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Lithospheric flexure due to prograding sediment loads: implications for the origin of offlap/onlap patterns in sedimentary basins

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ABSTRACT

Simple elastic plate models have been used to determine the stratigraphic patterns that result from prograding sediment loads. The predicted patterns, which include coastal offlap/onlap and downlap in a basinward direction, are generally similar to observations of stratal geometry from Cenozoic sequences of the U.S. Atlantic and Gulf Coast margins. Coastal offlap is a feature of *all* models in which the water depth and elastic thickness of the lithosphere, T_e (which is a measure of the long-term strength of the lithosphere), are held constant, and is caused by a seaward shift in the sediment load and its compensation as progradation proceeds. The coastal offlap pattern is reduced if sediments prograde into a subsiding basin, since subsidence causes an increase in the accommodation space and loading landward of a prograding wedge. The stratal geometry that results is complex, however, and depends on the sediment supply, the amount of subsidence, and T_e. If the sediment supply to a subsiding basin proceeds in distinct 'pulses' (due, say, to different tectonic events in a source region) then it is possible to determine the relationship between stratal geometry and T_e. Coastal offlap and downlap are features of most models where the lithosphere either has a constant T_e, slowly increases T_e with time, or changes T_e laterally; however, in the case where sediments prograde onto lithosphere that rapidly increases T_e with time, the offlap can be replaced by onlap. Lithospheric flexure due to prograding sediment loads is capable of producing a wide variety of stratal geometries and may therefore be an important factor to take into account when evaluating the relative role of tectonics and eustatic sea-level changes in controlling the stratigraphic record.

INTRODUCTION

One success of early attempts to model sedimentary basins was to show that tectonics in the form of thermal contraction and uplift, in combination with sediment loading, could explain the overall 'architecture' of many rift-type basins. A widely used model has been one (McKenzie 1978) in which basin subsidence is caused by a uniform heating, thinning and stretching of the crust and lithosphere during continental rifting. The stretching model explains the change in crustal thickness beneath rift-type basins such as the North Sea (Barton and Wood, 1984) and the symmetry of crustal structure across conjugate passive margins such as Newfoundland and Britain (Keen *et al.*, 1988). Steckler (1981), Watts *et al.* (1982) and Célérier (1986) determined the stratigraphic response to thermal subsidence by assuming that sediments fill a basin to some constant palaeo-bathymetric surface and, following the results of oceanic flexure studies (Watts, 1978), load a basement that increases its elastic thickness, T_e (which is determined by the flexural rigidity of the lithosphere), with age. The geometry that results is the so-called 'steer's head' (Dewey, 1982) which is a basin that increases its width through time and shows onlap at its edges. Many intra-cratonic basins appear to widen with time (Sloss & Schlerer, 1975). Furthermore, coastal onlap is a common observation of the post-rift development of Atlantic-type continental margins such as the East Coast, U.S. (Klitgord *et al.*, 1988), Brazil (Mohriak *et al.*, 1990) and West Africa (Spengler *et al.*, 1966) margins.

The close similarity between the shape of the coastal onlap pattern predicted by the steer's head model to observations lead Watts (1982) to speculate that tectonics is a primary control on the development of the stratigraphic record. This view has been challenged by Vail *et al.* (1977), among others, who have pointed out that the steer's head model is based on thermal and mechanical properties of the lithosphere such as conductive cooling/heating, increase in T_e with age and, stress relaxation which occur on too long a time-scale to be important in controlling the high-frequency of occurrence of unconformities in the stratigraphic record. These workers have appealed instead to rapid changes in eustatic sea-level.

Before ruling out a tectonic control in favour of an eustatic one, however, it is useful to examine the assumptions that were made in the steer's head model, especially with regard to the response of the lithosphere to sediment loads and the mechanisms of basin fill.

The steer's head model is based on oceanic flexure studies which predict that T_e of continental lithosphere would increase with age as it cools following a rifting event. Barton & Wood (1984), Watts (1988) and Fowler & Mckenzie (1989) have argued, however, that unlike oceanic lithosphere, T_e of rifted continental lithosphere, is small and may not increase with age. The reason for this is not presently understood but, it may be related to (a) the low strength of the continental lithosphere when it is subjected to extension, (b) the weakness of continental lithosphere in thermally activated regions due to the dependence of strength on the homologous temperature and the low melting temperature of granite (e.g. Fowler & McKenzie, 1989) or (c) the susceptibility of continental lithosphere to the effects of sediment blanketing, which in areas of rapidly deposited sediments may raise the temperature sufficiently to weaken it. Many basins are underlain by extended continental lithosphere, so, if it has little or no strength then, the results of these studies call into question whether flexure can account for coastal onlap.

It has been shown recently for the North Sea basin (White & McKenzie, 1988) and East Coast, U.S. margin (Watts, 1988) that a thermal model in which different amounts of extension occur in the crust and mantle can produce coastal onlap in rift-type basins—even if T_e is small or does not increase with age. If, for example, the mantle thinning extends over a broader area than the thinning of the crust then the basin edge will be emergent during the early post-rift development thereby limiting sediments to the basin depocentre. Eventually, as the mantle cools, the emergent region will subside, sedimentation will extent to the basin flanks and individual stratigraphic units will overstep or onlap the basement.

More difficult to explain by tectonics are basinward shifts in the pattern of onlap or offlap. Some workers have argued (e.g. Sleep, 1971; Sleep & Snell, 1976; Nunn & Sleep, 1984; Quinlan & Beaumont, 1984) that offlap is caused by stress relaxation in the lithosphere due to viscous effects. This model, however, is not supported by oceanic flexure studies which show that T_e depends on plate age rather than load age. The oceanic lithosphere does exhibit some relaxation on loading as it thins from its initial (seismic?) thickness to its long-term elastic thickness but, studies of seamounts and oceanic islands (e.g. Bodine *et al.*, 1981) suggest that this is probably complete within 1 to 10 Myr of loading. Watts *et al.* (1982) and Watts & Thorne (1984) have shown that offlap patterns in the East Coast, US margin and the Michigan basin can be explained by an elastic model with a T_e that increases with age—provided it was subject at some time during its evolution to either a) a widespread erosional 'event' or b) a long-term eustatic sea-level fall.

A more serious limitation of the steer's head model is the assumption that sediments fill in a depression in a 'layer cake' fashion up to some constant surface. Such a model of aggradation precludes other mechanisms of sediment fill, most notably progradation. Ager (1973), for example, follows the view of many 'sedimentologists' in arguing that sediments, rather than aggrading, in general build out laterally as a series of prograding wedges. Perhaps the best observational evidence for progradation is shown in the growth patterns of large river delta systems such as the Mississippi (e.g. Galloway, 1989) and Niger (e.g. Avbovbo, 1978) deltas where the sediments have built out several hundreds of km beyond the initial depositional shelf-break in several tens of Myr. Deltas are not the only systems to show progradation. Carbonate ramps and platforms (Sarg, 1988; Brooks & Holmes, 1988) also show progradation, as do continental shelves and slopes, especially during a sea-level rise.

Barrell (1914) was among the first to recognize the importance of river delta systems in better understanding the long-term ($>10^6$ years) mechanical properties of the Earth's lithosphere. The lateral transport of sediments by progradation is an example of a migrating load on the surface of the lithosphere which should respond by some form of flexure. Walcott (1972) and Cochran (1973) estimated the total amount of sediment added to the continental margin by the Mississippi and Amazon rivers and used it to compute the depression of the basement and gravity anomaly for different values of the effective elastic thickness of the lithosphere, T_e.

These early studies were useful in that they demonstrated that flexural loading could account for the thickness and overall 'architecture' of present-day river delta systems. They did not consider, however, the stratigraphy as the cumulative response to individual phases of the growth of a delta system. One difficulty with doing this is that, with the possible exception of the Mississippi and Niger deltas, the geometry of their growth is generally too poorly known.

Kenyon & Turcotte (1985), adopting a different approach, used a diffusion model to describe the evolution of river delta systems through time. They showed that such a model could explain (a) the exponential shape of depositional slope fronts, and (b) the rate of progradation in modern river delta systems such as the Mississippi. These workers did not evaluate, however, the flexural effects of adding each new load or its stratigraphic consequences.

A stratigraphic model that includes diffusion as well as flexural loading has recently been developed by Flemings & Jordan (1989) and Sinclair *et al.* (in press) and applied to the flexural sag basins that form in front of migrating thrust/fold loads. These workers showed that as time evolves, diffusion progressively reduces the slope of a thrust front such that the area of the crust affected by sedimentation broadens with time and downlap develops in the lower part of the prograding wedge. Flemings and

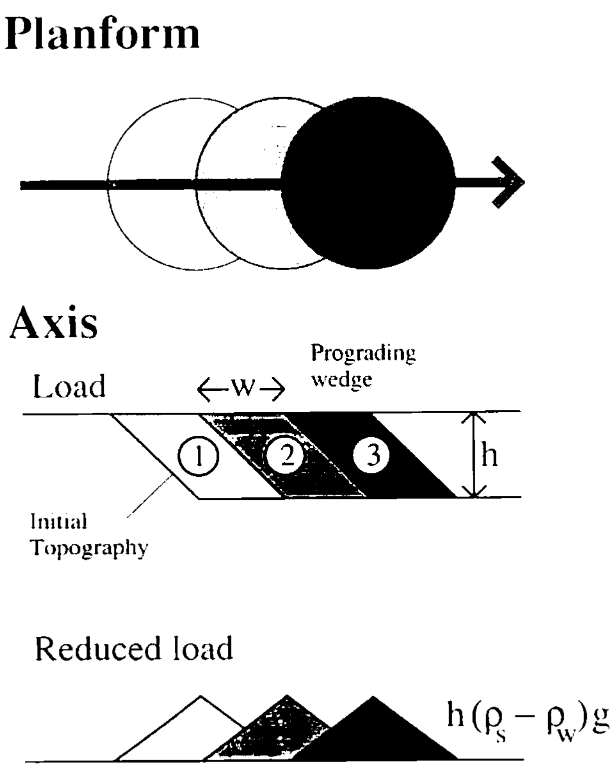

Fig. 1 Schematic diagram showing a prograding sediment wedge system. Each wedge is assumed to be lobate in planform. The prograding wedge has a slope that is parallel to the initial water-filled basin and is assumed to extend laterally as far as the base of the initial (or previous) depositional surface. w = the width of each prograding wedge, h = depth of water that is available, ρ_s = density of sediment, ρ_w = density of water and g = average gravity. In the case illustrated, the wedge loads reduce to a series of cone-shaped loads of peak offset, w.

Jordan (1989) and Sinclair *et al.* (in press) also noted that, depending on their rate of supply, sediments may also onlap the thrust wedge itself.

The purpose of this paper is to show that lithospheric flexure due to prograding sediment loads can have a profound effect on the overall 'architecture' of sedimentary basins and, in some cases, the details of their stratigraphic patterns. Coastal offlap and downlap are the most common stratigraphic features predicted by flexural loading models and do not require either (1) an increase in T_e with age, (2) post-depositional erosional 'events', or (3) eustatic sea-level changes to produce them. More difficult to explain by flexure is coastal onlap in a prograding system but, even this can be accounted for *if* sediment loading is accompanied by an increase in accommodation space in coastal areas due either to rapid increases in the (a) flexural response or (b) tectonic subsidence.

FLEXURE

The stratigraphic effects of prograding sediment loads have been computed in this study using a simple elastic plate model, similar to one used by previous workers to model the flexure at oceanic islands and seamounts, deep-sea trenches and outer rises. The main difference is that the plate is subject to multiple loads as they move across it with time. The resulting stratigraphy is obtained by summing the contributions of individual loads and 'updating' the depths to the pre-deformation surface, the load, and the flexure profile.

A general difficulty in stratigraphic modelling is being able to specify the shape of the sediment loads through time. The load could be infered from knowledge of the palaeo-bathymetry but, this requires detailed bio-stratigraphic data from deep wells. Another approach is to use a diffusion model since this model allows the amount of material that is removed by erosion from a topographic profile and the amount that is added to it by sedimentation to be calculated through time. Unfortunately, this model only applies to homogeneous hillslopes and may not be applicable to regional topography characterized by lithological variability (Culling, 1965).

In this study, the sediment loads have been divided into two parts: one a 'driving' load which causes the flexure and one an 'infill' load which amplifies it. The driving load is the sediment that accumulates between the initial water-filled topography (which is assumed to have been produced by some means prior to progradation and, for the purposes of the initial calculations, not to change its shape with time) and the new surface of deposition of the prograding wedge. By way of contrast, the infill load is the sediment that fills in the flexure up to the initial topography. Both loads act as an excess mass on the plate which responds by an amount that depends on its intrinsic strength.

The geometry of the driving loads is shown in Fig. 1. The loads step out into a water-filled basin at a constant rate and are assumed to move in the same direction such that any 'switching' from side to side has little or no affect on the resulting stratigraphy. The loads are lobate in platform (Fig. 1) and are assumed to be of uniform density and to have circular symmetry. In cross-section (Fig. 1), each load is assumed to (a) have a depositional slope that is parallel to the initial water-filled basin, (b) offset the previous wedge by a distance w, and (c) infill the basin to a depth h.

Figure 2 shows the results of a simple calculation based on three lobate loads, each offset by a constant distance. The flexure was computed using a Fast Fourier Transform technique (e.g. Ribe, 1982) and values of T_e of 10 km, a load and infill density of 2600 kg m^{-3}, and a load height, h, of 1000 m. Each profile in Fig. 2 represents a cross-section through the centre of the lobes. The upper profile shows the flexure that was obtained for a single lobe load. The deformation profile has a characteristic flexural form with a broad area of subsidence out to the first node (the point of no subsidence or uplift), r, and beyond that a peripheral bulge. The nodal distance, r, is determined mainly by T_e and, to a lesser extent, the geometry of the load. For the parameters assumed in Fig. 2, r = 69 km. The middle profile shows the modifying effects on the flexure,

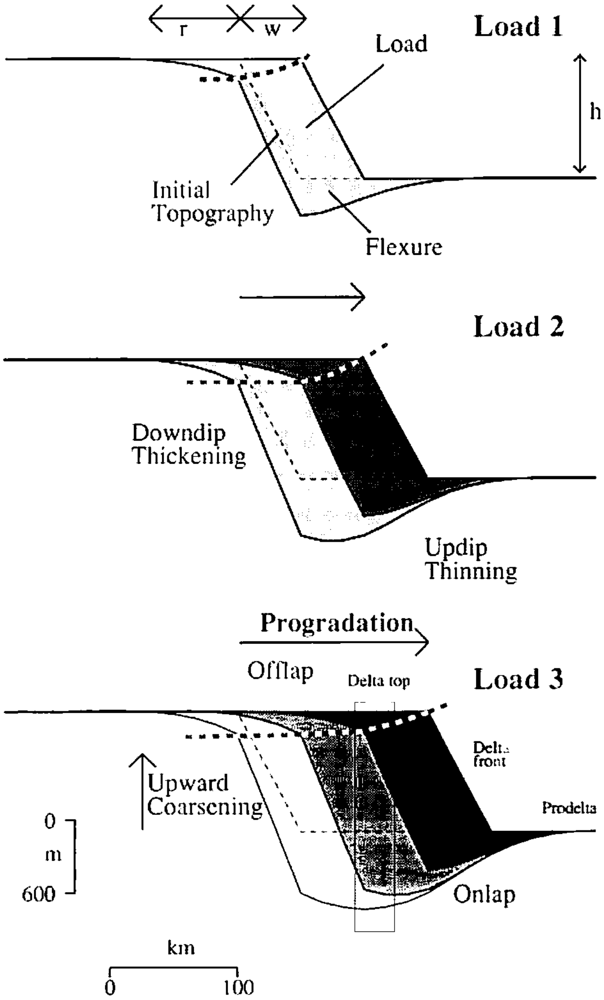

Fig. 2 Simple model for the progressive flexure of the lithosphere due to prograding sediment loads. The upper profile shows the initial load and the resulting flexure. $w = 10$ km, $h = 750$ m, $\rho_w = 1030$ kg m^{-3}, $\rho_s = 2600$ kg m^{-3}, and $T_e = 10$ km. The middle profile shows the effect of adding a second load. The initial load and flexure are deflected down by the new load in regions flanking the new load and uplifted in peripheral regions. The net effect is to cause an apparent downdip thickening and updip thinning of the infill material, coastal offlap and downlap in a basinward direction. The lower profile shows the effect of adding a third and final load. The Delta top, Delta front and Pro-delta refer to the type of sediments that would be expected to be deposited within the prograding wedge. The heavy dashed line traces the boundary between Delta top and Delta front sediments and therefore represents a facies boundary. A similar facies boundary (not drawn) would exist between the Delta front and Pro-delta sediments.

the loads, and the initial topography of adding a second lobe at an offset distance, $w = 40$ km. This offset distance is less than the nodal distance so the first load is now in the flexural depression of the new load causing it to subside. The subsidence results in an apparent up-dip thinning and down-dip thickening to the original load and infill material of the old wedge, a tilting of its once flat upper and lower surfaces, and an increase in steepness of its original slope. The lower profile shows the effect of adding a third wedge load at an offset distance $w = 80$ km from the first load. This offset distance exceeds the nodal distance so that the first load is now on the flexural bulge of the new load and so is gently uplifted. The second load is, however, within the nodal distance and so is subject to subsidence. The cumulative stratigraphy shows a characteristic pattern of downlap in the lower part of the wedge and a basinward shift in onlap or offlap in the upper part. The downlap occurs in the direction of progradation, but the offlap develops in an opposing direction because of a progressive basinward shift of the centre of gravity of the loads and their resulting flexure.

The model in Fig. 2 assumes the same T_e, load geometry and, water-filled 'edge' topography profile for each wedge load so the lateral extent of the offlap and downlap patterns (as, say, measured along a horizontal surface) are the same. The solid lines trace the movement of the depositional front of each new load and therefore can be considered to be chronostratigraphic horizons. Each depositional front is rotated by flexure to form a single horizon, sigmoidal in shape, which is curved in two directions. The sigmoidal flexures resemble the clinoformal structure that are commonly observed on seismic reflection profiles of prograding delta systems. The edge of the depositional front migrates at a constant rate and so the supply of sediments to the basin (i.e. the sum of the load and the flexure) will also be constant. The dash line (Fig. 2) tracks the position of the former shelf-break which separates shallow water deposits (e.g. sands) of the delta top from delta front sediments (e.g. muds) and can therefore be considered as a facies boundary. Figure 2 shows that the facies boundary cut the time lines at high angles, which reflects the gentle tilting of the basin margin due to flexure.

In constructing the stratigraphy in Fig. 2, it was assumed that before a new load is added, the flexural bulges are reduced by erosion to the initial water-filled surface and that there is no subsequent isostatic uplift of the crust and lithosphere. The bulges do not exceed a few tens of meters, however. Also, the uplift due to erosion will be limited by the strength of the crust and lithosphere. For these reasons, it is not thought that the effects of erosional rebound are likely to significantly modify the resulting stratigraphic patterns.

APPLICATION OF A PROGRESSIVE FLEXURE MODEL TO THE U.S. ATLANTIC MARGIN

The predictions of the progressive flexure model have been compared to two cases of Cenozoic sediment progradation: one of relatively small-scale at the East Coast, U.S. and the other of large-scale at the Texas Gulf Coast. These passive continental margins were formed by rifting and crustal extension during the Early to Middle Jurassic, and so in this case the Cenozoic sequences prograded out on to lithosphere that was subsiding by only a small amount. The

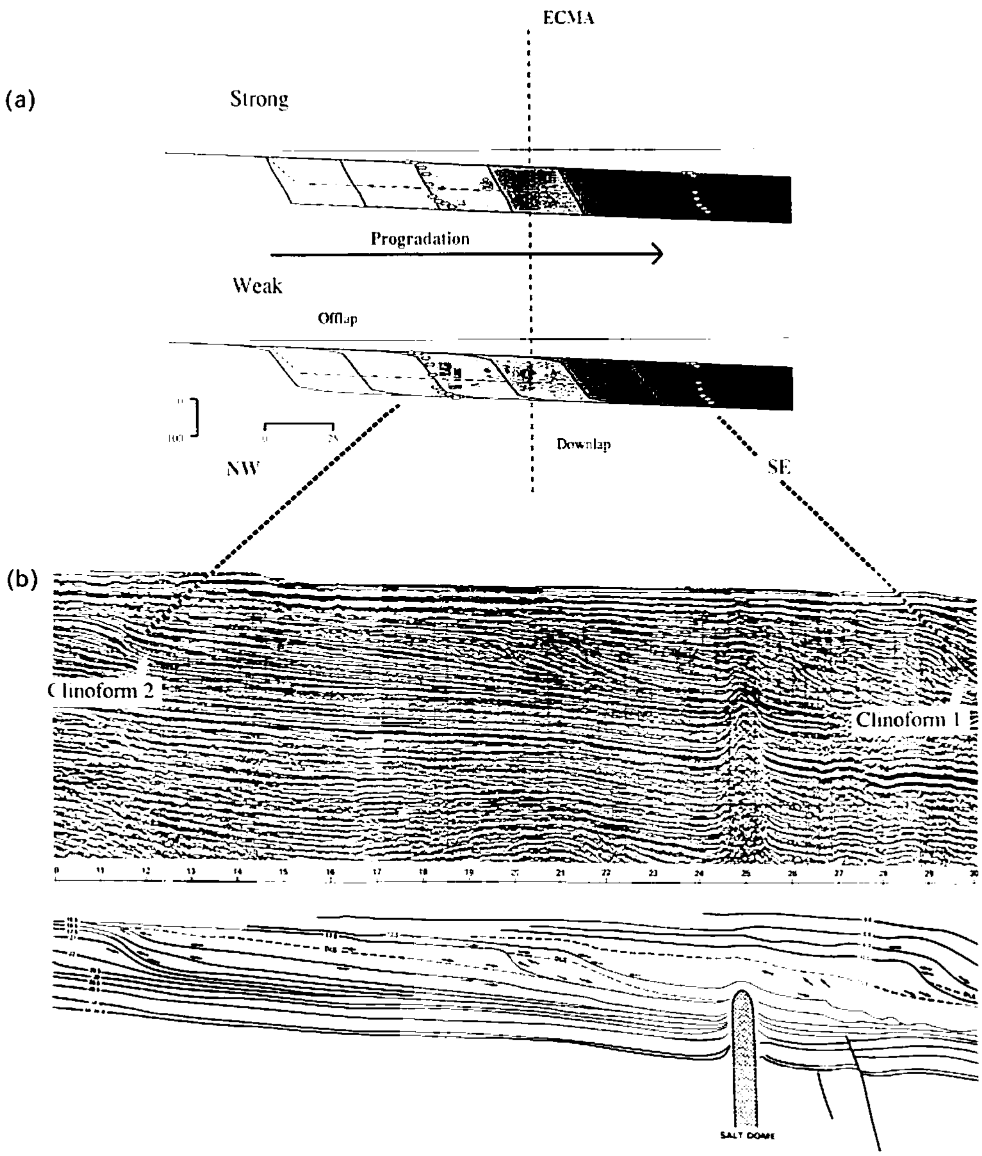

Fig. 3 Comparison of the calculated stratigraphy due to prograding loads model to clinoform structures (Greenlee and Moore, 1988) observed an seismic reflection profiles off the East Coast, U.S. (a) Calculated flexure for $w = 10$ km, $h = 400$ m and $T_e = 50$ km (Strong) and $w = 10$ km, $h = 250$ m and $T_e = 2$ km (Weak). Small open ellipses are digitized clinoforms 1 and 2. (b) Observed seismic reflection profile and line drawing showing clinoform structures. ECMA = East Coast Magnetic Anomaly which most authors consider to represent the boundary separating highly magnetized oceanic crust (oceanward side) from weakly magnetized stretched continental crust (landward). There is reasonably good agreement between the shape of the calculated and observed clinoforms.

East Coast, U.S. and Texas Gulf Coast margins have the additional advantage that they are among the most extensively studied and well-known in the world.

(a) East Coast, U.S.

Seismic reflection profile data off New Jersey (e.g. Greenlee & Moore, 1988; Fig. 3) suggest that during the Miocene a series of clastic sequences prograded out about 100 km across a large part of the shelf off the East Coast, U.S. The Miocene at this margin is divided into a number of sequences which on seismic reflection profiles are characterized by a series of stacked, 10 km wide, clinoformal structures. Each sequence shows a pattern of downlap in a basinward direction and coastal onlap and offlap in landward areas. Coastal onlap dominates the lower and upper Miocene sequences. The middle Miocene, however, shows a pronounced downward shift in the pattern of onlap (i.e. offlap) which according to Greenlee & Moore (1988) is the result of erosion of an otherwise onlapping sequence during a major eustatic sea-level fall.

The flexural effects of a series of prograding sediment loads of similar geometry to the Miocene wedges of the East Coast, U.S. is illustrated in Fig. 3 for values of $T_e = 2$ km (weak) and $T_e = 50$ km (strong). The width of each sediment wedge, w, is based on the clinoformal structures that have been identified on seismic reflection profiles and the range of water depth, h, is consistent with the results from palaeo-bathymetric studies of samples from deep stratigraphic test wells in the outer shelf. As was the case in Fig. 2, sigmoidal flexures develop which have a curvature that is a function of T_e—being small for a strong underlying lithosphere and large for a weak one. Figure 3 shows that the predicted flexure is similar in form to the observed clinoforms suggesting that these structures are also the result of loading due to sediment progradation. The inner shelf clinoforms seem to be more strongly curved than those of the outer shelf and it is speculated that this is because the lithosphere is weaker landward of the East Coast Magnetic anomaly than seaward of it, as has been suggested recently by Watts (1988) on the basis of gravity anomaly studies.

(b) Texas Gulf Coast

Perhaps the best known case of sediment progradation along the US margin occurs in the Texas Gulf Coast region. The stratigraphic sequence (e.g. Buffler & Sawyer, 1988) consists of thick sequences of Tertiary–Recent clastic sediments deposited in offlapping wedges over mainly carbonate beds of Cretaceous age. The Cretaceous

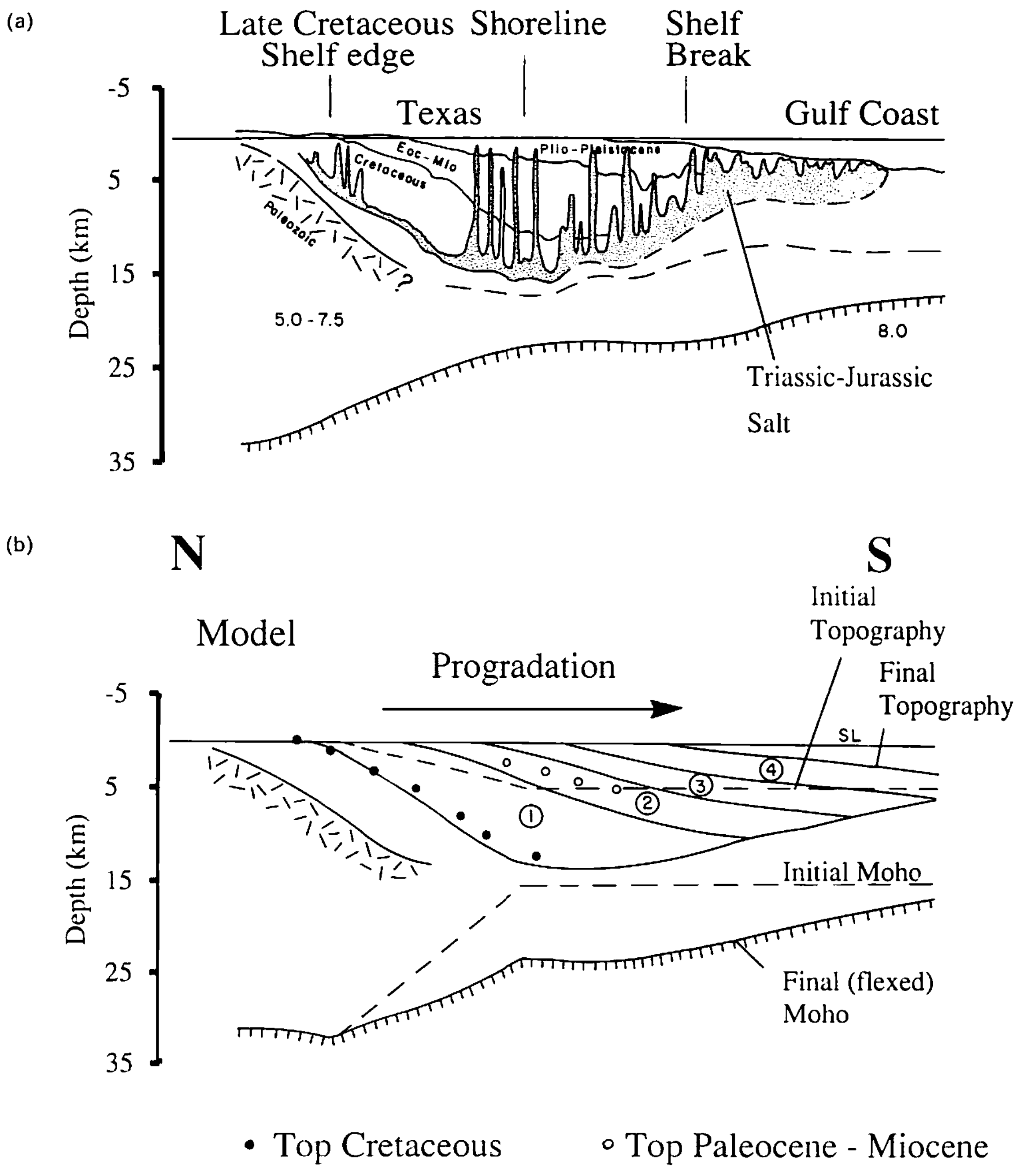

Fig. 4 Comparison of the calculated stratigraphy and crustal structure due to prograding loads (b) to a geological cross-section of the Texas Gulf Coast region based on the seismic reflection and refraction data of Antoine *et al.* (1974) (a).

section is underlain by paralic sediments, redbeds and evaporites of Pennsylvanian–Jurassic age.

In early Mesozoic time, the northern rim of the Gulf was characterized by shallow evaporite basins and a slow rate of subsidence. After the culmination of the Laramide orogeny in the Late Cretaceous/Early Tertiary, increased subsidence in the Gulf and the rapid influx of clastic materials from the north caused the depositional shelf-break to migrate seaward by as much as 400 km.

Seismic refraction and reflection studies (e.g. Antoine *et al.*, 1974) indicate that the Gulf Coast clastic wedge is underlain by crust which is about a factor of 3 thinner than normal thickness continental crust. The origin of the thinned crust is not clear but it probably formed (e.g. Buffler & Sawyer, 1988) by stretching as Africa/South America began separating during the Early Jurassic. Assuming a crustal thinning factor of 3, then McKenzie's stretching model predicts that by Early Tertiary time, a water depth of up to 4800 m would have been available for the prograding clastic wedges.

Figure 4 compares a crustal section of the Texas Gulf Coast to predictions of a simple model in which a lobate sediment body progrades out into a water-filled basin. A value of the wedge width (ϖ) of 270 m was assumed, based on the stratal geometry on geological transects of the margin (e.g. Antoine *et al.*, 1974). If the initial depth of the water-filled basin is assumed to be 4800 m then the crustal structure prior to progradation (dashed lines, Fig. 4b) can be computed by isostatically balancing a column of stretched crust with a mid-oceanic ridge column. Figure 4b shows that the main effect of progradation is to depress the initial crustal structure such that at the shoreline the top of the Lower Cretaceous and the Moho are up to 8 km deeper than they were originally.

There is good overall agreement between the computed and observed crustal structure of the Gulf Coast (Fig. 4). The model explains the coastal offlap patterns that outcrop in Texas, the occurrence of downlap on deep seismic reflection profiles of the Texas/Louisiana shelf and, the overall configuration of the top of the Lower Cretaceous and Moho as revealed by seismic refraction data. The offlap and downlap patterns are a feature of all constant T_e prograding models. The configuration of the top of the Lower Cretaceous is dependent, however, on the density of the sediment load and infill and the value of T_e assumed. In Fig. 4b a T_e of 45 km and sediment densities that systematically decreased from 2700 kg m^{-3} for the first wedge to 2200 kg m^{-3} for the last load were assumed. A basinward decrease in density is in accord with the results of seismic refraction studies (Antoine *et al.*, 1974) which show that the *P*-wave velocity of the upper sedimentary layer increases from 2.4 to 4.6 km s^{-1} in Texas to 1.8–2.2 km s^{-1} beneath the shelf.

SUBSIDING MARGINS

In the previous section it was argued that progressive lithospheric flexure due to prograding sediment loads can account for coastal offlap and downlap patterns in stratigraphic sequences at the East Coast, U.S. and Texas Gulf Coast margins. It is not necessary in the flexural models to appeal to other factors such as an eustatic sea-level fall and the associated 'switching' of sediments from the aggradational to progradational facies (e.g. Jervey, 1988) to explain these patterns. As pointed out previously, the East Coast, U.S. and Texas Gulf Coast margins are cases where progradation occurred relatively late in basin evolution when any change in subsidence due, for example, to thermal cooling was small. There are cases in the stratigraphic record, however, where sediments have prograded at relatively slow rates compared to those of the thermally driven subsidence (e.g. at young margins) and so the effects of subsidence on the stratal geometry also need to be taken into account.

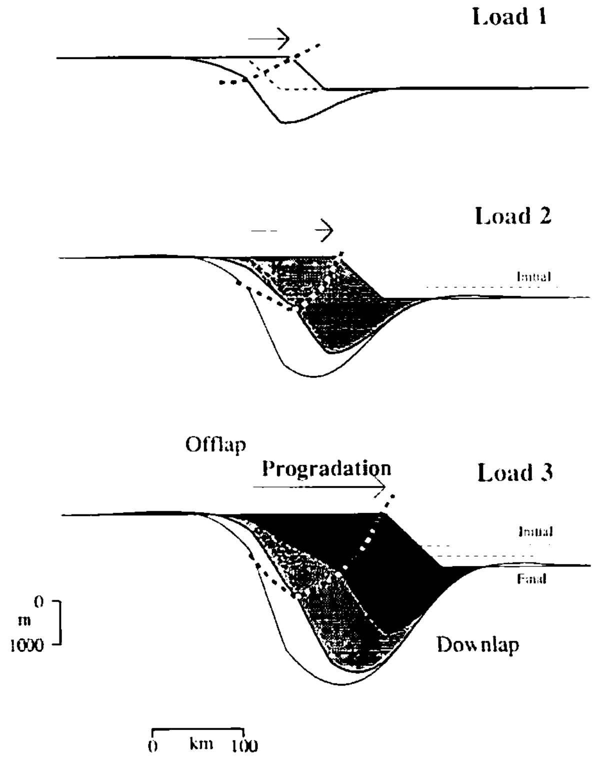

Fig. 5 Simple model for the progressive flexure of the lithosphere due to prograding sediment loads on a subsiding margin. The subsidence is assumed to increase linearly with time about a fixed hinge line. The initial subsidence (upper dashed line) is 750 m and the final subsidence (lower dashed line) is 1250 m. The stratigraphy in the subsiding margin case differs from that of the non-subsiding margin case in having (a) an increase in cross-sectional area of each stratigraphic unit (i.e. load + flexure) with time, (b) the facies boundary between top and slope delta sediments track more deeply in the wedge and (c) the pattern of coastal offlap is greatly reduced in lateral extent.

Figure 5 shows the synthetic stratigraphy for three prograding sediment wedges which build out from the edge of a subsiding basin. The model assumes a lithosphere of constant flexural strength and a tectonic subsidence that increases linearly from a fixed 'hinge line' located at the edge of the initial water-filled basin. After three loads have been added, the depositional sequence forms a characteristic pattern of coastal offlap and downlap in a basinward direction.

The predicted stratigraphic patterns in Fig. 5 are similar to the non-subsiding margin case (Fig. 2) but differ in that the lateral extent of coastal offlap is reduced, the shelf-break (dashed lines) tracks more deeply in the wedge, and the chronostratigraphic horizons (solid lines) progressively dip more steeply than the depositional front. These differences are largely the result of a progressive increase in accommodation space landward of the prograding wedge due to tectonics. Sediment loading behind the prograding wedge will drive the node landward thus competing with (but may not overcome) the effects of progradation. Figure 5 shows that the cross-sectional area between the time lines (solid lines) increases as progradation proceeds, implying an increase in the sediment supply through time.

Several studies (e.g. Sleep, 1971; McKenzie, 1978) have pointed out that the tectonic subsidence of rift-type basins is thermal in origin and is similar in form to that of a mid-oceanic ridge. According to the stretching model there is an initial subsidence, due to crustal thinning, followed by a thermal subsidence as the lithosphere cools. Irrespective of the amount of stretching, the subsidence during the thermal phase is exponential in form and increases linearly as the square root of time.

If the subsidence in Fig. 5 is thermally driven, the former positions of the depositional front (heavy lines) will be separated by the square root of time intervals. The rate of sediment supply to the basin (which can be estimated from the cross-sectional area of the load + flexure divided by the elapsed time) depends on the square root of time interval assumed. The rate is similar in form, however, for a wide range of square root of time intervals—decreasing initially and then slowly increasing with time.

DISCUSSION

The results of this paper suggest that flexural loading due to prograding sediment loads is capable of contributing to both coastal onlap and offlap as well as downlap patterns in a basinward direction. This is of interest since stratal geometry has been, and still is, the principle means by which Vail *et al.* (1977) infer sea-level changes in the geological past: onlap being generally indicative of an eustatic sea-level rise, and offlap a fall.

The actual control of sea-level changes on the development of stratigraphic patterns is complex, however, and depends on the relative rates of sea-level change, tectonic subsidence and, sediment supply (e.g. Pitman, 1978). Some insight into the role played by sea-level changes has recently been presented by Jervey (1988) and Reynolds *et al.* (in prep.). These workers constructed simple forward models for near-shore regions that take into account sea-level changes, tectonics and sediment supply. They argued that sea-level change is a major factor that controls stratal geometries and the alternating patterns of coastal onlap and offlap accompanied by downlap that are observed in continental margin basins. Jervey (1988) ignored isostasy but, Reynolds *et al.* (in prep.) took into account flexure. Both approaches, however, are based on the assumption that the supply of sediment to a basin is constant. The means that there is 'linkage' between aggradation and progradation (i.e. the amount of decrease in one determines the increase in the other) and that the relative proportion of these facies is determined by sea-level.

The supply of sediment to a basin is usually a poorly known parameter and it is unlikely that the sediment supply to a basin will be constant through time. Galloway (1989), for example, infers an 'episodic' supply of sediment to the Mississippi delta due to waxing and waning of different source regions in the hinterland.

In order to test Galloway's hypothesis, a simple model was constructed (Fig. 6) in which a subsiding margin of constant flexural rigidity was subject to repeated 'pulses' of sediment influx, progradation and flexure. During the first pulse, three sediment wedges prograde out a distance of about 150 km from the hinge 'line'. The stratal geometry that results is similar to Fig. 4, and shows limited coastal offlap accompanied by downlap in a basinward direction. In the next pulse, four wedges prograde. The first wedge of the second pulse loads back near the hinge line and therefore oversteps the older units of the first pulse. Subsequent wedges show limited offlap and downlap in a basinward direction. Finally, three wedges of the third pulse prograde out. The first of these wedges also oversteps the older units of the second pulse and the other two wedges show limited coastal offlap.

The synthetic stratigraphy in Fig. 6 resembles the model results of Jervey (1988), at least with regard to the occurrence of downlap in a basinward direction and the tracking of the shelf-break in the successive wedge loads. It differs through in predicting (a) depositional surfaces which steepen in a landward direction, and (b) coastal offlap. The Jervey (1988) model is characterized by depositional surfaces of constant dip and coastal onlap.

The constant dip of the former depositional surfaces is attributed to the fact that Jervey (1988) ignored the effects of isostatic compensation. In Fig. 6 the former depositional fronts show strikingly different dips from the present one. Usually the dips are shallower but sometimes, depending on the flexural wavelength, they are steeper. Depositional fronts that have a different dip to the present day front are a characteristic feature of the progressive flexural loading model, the slope difference being mainly a function of the T_e of the underlying basement.

Coastal onlap is not a feature of the prograding wedge model, yet it is widely observed in continental margin basins. In the Jervey *et al.* (1988) models, coastal onlap is the result of an increase in accommodation space due to rising sea-level landward of each prograding sediment wedges. The model in Fig. 6 omits the effects of eustatic sea-level changes, so it is of interest to determine whether there are any conditions whereby tectonics in the form of flexure can produce coastal onlap in prograding depositional systems.

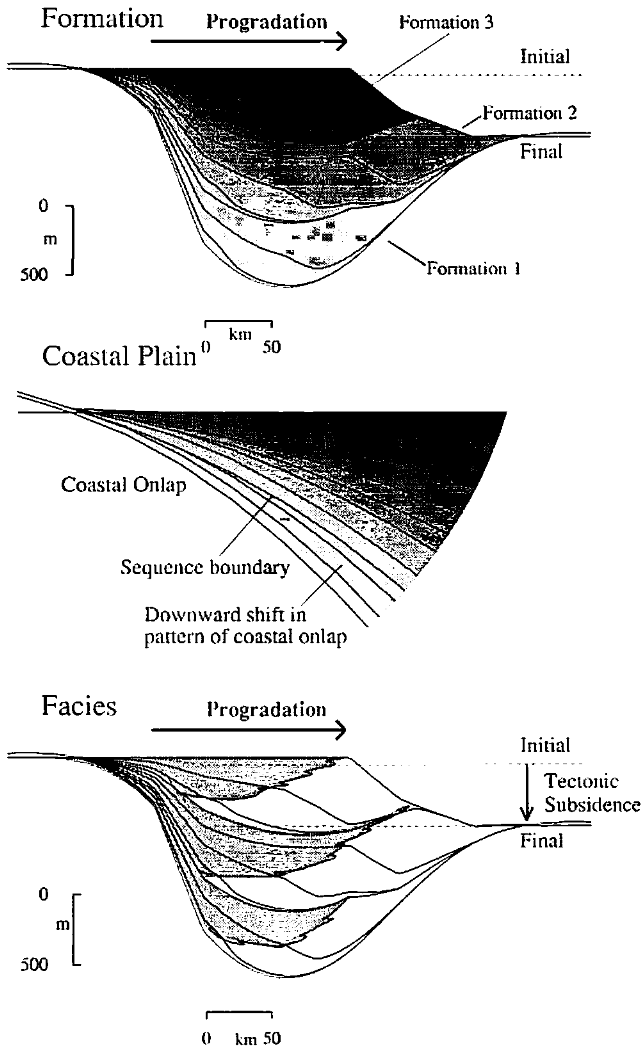

Fig. 6 Simple model for the progressive flexure of the lithosphere due to 'episodic' sediment loads on a subsiding margin. Three 'pulses' of sediment progradation that result in 3 formations are considered. The first involves 3 wedges (Formation 1) which prograde out a distance of about 150 km. The second involves 4 wedges (Formation 2) and the third 3 wedges (Formation 3). The lower profile shows the position of the shelf break in slope and represents the boundary between delta top and slope facies. The middle profile shows the detailed stratigraphy of the coastal plain region in the vicinity of the hinge line. The upper profile shows the stratigraphy of the margin after all 3 formations (i.e. a total of 10 sediment wedge loads) have been applied.

The tectonic requirements for coastal onlap are the same as they as those for sea-level changes namely: accommodation space needs to be created landward of a prograding wedge. In the flexural models discussed here this may be accomplished either by (a) increasing T_e of the lithosphere as each new wedge is added or (b) increasing the tectonic subsidence landward of the fixed hinge line.

Figure 7 shows the predicted stratigraphy for a simple model of a subsiding margin in which the prograding sediments load a lithosphere that is increasing its T_e with time. The figure shows the overall basin shape (Fig. 7a) and details of the stratigraphy at its edge (Fig. 7b). The model results are based on (a) an equal square root of time interval between the depositional fronts and, (b) a T_e that increases with age by an amount given by the depth to the 450°C oceanic isotherm. Three cases of change of T_e with time are shown: slow (lower profiles), medium (middle) and fast (upper).

The models in Fig. 7 show the stratigraphic consequences of the 'competition' that exists between progradation, which tends to drive the flexural node basinward, and a T_e increase, which drives it landward. The stratigraphy for a 'slow' increase in T_e is similar to what would be expected in the case of progradation at a relatively old continental margin, about 56–100 Myr after a rifting event. The T_e change in this case is not large enough to overcome the prograding load effect, and the stratal geometry shows offlap patterns, unconformities and limited *coastal onlap.* The stratigraphy for the 'fast' change in T_e is similar to what would be expected for a young margin 0–9 Myr after a rifting event. The change in T_e in this case is sufficiently large to overcome the prograding load effect and the stratal geometry is now dominated by coastal onlap. Both cases of T_e change with age show downlap in a basinward direction (Fig. 7a).

The stratal geometries in Fig. 7 resemble the model results of Jervey (1988, fig. 8) despite the fact that the predicted patterns of onlap and downlap were produced independently by different driving mechanisms. The coupled onlap and downlap pattern in the Jervey (1988) model is the result of an eustatic sea-level change whereas the patterns in Fig. 7 are simply the result of flexural loading of a prograding wedge system.

As was pointed out earlier a limitation of the Jervey (1988) model is that it assumes the supply of sediment to a basin is constant through time. There is therefore a linkage between the lateral extent of coastal onlap and downlap: an eustatic sea-level rise will produce an *increase* in onlap and a corresponding *decrease* in downlap while a fall will show a *decrease* in onlap and an *increase* in downlap. Stratigraphic sequences showing these relationships are apparently (Jervey, 1988) a common feature of rift-type basins, so it is of interest to see whether the prograding wedge model, which ignores eustatic sea-level changes, can also predict such a pattern.

There will be some relationship between coastal offlap/onlap and downlap in the prograding wedge model because of mechanical coupling. Figure 7a shows, for example, that the change in T_e from *slow* to *fast* is accompanied by a *decrease* in the pattern of downlap. This is because the slow change case has a high overall value of T_e which produces a broader basin and a wider pattern of downlap than the fast change case which occurs early on in the evolution of a basin when the average T_e is relatively low. Furthermore, Fig. 7b shows that the *slow* to *fast* case is accompanied by an *increase* in the tendency toward coastal onlap. Hence, the prograding wedge model can produce a *decrease* in downlap accompanied by an *increase* in coastal onlap—the

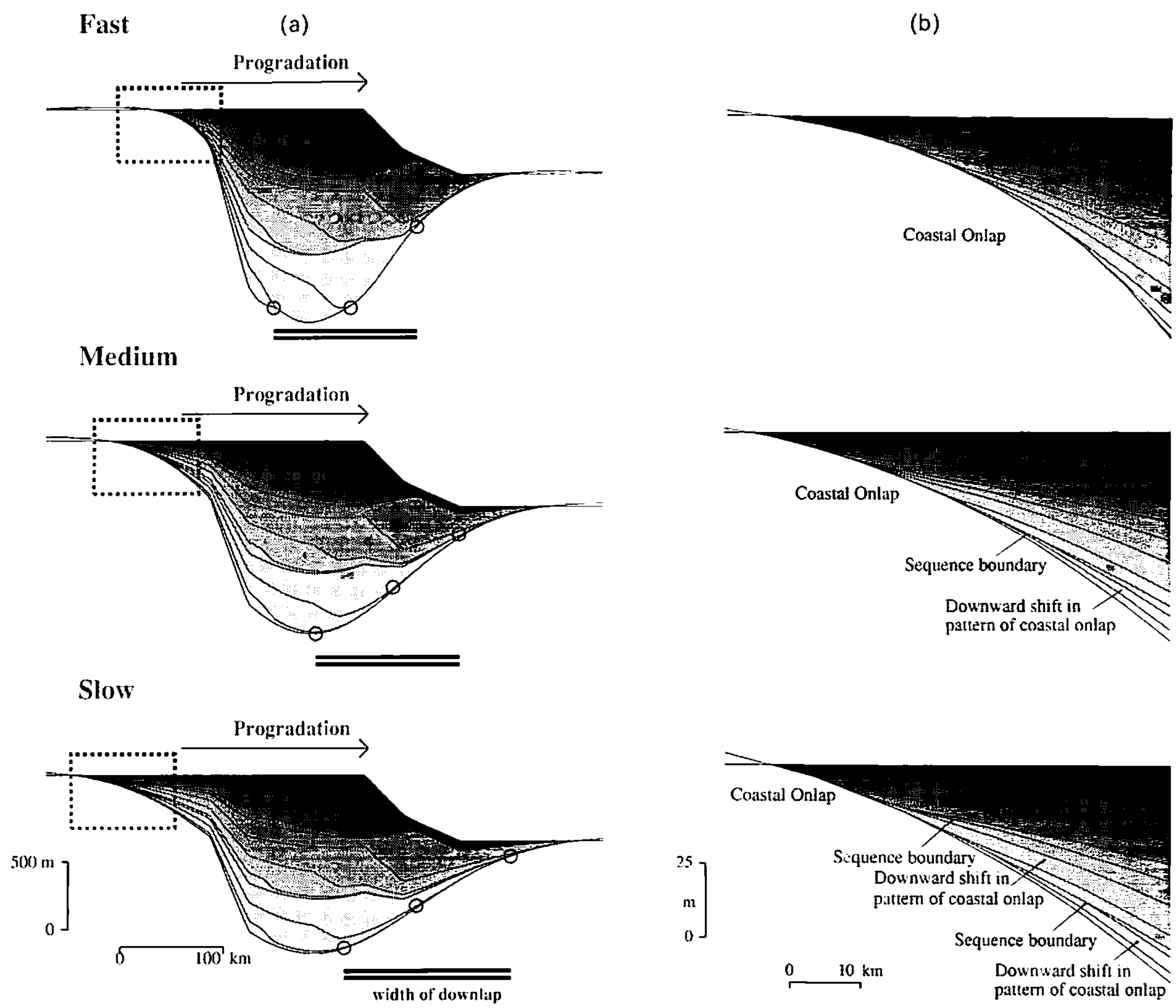

Fig. 7 Predicted stratigraphy of a prograding wedge system which loads a lithosphere with a T_e that is slowly changing with time (lower profile), rapidly changing with time (upper profile) and an intermediate case (middle profile). The same number of formations (and sediment wedge loads) have been applied as in Fig. 6. (a) Overall basin shape. The heavy dashed lines show the detailed stratigraphy presented in (b). Detailed stratigraphy of the basin edge. The figure shows an increased tendency for coastal onlap as the flexural rigidity changes more rapidly with time and a corresponding increase in coastal offlap as the flexural rigidity changes less rapidly.

same pattern predicted by Jervey's (1988) sea-level driven model.

The effects illustrated in Fig. 7 are further modified if we assume that rather than prograding over lithosphere that increases its flexural strength with age, sediments migrate across lithosphere that has lateral contrasts in flexural strength. Such contrasts may arise in continental margin basins due to fundamental differences in the long-term mechanical strength of continental and oceanic lithosphere (e.g. Steckler & ten Brink, 1986). They may also occur across boundaries between unstretched continental lithosphere, stretched lithosphere and oceanic lithosphere since it has been suggested, on the basis of gravity and geoid studies of rift-type basins in the North Sea (Barton & Wood, 1986), East Coast, U.S. (Watts, 1988) and Exmouth Plateau (Fowler & McKenzie, 1989) that stretched lithosphere may be weak during the post-rift development of extensional basins.

Figure 8 shows two cases of a prograding sediment wedge that migrates across a lithospheric plate which has a lateral contrast in flexural strength. The models assume that the strength contrast occurs across a narrow zone (thick vertical dashed line) and that a prograding wedge can 'sample' both strong ($T_e = 25$ km) and weak ($T_e = 5$ km) lithosphere. Since T_e is constant on either side of the narrow zone, the predicted stratigraphy shows both coastal offlap/onlap and downlap patterns. The pattern of offlap is best developed for the case (Fig. 8b) of sediments that begin on a strong lithosphere and poorly developed for the weak case (Fig. 8a). The pattern of downlap is greatest, however, for the weak case than the strong case. Figure 8 therefore suggests that increases in coastal offlap are accompanied by a *decrease* in the pattern of downlap, a result which, interestingly, is in the opposite sense to that of Fig. 7.

The results of this paper are not intended to imply that

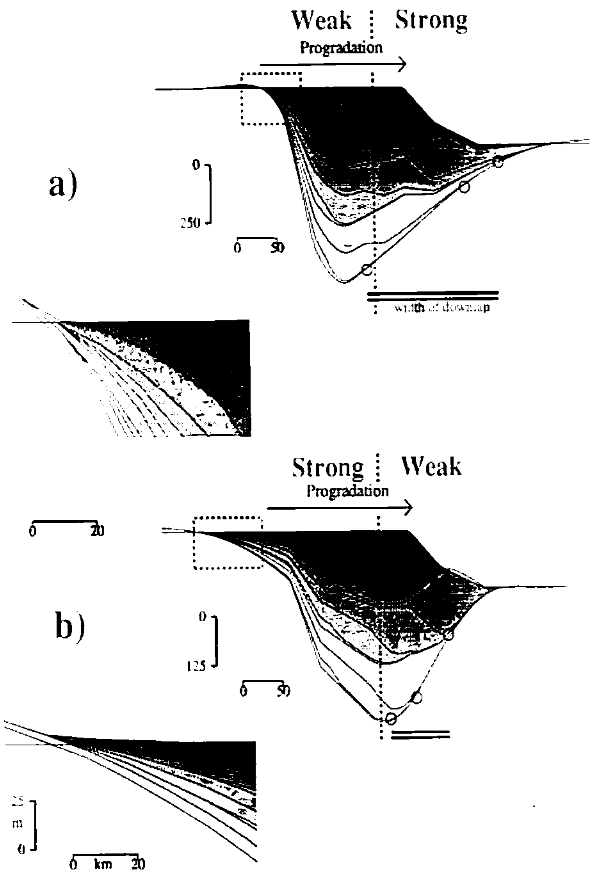

Fig. 8 Predicted stratigraphy for two cases of sediments that prograde across lithosphere that has lateral variations in flexural strength. The same number of formations (and sediment wedge loads) have been applied as in Fig. 6. The heavy dash line shows the boundary between the weak and strong zones. (a) weak–strong case, (b) strong–weak case.

flexural loading due to prograding sediment loads is the only mechanism that can generate coastal offlap/onlap and downlap patterns in sedimentary basins, or that factors such as eustatic sea-level changes and other tectonically induced movements of the land surface are unimportant. On the contrary, there are difficulties with the prograding wedge model that need to be explained before its role in controlling stratal geometries can be fully understood. The model, for example, does not account for sequence boundaries of large areal extent or the predominance of onlap over offlap in coastal sequences. Furthermore, the model requires a variable sediment supply that in some instances, due to tectonic or other events in the source region, can undergo large-scale, episodic, fluctuations. In spite of these difficulties, this paper has shown that progressive lithospheric flexure due to prograding sediment loads can produce a rich variety of stratigraphic patterns and may therefore be an important contributor to the geometry of sedimentary basins. Field-based studies of the sediment influx, facies, and clinoform geometry of 'case history' basins offer the most promise in the future in determining whether it is a primary factor or is secondary in importance to other factors such as eustatic sea-level changes and tectonic movements of the land surface.

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