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Fore-arc evolution and continental growth: a general model

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Abstract—The extended evolution of fore-arc regions which leads to their eventual incorporation into stable kratonic continental crust is elucidated by a general model based upon observations from the modern circum-Pacific and the Palaeozoic Tasman Geosyncline.

Fore-arc regions widen during subduction in the manner described by Karig & Sharman (1975). Their history, after subduction has ceased, depends upon the thickness of the accretionary prism formed during subduction. Where the prism is thick (*ca.* 20 km) kratonization is a single-step process. The fore-arc region remains above sea-level; post-arc silicic volcanics accumulate due to granitoid plutonism, the magmas being derived by melting of the subduction complex and from the oceanic lithosphere trapped beneath it. The volcanic arc subsides, becoming the site of a fore-deep.

Intermediate-thickness accretionary prisms (*ca.* 16 km) are kratonized in a two-step process. They remain at shelf depths, while their associated volcanic arcs sink to comparable depths. Both acquire a post-arc shallow marine sequence of typical platform-cover facies. They are then deformed and intruded by granitoids when the crust attains critical thickness (*ca.* 20 km).

Thin accretionary prisms ($\leq$ 12 km) require a three-step process for kratonization. They and their associated arcs sink to bathyal depths. They are overwhelmed by prograding post-arc flysch deposits of continental origin. Deformation of the post-arc flysch and plutonism occur when critical crustal thickness (*ca.* 20 km) is attained. A transitional tectonic regime ensues, with molasse-like transitional basins preferentially sited over the extinct volcanic arcs and the thinner parts of buried accretionary prisms.

The model satisfactorily explains the Late Proterozoic–Palaeozoic evolution of southeast Australia, where a 1000 km wide tract of continental crust was accreted to the Australian Kraton in 250–300 Ma, beginning as a S.W. Pacific-type oceanic terrain. It has been found useful for interpreting geosynclinal terrains in other continents.

According to the model, the dynamic processes that contribute to kratonization are systematically causally connected. Kratonization is a unified, internally deterministic and self-sustaining phenomenon. The model has implications for the origin, 'stratigraphy' and composition of upper and lower continental crust; the origins and tectonic settings of ophiolites, granitoids, paired metamorphic belts and transitional basins; and for the nature and causes of orogenesis.

INTRODUCTION

THE ORIGIN of stable ('kratonic') continental crust is one of the fundamental problems of geotectonics. I have previously argued (Crook 1974) that West Pacific-type geosynclines, which in their early stages comprise island arc-marginal sea oceanic terrains, are particularly efficient geodynamic 'machines' which transform oceanic terrains into kratonic continental crust. Here I explain how this transformation occurs, using a general model which traces the evolution of fore-arc regions from the stage of active accretionary growth during subduction, described by Karig & Sharman (1975) and Karig *et al.* (1976), through an extended post-subduction history which culminates in their transformation into kratonic continental crust of normal thickness.

In constructing the model, observations from modern circum-Pacific fore-arc regions have been combined with interpretations of certain terrains in the Palaeozoic Tasman Geosyncline of southeastern Australia, a fold belt which has been part of the Australian kraton since the end of the Palaeozoic. The model was developed during studies of the evolution of the continental crust of southeastern Australia. It has been tested, first, by detailed comparison with all terrains in that region. This comparison (Crook in press) shows a very high degree of correspondence between the geological data and the model.

The model has also been tested, less rigorously, by application to terrains in other continents — the Polar Urals, Southern Uplands of Scotland, Newfoundland, Eastern Desert of Egypt, South Island of New Zealand, and the East Klamaths of the U.S.A. Significant insights into tectonic evolution emerged in each case; in many there was close correspondence between geological data and the model. These results, together with the diverse general implications of the model which are summarized at the end of this paper, lead me to conclude that the model is a general one which will prove useful as a paradigm for understanding the evolution of continental crust.

THE MODEL

Terminology and summary description

The model which describes the evolution of fore-arc regions is shown schematically in Fig. 1. Young arc terrains (Fig. 1.1) consist of a volcanic arc or chain and a fore-arc region extending outwards to the trench. In the fore-arc region there are three stratotectonic units which together constitute the accretionary prism. These units are: the subduction complex, composed mainly of off-scraped sediment; an overlying upper-slope or fore-arc sedimentary sequence, derived partly from the volcanic

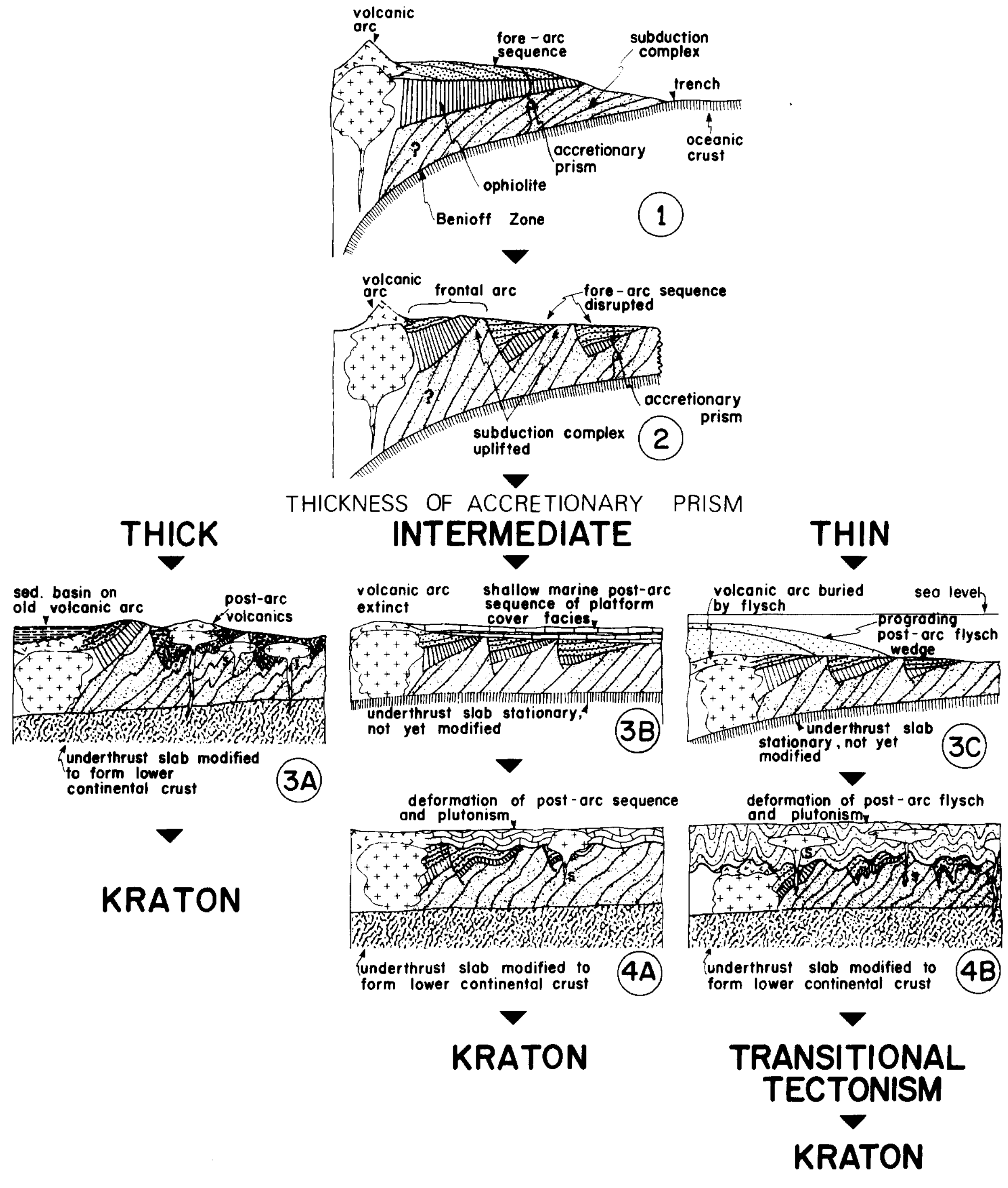

Fig. 1. Model of the extended evolution of fore-arc regions and their associated volcanic arcs leading to kratonization.

arc, which includes deposits in trench-slope basins; and, in the inner part of the fore-arc region adjacent to the volcanic arc, an ophiolite which represents the oceanic lithosphere that formed the substrate on which the oldest parts of the fore-arc sequence were deposited. The contact between the ophiolite and the subduction complex is tectonic. A back-arc apron is present behind the volcanic arc.

As subduction proceeds, the fore-arc region widens, due to outward growth of the accretionary prism in the manner described by Karig & Sharman (1975). The maximum width attained by modern fore-arc regions exceeds 400 km. Structural complications develop in the older parts of the accretionary prism, commonly producing a geologically complex elevated area, which Karig & Sharman term a frontal arc (Fig. 1.2).

After subduction has ceased the arc terrain acquires other stratotectonic units, notably a post-arc sequence, which may be partly volcanogenic, and granitoid intrusions (Figs. 1.3A and 1.4A, B). Metamorphic belts may be associated with the granitoids. Transitional basin sequences of molasse-like red-beds may also accumulate in certain areas after the terrain-wide deformation associated with granitoid intrusion. Eventual kratonization of the arc terrain is marked by the deposition upon it of a sedimentary platform cover.

This post-subduction history depends significantly upon the thickness of the accretionary prism. Thick (*ca.* 20 km) accretionary prisms become kratonized after subduction ceases in a single-step process involving deformation, metamorphism and plutonism. Post-arc sequences, in part volcanogenic, which accumulated on top of the accretionary prism after subduction has ceased, are of minor significance (Fig. 1.3A).

Intermediate-thickness (*ca.* 16 km) and thin ($\leq$ 12 km) accretionary prisms require two- and three-step processes respectively for kratonization. In each case the fore-arc region is initially below sea level. Significant thicknesses of largely non-volcanogenic post-arc sediment accumulate: neritic deposits on intermediate-thickness prisms (Fig. 1.3B) and flysch on thin accretionary prisms (Fig. 1.3C). Deformation, metamorphism and plutonism (Fig. 1.4A, B) apparently occur only when the total thickness of the accretionary prism and post-arc sequence reaches a critical value of *ca.* 20 km. Thereafter those regions with intermediate-thickness prisms act as kratonic blocks. A phase of transitional basin sedimentation and deformation generally precedes kraton formation in the case of thin prisms.

Thus, from the viewpoint of its extended evolution, an arc terrain comprises a volcanic arc and its associated accretionary prism, back-arc apron, post-arc sequence, granitoids, metamorphic belts and transitional basin sequences. From the viewpoint of continental growth, the fore-arc region is the key element in the arc terrain. Its accretionary growth generates a significant thickness of crustal materials of considerable areal extent. I have therefore chosen to highlight fore-arc evolution as the significant aspect in this model of continental growth.

The evolutionary pattern of arc terrains is well displayed in considerable diversity in the Tasman Geosyncline, the geology of which is described by Brown *et al.* (1968), Crook & Powell (1976), Douglas & Ferguson (1976), Packham (1969), Parkin (1969) and Scheibner (1978). The geosyncline appears to consist mainly of completely evolved island arc terrains with accretionary prisms of diverse thicknesses (Crook in press). These terrains have been sutured largely fore-arc to back-arc, but suturing events seem to play only a minor role in the deformation history of the fold belt.

Unlike many more widely known fold belts of the northern hemisphere, the Tasman Geosyncline lacks zones of nappes and thrust sheets: the style of deformation is predominantly upright rather than recumbent (Crook 1969, Crook & Powell 1976). This reflects the lack of continent–continent collisions during its evolution. The component tectonic units of the fold belt remained in their peri-continental oceanic setting throughout their evolution, being eventually accreted laterally onto the kraton rather than emplaced on top of it. This fold belt therefore displays rather completely the extended evolution of fore-arc regions.

Fore-arc evolution during subduction

In describing the model in detail, it is convenient to treat separately those features which form during subduction and those which form later, after subduction has ceased.

Fore-arc structure. The arc-trench gap, which is the distance between the volcanic arc and the trench, widens as subduction continues (Fig. 2) (Dickinson 1973). Karig & Sharman (1975) propose that the widening is very largely due to outward growth of the accretionary prism by two mechanisms: subduction and prograding sedimentation. Packets of sediment are incorporated into the subduction complex at the foot of the inner wall of the trench, by transfer from the subducting plate. Concurrently the fore-arc sequence progrades outwards over the growing subduction complex. Much, if not all, of the subducted sediment is trapped in the accretionary prism because of its low density.

In their early evolutionary stage, arc terrains are assumed to have the form shown schematically in Fig. 1.1. Note that, for the Benioff Zone to extend to the depth of the 150–180 km observed in young magmatic arcs, subduction must have commenced at a point some 90 km in front of the volcanic chain (Fig. 2) (cf. Dickinson & Seely 1979). This implies that some kind of oceanic lithosphere, i.e. ophiolite, occurs in the inner parts of accretionary prisms as a substrate on which the fore-arc sequence accumulated.

As the accretionary prism widens, component slabs within the subduction complex gradually rotate into much steeper attitudes (Fig. 1). There will be a concomitant 'down to the rear' movement within fore-arc sequences (Fig. 1.2), described by Karig & Sharman (1975). Dewatering, metamorphism and other structural adjustments also occur. This rearrangement may occur repeatedly during fore-arc accretion so that the inner parts of wide accretionary prisms will have an internal structure shown schematically in Fig. 1.2. The innermost part of such a fore-arc region corresponds to Karig & Sharman's 'frontal arc'.

Compressive rotational movements within the subduction complex are accompanied by extensional effects in the overlying fore-arc sequence. The areal extent of the fore-arc sequence may increase because new depositional sites are formed. Also the subduction complex may be protruded through the overlying fore-arc sequence leading to dismemberment of both the fore-arc sequence and, where present, its ophiolite substrate. This provides a mechanism for moving ophiolite-based fore-arc se-

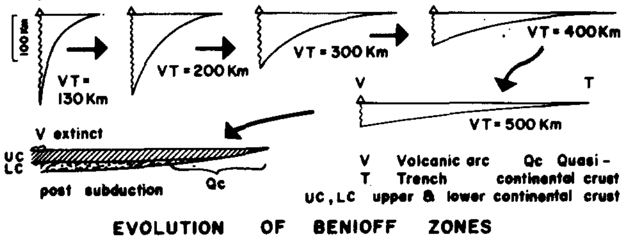

Fig. 2. Hypothetical evolution of Benioff Zones, showing shallowing of the Benioff Zone beneath the volcanic arc concomitant with widening of the arc-trench gap.

quences into positions somewhat further forward from the volcanic arc than those they originally occupied.

As a result of the extension the upper levels of the accretionary prism will be cut by major high angle faults which will typically mark disjunctions in age, structure and sedimentary facies. This appears to be the explanation for the faults shown in Fig. 2 of Karig *et al.* (1976).

Fore-arc magmatism and metamorphism. In addition to structural changes, fore-arcs may become sites of granitoid plutonism and silicic to intermediate volcanism, attributed by Marshak & Karig (1977) to the subduction beneath the accretionary prism of major heat sources such as triple junctions or spreading ridges. Examples from the New England Fold Belt, Australia show that this syn-subduction magmatism may be accompanied by significant low-pressure regional metamorphism and concomitant deformation of parts of the fore-arc sequence. This metamorphism is quite distinct from the burial metamorphism ubiquitous in accretionary prisms and from the high pressure metamorphism which occurs in subduction complexes.

Fore-arc morphology and crustal thickness. Differences between the geomorphic profiles of various fore-arc regions have been attributed by Karig & Sharman (1975) to variations in supply of material to the fore-arc basin and subduction complex. Evidently a spectrum of geomorphic profiles exists, resulting from the interaction of two independent variables. These are (a) sediment supply from the volcanic arc and frontal arc to the fore-arc region and the trench, and (b) sediment supply from non-arc sources (continental margins, pelagic sedimentation) onto the oceanic crust which is being subducted, or directly into the trench.

Three particular cases within this geomorphic spectrum can usefully be considered further. In the first case the fore-arc profile out to the trench-slope break is largely subaerial, with local tectonic depressions which may be flooded by the sea. This case is exemplified by the eastern part of the North Island of New Zealand, an active fore-arc region. The freeboard shows that the accretionary prism in this case has attained the 20 km thickness typical of upper continental crust.

In the second case a marked trench-slope topographic break is present either as the outer margin of a largely submarine terrace or as an elevated outer margin to a fore-arc topographic basin (Karig & Sharman 1975), as exemplified by their profile across Sumatra (Fig. 3). The 1 : 1 scale redrawing of this profile emphasizes its similarity, in terms of available space for sediment accumulation, to a simple submarine terrace profile. In either variant, much of the fore-arc region is at shelf or marginal plateau depths. This implies a thinner-than-normal ('transitional') upper crust, typically some 16 km thick.

The third case is exemplified by the Marianas profile in Fig. 3. The fore-arc profile slopes irregularly, with the trench-slope break usually poorly expressed. Much of the fore-arc region is at bathyal depths, implying an upper crustal thickness of some 12 km.

These differences in the inferred thicknesses of accretionary prisms are of fundamental importance to the subsequent evolution of fore-arc regions once subduction has ceased. The three cases appear to be tectonically significant points within a continuous spectrum of fore-arc geomorphic profiles and upper crustal thicknesses. An indication of the possible diversity within this spectrum is given by Dickinson & Seely (1979, fig. 6).

Fore-arc evolution after subduction has ceased

When subduction ceases the arc terrain includes a fore-arc accretionary prism which may be more than 400 km wide, the thickness of which reflects the nature and rate of sediment supply during subduction. The structure is likely to be complex; significant areas of granitoids and low pressure regional metamorphics may be present, particularly in the older parts of the accretionary prism.

Provided the arc terrain has not been emplaced tectonically onto continental or quasi-continental crust, its subsequent history will depend crucially on the thickness of the accretionary prism (Fig. 1). I will examine three cases, representative of a spectrum of thicknesses, using terrains in the Tasman Geosyncline as examples. In each case the thickness of the accretionary prism is inferred from the facies and thickness of the post-arc sequence, assuming that upper continental crust has a typical total thickness of 20 km.

Thick accretionary prisms. Where the accretionary prism is thick, *ca.* 20 km, kratonization is completed in a one-step process (Fig. 1). This is exemplified by the New England Fold Belt in northeastern New South Wales, Australia (Figs. 4 and 5), where a volcanic arc, active from at least the Early Devonian until the mid-Permian,

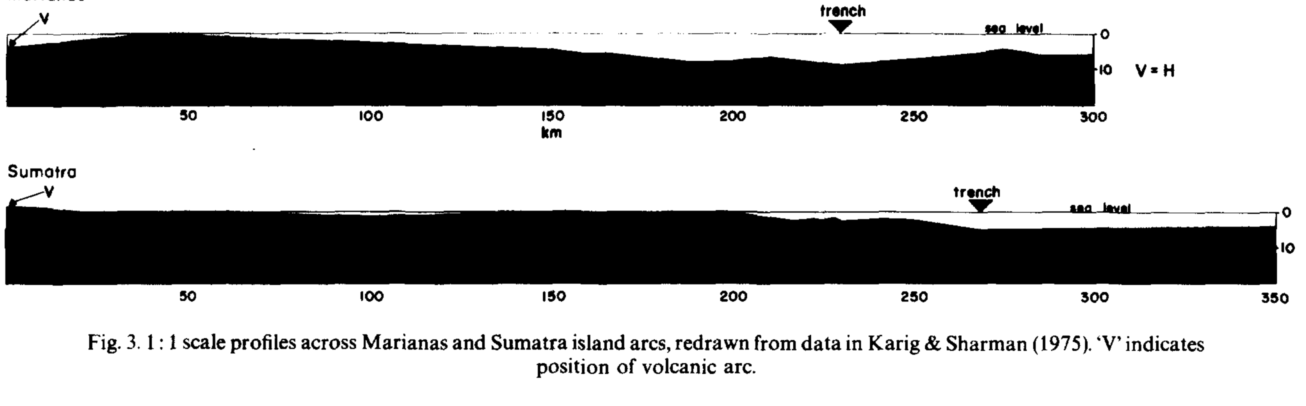

Fig. 3. 1 : 1 scale profiles across Marianas and Sumatra island arcs, redrawn from data in Karig & Sharman (1975). 'V' indicates position of volcanic arc.

formed the western margin of an extensive fore-arc region.

The salient features are as follows:

1. The geology and structure of thick accretionary prisms is already complex when subduction ceases. In the New England Fold Belt, many of the faults shown in the central part of Fig. 4 were already present when subduction ceased in the mid-Permian. This region is a mosaic of juxtaposed blocks of fore-arc and subduction complex rock associations. Blueschists occur in the Woolomin Association on the coast north of the Lorne Basin (Barron *et al.* 1976) (Fig. 4). Anatectic granitoids, S-type of Chappell & White (1974), the Bundarra and Hillgrove Suites of Fig. 4, had already been emplaced (Korsch 1977). Each suite consists of a line of plutons which decrease in age towards the volcanic arc. The Hillgrove Suite is associated with low pressure regional metamorphism of the fore-arc Coffs Harbour Association (Fig. 4). Origin of the granitoids by subduction of a major heat source, with consequent melting of the subduction complex, seems probable. If so, the lines of plutons define the orientation and magnitude of the vector of convergence between the arc and the subducting plate.

2. Thick accretionary prisms have already attained continental freeboard over much of their extent before subduction ceases. Post-arc materials accumulate in subaerial or shallow marine environments, except perhaps over the outermost thinnest part of the accretionary prism where some flysch could be expected. This phenomenon is a direct consequence of the accretionary prism having attained typical upper continental crustal thickness.

In the New England Fold Belt of New South Wales post-arc sedimentary and volcanic rocks — the 'Dummy Creek Association' and 'Upper Permian acid-intermediate volcanic rocks' of Fig. 4 — are subaerial to shallow marine in origin (Korsch 1977).

3. After subduction has ceased most of the volcanic arc sinks, presumably because of reduced head flow. It becomes the site of a sedimentary basin which derives much of the early part of its fill from erosion of the elevated accretionary prism. This basin is initially of transitional type; it has some of the characteristics of the classic fore-deeps that lie between kratons and the outer nappe zones of many fold belts.

The basin may be wider than the volcanic arc and is typically offset away from the fore-arc region so as to expose the oceanwards fringe of the volcanic pile. In some cases a substantial part of the volcanic arc remains unburied. These relationships suggest that the thinner crust of the back-arc region also contributes to the basin's location.

The volcanic arc that adjoins the accretionary prism shown in Fig. 4 is largely buried beneath the Gunnedah and Sydney Basins. Its outer parts, comprising coherent lavas, ignimbrites and sills, are intercalated within the predominantly volcaniclastic sediments of the western part of the Tamworth Belt of Fig. 4 (Crook & Powell 1976). North of the area shown in Fig. 4, in Queensland, the volcanic arc is more extensively exposed in the Connors Arch and Auburn Arch (Geological Society of Australia 1971). Upper Permian and Lower Triassic transitional basin sediments in the Sydney and Gunnedah Basins are largely derived from the accretionary prism of Fig. 4 (Branagan *et al.* 1976).

4. Magmatism occurs in thick accretionary prisms after subduction has ceased. Granitoid plutons derived from non-sedimentary protoliths, I-type granitoids of Chappell & White (1974), are emplaced into subduction complex and fore-arc sequences. Some rise sufficiently far to breach the surface, giving rise to post-arc silicic volcanics.

A subduction-related heat source can be invoked to generate these granitoids, if one assumes that the Benioff Zone stepped outwards away from the kraton. However this origin seems unlikely: intermediate-thickness and thin accretionary prisms display similar post-subduction plutonism (q.v.) which can in no way be related to subduction. I conclude therefore that post-subduction plutonism in thick accretionary prisms is likewise unrelated to subduction.

A deep-crustal origin can be inferred for these granitoids. I suggest that their emplacement marks the substantive ultrametamorphism of oceanic lithosphere previously altered by sea-floor processes, which had been trapped beneath the accretionary prism when subduction ceased. The ultrametamorphism, in addition to generating the granitoid magma, transforms the trapped oceanic lithosphere into materials characteristic of lower continental crust. This transformation contributes significantly to the kratonic character that is thereafter displayed by the whole fore-arc region.

The heat required for ultrametamorphism results from the interaction of four factors: (i) heat flow from the mantle; (ii) endothermic dehydration reactions in the now-stationary subducted oceanic lithosphere trapped beneath the accretionary prism; (iii) radiogenic heat from sediments within the accretionary prism, particularly in the subduction complex; (iv) the blanketing effects of sediments in the upper part of the accretionary prism.

The 'New England Batholith' and 'Stanthorpe Suite' of Fig. 4 are examples of post-subduction I-type granitoids, dated at 248–236 Ma and 225–222 Ma (Late Permian, Early Triassic), respectively (Korsch 1977). The 'Upper Permian acid-intermediate volcanic rocks' and the volcanogenic part of the 'Dummy Creek Association' of Fig. 4 are the volcanic expression of this magmatism.

The I-type granitoids transect and therefore post-date many of the faults shown in Fig. 4, attesting to the enhanced stability of this fore-arc region since granitoid emplacement.

5. With the conclusion of post-subduction magmatism, kratonization of thick accretionary prisms is complete. Thereafter they behave as kratonic areas, and acquire a veneer of platform cover sediments and eventually silcrete: representatives of the two final (kratonic) stages, 7

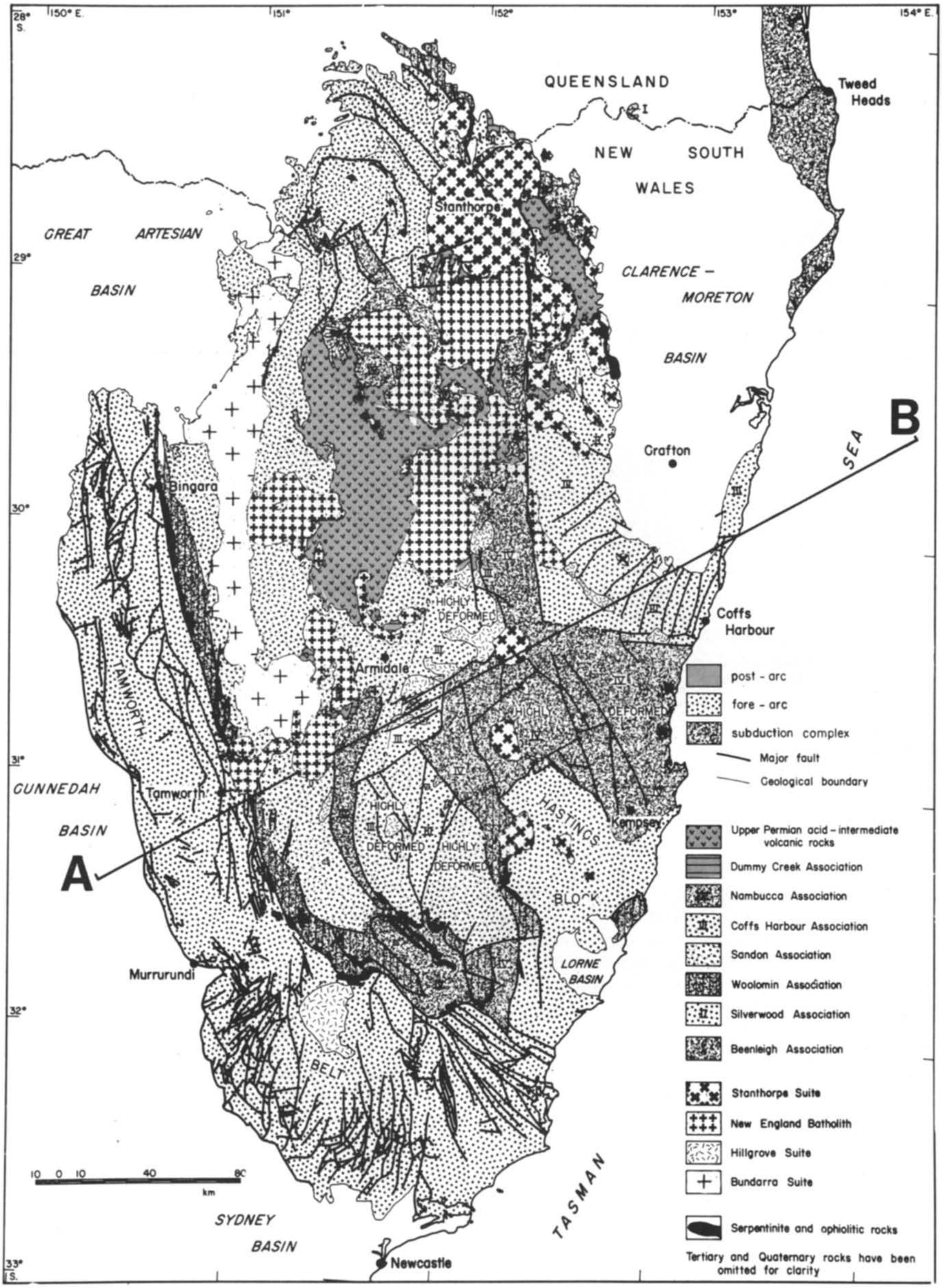

Fig. 4. Interpretation of the southern New England Fold Belt (13, 14 on Fig. 8). Geology after Korsch (1977).

and 8, in the kratonization sequence of Crook (1974). Earth movements are essentially epeirogenic in nature.

Because the outer parts of an accretionary prism are thinner than those closer to the magmatic arc, a somewhat different history may be expected in these areas. Intramontane transitional basins form. These are dominated by coarse fluvial clastics which are intercalated with, and succeed, the last acid to intermediate volcanics. The sequences in these basins are moderately deformed before being overlain by platform cover sediments.

That part of the New England Fold Belt shown in Fig. 4 has been kratonic in character since the mid-Triassic, experiencing epeirogenic uplift during the Cenozoic (Wellman & McDougall 1974). The Jurassic and Cretaceous sediments of the Great Artesian and Clarence–Moreton Basins represent platform cover. Silcretes were formed in Cenozoic weathering profiles on these sediments in the northwest, and in sub-basaltic fluvial sediments overlying Palaeozoic bed-rock elsewhere.

The Triassic rocks of the Lorne Basin and those in the lower part of the Clarence–Moreton Basin are examples of transitional basin sequences. They represent the 'transitional tectonic domains' of the Tectonic Map of Australia (Geological Society of Australia 1971, Plumb 1979).

6. The various features of thick accretionary prisms may be depicted by means of a schematic crustal cross-section (Fig. 1.3A). A specific example from the New England Fold Belt (Fig. 5) is based upon the surface geology of Fig. 4 and gravity data (B.M.R. 1976). It is schematized by omission of metamorphic overprint and by structural simplifications. The position of the former Benioff Zone shown in Fig. 5 is derived from that observed beneath the eastern Aleutians. However the New England fore-arc was evidently wider than the eastern Aleutians: some 400 km are preserved along the section line in Fig. 5 and an unknown additional width has been rifted away during the opening of the Tasman Sea in the Late Cretaceous and now lies in the Lord Howe Rise (Crook & Belbin 1978, fig. 2). Consequently the former Benioff Zone is shown at a somewhat deeper position below the eastern part of Fig. 4 than is observed beneath the eastern Aleutian fore-arc at corresponding distances from the volcanic arc. One may conjecture that the location of Late Cretaceous rifting that formed the Australian continental margin in this region was determined by significant thinning of the New England accretionary prism. From this perspective the exclusively submarine Cenozoic history of the Lord Howe Rise is not surprising.

The position of the former Benioff Zone shown in Fig. 5, while conjectural, gives an indication of the minimum extent to which material beneath the accretionary prism may have contributed to the formation of lower continental crust during the ultrametamorphism referred to above. However, if Benioff Zones rise beneath volcanic arcs as fore-arcs widen, in the manner shown in Fig. 2, the Benioff Zone beneath the New England fore-arc may have been much shallower when subduction ceased than is shown in Fig. 5. The contribution to the lower continental crust from beneath the accretionary prism would then have been correspondingly greater.

Accretionary prisms of intermediate thickness

The post-subduction history of accretionary prisms about 16 km thick differs from that of thick accretionary prisms. Kratonization is completed in a two-step process (Fig. 1). Examples in the Tasman Geosyncline include the Dundas Trough and related depositional and volcanic areas of northwestern Tasmania (Figs. 6 and 7).

The chief characteristics of this history are as follows:

1. When subduction ceases, the arc terrain is either already significantly below sea level or sinks to such depths. The depth attained depends upon the crustal thickness. To judge from the post-arc sediments, which are of neritic facies, the depths usually attained are those typical of continental shelves and marginal plateaux. However areas in the vicinity of the volcanic arc may initially sink deeper to become sites of flysch accumulation and only later acquire shallower-water deposits. This anomaly is probably related to the rapidly declining heat flow in this part of the arc terrain.

In northwest Tasmania, Upper Cambrian quartz-rich post-arc flysch up to 2 km thick overlies the ?Proterozoic-Middle Cambrian Mount Read Volcanics, which constitute the volcanic arc, both north and south of Queenstown (Figs. 6 and 7). Elsewhere the ?Proterozoic-Middle

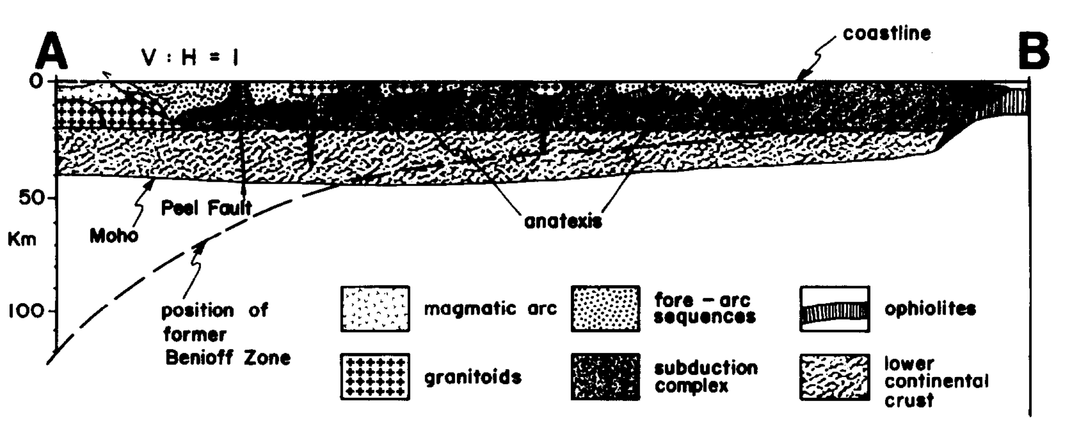

Fig. 5. Schematic 1 : 1 cross-section along A–B in Fig. 4. Structure is simplified and metamorphic overprint is omitted.

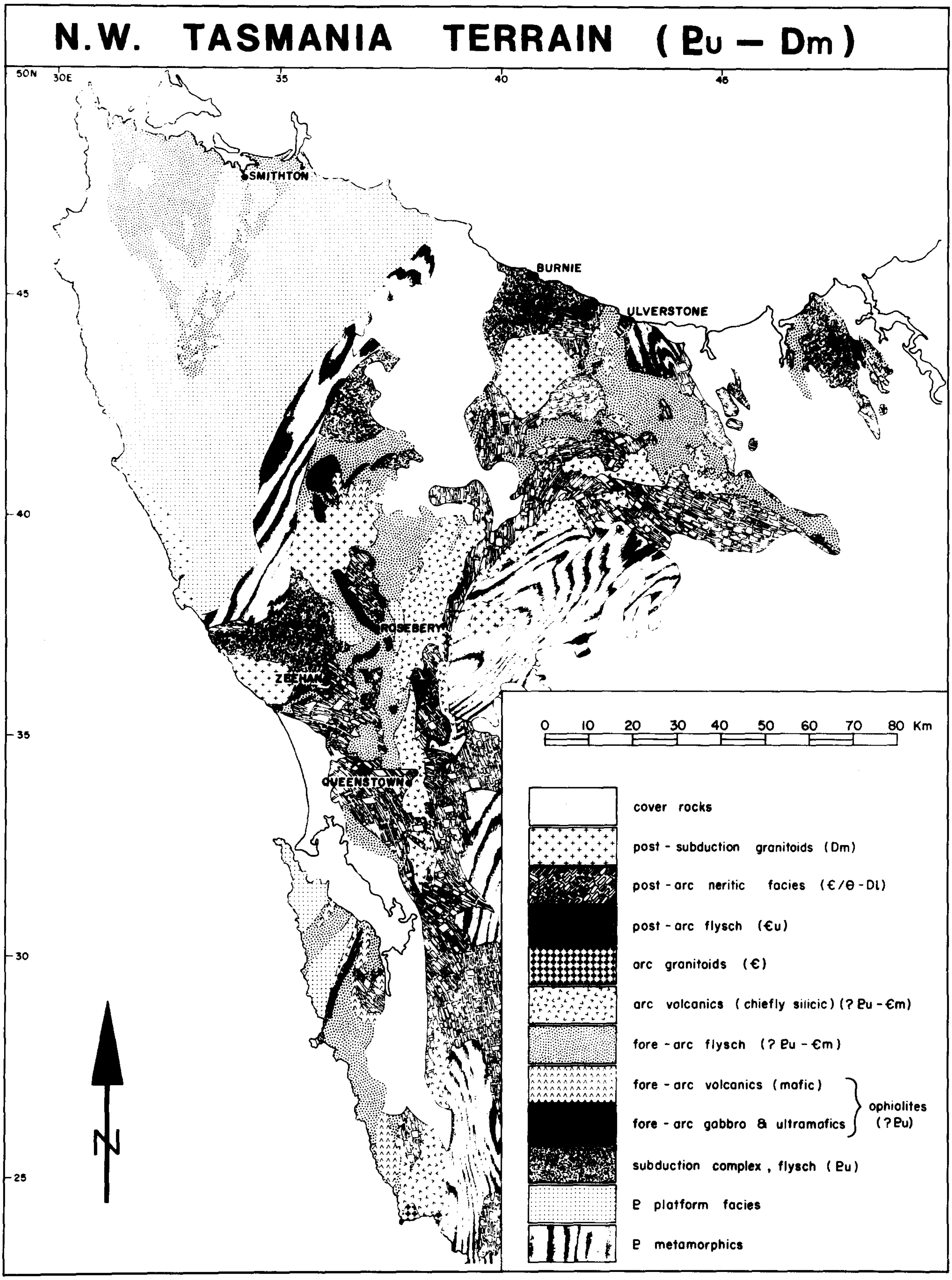

Fig. 6. Interpretation of the Upper Proterozoic to Middle Devonian Northwest Tasmania Terrain (15, 16 on Fig. 8). Geology after Williams *in* Scheibner (1978). Age ranges shown for Proterozoic (P), Cambrian (€), Ordovician (O) and Devonian (D), with lower (l), middle (m) and upper (u) divisions specified.

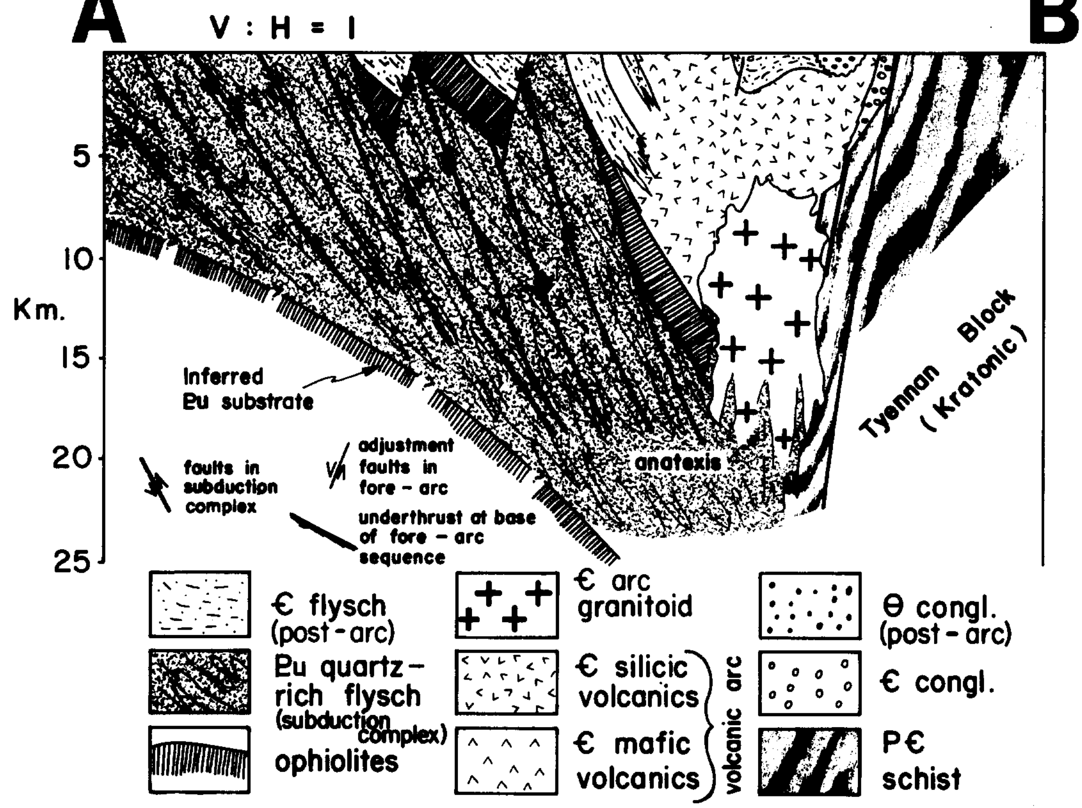

Fig. 7. Schematic 1:1 cross-section along a NW–SE transect located mid-way between Zeehan and Rosebery on Fig. 6. Metamorphic overprint omitted. Ages abbreviated as in Fig. 6.

Cambrian fore-arc flysch (Dundas Group), its ophiolite substrate, and parts of the Upper Proterozoic quartz-rich flysch (Oonah and Burnie Formations) that forms the subduction complex (Fig. 7), are all overlain by the Cambro-Ordovician Owen Conglomerate, the basal unit of the post-arc neritic sequence (Williams *in* Scheibner 1978). The Owen Conglomerate also overlies the post-arc flysch and in many places lies directly on the volcanics (Fig. 7). Both the conglomerate and the quartz-rich post-arc flysch were derived from the Tyennan Block, a Proterozoic metamorphic terrain lying to the east of the volcanic arc (Fig. 6).

2. From the onset of post-arc neritic sedimentation, arc terrains behave, for a considerable time, as quasi-platforms. They are tectonically quiescent sites of shallow marine-neritic sedimentation. The sedimentary association does not differ appreciably from that which forms the platform cover of kratons. The post-arc neritic sequence is typically 2–3 km thick.

Despite its quiescence during post-arc sedimentation, the region is not a true kratonic platform. Significant deformation and granitoid plutonism are merely deferred.

The post-arc neritic sequence in northwest Tasmania (Fig. 6) comprises the Cambro-Ordovician Owen Conglomerate (500–1200 m thick) and a Lower Ordovician to mid-Lower Devonian limestone–quartz sandstone–mudrock sequence, the Gordon Limestone and Eldon Group, 1800–2000 m thick (Williams *in* Scheibner 1978).

3. Post-arc sedimentation on intermediate-thickness accretionary prisms is terminated by regional deformation of the post-arc sequence and the accretionary prism. This is followed by emplacement of granitoid plutons. These two events constitute the second step in the two-step kratonization process (Fig. 1).

This deformation and the presence of the post-arc sequence make it difficult to determine how much structural rearrangement occurs in intermediate-thickness fore-arcs during subduction. However, metamorphic grades in all sequences are typically low: blueschists do not occur.

The apparent absence of high-grade low pressure metamorphics may simply reflect the absence of subducted heat sources during the evolution of the examples on which this part of the model is based. However the absence of blueschists suggests that the degree of complexity attained by the fore-arc during subduction may be less than that shown by thick accretionary prisms.

Deformation of the northwest Tasmanian post-arc sequence and the underlying arc terrain occurred in the Middle Devonian. Granitoids were emplaced later in the Middle Devonian (Williams *in* Scheibner 1978).

The terminal deformation and plutonism in northwest Tasmania does not seem to result from a collision between this adjacent terrain and another tectonic unit. Rather it seems to reflect the attainment, by the crust of the arc terrain, of a critical thickness corresponding to that of typical upper continental crust.

The emplacement of granitoids — of S-type in the northwest Tasmanian case, although I see no reason why I-type granitoids should not occur in other terrains — implies significant ultrametamorphic reconstitution of the lower part of the subduction complex, and inferentially of the underlying old oceanic lithosphere. This is borne out by the subsequent history of the fore-arc region.

4. Following granitoid emplacement, intermediate-thickness accretionary prisms and their overlying deformed post-arc sequences behave as kratonic blocks. Later sediments and volcanics accumulate in shallow marine to subaerial conditions and are of typical undeformed platform cover facies. In northwest Tasmania the platform cover comprises part of the Late Carboniferous to Triassic Parmeener Supergroup, Jurassic dolerite sills and younger strata (Williams *in* Scheibner 1978).

Thin accretionary prisms

Kratonization of thin accretionary prisms ($\leq$ 12 km thick) is a three-step process (Fig. 1) involving substantial post-arc sedimentation, deformation, metamorphism and plutonism. The volcanic arcs and accretionary prisms are typically deeply buried, being exposed only in the vicinity of major faults. Consequently the location, age and extent of each accretionary prism must in part be inferred from indirect evidence. Examples of thin accretionary prisms in the Tasman Geosyncline include the Palaeozoic terrain of central Victoria (3, 4, and 5, on Fig. 8), comprising the Stavely High, Ballarat Trough, Heathcote High and Melbourne Trough of Crook & Powell (1976, fig. 0.1).

Salient features of the post-subduction history of thin accretionary prisms are as follows:

1. After subduction ceases the arc terrains remain at bathyal depths (Fig. 3) or sink to such depths. They are subsequently buried beneath post-arc flysch. This substantially obscures the accretionary prism so that the determination of its complexity at the cessation of subduction is very difficult. However neither blueschists nor syn-subduction granitoids are known to occur in thin accretionary prisms. Granitoids may be present within the volcanic arc, as part of the subduction-related magmatic edifice.

The thickness of post-arc flysch is substantial, 6–8 km being typical. The duration of flysch sedimentation is related to the rate of sediment supply: flysch continues to accumulate until the upper crust reaches a critical thickness of about 20 km, whereupon the next step in the kratonization process ensues.

In western and central Victoria, part of the volcanic arc is exposed as the Stavely High (3, Fig. 8) of inferred Early to Middle Cambrian age. The subduction complex is nowhere exposed: its presence at depth is inferred from the widespread occurrence of post-subduction S-type granitoids. Exposures of the fore-arc sequence are quite limited, the chief being in the Heathcote High (between 4 and 5, Fig. 8) where a Lower to Middle Cambrian volcanogenic sequence (Heathcote Greenstone and Knowsley East Formation) occurs (Douglas & Ferguson 1976).

The post-arc flysch, which is quartz-rich, ranges in age from probably Late Cambrian to Late Ordovician west of the Heathcote High (4, Fig. 8) where the thickness is 4.5 km, and from Early Ordovician to Early Devonian in the Melbourne Trough (5, Fig. 8) east of the Heathcote High, where the thickness may reach 10 km. The flysch body prograded eastwards throughout the region: on the Heathcote High, some 200 km east of the volcanic arc, the earliest post-arc strata are Upper Cambrian cherty argillites (Goldie Shale). These are overlain by Lower Ordovician flysch.

The post-arc flysch in this region is part of an extensive body of quartz-rich flysch which forms the post-arc sequence on adjacent southeast Australian arc terrains (1, 2; 6, 7; 8a, 9c; Fig. 8). The whole body is probably an analogue of the modern Bengal Fan (Cas *et al.* in press).

2. Flysch accumulation is terminated by regional deformation which marks the attainment of continental freeboard. Emplacement of granitoids follows, not necessarily immediately. Intensity of deformation is variable: extensive areas display only weak to moderate deformation, with low grades of metamorphism. Other areas, some of significant regional extent, are strongly deformed; many but not all display low pressure metamorphism of intermediate to high-grade which is synchronous with granitoid emplacement. The granitoids are typically either all S-type or all I-type but both are present in some regions. Some granitoids breach the surface, producing the silicic volcanics which are typical of the third step in the kratonization process.

Deformation of the Ballarat Trough, Victoria (4, Fig. 8) may have occurred as early as the Late Ordovician (Crook & Powell 1976). In the western part of this trough where the buried accretionary prism was thickest, the granitoids, which are mostly S-type, are Silurian or Lower Devonian (VandenBerg *in* Scheibner 1978), deformation is commonly strong and metamorphic grade is locally high. In the eastern part where the accretionary prism is thinner, deformation is moderate to weak, metamorphic grade is low and the granitoids, mostly of S-type, are Upper Devonian. Weak deformation and metamorphism of the Melbourne Trough (5, Fig. 8) which overlies the thinnest part of the accretionary prism, occurred in the Middle Devonian. S-type and some I-type granitoids, with associated volcanics, were emplaced in the Late Devonian.

Inferentially the S-type granitoids of central and western Victoria are derived from the deeply buried subduction complex within the accretionary prism. Ultrametamorphism is implicated and, because some I-type granitoids occur as well, I therefore infer that reconstitution of the underlying old oceanic lithosphere also occurred. Because these granitoids were emplaced long after subduction had ceased, the heat required for their generation must have been unrelated to the subduction of major heat sources.

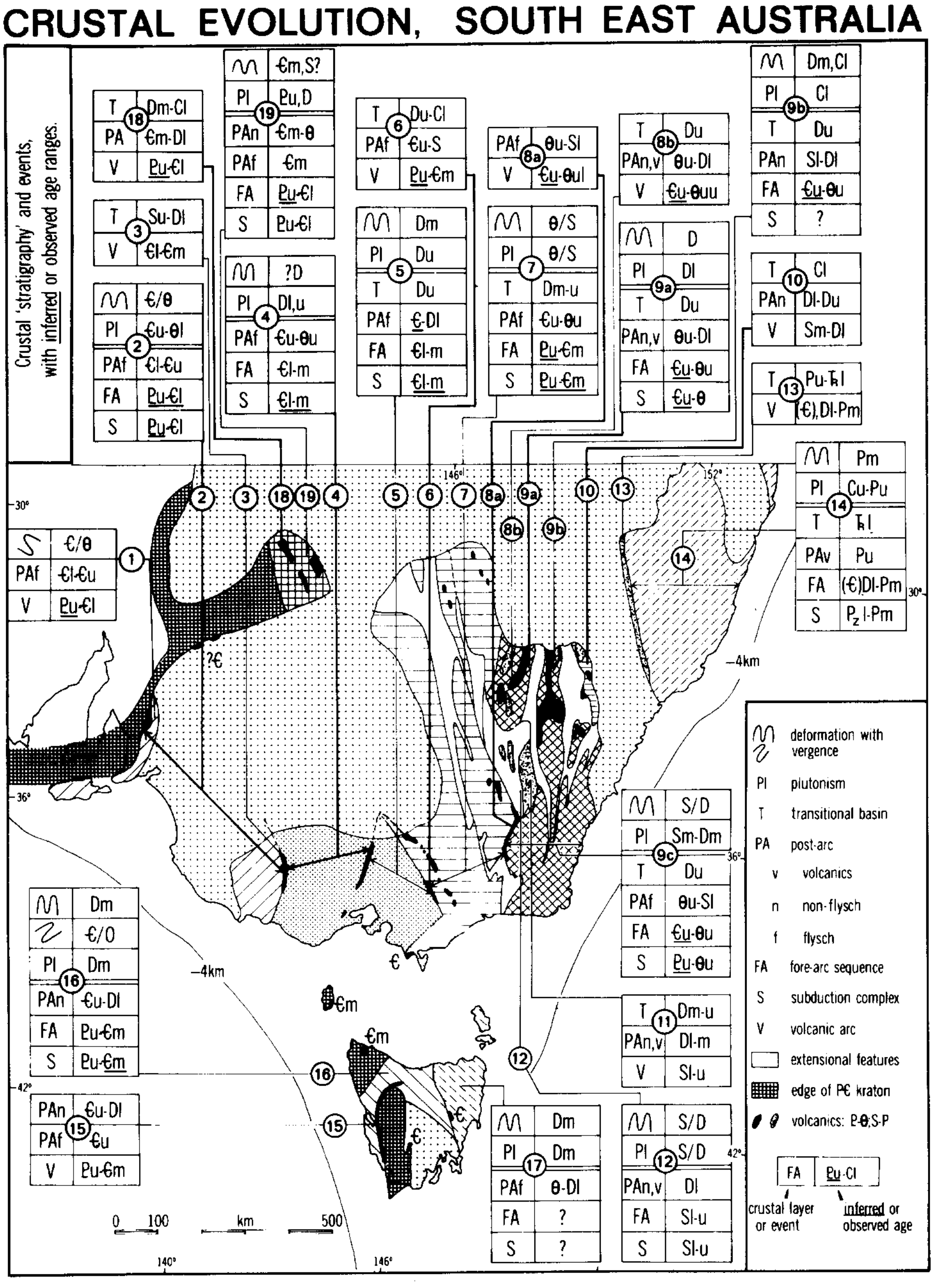

Fig. 8. Crustal 'stratigraphy' and evolution of southeast Australia, showing arc terrains (areas of uniform hachuring), extensional features and areas of continuous platform cover. Transitional basins are omitted. Numbered boxes summarize the 'stratigraphy' and age relationships of strato-tectonic units in the upper continental crust (lower parts of boxes) and the timing of granitoid plutonism and pervasive deformation (upper parts of boxes) for designated regions within each arc terrain. Key to the circled numbers is: 1: Padthaway Terrain-volcanic arc; 2: Padthaway Terrain-fore-arc; 3: Central Victorian Terrain-volcanic arc; 4: Central Victorian Terrain-inner fore-arc; 5: Central Victorian Terrain-outer fore-arc; 6: Riverina Terrain-volcanic arc; 7: Riverina Terrain-fore-arc; 8a: Maneroo Terrain-volcanic arc Kiandra segment; 8b: Maneroo Terrain-volcanic arc Parkes segment; 9a: Maneroo Terrain-fore-arc Parkes segment; 9b: Maneroo Terrain-fore-arc Molong segment; 9c: Maneroo Terrain-fore-arc Monaro segment; 10: Coolamigal Terrain; 11: Tumut Terrain-volcanic arc; 12: Tumut Terrain-fore-arc; 13: New England Terrain-volcanic arc; 14: New England Terrain-fore-arc; 15: NW Tasmanian Terrain-Mt. Read volcanic arc; 16: NW Tasmanian Terrain-fore-arc; 17: Mathinna Terrain; 18: Tibooburra Terrain-volcanic arc; 19: Tibooburra Terrain-fore-arc. Symbols in the right hand parts of each box represent age ranges from Proterozoic (P̲) through the Palaeozoic (Pz) periods (€, Θ, S, D, C, P) into the Triassic (Tr). The lower, middle and upper parts of these systems or erathems are indicated by l, m, and u. Other symbols are explained on the diagram.

3. Although the deformation, metamorphism and plutonism just described implies substantial reorganization of the crust, attainment of kratonic stability requires a third step, characterized by transitional tectonism (Scheibner 1976) (Fig. 1). This is partly exemplified by some of the transitional domains shown on the Tectonic Map of Australia (Geological Society of Australia 1971), which are transitional in character, and in space and time, between the rocks of the orogenic domain and the platform cover (Plumb 1979).

Typically, transitional basins develop about depocentres which preferentially overlie the buried volcanic arcs and the inferred thinner parts of buried accretionary prisms. Some of these basins are rift-like, commencing with significant effusions of silicic and lesser mafic volcanics which pass laterally and vertically into terrigeneous sediments (Fergusson *et al.* 1979). Other basins lack volcanics. Coarse fluvial sediments predominate; shallow marine deposits are less common. Red beds are typical. Thicknesses of several kilometres are not unusual. The transitional basin sequences are weakly to moderately deformed. Some are intruded by granitoids.

Away from the principal depo-centres, some volcanic piles develop in association with cauldron subsidences and subvolcanic granitoids. Fluvial sediments are a minor component of these accumulations.

Examples of transitional basin sequences include the Siluro–Devonian Grampians Group, 6 km thick, which overlies and substantially obscures the volcanic arc of the Stavely High (3, Fig. 8). This sequence is intruded by Lower Devonian granitoids. Several Upper Devonian volcanic cauldrons occur in the eastern Ballarat and Melbourne Troughs (4, 5, Fig. 8). An Upper Devonian–Lower Carboniferous transitional basin sequence 6 km thick, with abundant volcanics, the Mt. Howitt Province (Douglas & Ferguson 1976), overlies the volcanic arc of the next adjacent arc terrain in the Mt. Wellington High (7, Fig. 8). This sequence was deformed weakly during the Carboniferous period.

Unrelated but superficially similar features may develop on top of some buried thin accretionary prisms after continental free-board is attained. Thick shallow marine to subaerial silicic volcanogenic accumulations, termed 'volcanic arches' by Scheibner (1976), develop as linear belts which are in part fault-founded. Smaller examples, such as the southeastern extensional features of Fig. 8, resemble the lower parts of volcanogenic transitional basin sequences but lack their associated terrigenous clastics. However other examples, such as on the margins and at the southern end of the Hill End Trough (between 9b and 10, Fig. 8), are evidently related to the formation of a new inter-arc basin which was not kratonized until much later. Commonly these silicic volcanics are strongly deformed. Moreover, significantly younger transitional basin sequences are associated with the arc terrains that bear the silicic volcanics. I conclude therefore that these silicic volcanics are not formed as part of the normal process by which thin accretionary prisms are incorporated into continental crust. Rather they reflect a phase of rifting of incompletely kratonized crust that is analogous to the Cenozoic Basin and Range Province of the western United States.

4. Following deformation of their associated transitional basin sequences, the now-fully evolved arc terrains act as kratonic blocks and acquire a platform cover. In central Victoria the platform cover commences with undeformed Lower Permian shallow marine and subaerial glacigene sediments of the Bacchus Marsh district (Douglas & Ferguson 1976).

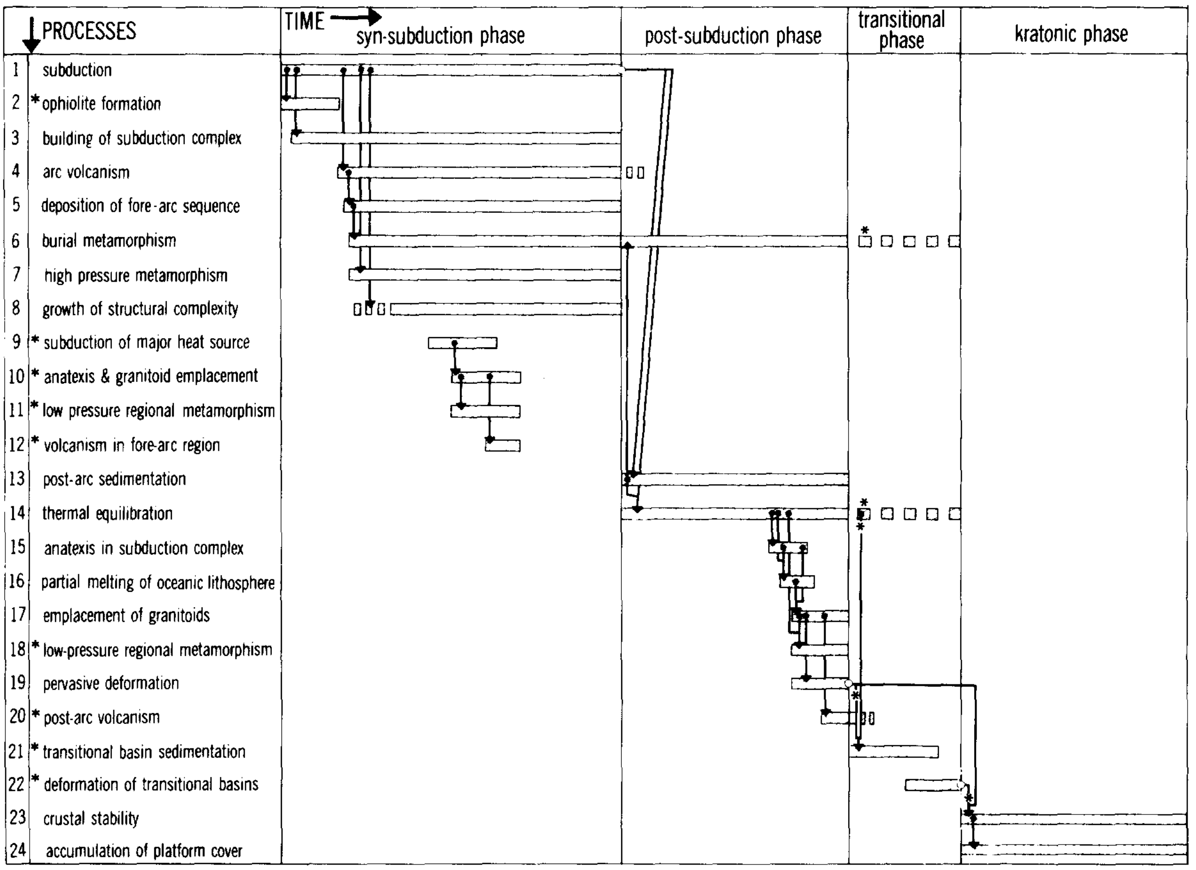

Fig. 9. Temporal relationships and causal connections between processes contributing to kratonization. Causal connections are indicated by arrows linking causative and resultant processes. Closed circles indicate activity of a process as a cause; open circles indicate cessation of activity of a process as a cause. Asterisks indicate process that do not invariably occur.

THE ORIGIN OF CONTINENTAL CRUST

The model of fore-arc evolution here described, implies a particular view of the origin and evolutionary dynamics of continental crust. Continental crust, of kratonic character and thickness, is pre-eminently a result of fore-arc evolution, the upper crust being derived from accretionary prisms and their associated volcanic arcs and overlying post-arc materials.

The various dynamic processes by which crustal materials are assembled and kratonized are not haphazardly related, as implied by Coney (1970) and Dewey (1975). Rather, they constitute an ordered sequence with causal connections which are quite systematic, and explicable without recourse to adventitious circumstances. Thus kratonization as a whole may be conceived of and modelled as a unified, internally deterministic and self-sustaining phenomenon, the duration of which is directly dependent upon the rates of the various contributing processes.

The succession of processes which results in the formation of kratonic continental crust may be summarized as follows, temporal relationships and causal connections between processes being shown in Fig. 9:

1. initiation of subduction;
2. ± development of ophiolite as the initial substrate of the fore-arc region;
3. building of the subduction complex;
4. arc volcanism;
5. deposition of the fore-arc sequence;
6. burial metamorphism of the subduction complex and fore-arc sequence;
7. high pressure, but not necessarily high grade, metamorphism of parts of the subduction complex;
8. development of structural complexity in the growing accretionary prism;
9. ± subduction of a major heat source;
10. ± anatexis in the subduction complex and emplacement of S-type granitoids at higher levels;
11. ± concomitant low pressure regional metamorphism;
12. ± volcanism in the fore-arc region if granitic magma breaches the surface;
13. post-arc (i.e. post-subduction) sedimentation;
14. gradual thermal equilibrium of assembled materials, as a consequence of cessation of subduction;
15. anatexis in the subduction complex;
16. partial melting of the oceanic lithosphere trapped beneath the accretionary prism;
17. emplacement of granitoids at higher levels in the arc terrain;
18. ± concomitant low pressure regional metamorphism;
19. deformation, particularly evident at higher levels in the arc terrain, as a consequence of emplacement of granitoids;
20. ± post-arc (i.e. post-subduction) silicic volcanism caused by granitoids breaching the surface;
21. ± development of and sedimentation in transitional basins;
22. ± deformation of transitional basin sequences;
23. onset of kratonic crustal stability;
24. accumulation of platform cover.

The ± symbols indicate processes which do not occur invariably.

Evidently, adventitious collisions between continental blocks are not a necessary part of kratonization. Indeed, the Wilson Cycle of ocean opening and closing is disruptive to the growth of continents by kratonization. Ocean opening provides new oceanic crust which must be processed; the closing of oceans emplaces fore-arc terrains, which are often incompletely evolved, onto continental margins where they become major sediment source areas rather than lateral additions to kratons.

Although the nature and origin of the kratonic upper continental crust formed by fore-arc evolution are evident, the same cannot be said for the associated lower continental crust. Clearly the lower continental crust must have formed during the kratonization of the upper crust, but this conclusion is compatible with at least two possible origins for the lower crust which are themselves not entirely mutually exclusive.

The first possibility is that the lower continental crust is derived directly from the mantle, being underplated onto the upper crustal materials. Underplating could be gradual; or it could be a relatively brief episode late in the kratonization process, in which the mantle responded to the presence of some critical thickness of upper crustal materials. A brief episode of underplating would provide an explanation for the post-subduction granitoid plutonism observed in the upper crust, because of the heat transfer from the mantle to the upper crust during underplating.

The second possibility is that the lower continental crust has subducted oceanic lithosphere as its protolith. This protolith could be emplaced in either of two ways. Peeling-off of the upper parts of the down-going lithosphere could occur beneath the accretionary prism. This would cause progressive lateral accretion of mafic and ultramafic materials at depth as the accretionary prism widened.

Alternatively, emplacement may occur in the final stages of subduction, the protolith simply being that part of the down-going oceanic lithosphere which was trapped beneath the accretionary prism when subduction ceased. Karig *et al.* (1976) have shown that Benioff Zones have shallow dips beneath wide accretionary prisms, so that an extensive slab of recently-subducted oceanic lithosphere lies beneath the accretionary prism at all times.

The cessation of subduction which may be required for emplacement of the protolith may be due to consumption of all available oceanic lithosphere and concomitant suturing of the fore-arc region to another crustal block — typically an adjacent back-arc region. Alternatively, if Benioff Zones become shallower beneath volcanic arcs as the arc-trench gap widens (Fig. 2), as data in Karig *et al.*

(1976) seem to suggest, subduction may cease because of mechanical interference between the down-going lithosphere and the lithosphere behind the volcanic arc after the asthenosphere beneath the volcanic arc has been removed.

Whatever its mode of emplacement, the protolith must be ultrametamorphosed if the crust is to attain kratonic stability. Ultrametamorphism is not immediate: it follows attainment of continental thickness by the upper crust and is marked by the emplacement of post-subduction granitoids. Consequently it is a response to the heat balance within the assembled crustal materials.

This origin for lower continental crust implies that this part of the crust is significantly more mafic than is customarily assumed. However this is consistent both with geophysical data, as summarized in Mueller's (1977) new model of continental crust, and with the observed composition of the Ivrea Zone of the Alps, which is the outcrop of the lower crust of northern Italy.

In constructing the model of fore-arc evolution and continental growth I have preferred the second possible origin for lower continental crust rather than the first for two related reasons, one conceptual and the other practical. Conceptually, derivation of lower continental crust from subducted oceanic lithosphere makes for a neater and more unified model. The protoliths for both lower and upper continental crust are formed and assembled as a direct consequence of the main driving process of the first major phase of fore-arc evolution (Fig. 9): subduction.

Derivation of lower continental crust by underplating poses a significant practical problem: disposal of the oceanic lithosphere which was trapped beneath the accretionary prism when subduction ceased. The post-subduction thermal histories of intermediate-thickness and thin accretionary prisms are inconsistent with any significant removal of the trapped oceanic lithosphere immediately subduction ceases or soon afterwards. If removal by sinking to depth occurs later it must be concomitant with underplating, in which case a physical reason for sinking is lacking, given the observed composition of the Ivrea Zone. Furthermore the delay in sinking must be explained.

Clearly the genesis of I-type granitoids is a problem if oceanic lithosphere is the protolith for lower continental crust, given the known composition of that lithosphere in the vicinity of modern spreading centres. However oceanic lithosphere trapped beneath accretionary prisms is likely to have been significantly altered by hydrothermal circulation and halmyrolysis which may be pervasive. Hydration of primary minerals and growth of phyllosilicates which increases the K_2O-content by a factor of 7 to 10 times (S. Hart personal communication) are likely. Although fore-arc related ophiolites probably experience a somewhat different alteration history, the common occurrence in them of plagio-granites suggests that oceanic lithosphere has some potential as a source of granitoids. However mass-balance problems are unresolved.

The widespread occurrence of silicic granulites is not decisive evidence against oceanic lithosphere being the protolith for the lower parts of 30–50 km-thick continental crust. A likely site for the formation of silicic granulites is the lower crust of continent–continent collision zones, where crustal thickness are such that the silicic upper crust of the over-ridden continental plate is depressed to depths appropriate for granulite-facies metamorphism.

IMPLICATIONS OF THE MODEL

This model, like other general tectonic models has diverse implications. Several implications are summarized below; they will be discussed in more detail elsewhere.

1. Fore-arc evolution of the type described above is pre-eminent among the diverse processes of continental growth or accretion. It was responsible for more than 1000 km lateral growth of the Australian kraton during the Palaeozoic. The Mongol–Okhotsk Mobile Belt (Crook 1974) and the north-east African–west Arabian Mobile Belt (Frisch & Al-Shanti 1977) appear to have formed in a similar manner. Archean kratonic nucleii may also have formed in this way (Crook 1974, Langford & Morin 1976).

2. The model implies a 'stratigraphy' for the upper continental crust of southeast Australia (Fig. 8) and for continental crust in general. This 'stratigraphy' is closely similar to that proposed independently, on geophysical grounds, by Mueller (1977, fig. 13), whose new model of continental crust corresponds, to a first approximation, to ultrametamorphosed versions of 3A, 4A and 4B on Fig. 1.

3. Upper continental crust is largely of sedimentary origin, the major component being flysch and its metamorphic and anatectic granitoid derivatives. Lower continental crust is an ultrametamorphic derivative of oceanic lithosphere. Its composition resembles that of the Ivrea Zone of northern Italy.

4. Granitoids are a characteristic rock type of fore-arc regions. Some, of S-type, are syn-subduction granitoids arising from the subduction beneath the accretionary prism of major heat sources which cause partial melting in the subduction complex. Others, both S- and I-type, are post-subduction granitoids emplaced into the accretionary prism and its overlying post-arc sequence after subduction has ceased. The source materials for these post-subduction granitoids lie in the subduction complex (S-type) and the old oceanic lithosphere which underlies it (I-type). The ultrametamorphism which generates the post-subduction magmas is a result of the *in situ* heat balance, rather than adventitious heat sources.

5. Silicic volcanism 'anomalously close' to the trench, as encountered for example in DSDP hole 439 off Japan (von Huene *et al.* 1978), is a surface manifestation of fore-arc granitoid plutonism.

6. Inter-arc basins and smaller extensional features in some cases form in incompletely kratonized arc terrains immediately following silicic volcanism, when the tensional effects of magmatism are sufficiently large to disrupt the accretionary prism. The Hill End Trough (between 9b and 10, Fig. 8) (Cas & Jones 1979), is an example.

7. Paired metamorphic belts, and some unpaired low-pressure metamorphic belts, are a product of fore-arc evolution. Paired belts occur in thick accretionary prisms, the high-pressure member being located in the subduction complex and the low-pressure member being associated with regional aureole granitoids (White *et al.* 1974), principally in fore-arc basin sequences. Unpaired low-pressure belts occur in both the accretionary prism rocks and post-arc sequences overlying thin accretionary prisms.

8. Ophiolites occur in two distinct settings in fore-arc terrains: (a) in the subduction complex where they are typically dismembered and incomplete and (b) as the substrate to fore-arc basin sequences, where their preservation is generally better.

9. Transitional basin sequences dominated by paralic and fluvial red-beds are characteristic late-stage features in the evolution of fore-arc terrains. They are preferentially sited on top of volcanic arcs and the inferred thinner parts of accretionary prisms. Those basins associated with thick accretionary prisms form soon after subduction ceases but are comparatively short-lived. Those associated with thinner accretionary prisms form long after subduction has ceased; thick sedimentary sequences are typical.

10. Orogeny in evolving fore-arc terrains seems only incidentally related to collisions between crustal blocks, which are generally island arcs that meet fore-arc to back-arc. The Wilson Cycle applies poorly, if at all, to these terrains. Orogeny is apparently a deformational response to thermal equilibration and mass transfer within the newly assembled continental crustal materials once they have attained a critical thickness of *ca.* 20 km.

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More generally I wish to acknowledge those who have to some degree shared my perception that the Tasman Geosyncline displays some rather unusual features. In particular Nikita Bogdanov, Clark Burchfiel, John Rodgers and Peter Wyllie have thereby encouraged my study of its peculiarities, without necessarily endorsing this particular outcome.

This model could not have been formulated at this time without the major advances in understanding of granitoids made by B. W. Chappell, L. A. I. Wyborn and A. J. R. White, whose sustained critique of my earlier attempts to interpret the geosyncline has been a constant stimulus. I hope that this paper and its more detailed sequels will convince them that the Tasman Geosyncline did not develop over a Precambrian crystalline basement.

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