GONDWANA AND TETHYS

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Gondwana and Tethys

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Introduction

M. G. Audley-Charles and A. Hallam

In the last few years it has become apparent that the concept of a single end-Palaeozoic supercontinent, Pangaea, breaking up progressively through the Mesozoic and Tertiary into the present continental array, is seriously oversimplified. While the Atlantic Ocean was opening up, causing disintegration of western Pangaea, a series of substantial continental fragments was being swept across Panthalassa, to dock on the western margins of North America and the eastern and south-eastern margins of Asia. Some of the more southerly continental blocks that now form part of eastern 'Laurasia' had their origin on the other side of Tethys, in the Gondwanan supercontinent; the process of accretion probably began in Palaeozoic times. There is no problem in geotectonics more challenging than obtaining a fuller understanding of this process. It requires collaborative research by tectonicians, palaeomagnetists, stratigraphers, and palaeobiogeographers, and with this in mind it was decided to hold a joint meeting, the first Lyell Meeting, between the Geological Society and the Palaeontological Association, on 7 and 8 May 1986. This volume represents the published proceedings, with the chapters corresponding substantially with the sequence of lectures presented.

The first part is concerned with two aspects of Gondwana and Tethys evolution: (1) the tectonic (structural and metamorphic) processes associated with the splitting and dispersal of continents; and (2) evaluation of the palaeomagnetic and stratigraphical records associated with the splitting of Gondwana and evolution of the Tethys oceans.

Price *et al.* find the popular mechanisms for plate movement, namely basal traction, slab pull, and ridge push inadequate to explain known plate movements. They argue that the driving force for plate motion is the body weight of the lithosphere, generating gravity glide on the lithosphere–asthenosphere boundary. They postulate that the continental splitting and subduction initiation result from this mechanism, as well as effects requiring high horizontal stress. Dewey, on the other hand, argues that Pangaea-type continental configurations generate deviatoric tension, and hence are predisposed to splitting along weak zones and where tension is locally enhanced by thermal uplifts. He recognizes episodic continental assembly, disruption, and reassembly associated with tectonic and magmatic events and sea-level changes.

Hall puts forward a case for considering that high-temperature/lower-pressure metamorphism in parts of the eastern Mediterranean was the result of thermal effects on the deep crust during lithospheric extension, and not the consequence of subduction-related compression. Furthermore, he shows how the generation of flysch deposits could result from the peripheral uplift associated with lithospheric stretching. The preservation of high-T and low-P metamorphic rocks associated with crustal extension will occur where subsequent collision thrusts the metamorphic basement over its former cover. He identifies an example from eastern Tethys, where the Gondwana margin (Australia) has collided with a volcanic arc in Timor.

Hall's paper provides the link between the extension processes of rifted Gondwana and collision activities associated with subduction of Tethys. Mann and Vita-Finzi are concerned with analysing the forces involved at such collisional plate boundaries. Using the geometry of the frontal folds, they have determined the horizontal force equivalent to 0.134–0.790 kbar (value is model-dependent) to have been responsible for these folds where the mountain front has advanced at 17 mm/yr.

Tarling reviews the palaeomagnetic data for the Mesozoic configuration of Gondwana and its breakup. He emphasizes the importance of matching ocean-floor magnetic anomalies that formed within the same transform segment. On this basis, he proposes modifications to the configuration and breakup of Mesozoic Gondwana put forward by Norton and Sclater (1979). He points out the need for many more palaeomagnetic data, especially from the collided continental fragments rifted from Gondwana.

The rifted continental margin of northern Australia has been well documented by the petroleum exploration companies, so that, together with the magnetic spreading lineations in the floor of the north-east Indian Ocean, there

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is a body of evidence indicating that a major continental block was rifted from northern Australia in the Jurassic. This rifted margin has been greatly modified during the Cainozoic by collision with the arc–trench systems. Nevertheless, this Jurassic rifting episode is traced by Audley-Charles in the stratigraphic record of the fold and thrust belt collision zone of the Banda Arc, Sulawesi, and New Guinea. The importance of this uplifted fold and thrust belt is that it exposes the stratigraphic sequences that record the rifting episode of Mesozoic Gondwana breakup, and it also presents an example of a collision suture at a rifted continental margin. This may be valuable in guiding our interpretation of the palaeogeographical significance of tectonic sutures. How valuable it is depends in part on whether or not this Banda–New Guinea suturing is typical of continental collision zones.

Metcalfe interprets the pre-Tertiary core of south-east Asia in terms of at least four continental blocks rifted from Gondwana. He sets out the stratigraphic, palaeontological, and palaeomagnetic evidence for his reconstruction that these blocks are now separated by tectonic collision sutures. The presence of glacial-marine deposits and associated cold-water faunas provides diagnostic evidence for these Asian continental blocks having been part of eastern Gondwana in the late Carboniferous and early Permian. Palaeomagnetic data are not adequate in quality or range to provide much guidance. The fossil flora, designated as warm Cathaysian (*Gigantopteris*) or cold Gondwanan (*Glossopteris*), are generally regarded as characteristic of distinct palaeogeographical provinces. However, the degree to which the warmer climate around parts of maritime Gondwana could have been associated with Cathaysian floras is unknown.

The evaluation of the palaeogeographical significance of structural sutures is one of the outstanding problems of Asian geology. Adjacent regions of contrasting facies and faunas require explanation, but what kind of suture should we expect to find where two continental blocks have collided? How does the obliquity of collision and the relative role of strike-slip movement, some of which may post-date collision, influence the suture? In the last chapter of this first part of the volume, Şengör *et al.* present their latest review and reappraisal of the palaeogeographic evolution of the whole Gondwana–Tethys belt. Based largely on regional geology derived from reviewing an extensive literature, they start from the basic tenet that major structural sutures containing ophiolite fragments represent tectonic sutures between continental blocks where oceanic crust has been subducted. They test this against available magmatic and structural evidence for subduction together with faunal, facies, or palaeomagnetic indications of large-scale movements of the blocks involved. Inevitably, in such a broad regional approach to this huge and complex region there will remain disagreements over local and not so local interpretations. Nevertheless, they offer the beginning of an explanation for the remarkable continuity of facies and faunas that can be observed, for example in the late Triassic, late Cretaceous, and Eocene: from Crete in the eastern Mediterranean to Timor and Seram in eastern Indonesia. A feature to emerge is the variety of continental blocks to have been transferred from Gondwana to Laurasia. Another conclusion that may be drawn from this review is that an enormous amount of ocean crust and its sedimentary cover must have been subducted, and an impressively large volume of continental margin deposits (slope and rise), as well as forearc deposits (perhaps with basement), seem to have disappeared in the process of reassembling these Gondwana continental fragments at the Laurasian margin.

The second part of this volume from Chapter 10 onward, is devoted principally to a presentation and evaluation of palaeontological evidence relevant to the Gondwana–Tethys theme, and is dealt with in stratigraphical order as far as is feasible.

Taking account of the distribution of Ordovician and Silurian marine invertebrates, Cocks and Fortey present a new base map for ancient Gondwana. Whereas the core comprising South America, Africa, peninsular India, Madagascar, and Antarctica has remained uncontroversial as a coherent supercontinent through much of Phanerozoic time, there has been dispute about peripheral regions along the Tethyan margin. In the Cocks and Fortey reconstruction, a belt extending from southern Europe and the Middle East to southern China and south-east Asia is incorporated into the early Palaeozoic Gondwana supercontinent.

Chaloner and Creber review the current status of the well-known late Palaeozoic *Glossopteris*, Cathaysia, and Angara floral provinces, urging caution in the interpretation of fragmentary leaf remains, misidentification of which has led to confusion in the past. Confining their attention to the more reliable data, they use it to test an early Permian plate reconstruction based on palaeomagnetic results. Note is taken by Waterhouse of the distribution of Permian marine faunas and sediments, and the climatic zones inferred from them, to throw into question some recent palaeogeographical reconstructions that demand a vast wedge-shaped Tethys in Asia, and movement of allochthonous terranes over substantial distances.

Turning to the Mesozoic, Kristan-Tollmann adds to her earlier published examples of Triassic fossils that appear strikingly similar, even at species level, throughout the length of the Old World Tethys. What is especially interesting is that the same species occur also in western North America. This has been interpreted as the consequence of free larval migration westwards across Panthalassa, by means of an equatorial palaeocurrent (Tollmann and Kristan-Tollmann 1985) but account should also be taken of the fact that the western margin of North America consists of a series of allochthonous terranes. Indeed, much of Panthalassa might have contained numerous large islands in Triassic times (Tozer 1982) which would have greatly facilitated marine invertebrate dispersal. However, crustacean coprolites similar to those described by Kristan-Tollmann have also been recorded by Senowbari-Daryan and Stanley (1986) from Peru, which is unlikely to have been composed of far-travelled terranes (Hallam 1986).

Jurassic ammonite distributional data are dealt with in two short articles by Thierry and Westermann, dealing respectively with the western and eastern Tethys. Thierry attempts to disentangle historical from ecological factors in accounting for ammonite distributions, in order to use these to test a palaeogeographical reconstruction. This, however, involves having to make assumptions about the life habits of particular groups. A conflict emerges concerning the interpretation of the Caucaso-Dobrodgean furrow. Another contributor to this book, Şengör, considers this to have been closed in mid-Jurassic times, whereas Thierry requires it to have remained open at that time, in order to account for the confinement of certain ammonites to the north of the Pontides. This conflict points out the need for a better understanding both of Jurassic palaeogeography and the factors controlling ammonite distributions. In contrast, Westermann finds that Jurassic ammonite distributional data support recent plate-tectonic reconstructions indicating a separation of the Lhasa and Quamdo blocks of Tibet from the Tethyan Himalaya.

Mesozoic brachiopod distributions suggest to Ager that Mesozoic Turkey should be placed on the northern side of Tethys rather than the southern side, as in some reconstructions, because the Turkish faunas are European rather than African or Middle Eastern in their affinities. Skelton's article raises once more the question of trans-Pacific migration, this time utilizing data from mid- to late-Cretaceous benthic organisms. He infers an island-studded ocean, related to extensive mid-plate volcanicity, which allowed Caribbean benthos to spread westward by larval drift into the eastern Old World Tethys, utilizing the islands as a series of staging posts.

The only article on terrestrial vertebrate distributions, by Rage, presents a comprehensive review of Mesozoic and Cainozoic data. Free dispersal was possible in Triassic times across the whole of Pangaea, and land communications between Gondwana and Laurasia persisted into the Jurassic. Subsequently, relations between the Pangaea components became more complicated, and after the Eocene, exchanges between the now-separated Gondwana continents disappeared, whereas intermittent exchanges continued between some Gondwana continents and the northern continents. Rage presents an interesting analysis of distributions changing with continental breakup. The outstanding problem remains the lack of proven late Cretaceous endemic fauna in India, although all plate-tectonic reconstructions agree that the subcontinent was isolated by ocean at this time.

Rosen and Smith present a new method of analysis of late Cretaceous to Recent reef coral and sea-urchin distributions, and argue the case for using cladistic biogeography as the most rigorous method available. By using parsimony analysis of taxa shared between sample localities, one should be able to determine the relative recency of faunal contact between different regions. The method provides a means of testing independently derived palaeogeographical reconstructions and could prove of great value for older periods, for which the inferred palaeogeographies are considerably more conjectural. Whitmore's concluding chapter deals with angiosperm distributions in the Malay Archipelago, which was created in mid-Miocene times by eastern Gondwana–Laurasia collision. Some plant taxa support the geological inference that Gondwana fragments moved north before the main collision, to provide island 'stepping-stones'.

As is characteristic of a symposium volume presenting a wide variety of techniques and viewpoints, it is perhaps most useful in providing a series of 'state of the art' reviews and identifying areas of potential or actual conflict that need to be resolved by further research, rather than in solving major problems. Clearly, one such area is the timing of separation and collision of particular continental blocks, particularly adjacent to the eastern Tethys. What does seem to be beyond serious dispute now, is that a succession of continental fragments has been peeled off from the northern rim of Gondwana, transported across Tethys, and plastered against the southern margin of Laurasia. (Owen's (1983) alternative, involving the creation of Tethys by Earth expansion since the Triassic, can be decisively ruled out (Weijermars 1986).)

Some sort of large-scale geotectonic explanation is required. Bearing in mind the gravity-gliding interpretation of plate tectonics favoured by Price *et al.* in this volume, the most useful way of proceeding might be to pursue further the stimulating idea of Anderson (1981), who notes the remarkable correspondence of the Wegenerian reconstruction of continents and the present location of geoid anomalies and hotspots. A supercontinent such as Pangaea could have acted to insulate a large part of the underlying mantle for periods of well over 10^8 years. The excess heat trapped beneath the supercontinent would have caused uplift by thermal expansion and partial melting; a large geoid high would have been generated by this expansion. Eventually, the supercontinent would disintegrate, and fragments disperse from the geoid anomaly. The main implication in the present context is that the anomaly would have been centred under Gondwana.

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Gravity glide and plate tectonics

Neville J. Price, Geoffrey D. Price, and Sarah L. Price

ABSTRACT: We briefly discuss the three groups of mechanisms which are generally held to be responsible for plate motions, namely basal traction, slab pull, and ridge push. We conclude that individually, or in combination, these mechanisms are insufficient to account for known plate behaviour. Instead, we reintroduce and expand upon an idea proposed by Hales (1969) that the driving force for plate motion is the body weight of the lithosphere, which gives rise to gravity glide on the inclined interface between the lithosphere and asthenosphere. It is shown that this mechanism is capable of producing an average horizontal stress of as much as 6 kbar and is able to explain (1) continental splitting, (2) the initiation of subduction, and (3) observable plate motions, as well as a variety of effects requiring high lateral stresses. This is the first model to be proposed that successfully provides a quantitatively and physically viable mechanism to explain plate-tectonic processes and the behaviour of the Earth's lithosphere.

Introduction

It is generally accepted that the Earth consists of an outer, rigid lithosphere, subdivided into several plates, overlying a dynamic, convecting asthenosphere. However, the theory of plate tectonics lacks a generally accepted, detailed, or quantitative model to describe the dynamics and kinematics of plate motion and mantle convection. We must identify and quantify the forces which determine plate motion, if we are to progress beyond completely qualitative models for major events such as the breakup of Gondwana.

In this paper we briefly discuss some current models of the various forces which are thought, by some, to determine the motion of the lithospheric plates, and hence the evolution of continental crust. We subsequently present a simple analysis which indicates that gravity glide, resulting from the body weight of the lithosphere itself, can generate large lateral stresses, acting normal to, and directed away from, the oceanic ridges and continental rifts. These stresses are of such a magnitude that they will play a vital role in determining the initiation of continental rifting, the development of subduction zones and the kinetics of plate motion.

Forces acting on lithospheric plates

Studies directed at determining the origin of the forces acting on plates, tend to be performed either from a fluid dynamical perspective or from a plate-tectonic point of view (Loper 1985). In the former approach, the plates are considered merely to behave passively as an upper, cold, thermal boundary layer of a convecting mantle, while in the latter approach, the forces and moments which act upon a plate are analysed in an effort to identify the dominant forces and hence the nature of the driving force for plate motion. Such plate-oriented studies invariably conclude that it is highly improbable for plates to be driven by basal, mantle-derived drag (e.g. Forsyth and Uyeda 1975; Richardson *et al.* 1979; Bott 1982), and that plates must, to some extent, be decoupled from the underlying convective mantle, as proposed by such workers as Parsons and Richter (1981). However, the behaviour of lithospheric plates is not expected to be totally unrelated to mantle convection, since subduction of rigid lithosphere will undoubtedly have an effect upon the dynamics of the mantle through local viscous coupling. Nevertheless, current thinking appears to favour the contention that the details of plate motion, particularly their translational velocities and rates of rotation, are determined by the forces acting at the margins of the plates, rather than by the flow of the underlying mantle.

The forces that act at the edges of plates have been discussed in a number of works (e.g. Forsyth and Uyeda 1975; Bott 1982; Dewey 1988), and can be divided into forces which might move plates and forces which would resist plate motion. The two main forces which act at the edge of plates to produce motion are a pushing force at the ridges, i.e. constructive plate boundaries, and a pulling force at the trenches. The latter force is thought to result from the

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negative buoyancy associated with the sinking of cold lithosphere at subduction zones. There are many forces which resist plate motion. The most significant are probably those associated with slab subduction, continental collision and transform-fault friction (Bott 1982).

The hypothesis that plates are driven by edge forces acting at trenches and ridges is acceptable on thermodynamic grounds (e.g. Runcorn 1980; Bott 1982). There have been several detailed, quantitative studies of the feasibility of driving plates by edge forces (e.g. Forsyth and Uyeda 1975; Chapple and Tullis 1977). From an analysis of current plate geometry and motion, these workers concluded that the forces acting on the downgoing slab are an order of magnitude larger than any other force, and so control the velocity of the oceanic plates. Forsyth and Uyeda (1975) suggested that oceanic plates which are attached to substantial amounts of subducting lithosphere, approach a terminal velocity. At this velocity the slab-pull force that results from the negative buoyancy of the plate is balanced by the viscous and frictional drag forces which resist plate subduction, irrespective of all other driving forces acting on the plate. This model, however, breaks down when extrapolated back in time. Firstly, there is the problem of initiating subduction, and secondly, their analysis of current oceanic plate velocities in terms of terminal velocities appears to be incompatible with the high velocity (>150 mm/yr) believed to be associated with the Indian plate before 55 Ma BP (Peel, 1982).

An analysis of intra-plate regional stresses led Richardson *et al.* (1979) to conclude that the forces within the mantle, which resisted slab subduction, approximately balance the large gravitational potential of the slab, and hence significant ridge-push forces must exist to explain the ubiquitously observed compressive intra-plate stresses. In addition, Price and Audley-Charles (1987) have argued that detachment of the downgoing slab has occurred below Timor and that, if it is now sinking under its own weight, it is doing so very slowly (<20 mm/yr). Thus it can be inferred that the large slab-pull forces are generally absorbed by the underlying mantle, by means of viscous drag and coupling, rather than being transmitted to the rest of the trailing plate, and hence the motion of surface plates must be largely determined by ridge-push forces.

The nature of the 'ridge-push' forces is as contentious as their role in plate tectonics. An early suggestion for the origin of the 'ridge-push' force (Hales 1969), was that it arose from the gravity sliding of young oceanic lithosphere down an inclined and 'lubricated' lithosphere–asthenosphere interface. This mechanism is not strictly a 'ridge-push' force, but is the result of the distributed body weight of the lithosphere. McKenzie (1972), Turcotte and Schubert (1982), and others concentrated on 'gravity glide' resulting from the topographic relief of ocean ridges. Bott (1982, p. 357) dismissed the suggestion that gravity glide can play any role as a plate driving force, and stated instead that the ridge-push force is caused by the progressive upwelling of low-density asthenosphere beneath ridge crests. His arguments against gravity gliding are aimed only at the attempts to explain ridge-push forces as resulting from ridge topography, and are not applicable to Hales' original suggestion that gravity glide is the result of sliding down the inclined surface at the base of the lithosphere. We later reintroduce and expand upon the idea briefly outlined by Hales, and deal with certain criticisms of his mechanism.

The various concepts so far put forward in the literature are piecemeal approaches to the problem of the evolution of a plate. Ideally, the whole history of the development of the plate, its motion and associated features should be explicable in terms of one or a combination of mechanisms. The main steps involved in such an evolution are represented in Fig. 1.

Initially, (Fig. 1a) we consider continental or oceanic lithosphere to extend around the Earth, without perturbations such as ridges or subduction zones. This uniform state is disturbed (Fig. 1b) by the rising of the thermal plume which causes an increase in temperature and results in an upward deflection of the geotherms. We take the junction between the lithosphere and the asthenosphere to be thermally defined, and to occur at such a temperature that there is a marked change in the rheological behaviour of the mantle material. Consequently, the plume causes an upward migration of the lithosphere–asthenosphere boundary. It is above the high point of the plume, we suggest, that a ridge will develop. The establishment of the ridge with a well-developed, low-velocity zone beneath is shown in Fig. 1c. However, it is not until subduction is initiated that a plate is defined and set free to move (Fig. 1d). At this stage the ridge becomes a 'spreading centre'.

It is our contention that the whole of this cycle of events can be explained in terms of gravity glide and associated gravitational effects. However, as we shall see, the horizontal stresses in the lithosphere, in the initial, or 'at rest', condition are extremely important to our thesis, so we shall discuss the factors which control the

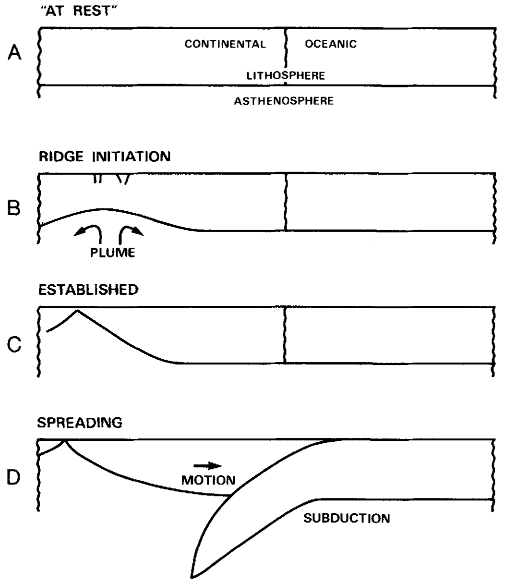

FIG. 1. (a) The 'at rest' configuration of lithosphere and asthenosphere prior to a perturbation caused by a plume. (b) The initiation of a ridge above a plume. (c) Ridge established, but significant plate motion inhibited. (d) Initiation of subduction and significant plate motion.

magnitude of these lateral stresses before we proceed to consider the gravity-glide mechanism.

Magnitude of lateral lithospheric stresses

In the initial state represented in Fig. 1a, the lithosphere will be in a state of zero lateral strain and the lateral stresses will be induced only as the result of gravitational loading (i.e. the total vertical stress S_z) where:

$$S_z = \varrho \cdot g \cdot z \tag{1}$$

and where ϱ is the density, z is the depth and g is the gravitational acceleration. If the lithosphere does not fail, the lateral stresses will be the result of elastic deformation, so that the horizontal stress (S_x) is given by:

$$S_x = S_z/(m-1) \tag{2}$$

where m is Poisson's number (Price 1966). This ranges from a minimum of 2.0 [for a liquid], to infinity for a completely rigid solid. The value we shall use for strong lithospheric material is $m = 4.4$. If failure occurs, it may be the result of brittle deformation (this mode is likely in the upper levels of the lithosphere) or, at high temperatures (in the lower continental crust or in the lower levels of the lithosphere), failure may occur in a ductile manner. For brittle shear failure (i.e. the development of normal faults), the ratio between the greatest and least principal stresses (S_1/S_3) is given by the relationship:

$$S_1/S_3 = (1 + \sin\phi)/(1 - \sin\phi) \tag{3}$$

where ϕ is the angle of sliding friction (Price 1966). In the environment under consideration, $S_1 = S_z$, and $S_3 = S_x$, so that S_3 can be readily estimated from the value of ϕ. If ductile failure occurs however, then in order to estimate the lateral stress, it is necessary to use the equations of state of the various rock types that are likely to deform by diffusion mechanisms (Fyfe *et al.* 1978).

The lateral stresses (S_x) for the elastic or brittle failure conditions can readily be assessed. The value of m for olivine is approximately 4.4, hence the increase in lateral stress with depth is as indicated by line B in Fig. 2a. The vertical stress (S_z) is represented by line A, so it will be seen that the lateral stress is 70 per cent smaller than the vertical stress at any level z.

The value of ϕ for many strong rocks is about 45°, so that if failure occurs by brittle shear, the ratio of S_1/S_3 is 5.8. The lateral stress for these rocks is represented by line C in Fig. 2a. It can be inferred from this figure that the lateral stresses, induced elastically, are too high to permit failure to occur in rocks with $\phi = 45°$. (In order for shear to occur ϕ must be less than 30°, for the lateral stress conditions represented by line B.)

The problem of assessing the differential stresses that are consistent with ductile deformation is much more difficult, for it requires quantifying temperature and strain rates, and it depends upon whether the rocks are wet or dry. In addition, the stresses depend upon lithology and grain size. Hence, the lateral stress/depth relationship in oceanic and continental lithosphere will be different.

The probable differential stress that wet and dry olivine can withstand at a strain rate of 10^{-15} sec^{-1} (derived from Dewey 1988) for a given depth, is indicated in Fig. 2b. The combined stress curves of Fig. 2a and b are represented in Fig. 2c, where the 'probable' lateral stress/depth relationship is shown as the bold solid line. The bold dashed line represents the probable lateral stress/depth relationship for continental crust. The shaded area indicates the amount by which the lateral stress falls short of the lithostatic pressure (for which condition the horizontal and vertical stresses are equal and given by line A in