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RIFTS AND SUTURES OF THE WORLD

by

Albany Global Tectonic Group
State University of New York at Albany

for

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Compiled for

Geophysics Branch
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0. INTRODUCTION

Wilson's recognition of plate structure in 1965 immediately proved powerful in analysis of the structure of the oceans and in studies of the instantaneous properties of the lithosphere but application of plate tectonic theory to understanding of the historical geology of the continents did not begin until Wilson (1968) pointed out, in an address to the American Philosophical Society, that recognition of the earth's plate structure required that we consider historical geology as recording complex interwoven cycles of ocean opening and closing.

These cycles can be considered as extending from intracontinental rifting through the opening of oceans like the Atlantic and the Pacific to the Mediterranean-Alpine condition and eventually to the style of continental collision represented in the Himalaya. Wilson elaborated this cyclical idea in the second edition of his text (Physics and Geology, Jacobs, Russell and Wilson, 1972) and Dewey and Burke (1974) suggested the name "Wilson cycle".

All oceans are short lived and ocean floor lithosphere is fated to be either obducted or subducted. At present no ocean floor on earth (with the possible exception of material underlying the North Caspian depression) is more than 200 m.y. old so that earth history before this time has to be studied within the continents.

Continents can be expected to contain both rift-systems marking places in which the Wilson cycle has not developed further than an initial phase and sutures marking places in which cycles of opening and closing have proceeded to completion. Although both rifts and sutures include structures made in diverse and complex ways (see reviews of

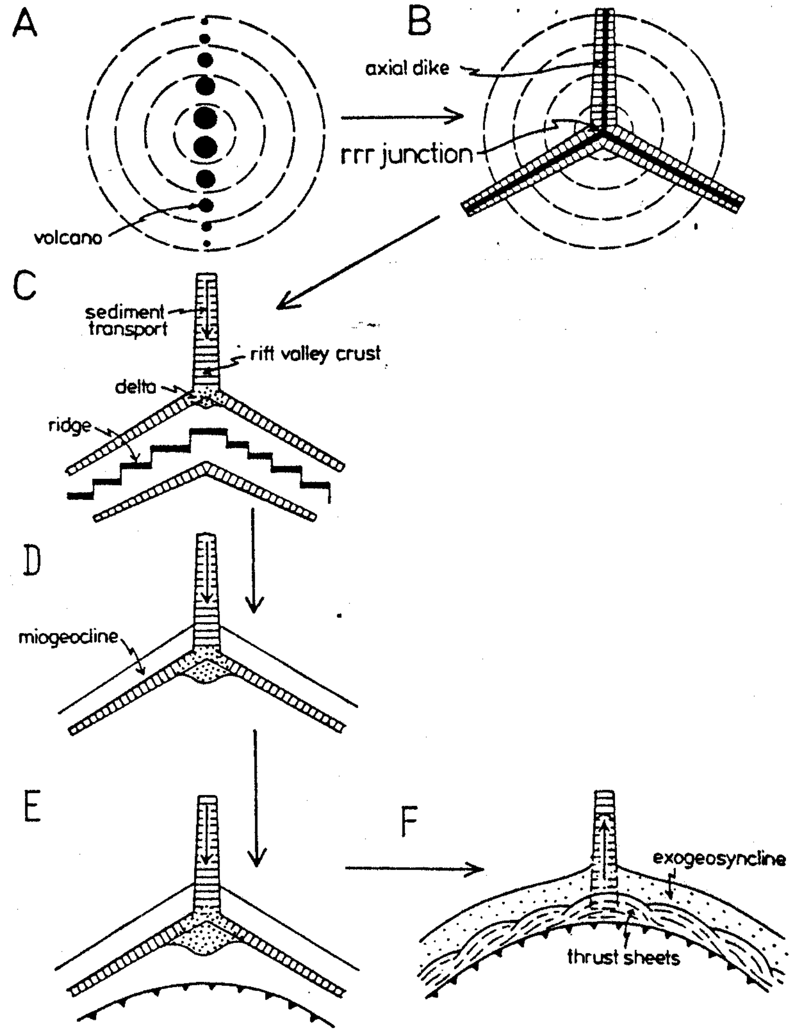

Figure 0.1. <u>Formation of Rifts and Sutures in the Wilson Cycle</u>. Intra continental vulcanism (A) is sometimes followed by intra-continental rift development (B). These rifts may all fail to develop and be preserved within the continent or two may develop to form an ocean (C) leaving only one as a failed rift at an Atlantic type ocean margin. After the ocean matures (D) a continent will be carried across the ocean closing it (E) and eventually colliding with the continent containing the failed rift (F) thus producing a suture and the kind of rift known as an aulacogen.

sutures by Dewey, 1976 and rifts by Burke, 1977) these two classes of structure form the basic framework of the continents. Rifts and sutures can be expected to have distinctive geophysical signatures as they extend, at the time of their formation, through the entire lithosphere. Moreover, sutures can be expected to separate areas within continents that have had different tectonic histories and therefore sutures can also be expected to separate areas with differing geophysical signatures.

In this report the distributions of rifts and sutures are plotted on maps and discussed in an explanatory text. The criteria used for identifying sutures and rifts are those discussed by Dewey (1976) (Burke, Dewey and Kidd, 1976) and Burke (1977) in their reviews of these features. Major responsibility for preparation of the report within the Albany Global Tectonic Group has been shared as follows.

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1.0. AUSTRALASIA

1.1. INTRODUCTION

Australia is unusual among continents in having no active plate boundaries within it. For this reason it is a particularly interesting and appropriate setting in which to study the geophysical effects of old plate boundaries. Both ancient rifts and old suture zones can be recognized within the continent, and differences related to their distribution can be sought on several different scales.

On the largest scale the eastern third of the continent may be characterized as the product of Phanerozoic accretion in contrast to the Precambrian western two-thirds. Within both the Precambrian and the Phanerozoic, domains of distinct physical properties may be mapped on scales ranging from thousands to hundreds to even tens of kilometers using the suture and rift distributions described in this report.

1.2. RIFT SYSTEMS

Rifts are defined as places where the entire thickness of the lithosphere has ruptured in extension (Burke, 1977a). They generally form where the lithosphere is unusually thin and, for this reason, are most common within oceans. Because all ocean floor is eventually destroyed by subduction or obduction, it is only the less common kinds of rifts, formed within continents, that are preserved in the older part of the geological record. In our review of the rift systems of Australasia the youngest, active rifts are described first and then progressively older structures. This corresponds rather closely to a progression from the active rifted plate boundaries around Australasia inward through the Mesozoic and Cenozoic rifts at the margins of the continent to the older rifts preserved in the interior.

1.2.1. ACTIVE RIFTS

1.2.1.1. Australia

A spreading ridge forming the boundary between the Indian and Antarctic Plates lies south and west of Australia (Weissel and Hayes, 1972). It extends through transform offsets to the Macquarie triple-junction and continues as a ridge separating the Pacific and Antarctic Plates (Molnar et al., 1975). Elsewhere (Figure 1.1) Australasia is bounded by convergent and transform boundaries, and rifting where it occurs close to these boundaries (e.g., in the Coral Sea to the north-east of Australia), is a secondary phenomenon related to imperfect convergent and transform motion.

Within the Australian continent there is a concentration of small

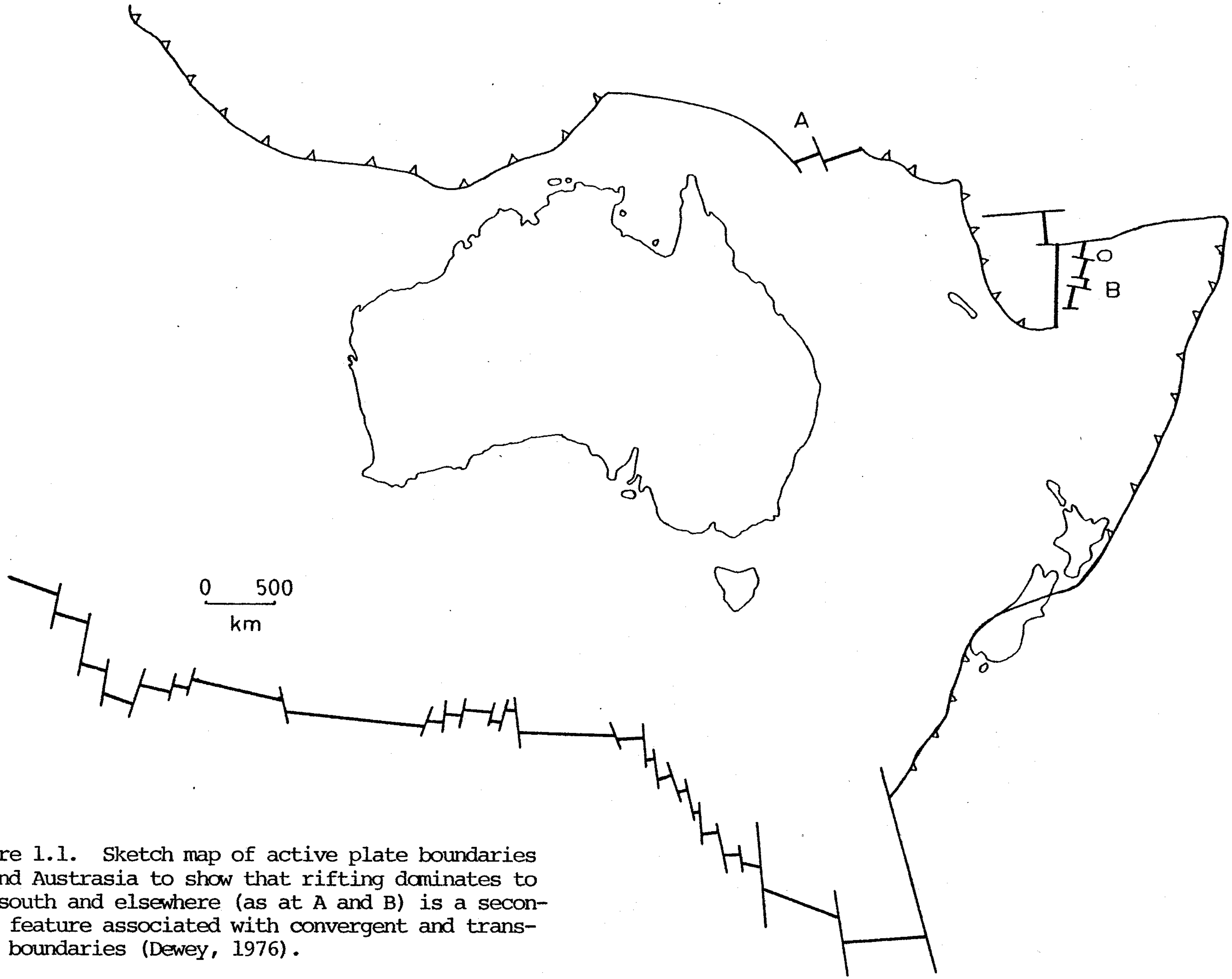


Figure 1.1. Sketch map of active plate boundaries around Austrasia to show that rifting dominates to the south and elsewhere (as at A and B) is a secondary feature associated with convergent and transform boundaries (Dewey, 1976).

and intermediate earthquakes along the trend of the Adelaide aulacogen (Figure 1.2). This concentration has been attributed to intra-plate stresses caused by differential spreading rates along the Antarctic Ridge (Fitch et al., 1973). Intra-plate stresses are apparently dissipated by faulting along prominent old structures in many continental areas. Old rifts, as structures that penetrate the entire lithosphere, are particularly vulnerable to this kind of reactivation. The extent of active rifting in the most seismically active area, "the Pirie-Torrens sunklands" of South Australia, is not great and is perhaps best indicated by the existence of a relatively small sub-sea level depression which is 12000 km^2 in area and ~10 m deep.

1.2.1.2. New Zealand

In New Zealand an active rift is occupied by the North Island volcanoes. This type of rift, seen elsewhere, parallels the length of the volcanic arc. It may mark a line of lithospheric weakness along which rupture and formation of a marginal basin could occur in the future (Karig, 1971).

1.2.2. MESOZOIC AND CENOZOIC RIFTS OF AUSTRALASIA

The formation of ocean-floor started at the northwestern margin of Australia in the Jurassic, at the western margin in the Cretaceous, on the east in the Late Cretaceous and on the south in the Eocene (Figure 1.3). Rifts that failed to develop into ocean were formed during and a few tens of millions of years before all these events. These are preserved around the coasts of Australia where they have been the subject of much oil and gas exploration and some production.

Figure 1.2. Sketch showing a concentration of intra-plate earthquakes in the general area of the Proterozoic Adelaide aulacogen. Associations of this kind of earthquake with old major structures are known in other continents (modified from Veevers and McElhinney, 1976).

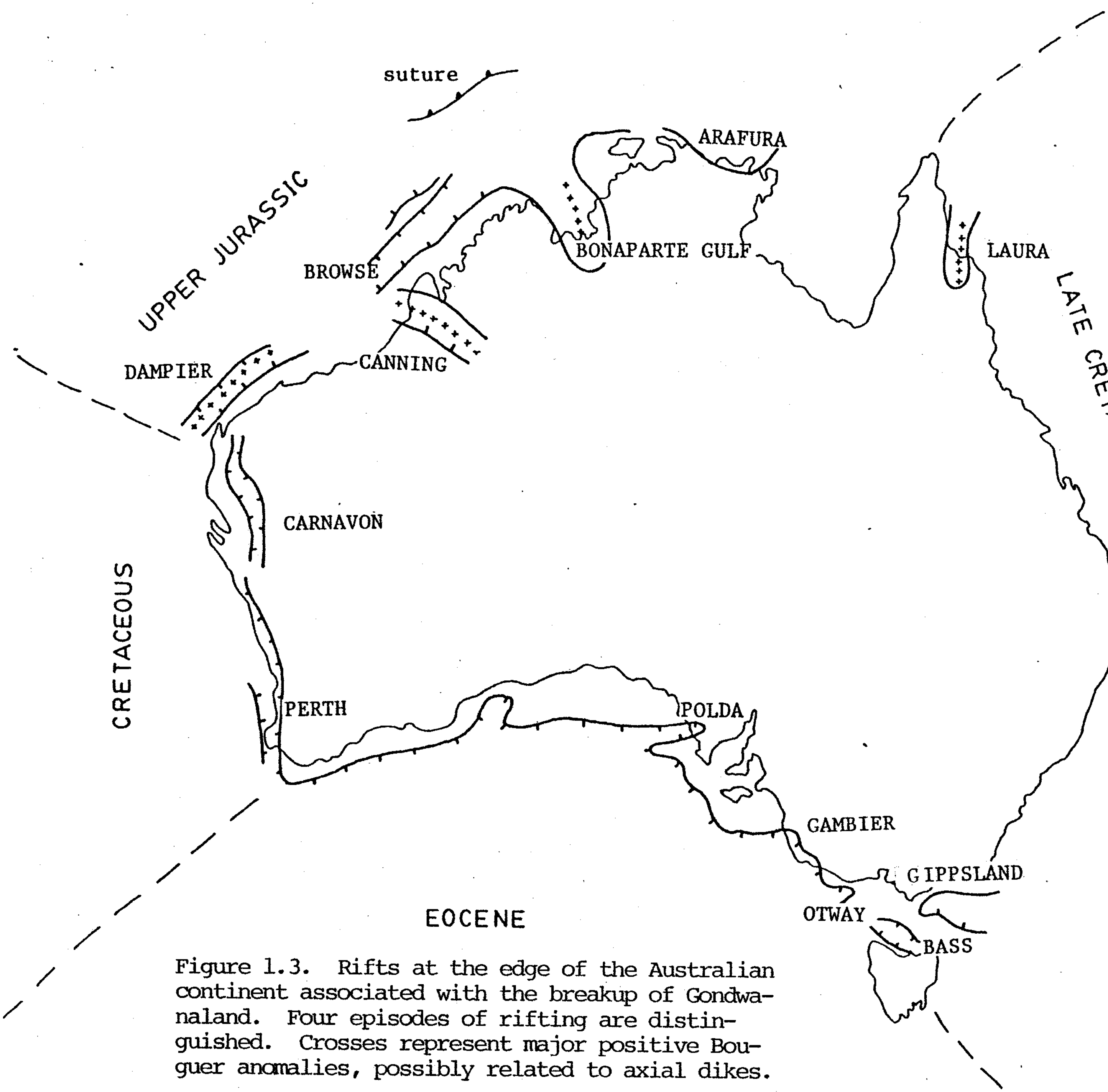


Figure 1.3. Rifts at the edge of the Australian continent associated with the breakup of Gondwanaland. Four episodes of rifting are distinguished. Crosses represent major positive Bouguer anomalies, possibly related to axial dikes.

1.2.2.1. Eocene Rifts at the South Coast of Australia

Thicknesses of up to a few hundred meters of Jurassic and Lower Cretaceous sediments are recorded at numerous places along the south coast of Australia, for example, in the Eucla Basin (Deighton, et al., 1976, Figure 1). However, these are shelf sequences and major rift development apparently took place in the Upper Cretaceous (perhaps 100 m.y. ago) when the thick sediments and volcanics of the Polda, Otway, and Gambier rift-basins were deposited. Talwani and Mutter (1977) found that there was an average separation of 50 km between the well-defined continental crust of southern Australia and the oldest magnetic anomaly (Anomaly 22) of the Southern Ocean. They concluded that the intervening zone was neither continental nor oceanic and embodied a "rift type" crust. Burke (1977b) reached a similar conclusion pointing out that mapping of failed rift-systems at Atlantic-type oceanic margins was a basic prerequisite to analysing Atlantic type continental margin development. Because the rifts in the intermediate zone are deeply buried there is, as yet, little evidence on the pattern of their distribution. Satellite derived geophysical data may prove helpful, if not in mapping the individual rifts, at least in characterizing zones at Atlantic-type margins in which deeply buried rift-valley lithosphere is developed. The northeastern seaboard of the U.S.A. is one such area and it would be revealing to compare satellite gravity and magnetic patterns across it with those across the southern coast of Australia.

After the formation of the first ocean-floor between Australia and Antarctica (at Anomaly 22 time, in the Early Eocene), the Australian margin appears to have developed in a normal way with clastic wedges prograding over and burying the marginal rifts (Veevers and McElhinny, 1976, Figure 5; Deighton et al., 1976). The absence of the commonly developed carbonate reef facies is perhaps to be ascribed to the high paleolatitude of South Australia (> 40° S) during and soon after rupture.

1.2.2.2. Late Cretaceous Rifts on the East Coast of Australia

An early analysis (Hayes and Ringis, 1973) showed magnetic anomalies in the Tasman Sea striking into the east coast of Australia thus indicating that the coast forms a convergent boundary. However, a recent reappraisal (Weissel and Hayes, 1977) has shown that the east coast was formed largely as a transform boundary and that it cuts off no anomalies.

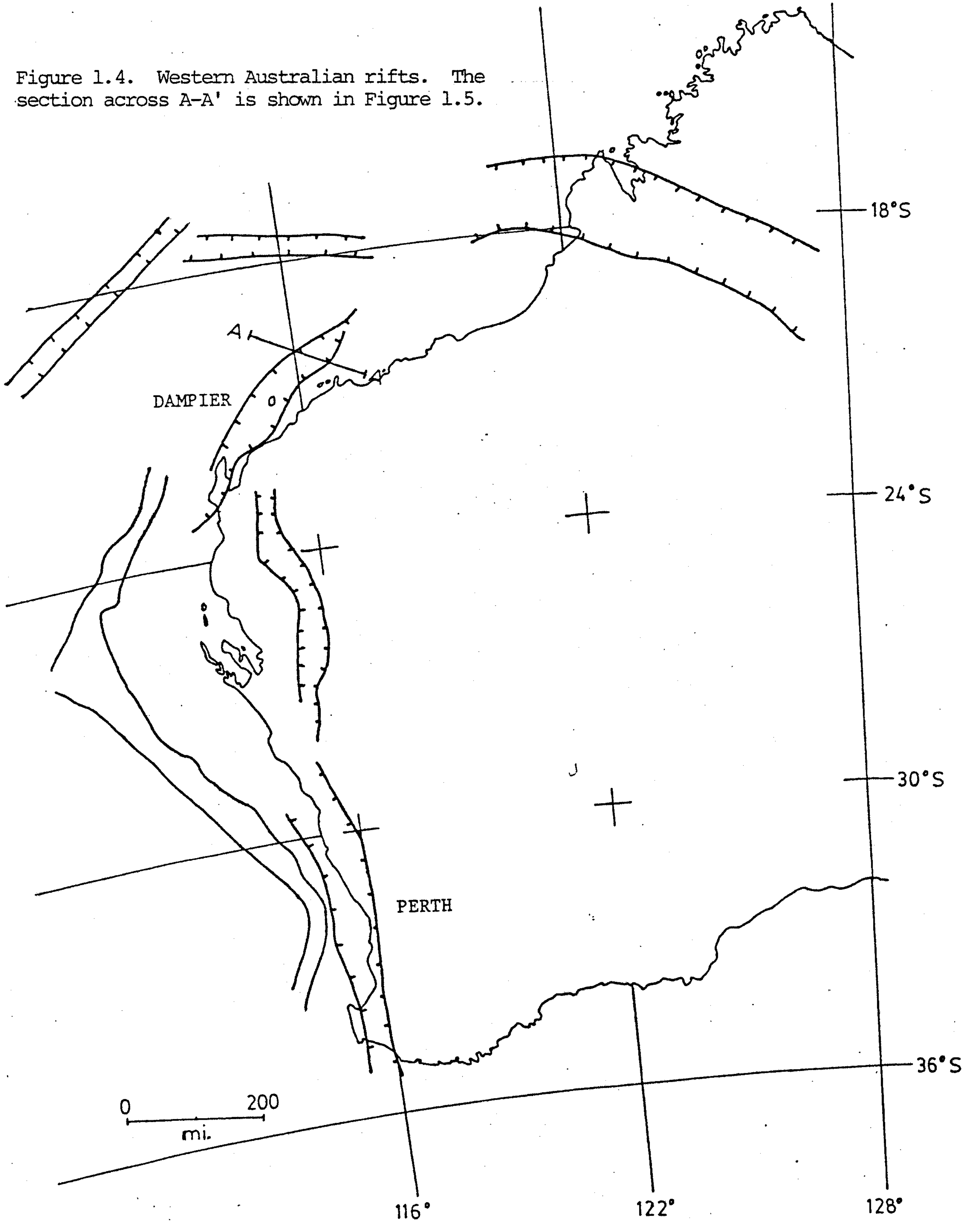
Only two well-developed failed rift systems, the Gippsland Basin in the south and the Laura Basin in the north, have been mapped in association with the opening of the Tasman and Coral Seas. The Gippsland Basin is separated from the coast of southern Australia by the Bass Strait. Opening of the Tasman Sea preceded that of the Antarctic Ocean south of Australia by approximately 30 m.y., but there is no clear stratigraphic evidence to indicate if the preceding rifting events took place at different times. The Bass Basin rift lying between Tasmania and Australia links the Gippsland rifts to those of the south coast of Australia. The Laura Basin (Figure 1.3) strikes at a high angle into the Coral Sea and Taylor (1975) has suggested a Late Cretaceous age for its rifting followed by Coral Sea spreading beginning in Paleocene or Eocene time.

1.2.2.3. Jurassic and Cretaceous Rifts of Western Australia

The rift pattern of Western Australia is shown in Figure 1.4. The oldest rifts indicated are in the Perth Basin system of Permian age. This was a time of major rifting in many parts of Gondwana although none of the rifts formed at that time underwent evolution to the oceanic condition. As is common, later rifting occurred in the same areas, and the continental margin of western Australia, formed in the Cretaceous, parallels the Permian Rift. Farther north Latest Triassic to Jurassic rifting is better developed. Figure 1.5 is a sketch cross-section across the Lewis trough rift in the Dampier system (for location, see Figure 1.4) illustrating how progradation of continental margin sediments followed Jurassic rifting. Two episodes of continental rupture, one during the Jurassic north of 23°S which formed the northwestern coast of Australia and a second during the Cretaceous south of 23°S which formed the west coast, took place in this area. The abundance of failed rifts on the continental shelf in the northwest has encouraged petroleum exploration and provided detailed stratigraphic information. The latter has permitted Veevers and Cotterill (1976) to make an ingenious comparison between the western margin of Australia and the East African Rift System both in horizontal scale and in relative timing of structural events.

The northernmost rifts of Jurassic age, the Bonaparte Gulf and Arafura basins, are of particular interest because to the north of them the continental margin of Australia, formed by rupture about 150 m.y. ago, has collided with the Indonesian Arc in Timor. Therefore, these are no longer failed rifts facing an ocean but aulacogens

Figure 1.4. Western Australian rifts. The section across A-A' is shown in Figure 1.5.



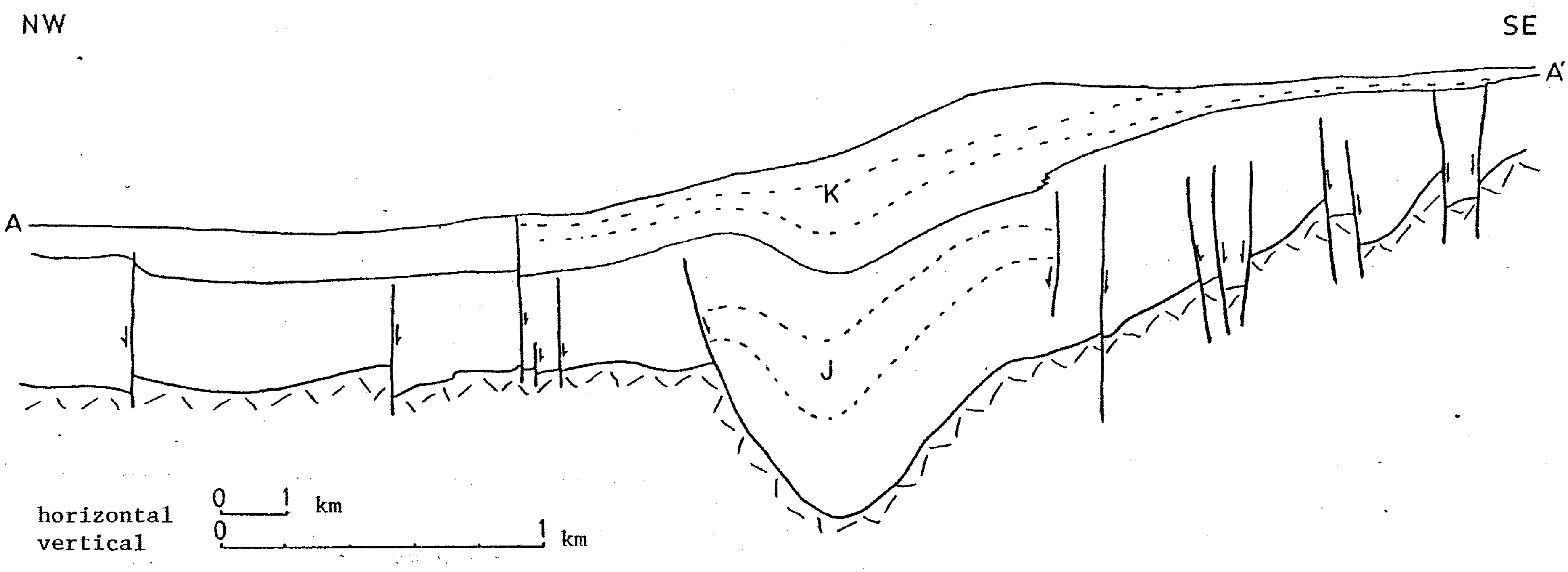


Figure 1.5. Section across the Dampier Rift (A-A' on Figure 1.4) of North-Western Australia to show how Cretaceous continental margin sediments (K) have prograded over Jurassic and Triassic filled rifts (J).

as defined by Shatsky (Burke, 1977a), that is rifts striking into fold belts.

The Bonaparte Gulf aulacogen is marked, as are many other rifts, by a prominent axial Bouguer anomaly (Gravity Map of Australia, 1976). These features are commonly attributed (Burke, 1977b) to the emplacement of axial basaltic dikes in the rift at the time of initial rifting.

1.2.3. OLDER PHANEROZOIC RIFTS

It has been suggested (Veevers and McElhinny, 1976, especially Figure 3) that a rifting event accompanied widespread eruption of flood basalt in Australia at the end of the Precambrian and that a continental object was removed at this time from the area now occupied by the Tasman orogenic belt (Figure 1.6). Evidence for this event seems less than compelling, and more detailed interpretation of rocks and structures dating from this time must be awaited. All identifiable rifts of this or earlier ages in Australia fall into one of two categories: failed rifts formed and preserved within continents, or rifts that have formed oceans and later been involved in arc or continental collisions. The second class are coincident with suture zones and for this reason they are not considered further in the discussion of rifts. Only the intra-continental or failed type are considered here.

1.2.4. YOUNGER PROTEROZOIC RIFTS

The Amadeus Basin has a long and complex history. It overlies the foreland trough of the Musgrave 1100 m.y. collision but its most important development took place during the interval 1000 to 800 m.y. ago when it appears to have formed a wide failed rift striking eastward

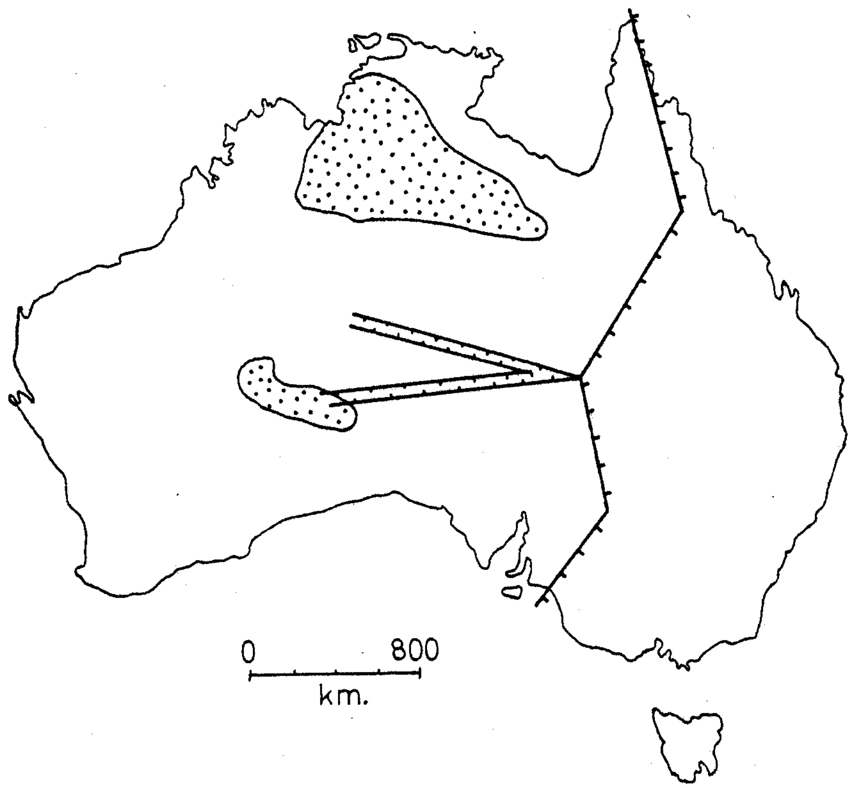


Figure 1.6. Late PreCambrian rifts of Australia. A continental object is suggested to have been removed to the east along the zig-zag boundary. Rifts penetrate the continent, and flood basalts (dotted areas) were widely erupted (based on Veevers and McElhinney, 1976, Figure 3B).

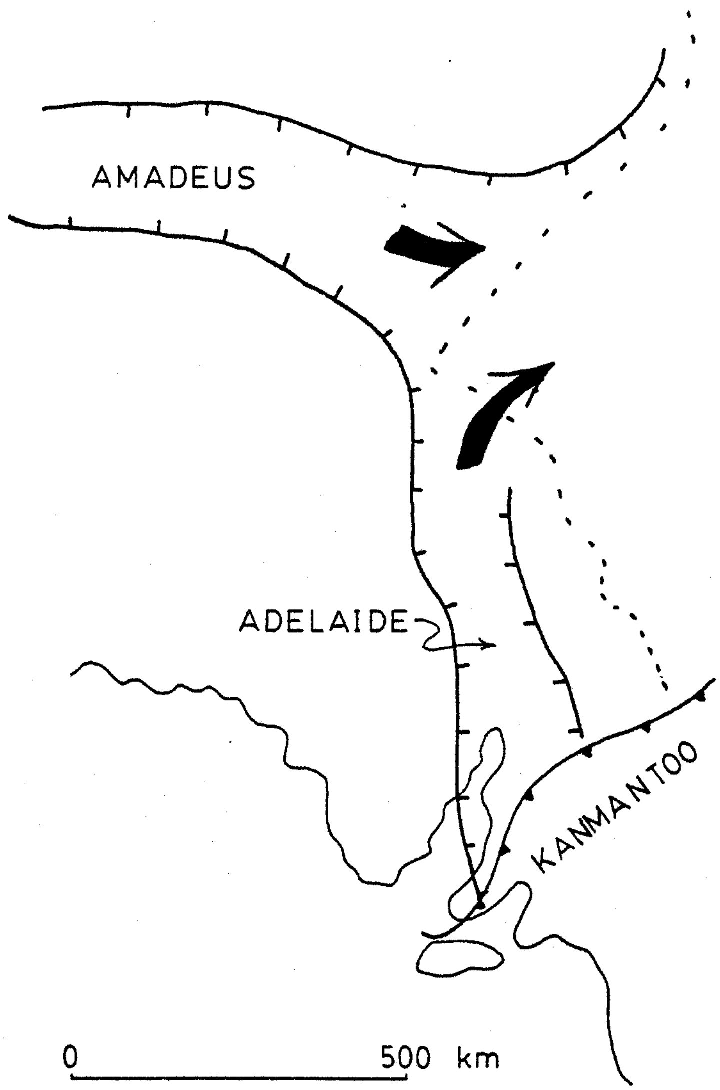


Figure 1.7. Younger Proterozoic rifts. The Amadeus and Adelaide Rifts opened into the ocean 1100 to 800 million years ago. In latest Proterozoic the Adelaide Rift opened to the south where it was involved in the Kanmantoo Arc collision which occurred in earliest Cambrian times. The dashed line marks the approximate continental margin.

into an ocean now occupied by rocks of the Tasman orogenic belt (Figure 1.7). At this time the site of the Adelaide 'geosyncline' (first identified as an aulacogen by Rutland, 1973) appears to have been occupied by a rift opening northward into the same ocean. The later sequences in the Adelaide structure are suggestive of opening to the south and a later rifting event (? 800 m.y. ago) appears to have been responsible for this change. The Kanmantoo folds in arc and Atlantic margin-type rocks show the typical curvature of a collision zone between an arc and an embayed continental margin.

However, the observation that the rocks of the Adelaide aulacogen can be traced into the deformed Kanmantoo belt, if correct, suggests that the arc actually developed at the open mouth of the failed arm and turned it into an aulacogen behind an Andean-type continental margin-orogen.

1.2.5. OLDER PROTEROZOIC RIFTS

Older Proterozoic Rifts have not been clearly discerned in Australia with the exception of the Batten trough, an aulacogen containing over 4 km of sediments and volcanics (Brown, Campbell and Cook, 1968, p. 13). The trough appears linked to the Pine Creek fold belt and Figure 1.8 shows a possible continental margin pattern about 1.7 b.y. ago with oceans on the sites of both the Pine Creek and the Mt. Isa orogenic belts.

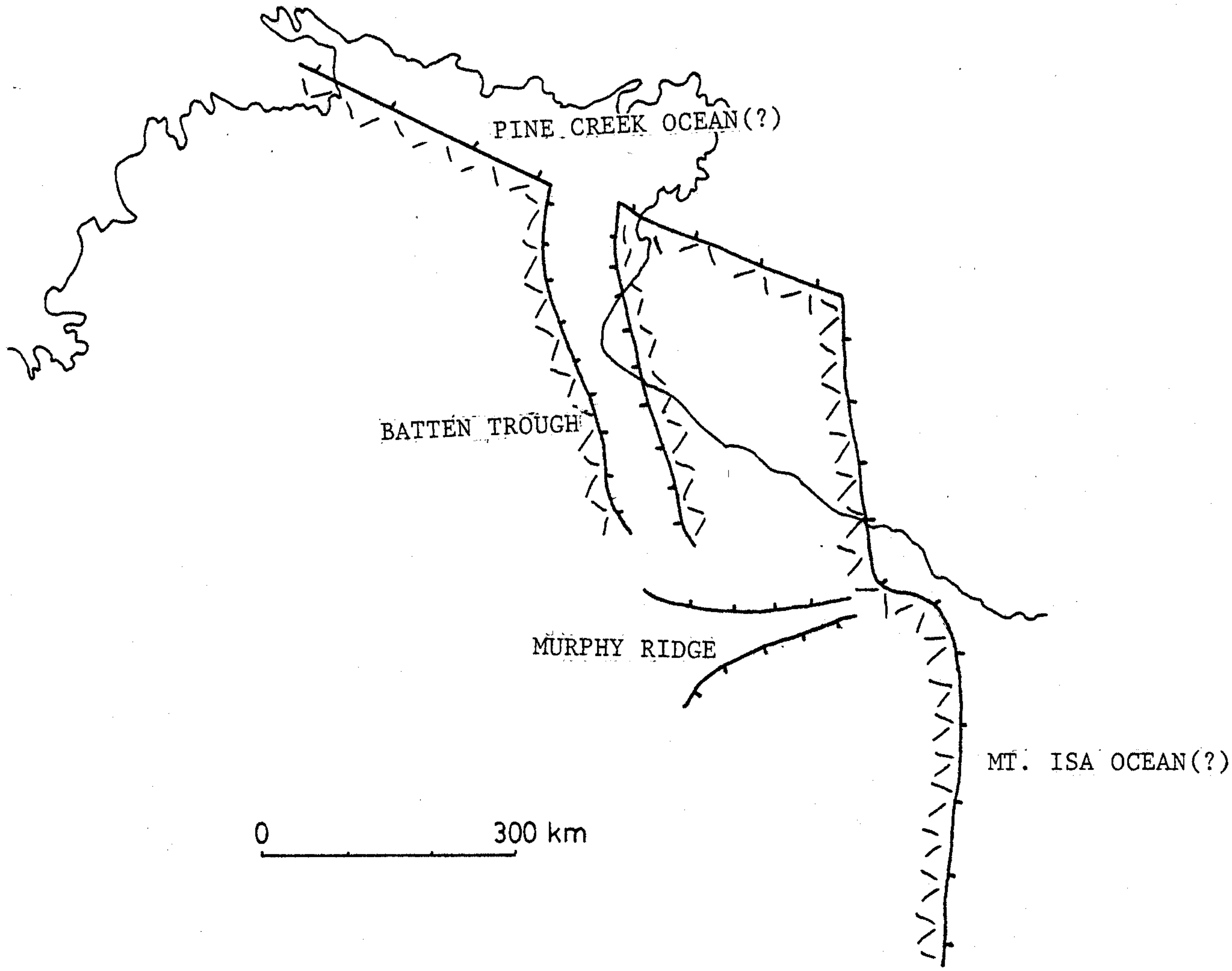


Figure 1.8. The Batten Trough, an older Proterozoic rift, forms an aulacogen as it strikes into the Pine Creek fold belt and former ocean. The Mt. Isa ocean may have been linked to the Pine Creek ocean as illustrated.

1.3. SUTURES

1.3.1. ACTIVE SUTURES

Active suturing in Australasia is confined to the Indonesian margin of the continent in Timor and New Guinea. The development of Indonesia is essentially part of Asian geological history rather than of Australian. A good summary of Indonesian tectonics appears in Hamilton (1977) and his maps (Hamilton, 1975a, b) are particularly useful.

1.3.2. MESOZOIC SUTURES

Mesozoic suturing is recorded nowhere on the Australian continent but is well developed in the two islands of New Zealand and in New Caledonia (Figure 1.9). These areas became isolated as a result of the rupturing of the Australasian continent to form the Tasman Sea about 80 m.y. ago, but during their convergent history they lay on or near the margin of the Gondwana continent.

In New Zealand the Mesozoic suture formed in the Rangitata orogeny between 100 and 200 m.y. ago has been largely obscured. In the South Island the suture has been cut by Cenozoic transform motion on the Alpine fault and in the North Island it has been buried beneath volcanics marking the southern end of the active Tonga Kermadec arc (Figure 1.9). Whether and how the New Zealand Mesozoic suture can be related to that of New Caledonia (Avias, 1973) is an outstanding problem of southwestern Pacific tectonic history. An initial approach to this problem is represented by Grant-Mackie, (1977).

Landis and Bishop (1972, especially Figure 6) suggested that the Torlesse facies rocks which form much of the eastern half of the South

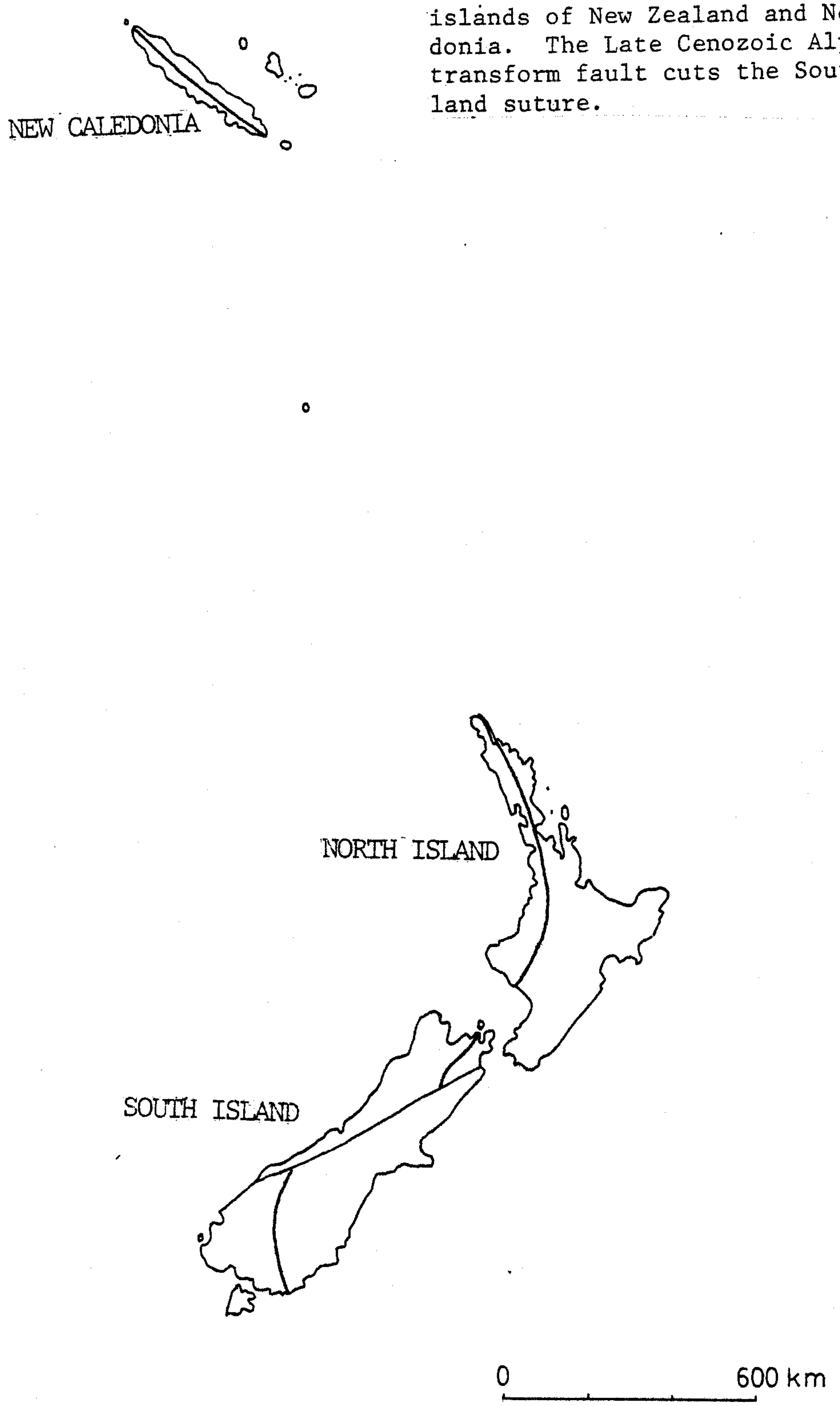


Figure 1.9. Mesozoic arc collisions are recorded in the sutures of both islands of New Zealand and New Caledonia. The Late Cenozoic Alpine transform fault cuts the South Island suture.

Island of New Zealand, were an accretionary melange. Therefore, although the Rangitata Orogeny marked a convergence, New Zealand had faced ocean to the east throughout the subduction episode. In that interpretation the ultramafic rocks that we suggest mark a suture are considered to be fragments in an accretionary melange. Figure 1.10 (based on a sketch in DeWit, 1977) shows that a substantial area of continental lithosphere lies oceanward of the Torlesse facies in the Campbell Plateau. It is for this reason that we interpreted the exposed relations in New Zealand as marking an arc or micro-continent collision during the Rangitata Orogeny rather than merely an accretionary prism.

1.3.3. PALEOZOIC SUTURES

1.3.3.1. Tasman Orogenic Belt

The Tasman belt occupies nearly half of Australia. It records a Paleozoic history dominated by arc and marginal basin phenomena. Incompleteness of our understanding of events in the Tasman results not only from the normal destructive processes that typify convergent margins but also from burial of large parts of the belt under the Great Artesian Basin of Mesozoic age.

1.3.3.2. Kanmantoo Arc

This feature was discussed in relation to the Adelaide aulacogen (1.2.4. above). Adelaidean sedimentary units interfinger with the folded Kanmantoo arc rocks in a foreland thrust-belt (Kleeman and White, 1956). The subduction zone active during collision appears to

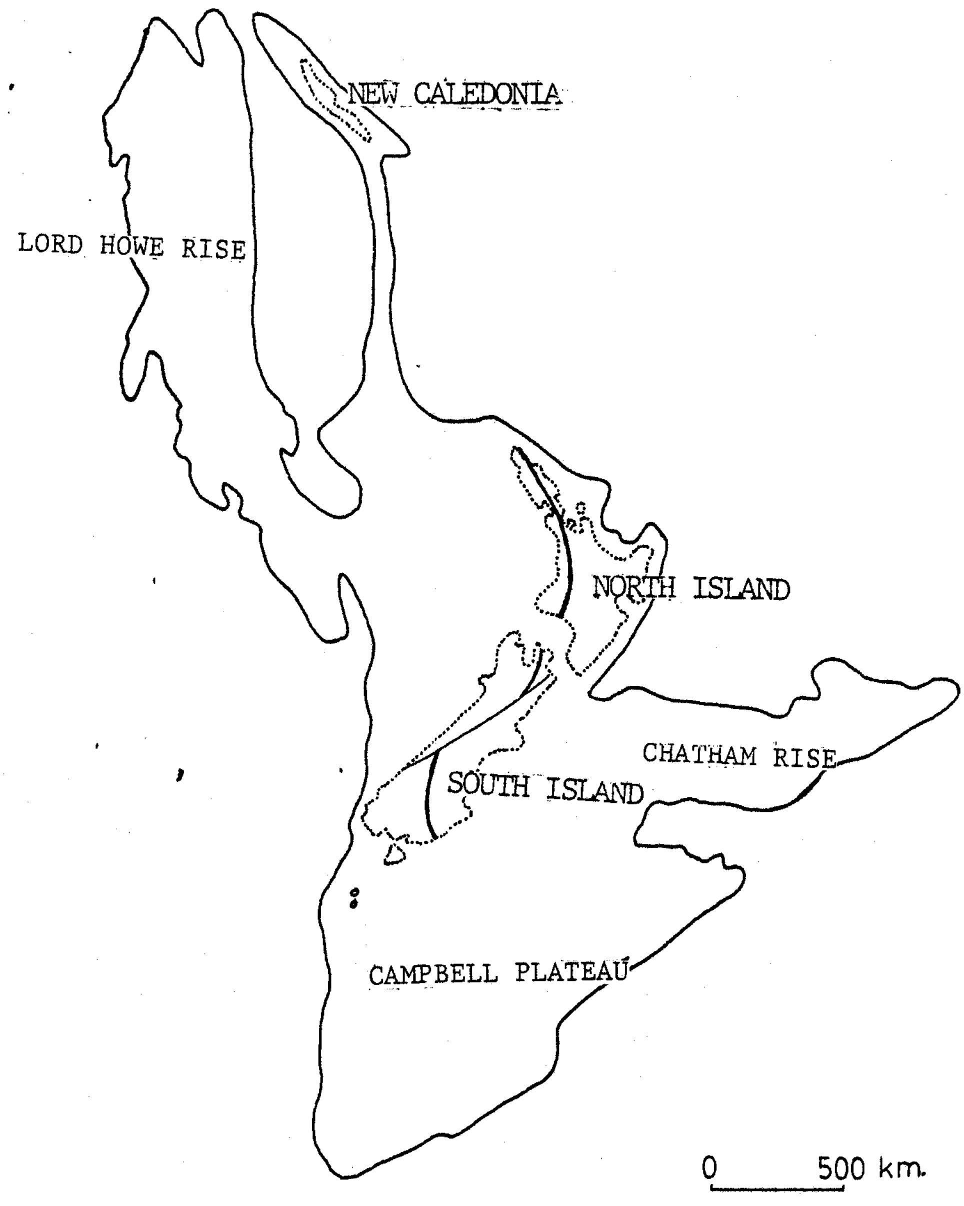


Figure 1.10. The Mesozoic suture of New Zealand and New Caledonia lies within a large area of continent. A continental object consisting of the Chatham Rise and the Campbell Plateau appears to have collided with New Zealand and the Lord Howe Rise which formed the eastern margin of Australia at the time of the collision.

have dipped toward the Adelaide aulacogen, if a part of the Kanmantoo arc is viewed as "exotic". An alternative interpretation is that the whole arc was an Andean-type orogen (see 1.2.4.) with a northwest dipping subduction zone beneath it (Burchfiel, pers. comm., 1977).

1.3.3.3. Paleozoic of Tasmania

Exact relations between the Tasman orogenic belt as exposed in Tasmania and that on the mainland of Australia have not been established. Northwestern and western Tasmania formed the Rocky Cape continental block in latest Precambrian and Cambrian times and were separated from the Tyennan continental block of Central Tasmania by the Dundas Trough underlain by ocean floor. Subduction of this material beneath the Tyennan nucleus led to formation of an island arc on its western flank (present coordinates) that was active until late Early Cambrian times.

Within the trough itself the stratigraphic succession is formed of quartzites-slates-dolomites and tuffs with a thickness of about 2500 m. Their age may be latest Adelaidean or earliest Cambrian (or possibly both). Some of the volcanics (keratophyre, rhyolite, pyroclastics with some welded tuffs) of the island arc on the Tyennan nucleus appear to have flowed into the easternmost sections of the Dundas Trough.

The trough probably closed during the late Early Cambrian, but the convergence across the suture continued until the Late Cambrian as shown by upward-coarsening flysch/molasse sequences. During the Late Cambrian marine deposition ceased entirely and fluviatile conditions prevailed over the Dundas Trough area (Brown et al., 1968, pp. 59-61). In the eastern part of Tasmania lower Paleozoic rocks are generally of platform type.

1.3.3.4. Lachlan Province

The Wagga Belt in the southeast of the Province is, according to Scheibner (1974), a Silurian island arc affected by Devonian collision. However, the entire Lachlan province may equally well be interpreted as a lowermost Paleozoic Andean-arc followed at the end of the Silurian by strike-slip tectonics. This sequence resulted in a N-S trending basin-and-range environment with bimodal volcanism and substantial extension and Devonian re-telescoping across the strike of the Basin-and-Range regime. In fact, the latter view is preferable, because many of Scheibner's (1974) "marginal basin floor ophiolites" may be nothing more than the basaltic component of the bimodal volcanism. Also, the extreme width of the "quartzose flysch" independently suggests the existence of N-S strike-slip faulting.

The Wagga Belt presents considerable difficulty for the Early Paleozoic paleogeography. During the Ordovician an island arc was present in the southern part of the Tasman Orogen (Figure 1.11).

During the late Silurian several basins opened in what is now the Wagga Belt and were floored by attenuated continental crust. This was apparently a time of major strike-slip activity that possibly "emplaced" the Wagga Belt where it is now. The small basins are characterized by bimodal volcanism. One or two of them may have opened enough to have ophiolite floors.

In the Early Devonian another island (or Andean) arc formed to the west of Wagga Belt (Figure 1.12). This indicates the recommencement of subduction activity and the Basin and Range type basin to the

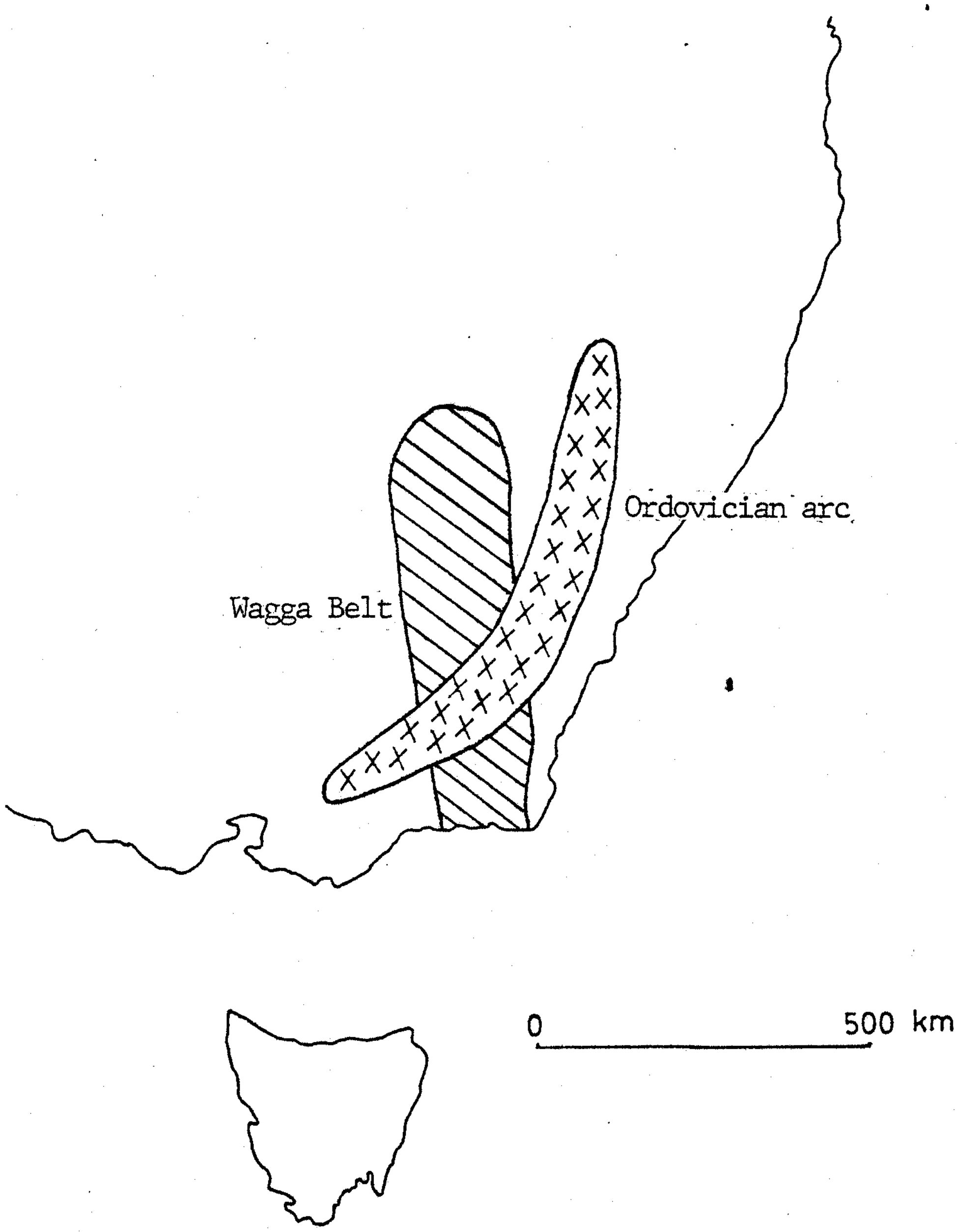


Figure 1.11. Relationship of the present Wagga Belt to the Ordovician arc. Late Silurian strike slip motion dismembered the arc to produce the present relations.

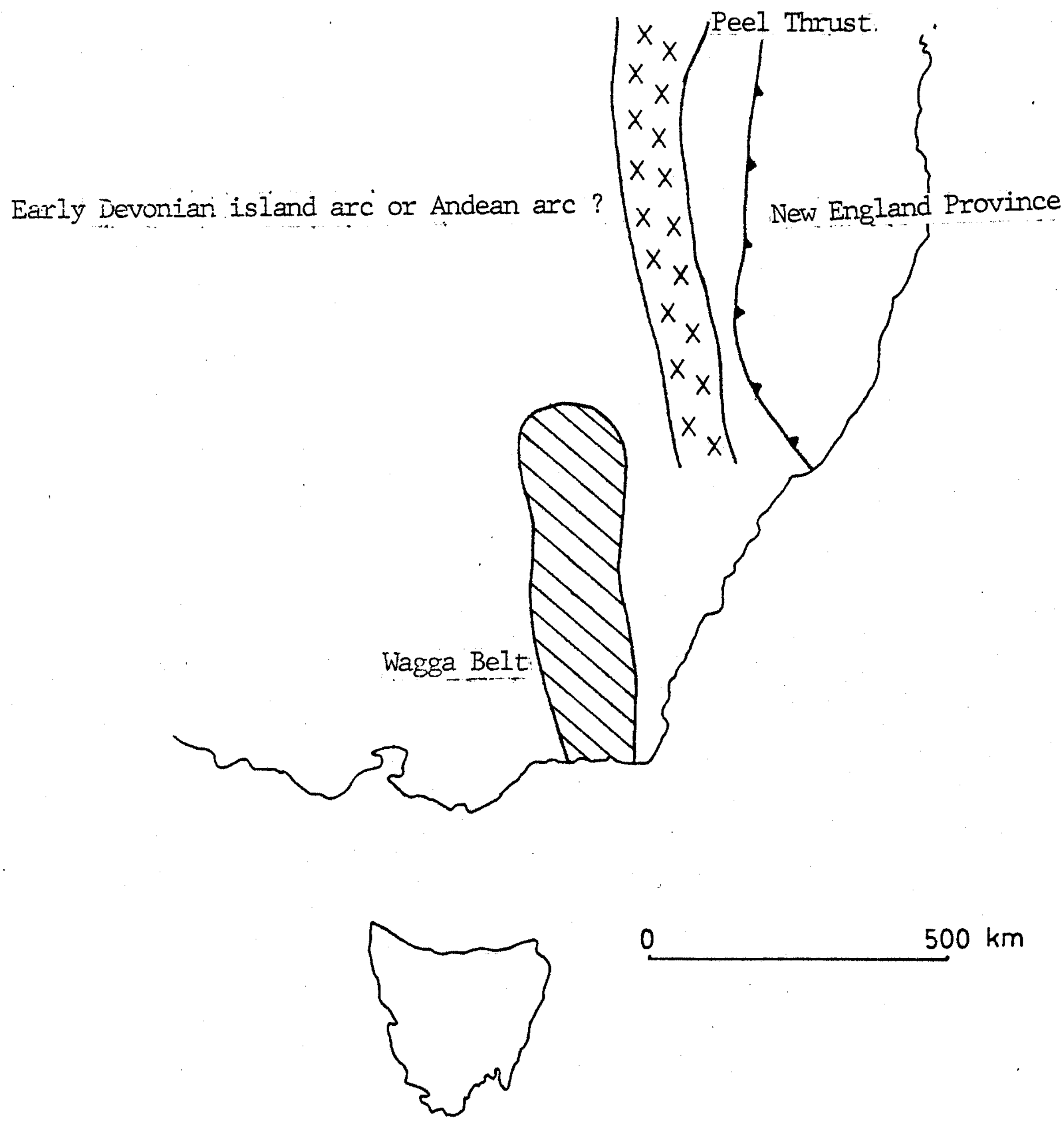


Figure 1.12. Renewed Early Devonian Arc activity to the east closed the Basin and Range Province of the Wagga Belt.

west began to close. The closing, however, was irregular. In the Wagga Belt the basins in the middle shut first, then those to the west and last those to the east. This is what Rutland (1976) refers to as "folding and plutonism spread outwards in the Wagga Belt".

In summary, the Lachlan Belt evolved from Cambrian up to the end of the mid-Devonian as an Andean margin. It was characterized, peculiarly, by upright almost chevron folds with axial planes striking north-south, the across-strike shortening being interrupted only during the Late Silurian basin-and-range event. The "main plutonism" of 400 m.y. ago reported by Rutland (1976) may reflect this episode.

Alternatively, after the Late Silurian Benambran Orogeny a few true marginal basins may have opened in the area and closed by the Tabberabberan orogeny during the Devonian.

1.3.3.5. New England Province

The New England Province represents an island arc with a well-developed fore-arc basin (Roberts and Oversby, 1973). The marginal basin which separated the arc from the main Lachlan province was probably a basin that opened after the "terminal" folding of the latter during the Tabberabberan orogeny. It closed during Early Permian times by westward subduction under the Lachlan and by eastward subduction under the New England Arc. This marginal basin is today represented by the ophiolites of the Peel Thrust. From Early Permian times on the Tasman became an Andean type arc and soon became inactive.

1.3.4. PROTEROZOIC SUTURES

After the consolidation as arcs and by arc and microcontinental collision of the Archean areas now represented by the Kimberley, Pilbara and Yilgarn blocks, oceans formed whose sites now lie to the north, south and east of these blocks.

1.3.4.1. King Leopold - Halls Creek - Pine Creek Belts

The oldest Proterozoic mountain belt in Australia is that of King Leopold-Halls Creek-Pine Creek orogenic belt (Figure 1.13). This belt was deformed about 1970 m.y. ago. It contains abundant ultramafics, tholeiites and spilites (Geology of Australia, 1/2,500,000, 1976). The orogenic belt may have continued along strike into the later Mt. Isa "geosyncline" as there are plutonic rocks of ~1800 m.y. age that appear in that region (Rutland, 1976). Other than this possibility, it is not possible to trace the orogenic belt any farther.

We interpret the distinctive angular S-shape of this belt (Figure 1.13) as revealing the angular shape of initial continental rupture preserved through collision. The Halls Creek sector has the most severely deformed looking outcrop patterns and may have been involved in more transform motion than the King Leopold and Pine Creek sections.

1.3.4.2. Gascoyne Belt

Another orogenic belt of Proterozoic age (1700 m.y.) is the Gascoyne Orogen, situated between the Pilbara and Yilgarn Archaean nuclei (Figure 1.13). About 1700 m.y. ago a major convergent event deformed the sediments of the Gascoyne "trough". As there are no ultramafics exposed in the orogen (however, Smith and Horwitz (1975)

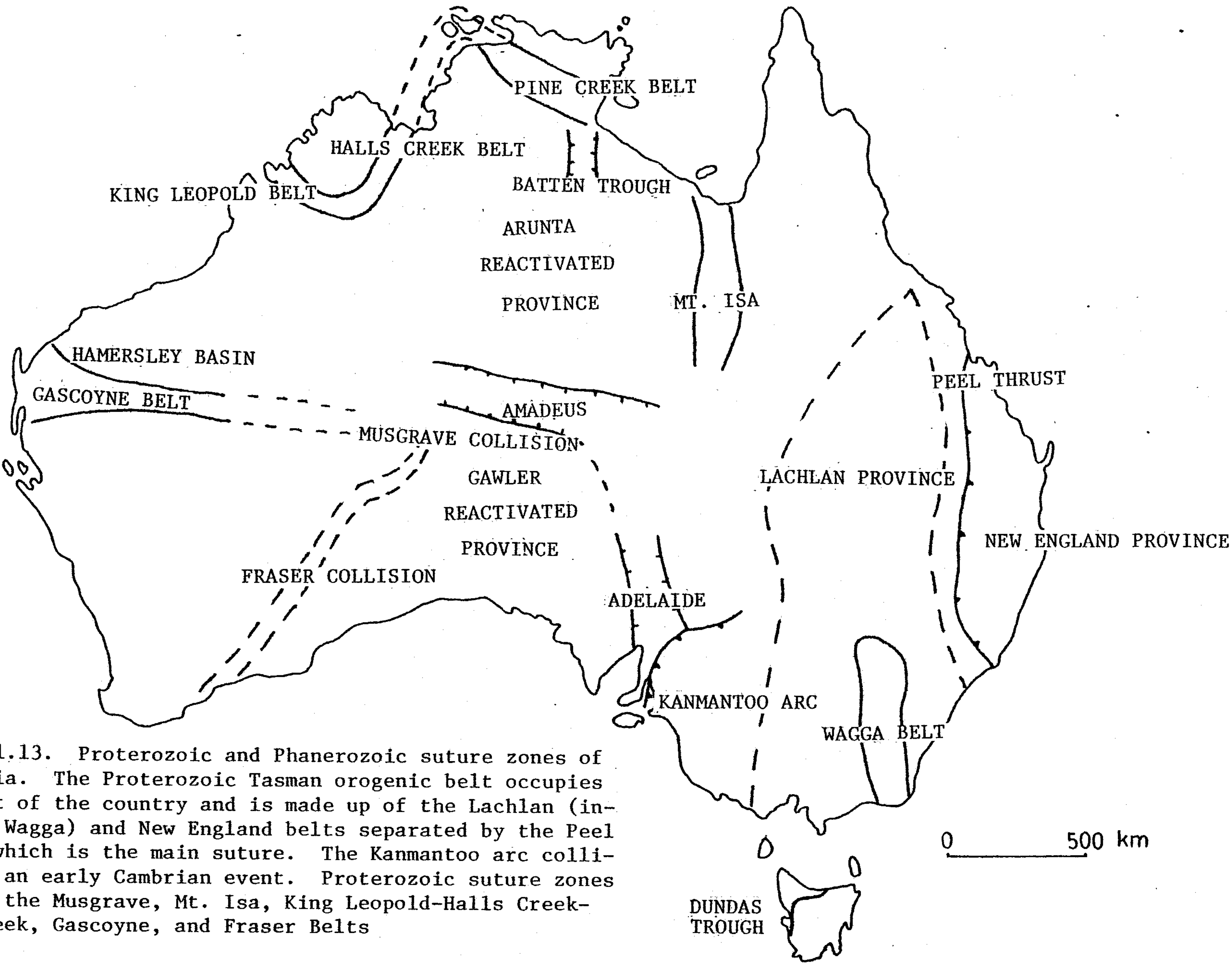


Figure 1.13. Proterozoic and Phanerozoic suture zones of Australia. The Proterozoic Tasman orogenic belt occupies the east of the country and is made up of the Lachlan (including Wagga) and New England belts separated by the Peel Thrust which is the main suture. The Kanmantoo arc collision is an early Cambrian event. Proterozoic suture zones include the Musgrave, Mt. Isa, King Leopold-Halls Creek-Pine Creek, Gascoyne, and Fraser Belts

presented some geophysical evidence to show that the Wyloo Group had an oceanic basement), the exact nature of the Gascoyne "trough" cannot be ascertained. It had miogeoclines on both its sides (as did the Alps) and underwent a major compressional deformation with basement thrusts and flysch deposition (clastic sections of the Wyloo Group). The observation that the two miogeoclines are preserved as such (with tectonic disturbance) may indicate that the trough was not very large. Syn-and post-deformational granites may be anatectic in origin and resemble the Bergell Complex of the Alps.

1.3.4.3. Arunta and Gawler Reactivated Provinces

The Arunta and Gawler Provinces (Figure 1.13), characterized by a major magmatic and tectonic event of about 1850-1700 m.y. (Rutland, 1976), represent a terrain of basement reactivation (Dewey and Burke, 1973). Although the Archaean basement has not been identified within the Gawler Province by isotopic means, it is definitely present, locally, within the Arunta Province. These two provinces today are separated by the Musgrave Orogen, and a definite statement about their original unity cannot be made. In fact, the observation that the two provinces have differing sedimentary covers may indicate that they were first united by the Musgrave collision event 1100 m.y. ago. However, they have features in common that may suggest the opposite. The isotopic ages decrease towards the east in both belts and both have very similar deformational histories. At the present level of understanding, this question cannot be easily resolved.

The Gawler and Arunta Provinces may have reworked Archaean as well as newly accreted material in their basement. The dominant tectonic style of largely upright, open to closed folds supports the view that the <u>last</u> event in the Province was basement reactivation. It would be foolish with present information to attempt to unravel the pre-reactivation processes. No actual sutures have so far been detected, but they probably lie between the Yilgarn and Gawler Blocks (obliterated by the later Fraser event) and the Arunta and Kimberley-Pilbara Blocks. Continental collision is presumed from the evidence of Arunta and Gawler reactivations. If the Gawler Block was added to Australia first during the Frasier-Musgrave events, then this suture may now be entirely outside the Australian continent, as is the case for a large portion of the Grenville suture and North America.

1.3.4.4. Fraser-Musgrave Suture(s)

The Arunta-Gawler region of the Australian continent assumed its present geometry after the Fraser and Musgrave events. The Musgrave Belt (Figure 1.13) sits along strike with the older Gascoyne Orogen and contains a good and representative ophiolite suite, indicating that an ocean did indeed close here. The ophiolitic rocks have tectonic contacts with the surrounding rock (Rutland, 1977, Figure 6). Numerous granitic plutons and metamorphics associated with this event give ages around 1100 m.y. (Rutland, 1976).

Whether the NNE-striking Fraser Belt, perhaps of somewhat younger age (1050), has a direct connection with the Musgrave Belt is uncertain as far as is presently known. It is entirely possible that the Fraser and Musgrave Belts represent a single paleo-plate boundary that was responsible for the final welding of the Gawler Block to the Australian continent.

1.3.4.5. Mount Isa

A persuasive "open and shut case" is represented by the Mt. Isa "geosyncline" (see Brown et al., 1968). The tholeiitic Marraba and Eastern Creek volcanics in the shallow water sedimentary sequence indicate an opening phase for the Mt. Isa "trough" (Figure 1.13). The existence of the "ophiolitic" assemblage (Geology of Australia, 1976, 1/2,5000,000) indicates that the extension went far enough to produce oceanic crust. This took place about 1700 m.y. ago (Wilson, et al., 1972; Derrick and Wilson, 1975; Rutland, 1976). Dunnet (1976) has interpreted the structure as an aulacogen as it disappears southwards. However, the evidence of multiple deformation with flow folding and massive thrusting indicate a considerable amount of shortening with the western limb overriding the eastern limb. The present expression of this is the "shear zone" of Blanchard and Hall (1942) with a westerly dip of ~70-80°. The extensive pegmatites in the southwest of the Mt. Isa "geocyncline" may be indicative of actual crustal melting to produce magmas. Altogether, Mt. Isa is too tectonized to be an aulacogen. The limestones and dolomites here may be the remnants of an old miogeocline. The closure of the Mt. Isa ocean, probably connected to the Musgrave ocean, took place about 1550 m.y. b.p. (Rutland, 1973, p. 1018).

1.3.4.6. Western Australia

By 1100 to 800 m.y. ago the central and western sections of the Australian continent had assumed a geometry very similar to that of the present. About 900 (?) m.y. ago a piece of continent was rifted away from the central section and this rifting event left two failed arms as witnesses. In the Amadeus and northern Adelaide thick graben facies lithologies were deposited. Later another rifting event, about 800 m.y. ago, affected the southern end of the Adelaide failed arm (similar to the situation in the Oslo Rift).

1.3.5. ARCHEAN SUTURING

There are three Archaean nuclei in Australia (Figure 1.14), Yilgarn, Pilbara and Kimberley domains. Only the Yilgarn and Pilbara are exposed to field examination, because the Kimberley is covered by partly Nullaginian (Early Proterozoic) and largely Carpenterian and Adelaidean (Later Proterozoic) sequences (Brown et al., 1968, Table 1.5). Therefore, no direct evidence for the existence of a Kimberley "nucleus" exists. However, the Carpenterian deposition was of platform-type on top of the Kimberley Block, in contrast to the deep-water facies in the Halls Creek Belt (1.3.4.1.) and a similarly shallow-water deposition may have prevailed on the block during the Nulliganian deposition (Brown et al., Figure 1.11). Basic and intermediate volcanism and granite intrusion may have occurred during or before Nulliganian times. There was, however, a cratonic block at the site of the Kimberley block before the evolution of King Leopold and Halls Creek orogenic belts (1970 m.y.). This therefore puts its formation into the Archaean. Nothing can be said about the tectonic style of the Kimberley due to the later cover. The exposed Archaean nuclei, the Pilbara and Yilgarn blocks, are both larger than the Kimberley block.

The Pilbara nucleus is situated between the King Leopold orogenic belt and Canning Basin in the northeast and the Gascoyne orogenic belt in the south. Within it are greenstone belts and

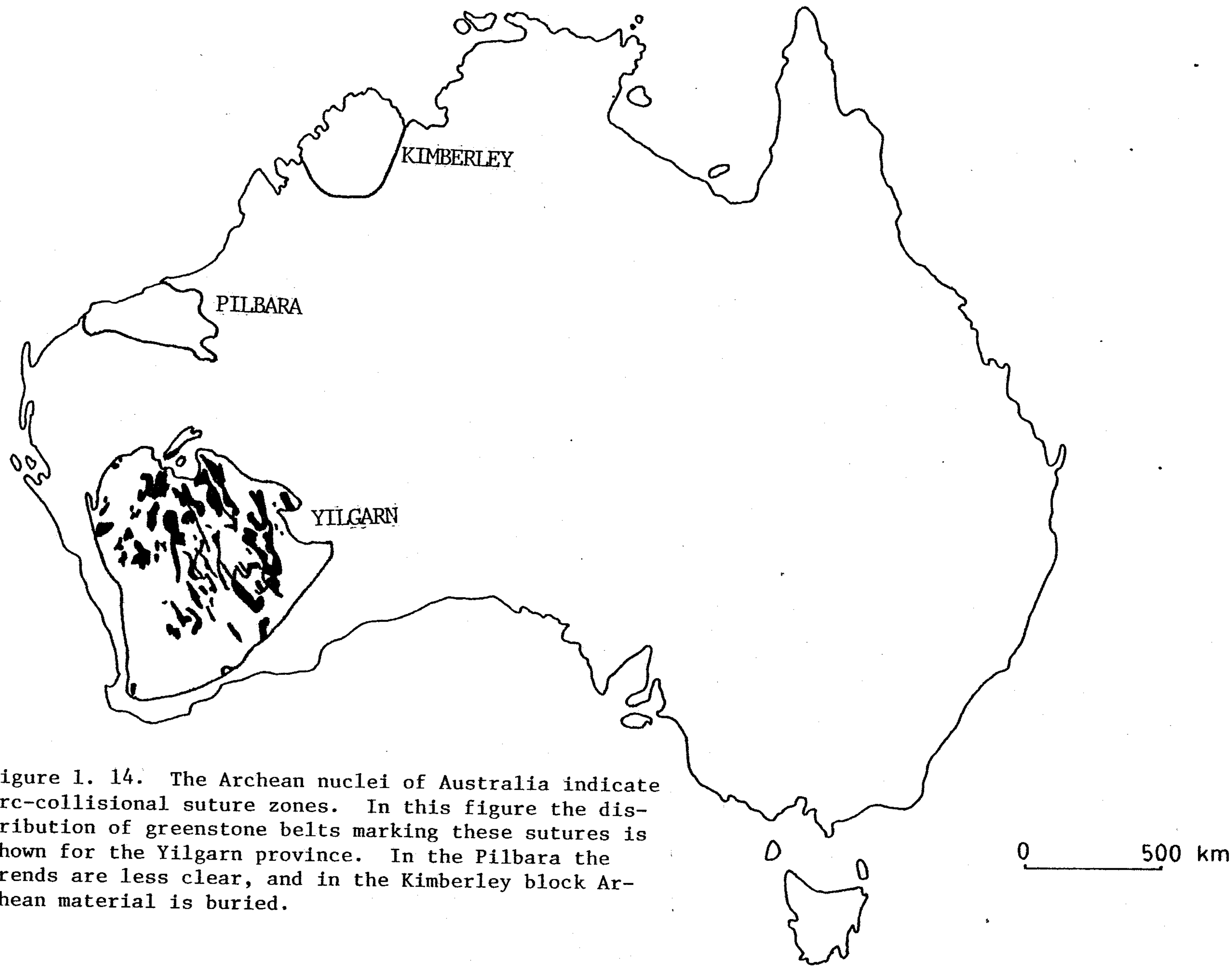


Figure 1. 14. The Archean nuclei of Australia indicate arc-collisional suture zones. In this figure the distribution of greenstone belts marking these sutures is shown for the Yilgarn province. In the Pilbara the trends are less clear, and in the Kimberley block Archean material is buried.

granites of 3000 m.y. age. These greenstone belts are strongly curved and lack the linear outlines of the later Yilgarn greenstone belts (Figure 1.14). There are chert sequences in the Pilbara and these have been used to correlate different sub-domains within the block. However, in the light of present understanding of Archaean greenstone-belt development (e.g. Burke et al., 1976; Tarney et al., 1976) it is not possible to assume that these sequences represent a single event and can thus be correlated. It is more likely that they represent repetitive eruptions in island-arc environments.

A later granitic event, with Sr^{87}/Sr^{86} ratios of ca. 0.73 (DeLaeter et al., 1975) occurred in the whole of the Pilbara region (~2629 ± 17 m.y., Roddick et al., 1976) and represents a basement reactivation, which, in turn, may imply continental collision with some other Archaean microcontinent. What that microcontinent might have been is, of course, an open question.

The Hamersley Basin, south of the Pilbara Block and resting presumably on the latter, is apparently a miogeocline representing a rifting event (on the site of the Gascoyne belt) between ~1970 m.y. ago and 1700 m.y. ago. The sequences here (Hamersley and Fortescue groups) are about 4000 m thick and contain basaltic volcanics apparently of rifting type.

The Yilgarn Group (or block) is represented by the linear greenstone belts of younger age than the Pilbara (~2689 ± 17 m.y., Rutland, 1976) and may indicate that by this time more sizable continental objects were produced than those represented in the Pilbara Block. The Yilgarn, in the north, has a miogeocline similar to the

Hamersley Basin in the Naberu basins. This may be correlative with the Hamersley but does not necessarily indicate that the Yilgarn and Pilbara rifted from each other.

Both Yilgarn and Pilbara blocks contain greenstone belts and intervening granodiorite terrains that we have elsewhere interpreted (Burke, Dewey and Kidd, 1976) as formed by processes similar to those acting today. The huge dikes of the Yilgarn (McCall and Peers, 1971) are of special interest because they indicate, like the Great Dike of Rhodesia, an early history of continental rupture.

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2.0 AFRICA

2.1. INTRODUCTION

Since the breakup of Gondwana began 200 m.y. ago, Africa has become the world's most active rift area. By contrast older rifts are as yet poorly defined in the continent. With respect to sutures the situation is reversed. The extensive Precambrian outcrops of Africa which result from its Neogene elevation reveal some of the best exposed Precambrian sutures of the world. Continental collisions in Late Proterozoic and Early Paleozoic times produced huge areas of reactivated terrain in the Pan-African realms. Satellite-determined geophysical fields may be controlled both by the geologically young rifting episodes and by the much older continental collision events.

2.2. RIFTS

2.2.1. Active Rifts

The African plate is the site of the best known and most widespread active intra-continental rifts. The reason for this has been widely discussed, with increasing recognition that the African plate has been at rest with respect to the underlying mantle convective pattern for the last 25 m.y. The magmarsis hypothesis of Krenkel (1927), which states that the rift and basin and swell structure of Africa is due to underlying convective circulation, is being revived in modern form.

The East African rift system (Figure 2.1) contains the majority of active rifts some of which are reactivated from Mesozoic times. Elsewhere in the continent Neogene tectonism has also reactivated older rifts.

2.2.1.1. Active Rifts in East Africa

One of the most striking features of this rift system is the elaborate areal pattern of the active rifts (Figures 2.1 and 2.2). If the African continent splits into two in the future by the formation of new ocean along a linked set of rifts, many rifts will be left as failed arms in one or the other of the two new continents.

The northern and southern Gregory rifts in Kenya form a three-rift system with the Kavirondo rift that is centered on the town of Nakuru and is at an early evolutionary stage. Studies of the elevation of erosion surfaces on the rift shoulders around the Nakuru rift junction have shown that the topographic dome centered on Nakuru at the crest of which the rifts meet developed before the rifts themselves (Baker and Wohlenberg, 1971). This is of interest because it has been suggested

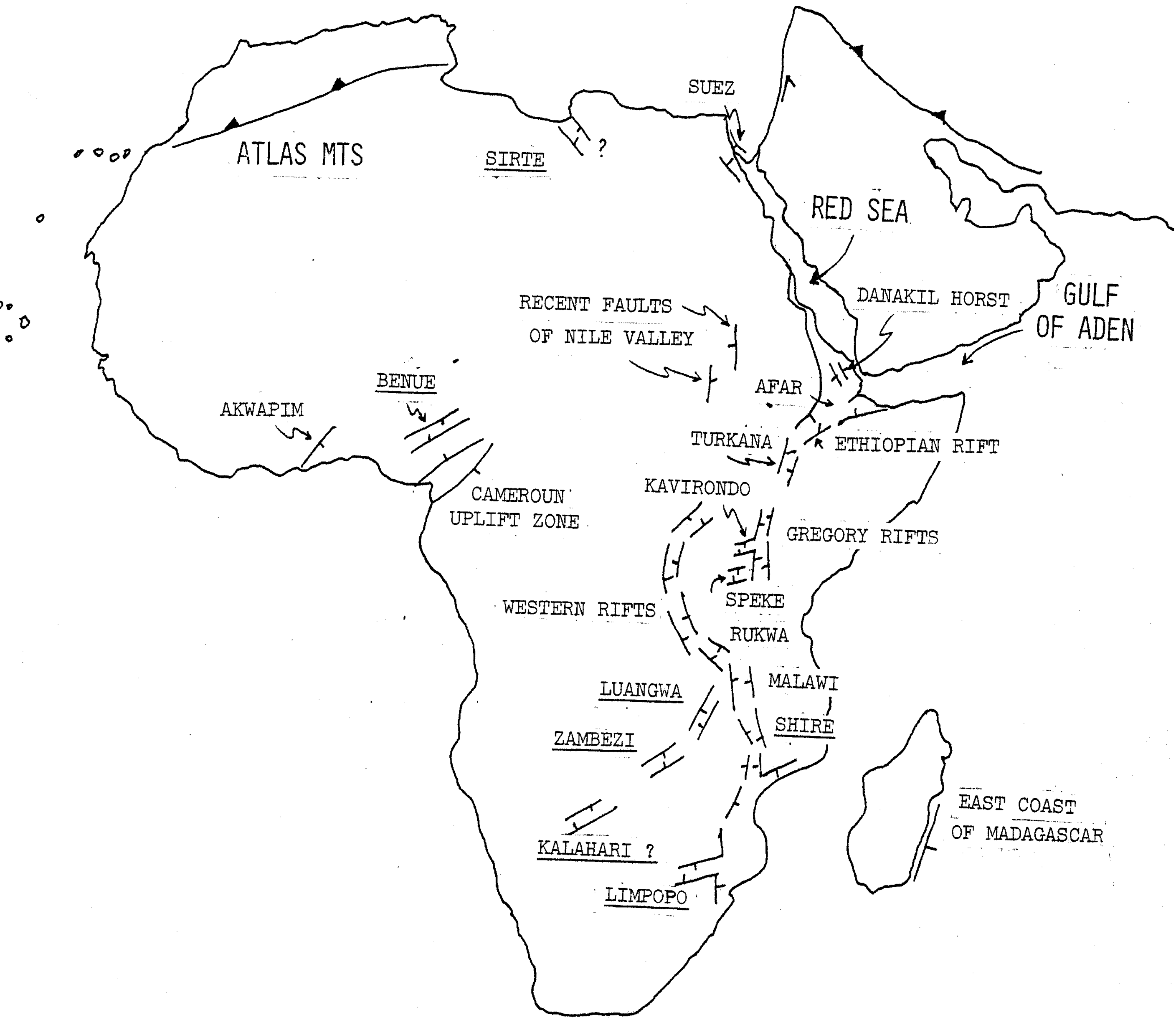

Figure 2.1. Major active rifts of Africa and selected active fault zones. Compressional tectonics are restricted to the Atlas Mountains in Africa. Over the rest of the plate a basin and swell structure with associated active uplift, faults and rifting has developed over the last 25 million years. The most prominent rifts are in the East African Rift System. Reactivated Mesozoic rifts are underlined on this sketch map.

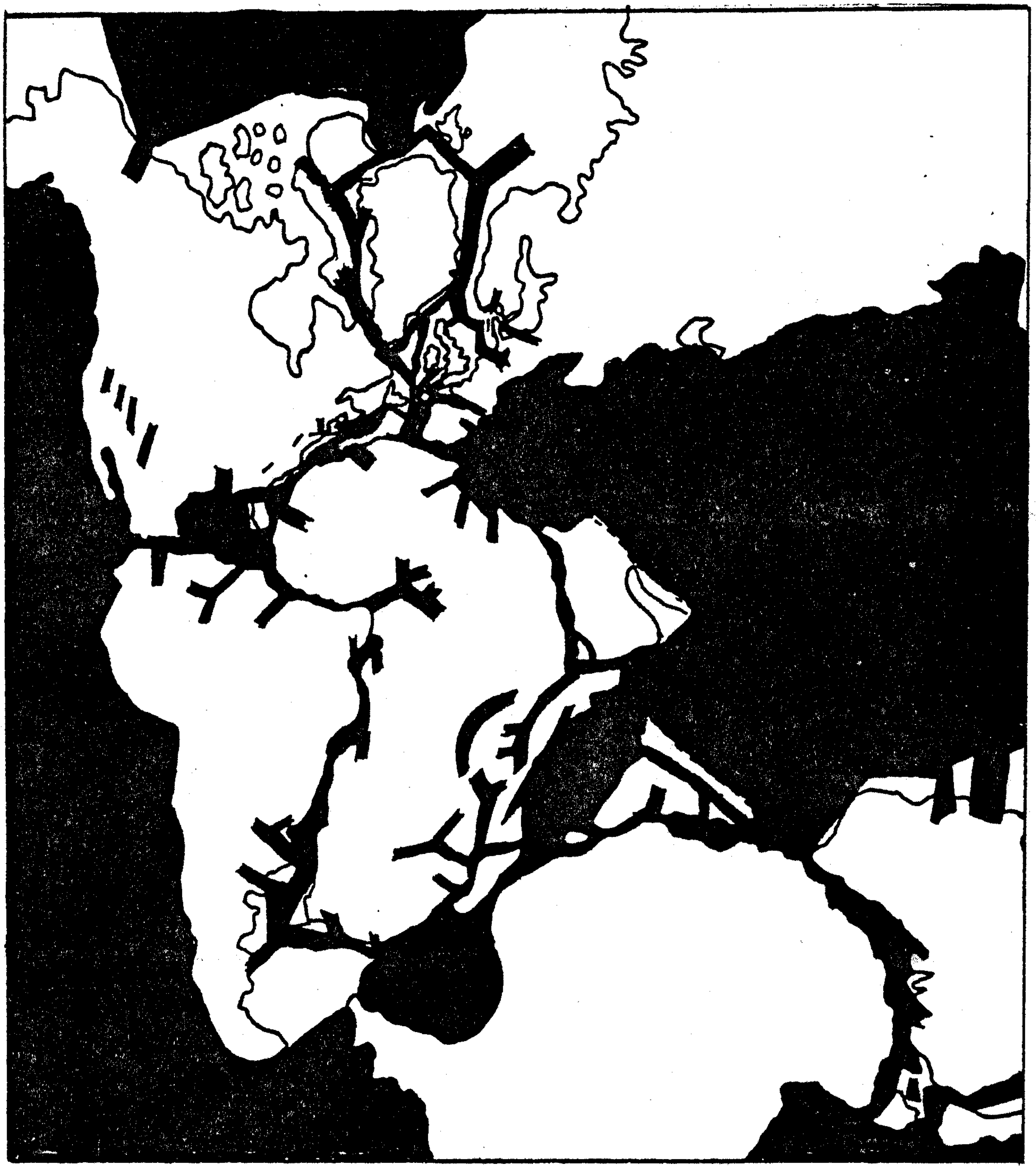

Figure 2.2. Mesozoic and Cenozoic rift systems and oceans are shown in black on a Triassic earth (based on Smith, Briden and Drewry, 1972) to illustrate the abundance of failed rift systems associated with continental breakup. In Africa the rift systems have been active twice, in Mesozoic and Cenozoic times.

that intra-continental rifts commonly develop by horizontal propagation of vertical lithospheric cracks with uplift following later as a response to the emplacement of hot rock in the crack. While this process certainly happens (Dewey and Burke, 1974), present evidence indicates that intra-continental rift systems have developed more commonly by the linking of uplifted areas on which individual crestal rift systems have formed (see discussion in Burke, 1977).

Detailed development of the rifts can be very complex even over a relatively short time. Figure 2.3 from Cerling and Powers (1977) illustrates how the rift axis has migrated almost 100 km east over the last 20 m.y. in the vicinity of Lake Turkana between the major Gregory and Ethiopian Rifts.

It is becoming widely appreciated that the structural peculiarities of the African Plate (its anomalous elevation, basin and swell structure, large amount of intra-plate or hot spot vulcanism, and active rift systems) are associated with Africa's position at rest for about the last 25 m.y. with respect to the underlying mantle and the earth's spin-axis (see, for example, Burke and Wilson, 1972; Briden and Gass, 1974; Burke and Dewey, 1974; Richter and Parsons, 1975; McKenzie and Weiss, 1975). Because there is no lateral motion between the lithosphere and the underlying asthenosphere, the present topography of the African Plate seems likely to bear an unusually simple relation to underlying mantle convection. Elevated areas, many of them capped by volcanoes (e.g., the Cape Verde Islands and the Ahaggar), are likely to overlie rising convection currents or plumes, and depressions (e.g., the Chad Basin) are likely to mark the areas between the rising currents. The eleva-

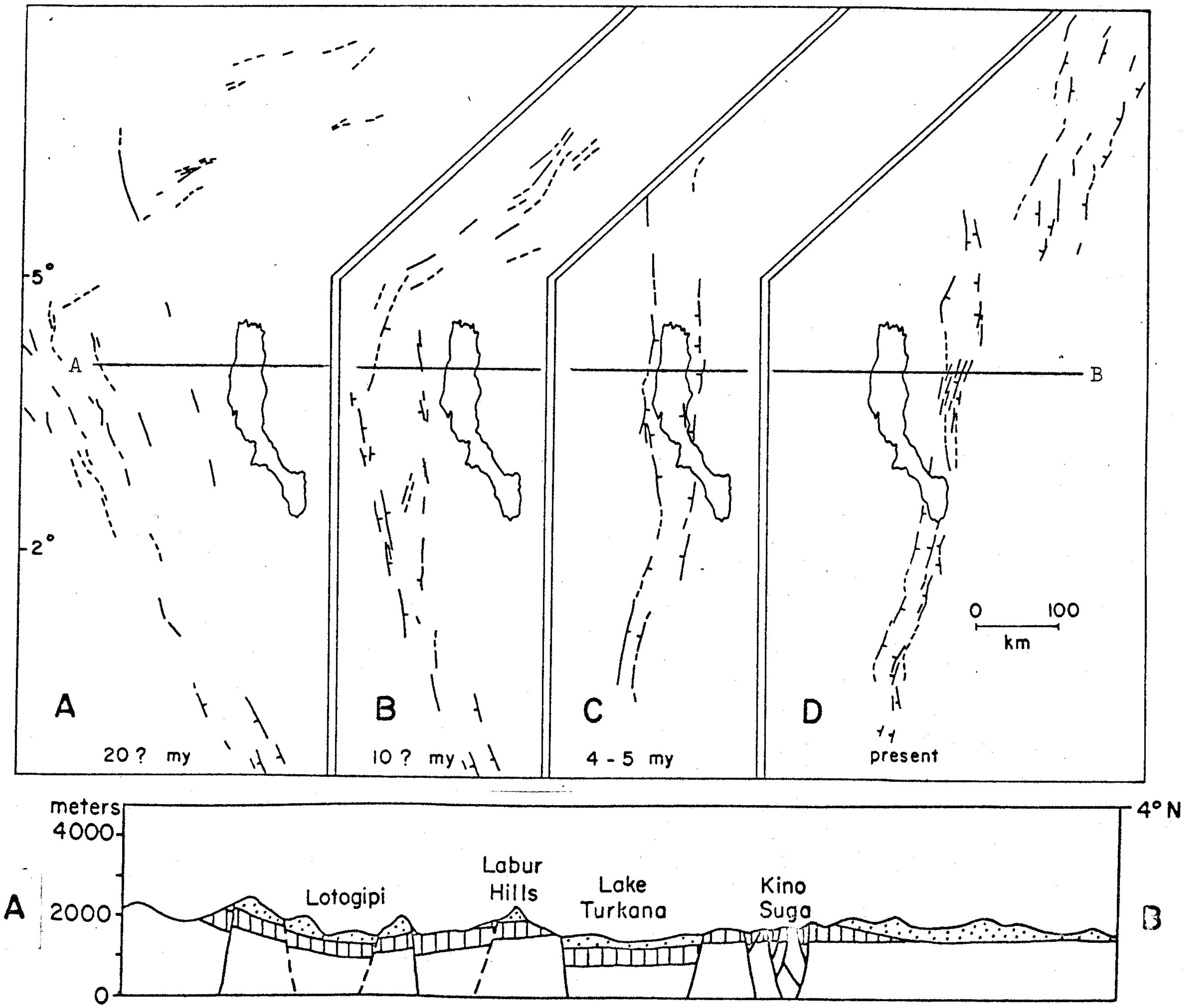

Figure 2.3. Maps and cross-sections across the Lake Turkana rift showing migration of the rift axis with time (after Cerling and Powers, 1977).

tions mark underlying areas of mass deficiency related to partial melting of the mantle produced by the hot rising currents. As long as the rising convection current continues to reach the base of the lithosphere in the same place, elevation is dynamically maintained as a response to underlying mass deficiency. Ages of volcanic rocks that date back to 25 m.y. ago on African uplifts indicate that these areas have been elevated, perhaps intermittently, throughout the Neogene.

The rift valleys of East Africa occur within a broad uplifted area (the East African swell) and are associated with discrete uplifts, as at Nakuru, within the swell. Geophysical observations on the rift system have suffered from difficulty in distinguishing effects related to the rift valleys from those related to the broad uplifts. The active rift systems of East Africa that overlie large volumes of hot, partly melted asthenosphere have their topographic manifestation in high ground and appear to be responsible for most teleseismic effects in the Rift System; for example, the poor transmission of Sn (Gumper and Pomeroy, 1970) is no doubt related to the presence of this material. Gravity observations are particularly ambiguous and susceptible to assumptions about regional fields as well as the effects of sediment accumulations in the rifts (Figure 2.4).

Perhaps the most structurally significant geophysical observations made in the East African Rift System were seismic refraction studies in the Kenya Rift reported by Griffiths, et al. (1971). Although their work is not unambiguous, Griffiths and his colleagues showed that their results could have been caused by an axial dike of basaltic material about 10 kilometers wide lying roughly in the middle of the Gregory rift,

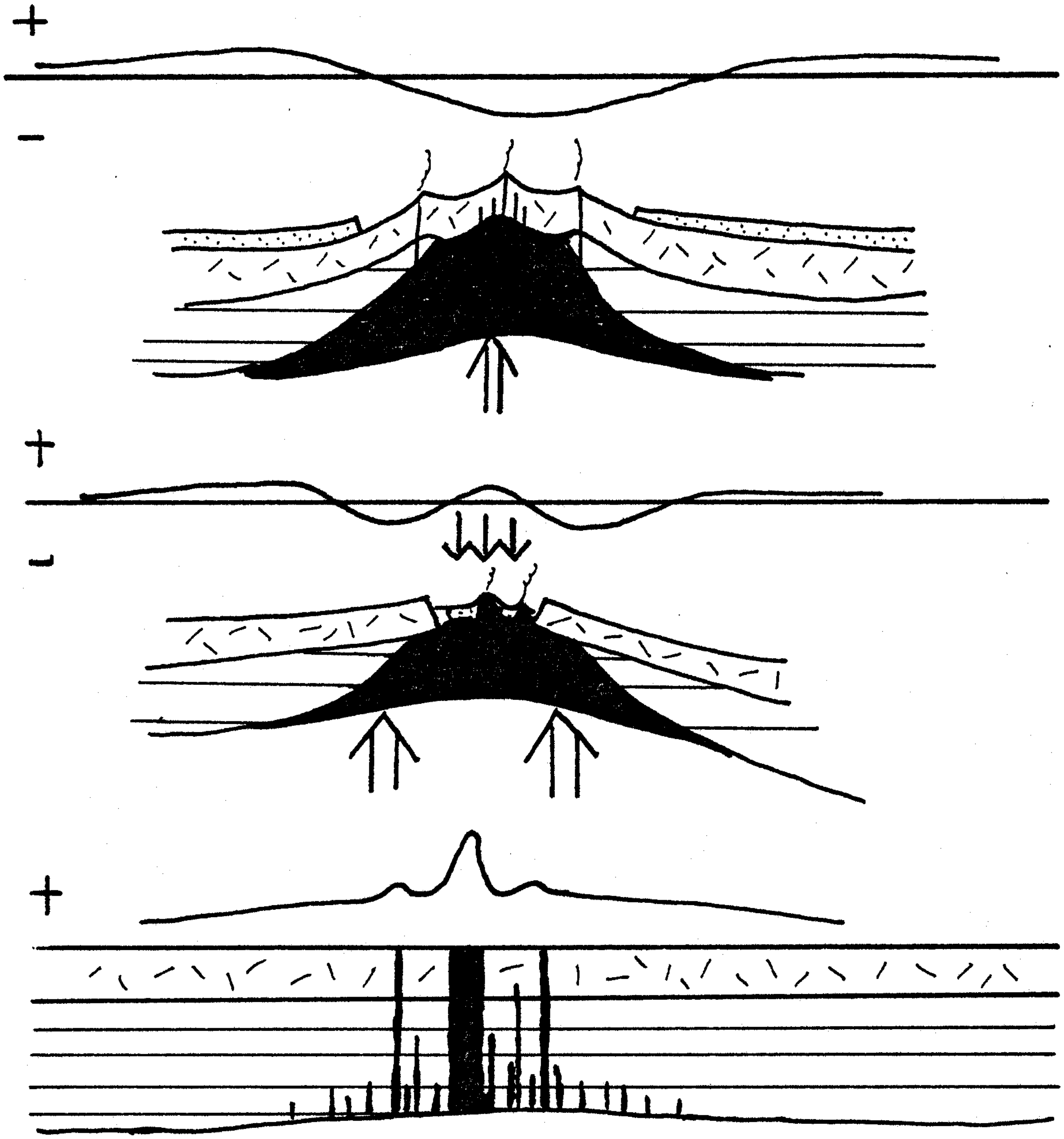

Figure 2.4. Bouguer anomalies at swell, active and failed rift structures. Bouguer anomaly over a swell (top) is negative and attributable to mass deficient, partly melted asthenosphere (black) rising (open arrow) through the lithospheric mantle (lined) and doming continental crust (dashes). Volcanoes crest the dome,and overlying sediments (dots) form a bordering scarp (based on the Ahaggar). A central positive Bouguer anomaly is related to axial dikes in an active rift (middle figure). A broad negative anomaly is attributed to the persistant asthenosphere swell (black, open arrows). The greater mass of the axial dike system due to mafic and ultramafic rocks makes the rift sink (solid arrows; based on the Kenya rift). In an inactive rift (bottom) with no sediments the axial dikes are exposed cutting the continental crust, but are more abundant in the lithospheric mantle. A sharp positive Bouguer anomaly marks the axial dike (based on the Great Dike of Rhodesia).

reaching nearly to the surface, and extending along the rift valley for a distance of over 100 kilometers. Khan and Mansfield (1971) showed that gravity observations in the Kenya rift were compatible with the presence of such an axial dike, and Searle and Gouin (1972) demonstrated that the gravity field in the 150 km wide Ethiopian rift at the latitude of Addis Ababa could be interpreted as related to three similar axial basaltic dikes together about 75 km wide.

Three other lines of evidence support the idea that axial dikes are an important element in rift valley structure. These are evidence from the igneous rocks of both ancient and modern rifts, and the outcrop pattern and gravity pattern of ancient rifts.

A great deal has been learned about the igneous rocks of ancient and modern rift systems. The most structurally significant conclusions from these studies are that rift igneous rocks have been produced by partial melting of the mantle, probably at depths in the range 60-120 km, and, although fractional crystallization and remelting have been important in producing the wide range of alkaline and to a lesser extent tholeiitic igneous rocks reported from the rifts, both isotopic and other geochemical evidence show that in general reaction with continental crustal rocks has been unimportant. The implication of these conclusions is that rift igneous rocks are discrete bodies with intrusive contacts and without significant reaction zones with their country rocks throughout the thickness of the continental crust. This result is entirely compatible with the axial dike model.

Gravity surveys over inactive rift systems are especially illuminating because the mass deficient part of the asthenosphere that dominates the gravity field over an active rift is not present under an inactive rift where the gravity field mainly results from the changes effected in the crust and lithosphere at the time of rift formation. Positive Bouguer anomalies attributable to axial dikes are known over other fossil rift structures (Figure 2.5) and the Great Dike of Rhodesia appears to be an outcrop of the igneous rocks responsible for the anomalies (Grantham, pers. comm., 1957). The Great Dike is about 10 km wide, has vertical margins (Weiss, 1940), and is composed of layered mafic and ultramafic rocks of approximately tholeiitic composition.

The floor of a rift valley will lie in isobaric equilibrium at a lower elevation than the shoulders because of the greater density of the basaltic and ultramafic rocks of its underlying axial dike system. The Neogene structure of Africa suggests that during active rifting, as sediments are eroded from elevated rift shoulders and deposited on the rift floor loading it and causing subsidence, the shoulders are elevated anew because of the persistent underlying mass deficiency in the asthenosphere. It is verly likely that this process has allowed the accumulation of sedimentary sections up to 10 km thick in such graben as the Triassic basins of the eastern United States. When the underlying mass deficient object is no longer there, general subsidence leaves the rift with its thick sediment section and a thinned underlying continental crust, the latter being an expression of the dike injection within the continent (Figure 2.5). The widely mapped variations in the thickness of continental crust (from 35-45 km over most continents) indicate that the lithosphere is strong enough to maintain the lateral stresses set up by the irregular mass distribution across the fossil

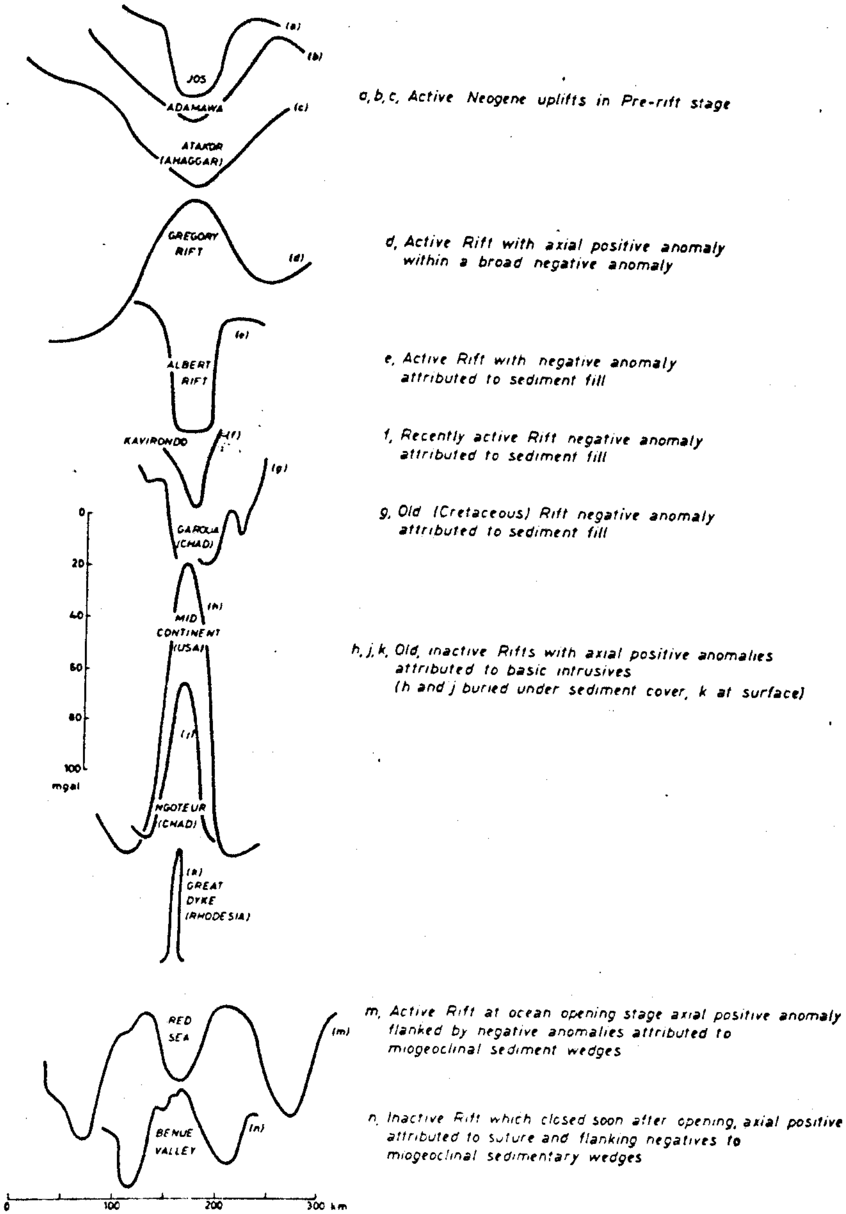

Figure 2.5. Selected gravity profiles across uplifts and rift structures. Active rifts and uplifts are associated with broad negative anomalies (a–f) which may be accentuated by sediment fill (e,f). An axial positive anomaly may also occur (d). Inactive rifts may display negative anomalies due to sediment fill (g) or narrow axial positives due to intrusives (h,j,k). Miogeoclinal sediment wedges cause flanking negative anomalies around the axial positive anomalies of an actively spreading young ocean (m) and a suture along which a young ocean has shut (n). For references see Burke and Whiteman, 1973, in Implications of Continental Drift to the Earth Sciences, Tarling and Runcorn, editors.

rift.

2.2.1.2. Other Active Rifts in Africa

Neogene tectonics of Africa away from the East African Rift system are dominated by the development of basin and swell topography (Burke and Wilson, 1972), but local rift activity is seen in reactivation of some Mesozoic structures. For example, the Zambezi rift, formed in Jurassic times, has steep slopes at its margins today which result from reactivation of border faults. The floor of the valley lies about 300 m below the Rhodesian plateau. Earthquakes were induced on faults in the valley as a result of the application of the load of water impounded behind the Kariba Dam (Gough and Gough, 1970). Other reactivated rifts are shown on Figure 2.1. All have topographic expressions comparable to that of the Zambezi rift. It is perhaps significant that renewed igneous activity does not appear to have accompanied the reactivation. Neogene vulcanism in a reactivated rift, as in the Benue trough, can generally be related to neighboring hot-spot volcanoes. Evidence of rift reactivation at the continental margins is not yet available but with increasing petroleum exploration in these areas some evidence of this may come to light.

Burke (1977, p. 380) has suggested that renewed subsidence in ancient rifts may result from triggering the basalt-eclogite transition as a result of changing thermal structure in the mantle. The arrest of Africa over the mantle convection pattern 25 m.y. ago may have produced an appropriate change in thermal structure.

2.2.2. RIFTS IN AFRICA ASSOCIATED WITH THE BREAK-UP OF GONDWANA

2.2.2.1. African Rifts at the Margin of the Central Atlantic

No rifts have been mapped striking into the African continent from the Atlantic coast except for the Casamance rift (Burke, 1976a, p. 97) and the rifts of Morocco (Figures 2.6 and 2.7). Burke distinguished the Casamance Rift from contours on pre-Mesozoic basement (Templeton, 1971, Figure 8). The salt domes at its mouth (Ayme, 1965) have been associated with the Lou-Ann and related salts of the Gulf of Mexico which are attributed to the spilling of Pacific Ocean waters into the young, dry oceanic floor of the Gulf of Mexico and its associated rifts (Burke, 1975).

The rifts of Morocco (Van Houten, 1977) are characterized by red-bed sequences, volcanics and evaporites. Only the Essaouira (Souss rift of Figure 2.6) opened into the Atlantic. The other rifts of Morocco (Figure 2.7) opened into the Tethys or developed only as intra-continental rifts.

Although the Moroccan rifts resemble the Casamance rift in containing thick saline deposits, the Tethys (not the Pacific) appears to have been the source of their salts. Saline waters do not appear to have spilled on a grand scale onto the graben-bottoms and ocean floor between Morocco and Senegal.

The virtual absence of marginal rifts from the bulge of Africa contrasts strongly with the numerous Triassic rifts of the North American Atlantic coast and this difference may be detectable in satellite gravity and magnetic patterns. We do not as yet recognize the reason for the extension of the continental rupture-path through the Triassic rift complexes along a line close to their southeastern (present coordinates) limit, rather than along some other path that would have left more rifts on the African coast and fewer on the American.

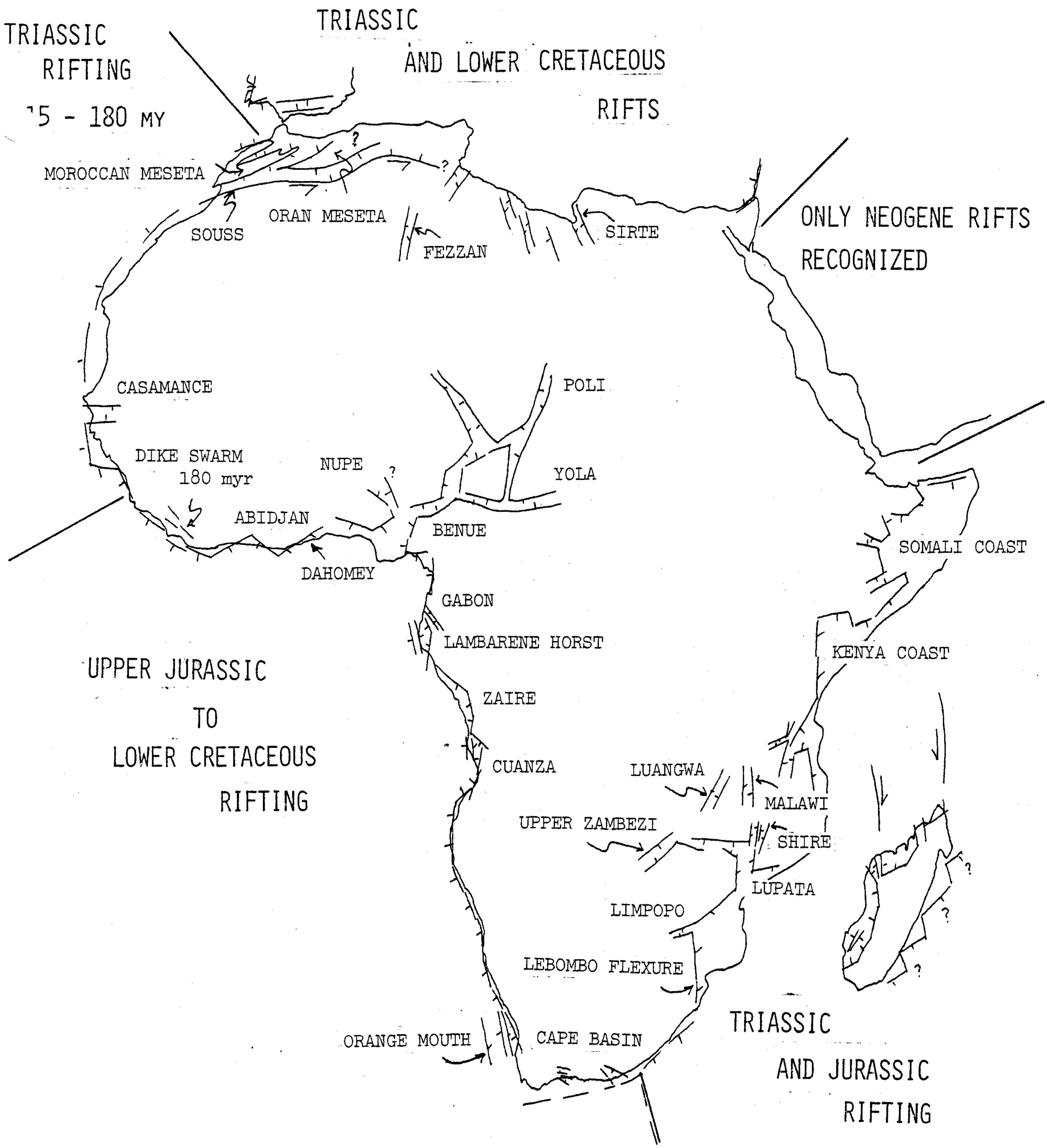

Figure 2.6. African rifts associated with the breakup of Gondwana.

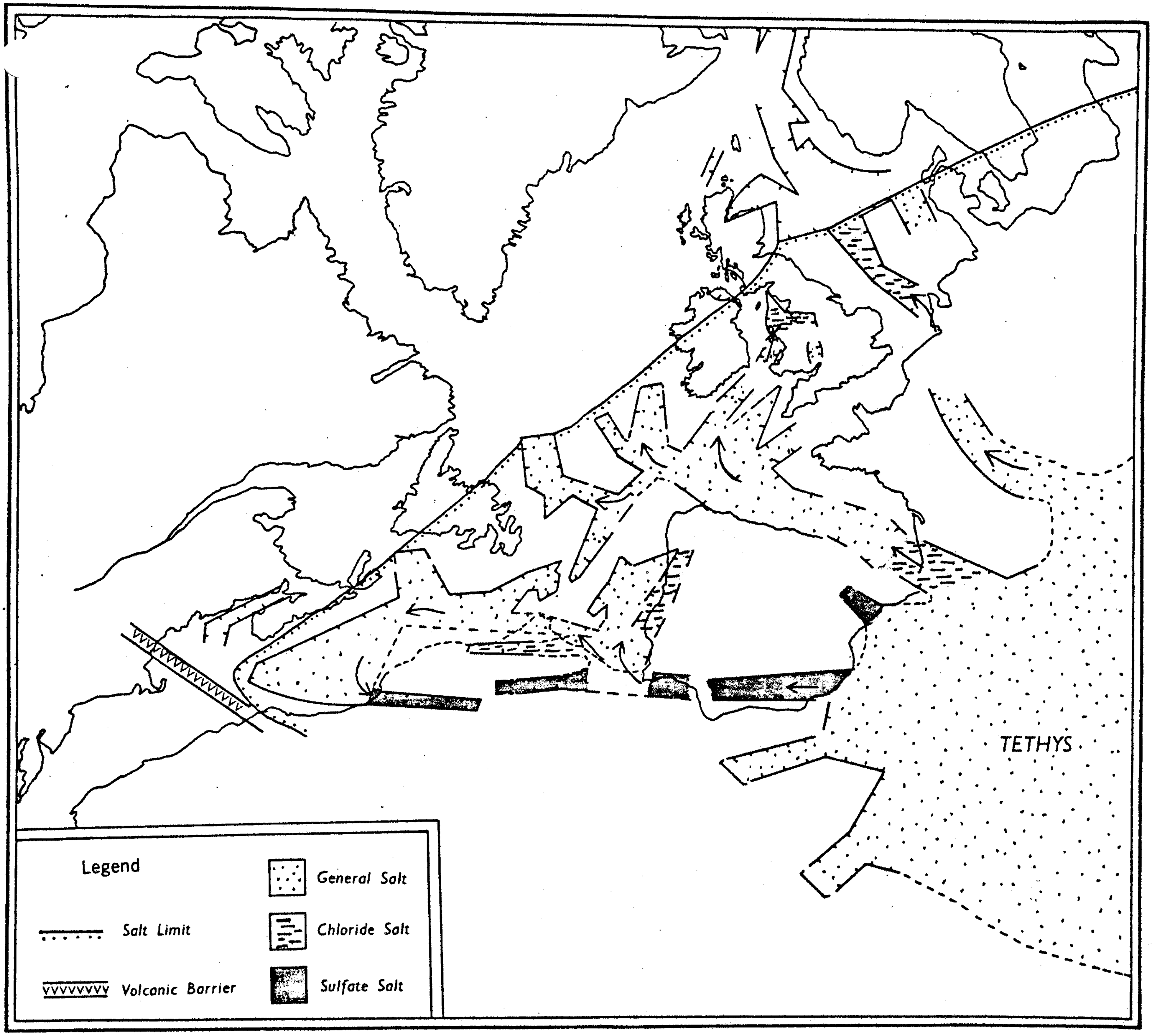

Figure 2.7. Rifts of Triassic age in North Africa, Canada, and Europe shown on a closed Atlantic. Salt deposits in these rifts were formed by evaporation of water spilled from the Tethys ocean. Deposits of mainly chloride salts are generally farther from the oceanic source than deposits of sulphate salts indicating that the spilling may have taken place in two stages. Salt waters failed to penetrate to the south-west through the volcanic barrier formed by the rocks of the White Mountain Magma Series in New Hampshire.

2.2.2.2. African Rifts Associated with the Opening of the South Atlantic

Four groups of rifts of Upper Jurassic to Lower Cretaceous age are distinguished (Figure 2.6). Those of the Gulf of Guinea formed in relation to the broadly transform motion of South America with respect to Africa, the Benue trough and its landward prolongations form a second population, those south of the Niger delta form a third, and the Cape Basin rifts form the fourth.

2.2.2.2.1. Gulf of Guinea

The Gulf of Guinea rifts (Figure 2.8) include some with evidence of strike-slip motion or compression parallel to the direction of opening (offshore Ghana) and others, striking perpendicular to the direction of opening, with evidence of extension (the Abidjan basin, Figure 2.8).

2.2.2.2.2. Benue Trough

The Benue Trough itself has been involved in convergence after rifting and is discussed in section 2.3.1.1., but the associated rifts northeast of it, being nearer the pole that describes convergence across the trough, have escaped substantial compression (Burk (sic) and Dewey, 1974). These rifts are best known from their positive Bouguer gravity fields (Louis, 1970) which are of the type reported over axial dikes. There seems to be some possibility of characterizing this area in a satellite-determined gravity field. Part of the area is buried under the Neogene sediments of the Chad basin, and because rifts resembling those of central Africa may underlie other intracontinental basins (Burke, 1976b), geophysical observations from satellites over this area may have distinctive characters.

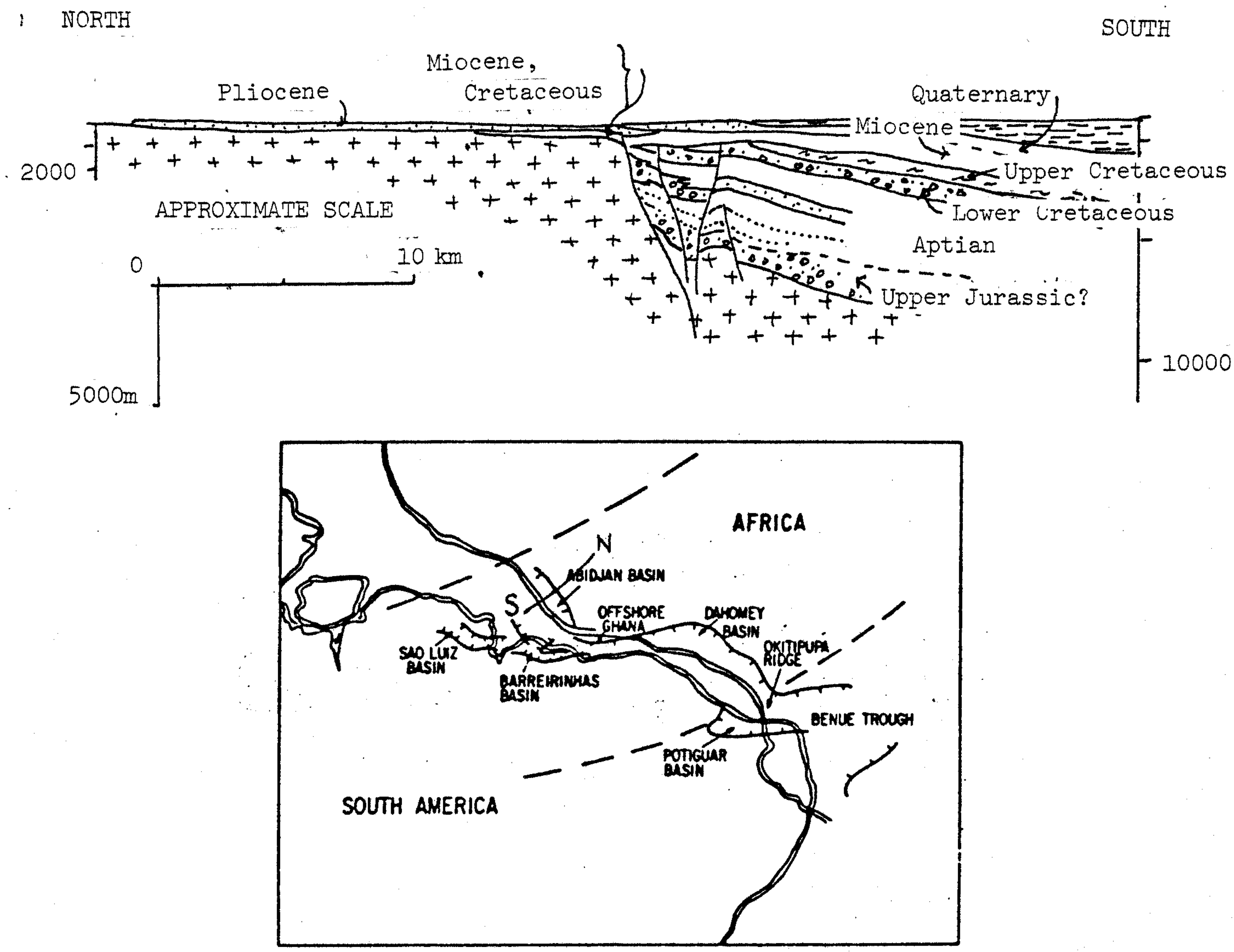

Figure 2.8. Section across the Abidjan basin rifted margin of the Gulf of Guinea (see Figure 2.6) to show the abrupt faulted boundary of the basin (based on Choubert and Faure-Muret, 1971).

2.2.2.2.3. South of the Niger Delta

Rifts to the south of the Niger delta are best known from published studies on the petroleum-bearing basins of Gabon (Belmonte, et. al., 1965) and Cuanza (Brognon and Verrier, 1966). Burke (1976a, p. 99) has drawn attention to prominent horsts within these rift basins that were mainly developed in the 10 million or so years preceding the earliest formation of South Atlantic ocean floor. The horizontal extent and vertical relief of the horsts are like those of the active Ruwenzori horst of the Western rift in the East African rift system. Apart from the well-documented rifts of the oil basins, the most prominent rift of the South Atlantic coast of Africa is a 300 km-long rift lying north of Cape Town that apparently formed in close relation to the Rio Salada and Colorado graben of Argentina (Burke, 1976a, Figure 2).

2.2.2.2.4. Rifts of the Cape Basin

The rifts of the Cape Basin (Burke, 1976a, p. 100 and Figure 2) are good examples of rifts that follow an earlier trend - in this case that of the Cape folds. They formed as the Malvinas (Falkland) plateau moved past South Africa in transform motion.

2.2.2.3. Rifts Associated with the Opening of the Indian Ocean

Rifts at the margin of the Indian Ocean are deeply buried by sediment deposited at the continental margin not only as a result of subsidence since the initial Jurassic rupture but also as a result of the addition of a second supply of sediment by erosion following elevation of the Neogene East African swell that embraces the East African Rift system. For this reason, the location and shape of the rift-boundaries (as shown on Figure 2.6) is relatively poorly known.

2.2.2.3.1. Mozambique Rifts

The southern-most rifts include the famous monoclinal flexure of the Lebombo (Cloos, 1939) where the continental margin is warped down for some kilometers in an area of contemporary dike emplacement and volcanic effusion. The rift occupied by the Limpopo River Valley strikes into Africa from the northern end of the Lebombo monocline and this rift not only contains late Triassic and Early Jurassic igneous rocks (Cox, 1970), but also substantial copper mineralization at Messina (Sawkins, 1976). North of the Limpopo, the rift and/or monoclinal flexure trends north-east and then north into the Lupata rift of the Lower Zambezi valley. East of this rift the monoclinal flexure trends nearly east-west to the coast. Burke and Whiteman (1973) were impressed by the resemblance of the trends of rifts and monoclines in Mozambique to those on the west coast of Madagascar, and for this reason suggested (1973, Figure 7) that Madagascar had rifted from Mozambique. Later research (Bunce and Molnar, 1977; Norton and Sclater, in prep.) has shown that Madagascar has rifted from a more northerly position.

Salt formed in Upper Jurassic time has been described from the Mozambique rift area (Kent, 1974, Figure 2) but is not known farther south. This limit is probably latitudinally controlled. A large part of the Mozambique Republic with thick sediment piles appears to be underlain by oceanic lithosphere.

2.2.2.3.2. Rifts of Somalia, Kenya and Tanzania

These rifts are known from petroleum exploration (see, for example, Kent, 1974). Evaporites, apparently of Lower Jurassic age, are known in the Tanzanian rifts and may be present off the Kenya coast. They are found again in the rift of Somalia. An as yet unresolved problem is that the earliest thick marine sequences are of Middle Jurassic age (Kent, 1974, p. 316) and this seems rather early when compared with the timing of continental rupture proposed by Norton and Sclater (in prep.).

2.2.2.3.3. Rifts of Madagascar

Two main rifting episodes of Karroo (Permian) and Upper Jurassic age appear to be recorded in the basins of Madagascar (Besairie, in Choubert and Faure-Muret, 1971, p. 552-558; Figure 2.9). The latter episode appears more likely to be related to the removal of the island from Africa.

2.2.2.4. Rifts of the Tethyan Coast of Africa

Two populations of rifts can be distinguished on the Mediterranean coast of Africa. The older population are of Late Triassic age and are associated with the formation of the western part of Tethys (Dewey et al., 1973). The younger group, including the oil-rich Sirte graben, formed during a Cretaceous rifting episode and are much better developed on the west coast of Africa (Burk (sic) and Dewey, 1974). No continental fragments are known to have left the African coast at this time.

2.2.2.5. Rifts of the African Interior

The Upper Zambezi and Luangwa rifts are of Late Triassic to Early Jurassic age, and it is possible that other poorly exposed rifts of central southern Africa may have been active at this time. Early Cretaceous rifting is best known from Southern Malawi and the Lupata Valley. Interior Africa thus appears to record the two main Mesozoic rifting

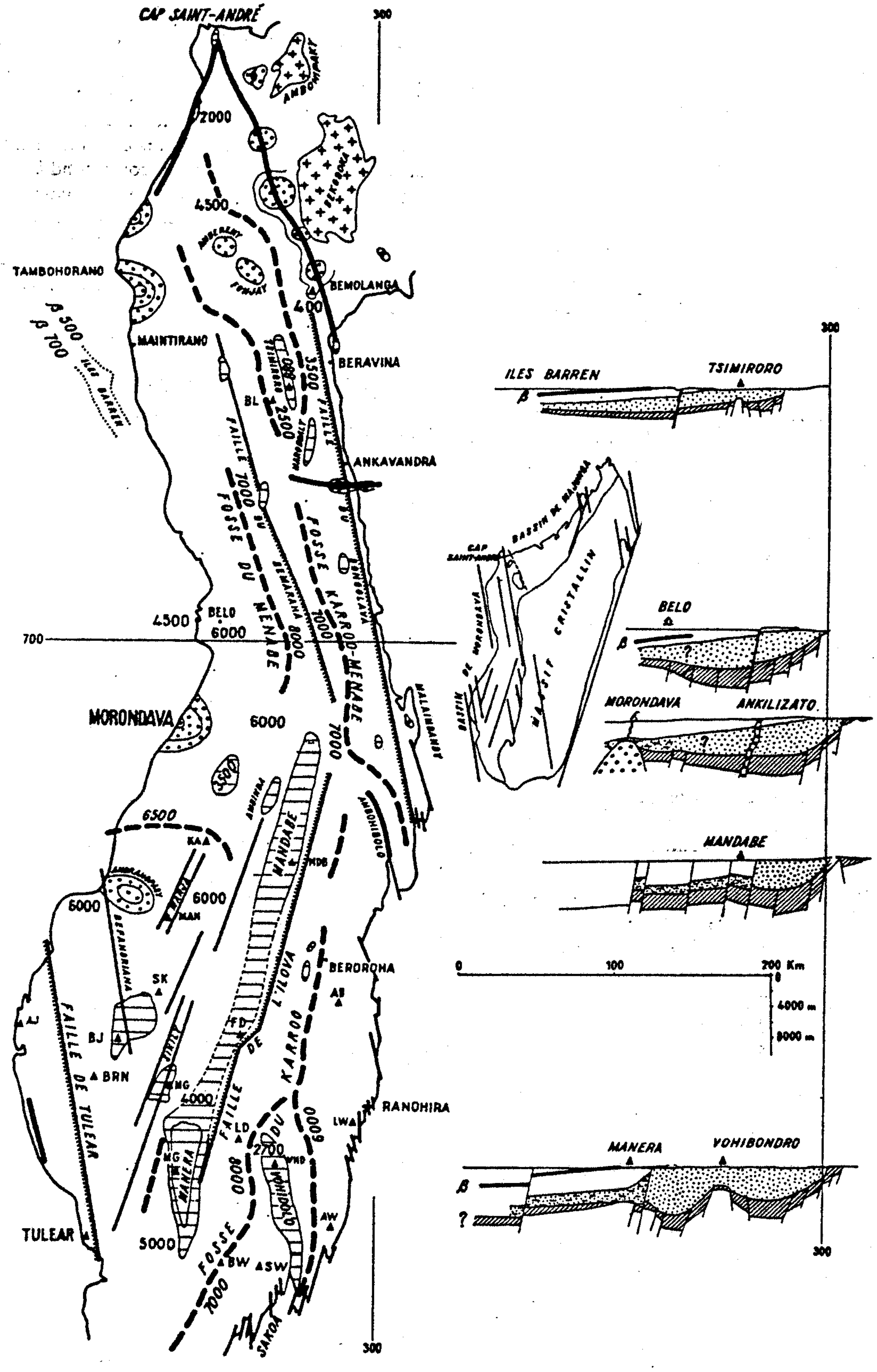

Figure 2.9. Rifts of the west coast of Madagascar. Inclined lines in the cross-section represent rocks of possible Permian age and are related to a first phase of rifting. A Late Jurassic rifting event is marked by the basalts (labelled β in cross-section). On the detailed map, solid heavy lines are positive Bouguer anomalies and dashed heavy lines are negative Bouguer anomalies (based on Besairie in Choubert and Faure-Muret, 1971).

intervals of Latest Triassic and Early Cretaceous that were associated with ocean formation and the breakup of Gondwana. The Latest Paleozoic rifting episode recorded in the Gondwanas of India has not been well identified in Africa where the contemporaneous Karroo beds are mainly deposited in a foreland trough associated with the Cape Fold collision orogeny.

2.2.3. OLDER RIFTS

Older rifts, especially Precambrian rifts, are not discussed in this review, not because they do not exist but because their recognition is extremely difficult at present. Analyses of the Precambrian geology of Africa in terms of the Wilson cycle have not yet developed to the extent that rift structures are clearly distinguished from fold belts.

2.3. SUTURES

2.3.1. PHANEROZOIC SUTURES

Africa has fewer sutures of Phanerozoic age than any other continent with the exception of the Indian peninsula. Only the Benue trough marks a convergent event less than 400 m.y. old and even this structure does not extend right through the continent.

Opening of the South Atlantic 130 m.y. ago removed the sutured arcs of South America from the Cape Fold Belt and thus robbed Africa of its only Paleozoic suture zone. The Pan-African events, which extend up into the early Paleozoic, are treated with Proterozoic phenomena.

It is an unresolved question whether or not the persistence of Africa for the last 600 m.y. as a continental object reflects large scale geophysical peculiarities. It is worth noting in comparison that the next largest continental object that has maintained its integrity for a long time is the interior of North America, which has been in one piece for about 1 b.y.

2.3.1.1. Benue Trough

Apart from Pan-African events that extend up from the Late Proterozoic into the Cambrian, mountain building activity in Africa during the Phanerozoic has occurred almost entirely around the edges of the continent. The most conspicuous exception is the highly folded, 100 km long and 150 km wide Benue trough of Cretaceous age. The trough is a fine example of the fact that fold belts within continents mark sites of ocean opening and closing. When the South Atlantic began to open at the beginning of the Cretaceous, the African continental mass formed two plates as an ocean began to open along the Benue trough. By 80 m.y. ago this ocean had sutured shut and Africa has since behaved as a single plate (Burke and Dewey, 1974). Many features of a small-scale Wilson cycle can be seen in the Benue trough (Burke, Dessauvagie and Whiteman, 1971, 1972). In early Cretaceous time rifting was accompanied by alkaline igneous activity; later, between the Aptian and the Cenomanian, thick montmorillonitic shale sequences apparently from basalt weathering occupied much of the trough, while deltas from the sides and the northeastern end deposited up to 3 km of sandstones. Folding and thrusting of Coniacian age produced large amounts of deformation which for a long time puzzled students of African geology (see Figure 2.10). The discovery that andesitic volcanic rocks and associated dioritic intrusions were widely developed both in outcrop and in the subsurface over an area of 10^4 square km in southeastern Nigeria (Burke, et al., 1972) was strong evidence that the embryonic ocean of the Benue had closed by the subduction of a southeasterly-dipping oceanic slab. Sulfide mineralization in the trough appears of rift-valley type (Burke and Sawkins, 1977) redistributed at the convergence, and the gravity field over the Benue suture (Figure 2.11) is highly distinctive (Ajakaiye and Burke, 1973).

2.3.1.2. Cape Fold Belt and Karroo Foreland Trough

At the southern tip of the African continent in the Cape Fold belt a Lower Paleozoic miogeocline went into a convergent mode in the Upper Paleozoic, and evidence indicating a continental collision to the south is well preserved in the Permo-Trias of the Karroo foreland trough or exogeosyncline.

The Cape folds resemble the Innuitian of the Canadian Arctic

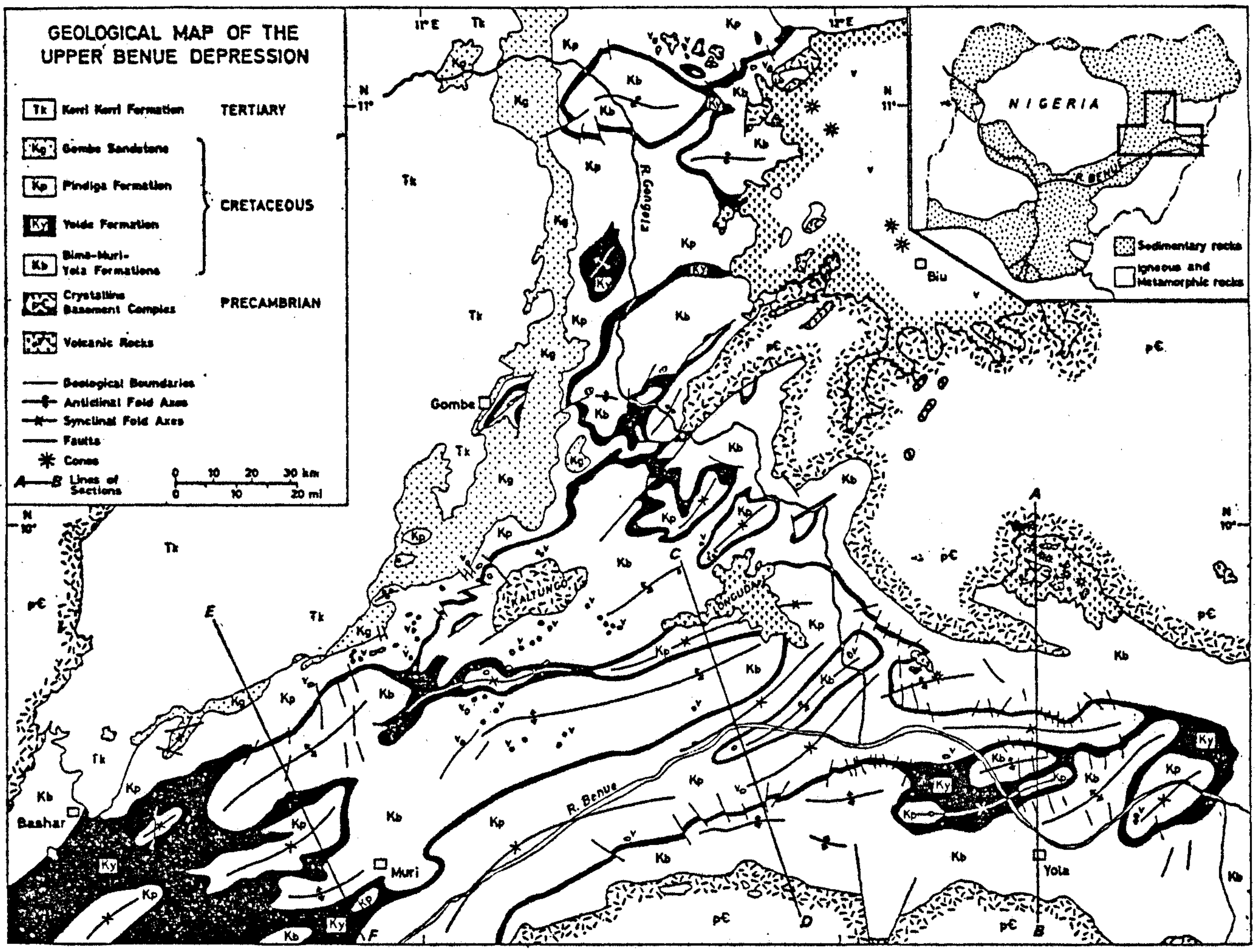

Figure 2.10. Geologic map of the Upper Benue Trough illustrating the extent and intensity of Late Cretaceous folding associated with the closing of the embryo Benue ocean (from Burke, Dessauvagie, and Whiteman, 1972).

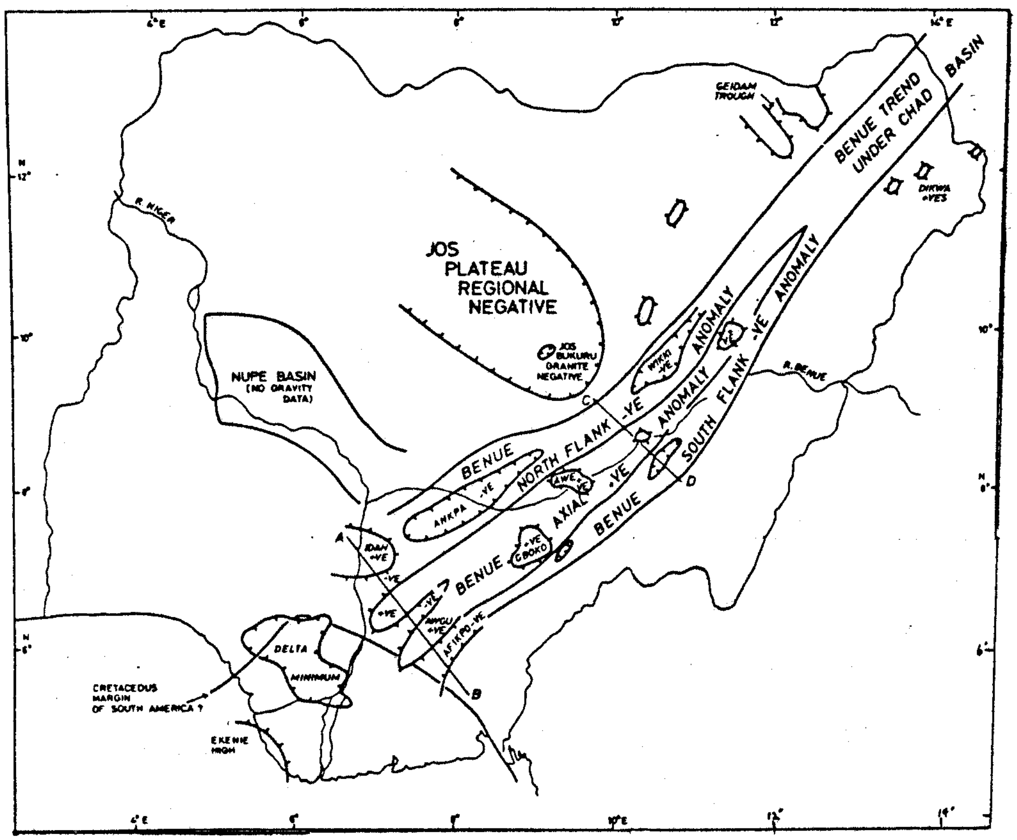

Figure 2.11. Gross features of Bouguer gravity field in Nigeria showing the Benue Trough with axial positive and flanking negative anomalies (from Ajakaiye and Burke, 1973).

Islands because in both the suture zone has been removed in the opening of a later ocean. In both cases operation of the Wilson cycle can be clearly inferred from the rocks that remain. Development of the Cape Province graben in the Lowest Cretaceous marked the opening of an ocean south of the Cape fold belt as the Falkland (Malvinas) plateau slid past Africa in transform motion (Burke, 1976a). The Cape suture zone may have been wide, resembling those of the Tasman belt and the South America Cordillera along strike. If it was wide, parts of it may underlie not only the Falkland plateau but also the concave side of the Antarctic peninsula.

2.3.2. PAN-AFRICAN SUTURES

Discussion of the fold belts of the Precambrian of Africa is presently in a state that may prove confusing for those not actively involved in these studies. A difference of opinion exists between those who interpret the fold belts as products of plate-tectonic processes and those who interpret them as products of other processes. Some writers compromise by attributing some African mountain belts to each group of processes. A recent summary in which both sets of opinions are presented is "The Past Distribution of Continents" (McElhinny, 1977).

In this review we interpret the mountain belts as products of plate-tectonic processes. We believe that these processes are the only ones known to be capable of making mountain belts, and it appears that the mountain belts of Africa are sufficiently similar to later collisional mountains (e.g., the Himalaya) so that the attribution of them to some different, necessarily less clearly defined, process seems unjustified.

Some writers on Precambrian orogeny have placed great faith in paleomagnetic studies (see, for example, articles in "The Past Distribution of the Continents") overlooking the fundamental limitation of paleomagnetic measurements, which is that only a paleolatitude can be inferred from them. Paleolongitude estimations always involve some assumptions and the widespread presence of sutures within the continents makes most of these assumptions invalid. For example, by assuming Africa has always been its present shape, paleolatitude studies have been interpreted to show that it was always in one piece (Briden, 1973). Although the presence of sutures within Africa invalidates this interpretation, and although an element of circularity may be perceived in the argument, it continues to receive attention. An exhaustive analysis of Precambrian paleomagnetic results shows that they are as compatible with the contemporary operations of plate tectonic processes as indeed they are with the operation of any other process of mountain building (Burke, et al., 1976).

2.3.2.1. Pan-African Suture Around the West African Craton

In 1971, Burke and Whiteman sketched the outline of Pan-African suture zones around the West African craton (Figure 2.12; see also Burke and Dewey, 1972). Leblanc (1976) has identified rocks lying in the suture zone at Bou Azzer in Morocco as ophiolite and this occurrence is of special interest as the oldest ophiolite so far identified and the only described complete (that is, not dismembered) Precambrian ophiolite. Similar, but dismembered, ophiolitic rocks occur in tectonic association with calc-alkaline volcanics in the residual hills ("boutonnieres") of the Wadi Saoura 500 km farther east, and we extend the Bou Azzer suture

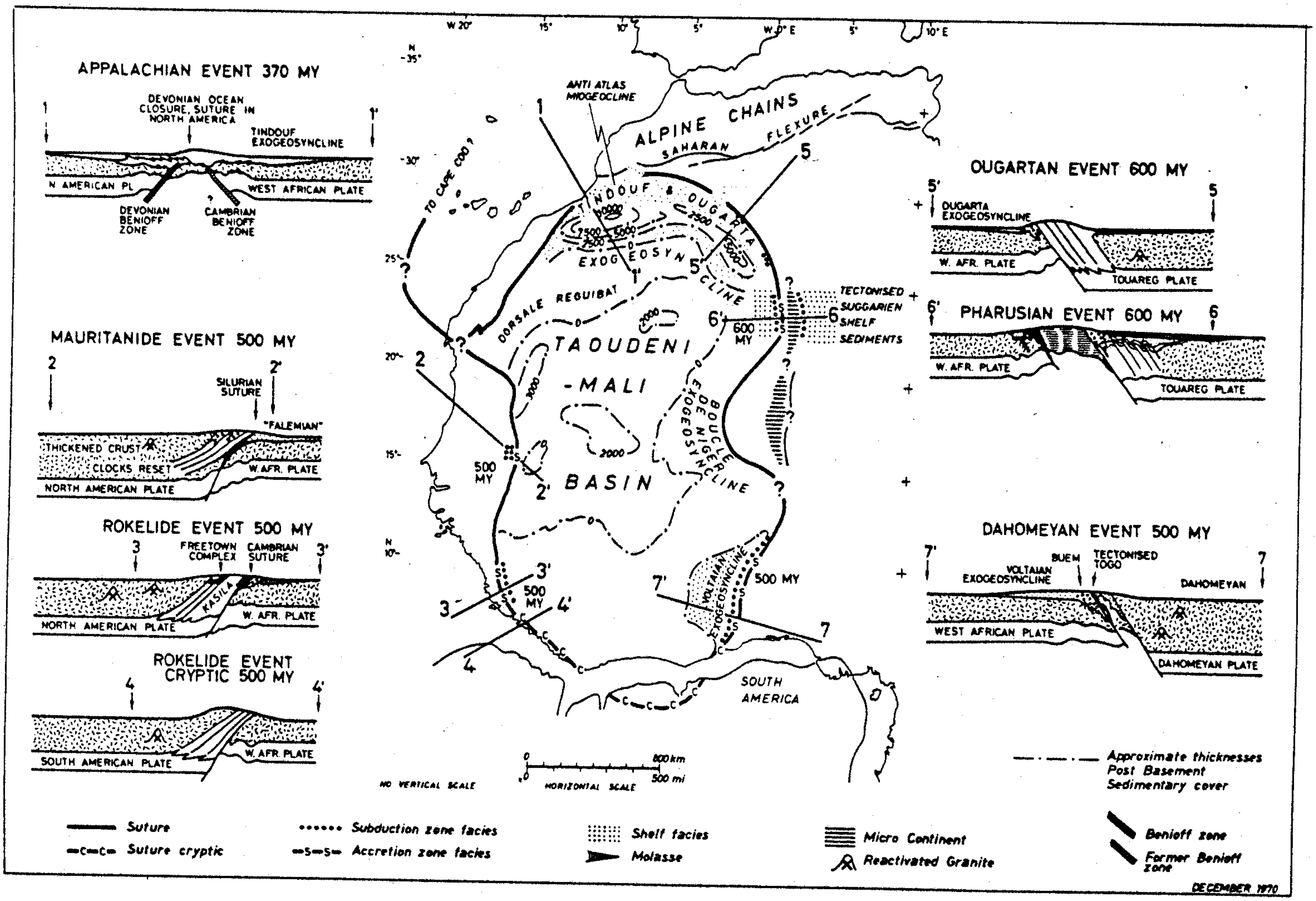

Figure 2.12. Pan-African collision orogenies around the West African craton. (From unpublished work of Burke and Whiteman). Exogeosynclinal or foreland basins are developed on the West African craton which is surrounded by reactivated basement interpreted as produced by continental thickening (Tibetan mechanism). The suture between these two groups of rocks shows "eugeosynclinal" developments in the Pharusien, the Buem and the Rokellides, but in some intervening areas it is cryptic,and reactivated basement abuts against unreactivated basement (from Burke and Dewey, 1973).

zone along a curving strike through these isolated outcrops (Figure 2.12, cross-section 5). The Tindouf and Ougarta basins, lying on the West African craton to the south of the suture zone, are excellent examples of foreland troughs (or exogeosynclines) formed during collision and overlying miogeoclines formed at the continental margin when the West African craton was ruptured. The strike of the axis of deposition in the two basins curves around parallel to the trend of the suture. An interpretation of the Tindouf basin as a marginal basin (Hurley et al., 1974) is not consistent with structural and stratigraphic evidence. Farther southeast in the Ahaggar are rocks described as dismembered ophiolites by Leblanc (1976, p. 34) and Caby (1972). Studies in this general area by Caby (1970) and others over the last decade have revealed a sequence of events readily explained in terms of the Wilson cycle. A continental shelf with a miogeoclinal wedge is represented by the 'Stromatolite Series' of Proterozoic age. Ophiolite obduction at the margin of the continent was associated with tectonism of island arc and marginal basin basalts, andesites and greywackes in a Pan-African continental collision. The later stages of this collision involved motion along enormous strike-slip faults (> 400 km long in Ouzzal) like those reported from Asia today by Molnar and Tapponnier (1975). The youngest products of the collision are potassic ignimbrites and glacial continental sandstones (of the 'Purple Series') which are analogous to those forming on the Tibetan Plateau as a result of continental collision today. Similar isolated occurrences of potassic calc-alkaline late, or post, tectonic volcanic rocks with high level granites have been reported from many areas of Pan-African reactivation (for example, Barbeau and Geze, 1957; McCurry, 1975) and we suggest that, like those in the Purple Series, they are surface expressions of the potassic acid magmas produced by widespread Pan-African partial melting of the continent resulting from thickening in post-collisional convergence. The lavas of the Tibetan Plateau (Hennig, 1915) are the modern analogue.

Burke and Dewey (1972) described the Pan-African suture farther south in Benin (Dahomey), Togo and Ghana, emphasizing the character of the reactivated terrain in Nigeria to the east; they also note that the suture becomes progressively cryptic southward so that although greywackes, serpentinites, basalts, and calc-alkaline volcanic rocks outcrop in Togoland and easternmost Ghana, shelf and miogeoclinal sediments from either side of the closed ocean are in contact south of the Volta river. In Brazil, around Sao Luis, the suture zone is completely cryptic with reactivated basement in thrust contact with unreactivated basement.

The suture reappears in Liberia where Thorman (1976) has recently described the Pan-African-Liberian age province boundary as a thrust junction with klippen at Gibi mountain. Thorman (1976, p. 856) rejects the idea that a suture zone passes through the coastal area of Liberia because he has mapped lithounits for tens of kilometers obliquely across the age boundary (Thorman, 1974) and because there is no mineral facies break at the Todi shear zone marking the age province boundary (granulites occur on both sides). He suggests (pers. comm.) that some type of event other than continental collision may have occurred, but we would point out that the evidence for continental collision is very strong provided enough strike length of the suture is considered. There are no rocks of obviously oceanic origin for about 1500 km along strike from the Buem Group of Ghana through Brazil and Liberia to the outcrop of the Rokel River group in Sierra Leone, although the suture zone is everywhere mappable as an age province boundary. As demonstrated by Dewey (1977), continental collision is sufficiently complex a process that a cryptic suture 1500 km long may readily be produced. The more unlikely phenomenon would be the closing of an ocean at two places 1500 km apart along strike and production of a mountain belt by a process unrelated to the Wilson cycle in between.

The geology of the Pan-African boundary in Sierra Leone, Senegal and Mauretania has been reviewed by Grant (1973) and Sougy and Dillon (1974). The latter authors concluded, as had Burke and Dewey (1972, p. 330) that the Rokelides, at the suture in Sierra Leone, might be related to operation of the Wilson cycle, and it is clear from Grant's descriptions and figures (Figure 2.13) that an interpretation of the Mauretanides in terms of operation of the Wilson cycle as Burke and Whiteman had suggested (in Burke and Dewey, 1973) is appropriate.

The Pan-African suture around the West African craton joins the suture system related to the closing of the Iapetus in Morocco. Although the details of the latter relationship have not been worked out, it is clear that the ocean closing processes produce a complex reticulate network of mountain belts. Grant (1973, p. 484) suggested that the reticulate pattern of Pan-African belts contrasts with the curvilinear shape of the Ural and Appalachian Phanerozoic collision zones. The contrast disappears once the nature of the Appalachians and Urals as parts of more extensive diachronous collision zones is appreciated. Again, the complex structure of the Alpine-Himalayan zone (Dewey, 1977) is a good guide.

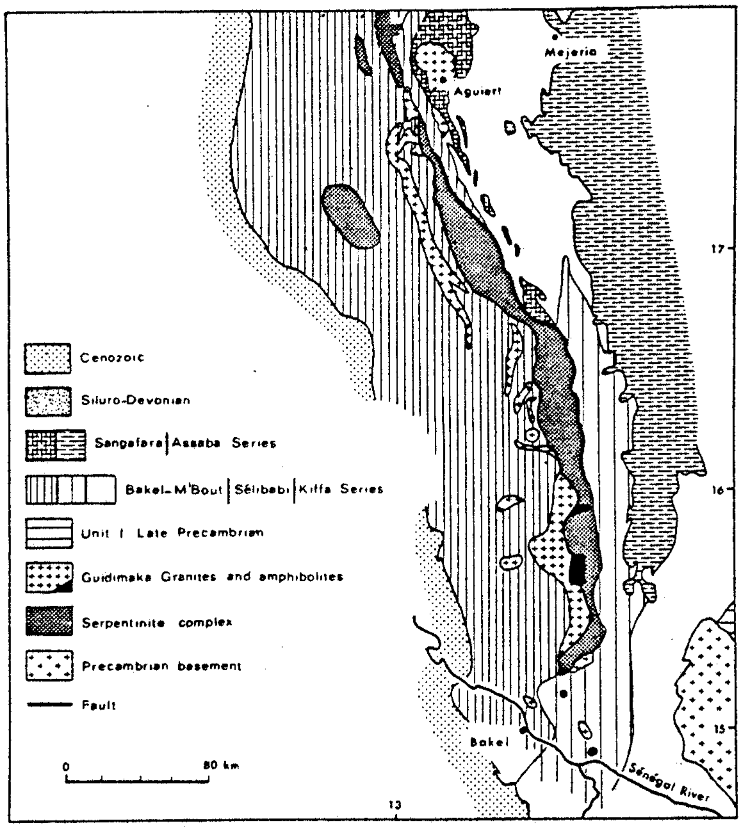

Figure 2.13. Late Pan-African suture zone in the Mauritanides of Senegal. The serpentinite complex is interpreted as a dismembered ophiolite (based on Grant, 1973).

The gravity signature over suture zones is typically that of a boundary or edge because suture zones mark places where continental objects with different histories are in contact. This is well illustrated at the margins of the West African Craton as shown in Figure 2.14.

2.3.2.2. Pan-African of Northeast Africa and Arabia

Reactivated Pan-African terrain stretches eastward for nearly 4,000 kilometers from the Ahaggar to Arabia. Within this huge area mineral ages are generally reset and there was intense late Proterozoic to early Phanerozoic granitic activity. There is, however, strong evidence that much of the material in the area had been continental crust for a long time. Nigerian basement at Ibadan yields whole rock Rb-Sr ages of 2.2 b.y. (Grant, 1970) and whole rock Pb ages of 2.75 b.y. (Oversby, 1976). Ages indicating events about 1.0 b.y. ago are also reported (Grant et al., 1972) from this area of Pan-African reactivation and a Kibaran suture may lie near Ile-Ife.

It seems unlikely that reactivation over the whole northeast of Africa can be related solely to collision at the edge of the West African craton and there is a likelihood that Pan-African cryptic sutures remain to be discovered in Tchad, Libya, the Central African Empire and Sudan Republic (see, for example, Nagy, et al., 1976). The existence of such sutures in the Arabian shield was reported by Brown and Coleman (1972) and has been elaborated on by Bakor, et al., (1976). These authors describe a dismembered ophiolite from Jebel al Wask, between Medina and the Red Sea, and conclude that it is likely to represent the floor of a marginal basin. By closing the Red Sea they are able to link (Figure 2.15) Jebel al Wask and some tens of other ultramafic occurrences into

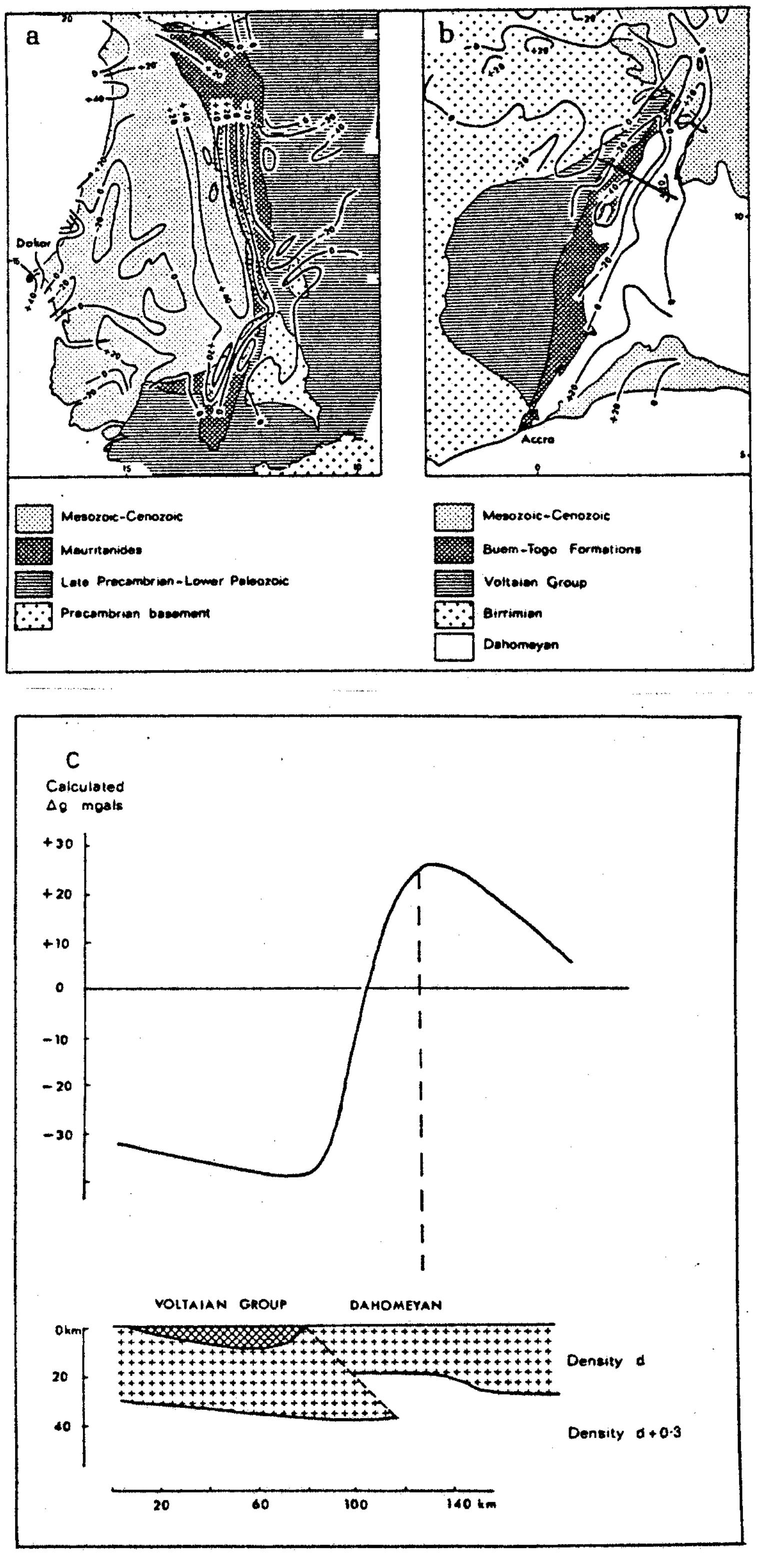

Figure 2.14. Maps (a,b) and cross-section (c) showing Bouguer anomalies across the Pan-African suture zones in Senegal-Mauritania (a) and Ghana-Togo-Benin (b and c). The juxtaposition of formerly separate continental objects in the suture zone yields a characteristic edge effect gravity anomaly (figures from Grant, 1973).

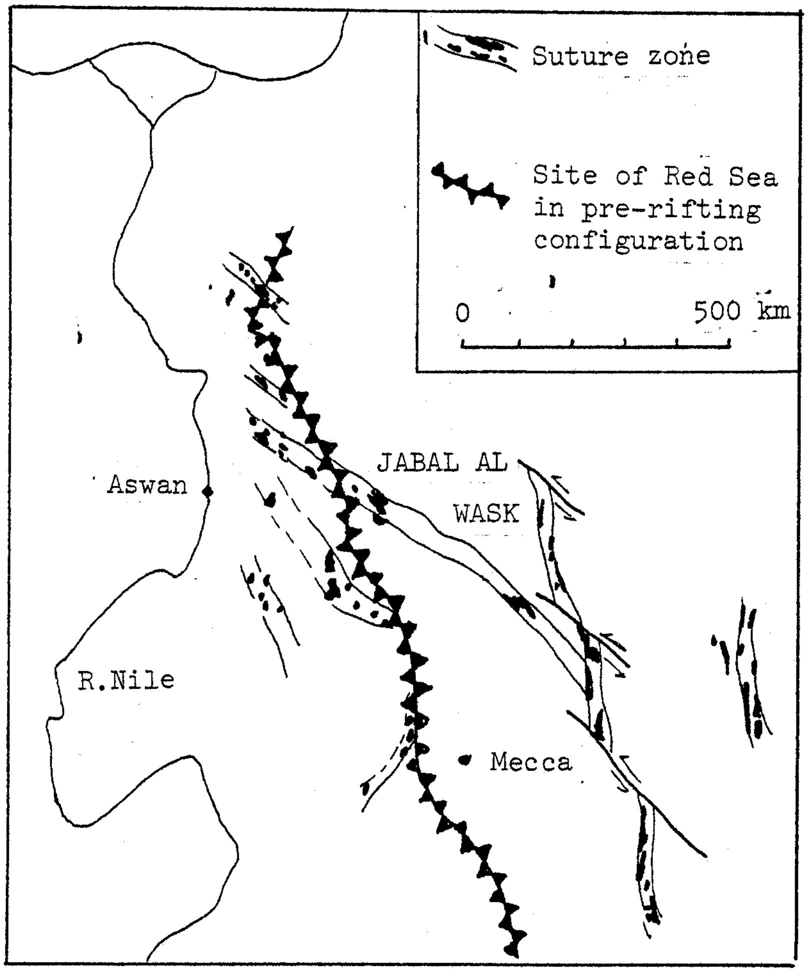

Figure 2.15. Distribution of Pan-African suture zones in western Arabia and north-east Africa (after Bakor, Gass and Neary, 1976).

belts in Arabia, Eygpt and Northern Sudan. They interpret the whole area with its association of oceanic and island arc volcanic type rocks as a site of Pan-African arc aggregation. Al Shanti and Mitchell (1976) interpreted the easternmost ultramafic belt in Arabia as a site of continental collision, and it may be that continental collision and reactivation followed episodes of arc and micro-continental collision.

An as yet unanswered question is how far to the west and south of Arabia this Pan-African area of arc collisions can be extended. Descriptions of rocks in the Precambrian of Sudan Republic (Whiteman, 1972) and Southern Ethiopia (pers. comm. from M.J. deWit, 1977) indicate that the Arabian arc-collision zone may extend that far, but the Pan-African farther south in East Africa appears to lack young island arc and ocean floor rocks.

2.3.2.3. Mozambique Belt

The reactivated Pan-African terrain of East Africa, known as the Mozambique Belt, contains no well described sutures south of those in Sudan (Bakor, et al., 1976). Its western border resembles the Grenville Front in being strongly sheared in some areas though not in others and in having structures mapped across it for many tens of kilometers. We interpret the western boundary of the Mozambique Belt as likely to be an analogue of the Grenville Front, that is, a zone within a continent that marks the limit of convergence, thickening and reactivation. Comparison of the Mozambique Belt with the margins of the Tibetan Plateau (Molnar and Tapponnier, 1975) suggests that these boundaries will be loci of strike-slip motion (like the Altyn Tagh fault) in some places and thrusting (as in the Nan Shan) in others.

2.3.2.4. Pan-African of Namibia and West Congo

Kröner (1974, 1975) has described the development of the Gariep area in terms of the Wilson cycle. Continental rupture and formation of graben facies 900-800 m.y. ago was followed by ocean and arc development with suturing at about the beginning of the Phanerozoic. Kröner (1974, p. 101) has reported one of the rare Precambrian occurrences of blueschist metamorphism in association with the convergent phase of this cycle in the Bogenfels area to the north.

The Pan-African rocks of Namibia and western South Africa marking a site of ocean closing lie close to the Atlantic coast, illustrating the familiar idea (Wilson, 1966) that oceans open in places where other oceans have previously closed. Followed northward along the coast Pan-African fold belts diverge as operation of the Wilson cycle requires they must. One structure persists along the coast through Angola, Zaire and Congo to Gabon and is best marked by an exogeosynclinal or foreland trough development (Schermerhorn and Stanton, 1960), the suture zone and mobile core being unexposed but lying to the west under the continental shelves of Africa and Brazil. The other structure turns inland to form the Damarides (Figure 2.16).

2.3.2.5. Pan-African of Damarides

The Damarides (Clifford, 1967; Kröner, 1977) are among the most interesting of Pan-African fold belts. Shelf carbonate facies of the Otavi and Naukluft sequences occur on both sides of the 400 km wide belt and give way to more clastic, tectonized and metamorphosed rocks closer

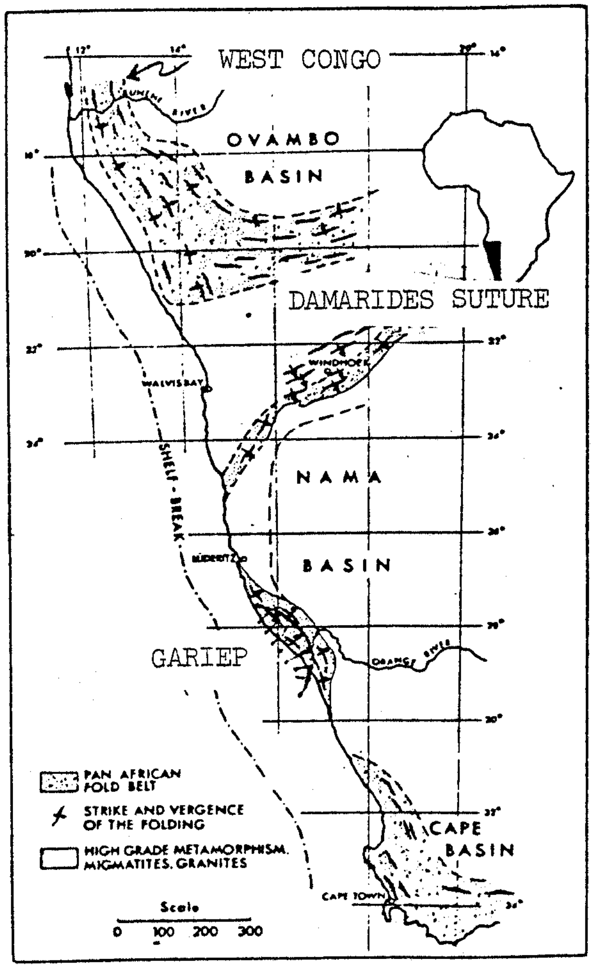

Figure 2.16. Pan-African folds of south-western Africa. The West Congo and Cape Basin-Gariep folds merge with the Damarides near Walvis Bay in a suture knot. The Atlantic Ocean opened along the north-trending West Congo and Cape Basin-Gariep sutures (based on Martin, 1975).

to the center (Figure 2.16). This relationship alone is strongly suggestive of two miogeoclines facing an area that was once ocean and has now become a fold belt. The area within the fold belt is currently being studied by various methods (Anon, 1975). It contains few mafic rocks but much granite. The best described mafic rocks are the Matchless amphibolites (Anon, 1975) with which minor ultramafics are associated and which lie somewhat to the south of the axis of the fold belt extending for 300 km along strike (Martin, 1965). Kröner (1975, 1977) rejects the idea that this mafic and ultramafic material might represent dismembered ophiolite because it occurs now in a 'flysch-type schist'. Presumably, if it reached its present position as a result of operation of the Wilson cycle, it has been tectonically emplaced so this argument is irrelevant. What is impressive about the Matchless amphibolite is its persistence along strike (Anon, 1975, Figure 10) and its location within an intracontinental fold belt between two miogeoclines. The strongest argument for ocean closure in the Damarides is that along strike in both directions in Gariep and the Zambesi valley there is good evidence of suturing.

Kröner (1977, p. 121) states that many features of the Damaride orogen are incompatible with the hypothesis of collision. These features are of two kinds: some, such as alleged continuity of lithostratigraphic units across the orogen, are unsupported assertions; others, such as the occurrence of basement within the belt, are not incompatible with the collision hypothesis but only with Kröner's limited understanding of the complexity of plate-tectonic processes.

2.3.2.6. Pan-African of Zambesi

To the northeast the Damarides disappear under the sands of the Kalahari and reappear as the Zambesi Belt. The Zambesi Belt is unique among Pan-African fold belts because a sketch-map has been published purporting to show continuity of older structures across it (Shackleton, 1973, his Figure 1).

Our conclusion that the Zambesi Belt may mark a site of operation of of the Wilson cycle is supported by evidence from the Urungwe Klippe (not depicted on Shackleton, 1973, Figure 1). This is an area of metasediments in Rhodesia thrust southward from the Zambesi Belt (Stagman, 1962; Shackleton, Vail and Wood, 1966). At the base of the main thrust slice there are local occurrences of banded ironstone, serpentinite and greenstone. We suggest that these might be material obducted from the Zambian suture during continental collision.

The Zambian suture extends eastward into the reactivated Mozambique Belt, but its exact location is unmapped, as it passes through country overlain by younger rocks.

2.3.3. OLDER PROTEROZOIC SUTURES

2.3.3.1. Namaqualand - Natal Fold Belt

In Southern Africa Matthews (1972) has reported a suture zone of one b.y. age with obducted dismembered ophiolite at the southern border of the Kaapvaal craton close to the Natal coast. When this boundary is pursued westward it becomes cryptic, but the Namaqualand-Natal fold belt to the south is a reactivated zone like those of the Grenville and the Pan-African, and we suggest that it represents, like them, a Tibetized collision zone. Watters (1976) has described a calc-alkaline magmatic arc from the northwestern corner of the Namaqua fold belt. The development of this feature, the Rehoboth magmatic arc, at a convergent margin requires operation of a Wilson cycle in the interval 1.3-0.9 b.y. roughly on the site of the cycle represented in the Gariep-Damaride suture zone.

2.3.3.2. Kibaride Fold Belt

The Kibaride fold belt of mid-Proterozoic age (Cahen, 1970, p. 100) contains an immense thickness of clastic sediments and minor volcanics over more than 300 km of strike length in Zaire. Graben facies rock associations are common but no well developed suture zone has yet been identified. A large proportion of highly potassic granites suggests that collision played a major part in development of the Kibarides.

2.3.3.3. Ubendian Fold Belt

The Ubendian is an Early Proterozoic fold belt (> 1800 m.y. old) about 100 km long by 200 km wide with a considerable amount of metavolcanic, metabasic and ultrabasic material (Pallister, 1971). Calcic anorthosites invite comparison with those of the Archaean of Greenland and the Limpopo which we have suggested may mark old ocean floor material (Burke et al., 1976, p. 123) although others (for example, Windley, 1976) see them as related to convergent plate margin calc-alkaline igneous activity.

2.3.3.4. Limpopo Fold Belt

The oldest collision zone in Africa, apart from the very numerous sutures of the Archaean and Proterozoic (Birrimian) greenstone belts, is the Limpopo belt. Coward et al., (1976) have suggested that the Limpopo belt sediments can be traced northwestward into the Matsitama greenstone belt of northern Botswana and we have plotted this curvature of strike. Their Figure 5 summarises the development of this area in terms of the Wilson cycle. Coward and his colleagues have analyzed in detail the polyphase deformation in a large area of southern Rhodesia in relation to the convergent plate margin processes of the Limpopo valley.

Opening and closing of the ocean on the site of the Limpopo may have been a very rapid process, as the emplacement of the Great Dike, apparently the site of the world's oldest failed rift system (Burke, 1977), postdated the events analyzed by Coward and others. The termination of the Great Dike in the Limpopo belt is associated with a 2.0 b.y. old orogenic event that reactivated much of the older Limpopo material and presumably represents a second Wilson cycle on a site close to that of the one described by Coward et al.

Interpreters of the Limpopo on a non plate-tectonic basis are numerous and Kröner (1977, p. 105-107) may be regarded as typical. He rejects the first order observation that the Limpopo is a fold belt between older continental objects and therefore likely to be a place where an ocean has closed because it has some properties -- for example, "symmetry", and an absence of molasse -- which he quite wrongly asserts collision orogens cannot have. The Limpopo belt contains large volumes of mafic and ultramafic rocks which today could only be made in an oceanic environment. No alternative mechanism has yet been suggested for producing these rocks.

2.3.4. ARCHAEAN SUTURES OF AFRICA

Old greenstone belts are widespread in Africa, those of the Kaapvaal craton having been described in particular detail. Although all sorts of non-plate tectonic interpretations of greenstone belts are current there is no need to seek peculiar non-uniformitarian processes that might have operated in the remote past, because the rocks and structures of the greenstone belts closely resemble those of island arc convergent environments (Burke, Dewey and Kidd, 1976). Because the island arc rocks of the greenstone belts are preserved within the continents they mark sites of ocean closing.

Greenstone belt provinces are relatively small areas up to a few thousand kilometers across, that may be geophysically coherent sources to satellite detectors.

The major greenstone belts of Africa are shown on Figure 2.17. It is worth pointing out that the largest single area, the Birrimian province of West Africa, was welded together as late as 2 b.y. ago, half a billion years after the traditionally accepted end of the Archean greenstone belt cycle of behavior.

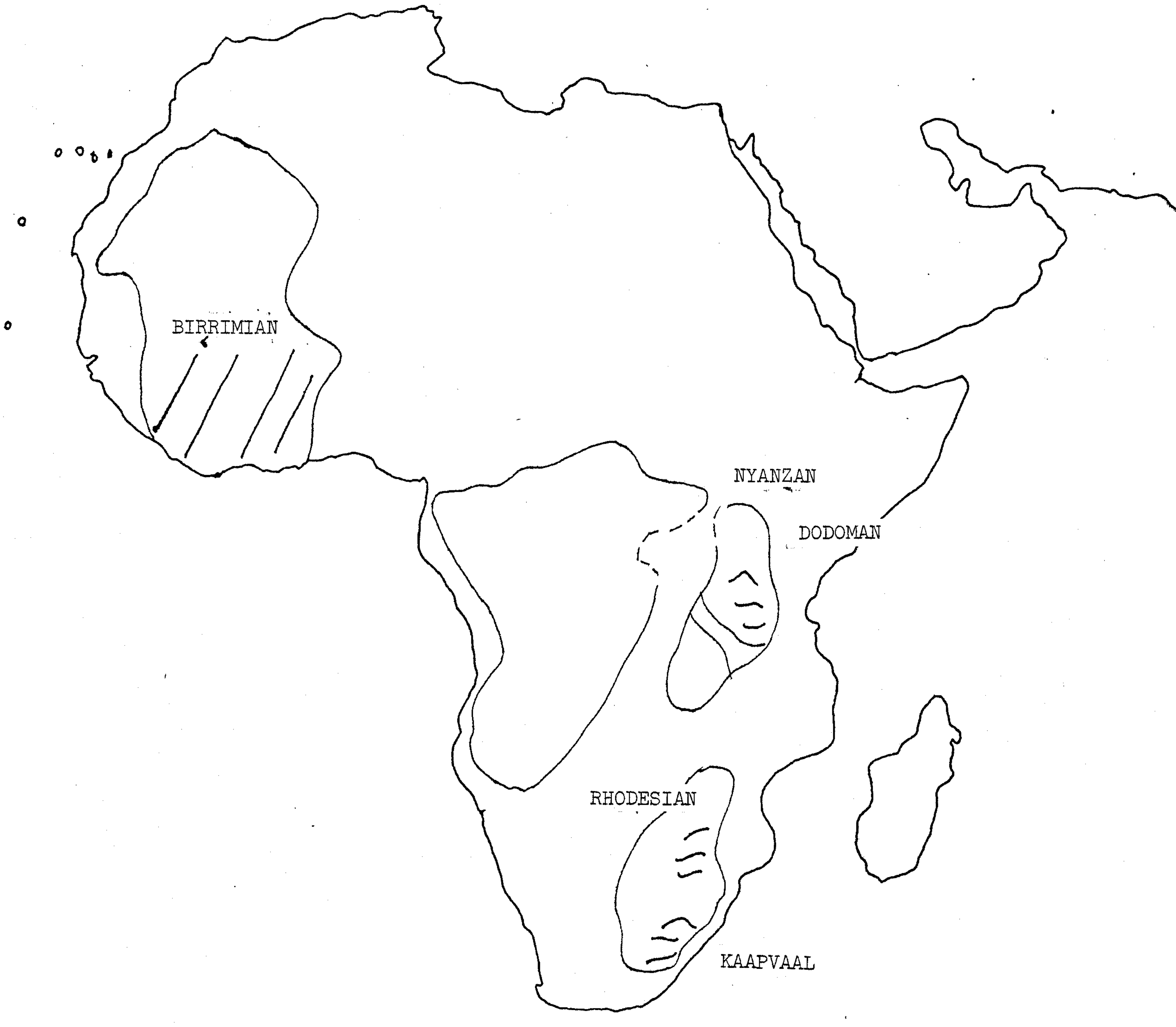

Figure 2.17. African cratons roughly 2.0 byr old, showing general trends of greenstone belts. Most, if not all, of the greenstone belts mark sutures. Greenstone belts in the Kaapvaal and Rhodesian cratons range between 3.6 and 2.5 byr old, but those in the Birrimian include some as young as 2.0 byr.

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3.0. ANTARCTICA

3.1. INTRODUCTION

Because of inaccessibility and limited exposure (Figure 3.1), little of Antarctica has been directly mapped, and it is therefore understood in less detail than the other continents. In addition, the ice and harsh climate prevent direct study during much of the year. If the ice sheets, four kilometers thick in places, melted entirely and isostatic equilibrium achieved, the continental configuration would consist of a large block in the east and a series of islands, including the archipelago of the Antarctic Peninsula, in the west (Figure 3.2.).

Geologically Antarctica consists of a large Precambrian block or group of blocks (East Antarctica) plus a series of younger rocks in accreted orogenic belts (Western Antarctica and the Antarctic Peninsula) which range in age from Late Precambrian to Cretaceous (Figure 3.3); in addition, Tertiary and Quaternary volcanism is seen locally. The geographic and geologic boundary is marked by a group of ranges known collectively as the Transantarctic Mountains (Figure 3.4).

Lack of accessible outcrop has encouraged the collection of indirect evidence from geophysical work; drilling is generally limited to ice studies. For example, the extent of the continent under the ice could only have been determined by seismic refraction. Gravity studies (Smithson, 1972) outlined deep structure under the Transantarctic Mountains, and magnetics allowed dating of the separate components of the boundary rifts. The dearth of direct observations has resulted in the rifts and sutures of Antarctica being more speculatively defined and less precisely located than in the other

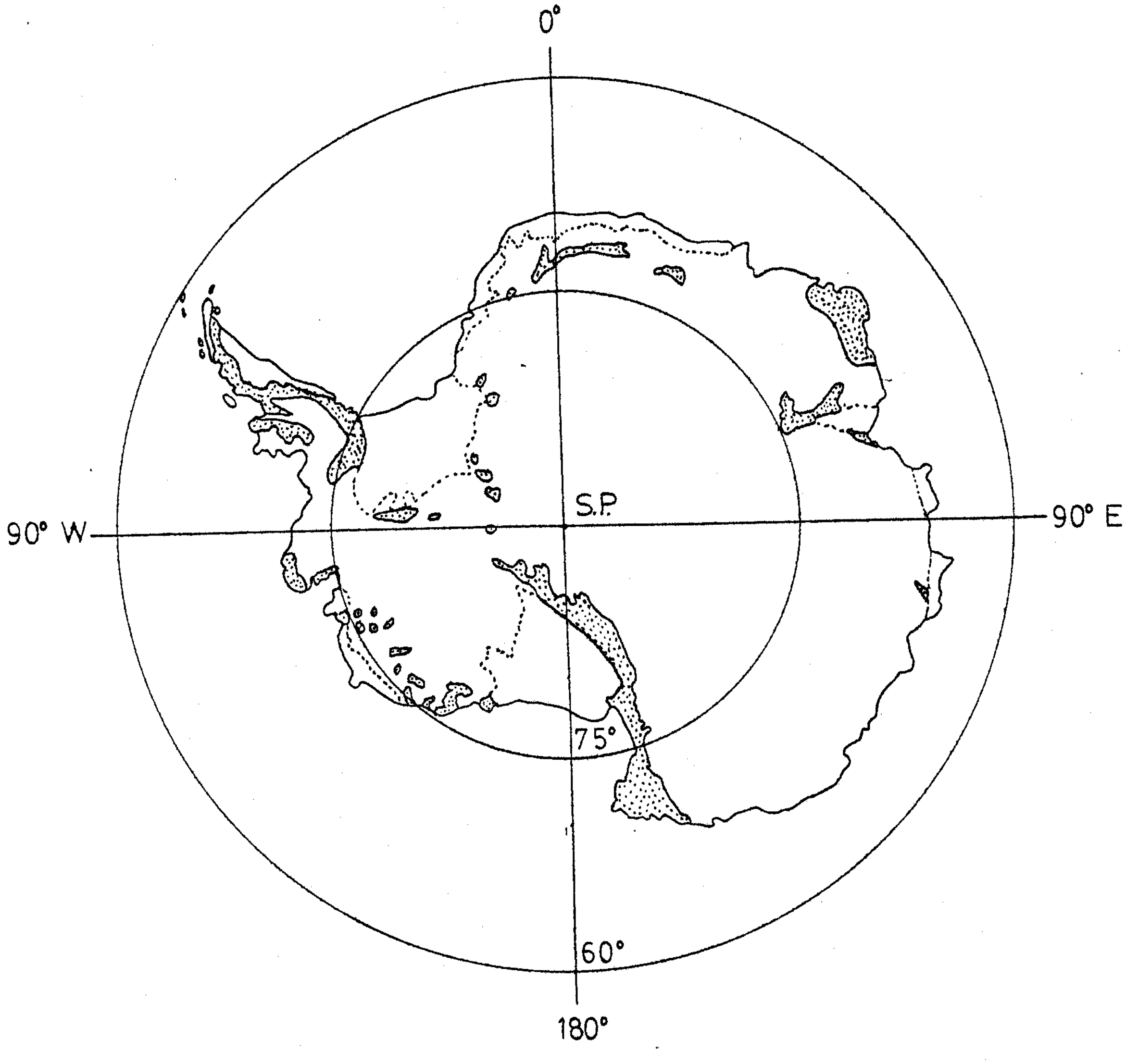


Figure 3.1. Present outcrop in Antarctica is designated by stippled areas. Shoreline, where it is not coincident with the ice cap boundary, is shown by a dashed line (modified from Elliot, 1975).

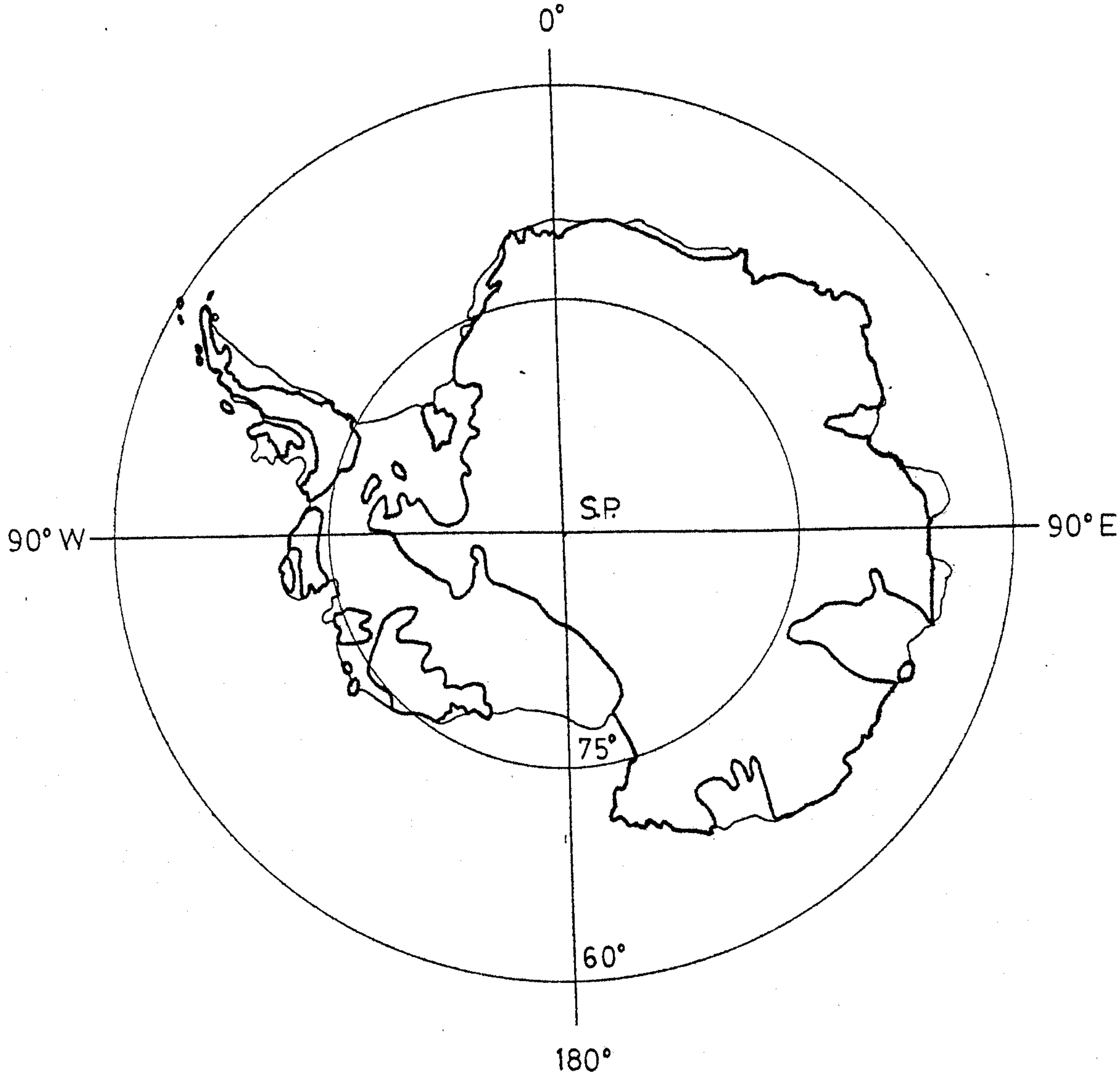


Figure 3.2. Land extent in Antarctica following isostatic adjustment if the ice cap melted entirely. The heavy line marks the boundary of land above sea level. The present outline of ice and land is shown by the narrow line (modified from Elliot, 1975).

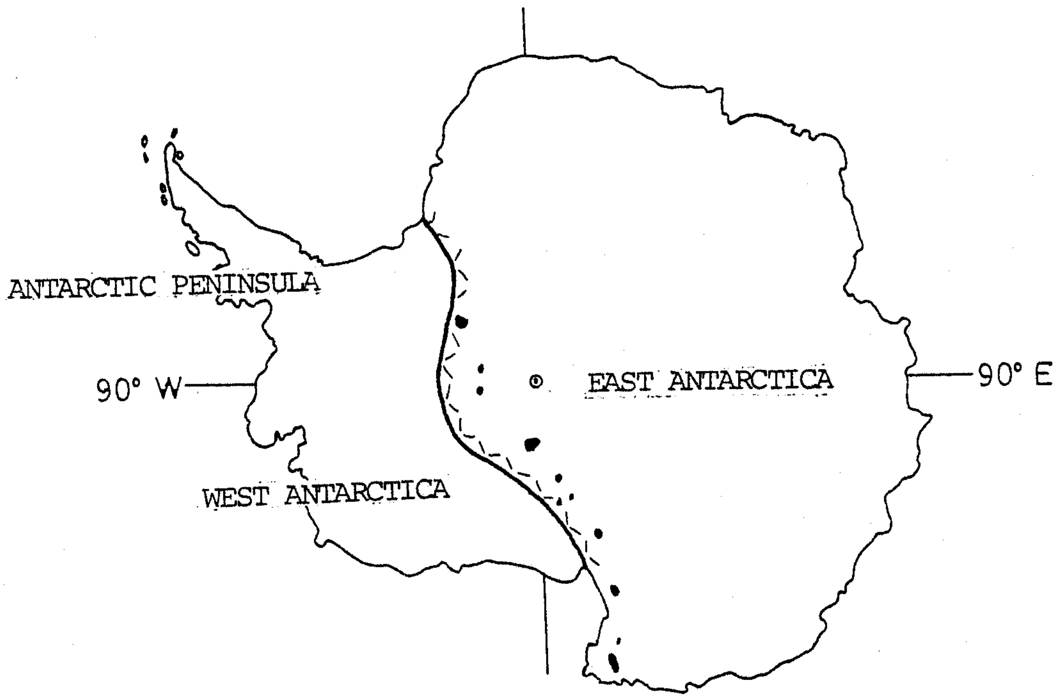


Figure 3.3. The major geographic divisions of Antarctica. The outcrop extent of the Ferrar Group, Jurassic mafic volcanics and intrusives, is shown in black. Their coincidence with the shield edge suggests attempted rifting along a weak older boundary.

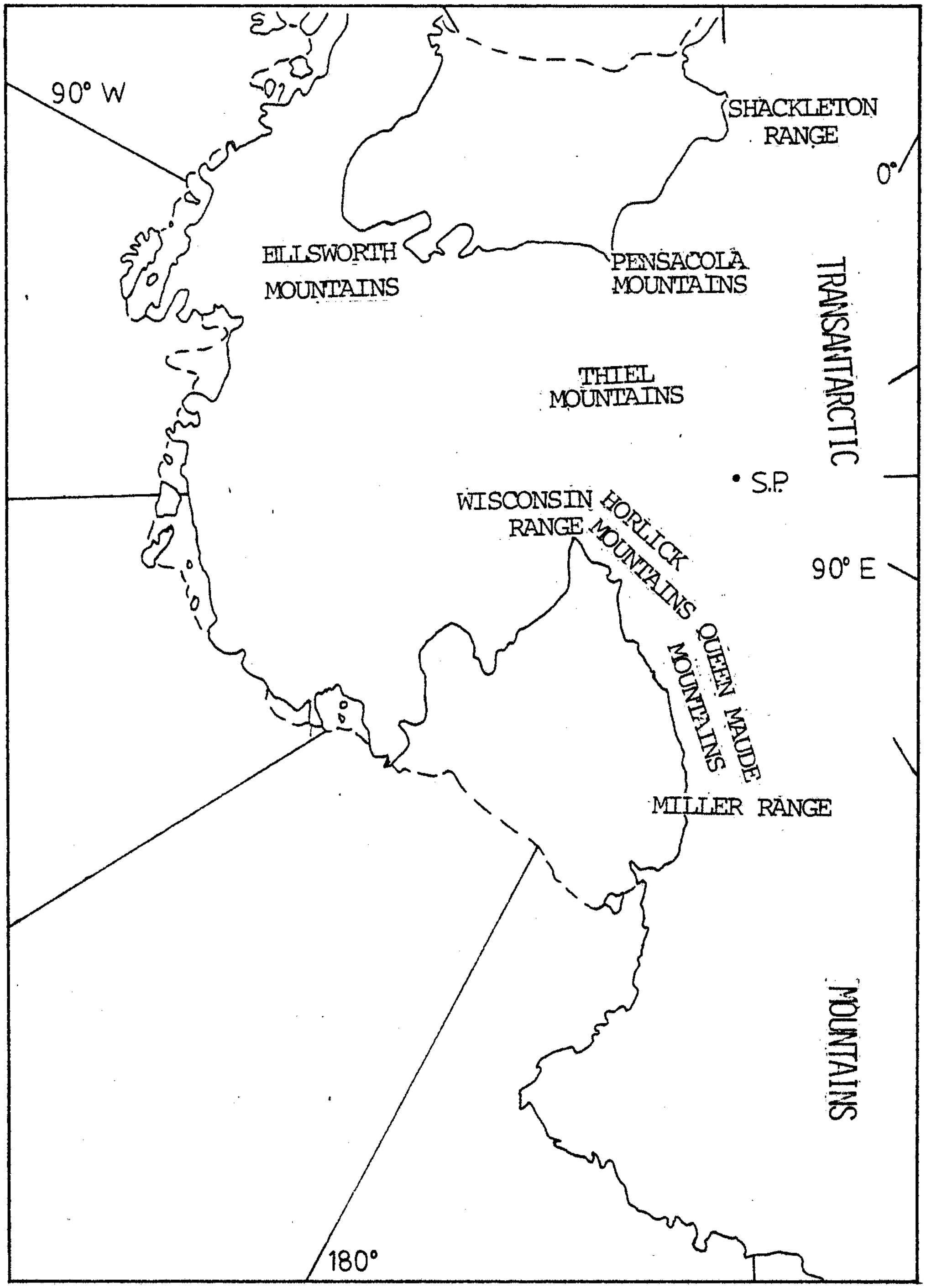
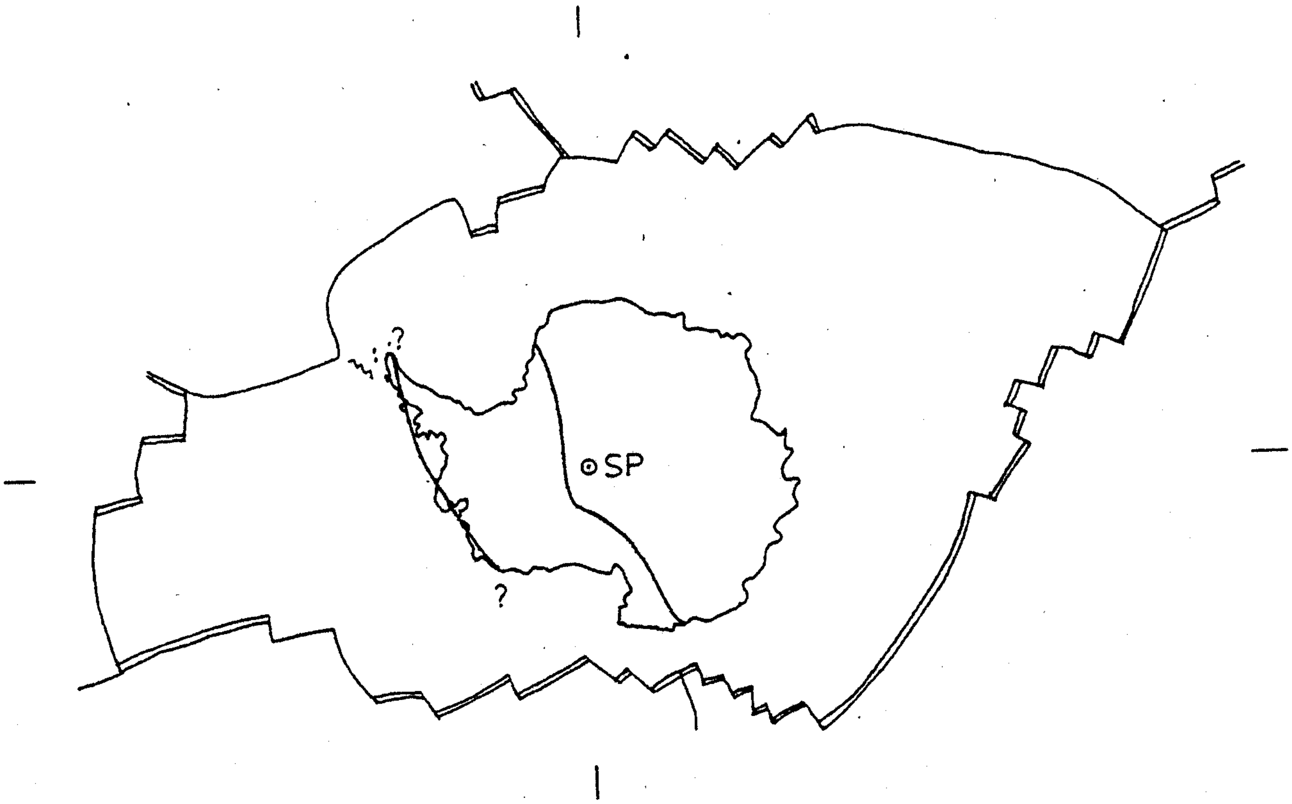


Figure 3.4. Location map for the mountains of West Antarctica and for the for the ranges that comprise the Transantarctic Mts.

continents (Figure 3.5).



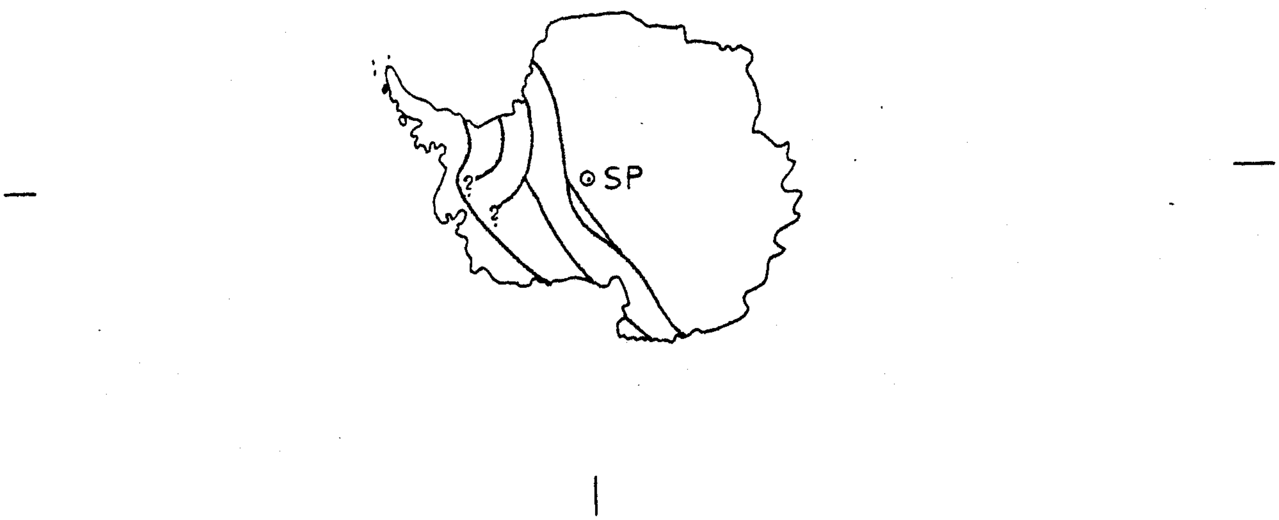


Figure 3.5. Maps of the rifts and sutures of Antarctica, both active and inactive.

3.2. RIFTS

3.2.1. ACTIVE RIFTS

Despite a nearly complete circuit of mid-ocean ridges (several segments of the plate boundary are transform faults), Antarctica is aseismic. This aseismicity perhaps results from the ice cap pressure which counteracts the effects of the boundary rifts so that there is no active internal rifting; (see 3.2.2.2. for a possible exception). Hamilton (1967) notes a similar situation in Greenland. All the active rifts of Antarctica are those bounding the plate and they are associated with the breakup of Gondwanaland (Figure 3.6).

3.2.1.1. Eocene Rifts

The plate boundary between Antarctica and Australia was the last of Antractica's margins to be rifted during the breakup of Gondwanaland. This occurred starting in the Eocene approximately 50 million years ago and resulted in the separation of Antarctica and Australia and the formation of the Southern Ocean (Weissel and Hayes, 1972) (see section 1.2.1.1.).

3.2.1.2. Cretaceous Rifts

Antarctica separated from India, Africa and South America in the Early Cretaceous, approximately 130 million years ago (Norton and Sclater, in press). This is the section of boundary rift which includes the two major transform fault segments (Figure 3.6).

3.2.2. INACTIVE RIFTS

3.2.2.1. Drake Passage Rift

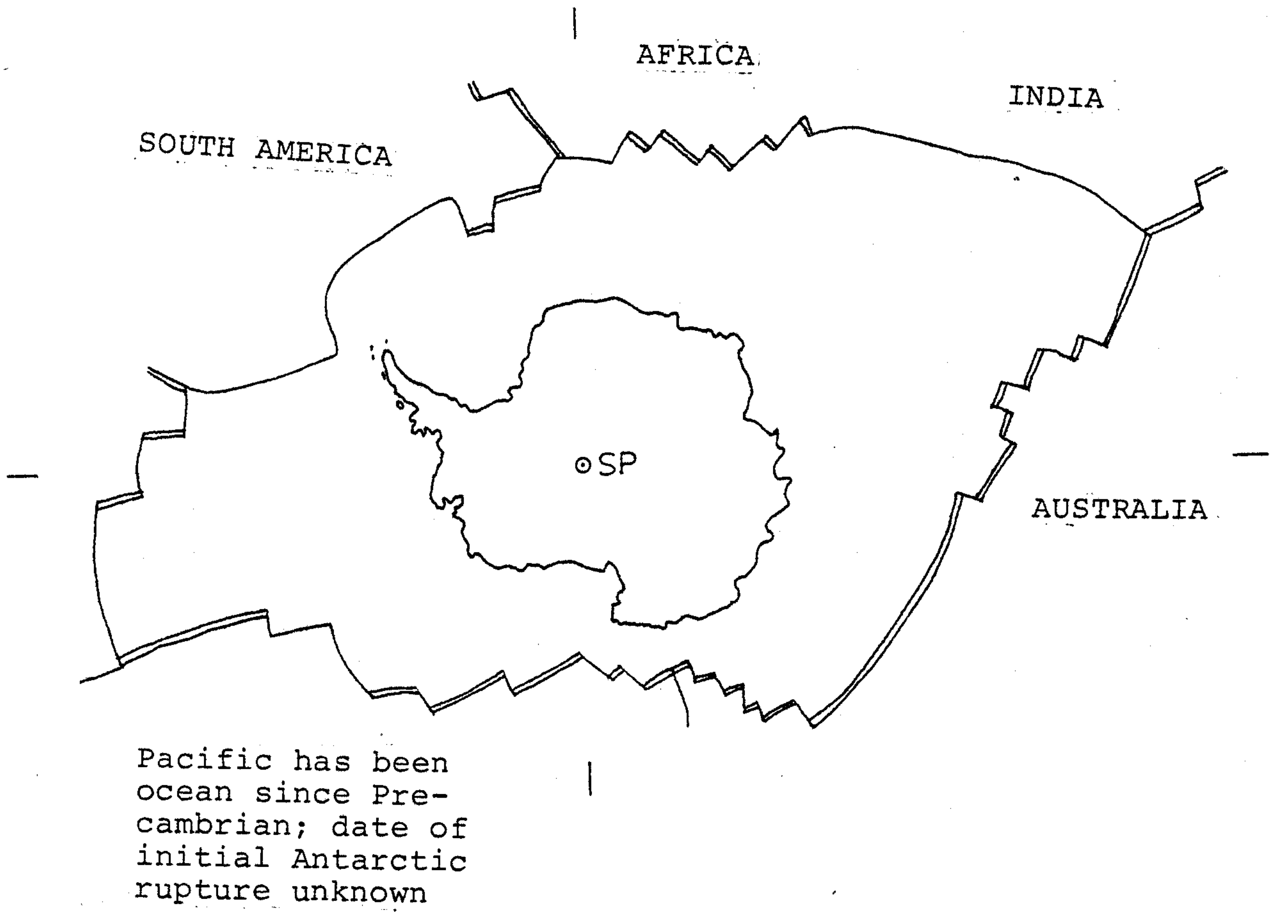


Figure 3.6. The plate boundaries of the Antarctic plate. All are essentially rifted segments except for major transforms located approximately 120-150 degrees west and east.

Although inactive today, rifting in the Drake Passage (Figure 3.7) is a very young phenomenon whose oldest magnetic anomaly is 20 million years or Miocene age (Barker, 1972). It is part of a plate configuration complexity in the Scotia Sea which is only beginning to be appreciated and understood (Dalziel and Elliot, 1972).

3.2.2.2. Antarctic Peninsula

Another possible, but less well defined example of Cenozoic (attempted?) rifting is marked by tholeiitic to alkaline volcanism in the Antarctic Peninsula and sections of West Antarctica (Figure 3.7). Some of these volcanics display mantle strontium isotope ratios (Jones and Walker, 1972). Despite the lack of seismicity noted earlier, this evidence could be considered as suggesting recently inactive or even current rifting in West Antarctica (Elliot, 1975).

3.2.2.3. Jurassic Rifts

The Pensacola and Queen Maude Mountains and sections of the Transantarctic Mountains bordering the Ross Ice Shelf and Sea (Figure 3.4) all have rocks of the Ferrar Group (Craddock, 1972). The Ferrar Group contains mafic volcanics and intrusives of Jurassic age, and they include basalt, diabase, and gabbro. The Dufek Massif, also part of the Group, is a layered gabbroic complex which comprises a large percentage of the exposed rock of the Pensacola Mountains. The exposures of the Ferrar Group coincide with the western boundary of the Precambrian shield area (Figure 3.3). This is a line of presumed older subduction and accretion and therefore a zone of probable weakness. Other internal or continental margin rifts, presumably resultant from

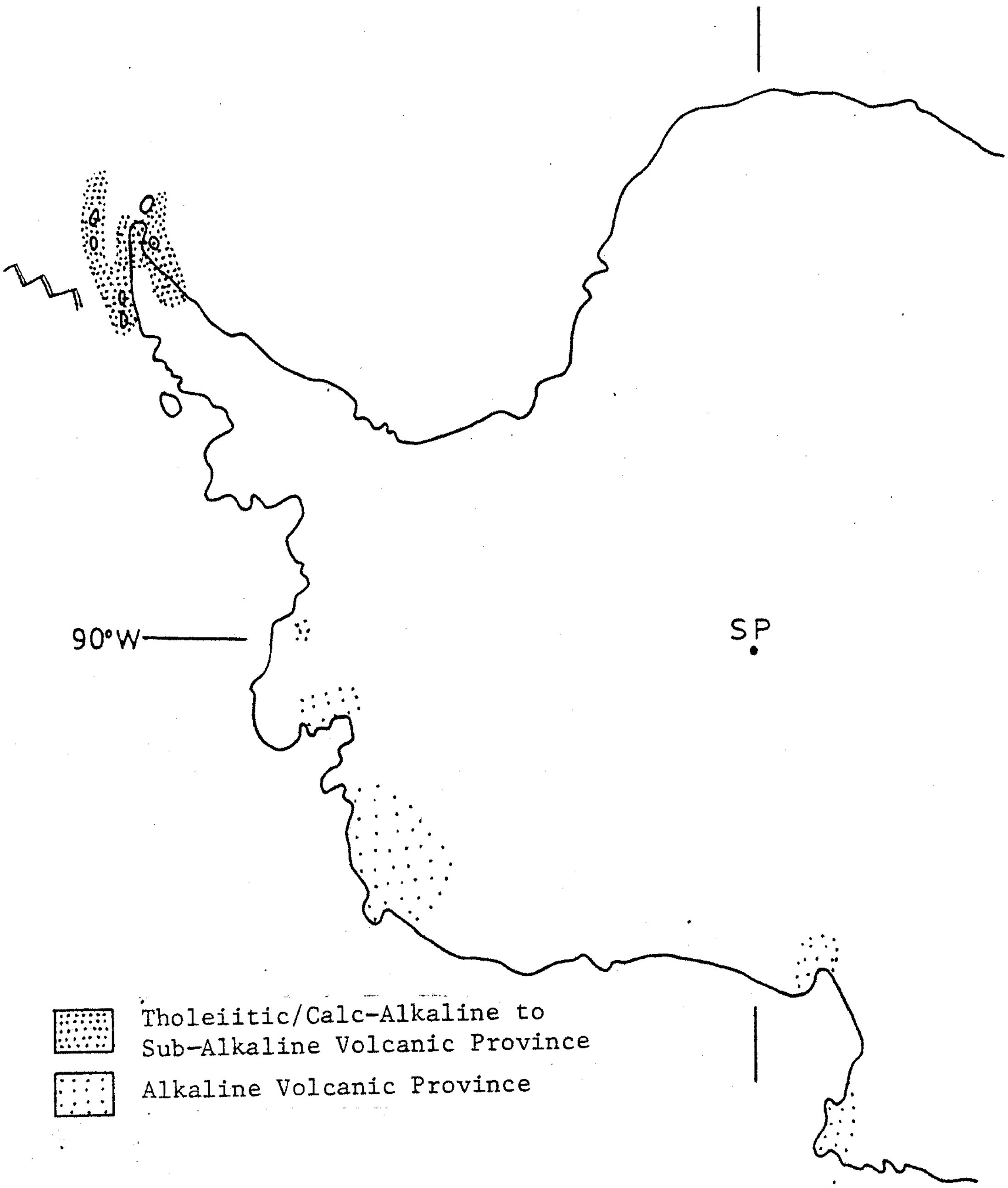


Figure 3.7. Cenozoic volcanism in West Antarctica and the Antarctic Peninsula suggests possible young or even active attempts at rifting. A rift active in the Cenozoic (8-20 my. ago) is present off the Antarctica Peninsula in the Drake Passage.

the Late Mesozoic-Early Cenozoic breakup of Gondwanaland, have not been identified in Antarctica as they have been elsewhere (see 1.2.2.1.).

3.2.2.3. Aulacogens

No aulacogens (failed rifts striking into orogenic belts) have been identified in Antarctica. However, this may only reflect a data scarcity as they are seen elsewhere in similar geologic settings (Burke and Dewey, 1973).

3.3. SUTURES

In other publications and in other sections of this report, sutures have been identified primarily by the presence of linear or curvi-linear bands of similar-age ultramafics which separate relatively cohesive blocks of rocks with different characters and ages. In some instances granitic rocks parallel the ultramafics (or ophiolites) suggesting arc or continent collision and accretion. Unfortunately, to date no ultramafic rocks have been discovered on Antarctica (Craddock, 1972), although some high grade metamorphosed ultramafic rocks have been studied on islands of the Scotia Arc (Dalziel and Elliot, 1972). Possible sutures have been picked as associated with a selection of orogenic belts which were accreted onto the Precambrian shield of Eastern Antarctica from Late Precambrian to Early Cenozoic (Figure 3.8). These belts were identified by calc-alkaline rocks and evidence of compressive deformation (Craddock, 1973). The number, ages and extents of these belts are disputed (Craddock, 1973; Elliot, 1975; DeWit, 1977). Stump (1973) believes in a long term compressive margin history, but contends that subduction was nearly continuous rather than episodic (Table 3.1).

3.3.1. ACTIVE SUTURES

There are no active trenches around the Antarctic continental margin at the present time. However, the South Sandwich Islands, located in the arcuate section of the Scotia Arc are seismically active (Figure 3.9) and mark a location of present day subduction of the American Plate.

3.3.2. INACTIVE SUTURES

	CA Intrusion	CA Volcanism	Compressional Deformation	Orogeny
Paleogene				Andean
Cretaceous	X	X	X	
Jurassic	X	X		
Triassic	X?	X	X	Gondwanian
Permian		X		
Carboniferous				Borchgrevink
Devonian	X	X		
Silurian			X?	
Ordovician	X		X	Ross
Cambrian		X		
Late Precambrian	X	X	X	Beardmore

Table 3.1. These indicators of a compressive margin have been identified in Antarctic rocks of the ages indicated. Data compiled from Stump (1973) and Elliot (1975).

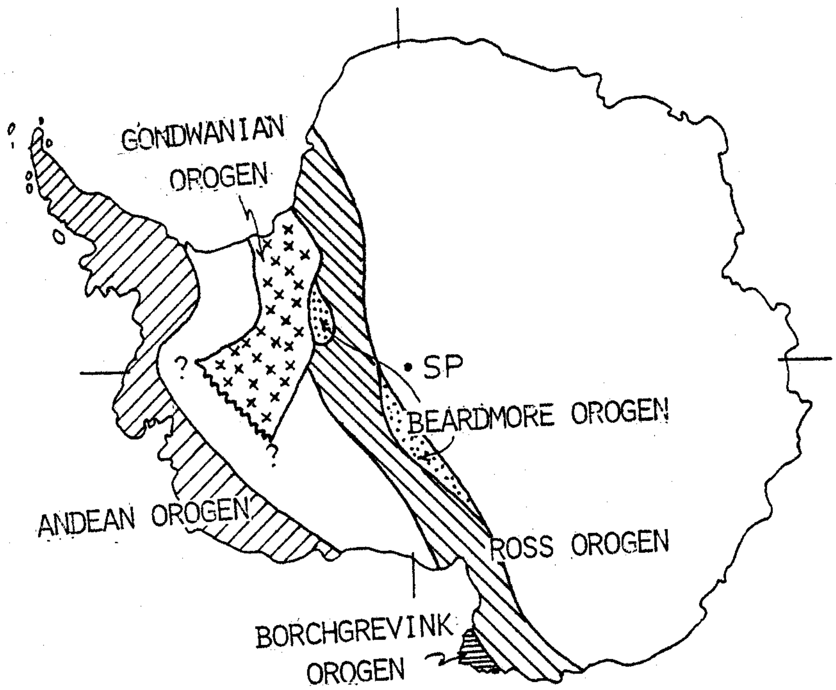


Figure 3.8. Orogenic belts of Antarctica. These may represent arcs and/or microcontinents which were accreted onto the East Antarctic shield. Although the boundaries of the orogens are poorly defined, suturing appears to become generally younger to the west.

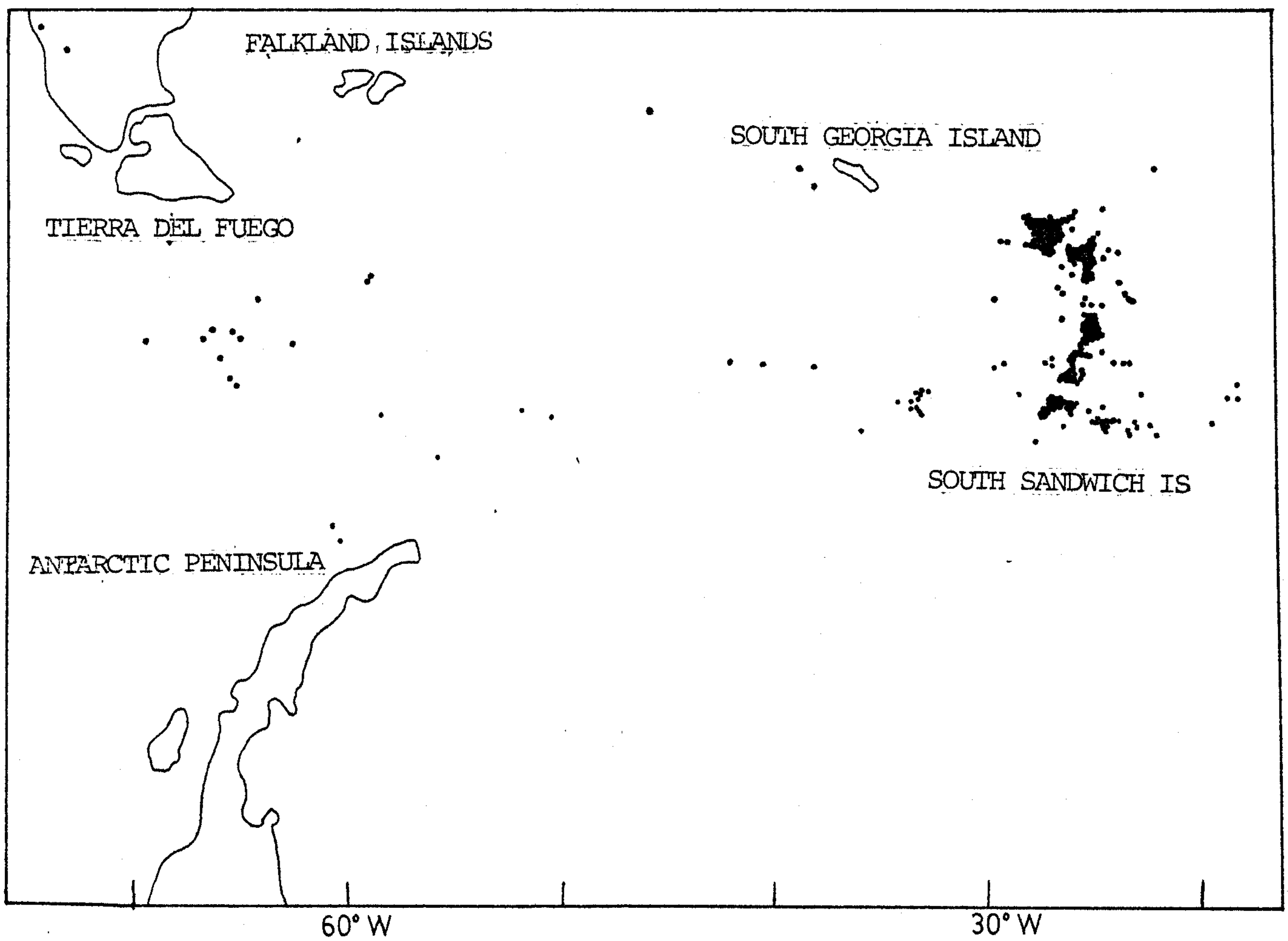


Figure 3.9. Seismicity in the Scotia Arc connecting the Antarctic Peninsula with southern South America. Earthquakes occur most heavily in and around the South Sandwich Islands which mark the location of the South Sandwich trench (modified from Dalziel and Elliot, 1975).

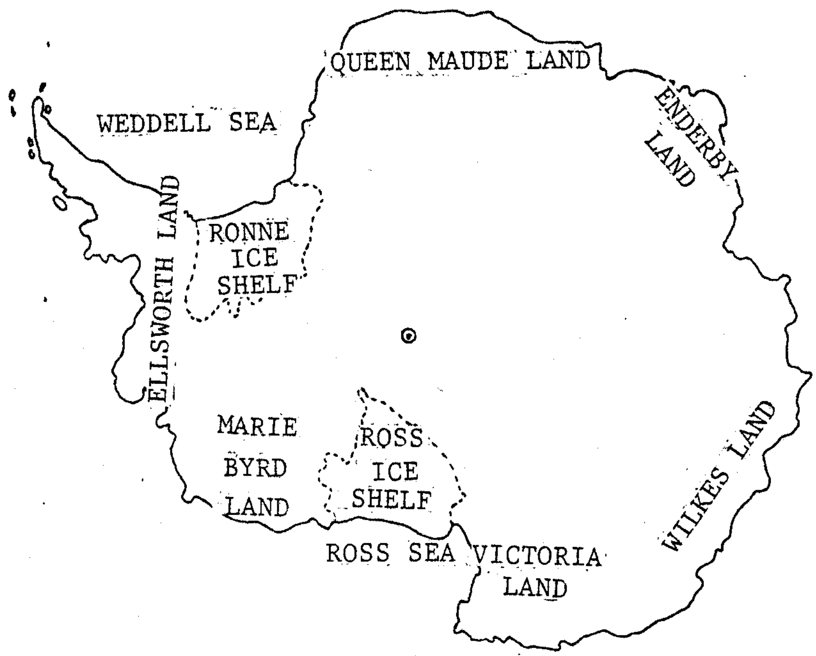


Figure 3.10. Location map for the "Lands" of Antarctica.

3.3.2.1. Andean Orogen (Late Mesozoic-Early Cenozoic)

Direct evidence of the Andean Orogeny (calc-alkaline plutonism and volcanism, shallow water marine sediments) is seen in Marie Byrd Land, Ellsworth Land, and sections of the Antarctic Peninsula (Figure 3.11). Indirect evidence (uplift and erosion) has been identified throughout the Transantarctic Mountains. "Andean" granitic plutons have been dated in Marie Byrd Land (88-166 m.y.), Ellsworth Land (102-109 m.y.), and the northern Antarctic Peninsula (45-110 m.y.). Related calc-alkaline volcanism strongly suggests active subduction in an "Andean style" despite the lack of deformed strata. Suarez (1974) locates this subduction zone seaward of the Antarctic Peninsula on the basis of sedimentary distribution. There is no continent or arc material to the west and it is therefore dubious that this period of subduction culminated in an actual suture.

3.3.2.2. Gondwanian Orogen (Early Mesozoic)

The Gondwanian Orogeny is dated in Marie Byrd Land and the Antarctic Peninsula from granitic intrusions and low grade regional metamorphism. In addition deformation evidence is recorded in the Pensacola and Ellsworth Mountains, and Upper Paleozoic-Lower Mesozoic shallow marine sediments are found in the Transantarctic Mountains from Victoria Land to the Horlick Mountains (Figure 3.12).

3.3.2.3. Borchgrevink Orogen (Middle Paleozoic)

Evidence of a Mid-Paleozoic orogeny is exposed only in Victoria Land and Marie Byrd Land with deformation, granitic intrusions and amphibolite grade regional metamorphism (Figure 3.13). However, other

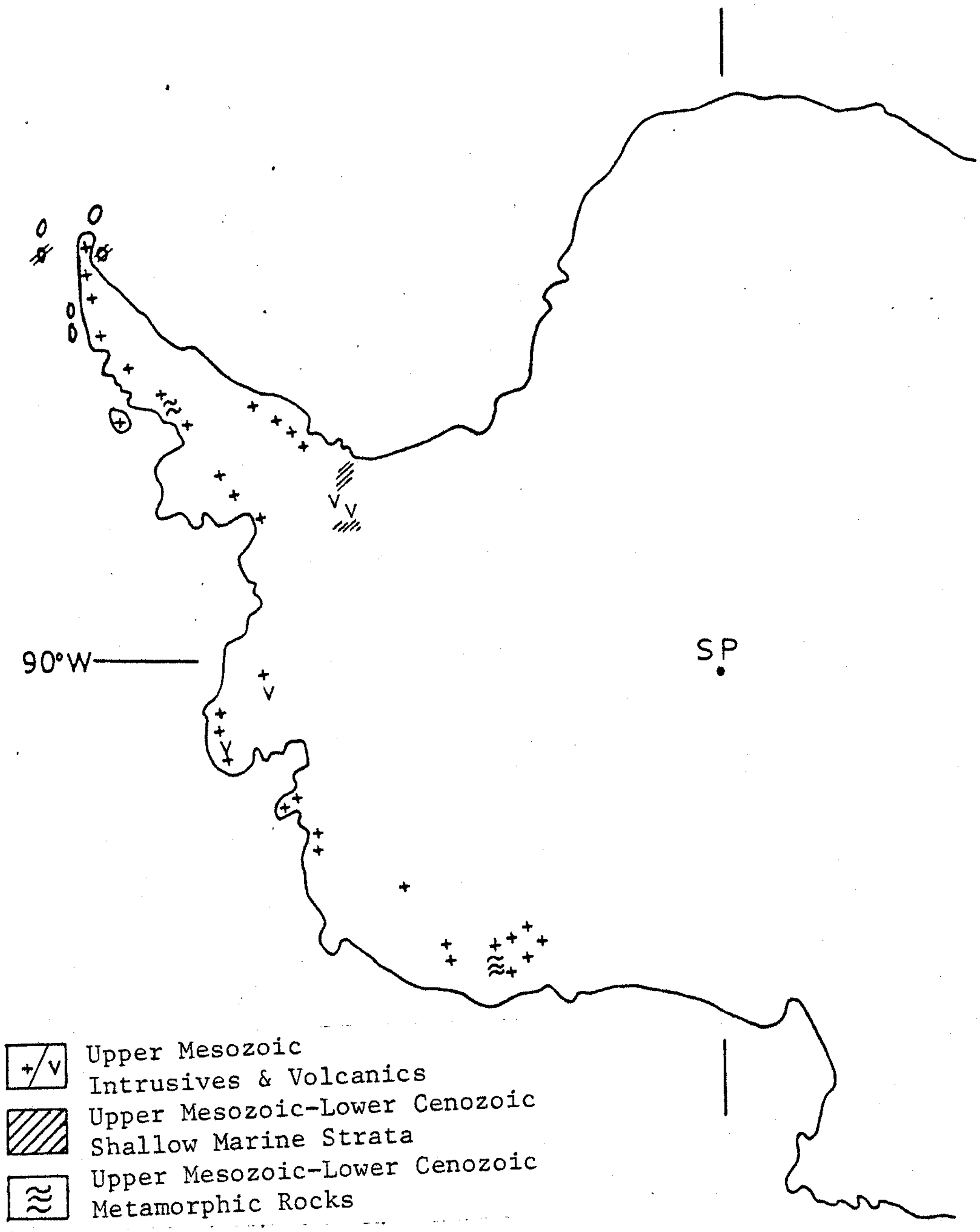


Figure 3.11. Geographic distribution of evidence for the Andean Orogeny involving convergent Andean arc phenomena in the Late Mesozoic and Early Cenozoic.

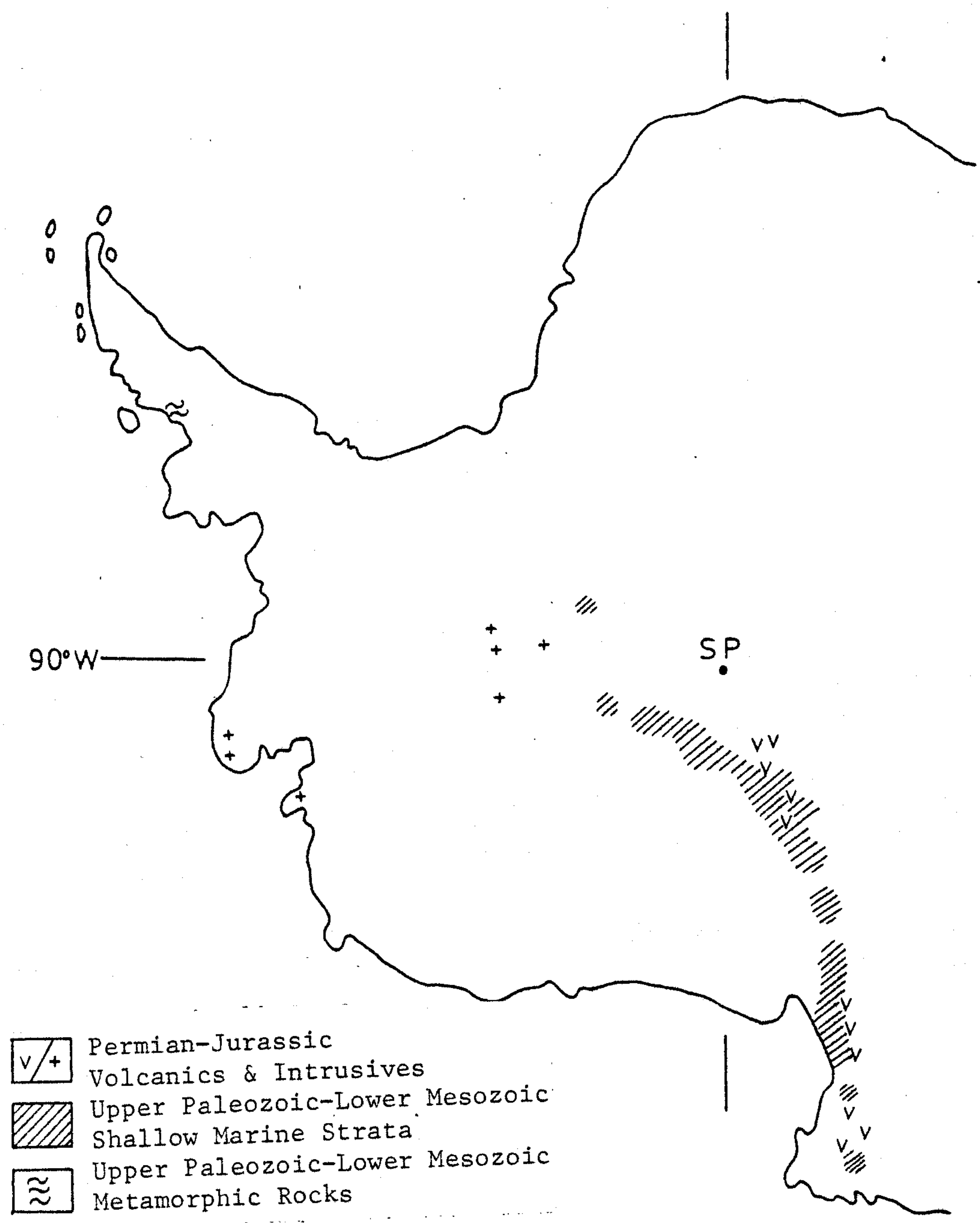


Figure 3.12. Geographic distribution of evidence for the Gondwanian Orogeny involving convergent motion of an Andean type in the Late Paleozoic and Early Mesozoic.

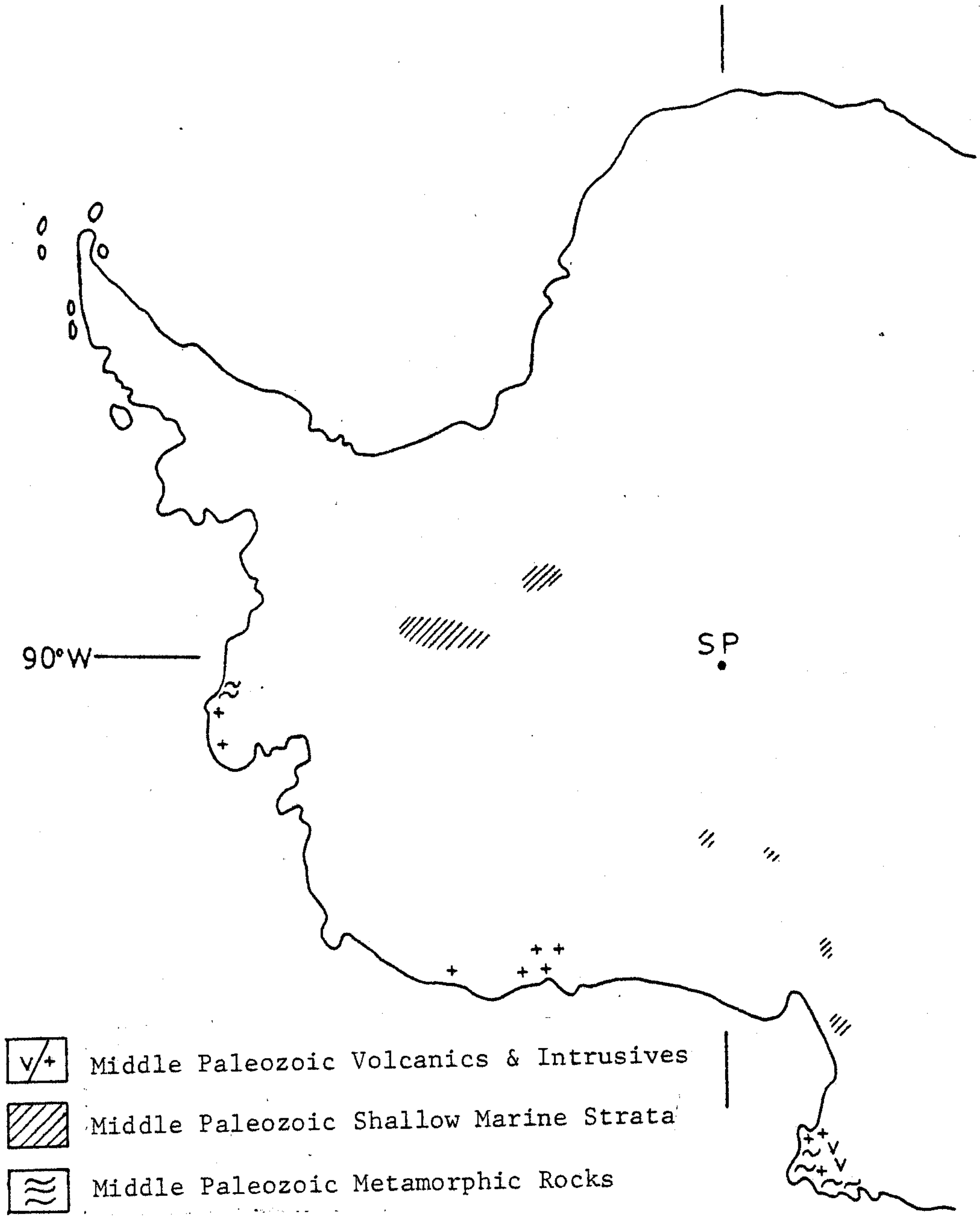


Figure 3.13. Geographic distribution of evidence for the Borchgrevink Orogeny. Radiometric dates give an age range of 300-375 m.y.

dates of a Mid-Paleozoic event are widely scattered in West Antarctica, and shallow marine sediments outcrop in a number of locations in the Transantarctic and Ellsworth Mountains.

3.3.2.4. Ross Orogen (Early Paleozoic)

A 500 m.y. thermal event is recorded across much of the East Antarctic basement (Elliot, 1975). Deformation, low to medium grade metamorphism, and granitic intrusions mark the Transantarctic Mountains from Victoria Land to the Pensacola Mountains. Lower Paleozoic shallow marine sediments also appear sporadically along this mountain belt (Figure 3.14). Elliot (1975) assumes collision with a micro-continent rather than an arc because of the absence of volcanic detritus in the pelitic sediments.

3.3.2.5. Beardmore Orogen (Late Precambrian)

The extent of evidence for the Beardmore Orogeny is geographically almost equivalent to that for the slightly younger Ross Orogeny (Figures 3.14 and 3.15). In a few places, e.g. Victoria Land, the Ross and the Beardmore have not yet been positively distinguished. The Beardmore did not produce appreciable volcanic or regional metamorphism except for felsic pillow basalts in the Pensacola Mountains (Elliot, 1975). The general absence of volcanic detritus favors an Atlantic-type margin much like the somewhat later Ross Orogeny. This is perhaps indicative of the rupture event forming the Pacific Ocean and comparable to the Belt rocks of North America (Burke, Dewey and Kidd, 1977).

No greenstone belts suggestive of sutures (Burke, Dewey and Kidd,

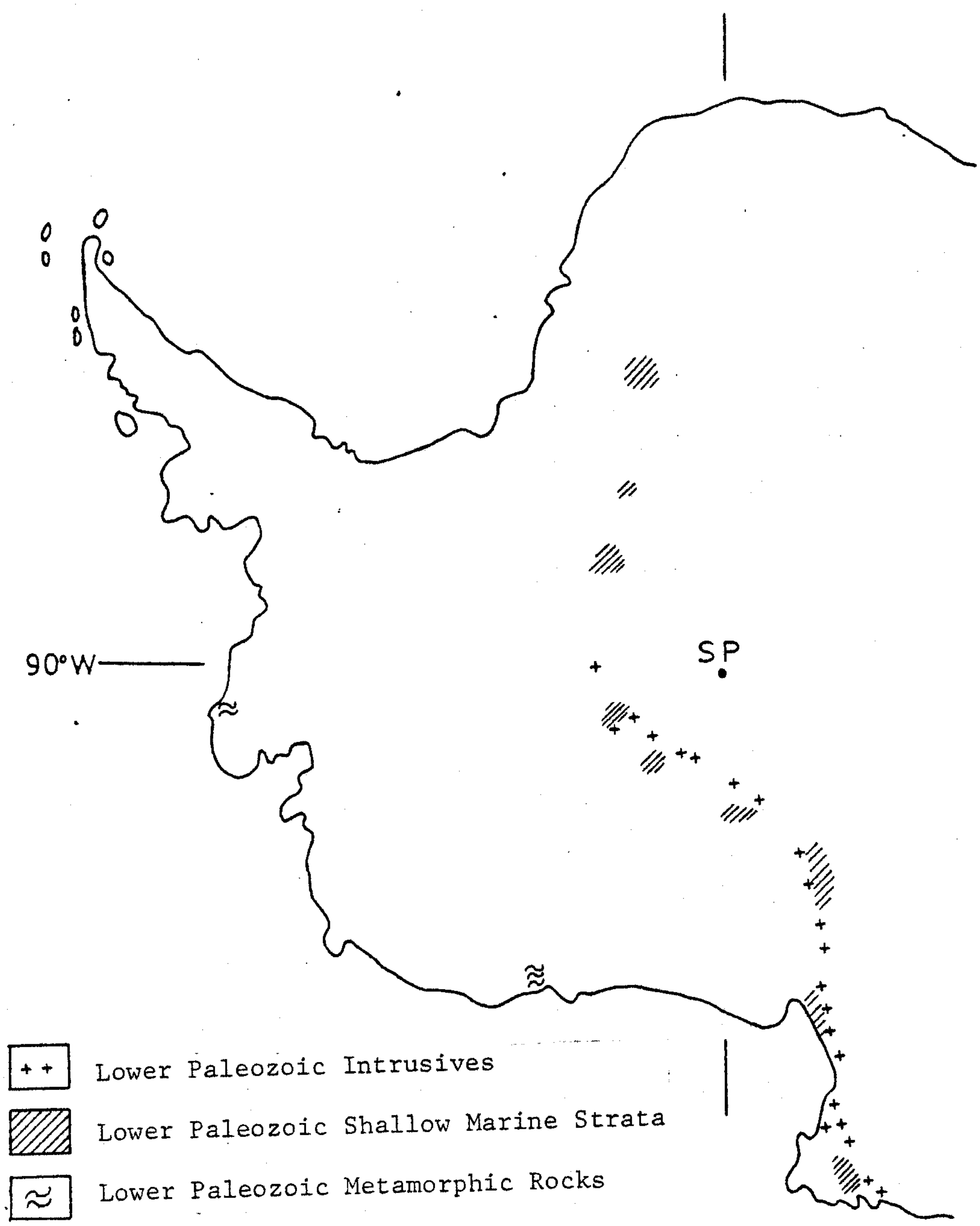


Figure 3.14. Geographic distribution of evidence for the Ross Orogeny involving convergent motion of an Atlantic type margin. Radiometric dates give an age range of 425-525 m.y.

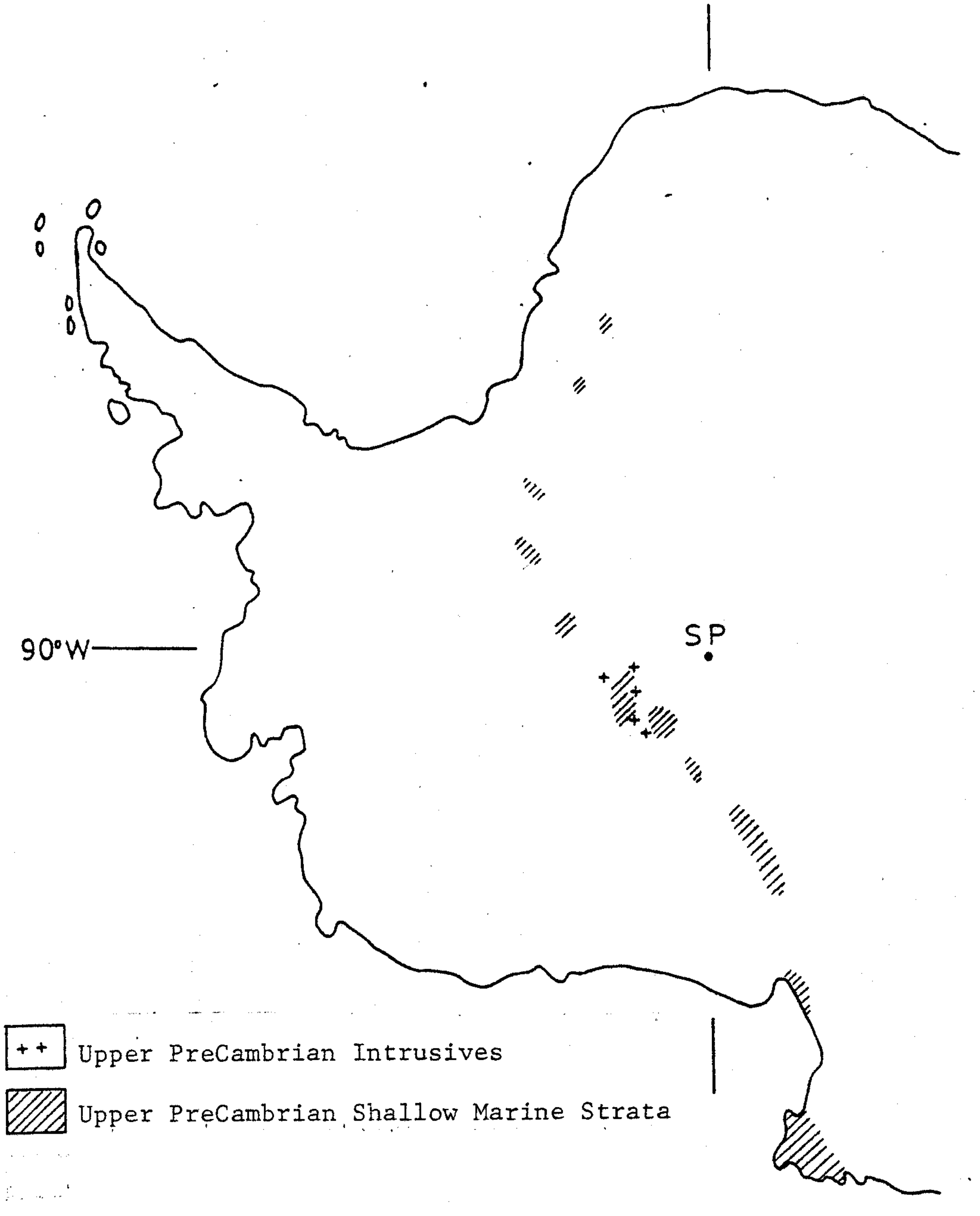


Figure 3.15. Geographic distribution of evidence for the Beardmore Orogeny involving convergent motion of an Atlantic type margin in the Late Precambrian.

1976) have yet been identified in the East Antarctic shield. This is not surprising when one considers that it is almost entirely covered with a thick ice cap. Because mapping of other similar shields (Australia, Asia, North America) has shown the existence of these belts, it is a reasonable speculation, although impossible to prove that they exist in Antarctica also. The Nimrod Orogeny, dated 1-2 b.y. old, is marked by deformation of some older Precambrian rocks (Ford, 1972).

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4.0. ASIA

4.1. INTRODUCTION

Asia has convergent plate boundary zones along its eastern and southern sides; the extreme complexity that can arise in the geology of such convergent zones in a short time is well illustrated by the Phillippine-New Guinea-Borneo area (Dewey, 1976). The Mesozoic and Paleozoic geology of the Malayan-Burman-Tibetan area, and of the areas between the Siberian platform west to the Urals, and south to the Tien Shan (West Siberian Plain, Kazakhstan, and Mongolia) appears to be the result of similar processes and has given rise to similar or worse complexity. The limited and often ambiguous information available on these areas means that the analyses of Mesozoic and Paleozoic sutures of these areas is tentative. It may be that the Central Asian area of Paleozoic accreted material is better treated for satellite gravity and magnetic correlation purposes as a block distinct from the shield areas to north and south, rather than attempting to look for the effect of one (out of many) sutural zones within it. Data on Asian rifts are more abundant and less ambiguous and the analysis of these features is therefore less tentative.

4.2. RIFTS (Figure 4.4.)

4.2.1. ACTIVE RIFTS

4.2.1.1. Intracontinental Active Rifts

Molnar and Tapponnier (1975) suggested that all three of the areas of active rifting described below are connected directly to the collision of India and Asia. Although the Yunnan graben (geographic features are located on Figure 4.1) are essentially on the margin of the collision zone, the other rifts are not, and the Baikal rifts are not in any obvious way linked physically to the collision zone by a continuous chain of fault features. It is not clear from the collision hypothesis why the rifts should have developed where they have. It is possible that the stresses transmitted from the collision zone are opportunistically exploiting areas where the lithosphere had already been thinned and perhaps rifts developed, prior to the collision. Alternatively, it is possible that one or both of these rift zones have no particular connection with the collision, representing instead a hotspot generated rift (Baikal) and the initiation (for other reasons) of a plate boundary between China plus Southeast Asia and the remainder of Asia (Ordos-Shansi rifts).

4.2.1.1.1. Ordos-Shansi-Gulf of Peking

In this area a complex of en echelon, mostly north-south trending graben are connected by east-west strike-slip faults and fault belts. Rifts are inferred to exist under the thick alluvial sediments of the North China Plain from sediment thickness data shown by Alverson et al. (1966). Only very minor vulcanism is found closely associated with

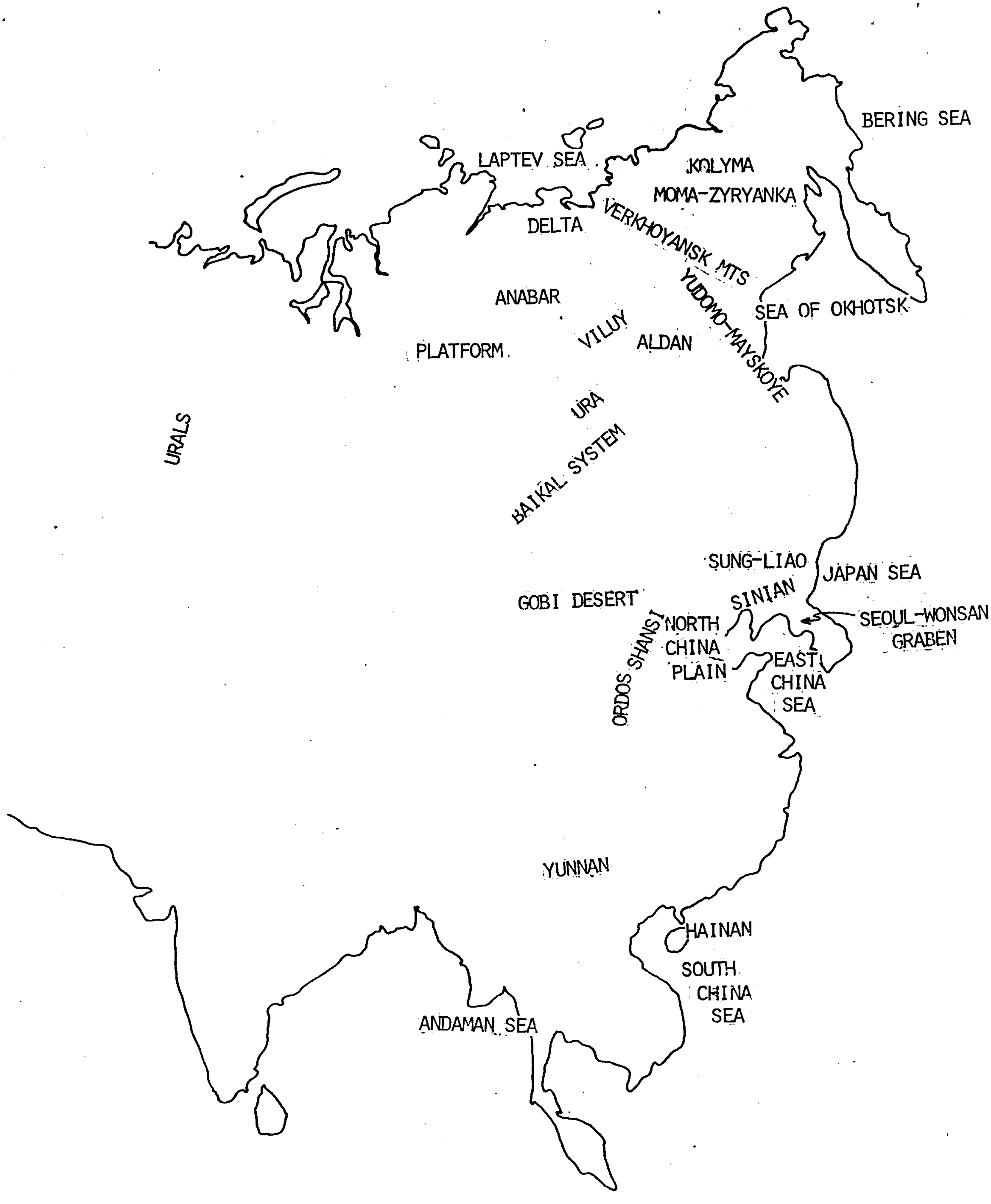


Figure 4.1. Location map of geographic features related to rifts in Asia north of the Himalaya.

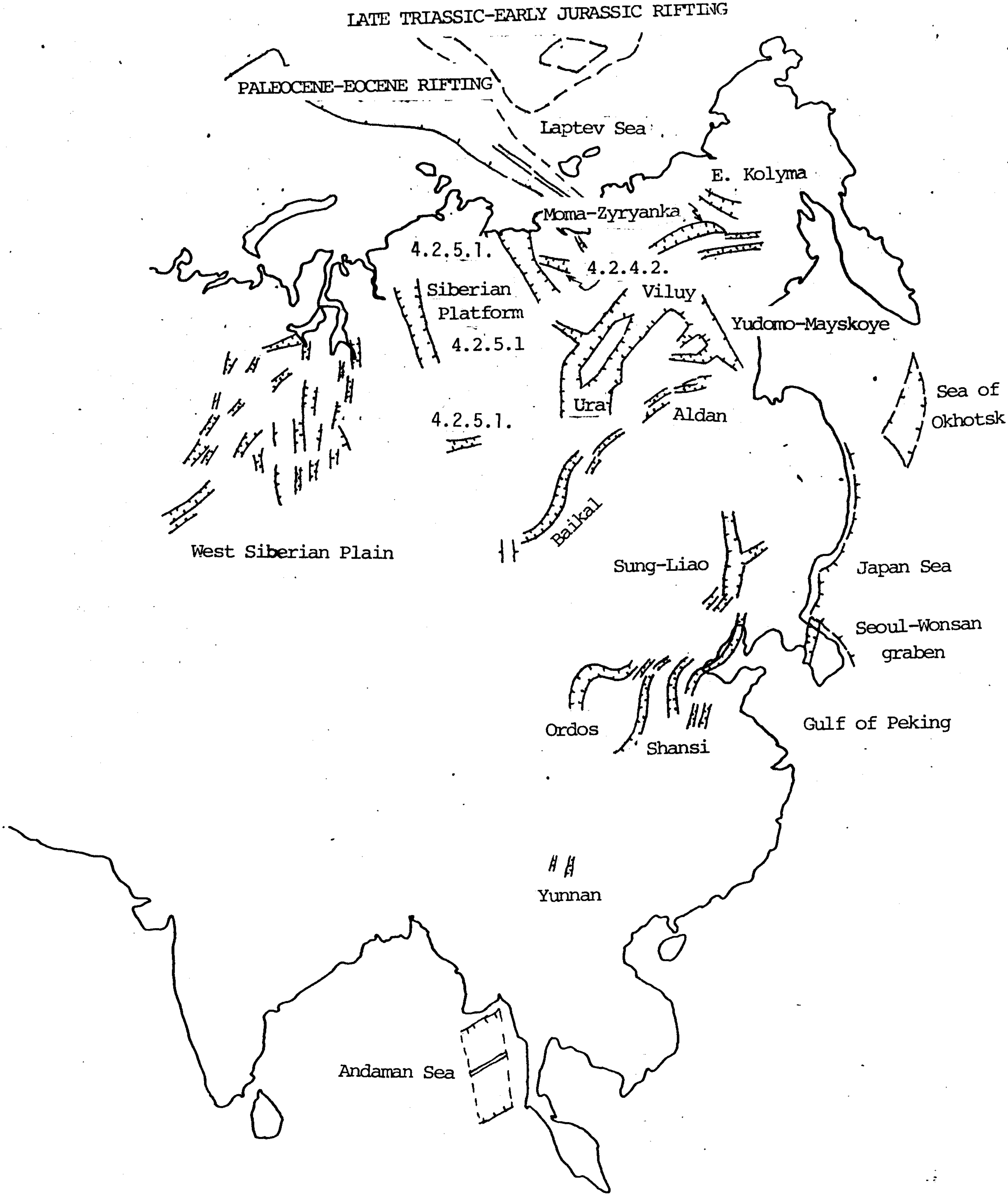


Figure 4.4. Rifts of Asia north of the Himalaya

these rifts. Sediment thicknesses are large (4 km) under the North China Plain, but are less thick in more western areas.

4.2.1.1.2. Baikal System

The well known active rift of Lake Baikal is accompanied by several smaller <u>en echelon</u> rifts and faults found mainly to the northeast and east of the lake itself. Most of the structures trend north-northeast to northeast. The rifts are all fairly narrow, contain minor thicknesses of sediment, and do not have any known active volcanism occurring within them, although some does occur nearby in one area.

4.2.1.1.3. Yunnan

Two small north-trending rifts occur in this area.

4.2.1.2. Active Oceanic Rifts

4.2.1.2.1. Lena Delta - Laptev Sea

The Gakkel ridge in the Arctic Ocean is an actively spreading oceanic rift. It strikes directly into the Siberian continental shelf in the region of the Laptev Sea, and, extrapolated, it would pass a short distance east of the Lena delta. The relative motion between Eurasia and North America is such that the pole of rotation is located on land in the area of the Verkhoyansk Mountains (Pitman and Talwani, 1972); therefore, divergent motion since the Early Eocene in the Arctic Ocean has been sufficient to form a considerable width of oceanic crust while the much lesser amount of divergence near the rotation pole has yet to lead to the formation of ocean crust. Two very small north-south trending Cenozoic rifts are known just south of the delta of the Lena River

(Alverson, et al., 1966) and are clearly related to this zone of extension. The continental shelf area of the Laptev Sea is poorly known, and we predict that it contains very deep, sediment-plugged rifts; these are likely to contain much basalt as flows, sills and dikes; some oceanic crust may be present.

4.2.1.2.2. Andaman Sea

The active spreading rift in the Andaman Sea (Curray and Moore, 1975) appears to be within an active rear-arc marginal basin in the sense of Karig (1971). The rifted margins of the Andaman Sea are probably not older than Early Miocene (25 m.y.). None of the other such marginal basins along the eastern periphery of Asia is known to be spreading.

4.2.2. INACTIVE CENOZOIC RIFTS

4.2.2.1. Seoul - Wonsan Graben and Japan Sea

This graben (Burke, 1977) forms one arm of a three-armed rrr junction (Burke and Dewey, 1973) with the other two arms now being the rifted western margin of the Japan Sea. The Japan Sea started to open in the Early Miocene (about 20-25 m.y. ago) as a rear-arc marginal basin (Sillitoe, 1977).

4.2.2.2. Sea of Okhotsk

The small, triangular, deep oceanic area at the southern end of the Sea of Okhotsk is likely (Karig, 1971) to be a Neogene (but not actively spreading) rear-arc marginal basin with rifted margins. The origin of the remainder of the Sea of Okhotsk is obscure; post-Cretaceous rifting may well have played a part in its formation, but no firm data are known to confirm this.

Of the other East Asian marginal basins, the Bering Sea and the East China Sea are not known to have been formed by rifting; the South China Sea is not known to have been formed by rifting, although it is possible that the Vietnamese margin south of Hainan is a rifted margin because the Mesozoic volcanic belt of East China stops south of Hainan.

4.2.2.3. Gobi Desert

A few, small, shallow Cenozoic rift structures occur in the area of the Gobi Desert according to Alverson, et al., (1966), and are not active at present according to Molnar and Tapponnier (1975). They are not plotted on the map because they are very small, discontinuous, and of irregular orientation.

4.2.2.4. Western Siberian Arctic Continental Margin

This margin was formed by rifting in the Eocene (~60 m.y.) (Herron, et al., 1974), sea floor spreading subsequently carrying away the Lomonosov Ridge, a continental fragment broken at this time from this margin. The position of this rift may have been defined previously by Triassic rifting, as is seen between Norway and Greenland. This is suggested because of the Triassic basalt that is abundant on Franz Josef Land and the islands near Spitzbergen, and from the rectilinear continuity of the very narrow Lomonosov Ridge continental fragment.

4.2.3. MESOZOIC RIFTS

4.2.3.1. Aldan Shield

Small, shallow, irregular but generally east-west trending rifts of

Late Jurassic to Early Cretaceous age occur in the Aldan shield (Alverson, et al., 1966).

4.2.3.2. Moma - Zyryanka Rift

This rift, in the western part of the Kolyma block, is a substantial, long structure with up to 4 km of sediments in it over a significant length. Churkin (1972) suggested that it is an active rift, connected with the plate boundary in the Arctic Ocean. Alverson et al., (1966), however, show it as a Jurassic-Cretaceous structure, a clearly preferable interpretation since it contains no sediments younger than this age (Geologic Map of Asia, 1976).

4.2.3.3. Eastern Siberian Arctic Continental Margin

This margin, running east from the Laptev Sea, was formed by rifting from the Canadian Arctic margin about 180 m.y. ago (Early Jurassic) (Herron, et al., 1974). The subsequent sea floor spreading formed the western Arctic Ocean.

4.2.3.4. West Siberian Plain

Many small to medium-sized Triassic rifts of moderate to poor continuity exist buried under an average of 3 km of sediments over much of the West Siberia Plain (Rudkevich, 1976). These rifts may have up to 3 km of sediments and basalts in them. They are of the same age as the Siberian (or Tunguska) Traps, which occur in the area of the Siberian Platform adjoining the east side of the West Siberian Plain, and which are the largest preserved flood basalt occurrence on earth. The rifts have a general north-south trend, although there are exceptions, and there appears to be a divergence to northeast and southeast trends in the southern part of the province. These rifts have been compared (Milanovsky, oral. comm., 1977) to the Basin and Range rifts of the western United States. The comparison is not convincing since the volcanics of the West Siberian rifts are reported to be basaltic, unlike the bimodal volcanics of the Basin and Range. Also, the rifts formed after the Uralian Ocean had closed (Hamilton, 1970) unlike the Basin and Range rifts which have formed while the adjacent continental margin was open to ocean and involved in convergent or transform plate tectonics. It is noteworthy that this episode of widespread rifting and voluminous flood basalt volcanism did not result in any rift developing into an ocean.

4.2.3.5. Sung - Liao (Harbin) Rift

Isopachs plotted by Alverson et al. (1966) indicate that this structure is an example of a completely failed three-armed rift, with subsidiary, smaller rifts to the southwest. It is most likely to be Early Triassic in age, although it might have been initiated in the Permian. It is unusual as it formed within continental crust that had only shortly before been consolidated by orogeny. The main rift, trending north-northeast, contains a large (~5 km) thickness of sediments. In addition, later extensive subsidence over a broad basinal area centered on this rift suggests that two or perhaps all three arms of the rift have a large basaltic axial dike in them.

4.2.4. PALEOZOIC RIFTS

4.2.4.1. Viluy Rifts

This twin rift system, of substantial width, and containing a very thick section of sediments (8 km ±), is an aulacogen with respect to the Verkhoyansk-Chersky orogenic belt, the site of an oceanic tract closed by Mid-Cretaceous. The age of the Viluy rifts must be the time when the eastern side of the Siberian Platform was defined by rifting and started evolving into a miogeoclinal (aseismic) continental margin. There appears to be disagreement over this age; Kosygin and Parfenov (1975) say that it (and other aulacogens of the Siberian Platform) are "Late Precambrian" in age, which appears to mean anything between 1500 and 600 m.y. ago. However, Bazanov, et al., (1976) state that it formed in the "Middle Paleozoic", and that basaltic dikes and sills are commonly found cutting rocks up to this age on the margins of, and within, the rifts. Reports of Devonian gypsum and salt from the rift and further north along the Verkoyansk belt (Sborshchikov and Natapov, 1968) strongly suggest that this margin and rift are Devonian structures. The large amount of subsidence centered on the area of these rifts (Viluy "syneclise"), which continued long after rift formation, suggests that both rifts possess a substantial basaltic axial dike.

4.2.4.2. Northeast Siberian Platform

A restricted zone of salt domes striking at a high angle to the northern Verkhoyansk fold belt (Sborschchikov and Natapov, 1968) is probably the expression of a small buried aulacogen of Devonian age.

4.2.4.3. Eastern Kolyma

East of the Kolyma uplift in the eastern Kolyma block, a north-south oriented rift (Alverson et al., 1966) is indicated by the involution of the fold trends of the rocks to the south of the stable Kolyma block. This rift is probably of Devonian age, but could be older (Geologic Map of Asia, 1975).

4.2.4.4. Yudomo - Mayskoye

This north-south zone of mild folding and westward thrusting (Alverson et al., 1966) strikes north from the Sea of Okhotsk and widens into the Verkoyansk fold belt. It lies between the southeastern part of the Siberian Platform and the Okhotsk stable block. This zone may be a suture (Section 4.3.2.1.), but alternatively it may be one of the rare cases where a rift has "succeeded" to the extent of forming a small (50-200 km) width of oceanic crust before motion across it stops. Later convergence removes some of this oceanic crust by subduction producing mild folding and faulting in the sedimentary fill of the "proto-ocean". Salop and Scheinmann (1968) show this as a questionable aulacogen, but it is clearly more than just a rift. If it is the kind of "open and shut case" indicated above, the age of rifting is probably not younger than Early Cambrian and may be much older. This age is suggested on the basis of the age of sedimentary rocks exposed by the thrusting on the western side of this structure.

4.2.5. PRECAMBRIAN RIFTS

4.2.5.1. Rifts of the Siberian Platform

Apart from the Devonian (Sections 4.2.4.1. and 4.2.4.2.) and possible Cambrian (Section 4.2.4.4.), rifting along the eastern margin of the

Siberian Platform, all other aulacogens within it and its other three margins are thought to have been formed by rifting at the same time in the Proterozoic. Kosygin and Parfenov (1975) give data that suggest the age of rifting was about 1500-1550 m.y. ago. Three of the aulacogens of this group are wide and deep, major structures. These are the Ura aulacogen (between Lake Baikal and the Viluy area), and the two that run north-south and flank the Anabar Shield, opening toward the northern margin of the Platform. All three of these aulacogens are likely to contain basaltic axial dikes because they have localised significant subsidence long after their formation as rifts. In addition, the Ura aulacogen has had a major influence on the position of the margin of folded rocks in the Baikalian miogeocline, resulting in a large curved re-entrant in the fold front, like the Viluy and Coppermine aulacogens (q.v.). The Ura aulacogen passes directly along strike into the Viluy aulacogen, and its position clearly controlled the position of the latter. Information on these rift structures has also been obtained from Alverson et al., 1966; Salop and Scheinmann, 1969; and Bazanov et al., 1976.

4.2.5.2. Sinian Rifts

A zone 1000 km long averaging 100 km wide of miogeoclinal sedimentary rocks of Sinian age runs from about 100 km west of Peking through the Gulf of Liaotung area and on to the central part of the North Korea - China border. These Proterozoic rocks are, in places, folded and faulted, although it is known that original thicknesses of over 4-5 km are preserved in places (Alverson et al., 1966). It is possible that this zone represents one or more Proterozoic rifts (aulacogens). Because they lie in a zone of Cenozoic active tectonics, because they are somewhat internally deformed, and because of fairly extensive younger cover, the rifts cannot be precisely defined and are therefore not shown on the map.

4.3. SUTURES

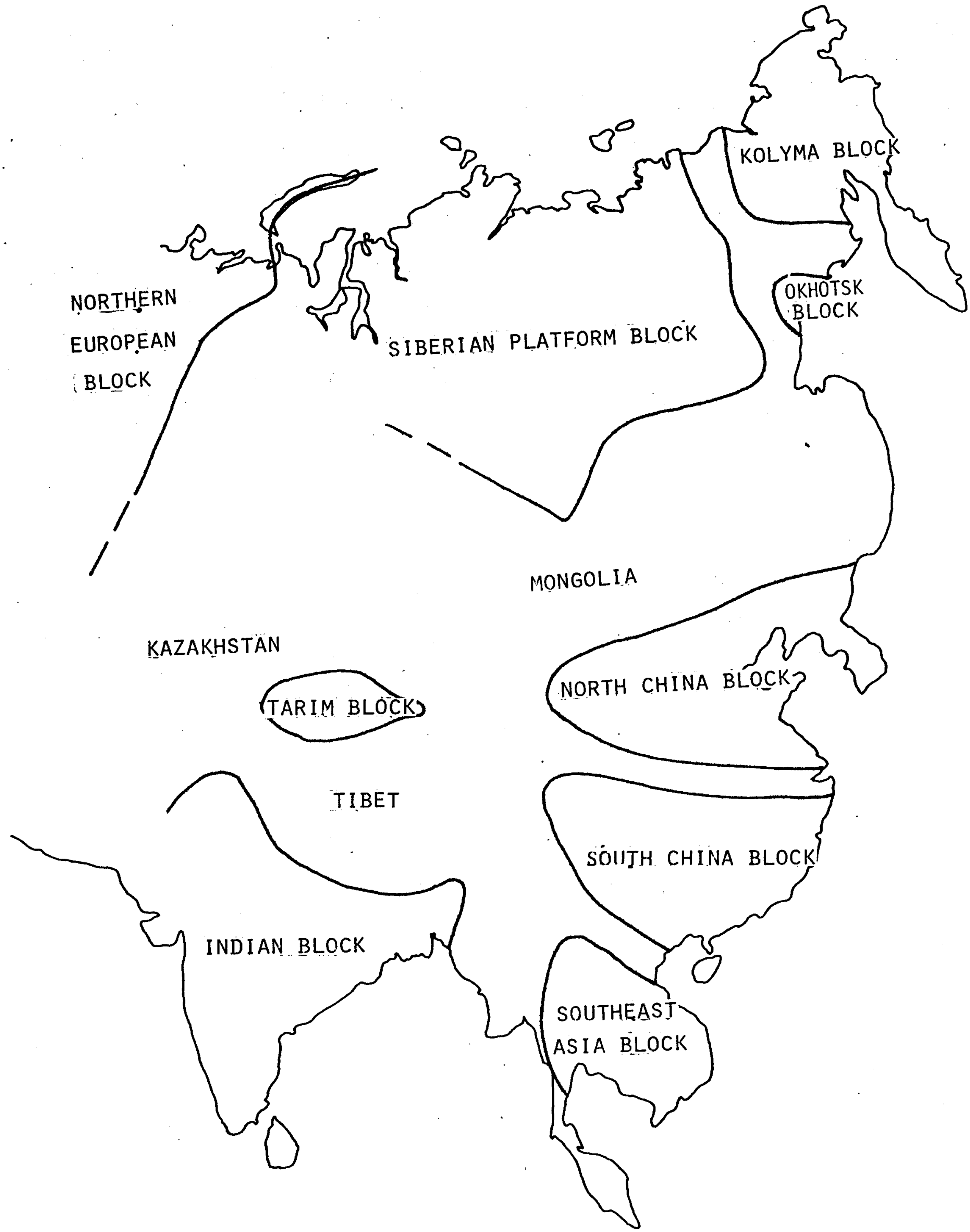
Asia is the complex product of collisional processes occurring from Proterozoic through Cenozoic times. Unfortunately, at present the geometry and timing of these events are poorly understood. In this report we lean heavily on Burrett's (1974) analysis augmented by our own study of the Chinese (1975) Geologic Map of Asia, Terman's (1973) Tectonic Map of China and Mongolia, and a perusal of the Russian literature.

Figure 4.2. shows the approximate outlines of the major continental blocks that assembled to form the present Asian landmass. Broadly speaking, the areas separating these blocks mark the major suture zones within the Asian continent. Many, if not all of these; however, are themselves the product of protracted arc accretion events, and hence contain a number of "lesser" sutures predating the time of final closure in each area. For the most part this earlier activity is beyond our resolution. In Figure 4.3. all of the belts of roughly age-equivalent ophiolites in Asia that we have so far been able to define have been plotted. This serves to illustrate at least some of the possible complexity.

4.3.1. CENOZOIC SUTURES

4.3.1.1. Himalaya

The impingement of India against Tibet that formed the Himalaya is the only Cenozoic collision recorded in Asia. This suture is marked by a well defined ophiolite belt running from the Pamir along the Indus Suture and down through Burma.



'igure 4.2. Stable blocks with known or presumed pre-Upper Proterozoic ontinental basement.

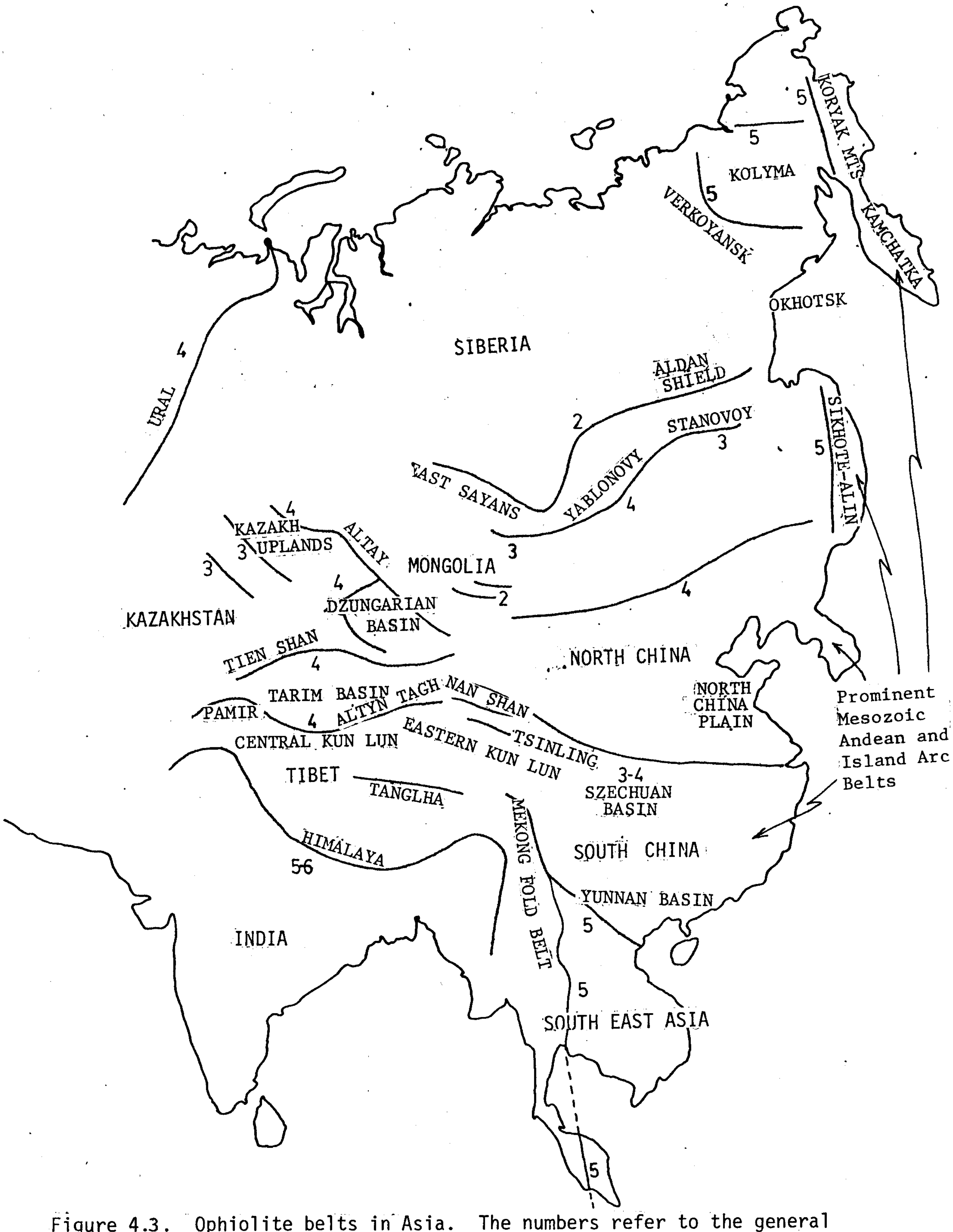


Figure 4.3. Ophiolite belts in Asia. The numbers refer to the general age of ophiolites, not necessarily the time of suturing: 2 - Proterozoic, 3 - Lower Paleozoic, 4 - Upper Paleozoic, 5 - Mesozoic, 6 - Cenozoic.

4.3.1.2. Indonesia

Complex current pole motions have as yet produced no modern sutures except for arc collisions between Timor and continental Australia and northern and southern New Guinea (also part of continental Australia). Active rifting secondary to convergence is going on east of Borneo. Older microcontinental objects have been sutured onto Indonesian arcs (Hamilton, 1977, especially Figures 2 and 3).

4.3.2. MESOZOIC SUTURES

4.3.2.1. Northeast Asia

The Kolyma block collided with the Siberian Platform in the Late Mesozoic forming the 600 km wide Verkoyansk range on the Siberian side of the suture (Herron et al., 1974). This is the widest foreland fold-thrust belt recognized in the world. The youngest sediments involved in the folding in the Verkoyansk are of Jurassic-Cretaceous(?) age, and granites[1] associated with the suture are also dated Late Mesozoic, suggesting a Cretaceous age for this event.

In the Okhotsk region the Verkoyansk structures split to envelop a relatively small stable area with Archean rocks outcropping at its center. This probably represents a continental fragment which was sutured either to Kolyma or Siberia shortly before they collided.

On the eastern side of the Kolyma block there is a diffuse belt of Late Mesozoic ultramafics. These occur within an accretionary melange

[1] In this section we use the term granite in its most general sense. Although we would like to distinguish between true granite and granodiorite, at present the data does not allow us to do so.

wedge which was formed during underthrusting of the Kamchatka-Koryak region in the Mesozoic (Scholl et al., 1977). This belt still faces open ocean and hence does not yet represent a continental suture zone

The South Anyui Trough (Seslavinsky, 1970) is a suture zone that also closed in Early Cretaceous time, and it represents the break between the Kolyma block and the Chukotsk block. Seslavinsky gives an unusually clear description that permits identification of the volcanic arc to the northeast, succeeded to the southwest by the ophiolitic melange and flysch trough (sutural zone) which is thrust over platform (miogeoclinal) sediments of the Kolyma block.

4.3.2.2. China

We locate the suture between North and South China along the ophiolite belt in the eastern Tsinling. To the west, the suture probably coincides with the continuation of this belt in the western Tsinling and Nan Shan ranges. However, data from this area is exceedingly sparse and we note the possibility that the western extension of this feature may instead lie within the thrust belt bordering the western margin of the Szechwan basin. A few ophiolitic bodies are reported from this area also (Chinese Geological Map of Asia, 1975). To the east of the Tsinling, the position of the suture becomes obscure under sediments of the North China Plain. Burrett (1974) notes that the main folding event in the eastern Tien Shan, Nan Shan and Tsinling is of Yenshanian (E. Jur.?) age, as are the majority of granites in the Tsinling. He also notes that in the eastern Tsinling tightly folded Lower Triassic flysch is overlain unconformably by Middle Jurassic terrestrial sediments. Together these data suggest an Early Jurassic age for this event.

4.3.2.3. Southeast Asia

A probable Triassic suture exists between Southeast Asia and South China in the ophiolite melange belt paralleling the Red River in Vietnam, and northwestward along the west side of the Yunnan basin in the Mekong Foldbelt. Throughout this area the main deformation is of Carnian-Norian age (Burrett, 1973).

Another suture of almost exactly the same age seems to exist between the Indochina block and the Burman block. Izokh and Tien (1964) describe pre-Jurassic, post Early Triassic ultramafic tectonic emplacement in the "Dien Bien Phu" zone. The Geologic Map of China (1975) contains indications that this zone may contain several sutures of about this age between various volcanic arcs. It is suggested that this sutural zone narrows to the south and crosses Malaya, where a zone of ophiolites and flysch was deformed and emplaced in the Triassic (Gobbet and Hutchinson, 1973).

4.3.2.4. Amur Region

In the Amur region there is a well defined north-south trending belt of ultramafic bodies running along the west side of the Sikote-Alin mountains. Previously, we interpreted this zone as a Precambrian suture (Burke, Dewey, and Kidd, 1977); however, mafic volcanics, and marine sediments associated with these bodies are of Upper Paleozoic and Mesozoic age; also the parallel belt of granites within the Sikhote-Alin are dominantly of Middle-Upper Mesozoic age (Chinese Geologic Map of Asia, 1975). We therefore suggest that this is, in fact, a Jurassic or Cretaceous suture, probably formed by an arc collision.

4.3.2.5. Tibet

Chang and Zeng (1973) indicate that there is a Mesozoic arc suture within the Tibetan Plateau. They note that the metamorphic ages increase progressively as one proceeds north across the Himalaya, Kangar Tesi, Tanglha, Kun Lun and Altyn Tagh ranges. In particular, the dates in the Tanglha mountains range from 107-210 m.y. suggesting the possibility of an arc collision during the Middle Mesozoic. Blueschists and serpentinites have been collected on geologic traverses across Tibet indicating that there may indeed be one or more sutures within the basement of the plateau (Henning, 1915). Alternatively, Şengör and Kidd (1978) suggest that the entire plateau is underlain by accretionary melange of Paleozoic-Lower Mesozoic age, produced during the closure of Tethys. If so, there are no sutures between the Kun Lun and the Himalaya. At present, we are unable to choose conclusively between these two hypotheses. We therefore plot a tentative suture of Middle Mesozoic age along the Tanglha.

4.3.3. PALEOZOIC SUTURES

4.3.3.1. Tarim Block

There is a probable upper Paleozoic suture separating northern Tibet from the Tarim block. This boundary is marked by an ophiolite belt associated with Upper Paleozoic granite that runs along the northern Pamir, central Kun Lun and Altyn Tagh ranges. It should be noted, however, that the published age of ophiolites in the Altyn Tagh is Proterozoic (Burrett, 1974). Hence, there is the possibility that this is, in fact, a Precambrian belt that was later remobilized during the

Upper Paleozoic, possibly as a result of subduction along the southern margin of the Tarim block.

The Carboniferous-Early Permian suture between Tarim and Kazakhstan runs along the southern Tien Shan. Norin (1937, 1941) has described ophiolites, blueschists, and colored melange from this area. Burrett (1974) suggests that at least some of these ophiolites were obducted in the Early Carboniferous; however, he also notes that deformation within the Tien Shan does not culminate until Early Permian. We take the latter date for the time of final suturing.

4.3.3.2. Kazakhstan

Kazakhstan is sutured to Siberia along an ophiolite melange belt bordering on the southwest margin of the Altay. In the northwest this corresponds to the Irytsh Crush Zone. Map relations indicate that this belt extends southwestward along the northern side of the Dzungarian basin and into western Mongolia, where relations become obscure. To the northwest the Irytsh disappears under sediments of the West Siberian Plain, and hence its relationship to the Ural suture is unknown. The Irytsh corresponds to the faunal boundary between Kazakhstan and Siberia until Upper Carboniferous, at which time the faunas in these two areas become indistinguishable (Ziegler et al., 1977). Burrett (1974) also notes that the main deformation north of the Dzungarian basin is Carboniferous, and that by Early Permian continental sedimentation dominates in the Altay, all indicating a Late Carboniferous date for collision.

Although we have so far treated Kazakhstan as a single entity during the Paleozoic, this is an oversimplification. Our compilation of ophiolite belts in central Asia suggests that there are at least two sutures of Lower-Middle Paleozoic age in the Kazakh Uplands, and another of probable Upper Paleozoic age separating the Kazakh Uplands from the Dzungarian basin.

4.3.3.3. Mongolia

The wide Mongolian belt separating the North China block from Siberia was the site of repeated collisional events during the Late Proterozoic and Paleozoic. It is probably the most convincing example of continental growth through arc accretion existing anywhere. Within this zone, which is greater than 1000 km. wide south of Lake Baikal, belts of granite show a progressive decrease in age southward from Proterozoic in the Baikal-Sayan region to Upper Paleozoic in southern Mongolia (Peive et al., 1972). Our plot of ultramafic bodies also indicates that there are a number of ophiolite belts within this region, and they also show a general decrease in age southward.

Given the width and overall complexity of this region, it is difficult to resolve the exact location of the final closure between North China and Siberia. We tentatively locate it in a relatively well defined Upper Paleozoic ophiolite belt extending from the western end of the Altay, across the southern part of Mongolia and into Manchuria. In southeast Mongolia this corresponds to the Solonker zone of Borzakovskiy et al., (1970). The main deformation in southern Mongolia is Upper Permian, after which sedimentation is dominantly continental, suggesting collision in the Upper Permian (Burrett, 1974).

4.3.3.4. Yablonovy and Stanovoy Mountains

We also plot a Paleozoic suture along a relatively diffuse ophiolite belt extending from south of Lake Baikal eastward along the southern margins of the Yablonovy and Stanovoy mountains.

4.3.4. PRECAMBRIAN SUTURES

4.3.4.1. Lake Baikal

At present the only Precambrian collision we can resolve in Asia is the Baikalian event. This is marked by a well defined ophiolite belt extending from the eastern Sayans, through the Lake Baikal region and thence along the southern margin of the Aldan Shield. This is accompanied by an equally well defined belt of Proterozoic granites (Chinese Geologic Map of Asia, 1975). Inliers of Archean basement occurring immediately to the south of the suture in the Yablonovy and Stanovoy ranges, suggest that it resulted from the impingement of a continental fragment in the Late Proterozoic. Kosygin and Parfenov (1975) note that along the southern margin of the Siberian Platform Upper Proterozoic deformed rocks are overlain unconformably by continental sediments that are continuous with overlying Cambrian carbonates. They place the main deformation in the Baikal region at about 850 m.y.

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5.0. INDIA

5.1. INTRODUCTION

Peninsular India is made up of three main cratonic areas (Naqvi, 1974). They are the Singhbhum, Dharwar and Aravalli blocks (Figure 5.1). Because of their antiquity they are considered the Archean proto-continents from which the present day Peninsular India has evolved. These masses have existed from at least as far back as 3.4-3.8 b.y. ago and are among the oldest of any rocks known to be preserved anywhere on the earth. The tectonic framework of India has been determined by the interactions between and within these ancient shield areas. The array of rifts and sutures results from three settings of activity which can be considered as follows: 1) the growth and stabilization of the proto-continental blocks and the resulting interaction in the Archean and Precambrian; 2) the rifting of India from its Gondwana neighbors during the Mesozoic; 3) the collision of northern India with Asia across the Alpine-Himalaya Mountain Belt from the Eocene to the present.

The jostling of these early Archean pre-plates is evidenced by deep main faults (Naqvi, 1974) dating from the permobile regime (~3.5-2.5 b.y. ago) of Burke et al., (1976). This provides the sites for subsequent episodes of both complete and incomplete (failed arm) rifting.

The major zones active since Archean times are the Narmada-Son-Damodar lineament (5.2.2.1.), the Mahanadi rift (5.2.3.2.) which divides the Singhbhum proto-continent into two nucleii and the Aravalli Main Boundary fault (Figure 5.1), dividing that proto-continental mass (Naqvi, 1974).

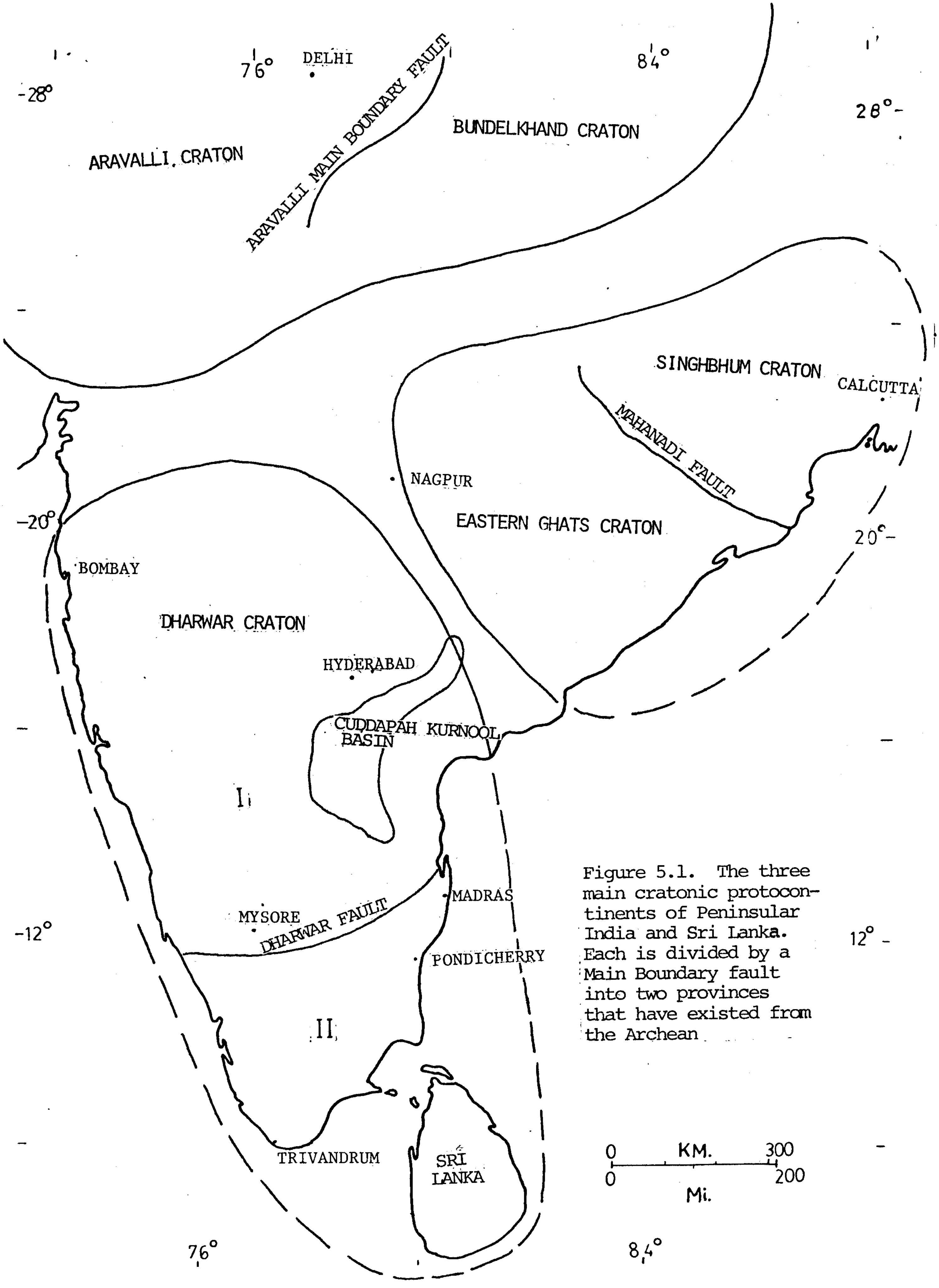

Figure 5.1. The three main cratonic protocontinents of Peninsular India and Sri Lanka. Each is divided by a Main Boundary fault into two provinces that have existed from the Archean

The southern Dharwar deep-seated lineament is related to the east-west axis of metamorphism (Figure 5.2) in southern India but is not considered a sutured boundary since it lacks a parallel trend of ultramafics or obducted ophiolites.

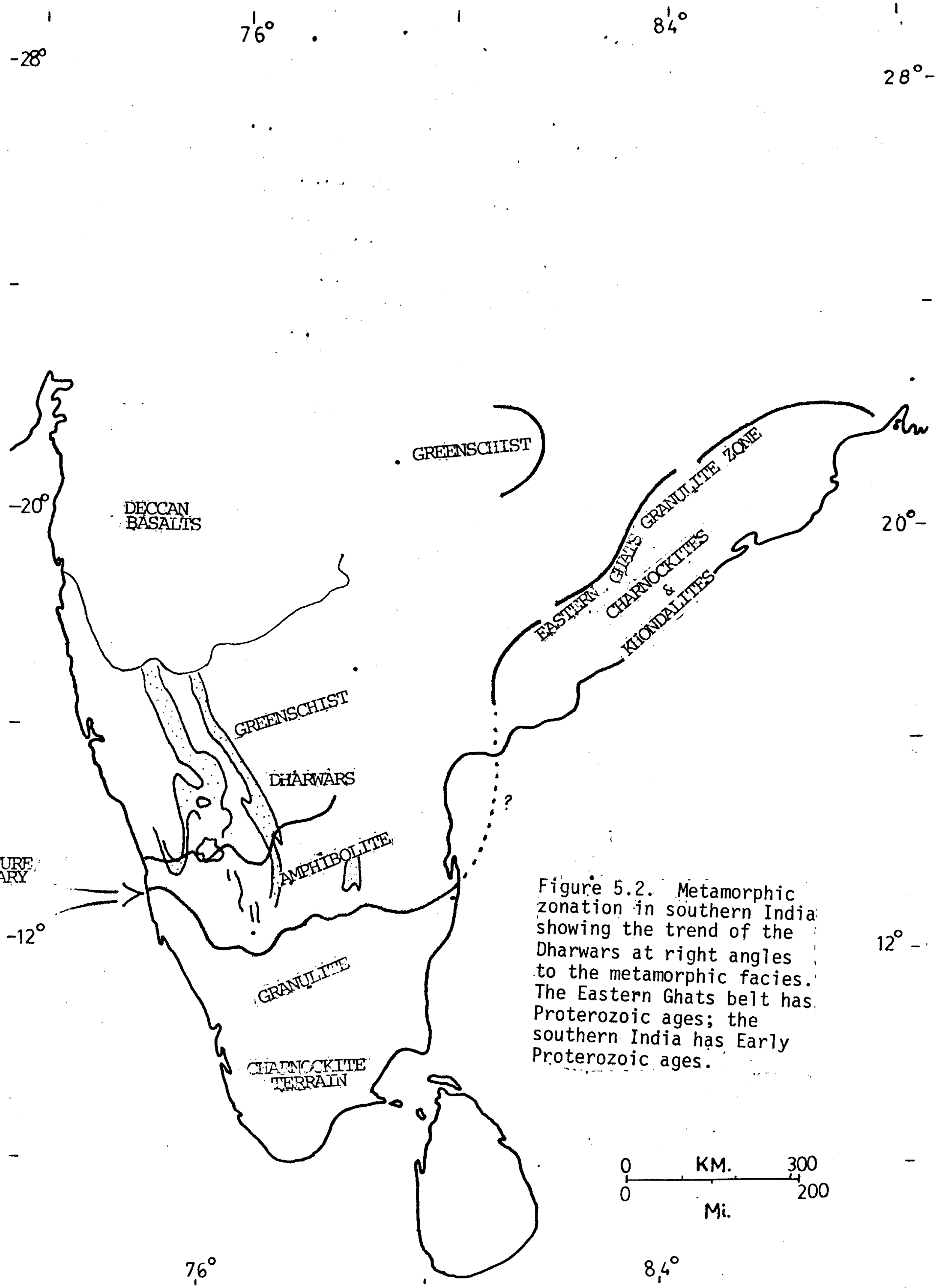

Figure 5.2. Metamorphic zonation in southern India showing the trend of the Dharwars at right angles to the metamorphic facies. The Eastern Ghats belt has Proterozoic ages; the southern India has Early Proterozoic ages.

5.2. RIFTS

5.2.1. ACTIVE RIFTS

5.2.1.1. Thakkhola

The only known location of active rifting is in the Thakkhola River valley located north of Annapurna in Nepal. Gansser (1964) shows that this N-S, fault bounded valley graben flanked by Cretaceous sediments with Jurassic limestones has a total vertical displacement of 2700 m. The rift appears related to late extension in the Himalayas parallel to their length. Perhaps the Devonian rifts of Norway are similar.

5.2.2. PALEOCENE-EOCENE RIFTS

5.2.2.1. Narmada-Son Lineament

A common occurrence in India as elsewhere is the reactivation of rifting along previous crustal weaknesses (faults, rifts or sutures). Associated with the breakup of Gondwana which defined the presently observed continental margins of India, is the existence of failed arms (Burke and Dewey, 1973) and renewed motion on deep main, pre-existing faults (Naqvi, 1974).

The most conspicuous and well-defined of these features is the Narmada-Son lineament. This feature is the valley in which flow the Narmada and Son rivers. It bisects Peninsular India from near Broach in the west to West Bengal in the east trending in a ENE-WSW direction. Though overlain by the vast Deccan basalts [of Late Cretaceous age (Krishnan, 1968; Wadia, 1939)], this zone is clearly evident in seismic (Naqvi, 1974) and gravity profiles (Qureshy, 1968). In the east the Son valley is flanked by Upper Gondwana (Jurassic) sediments.

The Narmada-Son lineament is a persistent zone of weakness with activation in the Permian (Gondwana) as well as a pre-Gondwana history of motion.

Naqvi (1974) suggests the Narmada-Son as a lineament separating the proto-continental nucleii of Aravalli and Dharwar in Archean time (5.3.2.3.).

5.2.2.2. Broach

The west coast of India was formed by Late Paleocene to Eocene rifting occurring at a four arm junction consisting of the Broach rift, the Narmada-Son lineament, the Gir coast and the N-S trending Indian coast parallel to the trend of the Panvel flexure. Broach is a N-S trending failed arm (Burke, 1977) of the Khambat rift system whose opening is synchronous with the Deccan basalt flows and the removal of the Seychelles from India (Davis, 1968). Downwarping of the western continental margin (the Panvel flexure), and dikes parallel to the coast are associated with this event.

5.2.3. PALEOZOIC-MESOZOIC RIFTS ASSOCIATED WITH THE BREAKUP OF GONDWANA

5.2.3.1. Godavari Rift

Early Cretaceous rifting associated with the formation of the Bay of Bengal reactivated this structure dating from at least the Permian (Veevers et al., 1975). The trend of this zone is NW-SE and its length is about 500 km. It joins the major lineament of the Narmada-Son north of Nagpur in the center of the Indian Peninsula (Figure 5.3). At its southeastern end it meets the coast of India and is the location

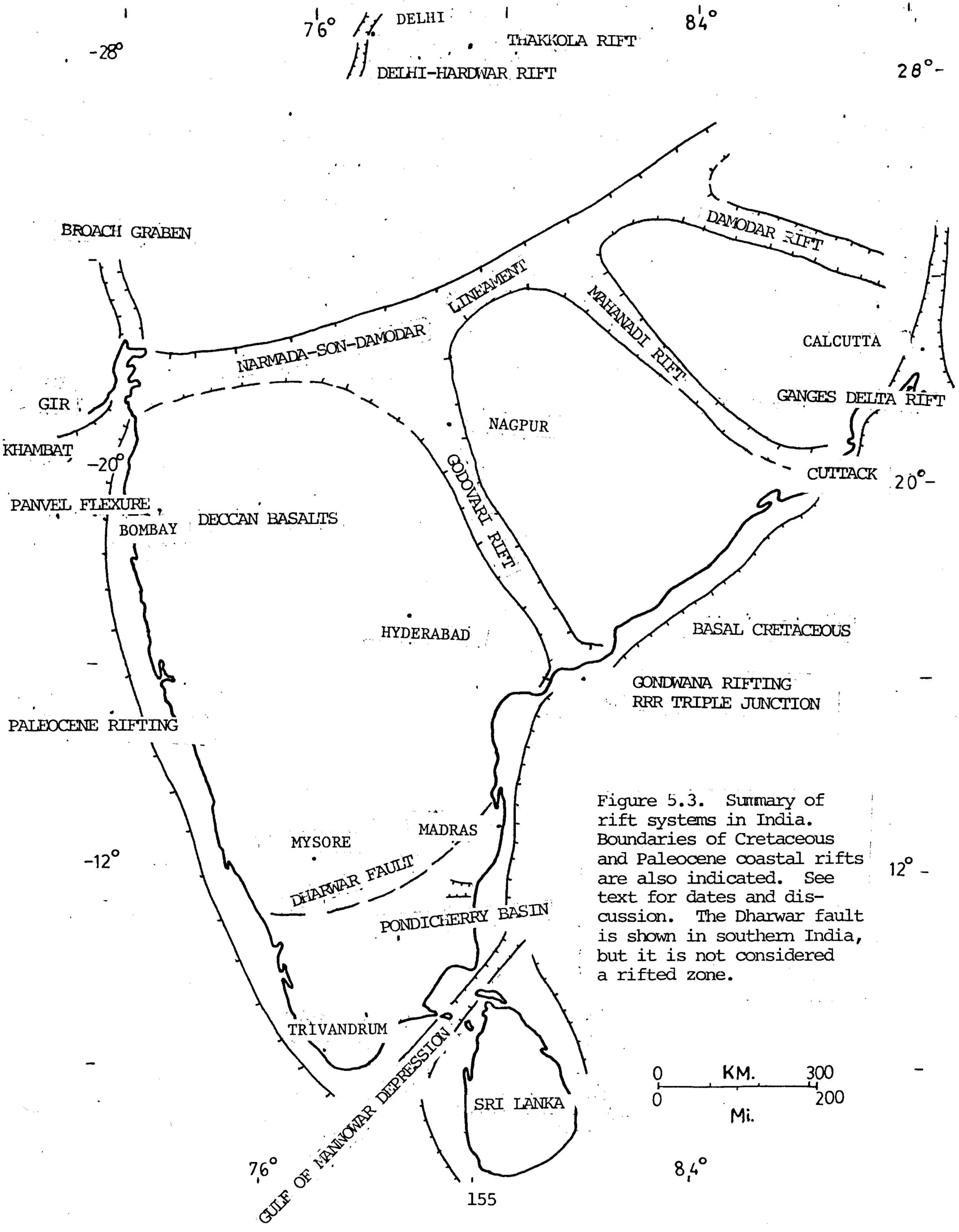

Figure 5.3. Summary of rift systems in India. Boundaries of Cretaceous and Paleocene coastal rifts are also indicated. See text for dates and discussion. The Dharwar fault is shown in southern India, but it is not considered a rifted zone.

of a major bend in the coastline from N-S to NE-SW. The Godavari river has built an extensive delta with a thick (4 km) sequence of sediments of Cretaceous age at the base indicating that an older river flowed down the structure.

5.2.3.2. Mahanadi Rift

The Mahanadi rift (Figure 5.3) is also a Cretaceous reactivated rift like the Godavari rift. Its junction with the coast of India and the Bay of Bengal is similarly marked by a large sedimentary wedge (Sastri et al., 1974). The best control on the timing of the opening of the Bay of Bengal by the removal of Australia is provided by the Lower Cretaceous magnetic anomalies off southwestern Australia (Markl, 1974). Krishnan (1968) lists Carboniferous shales cut by basalt dikes near Cuttack at the head of the Mahanadi river where the delta contains fossils of a similar age.

5.2.3.3. Ganges Delta Rift

The Ganges delta rift (Figure 5.3) now contains a vast prograding sediment wedge with an 8 km downwarped oceanic crust (Curray and Moore, 1974). It strikes N-S into East Pakistan and is considered by Burke and Dewey (1973) to mark the site of incomplete Mesozoic rifting. Ages for up to 600 m of basaltic lava flows are possibly younger than Upper Jurassic (Sahni, 1928), (Krishnan, 1968).

5.2.3.4. Damodar Rift

The Damodar rift (Figure 5.3) branches ESE-WNW from the Narmada Son lineament (5.2.2.1.) which also shows evidence of activity in this period. The Damodar river follows the axis of the rift and joins the coast of India at the western edge of the Ganges delta (5.2.3.3.) where these two rifts may meet (Burke and Dewey, 1973). Coalfields with Permian ages occur in the Damodar valley as well as in the Mahanadi and Narmada-Son valleys. This is consistent with spreading initiation. A common feature in all of these rift structures is a half-graben type of faulting (Ghosh, 1973) with draping of later sediments along one edge of the zone and monoclinal folding on the other side of the graben.

5.2.3.5. Pondicherry Rift

The Pondicherry failed arm is of Mesozoic age and is associated with the Gulf of Mannowar basin which contains 3 km of Mesozoic and younger sediments. Left lateral motion (Burke and Dewey, 1973) is responsible for this separation of Sri Lanka from Peninsular India. The Gulf of Mannowar joins with the rifted coasts of India at Cape Comorin in the west and Pondicherry in the east which are possible spreading triple junctions (Burke op. cit.).

5.3. SUTURES

5.3.1. PALEOCENE-EOCENE SUTURES

5.3.1.1. Indus Suture

The absence of seismic activity throughout Peninsular India, and the concentration of such present-day activity along the disturbed belt of the Himalayas in Extra-Peninsula India support the concept of the collision of the triangular wedge of the Indian sub-continent northwards against the Asian continental mass.

Ophiolites, serpentinites and altered peridotites mark a zone in the Baluchistan Arc and occur along a major fault axis between Karachi and Quetta. The Tsangpo river valley contains Cretaceous flysch derived from a rapid orogenic uplift and the age of the limestones in the High Himalaya date the suturing as Paleocene-Eocene (Gansser, 1964).

5.3.2. PROTEROZOIC SUTURES

5.3.2.1. Dalma Thrust

The Dalma thrust (Figure 5.4) is a suture in the Singhbhum craton that is associated with the closing of an ocean of 1700-1600 m.y. age (Yellur, 1977). The time of closing is related to the intrusion of the Singhbhum granite dated approximately 950 m.y. (Holmes, 1955). This is contemporaneous with the Satpura orogeny of the Bihar Mica Belt (Krishnan, 1968) which contains chromite-bearing lherzolite. In south Bihar there is a lens of altered peridotite also of mid-Proterozoic age. The Singhbhum Copper Belt, with its two parallel

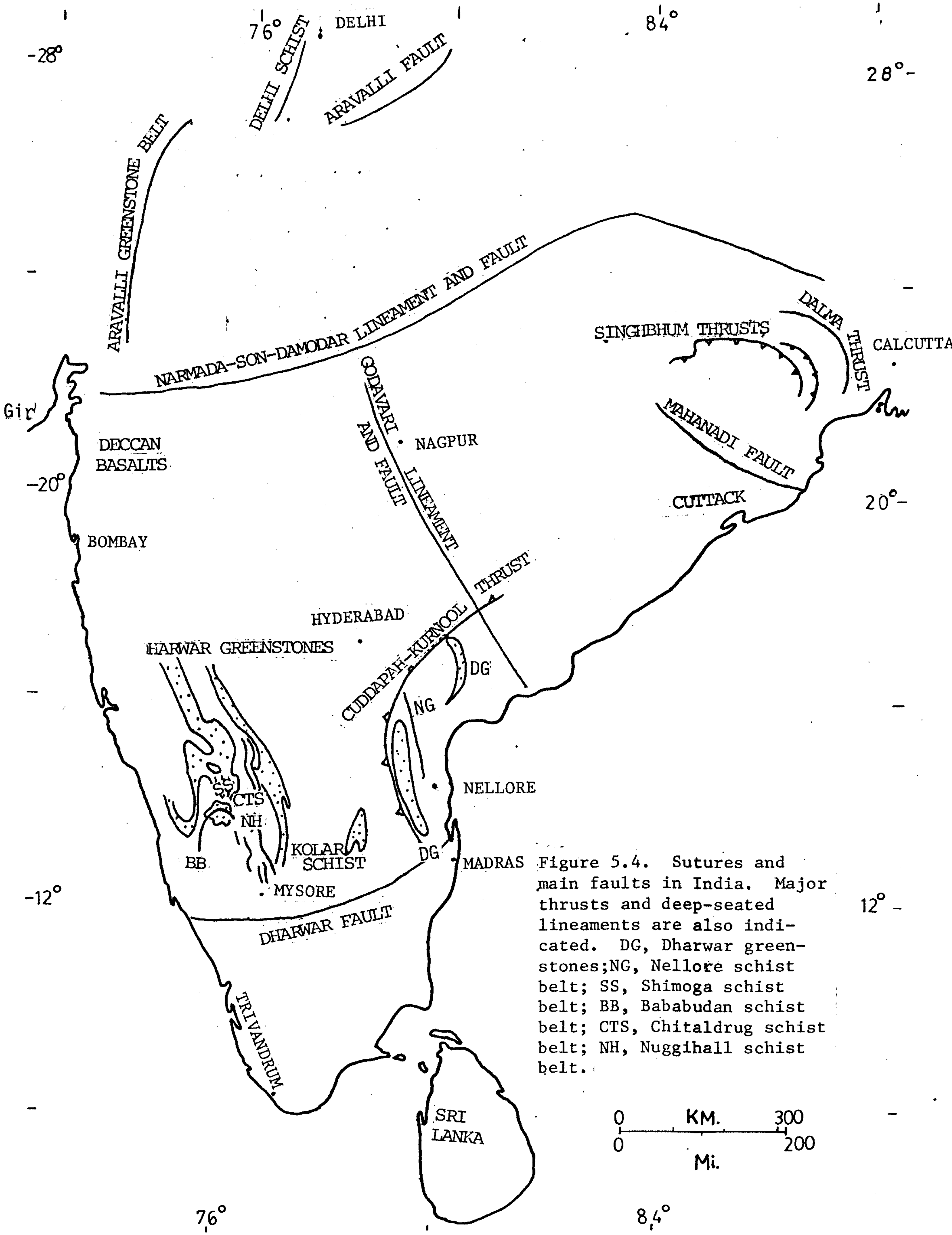

Figure 5.4. Sutures and main faults in India. Major thrusts and deep-seated lineaments are also indicated. DG, Dharwar greenstones; NG, Nellore schist belt; SS, Shimoga schist belt; BB, Bababudan schist belt; CTS, Chitaldrug schist belt; NH, Nuggihall schist belt.

south-facing thrusts is well marked by shearing and is a source of ores of iron, copper and uranium. The Dalma Traps slightly to the north are a zone of schistose talc and chlorite rocks with accompanying mineral deposits. These three thrust zones are parallel to each other in North Singhbhum and follow the Satpura strike which trends E-W over southern Bihar.

5.3.2.2. Delhi

Delhi belt (Proterozoic) is metamorphosed in a similar fashion to the Aravalli belt (5.3.2.3.) and is considered slightly younger. Mafic igneous rocks are associated with sedimentary and thin metamorphic equivalents (Pichamuthu, 1968). There is a major unconformity separating Delhi and Aravalli rocks with ages of 900 m.y. given for the Delhi system (Holmes, 1955). Delhi folding is superimposed on the Aravalli system since their trends are parallel the two belts may represent successive, or reactivated stages of suturing over a period of time of hundreds of millions of years.

5.3.2.3. Aravalli Greenstones

Similar to the Dharwar (5.3.2.4.) greenstone belt, but occurring in more limited areas, the Aravalli (Figure 5.4) greenstones have been assigned to the Early Proterozoic (Windley, 1977; Pichamuthu, 1968). Their trend in Rajasthan is generally NE-SW with a more distinctly N-S trend in the south nearer the coast. Naqvi (1974) considers this zone to be one of the nucleii of the Aravalli proto-continent with ages up to 3.5 b.y. but averaging approximately 1600 m.y. Along with the wide age range, the Aravalli system appears to link up with the Dharwars across the Deccan plateau to the south or more likely, with the ENE-WSW-trending Satpura belt though this is still conjectured and neither confirmed nor ruled out by the lack of cores penetrating the plateau basalts (Dar, 1972). The deposition of the quartzite bands in the Aravallis indicates that the folding here was very complex and was affected by intense shearing and thrust faulting. There is a decrease in metamorphic grade from the northwest toward the southeast, i.e., away from the axis of the Aravalli range (Pichamuthu, 1968), whose rocks are chiefly argillaceous.

5.3.2.4.

The Nellore greenstones are part of the Dharwar system but are located in a NNW-SSE-trending belt north of Madras near Nellore. This mica-bearing schist belt shows ages of 1630 ± 100 m.y. (Holmes, 1955) and may be related to the later Proterozoic Eastern Ghats system though the strike is at a high angle to the trend of the charnockite and khondalite series along the eastern margin of India. The limited size of this and other Dharwar greenstones in southern India may signify the closing of smaller oceanic basins or rear arc, thin crustal basins caught up in larger subduction systems (Kidd, 1974).

5.3.2.5.

The Eastern Ghats ultramafics (Figure 5.2) may mark the remains of an obducted oceanic crust of Archean age. Chromite and magnesite are associated with the Kondapalle hills of the Krishna district and several locations in Orissa trending in zones parallel to the existing

Gondwana rifted coast (Krishnan, 1968). Charnockite bodies (Krishan, 1968; De, 1969) are also found in the granulite facies metamorphosed Eastern Ghats. Whether this is a site of suturing or a non-sutured high grade metamorphic zone is not certain.

5.3.3. ARCHEAN

5.3.3.1. Dharwars

Among the oldest rocks in India showing the traces of suturing are the Dharwar greenstones (Figure5.4) (Windley, 1977; Katz, 1977; Glikson, 1977). The Dharwars of southern India occur in isolated and fairly well-defined bands with structural trends ranging from NNW-SSE to NW-SE. However, they span an enormous overall time span and may represent successive episodes of intercratonic suturing throughout the period 3.4-2.7. b.y. (Tarney et al., 1976; Burke et al., 1976).

Two Archean sutures are identified. They are the Chitaldrug Schist belt (Naqvi, 1975) and the Shimoga Schist belt (Naqvi and Hussain, 1973a) containing massive schistose greenstones (chlorite, hornblende and mica schists) and banded ferruginous quartzite respectively.

Since there is increasing metamorphic grade from north to south in southern India, the generally N-S Dharwar trends are overprinted by metamorphic events. Figure 5.2 shows this metamorphic zonation though the highest grade boundary does not represent a suturing event at the time of occurrence.

5.3.3.2. Pre-Dharwar

The pre-Dharwar greenstone belts in southern India (Figure 5.4) are possible sites of ocean closing, though the exact nature of oceanic and continental crust is somewhat speculative for times greater than 3.0 b.y. ago. Burke et al., 1976, suggest that continental blocks in the Archean were similar in composition and thickness to present-day continents. However, they suggest a higher overall heat flow and a greater number of smaller continents than exists now. Accordingly, the Kolar schist belt 3.4-3.0 b.y. age, the Bababudan belt and the Nuggihalli belt are suggested as sutured boundaries in the Archean (Windley, 1977). In the Bababudan belt, which is the oldest of the three, there are unoxidized detrital pyrites suggesting an atmosphere deficient in oxygen at the time of deposition. In both the Kolar schist belt and the Nuggihalli schist belt ultramafic and mafic assemblages (serpentinized peridotites, komatites and high Mg, low-K tholeiites) suggest sutured and metamorphosed (to amphibolites) zones (Viswanathan, 1974b).

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6.0. THE NEAR EAST

6.1. INTRODUCTION

This report deals with the Near East sector of the Alpine-Himalaya System, the evolution of which resulted in the welding of Eurasian and Gondwana cratonic nuclei. The western limit of the area dealt with is central Turkey; to the west of this limit the geology is covered by the European report (Section 7.0). The eastern limit is the Afganhistan region and beyond this the area of responsibility of the Asian and Indian report (Sections 4.0 and 5.0) begins. The sutures within the forelands of the Alpine-Himalaya System of the Near East have been dealt with in the reports on Africa (Section 2.0) and Europe (Section 7.0).

The region under investigation abounds in Neogene to Recent depressions, but as their nature and origin are as yet not clear, and as they differ from rifts proper in having roughly equant shape (Şengör, in prep., Şengör and Kidd, in prep.), we shall not consider them in this report. There are only two rifts in the area that resemble other rifts in the world and both are currently active.

The region covered in this report has two main 'old' suture lines within it. However, both of these are Tertiary in age and the southern suture is still active, at least locally. The continental collision that gave rise to the younger of the two sutures is still in progress, which results in the high seismic activity, throughout the orogenic belt. Şengör and Kidd (in prep.) have shown that the whole orogen is actively shortening across strike which result in the continued uplift and intensive volcanicity of the area. While considering the geophysical signatures of old as well as young sutures it is of critical importance to consider the present tectonic regime in the area under investigation.

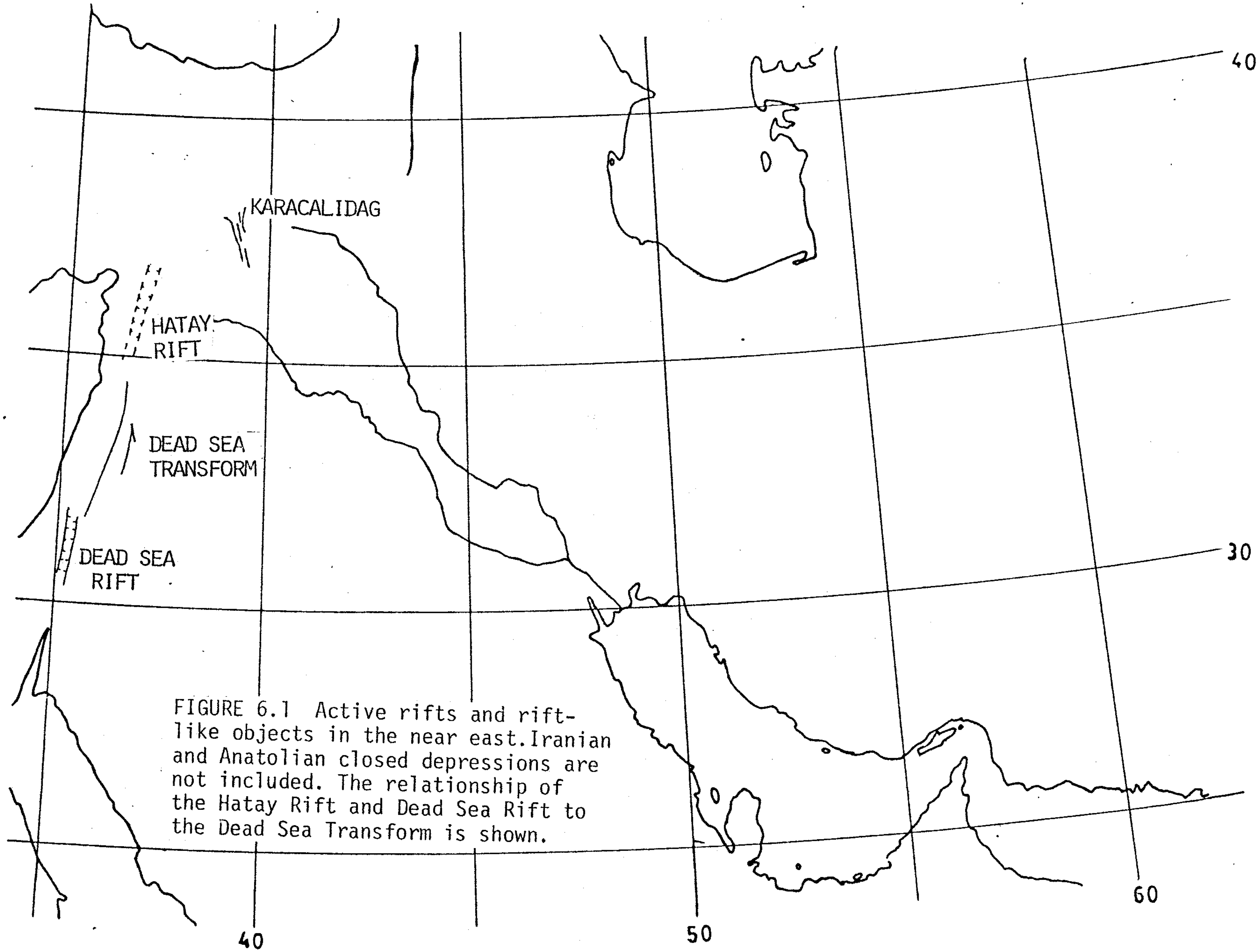

FIGURE 6.1 Active rifts and rift-like objects in the near east. Iranian and Anatolian closed depressions are not included. The relationship of the Hatay Rift and Dead Sea Rift to the Dead Sea Transform is shown.

6.2. RIFTS

6.2.1. ACTIVE RIFTS

Burke (1978) defined rifts as 'places where the entire thickness of lithosphere has ruptured under extension'. In this area there are two places that fit this definition at the present time.

6.2.1.1. Dead Sea Rift

The Dead Sea Rift located in the southern section of the Dead Sea Transform (Figure 6.1) is a pull-apart basin (Carey, 1958; Quennell, 1958; Frund, 1965) that has been evolving since the beginning of the Tertiary. Basaltic vulcanism, as typical of such basins, is in this case confined entirely to the southeastern shoulder of the rift.

6.2.1.2. The Hatay Rift Valley

A roughly NNE-striking narrow graben, the Hatay Rift Valley is located at the northern termination of the Dead Sea Transform (Figure 6.1). The region has a typical rift morphology and Quaternary basalt eruption within the rift valley (Yüngül, 1951). Although the age of initiation of this structure is not precisely known, it is likely to have originated during the Late Miocene and is still active.

6.2.1.3. Active Rift-like Objects

In 1947 McCallien first pointed out that a number of roughly N-S striking lineaments were present in eastern Anatolia (Figure 6.1). He thought that these were extensional features resulting from the 'impact' of Arabia. Recently Adaima et al., (1977) argued that many of the alkaline basaltic volcanoes in eastern Anatolia and Caucasus are related to such N-S extensional lineaments. Şengör et al., (1978) argued that the enormous basaltic shield volcano of Karacalidag in SE Anatolia is the result of basaltic eruptions along N-S fissures. The age of this event, late Pliocene-Pleistocene, is consistent with its interpretation as an extensional feature related to Arabia/Anatolia collision.

6.2.2. ANCIENT RIFTS

There are no recognized ancient inactive rifts in this area that have been preserved as such. The rifts that were related to the opening of the ocean in the area must have been totally obliterated by the later collisions. No aulacogens are present, except for von Gärtner's (1969) identification of the Greater Caucasus as an aulacogen. The southern Caspian Sea may be an oceanic reëntrant of the now obliterated Tethys, or an aulacogen. However, the possible oceanic substratum of the North Caspian Depression (outside of this area) complicates the interpretation of the Caspian Sea (Garetsky et al., 1972).

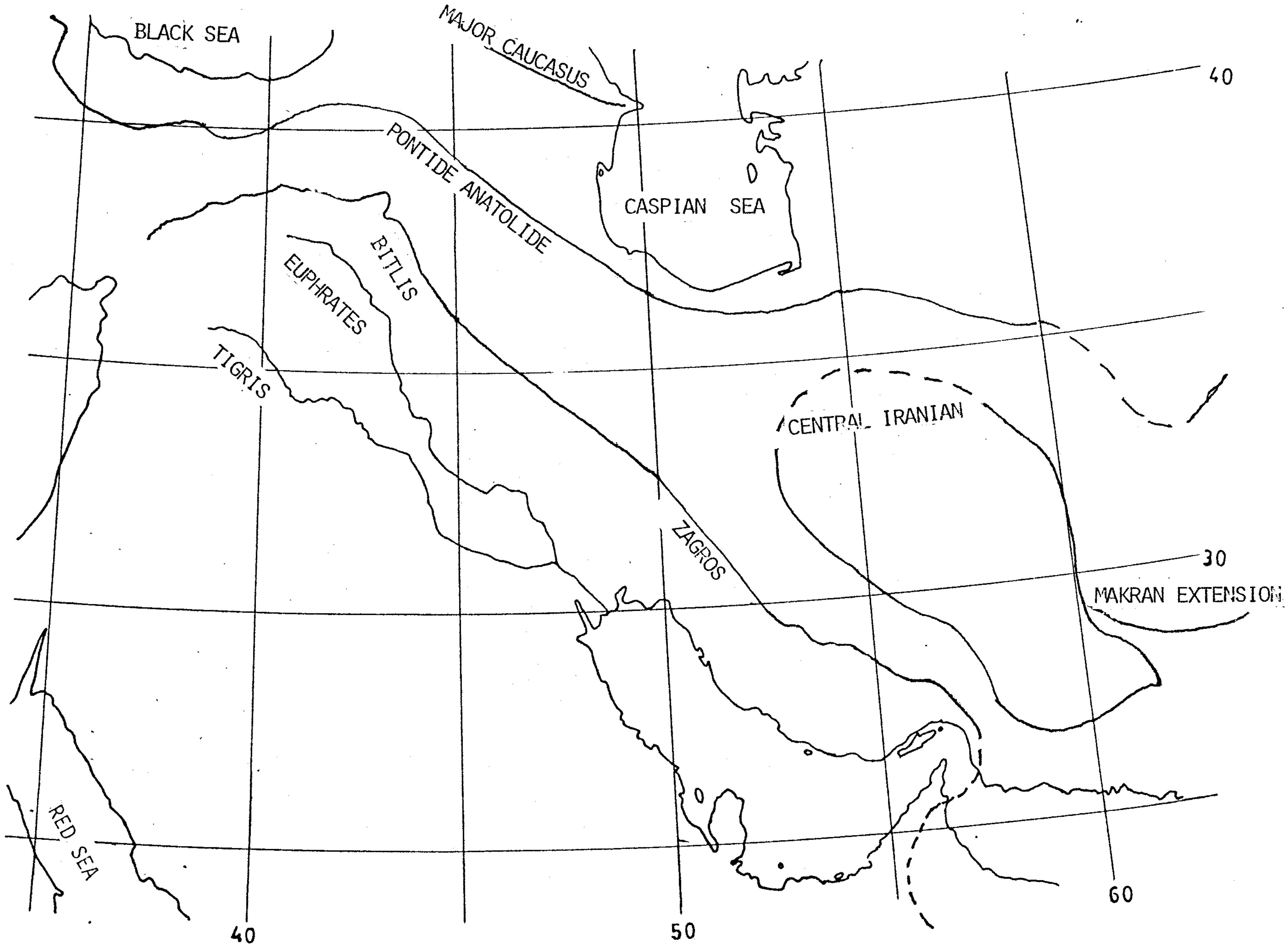

Figure 6.2. Sutures of the Near East.

6.3. SUTURES

6.3.1. CENOZOIC AND MESOZOIC SUTURES

6.3.1.1. Zagros Suture

As the direct continuation of the Bitlis suture to the southeast (Braud and Ricou, 1975), the Zagros suture (Figure 6.2) also closed during the Late Miocene (Dewey et al., 1973; Bird et al., 1975). Based on the isopach studies (Takin, 1972) and the age of ophiolite obduction (Stöcklin, 1974) some authors take the view that final closure here took place in pre-Maestrichtian time. However, the overall tectonics and the fact that Eocene deep-sea sediments and basaltic volcanics have been found in the Bitlis Suture (Arpat, 1977) argue against the idea of a late Cretaceous closure here. Gealy (1977) has argued that the ophiolite obduction in Oman, a direct continuation of the Zagros ophiolite belt into the Arabian peninsula (Glennie et al., 1973), has occurred as a result of arc-continent collision. Brookfield (1977) believed that a transform fault that paralleled the continental margin here may have controlled the ophiolite obduction here.

The Late Miocene Zagros suture merges with the Makhran active arc terrain (Farhoudi and Karig, 1977) along a north-south striking fault zone (Oman Line, Gansser, 1956, Sherman, 1976).

6.3.1.2. The Bitlis Suture

Bitlis suture is located in southeast Turkey (Figure 6.2) where the Bitlis Ocean (Dewey and Şengör, in press), which began opening during the late Triassic (Dewey et al., 1973; Bein and Gvirtzman, 1977) closed during the late (middle?) Miocene (Dewey et al., 1973; Hall, 1976;

Yalcin, 1977; Dewey and Şengör, in press). The French workers (e.g., Ricou et al., 1975) and the M.T.A. (Turkish Geological Survey, see Baştug, 1976) defend the view that the Bitlis Suture is related to an Eocene ocean-opening and late Miocene closure event. They believe that the Cretaceous ophiolite obduction here was governed by the Pontide/Anatolide collision to the north. However, this view is largely contradicted by the field evidence.

6.3.1.3. Pontide/Anatolide Suture

The Pontide/Anatolide suture represents the closure of the Pontide ocean (Figure 6.2) during the Oligocene-early Miocene in eastern Turkey (Seymen, 1975). The ocean disappeared along a north-dipping subduction zone as seen from the vergence of the present suture structures (Ataman et al., 1975; Seymen, 1975) and from the arc volcanics on the Pontides and Lesser Caucasus (Adamia et al., 1977). This suture continues along strike into the suture bounding the southside of the Turan Block which also has an age of middle Cenozoic (Dewey et al., 1973).

6.3.1.4. Major Caucasus Suture

This rather poorly defined suture (Figure 6.2) represents a marginal basin that closed in the Caucasus area sometime during the period between the Cretaceous and the Eocene (Adamia et al., 1977).

6.3.1.5. Central Iranian Suture

The circum-Central Iranian Micro-continent suture (Figure 6.2) is characterized largely by Cretaceous-Eocene ophiolitic melange (Stöcklin, 1974). The suture has generally steep structures and the polarity of subduction along it is not easy to determine. Takin (1972) argues for a late Cretaceous closure of this suture, but the completion of it may have lasted into the Oligocene.

6.3.1.6. Possible Makran Extension Suture

This suture (Figure 6.2) may in fact be the northernmost expression of the Makhran arc terrane (Farhoudi and Karig, 1977) and record an island-arc continent collision of Cretaceous age.

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7.0. EUROPE

7.1. INTRODUCTION

Phases of rifting related to the opening of adjacent oceans comprise the following; Paleocene/Eocene related to the opening of the North Atlantic (rifts 1-6; rifts are numbered in Figure 7.1); Late Jurassic/Early Cretaceous related to the separation of Spain from France and the Grand Banks of Newfoundland (rifts 9 and 10); Triassic/Jurassic (rifts 11-25) over a wide area of western and northwestern Europe related to the complex rift/transform zone that connected the Proto-Central Atlantic to the rift/transform systems within the Tethys Alpine system; Late Precambrian related to the opening of the Uralian Ocean (rifts 26-32) and the Caledonian Ocean (rift 33). The Paleogene Bresse/Limagne and Rhine Grabens (rifts 7 and 8) have more complex origins to be discussed below.

Sutures resulting from the closing of oceans and associated intracontinental foreland-margin thrusting (A-type subduction zones of Bally, 1975), fall into five phases: Alpine (sutures 1-13; sutures are numbered in Figure 7.2) related to the complex diachronous demise of several oceanic tracts from Cretaceous to Miocene times; Hercynian/Uralian (sutures 14-22) resulting from the closure of large ocean basins (Urals), probable back arc basins (suture 18), and narrow oceanic tracts (sutures 14 and 15) during Late Carboniferous/Earliest Permian times; Caledonian (sutures 23-32) resulting from ocean closures from Latest Precambrian to Early Devonian times; sutures 34 and 35 of Late Middle Proterozoic age (1200) and Laxfordian/Karelian (sutures 36-40) Middle

Proterozoic age (about 1700 m.y.). The main cartographic source is the Tectonic Map of Europe (1962).

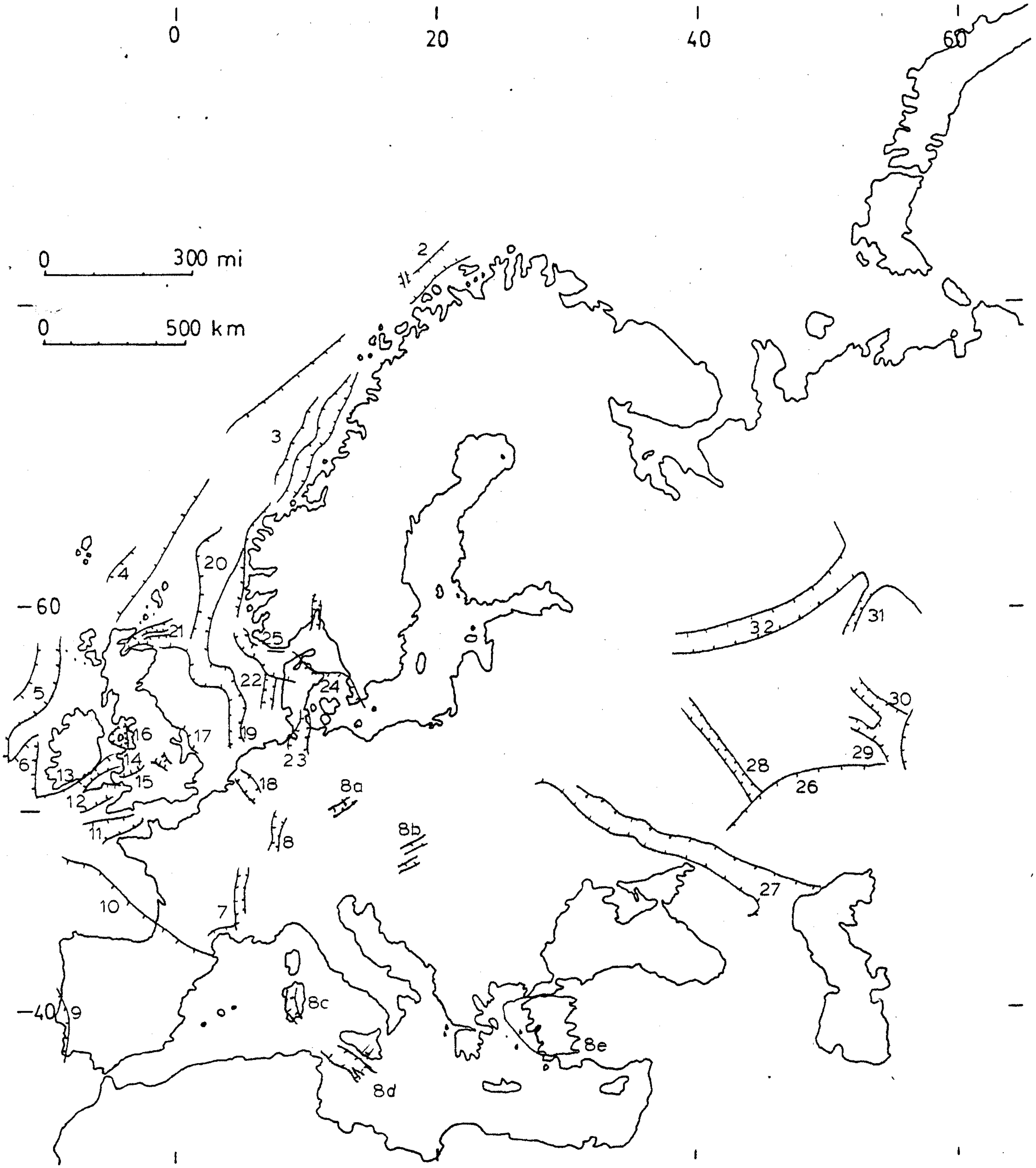

Figure 7.1. Rifts of Europe. Numbers are referenced in the text.

7.2. RIFTS

7.2.1. FORELAND GRABEN OF THE ALPS

The Limagne Graben (rift 7) of Oligocene age contains northwest striking normal faults and may have developed as a north-striking left lateral shear zone associated with north-south compression in the Alps. The Rhine Graben (rifts) began during the Lutetian stage of the Eocene at precisely the time when the Upper East Alpine nappe complex was overriding the Pennides and has been interpreted as related to continental collision following the subduction of the Alpine Ocean (Şengör, Burke and Dewey, 1978). The Bohemian Rifts (rift 8a) of Oligocene age (Illies and Gieiner, 1976) are probably also collisional in origin.

7.2.2. INTRA-ALPINE RIFTS

Rift systems resulting from complex intra-orogenic collisional and rotational strains are as follows: the Pannonian Rifts of Miocene age (8b), the Sardinian Rift of Oligocene/Miocene (8c), the Pantellerian Rifts of Lower Pliocene age (8d) and the Aegean Rift Complex of Late Miocene to Recent age (8e).

7.2.3. OPENING OF THE NORTH ATLANTIC

The Spitzbergen Fracture Zone (rift 1) is the northern boundary of the North Atlantic along which Greenland slid northwestwards. During the Paleogene the slip vector was slightly oblique to the fracture zone so that compression resulted in the formation of the Spitzbergen Alpine Fold Belt. From the southeastern end of the Spitzbergen Fracture to just west of Iceland a series of faults and grabens are associated with the Early Paleocene opening of the North Atlantic (Hammerfest

Basin (2), Helgeland Trough (3), Shetland Trough (4), Rockall Trough (5), Porcupine Bight Trough (6).

7.2.4. LATE JURASSIC/EARLY CRETACEOUS ROTATION OF SPAIN

During this phase the Lusitanian Basin (9) and other basins associated with the rifted margin of the Bay of Bascay (10) were formed.

7.2.5. LATE TRIASSIC/EARLY JURASSIC RIFTING OF GREAT BRITAIN AND THE NORTH SEA

At this time Africa was beginning to separate from North America and slide eastwards past southern Europe. During this relative motion, associated extensional strain was distributed over a wide zone of northwest Europe within which occurred the following major rifts: English Channel western approaches (11), Bristol Channel (12), Wexford (13), Cardigan Bay (14), Manse (16), Solepit Basin (17), Lower Rhine Graben (18), Central (19), Viking (20), Moray Firth (21), Horn (22), Gluckstadt (23), Danish (24), and Ekesurd (25) (Woodland, 1975). Many of these rifts were later reactivated particularly during the Meso-Alpine (Eocene) and Neo-Alpine (Miocene) compressional phases in the Alps.

7.2.6. PALEOZOIC AND LATE PRECAMBRIAN AULACOGENS

Because of widespread suturing rifts are poorly represented within the Paleozoic of Europe. Rifts associated with an Hercynian ocean-forming event are best represented by the type aulacogen (27) of Shatsky, the Dnieper-Donetz Basin. The Pre-Caspian depression (26) into which the Dnieper-Donetz and Pachelma rifts open, is characterized by extensive salt diapirism and may be a remnant ocean tract left by the incomplete closure of the Uralian/Hercynian oceanic tract.

Late Precambrian aulacogens (Nalivkin, 1976) that developed as failed arms of rift systems were associated with the opening of the Caledonian Ocean of Britain (Charnwood Forest Aulacogen, 33) and with Uralian Ocean (Dnieper-Donetz, 27; Pachelma, 28; Sernovodsk, 29; Kaltasa, 30; Vyatka, 31; and Moscowian, 32).

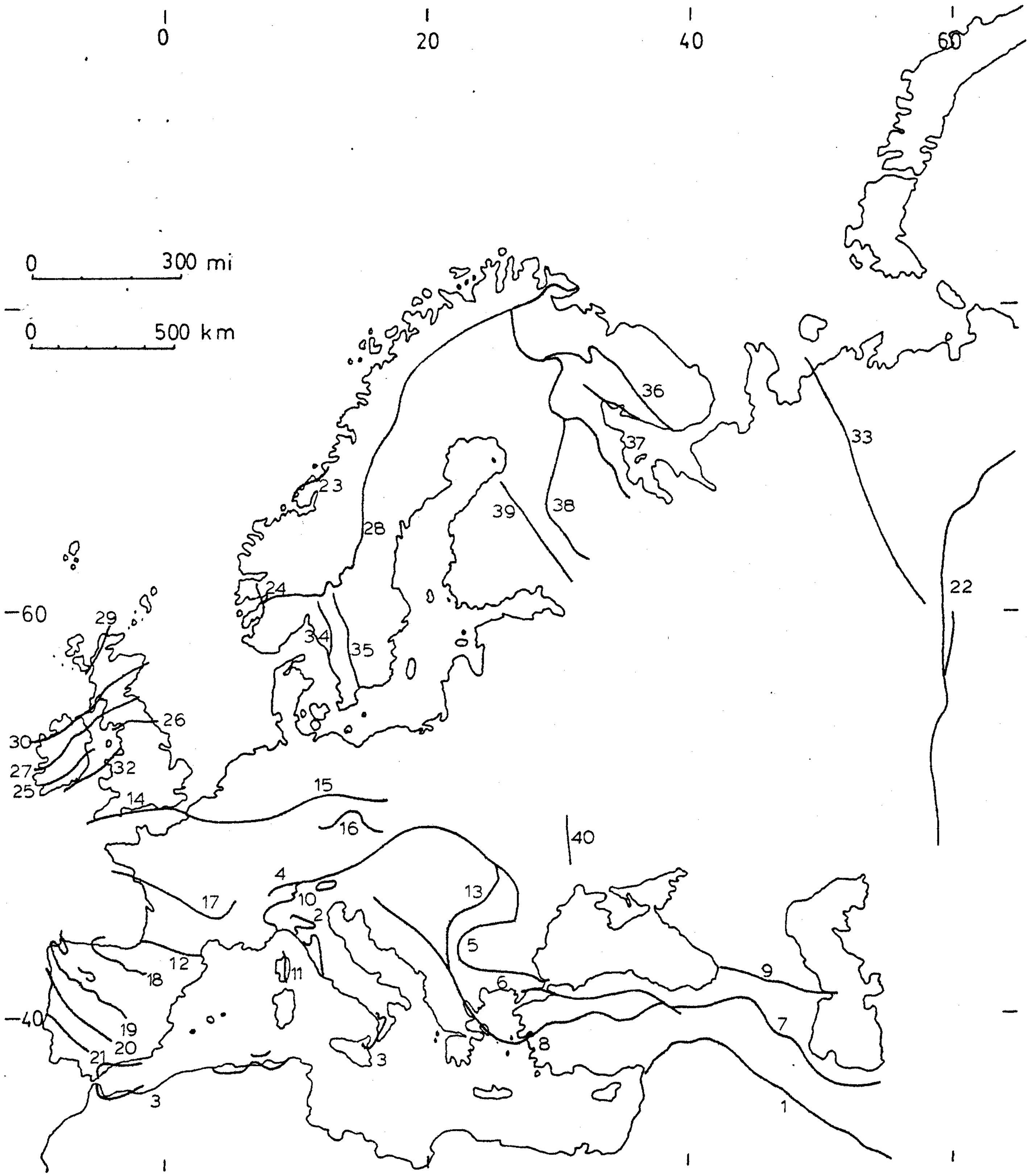

Figure 7.2. Sutures of Europe. Numbers are referenced in the text.

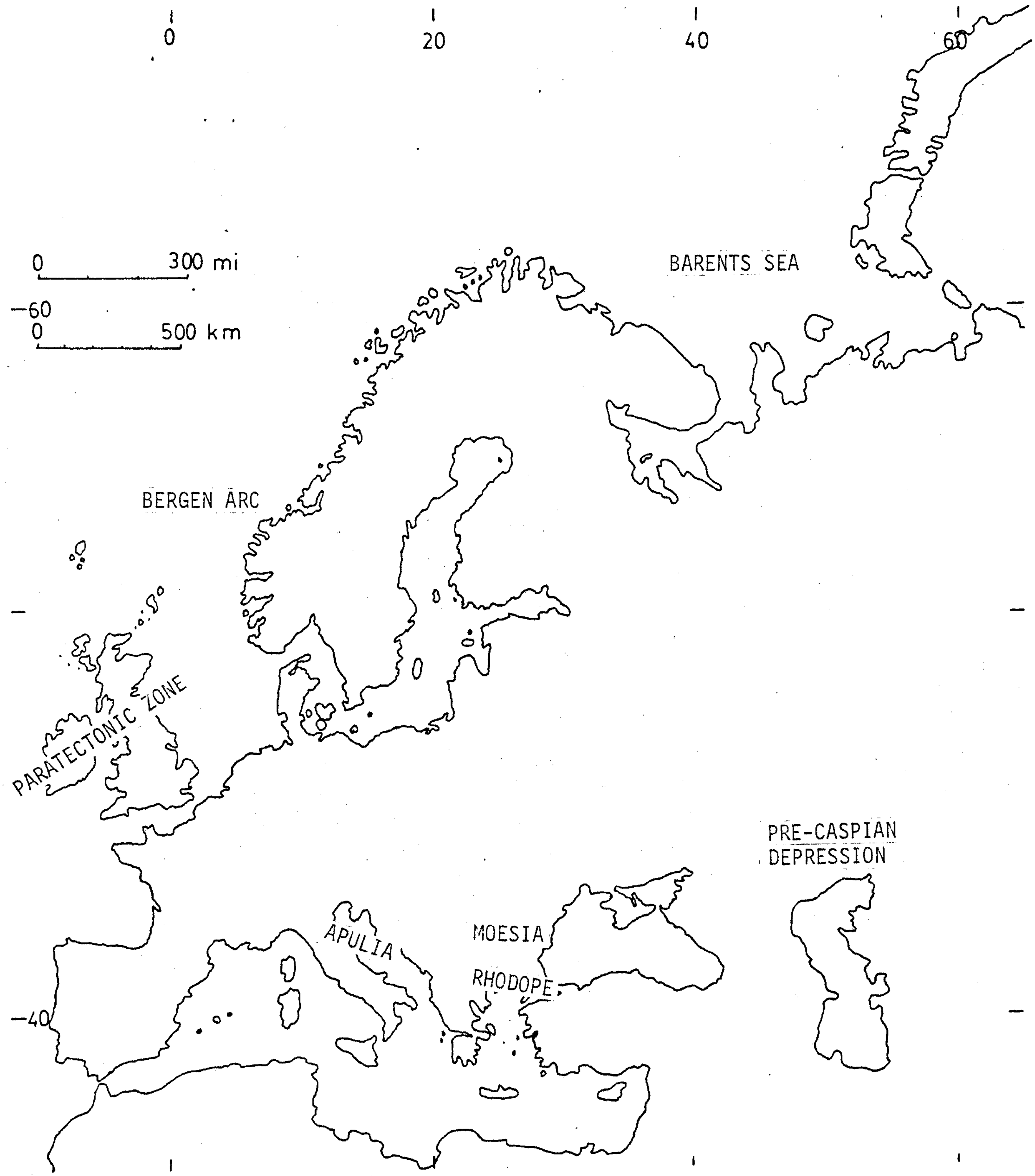

Figure 7.3. Location Map for Europe.

7.3. SUTURES

7.3.1. ALPINE SUTURES OF MESOZOIC/CENOZOIC AGE

The Bitlis/Zagros Suture (1) is a southward verging overthrust zone of shredded ophiolites, flysch, and colored melange, along which ophiolite obduction occurred during the Late Cretaceous and terminal suturing took place during Late Miocene/Early Pliocene times. The Bitlis Zone consists mainly of gently northward dipping thrusts which steepen eastward along strike into the Zagros suture. The Liguride Suture (2) in the northern Apennines was developed by the eastward overthrusting of an ophiolite melange mainly during Miocene times, although thrusting continued into the Pliocene/Pleistocene. The Liguride Complex was mainly developed by subduction accretion prior to terminal suturing during the subduction of an ocean that had been developed by plate accretion during the Jurassic. The origin of the Betic/Rif/Atlas Suture (3) is somewhat obscure. It appears to represent a thin shovel-shaped flake of exceedingly thin continental crust that was imbricated and squeezed between Spain and North Africa during the Miocene. In several places along the suture (Ronda, Beri Bouchera, Cap Bougeroun) ultramafic complexes are intimately associated with two-pyroxene granulites and probably represent the subcontinental mantle of continental crust that was attenuated during the Jurassic. The Helvetic subduction zone (4) of Miocene age lies between the Aar and Gotthard Massifs, fades out westward between the Aiguilla Range and MontBlanc Massifs, and merges eastward with the higher and earlier base of the Upper East Alpine Thrust Complex. The Helvetic subduction zone is wholly intracontinental, although it may be nucleated in a Jurassic rift valley with basified cores and thinned continental crust, and represents the zone along which the original continental basement of the Helvetic nappes slid southward beneath the Pennides, probably along the crustal low velocity zone at a depth of about 20 km. The Eastern Carpathian suture (5) is the zone, possibly intracontinental, along which the Moesian basement appears still to be sliding beneath the Transylvanian nappes. The Balkan suture (6) may not have been the site of extensive oceanic subduction, but it is probably the lone along which the Rhodope Massif (Figure 7.3 is a location map) was welded to the Moesian Platform during the Neogene and may now be a minor strike-slip fault system. The Pontide/Minor Caucasus suture (7) is of Oligocene/Early Miocene age, continues westwards into the Eocene Vardar Suture Zone of Greece (8) and was probably the zone along which the pre-Late Triassic Tethys was subducted (as is necessitated by a Pangea reconstruction). During the Early Cretaceous, ophiolite complexes were obducted from the Vardar Zone southwestwards across the adjacent Pelagonian Massif. The Khoura/Khourala Suture Zone (9) between the Major and Minor Caucasus is the Late Cretaceous line of terminal suturing along which the floor of a Late Jurassic/Early Cretaceous rear-arc basin was subducted. The remnants are preserved in the eastern Black Sea and the south Caspian Sea. The Zermatt/Arosa/Matrei Suture of the Alps is the zone containing shredded ophiolite and Cretaceous blueschist. Along this zone the continental basement of the southern Alps, which had been extensively thinned during the Early Jurassic, was thrust northward across the southern edge of northern Europe. This motion followed the Late Cretaceous/Paleocene subduction of the 'oceanic' floor of the Alpine realm. The Eocene Corsican (11) and the Pyreneean (12) sutures are an original continuation of the Zermatt suture prior to the rotation of Corsica/Sardinia away from France during the Oligocene. The Apuseni 'Suture' (13) is of Cretaceous age and involved extensive ophiolite obduction.

7.3.2. URALIAN/HERCYNIAN SUTURES OF LATE CARBONIFEROUS AND EARLIEST PERMIAN

The Lizard Suture (14) involves the northward overthrusting of a complicated, highly deformed ophiolite complex northwards over a melange and a series of Devonian/Carboniferous pelite and flysch deposits. The North German Phyllite Zone (15) is the highly deformed contact zone between the Middle German Swell to the south and the Rheinische Schiefergebirge/Harz complexes to the north. The latter consists of a listrically imbricated stack of Devonian/Carboniferous pelite/flysch, which may be a Carboniferous subduction accretion prism. Ophiolites are known in the suture zone to the southeast of the Harz Mountains. The Bohemian suture (16) is known from work in East Germany (Behr, pers. comm.), and it appears to be a northeastward-verging overthrust of gigantic proportions that emplaces granulites of probable lower crustal origin over high level greenschists and low-grade amphibolite. The Isle de Groix/Massif Centrale suture (17) is a south-verging overthrust that, in the Massif Centrale appears to be a similar kind of structure to the Bohemian suture, but northwestwards in the Isle de Groix contains Upper Paleozoic blueschists. The Isle de Groix/Bohemian suture system is believed by Professor P. Matte (pers. comm.) to be the zone along which the wide Hercynian ocean, predicted by paleomagnetic data, was subducted leading to continental collision and terminal suturing possibly as early as Early Carboniferous. In Spain four sutures (18-21) are clearly present and characterized by shredded ophiolite remnants, but it is still not clear which of them is the line along which the large Hercynian ocean was subducted and which represent closed rear-arc or intra-arc basins. The Uralian suture (22) is marked by giant ophiolite complexes and is the zone along which the Moscow platform collided with the western edge of the Siberian Block during Late Carboniferous/Early Permian times following the subduction of a very large oceanic area indicated by paleomagnetic data on Lower Paleozoic rocks.

7.3.3. CALEDONIAN SUTURES OF LATE PRECAMBRIAN THROUGH EARLY DEVONIAN

The Trondheim (23) and Bergen arc (24) sutures involve probable ophiolite obduction during the Ordovician and terminal suturing during the Lower Devonian. The Leinster (25), Cumbrian (26), and Strokestown/Southern Uplands (27) sutures are marked along the lines of Early Paleozoic subduction zones that bound the paratectonic zone of the British Caledonides, which is believed to be an incompletely closed Ordovician/Silurian ocean. The Scandinavian suture (28) is the eastern overthrust boundary of the Scandinavian Caledonides along which at least 250 km of Baltic shield continental basement has slipped westward beneath the Caledonides. It may have a similar origin to the Helvetic suture of the Swiss Alps. The Moine Thrust (29) is a similar but westward-verging structure along which both Early Ordovician and Early Devonian movement occurred. The Clew Bay/

Antrim/Highland Boundary suture is marked by ophiolites and appears to have resulted from the closure of a narrow oceanic tract during medial Ordovician times. The Spitzbergen suture (31) is suggested on the basis of the discovery of blueschist in a highly deformed zone in West Spitzbergen. The Permynnyd Suture (32) in Anglesey is a high strain zone of Late Precambrian age containing lawsonite-bearing blueschists and ophiolites.

7.3.4. PROTEROZOIC SUTURES

The Barents Sea suture (33) is based upon a zone of strongly deformed Proterozoic shelf rocks and may represent a subducted ocean or a compressed aulacogen. Either or both of two massive mylonitic high strain zones (34, 35) 1200-1300 m.y. old, within and bounding the north-striking Gothides in Sweden, may be sutures. The Kola suture (36) forms the south-verging overthrust (1700 m.y. old) of reactivated Archean over less reactivated Archean rocks. To the south, south-verging thrusts (38) may be intracontinental imbrication related to progressive shortening following continental collision along the Kola Suture. The 1700 m.y. mega shear zone (39) may be analogous to the Neogene Altyn Tagh Fault of Tibet and represent the lateral escape of a continental sliver from a 1700 m.y. continental collision zone. A 1700 m.y. suture, the Ingul Suture (40) is suggested in the Ukraine by a north striking boundary with deformed ultramafic rocks between the unreactivated Bug Podolian Archean basement to the west and the reactivated Ingul Archean basement to the east.

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8.0. NORTH AMERICA

8.1. INTRODUCTION

This continent contains a great number of relatively well studied rifts, particularly those of Triassic/Jurassic age connected with the opening of the central Atlantic (Figure 8.1), those of early Cambrian age, associated with the opening of the Proto-Atlantic or Iapetus ocean (Figure 8.2) and those formed about 1200 m.y. ago, when the western margin of North America was defined by removal of another large continental fragment (Figure 8.3). Although rifts older than 1200 m.y. are not so numerous, a particularly well exposed and well studied rift (aulacogen) more than 2000 m.y. old is present (Athapuscow aulacogen). Well exposed and relatively well-documented examples of sutures 300 to 400 m.y. old (Appalachians), and 1700-1800 m.y. old (Labrador Trough-Nelson Front) (Figure 8.4) are also noteworthy features of this continent.

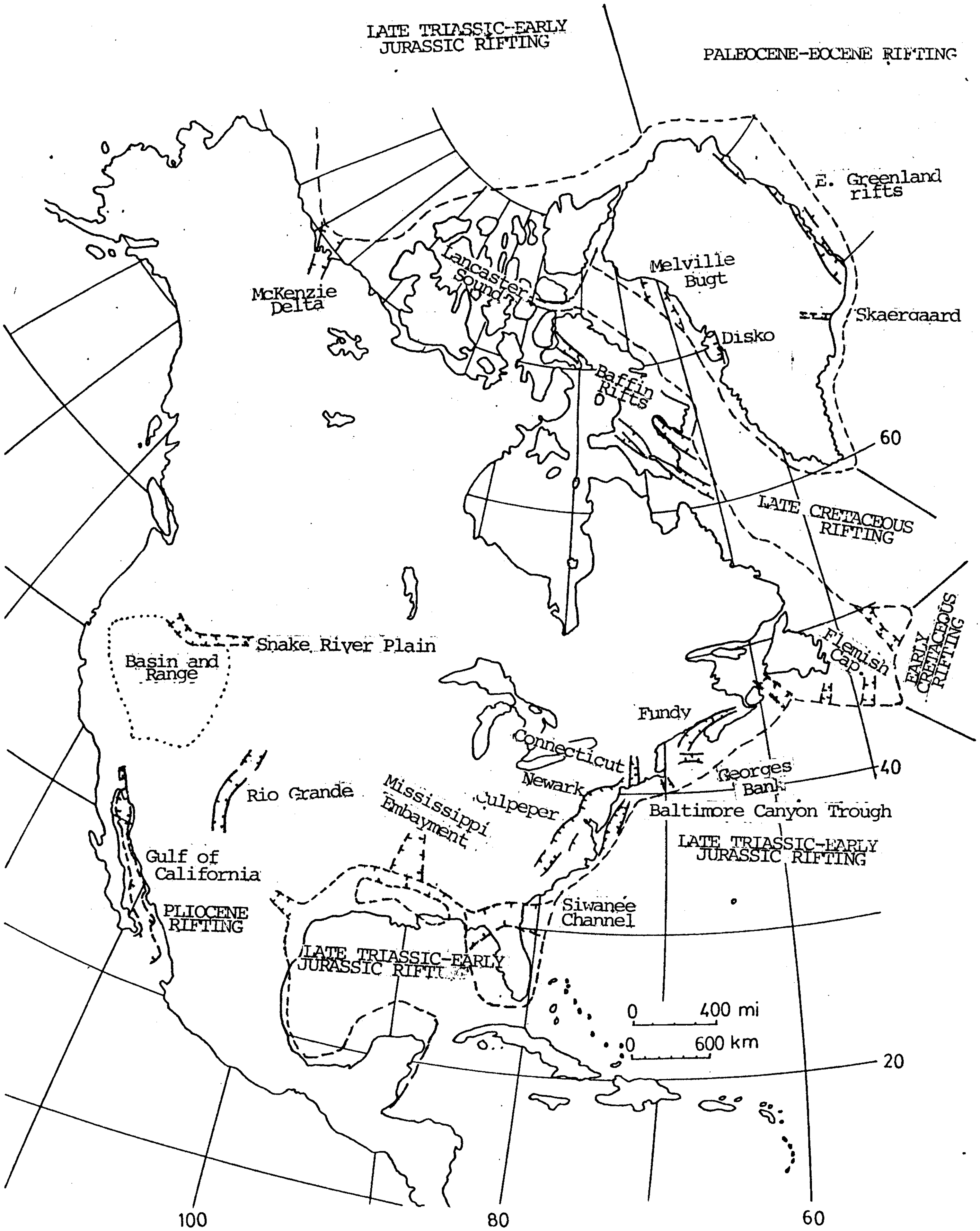

Figure 8.1. Active, Cenozoic, and Mesozoic rifts of North America

8.2. RIFTS

8.2.1. CENOZOIC RIFTS, ACTIVE AND INACTIVE

8.2.1.1. Rio Grande Rift

This is the only active rift in North America (Figure 8.1) that is relatively wide and occurs in the stable platform. These two identifying features and its associated alkali basalt volcanism make it resemble such active rifts as the East African rift much more than the other active North American rifts. The Rio Grande rift started to form about 29 m.y. ago and the volcanism associated with it has continued up to very recent times (Chapin and Seager, 1975). It is not clear whether it should be regarded as part of the Basin and Range Province since it is geographically isolated from it.

8.2.1.2. Snake River Plain Rift

This is also an active rift and of comparable size to the Rio Grande rift, but it is located within the area of the Cordilleran Orogen (Figure 8.1). Volcanism associated with this rift is up to 15 m.y. old (King, 1977), and its focus has apparently migrated with time to both northwest and northeast (to Yellowstone) away from the apex of this V-shaped (in plan) rift. The rift lies transverse to and essentially marks the northern limit of the Basin and Range faulting. The significance of this relationship will probably not be understood until the causes of Basin and Range rifting are also better understood.

8.2.1.3. Basin and Range Rift Province

The numerous rifts of this area (Figure 8.1) are, without exception, structures of small width; many have limited continuity along their length. In addition, the area affected by rifting is about as wide as it is long. All these properties contrast markedly with rifts, such as the East African rift, formed at continental breakup in areas not adjacent to or within areas of active or recently active convergent tectonics. Volcanism associated with the Basin and Range rifts is strongly bimodal and alkalic (Christiansen and Lipman, 1972), and a remarkably large proportion of the volcanics are of silicic composition, another dissimiliarity to continental breakup rifts. Rifting and volcanism started perhaps as long ago as 25 m.y., but most are younger than 15 m.y. old, and have continued in places until very recently. There is some evidence that rifting started in the center of the area and has progressively migrated to its margins. The province extends a considerable distance into Mexico, but available maps are indequate to define the indivicual rifts in this area; as a consequence they are not plotted on the map in this area.

This rift province does not have any convincing present or known past analogues; the great lengths of orogenic belts that have not had such rift provinces formed in them, prior to continental collision, suggest that a Basin and Range event is an unusual occurrence. The Basin and Range may be a huge zone of right-lateral shear associated with the San Andreas Fault.

8.2.1.4. Margins of the Gulf of California/Salton Sea

These margins formed about 4 m.y. ago (Larson, 1972), although they may have been outlined by somewhat earlier rifting. The Gulf of California (Figure 8.1) has subsequently been widened by sea-floor spreading. The rifts of the spreading ridge segments are sufficiently close to the continent that we have plotted them in the northern part of the Gulf. The northern rift-defined basins, including the Salton Sea rift, have been plugged with sediments transported in by the Colorado River. Rifts completely filled with sediments so early in their history are unusual, although closely similar structures probably exist where the Gakkel Ridge meets the Siberian continental margin in the Laptev Sea (Section 4.2.1.2.1.).

8.2.1.5. East Greenland Continental Margin

This rifted margin (Figure 8.1) was outlined by successful sea-floor spreading in the northern Atlantic from about 63 m.y. ago (Pitman and Talwani, 1972). However, it was the site of previous rifting of several ages in the region north from Scoresby Sund (q.v. Mesozoic, Palaeozoic rifts, below). The only episode of rifting in this area known to have had abundant magmatism associated with it was, however, this early Cenozoic event which resulted in successful sea-floor spreading. Burke and Dewey (1973) suggested that the Kangedlugssuak lineament, marked by a depression in bedrock under the icecap, is a rift dating from this time.

8.2.2. MESOZOIC

8.2.2.1. Late Cretaceous Rifting of Baffin Bay-Labrador Sea-N. Newfoundland Continental Margin

This episode of rifting (Figure 8.1) taking place about 80 m.y. ago (Pitman and Talwani, 1972) resulted in seafloor spreading in Labrador Sea and Baffin Bay. The rifted continental margins of western Greenland,

and Eastern North America along Baffin Island, to Flemish Cap northeast of Newfoundland were formed at this time, although it is possible that earlier rifting episodes, that did not lead to sea-floor spreading, had occurred along part of these margins. Several rifts that did not develop to the stage of sea-floor spreading are preserved from this episode, although most are poorly dated from direct evidence. These are the rifts that lie under the Greenland continental shelf in Baffin Bay (Burke, 1977); the faults that define the eastern margin of the thick sequence of sediments and basalt at Disko Island; Lancaster Sound (see discussion in Burke and Dewey, 1973); the rifts of northern and southern Baffin Island that contain Ordovician and some Silurian sedimentary rocks (Geological Map of Canada, 1969); the two gulfs of southern Baffin Island, and, possibly, the Flemish Cap rift although this more probably originated in the 135 m.y. old rifting episode that defined the Eastern Newfoundland continental margin (Section 8.2.2.2.).

8.2.2.2. Eastern Newfoundland Continental Margin

This short segment of continental margin (Figure 8.1) formed just prior to the start of sea-floor spreading at about 135 m.y. ago (Pitman and Talwani, 1972) between this part of North America and the Spanish/Portugese block. Evidence from Portugal demonstrates that this rifting selected the site of a Triassic/Jurassic (180 m.y. old) rift (Burke, 1977). The rift between the northern Grand Banks shelf and Flemish Cap (Flemish Cap rift) probably was active during this 135 m.y. old event (Late Jurassic-Early Cretaceous), although it too may have been defined earlier.

8.2.2.3. Eastern North American Continental Margin, Southern Grand Banks to Florida, and Associated Rifts

This rifted continental margin (Figure 8.1) was outlined by rifting between about 200-180 m.y., and sea-floor spreading forming the central Atlantic followed (Dewey et al., 1973). Numerous rifts of this age are preserved within the continent near this continental margin, of which the on-land Late Triassic/Earliest Jurassic graben of the eastern U.S. and Canada are the best known, although these are not the largest rift structures of this age.

8.2.2.4. Gulf of Mexico Margin and Associated Rifts

This margin (Figure 8.1) was formed by rifting at approximately the same time as the East Coast margin, 200-180 m.y. ago. The rift of the Mississippi embayment was apparently located on the site of an older Precambrian or Early Cambrian rift (Ervin and McGinnis, 1975).

8.2.2.5. Arctic Continental Margin and McKenzie Delta Rift

This margin formed by rifting about 180 m.y. ago (Early Jurassic) with the removal of the Kolyma block (Herron et al., 1974). The McKenzie delta lies above a rift (Figure 8.1) formed at this time.

8.2.2.6. East Greenland

Rifting in this area (Figure 8.1), which did not lead to sea-floor spreading, occurred in the Late Jurassic (Kimmeridgian) and in the Early Triassic. In each case rifting followed the old rift directions established in the Carboniferous (Haller, 1971).

8.2.3. PALAEOZOIC

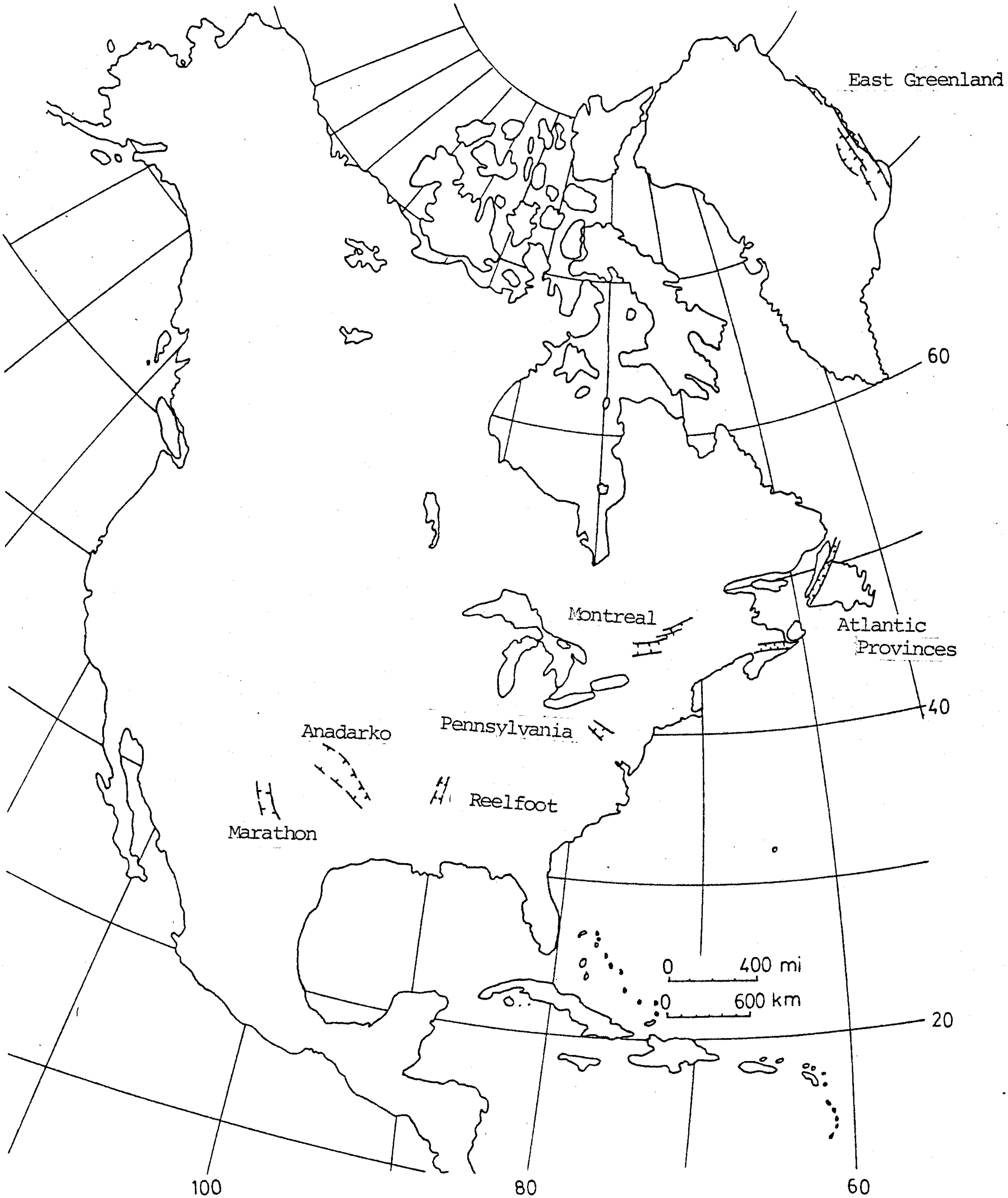

Figure 8.2. Palaeozoic rifts of North America

8.2.3.1. Carboniferous Grabens of the Atlantic Provinces

These grabens (Figure 8.2), associated with large strike-slip faults, were initiated in the early Carboniferous (Belt, 1969, Webb, 1969).

8.2.3.2. East Greenland

Rifting along north-south lines was extensive in this area of Greenland (Figure 8.2) during the Early and Middle Carboniferous (Haller, 1971). These rifts are subparallel to an earlier, Devonian rift that formed during the continental collision between Greenland and Scandinavia in Late Silurian-Early Devonian time.

8.2.3.3. Aulacogens Associated with the Opening of the Appalachian Ocean

These include the Ottawa-Bonnechere graben (or Montreal aulacogen) (Kay, 1942, Burke and Dewey, 1973), the Anadarko aulacogen (Ham and Wilson, 1967; Wickham et al., 1976), the Marathon aulacogen, a rift under the site of the Mississippi embayment (Ervin and McGinnis, 1975), and a possible rift in Pennsylvania (Rankin, 1976). All of these rifts (Figure 8.2) were active in the Latest Precambrian-Early Cambrian, although some may have suffered an earlier episode of rifting about 800 m.y. ago (Rankin, 1976).

8.2.4. PRECAMBRIAN RIFTS

8.2.4.1. Late Proterozoic Rifts on the Future Site of the Appalachians/ Caledonides

Rankin (1976) suggests that thick sediments and volcanics about 800 m.y. old in the Southern Appalachians were deposited in a rift

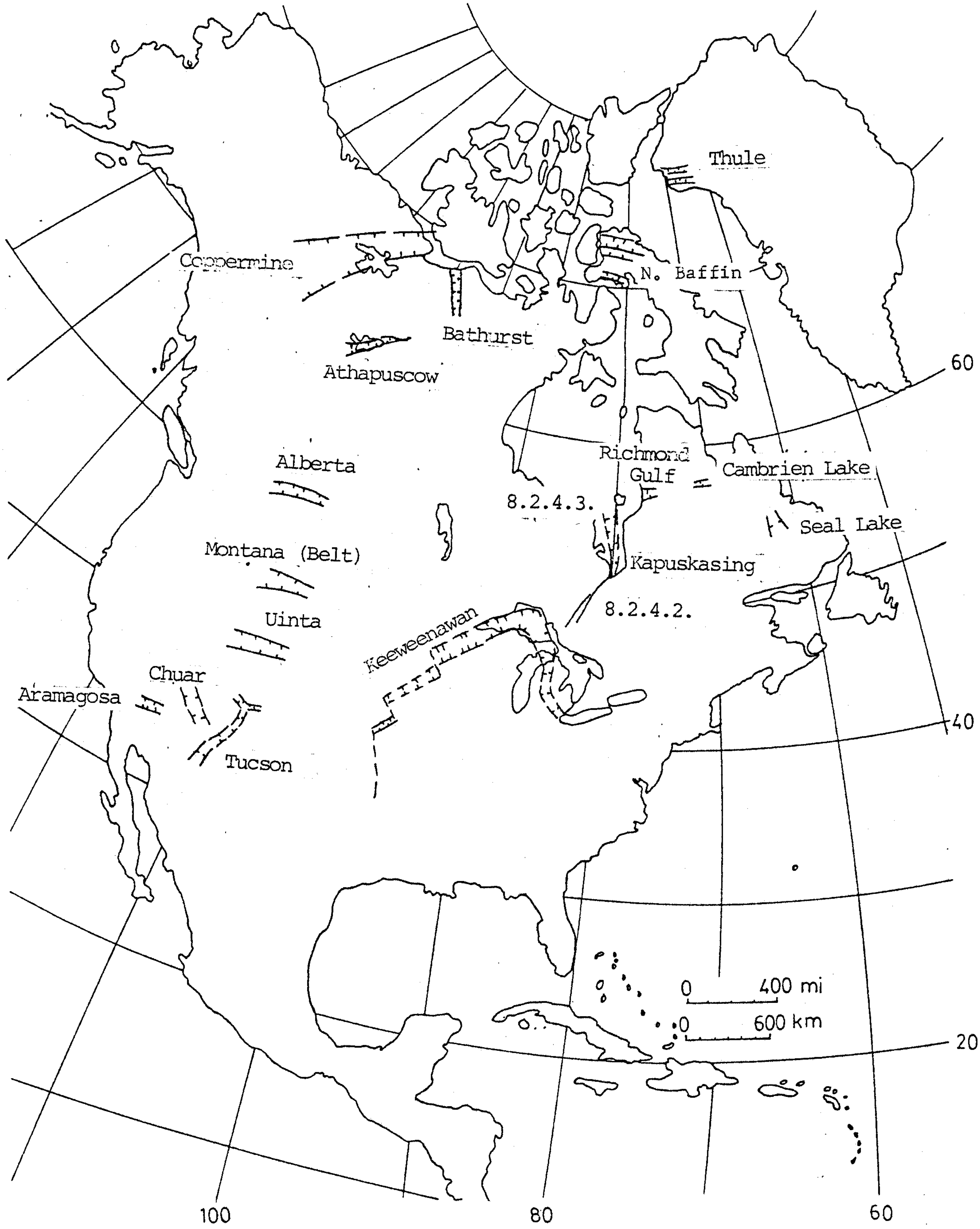

Figure 8.3. Precambrian rifts of North America

complex that subsequently became the site of the Appalachian Ocean. These are not shown on the map because they have been strongly deformed and are no longer in a more-or-less undisturbed rift setting.

A rift of about the same age is indicated by a 15 km thick sedimentary accumulation of the Eleanore Bay Group in East Greenland (Haller, 1971). This also has been deformed to varying degrees in the Caledonian orogeny and so cannot be thought of as an undisturbed rift. It lies on the site of subsequent rifts of Devonian through Jurassic age and for both these reasons is not plotted on the map.

8.2.4.2. Keeweenawan-age Rifts

Where well-dated (Chase and Gilmer, 1973; Obradovich and Petermann, 1968; Ford et al., 1972) these rifts (Figure 8.3) are usually found to be fairly close to 1150 m.y. old, although some may have had activity up to 1.3-1.4 b.y. ago. Burke and Dewey (1973) summarised information on these rifts and suggested that they occurred at the time the Pacific margin of North America was defined by rifting and removal of another (unidentified) continental fragment. Rifts of this episode include the Keeweenawan and its extensions under Michigan and the mid-continent gravity high. The southern Kapuskasing structure may form a "third arm" to these two rift arms (Burke and Dewey, 1973). The Gardar igneous province of southern Greenland is clearly linked to this rifting episode, but there is no well-defined rift present in this area. Other rifts of this age (Figure 8.3) are Seal Lake (Labrador), the rifts of northern Baffin Island and northwestern Greenland, the Coppermine-McKenzie Mountains rift, the Alberta buried rift (Kanasewich et al., 1969), the Belt and Uinta rifts (Stewart, 1972, 1976), the Aramagosa rift of Death Valley area (Wright et al., 1974) and the Chuar and Tucson rifts (Stewart, 1972).

8.2.4.3. Rifts About 2150 M.Y. Old

The evidence for rifting and establishment of two separate continental margins of this age is unusually well-preserved in the North American shield. The aulacogens leading into the Coronation Orogen comprise one such area; the Athapuscow aulacogen in the East Arm of Great Slave Lake, and the Bathurst aulacogen at the other (northern) end of the exposed part of the orogenic belt have been described by Hoffman (1973) and Hoffman et al., (1974). Along the other well-preserved orogenic belt of this age, the Labrador trough and its extensions, two small rifts exist at Cambrien Lake (Fahrig, 1969) and Richmond Gulf (Stevenson, 1968) striking at a high angle to the adjacent orogenic belt. Another rift of this age is postulated to underlie James Bay as a northern extension of the Kapuskasing structure (Figure 8.3).

8.3. NORTH AMERICAN SUTURES

8.3.1. MESOZOIC-CENOZOIC SUTURES

8.3.1.1. Guatemala

This suture (Figure 8.4), a continuation of 8.3.1.2. (q.v.), is marked by a prominent line of obducted ophiolites (see Section 10.3). It formed by collision between the Yucatan and Honduras blocks in the Latest Cretaceous-Earliest Cenozoic (Burke and Fox, 1977). It has subsequently been used by more recent transform fault motion as the Caribbean-North American plate boundary.

8.3.1.2. Cuba-Puerto Rico

This suture (Figure 8.4) involved collision in the Eocene between the Bahama-Yucatan platform, (a carbonate accumulation built, in the case of the Bahamas, on a hot-spot volcanic accumulation over oceanic crust; Dietz, 1973), and a Cretaceous island arc that forms the backbone of Cuba and Hispaniola. This part of this suture zone involved southward subduction of ocean under the arc; Burke, Fox and Şengör, (1978) have suggested that the opposite polarity of subduction was responsible for the arc rocks of Jamaica, southern Hispaniola and Puerto Rico.

8.3.2. MESOZOIC-LATE PALEOZOIC SUTURES

8.3.2.1. Klamaths-Sierras-Franciscan, British Colombia

Convergent tectonic conditions have existed along the western margin of North America (Figure 8.4) from the Devonian onwards. Highly complex geology has resulted from the construction of island

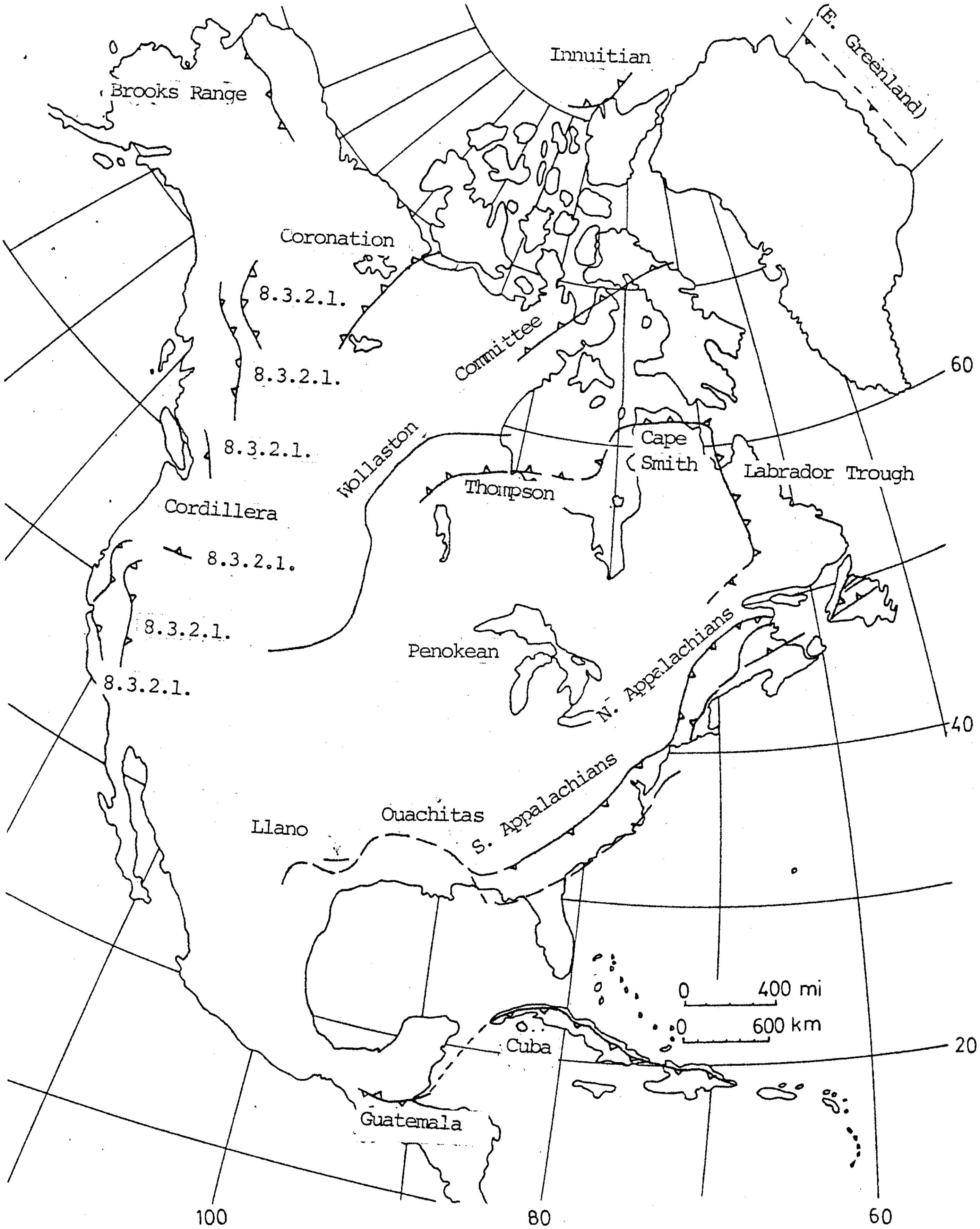

Figure 8.4. Sutures of North America

and Andean-type arcs, and their dissection by transcurrent faulting of different ages, and of large magnitude, subparallel to the arcs. Ophiolite belts within the Cordillera of western North America are all thought to represent sutures between arcs, or between an arc and the continental margin; no major sutures between significant size pieces of continent, or arcs far travelled across the paleo-Pacific Ocean, are known to exist here. The sutures are therefore the product of the closure of marginal basins (Burchfiel and Davis, 1972, Monger, et al., 1972, Schweikert and Cowan, 1975). Some of these "mini-sutures" have produced geology that greatly resembles that of Archean greenstone belts (Burke et al., 1976). Most of these mini-sutures are of Permian to Jurassic age.

8.3.2.2. Brooks Range, Alaska

A suture of early Cretaceous age (Figure 8.4), exists along the northern margin of the Brooks Range (J. Martin, pers. comm.). This is between the continental block to the north that Herron et al. (1974) suggested was part of the Kolyma block prior to the Jurassic, and the small continental fragment of the Seward Peninsula and Chukotsk. The trend of the suture west of the Bering Strait under the Siberian continental shelf is unknown, but it does not appear to come on land.

8.3.3. PALEOZOIC SUTURES

8.3.3.1. Southern Appalachians-Ouachitas

Suturing of Africa plus South America to North America (plus Europe) took place in the latest Carboniferous (Dewey and Burke, 1973). We have drawn the line of suturing down the Brevard Zone of the

Southern Appalachians (Figure 8.4); this could be an Acadian sutural structure and the main suture may lie eastward under the coastal plain or continental shelf. The voluminous Carboniferous flysch of the Ouachitas, interpreted by Graham et al., (1975) to have been deposited on ocean floor, suggests that the suturing in the Ouachitas may have been a little later than in the Southern Appalachians, if, as seems likely, the sediment was derived from the latter. The occurrence of tectonic serpentinite slivers in the flysch (Sterling and Stone, 1961) support the idea that the flysch was deposited on oceanic crust. The suture here, like the Siluro-Devonian suture in southern Scotland (Section 7.3.3.), is not a sharp feature, but is buried under the very thick deformed flysch that filled the closing ocean in its final stages of existence.

8.3.3.2. Northern Appalachians

Suturing occurred within the northern Appalachians (Figure 8.4) between North America and a prong extending from Western Europe in the Middle Devonian (Bird and Dewey, 1970; McKerrow and Ziegler, 1972; Dewey and Kidd, 1974). This followed a number of marginal basin opening events that produced mini-sutures in the Ordovician and Middle Devonian. The main suture can be distinguished from these by its continuity down the center of the orogen.

8.3.3.3. East Greenland

The rifting site on which the northern Atlantic opened between Greenland and Norway closely followed the site of the Late-Silurian-Early Devonian suture between the same two continental parts (Figure

8.4). Although this suture can be followed across Ireland and Britain, and, on faunal evidence, appears in a small zone around Bergen, Norway, it is not clear whether the Early Cenozoic rifting left it on the Greenland or on the Scandinavian side north of the position of Bergen. It may be that segments were left on both sides, but as they are buried under the thick sediments of the continental shelves, this cannot be established by direct observation.

8.3.3.4. Innuitian Orogen

Trettin et al., (1972) pointed out that a collision with a continental block in Mid-to Late-Devonian time is required for this orogenic belt. The suture zone may be just exposed where ultramafic rocks occur on the northernmost parts of Ellsmere Island. Alternatively, these may have resulted from a "minisuture" involving an arc and marginal basin closing, and the main suture may be buried under the Arctic continental shelf, or partly or wholly removed with the Kolyma block in the Jurassic (Herron et al., 1974) and now be buried under the Siberian continental shelf.

8.3.4. PRECAMBRIAN

8.3.4.1. Llano Serpentinite Belt

This structure (Figure 8.4) is probably a suture (Şengör, pers. comm.), but its age is not well constrained. It could be a short length of the main Grenville suture (~1000 m.y. old), or it could be an older suture zone caught within the area of Grenville age reactivation.

8.3.4.2. Labrador Trough-Cape Smith Belt-Nelson/Thompson Front

This belt (Figure 8.4) has been interpreted as containing a suture by Wilson (1968), Gibb and Walcott (1972) and Burke and Dewey (1973). Geological data for parts of this zone are given by Dorr and Laurin (1971), Dimroth, (1972), Baragar (1974) and Coats et al., (1972). A discussion of critical parts of this evidence is given by Burke et al., (1977); the most convincing aspect is the occurrence on the west side of the Labrador trough of a series of shallow water marine sediments in a miogeoclinal wedge that overlie a thick sequence of arkosic clastics and basalt. This sequence is exactly like that formed by rifting and establishment of an Atlantic-type aseismic continental margin. The rocks were subsequently involved in a flat thrust and fold belt, such as is found for the miogeoclinal parts of younger orogenic belts across which continental collision has occurred. Dolerite dikes paralleling the northern Labrador trough and swinging to follow the Cape Smith belt have ages of 2150 m.y. (Stevenson, 1968), which we take to be the age of opening of an ocean on the site of this orogenic belt. The suturing seems to have taken place, probably at slightly different times in different sectors of this belt, between 1700-1800 m.y. ago. This resulted in the reactivation of the colliding continent to the north and east of this zone. The other continental piece, the Archean Superior Province, remained unreactivated in the part that remains today after escaping the effects of younger orogenies.

8.3.4.3. Committee Fold Belt

This fold belt, running through northern Baffin Island and Melville Peninsula (Figure 8.4), consists dominantly of miogeoclinal type rocks but with a narrow belt of ultramafic and mafic rocks, that we interpret as a suture, along its northern side (Jackson and Taylor, 1972). It is poorly dated, except that it lies within the Hudsonian province that gives reactivation ages of 1700-1800 m.y. Suturing across this belt between two continental fragments whose extent is not presently determinable probably took place during this time interval. Other sutures are very likely to exist within the very extensive Hudsonian terrain of the Canadian shield but present knowledge is not adequate to identify any others except the Wollaston Lake belt (Section 8.3.4.4.). In this respect the Hudsonian terrain of North America is rather similar to the younger extensive Pan-African reactivated terrain of Africa.

8.3.4.4. Wollaston Lake Belt

Camfield and Gough (1977) identify this as a suture (Figure 8.4) and trace it south beneath platform sediment cover from its geophysical signature. We agree with its identification as a suture and suggest that it can be extended further to a known age-province boundary in Wyoming. Its age is also Hudsonian, that is 1700-1800 m.y. old. The continental fragments colliding to form this suture, if it is as long as we propose, are likely to have been large.

8.3.4.5. Penokean

VanSchmus (1976) has proposed that the Hudsonian-age terrain of the Penokean in Wisconsin was due to Andean arc activity. Whether it was due to this, or to Tibetan-style reactivation, there must be a suture of about this age to the south as there does not seem to be one against the Superior Province block to the north. However, the position of such a suture is very poorly constrained as all possible sites are covered by platform sediments. It has therefore not been plotted on the map.

8.3.4.6. Coronation Orogen-Wopmay Suture

A very fine example of an Atlantic-type miogeoclinal sediment sequence developed after rifting 2100 to 2300 m.y. ago has been described from the area east of Great Bear Lake (Figure 8.4) by Hoffman (1973). Burke et al., (1977) suggested that a suture exists to the west of a metamorphic belt that adjoins the now-deformed miogeoclinal sequence, along the line of the Wopmay fault. Hoffman et al., (1978) favor a site for a suture within the metamorphic belt.

8.3.4.7. Superior Province Greenstone Belts

These belts run in a generally east-west direction within the western and southern Superior Province. Burke et al., (1976) suggested that many of these belts represent small sutures due to driving out of marginal basins by arc collisions. Because they are numerous, and definitive data is not available for most of them, they are not shown on the map.

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9.0. SOUTH AMERICA

9.1. INTRODUCTION

The western margin of South America has faced ocean since the Early Cambrian and, like the corresponding margin of North America, has been involved in convergent tectonics of Andean and island arc type since the Palaeozoic without the occurrence of any major continental collisions.

The eastern and northeastern margins of South America were blocked out by rifting in the Early Cretaceous, and numerous rifts of this age exist near this margin. This rifting episode was the start of the spreading in the South Atlantic that separated South America and Africa.

The pre-Mesozoic geology of the South American platform is poorly known, mostly because of the extensive sedimentary cover.

9.2. RIFTS (Figure 9.1)

9.2.1. ACTIVE RIFTS

9.2.1.1. The Chiquitos Graben

Wilson (in Jacobs et al., 1959, p. 301) drew attention to this structure trending away from a bend in the Andes. Whether it is active now and whether it is an aulacogen or an impactogen is not clear from available information.

9.2.1.2. Andes

Active rift zones among the Andean volcanoes have not been identified but these features are sufficiently common in other Andean arcs (e.g., New Zealand, Central America) that they are likely to be identified in the future.

9.2.2. MESOZOIC RIFTS

Associated with the breakup of Pangea no Cenozoic rifts, other than those currently active, are identified in South America. Mesozoic rifts of the east coast are associated with the rupture of Gondwana and the separation of Africa from South America. Those of the north coast include an eastern set associated with the opening of the Gulf of Guinea and a western set related to the formation of the Caribbean. Mesozoic rifts of the west coast have been involved in later convergence and are not considered in this section.

9.2.2.1. Cretaceous Rifts of the East Coast

The southernmost rifts are those of Argentina (the Malvinas and San Jorge plotted as 9.2.2.1.) (Urrien and Zambrano, 1973) and as

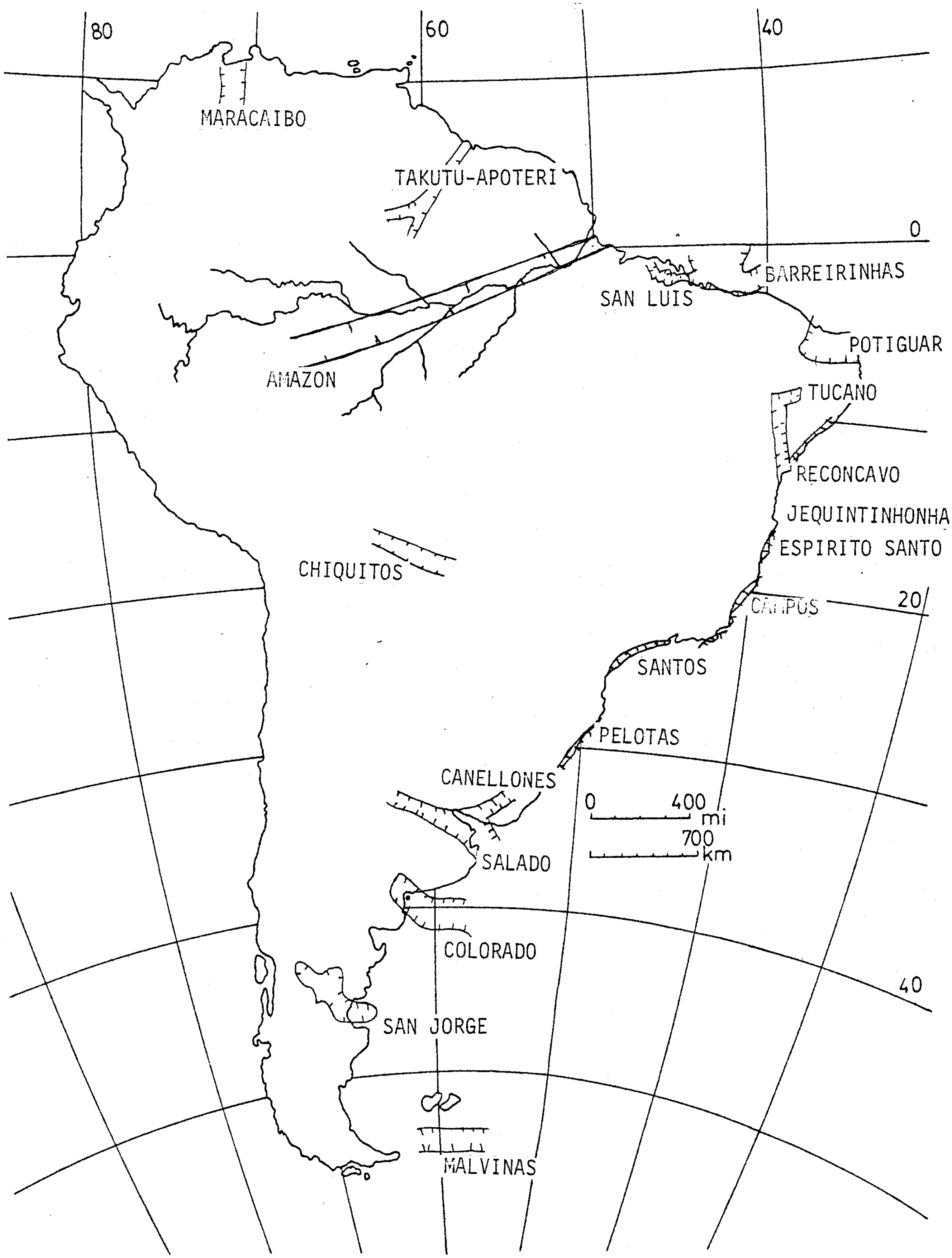

Figure 9.1. Rifts of South America.

Menard (1973) pointed out two of these, the Colorado (9.2.2.1.1.) and Rio Salada (9.2.2.1.2.) graben, trend at right angles to the continental margin as would be expected on the triple-rift model of Burke and Dewey (1973). Newton (1976) has made the same suggestion about a third graben. When Africa and South America are brought close together the mouths of two graben, each containing over 5 km of sediment, match closely the ends of an offshore elongate rift-like basin mapped by Emery et al., (1975) on the South African side. Bolli et al., (1975) reported drilling into this structure and penetrating Lower Cretaceous sapropel-rich rocks. A fourth rift--The Canellones graben (9.2.2.1.3.) (Jones, 1956)--completed the pattern of rifts marking the site of the southern end of the future Atlantic Ocean. Jones (1956) reported the results of gravity surveys and exploratory drilling in the Canellones rift. His figures show that it is a classic example of a rift which has reached the stage of axial dike emplacement but has not spread (Burke and Dewey, 1973, p. 406). More than 2 km of non-marine sediments overlying basalt were encountered in the Sauce well on the rift flank, but the Santa Rosa well, drilled over the axial positive gravity anomaly, encountered only a few hundred meters of intercalated lavas and sediments.

Along 200 km of the east coast of Brazil seven sedimentary basins have been distinguished that include graben complexes formed at rifting. From south to north these are the Pelotas (9.2.2.1.4.), Santos (9.2.2.1.5.), Campos (9.2.2.1.6.), Espirito Santo (9.2.2.1.7.), Jequitinhonha (9.2.2.1.8.), Reconcava-Tucano Sul (9.2.2.1.9.) and Sergipe-Alagoas (9.2.2.1.10.) basins (Asmus and Ponte, 1973; Campos et al., 1974). The last two are the best known and maps of their basement structure (in Asmus and Ponte, 1973) based on oil well and reflection seismic data resemble the structure of the Ethiopian Afar and the East African rifts with internal horsts resembling the Danakil and step-faulting within the graben and perhaps monoclinal warping or topographic burial yielding slopes locally exceeding 5 km in 20 km at the basin margins. The Reconcavo and Tucano Sul graben meet at a triple-rift junction. All the graben complexes of eastern Brazil lie in the area of deposition of Aptian salt north of the Rio Grande ridge (Burke, 1975) and in all, with the exception of the Reconcavo basin (where there is no salt), Aptian salt overlies the volcanic and non-marine sedimentary graben-fill with some degree of unconfirmity. Graben faulting did not generally persist after the deposition of the Aptian salt and, as mapped in the Sergipe basin, the salt overlies both horsts and graben.

This pattern of salt distribution has been attributed to the existence of the graben as sub-sea level depressions flooded episodically by salt water spills from the southern ocean (Burke, 1975).

9.2.2.2. Cretaceous Rifts of the North Coast of Brazil

These rifts are associated with the opening of the Atlantic and the Atlantic opening of this sector has been dominated by transform movement associated with the giant equatorial fracture zones. This is one of the best places in the world to seek distinctive features of continental margins associated with transform motion.

Asmus and Ponte (1973) illustrated compressional folds and reverse faults in the Barreirinhas basin (9.2.2.2.1.). These structures are consistent with the model of LePichon and Hayes (1971) in which movement in the Gulf of Guinea was constrained as long as transform motion was intra-continental but became freer, requiring a new pole to describe it, once transform motion was entirely within oceanic lithosphere.

The structural complexity associated with intra-continental transforms is familiar in such features as the mountains associated with compressional segments along the San Andreas fault (the Transverse Ranges) and the Dead Sea transform (the Anti-Lebanon). Pull-aparts such as those of the Dead Sea and Death Valley indicate that very rapid basin filling is to be expected in the local tensional segments along the transforms. The thick non-marine section above the volcanics in the Barreirinhas basin may have accumulated in this kind of environment. The sedimentary and igneous rocks accumulated in a transform pull-apart are unlikely to be very different from those in other graben. One possible means of distinction is that the deposits associated with transforms are likely to suffer compression shortly after deposition.

The Gulf of Guinea and North Brazilian basins, apart from the Barreirinhas basin are generally without volcanics and this makes analysis of the timing of their development difficult. Earliest Cretaceous seems the most likely time of initial rifting. The Potiguar (9.2.2.2.3.) basin although trending parallel to the early transform direction appears structurally less complex than its neighbors. This may be a result of its relatively late initiation in Aptian-Albian times. In general, the oldest marine rocks in the Gulf of Guinea and northern Brazil are Albian in age and there is no Aptian salt (Burke, 1975).

9.2.2.3. North Coast Jurassic Rifts

Only two rift systems are plotted related to this opening event. Cretaceous convergence and Cenozoic transform motion in Venezuela and Colombia have probably helped to obscure much of a complex history.

The occurrence of evaporite in the Gulf of Mexico leads to the idea that there might be similar evaporites off northern South America but none have yet been recognized except in the Gulf of Paria and near Port-of-Spain, Trinidad.

9.2.2.3.1. Takatu-Apoteri Rift System

The Takatu-Apoteri rift system (Berrange and Dearnley, 1975) is better known within the continent than close to the coast. Its dating is poorly defined within the later Mesozoic, but a Jurassic age seems likely.

9.2.2.3.1. Maracaibo Rift

The Maracaibo rift is very poorly defined but its occurrence distinguished here for the first time seems required by both the shape of the fold belt and the occurrence of Jurassic red beds.

9.2.2.4. Paleozoic Rifts

Most Paleozoic rifts in South America have been involved in later convergence. The Amazon valley separating the Brazilian and Guyana shields appears to be the only prominent rift-like feature preserved.

9.2.2.4.1. Amazon Valley

This is considered to be a rift with Silurian to Pennsylvanian activity because modern subsurface studies (Bigarella, 1973) show that many units with ages in this range are very thick and were deposited in an elongate depression roughly parallel to the modern Amazon river valley. Virtually, no accompanying igneous activity has been reported and for this reason the rift nature may be challenged. However, subsidence rates were fast enough to imply a rift-like thermal structure and there is some evidence of unusual gravity high along the valley (deLoczy, 1970).

Mesozoic reactivation close to the Amazon mouth is related to the opening of the South Atlantic.

9.3. SOUTH AMERICAN SUTURES (Figure 9.3)

9.3.1. CENOZOIC SUTURES

No sutures of Cenozoic age are known within South America.

9.3.2. MESOZOIC SUTURES

9.3.2.1. Caribbean Mountains

This suture represents the line of closure of a marginal basin developed between the northern margin of South America and an island arc that subducted Caribbean oceanic crust. The closure of this marginal basin occurred in mid-Cretaceous time (Maresch, 1974).

9.3.2.2. Rocas Verdes (Southern Chile)

A narrow belt of ophiolitic rocks lying east of the batholith belt of southern Chile represents the remains of a marginal basin that opened behind the volcanic arc in the earlist Cretaceous and closed again by mid-Cretaceous times (Dalziel et al., 1974). This basin was originally very narrow and almost rift-like at its northern end but much wider to the south.

9.3.3. PALAEOZOIC SUTURES

A Palaeozoic orogenic belt forms the basement to much of Argentina, although it is almost wholly covered by younger sedimentary rocks. DuToit (1937) showed that the northern boundary of this belt runs through the Sierra de la Ventana trending WNW; when it meets the eastern Andes it turns and runs north parallel to them. This orogenic belt represents accretion to the western margin of South America; it continued originally through the Palaeozoic orogenic belts of West

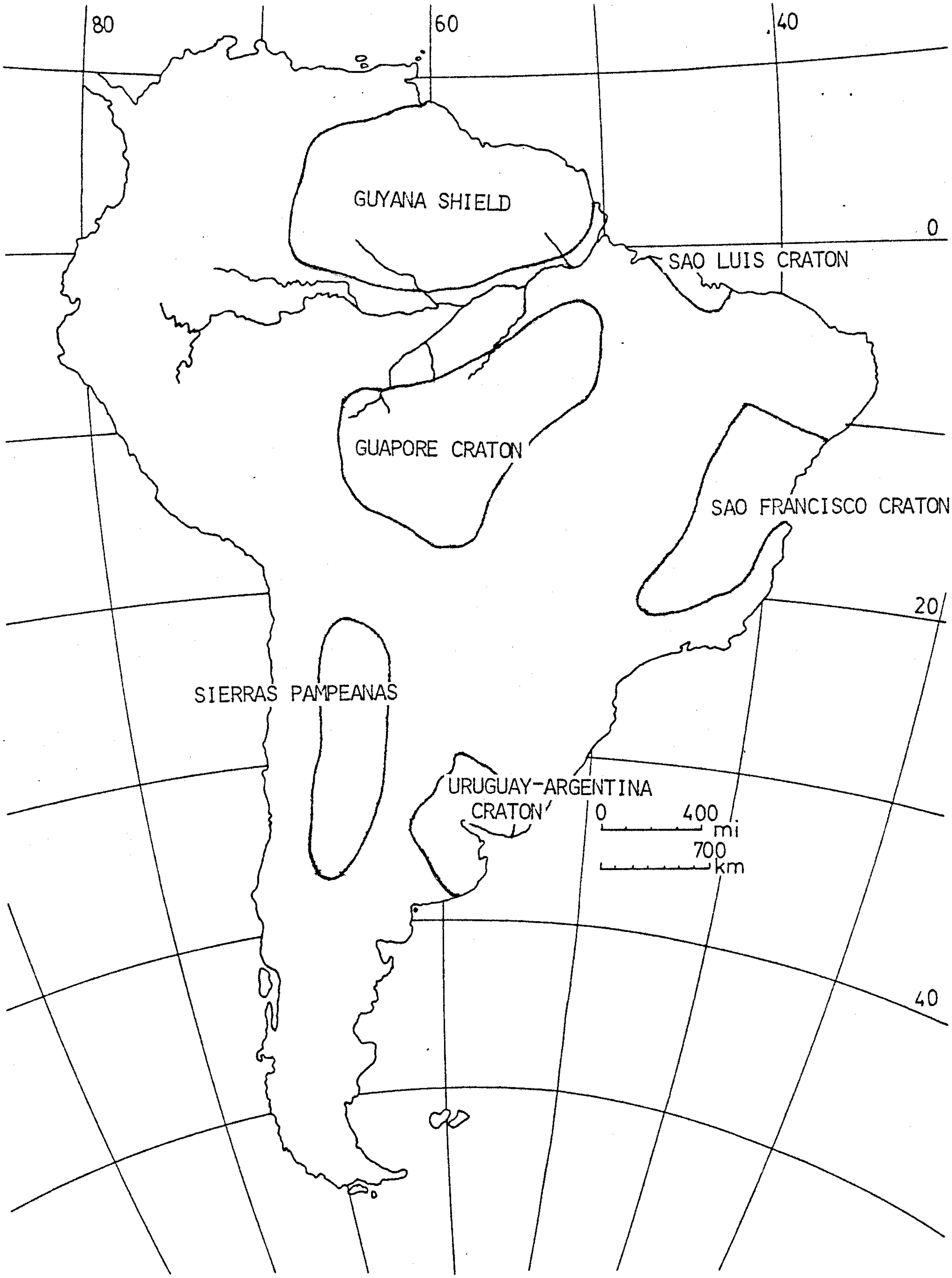

Figure 9.2. Cratons of South America.

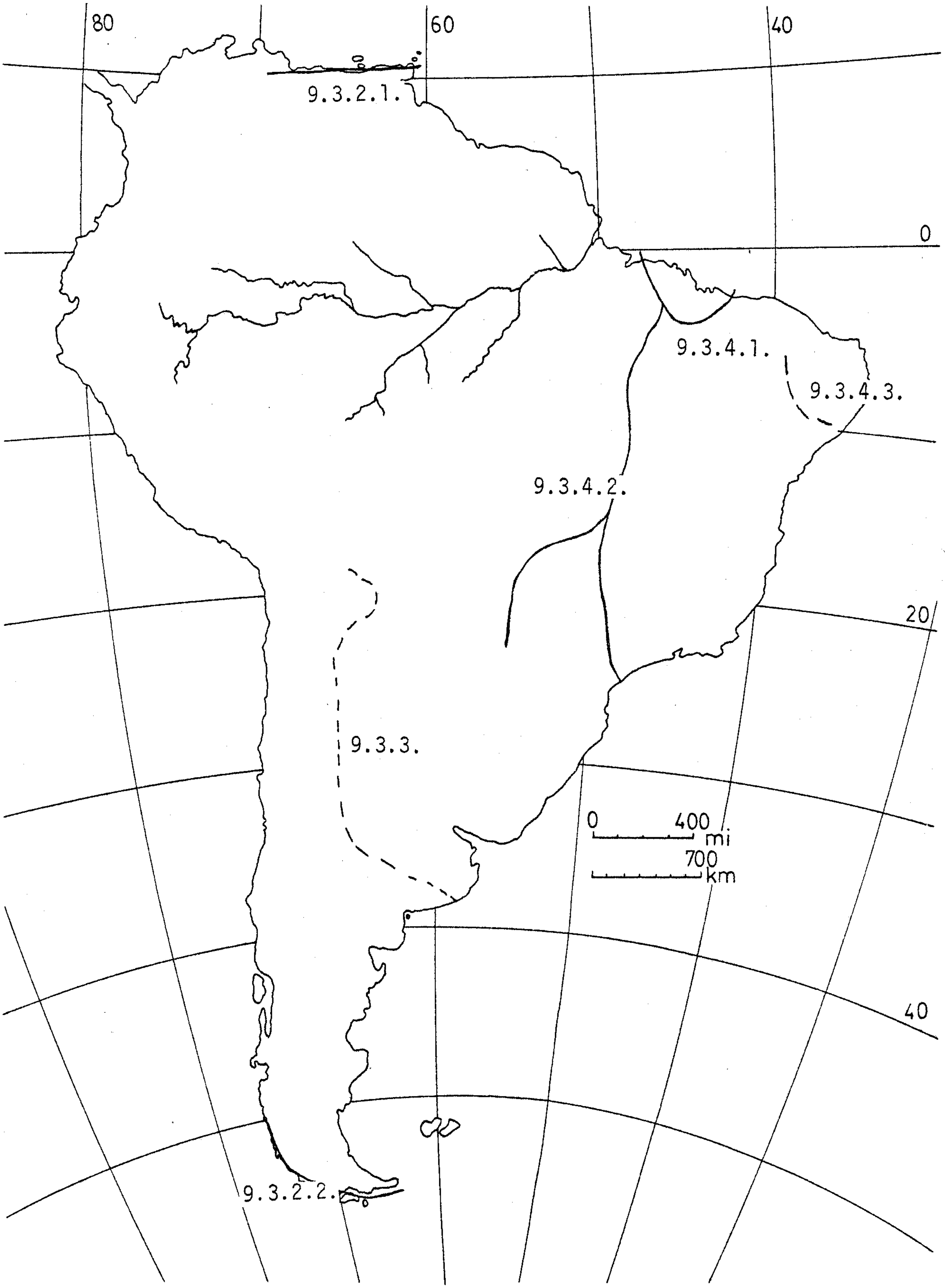

Figure 9.3. Sutures of South America.

Antarctica and the Tasman belt of eastern Australia. Like the latter, it contains, where it is exposed, dismembered ophiolites (Borello, 1972); that probably represent marginal basin sutures. As so very little of this terrain is exposed, such sutures cannot be plotted, but they probably run approximately parallel to the northern boundary of this terrain, which is shown as 9.3.3. on the map (please note that 9.3.3. is not itself a suture).

9.3.4. PRECAMBRIAN SUTURES

9.3.4.1. Sao Luis Craton Boundary (Figure 9.2)

This boundary can only be approximately defined (Cordani, et al., 1968), because of young sedimentary cover. It is known that the continuations of this boundary found in West Africa around the West African craton (Section 2.3.2.1.), are both sutures and therefore it must be also. The age of suturing is most probably Late Precambrian; radiometric ages up to Early Ordovician obtained from the adjacent orogenic belt (Cordani et al., 1968) are probably measuring post-collisional uplift.

9.3.4.2. Brasilia Orogenic Belt and Extensions

An orogenic belt of Late Precambrian age runs near north-south to about 16°S between the Guaporé craton and the San Francisco craton of eastern Brazil. It contains a prominent belt of dismembered ophiolites (Quade and Stache, 1972; Cordani et al., 1972) that we interpret as a suture. This suture must meet that around the Sao Luiz craton at a suture 'knot'. South of about 16°S, the center of this orogenic belt is hidden by sediments of the Parana basin. However, the margins of the belt diverge widely, and both are thrust fronts verging away from the belt. This suggests that two main sutures exist from this point, another suture 'knot', and follow the same trends as the thrust margins of the belt. These must be in places that are now largely covered with younger sedimentary rocks of the Parana basin, and their location is accordingly imprecise. The eastern suture runs SE into an orogenic belt, situated along the coast, also of Late Precambrian age, that cuts the fold trends associated with the eastern suture. We conclude that this coastal fold belt is due to a slightly younger continental collision; its suture is buried either under the continental shelf of eastern South America, or of Africa (Section 2.3.2.4.).

9.3.4.3. Northeast Brazil

Thrusts verging onto the northeastern margin of Sao Francisco craton are shown on the tectonic map of Brazil (1971), together with some probably ophiolitic ultramafic bodies nearby in the Late Precambrian (600-700 m.y.) orogenic belt. Jordan (1972) gives further data on this belt, which suggests a suture trending WNW is present.

9.3.4.4. Sierras Pampeanas

The Sierras Pampeanas are Late Precambrian (600+m.y.) orogenic rocks (Borello, 1972, Cordani et al., 1972) that may be related to the westernmost suture of the southern Braziliane terrain (9.3.4.2.), or to another unknown suture. It is probably less likely, although not impossible, that they are related to subduction along the Pacific margin of South America (Burke et al., 1977), because they are overlain by a miogeoclinal sequence (Borello, 1972) like that of other present and past rifted margins.

9.3.4.5. Older Precambrian Cratons

The three large cratons of South America contain rocks that mostly yield ages of 1800-2000 m.y. In the cratons, the tectonic grain trends north to northwest. The geology of these areas is poorly known; no good evidence is yet known of any sutures within them, although ultramafic rocks plotted in the southern part of the Sao Francisco craton (Tectonic map of Brazil, 1971) may belong to a suture zone which trends approximately N-S.

9.4. BIBLIOGRAPHY

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10. CENTRAL AMERICA

10.1. INTRODUCTION

Central America and the Caribbean islands are mainly made up of arc material of Cenozoic and Mesozoic age. This material lies between continental North America and continental South America. Only one continental object of moderate size, the Chortis block (Figure 10.1), has been recognized within Central America.

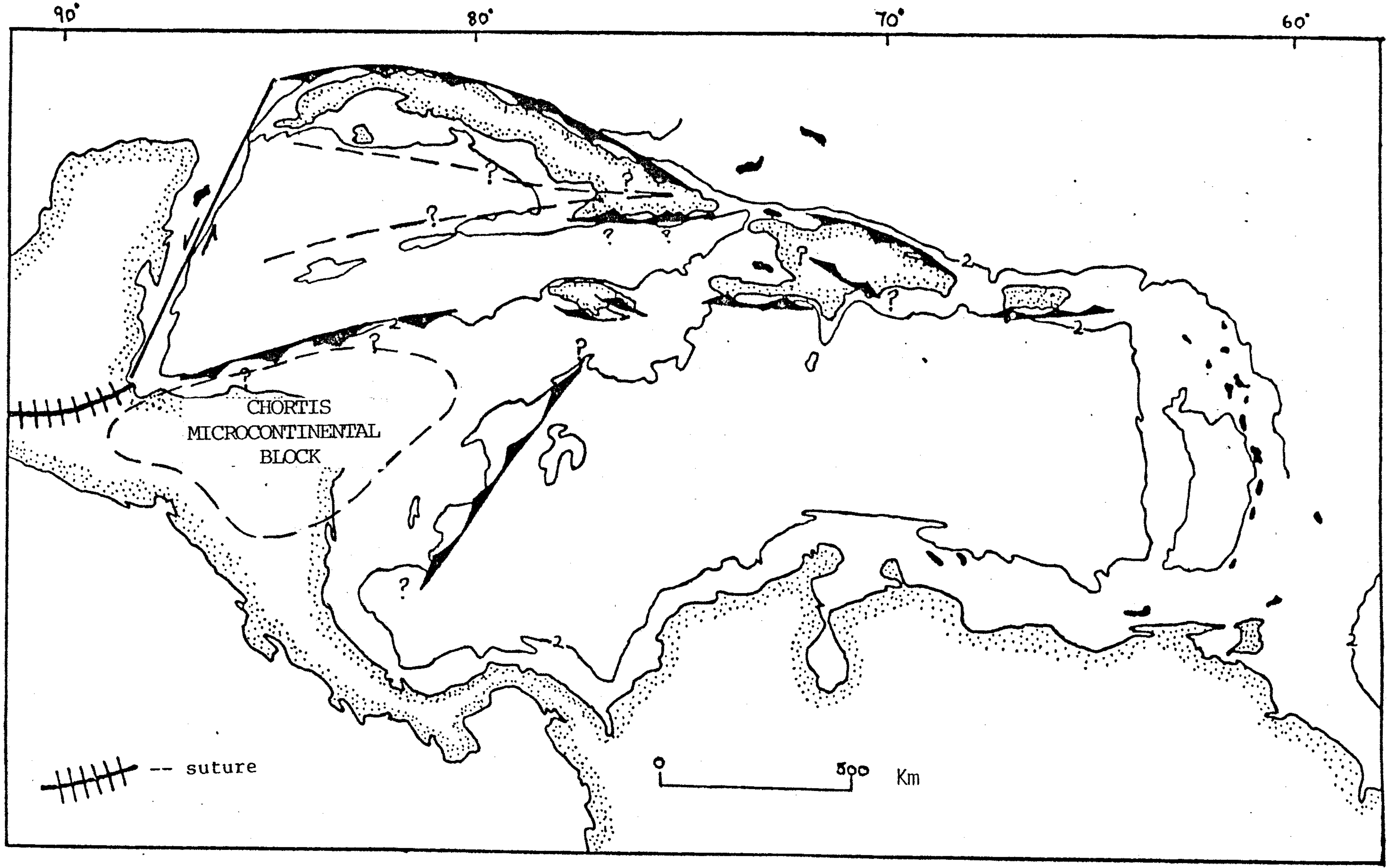

Figure 10.1. The Chixoy-Polochie-Motagua-Jocotan-Chamelecan suture zone of Guatemala separates the North American continent from the Chortis microcontinental block. Suturing occurred in Late Cretaceous time. Barbed lines indicate subduction zones active in the northern Caribbean during the Late Cretaceous.

10.2. RIFTS

10.2.1. ACTIVE RIFTS

Very numerous small active rifts occur in the northern part of the Caribbean plate, and they are interpreted as products of internal deformation (Figure 10.2). These rifts are best known in Guatemala where they have been sites of earthquakes, but they also occur as far east as Hispaniola where alkali basaltic volcanism is associated with them. Active spreading is restricted to the mid-Cayman rise (Figure 10.2).

10.2.2. OLDER RIFTS

Some Cenozoic rifts are reported from Cuba (Iturralde-Vinent, 1977) but little is currently known. Most older rifts, e.g., the Wagwater trough, (Burke and Fox, 1977) have been involved in later compression.

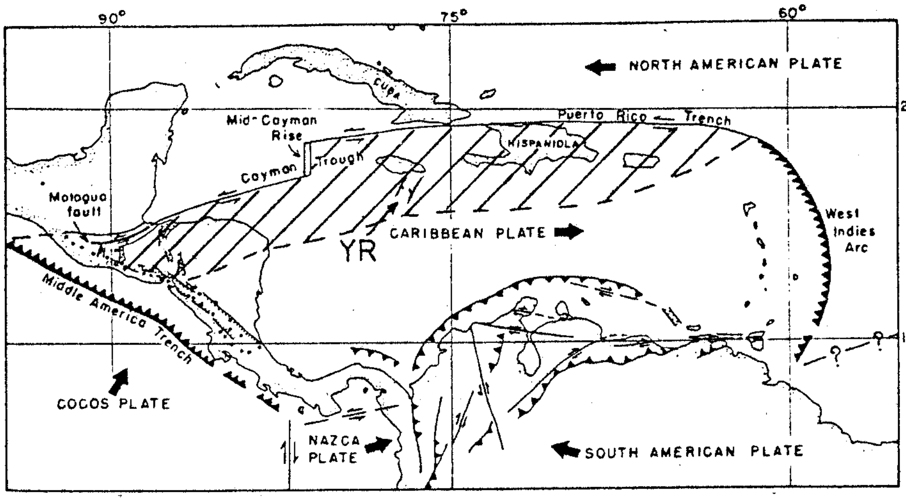

Figure 10.2. Although the North American plate behaves rigidly as it slides past the Caribbean plate, the northern part of the Caribbean plate is suffering internal deformation especially in the shaded area. Also, active rifts are common around 90 W. and in many other areas; only the Yallahs rift (YR) is shown. These active rifts may be interpreted as Riedel shears. The rift trending northwest along the volcanic arc in Nicaragua and El Salvador (at 10 N., 85 W. on map) is a volcanic crestal rift like that in New Zealand (Section 1.2.1.2.).
(based on Plafker, 1976)

10.3. SUTURES

10.3.1. MESOZOIC SUTURES

The collision of the Chortis block (Dengo and Bohnenberger, 1969) with North America in the latest Cretaceous (Figure 10.1) forms the only known suture, the Chixoy-Polochie-Motagua-Jocotan-Chamelecan suture, in the area (Burke and Fox, 1977). The southern margin of the Chortis block has accreted Late Cretaceous arc material, and we do not consider this structure a suture.

The northern boundary of Cuba where the Cuban island arc collided with the Bahama limestone block is a pseudo or mini-suture, not a true inter-continental suture.

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