateau occurred at times when the inferred stress field was in compression, with the first phase coinciding with a change in the stress gradient during late Miocene time.

Figure 21B presents a comparison of the timing of tectonic phases during Tertiary time in the Indian Ocean region, with the timing of major tectonic phases in the other two large ocean basins. Inspection of Figure 21B demonstrates that most of the major changes in plate dynamics in the Indian Ocean occurred simultaneously with major plate-tectonic reorganizations in the Atlantic and Pacific oceans. This feature suggests a strong mechanical coupling between the different lithospheric plates, an observation also supported by recent studies of plate motions (Pollitz, 1988; Engebretson et al., 1985). In this way, stresses may be transmitted over large distances that exceed the dimensions of individual plates to even global scales. Thus, changes in plate stress may have occurred at times of major plate reorganizations (Rona and Richardson, 1978; Schwann, 1985) or might have taken place on a global scale, which explains the major changes observed in apparent sea level. Therefore, global synchroneity in

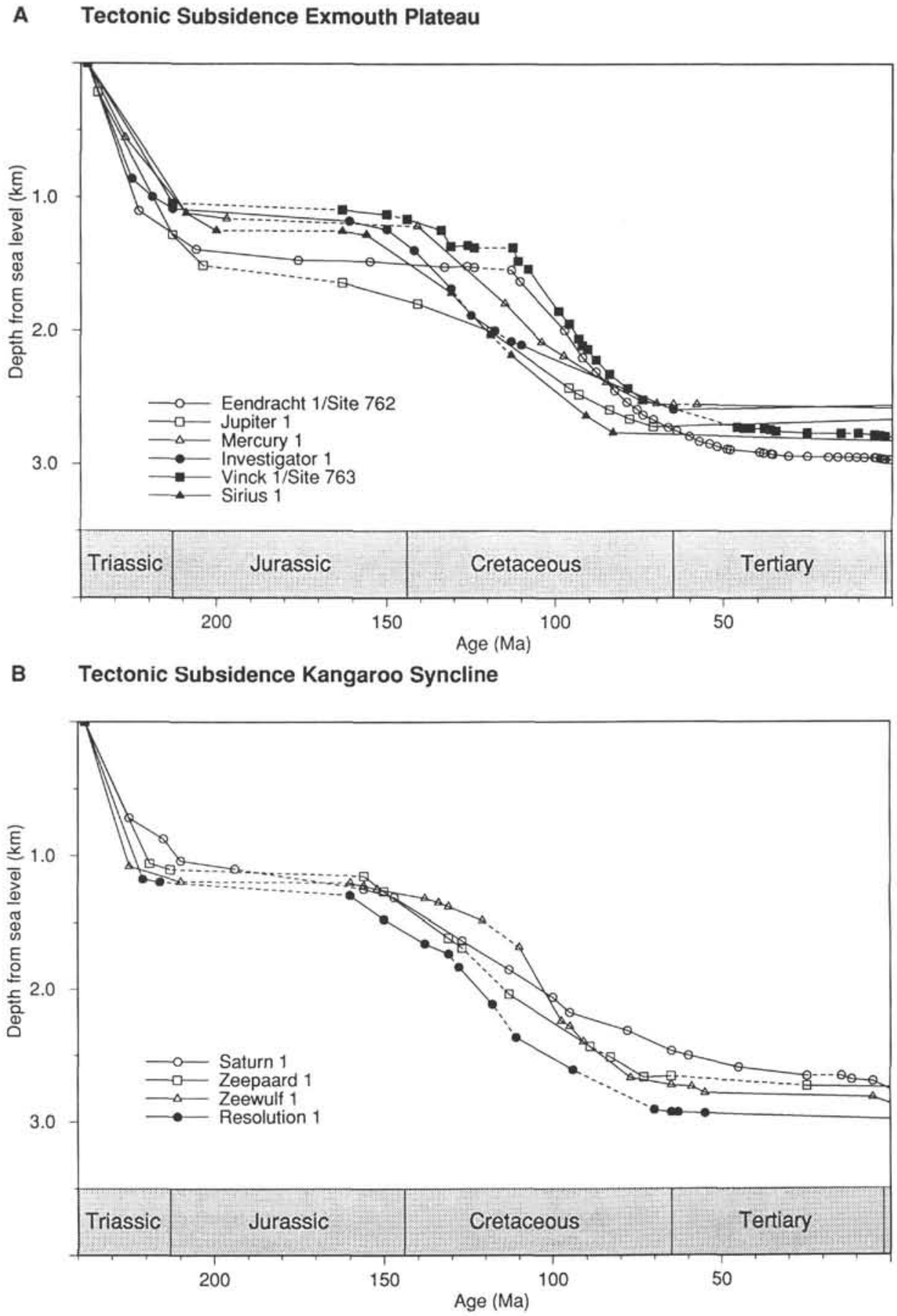
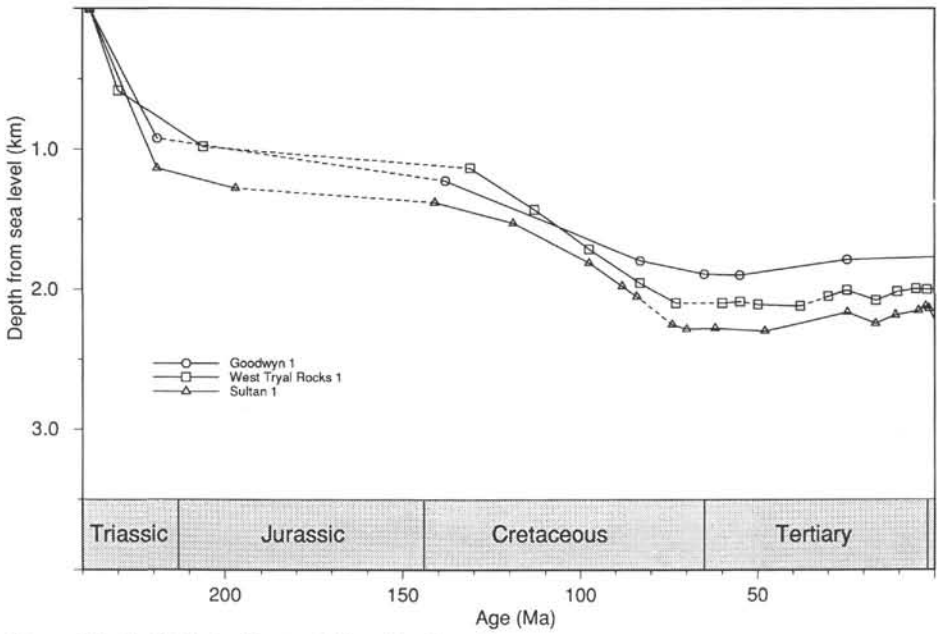


Figure 18. Tectonic subsidence curves for different domains of the Northwest Australian Shelf. **A.** Exmouth Plateau. **B.** Kangaroo Syncline. **C.** Rankin Trend. **D.** Brigadier Trend. **E.** Barrow-Dampier Sub-basin. **F.** Wombat Plateau. **G.** Abyssal plains.

apparent sea level cannot be used as a criterion to discriminate a global eustatic cause from a tectonic origin of the third-order sea-level cycles.

Tectonics and eustasy often occur simultaneously. As Cloetingh (1986) pointed out, the occurrence of a major plate reorganization during late Oligocene time simultaneously with the occurrence of a well-documented glacio-eustatic event might have caused the exceptional magnitude of the decrease in sea level at that particular time. In contrast, smaller changes in intraplate stress are by their nature limited to a more regional expression. This may explain the deviations observed along the northwestern Australian margin between regional sea-level charts (Fig. 21B) and predictions based on "global" sea-level cycles (Haq et al., 1987). In general, the regional curves lack the resolution displayed in Haq et al.'s curve. Within the limits of stratigraphic resolution, the curves for the Northwest Australian Shelf and northwestern Australia show a reasonable similarity in the timing of major lowstands and highstands of sea level. With the exception of the Pliocene unconformities, the major unconformities inferred from Leg 122 coincide with lowstands on the regional curves of Quilty (1980) and Apthorpe (1988). In contrast, the major unconformities from Leg 116 show a lack of correlation

C Tectonic Subsidence Rankin Trend



D Tectonic Subsidence Brigadier Trend

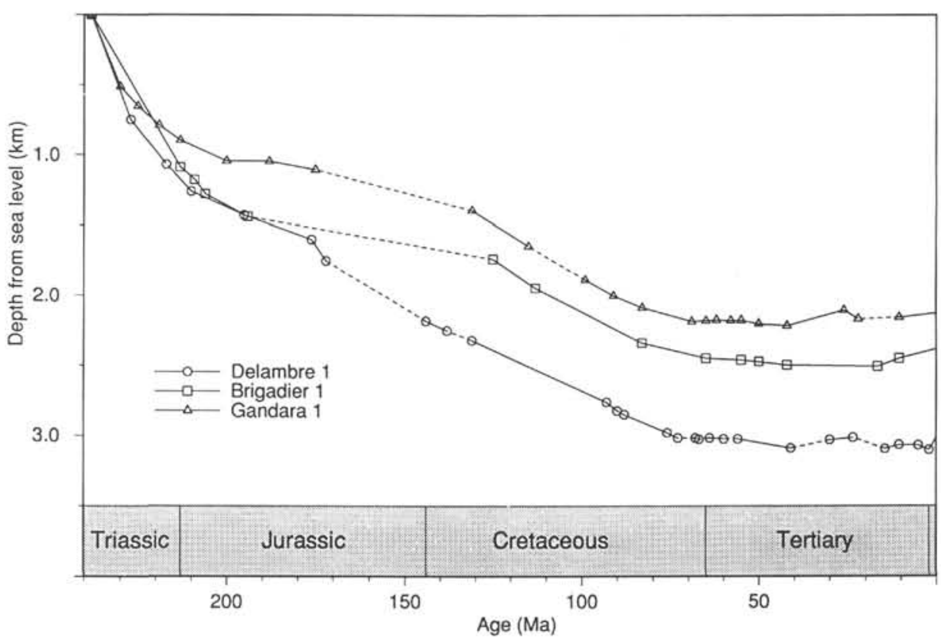
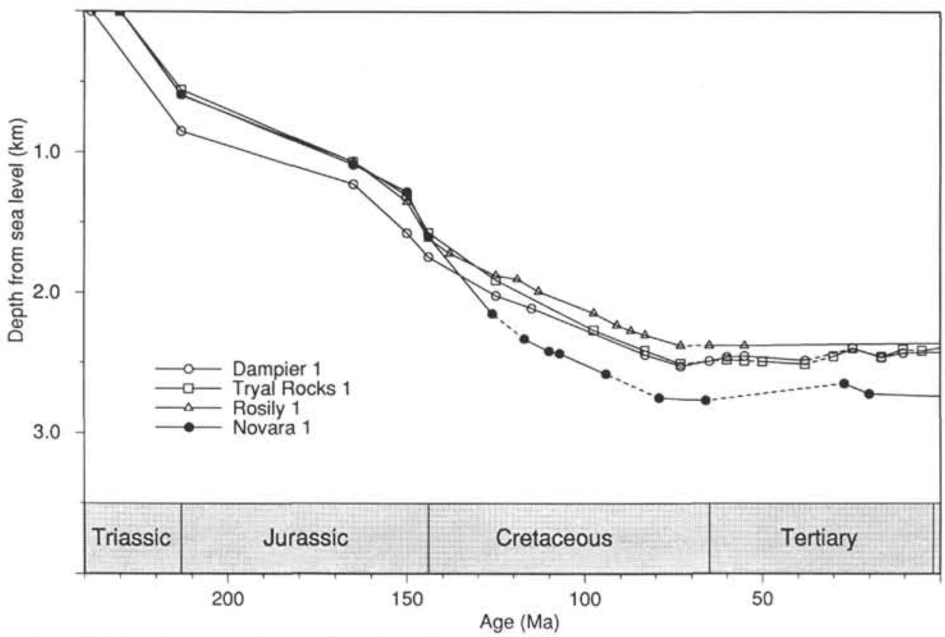


Figure 18 (continued).

with lowstands in the regional curves for the northwestern Australian region. Notwithstanding their sometimes substantial uncertainties in timing, the correlation of the unconformities found during Legs 116 and 122 with the lowstands in Haq et al.'s curve (1987) is almost perfect. Inspection of Figure 21C shows, however, that in general the regional curves deviate strongly from the "global" cycle chart. For example, the mid-Oligocene lowstand on the northwestern Australian margin correlates with a highstand in Haq et al.'s curve (Haq et al., 1987; Chaproniere, 1984; Carter, 1985), and the late Oligocene highstand in the northwestern Australian region coincides with a lowstand in Haq et al.'s curve.

Interestingly enough, the major unconformities drilled during Legs 116 and 122 coincide with plate-tectonic changes of a global character. Breakup phases in the Indo-Australian region mark the transition from lowstands to highstands in the regional sea-level curve of Quilty (1980). Changes in oceanic age distribution (Angevine et al., 1988) and the associated thermally induced increase in flexural rigidity provide a mechanical explanation (Watts, 1982) for the apparent increase in sea level, whereas the associated stress change can explain the subsequent lowstand, which occurs too quickly to be explained by a thermo-mechanical mechanism. The same is true for the correlation between breakup phases that occurred on a more global scale and most of the major decreases in apparent sea level given in Haq et al.'s curves (1987). Such a correlation between major tectonic phases and sea level decreases in Vail et al.'s curves (1977) was noted also by Bally (1982).

E Tectonic Subsidence Barrow-Dampier Sub-basin



F Tectonic Subsidence Wombat Plateau

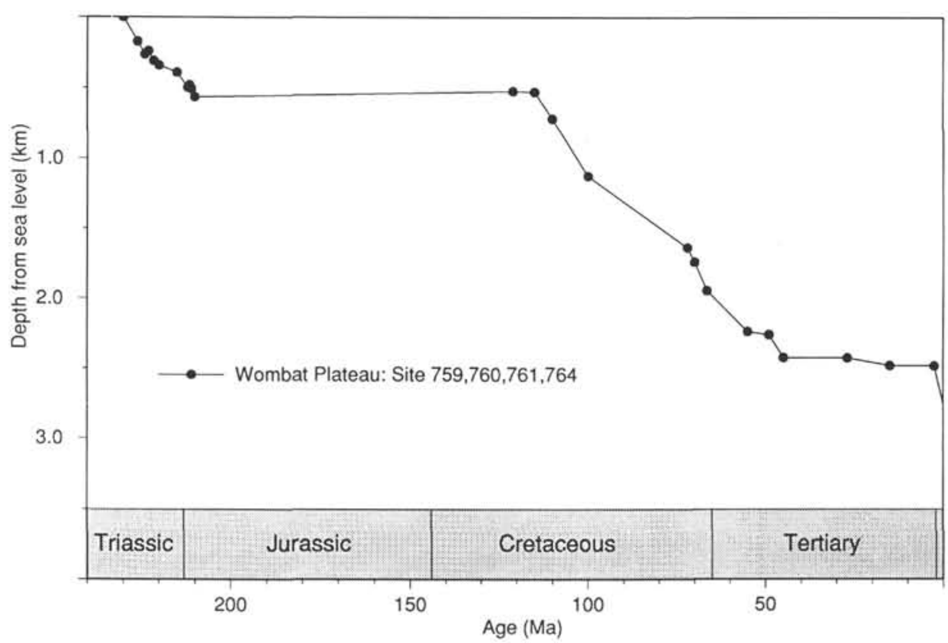


Figure 18 (continued).

DISCUSSION AND CONCLUSIONS

The timing of changes in the Neogene stratigraphic record and the inferred changes in the paleostress field of the Northwest Australian Shelf region studied using Leg 122/123 data are partly reflected in the tecto-stratigraphic evolution of the northeastern Indian Ocean, also studied using Leg 116 data, and in the record of the Indian continental margin. Unfortunately, little information has been published in a format useful for analysis, and the Indian margin studies of sea level are complicated by the Himalayan uplift and fan deposition.

Pandey (1986) suggested that the East Indian Shelf experienced major uplift and regression ~12 Ma. Given that the stress required to produce significant relative changes in sea level at passive margins is an order of magnitude less than that required to produce folding in the Indian Ocean lithosphere, if stress slowly increased with time, the continental margins should record the stress changes prior to ~7.5 Ma (Cochran, 1990) onset of deformation in the seafloor. The changes in intraplate stress in the northeastern Indian Ocean partly overlap with changes in stress associated with the onset of the Banda Arc collision. This collision probably resulted in an increase in the level of compressional stress. Both analyses of earthquake focal mechanism data and stress modeling suggest that the northwestern margin of Australia is under northeast- to southwest-oriented compression. Additional stress measurements will be needed, as well as paleostress

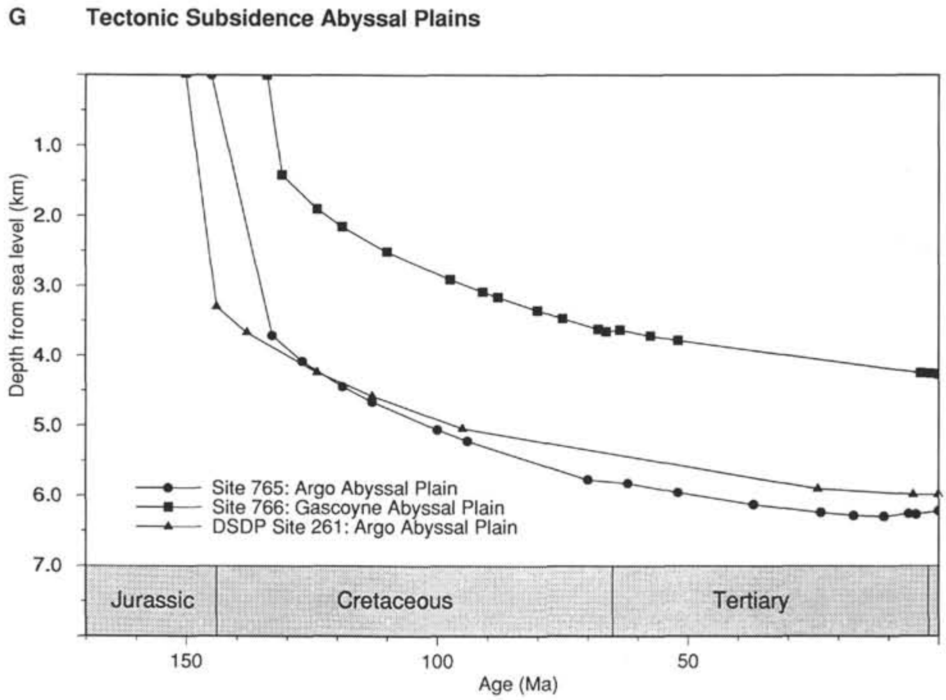


Figure 18 (continued).

data from microstructural studies, to test predictions of our model. A direct comparison of the predictions of stratigraphic modeling with stress data is further complicated by the effects of depth-dependent rheological models of the lithosphere and lateral variations in plate strength, as discussed above (see Figs. 5 and 6), and by the dynamic effects of the mechanics of basin formation on the subsidence record itself (Braun and Beaumont, 1989). The actual mechanics of basin formation primarily affects the record of vertical motions in a tensional stress regime during the rifting phase, but is of minor importance during the late-stage phase of rifted margin evolution. In this context, it is interesting to note that the paleostress field inferred from the stratigraphic modeling of the Northwest Shelf is consistent with observed tectonic activity, including the middle Miocene fault reactivation in the Browse and Carnarvon basins and Pliocene tilting events at the Wombat Plateau.

Rapid changes in basin subsidence during the late stage of basin evolution are not unique to the northwestern margin of Australia, but also have been observed for the U.S. Atlantic margin (Heller et al., 1982) and the rifted margins around the northern Atlantic Ocean (Cloetingh et al., 1990). Stratigraphic modeling has shown that the increase in the level of intraplate compression, associated with a major plate reorganization in the northern Atlantic (Klitgord and Schouten, 1986) during Pliocene time can explain the rapid subsidence phase and the presence of a thick Pliocene–Quaternary sediment wedge hitherto commonly attributed to glacio-eustasy. This study has shown that global tectonics during Neogene time might have significantly contributed to the record of changes in sea level.

Much future work will be needed to understand the role of stress changes on relative sea level changes and stratigraphy. The northwestern Australian margin is responding to a complex signal of eustatic sea level changes and far-field effects of stresses generated at the region's plate boundaries. This study has shown that the stratigraphy of the northwestern Australian continental margin was clearly affected by the onset of deformation in the Indian Ocean, with the deflections expected from the location of the region within the Indo-Australian Plate (Cloetingh and Wortel, 1985, 1986). As more data become available regarding Neogene stratigraphy of the carbonate platforms and continental margins, we can better understand the effects of intraplate stress changes and the causes of third-order changes in sea level.

ACKNOWLEDGMENTS

This research was supported by grants from Shell Research, the Petroleum Research Fund, administered by the American Chemical Society, and NATO Grant No. 0148/87. Exon and Williamson publish with the permission of the Executive Director, Bureau of Mineral Resources, Canberra, Australia. We thank Henk Kooi and Fred Beekman for discussions and Brett Mudford and an anonymous reviewer for thoughtful reviews.

REFERENCES

Angevine, C. L., Linneman, S. R., and Heller, P. L., 1988. Supercontinent breakup: effect on eustatic sea level and the oceanic heat flux. *In* Vlaar, N. J., Nolet, G., Wortel, M.J.R., and Cloetingh, S.A.P.L. (Eds.), *Mathematical Geophysics: A Survey of Recent Developments in Seismology and Geodynamics:* Dordrecht (Reidel), 389–400.

Apthorpe, M., 1988. Cainozoic depositional history of the North West Shelf. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia.* Proc. Pet. Expl. Soc. Aust. Symp., 55–84.

Audley-Charles, M. G., 1986. Timor-Tanimbar Trough: the foreland basin of the evolving Banda orogen. *In* Allen, P. A., and Homewood, P. (Eds.), *Foreland Basins.* Int. Assoc. Sed. Spec. Publ., 8:91–102.

Bally, A. W., 1982. Musings over sedimentary basin evolution. *Philos. Trans. R. Soc. London A,* 305:325–338.

Barber, P. M., 1988. The Exmouth Plateau deep water frontier: a case history. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia.* Proc. Pet. Expl. Soc. Aust. Symp., 173–187.

Bergerat, F., 1987. Stress fields in the European platform at the time of Africa-Eurasia collision. *Tectonics,* 6:99–132.

Bergman, E. A., and Solomon, S. C., 1984. Source mechanisms of earthquakes near mid-ocean ridges from body waveform inversion: implications for the early evolution of oceanic lithosphere. *J. Geophys. Res.,* 89:1415–1441.

_______, 1985. Earthquake source mechanisms from body-waveform inversion and intraplate tectonics in the northern Indian Ocean. *Phys. Earth Planet. Inter.,* 40:1–23.

Bond, G. C., and Kominz, M. A., 1984. Construction of tectonic subsidence curves for the early Paleozoic miogeocline, southern Canadian Rocky Mountains: implications for subsidence mechanisms, age of breakup and crustal thinning. *Geol. Soc. Am. Bull.*, 95:155–173.

Bradshaw, M. T, Yeates, A. N., Beynon, R. M., Brakel, A. T., Langford, R. P., Totterdell, J. M., and Yeung, M., 1988. Paleogeographic evolution of the North West Shelf Region. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia*. Proc. Pet. Expl. Soc. Aust. Symp., 29–54.

Braun, J., and Beaumont, C., 1989. A physical explanation of the relation between flank uplifts and the break up unconformity at rifted continental margins. *Geology*, 17:760–764.

Byerlee, J. D., 1978. Friction of rocks. *Pure Appl. Geophys.*, 116:615–626.

Carter, N. L., and Tsen, M. C., 1987. Flow properties of continental lithosphere. *Tectonophysics*, 136:27–63.

Carter, R. M., 1985. The Mid-Oligocene Marshall paraconformity, New Zealand: coincidence with global eustatic sea-level fall or rise. *J. Geol.*, 93:359–371.

Chaproniere, G.C.H., 1984. Oligocene and Miocene larger foraminiferida from Australia and New Zealand. B.M.R. Geol. Geophys. Bull., 188:1–98.

Cloetingh, S., 1986. Intraplate stresses: a new tectonic mechanism for fluctuations of relative sea level. *Geology*, 14:617–621.

_______, 1988. Intraplate stresses: a new element in basin analysis. *In* Kleinspehn, K., and Paola, C. (Eds.), *New Perspectives of Basin Analysis:* New York (Springer-Verlag), 205–230.

Cloetingh, S., and Wortel, R., 1985. Regional stress field of the Indian plate. *Geophys. Res. Lett.*, 12:77–80.

_______, 1986. Stress in the Indo-Australian plate. *Tectonophysics*, 132:49–67.

Cloetingh, S.A.P.L., Wortel, M.J.R., and Vlaar, N. J., 1982. Evolution of passive continental margins and initiation of subduction zones. *Nature*, 297:139–142.

Cloetingh, S., McQueen, H., and Lambeck, K., 1985. On a tectonic mechanism for regional sealevel variations. *Earth Planet. Sci. Lett.*, 75:157–166.

Cloetingh, S., Kooi, H., and Groenewoud, W., 1989a. *Intraplate Stresses and Sedimentary Basin Evolution*. Am. Geophys. Union, Geophys. Monogr., 48:1–16.

Cloetingh, S., Tankard, A. J., Welsink, H. J., and Jenkins, W.A.M., 1989b. Vail's coastal onlap curves and their correlation with tectonic events, offshore eastern Canada. *AAPG Mem.*, 46:283–293.

Cloetingh, S., Gradstein, F., Kooi, H., Grant, A. C., and Kaminski, M., 1990. Plate reorganization: a cause of rapid late Neogene subsidence and sedimentation around the north Atlantic? *J. Geol. Soc. London*, 147:495–506.

Cochran, J. R., 1983. Effects of finite rifting times on the development of sedimentary basins. *Earth Planet. Sci. Lett.*, 66:289–302.

_______, 1990. Himalayan uplift, sea level and the record of Bengal Fan sedimentation at the ODP Leg 116 Sites. *In* Cochran, J. R., Stow, D.A.V., et al., *Proc. ODP, Sci. Results*, 116: College Station, TX (Ocean Drilling Program), 397–414.

Cochran, J. R., Stow, D.A.V., et al., 1989. *Proc. ODP, Init. Repts.*, 116: College Station, TX (Ocean Drilling Program).

Collwell, J. B., and von Stackelberg, U., 1981. Sedimentological studies of Cainozoic sediments from the Exmouth and Wallaby Plateaus of northwest Australia. *BMR J. Aust. Geol. Geophys.*, 6:43–50.

Darwin, C., 1842. *The Voyage of the Beagle:* Tucson (Univ. of Arizona Press), republished 1959.

DeMets, C., Gorden, R. G., Argus, D. F., and Stein, S., 1990. Current plate motions. *Geophys. J. Int.*, 101:425–4780.

Dewey, J.F., 1988. Lithospheric stress, deformation and tectonic style: the disruption of Pangea and the closure of Tethys. *In* Audley-Charles, M. G., and Hallam, A. (Eds.), *Tethys and Gondwana*. Geol. Soc. Spec. Publ. London, 37:23–40.

Eberli, G. P., 1991. Growth and demise of isolated carbonate platforms: Bahamian controversies. *In* Müller, D., McKenzie, J., and Weissert, H. (Eds.), *Controversies in Modern Geology:* London (Academic Press), 231–248.

Embry, A. F., 1990. A tectonic origin for third-order depositional sequences in extensional basins: implications for basin modeling. *In* Cross T. A. (Ed.), *Quantitative Dynamic Stratigraphy:* New York (Prentice-Hall), 491–501.

Embry, A. F., and Podruski, J. A., 1988. Third-order depositional sequences of the Mesozoic successions of Sverdrup Basin. *Can. Soc. Petrol. Geol. Mem.*, 15:73–84.

Engebretson, D. C., Cox, A., and Gordon, R. G., 1985. Relative motions between oceanic and continental plates in the Pacific Basin. *Spec. Pap.—Geol. Soc. Am.*, 206.

Exon, N. F., von Rad, U., and von Stackelberg, U., 1982. The geological development of the passive margins of the Exmouth Plateau off northwest Australia. *Mar. Geol.*, 47:131–152.

Exon, N. F., and Willcox, J. B., 1978. Geology and petroleum potential of the Exmouth Plateau area off Western Australia. *AAPG Bull.*, 62:40–72.

_______, 1980. The Exmouth Plateau: stratigraphy, structure and petroleum potential. *Bull.—Bur. Miner. Resour., Geol. Geophys. (Aust.)*, 199.

Exon, N. F., Williamson P. E., et al., 1988. Preliminary post-cruise report, *Rig Seismic* Cruises 7 and 8: sedimentary basin framework of the northern and western Exmouth Plateau. *BMR Rec.*, 1988/30.

Falvey, D. A., 1972. Seafloor spreading in the Wharton basin (northeast Indian Ocean) and the breakup of eastern Gondwanaland. *APEA J.*, 12:86–88.

_______, 1974. The development of continental margins in plate tectonic history. *APEA J.*, 14:95–106.

Fowler, S., and McKenzie, D., 1990. Gravity studies of the Rockall and Exmouth Plateaux using Seasat *Basin Res.*, 2:27–34.

Galloway, W. E., 1987. Depositional and structural architecture of prograding clastic continental margins: tectonic influence on patterns of basin filling. *Norsk Geol. Tidsk.*, 67:237–251.

Geller, C. A., Weissel, J. K., and Anderson, R. N., 1983. Heat transfer and intraplate deformation in the central Indian Ocean. *J. Geophys. Res.*, 88:1018–1032.

Goetze, C., and Evans, B., 1979. Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics. *Geophys. J. R. Astron. Soc.*, 59:463–478.

Gordon, R. G., DeMets, C., and Argus, D. F., in press. Present-day motion between the Australian and Indian Plates: kinematic constraints on the distributed lithospheric deformation in the equatorial Indian Ocean. *Tectonics*.

Govers, R., Wortel, M.J.R., Cloetingh, S.A.P.L., and Stein, C. A., in press. Stress magnitude estimates from earthquakes in ocean plate interiors. *J. Geophys. Res.*

Greenlee, S. M., Schroeder, F. W., and Vail, P. R., 1988. Seismic stratigraphy and geohistory analysis of Tertiary strata from the continental shelf off New Jersey: calculation of eustatic fluctuations from stratigraphic data. *In* Sheridan, R. E., and Grow, J. R. (Eds.), *The Atlantic Continental Margin:* Alexandria (Geol. Soc. Am.), 437–444.

Gutenberg, B., and Richter, F. C., 1954. *Seismicity of the Earth:* Princeton (Princeton Univ. Press).

Hallam, A., 1988. A reevaluation of Jurassic eustasy in the light of new data and the revised Exxon curve. *Spec. Publ.—Soc. Econ. Palaeontol. Mineral.*, 42:261–274.

Haq, B. U., Hardenbol, J., and Vail, P. R., 1987. Chronology of fluctuating sea levels since the Triassic. *Science*, 235:1156–1167.

Haq, B. U., von Rad, U., O'Connell, S., et al., 1990. *Proc. ODP, Init. Repts.*, 122: College Station, TX (Ocean Drilling Program).

Harland, W. B., Cox, A. V., Llewellyn, P. G., Pickton, C.A.G., Smith, D. G., and Walters, R., 1982. *A Geologic Time Scale:* London (Cambrige Univ. Press).

Hegarty, K. A., Weissel, J. K., and Mutter, J. C., 1988. Subsidence history of Australia's southern margin: constraints on basin models. *AAPG Bull.*, 72:615–633.

Heller, P. L., Wentworth, C. M., and Poag, C. W., 1982. Episodic post-rift subsidence of the United States Atlantic continental margin. *Geol. Soc. Am. Bull.*, 93:379–390.

Hocking, R. M., 1988. Regional geology of the northern Carnarvon Basin. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia*. Proc. Pet. Expl. Soc. Aust. Symp., 97–114.

Houseman, G., and England, P., 1986. A dynamical model of lithosphere extension and sedimentary basin formation. *J. Geophys. Res.*, 91:719–729.

Howell, E. A., 1988. The Harriet oilfield. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia*. Proc. Pet. Expl. Soc. Aust. Symp., 391–401.

Hubbard, R. J., 1988. Age and significance of sequence boundaries on Jurassic and early Cretaceous rifted continental margins. *AAPG Bull.*, 72:49–72.

Jarvis, G. T., and McKenzie, D. P., 1980. Sedimentary basin formation with finite extension rates. *Earth Planet. Sci. Lett.*, 48:42–52.

Karig, D. E., Barber, A. J., Charlton, T. R., Klemperer, S., and Hussong, D. M., 1987. Nature and distribution of deformation across the Banda Arc—Australian collision zone at Timor. *Geol. Soc. Am. Bull.*, 86:18–32.

Karner, G. D., 1986. Effects of lithospheric in-plane stress on sedimentary basin stratigraphy. *Tectonics*, 5:573–588.

Karner, G. D., and Watts, A. B., 1982. On isostasy at Atlantic-type continental margins. *J. Geophys. Res.*, 87:2923–2948.

Kirby, S. H., 1985. Rock mechanics observations pertinent to the rheology of the continental lithosphere and the location of strain along shear zones. *Tectonophysics*, 119:1–27.

Klitgord, K. D., and Schouten, H., 1986. Plate kinematics and the Central Atlantic. *In* Vogt, P. R., and Tucholke, B. E. (Eds.), *The Geology of North America: The Western North Atlantic Region.* Geol. Soc. Am., 351–378.

Kooi, H., Burrus, J., and Cloetingh, S., in press. Lithospheric necking and regional isostasy at extensional basins: part 1, subsidence and gravity modeling with an application to the the Gulf of Lions margin (SE France). *J. Geophys. Res.*

Kooi, H., and Cloetingh, S., 1989a. Some consequences of compressional tectonics for extensional models of basin subsidence. *Geol. Rundsch.*, 98:183–195.

_______, 1989b. Intraplate stresses and the tectono-stratigraphic evolution of the Central North Sea. *AAPG Mem.*, 46:541–558.

Lambeck, K., Cloetingh, S., and McQueen, H., 1987. Intraplate stresses and apparent changes in sea level: the basins of north-western Europe. *In* Beaumont, C., and Tankard, A. T. (Eds.), *Basins of Eastern Canada and Worldwide Analogs.* Mem.—Can. Soc. Pet. Geol., 12:259–268.

Leg 123 Shipboard Scientific Party, 1989. The birth of the Indian Ocean. *Nature*, 337:506–507.

Letouzey, J., 1986. Cenozoic paleo-stress pattern in the Alpine foreland and structural interpretation in a platform basin. *Tectonophysics*, 132:215–231.

Letouzey, J., Werner, P., and Marty, A., in press. Fault reactivation and structural inversion. Backarc and intraplate compressive deformations, example of the eastern Sunda Shelf (Indonesia). *Tectonophysics.*

Liu, C.-S., Curray, J. R., and McDonald, J. M., 1983. New constraints on the tectonic evolution of the eastern Indian Ocean. *Earth Planet. Sci. Lett.*, 65:331–342.

Lucazeau, F., and Le Douaran, S., 1985. The blanketing effects of sediments in basins formed by extension: a numerical model. Applications to the Gulf of Lyon and Viking graben. *Earth Planet. Sci. Lett.*, 74:92–102.

Ludden, J. N., Gradstein, F. M., et al., 1990. *Proc. ODP, Init. Repts.*, 123: College Station, TX (Ocean Drilling Program).

McAdoo, D. C., Martin, C. F., and Poulouse, S., 1985. Seasat observation of flexure: evidence for a strong lithosphere. *Tectonophysics*, 116:209–222.

McAdoo, D. C., and Sandwell, D. T., 1985. Folding of oceanic lithosphere. *J. Geophys. Res.*, 90:8563–8569.

McKenzie, D. P., 1978. Some remarks on the development of sedimentary basins. *Earth Planet. Sci. Lett.*, 40:25–32.

McNutt, M. K., Diament, M., and Kogan, M. G., 1988. Variations of elastic plate thickness at continental thrust belts. *J. Geophys. Res.*, 93:8825–8838.

Miall, A. D., 1986. Eustatic sea level changes interpreted from seismic stratigraphy: a critique of the methodology with particular reference to the North Sea Jurassic record. *AAPG Bull.*, 70:131–137.

Miller, K. G., Fairbanks, R. G., and Mountain, G. S., 1987. Tertiary oxygen isotope synthesis, sea level history, and continental margin erosion. *Paleoceanography*, 2:1–19.

Nemec, W., 1988. Coal correlations and intrabasinal subsidence: a new analytical perspective. *In* Kleinspehn, K. L., and Paola, C. (Eds.), *New Perspectives in Basin Analysis:* New York (Springer-Verlag), 161–188.

Neprochnov, Y. P., Levchenko, O. V., Merklin, L. R., and Sedov, V. V., 1988. The structure and tectonics of the intraplate deformation area in the Indian Ocean. *Tectonophysics*, 156:89–106.

Pandey, J., 1986. Some recent paleontological studies and their implications on the Cenozoic stratigraphy of the Indian subcontinent. *ONGC Bull.*, 23:1–44.

Parkinson, N., and Summerhayes, C., 1985. Synchronous global sequence boundaries. *AAPG Bull.*, 69:685–687.

Parsons, B., and Sclater, J. G., 1977. An analysis of the variation of ocean floor bathymetry and heat flow with age. *J. Geophys. Res.*, 82:803–828.

Patriat, P., and Segoufin, J., 1988. Reconstruction of the Central Indian Ocean. *Tectonophysics*, 155:211–234.

Petroy, D. E., and Wiens, D. A., 1989. Historical seismicity and implications for diffuse plate boundary convergence in the northeast Indian Ocean. *J. Geophys. Res.*, 94:12301–12319.

Philip, H., 1987. Plio-quaternary evolution of the stress field in the Mediterranean zones of subduction and collision. *Ann. Geophys.*, 5B:301–320.

Pickering, K. T., Hiscott, R. N., and Hein, F. J., 1989. *Deep-marine Environments, Clastic Sedimentation and Tectonics:* London (Unwin Hyman).

Pitman, W. C., III, and Golovchenko, X., 1983. The effect of sea level change on the shelf edge and slope of passive margins. *Spec. Publ.—Soc. Econ. Paleontol. Mineral.*, 33:41–58.

Pollitz, F. F., 1988. Episodic North America and Pacific plate motions. *Tectonics*, 7:711–726.

Powell, T. S., and Luyendijk, B. P., 1982. The geological evolution of the continental margin of northwest Australia. *APEA J.*, 16:13–23.

Purcell, P. G., and Purcell, R. R. (Eds.), 1988. *The North West Shelf, Australia.* Proc. Pet. Expl. Soc. Aust. Symp.

Quilty, P. G., 1980. Sedimentation cycles in the Cretaceous and Cenozoic of western Australia. *Tectonophysics*, 63:349–366.

_______, 1981. Early Jurassic Foraminifera from the Exmouth Plateau, Western Australia. *J. Paleontol.*, 55:985–995.

Richardson, R. M., Solomon, S. C., and, Sleep, N. H., 1979. Tectonic stress in the plates. *Rev. Geophys. Space Phys.*, 17:981–1019.

Rona, P. A., and Richardson, E. S., 1978. Early Cenozoic plate reorganization. *Earth Planet. Sci. Lett.*, 40:1–11.

Royden, L., and Keen, C. E., 1980. Rifting processes and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. *Earth Planet. Sci. Lett.*, 51:343–361.

Schlager, W., 1988. Drowning unconformities on carbonate platforms. *Soc. Econ. Paleo. Mineral. Spec. Publ.*, 44:15–25.

Schlager, W., and Gamber, O., 1986. Submarine slope angles, drowning unconformities, and sef-erosion of limestone escarpments. *Geology*, 14:762–765.

Schlanger, S. O., and Stein, S., 1987. Charles Darwin and Captain Moresby on the drowning of the Great Chagos Bank: 19th century discovery of "aseismic" ridge seismicity in the Indian Ocean. *Eos*, 68:140–141.

Schwann, W., 1985. The worldwide active middle/Late Eocene geodynamic episode with peaks at 45 and 37 My b.p. and implications and problems of orogeny and sea-floor spreading. *Tectonophysics*, 115:197–234.

Sclater, J. G., and Celerier, B., 1987. Extensional models for the formation of sedimentary basins and continental margins. *Norsk Geol. Tidsk.*, 67:253–267.

Sclater, J. G., and Christie, P.A.F., 1980. Continental stretching: an explanation of the post-mid-Cretaceous subsidence of the Central North Sea Basin. *J. Geophys. Res.*, 85:3711–3739.

Shcherbakov, V. S., et al., 1989. A study of intraplate deformations by seismic methods in the Indian Ocean. *All-Union Conf. Plate Tectonics and Related Mineral Deposits,* Moscow, 11. (Abstract)

Shipboard Scientific Party, 1989a. Intraplate deformation and Bengal Fan sedimentation: background and objectives. *In* Cochran, J. R., Stow, D.A.V., et al., *Proc. ODP, Init. Repts.*, 116: College Station, TX (Ocean Drilling Program), 3–11.

_______, 1989b. ODP Leg 116 Site Survey. *In* Cochran, J. R., Stow, D.A.V., et al., *Proc. ODP, Init. Repts.*, 116: College Station, TX (Ocean Drilling Program), 197–210.

Steele, R. J., 1976. Some concepts of seismic stratigraphy with applications to the Gippsland basin. *APEA J.*, 16:67–71.

Stein, C. A., 1984. Part I: Heat transfer, seismicity and intraplate deformation in the central Indian Ocean. Part II: The transition between the Sheba Ridge and the Owen Basin: rifting of old oceanic lithosphere [Ph.D. dissert.]. Columbia Univ., New York.

Stein, C. A., Cloetingh, S., and Wortel, R., 1989a. SEASAT-derived gravity constraints on stress and deformation in the northeastern Indian Ocean. *Geophys. Res. Lett.*, 16:823–826.

_______, 1990. Kinematics and mechanics of the Indian Ocean diffuse plate boundary zone. *In* Cochran, J. R., Stow, D.A.V., et al., *Proc. ODP, Sci. Results*, 116: College Station, TX (Ocean Drilling Program), 261–277.

Stein, S., 1978. An earthquake swarm on the Chagos-Laccadive ridge and its tectonic implications. *Geophys. J. R. Astron. Soc.*, 55:577–588.

Stein, S., Cloetingh, S., Sleep, N. H., and Wortel, R., 1989b. Passive margin earthquakes, stresses and rheology. *In* Gregersen, S., and Basham, P. W. (Eds.), *Earthquakes at North Atlantic passive margins: Neotectonics and postglacial rebound.* NATO ASI Series (Kluwer Math. Phys. Sci.), 231–259.

Stein, S., Cloetingh, S., Wiens, D. A., and Wortel, R., 1987. Why does near ridge extensional seismicity occur primarily in the Indian Ocean? *Earth Planet. Sci. Lett.*, 82:107–113.

Stein, S., and Okal, E. A., 1978. Seismicity and tectonics of the Ninetyeast Ridge area: evidence for internal deformation of the Indian Plate. *J. Geophys. Res.*, 83:2233–2246.

Stephenson, R. A., and Cloetingh, S.A.P.L., in press. Some examples and mechanical aspects of continental lithospheric folding. *Tectonophysics.*

Stephenson, R. A., Ricketts, B. D., Cloetingh, S.A.P.L., and Beekman, F., 1990. Lithospheric folds in the Eurekan Orogeny, Arctic Canada? *Geology*,. 18:603–606.

Swift, M. G., Stagg, H.M.J., and Falvey, D. A., 1988. Heat flow regime and implications for oil maturation and migration in the offshore northern Carnarvon Basin. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia.* Proc. Pet. Expl. Soc. Aust. Symp., 539–551.

Sykes, L. R., 1970. Seismicity of the Indian Ocean and a possible nascent island arc between Ceylon and Australia. *J. Geophys. Res.*, 75:5041–5055.

Tankard, A. J., and Balkwill, H. R. (Eds.), 1989. Extensional tectonics and stratigraphy of the north Atlantic margins. *AAPG Mem.*, 46.

Turcotte, D. L., and Ahern, J. L., 1977. On the thermal and subsidence history of sedimentary basins. *J. Geophys. Res.*, 82:3762–3766.

Vail, P. R., Mitchum, R. M., Jr., and Thompson, S., III, 1977. Global cycles of relative changes in sea level. *AAPG Mem.*, 26:83–97.

Veenstra, E., 1985. Rift and drift in the Dampier Sub-basin, a seismic and structural interpretation. *APEA J.*, 25:177–189.

Veevers, J. J., 1988. Morphotectonics of Australia's northwestern margin. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia.* Proc. Pet. Expl. Soc. Aust. Symp., 19–27.

Veevers, J. J., Heirtzler, J. R., et al., 1974. *Init. Repts. DSDP*, 27: Washington (U.S. Govt. Printing Office).

Verwer, J. G., 1977. A class of stabilized three-step Runge-Kutta methods for the numerical integration of parabolic equations. *J. Comp. Appl. Math.*, 3:155–166.

Vink, G. E., Morgan, W. J., and Zhao, W. L., 1984. Preferential rifting of continents: a source of displaced terranes. *J. Geophys. Res.*, 89:10072–10076.

von Rad, U., and Exon, N.F., 1983. Mesozoic-Cenozoic sedimentary and volcanic evolution of the starved passive margin off northwest Australia. *In* Watkins, J. S., and Drake, C. L. (Eds.), *Studies in Continental Margin Geology.* AAPG Mem., 34:253–281.

von Rad, U., Schott, M., Exon, N. F., Quilty, P. G., Mutterlose, J., and Thurow, J., 1990. Mesozoic sedimentary and volcanic rocks dredged from the northern Exmouth Plateau: petrography and microfacies. *BMR J. Aust. Geol. Geophys.*, 11:449–476.

von Stackelberg, U., Exon, N. F., von Rad, U., Quilty, P., Shafik, S., Beiersdorf, H., Seibertz, E., and Veevers, J. J., 1980. Geology of the Exmouth and Wallaby Plateaus off northwest Australia: sampling of seismic sequences. *BMR J. Aust. Geol. Geophys.*, 5:113–140.

Watts, A. B., 1982. Tectonic subsidence, flexure and global changes of sea level. *Nature*, 297:469–474.

Watts, A. B., Karner, G. D., and Steckler, M. S., 1982. Lithospheric flexure and the evolution of sedimentary basins. *Philos. Trans. R. Soc. London A*, 305:249–281.

Watts, A. B., and Thorne, J., 1984. Tectonics, global changes in sea level and their relationship to stratigraphical sequences at the U.S. Atlantic continental margin. *Mar. Pet. Geol.*, 1:319–339.

Weissel, J. K., Anderson, R. N., and Geller, C. A., 1980. Deformation of the Indo-Australian plate. *Nature*, 287:284–291.

Wiens, D. A., 1986. Historical seismicity near Chagos: a complex deformation zone in the equatorial Indian Ocean. *Earth Planet. Sci. Lett.*, 76:350–360.

Wiens, D. A., DeMets, C., Gordon, R. G., Stein, S., Argus, D., Engeln, J. F., Lundgren, P., Quible, D., Stein, C., Weinstein, S., and Woods, D. F., 1985. A diffuse plate boundary model for Indian Ocean tectonics. *Geophys. Res. Lett.*, 12:429–432.

Wiens, D. A., and Stein, S., 1983. Age dependence of oceanic intraplate seismicity and implications for lithospheric evolution. *J. Geophys. Res.*, 88:6455–6668.

_______, 1984. Intraplate seismicity and stresses in young oceanic lithosphere. *J. Geophys. Res.*, 89:11442–11464.

Willcox, J. B., and Exon, N. F., 1976. The regional geology of the Exmouth Plateau. *APEA J.*, 16:1–11.

Winterer, E. L., 1991. Pelagic sediments, seismic reflectors, and changes in sea level. *In* Muller, D., Weissert, H., and McKenzie J. (Eds.), *Controversies in Modern Geology:* London (Academic Press), 211–230.

Withjack, M. O., and Gallagher, J. J., 1983. The rifted basins of eastern India. *SEAPEX Proc.*, 6:41–57.

Woods, E. P., 1988. Extensional structures of the Jabiru Terrace, Vulcan Sub-basin. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia.*, Proc. Pet. Expl. Soc. Aust. Symp., 311–330.

Woodside Offshore Petroleum Pty. Ltd, 1988. A review of the petroleum geology and hydrocarbon potential of the Barrow-Dampier Sub-basin and environs. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia.* Proc. Pet. Expl. Soc. Aust. Symp., 115–128.

Wormald, G., 1988. The geology of the Challis oilfield—Timor Sea, Australia. *In* Purcell, P. G., and Purcell, R. R. (Eds.), *The North West Shelf, Australia.* Proc. Pet. Expl. Soc. Aust. Symp., 425–437.

Wortel, R., and Cloetingh, S., 1983. A mechanism for fragmentation of oceanic plates. *AAPG Mem.*, 34:793–801.

Wright, A. J., and Wheatley, T. J., 1979. Trapping mechanisms and the hydrocarbon potential of the Exmouth Plateau, Western Australia. *APEA J.*, 19:19–29.

Zachariasse, W. J., 1992. Neogene planktonic foraminifers from ODP Sites 761 and 762 off northwest Australia. *In* von Rad, U., Haq, B. U., et al., *Proc. ODP, Sci. Results*, 122: College Station, TX (Ocean Drilling Program), 000–000.

Ziegler, P. A., 1987. Late Cretaceous and Cenozoic intra-plate compressional deformations in the Alpine foreland—a geodynamic model. *Tectonophysics*, 137:389–420.

_______, (Ed.), 1988. Evolution of the Arctic-North Atlantic and the western Tethys. *AAPG Mem.*, 43.

Zoback, M. L., and the World Stress Map Team, 1989. Global patterns of tectonic stress. *Nature*, 341:291–298.

Zuber, M. T., 1987. Compression of oceanic lithosphere: an analysis of intraplate deformation in the Central Indian Ocean Basin. *J. Geophys. Res.*, 92:4817–4825.

Date of initial receipt: 10 July 1990
Date of acceptance: 14 May 1991
Ms 123B-144

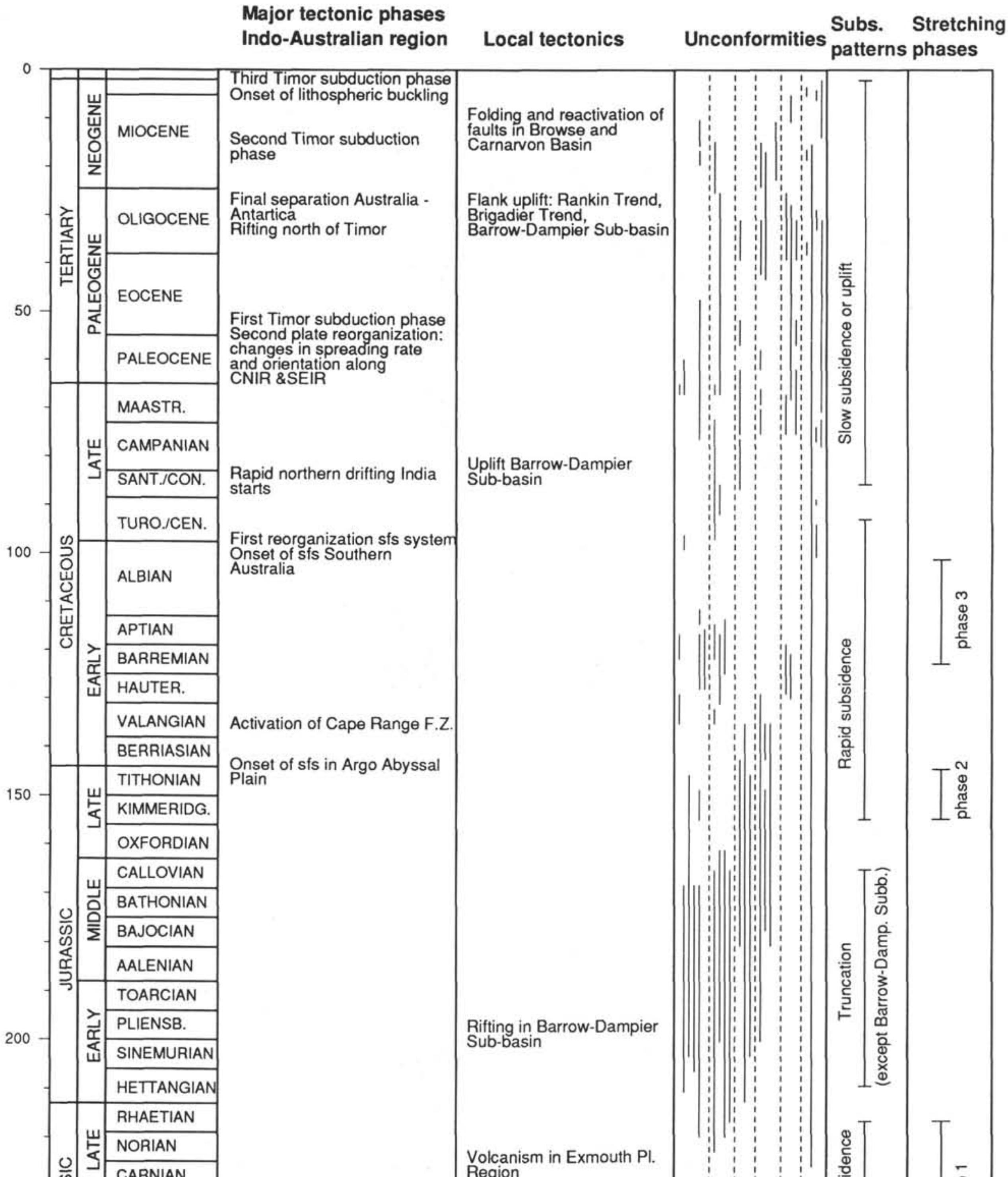


Figure 19. Tectonic phases and subsidence patterns of the Northwest Australian Shelf region. Columns from left to right give timing of major tectonic phases in the Indo-Australian region (After Exon et al., 1982; Patriat and Segoufin, 1988; Barber, 1988; Audley-Charles, 1986; Hegarty et al., 1988; Cochran, 1990), timing of local tectonics (after Veevers, 1988; Barber, 1988; Woods, 1988; Hocking, 1988), dating of major unconformities from well data, characteristic subsidence patterns, and timing of stretching phases inferred from subsidence analysis. Unconformities are grouped in accordance to structural domains: 1 = Exmouth Plateau, 2 = Kangaroo Syncline, 3 = Rankin Trend, 4 = Barrow-Dampier Sub-basin, 5 = Brigadier Trend, 6 = Wombat Plateau.

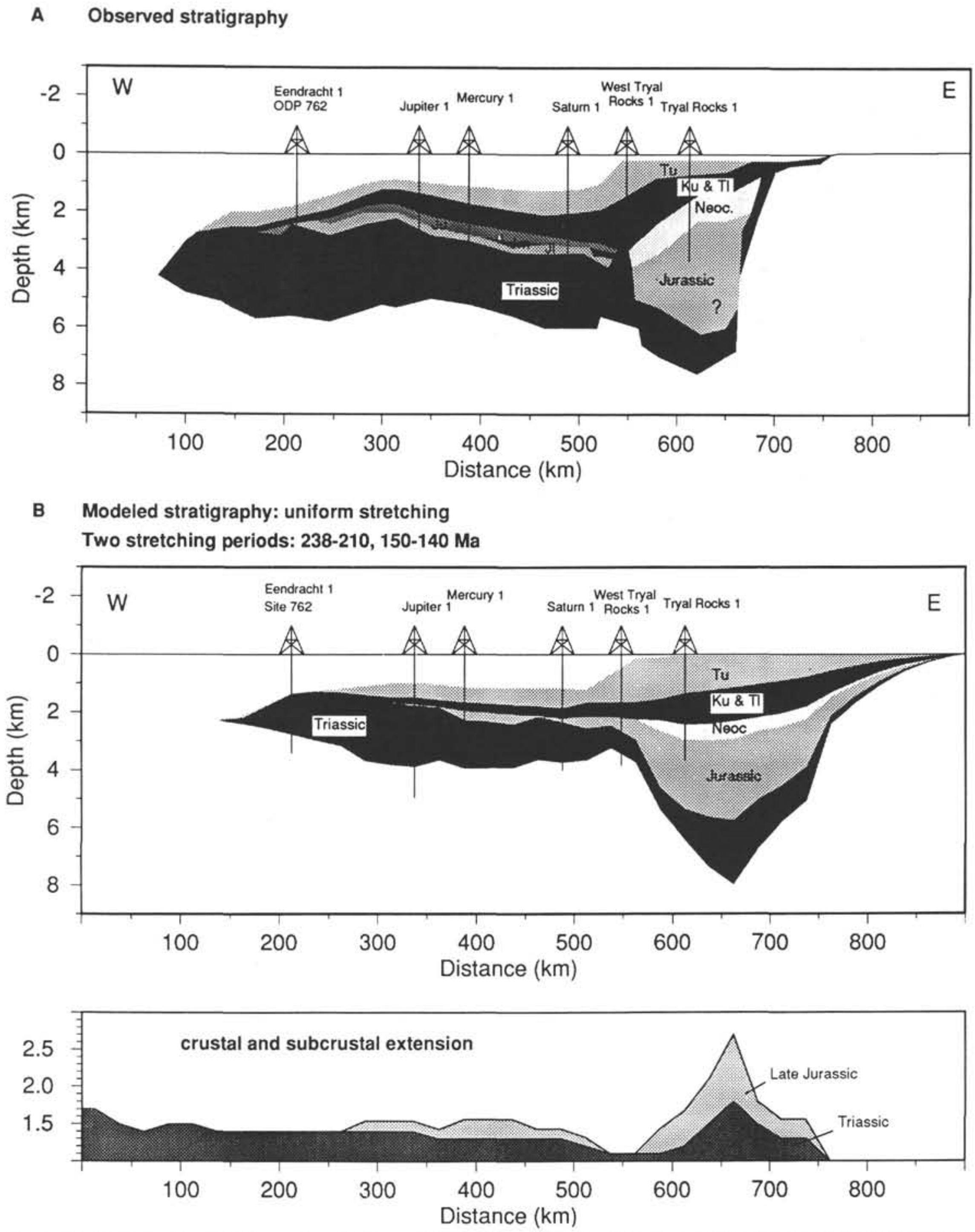
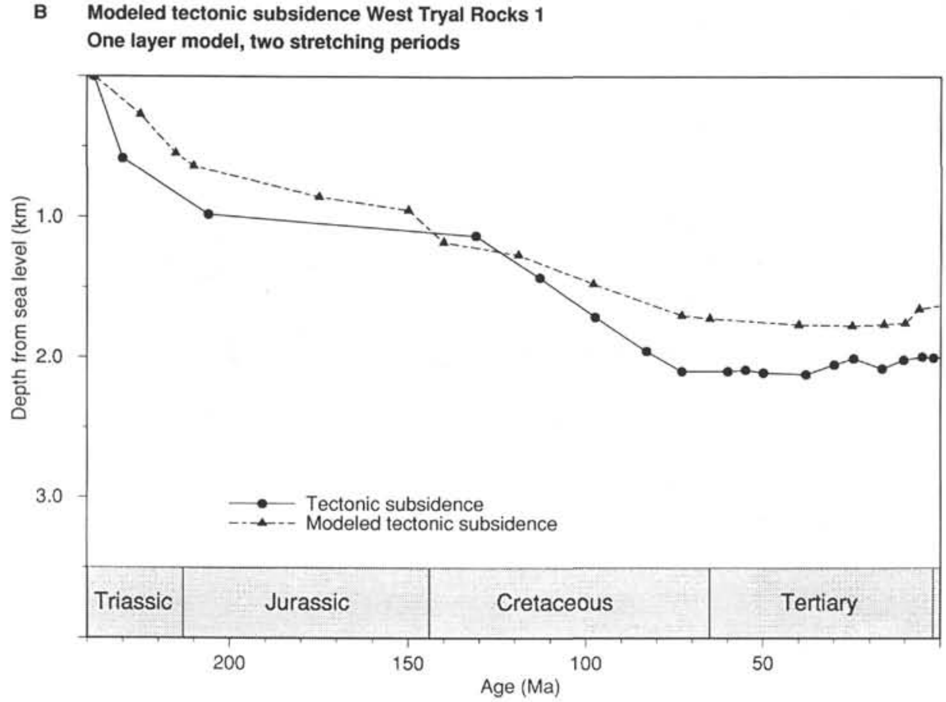


Figure 20. Stratigraphic modeling and synthetic subsidence curves for two wells at the Rankin Trend (West Tryal Rocks 1) and the Kangaroo Syncline (Saturn 1) and comparison with generalized observed stratigraphy and backstripped tectonic subsidence curves for the Northwest Shelf. **A.** Documented stratigraphy. **B.** Modeled stratigraphy, distribution of cumulative stretching factors and comparison of predicted with observed tectonic subsidence for uniform stretching model with two finite stretching phases. **C.** Modeled stratigraphy, distribution of cumulative stretching factors, and comparison of predicted with observed tectonic subsidence for uniform stretching model with three finite stretching phases. **D.** Modeled stratigraphy, distribution of cumulative stretching factors, and comparison of predicted with observed tectonic subsidence for two-layered stretching model with three phases of finite extension. **E.** Modeled stratigraphy, distribution of cumulative stretching factors, and comparison of predicted with observed tectonic subsidence for model given in Figure 20D, incorporating a fluctuating stress field in the analysis. Inset shows a blow-up of the last part of the subsidence curve to illustrate the incorporation of intraplate stresses.



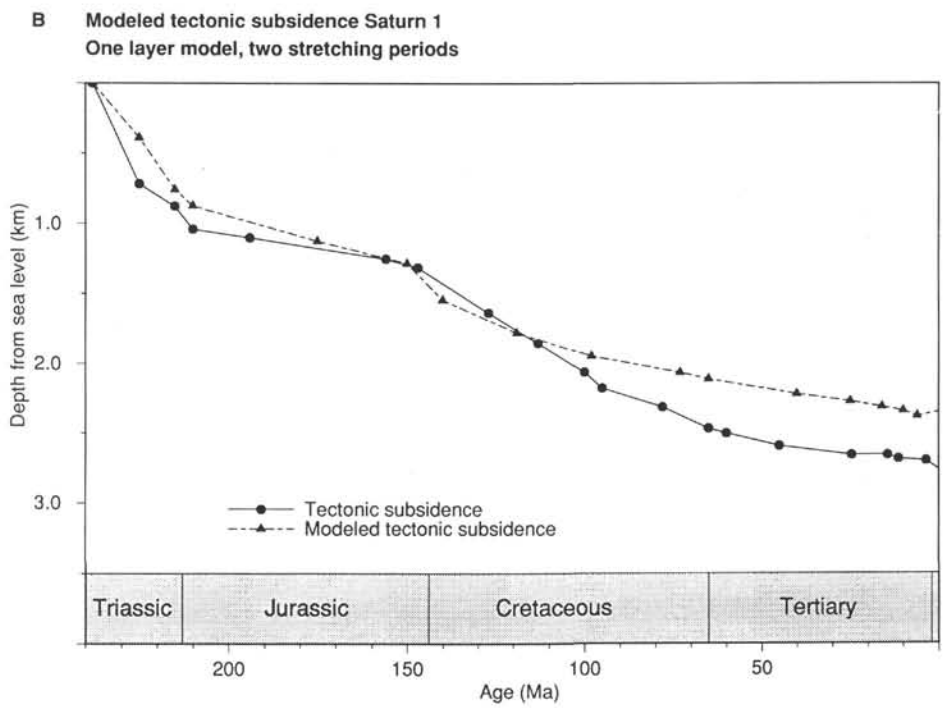


Figure 20 (continued).

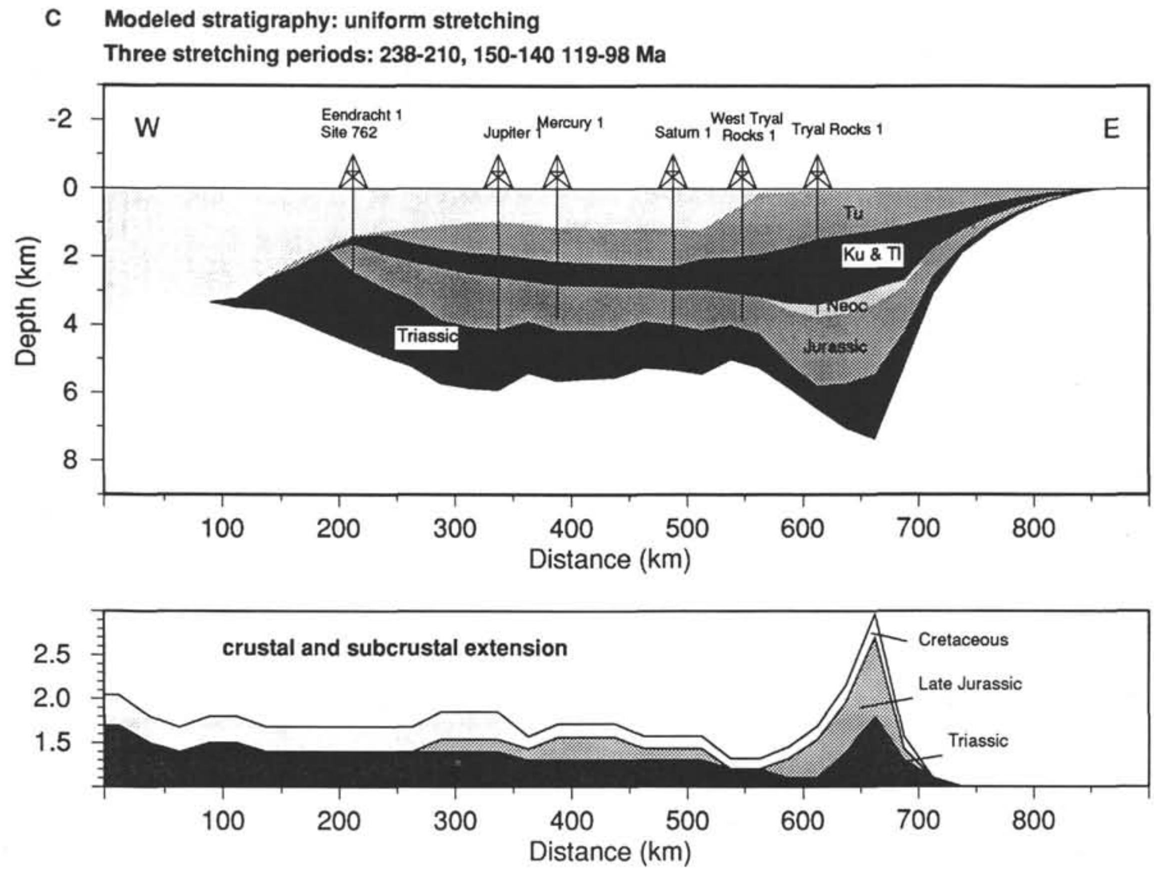
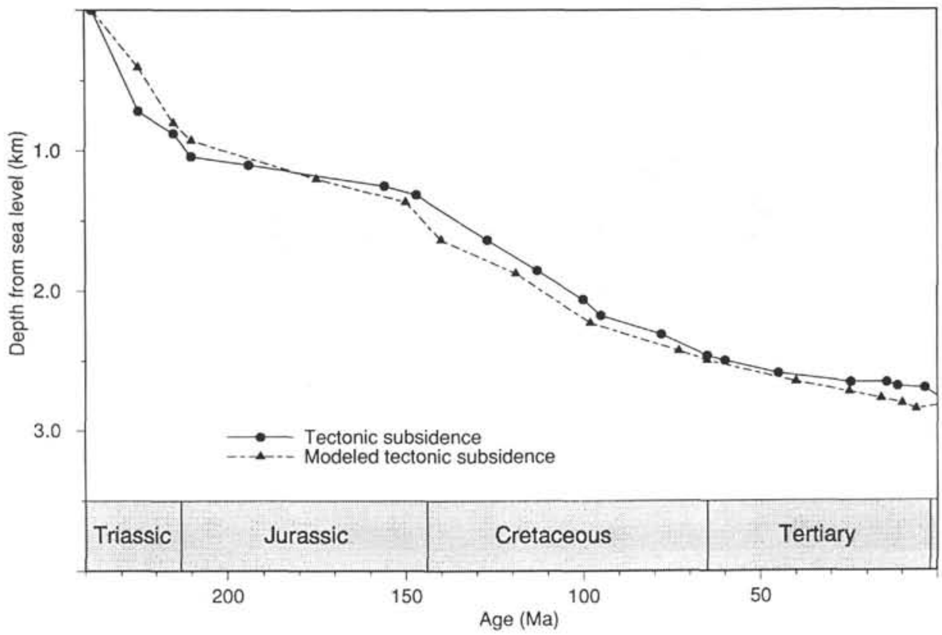


Figure 20 (continued).

C Modeled tectonic subsidence Saturn 1
One layer model, three stretching periods



C Modeled tectonic subsidence West Tryal Rocks 1
One layer model, three stretching periods

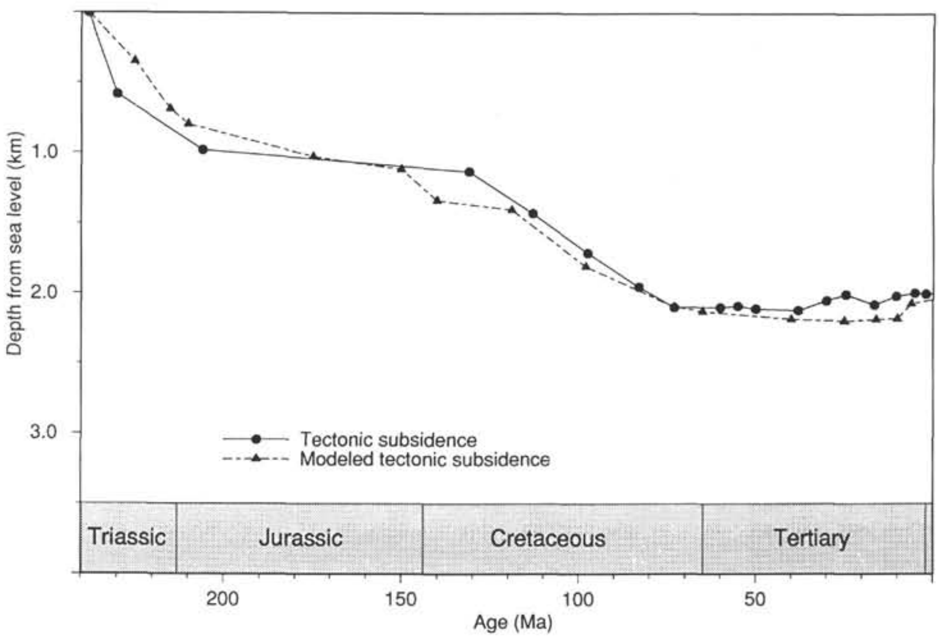


Figure 20 (continued).

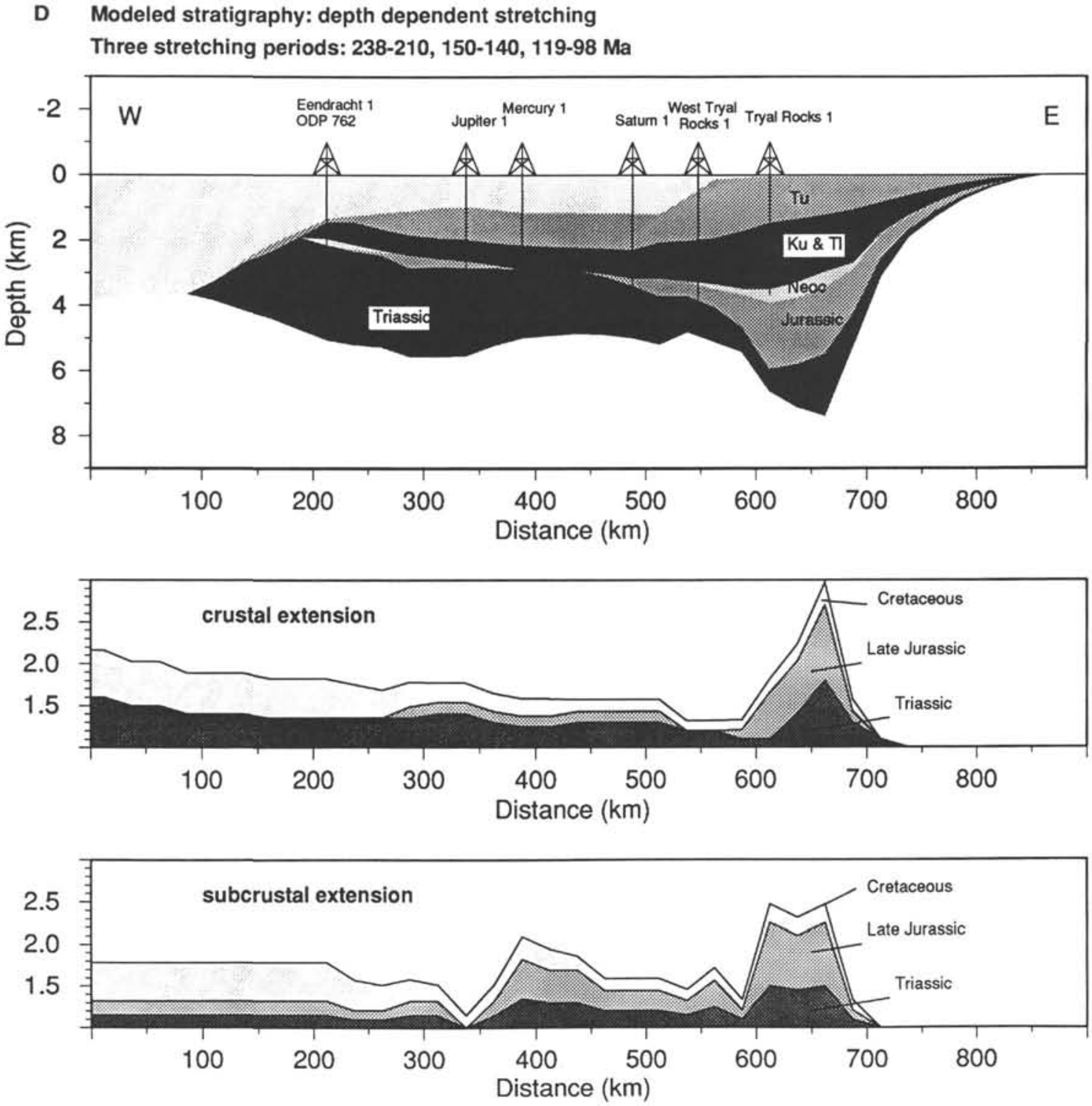
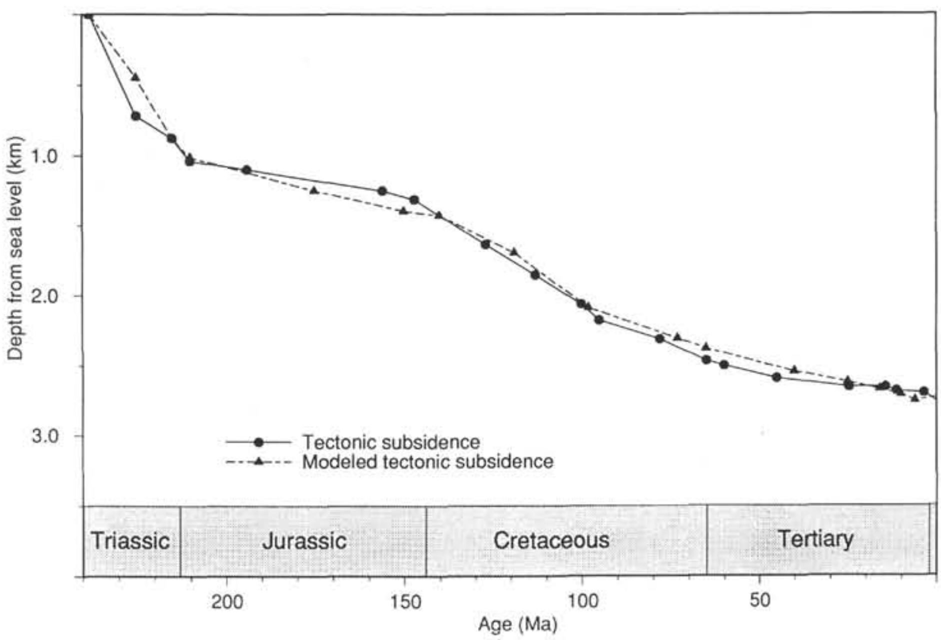


Figure 20 (continued).

D Modeled tectonic subsidence Saturn 1
Two layer model, three stretching periods



D Modeled tectonic subsidence West Tryal Rocks 1
Two layer model, three stretching periods

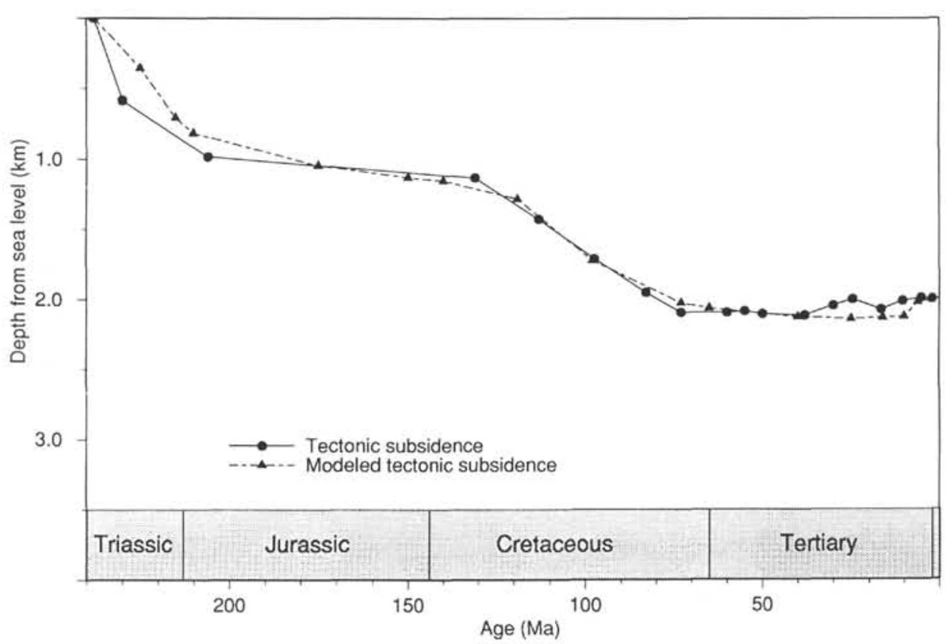


Figure 20 (continued).

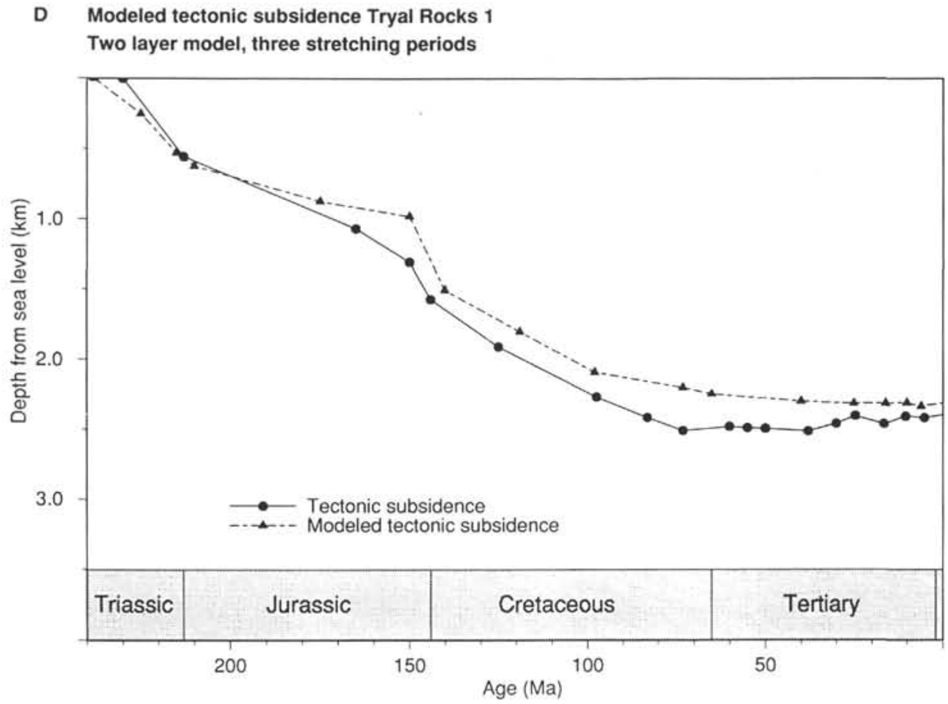


Figure 20 (continued).

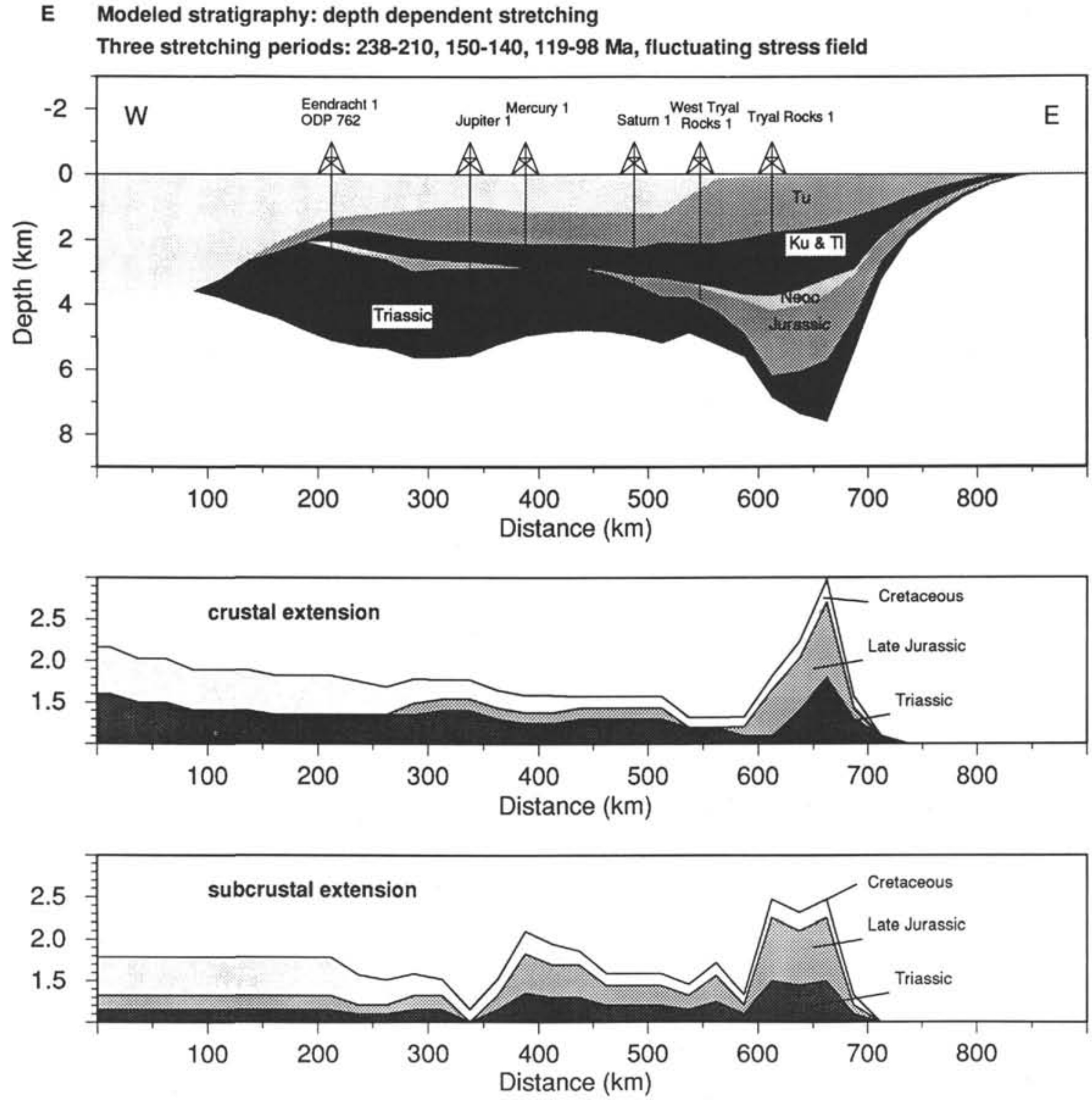
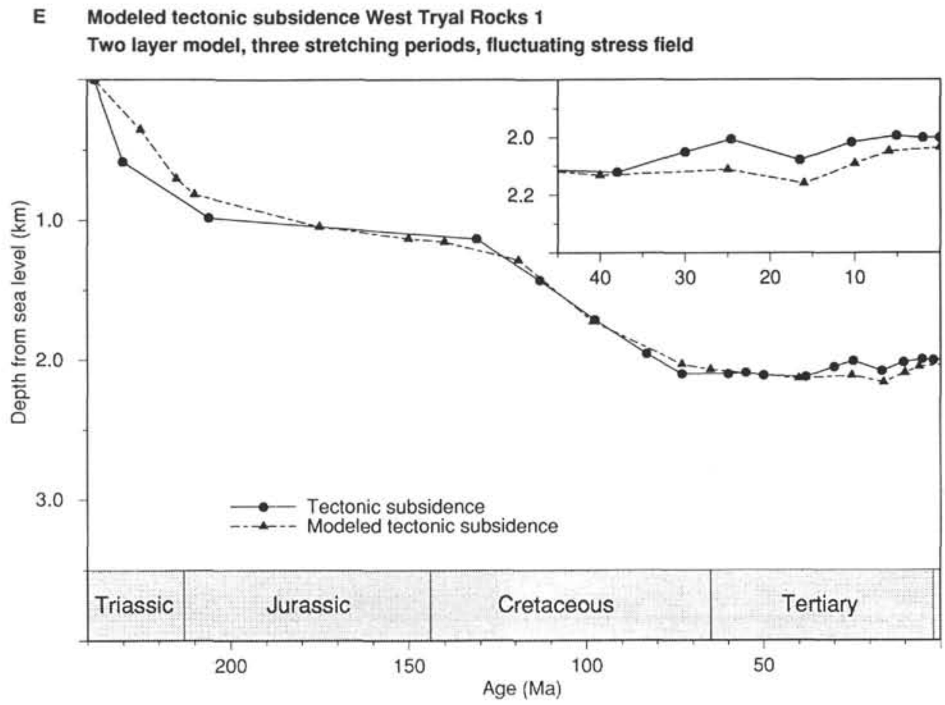


Figure 20 (continued).



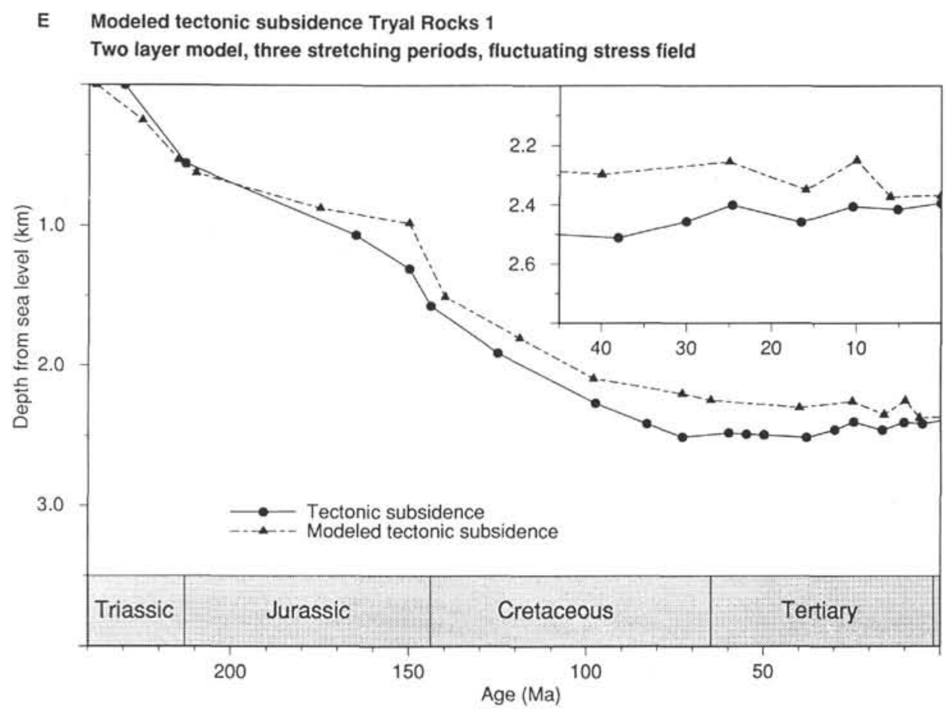


Figure 20 (continued).

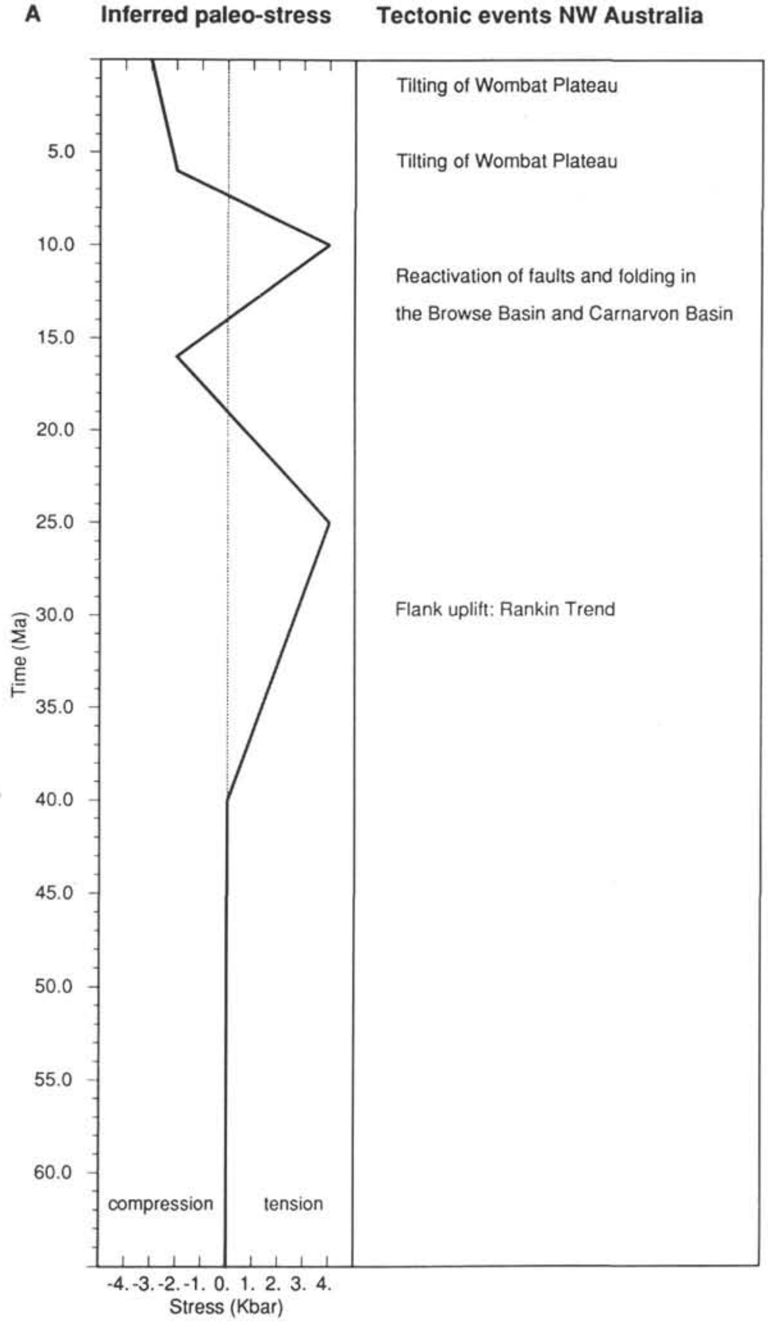


Figure 21. **A.** Paleostress inferred from stratigraphic modeling of the Northwest Shelf of Australia (Fig. 20E) and comparison with timing of local tectonic events (after Bradshaw et al., 1988; Woods, 1988; Zachariasse, 1991). **B.** Timing of tectonic phases documented for the whole Indo-Australian Plate (after Liu et al., 1983; Patriat and Segoufin, 1988; Letouzey et al., 1990; Audley-Charles, 1986; Hegarty et al., 1988; Withjack and Gallagher, 1983; Barber, 1988; Cochran, 1990) and timing of major tectonic events in the Atlantic and Pacific oceans (after Klitgord and Schouten, 1986; Pollitz, 1988). **C.** Different sea-level charts proposed for the region of the Northwest Australian Shelf and timing of major unconformities derived from ODP sites in the Indian Ocean. Curves from left to right include, the "global" sea-level chart of Haq et al. (1987) and the regional curves of relative sea level for western Australia (Quilty, 1980) and the Northwest Shelf (Apthorpe, 1988). Solid circles give the timing of major unconformities in the northeastern Indian Ocean from Leg 116. Solid squares and triangles show timings of major unconformities drilled during Leg 122, at sites located on the Exmouth and Wombat plateaus, respectively. Bars indicate time span of unconformities.

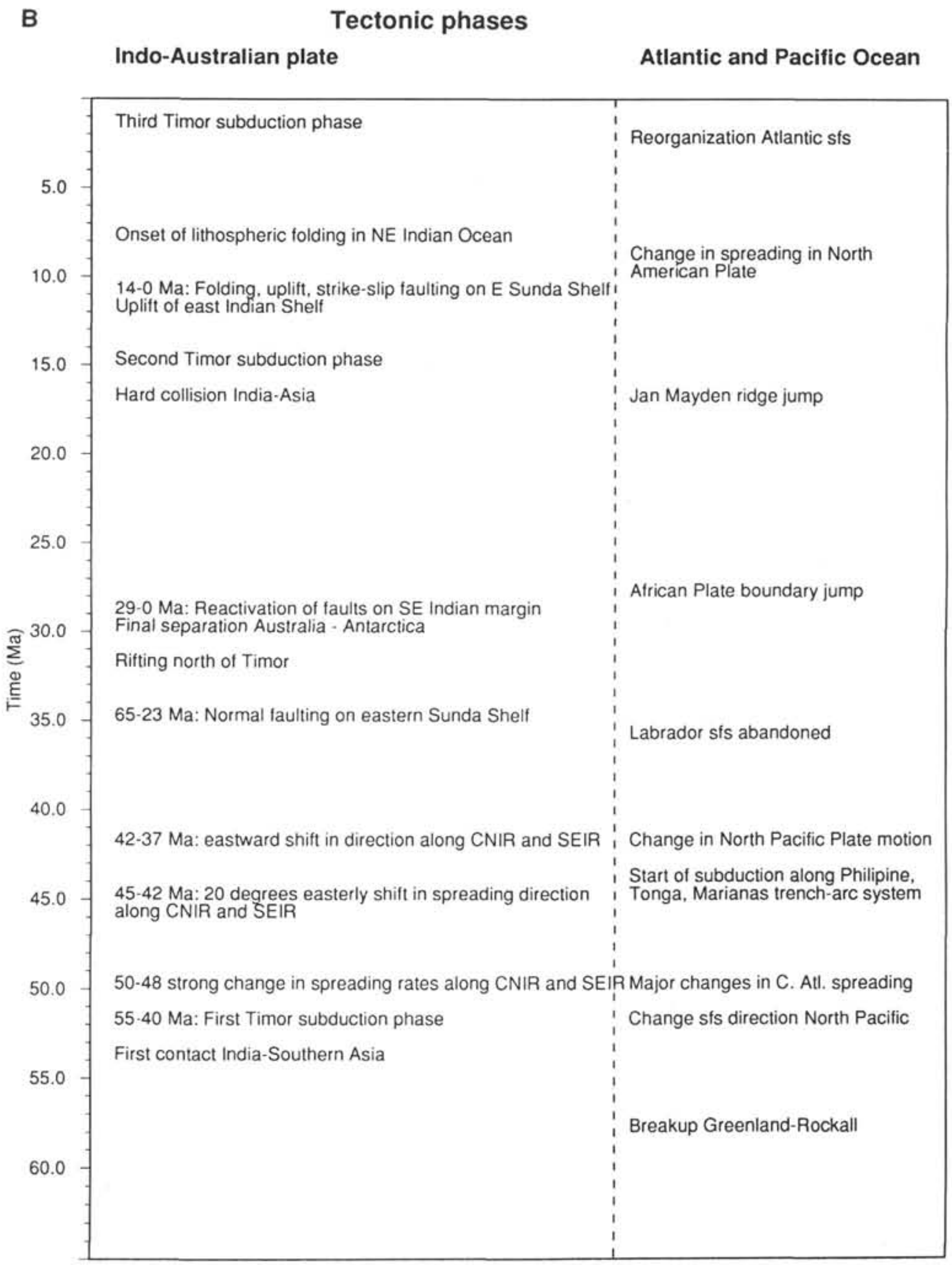


Figure 21 (continued).

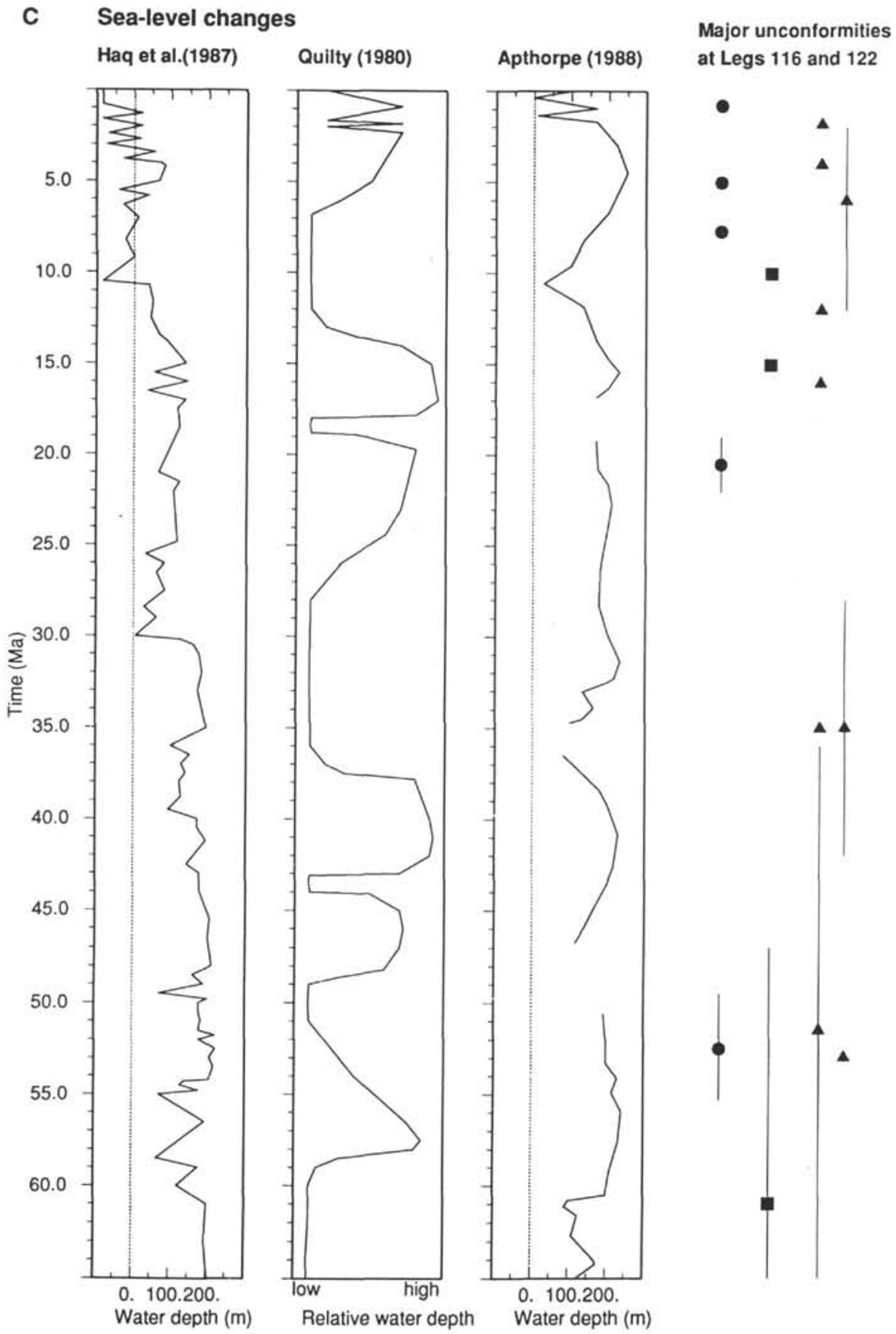


Figure 21 (continued).