Limited Lithosphere Separation as a main cause of Continental Basins, Continental Growth and Epeirogeny

Foreword

The original paper contained illustrative diagrams as Figs 1 to 3a. These have been omitted from this reprint but their descriptive captions have been retained as space-holders to show the adopted sequence of reasoning.

Under the heading 'Geological Reconstructions' three of these, namely *Scotland-Norway* (Fig 4), *Northern Appalachians-Newfoundland* (Fig 5), and *Gaspé-Long Range Peninsula* (Fig 6) have been omitted because subsequent exploration has rendered them insecure. But the one illustrated as Fig 7 *New Zealand/Campbell Plateau and the Ross Sea Basin* (involving Australia, New Zealand and Antarctica) has been retained with its related text.

M. F. OSMASTON

The White Cottage, Sendmarsh, Ripley, Woking, Surrey, England

Limited Lithosphere Separation as a main cause of Continental Basins, Continental Growth and Epeirogeny

Introduction

Recent work in the study of large scale horizontal motions of the lithosphere has led to progress in understanding the vertical movements of the ocean floor. This progress serves to emphasize how very little is yet understood about the vertical movements that affect the continental crust. Vertical movements have been of primary importance in the formation of many continental basins and geosynclines and in elevating the plateaux from which the sediments to fill them have often been derived. The geological record confirms the major, and at times even dominant, role that such movements must have played in the evolution of much of the continental crust to its present state. By extending the study of lithosphere separation to environments intermediate between those of ocean ridges and continental rift valleys, the opportunity occurs to make valuable progress with this fundamental problem. The paper concludes with brief examples relevant to the origin of the North Sea basin, the evolution of the northern Appalachians, and the origin of the Ross Sea basin in Antarctica.

Recent Work on Continental Epeirogeny and Growth

The following outline makes no claim to completeness, being intended to draw attention only to those aspects upon which this paper bears.

Compressive folding is directly responsible for only a small proportion of the present elevated land areas of the world, most such areas owing their present elevations to events having no obvious genetic connection with earlier periods of folding (Billings, 1960; Belousov, 1962; King, 1967). Itis widely recognized that subsidence mechanisms, other than mere isostatic adjustments due to sedimentary loading, are required to account for observed geosynclinal sediment thicknesses and for the large and long-continued subsidence of many continental basins and shelves (Jeffreys, 1959, p. 336 ; Belousov, 1962, 1966; Stoneley, 1969). Sub-crustal erosion, phase changes within the crust or underlying mantle, or combinations of these, have been invoked as subsidence mechanisms by various authors (Belousov, 1960, 1966; Gilluly, 1964; Van Bemmelen, 1966; Joyner, 1967; Van de Lindt, 1967; Collette, 1968; Magnitinsky & Kalashnikova, 1970; Sleep, 1971), but such hypotheses do not explain why subsidence should occur so selectively, frequently cutting sharply across structural belts, the physical state of whose underlying lithosphere would seem unlikely to vary so greatly or so sharply along the belt concerned. Also, the existence of continuing and appropriately intricate sub-crustal erosive movements within the continental lithosphere would hardly seem compatible within the increasingly secure framework of studies in plate tectonics.

Bott and his co-workers (Bott, 1964, 1971; Bott & Johnson, 1967; Bott & Dean, 1972) have confined their attention to differential vertical movements between crustal 'Units of differing character, proposing that the stress differences which arise result in the horizontal flow transfer of lower crustal or uppermost mantle material from one unit to the other. Such proposals were not offered as relevant to geosynclinal subsidence, nor, it seems, could they apply to the many basins which interrupt structural belts. Collette (1968) has also pointed out that, as the Moho of basins is shallower near the middle, density considerations would prevent migration of deep crustal material towards the margins.

Many authors have accepted that limited addition to the continental crust occurs by sedimentary outbuilding over oceanic crust, notably at river deltas. It is generally accepted that some additions to the area of the continental crust also occur along lines where oceanic lithosphere turns downward along Benioff zones, but this involves a characteristically intense folding and disruption ofthe material at the time of addition. While there is good evidence that this process has occurred in many fold belts (e.g. Dewey & Bird, 1970) we shall argue that, in areal terms, limited lithosphere separation with sedimentary and volcanic infilling may be a far more significant means of addition to continents.

Carey (1958) was the first to argue that sedimentary basins might be formed by crustal separation. Osmaston (I969) suggested briefly that the Verkhoyansk geosyncline might have been formed by separation of the Kolyma platform from the Angara and Aldan shields, and McGinnis (I970) proposed that limited separation could have caused some intracontinental basins in North America. Impetus to this approach has come from evidence that the small ocean basins behind island arcs, the crusts of which are intermediate between oceanic and continental in character (Menard, 1967) and exhibit high heat flow (Sclater & Menard, 1967; McKenzie & Sclater, 1968), have been produced by lithosphere separation (Karig, 1970, 1971; Matsuda & Uyeda, 1971; Packham & Falvey, 1971) as orginally suggested by Carey (1958). The possibility that sedimentary fillings of such basins could become folded during continental collision has been accepted in recent discussions ofgeosynclinal development (Mitchell & Reading, 1969; Bird & Dewey, 1970; Dickinson, 1971).

In view of this interest we discuss here the geophysics and some of the geological consequences of lithosphere separation in an essentially continental environment. The amount of separation envisaged in what follows is greater than that of continental rift valleys (and therefore forms a logical extension of a previous study (Osmaston, 1971», but is limited in the sense that separation ceases before separation has attained oceanic dimensions. A very important feature of our discussion is the attention given to the effects of heat conducted laterally into the older bounding lithosphere. Throughout this paper we ignore departures from isostasy such as may arise from differences between the time constants of the agent and the response processes. We accept, however, the evidence (James & Steinhart, 1966) that crustal thickness, crustal density, and mantle density, perhaps to at least the depth of the base of the lithosphere, all contribute to the attainment of isostatic balance.

Terminology - Chasmic Faults and the Lithosphere

When Carey (1958) proposed the formation of sphenochasms and rhombochasms by crustal separation he gave no name to the faulted boundaries thus produced. Wilson (1965), in introducing transform faults and their variations, seems to have been preoccupied with continuous separative movements. The term chasmic fault is therefore proposed to define any major age discontinuity extending through the lithosphere (Fig. I). Each such 'fault' might be single or a complex fault zone, depending upon the tectonic details of early separation. Thus the search for fits between continental shelf outlines has as its purpose the delineation of the corresponding pairs of chasmic faults. Similarly, if continental basins are formed by separation, chasmic faults may be a common feature of continental interiors. This is a significant matter for the study of epeirogenesis because, not only will chasmic faults initially divide young and hot lithosphere from that which is older and cooler, but also they will mark boundaries between areas of lithosphere that may differ greatly in composition and physical state and will therefore respond differently to changes in temperature at depth.

Nothing certain is known about motions and horizontal forces in the material beneath the lithosphere. Consequently, even if the seismic evidence used to define the base of the lithosphere (e.g. Kanamori & Press, 1970) is in fact relevant to the longterm creep properties of the material (which is disputable), it is at least possible that substantial thicknesses of sub-lithosphere material move integrally with the lithospheric plates they underlie. Therefore, although we shall use the term lithosphere separation, it should be borne in mind that this may affect material to depths beyond the base of the lithosphere as currently defined.

Crustal Thicknesses

Analysis of crustal seismic refraction measurements (Lee & Taylor, 1966) shows that the oceanic crust is 6-8 km thick and that the continental crust averages 35 km thick, but with wide variations (James & Steinhart, 1966) in which it appears that the crust of Precambrian shields and platforms tends to be at least 40 km thick.

The fact that the oceanic and the continental lithosphere are substantially in isostatic equilibrium suggests that if the mantle parts of both are not very different in composition, as the most recent studies suggest (Sclater & Francheteau, 1970), the emplacement ofa sufficient amount of sialic (and some basic) material on top ofoceanic crust might produce a suitable initial basis for the development of continental crust. Our concern here is with the means of such emplacement and with its consequences, and not with processes that may have the effect of adding material to the base of the crust. It has been argued (James & Steinhart, 1966; Drake & Nafe, 1968; Kosminskaya & Zverev, 1968; Sheridan, 1969; Osmaston, 1971, p. 397) that, at the base of the crust, material exhibiting Vp in the range 7·2-7·9 km/sec is probably altered mantle (e.g. partially amphibolized or serpentinized (Sheridan, 1969)) and may not be permanent. In particular, such material may be restored to more normal mantle velocity by sufficient orogenesis to cause upward loss of its volatile content. For this reason we shall regard V p > 7·2 km/sec as applying to subcrustal material.

Seismic refraction data, especially where these have been tied in with gravity data, show that the crust of major continental basins and continental shelves is markedly thinner than the overall continental average. Crustal thickness studies of small and land-surfaced sedimentary basins are limited both by the refraction line lengths required and by the need to place large explosive charges in bore-holes, so the relevant data are correspondingly under-represented.

Ocean-floor Subsidence

Plate accretionary models of ocean-floor spreading (Langseth et al., 1966; Vogt & Ostenso, 1967; Morgan, 1968; Sclater & Francheteau, 1970; Le Pichon & Langseth, 1970; Sclater et al., 1971; McKenzie & Sclater, 1971) show conclusively that the 3-4 km subsidence of the oceanic lithosphere (Fig. 1) during the first 100 m.y. after its emplacement at ocean ridge crests is due to its increasing density as it loses the extra heat it possessed at the time of emplacement. This density increase involves solid-state phase changes as well as thermal contraction. Estimates of oceanic plate thickness (which, as we have noted, may not be synonymous with lithosphere thickness) based on these studies are about 100 km, but this figure is sensitive to the considerable uncertainty in mantle thermal conductivity and water content and to the condition assigned to the sub-plate material.

Sclater & Francheteau (1970) argue that the continental parts of tectonic plates may be twice as thick as the oceanic parts, and petrogenetic evidence (Ringwood & Lovering, 1970) indicates that more than 300 km may be attained beneath shields. If the present argument is correct, that continents grow by incorporation of young lithosphere, major variations in lithosphere thickness within continents are to be expected.

FIGURE 1 Generalized vertical relationships in an ocean of Atlantic type, with wide continental shelves, after symmetrical lithosphere accretion has been proceeding for about 100 m.y. Note the large vertical

FIGURE 2a & FIGURE 2b

FIGURE 2a Direct result of arbitrarily closing up the ocean shown in Fig. 1 to represent the situation after less than 10 m.y. Oceanic crust, vertical ruling. M is base of crust. No allowance has been made for lateral heat transfer across the chasmic faults (see text).

FIGURE 2b Simplified result of separation similar to Fig. 2a but in a more continental environment providing abundant sediments, and taking into account the marginal upwarping resulting from heat transfer across the chasmic faults. Sediments (stippled) may include much additional volcanic material which would increase the density and thickness of the accumulation.

In succeeding sections of this paper we make the simplifying initial assumption that separation of any lithosphere has the consequences already established for the ocean floor, namely that fresh lithosphere is emplaced with an oceanic crust 7 km thick in a water depth of 2·5 km and subsides 3·5 km in 100 m.y. through loss of heat from its upper surface. It may be noted, however, that separation of lithosphere thicker than that of mid-oceans will draw material from greater depth, giving, in general, an increased heat content for the fresh material, and resulting in a shallower ridge crest, but that subsidence will eventually attain the same water depth as for an initially thinner lithosphere. The decrease of ridge crest depth with ridge spreading rate (Sclater et al., 1971) might be due to this effect if slow spreading produces a thicker lithosphere.

Early Lithosphere Separation

It is not at all clear that major downfaulting of the margins is an inevitable feature of the earliest stage of lithosphere separation, for dykes intrude the crust without being associated with such faulting and it has been shown (Osmaston, 1971) that on ocean ridges important support forces are provided by the adjacent new upwelling material. Moreover, in the presence of heavy sedimentation, the sediments will buttress the original margins (Osmaston, 1969). It seems desirable, therefore, to keep an open mind on this point until the facts can be ascertained in a sufficient range of geological situations. Studies of structural continuity across restored chasmic faults could do much to resolve this problem.

Figure 2a has been obtained by simply closing up Fig. 1 to represent an earlier stage of separation (e.g. 100-200 km). Sedimentation on to the new floor is assumed small, due either to wide shelves or to low erosion rates. It is, however, quite certainly an incorrect picture, because at this early stage the fresh lithosphere will be much hotter than the older lithosphere on either side. Major heat transfer across the chasmic faults will therefore take place and the shelf margins will become upwarped isostatically by the density reductions thus induced at depth beneath them.

An upwarp of this kind is in fact a well documented and persistent feature of the edge of the western North Atlantic continental shelf (Officer & Ewing, 1954; Emery, 1965; Berger et al., 1966; Drake et al., 1968 ; Sheridan, 1969). In most cases the up-warping affects Upper Cretaceous strata and has caused substantial ponding of later sediments. The loading provided by these would tend to perpetuate this structural relationship even after general subsidence had become re-established. Studies of the eastern shelf of the Atlantic from Lisbon to the Faeroe Isles (Stride et al., 1969) have revealed massive Late Cretaceous-Early Tertiary erosion of the shelf edge, before re-establishment of subsidence. North of the Azores fracture zone the age of this upwarping correlates well with the initiation of separation inferred from oceanic magnetic anomalies (Heirtzler, 1969; Pitman et aJ., 1971). Further south, a pre-Late Jurassic upwarp east of Florida (Sheridan, 1969) may also correlate with early separation there.

In Fig. 2b we illustrate separation similar in amount to that of Fig. 2a, but in an essentially continental environment. The upwarping of the margins has also been taken into account. It shows the sediment thickness required to replace the sea-water (2·5 km depth assumed) in Fig. 2a, without allowing for any heat-loss-dependent subsidence. A normal erosion rate of 80 m/m.y. (Schumm, 1963; Judson, 1968; Gilluly et aJ., 1970) from a belt 500 km wide would keep such a trough filled to sea level at a lithosphere separation rate of 0·5 em/yr. If allowances are made for increased erosion due to uplift, or due to the separation being located in a young mountain belt, and for volcanic contributions to the filling (see below), separation at perhaps ten times this rate could be accommodated without deep-water sedimentation being involved. Thus it appears entirely possible that in small basins the whole of this initial filling could be of continental facies. In larger basins this would also apply to the early-formed marginal areas of the basin.

At this stage in our analysis we have produced a simplified crust some 15 km thick, with the effects of thermal subsidence still to be considered. We may note here, however, that the effect of heavy sedimentation during lithosphere separation would certainly greatly modify the magmatic aspects of the process as compared with normal oceanic crust production. The high cooling rates at ridge crests (Oxburgh & Turcotte, 1969; Osmaston, 1971) would be much reduced by the sedimentary blanket and volcanic activity, producing sills, dykes and perhaps larger intrusions within the sediments, would probably continue for a considerable time after leaving the axial zone. The generation of oceanic-type magnetic anomalies would be most improbable owing to the maintenance of high temperatures for periods much longer than those between geomagnetic reversals. The result would be a crust having high geothermal gradients and probably somewhat thicker (basic intrusions being denser than sediments) than in the simplified model of Fig. 2b.

Post-separative Subsidence

FIGURE 3a & FIGURE 3b

FIGURE 3a Intermediate stage of post-separative subsidence of the basin floor, showing the effect on crustal thickness of accelerated subsidence near the margins, due to lateral heat loss across the chasmic faults.

FIGURE 3b Final stage of basin subsidence, showing the total resulting crustal thickness, on the simplified assumptions discussed in the text. This is probably a minimum value for real conditions. Note the structural sag resulting from the later completion of subsidence near the middle of the basin. Note also the subsidence of and possible marine transgression across the bounding margins of the basin.

If cooling of the new lithosphere proceeds to the same condition as that attained beneath the ocean floor after about 100 m.y. (but generally more slowly owing to the thick crustal blanket), the subsidence would permit the accumulation of a further 11 km of sediments, yielding a simplified crust with a total thickness of 26 km (Fig. 3b). In this calculation the density of the mantle was assumed to have risen to 3·33 g/cm^3 during cooling, and the overall mean 'sediment' density to 2·6 g/cm^3, owing to compaction and thermal metamorphism of the lower parts of the accumulation. .

We have thus already attained a crustal thickness typical of many basins, particularly if the probable presence of an additional igneous contribution is also taken into account (see preceding section). Some of the details of the basin subsidence stage, however, appear relevant to comparison with actual basins and will now be discussed briefly.

Most continental basins are of dimensions such that the separative phase may have lasted only a few million years at typical average rates of relative plate movement. The opening of the 500 km wide North Sea basin, for example, might have been accomplished as a single movement in under 10 m.y. Therefore, although subsidence due to heat loss begins to affect new lithosphere from the very moment of its emplacement, subsidence will in general continue for so much longer than it took to open the basin that we may regard the lower crust of the basin as being of substantially uniform age across the basin.

Figure 3a shows an intermediate stage in the subsidence process. This draws attention to the effects of lateral heat loss across the bounding chasmic faults. This loss will be most rapid at the margins, decreasing towards the middle of the basin, with the result that the subsidence of the basin crust near the margins will initially be faster, running toward completion sooner than further out in the basin. At this stage, therefore, the basin crust will be thinner near the middle of the basin (Fig. 3a), even if sedimentation has maintained a substantially level upper surface. Notice that outward transfer of heat across the chasmic faults will involve simultaneous subsidence of the basin and (except insofar as the heat merely maintains previously-elevated temperatures) uplift of the bounding areas. This is one of the most widely evident geological features of basin formation. The chasmic faults will be the main sites of such differential movement, resulting in displacements of a normal fault character extending throughout much of the subsidence life of the basin.

As the heat from beneath the new basin spreads into the older lithosphere on either side, the upwarp, initially of the form shown on the left in Fig. 3a, may be expected to extend to greater distances and take the form shown on the right wherever the lithosphere is especially thick.

Dissipation of the excess heat present in the lithosphere of the new basin will proceed much more slowly as the process nears completion. Subsidence rates will remain greater, however, where the largest amount of undissipated excess heat still remains, namely at the centre of the basin (Fig. 3b), thereby developing the characteristic structural sag which we believe has proved so misleading in previous attempts to understand basin formation. At about this time, a substantial measure of thermal equilibrium across the chasmic faults having been attained, both old and new lithosphere at the basin margins will subside together, yielding a marine transgressive phase in appropriate circumstances.

Subsidence Rates

We now consider briefly the four types of epeirogenic process in which an increase of density results from the extraction of heat. These are (1) thermal contraction, (2) liquid-to-solid phase change, (3) solid-state phase changes, and (4) reactions involving combination of the rock with a volatile phase (usually $H_2 0$) that was already present in its interstices. The first and third are strictly reversible, but the second and fourth may be only partly so, owing to the possibility of the released liquid or volatile component migrating toward the surface on reheating. It should be noted particularly in relation to hydroxylation processes that in the study of epeirogenic movement we are concerned with changes in volume of columns of material extending to great depths, and commonly of considerable horizontal extent, so hypotheses involving introduction of the $H_2 0$ from 'elsewhere' are quite inappropriate to the problem, except, perhaps, in the case of the top kilometre or two of the oceanic crust. The horizontal migration of sedimentary pore fluids, familiar to the petroleum industry, and fluid loss to the surface during sediment compaction, do, of course, contribute to observed vertical movements but will not be considered explicitly here.

Epeirogenic processes differ widely in the amount of heat that has to be extracted to produce a given amount of contraction. To evaluate this we define the thermal sensitivity (B) of the process as the volume change resulting from unit heat loss. Then, for thermal contraction we have

$$\mathbf{B} = \frac{\alpha_v}{\sigma_v} \qquad (i)$$

where α_v is the volume coefficient of thermal expansion and σ_v is the volumetric heat capacity.

For processes of types (2), (3) and (4) we use a form of the Clausius-Clapeyron thermodynamic relation (Ramberg, 1952, p. 16; Fyfe et al., 1958, p. 115; Turner & Verhoogen, 1960, p. 22) to yield

$$\mathbf{B} = \frac{1}{\mathrm{T_e}} \cdot \frac{\mathrm{dT_e}}{\mathrm{dP}} \qquad (ii)$$

in which dT_e/dP is the slope of the equilibrium boundary between the two states on a pressure-temperature diagram, and T_e is the absolute temperature of transition at the chosen pressure. The first thing to notice about this relation is that, because dT_e/dP is almost invariably positive, the inverse dependence on T_e results in values of B that tend to decrease with increasing depth. Relation (ii) is applicable to multicomponent mineral assemblages provided that the equilibrium slope has been determined in the presence of all participant phases. This proviso is a substantial cause of uncertainty owing to uncertainties in the compositions of the mantle and crust. The relevant characteristics of the process can be changed markedly by the presence or absence of quite small proportions of some constituents.

For this reason the values of B given in Table I have been selected from a fuller review of thermal epeirogenic processes being compiled by the present author and which will be published separately in due course. Such a procedure is to some extent justified by the differences in the magnitude of B for the four types of process.

TABLE 1 Some Properties of Thermal Epeirogenic Processes

Process	B 10^{-3} cm^3/cal	Specific Subsidence Rate* m/m.y. per μcal/cm^2 sec	
		Under water	Under sedimentation to constant level
Thermal contraction	0.032	15	47
Solidification	0.25	120	380
Hydroxylation reactions	0.5	240	760
Solid-state phase changes	4.0	1900	6000
Real ocean floor 0-10 m.y.		30-60	95-195

*Assumed densities: sediments, 2'5; mantle, 3'2; sea-water, 1'03.

The value of B given for thermal contraction relates to mantle material and is probably a good average down to 75 km depth, beyond which a decreasing value is to be expected. Within the continental crust the value may be as much as 30% lower but this depends on the composition. The value for solidification is affected by changes in the water content of the melt and therefore upon the degree of melting. The figure quoted has been estimated for mantle material at a depth of 50-100 km and a degree of melting of 10 wt %. The value for hydroxylation reactions relates to depths of 15-20 km, but individual reactions differ from this by factors as large as two. Note also that the low density of steam at low pressures results in a rapid increase in B at shallower depths, but at greater depths B will in most cases decrease toward zero on approaching the depth limit for stability of the hydroxylated mineral concerned (probably less than 75 km for most of those likely to occur in sufficient abundance to be epeirogenetically significant).

Compositional uncertainties are of particular significance to the value of B for solid state phase changes. Most such changes are complex reactions between many participant minerals and involve solid solution effects, the limits of which are sometimes intimately dependent upon even quite minor constituents of the assemblage. Moreover, reaction rates are commonly extremely slow below about 900°C and extrapolation of laboratory results to lower temperatures (relevant, perhaps, to conditions in the lower part of the continental crust) is fraught with uncertainty. These problems are well illustrated by recent discussions (MacGregor, 1970; O'Hara et al., 1971). The expected presence of H20 or hydrated minerals in many situations is a further complication. From data currently available it appears that phase changes relevant to the crust and mantle to depths of 150 km may have values of B in the range (0.5-8.0) X 10-3 cm3/cal. The value given in Table I should therefore be treated with corresponding caution. It is also important to stress that most phase changes require a considerable change of temperature for completion, so their apparently high thermal sensitivity for epeirogenetic purposes is in fact much reduced by the heat which has to be extracted to achieve the required cooling.

We now consider the surface subsidence rate that would occur if heat flow were entirely vertical and entirely attributable to one of these four types of subsidence process. For unit heat flow (10-6 cal/cm2 sec) we then have the specific subsidence rate for the process under the stated conditions. In Table 1 the values of this parameter have been calculated for subsidence under sea-water and under constant-level sedimentation.

In an actual case, the measured heat flow at a point on the Earth's surface may be regarded as having two principal components, one of which is continually replaced by radioactive heat generation within the entire underlying column of material, and therefore involves no subsidence.* The other component may be termed the cooling component and involves subsidence due to some combination of the processes listed in Table 1 appropriate to the distributions of temperature and composition within the column concerned. Since it is a characteristic of rock columns extending through a large depth range that, on cooling, they undergo contractions in addition to pure thermal contraction, it follows that after subtraction of the radioactive component the calculated specific subsidence rate (still assuming vertical heat flow) must always exceed that for pure thermal contraction. In Table I, the approximate figure for the specific subsidence rate of 'real ocean floor' during its first 10 m.y. after emplacement allows an arbitrary 0.9 X 10^{-6} cal/cm^2 sec for the radioactive component and is based on subsidence rates, heat flow, topographic profiles and spreading rates taken from the literature (Lee & Uyeda, 1965; Von Herzen, 1967; Vogt & Ostenso, 1967; Sclater & Francheteau, 1970; Sclater *et al.,* 1971), and we may note that it does in fact exceed the pure thermal contraction rate by a factor of two to four.

* This statement would be strictly correct only if α_v did not vary with depth or if the heat loss and its replacement by radioactivity occurred at the same levels. In young lithosphere, the heat loss will generally occur at shallower depth than that at which it is replaced radioactively, the differences in α_v thus resulting in a small subsidence rate, here neglected.

For our present purpose, however, the significant figure is the one in the right-hand column purporting to represent the specific subsidence rate of young 'ocean floor' if it were subjected to continuous sedimentation to sea-level. It has been derived from that for submarine ocean floor by applying the appropriate isostatic correction factor (3.15) for sediment loading. Implicit in this calculation is the assumption that the same proportionate mixture of epeirogenic processes would occur in both circumstances. This is clearly improbable owing to the effects of sedimentary blanketing and the depression of the crustal base to deeper levels. However, it is at least a first approximation to the maximum specific subsidence rate that is likely to result from heavy sedimentation on to young lithosphere. Ocean floor heat loss rates and subsidence rates fall off rapidly with time but it is conceivable that under sedimentation a heat loss rate through the sediments might be maintained at up to 3 X 10^{-6} cal/cm^2 sec for the first 10 m.y. Taking the maximum specific subsidence rate (195 m/m.y. per unit heat flow) the total sediment thickness after 10 m.y. would be nearly 14 km, and the temperature at the base of the new crust would exceed 1000°C, if allowance is made for radioactive heat production in the sediments. This is clearly stretching the figures to rather improbable limits. We tentatively conclude, therefore, that the maximum rate of subsidence under sedimentation, sustainable over a 10 m.y. period, and (it must be remembered) on the assumption that all heat loss from the new lithosphere is vertical, is of the order of 400 m/m.y.

Arguments have already been given for the occurrence of substantial heat transfer across chasmic faults. The amount of such heat transfer is difficult to quantify at present but it is clear that it will have most effect on the rate of subsidence of the new lithosphere when the amount of separation is small, with a fairly rapid diminution of the effect when the separation is comparable with or exceeds the thickness of the old lithosphere on either side. Accordingly, in narrow sedimentary troughs formed by lithosphere separation very considerable enhancement of the subsidence rate inferred above is to be expected.

These conclusions appear remarkably consistent with geologically deduced rates of accumulation in geosynclines (Kay, 1955; Sutton, 1969; Ziegler, 1970). Such rates are not necessarily the rates of subsidence except in the few cases where depositional depths at both ends of the sequence have been ascertained. It appears that accumulation rates of the order 150 m/m.y. for 50 m.y. are fairly common; for periods of 20 m.y., rates up to 600 m/m.y. are found, and the highest rate so far recorded is 1400 m/m.y. for 7 m.y. in the Fossa Magna of Japan.

In view of this accord it appears that the subsidence responsible for many geosynclinal accumulations could have been the direct consequence of lithosphere separation. It should be noted that the initial 8 km, or so, of sediments and/or volcanics (Fig. 2b) needed to bring the sedimentation surface up to sea level will accumulate at rates unrelated to thermal epeirogenic subsidence rates, but it seems possible that this deep part of the crust is as yet unexposed in any of the geosynclines of Phanerozoic age, to which data on accumulation rates are at present restricted.

Folding

Compressive folding of the crust of a basin formed by lithosphere separation would thicken the crust, and the associated metamorphism would increase its mean density. Thickening the crust would tend to raise its surface above sea-level, but the increase in density would offset this to some extent. There is thus no obvious reason why basin crust should not ultimately develop into the thick, dense and complexly deformed crust typical of Precambrian shields.

At present, almost nothing is known about the forces either available or required for the folding of continental lithosphere. The absence of any evidence of folding of the oceanic crust implies either (a) that young continental lithosphere is much softer than young oceanic lithosphere or (b) that compressive forces (if any) developed by the ocean ridge/Benioff zone system are small and perhaps independent of those responsible for the compression of wide basins (e.g. the Late Mesozoic folding of the Verkhoyansk geosyncline (Nalivkin, 1960)). We cannot discuss here the genesis of horizontally compressive geotectonic stress fields, but would emphasize that the detailed distribution of folding of anyone age in actual fold belts often suggests that the response to such stress fields has been greatly influenced by sharp variations in lithosphere condition such as might have been inherited from preceding separative movements.

At some stage in its cooling the continental lithosphere probably becomes too stiff to be folded by available forces unless some form of thermal 'softening-up' process takes place first. This might be achieved either by separative formation of an adjacent geosynclinal trough or by Benioff zone volcanism.

In other cases the continental basin crust will be folded before its lithosphere becomes too stiff, which will mean that cooling is incomplete. This will have two important consequences. Firstly, if the crust is still thin and the foldjng involves only minor crustal shortening, this part of the fold belt may never appear above sea level. Secondly, even if the crust is thick enough to result in an elevated mountain range, the continuing cooling of its lithosphere will result in continuing subsidence. It is easily shown that isostatic adjustment for removal of crust by surface erosion would require that 500-1200 m of material (depending on densities assumed) be removed in order to lower the surface by 100 m. The widely-known fact that many mountain ranges have been eroded to near sea-level, but have obviously not suffered the removal of anything like the isostatically requisite amount of overburden, seems good evidence for the continuance of thermal subsidence after folding.

A succession of lithosphere jostling movements, involving alternate separation and approach, could result in a belt of tightly folded crust interpretable as having originally thinly floored a basin of very great width, when in fact the maximum separation between the tectonic plates concerned may at no me have exceeded a fraction of this amount. This might explain how some wide belts of Precambrian rocks possessing rather uniform strike across their entire width may have been formed.

Limited separation, followed by compression, has two other important characteristics. The first is that the alignment of older tectonic trends on opposite margins of the new folded basin may be substantially preserved. The frequent occurrence of such relationships in the Precambrian of Africa does not encourage the view that such fold systems were produced by oceanic closure because of the improbability that this would restore pre-existing relationships so closely (Hurley, this volume, p. 1083; Shackleton, this volume, p. 1091). The second characteristic concerns the probable non-development of a Benioff zone and its corresponding crustal descent line, or suture, owing to the presence of relatively thick continental basin crust, not oceanic crust, between the convergent plate margins. Crustal shortening undoubtedly requires downward disposal of excess subcrustal lithosphere, but the mere occurrence of folding within the basin presumably implies that the mantle part of the basin lithosphere is still soft enough to be extruded downwards as compression proceeds. The difficulty of finding suture lines in some Precambrian fold belts (Burke & Dewey, this volume, p. 1035) may thus be due to their non-occurrence.

Lateral Heat Flush and the Genesis of Granites

Various consequences of the transfer of heat across chasmic faults in simple separative situations have already been mentioned. If, however, the lithosphere break-up is complex, involving complete detachment of 'islands' of old lithosphere, these will appear as inliers within the basin. Because, unlike the main margins of the basin, they are completely detached from an extensive heat sink consisting of old lithosphere, the lithosphere of these inliers will attain higher temperatures, causing their tops to be uplifted even more than the main margins.

It is now well established from theoretical studies of the separation of 100 km thick oceanic lithosphere that, at a depth of 30 km, temperature excesses (relative to those inferred for oceanic lithosphere more than 100 m.y. old) remain in the range 550200°C during the first 30 m.y., or so, after emplacement. These figures illustrate the character of the heat input to the lower part of any mature continental crust bounding a new basin. The greater thickness of continental lithosphere would, as already mentioned, probably increase the heat content of the young lithospheric material, but, on the other hand, lateral heat losses will cause the excess temperatures near the basin margins to fall away more quickly. Nevertheless it is quite clear that these conditions imply that any crust of mature continental thickness (say 35-45 km) bordering a new basin will experience a major thermal event and that thermal metamorphism, with the rise of granites and pegmatites, may well occur. These effects are particularly to be expected in the case of wholly detached blocks of lithosphere.

It is proposed, therefore, that many post-tectonic granites, and others whose initial mobilization occurred in the absence of compressive stress, may owe their occurrence to the thermal effects of nearby lithosphere separation. This is not to deny that some granitic plutons (e.g. the Tertiary plutons of western Scotland) may result from heat brought in by local volcanism. On the other hand there is now some reason to question whether the undoubted connection between some granites and the presence of a Benioff zone is the whole story, or whether many of these granites are in fact products of the recently recognized tendency (mentioned briefly at the beginning of this paper) for separative movements to occur at the rear of such zones.

Geological Reconstructions

An obvious test of whether continental basins have in fact been formed by lithosphere separation is that the movement proposed in any particular case should not only account for the existence and subsidence of the basin but that reconstructions on this basis should be consistent with basin geometry and with pre-basin geological structures. We now discuss briefly three examples, selected to illustrate different aspects of the arguments given in preceding sections. To justify each fully from all the data available is clearly impossible within the compass of the present paper and no attempt will be made to do so.

FIGURE 4 Proposed relationship of northeast Scotland to southwest Norway before formation of the North Sea basin. Reconstruction is not drawn to a common projection but from separate maps at I: 1,000,000 scale (Oxford Atlas, 1957 for Scotland; Geological Map of Norway, Holtedahl & Dons, 1960). Black, basic intrusions; stipple, granites; oblique ruling on Karm0Y, augen gneiss and breccia. Cross-barred lines are limestone outcrops in Caledonian fold belt. Places marked are: Aberdeen (A), Bergen (B), Stavanger (S), and Utsira island (U). Utsira is shown in its present relationship to Norway, with an arrow suggesting possible displacement. North vectors indicate present north direction on Norway and northeast Scotland. Heavy broken line is outer Bergen arc. Heavy continuous line is northwest limit of Cambro-Silurian 'allochthon' in Norway. Dotted line south of Stavanger shows limit of Cambro-Silurian rocks.

(a) Northeast Scotland and Norway (Fig. 4). The subsidence of the North Sea basin has been spectacularly proven by the recent search for oil and gas. The base of the Permian now lies at more than 5 km depth near the middle of the basin but crops out above sea level in northern England (Kent, 1967; Birch, 1969). The uplift, metamorphism and extensive granitic intrusion of the Dalradian rocks of the Scottish central highlands present a major geochronological problem. Recent discussions have been given by Brown & Miller (1969), and by Dewey & Pankhurst (1970). The gabbros in northeast Scotland have been well dated by Rb-Sr whole rock isochron at 486 ± 17 m.y. (Pankhurst, 1970) and approximates the date of main folding. Geochronological studies in southwestern Norway are still at a very early stage (M. R. Wilson, I. Pringle & I. Bryhni, personal communications, 1972) so the correlation suggested by Fig. 4 cannot yet be discussed.

Dewey & Pankhurst (1970) relate the apparently continuing rise of granites in the Scottish Central Highlands during the period 460 m.y. to 380 m.y. to Benioff zone activity during closure of a proto-Atlantic ocean. The present proposal is that, whether or not there was a Benioff zone there at that time, many of the granites and the uplift responsible for their successive cooling ages were due to lateral heat flush caused by several isolatory separative movements, of which the separation from Norway was one of the last. In Fig. 4 the only Scottish granites shown are those with concordant Rb-Sr and K-Ar ages of 404 m.y. (uncertainties ± 13 m.y. and ± 5 m.y. respectively) (Dewey & Pankhurst, 1970, p. 380). All are within 40 km of the Aberdeenshire coast. Similarly, of the Norwegian granites shown in Fig. 4, only those within 40 km of the coast are mapped (Holtedahl & Dons, 1960) as post-tectonic. It is suggested, therefore, that separation from Norway started shortly before 404 m.y. ago. The breccia on Karm0Y might be related to this event. Even the 15-20 km of erosion of the Scottish unit, envisaged by Dewey & Pankhurst (1970, p. 383), could have been accomplished by 2-3 km of thermal uplift, as is typical of present-day rift-valley upwarps, accompanied by isostatic adjustment as erosion proceeded.

The details of the relative positions shown in Fig. 4 were decided quite arbitrarily, although with a view to the eventual incorporation of the Scottish Southern Uplands in a suitable position. Offshore marine geology for the Norwegian coast was not available to the writer but could obviously improve control if inliers of a dislocated character can be identified as such. The relative rotation shown (33°) is larger than, but in the same sense as, that (21°) obtained from the palaeomagnetic results for the Arrochar igneous complex in Scotland (dated at 418 m.y.) (Briden, 1970) and the Late Ludlow/Early Downtonian Ringerike sandstone of the Oslo region (Storetvedt *et al.*, 1968).

(b) *Basins in the northern Appalachians* (Fig. 5). The somewhat unconventional nature of this map is explained in the caption. Its main purpose is to suggest the formation of a complex of basins, notably the Gulf of Maine and the New Brunswick Basin and parts of the Gulf of St. Lawrence, by a general east-northeasterly movement relative to the Canadian shield, and probably derived from the motion of a major lithospheric unit originally forming the southeastern margin. Such a map obviously omits a number of small areas in which old rocks are known to underlie the cover but these are unlikely greatly to affect the general picture.

The first major problem is to infer when the movements might have occurred. Black (1964) on limited palaeomagnetic evidence, inferred a 30° anticlockwise rotation of Newfoundland at the end of Devonian time. This, and the extensive Carboniferous sedimentation in the basin areas, sets the minimum age of movement. A maximum age is suggested by Fig. 6 in which the mafic, ultramafic and cYastic Taconian allochthon of western Newfoundland (Kay, 1969; Williams, 1971) was not in fact transported from the east but is the endwise overflow of the Gaspé Peninsula Taconian fold belt (which also contains ultramafics). Figure 6 shows a rotation of $22\frac{1}{2}°$ for the Long Range Peninsula of Newfoundland, but Belle Isle, shown in black in its present position relative to the peninsula, then overlaps the mainland, implying either that Belle Isle was displaced during the movements or that the peninsula plus Belle Isle should be rotated rather more than shown in Fig. 6.

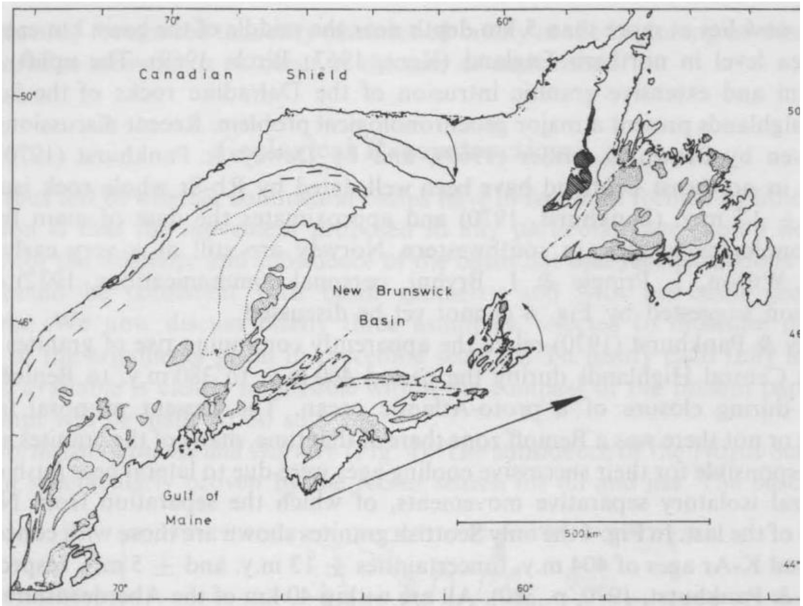

FIGURE 5 Greatly simplified geological map of the Northern Appalachian-Newfoundland region. Areas affected by post-Lower Devonian subsidence have been excluded from the outlines. Acadian granites and granodiorites, stippled. Principal steep or vertical faults, dash-dot lines. Dashes show Taconian (mid-Ordovician) tectonic trends along Appalachian front. Oblique ruling indicates Humber Arm Taconian allochthon in Newfoundland. Data taken from Tectonic Map of North America (King, 1969) and replotted on a Bonne's projection of the area (Debenham, 1962, p. 82) to minimize distortion. Approximate outline of the Matapedia basin, interpreted by Bird & Dewey (1970, p. 1049) as an inter-arc oceanic basin of post-Taconian age, is shown dotted and discussed in the text. The large arrow shows suggested total (Devonian?) motion of a tectonic plate initially attached to the entire southeastern margin of the region. Small arrows show inferred subsidiary motions.

If Fig. 6 is accepted in principle it is at once obvious from Fig. 5 that the western boundary of the V-shaped New Brunswick basin also needs to be rotated clockwise if it is to accommodate at this time the rest of Newfoundland (assuming any necessary slippage along the Longe Range fault system traversing the island). Closure of the Matapedia basin, which appears to be of immediately post-Taconian age (Bird & Dewey, 1970; McKerrow & Ziegler, 1971), could make such a fit possible, however.

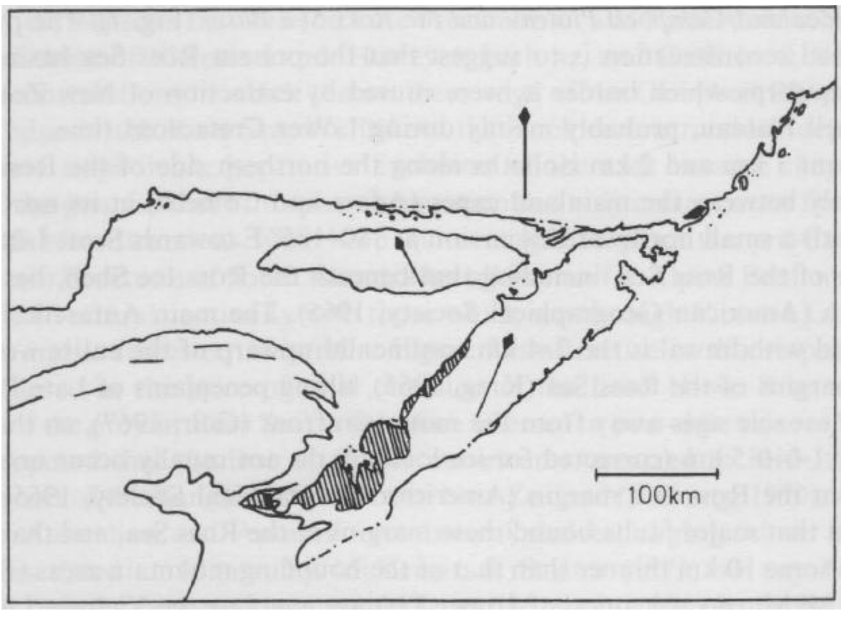

FIGURE 6 Proposed reconstruction of part of Fig. 5 for Mid-Ordovician time, showing relationship between the western Taconian allochthon of Long Range Peninsula and the end of the Gaspe Taconian fold belt. North vectors show present north directions on units involved . Position of Anticosti Island is not closely controlled but that shown would result in basin formation only to the north of the island, witi1 movement on the south side being along a curved fault. The position of Belle Isle (black, superimposed on mainland) is discussed in the text.

Accordingly, as a working hypothesis for closer investigation it is suggested that the movement of Newfoundland took place in two stages. The first, of immediately post-Taconian (i.e. Upper Ordovician) age, opened the Matapedia basin, thus rotating Newfoundland anticlockwise and detaching its western coast from the end of the Gaspé peninsula. The second, of possibly Lower to Middle Devonian age, may have been the main movement which extracted Newfoundland from the New Brunswick basin, and probably also produced the Gulf of Maine. If this is correct, then protoAtlantic closure (Wilson, 1966 ; Dewey, 1969) must have been completed in or by Lower Devonian time because of the difficulty, that would otherwise arise, of producing identical movements on opposite sides of an ocean basin. The continuance of granite dates far into the Devonian could then be the result of lateral heat flush. In this connection it is interesting to note the marginal locations of many of the Acadian granites.

Finally, we return to the matter, mentioned on pp. 643-4, of the low subcrustal seismic velocity (7.2-7.8 km/sec) found beneath much of the basin areas discussed here. Our inference that the crust of these basins has not been subject to folding and metamorphism since formation gives weight to the idea that the presence of volatiles, either combined or free interstitially, may cause the low velocity. The fact that such velocity does not usually occur beneath deep oceans, however (Drake & Nafe, 1968; Maynard, 1970) where folding and metamorphism are also absent, but is found near the axes of ocean ridges, suggests that temperature is the significant factor. Retention of the low velocity beneath basins and continental margins would then be due to the higher sub-crustal temperatures resulting from the greater depth to which the material is depressed. Accordingly, it is suggested that the low velocity may be due to the presence of small amounts of intergranular volatiles. The work of Anderson & Spetzler (1970) on the effects of inter granular films suggests that the amounts required might be volumetrically quite trivial. Nevertheless, if this inference is correct, low-temperature hydroxylation processes may play a rather less important part in basin subsidence than has been suggested.

(*c) New Zealand/Campbell Plateau and the Ross Sea Basin* (Fig. 7). The purpose of this simplified reconstruction is to suggest that the present Ross Sea basin, and the mountain upwarps which border it, were caused by extraction of New Zealand and the Campbell Plateau, probably mainly during Lower Cretaceous time.

The present 1 km and 2 km isobaths along the northern side of the Ross Sea run fairly directly between the mainland capes (Adare and Colbeck) at its northern corners, but with a small northward excursion at 180-185° E towards Scott Island. Most of the floor of the Ross Sea, including that beneath the Ross Ice Shelf, lies at about 500 m depth (American Geographical Society, 1965). The main Antarctic feature of the proposed withdrawal is the 2-4 km continental upwarp of the entire western and southern margins of the Ross Sea (King, 1965), tilting peneplains of Late Palaeozoic and Mid-Mesozoic ages away from the mountain front (Gair, 1967), so that sub-ice altitudes of 1.0-0.5 km (corrected for ice loading) do not usually occur until at least 300 km from the Ross Sea margin (American Geographical Society, 1965). Gravity data suggest that major faults bound these margins of the Ross Sea, and that the Ross Sea crust is some 10 km thinner than that of the bounding mountain areas (Robinson, 1964). The McMurdo volcanics, of Upper Tertiary age, line the Victoria Land coast, abutted by New Zealand's South Island in Fig. 7, and Wright (1966) has given important correlations between the Late Precambrian and Palaeozoic rocks of South Island and those of Victoria Land. However, both he, and Griffiths & Varne (1972), offered reconstructions in which the southern edge of the Campbell Plateau adjoined the shelf edge of the Ross Sea. We believe that such reconstructions depict only an intermediate stage in the separative evolution of the area because they have been based on the premise that lithosphere separation can only create ocean floor. Accordingly, such reconstructions have been quite unable to place the pre-Carboniferous (Caledonian) structures of South Island, after allowance for Alpine Fault movement, in reasonable structural continuity with those which run through Victoria Land, western Tasmania and southeastern Australia. Figure 7 is attractive in this respect, however, but is achieved at the price of total omission of the Lord Howe Rise, whose crust (like that of the present Ross Sea) is therefore inferred to post-date this reconstruction.

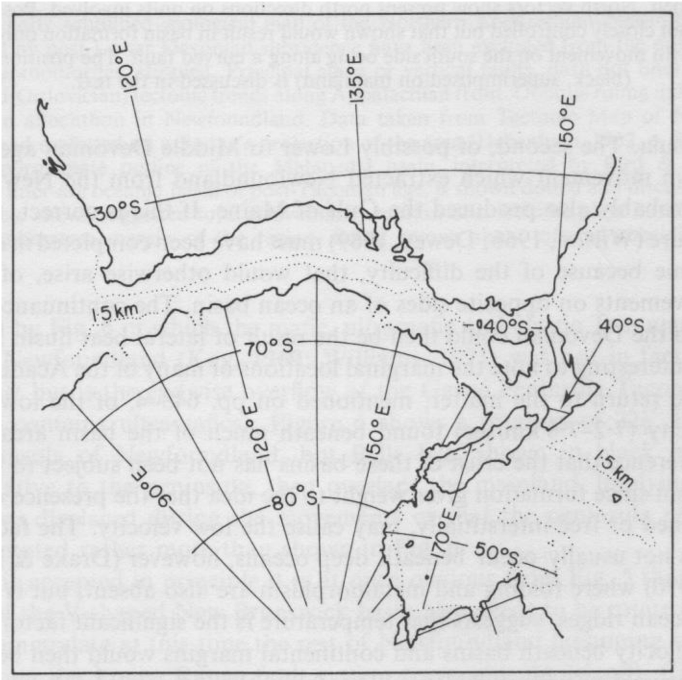

FIGURE 7 Crude reconstruction of Antarctica, Australia, New Zealand and the Campbell Plateau to suggest genesis of the present Ross Sea basin. Drawn from a reconstruction on a 34 em diameter globe using thin spherically-deformed plastic shells. 1·5 kIn isobaths taken from Tectonic Map of Australia (1960), American Geographical Society Map of Antarctica (1965) and from Brodie (1964) and Brodie & Dawson (1965). Eastern margin of Ross Sea has poor topographic expression, except at northern corner. The margin shown is the +200 m ice surface contour. Bathymetry on the west side of New Zealand, including the Lord Howe Rise, has been omitted (see text). Dashed line marks the Alpine Fault in South Island, New Zealand.

It is proposed that the first extractive movements occurred in Mid-Jurassic time, with the onset of the Rangitaka Orogeny in New Zealand (Fleming, 1970). A computer fit of the Australian and Antarctic continental shelves, slightly closer than Fig. 7 shows, was dated at 45 m.y. by McKenzie & Sclater (1971) so any Mid-Jurassic separative troughs between them would have had about 100 m.y. to fill before finai separation got under way. The Rangitaka Orogeny deformed New Zealand by dragging the northern part clockwise with respect to a fixed (?) southern end, and was accompanied by vigorous uplift, particularly on the west. It is inferred that this movement initiated a Late Jurassic trough to the west of New Zealand, and eventually left ellough room between New Zealand and Tasmania/Victoria Land for deposition of what is now the southern part of the Lord Howe Rise. The same movements may also have drawn Tasmania southeastwards to form the basins in the Bass Strait. The many horst and basin structures of New Zealand itself, developed during this time and later, appear ideally susceptible to interpretation in terms of limited separation but cannot be discussed here. For this reason all forms of reconstruction within New Zealand, including restoration of the Alpine Fault movement, have been ignored in Fig, 7. Extraction of the Campbell Plateau from the Ross Sea must have been complete by about 80 m.y. B.P., when there begins to be evidence of ocean-floor spreading in the Tasman Sea and south of the Campbell Plateau (Griffiths & Varne, 1972).

Perhaps the most intriguing implication of Fig. 7 is the fact that the Campbell Plateau is so different from Antarctica in its geology (Brodie, 1964), its crustal thickness of 17-23 km (Adams, 1962), and its apparent epeirogenic response to the proposed separation. The islands of the Campbell plateau all lie near its edges, suggesting that they may be inliers based on older crust detached from the Antarctic boundaries and/or that marginal upwarping of this lithosphere unit is responsible. Both are certainly true for South Island of New Zealand. On the other hand, along the southern margin of the Ross Sea, Precambrian and Early Ordovician (Ross Orogeny) folds of metagreywackes and phyllites with north and northwest axes terminate abruptly at the coast (Grindley et al., 1964). There is abundant evidence, however, from the rocks of the eastern side of New Zealand and from the Campbell Plateau and Chatham Rise, that this area was a major basin in Late Palaeozoic and Mesozoic times. Whatever separated from this area to produce this Mesozoic basin may exhibit the continuation of the ancient fold belt of the southern Ross Sea margin. Thus, if Fig. 7 is correct in its implication, the present site of much of the Ross Sea has been the site ofat least two major separative events during Phanerozoic time. It is suggested that chasmic faults remain sites of weakness for long periods owing to the protracted differential vertical movement on them which we have shown may persist for upwards of 100 m.y. The important conclusion drawn from this is that, owing to the repeated occurrence of chasmic faulting during geological time, the reconstruction of pairs of chasmic faults of a particular age should not necessarily result in the matching across them of geological structures of appreciably earlier age.

Conclusions and Discussion

We have shown that limited lithosphere separation, in circumstances that ensure an abundant sediment supply, will produce crust having characteristics which compare closely with known features of continental basins (not lying on shields), continental shelves and many geosynclines. In particular, the total crustal thickness, duration of subsidence, and the early (geosynclinal) subsidence rate appear able to attain observed values without difficulty. It is also clear that heavy sedimentation would result in production of a lower crust very different from ocean ridge crust and that the magnetic anomalies so characteristic of ocean-floor spreading would not be produced. This result rests solely on consideration of the thermal and other consequences of lithosphere separation and new lithosphere emplacement, on the lines now well established by studies of the cooling and subsidence of the oceanic lithosphere. No attention has had to be given to the separative agency except that which is implicit in the assumption of limited separation, i.e. that separation ceases when it has attained values representative of actual continental basins. Extinctions of formerly active ocean ridges are already known, so we have not added to existing constraints upon theories concerning lithosphere separative agencies.

The presence of chasmic faults (Figs. 1-3) marking the boundary between new and old lithosphere is important. The resulting lateral transfer of heat across chasmic faults at depth is a uniquely effective way of accounting for closely juxtaposed vertical movements of opposite sign. Such heat transfer accelerates the subsidence of narrow troughs and the marginal parts of wide basins, whose later subsidence is thus confined to the middle area. Surface expression of chasmic faults will continue as movements of normal fault character throughout much of the subsidence life of the basin. Moreover, because chasmic faults separate lithosphere having different histories and compositions, they may retain almost indefinitely a sensitivity to thermal rejuvenation, even after the adjacent basin crust has been folded. Rejuvenation of 'ancient lines of weakness' is a well-known geological phenomenon and may indicate the presence of ancient chasmic faults.

Lateral heat flush into the lithosphere bounding a new basin will vary with the prior condition of that lithosphere. The comparatively minor upwarping of the Atlantic shelf margins and their early reversion to subsidence would be consistent with these margins being fairly young lithosphere themselves. Our discussions of the origin of the North Sea and of the New Brunswick and Gulf of Maine basins suggest that much of the North Atlantic continental shelves north of the Azores may be an elongate basin system formed by a Mid-Palaeozoic limited separation of Europe from North America, immediately following closure of the proto-Atlantic. However the vigorous Jurassic upwarping of Jameson Land in southeast Greenland (Haller, 1970) suggests that initial opening of the Atlantic south of the Azores about 180 m.y. ago

(Francheteau, Vol. I, p. 195) was accompanied by important separative movements further north but that these troughs became added to the shelf area by being filled with sediments.

Where lateral heat flush enters old lithosphere possessing a thick crust, there is evidence not only that the upwarping can be as much as 3 km (largely maintained by isostasy in the face of heavy erosion) but that the heat can cause granites to rise within the crust. Such features, accompanied by the rapid subsidence of troughs and the possibility of flood or other volcanism, are commonly and, we suggest, appropriately regarded as orogenic in character. However, it is obyiously essential for the progress of plate tectonics that the effects of lithosphere separation should be clearly distinguisheo from those of plate convergence, so the terms separative orogeny and convergent orogeny are proposed for this purpose. In practice the correct attribution of observed structures may be rather less simple because, for example, differential epeirogenic movements can probably give rise to low-angle reverse faulting and apparent thrusting (Sanford, 1959; Osmaston, 1971).

Folding of the continental crust of basins of separative origin has been discussed at some length on pp. 652-3. Such crust is presumably more prone to folding by virtue of its younger and hotter lithosphere. Identification of such areas might help in elucidating the force distribution responsible for horizontal tectonic movements.

In conclusion, there is much to support the proposition that limited lithosphere separation has long been a major factor in producing continental basins and adding to the area of the continental crust. The matter clearly merits further study. It is appropriate, however, to remind ourselves of the central issue of the origin of epeirogenic movements. All such movements-even downward ones to provide more space for sediments-expend energy, and sources for this have to be found. Freshly emplaced lithosphere possesses adequately large amounts of thermal, chemical, and potential energy, and is a source wholly compatible with plate tectonics. Its apparent ability to distribute some of this energy to distances of several hundred kilometres into the surrounding lithosphere suggests that, if the long thermal time constants involved can be successfully taken into account, more detailed epeirogenic problems, such as the post (?) Mesozoic epeirogenetic differentiation and eastward tilting of crustal units within the British Isles, may ultimately become explicable in terms of distant separative orogeny. The large horizontal extent of individual epeirogenic movements and their frequently sharp geographical boundaries already support the suggestion made on page 643 and by Osmaston (1971), that much material currently assigned to the asthenosphere may in fact move integrally with tectonic plates and thus contribute to the scale of epeirogeny.

Acknowledgements

The possibility that lithosphere separation might be relevant to the formation not only of ocean basins but also of continental basins was first pointed out and brought to my notice by S. W. Carey's famous paper (Carey, 1958). In that the development of that hypothesis into the form presented here has been financed entirely from personal resources I am particularly grateful to Professor J. Sutton and to many other former colleagues at Imperial College, London, for their continuing interest and for many hours of stimulating discussion. Professor M. Blackman gave helpful advice on thermal expansion, Professor W. S. Fyfe kindly reminded me of the broad utility of the Clapeyron equation, and I thank Dr. S. Moorbath for advice on geochronology. Finally, I would thank especially Drs. W. S. McKerrow, H. G. Reading and A. Richardson for their critical advice and enthusastic encouragement over a long period.

References

Adams, R. D. 1962. Thickness of the Earth's crust beneath the Campbell Plateau, *N.Z. J. Geol. Geophys.,* 5, 74-85.

American Geographical Society, 1965. Map of Antarctica, 1: 5,000,000.

Anderson, D. L. and Spetzler, H., 1970. Partial melting and the low-velocity zone, Phys. Earth Planet. Interiors, 4, 62-64.

Belousov, V. V., 1960. Development of the Earth and tectonogenesis, J. Geophys. Res., 65, 4127-46.

Belousov, V. V., 1962. Basic Problems in Geotectonics, edited by J. C. Maxwell, transl. by P. T. Broneer. McGraw-Hili, New York. 816pp.

Belousov, V. V., 1966. Modern concepts of the structure and development of the Earth's crust and the upper mantle of continents, Q. J. Geol. Soc. Lond., 122, 293-314.

Berger, J., Cok, A. E., Blanchard, J. E. and Keen, M. J., 1966. Morphological and geophysical studies on the eastern seaboard of Canada: the Nova Scotian Shelf, Roy. Soc. Can. Spec. Pub., 9, 102-13.

Billings, M. P., 1960. Diastrophism and mountain building, Geol. Soc. Amer. Bull., 71, 363-98.

Birch, R. L., 1969. The search for gas in the North Sea, Geology, 1, 31-37.

Bird, J. M. and Dewey, J. F., 1970. Lithosphere plate: continental margin tectonics and the evolution of the Appalachian orogen, Geol. Soc. Amer. Bull., 81, 1031-60.

Black, R. F., 1964. Palaeomagnetic support of the theory of rotation of the western part of the island of Newfoundland, Nature, Lond., 202, 945-8.

Bott, M. H. P., 1964. Formation of sedimentary basins by ductile flow of isostatic origin in the upper mantle, Nature, Lond., 201, 1082-4.

Bott, M. H. P. 1971. Evolution of young continental margins and formation of shelf basins, Tectonophysics, 11, 319-27.

Bott, M. H. P. and Dean, D. S. 1972. Stress systems at young continental margins, Nature Phys. Sci., 235, 23-25.

Bott, M. H. P. and Johnson, G. A. L., 1967. The controlling mechanism of Carboniferous cyclic sedimentation, Q. J. Geol. Soc. Lond., 122, 421-41.

Bott, M. H. P. and Watts, A. B., 1970. Deep structure of the continental margin adjacent to the British Isles. In: The Geology of the East Atlantic Continental Margin, edited by F. M. Delaney, Vol. 2, pp. 93-109. ISCU/SCOR Symposium, Cambridge 1970. H.M.S.O., London. Briden, J. c., 1970. Palaeomagnetic results from the Arrochar and Garrabal Hill-Glen

Fyne igneous complexes, Scotland, Geophys, J. R. astr. Soc., 21, 457-70. Brodie, J. W., 1964. Bathymetry of the New Zealand region. New Zealand Oc. Inst. Mem. 11. N.Z.D.S.I.R. Bull, 161. Brodie, J. W. and Dawson, E. W., 1965. Morphology of North Macquarie Ridge, Nature, Lond., 207, 844-5.

Brown, P. E. and Miller, J. A., 1969. Some aspects of palaeozoic geochronology of British Isles. In: North Atlantic: Geology and Continental Drift, edited by M. Kay. Amer. Assoc. Pet. Geol. Mem., 12,

363-74.
Bunce, E. T., Crampin, S., Hersey, J. B. and Hill, M. N., 1964. Seismic refraction observations on the continental boundary west of Britain, J. Geophys. Res., 69, 3853-63.
Carey, S. W., 1958. The tectonic approach to continental drift. In: Continental Drift: a Symposium, 1956, pp. 177-355. University of Tasmania, Hobart.
Collette, B. J., 1968. On the subsidence of the North Sea area. In: Geology of Shelf Seas. edited by D. T. Donovan. Oliver and Boyd, Edinburgh. 15pp.
Collette, B. J., Lagaay, R. A., Ritsema, A. R. and Schouten, J. A., 1970. Seismic investigations in the North Sea, 3 to 7, Geophys. J. R. astr. Soc., 19, 183-99.
Dainty, A. M., Keen, C. E., Keen, M. J. and Blanchard, J. E., 1966. Review of geophysical evidence on crust and upper-mantle structure on the eastern seaboard of Canada. In: The Earth Beneath the Continents, edited by J. S. Steinhart and T. J. Smith. Amer. Geophys. Un. Geophys. Monogr., 10, 349-69.
Debenham, F. (editor), 1962. The Reader's Digest Great World Atlas. Reader's Digest Association, London. 179pp.
Dewey, J. F., 1969. Evolution of the Appalachian/Caledonian orogen, Nature, Lond., 222, 124-9.
Dewey, J. F. and Bird, J. M., 1970. Mountain belts and the new global tectonics, J. Geophys. Res., 75, 2625--47.
Dewey, J. F. and Pankhurst, R. J., 1970. The evolution of the Scottish Caledonides in relation to their isotopic age pattern, Trans. Roy. Soc. Edinb., 68, 361-89.
Dickinson, W. R., 1971. Plate tectonic models of geosynclines, Earth Planet. Sci. Let., 10, 165-74.
Drake, C. L. and Nafe, J. E., 1968. The transition from ocean to continent from seismic refraction data. In: The Crust and Upper Mantle of the Pacific Area, edited by L. Knopoff, C. L. Drake and P. J. Hart. Amer. Geophys. Un. Geophys. Monogr., 12, 174-86.
Drake, C. L., Ewing, J. I. and Stockard, H., 1968. The continental margin of the eastern United States, Can. J. Earth Sci., 5, 993-1010.
Emery, K. 0., 1965. Geology of the continental margin off eastern United States. In: Submarine Geology and Geophysics, edited by W. F. Whittard and R. Bradshaw, pp. 1-20. Butterworths, London.
Fleming, C. A., 1970. The Mesozoic of New Zealand: chapters in the history of the circumPacific Mobile Belt, Q. J. Geol. Soc. Lond., 125, 125-70.
Fyfe, W. S., Turner, F. J. and Verhoogen, J., 1958. Metamorphic reactions and metamorphic facies, Geol. Soc. Amer. Mem., 73.
Gair, H. S., 1967. The geology from the Upper Rennick Glacier to the coast, northern Victoria Land, Antarctica, N.z. J. Geol. Geophys. 10, 309--44.
Gilluly, J., 1964. Atlantic sediments, erosion rates and the evolution of the continental shelf: some speculations, Geol. Soc. Amer. Bull., 75, 483-92.
Gilluly, J., Reed, J. C. Jr. and Cady, W. M., 1970. Sedimentary volumes and their significance, Geol. Soc. Am. Bull., 81, 353-76.
Griffiths, J. R. and Varne, R., 1972. Evolution of the Tasman Sea, Macquarie Ridge and Alpine Fault, Nature Phys. Sci., 235, 83-86.
Grindley, G. W., McGregor, V. R. and Walcott, R. I., 1964. Outline of the geology of the Nirnrod-Beardmore-Axel Heiberg Glaciers region, Ross Dependency. In : Antarctic Geology, edited by R. J. Adie, pp. 207-18. North-Holland, Amsterdam.
Haller, J., 1970. Tectonic map of East Greenland, I: 500,000, Medd. om Gronland, 171 (5), 286pp.
Heirtzler, J. R., 1969. Geomagnetic studies in the Atlantic Ocean. 111: The Earth's Crust and Upper Mantle, edited by P. J. Hart. Amer. Geophys. UI1. Geophys. MOl1ogr., 13,430-6.
Holder, A. P. and Bott, M. H. P., 1971. Crustal structure in the vicinity of Southwest England, Geophys. J. R. astr. Soc., 23, 465-89.
Holtedahl, O. and Dons, J. A. (compilers), 1960. Geological map of Norway, I: 1,000,000, Norges Geologiske Undersøkelse, Oslo.
James, D. E. and Steinhart, J. S., 1966. Structure beneath continents: a critical review of explosion studies, 1960-1965. In: The Earth beneath the Continents, edited by J. S. Steinhart and T. J. Smith. Amer. Geophys. Un. Geophys. Monogr., 10, 293-333.
Jeffreys, H., 1959. The Earth. 4th Edition. Cambridge University Press, 420pp. Joyner, W. B., 1967. Basalt-eclogite transition as a cause for subsidence and uplift, J. Geophys. Res., 72, 4977-98. Judson, S., 1968. Erosion of the land, or what's happening to our continents? Amer. Sci., 56, 356-74.
Kanamori, H. and Press, F., 1970. How thick is the lithosphere? Nature, Lond., 226, 330-1.
Karig, D. E., 1970. Ridges and basins of the Tonga-Kermadec island arc system, J. Geophys. Res., 75, 239-54. Karig, D. E., 1971. Structural history of the Mariana island arc system, Geol. Soc. Amer. Bull., 82, 323-44. Kay, M., 1955. Sediments and subsidence through time, Geol. Soc. Amer. Spec. Pap., 62, 665-84.
Kay, M., 1969. Thrust shuts and gravity slides of western Newfoundland. In: North Atlantic-Geology and Continental Drift, edited by M. Kay. Amer. Assoc. Pet. Geol. Mem. 12,665-9.

Kent, P. E., 1967. Progress of exploration in North Sea, Amer. Assoc. Pet. Geol. Bul/., 51, 731-41.

King, L. C., 1965. Since du Toit: Geological relationships between South Africa and Antarctica. Alex. L. du Toit Mem. Lec!. 9. Geol. Soc. S. Africa. Annexure to Vol. 68, 32pp.

King, L. C., 1967. Morphology of the Earth. 2nd Ed. Oliver and Boyd, Edinburgh, 726pp.

King, P. B., 1969. Tectonic map of North America, 1: 5,000,000, Dept. of the Interior, U.S. Geological Survey, Washington, D.C.

Kosminskaya, I. P. and Zverev, S. M., 1968. Deep seismic soundings in the transition zones from continents to oceans. In: The Crust and Upper Mantle of the Pacific area, edited by

L. Knopoff, C. L. Drake and P. J. Hart. Amer. Geophys. Un. Geophys. Monogr., 12, 122-30.

Langseth, M. G., Le Pichon, X. and Ewing, M., 1966. Crustal structure of the mid-ocean ridges. 5. Heat flow through the Atlantic Ocean floor and convection currents, J. Geophys. Res., 71, 5321-55.

Lee, W. H. K. and Taylor, P. T., 1966. Global analysis of seismic refraction measurements, Geophys. J. R. astr. Soc., 11, 389-413.

Lee, W. H. K. and Uyeda, S., 1965. Review of heat flow data. In: Terrestrial Heat Flow, edited by W. H. K. Lee. Amer. Geophys. Un. Geophys. Mongr., 8, 87-190.

Le Pichon, X. and Langseth, M. G. Jr., 1970. Heat flow from the mid-ocean ridges and seafloor spreading, Tectonophysics, 8, 319-44 . .

MacGregor, I. D., 1970. The effect of CaO, Cr203, Fe20 3 and Ah03 on the stability of spinel and garnet peridotites, Phys. Earth Planet. Interiors, 3, 372-7.

Magnitinsky, V. A. and Kalashnikova, r. V. 1970. Problems of phase transitions in the upper mantle and its connection with the earth's crustal structure, J. Geophys. Res., 75, 877-85.

Matsuda, T. and Uyeda, S., 1971. On the Pacific-type orogeny and its model-extension of the paired belts concept and possible origin of marginal seas, Tectonophysics, 11, 5-27.

Maynard, G. L., 1970. Crustal layer of seismic velocity 6·9 to 7·6 kilometers per second under the deep oceans, Science, 168, 120-1 .

McGinnis, L. D., 1970. Tectonics and the gravity field in the continental interior, J. Geophys. Res., 75, 317-31.

McKenzie, D. P. and Sclater, J. G., 1968. Heat flow inside the island arcs of the northwestern Pacific, J. Geophys. Res., 73, 3173-9.

McKenzie, D. P. and Sclater, J. G., 1971. The evolution of the Indian Ocean since the Late Cretaceous, Geophys. J. R. astr. Soc., 24, 437-528.

McKerrow, W. S. and Ziegler, A. M., 1971. The Lower Silurian palaeogeography of New Brunswick and adjacent areas, J. Geol. , 79, 635-46.

Menard, H. W., 1967. Transitional types of crust under small ocean basins, J. Geophys Res., 72, 3061 -73.

Mitchell, A. H. and Reading, H. G., 1969. Continental margins, geosynclines and ocean-floor spreading, J. Geol., 77, 629-46.

Morgan, W. J., 1968. Rises, trenches, great faults, and crustal blocks, J. Geophys. Res., 73, 1959-82.

Nalivkin, D . Y., 1960. The Geology of the USSR, transl. by S. I. Tomkeieff. Pergamon, Oxford.

Officer, C. B. and Ewing, M., 1954. Geophysical investigations in the emerged and submerged Atlantic coastal plain, 7, Continental shelf, continental slope, and continental rise south of Nova Scotia, Geol. Soc. Am. Bul/., 65, 653-70.

O'Hara, M. J., Richardson, S. W. and Wilson, G., 1971. Garnet-peridotite stability and occurrence in crust and mantle, Contr. Mineral. and Petrol., 32, 48-68. =Osmaston, M. F., 1969. Discussion on paper by R. Stoneley 'Sedimentary thicknesses in orogenic belts.' In: Time and Place in Orogeny, edited by P. E. Kent, G. E. Satterthwaite, and A. M. Spencer, Geol. Soc. Lond., Spec. Pub. No.3, 306-7.

Osmaston, M. F., 1971. Genesis of ocean ridge median valleys and continental rift valleys, Tectonophysics, 11, 387-405.

Oxburgh, E. R. and Turcotte, D. L., 1969. Increased estimate for heat flow at oceanic ridges, Nature, Lond. 223, 1354-5.

Packham, G. H. and Falvey, D. A., 1971. An hypothesis for the formation of marginal seas in the western Pacific, Tectonophysics, 11, 79-109.

Pankhurst, R. J., 1970. The geochronology of the basic igneous complexes, Scot. J. Geol., 6, 83-107.

Pitman, W. C., III, Talwani, M. and Heirtzler, J. R., 1971. Age of the North Atlantic Ocean from magnetic anomalies, Earth Planet. Sci. Letts., 11, 195-200.

Ramberg, H., 1952. The Origin of Metamorphic and Metasomatic Rocks. Univ. Chicago Press, 317pp.

Rezanov, J. A. and Chamo, S. S., 1969. Reasons for absence of a granitic layer in basins of the South Caspian and Black Sea type, Can. J. Earth Sci., 6, 671-8.

Ringwood, A. E. and Lovering, J. F., 1970. Significance of pyroxeneilmenite intergrowths among kimberlite xenoliths, Earth Planet. Sci. Letts., 7, 371-5.

Robinson, E. S., 1964. Regional geology and crustal thickness in the Transantarctic Mountains and adjacent ice-covered areas in Antarctica. (Abstract), Geol. Soc. Amer. Spec. Pap., 76, 316.

Sanford, A. R., 1959. Analytical and experimental study of simple geological structures, Geol. Soc. Am. Bull., 70, 19-52.

Schumm, S. A., 1963. The disparity between present rates of denudation and orogeny. U.S. Geo!. Surv. Prof. Pap. 454-H, 13pp.
Sclater, J. G., Anderson, R. N. and Bell, M. L., 1971. Elevation of ridges and evolution of the Central Eastern Pacific, J. Geophys. Res., 76, 7888-915.
Sclater, J. G. and Francheteau, J., 1970. The implications of terrestrial heat flow observations on current tectonic and geochemical models of the crust and upper mantle of the Earth, Geophys. J. R. astr. Soc., 20, 509-42.
Sclater, J. G. and Menard, H. W., 1967. Topography and heat flow of the Fiji Plateau, Nature, Lond., 216, 991-3.
Sheridan, R. E., 1969. Subsidence of continental margins, Tectonophysics, 7, 219-29.
Sheridan, R. E. and Drake, C. L., 1968. Seaward extension of the Canadian Appalachians, Can. J. Earth Sci., 5, 337-73.
Sleep, N. H., 1971. Thermal effects of the formation of Atlantic continental margins by continental break-up, Geophys. J. R. astr. Soc., 24, 325-350.
Stoneley, R., 1969. Sedimentary thicknesses in orogenic belts. In: Time and Place in Orogeny, edited by P. E. Kent, G. E. Satterthwaite and A. M. Spencer, Geol. Soc. Lond., Spec. Pub. 3, pp. 215-38.
Storetvedt, K. M., Halvorsen, E. and Gjellestad, G., 1968. Thermal analysis of the natural remanent magnetism of some Upper Silurian red sandstones in the Oslo region, Tectono-physics, 5, 413-26.
Stride, A. H., Curray, J. R., Moore, D. G. and Belderson, R. H., 1969. Marine geology of the Atlantic continental margin of Europe, Phil. Trans. Roy. Soc. Lond. A., 264, 31-75.
Sutton, J., 1969. Rates of change within orogenic belts. In: Time and Place in Orogeny, edited by P. E. Kent, G. E. Satterthwaite and A. M. Spencer, Geol. Soc. Lond. Spec. Pub. 3., pp. 239-50.
Tectonic Map of Australia, 1960. 1st. edition. 1: 2,534,400. Bur. Min. Resources, Geology and Geophysics, Canberra. .
Turner, F. J. and Verhoogen, J., 1960. Igneous and Metamorphic Petrology. 2nd Edition. ' McGraw-HilI, New York. 694pp.
Van Bemmelen, R. W., 1966. On mega-undations: a new model for the Earth's evolution, Tectonophysics, 3, 83-127.
Van de Lindt, W. J., 1967. Movement of the Mohorovicic discontinuity under isostatic conditions, J. Geophys. Res., 72, 1289-97.
Vogt , P. R. and Ostenso, N. E., 1967. Steady state crustal spreading, Nature, Lond. , 215, 810-17.
Von Herzen, R. P., 1967. Surface heat flow and some implications. In : The Earth's Mantle. edited by T. F. Gaskell, pp. 197-230. Academic Press, London.
Williams, H ., 1971. Mafic-ultramafic complexes in western Newfoundland Appalachians and the evidence for their transportation : a review and interim report, Geol. Ass. Canada. Proc., 24, No. I, 9-25.
Wilson, J. T. , 1965. A new class of faults and their bearing on continental drift, Nature, Lond. 207, 343-7.
Wilson, J. T ., 1966. Did the Atlantic close and then reopen? Nature, Lond., 211, 676-81.
Wright, J. B., 1966. Convection and continental drift in the Southwest Pacific, Tectono-physics, 3, 69-81. Ziegler, A. M., 1970. Geosynclinal development of the British Isles during the Silurian period.
J. Geol., **78,** 445-79.