Tectonic subsidence, flexure and global changes of sea level

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Tectonic models for the evolution of passive continental margins predict that following rifting, sediments should progressively onlap basement at the edge of a margin as the lithosphere cools and increases its flexural rigidity with age. The pattern of modelled onlap is strikingly similar to that used by Vail and colleagues to estimate sea-level rise through geological time. This similarity suggests that major portions of stratigraphical sequences at margins may have a tectonic, rather than eustatic, control. The patterns of onlap used by Vail and colleagues may be widespread, however, because several widely separated passive margins rifted at similar times, but they are unlikely to be worldwide.

THE distribution of seas through geological time is the result of two principal factors: global or eustatic changes of sea level and tectonic movements of the Earth's crust[1]. One or both of these factors may cause shorelines to shift and sedimentary facies to change. It may be expected that a relative rise in sea level, due either to a global rise in sea level or to crustal subsidence, would produce a transgressive stratigraphical sequence, while a relative fall of sea level, due either to a global fall in sea level or to crustal uplift, would produce a regressive sequence. The occurrence of a transgressive or regressive sequence at a particular locality in a sedimentary basin, however, is controlled by the rate of sediment influx and the rate and direction of change in the relative position of sea level and the sea floor[2]. It appears that global changes in sea level, due to changes in the volume of land ice or mid-ocean ridge crests, proceeded rapidly enough to explain some of the transgressive and regressive sequences at the edges of epeiric seas during the geological past. But there has been considerable debate[3] on the origin of the transgressive and regressive sequences that occurred during periods of no extensive land ice and relatively constant ridge crest volumes.

Sloss and Speed[4] argued that the major control on stratigraphical sequences in the continental interiors[5] were tectonic movements. They recognized three main episodes or 'modes' during the development of stratigraphical sequences in the continental interiors: 'oscillatory', in which land areas were subject to differential uplift and subsidence; 'emergent', in which land areas were progressively uplifted; and 'submergent', in which land areas progressively subsided. Sloss and Speed[4] noted a similar timing of these modes between widely spaced continents and argued for a synchronous tectonic control over broad regions.

Vail *et al.*[6,7], on the other hand, argued that the major control of sedimentary sequences in continental interiors and margins was global changes of sea level. They recognized several depositional cycles during the development of a stratigraphical sequence, each of which was bounded by surfaces of discontinuity. By identifying these surfaces, using characteristic patterns of seismic reflectors of 'onlap' and 'offlap', they constructed cycle charts at individual localities. Vail *et al.*[6] noted that similar cycles could be recognized at localities in widely separated margins and argued for a eustatic control. They used estimates of the amounts of onlap and offlap to construct a global sea-level curve for the Phanerozoic that was characterized by several short-term fluctuations superimposed on a broad long-term change. Vail *et al.*[7] used the pattern of onlap and offlap to infer sea-level rise and fall respectively. By calibrating their 'global' curve with other estimates of long-term sea-level changes[8,9] they argued that the short-term fluctuations could have magnitudes of up to a few hundred metres[10].

Pitman[9] pointed out that the occurrence of a transgressive or regressive sequence at the edge of a gently sloping shore depends on the rates of change of tectonic movements and global sea level. For example, a transgressive sequence could result from a sea-level fall, if the rate of subsidence of a basin exceeds the rate of sea-level fall. Pitman[9] showed that a major transgression during the Eocene in the North American and African margins could result simply from the interaction of changes in the rates of long-term sea-level fall with the steady, normal subsidence of these margins.

Vail and Todd[11] therefore modified the original statements[7] that coastal onlap and offlap could be directly equated to sea-level rise and fall. Their amended global sea-level curve for the Jurassic, however, was strikingly similar to the original curve, the main difference being the less abrupt sea-level falls.

The nature of the control of stratigraphical sequences has subsequently been widely debated. Hallam[12] argued that a tectonic control on the scale proposed by Sloss and Speed[4] was unlikely, as there was no satisfactory mechanism to explain why widely separated continents would be affected by tectonic movements at similar times. He suggested[13] that the main control on stratigraphical sequences were global changes of sea level and constructed a curve for the Jurassic, similar in overall shape to the Vail *et al.*[6] curve. But as Donovan and Jones[3] have pointed out, without a satisfactory mechanism, the

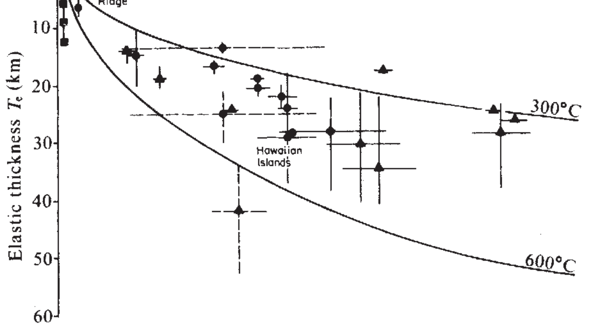

Fig. 1 Plot of elastic thickness T_e against age of the oceanic lithosphere at the time it was loaded[48]. The elastic thickness is determined from observations of the flexural rigidity and assumes Young's modulus = 10^{12} dyn cm^{-2} and Poisson's ratio = 0.25. ●, Seamounts and oceanic islands; ▲, deep-sea trench–outer rise systems; ■, ridge crests; ◆, river delta. Solid lines indicate the 300 °C and 600 °C oceanic isotherms based on the cooling plate model[49].

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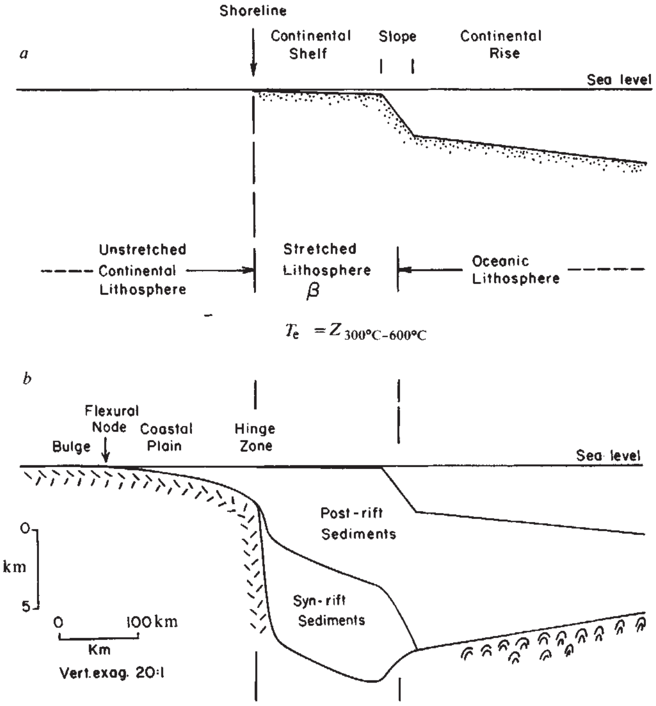

Fig. 2 Thermal and mechanical model for the tectonic evolution of a passive continental margin. *a*, Initial conditions. The tectonic subsidence of the margin is due to thermal contraction following heating and thinning of the lithosphere and crust at the time of rifting. Sediments are assumed rapidly to infill continental shelf, slope and rise regions and to maintain a constant bathymetric profile through time. *b*, Cooling and flexure. The sediments represent a load on the cooling lithosphere which responds by flexure. Sediments that form early in margin history load a relatively weak lithosphere while sediments that form later in its history load a strong lithosphere. The sediments that infill the initial subsidence are referred to as syn-rift while the sediments that infill the thermal subsidence are referred to as post-rift[31]. The model calculations assume an initial lithospheric thickness of 125 km, an initial crustal thickness of 31.2 km, a coefficient of volume expansion of 3.4×10^{-5} °C^{-1}, a mantle temperature of 1,333 °C and initial densities of 2.8 and 3.33 g cm^{-3} for the crust and lithosphere respectively. A uniform density of 2.5 g cm^{3} is assumed for the sediment infill.

large amplitude sea-level changes inferred by Vail *et al.*[6] and Hallam[12] must also remain in doubt. Bally[14] suggested, in fact, that the effects of global sea-level changes are subordinate to tectonics and that the major control on stratigraphical sequences are widespread and correlatable tectonic events associated with major plate reorganizations.

There is little doubt, however, that the sea-level curve of Vail *et al.*[6] represents an important synthesis of high-quality stratigraphical data that apparently has been useful in correlation studies and petroleum exploration. The outstanding question is the interpretation of these curves and whether they represent global sea-level changes or widespread tectonic events. Unfortunately, many of the data used by Vail *et al.*[6] are not generally available so it is difficult to determine whether they have satisfactorily separated eustatic from tectonic effects. This study uses recently developed models for the tectonic evolution of passive margins[15,16] to estimate how tectonics has contributed to stratigraphical sequences.

Tectonic subsidence at passive margins

The knowledge of the tectonic evolution of passive margins has progressed rapidly during the past decade. Backstripping studies[17–21], using data from commercial exploratory and COST-type wells, have shown that the principal factors affecting the post-rift subsidence of passive margins are thermal contraction and sedimentary loading. Factors such as chemical and mechanical compaction, palaeobathymetry, amount of sediment influx and long-term changes in sea level all contribute to the subsidence but their combined effects are small compared with thermal contraction and sediment loading. By correcting the well data for the effects of sedimentary loading it has, therefore, been possible to estimate the form of the thermal contraction and how it may vary across a margin.

Most modelling studies now assume that thermal contraction at passive margins arises from heating and thinning of the crust and lithosphere at the time of intial rifting[22,23]. The main differences between the various models is the manner by which the heating and thinning occurs. Sleep[22] proposed that the thinning was caused by uplift and subaerial erosion while McKenzie[23] proposed that it was caused by a passive, uniform extension of the crust and lithosphere.

There is good observational evidence for crustal extension during the early rifting history of passive margins, particularly at the sediment-starved margins of the eastern Atlantic and portions of the margin off eastern North America. For example, de Charpal *et al.*[24] have mapped listric faults and tilted fault blocks in northern Biscay, and Given[25] and Schlee[26] have mapped graben systems off Nova Scotia and New England. The amount of extension cannot easily be estimated from fault geometry, but seismic refraction data are consistent with a stretching factor[23] of $\beta = 2$–3 over distances of 100–200 km in northern Biscay[27,28].

The stretching model[23] has also been applied to the thickly sedimented margins of the western Mediterranean off southern France[21] and portions of the margin off eastern North America[29,30]. The main difficulty is the lack of information at these margins on the thickness and depositional environment of the sediments formed during rifting and, the tectonic fabric of the underlying basement rocks. For example, the relative proportion of syn-rift to post-rift sediments, an important constraint in the stretching model[31], is not precisely known at these margins. The available evidence[29,30] suggest a stretching factor in the range $\beta = 1.5$–6.0. Unfortunately, there is too little seismic refraction data to constrain these estimates satisfactorily but geoid data are consistent with large amounts of stretching ($\beta = 3$–4) over distances of 100–200 km along portions of these margins (ref. 20 and M. S. Steckler, personal communication).

The model of McKenzie[23] therefore appears to give a general explanation of the available geological and geophysical data at passive margins. Some problems still remain, however, the most important of which are: the discrepancies between estimates of stretching based on fault geometry and deep crustal seismic refraction data[28], the occurrence of seaward-dipping reflectors of possible subaerial volcanic origin at some margins[32] and the relative role of passive and active heating in the thermal evolution of a margin[21].

Flexure

The response of the lithosphere to sediment loads at a passive margin has been modelled using either an Airy or flexure model of isostasy[17,18]. Seismic reflection profiles indicate that active faulting accompanies the early stages of rifting[24–26]. This suggests that an Airy model, in which sediments are locally compensated, is most applicable early in margin evolution. The presence of gently dipping post-rift sediments and a broad coastal plain, however, suggests the flexure model becomes more important later in margin evolution.

The actual lithospheric response to sediment loads is difficult to determine at a passive margin because sediments are continually added to it during long periods of time. The best evidence has come from studies of seamounts and oceanic islands, as they formed during relatively short time periods (1–1.5 Myr). These studies show that the flexural rigidity and the equivalent elastic thickness of oceanic lithosphere, T_e, increases with the age of the lithosphere at the time it was loaded (Fig. 1). Loads formed on young oceanic lithosphere, such as the Walvis ridge, are associated with small values of T_e while loads formed on old oceanic lithosphere, such as the

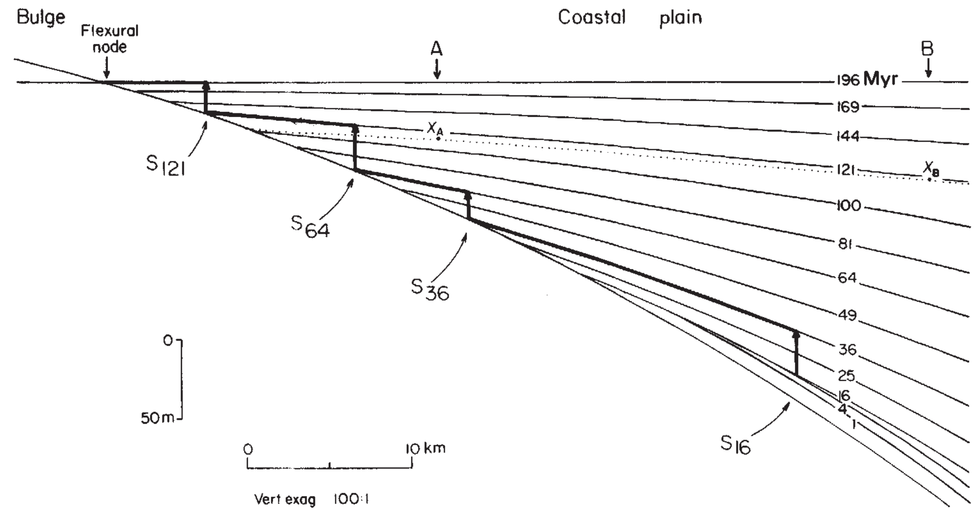

Fig. 3 Coastal plain stratigraphy calculated using the thermal and mechanical model in Fig. 2. The model calculations assume $\beta = 3.0$ and $T_e = Z_{450}$ °C. Solid lines represent individual stratigraphical horizons at different times following rifting. The effects on the stratigraphy of compaction have not been included. The heavy solid lines illustrate the procedure used by Vail *et al.*[7] and this study to estimate the vertical component of coastal onlap. The dotted line indicates the equilibrium shoreline where the rate of change of long-term sea-level fall[9] equals the subsidence rate. Because the coastal plain sediments subside at a slower rate than long-term sea-level rises or falls, there is a tendency, depending on the rate of sediment influx, for transgression below the dotted line (enhancing the effects of flexure) and regression above the line (competing with flexure).

Hawaiian islands, are associated with large values. Figure 1 shows there is good general agreement between T_e and the depth, Z, to the 300–600 °C oceanic isotherms. Thus, flexure studies in the ocean basins suggest that as the oceanic lithosphere increases in age and cools, it becomes more rigid in its response to loads.

Because the lithosphere at passive margins is extensively heated during rifting[22,23], it should therefore become progressively more rigid in its response to sediment loads as it cools following rifting. Sediments formed soon after rifting would be expected to load a relatively hot and weak lithosphere while later sediments would load a relatively cool and strong lithosphere. A flexure model, in which the rigidity of the lithosphere increases with time, explains certain tectonic–stratigraphical features of well sedimented passive margins such as an outer stratigraphical high, a hinge zone and a coastal plain in which younger sediments progressively onlap basement[33].

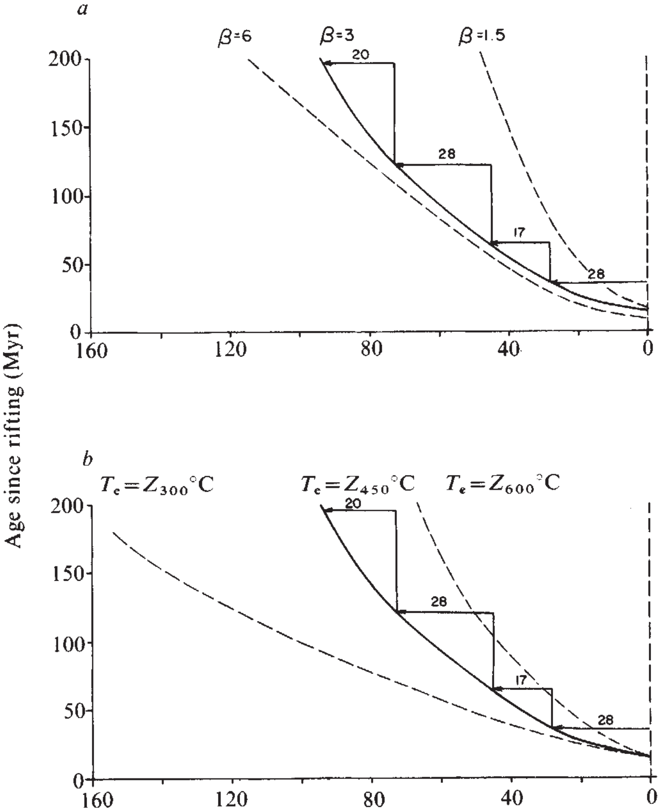

Fig. 4 The vertical component of coastal onlap estimated from the thermal and mechanical model in Figs 2 and 3. Solid lines indicate the individual measured increments of coastal onlap and the heavy solid lines indicate the smoothed change in onlap following rifting. *a*, $T_e = Z_{450}$ °C and β variable. *b*, $\beta = 3.0$ and T_e variable.

Thermal and mechanical models

Thermal and mechanical models have now been constructed[15,16,34] for the stratigraphy of passive margins that combine the effects of thermal contraction and sedimentary loading. In these models, thermal contraction occurs following crustal and lithospheric extension at the time of rifting while sedimentary loading occurs by flexure of a progressively more rigid basement following rifting. Since T_e is a strong function of temperature (Fig. 1), thermal and mechanical effects can be coupled. Thus, T_e can be calculated as a function of time and position following rifting.

Figure 2 shows a simple model for the tectonic evolution of a passive margin that formed by crustal and lithospheric extension at the time of rifting. The region of extension is limited to beneath the shelf and slope and to have a magnitude ($\beta = 3$) and horizontal extent (175 km), similar to present day margins. Sediments are assumed to infill uniformly the cooling margin, maintaining a constant bathymetric profile with time. The model includes the effects of lateral heat conduction across the stretched region and flexure that varies as a function of time and position ($T_e = Z_{450}$ °C).

Figure 3 shows the stratigraphy predicted by the thermal and mechanical model at the edge of the margin. The solid lines indicate stratigraphical horizons at equal increments of the square root of time since rifting. Figure 3 shows that there is initially a significant onlap of sediments onto the basement, due to the abrupt transition in the stretching model from fault-controlled Airy-type subsidence to flexural controlled subsidence. Soon after rifting, the pattern of onlap is abruptly terminated because lateral heat flow from the stretched region to unstretched continental lithosphere causes the coastal plain to remain emergent[15,16]. Beginning about 16 Myr after rifting, however, flexure overcomes the effects of lateral heat flow and younger sediments progressively onlap the basement, due to its increase in flexural rigidity with age.

The model in Fig. 3 is simplified as it assumes an abrupt increase in β across the hinge zone and a sediment influx that

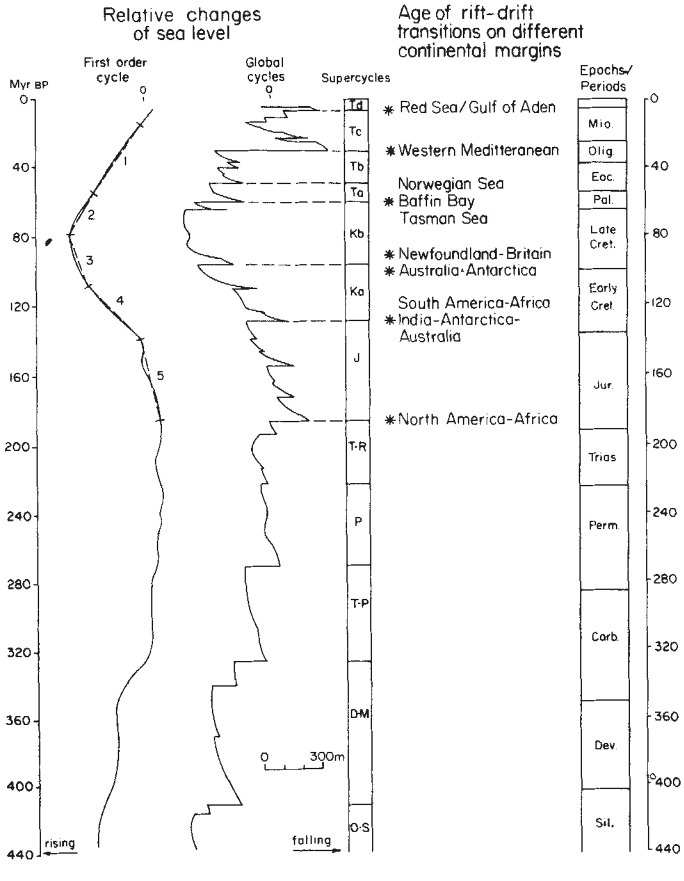

Fig. 5 Global cycles of relative changes of sea level since the Devonian[6] compared with the age of the rift–drift transition at different continental margins[48]. The numbered dashed lines on the 'first-order cycle' are construction lines used to estimate the rates of change of sea-level rise and fall in Fig. 6.

uniformly infills shelf, slope and rise regions. Although there is little information on how β varies across a hinge zone, it seems likely that it would change more smoothly than assumed in the model. An abrupt change in β maximizes the lateral effect of heat flow across the hinge zone so the coastal plain would remain emergent for a longer time following rifting. However, a more gradual transition in β across the hinge zone only slightly decreases the pattern of onlap in Fig. 3. The pattern of onlap would be expected to change if sediments by-passed shelf regions due, for example, to vigorous submarine erosion[26]. A prolonged absence of shelf loading would cause the flexural node (Fig. 2*b*) to migrate seaward towards loaded slope and rise regions, resulting in offlap in the coastal plain. Changes in the pattern of sediment influx seem unlikely, however, to compete with the modelled onlap in Fig. 3 unless it, like flexure, varies systematically during margin evolution.

The model studies therefore suggest that coastal onlap is a characteristic feature of stratigraphical sequences at a cooling margin. This result is important since measurements of the amount of onlap at present day and ancient passive margins is the principal means by which Vail *et al.*[6,7] estimate sea-level changes through time. The models in Figs 2 and 3 have been constructed assuming no sea-level changes through time. Thus, by measuring the amount of onlap in the models we can estimate the contributions of tectonics to stratigraphical sequences.

Vail *et al.*[7] measured onlap by estimating the 'coastal aggradation', or 'vertical component of onlap', in a stratigraphical sequence. Beginning with the lowest point of onlap in a sequence, they successively summed each increment of onlap until the highest point of onlap is reached. Figure 3 illustrates, on the modelled stratigraphy, the procedure used by Vail *et al.*[7]. Starting with S_{16}, the lowest point of onlap in the sequence, the first increment of onlap is 28 m and the time interval is 16–36 Myr. Successive increments of 17, 28 and 20 m give a total amount of onlap for the sequence of 93 m. Figure 4 shows the largest amount of onlap (100–160 m) occurs for the greatest amount of stretching and weakest plate, while the smallest amount (45–65 m) occurs for the least stretching and strongest plate. The form of the onlap is similar in each model, generally following the shape of an exponential curve, with an initial rapid increase followed by a more gentle increase.

Although they did not detail their method, Vail *et al.*[7] used coastal onlap to estimate sea-level rise through time. They used a modal average of three or more correlative cycles of onlap from different continents to construct a global sea-level curve for the Phanerozoic. But, as Fig. 4 shows, coastal onlap is a characteristic feature of the tectonic evolution of a cooling passive margin and does not require sea-level changes to produce it. Therefore, if the model predictions are correct, some correlation should exist between the beginning of major cycles of onlap in the Vail *et al.*[6] curve and the age of the rift–drift transition in passive margins, formed as a result of continental break-up.

Figure 5 shows there is an excellent visual correlation between the beginning of the supercycles of Vail *et al.*[6] and the age of the rift–drift transition at margins formed by the breakup of the Pangea supercontinent. There is a striking similarity between the form of the modelled onlap and individual supercycles (Fig. 5). This similarity suggests, therefore, that major portions of the Vail *et al.*[6] curve may have a tectonic, rather than eustatic, control.

Vail *et al.*[6], however, terminate their supercycles by rapid sea-level falls (Fig. 5). The tectonic models predict regressive sequences early in the development of a margin (Fig. 3) but, they do not predict them at higher levels in a stratigraphical sequence. Therefore, if the magnitudes of the sea-level falls deduced by Vail *et al.*[6] are correct then, factors other than flexure must be invoked to explain them.

There is some doubt that Vail *et al.*[6] have correctly estimated the amount of sea-level fall from stratigraphical sequences[14,35]. The main problem is that they determine sea-level fall by measuring the vertical component of onlap between the highest point of onlap in an underlying sequence directly to the lowest point of onlap in an overlying sequence. Because the record is largely removed by erosion during a sea-level fall, they cannot use the 'incremental method' that was used to measure onlap. Thus, the effects of tectonics during the sea-level fall are not satisfactorily accounted for.

There is good evidence[36], however, for major periods of regression in the geological record, although their origin is in some doubt. For example, Grasty[37] has related major regressive events to crustal thickening associated with orogeny while Ager[36] has related them to long-term changes in global sea level, due to changes in mid-ocean ridge crest volumes.

The effects of long-term changes in sea level[9,6] on the modelled stratigraphy are illustrated in Fig. 6. The open triangles (Fig. 6) indicate the age since rifting when the rate of sea-level fall equals the rate of model subsidence and the shoreline would stabilize. For small times following rifting the seas would rise faster than the model subsides, causing a transgression, while for longer times the seas would fall faster, causing a regression. The actual effect of long-term sea-level changes depends, however, on the relative rates of sediment influx and subsidence[2]. For the modelled stratigraphy in Fig. 3, a transgressive sequence would be expected to develop soon after rifting, enhancing the pattern of onlap produced by flexure, while for later times a regressive sequence would develop, competing with flexure. The timing of the resulting seaward shift in onlap would depend on position, but for the model in Fig. 3 it occurs about 112–117 Myr following rifting.

The model predictions are difficult to compare with the Vail *et al.*[6] curve as this curve represents the summation of several shifts in the pattern of onlap from different positions in widely

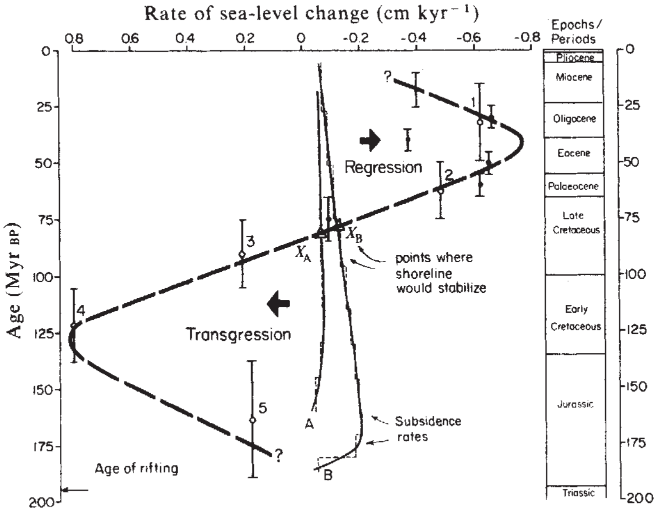

Fig. 6 Comparison of the rate of change of sea level to the subsidence rate at two points (A, B) in the thermal and mechanical model in Fig. 3. The rate of change of sea level is based on the long-term sea-level curve of Pitman[9] and the first-order cycle of Vail *et al.*[6]. ●, Based on Table 2 of ref. 9; ○, a straight line fit to the first-order cycle of Vail *et al.*[6] (Fig. 5). △, The time in Myr BP that the rate of sea-level fall equals the rate of subsidence and the shoreline, in the absence of other effects, would be in equilibrium.

spaced margins. There is evidence though, from published geological cross-sections of the US Atlantic coastal plain[38], of a progressive onlap of Jurassic to early Cretaceous sediments onto basement that is terminated by a seaward shift in the overall pattern of onlap about 120 Myr after rifting. Thus, the model in Fig. 3 seems to agree with the overall stratigraphy of this relatively old margin.

The tectonic models are still unable, however, to explain the widespread rapid short-term falls of sea level in the Vail *et al.*[6] curve. Thus, without a satisfactory mechanism, the origin of these falls remains in doubt.

Discussion

The model studies suggest that tectonics is an important control on the development of stratigraphical sequences at passive margins. Flexure produces patterns of onlap, remarkably similar in form to those used by Vail *et al.*[7] to infer sea-level rise. Although Vail *et al.*[7] attempted to remove tectonic effects by summing increments of coastal onlap in a sequence, they did not correct for flexural effects that vary as a function of time and position. They considered only a limited role for tectonics that was generally equivalent to assuming sedimentary loading occurs by flexure of a lithosphere of uniform flexural rigidity.

As has already been pointed out, however, flexure cannot explain all the occurrences of onlap in stratigraphical sequences; neither can it explain the occurrence of offlap at high levels in stratigraphical sequences. The Cenomanian and Callovian transgressive sequences[36] extended too far into the interiors of the continents to be explained by flexure, so other factors, such as long-term changes in global sea level due to changes in mid-ocean ridge crests, are required to explain them. Similarly, the Oligocene regressive sequences of the eastern US[35], western Australia[39], and the northern North Sea[6] cannot be explained by flexure. These relatively old margins were subsiding at relatively slow rates during the Oligocene so the relatively rapid long-term fall in global sea level[9] since 85 Myr BP may have contributed to these sequences. Furthermore, flexure cannot explain transgressive sequences such as the Hettangian[40] of south-west Britain, which were too far north to have been affected by the Tethys passive margin[41], or regressive sequences such as the Oligocene of southern Africa[42]. Tectonics, but not necessarily flexure, may have contributed to these sequences[40,42].

Figure 7 shows the overall pattern of stratigraphical sequences predicted by the tectonic models for different age margins. Following rifting each margin would be expected to show its own pattern of onlap as it cools and becomes progressively more rigid with age. But as each margin increases in age, there is a greater tendency to be affected by long-term changes in sea level. For example, the fall in long-term sea level from 85 to 15 Myr BP (ref. 9) would cause a shift in the pattern of onlap at approximately similar times in each margin. The patterns of onlap would also be affected, of course, by the rise in sea level before 85 Myr, but this would only enhance the effect of flexure. Figure 7 shows that although there is a similarity in the timing of the shift in onlap at each margin, the overall pattern of onlap is strikingly different between each margin. The models predict, in fact, that there would be little correlation in patterns of onlap between margins unless they rifted at similar times. Thus patterns of onlap, of the type used by Vail *et al.*[7] may be widespread, because many widely spaced margins rifted at similar times, but they are unlikely to be worldwide.

Unfortunately, there is presently too little stratigraphical data available to test the model predictions in Fig. 7. Vail *et al.*[7] only published a few sea-level curves from separate localities and most of these extend for the duration of only one supercycle. However, Todd and Mitchum[43] published sea-level curves for Texas Gulf Coast and West Africa; localities associated with the rifting of North America from Africa at ~196 Myr BP (ref. 44). Although it is not possible to compare their curves directly with the model predictions, both curves show a steady increase in the amount of onlap during the Jurassic. The main difference is that the models do not predict the short-term falls in sea level, the largest being in the Valanginian (125–130 Myr BP). As Todd and Mitchum[43] pointed out, however, a restricted Valanginian sequence can only be inferred from seismic profiles in the Gulf Coast.

This article has focused on the nature of the control on stratigraphical sequences in present day passive margins but tectonics, in the form of lithospheric flexure, may also have been a major control in the continental interiors. For example, Vail *et al.*[6] constructed a sea-level curve for the Devonian and Cambrian; presumably based on data from the Michigan basin and the western and eastern US orogenic belts. They show a

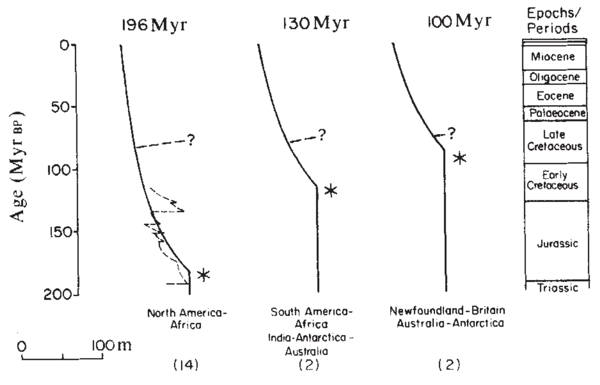

Fig. 7 The pattern of onlap predicted for passive margins of different ages. The heavy solid line represents the pattern of onlap predicted from the thermal and mechanical models with $\beta = 3$ and $T_e = Z_{450}$ °C. The heavy dashed line indicates the shift in the pattern of onlap caused by a fall in long-term sea level since 85 Myr BP. The actual shift in the pattern would vary across a margin. The fine dashed line indicates the sea-level curve for the Texas Gulf Coast based on Todd and Mitchum[44]. This curve has not been calibrated[44] and is shown for comparison only. There is a significant difference in the patterns of onlap between each margin because they rifted at separate times. The numbers in parentheses indicate the number of localities that Vail *et al.*[6] apparently used in each margin. Because they used a modal average of patterns of onlap from individual localities, the pattern of onlap from the North America–Africa margin should dominate their global sea-level curve.

prominent pattern of onlap for each of these periods which they interpreted as a sea-level rise. The increase in width of the Michigan basin during Devonian[45] can be explained[46], however, by an increase in flexural rigidity of the lithosphere with age, and does not require sea-level changes to produce it. Furthermore, the apparent progressive onlap of sediments onto the western and eastern margins of the North American craton during the Cambrian[47] can be explained by flexure, although the large horizontal extent of the Albertan may require long-term sea-level changes in addition.

This is not to imply that a completely satisfactory model exists yet for the tectonic evolution of sedimentary basins in present day and ancient passive margins. There is too little seismic reflection and refraction data at present day margins that satisfactorily constrains the deep crustal structure implied by the thermal models. The occurrence of coastal onlap at a passive margin, however, is the consequence of the cooling of the lithosphere following rifting and does not depend on a particular type of thermal model. The actual separation of tectonic effects from stratigraphical sequences remains a difficult problem that will require careful studies of seismic, geological and palaeontological data from localities in widely separated margins. The problem is important for future consideration, however, as it has major implications for tectonics, lithospheric mechanics, correlative stratigraphy and the thermal history of the Earth.

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Human EJ bladder carcinoma oncogene is homologue of Harvey sarcoma virus *ras* gene

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Examination of homologies between retroviral oncogenes and transforming sequences defined by transfection reveals that the human bladder carcinoma (EJ) oncogene is homologous to the Harvey sarcoma virus oncogene (ras). *Structural analysis limits the region of homology to a 3.0-kilobase* SacI *fragment of the EJ oncogene. Both EJ and* ras *DNA probes detect similar transcripts in transfectants derived from bladder carcinoma cell lines.*

TWO groups of cellular oncogenes have been discovered during the past decade. The first consists of genes that were characterized by virtue of their association with retroviruses. The prototype of this class is the *src* gene of avian sarcoma virus. Several experiments have indicated that this gene was acquired from the chicken genome by an avian retrovirus, and has been exploited by the chimaeric virus to transform cells[1,2]. Once incorporated into the viral genome, expression of the *src* gene is driven by viral controlling elements and is no longer responsive to control mechanisms that governed its expression while in the cellular chromosome. In addition to the *src* gene, this group includes at least 12 other gene sequences, each associated with a different chimaeric retrovirus[3,4]. These genes are conserved over great evolutionary distances[2,5] implying that they mediate essential cellular or organismic functions.

A second class of cellular transforming genes has been detected by the experimental route of DNA transfection. Recent reports have indicated that the DNAs of some non-virally induced tumour cell lines can induce transformation when applied to mouse fibroblast monolayers. These tumour cell lines are derived from chemically induced animal tumours[6–9] and from human tumours of spontaneous origin[7,10–12].

The two classes of oncogenes have many properties in common, the most striking of which is the apparent origin of both types of genes from normally benign, cellular genetic elements. Because of this and other parallels, we undertook a search to determine whether the two groups of genes shared any members in common. Such overlap would have far-reaching consequences for our understanding of the mechanisms of viral and non-viral carcinogenesis.

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