PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Problems of Passive Margins from the Viewpoint of the Geodynamics Project: A Review [and Discussion]

M. H. P. Bott, R. A. Scrutton and M. F. Osmaston

Phil. Trans. R. Soc. Lond. A 1980 **294**, 5-16
doi: 10.1098/rsta.1980.0007

Phil. Trans. R. Soc. Lond. A **294**, 5–16 (1980) [5]
Printed in Great Britain

Problems of passive margins from the viewpoint of the geodynamics project: a review

By M. H. P. Bott, F.R.S.
Department of Geological Sciences, University of Durham, Durham DH1 3LE, U.K.

Passive margins form by continental splitting which may follow an early pre-split graben stage of development. Such margins are divisible into rifted and sheared types. After formation the margins develop by predominantly vertical tectonics as the intervening ocean widens by seafloor spreading.

The plate splitting mechanism is not yet understood, but evidence from aseismic ridges and associated continental volcanism is relevant. Location of the original continent–ocean contact is also problematical in many regions, especially where quiet magnetic zones occur above crust of uncertain status.

Rifted margins are notable for great subsidence resulting from early graben formation and later flexural subsidence of the shelf and slope regions. Graben formation is attributed to wedge subsidence of the upper crust in response to crustal stretching, possibly associated with doming before splitting. Four factors have been recognized as contributing to flexural subsidence: gravity loading, thermal subsidence of the adjacent oceanic lithosphere, possible thermal subsidence of the continental lithosphere following heating and erosion at the time of break-up, and thinning of the continental crust by seaward creep of lower crustal material. The relative importance of these four mechanisms needs clarification.

Introduction

Passive continental margins occur at the junction of continental and oceanic types of crust within plate interiors. They form as a result of continental splitting by rifting or by transform faulting. Such margins are seismically inactive except at the time of formation, in contrast to the active margins characteristic of the Pacific Ocean. Their tectonic development involves strong differential vertical movements both during the period of formation and subsequently. The Atlantic Ocean is bordered by passive margins except locally in the Caribbean and Scotia arc regions.

Rifted margins form as a result of continental splitting at a new ocean ridge. They undergo a varied history of development which may involve four stages as follows: (1) a rift valley stage, not necessarily ubiquitous, may involve thermal uplift and graben formation before continental splitting, as possibly now exemplified by the East African rift system and the Baikal rifts; (2) a youthful stage lasting about 50 Ma after splitting when thermal effects of the split are strongly felt, exemplified by the present Red Sea margins; (3) a mature stage during which more subdued development continues, characteristic of the present Atlantic margins; (4) a fracture stage when subduction starts, terminating the history as a passive margin. Two main types of subsidence affect rifted margins. Fault-controlled graben subsidence occurs during the rift valley stage and may extend into the youthful stage. Flexural downwarping towards the ocean is characteristic of the youthful and mature stages. The combined effect of graben and flexural

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

subsidence may produce sediment piles of up to 15 km thickness beneath outer shelf and slope.

Offset (or *sheared*) *margins* form by continental separation along transform faults. They undergo an initial active period as a transform fault followed by a subsequent history of greater stability than typical rifted margins. They have not yet been studied in such detail as passive margins.

The main problems common to passive margins of both rifted and offset types are location of the continent–ocean crustal contact, nature and development of the deep crustal and upper mantle transition, and the mechanisms of vertical movements of graben and flexural types (figure 1). An earlier review of these problems was given by Bott (1976*a*).

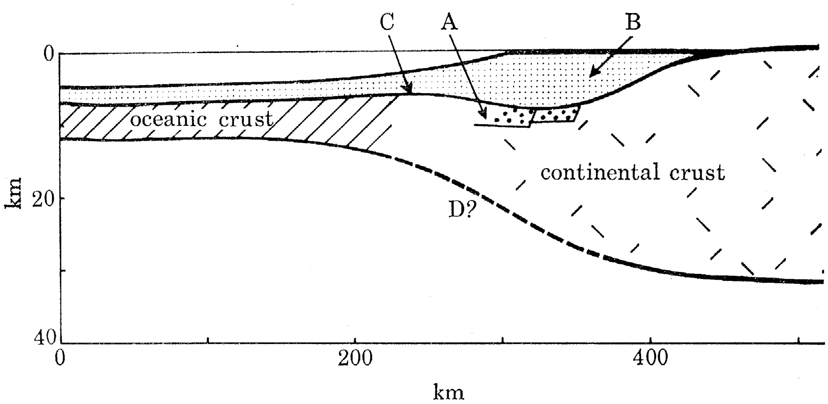

FIGURE 1. Structure of a typical passive margin of rifted type showing some of the characteristic features of sediment and crustal structure. A, Pre-split graben sediments; B, post-split sediments associated with flexural subsidence; C, the problematical position of the continent–ocean crustal contact buried deep beneath sediments; D, the apparent gradational transition between deep continental crust and oceanic lithosphere. After Bott (1979).

MECHANISM OF CONTINENTAL BREAK-UP

The present rift valley systems mark the possible locations of newly forming accretionary plate boundaries. The passive margins mark the locations of an extensive series of ancient splits which occurred in the datable past. Study of present rift systems and passive margins are thus complementary to each other in studying the plate splitting mechanism.

Regions such as the early Tertiary North Atlantic igneous province, where extensive continental volcanism occurs at approximately the same time as continental splitting, are particularly relevant to this problem. Such regions may be associated with development of anomalously thick oceanic crust, examples being the Icelandic Transverse Ridge (Bott 1974) and the Rio Grande Rise (Campos *et al.* 1974). The origin of both the volcanism and subsequently the thick oceanic crust has generally been attributed to an underlying hot upper mantle (Wilson 1963) or, more controversially, to a mantle plume (Morgan 1971). In more typical passive marginal regions the igneous activity at the time of splitting is less conspicuous, partly because it is less extensive and partly because it may be masked by the thick marginal sediments.

These observations lead to the speculation that the hot spot regions such as Iceland may be the foci of new continental splits by a mechanism related to the underlying hot, magma saturated upper mantle. Splitting may occur once an interconnected set of dykes and transform faults related to such a focus can connect two existing plate boundaries, normally following lines of pre-existing weakness. The nature and intensity of igneous activity associated with the splitting of more normal, less active, segments of passive margins now needs to be studied in regions where sediments can be penetrated into the crust.

Junction between continental and oceanic crust

It is basic to the geodynamic study of passive margins and to continental reconstructions to be able to locate the contact at shallow crustal depths between oceanic and continental crust. The contact has commonly been assumed to occur beneath the slope; for instance, Bullard *et al.* (1965) used the 900 m depth contour in their Atlantic reconstruction. Local misfits in this and other reconstructions show that this criterion is not universally applicable. Another early approach was to use the gravity anomaly to locate the change in crustal thickness from continental to oceanic, but this criterion is at best insensitive and may be wrong. An example from the southeastern Greenland margin (figure 2) shows that neither the slope nor the main change in crustal thickness are necessarily even approximately coincident with the continent–ocean contact (Featherstone *et al.* 1977). The location of the slope here depends on erosional activity of contour currents, and extreme thinning affects the continental crust now occurring beneath the rise.

The contact between the oceanic layer 2 and the continental basement can be located precisely by magnetic anomalies in some regions such as the western margin of the Rockall micro-

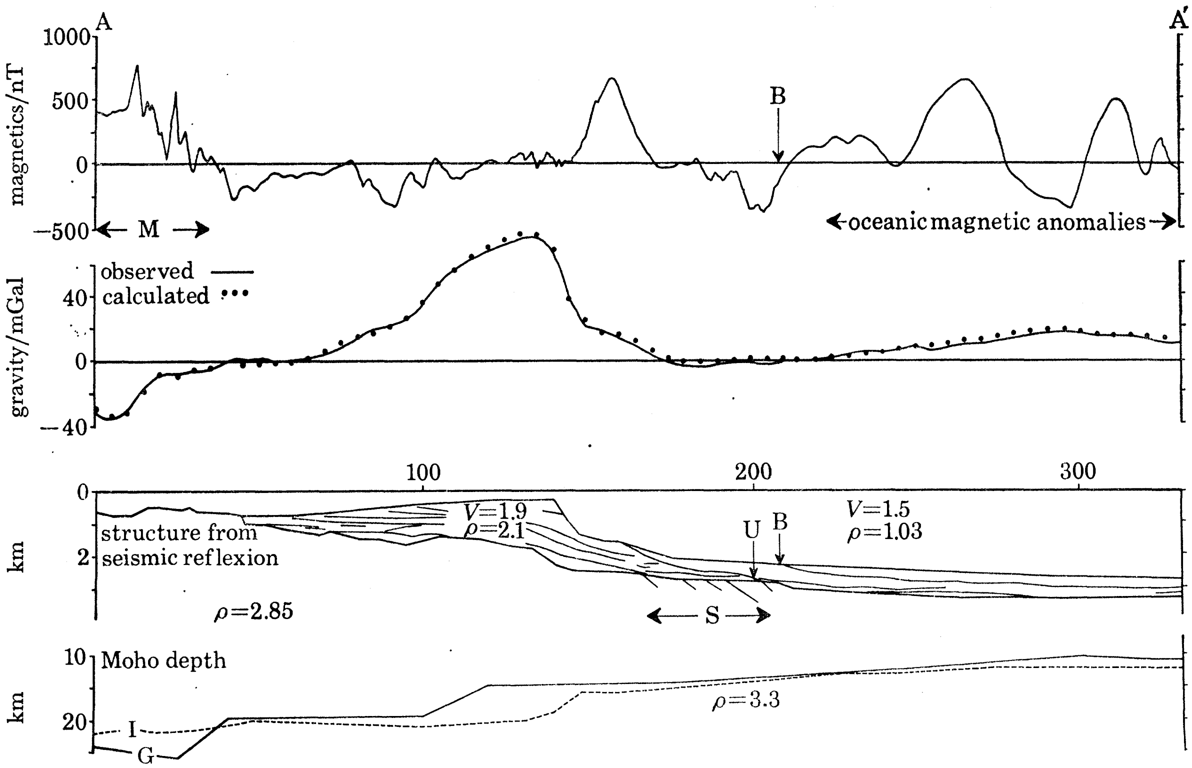

Figure 2. The structure of the southeastern Greenland continental margin at 63–64° N, showing the interpreted continent–ocean crustal contact marked B recognized from the magnetic anomalies and seismic profiling, the steep slope moulded by sedimentation and contour current erosion, and the change in crustal thickness predominantly occurring beneath the shelf. Reflectors interpreted as pre-split sediments are marked S, and an unconformity attributed to contour current activity is marked U. After Featherstone *et al.* (1977).

continent (Vogt & Avery 1974) and some South African margins (Talwani & Eldholm 1973). The magnetic method, however, fails over much of the central part of the Atlantic Ocean and in Rockall Trough because of the widespread occurrence of marginal magnetic quiet zones (Heirtzler & Hayes 1967). A major problem in the Atlantic is to determine whether such quiet zones represent oceanic crust (the majority view) or subsided continental crust. For instance, the quiet zones adjacent to the matching margins of North America and North Africa are normally considered to be underlain by oceanic crust formed during the constant polarity period of lower to middle Jurassic age, the marginal contact occurring beneath the slope and marked by a strong magnetic anomaly (Keen & Keen 1974; Sheridan 1974; Mayhew 1974), but Rabinowitz (1974) placed the contact within the North American quiet zone, regarding the inner part of the rise as subsided continent. Similarly, there is controversy as to whether Rockall Trough is underlain by oceanic crust formed during a constant polarity period (Cretaceous?) or by subsided continental crust. The marginal quiet zones may originate in more than one way and clarification of their status is an important outstanding problem.

Where the magnetic method of locating the contact fails, the seismic methods probably offer the best approach. The main problem is that the crustal contact is typically buried beneath up to 15 km of sediments making resolution outside present capabilities. The so-called 'starved margins' where sediments are relatively thin are an exception where the seismic methods are now applicable. For instance, beneath the margins of Rockall Trough, the contact can be located at the foot of the slope or beneath the rise by the change in character of the seismic basement revealed by reflexion surveying, a much rougher basement being characteristic of the oceanic side of the contact (Roberts 1975; Dingle & Scrutton 1977). Reflexion surveying similarly reveals the contact beneath the south-eastern Greenland margin at a position consistent with the magnetic anomalies (Featherstone *et al.* 1977). The refraction method also has potential, at present unexploited, for locating the contact in such regions.

Difficulty in recognizing the contact has also been encountered in the northeastern North Atlantic because of the lack of clear distinction between Icelandic type oceanic crust, which reaches a 30 km thickness beneath the Iceland–Faeroe Ridge (Bott 1974), and continental crust overlain by basaltic lavas. Talwani & Edholm (1972) regarded the outer Voring Plateau and the Faeroe Block as underlain by thick oceanic crust, but Bott *et al.* (1974) presented seismic evidence suggesting the Faeroe Block is underlain by continental crust beneath the surface lavas. Fresh evidence that there is an ocean–continent crustal contact between the Iceland–Faeroe Ridge and the Faeroe Block comes from observation of converted seismic phases produced at the contact in a crustal structure project crossing it (Bott *et al.* 1976). Such converted phases may offer a further method of approximately locating the crustal contact beneath passive margins.

Rabinowitz *et al.* (1976) have shown that the continent–ocean crustal contact for the southern Atlantic, as revealed by magnetic anomalies, is also associated with a characteristic signature in the Airy isostatic gravity anomaly. A steep gradient in the isostatic anomaly occurs above the contact, with a positive anomaly on the seaward side and a negative anomaly on the landward side. This anomaly pair is about 100–200 km wide and the amplitude is about 30–70 mGal.† Rabinowitz & LaBrecque (1977) attributed the positive anomaly on the seaward side to uncompensated oceanic basement highs near the margin. Alternatively the isostatic anomaly pair may be explained by a model of Turcotte *et al.* (1977) where the continental and oceanic lithosphere are locked together while the oceanic lithosphere subsides differentially on cooling.

† 1 Gal = 10^{-2} m s^{-2}.

I prefer this latter explanation. Whatever the origin of the isostatic anomaly pair, the steep gradient may provide a method of locating the crustal contact but with less precision than the magnetic and seismic methods.

Exact location of the crustal contact at passive margins is thus still problematical and controversial. Deep drilling, deep reflexion studies to basement and carefully planned refraction investigations, used in association with gravity and magnetic investigations, seem to offer most promise of better definition and clarification in normal and anomalous regions.

Turning to the transition at greater depths, the gravity anomalies show that passive margins are in approximate isostatic equilibrium (see above for small deviations), and that the crustal transition occurs over *ca.* 50–200 km. It is clear that there is normally some thinning of the continental crust, and some thickening of the oceanic crust, towards the contact. Further evidence on the nature of the deep crustal and upper mantle transition beneath passive margins is particularly scanty, although one would expect a recognizable transition between the cool old continental and hotter relatively new oceanic lithosphere. Seismological and thermal investigations of the deep transition are urgently needed if the geodynamic processes at passive margins are to be properly understood.

Early graben-type subsidence

The outstanding geodynamic problem of passive margins concerns the mechanism of subsidence. Such subsidence is of two main types: (1) early graben subsidence, possibly pre-dating the split; and (2) post-split flexural subsidence. The graben structures are difficult to study because most of them have been deeply buried beneath the later marginal sediments. Subsidence of graben type associated with the initial rupture is characteristic of much of the Atlantic region (Burke 1976) and has been well defined beneath the eastern margin of North America (Sheridan 1976). Such graben are not universally present adjacent to the crustal contact, for example, they are apparently absent from the Rockall Trough margins although some normal faulting is observed (Roberts 1975; Dingle & Scrutton 1977). In the northeastern North Atlantic region, graben-type subsidence of Triassic and Jurassic to early Cretaceous age has affected a very extensive region including Britain and its shelf regions (Dingle 1976), starting long before the continental splitting in this region.

The graben subsidence can be explained by a crustal stretching mechanism in response to a horizontal deviatoric tension. Such tension may be the result of domal uplift caused by heating of the underlying lithosphere, such as that of East Africa and the Basin and Range province in the western U.S.A. It can be shown that the additional surface load of the topography and the opposite upthrust of the low density lithosphere produce a horizontal deviatoric tension in the crust and that this tension is increased if the lower parts of the lithosphere can flow by slow viscoelastic creep (Kusznir & Bott 1977).

Modern concepts of graben formation start with the hypothesis of Vening Meinesz (1950) in which a downward narrowing wedge of continental crust forms by normal faulting and subsides isostatically to form a rift valley between flanking uplifts. This hypothesis fails in practice because the crustal root formed by the base of the wedge is not detected. However, the Vening Meinesz wedge subsidence idea has recently been applied to the brittle upper 10–20 km of the crust rather than to the crust as a whole (Bott 1971; Fuchs 1974; Bott 1976*b*) (see figure 3). This adaptation overcomes the difficulties and provides a viable explanation of graben formation by crustal tension within the setting of an incipient continental split.

It has also been suggested that extreme thinning of the continental crust within the rift valley setting can occur by plastic necking of the crust (Artemjev & Artyushkov 1971). Subsequent splitting of the rift zone could thus give rise to a zone of crustal transition beneath the slope (Kinsman 1975). Adequate extension of the brittle upper part of the crust cannot be accomplished by normal faulting of observed intensity but may occur by extensive igneous invasion and disruption. Such igneous activity is characteristic of the present continental rift valleys.

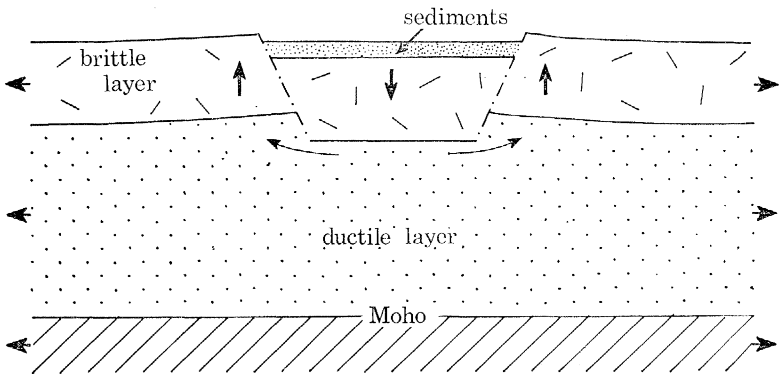

FIGURE 3. The mechanism of graben formation by wedge subsidence affecting the upper continental crust with outflow in the lower crust. After Bott (1976 *b*).

MECHANISMS OF REGIONAL SUBSIDENCE

Regional downwarping of passive margins following a possible graben stage is characteristic of the youthful and mature stages of development. Theories fall into three main groups depending on whether gravity loading, thermal phenomena, or stress-controlled crustal thinning is the dominant process.

The *gravity loading hypotheses* (figure 4) attribute subsidence to sediment loading. Assuming local Airy isostasy, marine sediment thicknesses of about twice the initial water depth can form if their mean density is about 2150 kg m^{-3}, or of three times the initial depth for a mean density of 2550 kg m^{-3}. This mechanism therefore requires substantial initial water depths if thick sediment piles are to be formed. A more sophisticated and realistic approach is to treat the lithosphere as a thin elastic plate and to investigate its flexural response to the load by elastic beam theory (Walcott 1972), but this does not greatly alter the overall pattern of predicted subsidence apart from extending the subsidence beyond the immediate confines of the load. The main success of the gravity loading hypothesis is in explanation of thick sediment wedges deposited on the slope and rise in delta regions such as that of the Niger. However, as shown by Watts & Ryan (1976), gravity loading cannot be the primary mechanism of formation of the substantial sediment thicknesses typically found beneath the shelf regions adjacent to passive margins. The gravity loading effect is a contributory factor in such subsidence, but is not the primary cause. A modification of the hypothesis applied by Collette (1968) to the North Sea suggests increased subsidence by the basalt to eclogite phase transition.

A second factor which must contribute to the development of a passive margin is the differential subsidence of the oceanic lithosphere relative to the adjacent continent caused by the progressive cooling of the oceanic lithosphere as it ages (Sleep 1969) (see figure 5). The differential vertical movement, amounting to about 3 km over 100 Ma, may be taken up by a single major

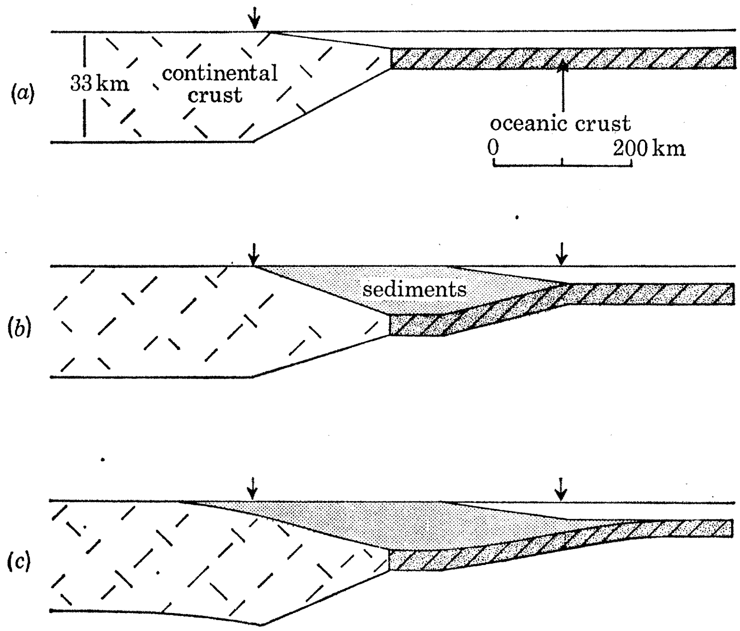

FIGURE 4. The gravity loading hypothesis, after Bott (1979). (*a*) The initial situation following Walcott's (1972) model. (*b*) The result of local Airy sediment loading, assuming the density of the sediments to be 2450 kg m^{-3} and of the mantle to be 3300 kg m^{-3}. (*c*) The result of flexural loading, adapted from Walcott (1972) with change in sediment density to that in (*b*).

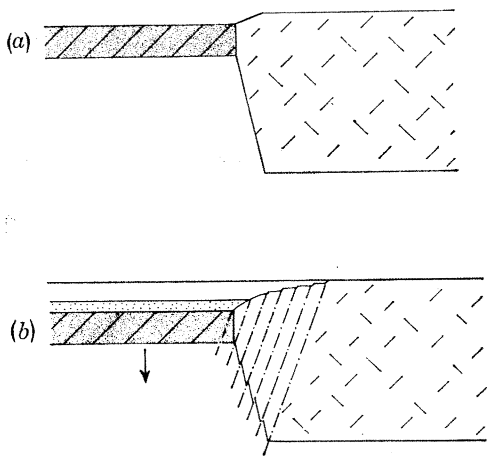

FIGURE 5. Downdrag of the continental slope by normal faulting associated with cooling and subsiding oceanic lithosphere. After Bott (1979).

fault or by a series of step faults as exemplified by the Rockall margins (Bott 1978); or the continental and oceanic lithospheres may become locked together at some stage causing local flexure and producing an isostatic gravity anomaly pair of the type observed by Rabinowitz (1974); or the continental side may also subside by an independent process reducing the relative vertical movement near the crustal contact. Turcotte *et al.* (1977) have numerically modelled the combined effect of such differential subsidence and sediment loading. Their mechanism explains thick sediments beneath the continental rise but still fails to account for thick shelf sediments without an independent mechanism of shelf subsidence, such as that of Sleep (1971).

The basic thermal hypothesis of shelf subsidence (Hsu 1965; Sleep 1971, 1973, 1976) assumes that the splitting continental lithosphere is heated at the time of continental break-up, causing reduction in density and consequent isostatic uplift (figure 6). The continental crust beneath the embryo margins is then thinned by surficial erosion or by some other process. After the split, and as the newly formed ocean widens, the lithosphere will cool with a time constant of about 50 Ma, causing isostatic subsidence of the shelf, which may be amplified by sediment load. This mechanism would be expected to give rise to an exponentially decaying rate of shelf subsidence upon which the effects of eustatic sea level changes may be superimposed.

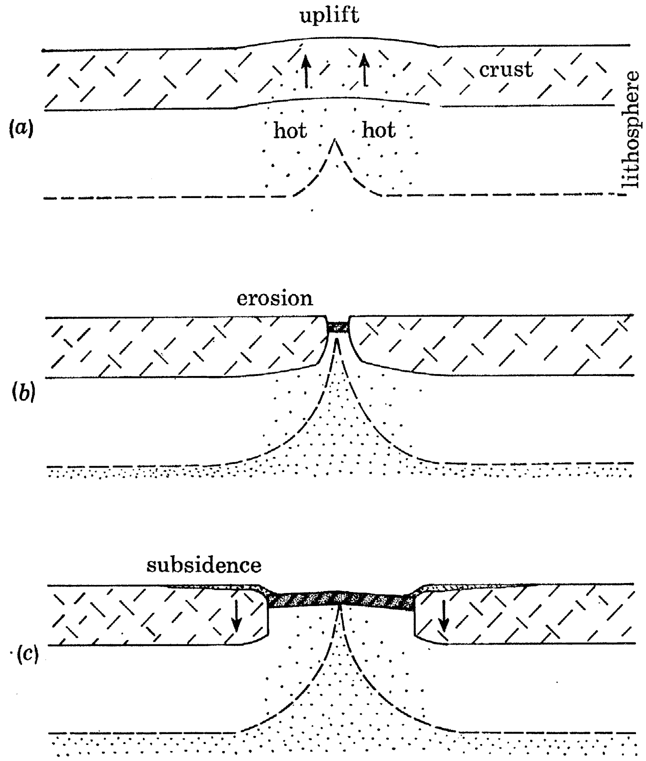

Figure 6. The thermal hypothesis of Sleep (1971). (*a*) Uplift following heating of the lithosphere. (*b*) Initiation of new ocean with erosion of uplifted continental regions causing crustal thinning. (*c*) Subsidence of continental margins as underlying continental lithosphere cools. Note the nearly vertical edge of the continental crust predicted by this hypothesis. After Bott (1979).

Modifications of the simple thermal hypothesis have been suggested in which the thermal event causes an irreversible increase in lower crustal density by metamorphism (Falvey 1974) or by igneous intrusions (Beloussov 1960; Sheridan 1969). According to Falvey, the rise in temperature causes a transition in the lower crust from greenschist to amphibolite facies.

Thermal hypotheses provide an elegant explanation of marginal shelf subsidence accounting for sedimentary thicknesses of up to about 4 km at the maximum, allowing for the gravity loading effect. They fail to explain some observed shelf sediment thicknesses of 5–15 km and are thus not the only factor. Sleep's version predicts a possible gap of about 50 Ma between the onset of spreading and the first marine sediments, followed by a decaying rate of subsidence. The deep crustal transition remains unmodified during subsidence.

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES
THE ROYAL SOCIETY A
PHILOSOPHICAL TRANSACTIONS OF

The crustal thinning hypothesis (Bott 1971) postulates that the continental crust beneath the shelf may progressively thin by oceanward flow of lower continental crustal material, causing the deep crustal transition to become progressively gradational (figure 7). This hypothesis depends on two premises, first that the lower crust is ductile rather than brittle as is indicated by experimental and theoretical considerations, and secondly that there is a stress system able to drive the flow. Bott & Dean (1972) showed that passive margins are associated with such a stress system as a result of the differential surface loading caused by seawater on one side and rock on the other, and the upthrust of the compensating root of low density crust on the continental side. Horizontal deviatoric tensions of up to 200 bar consequently occur in the continental crust, peaking at middle crustal depths and falling off towards the surface and the Moho.

(a)

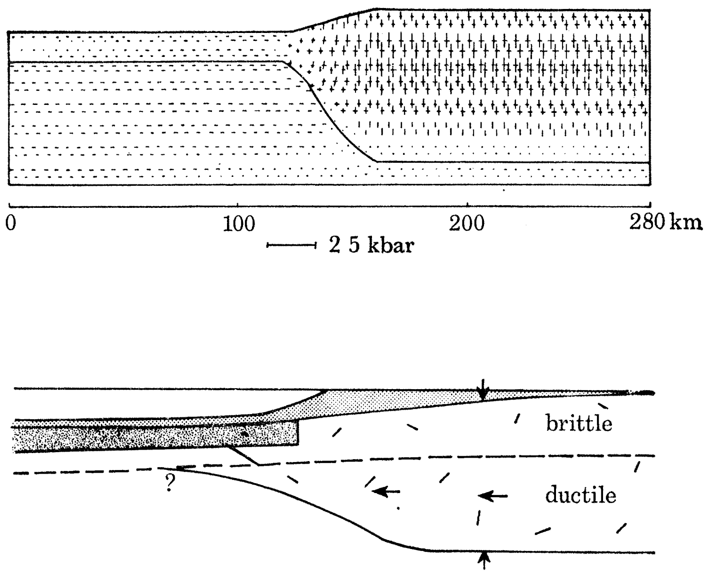

Figure 7. The crustal flow hypothesis of Bott (1971). (*a*) Supplementary compressive stresses resulting from the differential gravitational body forces across a passive margin; after Bott & Dean (1972). (*b*) Seaward flow of the lower and middle continental crust in response to this stress system, causing thinning of the continental crust and progressive gradation of the continent–ocean transition at depth. After Bott (1979).

The crustal thinning mechanism provides a ready explanation of shelf sediment thicknesses exceeding about 4 km. The mechanism is partly thermally controlled, in that crustal creep is a thermally activated process; it is partly stress controlled in that superimposed stress systems may alter the rate of flow, thus accounting for irregularly varying rates of subsidence with time. In contrast to the thermal hypothesis, crustal thinning should be marked by a gradational rather than sharp crustal transition at depth. Both the thermal and thinning mechanisms reduce the effective differential subsidence of the oceanic lithosphere as it cools.

Solving the puzzle

There has been much progress over the last ten years in understanding the formation and geodynamic development of passive margins, but several basic problems remain unsolved. These include (1) the mechanism of continental splitting and the nature of the associated igneous and tectonic activity, (2) the location and nature of the continent–ocean transition, and (3) the mechanisms of subsidence.

In solving these problems, deep drilling is likely to be of key importance in revealing the subsidence history of passive margins and in investigating the crustal transition just beneath the sediments. Complementary geophysical investigations are needed to map the structure of the sediment piles and to study the nature of the crustal and deep lithospheric transitions. This joint approach should provide further important clues to the puzzle of passive margins.

References (Bott)

Artemjev, M. E. & Artyushkov, E. V. 1971 *J. geophys. Res.* **76**, 1197–1211.
Beloussov, V. V. 1960 *J. geophys. Res.* **65**, 4127–4146.
Bott, M. H. P. 1971 *Tectonophysics* **11**, 319–327.
Bott, M. H. P. 1974 In *Geodynamics of Iceland and the North Atlantic area* (ed. L. Kristjansson), pp. 33–47. Dordrecht: D. Reidel.
Bott, M. H. P. 1976*a* *Anais Acad. bras. Cienc.* **48**, (Suplemento), 37–42.
Bott, M. H. P. 1976*b* *Tectonophysics* **36**, 77–86.
Bott, M. H. P. 1978 In *Crustal evolution in northwestern Britain and adjacent regions* (ed. D. R. Bowes & B. E. Leake), *Geol. J. spec. Issue* no. 10, pp. 377–392. Liverpool: Seel House Press.
Bott, M. H. P. 1979 In *Geological and geophysical investigations of continental margins* (ed. J. S. Watkins, L. Montadert & P. W. Dickerson). *Mem. Am. Ass. Petrol. Geol.* no. 29, pp. 3–9.
Bott, M. H. P. & Dean, D. S. 1972 *Nature, Phys. Sci.* **235**, 23–25.
Bott, M. H. P., Nielsen, P. H. & Sunderland, J. 1976 *Geophys. J. R. astr. Soc.* **44**, 229–238.
Bott, M. H. P., Sunderland, J., Smith, P. J., Casten, U. & Saxov, S. 1974 *Nature, Lond.* **248**, 202–204.
Bullard, E., Everett, J. E. & Smith, A. G. 1965 *Phil. Trans. R. Soc. Lond.* A **258**, 41–51.
Burke, K. 1976 *Tectonophysics* **36**, 93–112.
Campos, C. W. M., Ponte, F. C. & Miura, K. 1974 In *The geology of continental margins* (ed. C. A. Burk & C. L. Drake), pp. 447–461. New York: Springer–Verlag.
Collette, B. J. 1968 In *Geology of shelf seas* (ed. D. T. Donovan), pp. 15–30. Edinburgh: Oliver & Boyd.
Dingle, R. V. 1976 *Anais Acad. bras. Cienc.* **48**, (Suplemento), 67–80.
Dingle, R. V. & Scrutton, R. A. 1977 *Nature, Lond.* **268**, 720–722.
Falvey, D. A. 1974 *Aust. Petrol. Explor. Assoc. J.* **14**, 95–106.
Featherstone, P. S., Bott, M. H. P. & Peacock, J. H. 1977 *Geophys. J. R. astr. Soc.* **48**, 15–27.
Fuchs, K. 1974 In *Approaches to taphrogenesis* (ed. J. H. Illies & K. Fuchs), pp. 420-432. Stuttgart: Schweitzerbart.
Heirtzler, J. R. & Hayes, D. E. 1967 *Science, N.Y.* **157**, 185–187.
Hsu, K. J. 1965 *Am. J. Sci.* **263**, 97–109.
Keen, C. E. & Keen, M. J. 1974 In *The geology of continental margins* (ed. C. A. Burk & C. L. Drake), pp. 381–389. New York: Springer-Verlag.
Kinsman, D. J. J. 1975 In *Petroleum and global tectonics* (ed. A. G. Fischer & S. Judson), pp. 83–126. Princeton University Press.
Kusznir, N. J. & Bott, M. H. P. 1977 *Tectonophysics* **43**, 247–256.
Mayhew, M. A. 1974 In *The geology of continental margins* (ed. C. A. Burk & C. L. Drake), pp. 409–427. New York: Springer–Verlag.
Morgan, W. J. 1971 *Nature, Lond.* **230**, 42–43.
Rabinowitz, P. D. 1974 In *The geology of continental margins* (ed. C. A. Burk & C. L. Drake), pp. 67–84. New York: Springer–Verlag.
Rabinowitz, P. D., Cande, S. C. & LaBrecque, J. L. 1976 *Anais Acad. bras. Cienc.* **48**, (Suplemento), 241–251.
Rabinowitz, P. D. & LaBrecque, J. L. 1977 *Earth planet. Sci. Lett.* **35**, 141–150.
Roberts, D. G. 1975 *Phil. Trans. R. Soc. Lond.* A **278**, 447–509.
Sheridan, R. E. 1969 *Tectonophysics* **7**, 219–229.
Sheridan, R. E. 1974 In *The geology of continental margins* (ed. C. A. Burk & C. L. Drake), pp. 391–407. New York: Springer-Verlag.
Sheridan, R. E. 1976 *Tectonophysics* **36**, 113–132.
Sleep, N. H. 1969 *J. geophys. Res.* **74**, 542–549.
Sleep, N. H. 1971 *Geophys. J. R. astr. Soc.* **24**, 325–350.
Sleep, N. H. 1973 In *Implications of continental drift to the Earth sciences* (ed. D. H. Tarling & S. K. Runcorn), vol. 2, pp. 685–692. London: Academic Press.
Sleep, N. H. 1976 *Tectonophysics* **36**, 45–56.
Talwani, M. & Eldholm, O. 1972 *Bull. geol. Soc. Am.* **83**, 3575–3606.
Talwani, M. & Eldholm, O. 1973 *Nature, Lond.* **241**, 325–330.

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES
THE ROYAL SOCIETY A
PHILOSOPHICAL TRANSACTIONS OF

Turcotte, D. L., Ahern, J. L. & Bird, J. M. 1977 *Tectonophysics* **42**, 1–28.
Vogt, P. R. & Avery, O. E. 1974 *J. geophys. Res.* **79**, 363–389.
Walcott, R. I. 1972 *Bull. geol. Soc. Am.* **83**, 1845–1848.
Watts, A. B. & Ryan, W. B. F. 1976 *Tectonophysics* **36**, 25–44.
Wilson, J. T. 1963 *Nature, Lond.* **197**, 536–538.

Discussion

R. A. Scrutton (*Grant Institute of Geology, West Mains Road, Edinburgh EH9 3JW*). One of the mechanisms proposed to explain the subsidence of continental basement during the post-break-up phase of passive margin development is the down-dragging effect of the cooling, subsiding oceanic lithosphere. In the case where oceanic lithosphere is fused to continental lithosphere, Professor Bott suggests that the down-dragging stresses are released in the continental crust by extensive faulting (see figure 5). Because the oceanic lithosphere cools and subsides rapidly for 50–100 Ma following its formation, most of the faulting, which may have a total throw of 3 km, takes place during the initial post-break-up period.

As a result of this process, we would expect to see extensive faulting propagating upwards through the post-break-up, or drifting, phase sedimentary sequence that accumulates during the 50–100 Ma following break-up. This is not observed, however, since Professor Bott, Sir Peter Kent (this meeting) and others note that such extensive faulting is restricted to the rifting-phase sedimentary rocks, or to rocks older than Upper Jurassic or Lower Cretaceous. Passive margins that originated in or subsequent to the Cretaceous Period show very limited amounts of faulting in the drifting-phase sediments.

Does Professor Bott think that the observations fail to support the hypothesis?

M. H. P. Bott. This mechanism probably only applies exceptionally where shelf subsidence is anomalously small, such as is observed along parts of the margins of Rockall Trough. By analogy to the Carboniferous hinge lines of North Britain where differential subsidence of several kilometres occurs in probable response to basement faulting, it is possible that some or all of the basement faults do not penetrate into the sediments. The sediments may respond to such basement faulting by monoclinal flexure, but further investigation of this problem is needed.

M. F. Osmaston (*The White Cottage, Sendmarsh, Ripley, Woking, Surrey GU23 6JT*). Professor Bott has, here and on previous occasions, appealed to crustal thinning, by the oceanward creep of the lower part of the continental crust, as one of the possible causes of the observed subsidence of passive-type margins. Within such shelves, however, there are now many known examples of horst blocks, remarkably sharply defined by long-lived differential epeirogenic movement. This long-term differential movement must almost certainly stem from constitutional differences and changes in the lower crust (and below) and implies that the upper crust has remained in perfect registration with the material below, whereas the substantial differential horizontal flowage hypothesized by Professor Bott would generally have destroyed this registration.

The visual impression, given by geophysical sections, that the lower continental crust extends oceanward further than the upper continental crust, appears mainly attributable to the isostatic principle that large changes in Moho depth accompany comparatively slight changes in surface elevation. Consequently, what we chiefly need is a means of generating a magnetically quiet strip of crust of intermediate thickness along the continent–ocean boundary. As has been shown see, for example, Osmaston 1973), heavy sedimentation onto the strip of new floor, while continental separation was yet quite limited (and perhaps had even ceased for a while), would probably fulfil this need in the normal course of events.

Thus, while the oceanward creep of *upper* crustal materials (mainly through the mobility of salt and argillaceous sediments) is a well-established phenomenon, I believe the proposed oceanward creep in the *lower* crust to be both unnecessary and in substantial conflict with the evidence. I submit that the picture is not one of crustal *thinning*, but of substantial crustal *thickening* upon young lithosphere.

Reference

Osmaston, M. F. 1973 Limited lithosphere separation as a main cause of continental basins, continental growth, and epeirogeny. *Implications of continental drift to the earth sciences* **2** (ed. D. H. Tarling, & S. K. Runcorn), pp. 649–674. London: Academic Press.

M. H. P. Bott. The creep hypothesis involves ductile flow in the lower part of the crust, but not in the brittle upper part; thus flow in the lower crust should not give rise to any discordance between the sediments and the underlying basement as suggested by Mr Osmaston.

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES
THE ROYAL SOCIETY A
PHILOSOPHICAL TRANSACTIONS OF