CHAPTER 6

EXOTIC MEMBERS OF THE PBOC

In addition to relatively abundant fragments of ophiolitic lithologies (Chapter 5), the PBOC serpentinite-matrix melange also contains: (1) blocks which appear to have been derived tectonically from the adjacent rock associations; and (2) a range of amphibole-rich, relatively higher-grade metamorphic lithologies. Of these exotic blocks, only a selection of the fresher, relatively higher-grade types are examined in any detail (Section 6.2), and these only in a reconnaissance fashion.

6.1 LOW-GRADE, LOCALLY-DERIVED BLOCKS:

These include: (i) basaltic rocks, chert, jasper, siliceous argillite and rare volcanogenic epiclastics (e.g. GR634,823) which appear to have been derived from the Myra beds or the Glen Ward beds; (ii) tonalitic and hornblendite blocks from Pola Fogal Suite; and (iii) rare intermediate-silicic volcanics (e.g. GR650,815) which are probably derived from the Pitch Creek Volcanics. The locations of many of the more substantial tectonic blocks (> several tens of square metres in outcrop) are marked on Map 1. Some of these blocks have ostensibly eroded out of the surrounding schistose serpentinite, but most remain at least partially embedded in a matrix of highly schistose serpentinite (e.g. Plate 6.1A,B). Altered basaltic rocks are by far the most abundant of the block lithologies listed above and these are especially common in the PBOC schistose serpentinites lying along the Peel Fault System (Map 1).

Many of the locally-derived blocks have experienced at least some low-grade alteration over and above that developed in similar lithologies more-or-less adjacent to the PBOC. The basaltic and tonalitic blocks, in particular, have commonly experienced significant kaolinization (e.g. 318, 319; Plate 6.1A), chloritization (e.g. 320), uralitization (e.g. 317) and, on occasion, epidotization (e.g. 316; Plate 6.1B). In fact, many of the basaltic inclusions have a 'bleached' appearance in outcrop (e.g. Plate 6.1A) and this is largely due to pervasive kaolinization of the plagioclase and the replacement of pyroxene by pale greenish amphibole and/or carbonate ± chlorite ± ?talc. For the most part, these basaltic inclusions are microdoleritic-doleritic types and their relict textures are similar to the majority of Glen Ward and Myra *Type 1* doleritic intrusives.

On present levels of exposure, uralitized basaltic rocks are rare in the Myra beds and exceedingly rare in the Glen Ward beds (Chapter 3, Section 3.3). Most of the PBOC low-Ti doleritic intrusives are highly uralitized (Section 5.6) but, unlike the uralitized doleritic inclusions in the serpentinite, these low-Ti types are usually quartz-bearing (Section 5.6) and their secondary amphibole is relatively deeper-green in colour. It is unclear whether the uralitized inclusions are fragments of Myra or (perhaps less likely) Glen Ward basaltic rocks originating from relatively deeper structural levels where uralitization might be widespread, or whether uralitization occurred after these blocks were incorporated in the serpentinite at depth (see Section 6.3).

Uralitized and lower-grade, and essentially unaltered "locally-derived" tectonic blocks are widespread throughout many schistose serpentinites in the NEO (e.g. Leitch, 1980a). Their modes of origin and alteration histories usually reflect: (i) relatively minor and apparently localized metamorphic and/or metasomatic events; and (ii) relatively high-level structural and tectonic events of small magnitude and rather localized significance. Secondary assemblages in blocks of this type have not been examined in any detail during the course of this study.

6.2 RELATIVELY HIGHER-GRADE METAMORPHIC BLOCKS

These are predominantly plagioclase-poor amphibolites and plagioclase-free hornblende schists, with minor relatively plagioclase-rich amphibolites, blueschists, mica (?stilpnomelane) schists and greenschists. Blocks of mica schist (e.g. 296, GR6996,8316; albite+quartz+stilpnomelane+muscovite+chlorite+zoisite) and greenschist (e.g. 294, GR6575,8198; ?actinolite+albite+chlorite+opaque oxide+sphene) are typically small (outcrops < 10 sq.m.), rare, and relatively insignificant. However, greenschist (314, albite+actinolite+chlorite+quartz) is locally abundant in association with amphibolites at GR707,829.

Fig. 6.1: Selected major and trace element relations in PBOC exotic rocks, low-Ti basalts and low-Ti doleritic intrusions. Low-Ti rocks from the Woolomin beds (Offler, 1982) plotted for comparison. For key to magma-type abbreviations see Appendix L.

a. Ti:Zr:Y relations. Fields from Pearce and Cann (1973) IAT = A+B, MORB = B, CAB = B+C, WPB = D.

b. Ti:Zr:Sr relations. Fields from Pearce and Cann (1973) IAT = A, CAB = B, MORB = C.

c. $\Sigma FeO:MgO:Al_2O_3$ relations. Fields from Pearce *et al.* (1977) 1 = MORB, 2 = OIT, 3 = CTB, 4 + Spreading Centre Island, 5 = Orogenic (IAT). Line a - b applies only to within-plate ocean island basalts and discriminates between tholeiitic (OITH) and alkaline (OIA) types.

d. Zr/TiO_2:Nb/Y relations. Fields from Winchester and Floyd (1977) S-AB = subalkaline basalt, A,B = subalkaline andesite-basalt, A = andesite, R,D = rhyodacite - dacite, TA = trachyandesite, AB = alkaline basalt.

e. $Nb/Y:Zr/P_2O_5$ relations. Basalt fields from Floyd and Winchester (1975), Winchester and Floyd (1976). ALK = alkaline, TH = tholeiitic.

f. Ni:Y relations. MORB and IAT fields from Crawford and Keays (1978) NOTE sample 313 contains 292 μg/g Ni.

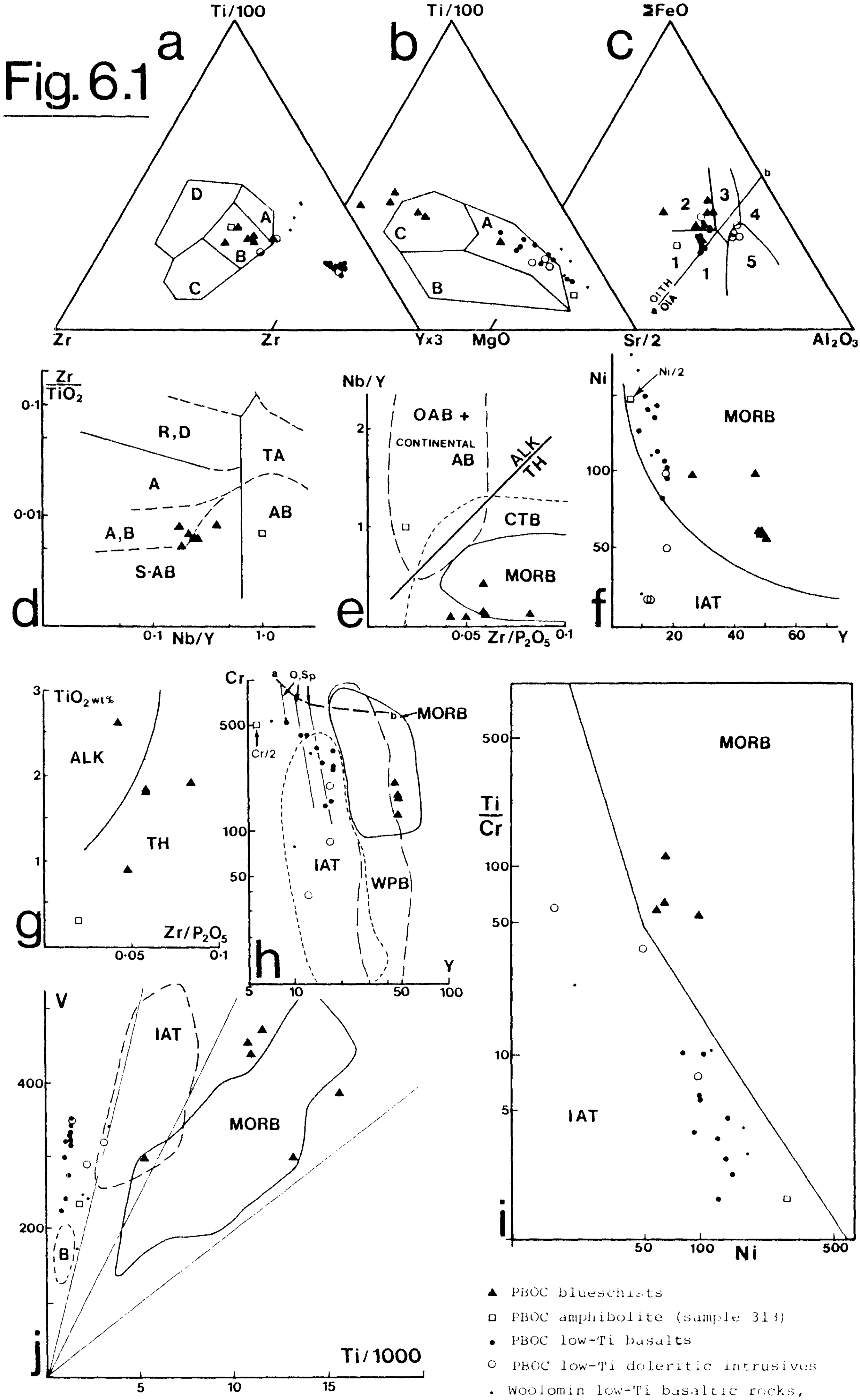
g. $TiO_2:Zr:P_2O_5$ relations. Fields for alkaline (ALK) and tholeiitic (TH) basalts from Winchester and Floyd (1976).

h. Cr:Y relations. Basalt fields from Pearce (1980). Line a - b is the locus of Cr:Y relations in partial melts derived by small (b,5%) to large (a,60%) degrees of partial melting of a hypothetical plagioclase lherzolite source (Pearce, 1980). O,Sp = olivine - spinel fractionation vectors (liquid lines of descent). NOTE sample 313 contains 1057 μg/gCr.

i. Ti/Cr:Ni relations. Fields for MORB and IAT from Beccaluva *et.al.* (1979).

j. Ti:V relations. Fields for MORB and IAT from Shervais (1982). B = boninite field compiled from references listed for Fig. 5.13.

NOTE: Low-Ti basaltic rocks are not plotted on diagrams d,e, and g. They fall well within the fields for subalkaline (tholeiitic) basalts on these three diagrams.



▲ PBOC blueschists

□ PBOC amphibolite (sample 313)

• PBOC low-Ti basalts

○ PBOC low-Ti doleritic intrusives

. Woolomin low-Ti basaltic rocks, Glenrock Station (Offler, 1982)

6.2.1 Amphibolites

(i) Plagioclase-poor types

Blocks of plagioclase-poor and plagioclase-free 'amphibolite' (e.g. 298, 311, 312) are relatively abundant in the PBOC schistose serpentinites. Most blocks of this type are clearly amphibolitized fragments of PBOC intrusives, and these have been discussed in Chapter 5. However, the mineral and/or whole-rock chemistries of a number of these blocks suggests that they were ultimately derived from precursors probably foreign to the PBOC (see below). For the most part, all three of the larger PBOC exotic amphibolite blocks (see Map 1) are poorly exposed. This poor exposure appears to be largely a function of widespread intense low-temperature shearing and subsequent pervasive weathering.

The largest of the predominantly plagioclase-poor amphibolite exotic blocks (block A) lies along the southeastern margin of a PBOC serpentinite in the vicinity of GR705,827 (see Map 1). Both block A and block B (see below) appear to be in fault contact with the adjacent Pola Fogal intrusives, and neither of these blocks are hornfelsed by the intrusives. In the vicinity of GR7011,8260, at least parts of this blcck consist of sedimentary rocks containing variable amounts of metamorphic (?metasomatic) amphibole. Partially amphibolitized interbedded argillites and sandstones are well-exposed at this locality. The thinner (<20cm) sandstone beds display well-developed graded-bedding but the thicker variants appear to be entirely amphibolitized, and recrystallization has obliterated sedimentary fabrics. Elsewhere in this block planar fabrics are relatively rare and, where developed they are often highly deformed (e.g. Plate 6.1C), and appear to be intrinsically metamorphic in origin. Some outcrops are massive although most display a well-developed, relatively low-temperature (cataclastic) shear foliation (Plate 6.1C) and, on occasion, an autoclastic melange fabric (Plate 6.1E). Disrupted albite±quartz±epidote (?metamorphic segregation) veins may be locally abundant, especially in the relatively finer-grained, strongly foliated variants (Plate 6.1D) and the greenschists.

The amphibolite block (block B) lying immediately to the southwest of block A (see Map 1) is also plagioclase-poor. Variants in this block are generally coarser-grained than those in block A, and premetamorphic fabrics are not recognizeable. Relatively small blocks (tens of sq.m., or less) of similar amphibolites occur in schistose serpentinites immediately to the north of this large block (e.g. GR6881, 8205), and at numerous other localities throughout the PBOC (e.g. GR658, 822; GR690,888; GR707,832).

The plagioclase-poor amphibolites consist almost entirely of pleochroic pale-deep green, Mg-rich, relatively Al- and Na-poor calcic amphiboles (Table 6.1, analyses 9-13). In some samples (e.g. 309, 315) these 'hornblendes' are zoned to actinolitic rim compositions (Table 6.1, analyses 10-13). Modal plagioclase rarely exceeds 10% and is usually albitized and kaolinized. Accessory phases include prograde Ca-rich pyroxene (e.g. Table 6.2, analysis 6), relatively rare oxides, sphene and, in some samples, Mg-poor Cr-Fe spinel (Table 6.2, analyses 7 and 8). Vein- and minor secondary alteration phases include albite, actinolite, anthophyllite (e.g. 312), prehnite, epidote, chlorite, carbonate and sericite (after plagioclase, e.g. 312).

'Hornblendes' in these amphibolites are relatively Al-poor (Al_2O_3 2.5% - 6.3%, Table 6.1) and contain significantly less Al^{VI} (Fig. 5.8) compared to those in PBOC amphibolitized intrusives (Al_2O_3 5.3% - 12.0%, Table C-6). The relatively higher Al^{VI} and Na^{M4} contents of the secondary amphiboles in the PBOC olivine norites (e.g. Fig. 5.8) might suggest that these crystallized at somewhat higher pressures than those in the amphibolites. Recently, however, Hynes (1982) has demonstrated that these parameters can be unreliable indicators of pressures of amphibole crystallization in metabasites. 'Hornblendes' in both rock-types under discussion are highly depleted in TiO_2 (<0.12%). Although this is also consistent with relatively low pressures of amphibole crystallization (*cf*. Raase, 1974; Hynes, 1982), in these rocks bulk composition was probably the controlling factor in amphibole TiO_2-, and to a lesser extent, Al_2O_3 contents (see Table 6.3a, analyses 8 and 9; *cf*. Appendix B). Consequently, the data available do not closely constrain estimates of the pressures at which these plagioclase-poor amphibolites might have crystallized, especially as experimental and other data pertaining to relatively more Na-, Ti- and Al-rich 'basaltic' compositions are probably not strictly applicable to these particular lithologies.

The bulk chemistries of two of the PBOC plagioclase-poor amphiboles (311, block A, and 312, block B; Table 6.3a analyses 8 and 9 respectively) suggest that these might be metamorphosed mela-gabbros. They are significantly enriched in *di* Table 6.3b), Ba, Zr and LREE, compared to PBOC ophiolitic intrusives with comparable MgO contents (*cf.* Tables 5.3 and B-7). Unlike the PBOC amphibolitized intrusives these amphibolites contain *prograde* (only) Ca-rich pyroxene which is significantly depleted in Cr and Al relative to relict pyroxenes in the former (Table 6.2, analysis 6; *cf.* Fig. 5.6). The Cr-Al spinels which occur in some of these amphibolites (e.g. 311, 315) are also chemically distinctive, being significantly enriched in Fe and Mn, and depleted in Al and Mg relative to those in the PBOC ophiolitic intrusives (Table 6.2, analyses 7 and 8; *cf.* Fig. 5.3, Table C-2).

The presence of chromite-bearing amphibolites (311), Cr-rich greenschists (314, 525 ug/g Cr, Table 6.3a, analysis 11) and partially amphibolitized argillites and sandstones (see below) in block A raises the possibility that this block, at least, might largely consist of metamorphosed mafic - ultramafic detritus which may or may not include a component derived from PBOC ophiolitic lithologies. Although the partially amphibolitized argillites and sandstones at GR7011,8260 are well-exposed, they were not sampled because they appeared to be too weathered and too friable for petrographic examination. Consequently their provenance is unknown but, in the light of the above discussion, they warrant further investigation.

(ii) Plagioclase-rich types

The largest block of relatively plagioclase-rich amphibolite in the Pigna Barney - Curricabark area (block C) crops out in the vicinity of GR675,821 (see Map 1). Small plagioclase-rich amphibolite blocks (tens of sq.m. or less) are embedded in schistose serpentinites at GR6895,8225 (306), and at GR577,945 (307). For the most part these amphibolites consist of highly irregular, mesoscopic, alternating amphibole-rich and plagioclase-rich layers, although relatively homogeneous and diffusely-banded types are common in block C (e.g. sample 313). Amphibolites from each of these three blocks are quite distinctive and some preliminary petrological data are summarized below.

Sample 306: a strongly foliated amphibolite containing approximately

60% pale olive-brown, Ti-bearing (0.8% TiO_2) magnesiohornblende (Table 6.1, analysis 7) and approximately 40% saussuritized plagioclase. Relict plagioclase is relatively calcic and uniform in composition (An_{82-78}, Table 6.1). Fe-Ti oxides are rare.

Sample 307: consists of highly irregular, elongate, medium-grained amphibole-rich patches and lenses in a banded matrix rich in altered (largely kaolinized) plagioclase (An_{30-25}, Table 6.1). Two types of amphibole are present: (i) a slightly pleochroic pale blue-green to pale brown to colourless amphibole (not analysed) forming relatively coarse-grained monomineralic patches; and (ii) a strongly pleochroic deep olive blue-green to pale olive-brown magnesio-alumino-kataphorite (Table 6.1, analysis 8; *cf.* Leake, 1978) which displays a granoblastic-elongate texture in association with plagioclase and relatively minor pale green Ti-poor salite (Table 6.2, analysis 5). Modal abundances of these three phases vary considerably, although salite rarely exceeds *ca* 10% of the mode. Sphene and rare opaque oxides are the only significant accessory phases.

Block C: These amphibolites largely consist of highly variable proportions of blue-green to pale olive-brown amphibole, and altered plagioclase. The latter is largely replaced by zoisite, clinozoisite, chlorite, prehnite±quartz±carbonate (more commonly in veins) assemblages. Mesoscopic patches of decussate pale-green amphibole are also relatively common in amphibolites forming this block (e.g. 293). The bulk chemistry of a composite sample (313) from a diffusely-laminated type (292) and a relatively massive type (291) displays some characteristics in common with relatively Mg-rich (MgO = 12.3%) 'basaltic' rocks. These include moderate Al_2O_3 (13.4%), high Ni (292 ug/g), high Cr (1057 ug/g), and low abundances of the relatively immobile incompatible elements (e.g. Zr, Nb, Y and, to a lesser extent P; Table 6.3a, analysis 10). However, the overall relatively immobile element relationships in this sample are not entirely consistent with any of the more common magma types. Thus Figure 6.1a,cj suggests subalkaline characteristics whereas Figure 6.1d,e suggests affinities with alkaline basaltic types. Rocks with comparable chemistry are unknown in the Tamworth Belt and the Woolomin Association. Also, this amphibolite is significantly enriched in *di*, TiO_2, P_2O_5, Zr, Ba LREE and Sr, and depleted in Al and *hy* relative to PBOC intrusives

PLATE 6.1

A. Tectonic block of highly kaolinized (?) microdolerite embedded in schistose PBOC serpentinite [sample 318, GR6205,8267].

B. Sub-rounded tectonic block of epidotized basalt in highly schistose PBOC serpentinite. Pale-coloured vein predominantly consists of prehnite + epidote [GR6832,8216].

C. Highly deformed PBOC amphibolite displaying a well-developed planar fabric. Note intense shearing in the upper right and left sides of the outcrop. [GR7065,8289].

D. Disrupted albite ± quartz ± epidote segregation veins in strongly foliated, fine-grained, PBOC plagioclase-poor greenschist [GR7066, 8288].

E. Autoclastic melange fabric in highly-sheared PBOC plagioclase-poor amphibolite. Pale patches (predominantly on phacoids) are moss or lichen [GR7066,8291].

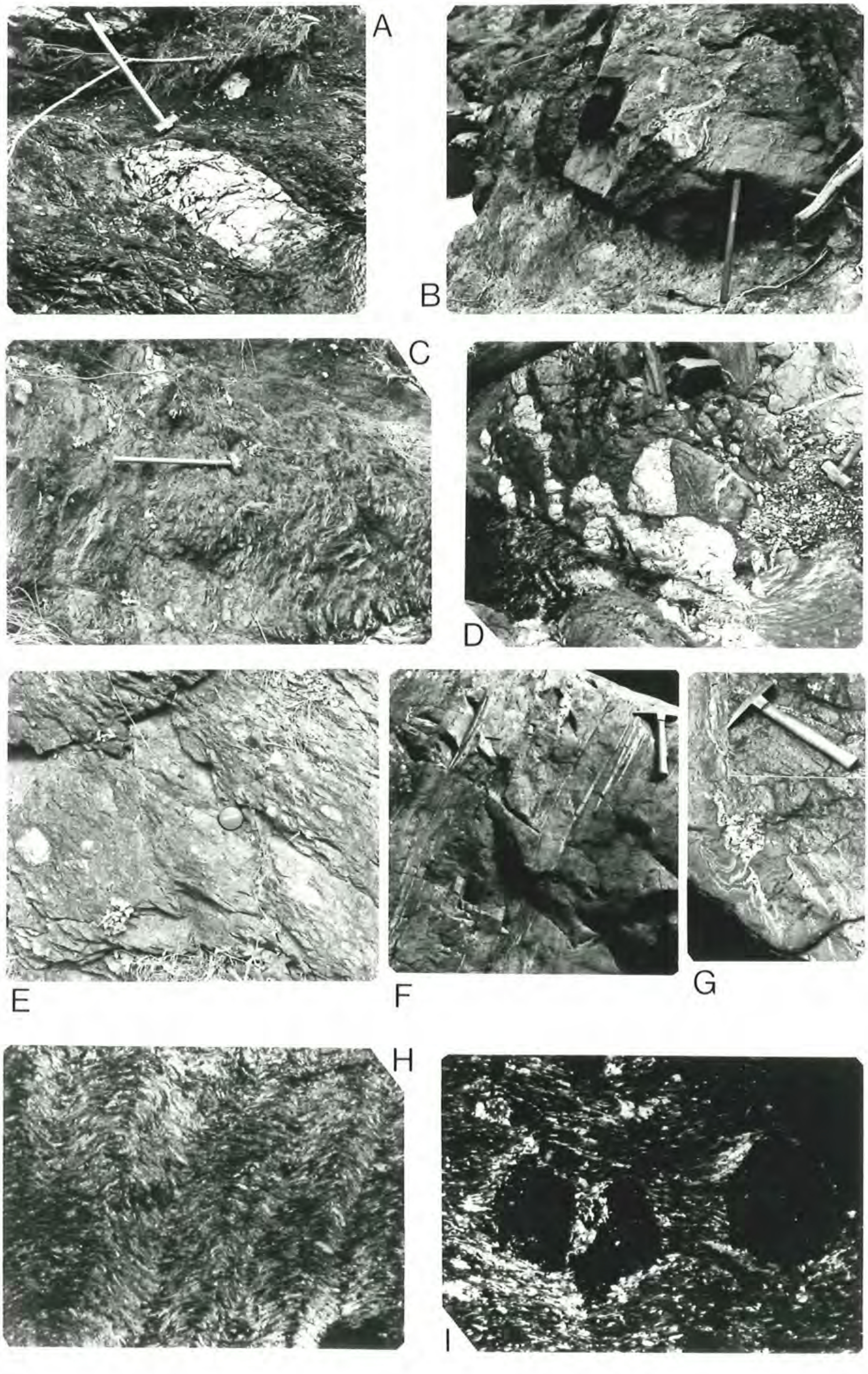
F. Planar fabric (weakly developed glaucophane ± muscovite foliation) in PBOC blueschist delineated by thin segregation veins. These veins largely consist of albite + minor actinolite and muscovite [GR6567, 8189].

G. Small-scale disharmonic folds in segregation veins in PBOC blueschist [GR6567,8189].

H. Crenulations in foliated PBOC blueschist. The field of view consists of glaucophane (elongate) plus accessory sub-equant grains of sphene (dark grey-black).[sample 305, mag. = 35x , crossed nicols].

I. Glaucophane foliation wrapping around relatively early-formed garnets (large black grains) in PBOC blueschist. Pale alteration products after garnet are muscovite and pumpellyite. The garnets are largely altered to chlorite which is not resolvable in this photograph. [sample 299, mag. = 9x , crossed nicols].

PLATE 6·1



with similar MgO contents (*cf*. Table 5.4a,b, analyses 1-3) and, for that matter, practically all other PBOC ophiolitic lithologies (with the exception to the low-Ti doleritic intrusives which are, however, relatively depleted in Mg, Ni and Cr; Table 5.5a).

6.2.2 Blueschists

Blocks of glaucophane schist occur at three localities in the PBOC (1, GR6567,8190; 2, GR6573,8195; 3, GR6922,8271). The largest of these blocks (1) is marked on Map 1. It occupies the bed of the Pigna Barney River for approximately 200 metres and it ranges in width from approximately 20m to 40m. It displays a well-developed planar fabric [glaucophane (-muscovite) foliation; Plate 6.1F] which commonly strikes sub-perpendicular to the length of the block. However, this occurrence appears to be an aggregate of relatively small, somewhat jumbled sub-blocks and,consequently, present structural attitudes are of uncertain significance. Block 2 is small (1m. x 0.4m.), lensoidal, partially embedded in schistose serpentinite, and also in the bed of the Pigna Barney River. "Block" 3 consists of several relatively equant blocks ranging up to 20m in width which appear to have eroded out of the adjacent schistose serpentinite.

Overall assemblages in each of these occurrences include:

1. glaucophane (Table 6.1, analyses 1,2) + minor muscovite (Table 6.2, analysis 3), pumpellyite, garnet (Table 6.2, analyses 1,2) albite (An_{2-0}), calcite, relatively rare actinolite and/or chlorite, and accessory sphene, apatite, zircon and pyrite.

2. glaucophane + minor sphene, and accessory pumpellyite, pyrite and apatite.

3. glaucophane (Table 6.1, analyses 3-5) + sphene, minor albite, actinolite (Table 6.1, analysis 6), pumpellyite (Table 6.2, analysis 4), and accessory pyrite.

In block 1 albite and actinolite are largely confined to thin (?)segregation veins and lenses (Plate 6.1F) which usually lie within the plane of the glaucophane(-muscovite) foliation. Muscovite is also irregularly distributed. Although some of the segregation veins are disharmonically-folded on a relatively small scale (Plate 6.1G), there is little evidence for strong multiple deformation in block 1. On the other hand, the glaucophane foliation in block 3 displays well-developed crenulations (Plate 6.1H), and has presumably experienced at least two distinct deformational episodes followed by cataclasis during (?) final emplacement.

Relict (largely chloritized) garnets in block 1 are almandine-grossular solid solutions (Table 6.2, analyses 1,2) containing ~20% pyrope+spessartine+andradite end-members. In general terms these are compositionally similar to garnets characteristic of relatively high-grade blueschist assemblages (e.g. Newton and Fyfe, 1976; Turner, 1981) and Group C (alpine-type) eclogites (Coleman *et al.* 1965; Raheim and Green, 1975; Ryburn *et al.* 1976). However, the PBOC blueschists appear to be devoid of the more diagnostic high-pressure phases such as jadeite, lawsonite, aragonite and omphacite, and the presence of relatively abundant sphene + calcite rather than rutile also suggests relatively low-grade, blueschist facies conditions of metamorphism (Ernst, 1972; Blake and Morgan, 1976; Hunt and Kerrich, 1977; Itaya and Banno 1980). Textural relations suggest that the garnet predates the glaucophane foliation (Plate 6.1I) and, as such, it might be relict from an earlier, largely obliterated, higher-grade blueschist facies assemblage. Constraints on the stability of glaucophane (Maresch, 1977), and recent generalized blueschist facies P-T phase diagrams of Wood (1979), Brothers and Yokoyama (1982) and Brown and O'Neil (1982), all suggest that the typical PBOC assemblage: glaucophane+sphene+albite+muscovite, in the absence of jadeite, crossite and lawsonite (or its high temperature breakdown product *viz* zoisite± pyrophyllite; Liou, 1971; Nitsch, 1972; Kushev, 1980), probably crystallized under P-T conditions in the vicinity of 400°C and 6-8kb. Rb-Sr isotopic data on muscovite separated from block 1 (Table H-2) suggest that this particular assemblage is Early Devonian in age (378 m.y.).

Although metamorphic segregation processes have probably produced significant chemical heterogeneities in these PBOC blueschists, for the most part their bulk chemistries (Table 6.3a, analyses 1-6) strongly suggest that they were originally basaltic rocks. Apart from their relatively low Al_2O_3:MgO + ΣFeO + Al_2O_3 ratios (Fig. 6.1c), which most probably reflect some segregation of Al into albite veins, and the relatively high TiO_2 (2.6%) and P_2O_5 (0.4%) abundances in sample 303 (Fig. 6.1g), most relatively immobile element characteristics of these blueschists resemble those typically displayed by subalkaline basalts, and in particular, E-type MORB (e.g. Figs 5.16, 5.17, 6.1a,b,d-j). In many of these characteristics blueschists from blocks 1 and 3 resemble Myra *Type 1* basaltic rocks (Table 3.5a, analyses 1,2; compare Figs. 5.17, 5.18, 6.1 with Figs 3.2-3.33). On this basis, and bearing in mind that the Myra beds are probably part of a relatively widespread subduction complex (see Chapter 1), it is conceivable that the TiO_2-rich blueschists might have been ultimately derived from subducted Myra *Type 1* basaltic rocks or their equivalents. On the other hand, Ti, Zr, Nb and Y abundances in the block 2 blueschist (300) compare more closely to those in some Glen Ward basaltic extrusives (compare analysis 1, Table 6.3a with analyses 2 and 3, Table 3.7) than abundances of these elements in Myra basaltic rocks. However, the latter have significantly greater $\Sigma FeO/MgO$ ratios than the former (300) and, by analogy with other Glen Ward basaltic rocks (Table 3.6a), they may well be significantly enriched in P, V, LREE, and depleted in Ni and Cr relative to this blueschist. Thus, the available chemical data impose few tangible constraints on the ultimate origin of block 2, especially in view of the relatively large range of basaltic rock types known to occur in the Woolomin Association as a whole (see Chapter 3, Section 3.5).

6.3 DISCUSSION

Rare, isolated exotic blocks of blueschist and higher-grade metamorphics occur in a number of serpentinites in the NEO. These appear to be most abundant in the Pigna Barney - Curricabark, Glenrock Station and Port Macquarie areas (this thesis; Barron *et al.*, 1976; Leitch 1980c; Offler, 1982b), although blocks of amphibolite have been reported from serpentinites in the Mount George and Toms Creek areas (Leitch, 1980a) and in the Woodsreef area (Glen and Heugh, 1973). Other exotic metamorphic blocks associated with serpentinites in the Peel Fault System and in Zone B include: (i) mafic eclogites near Attunga (Shaw and Flood, 1974) and at Port Macquarie (H.D. Hensel, pers. comm.); (ii) possibly the nephrite occurrences approximately 25km southeast of Tamworth (Hockley, 1974; Lanphere and Hockley, 1976; and (iii) blueschists at several other unspecified localities (H.D. Hensel, pers. comm.).

Because these exotic blocks typically occur in serpentinite-matrix melanges delineating major fault systems in the NEO, their tectonic implications must remain uncertain. However, by analogy with other ophiolites, at least some of these metamorphic blocks might be tentatively interpreted as remnants of sub-ophiolitic metamorphic soles ('dynamothermal aureoles' *cf.* Church and Stevens, 1971; Williams and Smyth, 1973; Graham and England, 1976; Malpas, 1979) fragmented during the emplacement of NEO serpentinites and related rocks.

Many ophiolites are directly underlain by thin (<<1km) highly deformed granulite- or amphibolite facies assemblages which pass rapidly into low-grade and/or unmetamorphosed rocks further away from the base of the overlying ophiolite. In various ophiolites such metamorphics have been recently interpreted to have formed: (i) within the oceanic crust prior to obduction [e.g. Ballantrae Complex, Spray and Williams, 1980; Treloar *et al.* 1980; Mamonia Complex, Cyprus, Spray and Roddick 1981; Hellenic (Greek) ophiolites, Spray and Roddick, 1980]; or (ii) from MORB and/or sedimentary rocks accreted to the base of the ophiolite during obduction (e.g. Bay of Islands and related Newfoundland ophiolites, Dallmeyer and Williams, 1975; Dallmeyer, 1977; Malpas, 1979; Jamieson, 1980; Feininger, 1981), or accreted during subduction (e.g. some Tethyan ophiolites, Hall, 1980; Searle and Malpas, 1980,1982; Thuizat *et al.*, 1981; and some ophiolites in the southwestern Pacific, Parrot and Dugas, 1980). On occasion, one or other of all three of the above interpretations have been specifically applied to individual dynamothermal aureoles beneath a number of the Tethyan ophiolites (compare Woodcock and Robertson, 1977; Hall, 1980; Thuizat *et al.*, 1981). The common lack of consensus highlights the uncertainties commonly encountered in evaluating the mode(s) of origin of such rocks.

Metamorphic rocks which have presumably originated *via* one or other of the above mechanisms may become incorporated in serpentinite-matrix melange zones within the ophiolite itself. In some parts of the Semail ophiolite of Oman, for instance, 'exotic' metamorphic blocks within serpentinite-matrix melanges may be correlated with metamorphic 'soles' underlying the ophiolite elsewhere (Searle and Malpas, 1980, 1982). However, in highly disrupted ophiolites it is not uncommon for a range of such metamorphic associates to be preserved only as rare and widely-scattered exotic blocks in ophiolitic melange [e.g. Taurus suture zone, Turkey, Hall, 1976,1980; Kings - Kaweah ophiolite, California (largely oceanic fracture zone metamorphism), Saleeby, 1977,1978,1979; Oman, Searle and Malpas, 1980,1982; New Caledonia, see Parrot and Dugas (1980) for a summary] and, like those in the PBOC, their origin(s) may be difficult to assess.

The limited petrological data available on the PBOC exotic metamorphic blocks indicate that they are a polygenetic group largely derived from mafic igneous parents and, rarely, sedimentary rocks. The PBOC blueschists, at least, and blocks of blueschist and eclogite in other NEO serpentinites (see above), are probably subduction-related (*cf.* Ernst, 1971,1973,1977) and it is conceivable that these might have been ultimately derived from a range of somewhat MORB-like Woolomin Association basaltic rocks (see Sections 1.1, 3.6.1,6.2.2). Isotopic ages obtained on PBOC and Port Macquarie blueschist blocks (Rb-Sr 378 m.y., this study, Appendix H; K-Ar 383=5 m.y., Port Macquarie, Pogson and Hilyard, 1981) are similar and suggest that subduction and perhaps accretion occurred in Zone B at least in the Early Devonian. On the other hand, nephrites from serpentinites in the Peel Fault System 25km to the south of Tamworth have Early Permian $^{40}Ar/^{39}Ar$ ages (273 ± 6m.y., 280 ± 6 m.y., Lanphere and Hockley, 1976) but the precise geological relevance of these 'nephrite ages' is uncertain. The data are obviously too few to establish whether the blueschists and nephrites were formed during separate metamorphic events or whether they simply reflect piecemeal sampling of metamorphic rocks generated during a prolonged subduction event (*cf.* Cawood, 1982a).

If, as appears likely (see Chapter 7), the PBOC was generated in an outer-arc or remnant arc setting, conventional concepts of ophiolite emplacement (see above) might not be entirely applicable. Exotic amphibolite blocks in the PBOC are chemically distinct from MORB (see Section 6.2), as is at least one amphbolite block at Port Macquarie (Table 6.3a, analysis 7), and consequently these are probably not metamorphosed fragments of oceanic crust. Perhaps some of these amphibolite blocks and associated greenschists were initially generated as metamorphic soles to ophiolitic slabs accreted at depth in the Woolomin Association. Similarly, the blueschists and uralitized basaltic blocks (see Section 6.1) in the PBOC and elsewhere in the NEO might represent altered fragments of such ophiolitic slabs themselves. Because relatively higher-temperature metamorphic overprinting of the PBOC blueschist assemblages is minimal (minor actinolite), it is likely that these exotic blocks, at least, were transported to relatively high structural levels within a geologically short period (?<50 m.y., *cf*. Draper and Bone, 1981) after their generation. However, the data available impose few constraints on mechanisms whereby these blocks might have been finally mixed with PBOC ophiolitic lithologies. A complex interplay of strike-slip and thrusting movements during, and perhaps following, accretion in Zone B is conceivable (see Chapter 1, *cf*. Karig, 1980).

TABLE 6.1

Representative Analyses of Amphiboles from Metamorphic Rocks in the PBOC

ANALYSIS No.	1	2	3	4	5	6	7	8	9	10	11	12	13
	Blueschists						**Amphibolites →**			CORE	RIM	CORE	RIM
SAMPLE	301	302	305	305	305	305	306	307	311	309	309	315	315
SiO_2	56.46	54.16	53.92	54.45	56.17	51.96	48.99	45.64	52.42	51.55	55.26	50.70	55.06
TiO_2	0.19	-	-	0.32	0.13	-	0.78	0.66	0.05	-	-	0.12	-
Al_2O_3	10.57	11.17	7.18	11.19	8.87	3.85	6.12	8.75	4.84	4.66	0.33	6.26	2.50
Cr_2O_3	-	-	-	-	-	-	0.21	-	0.09	0.19	-	0.38	0.06
ΣFeO	12.83	13.27	16.39	15.48	16.93	15.39	14.46	15.14	7.05	10.43	8.95	7.84	6.53
MnO	0.11	0.12	0.16	0.18	0.16	0.23	0.08	0.25	0.05	0.23	0.52	0.05	-
MgO	9.03	9.19	9.53	7.60	7.69	13.18	14.45	12.89	19.39	17.18	17.87	18.53	20.46
CaO	1.23	2.63	4.89	2.02	1.32	11.53	11.34	11.49	12.24	12.28	12.47	12.22	12.47
Na_2O	6.83	7.39	4.63	6.46	6.78	1.19	0.96	3.51	0.88	0.52	0.29	1.08	0.44
K_2O	-	0.08	0.05	-	-	0.09	-	0.10	-	0.08	-	0.14	0.05
TOTAL	97.25	98.01	96.75	97.70	98.05	97.42	97.39	98.43	97.01	97.12	96.19	97.32	97.57
Fe_2O_3	1.37	0.22	2.20	1.72	2.20	1.10	5.27	1.37	3.58	4.55	0.51	3.96	2.43
FeO	11.60	13.07	14.41	13.93	14.95	14.40	9.72	13.91	3.83	6.33	8.49	4.27	4.35
New Total	97.39	98.03	96.97	97.37	98.27	97.53	97.92	98.57	97.37	97.57	98.24	97.71	97.82
Numbers of cations on the basis of 23 (O,OH)													
Si	7.888	7.638	7.792	7.699	7.936	7.592	7.087	6.724	7.363	7.341	7.925	7.153	7.661
Al^{IV}	0.112	0.362	0.208	0.301	0.064	0.408	0.913	1.276	0.637	0.659	0.075	0.847	0.339
Al^{VI}	1.629	1.495	1.015	1.564	1.413	0.255	0.131	0.243	0.164	0.123	0.065	0.194	0.071
Ti	0.020	-	-	0.034	0.014	-	0.085	0.073	0.005	-	-	0.013	-
Cr	-	-	-	-	-	-	0.024	-	0.010	0.021	-	0.042	0.007
Fe^{3+}	0.144	0.023	0.239	0.183	0.234	0.121	0.574	0.152	0.378	0.488	0.055	0.420	0.254
Fe^{2+}	1.355	1.542	1.741	1.648	1.767	1.760	1.176	1.713	0.450	0.754	1.019	0.504	0.506
Mn	0.013	0.014	0.020	0.022	0.019	0.029	0.010	0.031	0.006	0.028	0.063	0.005	-
Mg	1.880	1.932	2.052	1.602	1.619	2.870	3.116	2.830	4.060	3.647	3.619	3.897	4.244
Ca	0.184	0.397	0.757	0.306	0.200	1.805	1.758	1.814	1.842	1.374	1.916	1.847	1.859
Na	1.850	2.021	1.297	1.771	1.857	0.337	0.269	1.003	0.240	0.144	0.081	0.295	0.119
K	-	0.014	0.008	-	-	0.017	-	0.019	-	0.015	-	0.025	0.008
Σ	15.075	15.438	15.129	15.130	15.123	15.194	15.143	15.878	15.155	15.094	15.018	15.242	15.068
M	0.581	0.556	0.541	0.493	0.478	0.620	0.726	0.623	0.900	0.829	0.739	0.885	0.893
Fe^*	0.081	0.015	0.191	0.105	0.142	-	-	-	-	-	-	-	-
An	<2	<2	n.f.	n.f.	n.f.	n.f.	78-82	25-30	n.f.	n.f.	n.f.	<2	<2

Fe^{3+} calculated following the method of Papike *et al.* (1974)
$M = Mg/(Mg+Fe^{2+})$
$Fe^* = Fe^{3+}/(Fe^{3+}+Al^{VI})$
An = Anorthite content of co-existing plagioclase
n.f. = not found

Analyses 1-5 = glaucophane
6,11 = actinolite
7,12 = magnesio-hornblende
8 = magnesio-alumino-kataphorite
9,13 = actinolitic-tremolitic hornblende
10 = actinolitic hornblende
Amphibole nomenclature after Leake (1978).

Plagioclase-rich amphibolites: 306 & 307
Plagioclase-poor amphibolites: 309,311,315

TABLE 6.2

Microprobe Analyses of some Minor Mineral Phases in
Metamorphic Rocks from the PBOC

ANALYSIS No.	1	2	3	4		5	6	7	8
	Blueschists					Amphibolites			
SAMPLE	299	299	301	305		307	311	311	315
SiO_2	37.13	37.36	48.79	37.09		51.71	52.60	-	-
TiO_2	0.33	-	0.41	-		-	0.12	0.20	0.40
Al_2O_3	21.39	21.18	27.84	25.88		1.34	1.29	7.21	5.18
Cr_2O_3	0.18	-	-	-		-	-	51.43	34.19
$^*Fe_2O_3$	1.13	1.79	n.d.	n.d.		7.51	1.04	9.10	28.46
FeO	23.85	24.19	3.40	3.67		1.89	8.35	30.12	30.09
MnO	3.06	2.54	-	0.61		0.26	0.13	0.76	1.52
MgO	2.06	2.19	3.16	3.00		12.75	14.20	1.32	0.53
CaO	10.98	10.93	-	22.57		22.90	22.24	-	0.15
Na_2O	-	-	0.50	0.40		1.64	0.16		-
K_2O	-	-	10.61	-		-	-	-	-
TOTAL	100.11	100.18	94.71	93.24		100.00	100.13	100.14	100.52
Oxygens	24	24	22	28		6	6	32	32
Si	5.884	5.917	6.617	6.807		1.925	1.959	-	-
Al^{IV}	0.116	0.083	1.383	0.117		0.059	0.041	2.415	1.771
Al^{VI}	3.800	3.870	3.067	5.481		-	0.016		
Ti	0.039	-	0.042	-		-	0.003	0.043	0.087
Cr	0.023	-	-	-		-	-	11.556	7.843
Fe^{3+}	0.135	0.213	-	-		0.210	0.029	1.946	6.214
Fe^{2+}	3.162	3.205	0.386	0.564		0.059	0.260	7.168	7.311
Mn	0.411	0.341	-	0.095		0.008	0.004	0.183	0.374
Mg	0.487	0.517	0.639	0.821		0.707	0.788	0.559	0.229
Ca	1.864	1.855	-	4.438		0.913	0.888	-	0.047
Na	-	-	0.131	0.142		0.118	0.012	$(0.130)^+$	$(0.124)^+$
K	-	-	1.836	-		-	-	-	-
Σ	15.921	16.001	14.101	18.465		3.999	4.000	24.000	24.000
mg	11.6	12.1	62.3	55.5		71.8	72.9	5.7	1.6
Cr	n.d.	n.d.	n.d.	n.d.		n.d.	n.d.	82.7	81.6
AL	53.2	54.2			*Ca*	48.1	45.1		
SP	6.9	5.8			*Mg*	37.3	40.0		
PY	8.2	8.7			*Fe'*	14.6	14.9		
AN	4.4	5.4							
GR	26.4	25.9							
UV	0.6	-							

$^*Fe_2O_3$ calculated $^+$cation ratios for Zn   Analysis 7 includes 0.62% ZnO

n.d. = not determined   Analysis 8 includes 0.58% ZnO

Analyses 1,2 = garnet; 3 = muscovite; 4 = pumpellyite; 5,6 = salite; 7,8 = chromite.

TABLE 6.3a

Major and Trace Element Analyses of Metamorphic Rocks Associated with the Pigna Barney Ophiolitic Complex

	Blueschists						Amphibolites				G'schist
ANALYSIS No.	1	2	3	4	5	6	7	8	9	10	11
SAMPLE	300	301	302	303	304	305	310*	311	312	313	314
SiO_2	49.70	49.17	49.48	50.64	46.54	50.25	56.61	52.24	52.65	46.63	61.43
TiO_2	0.84	2.20	1.83	2.62	1.94	1.82	0.01	0.08	0.04	0.30	0.17
Al_2O_3	14.24	12.58	13.10	12.86	14.93	12.90	1.35	4.76	5.77	13.37	11.30
V_2O_3	0.04	0.04	0.06	0.06	0.07	0.07	0.01	0.01	0.02	0.03	0.02
Cr_2O_3	n.d.	0.03	0.03	0.02	0.02	n.d.	0.23	0.35	n.d.	0.15	0.08
Fe_2O_3	1.59	2.16	2.66	2.79	2.66	2.90	1.18	2.13	1.96	1.73	1.75
FeO	8.07	8.61	10.05	9.72	10.77	10.35	4.54	5.14	7.82	7.20	5.29
NiO	0.01	0.01	0.01	0.01	0.01	0.01	0.16	0.09	0.04	0.04	0.01
MnO	0.19	0.18	0.20	0.18	0.22	0.20	0.19	0.16	0.30	0.17	0.03
MgO	8.73	8.44	7.30	6.81	6.94	6.24	21.14	19.33	15.06	12.28	9.15
CaO	10.70	7.41	8.55	6.56	9.06	7.30	11.02	11.50	11.82	12.23	3.06
Na_2O	2.17	3.54	3.49	4.82	2.83	4.34	1.24	0.93	0.93	1.64	3.12
K_2O	0.47	1.13	0.35	0.48	0.37	0.13	0.20	0.30	0.61	0.33	0.45
P_2O_5	0.09	0.28	0.19	0.41	0.14	0.19	0.02	0.06	0.05	0.10	0.07
ΣVol[1]	3.01	3.73	3.06	3.28	3.87	3.54	2.69	2.81	2.40	4.15	3.54
TOTAL	99.85	99.51	100.36	101.26	100.37	100.24	100.59	99.89	99.47	100.35	99.47
ΣFeO/MgO	1.09	1.25	1.70	1.80	1.90	2.08	0.26	0.37	0.64	0.71	0.75
TRACE ELEMENTS (μg/g)											
Rb	11	15	<2	5	4	<2	<2	<2	16	<2	5
Ba	109	388	55	120	73	27	-	91	131	116	64
Sr	182	25	115	90	146	62	40	323	29	257	82
Li	10.5	10.5	9.9	12.7	9.9	12.7	2.2	4.4	1.1	7.2	8.3
Zr	43	163	111	174	117	110	<2	27	13	19	43
Nb	5	16	10	9	11	12	<3	7	3	6	8
Y	26	46	48	47	48	50	3	5	4	6	5
Ti[1]	5240	11702	10059	15285	10796	11028	163	267	210	1798	1175
Cu	32	18	70	58	87	55	20	3	8	33	8
Zn	65	105	120	118	128	118	60	45	80	78	65
Ni	92	99	60	66	66	57	1231	719	302	292	94
Co	55	n.d.	n.d.	n.d.	n.d.	63	71	87	79	71	55
V	298	301	439	387	468	453	53	58	123	233	143
Cr	n.d.	207	171	129	166	n.d.	1560	2364	n.d.	1057	525
La	4	16	-	-	12	14	-	5	4	6	6
Ce	-	28	-	-	21	18	-	9	6	15	12
Nd	-	18	-	-	14	14	-	4	-	4	7

* FROM Pt. MACQUARIE N.S.W.

[1] See Appendix G

n.d. = not determined.

PBOC

Blueschists

Block 1 : analyses 2-5
Block 2 : analysis 1
Block 3 : analysis 6

Amphibolites & Greenschist

Block A : analyses 8 & 11
Block B : analysis 9
Block C : analysis 10

TABLE 6.3b

C.I.P.W. Normative Mineralogy of Metamorphic Rocks Associated with the Pigna Barney Ophiolitic Complex

	Blueschists						Amphibolites				Greenschist
ANALYSIS No.	1	2	3	4	5	6	7	8	9	10	11
SAMPLE	300	301	302	303	304	305	310*	311	312	313	314
qz	-	-	-	-	-	-	2.95	-	0.71	-	16.85
or	2.78	6.68	2.07	2.84	2.19	0.77	1.18	1.77	3.60	1.95	2.66
ab	18.36	29.95	29.53	40.87	23.95	36.81	5.83	7.8	7.87	13.88	26.40
an	27.75	15.13	19.10	12.05	27.00	15.34	-	7.95	9.80	28.17	14.72
c	-	-	-	-	-	-	-	-	-	-	0.29
di	20.21	16.18	18.26	14.72	14.14	16.35	43.20	38.82	39.12	25.74	-
hy	20.80	7.51	12.13	4.76	8.76	10.91	41.88	33.46	34.21	4.14	33.25
ol	3.57	13.72	10.18	14.74	14.24	10.45	-	5.27	-	19.87	-
mt	1.54	1.66	1.97	1.95	2.09	2.09	-	1.13	1.55	1.41	1.10
chr	-	0.04	0.03	0.02	0.03	-	0.44	0.52	-	0.22	0.12
il	1.60	4.18	3.48	4.98	3.68	3.46	0.02	0.15	0.08	0.57	0.32
ap	0.21	0.65	0.44	0.95	0.32	0.44	0.05	0.14	0.12	0.23	0.16
Vol	3.01	3.73	3.06	3.28	3.87	3.54	2.69	2.81	2.40	4.15	3.54
Σ	99.83	99.43	100.25	101.16	100.27	100.16	100.64	99.82	99.46	100.33	99.41
100 an/(ab+an)	60.2	33.6	39.3	22.8	53.0	29.4	-	50.3	55.5	67.0	35.8

$Fe^{3+}/\Sigma Fe = 0.1$

* From Port Macquarie, New South Wales
total includes:- 1.79 acmite
0.61 Na metasilicate

PART III

CHAPTER 7

ORIGINAL TECTONIC SETTING OF THE PBOC AND ASSOCIATED BASALTIC ROCKS

The present geological setting of the PBOC and associated basaltic rocks has been described in both a regional and a local context in the General Introduction (Sections 1 and 4) and Part I. On the basis of these data it was suggested, in Chapter 3 (Section 3.6), that:

(i) Basaltic rocks in the Myra beds probably represent relatively evolved E-type MORB-like intrusives, extrusives, or intrusive-extrusive complexes which were originally emplaced within or erupted on oceanic layer 1, perhaps well away from active spreading centres. Subsequently these rocks were incorporated in the Woolomin-(?)Sandon accretionary prism (see Chapter 1) along with a diverse range of other basaltic types (? including MORB, IAT, WPA, and minor calc-alkaline basalts-dacites and tholeiitic andesites) which occur elsewhere in the Woolomin Association (see p. 155).

(ii) Basaltic rocks in the Tamworth Belt (i.e. in the Tamworth Group and the Glen Ward beds) were originally emplaced within or erupted on a shallow- to moderately deep-water, poorly-lithified sedimentary sequence. This largely consists of proximal arc-derived 'andesitic' epiclastics and associated argillites, with localized limestones and intermediate-silicic volcanics. The Tamworth basaltic rocks display a spectrum of compositions ranging from: (1) variants resembling evolved IAT and alleged BAB of Pindos type, and whose occurrence is apparently restricted to the general vicinity of the PBOC (see p. 157; *cf.* Offler, 1982); to (2) relatively Ti- and Fe-rich variants from the Nundle-Morrisons Gap area which resemble E-type MORB from propagating rift systems and, to a variable extent, transitional basalts from a number of other oceanic settings (see p. 158).

Most recent reconstructions of the Early-Middle Palaeozoic evolution of the NEO envisage the (proto-) Tamworth Belt as a fore-arc basin adjoining, and perhaps in part overlying, an embryonic accretionary terrane (Woolomin Association - ? ultimately most of Zone B) lying oceanward (e.g. Scheibner, 1973, 1976; Leitch, 1974, 1975, 1979, 1980a, 1982; Crook and Powell, 1976; Day *et al.*, 1978; Crook 1980a,b; Leitch and Cawood, 1980; Cawood 1982a,b,c, in press).

Although this study reinforces an interpretation of the Woolomin Association as an accretionary terrane, it suggests that the evolution of the Tamworth Belt may well be more complex than that commonly envisaged for simple fore-arc basins (e.g. Dickinson, 1970, 1973, 1977; Dickinson and Seely, 1979; Karig *et al.*, 1980).

In particular, the widespread occurrence of substantial basaltic units (which may dominate the lithologies of individual sections) and, locally, intermediate-silicic volcanics [e.g. Copes Creek Keratophyre (Crook, 1961a; Cawood, 1982b, in press) and ? Pitch Creek Volcanics (see Chapter 4)] in the pre- Upper Devonian Tamworth Belt succession contrasts strongly with the virtual absence of igneous activity within most modern fore-arc basins (which are characterized by relatively low heat flow, e.g. MacDonald *et al.*, 1973; Sugimura and Uyeda, 1973) and their ancient analogues. Wherever present within well-established fore-arc basins, penecontemporaneous (w.r.t. fore-arc sedimentary rocks) igneous rocks are negligible in volume [e.g. mafic extrusives in the Vester Formation, Central Oregon (Dickinson, 1979), and in the Palmer Land-Graham Land fore-arc basin, Antarctica (Hyden and Tanner, 1981; Smellie, 1981)]. On the other hand, a variety of 'basaltic' eruptives does occur in the Mariana fore-arc region in the vicinity of the trench slope break* and inner trench wall. These include N-type MORB, IAT, and Ti-poor (boninitic) compositions, and a range of mafic and ultramafic intrusives (which includes gabbronorites) from the trench axis itself (e.g. Dietrich *et al.*, 1978; Rudnik *et al.*, 1979; Meijer, 1980; Bloomer, 1981; Sharaskin *et al.*, 1981).

Recent models proposing the evolution of the Tamworth Belt solely as a fore-arc basin (see references cited above) all envisage an arc-trench system

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* following the terminology of Dickinson (1973, 1974) and Karig and Sharman (1975): equivalent to the 'forearc structural high' of Dickinson and Seely (1979) and Seely (1979), and the 'non-volcanic outer arc' of Van Bemmelin (1954).

of 'normal' polarity (i.e. volcanic arc closer to the Australian continent than the associated Tamworth 'fore-arc' basin *cf.* nomenclature of Dickinson and Seely, 1979). From this configuration it follows that the Tamworth Group basaltic-silicic volcanism occurred within the postulated Tamworth fore-arc basin, but in the vicinity of the trench slope break or, in the absence of such a feature (i.e. a simple sloped fore-arc, see Dickinson and Seely, 1979), near the oceanward margin of the basin.* The Mariana-Bonin fore-arc region is the only reasonably well-documented - and perhaps the only modern example - where significant volcanism occurs in this particular setting, although some ancient analogues might exist. Thus, Crawford *et al.* (1981) propose a general model for oceanic fore-arc evolution based solely on the West Philippine-Mariana region which leads to their suggestion that ophiolites containing Ti-poor basaltic rocks [e.g. the Betts Cove, Papuan, Rambler, Pindos and Troodos ophiolites, and the Cambrian greenstone belts, Victoria (Crawford and Keays, 1978; Crawford, 1980)] may well be remnants of fossil fore-arcs of general Mariana type.

At first sight the general association of Tamworth Belt-PBOC basaltic rocks resembling MORB, IAT and Ti-poor types in an assumed fore-arc setting readily invites a comparison to the Mariana fore-arc region and its particular evolutionary history. However, any apparent similarities between the Tamworth Belt and the Mariana fore-arc region are in essence, only superficial. Fundamental differences include:

(i) In contrast to the Mariana fore-arc region the Tamworth Belt appears to display a well-developed trough or basin structure which acted as a sediment-trap during much of the Palaeozoic.

(ii) The Mariana trench and trench slope is essentially devoid of accreted sedimentary rocks and only minor amounts of 'ponded' sediment occur (including volcaniclastics and limestones, and argillites and sandstones derived locally from the disintegration of mafic-ultramafic 'ophiolitic' lithologies; e.g. Dietrich *et al.*, 1978; Rudnik *et al.*, 1979). In fact, Uyeda (1981, 1982) and Uyeda and Kanamori (1979) cite the Mariana fore-arc system as the type example of an arc-trench system typically lacking an accretionary prism. On the other hand, the comparatively well-

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* Earlier suggestions that Tamworth basaltic rocks might represent uplifted fragments of ophiolite basement to the Tamworth Belt (e.g. Glen and Heugh, 1973; Cross, 1974; Crook and Felton, 1975; Crook and Powell, 1976) were discounted in Chapter 3.

developed Woolomin-(?)Sandon accretionary prism is a fundamental component of current models envisaging a fore-arc setting for the Tamworth Belt (e.g. Cawood, 1982a).

(iii) Although similar in some respects (e.g. characteristically low abundances of incompatible elements) the PBOC low-Ti basaltic rocks and Mariana boninites (and indeed, boninites in general) differ significantly in numerous aspects of their respective mineral and bulk chemistries (e.g. in *Cr* value of spinels, normative *en* and *qz*, trace element ratios, see Sections 5.7.3 and 5.7.4). Moreover, inferred aH_2O in the PBOC melts appears to have been, at the very least, as low as is envisaged for primitive MORB (*cf*. Moore, 1970; Delaney *et al*., 1978; Chaigneau *et al*., 1980) and in any case, substantially (perhaps more than an order of magnitude) lower than that generally envisaged for boninitic melts. It is likely that these may have contained more than several wt % H_2O (e.g. Cameron *et al*., 1979; Cameron, 1980; Meijer, 1980; Hickey and Frey, 1982; see however, Duncan and Green, 1980b). These data imply fundamental differences in the generation and subsequent evolution of boninitic and PBOC melts. However, until a more substantial (and internally consistent) body of data accumulates on Ti-poor basaltic rocks, attempts to integrate aspects of their petrogenesis with models for specific tectonic environments (Crawford *et al*., 1981) may be premature. Notable in this context is the Kopi boninite, New Zealand (Wood, 1980) which erupted in association with alkaline basalts in a relatively stable continental shelf setting.

(iv) The Tamworth Basaltic rocks are intercalated with voluminous, relatively proximal arc-derived epiclastics, whereas in the Mariana example a somewhat analogous sedimentary (?) apron is restricted to the back-arc region (e.g. Karig and Moore, 1975; Rodolfo, 1981; *cf*. Dickinson, 1974).

(v) Tamworth Belt basaltic rocks resemble E-type MORB and evolved IAT (see Section 3.6.2) - an association more akin to the Mariana- and, for example, South Sandwich and Caroline (Sorol Trough) back-arc basins (e.g. Fornari *et al*., 1979; Tarney *et al*., 1981). The Marianas fore-arc region is characterized by N-type MORB, relatively primitive IAT, and boninite (e.g. Dietrich *et al*., 1978; see Crawford *et al*., 1981; and Sharaskin *et al*., 1981 for reviews).

(vi) Intermediate-silicic volcanics (? calc-alkaline, but apparently devoid of orthopyroxene, hornblende ? and biotite; see Chapter 4; Cawood, 1980) are locally abundant near the eastern (oceanward) margin of the Tamworth Belt. At least in part these appear to have erupted subaerially or in relatively shallow water (Crook, 1964) and may be associated with (? genetically related) tonalitic-trondhjemitic intrusives (e.g. Pola Fogal Suite). Although some arc-related volcanics (IAT and boninitic types) do occur in the Mariana fore-arc region, relatively evolved calc-alkaline extrusives and associated/related intrusives appear to be absent. However, calc-alkaline extrusives and/or their volcaniclastic equivalents are abundant in the arc itself (e.g. Shiraki *et al.*, 1978; Meijer and Reagan, 1981) and on the remnant arc (i.e. West Mariana Ridge) in the Mariana system (e.g. Karig and Glassley, 1970; Mattey *et al.*, 1981; Tarney *et al.*, 1981). Conceivably analogous volcanics, with or without associated tonalitic-trondhjemitic intrusives, occur in a relatively large number of other volcanic arcs and, for that matter, remnant arcs in the Western Pacific and elsewhere (e.g. Ishizaka and Yanagi, 1977; Gill, 1981; *cf.* Kesler *et al.*, 1977; Citron *et al.*, 1980). Other examples of remnant arcs displaying these characteristics include the Lau (-Colville) Ridge on the western margin of the Lau basin (Sclater *et al.*, 1972; Gill, 1976; Gill and Stork, 1979) and the Tobi Ridge area on the western margin of the Ayu Trough (Fornari *et al.*, 1979).

Thus the pre-Parry Group (pre-Frasnian, *cf.* Fig. 1.2) Tamworth Belt displays some general characteristics in common with modern arc - back-arc systems (e.g. points iv-vi above), suggesting that, at least during its early evolution, the Tamworth Belt occupied a back-arc setting.

Inter-arc - Back-arc Rifting : General Concepts

In view of a degree of 'woolliness' (*cf.* Partridge, 1977) in current applications of back-arc terminology, it is necessary to define the usage of certain terms employed in the following discussion.

The terminology adopted generally conforms to that of Karig (1970, 1974) as modified by Dickinson and Seely (1979). However, to readily distinguish between three significantly different back-arc environments the terms (1) marginal basin, (2) inter-arc basin, and (3) back-arc basin are respectively restricted to: (1) typically inactive 'back-arc' (in the sense of

Dickinson and Seely, 1979) basins bordering passive continental margins (e.g. Japan Sea, Sea of Okhotsk; *cf.* Packham and Falvey, 1971), including some more-or-less ensialic rifts* (e.g. Gulf of California, Saunders *et al.*, 1982); (2) basins situated between two active volcanic arcs (e.g. Waitemata Group basin, New Zealand; Ballance, 1974; *cf.* Fig. 2D of Dewey and Bird, 1970) which are typically actively spreading and perhaps most commonly reflect the embryonic and immature stages of 'back-arc' basin development where the (ultimately) remnant arc might remain active for a (?)short period, and (3) oceanic basins lying between active arc - fore-arc regions and inactive remnant arcs, and which presumably reflect intra-arc rifting and subsequent back-arc spreading (Karig, 1971a, 1972). Where necessary and wherever original data sources permit, the published terminology of certain 'back-arc' basins (i.e. those referred to for comparative purposes) has been altered to conform to the above constraints.

Although it has long been recognized that most modern inter-arc, back-arc and marginal basins have been generated by rifting processes (e.g. Karig, 1970, 1971a,b, 1972; Sclater *et al.*, 1972; Luyendyk *et al.*, 1973) only recently have adequate data become available on the nature of the basaltic volcanism occurring in the initial stages of the development of such rifts. Where these involve the splitting of relatively immature volcanic arcs (e.g. Western Pacific and Southern Atlantic types), initial and early rift volcanism may include mixed populations of 'depleted' and 'enriched' IAT- and MORB-like basalts and basaltic rocks transitional between these types, and, on occasion, boninitic types (e.g. Crawford *et al.*, 1981; Tarney *et al.*, 1981; Sharaskin *et al.*, 1981)**. The spatial and temporal distributions of these various basalt types are not clearly defined. However, it is conceivable that these early rift volcanics might display bimodal and/or mixed characteristics of the pre-rift IAT and the 'enriched' MORB-like basalts (*cf.* Tamworth basaltic rocks) typical of some youthful propagating rift systems in major ocean basins (e.g. Atwater, 1981; Delaney *et al.*, 1981; le Roex *et al.*, 1982; Sinton *et al.*, 1982) and embryonic marginal basins such as the Guaymas Basin, Gulf

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* NOTE: ensialic basins developed behind Andean-type continental margin volcanic arcs are excluded from this discussion.

** Transitional and alkaline volcanism characterizes the initial stages of rifting in mature Mediterranean arc systems (e.g. Lordkipanidz *et al.*, 1979) and marginal basins (e.g. Hynes, 1974; Menzies, 1976).

of California (Saunders *et al.*, 1982). Nevertheless, once the arc-related rifts become well-established, rifting appears to proceed intimately associated with the eruption of N-type MORB (e.g. Hawkins, 1976; Mattey *et al.*, 1981) and/or unusually volatile-enriched equivalents (e.g. Delaney *et al.*, 1978; Garcia *et al.*, 1979; Dick, 1980) at spreading centres more-or-less analogous to the mid-ocean ridges (e.g. Karig, 1971a; Moberly, 1972; Barker and Hill, 1981; Weissel, 1981). Intra-basin volcanism may, however, give rise to some arc-like, transitional ('enriched') and, on occasion, distinctly alkaline basaltic types (e.g. Hart *et al.*, 1972; Saunders and Tarney, 1979; Dick, 1982).

Although basaltic rocks apparently generated at the inception of presently active inter-arc or back-arc rifts largely conform in gross chemistry to the above generalizations, inter- and intra-rift variation is such that it is not possible to define in any detail basalt compositions which are specifically diagnostic of these settings. Variations in basalt type might be related to:

(i) The evolutionary stage of the arc at the time of initial rifting and the nature of any arc-like chemical characteristics it might contribute to the early rift basalts (*cf.* Tarney *et al.*, 1981).

(ii) Differing mechanisms of rift generation - the literature abounds with hypotheses and notions for the origin of arc-related rift basins. Some recent examples which review earlier hypotheses and provide a range of additional concepts include Dubois *et al.* (1978), Molnar and Atwater (1978), Poehls (1978), Sydora *et al.* (1978), Toksoz and Hsui (1978), Uyeda and Kanamori (1979), Nakamura and Uyeda (1980), Hsui and Toksoz (1981) and Uyeda (1981, 1982). These models generally appeal to variations in subduction rates and geometries, and (?)associated 'convection' in the overlying mantle wedge (*cf.* Roeder, 1975; Stevenson and Turner, 1977; Hager and O'Connell, 1978; Tovish and Schubert, 1978; Yokokura, 1981) and/or relative buoyancy of the subducting slab. Of these parameters, usually only the latter can be constrained with any confidence for only some fossil subduction-related systems (because it is related to the age of the subducting lithosphere relative to that of the over-riding plate; e.g. Molnar and Atwater, 1978), and even then only within broad and perhaps non-diagnostic limits. Consequently, applications of these concepts to the mechanism of initiation and subsequent evolution of ancient 'back-arc' settings must remain highly speculative (see below).

(iii) Heterogeneities in mantle source regions, either pristine [see Section 3.5.1(2)], or introduced as a result of: (1) locally subducted material (e.g. Tarney *et al.*, 1981); (2) regional (?)upper mantle metasomatism (e.g. Bailey, 1982); or (3) partial melting processes with subsequent segregation and perhaps removal of melts.

Despite occasional differences in chemistry largely in trace element abundances ('relatively depleted') IAT-'enriched' MORB association characterizing initial rift volcanism in some modern inter-arc and back-arc basins is also a common feature of many ophiolitic and related assemblages (*cf.* Sun and Nesbitt, 1978) cropping out at the margins of ancient basins - many of which have been interpreted as marginal, back-arc or inter-arc rifts on the basis of similar (and other) criteria. In particular, there is a close spatial association between members of arc-related rift assemblages of this type (i.e. IAT-like basalts + basalts transitional between these and E-type MORB ± Ti-poor basaltic rocks and, in the originally slightly more evolved arc settings, andesitic and silicic volcanics) in: (1) the Aspropotamos sequence of the Pindos Ophiolite (Capedri *et al.*, 1980; Pe-Piper, 1982); (2) the Karmoy Ophiolite (Furnes *et al.*, 1980), the Gullfjellet and Lykling ophiolitic complexes (Furnes *et al.*, 1981) and the Likken Greenstones (Gale and Roberts, 1974; Gale and Pearce, 1982), Norway; (3) the Semail Ophiolite, Oman (Searle *et al.*, 1980; Pearce *et al.*, 1981); (4) the Khan Taishir Ophiolite, Western Mongolia (Zonenshain and Kuzmin, 1978a,b); (5) the Sarmiento Complex (e.g. Saunders *et al.*, 1979; Stern, 1979, 1980); (6) ophiolitic basalts from the Eastern Maryland Piedmont (Handy and Wagner, 1982); and (7) the Kure basaltic rocks, Pontid Ranges (Anatolian Trough), Turkey (Guner, 1983).

On the basis of their contentious assumptions (*cf.* Duncan and Green, 1980a,b; Jenner, 1981) that most or all Ti-poor basaltic rocks are analogous in their mode of origin to the Mariana boninites, and that the Mariana Arc provides a genetic model widely applicable to the original tectonic setting of rocks of this type, Crawford *et al.* (1981) suggest that a number of the more well-known Ti-poor 'basalt'-bearing ophiolites (e.g. Troodos, Betts Cove, Rambler, and possibly Pindos and Cape Vogel) represent fossil fore-arc associations. Alternative interpretations for the original setting of these ophiolites drawing on more wide-ranging geological data (in addition to the gross MORB-Ti-poor 'basalt' association employed by Crawford *et al.*, 1981) typically favour an arc, inter-arc or back-arc environment (e.g. Norman and Strong, 1975; Pearce, 1975; Upadhyay and Neale, 1979). For the most part, back-arc settings appear to be more feasible for these ophiolites because, with the possible exception of Troodos, the geological environments indicated by the overlying volcanic and/or clastic rocks are not definitive of an active (rejuvenated) fore-arc (*cf.* Fig. 3E of Crawford *et al.*, 1981). On the other hand, these opposing views on fore-arc-back-arc ophiolite settings can be reconciled within the broader terms of Crawford *et al.*'s (1981) model if Ti-poor basaltic rocks are erupted in the embryonic inter-arc rift in close association with early rift (?'enriched', see above) extrusives. This does not preclude the possibility of more-or-less coeval eruption of Ti-poor types in the associated fore-arc region. Moreover, this situation allows for the possible formation of 'Ti-poor ophiolites', such as the PBOC and perhaps somewhat similar examples in Tasmania (Brown *et al.*, 1980), Mongolia (Zonenshain and Kuzmin, 1978a,b) and Papua New Guinea (Jaques and Chappell, 1980; Jaques, 1981), as an integral part of the initial development of inter-arc and back-arc basin crust.

In this general context it is noteworthy that the predominant basalt type in most ophiolites usually displays closer affinities with E-type MORB (Cameron *et al.*, 1980) than with the N-type MORB typical of the major ocean basins, and that they are typically exposed at the margins of their host basins. Irrespective of the development of Ti-poor basaltic types in these ophiolites, it is conceivable that they represent assemblages emplaced during the initiation of rifting and that by their mode of origin, basin-margin ophiolites of this type are more readily exposed than examples which might resemble more closely 'normal' oceanic crust. These, in turn, might be expected to characterize the basin floors (by analogy with modern 'back-arc' basins, see above).

Tamworth Belt

If the general characteristics of the pre-Parry Group Tamworth Belt succession are critically examined in the light of this discussion of arc-related settings, two principal alternatives for the pre-Famennian history of the belt emerge.

1. The Tamworth Belt evolved as an anomalous (perhaps unique) fore-arc basin within which large volumes of 'basalt' and lesser volumes of intermediate-silicic extrusives/intrusives were erupted/emplaced, apparently after the basin was relatively well-established. In a broad sense, this model is favoured by Cawood (1980, 1982a,b,c, in press) and Leitch (1975, 1979, 1982) who also propose or imply that: (i) arc-related volcanism and

associated clastic sedimentation occurred at or near the present eastern margin of Zone A from the (?)Cambrian until the Early-Mid Devonian; (ii) the locus of arc volcanism migrated westward (towards the Australian continent) with further evolution of the basin until the Early Permian. Alternatively, arc volcanism 'jumped' eastward into the 'Tamworth fore-arc basin' in the Early Devonian following eastward migration of the associated subduction zone, then returned to the western margin of the basin for reasons which are not obvious from this general fore-arc model; and (iii) subduction and (?)accretion occurred in proto-Zone B intermittently or continuously throughout almost the entire Palaeozoic Era! These interpretations may be significantly modified in the light of the new data available following this study:

2. The PBOC and basaltic rocks in the Tamworth Belt were emplaced or erupted in an arc-related rift setting at least in some respects analogous to that envisaged for the 'enriched' MORB-Ti-poor 'basalt' association which may characterize some inter-arc-back-arc rift assemblages (see above). Figure 7.1 schematically illustrates a tentative back-arc model for the early evolution of the Tamworth Belt. This model encompasses a rift-related origin for the Tamworth 'basalts' and the PBOC, and the intercalation of the former with voluminous arc-derived clastics (*cf.* rifting processes in the Guaymas marginal basin, Gulf of California, Saunders *et al.*, 1982).

Stages A-C of Figure 7.1 involve:

A. Migration of a 'reversed polarity' (*cf.* Dickinson and Seely, 1979) active arc (a) towards the Australian 'continent' in the Early Palaeozoic - and the accumulation of moderately deep-shallow water volcanogenic epiclastics and limestones on its flanks (i.e. the ?Cambro-Ordovician strata of Cawood, 1976, *in press*).

B. Collision between arc (a) and the eastern margin of the proto-Lachlan Fold Belt (b) (? frontal arc of Scheibner, 1973: *cf.* Cas *et al.*, 1981) followed by the development of a 'normal polarity' Mariana-type fore-arc system (*cf.* Uyeda, 1981, 1982), initiation of back-arc (?inter-arc) spreading which involved formation of the PBOC, (?)subsequent Tamworth Belt basaltic volcanism associated with the developing rift and its relatively coarse-grained volcaniclastic in-fill, sundering of the remnant arc (a), and emplacement/eruption of Pola Fogal-type intrusives and Pitch Creek-type volcanics in the volcanically active fore-arc region (c).

C. Gradual transformation of the convergent plate boundary from a Mariana-type to a Chilean-type; perhaps (?) due to the subduction of progressively younger and therefore more buoyant oceanic crust (see below) with consequent shallowing of the dip of the subducted plate (*cf.* Molnar and Atwater, 1978) and relatively rapid westward migration of the arc (d), concomitant development of the Woolomin-(?)Sandon accretionary prism (e) and (?) initiation of (?) related blueschist metamorphism (∿ 378 m.y.). Arc (d) presumably became well-established in the early Late Devonian, shedding andesitic detritus across the entire width of the Tamworth fore-arc basin (e.g. Baldwin Formation; Chappell, 1968; *cf.* discussion in Korsch and Harrington, 1981a).

The interpretation of a Chilean-type convergent plate margin setting for the Middle-Upper Palaeozoic evolution of the Tamworth Belt is consistent with the overall regressive sedimentary sequence (see General Introduction) and the major development of Middle-Late Carboniferous calc-alkaline volcanics (dominated by silicic pyroclastics) along the western margin of the belt (e.g. Leitch, 1974; Korsch and Harrington, 1981a; McPhie, 1982). Little direct evidence is available to evaluate the suggestion that subduction of progressively younger oceanic crust might have caused the transition from a Marianas-type to a Chilean-type convergent plate margin. However, it is interesting to speculate that this might have led ultimately to the subduction of a palaeo-Pacific spreading ridge resulting in widespread and voluminous intrusion of granitoids throughout Zone B (and to a minor extent in Zone A) in the (?)Late Carboniferous and Permian (*cf.* Marshak and Karig, 1977; DeLong *et al.*, 1978; Hill *et al.*, 1981).

Although a Mariana-type convergent plate boundary appears to characterize Stage B in the suggested early history of the Tamworth Belt, it is emphasized that the PBOC melts emplaced/erupted during this stage differ significantly from the boninitic types erupted in the Mariana fore-arc system. Whereas the latter appear to have been generated by hydrous partial melting of depleted upper mantle source rocks overlying the subducting (?dehydrating) plate, the PBOC melts appear to reflect anhydrous partial melting of depleted (but not highly refractory) upper mantle (?)diapirs (*cf.* Duncan and Green, 1980a,b). These might have risen significantly in the (?)tensional stress regime accompanying the initial stages of rifting in the proto-Tamworth Belt.

The original areal extent of the PBOC is largely unknown. Altered Ti-poor basaltic rocks which are similar in some respects to the PBOC examples

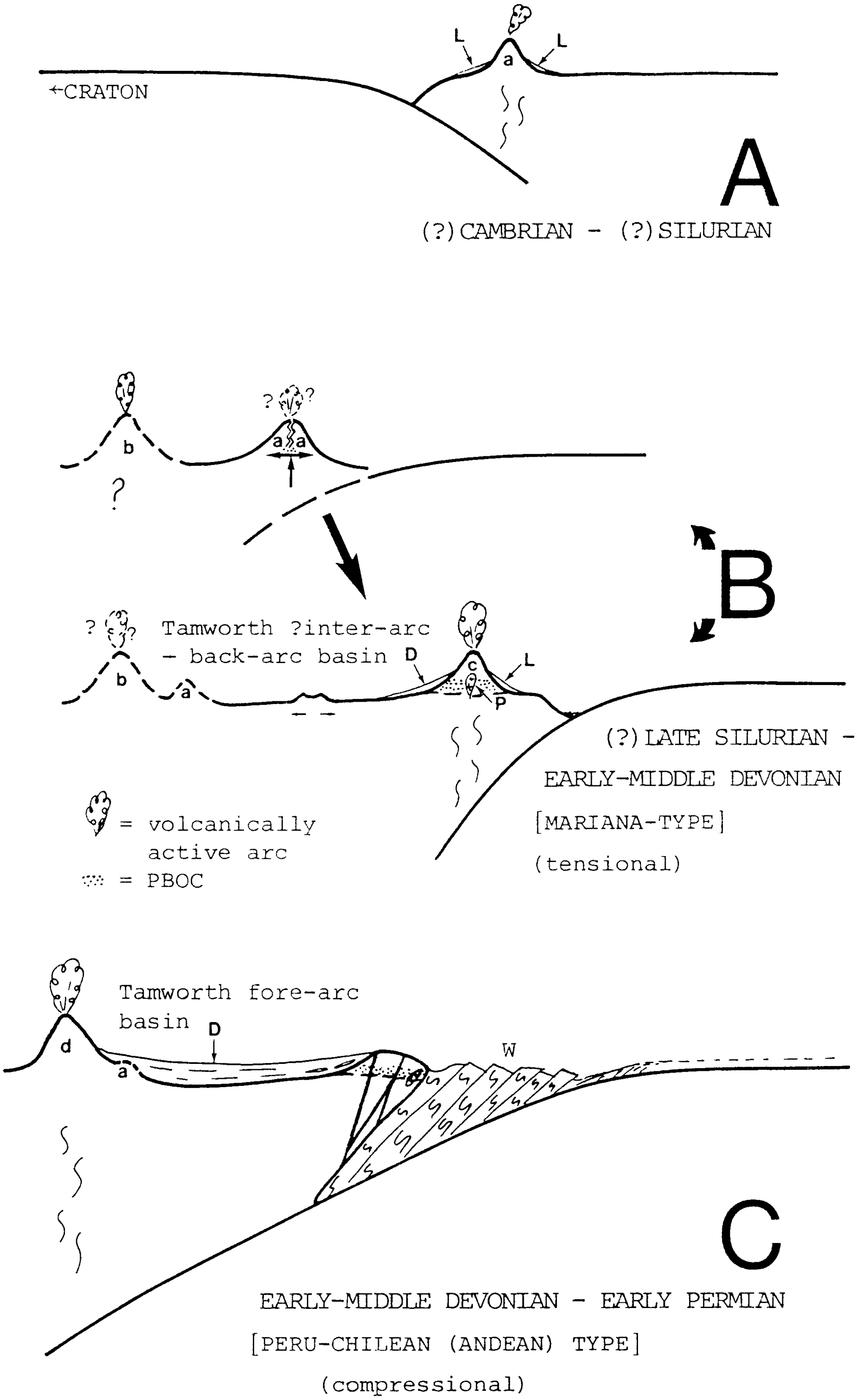

Fig. 7.1: Schematic outline of the three principal stages envisaged for the pre-Early Permian evolution of the Tamworth Belt - PBOC - Woolomin Association.
Cross-sections are not drawn to scale.
See text for discussion.
a = Lower Palaeozoic volcanic arc; b = frontal ?volcanic arc of the proto-Lachlan Fold Belt; c = volcanically active fore-arc region; d = post-Mid Devonian Andean-type volcanic arc; L = ?Lower Palaeozoic sediments; D = devonian sediments; P = Pola Fogal intrusives/?Pitch Creek Volcanics; W = Woolomin Association.

(see Section 5.7) occur in the Woolomin beds in the Glenrock Station area (Offler, 1982a). However, few data are available on the field relations of these particular rocks (tectonic blocks?; pre-prism ocean floor- or intraprism intrusives/extrusives?) and any implications they might have for the origin and evolution of the PBOC remain enigmatic. In contrast with the PBOC, cumulates associated with the Peel Fault System elsewhere are typically clinopyroxene-rich, and orthopyroxene-poor or orthopyroxene-free (e.g. Appendix F). Although few data are available on these intrusives, they are possibly more akin to MORG than the PBOC cumulates. Compared with those in the PBOC, basaltic rocks elsewhere in the NEO are significantly enriched in Ti and other incompatible elements.

If PBOC-type ophiolitic rocks were originally widespread along the eastern margin of the Tamworth Belt, it is conceivable that westerly-directed overthrusting by the Woolomin Association at some stage in the post-PBOC evolution of the NEO (*cf.* Scheibner and Glen, 1972; Scheibner, 1976) might be largely responsible for the restriction of present exposures to the predominantly east-west trending PBOC (where concurrent faulting might have been largely strike-slip). In relation to the post-Devonian evolution of other elements of the NEO, geological relationships in the Pigna Barney-Curricabark and Morrisons Gap areas impose few tangible constraints over and above those already incorporated in more general models and syntheses of earlier workers (see references cited previously).