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A comparison of foreland and rift margin sedimentary basins

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Foreland and rift margin basins are compared on the basis of (1) their tectonic setting, (2) reasons for subsidence, and (3) their large-scale geophysical and geological characteristics. The thermal and mechanical properties of the underlying lithosphere are shown to be fundamental to the form of tectonic subsidence. The lithosphere beneath foreland basins is flexurally downwarped by the loading of the adjacent fold–thrust belt, whereas tectonic subsidence at rifted margins is caused by mass replacement at depth, during lithospheric extension on rifting, and subsequent thermal contraction as the lithosphere cools.

The effects of rheology, thermal maturity, and lateral changes in properties of the lithosphere are outlined for foreland basins, as is the topographic effect of possible phase changes beneath the fold–thrust belt. The thermal and rheological consequences of lithospheric extension at rift margins makes flexural subsidence relatively less important than in foreland basins. Flexure may, however, be partly responsible for uplift landward of the hinge line that is associated with rifting. Other mechanisms that could cause such uplift include depth-dependent extension and thermal expansion due to the lateral diffusion of heat.

The models describing the evolution of these basins are shown to predict characteristics that are in accord with observations. The superposition of foreland and rift margin basins as a result of ocean closure can lead to an overall basin stratigraphy that is complex. Such phases of basin subsidence must be separated according to the tectonic environment in which they formed in any analysis of the cause and consequences of basin evolution.

Introduction

The intent of this paper is to provide a descriptive geological interpretation of some recent theoretical models of the origin and evolution of foreland and rifted margin sedimentary basins, a discussion of their largest-scale characteristics, and a description of the geological and geophysical evidence that is needed to test these theoretical models. The models, based on simple concepts of continuum mechanics and conductive heat transport, appeal to the fundamental properties of the lithosphere under compression and tension to explain the formation of these two basin types. The models are also shown to fit within the plate tectonic framework and to be a direct consequence of horizontal motions and interactions of lithospheric plates that is central to the plate tectonic model. The only additional assumption is that the plate boundaries must be deformable.

The theme of the paper is to propose two archetypal models for the tectonic subsidence of these basins and to develop the models with regard to lithospheric rheology, lateral changes in lithospheric properties, and the response to sediment and water loading. 'Tectonic subsidence' implies subsidence that would occur without loading of the lithosphere by sediment or water infill of the basin. Predicted characteristics of each basin type and underlying lithosphere are then compared, some simple examples listed, and the consequences of superposition outlined. The content of this paper is partly based on earlier papers by Beaumont (1978, 1979, 1981), Beaumont *et al.* (1982) and Keen *et al.* (1981 *a*, *b*), to which the reader is referred for details of the methodology. It is intended as an introduction to, and an amplification of, concepts and results presented in those papers.

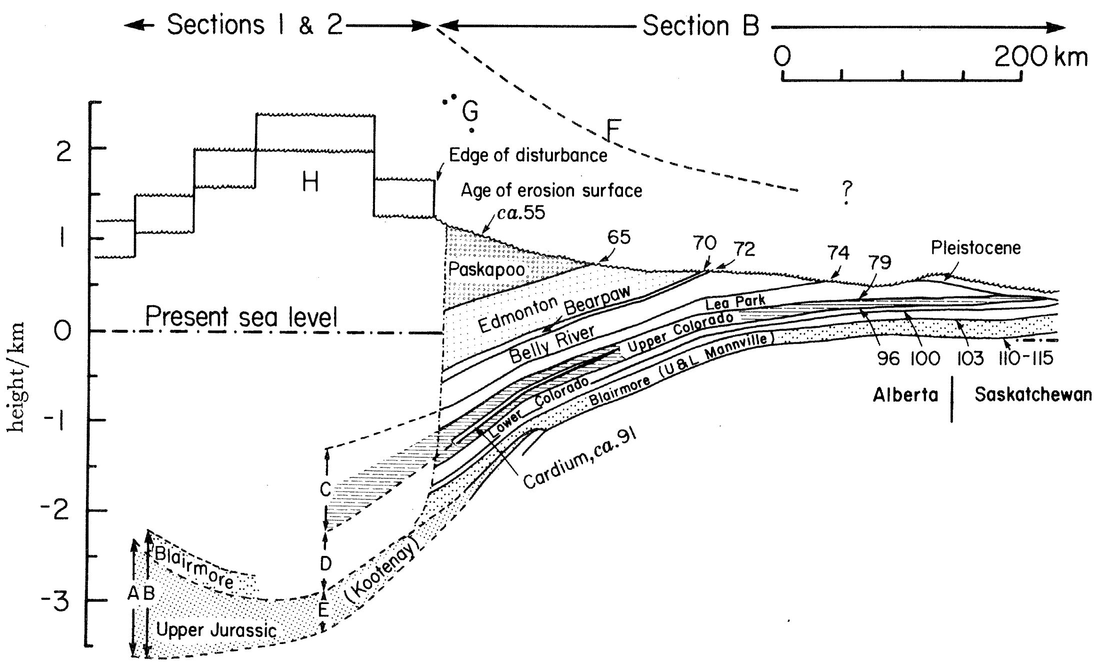

FIGURE 1. Simplified cross section of the post-mid-Jurassic foreland basin sequence of the Alberta Basin and adjoining Canadian Cordillera. The sections 1 and 2 and B extend from approximately 52° N, 118° W to 55° N, 105° W, as shown in Figure 6 of Beaumont (1981). A–E are palinspastically restored stratigraphic units; F and G are estimates of the position of the depositional baseline before erosion from coal moisture content of near-surface coals and shale compaction. The distance from F to the present surface is therefore an estimate of the amount of post-Laramide erosion that must be taken into account in any quantitative model of basin evolution. H are upper and lower bounds on topography of the adjacent 300 km of the Cordillera averaged on 50 km wide strips. The numbers are approximate ages in megayears of the important stratigraphic horizons. Note that much of the basin is above present sea level. The thickness of the crust beneath the sediments is not shown but is believed to be relatively uniform, particularly when compared with that at rifted continental margins.

Figures 1 and 2 illustrate cross sections of the Alberta Basin and Nova Scotian continental margin, which may be regarded as mature examples of foreland and rifted margin basins. Both basin types are asymmetric, a property inherited from the plate interactions responsible for their formation.

Foreland basins, which reach a maximum depth of *ca.* 6 km, are always bordered by an overthrust belt, a region of allochthonous sedimentary or crystalline crustal rocks, or both, that has been emplaced on the neighbouring lithosphere during overthrusting. This belt forms a topographically high mountain range of the Cordilleran or Himalayan type. The basin is wedge-shaped in cross section, with maximum depth adjacent to the mountains. The stratigraphy usually comprises similar laterally thinning units that may be truncated by erosion to expose sediments of increasing age with increasing distance from the mountains. In other examples successive sedimentary units appear either to thin to a common point, or to overstep each other, or are disrupted by the underlying sedimentary or basement movements that are believed to be contemporaneous with the emplacement of the fold–thrust belt. Along strike the basin depth and height of mountains vary, but the relation between basin and fold–thrust belt is maintained to give many examples of a sinuous, fundamentally two-dimensional character. Another universal feature is the lateral advance of the fold–thrust belt driving the foreland

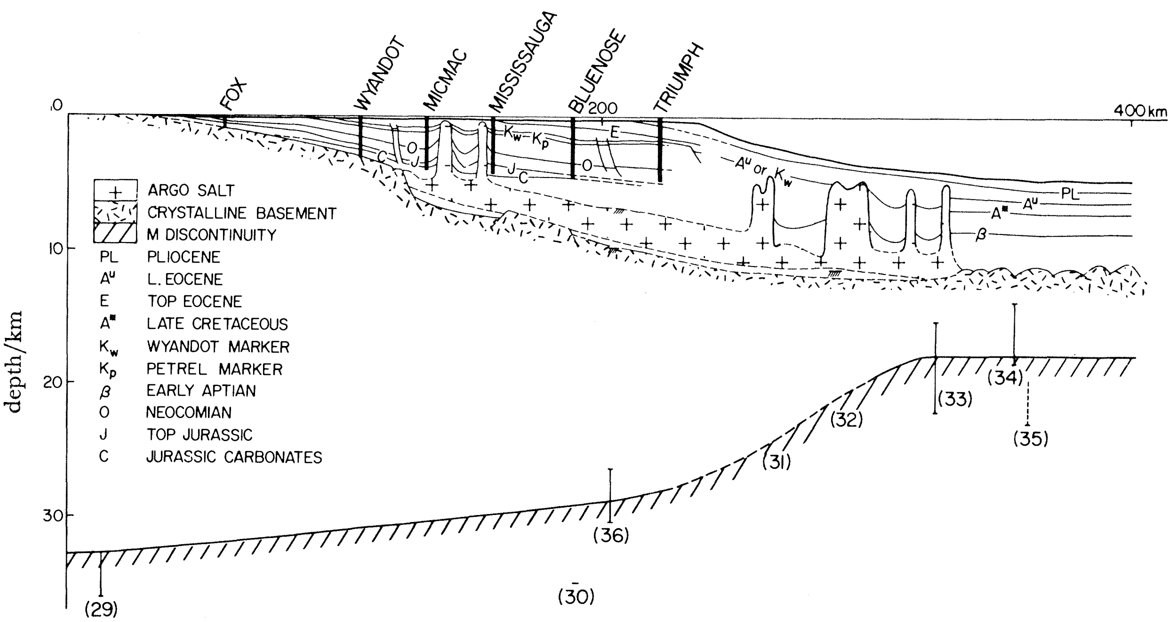

FIGURE 2. Simplified cross section of the Nova Scotian margin extending from approximately 45.5° N, 60.5° W to 42° N, 58.5° W, as shown in Beaumont *et al.* (1982). The depth to the Moho at the ends of seismic refraction lines adjacent but normal to the profile are shown projected onto the profile as vertical bars, which therefore represent the variability of crustal structure along strike. Regions where the positions of interfaces remain somewhat uncertain are shown as broken lines. The position of the basement–sediment interface beneath the basin is specified on the basis of wells that intersect it (Fox and Wyandot), seismic reflexion data, and seismic refraction data (shown as baseline with hatching). This basement is thought to be that of syn- and post-rift sediments; older sediments may extend to greater depth. Stratigraphy of the basin is based on evidence from deep exploratory wells (shown as vertical bars) and seismic reflexion data. Lithologies are regional averages representing the dominant component. Where unshaded, the sediments are thought to be mainly clastics. Stratigraphic and seismic marker horizons are labelled according to the key.

subsidence in front of it and incorporating successive basin units as the advance continues. These units are detached and transported basinward as thrust sheets. Detachment may be confined to the sediment column or extended into the underlying crystalline crust. Unroofing of the fold–thrust belt and uplift of the basin are also common features once the thrusting has ceased.

Rifted margin basins are usually deeper than foreland basins, sometimes reaching depths of *ca.* 16 km. They occupy tensional environments in the transitional region between continent and ocean and are not associated with fold–thrust belts. The shape of the basin is determined by the sedimentation: high rates of sediment influx will produce a prograding deltaic wedge, but a simple basement ramp will occur in a sediment-starved basin. The stratigraphy is also variable, depending on sediment input to the evolving basin. Rarely is there the overabundance of sediments that is often seen in foreland basins. This is because the basin can continue to build seaward.

Foreland basins and the adjacent fold–thrust belt

The term foreland basin is used here in a manner that is descriptively synonymous with exogeosyncline (Kay 1951), marginal basin (Krumbein & Sloss 1963), foredeep (Aubouin 1965), foreland and foredeep trough (Price 1973), peripheral and retroarc foreland basin (Dickinson 1974), and the marginal down-flexure of Russian authors, although the genetic explanations provided by these authors are not necessarily adopted.

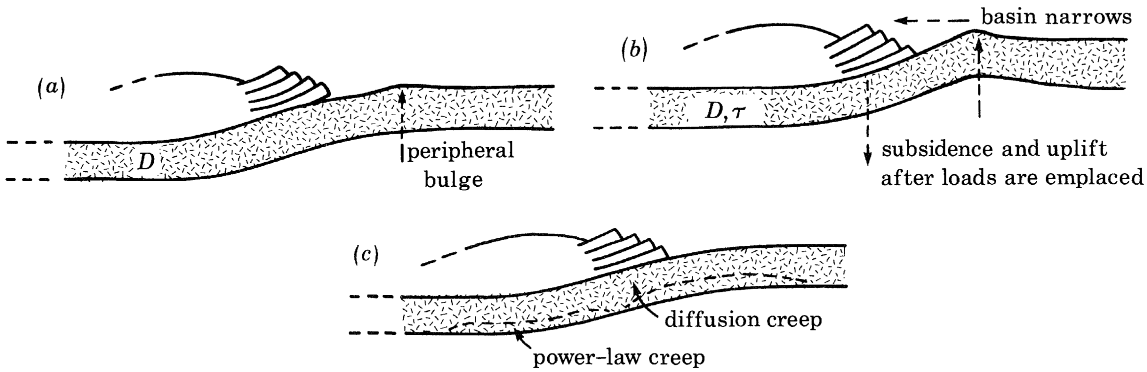

Figure 3. Diagrams illustrating flexure of the lithosphere beneath the fold–thrust belt, shown as a stack of thrust sheets. The foredeep is shown before sediment filling. In (*a*) the patterned region is that part of the lithosphere that appears to undergo elastic flexure. Its thickness is typically much less than the thickness of the thermal lithosphere. It is also assumed that the lithosphere is laterally uniform and can be characterized by its flexural rigidity, D. The peripheral bulge, which is typically a small percentage of the downwarp in amplitude, is exaggerated. It migrates to the right at a fixed distance ahead of the advancing thrust sheets. In (*b*) the patterned region is again laterally uniform but may be somewhat thicker than in (*a*). D is again the flexural rigidity and τ is the viscous relaxation time constant for the lithosphere treated as a whole. In this case the peripheral bulge tends to migrate to the left and to grow with time after a thrusting event. This leads to a coupled additional subsidence and uplift as the foredeep narrows. In (*c*) an attempt is made to outline the regions in which diffusion and power-law creep would occur were the lithosphere to have a microrheology that is the same as that determined from laboratory experiments on olivine. There will be some tendency for the basin to develop a peripheral bulge that grows and migrates to the left.

The scope of the discussion is limited to the depositional and erosional history of the basin that is directly controlled by events in the orogenic belt; that is, the flysch and molasse phases. Here, the concern is the response of the continental margin to the accretionary events associated with plate convergence and collision (Dewey & Bird 1970; Dickinson 1974) and the intraplate crustal shortening.

The working hypothesis is that foreland basins form at the site of downward-flexed lithosphere, and that this downward flexure is in response to passive loading by supra-lithospheric mass loads superimposed during formation of the fold–thrust belt. This hypothesis has more widespread application than to foreland basins only. Whenever there is lateral transfer of a rock mass over an adjacent part of the lithosphere that responds by flexure, a coupled trough is created in which sediments can accumulate. The form of this trough will depend on the timing and amount of the mass movements, the properties of the underlying lithosphere, the amount of sedimentary infill, and the role that other processes, for example subduction, under-thrusting and thermal cooling, play in warping the lithosphere at the site of the trough. The influence of each of these factors on the formation of foreland basins can be estimated from the tectonic and thermal environments in which foreland basins form.

The origin of tectonic subsidence of the foredeep by lithospheric flexure under the overthrust loads (figure 3) is analogous to the bending of a leaf spring. This, the simplest form of the model, assumes that the 'spring' is continuous and laterally uniform in its thermal and rheological properties. It is also assumed to be thermally mature, several lithospheric thermal time constants having elapsed since the last significant thermal perturbation. Such a model was shown to give a reasonably accurate account of the evolution of the Alberta Basin (Beaumont 1981) if it is assumed that the foredeep was continuously filled with sediments eroded from the adjacent Rocky Mountains and that eustatic sea level changes also occurred. It was, however, noted that the predicted average height of the Rockies was probably too large, although no quantitative estimates of palaeotopography exist. For this reason it is worthwhile considering, in a qualitative way, possible variations in the form of the tectonic subsidence, palaeotopography and basin development if the lithosphere were to have more complex properties. It is stressed that all the models proposed are fundamentally 'leaf-spring' models but that they include some second-order effects that may be detectable when other foreland basins are investigated. These complexities are briefly discussed to document the inevitable difficulties that will occur when trying to prove that any, some, or all are operative.

Lithospheric rheology: continuous uniform time-invariant elastic lithosphere

The continuous uniform time-invariant elastic lithosphere model is characterized by a lithosphere of constant flexural rigidity, D, overlying an inviscid fluid asthenosphere. The uniformity dictates that the distance between an applied load and the peripheral bulge (shown

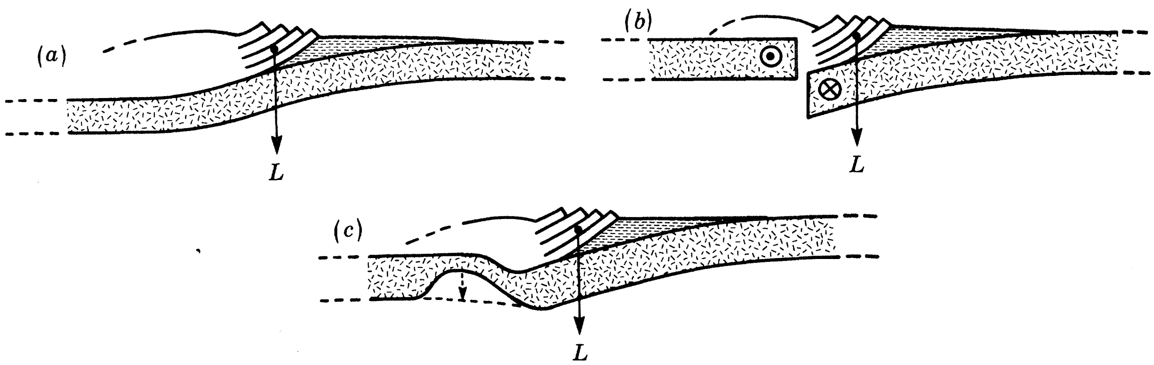

FIGURE 4. Diagrams illustrating the effects of lateral changes in properties of the lithosphere (random pattern) on the form of the foreland basin (dashed horizontal pattern), compared with a continuous uniform plate (*a*). The basin remains largely unaltered because the postulated strike-slip motions (*b*) and thermal thinning of lithosphere (*c*) occur within the orogen. The loading of the overthrusts is depicted as the cause of lithospheric flexure. The main effects of lateral changes that weaken the lithosphere will be to increase the amplitude of downwarping at the expense of reducing the width of the basin. The degree to which this occurs depends on the relative positioning of the weakened area and the fold–thrust belt.

in exaggerated form in figure 3*a*) is constant and can, in theory, be used to infer D and to provide estimates of effective lithospheric thickness. The additional growth of the basin when filled or partly filled by sediment (figure 4) must also be computed. This poses no problems if the amount of filling and sediment density distribution are known. Similarly, the effects of changes in the sea level, which lead to further onlap or offlap of the basin, can be predicted if the sea level changes are known. Pre-existing topography can also be accounted for if it is known.

We can already appreciate that the calculations are manageable for quite complex geometries but that the effects of inaccurately known quantities, such as sediment density, eustatic sea level changes and palaeotopography, will affect the model predictions. The position of the feather edge of the basin, the edge furthest from the thrust–fold belt, is very sensitive to these quantities. Consequently it should not be used to measure basin width for flexural rigidity estimates. A more accurate method is to compare overall basin shape, or the position of individual stratigraphic horizons, with model predictions. Even these require a reasonable estimate of the loading history in the fold–thrust belt. One generalization is, however, possible. If the fold–thrust belt advances basinward, then, in the absence of major sea level changes or high pre-existing topography, the feather edge of the basin will advance at the same rate. Consequently, the continuous uniform time-invariant elastic lithosphere model predicts a history of basin onlap. Furthermore, erosion of the fold–thrust belt cannot create offlap because distance from the load to the peripheral bulge is constant.

Lithospheric rheology: continuous uniform time-invariant viscoelastic lithosphere

This model casts the lithosphere as a uniform viscoelastic (Maxwell) layer overlying an inviscid fluid substratum (figure 3*b*). The elastic properties are again described by the flexural rigidity, D, but stress relaxation is permitted by viscous flow with a time constant τ. 'Time-invariant' implies that D and τ do not change with time; however, the lithospheric response to loading is time-dependent (Beaumont 1978). The response to loading is an immediate elastic flexure, the same as for the elastic models, followed by further progressive subsidence of the basin adjacent to the load and uplift and inward migration of the peripheral bulge. This, when combined with erosion of the uplifted basin strata, leads to a progressive narrowing of the basin and steepening of the dip on stratigraphic horizons. The relaxation induced by successive loads will be combined with the elastic and viscous response of subsequent and earlier loads. Therefore, the stratigraphy of the basin reflects a subtle interplay of the loading history with the viscoelastic relaxation of the lithosphere. In a general way viscous relaxation mitigates the tendency of the advancing fold–thrust belt to drive the feather edge of the basin before it. It also leads to the development of significant arches and erosional angular unconformities along the path of the inwardly migrating peripheral bulge if the amount of relaxation is large (Beaumont 1981).

When a loading phase is followed by erosion of the fold–thrust belt, the response to unloading of the 'leaf spring' is uplift of the basin. This is true for both elastic and viscoelastic models. The difference is that the immediate uplift of the basin for viscoelastic models in response to erosion has an immediate flexural wavelength that is longer than the width of the basin, simply because relaxation has had time to shrink the basin since the last loading phase. Instead of uplift and erosion that strips off stratigraphic layers much as they were deposited, as predicted by the elastic model, the viscoelastic response is therefore to truncate older stratigraphy, thereby producing a characteristic angular unconformity that exposes progressively older strata with increasing distance from the fold–thrust belt (Beaumont 1981). This is exactly as observed in the cross section of figure 1. Such an angular unconformity cannot be produced by any reasonable model that includes only loading of a continuous uniform time-invariant elastic lithosphere, although bevelling in response to an additional significant lowering of sea level can produce a similar effect. Angular unconformities do not, however, provide a conclusive case for viscoelasticity because other rheologies and lateral changes in lithospheric properties can also mimic these effects. Nevertheless, the viscoelastic model is the simplest model that produces an acceptable explanation of the Alberta Basin (Beaumont 1981).

Lithospheric rheology: inferences from laboratory experiments

Deformation diagrams for creep in olivine (see, for example, Ashby & Verrall 1977; Ranalli 1980) have been used by Beaumont (1979) to predict a first-order model of rheological zonation of the lithosphere under flexure. These models, although crude, include the effects of a uniform increase in temperature with depth in the lithosphere and a power law for the viscous creep relaxation of stress. It is unlikely that stresses induced by the distributed load of the fold–thrust belt and sediments alone will be sufficiently large to cause either brittle failure (cataclastic flow) or ductile (plastic) flow of the lithosphere if it was previously intact and without zones of local weakness. Therefore, zones of diffusion (linear) and dislocation (power law) creep similar to that shown diagrammatically in figure 3*c* can be expected. The terms diffusion and dislocation creep are used in the sense given by Beaumont (1979). Such a model is certainly more acceptable to the microrheologist than either of the previous two. The overall macro response of the lithosphere, which determines the development of the basin, is, however, likely to be very similar to the viscoelastic model. The only difference that might be detectable is that relaxation, which leads to tilting and narrowing of the basin, will rapidly decrease once the hotter regions of the lithosphere have undergone stress relaxation. Consequently, the response tends to that of a lithosphere that is apparently elastic long after loads have been added or removed. This is certainly true if thermally activated diffusion creep dominates the overall response (R. Courtney, personal communication). Were this not so, such simple rheologies as elasticity and viscoelasticity would long ago have been shown to be unacceptable. We believe that the response of the lithosphere to loading is primarily a function of its temperature distribution and that the effect of flexurally induced stress in promoting power-law creep is secondary. The geologist who is interested in basin analysis should therefore concern himself with the two simple models of lithospheric flexure and the underlying question of the thermal state of the lithosphere.

Lateral variations in lithospheric properties

There is good reason to believe that the lithosphere is not continuous and uniform beneath the whole of an orogen, as assumed in the simple models (figures 3 and 4*a*). For example the plate tectonic model explains the assembly of an orogen by subduction followed by either or both of collision and transform strike-slip motions, both of which produce lateral inhomogeneities in lithospheric structure and suggest models like those shown in figure 4*b*, *c*. Strike-slip plate motions with a component of convergence may best be represented by a broken plate model. Corresponding thermal attenuation beneath the core zone of an orogen would result in partial or total decoupling of this region. In both cases the amount of decoupling may vary with time in response to cooling of the lithosphere after the cessation of subduction and completion of suturing.

Simple elastic models in which variations in lithospheric thickness are included in the hinterland of the orogen indicate that such variations cause only second-order changes in the form of the foreland basin. The extremes (figure 4*a*, *b*) are represented by an end-loaded discontinuous but buoyantly supported beam and a point-loaded continuous buoyantly supported beam (Walcott 1970). The former predicts a somewhat narrower and deeper foreland basin than the latter for the same flexural rigidity. This reduces the predicted topography of the fold–thrust belt, as would the thermally thinned plate model (figure 4*c*), whose behaviour will lie between the two extremes. A basin of equivalent width to that predicted by a continuous uniform plate model is obtained by increasing the flexural rigidity.

It is important to ask whether the effect of lateral variations in lithospheric properties in an elastic model can mimic the effect of stress relaxation as shown by the viscoelastic model. The fundamental property of the viscoelastic model is that the plate appears weaker, thinner and to have decreasing flexural wavelength with increasing time after loading. Successive loads all show the same pattern of evolution no matter when they were added. Similar effects cannot occur in elastic models with lateral variations in properties if these lateral variations do not change with time. That is, a broken plate model will have approximately constant properties as long as the break remains. To appear weaker the plate must break, or be thermally thinned, as the fold–thrust belt evolves. To produce an angular unconformity in the basin the plate must be rewelded, or cool, before uplift and erosion of the fold–thrust belt. Although complex, such events cannot be ruled out *a priori*. Therefore it is possible, but unlikely, that time-varying lateral changes in lithospheric properties could mimic stress relaxation.

The simple concept of the fold–thrust belt advancing further onto a broken plate (figure 4*b*), so that the break recedes ever further into the hinterland until the plate appears continuous and uniform, produces an exactly opposite effect, that of an apparently increasingly strong plate when viewed from the perspective of basin stratigraphy. Post-tectonic erosion cannot produce an unconformity of the type observed because the basin strata will be stripped in much the same way as deposited.

In summary, it can be seen that lateral changes in lithospheric properties do produce second-order changes in the form of the foreland basin but that a detailed analysis, including a palinspastic reconstruction of the fold–thrust belt, will be necessary before individual lithospheric features can be demonstrated for a particular foreland basin.

Thermal maturity of the lithosphere as a key to the size of foreland basins

It is explained in the section on rifted continental margins that the flexural properties of the lithosphere are primarily determined by its internal temperature distribution. Hot, thin, thermally young lithosphere has a small flexural rigidity and wavelength, whereas old, cold, thermally mature lithosphere has a correspondingly larger flexural rigidity and wavelength. This view is confirmed by studies of the flexural response of oceanic lithosphere as it increases in age (Watts 1978). For this reason it is to be expected that, if the leaf-spring model of foreland basins is correct, the size of a foreland basin will be directly related to the thermal age of the underlying lithosphere at the time of basin formation. The thermal age is the elapsed time since the last major thermal event that thinned the lithosphere. The difference between basins that develop on thermally young and old lithosphere is illustrated in figure 5. In the limit of total thinning of the lithosphere, local, or Airy, isostatic equilibrium prevails and no foreland basin is generated.

If significant lithospheric cooling occurs during the formation of the basin (figure 5*a*) the flexural rigidity will increase, as will the width of the basin. This will accelerate the advance of the feather edge of the basin ahead of the fold–thrust belt, leading to a pattern of progressive onlap. Erosion of the fold–thrust belt and the basin can then, in theory, produce an angular unconformity because the lithosphere achieves its maximum flexural wavelength during the terminal erosion phase. This represents an additional way in which angular unconformities can arise. When invoking this model, however, evidence for a thermal event, which thins the lithosphere, is necessary.

The main point to be made in this section is that we should not expect all foreland basins to look like the Alberta Basin archetype. The reason that this basin is so pronounced is that it formed on cratonic lithosphere of the Canadian Shield that had not been subject to a major tectonic or thermal event for at least 500 Ma and probably longer. The tectonic setting of a foreland basin is therefore seen to be a critical factor in determining its final extent.

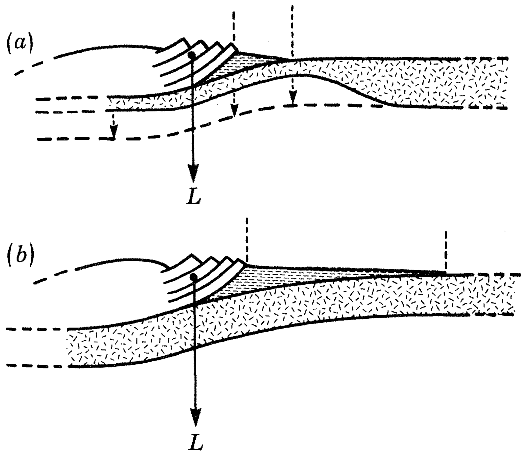

FIGURE 5. Diagrams illustrating the difference between foreland basins that form on thermally young (*a*) and old (*b*) margins. Young, hot lithosphere (less than about 30 Ma old) is flexurally weak; therefore the basin will be narrow but deep. The converse is true for thermally old (more than about 300 Ma) lithosphere. If the lithosphere cools appreciably while the basin forms, (*a*), the flexural rigidity and corresponding width of the basin will increase.

Lithospheric phase changes and topography of the fold–thrust belt

Beaumont (1981) shows the final configuration of a series of models for the development of the Alberta Basin. The evolution of the preferred model for six of the eleven time steps is shown in figure 6. The topography of the fold–thrust belt at the end of the Laramide Orogeny, *ca.* 35 Ma ago in the model, is certainly excessive in suggesting a range of mountains larger than any that currently exist, including the Himalayas. The need for such a large load is based on estimates of depth of burial of coals and shales beneath the now largely eroded Paskapoo formation, which was deposited during the Laramide Orogeny. The evidence in favour of the model load is therefore not as conclusive as it would be were no erosion to have occurred in the last *ca.* 35 Ma. Nevertheless, even minimum loads still give apparently excessive topography and there remains a need to reduce topography while preserving the load.

The topography shown diagrammatically as h_0 (figure 7*a*) would be reduced to h_1 (figure 7*b*) were the lithosphere broken or thinned or the thickness of the load reduced to reflect a load density greater than the 2400 kg m^{-3} used, but estimates suggest that such reductions would still be insufficient. The problem is not peculiar to the Canadian Cordillera. Oxburgh (1968) and Hunziker (1974), among others, have noted that for the Alpine chain, where thrust sheets tens of kilometres thick have been overthrust onto the continent, the initial topography must have been low because relatively little erosion occurred for *ca.* 30 Ma after emplacement. Richardson & England (1979) present an attractive model to explain this reduced topography by invoking amphibolite or granulite facies to eclogite facies phase transitions in the lower crust in response to the increased pressure beneath the thick overthrusts (figure 7*c*). They estimate an accompanying density change from 2850 to 3150 kg m^{-3} with a corresponding volume decrease of *ca.* 10%. Such a change within a 20 km thick region of the crust, in addition to an increase in the fold–thrust belt load density to 2600 kg m^{-3}, to include possible crystalline

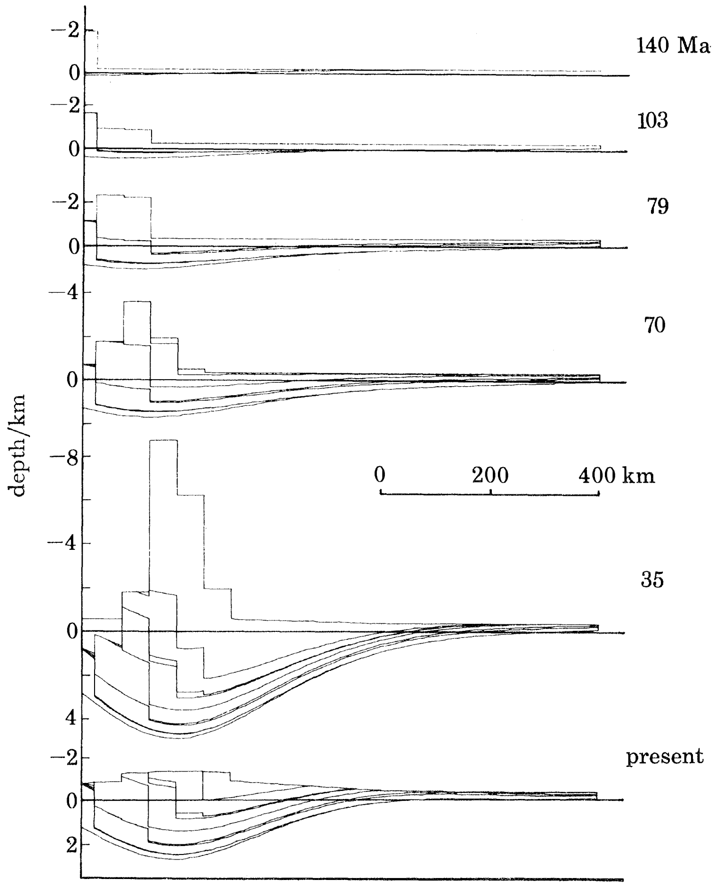

FIGURE 6. Six of eleven steps in a model of the evolution of the Alberta foreland basin (Beaumont 1981). The dates are model ages. The main points to note are: the lateral migration of the fold–thrust belt, here averaged over 50 km wide strips; that sea level changes, which widen the basin, have been included; the excessive height of the mountains at 35 Ma age; and the truncation of the basin strata during erosion between 35 Ma ago and the present. The datum is present sea level.

involvement, would reduce the average Laramide topography of the Rockies to *ca.* 2 km. The phase change is, however, temporary because relaxation of the temperature profile to the stable geotherm inverts the phase transition. This heating may take approximately 20 Ma. The time constant could vary by a factor two depending on the erosion rate, the thickness of the over-thrust sheet, and the gradient of the stable geotherm. This result suggests that the final configuration of the model (figure 6) is probably correct because thrusting ceased at the end of the Eocene Laramide Orogeny and sufficient time has elapsed for up to 10 km of erosion as proposed by the model and a return to a stable geothermal gradient.

The phase transition model (figure 7*c*) therefore provides a mechanism whereby topography can be reduced but in a temporary way. Phase transitions have often been invoked to rectify problems with models without justifying why the phase transition should occur. In this case, however, the model has a firmer basis. The forward phase change is a direct consequence of the pressure change induced by the loads, and the inversion is in response to a return to the normal geothermal gradient, both of which are predicted by equilibrium thermodynamics. No thermal pulse or inversion from a pre-existing metastable facies need be hypothesized.

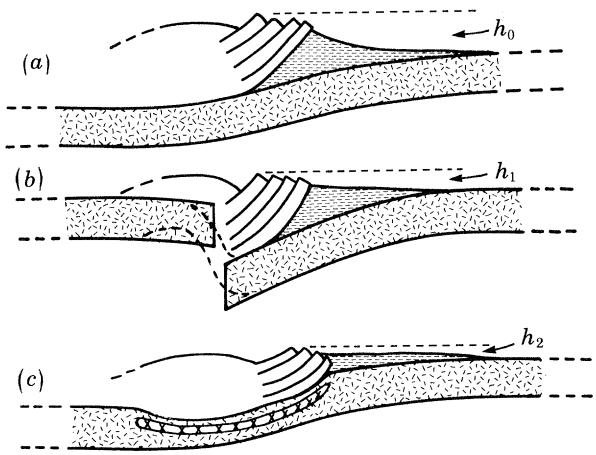

FIGURE 7. Diagrams illustrating reduction of the topography of the fold–thrust belt in response to a thinner or decoupled lithosphere, or as a result of volume changes in the crust due to phase changes (X-patterned area). (*a*) Continuous uniform lithosphere; (*b*) broken or thinned lithosphere; (*c*) phase change in lithosphere under fold–thrust belt.

It is important to note that the model (figure 6) is conservative in its load estimates. Larger loads are probably necessary to obtain the correct thickness of palinspastically restored stratigraphic units of the basin (R. A. Price, personal communication). These arguments imply the need for inclusion of phase changes in the leaf-spring model.

LITHOSPHERIC EXTENSION AT RIFTED CONTINENTAL MARGINS

Most continental margins created during rifting of a continental lithospheric plate evolve into deep sedimentary basins occupying the transitional region between oceans and continents. These rifted margins, sometimes called Atlantic-type or passive margins, are the subject of intense research interest because they may contain significant hydrocarbon resources. Of the possible mechanisms responsible for their formation and subsequent evolution, the proposal that there is significant horizontal extension during rifting which thins both the crust and subcrustal lithosphere can explain most of the first-order properties of the margins and can be tested by modelling. These properties include: the thinning of the crust landward of the ocean–continent boundary (Sheridan *et al.* 1979; Montadert *et al.* 1979; Keen & Hyndman 1979), subsidence (Sleep 1971; Watts & Steckler 1979; Watts & Ryan 1976; Keen 1979; Royden & Keen 1980), and listric faulting of the upper crust (Montadert *et al.* 1979).

Keen *et al.* (1981*a*, *b*) and Beaumont *et al.* (1982) have recently explored the consequences of extensional models of rifting with particular reference to the rifting process, the post-rift cooling, thermal contraction and subsidence of the lithosphere, and the amplification of this subsidence by sediment and water loads. Three kinematic models of rifting, based primarily on extension and thinning of the lithosphere, were examined in detail to determine the degree to which second-order effects during the extension and rifting process dictate the evolution of the margin. To first order, details of the rifting model were found to be unimportant. Therefore, for the purposes of this paper we shall initially confine the discussion to thermomechanical models that are a direct consequence of the uniform extension model proposed by McKenzie (1978). The amount of lithospheric extension, $\beta(x)$, is assumed to be uniform with depth and to increase progressively with position, x, towards the point at which oceanic lithosphere is generated. At that point the process of tensional stress release changes from stretching of the continental lithosphere to rupture and accretion of oceanic lithosphere.

The consequences of extension are conceptually simple. The crust and subcrustal lithosphere are initially thinned by $\beta(x)$ and the space created is filled by the passive upwelling of hot asthenosphere. Extension will therefore be accompanied by isostatic elevation changes due to the replacement of crust by denser mantle lithosphere and density changes due to heating and thermal expansion. This normally results in rapid subsidence during extension. Longer-term subsidence will occur as the extended region of the margin cools by conduction and undergoes thermal contraction after extension ceases. This subsidence is analogous to that of oceanic lithosphere as it migrates from an oceanic ridge (Parsons & Sclater 1977). Extension is assumed (1) to occur on a sufficiently short geological timescale that it can be modelled as an instantaneous process without significant error (Jarvis & McKenzie 1980) and (2) to produce sufficiently small horizontal thermal gradients that horizontal heat transport can be ignored in comparison with vertical conductive cooling (Beaumont *et al.* 1982). Local isostatic equilibrium is assumed during the rifting process.

Thermal and rheological evolution of the lithosphere

An immediate consequence of the proposed rifting model is that, during the evolution of the margin, the lithosphere will undergo changes in thermal and rheological properties in both time and space as it cools. Unlike the lithosphere beneath foreland basins, no simplifying model of a uniform continuous time-invariant elastic or viscoelastic lithosphere can be assumed. This is particularly true because rheology is a strong function of temperature, as we have previously argued. A model (figure 8) for the evolution of the lithosphere beneath the Nova Scotian margin (figure 2) demonstrates how radically the lithosphere is thinned from an original thickness of 125 km during extension, and how, over the next 185 Ma after rifting, conductive cooling restores most of the original thermal regime.

For the purposes of modelling the loading response to sediment and water it is assumed that the lithosphere can be subdivided into two regions: one an elastic core in which no stress relaxation occurs; the other a hotter underlying region that releases stress rapidly in comparison with the time for evolution of the margin as a whole. This lower region is therefore treated as an inviscid fluid that is undistinguished from the asthenosphere. The boundary between the elastic and fluid regions is taken to be an isotherm, termed the relaxation isotherm, the position of which varies in space and time as the margin evolves but can be predicted by considering both the thermal and mechanical evolution of the margin.

The post-rifting history of the margin is therefore traced by a thermomechanical model in which subsidence and temperature distribution are the result of conductive cooling and in which the isostatic response to loading by sediment and water is found by assuming that those loads are supported by this elastic layer underlain by an inviscid fluid. The model is coupled in that both the space available for sediments and water and the thickness of the elastic layer, termed the rheological lithosphere, are determined by the temperature distribution. The calculations are made by using finite difference and finite element numerical models for the thermal and mechanical aspects of the problem respectively (Beaumont *et al.* 1982).

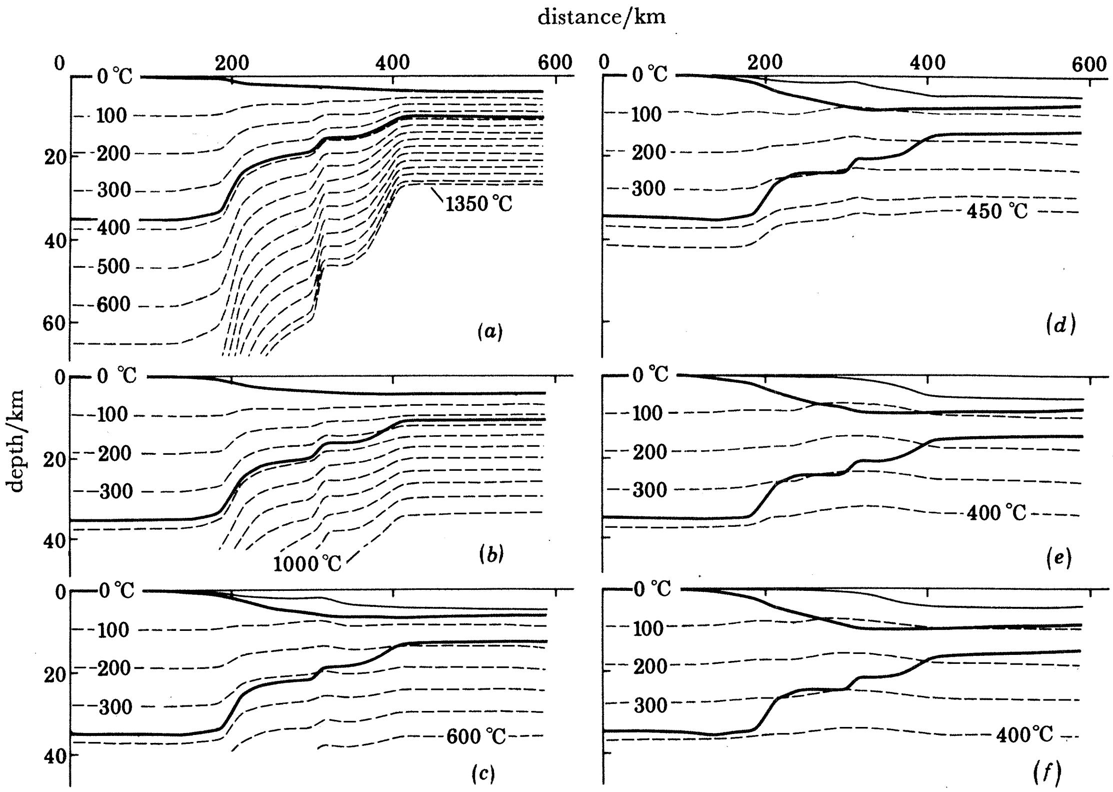

FIGURE 8. Thermal evolution of a uniform extension model of the Nova Scotian margin with a relaxation isotherm of 250 °C. (*a*) 185 Ma B.P., rifting; (*b*) 175 Ma B.P.; (*c*) 135 Ma B.P., end of Jurassic; (*d*) 100 Ma B.P., end of early Cretaceous; (*e*) 60 Ma B.P., end of late Cretaceous; (*f*) present. The lithosphere was 125 km thick before extension. Its initial thinning and that of the crust can be seen from the uplift of the 1350 °C isotherm (broken line in (*a*)) and the Moho (lower continuous bold line in (*a*)). The upper continuous bold line is the sediment–basement interface, and the continuous fine line is the sediment–water interface. Note that within *ca*. 63 Ma (the lithospheric thermal time constant for this model) after rifting, the thermal régime is largely restored to its initial state. Upwarps in the isotherms that are predicted for the present are the result of thermal blanketing by the low thermal conductivity sediments. The rheological lithosphere is that region between the sediment–basement interface and the 250 °C isotherm.

The simplicity of the rheology employed needs justification particularly by comparison with the possible rheologies suggested for the foreland basin models. Our view is that during the evolution of rifted margins, lithospheric rheology is so strongly dominated by cooling effects that it will be difficult to detect any properties that can be directly ascribed to viscous or viscoelastic relaxation. Evidence of rapid stress relaxation in the lower lithosphere will probably not be preserved in the short-term sediment record. Such relaxation is, however, allowed to occur instantaneously in the model by choosing a relaxation isotherm less than the temperature of the base of the thermal lithosphere. The same approach is implicit in the elastic lithosphere models for foreland basins because the lithospheric thickness so determined is 50–70 % less than that estimated for the thermal lithosphere. Viscous flow at depth beneath continental margins will be further inhibited by cooling that 'freezes in' pre-existing stress before relaxation is complete. Emphasis is therefore placed on the spatial and temporal variations of the elastic region at the expense of more complex rheologies. This approach can also be justified on a heuristic basis by comparison with olivine microrheology (Beaumont *et al.* 1982). If simple models of this type fail to predict correctly the evolution of rifted margins, the next step will be to introduce a rheology where the temperature dependence of viscosity is explicitly included.

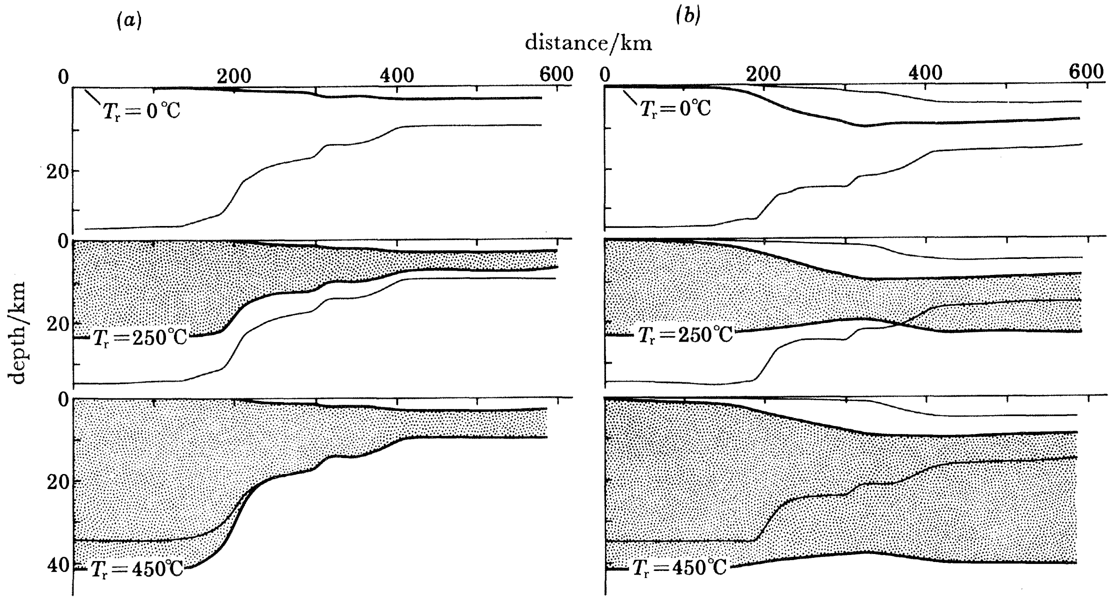

FIGURE 9. Evolution of the rheological lithosphere (stippled area) from rifting at 185 Ma B.P. (*a*) to the present (*b*) for the Nova Scotian margin cross section, illustrating the change in thickness as a function of the relaxation isotherm, T_r = 0 °C, 250 °C, 450 °C. The thin line at depth is the position of the Moho. Note that for T_r = 0 °C, local 'Airy' isostatic equilibrium exists and the rheological lithosphere has zero thickness. For finite T_r the thickness of the rheological lithosphere remains constant in the unextended region and thickens with time beneath the oceanic region. The thinning beneath the shelf edge results from upwarp of the isotherms beneath that region of the basin and depression of the surface beneath the thick sediments.

By comparison with foreland basins it is seen that the tectonic subsidence of rifted margins is not caused by supracrustal loads. Instead, rapid subsidence results from isostatic adjustment in response to what is effectively the replacement of continental crust by subcrustal lithosphere during extension. The longer-term subsidence results from thermal contraction during cooling, which re-establishes a stable geothermal gradient. The overall tectonic subsidence can be predicted if the extension $\beta(x)$ can be estimated. One method is to assume that the continental crust was uniformly thick before extension and to estimate $\beta(x)$ by measuring present crustal thickness in the belief that crustal thickness changes are solely caused by stretching. This will probably be shown to be an oversimplification but must be tested against observations.

In the same way that loading by sediments and water in foreland basins results in their lateral growth, equivalent loading of rifted margins also causes flexure of the underlying lithosphere. However, the flexural contribution to basin growth depends on the timing of loading because it is only later in the evolution of the basin that the lithosphere achieves sufficient strength to enlarge the basin significantly. This is the same thermal maturity property that determines the size of the foreland basins. In this case, however, the properties of unextended lithosphere on the continental side of a margin do not change appreciably with time (figure 9).

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Flexure of this region is primarily determined by the degree of decoupling from the loaded part of extended margin. This decoupling decreases as the margin evolves but does not necessarily result in onlap of the feather edge of the basin (figure 2), particularly if the locus of sedimentation progrades seaward as the basin evolves.

Alternative lithospheric models

In an earlier model of isostasy and flexure at continental margins, Walcott (1972) demonstrated the form that flexure and gravity anomalies would take were a uniform continuous time-invariant elastic lithosphere of assumed flexural rigidity 10^{23} N m. The results suggest rather too much flexure and, as Walcott himself notes, there are good reasons for considering variations in flexural rigidity across the margin. More recently Karner & Watts (1982) have extended the analogy of the isostatic response of cooling oceanic lithosphere to rifted margins. They assume the margin to be laterally uniform and to have a lithospheric thickness that is constant for each margin studied. The results indicate that the lithosphere becomes increasingly thick for successively older margins, but does not thicken as rapidly as oceanic lithosphere. Karner & Watts (1982) interpret this result to be evidence for a continuous increase in thickness of the elastic region of the lithosphere with the time after rifting in an environment of continuous sedimentation. That such an analysis produces reasonable results with respect to isostatic response functions may be fortuitous, for, as Beaumont *et al.* (1982) show, a large part of the gravity anomaly for young margins results from transient thermal expansion that is not related to flexure of the lithosphere when loaded. This thermal component is not included in the analysis by Karner & Watts (1982).

Uplift of the adjacent continent

The deep structure of rifted margin basins remains poorly known, but the shallower structure adjacent to the continent has in many cases been drilled to basement so that the structure, lithology and age of the sedimentary sequence are well defined. Evidence from these regions and from contemporary rift zones provides a more detailed, if incomplete, test of the extension model, one that the simplest uniform extension version of the model apparently fails because no peripheral uplift to the rift zone is predicted for reasonable initial crustal thicknesses (figure 10*a*). Uplift is certainly a feature of modern rifts and young continental margins, and there is evidence (Royden & Keen 1980) that flanking highs 1–2 km high and 50–100 km wide existed at some stage during the evolution of older margins.

At least two explanations, those of lateral heat flow and depth-dependent (non-uniform) extension, have been proposed, to which we add a third, that of regional, as opposed to local, isostatic adjustment during rifting. All three variations (figure 10) cause only second-order changes to the predictions of the basic model; nevertheless, the differences among them may be sufficient that they can be tested by comparison with stratigraphy landward of the hinge line.

The depth-dependent extension model (Sclater *et al.* 1980; Royden & Keen 1980; Beaumont *et al.* 1982) predicts uplift as a result of increased extension, δ, in the lower lithosphere with respect to that, β, in the upper lithosphere. This is shown in figure 10*b*, which diagrammatically places the boundary between the two regions at the Moho, but this is not necessarily so. A more probable decoupling horizon is that corresponding to the brittle–ductile transition, which may also coincide with the level at which listric faults sole. Realistic uplifts can be achieved in this manner when $\beta \approx 1$ and δ is large. The existence of depth-dependent extension is difficult to prove beneath old margins because cooling has largely restored the lower lithosphere to its original state. The signature of the model (figure 11) is one in which uplift is concurrent with extension, depicted as an instantaneous process, and decays as the lithosphere cools and contracts. The sedimentary record would show this as an erosional unconformity on uplift, then a non-depositional interval while the high remained, followed by sediment onlap as subsidence proceeded. In the absence of erosion the original base level will be restored on cooling, for regions where only subcrustal lithosphere was thinned. Depth-dependent extension is capable of producing uplift of 1–2 km, depending on the chosen values of δ and β and on the depth of decoupling (Royden & Keen 1980).

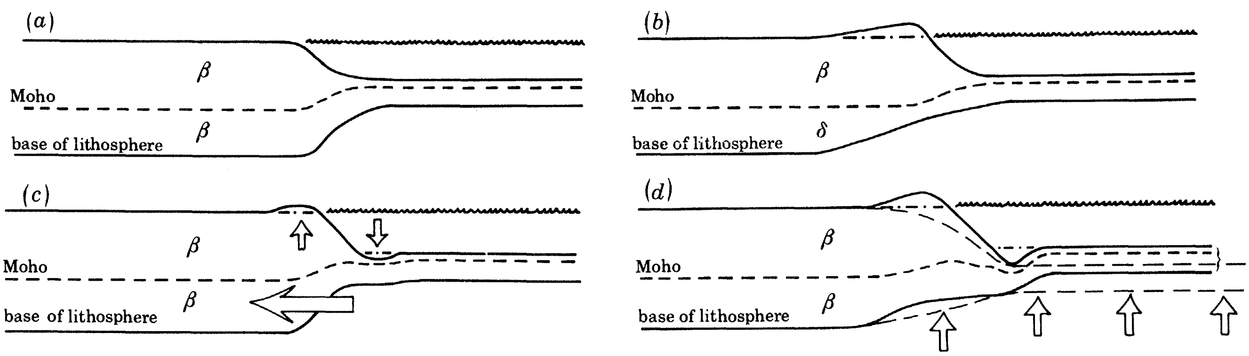

FIGURE 10. Diagrams illustrating the uniform extension model (*a*) and the three proposed models for uplift of the continent during formation of rift margin basins: (*b*) depth-dependent extension and local isostatic equilibrium; (*c*) lateral conduction of heat and local isostatic equilibrium; (*d*) uniform extension and regional isostatic equilibrium.

A second mechanism for uplift is that of lateral heat flux, a second-order effect ignored in the simple rifting models. Where the amount of extension changes rapidly across the margin, lateral temperature gradients might be sufficient to cause significant lateral heat flux. This causes enhanced cooling, contraction and subsidence of the highly extended region and heating, expansion and uplift of the unextended region (figure 10*c*). The sedimentary record would show evidence of an uplift that grew after rifting, achieving its climax after 60 Ma, and then decaying by 120–150 Ma (Beaumont *et al.* 1982) (figure 11). Deposition can, in theory, be delayed for a much longer interval than that predicted by the depth-dependent extension model. The model does, however, appear incapable of generating uplifts that exceed a few hundred metres (Beaumont *et al.* 1982; Steckler 1981).

The model that appeals to regional isostatic adjustment during rifting (figure 10*d*) extends the concept of lithospheric flexure to the rift phase. If the lithosphere retains some structural integrity during extension, lateral coupling will remain a factor in determining the final regional isostatic balance of the rifted margin. If we consider the necking of the lithosphere under tension, but in the absence of gravitational or buoyancy forces, the shape of the extended region, what we term the 'free extension shape', may be similar to that shown by the long broken lines in figure 10*d*. The depth about which necking was centred is determined by the rheology of the lithosphere and cannot be predicted without an analysis that includes the dynamics of extension and realistic rheologies, properties that are currently not well known. It is, however, highly unlikely that the necking will be symmetrical about the mid-depth of the lithosphere, or that the free extension shape will exactly coincide with that of a margin in local isostatic equilibrium. For example, figure 10*d* suggests that the free extension upper surface is at a greater depth than it would be if local isostatic conditions prevailed.

Application of buoyancy and gravitational forces to the free extension configuration will tend to raise the surface to the position of isostatic equilibrium. This is easily accommodated in regions where the lithosphere has little lateral strength. At the margin, however, where the rheological and thermal lithospheres thicken, this uplift will be flexurally resisted by the unextended region, which is already in equilibrium. Some flexural upwarp of the unextended region will occur. Correspondingly, flexural resistance to the buoyancy forces ensures that part

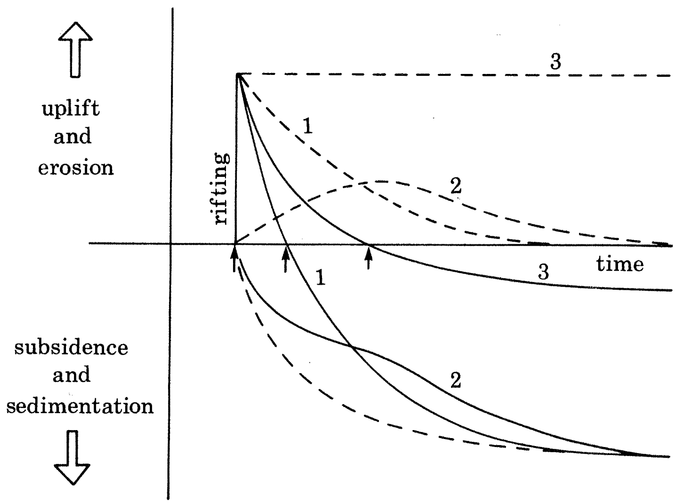

FIGURE 11. Illustrations of the form of uplift and subsidence predicted by the proposed rift models (figure 10) for a point landward of the hinge line where $\beta = 1$. Models 1 (the depth-dependent extension model) and 3 (regional isostatic adjustment during rifting) predict immediate uplift on rifting. This uplift decays for model 1 but remains for model 3 (short broken lines). The effect of lateral heat transport (model 2) is to create a smaller uplift which grows and then decays (short broken line). The solid lines illustrate the superimposed effects when flexural downwarp landward of the hinge line (long broken line) is also included. Note that models 1 and 3 predict uplift followed by subsidence, but that the subsidence of model 3 is much less. Model 2 predicts subsidence but a period of relative quiescence may exist as the thermal uplift grows. The form of the curves will vary with position across the uplifted region and with relative intensity of the competing effects.

of the thinned lithosphere remains relatively depressed. The flexural unwarp–downwarp pair balance to maintain a regional isostatic balance. The sense of the pair would be reversed were the upper surface of the free extension shape above the local isostatic depth. That this is not so may provide some insight into the rheology of lithospheric necking. By analogy with the more familiar concept of flexural downwarp, upwarps 1–2 km high and 50–100 km wide should not require excessive flexural stresses and are within the bounds of flexural rigidity estimates for the lithosphere. The signature of such flexural uplifts is that they occur concurrently with lithospheric extension and that, in the absence of erosion, they do not decay (figure 11). The sedimentary record near the peripheral bulge would be one of uplift and erosion.

One additional factor, that of flexural downwarp landward of the hinge line by sediment and water loads in the subsided areas, must also be superimposed on the simple vertical movement patterns shown as short broken lines in figure 11. In all cases this effect will accelerate onlap of the basin. The solid lines in figure 11 diagrammatically illustrate the combined effect for a point where $\beta = 1$. The long dashed line is the hypothetical flexural downwarp contribution.

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The important point to note is that each of the three models predicts a sufficiently distinct pattern of non-deposition, deposition and final depth that there is hope that their relative importance can be distinguished by a careful study of the stratigraphy of the shallow regions of rift margin basins. The unknown effects of erosion and sea level change inevitably militate against this optimism.

A comparison of basin characteristics

Foreland and rifted margin basins are created in predominantly compressional and tensional environments, respectively. The key to their difference is, as previously explained, the source of tectonic subsidence, supracrustal loads versus mass replacement at depth, 'a thickening versus thinning lithosphere' during their early history, and the lithospheric response to loading. By the very nature of changes that occur in the lithosphere during extension, the lithosphere will on average appear weaker beneath rifted margins than beneath foreland basins. The properties of the two basin types are therefore to a large extent mutually exclusive.

Where it is not clear what caused a basin to subside, some of the following characteristics may prove to be useful diagnostic features for simple cases. Other features are quite similar.

1. *Subsidence history*. Rifted margins are often characterized by an initial phase of rapid subsidence, which is concurrent with extension, followed by a thermal contraction phase in which subsidence is proportional to $t^{\frac{1}{2}}$ or decays exponentially. Conversely, foreland basins exhibit episodes of very rapid subsidence followed by quiescent periods. Rapid subsidence signals the emplacement and stacking of thrust sheets at a rate that outstrips sedimentation. The basin deepens in a pulse-like manner, only to be filled as erosion catches up. Commonly, several such pulses occur, indicating more than one discrete thrusting event. Foreland basins are also characterized by episodic and terminal uplifts and erosion that appear as major unconformities.

2. *Basin shape*. Both basin types are asymmetric. They differ in that foreland basins seldom exceed a depth of *ca*. 6 km, have a characteristic convex upward flexural shape, and are bordered by a fold–thrust belt, whereas rifted margin basins can achieve depths of *ca*. 16 km, show little evidence of flexure, except landward of the hinge line, and are not bordered by a fold–thrust belt.

3. *Topography of the basin basement*. The simplest of models suggest that, when a region has been one of subdued topography before the formation of a foreland basin, the topography would undergo little deformation save that of regional flexural warping. Conversely, the process of lithospheric extension during rifting creates rugged relief by near-surface block faulting and tilting, which will be preserved as basement topography beneath the basin. Whether such listric block faulting (Proffett 1977; Montadert *et al.* 1979; Le Pichon & Sibuet 1981; Le Pichon *et al.*, this symposium) can be used to estimate extension remains controversial. Its existence, or that of a horst and graben régime, are however, first-order indicators of a tensional environment in which brittle failure of the upper crust occurred. In contrast, brittle faulting would not be expected beneath flexurally downwarped, but unextended, regions of either rift margins or foreland basins.

A pre-existing faulted upper crust can, however, be inherited. Although bevelled by erosion before basin formation, it can be reactivated during flexure, which is controlled by the underlying intact part of the lithosphere. This, together with detachments that propagate within both the older sediments and crystalline basement beneath the basin during thrusting, can lead to an irregular and faulted basement beneath foreland basins. Such a response appears more common than that exemplified by the Alberta foreland basin, where pre-Jurassic sediments remained largely unfaulted during flexure. Moreover, there is no reason why relative motion of these faulted blocks beneath either type of basin should not continue during basin evolution, thereby generating growth faults that propagate into the overlying basin sediments.

4. *Crustal structure.* If it is assumed that crustal structure is not a relic of previous events, the models predict that the crust beneath rifted margins will be attenuated whereas that beneath foreland basins remain substantially uniform. Whether this is a useful guide to the mode of basin formation depends on the degree to which changes associated with basin formation overprint and obliterate inherited structure.

5. *Gravity anomalies.* Flexural downwarp is a major feature of foreland basins, they are therefore far from local isostatic equilibrium. This is reflected in the large (*ca.* -100 mGal; -10^{-4} m s^{-2}) free air gravity anomalies over the Alberta basin. Rifted margins are seldom characterized by such long wavelength anomalies because they are closer to a state of local isostatic equilibrium.

6. *Sediment lithology.* Coarse-grained clastic sediments, which form the molasse sequences of foreland basins, are very characteristic. Interleaving finer-grained flysch sequences mark phases of rapid subsidence and basin deepening. Black shale and coal sequences are also characteristic of the foreland environment. Other organically related sediments are relatively rare, either because foreland basins fail to achieve sufficient stability for major reef growth, for example, or the influx of clastics gives an unsuitable environment. Evaporites are also relatively less common than in rifted margins, where they often occur as early post-rift deposits. This is because foreland basins spend little time in a starved shallow marine state.

Coarse-grained clastics are also a feature of the rift phase of rifted margins reflecting topographic elevation changes and block faulting. These neighbouring topographic highs decay as the basin evolves. Consequently, there is normally a progressive change to finer-grained sediments or to a deltaic sand shale environment when there is a large influx of fluvial sediment. In suitable environments the later phases of slower subsidence are often marked by reef carbonates. Deltas can also be built into foreland basins, particularly when rapid deepening of the basin is followed by rapid erosion of the fold–thrust belt. Palaeocurrent indicators may in this case provide evidence of the existence of a distal feather edge to the basin at the peripheral uplift, though the effect of a conjugate margin in young rift basins may be very similar.

7. *Thermal history of sediments.* If it is assumed that conduction dominates heat transport in both types of sedimentary basin, sediments deposited in rift basins during the early stages of their evolution will be subject to a significantly enhanced non-equilibrium geothermal gradient. In contrast, the geothermal gradient in foreland basins that form on thermally mature lithosphere will be normal throughout the evolution of the basin. The enhanced geothermal gradient in rift margin basins is approximately proportional to the amount of extension, and decays with a time constant that is proportional to the unperturbed thickness of the thermal lithosphere (Royden & Keen 1980; Royden *et al.* 1980; Beaumont *et al.* 1982). Because thermal metamorphism is directly related to temperature history, all other things being equal, equivalent levels of thermal maturation will be reached at smaller depths in rift margin basins than in foreland basins. The effect of erosion, which reduces the depth of burial and the temperature of individual strata, must also be taken into account (Beaumont 1981), as must the contributions to the geothermal gradient from radioactive heat generation in the crust and sediments (Beaumont *et al.* 1982). Some caution is needed in applying these basic ideas because

temperature gradients in both types of basin can be perturbed by local conditions, for example by the effect of salt diapirs of high thermal conductivity, or on a more regional scale by heat transport by fluid flow.

Superposition of basins

In previous sections, foreland and rift margin basins have been considered to occupy very different spatial and tectonic environments. While this is an appropriate way to discuss type examples, the cycle of opening and closing ocean basins and accompanying interactions at plate margins leads to the interchange of tensional and compressional environments, which can,

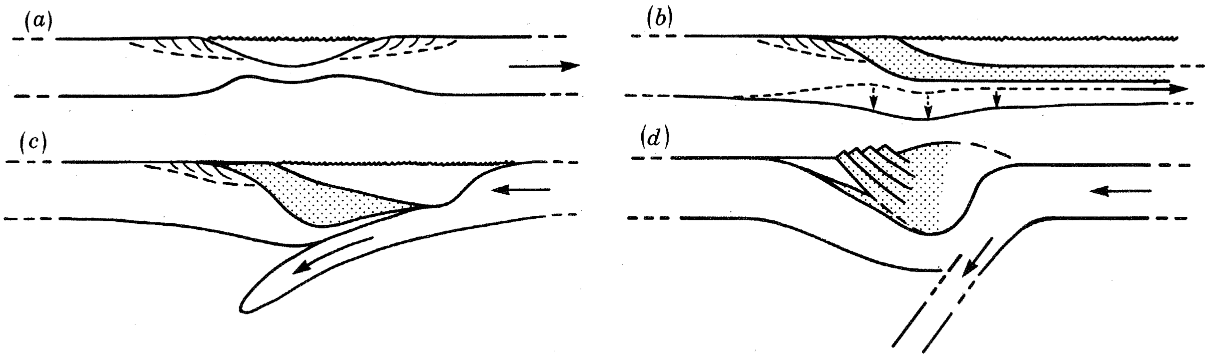

Figure 12. Diagrams illustrating the way in which a foreland basin can be superimposed on a rifted margin basin as a result of ocean closure. (*a*) Phase 1: rifting and initial subsidence. (*b*) Phase 2: drifting and thermal subsidence. (*c*) Phase 3: closure and stabilized margin. (*d*) Phase 4: collision and flexural subsidence.

in turn, lead to superposition of these basins at the same location. The simplest example, which involves only extension and compression (figure 12), illustrates this cycle for an ocean basin that existed for a timespan that was longer than several lithospheric thermal time constants. Phase 1 is that of rifting, extension and attenuation of the lithosphere. The brittle faulting of the upper crust, illustrated by listric normal faults, also serves to thin that part of the lithosphere, which acts as an intact elastic core during flexure. It is likely that these faults are not subsequently annealed and that they can be reactivated by added extension, or reversed by compression. In so far as the faulted region may represent a substantial part of the rheological lithosphere (Beaumont *et al.* 1982), we expect that the faulted lithosphere will remain flexurally weak even when thermally mature. Moreover, unfaulted zones on either side will probably become relatively stronger on cooling. Phase 2 illustrates the thickening of the thermal lithosphere during the evolution of the rifted margin basin in what is often termed its passive margin phase. Phase 3 shows the initiation of subduction at a hypothesized location sufficiently seaward that the margin remains intact during ocean closure. This is obviously a simplified view, but one that is adequate to illustrate the concept of superposition. Essentially the same development would follow were the polarity of subduction reversed. Phase 4 is that of encounter and collision with the conjugate margin, which resists subduction. The lateral forces provide the driving mechanism that disrupts the rift margin sequence and transports it landward as a succession of thrust sheets. As these are stacked onto the old rift margin, their load creates the foredeep and foreland basin. Depending on the level at which detachment occurs and on the amount of compression, the foreland sequences is superimposed on all or part of the rift margin sequence.

The foreland basin is, from this point of view, a 'successor basin', a term that we do not favour because it has no genetic connotation.

More complicated examples, in which there is more than one collision event, or there are interactions with major edifices on oceanic lithosphere, can lead to successive periods of thrusting that are recorded in the foreland sequence. If strike-slip motions are also considered, some indication of the way in which an orogen was assembled can be deduced. Purely strike-slip motions cause no overthrusting; therefore no foreland basin is created. With increasingly normal convergence the importance of overthrusts and the foreland basin increases. This view adds further credence to, for example, interpretations that suggest that the Acadian and Alleghenian Orogenies in the Appalachian fold belt were the result of largely strike-slip and normal movements in the northern and southern regions respectively.

The central Appalachians also provide an ideal example of the superposed nature of foreland sequences. At least three separate molasse wedges, separated by unconformities and corresponding to the major orogenies, have been superimposed on the Cambro–Ordovician rift sequence that corresponds to the opening of the Iapetus Ocean. Each of these sequences must be considered as a separate entity in any analysis of basin subsidence, not forgetting those parts of the stratigraphic sequence lost during uplift and erosion.

In so far as stresses acting within the lithosphere during collision are responsible for reactivation of inherited fault structures, many of the local structures of the fold–thrust belt and the foreland may reflect these stresses. We suggest that where previous basement faults exist, these will control the pattern of overthrusting, with the faults breaking upward through the overlying intact rift margin sediments. The telescoping of the margin restores much of the extension that occurred on rifting. In regions landward of pre-existing extensional faults the detachment surface will be confined to weak zones in the sedimentary sequence. This view appears to be corroborated by evidence from the Rocky Mountains in Canada (R. A. Price, personal communication) in which inferred basement involvement in thrusting does not occur eastward of the palinspastically restored position of the hinge line to the underlying rift margin sequence. Collisional stresses may also be responsible for the reactivation of horizontal motions further into the interior of the continents. It is probably no accident that the initiation of subsidence on the Hudson's Bay, Michigan and Illinois cratonic basins is coeval with the Taconic Orogeny in the Appalachians.

The implication of this discussion is that a number of factors influence the structure of foreland basins, the thermal age of the margin, the amount of overthrusting and angle of convergence, the existence and amount of reactivation of inherited basement faults, and the amount of uplift and erosion of the fold–thrust belt. The diversity of foreland basins must reflect changes in the relative importance of these factors in each case.

Conclusions

The conclusions of this descriptive comparison of foreland and rifted margin basins are simple and can be summarized as follows.

1. These basins form in respectively compressional and extensional environments within what may be regarded as lithospheric plate boundaries and as a consequence of lateral motions of the plates.

2. Tectonic subsidence of foreland basins is primarily the result of flexural downwarp of the lithosphere under supracrustal loads of the adjacent fold–thrust belt.

3. Tectonic subsidence of rifted margin basins primarily results from mass replacement at depth, during lithospheric extension on rifting, and the subsequent conductive cooling and contraction of the extended lithosphere.

4. Lateral extension of the basins beyond areas immediately influenced by the tectonic subsidence is controlled by lithospheric flexure under sediment and water loads. The flexural wavelength is primarily a function of the thermal age of the underlying lithosphere, which implies that flexure will be less important in rifted margin basins than in foreland basins.

5. The lithosphere may be modelled as a uniform elastic or viscoelastic layer supported by an inviscid fluid. Viscoelastic (Maxwell) models predict basins that shrink in size as stress is relaxed, whereas elastic models having a time-invariant flexural rigidity predict basins with constant width. More complex rheologies are considered likely to have intermediate extrinsic properties.

6. Second-order effects, such as lateral changes in lithospheric properties or some of the consequences of lithospheric cooling, will modify basin evolution.

7. Basin characteristics reflect a subtle interplay of tectonic subsidence, lithospheric flexure, sea level change, surface topography and inherited properties. All these factors must be considered in an analysis of subsidence history.

8. Changes in tectonic environment that accompany the change from opening to closing of an ocean basin, can lead to the superposition of foreland and rifted margin basins.

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