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MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Evolution of Regional Metamorphism during Back-Arc Stretching and Subsequent Crustal Shortening in the 1.9 Ga Wopmay Orogen, Canada [and Discussion]

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Evolution of regional metamorphism during back-arc stretching and subsequent crustal shortening in the 1.9 Ga Wopmay Orogen, Canada†

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[Plates 1 and 2]

The stratigraphic units, structural elements and metamorphic mineral assemblages of a regional metamorphic culmination in the 1.9 Ga Wopmay Orogen are exposed over greater than 30 km of composite structural depth, in a series of oblique sections produced by cross folding. Regional metamorphism developed continuously in three sequential, rapidly changing thermo-tectonic régimes within an evolving continental magmatic arc. At *ca.* 1900 Ma, stretching of intra-arc crust resulted in the accumulation of clastic sediment and bimodal volcanic rift-fill deposits. The onset (first stage) of regional metamorphism is marked by high-*T*–low-*P* mineral assemblages, condensed metamorphic zonal sequences and extensive areas of high-grade gneisses devoid of associated plutons. These features are interpreted in terms of a high thermal gradient related to stretching and thinning of the continental lithosphere. Five to ten million years after stretching, following deposition of a west-facing sedimentary prism, a suite of 1896–1878 Ma plutons was emplaced into the rift and margin deposits as they underwent subhorizontal shortening and deformation during the Calderian Orogeny. Thrusted and folded syn-orogenic foredeep deposits are also intruded by the syn-tectonic plutons. At high and intermediate structural levels, syn-tectonic metamorphic mineral growth and metamorphic zonal sequences which are spatially related to the plutons, document heat advection into the deforming marginal prism and mark a second stage of regional metamorphism related to the emplacement of the plutonic bodies. Inverted mineral isograds in autochthonous Proterozoic units beneath a basal décollement record downward thermal relaxation of isotherms following east-directed Calderian transport of the deformed, thickened, and still hot marginal prism over a relatively cold basement. Derivation of multi-point *P*–*T* trajectories from post-tectonic, poikiloblastic garnets charts metamorphic mineral growth during uplift and erosion of the internal zone, documenting the third (final) stage of regional metamorphism in Wopmay Orogen. The short erosional time interval (less than 11 Ma) between tectonic thickening and the end of uplift constrains the heat required for this last metamorphic stage to be inherited from the two preceding thermo-tectonic régimes: epicontinental stretching and the emplacement of the syn-tectonic plutonic suite.

1. Introduction

Growth of succeeding and contrasting metamorphic mineral assemblages is a sensitive monitor of evolving *P*–*T*–deformation conditions within continental crust which is undergoing changes in thermo-tectonic régimes (Thompson & England 1984). In a rapidly evolving tectonic setting, metamorphic minerals have the potential of documenting changing physical conditions of successive thermo-tectonic regimes. This potential is increased if there is continued addition of heat or insufficient time for the crust to undergo significant cooling following establishment of the initial thermal perturbation.

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In the early Proterozoic Wopmay Orogen, field and laboratory studies have resulted in the identification of three such successive stages in the evolution of regional metamorphism. Each of the three stages can be linked to a distinct thermo-tectonic regime which developed within an evolving continental magmatic arc. These regimes have been identified and integrated into the evolutionary tectonic model of Hildebrand *et al.* (1986*a*, *b*) by using stratigraphic (Hoffman 1980; Easton 1982), structural (King 1986), petrological (St-Onge 1981, 1984*a*, *b*, 1986), and geochronological (Bowring & Van Schmus 1986) data. The onset of regional metamorphism is related to a phase of epicontinental back-arc rifting. Further growth of the metamorphic culmination is a consequence of magmatic advection of heat by plutons emplaced into rift and margin deposits undergoing subhorizontal shortening and deformation. Following tectonic thickening, the final configuration of the regional metamorphic culmination is achieved by relaxation of isotherms during uplift and erosion. Each stage in the evolution of the regional metamorphism is characterized by mineral assemblages and P–T trajectories that are specific to the distinct thermo-tectonic regimes. The integration of the metamorphic and geochronological data allows a qualitative evaluation of the contribution to the regional metamorphism by differing heat sources within the evolving magmatic arc tectonic environment.

2. Wopmay Orogen

Wopmay Orogen (McGlynn 1970) is a 1.9 Ga (early Proterozoic) orogenic belt (Hoffman 1980; Hoffman & Bowring 1984) located in the northwest corner of the Canadian Precambrian Shield (figure 1). The northern part of Wopmay Orogen, between latitudes 65° N and 67° 45′ N (figure 1), is composed of three major tectonic elements (Hoffman & Bowring 1984). On the east is a west-facing depositional prism, the Coronation Supergroup (Hoffman 1981), most of which has been detached from its basement, shortened and tectonically thickened as it was displaced eastward, relative to the underlying 3.5–2.5 Ga (Henderson 1981; Bowring & Van Schmus 1986) Slave craton (figure 1) during the *ca.* 1885 Ma Calderian Orogeny (Tirrul 1982; Hoffman *et al.* 1983; St-Onge *et al.* 1983; Tirrul 1983; Hoffman & Bowring 1984; Hoffman *et al.* 1984; St-Onge *et al.* 1984; King 1986). The easternmost part of the Coronation Supergroup is autochthonous with respect to the underlying Archaean basement. A frontal thrust marks the eastern boundary of the Asiak Thrust-Fold Belt (figure 1), the low-taper, east-verging foreland of the Calderian Orogeny that developed in the shelf stratigraphy of the Coronation Supergroup (Hoffman 1973, 1980; Tirrul 1984*b*). Down-plunge cross sections and bed-length balancing show that the shelf strata were translated a minimum of 45 km eastwards towards the craton (Tirrul 1984*b*). West of the displaced Coronation Supergroup continental-shelf edge (figure 1) the structural style changes from typical foreland thrust-fold geometry to polydeformed and metamorphosed units in the Hepburn Metamorphic-Internal Zone (Hoffman *et al.* 1978; Hoffman 1980; St-Onge *et al.* 1982, 1983; King 1984; St-Onge *et al.* 1984; King 1986). The Coronation Supergroup accumulated between 1.9 Ga and 1885 Ma (Hoffman & Bowring 1984). A description of the stratigraphy of the Supergroup within the metamorphic-internal zone is given below (see §4*a*).

Hottah Terrane (McGlynn 1976; Hildebrand 1981) (figure 1) is the westernmost tectonic element of the exposed Wopmay Orogen. The terrane is a complex of amphibolite facies sedimentary and intermediate volcanic rocks cut by 1914–1902 Ma biotite–hornblende

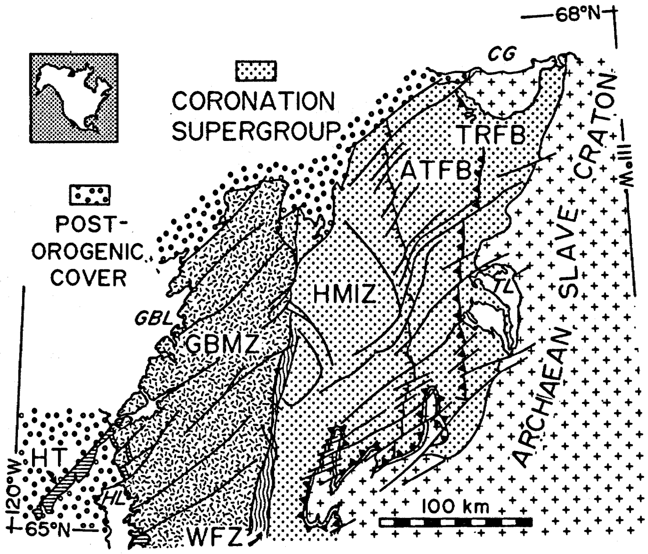

FIGURE 1. Location and major tectonic elements of the northern exposed Wopmay Orogen (modified from Hoffman & Bowring 1984). Symbols: line with ticks, Coronation margin shelf edge; line with solid triangles, Calderian frontal thrust and décollement; heavy lines, late transcurrent faults. Abbreviations: ATFB, Asiak Thrust-Fold Belt; GBMZ, Great Bear Magmatic Zone; HMIZ, Hepburn Metamorphic-Internal Zone; HT, Hottah Terrane; TRFB, Tree River Fold Belt; WFZ, Wopmay Fault Zone. Prominent bodies of water: *CG*, Coronation Gulf; *GBL*, Great Bear Lake; *HL*, Hottah Lake; *TL*, Takijuq Lake.

bearing calc-alkaline plutons (Hildebrand *et al.* 1986*a*). The volcanic rocks and the plutons are interpreted to be part of an Andean-type continental magmatic arc built on crust contiguous to the Archaean Slave Craton (Hildebrand & Roots 1985).

Between the Coronation Supergroup and the Hottah Terrane is the third tectonic element of the orogen, the Great Bear Magmatic Zone (figure 1). This zone is an 1875–1840 Ma (Bowring & Van Schmus 1986) volcano-plutonic depression (Hoffman & McGlynn 1977; Hildebrand 1983; Hildebrand *et al.* 1983; Hildebrand & Bowring 1984) that unconformably overlies the Hottah Terrane and the Archaean rocks that are infolded with the western part of the deformed Coronation Supergroup in the Wopmay Fault Zone (figure 1).

3. Down-plunge section of the metamorphic internal zone

Post-Calderian cross folding (see §4*d*) of the metamorphic-internal zone has resulted in a wide range of (Calderian) structural levels being exposed at the present erosion surface. To examine the component stratigraphic, structural and metamorphic elements of differing structural levels within a meaningful regional framework, the hinge zones and limb dips of the late cross folds were defined from a compilation of plunges of Calderian fold axes (figure 2) (King 1986). In the eastern internal zone, where stratigraphy is well constrained (Hoffman 1984) and Calderian structures simple (see §4*b*), the Calderian plunges define consistent, macroscopic domains of the north and south plunges (figure 2). Together with the great lateral continuity of stratigraphic units, thrusts and macroscopic folds (figures 3 and 4; Hoffman 1984), this is taken to indicate that Calderian structures in the Proterozoic cover are approximately cylindrical *at the regional scale* (King 1986). Macroscopic deformation of the basement is shown

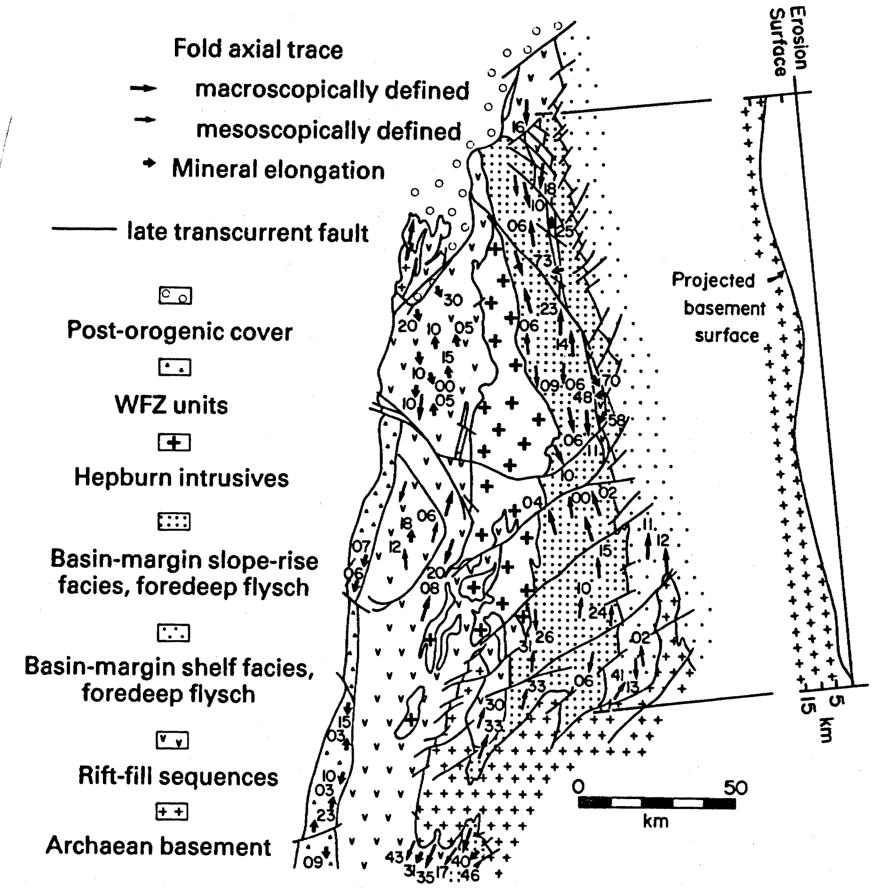

FIGURE 2. Compilation of Calderian linear fabric elements within the metamorphic-internal zone. On the right is the length section of the eastern internal zone constructed from the undulating Calderian plunges. Where plunge value is not indicated, the trend was derived from stratigraphic relations, but structural data were insufficient to derive plunge.

by King (1985) to be mainly a result of interfering fold structures, and to be projectable. West of the Hepburn Batholith (figures 3 and 4), relatively discontinuous stratigraphy, and the occurrence of more complex Calderian structures preclude a simple assumption of projectable, cylindrical structure. Domains of mainly north or south plunges are, however, broadly related to those east of the batholith (figure 2).

Several east–west transects of the internal zone were chosen, at widely spaced latitudes (figures 3 and 4) on which to construct locally controlled down-plunge or vertical cross sections. The parameters and techniques of construction of individual sections are given in King (1985). These independent cross sections were 'stacked' according to their relative structural position predicted by the strike-parallel or length profile of figure 2. From this, a composite cross section, incorporating *all* exposed structural levels of the internal zone was constructed (figure 5) (King 1986). This cross section represents over 30 km of composite structural relief. Different transects within the cross section that were predicted to represent approximately the same structural level are, although widely separated in map view (e.g. sections F–F′, figure 4; and G–G′, figure 3), compatible in terms of major structures and stratigraphic level. The method of construction is therefore thought to be internally consistent.

The isograd surfaces are considered to be sufficiently continuous and predictable in a suitable orientation to be projected onto the composite cross-section because of: (*a*) the mirrored repetition of isograd traces in the north and south-plunging hinge zones of large basement-cored folds (figure 4) suggesting continuity over 50 km in map view and (*b*) the distinct linearity, and coparallelism with Calderian fold axial traces, of the isograd traces for over 170 km (St-Onge 1981; Hoffman 1984). Deflections in the isograd traces have been shown to be concentric folds of the isograd surfaces about Calderian fold axes (St-Onge 1981).

The composite down-plunge section (figure 5) provides a framework in which to examine the stratigraphic and structural elements of the metamorphic-internal zone, as presented in the following section, and to constrain the evolution of the thermo-tectonic regimes of metamorphism (see §5).

4. Hepburn metamorphic-internal zone

(*a*) *Stratigraphy*

The allochthonous early Proterozoic stratigraphy of the Hepburn Metamorphic-Internal Zone comprises the clastic rocks and bimodal volcanics of the Akaitcho Group, the continental slope-rise facies units of the Epworth Group and the syn-orogenic flysch deposits of the Recluse Group (Hoffman 1980, 1981) (figures 3 and 4).

Structurally, beneath a basal décollement in the internal zone, a mantle of basal, transgressive deposits of the Epworth Group overlies Archaean units of the structural basement. The autochthonous basal Proterozoic units and the underlying Archaean basement are extensively exposed in a series of adjoining structural culminations in the southern part of the map area (figure 4). The culminations are structurally continuous with the main part of the Slave craton and are named from east to west, the Carousel, Scotstoun and Exmouth culminations (figure 4).

(i) *Archaean basement*. Units of the Archaean basement include a granite, which is massive, foliated or gneissic (Carousel and Scotstoun culminations, figure 4), a basic metavolcanic belt (Redrock Lake Volcanic Belt, figure 4) and an assemblage of biotite and hornblende quartzo-feldspathic gneisses and amphibolites (Exmouth culmination, figure 4).

(ii) *Akaitcho Group: epicontinental back-arc rift facies*. The western part of the internal zone (figures 3 and 4) is characterized by the complex assemblage of metasediments and bimodal igneous rocks of the Akaitcho Group (Hoffman *et al.* 1978). The assemblage is dominated by arkose, feldspathic quartzite and semipelite, gabbroic sills, pillowed and massive basalt, and subalkaline felsic flows, tuffs and sills. Lesser amounts of pelite, quartzite, dolomite and conglomerate are also present. Detailed descriptions of these units can be found in Easton (1980, 1981 *a–c*, 1982, 1983). U–Pb (zircon) dates from the western part of the Akaitcho Group cluster at 1900 Ma (Bowring & Van Schmus 1986).

The rocks of the Akaitcho Group have been interpreted, on the basis of stratigraphic character and geochemistry, to be epicontinental rift-fill deposits (Hoffman *et al.* 1978; Easton 1981 *a, b, c*, 1982, 1983; King 1986). Hildebrand *et al.* (1986 *a*) have suggested that the rift basin (Coronation marginal basin) developed on subsided, intra-arc continental crust following the establishment of an Andean-type arc in Hottah Terrane between 1914 and 1902 Ma.

(iii) *Epworth Group: basal transgressive shelf and slope-rise facies*. The Epworth Group is a west-facing sedimentary prism that overlies both the rift-fill units of the Akaitcho Group and the non-rifted part of the Slave craton (Hoffman 1972, 1973, 1980; Grotzinger & Hoffman

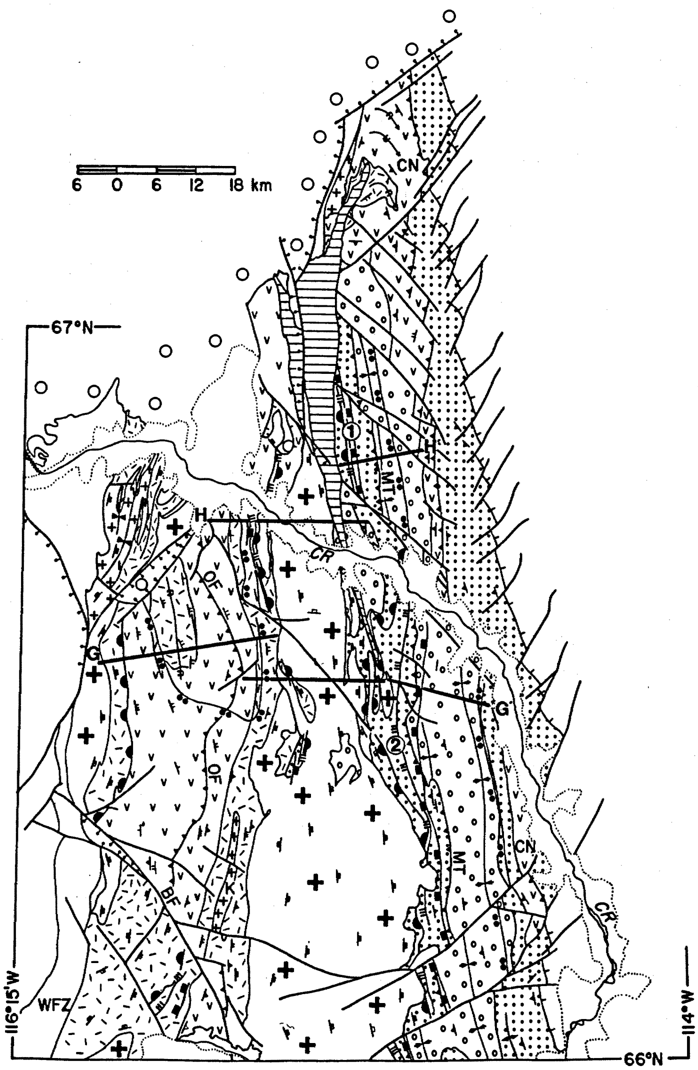

FIGURE 3. Simplified geological map of the metamorphic-internal zone of Wopmay Orogen between 66° N and 67° 45′ N (after Hoffman 1984). Abbreviations: BF, Belleau Fault; CN, Cloos Nappe; K, Kapvik granite; MT, Marceau Thrust; OF, Okrark Fault; WFZ, Wopmay Fault Zone. Bodies of water: *CR*, Coppermine River. Heavy lines identified with letter pairs correspond to lines of section for figure 5. Circled numbers identify locations of metamorphic geothermobarometry samples.

1983; Hoffman & Bowring 1984). Locally, units in the lower Epworth Group interfinger with units of the Akaitcho Group (King 1986). The contact with the Akaitcho Group is only locally preserved (figure 2) but the transgressive unconformity on Archaean units is extensively exposed in the metamorphic–internal zone. The basal units include supermature quartz-arenite, local polymictic conglomerate and stromatolitic marble mounds. The basal unconformity is not faulted and there are no dyke swarms cutting basement. The exposed unconformity is therefore interpreted to have developed on non-rifted basement.

Slope-rise facies units of the Epworth Group, consisting predominantly of graded, graphite argillite beds with thin, graded beds of white quartzite, characterize the eastern part of the metamorphic–internal zone (figures 3 and 4) above the basal décollement. In the more western part of the zone, feldspathic sandstones compose most of the section. The slope-rise facies units

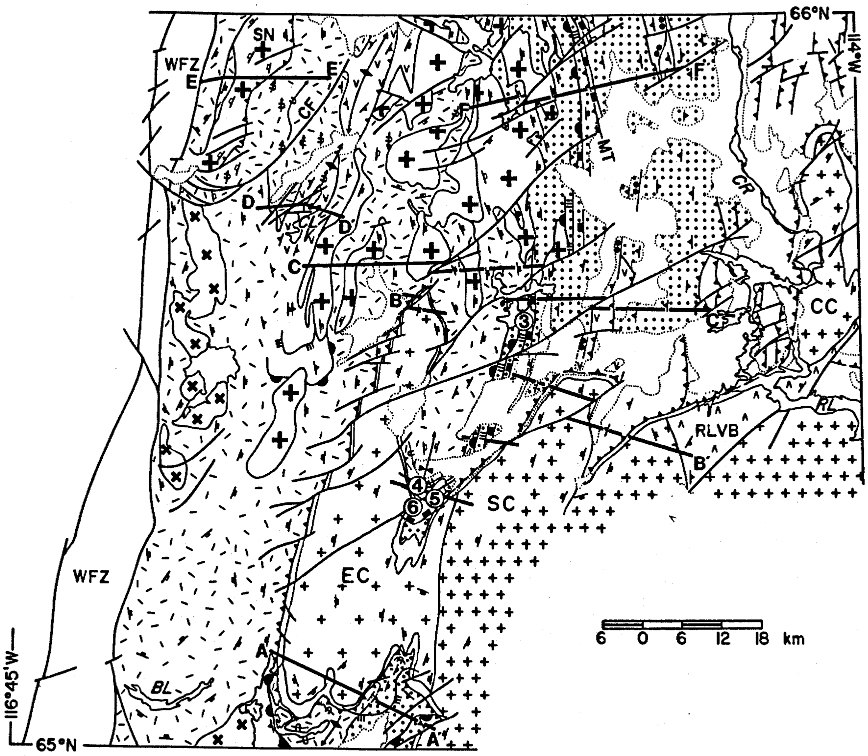

FIGURE 4. Simplified geological map of the metamorphic-internal zone of Wopmay Orogen between 65° N and 66° N. Abbreviations: CC, Carousel culmination; CF, Calder fault; EC, Exmouth culmination; MT, Marceau Thrust; SC, Scotstoun culmination; SN, Sitiyok Nappe; WFZ, Wopmay Fault Zone. Bodies of water: *BL*, Bent Lake; *CL*, Calpyso Lake; *CR*, Coppermine River; *RL*, Redrock Lake. Heavy lines identified with letter pairs correspond to lines of section for figure 5. Circled numbers identify locations of metamorphic geothermo-barometry samples.

KEY TO SYMBOLS IN FIGURES 3 AND 4

middle Proterozoic
- cover rocks
- post-orogenic gabbro

early Proterozoic
- Great Bear Batholith
- Hepburn Batholith
- Recluse Group
- Epworth Group
  - basal-transgressive and shelf facies
  - slope-rise facies
- Akaitcho Group
  - metasediments, minor metabasites, metafelsites
  - metabasites, minor metasediments, metafelsites

Archaean
- massive to gneissic granitoids
- metabasalts
- quartzo-feldspathic gneisses, amphibolite

structures
- gneissic layering (dip 0–30, 31–60, 61–90)
- crenulation cleavage (dip 61–90)
- schistosity (dip 0–30, 31–60, 61–90)
- bedding (dip 0–30, 31–60, 61–90)
- fold of bedding (anticline, syncline)
- fold of tectonic layering (recumbent, upright antiform, upright synform)
- thrust fault
- normal fault
- shelf edge

isograds
- Kfeldspar
- sillimanite
- andalusite/kyanite
- biotite

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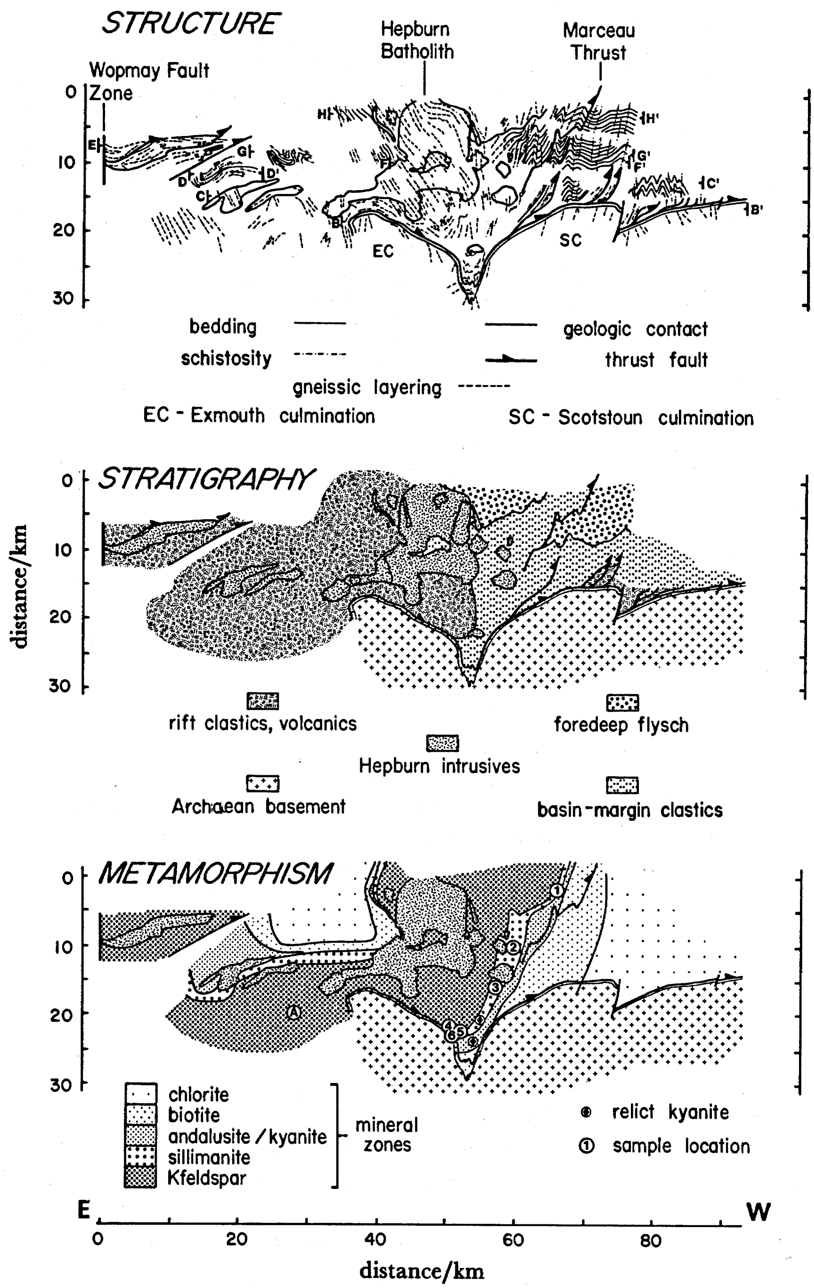

FIGURE 5. Composite down-plunge cross section of the metamorphic-internal zone of Wopmay Orogen presented in three panels for reasons of clarity. The letters B–B′ to H–H′ mark the section lines (also shown in figures 3 and 4) for the independent cross sections that were compiled to form the composite section. The method of construction is described in King (1986). The right side of the diagram is continuous with the foreland thrust-fold belt (see text).

of the Epworth Group are continuous with those of the open-shelf facies exposed within the Asiak Thrust-Fold Belt to the east (Hoffman *et al.* 1983, 1984).

(iv) *Recluse Group: foredeep flysch facies.* Hemipelagic shales and turbidite greywackes of the Recluse Group (Hoffman 1980; Hoffman & Bowring 1984) overlie conformably both the slope-rise facies and shelf-facies deposits of the Epworth Group. The turbidites were derived

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from a metamorphic–plutonic provenance to the northwest (Jeletsky 1974; Hoffman 1980). The Recluse Group is interpreted to be a syn-orogenic foredeep flysch facies which marks the onset of tectonic loading and the progressive eastward encroachment of Calderian orogenic activity (Hoffman 1980; Hoffman & Bowring 1984).

(*b*) *Calderian structures*

The Coronation Supergroup was deformed at *ca.* 1885 Ma during a major episode of regional subhorizontal shortening. This event, named the Calderian Orogeny (Hoffman & Bowring 1984) resulted in the main east–west shortening and east-directed transport and thickening of the Coronation Supergroup over the Slave craton, the collapse and infilling of the foredeep (Hoffman 1980), the emplacement of syn-tectonic intrusions (Hepburn Batholith) and the (late) folding of autochthonous basement (King 1985). In the metamorphic–internal zone, the shortening and transport of the Coronation Supergroup was accommodated by thrusting, folding and penetrative shortening of strata above a basal décollement. In contrast the autochthonous basement and the thin Proterozoic sequence below the décollement incurred relatively little, relatively inhomogeneous deformation during the main Calderian shortening (King 1986). The Calderian Orogeny is interpreted to have resulted from closing of the Coronation marginal basin (Hildebrand *et al.* 1986*a*).

(i) *Autochthon.* The thin mantle of Proterozoic units below the basal décollement are demonstrably 'glued' to basement, and are therefore autochthonous (King 1986). The mantle of Proterozoic units has been, however, affected by inhomogeneous, layer-parallel shortening and discontinuous shear zones that reflect very local, top-eastward displacement parallel to the décollement. King (1986) has suggested that this deformation may be the result of an imposed stress field related to movement on the décollement. The intensity of this fabric increases west of the Scotstoun culmination (figure 4) resulting in an inhomogeneously developed basement-contact-parallel schistosity. The inhomogeneous character of the strain in the autochthon between Scotstoun and Exmouth culminations is highlighted by the presence of a virtually unstrained stromatolitic marble (figure 6, plate 1). It is exceptional that this low-strain domain is preserved beneath more than 30 km of overlying allochthonous Proterozoic units.

(ii) *Calderian basal décollement.* Several hundred metres above the basal Proterozoic unconformity, and above the relatively low-strain autochthon, units of the Coronation Supergroup are affected by thrusts and upright and recumbent folds. The change in structural style occurs across a narrow zone (or surface?), which is not exposed but which is commonly marked by a narrow valley filled with glacial drift (figure 7, plate 1). This narrow zone is interpreted to correspond with the position of the basal décollement of the overlying allochthonous units (King 1986). As defined, the décollement can be traced continuously from the frontal thrust westward beneath the Asiak Thrust-Fold Belt and the metamorphic–internal zone, to the western limb of the Exmouth culmination (figures 4 and 5) where it dips downward to unexposed structural levels.

(iii) *Calderian allochthon.* Above the basal décollement the structural front of the internal zone is a transitional domain, which represents a change from foreland thrust-fold deformation (imbricate thrusts, upright chevron-style folds) in the east to complex polyphase recumbent folds of bedding, schistosity and gneissic layering in the west (King 1986). The transition in structural style is spatially coincident with: (1) the stratigraphic transition from the semipelites, quartzites and carbonate platform of the continental shelf through the slope-rise facies pelites to the initial-rift facies units and (2) a suite of mineral isograds (figures 3–5) which indicate increasing metamorphic grade westward.

At intermediate and high structural levels in the eastern internal zone (G–G′, H–H′, figure 5), the structural style is characterized by upright, open folds whose axial surfaces dip vertically or steeply to the west, and trend north–south. Regional stratigraphic relationships, stratigraphic top directions and bedding-cleavage relationships indicate that the folds are upward facing. Alignment of chlorite, muscovite and/or biotite (depending on metamorphic grade) defines a cleavage that is axial planar to these folds.

Marceau Thrust (Hoffman 1973; Hoffman *et al.* 1978; Hoffman 1980; St-Onge 1981) is a major east-verging thrust fault in the eastern metamorphic-internal zone (figures 2 and 3), and is traceable at high and intermediate structural levels where it places Epworth Group on Recluse Group (F–F′, G–G′, H–H′, figure 5). At intermediate structural levels, Marceau Thrust is folded coaxially with the foreland thrust-ramp folds (St-Onge 1981). The regional cleavage is axial planar to the folded thrust. Thrusting, progressive folding, and a phase of metamorphic mineral growth (see §5*b*) were therefore essentially coeval at these structural levels.

In contrast, a markedly different structural style is developed at lower structural levels within the Calderian allochthon (B–B′, figure 5). Early isoclinal folds of bedding and associated axial–planar schistosities are refolded about nearly coaxial mesoscopic to macroscopic, recumbent isoclinal folds (figure 7, plate 1) recording polyphase deformation. The successive phases of folding are not clearly separable in time and are considered part of a progressive deformation (King 1986).

West of the Exmouth culmination, at low structural levels, high-grade rocks of the Akaitcho Group (figure 4) are characterized by a gneissic layering which contains transposed axial surfaces of earlier-formed isoclinal folds. Near the Exmouth culmination the transposed gneissic layering is parallel to the basal décollement (figure 5). To the north, at intermediate structural levels down-grade from the Kfeldspar–sillimanite isograd (figures 4 and 5), all observed asymmetric recumbent folds of bedding indicate eastward vergence (figure 8, plate 1). Bedding and cleavage are commonly parallel, but where isoclinal fold closures are recognized, the cleavage is axial planar.

(iv) *Basement-cored culminations.* The three basement-cored structural culminations and their thin mantle of autochthonous basal Proterozoic units occur at the lowest exposed structural levels of central Wopmay Orogen. The north–south axes of the culminations, and of associated

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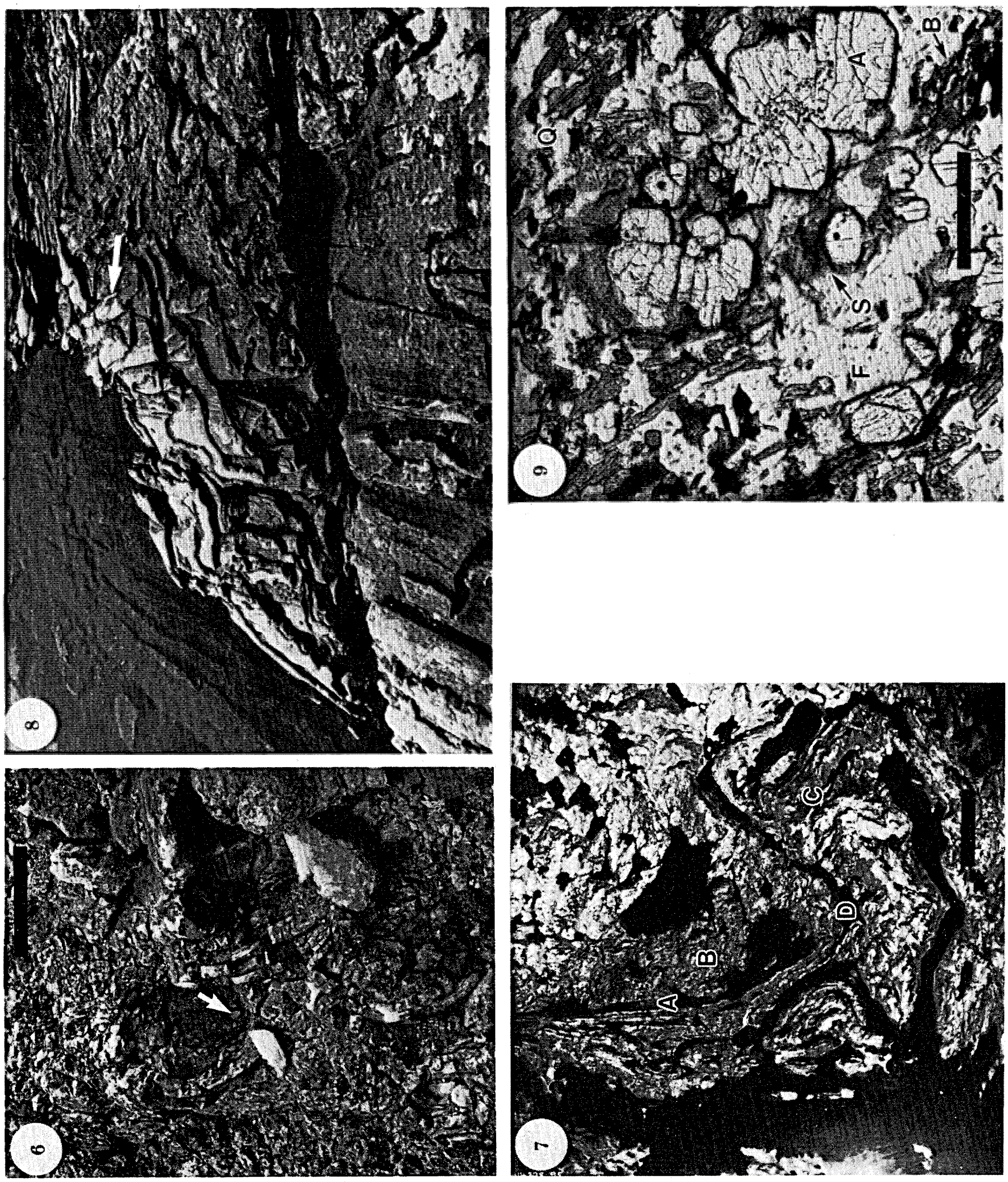
Description of plate 1

Figure 6. Stromatolitic marble from the autochthonous Proterozoic sequence in the southern hinge zone between Scotstoun and Exmouth culminations. Note that the stromatolites are virtually unstrained although they occur structurally *beneath* more than 30 km of allochthonous rocks. The metamorphic assemblage is forsterite (arrow)–dolomite–quartz. Horizontal bar represents 15 cm.

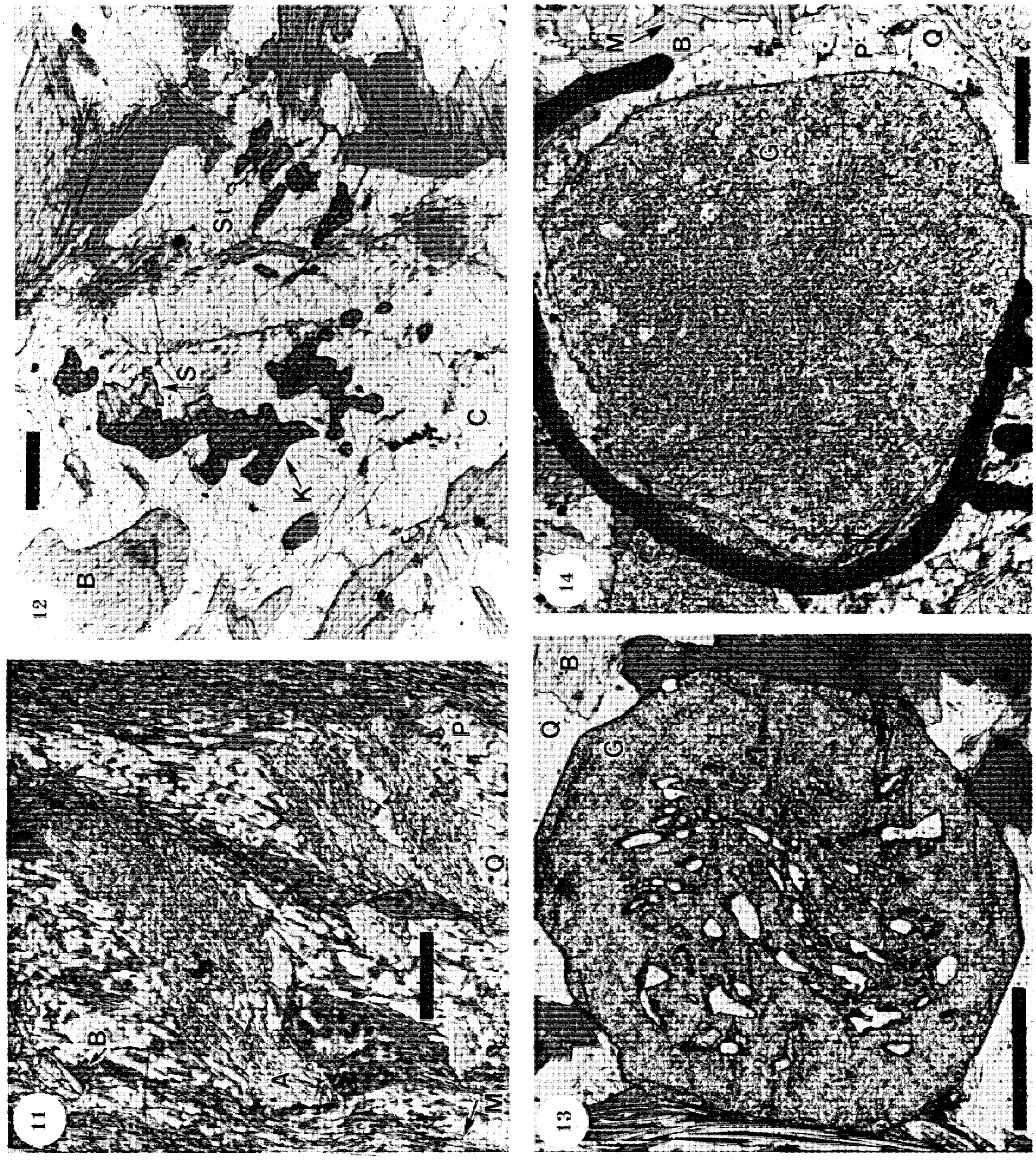
Figure 7. Airphotograph of the hinge zone of the southern part of the Exmouth culmination. Abbreviations: A, autochthonous Proterozoic cover; B, Archaean basement; C, Calderian allochthon; D, trace of basal décollement. Horizontal bar represents 2 km.

Figure 8. East-verging recumbent folds of bedding near Calypso Lake (figure 4). Photograph taken facing south. Arrow points at 15 cm pen.

Figure 9. Photomicrograph of relict andalusite in Kfeldspar–sillimanite gneiss. Abbreviations: A, andalusite; B, biotite; F, Kfeldspar; Q, quartz; S, sillimanite. Scale bar represents 0.25 mm.



Figures 6–9. For legend see opposite.



FIGURES 11–14. For legend see opposite.

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second-order folds, are coaxial with Calderian structures. The map pattern of the mineral isograds (see §5*c*) show that at least the north–south trending component of the culminations formed during the syn-Calderian thermal culmination. The east–west axis of the series of culminations is parallel to a set of regional cross folds (see §4*d*) and coincides with the crest of a first-order cross antiform. The regional structure therefore corresponds to a dome-and-basin (type 3 of Ramsay (1967)) interference pattern (King 1985).

(*c*) *Hepburn Batholith*

Plutons of the Hepburn Batholith discordantly intrude all *allochthonous* units of the Coronation Supergroup within the metamorphic-internal zone (figures 3–5). Hepburn plutons have not been found beneath the basal décollement in either the autochthonous Proterozoic sequence or in the Archaean basement. Older plutons of the batholith that contain a well developed foliation continuous with that of the country rocks, are syn-tectonic. Smaller, less strongly foliated intrusions cut both the older plutons and the host rock. Because of the syn-tectonic character of the plutons, and because they are not found below the décollement, the Hepburn Batholith is interpreted to have been transported, together with its host rocks, to its present position. The batholith consists mostly of peraluminous biotite–granite and tonalite but also includes significant amounts of gabbro and diorite (Hoffman *et al.* 1980). This range in composition suggests that there was both a crustal and a mantle contribution to the generation of the batholith (Lalonde 1984). U–Pb (zircon) ages for the Hepburn plutons range between 1896 and 1878 Ma (Bowring & Van Schmus 1986). R. S. Hildebrand (personal communication) has pointed out that the Hepburn Batholith is geochronologically and geochemically continuous with two phases of adjacent continental arc magmatism (based on Bowring & Van Schmus (1986) and unpublished data from A. E. Lalonde). This has led to the suggestion that the Hepburn Batholith is a product of continuing magmatism in the closing back-arc basin (Hildebrand *et al.* 1986*a*).

Because the plutons intrude the syn-orogenic Recluse Group (figures 3 and 5) but are themselves deformed and allochthonous, the ages for the plutons date the Calderian Orogeny (Hoffman & Bowring 1984).

(*d*) *Post-Calderian structures*

Two distinct episodes of regional, subhorizontal shortening affected all of the units and Calderian structures of the metamorphic–internal zone. The first episode of post-Calderian deformation is expressed as the northeast-trending cross folds, which provide the informative,

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Description of plate 2

Figure 11. Syn-tectonic andalusite with sigmoidal inclusion trails continuous with exterior (Calderian axial planar) schistosity from high structural levels (H–H′, figure 5). Scale bar represents 0.7 mm. Abbreviations: A, andalusite; B, biotite; M, muscovite; P, plagioclase; Q, quartz.

Figure 12. Photomicrograph of relict high-pressure mineral assemblage, mantled by late cordierite, in autochthonous Proterozoic pelites. Scale bar represents 0.3 mm. Abbreviations: B, biotite; C, cordierite; K, kyanite; S, sillimanite; St, staurolite.

Figure 13. Photomicrograph of garnet from Calderian allochthon showing a syn-tectonic core, with sigmoidal inclusion trails, mantled by a post-tectonic massive rim. Scale bar represents 0.25 mm. Abbreviations: B, biotite; G, garnet; Q, quartz.

Figure 14. Photomicrograph of post-tectonic poikiloblastic garnet in metamorphic sample 3 from the Calderian allochthon (sample location shown in figures 4 and 5). Scale bar represents 1 mm. Abbreviations: B, biotite; G, garnet; M, muscovite; P, plagioclase; Q, quartz.

oblique sections of Calderian structures. These cross folds are correlative in time and geometry with folds in the Tree River Fold Belt (Hoffman *et al.* 1984) of the northeast Calderian foreland (figure 1). First-order cross folds have a wavelength of 75–150 km and amplitudes of over 15 km (St-Onge 1984*a*; King 1986). Mesoscopic cross folds are only rarely observed in the internal zone. Autochthonous basement is clearly affected by the cross folding, but is nowhere thrusted (Hoffman *et al.* 1984).

The second post-Calderian deformation produced a regional system of conjugate transcurrent faults (figures 3 and 4). Bulk strain associated with the faulting is consistent with about 20% east–west shortening approximating horizontal plane strain (Tirrul 1984*a*).

5. Evolution and tectonic setting of metamorphism

A major thermal–metamorphic culmination affects all units of the Coronation Supergroup in the metamorphic–internal zone of Wopmay Orogen (figures 3–5). The thermal culmination corresponds to an increase in metamorphic grade, as documented in pelitic units, from slates with the maximum-phase assemblage muscovite–chlorite–plagioclase–quartz–graphite to gneisses with the maximum-phase assemblage garnet–sillimanite–cordierite–biotite–Kfeldspar–plagioclase–quartz–graphite–granitic melt. The last phase is inferred from the granitic pods present in metasediments as discussed in St-Onge (1984*a*). The final form of the thermal culmination is outlined at the present erosion surface by suites of mineral isograds mapped in pelitic and semipelitic units (St-Onge & Carmichael 1979; St-Onge 1981, 1983, 1984*a*,*b*; St-Onge *et al.* 1984). The mineral isograds transect, and therefore postdate, Calderian thrusts and folds of bedding. The isograds are, however, affected by late Calderian upright folds of the tectonic fabric (St-Onge 1981).

The final form of the regional thermal culmination, as defined by the regional mineral isograds, is elongate parallel to the structural grain of the orogen in map view (figures 3 and 4). In cross section, the metamorphic culmination has an asymmetric lobate form (figure 5). This form is spatially related to plutons of the Hepburn Batholith at intermediate to high structural levels (G–G′, H–H′, figure 5). At low structural levels, beneath the main cluster of Hepburn Batholith plutons, the mineral isograds are inverted (figure 5). This inverted configuration is mirrored in both the north-plunging and south-plunging hinge zone between the Scotstoun and Exmouth culminations (figure 4). The inverted isograds transect the basal décollement and the eastern autochthonous Proterozoic rocks (figure 5). At low structural levels in the western part of the internal zone, a large part of the thermal culmination has no obvious relation to plutonic rocks (figures 4 and 5). This domain is composed entirely of high-grade Akaitcho Group gneisses.

The control on these spatial relations within the regional thermal-metamorphic culmination is provided by the exceptional, exposed structural relief. This relief also enables the study of metamorphic mineral assemblages, mineral chemistry and metamorphic–structural relation over 30 km of structural depth. The synthesis of this metamorphic data with the stratigraphic, structural and geochronological data outlined in §3, allows the evolution of the thermal culmination through a variety of thermo-tectonic régimes to be documented. The metamorphic data and the thermo-tectonic régimes of metamorphism are discussed in order of development in the next three subsections.

(a) *Epicontinental rifting: onset of regional metamorphism*

Three features of the regional metamorphism are unique to the Akaitcho Group and are indicative of a distinct early stage of metamorphism in the internal zone. The most striking feature, both in map view (figures 3 and 4) and in composite down-plunge section (figure 5), is the metamorphism of the greater part of the Akaitcho Group to grades beyond the stability of muscovite plus quartz given few obvious local heat sources such as plutons (figures 4 and 5). Most of the Hepburn plutons are clustered at higher structural levels in, or adjacent to, metasediments of the Epworth and Recluse groups (figure 5). These relations show that T conditions in excess of 650 °C (Skippen 1977) were achieved regionally, with no significant addition of heat by plutonic advection. A large conductive thermal gradient (with or without volatile-phase convection) is therefore required for the regional metamorphism of the Akaitcho Group.

A second feature is the spacing of the mineral isograds at intermediate structural levels in the Akaitcho Group west of the Hepburn Batholith (figure 5). The metamorphic transition from biotite-grade slates to sillimanite–Kfeldspar migmatites occurs over less than 2 km of structural depth (figure 5). Although these relations were undoubtedly modified during later Calderian shortening and uplift (see §§5*b*, *c*), the close spacing of the Akaitcho Group isograds imply a high thermal gradient on the order of 100 K km^{-1}.

A third feature is the occurrence of the metastable assemblage sillimanite–andalusite–biotite–Kfeldspar–plagioclase–quartz–granitic melt in the pelitic gneisses of the Akaitcho Group which are located in the high-grade, pluton deficient domain west of the Exmouth culmination (Bent Lake area, figures 4 and 5). Although rare, the known occurrences of this assemblage are widely distributed. The andalusite in the assemblage is massive, corroded and mantled by sillimanite in contact with Kfeldspar (figure 9, plate 1) and is interpreted to be a relict phase. The occurrence of relict andalusite (requiring a $P \leqslant 3.76$ kbar†) in the high-grade gneisses is considered to reflect the high-T–low-P trajectory followed by this part of the Akaitcho Group during metamorphism. The inferred P–T trajectory for the andalusite-bearing gneisses is shown on figure 10. This P–T path is an expression of abnormally high temperatures at shallow crustal depths and hence of an exceptionally steep thermal gradient.

Direct structural–metamorphic relations constraining the timing of the high-T–low-P metamorphism in the Akaitcho Group were not observed. However, the spatial restriction of the andalusite-bearing gneisses to the Akaitcho Group and the occurrence of the prograde, low-P assemblage at low structural levels in the Calderian allochthon (figure 5) indicate that this phase of mineral growth occurred before major Calderian overthrusting and tectonic thickening.

The high-T, regional metamorphism of rift-fill units, the lack of plutonic heat sources, the development of a steep thermal gradient and the evidence of metamorphic mineral growth before major tectonic thickening are key observations, which suggest that regional, high-T metamorphism in Wopmay Orogen was initiated during epicontinental rifting. The thermal models presented in Thompson (1981) show that a rifting environment with a related temperature increase, caused by lithospheric thinning, at the base of the crust is a plausible

† 1 bar = 10^5 Pa.

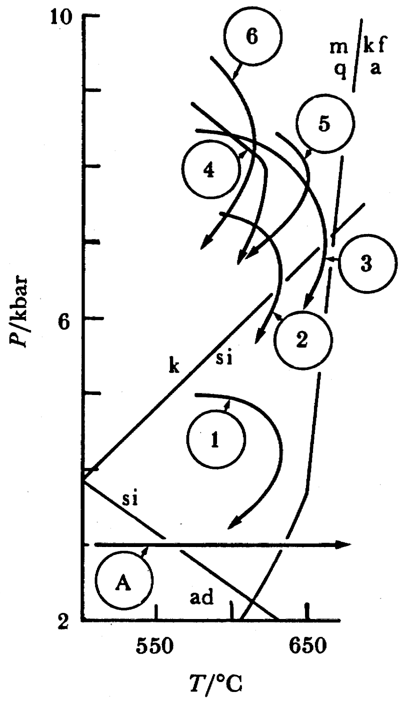

FIGURE 10. Pressure–temperature diagram with P–T trajectories calculated from core-to-rim analyses of zoned poikiloblastic garnets (1)–(6) (see text) and inferred for andalusite-bearing gneisses of the Akaitcho Group (A). References for positions of univariant curves is given in St-Onge (1984*b*). Abbreviations: a, aluminosilicate; ad, andalusite; k, kyanite; kf, Kfeldspar; m, muscovite; q, quartz; si, sillimanite.

setting for regional metamorphism. Wickham and Oxburgh (1985) have proposed a similar tectonic setting to account for the Hercynian high-T–low-P regional metamorphism in the Pyrenees.

(*b*) *Syn-tectonic magmatic arc: addition of heat by advection*

Within 5–10 Ma of Akaitcho rifting and coeval onset of regional metamorphism, the rift-fill deposits and the sedimentary prism underwent east–west, subhorizontal shortening and vertical thickening during the Calderian Orogeny. Closure of the volcano-sedimentary basin was accompanied by deposition of the axial-foredeep flysch facies of the Recluse Group, which was also involved in the Calderian deformation. The emplacement of the Hepburn Batholith into the Coronation Supergroup marginal prism occurred synchronously with regional shortening. Where the Hepburn plutons are clustered, at high and intermediate structural levels within the Calderian allochthon (F–F′, G–G′, H–H′, figure 5), a distinct thermal culmination developed. This 'local' thermal perturbation is characterized by mineral growth in metamorphic zonal sequences representative of low to intermediate pressures. The sequences contain, at medium-grade, andalusite–sillimanite and staurolite–sillimanite assemblages and, at higher grades, low-pressure to intermediate-pressure anatectic–granite–pod mineral associations (St-Onge 1981, 1984*b*). Most, but not all (see §5*c*) porphyroblasts (staurolite, garnet, andalusite, sillimanite) show evidence of syn-tectonic growth (figure 11, plate 2).

The spatial association of the low-pressure to medium-pressure metamorphic zonal sequences with the syn-tectonic plutons of the Hepburn Batholith and the syn-tectonic growth of the

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metamorphic minerals at high and intermediate structural levels within the Calderian allochthon document a stage in the evolution of the regional metamorphic culmination that is related to magma emplacement. The distinct episode of mineral growth in a syn-tectonic, magmatic arc setting is separated from the onset of metamorphism during Akaitcho rifting by accumulation of the Epworth and Recluse groups. However, because of the very short time interval (5–10 Ma) between rifting and emplacement of the Hepburn Batholith, the heat being added to the deforming Calderian allochthon by magmatic advection was probably introduced before any significant cooling of the syn-rifting metamorphic complex had occurred.

(*c*) *Syn-metamorphic uplift and erosion*

At low structural levels, the metamorphic-internal zone is characterized by a suite of *inverted* mineral isograds, mapped both in the Calderian allochthon and in the underlying autochthonous Proterozoic sequence (figure 5). As the upward-facing stratigraphy of the autochthon precludes structural overturning of the metamorphic zonal sequences, the occurrence of 'hot-side-up' isograds requires that a negative thermal gradient was established during the Calderian Orogeny. The negative thermal gradient at the base of the Calderian allochthon, together with the absence of Proterozoic plutons both in the autochthonous cover and in the underlying Archaean basement (figure 5), indicate that the Calderian allochthon was transported while still hot, after the emplacement of the Hepburn Batholith, over a relatively cold autochthon.

Downward relaxation of isotherms and continued metamorphic mineral growth following emplacement of the hot allochthon is documented in the underlying autochthon by the inverted metamorphic zonal sequence (kyanite–sillimanite) (figure 12, plate 2). Within the Calderian allochthon, post-emplacement, static mineral growth is illustrated by massive rims on syn-tectonic mineral cores (figure 13, plate 2) and by metamorphic mineral phases that overgrow Calderian fabrics at all exposed structural levels. Following emplacement of the Calderian allochthon, the autochthonous basement became involved in the ongoing Calderian shortening (King 1985). This involvement is expressed as the high-amplitude Carousel, Scotstoun and Exmouth culminations. Because the mineral isograds both cross-cut and are deformed by the basement-cored folds (B–B′, figure 4), late metamorphic mineral growth is considered to overlap in time with basement deformation.

The occurrence of large, post-tectonic, poikiloblastic garnets at all exposed structural levels in the metamorphic internal zone has allowed direct petrological derivation of metamorphic *P–T* paths (St-Onge 1986) related to the late mineral growth. The derivation is based on microprobe analyses, from core to rim, of zoned, poikiloblastic garnets which contain abundant inclusions of biotite, plagioclase, quartz and $\pm Al_2SiO_5$ (figure 14, plate 2). Independent, multipoint *P–T* determinations can be derived from within single, poikiloblastic garnets by applying appropriate calibrated geothermometers and geobarometers (Ferry & Spear 1978; Newton & Haselton 1981; Hodges & Spear 1982; Hodges & Royden 1984) to subassemblages composed of entrapped inclusions and adjacent host garnet (St-Onge 1986). The technique for direct petrological derivation of metamorphic *P–T* paths and the analytical data are given in St-Onge (1986). Minerals from six samples of pelitic schists that contain post-tectonic poikiloblastic garnets were analysed by electron microprobe. The samples are from within the muscovite–sillimanite zone on the eastern and southern sides of the Hepburn Batholith (figures 3 and 4). In down-plunge projection the locations of the samples span the full range of structural relief (figure 5) now exposed in oblique section in the metamorphic-internal zone.

Five of the samples, numbered from 1 (highest) to 5 (lowest structural level) (figure 5), are from pelitic schists situated structurally above the basal décollement. Sample 6 is from the underlying, autochthonous Proterozoic sequence (figure 5).

The individual *P–T* paths shown on figure 10 are all of similar form and each is compatible with England & Richardson's (1977) model of a tectonically thickened crust undergoing uplift and erosion. The calculated *P–T* paths show that post-tectonic poikiloblastic garnet growth occurred during decompression and therefore unroofing of the hinterland. Garnet growth was originally concomitant with continued prograde metamorphism and, later, with the cooling of the metamorphic culmination. The similar forms of the derived *P–T* paths for samples from the autochthon (sample 6) (whose *P–T* path cannot be tectonically transported) and from the allochthon (samples 1–5), indicate that the samples from the allochthon followed their decompression *P–T* trajectories after the emplacement of the Calderian allochthon, essentially in the tectonic framework and structural position shown on figure 5. This is corroborated by the excellent agreement between the structural level of the six samples indicated by the down-plunge projection (figure 5) and the maximum pressure documented by paths 1–3 and inferred by paths 4–6.

In figure 15, the calculated *P–T* paths are plotted on a strike-parallel cross section of the internal zone. This length section illustrates the large cross folds of both basement and cover in the metamorphic zone. From this figure, it is clear that initially higher-*P* samples (e.g. sample 6) are associated with cross antiforms and that initially lower-*P* samples (e.g. sample 1) correspond to cross synforms. Because of this correlation and because the various *P–T* paths are not 'nested', contrary to the pattern predicted for the uplift of different structural levels with a common uplift rate (see, for example, England & Thompson 1984), the syn-metamorphic uplift is interpreted to have occurred at different rates in different places. The variation in rate of uplift was probably largely controlled by the cross folding. Uplift was completed by 1870 Ma, the age (U–Pb zircon) of a portion of the Great Bear Magmatic Zone that unconformably

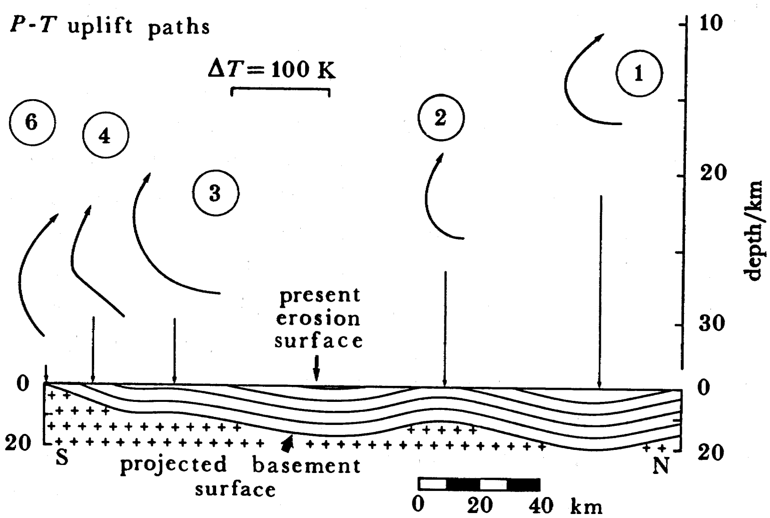

FIGURE 15. Length-section of internal zone drawn for north–south transect west of section shown in figure 2. Section illustrates late cross folds of both allochthon (ruled pattern) and Archaean basement (crosses). Positions of individual metamorphic samples and derived *P–T* trajectories are shown by arrows.

overlaps the western boundary of the internal zone (Bowring & Van Schmus 1986). The range in uplift rates is from 1.5 to 2.7 mm a^{-1} (St-Onge 1986).

The calculated *P–T* paths relate the late, prograde metamorphic mineral growth documented in both the Calderian allochthon and underlying autochthon to a period of uplift and erosion which followed subhorizontal shortening and deformation related to the Calderian Orogeny. Final mineral equilibration (St-Onge 1984*a*, 1986) and the final configuration of the mineral isograds (figure 5) were achieved in this late thermo-tectonic régime and mark the last stage in the evolution of the regional metamorphic culmination of the early Proterozoic Wopmay Orogen. Metamorphic mineral growth during uplift and erosion in Phanerozoic overthrust belts has been documented or modelled by several authors (see, for example, Richardson & England 1979; Oxburgh & England 1980; Hodges & Royden 1984; Selverstone *et al.* 1984). These studies have yielded similar *P–T* trajectories to those obtained for the garnet-bearing samples of the Hepburn Metamorphic–Internal Zone.

6. Discussion

In Wopmay Orogen, the last stage documented in the evolution of regional metamorphism consists of mineral growth during uplift and erosion. This stage is documented by the derivation of precise syn-metamorphic uplift *P–T* trajectories and evidence, at all exposed structural levels, of post-tectonic mineral growth. In the Alps a similar thermo-tectonic régime is used to explain the main Tertiary (Alpine) metamorphism (see, for example, Oxburgh & Turcotte 1974; Bickle *et al.* 1975; England 1978; Oxburgh & England 1980). A fundamental difference, however, between Wopmay Orogen and the Alps is the time interval between the tectonic thickening of the crust and the onset of rapid uplift and erosion. For the Eastern Alps, the time lag appears to have lasted over 20 Ma (Oxburgh & Turcotte 1974; Bickle *et al.* 1975; England 1978; Oxburgh & England 1980). In contrast, the *total* time interval between tectonic thickening and the end of uplift of the metamorphic–internal zone of Wopmay Orogen is at most 11 Ma. This brief time interval strongly suggests that very little or no 'heating-without-erosion' took place in Wopmay Orogen.

England & Thompson (1984) have shown that the time interval covering the period of no erosion is an important parameter in the thermal evolution of thickened crust since conductive heating is greatest immediately after overthrusting. A short or non-existent time interval of no erosion results in significantly lower maximum metamorphic temperatures being attained. A short total erosion timespan also results in less total heating of the buried rocks. If no other heat source besides conductive heating of a thickened crust was present during the total 11 Ma timespan of erosion, the maximum temperatures reached in the metamorphic–internal zone should, according to the numerical models of England & Thompson (1984), be less than 540 °C. This is clearly inconsistent with field and petrologic data from Wopmay Orogen, which show that a great part of the internal zone has reached temperatures within the stability field of Kfeldspar plus sillimanite (over 650 °C) (Skippen 1977) (figures 3–5). It is therefore concluded that a large part of the development of regional metamorphism is related to addition of heat to the early Proterozoic crust during two, thermally overlapping tectonic régimes that preceded the period of uplift, erosion and final equilibration. These include heating during crustal stretching at *ca.* 1900 Ma and advective heating during Calderian shortening and thickening between 1896 and 1878 Ma.

Such an incremental, thermally additive development of metamorphism in Wopmay Orogen is undoubtedly due to the rapid evolution (cf. Hoffman & Bowring 1984) of the orogen. It is probable that evidence from field and laboratory studies in other rapidly evolved orogenic belts will similarly highlight the contributions of early thermal perturbations to the form and *P–T–t* relations of final, uplift-related, regional metamorphism.

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Discussion

I. W. D. Dalziel (*Institute for Geophysics, University of Texas at Austin, Texas, U.S.A*). The interpretation by Dr St-Onge and Dr King of the Wopmay Orogen as a deformed and uplifted back-arc basin is in contradiction to earlier interpretations of Dr Paul Hoffman. Is there disagreement among workers in the region with regard to this? What is the basis of the authors' back-arc basin interpretations? Also, while the existence of regional metamorphism pre-dating compressional collapse of a back-arc basin is well documented in the Andes, the authors' interpretation of relatively high *T–P* metamorphism (including the crystallization of andalusite) as occurring before compressional deformation in the Wopmay Orogen was not clear in the talk.

M. R. St-Onge and J. E. King. The interpretation of part of Wopmay Orogen as a deformed and uplifted back-arc basin is not a direct contradiction of earlier interpretations. It simply reflects a new understanding of the tectonic evolution of the orogen given constraints imposed by newly acquired geochronological and petrological data. As referenced in the paper which accompanies the talk (St-Onge & King, this symposium) the basis for the back-arc basin interpretation is presented in Hildebrand & Roots (1985). There is a general consensus among workers in the region on this subject.

The high-*T*–low-*P* metamorphism, which predates the shortening and uplift in the metamorphic-internal zone of Wopmay Orogen, is documented (St-Onge & King, this symposium) by the following.

(1) The high-*T* regional metamorphism, at low structural levels, of mostly rift-fill units, to grades beyond the stability of muscovite plus quartz, in the absence of obvious heat sources such as plutons.

(2) The close spacing of mineral isograds at the higher structural levels in the rift-fill units suggestive of steep thermal gradients.

(3) The occurrence of the assemblage sillimanite–andalusite–biotite–Kfeldspar–plagioclase–quartz–granite melt only in the metamorphosed rift-fill units, an assemblage which requires

a high-*T*–low-*P* trajectory (path A, figure 10, St-Onge & King, this symposium.) This path is an indication of abnormally high temperatures reached at low crustal depths, in the absence of plutonic heat sources.

(4) The occurrence of the low-*P* andalusite-gneisses at low structural levels in the early Proterozoic allochthon (figure 5, St-Onge & King, this symposium), which indicates that the assemblage must have developed before any significant overthrusting and tectonic thickening.

Reference

Hildebrand, R. S. & Roots, C. F. 1985 *Geol. Surv. Pap. Can.* **85-1** A, 373–383.

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

THE ROYAL SOCIETY A

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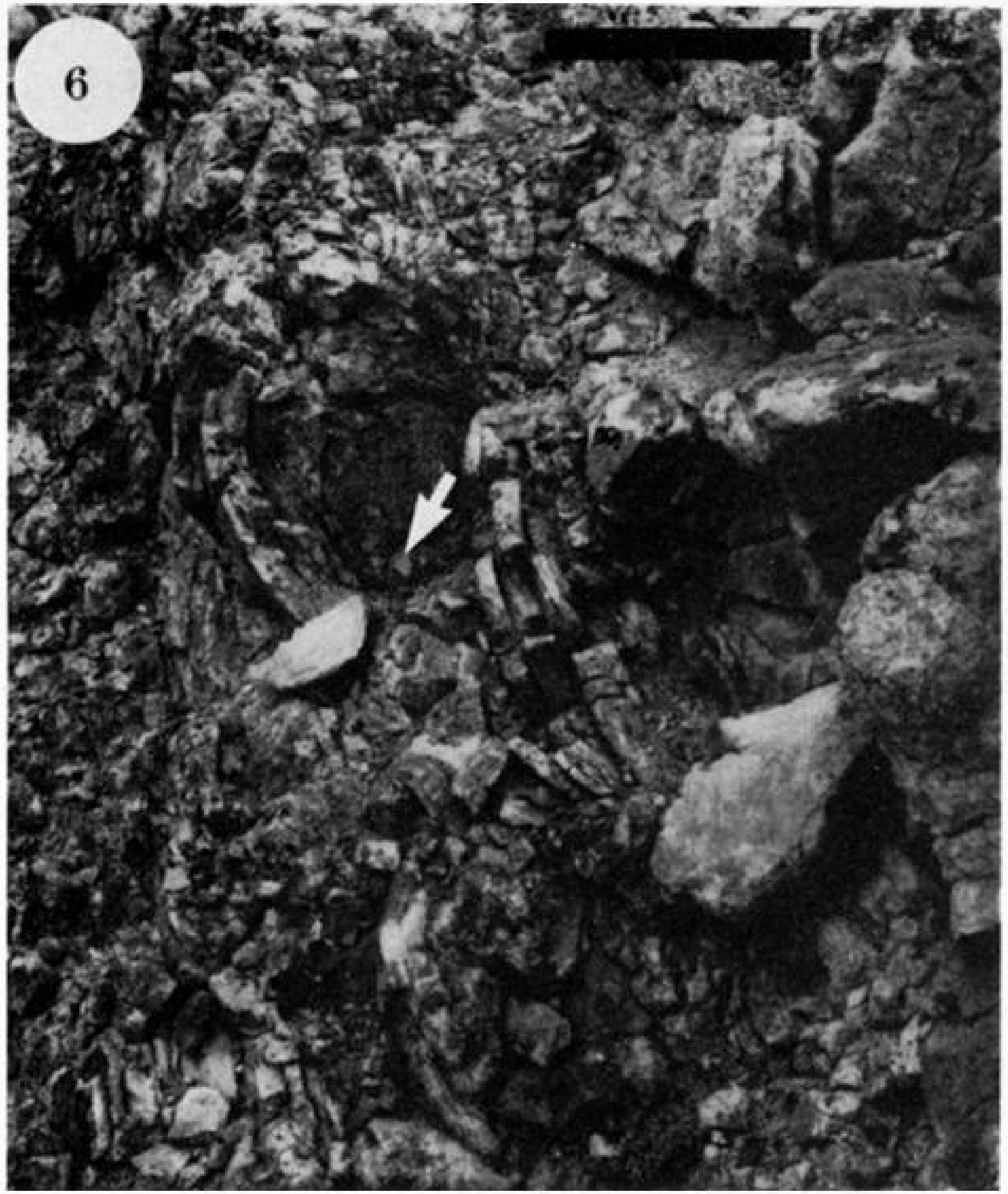

IGURE 6. Stromatolitic marble from the autochthonous Proterozoic sequence in the southern hinge zone between Scotstoun and Exmouth culminations. Note that the stromatolites are virtually unstrained although they occur structurally *beneath* more than 30 km of allochthonous rocks. The metamorphic assemblage is forsterite (arrow)–dolomite–quartz. Horizontal bar represents 15 cm.

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GURE 7. Airphotograph of the hinge zone of the southern part of the Exmouth culmination. Abbreviations: A, autochthonous Proterozoic cover; B, Archaean basement; C, Calderian allochthon; D, trace of basal décollement. Horizontal bar represents 2 km.

GURE 8. East-verging recumbent folds of bedding near Calypso Lake (figure 4). Photograph taken facing south. Arrow points at 15 cm pen.

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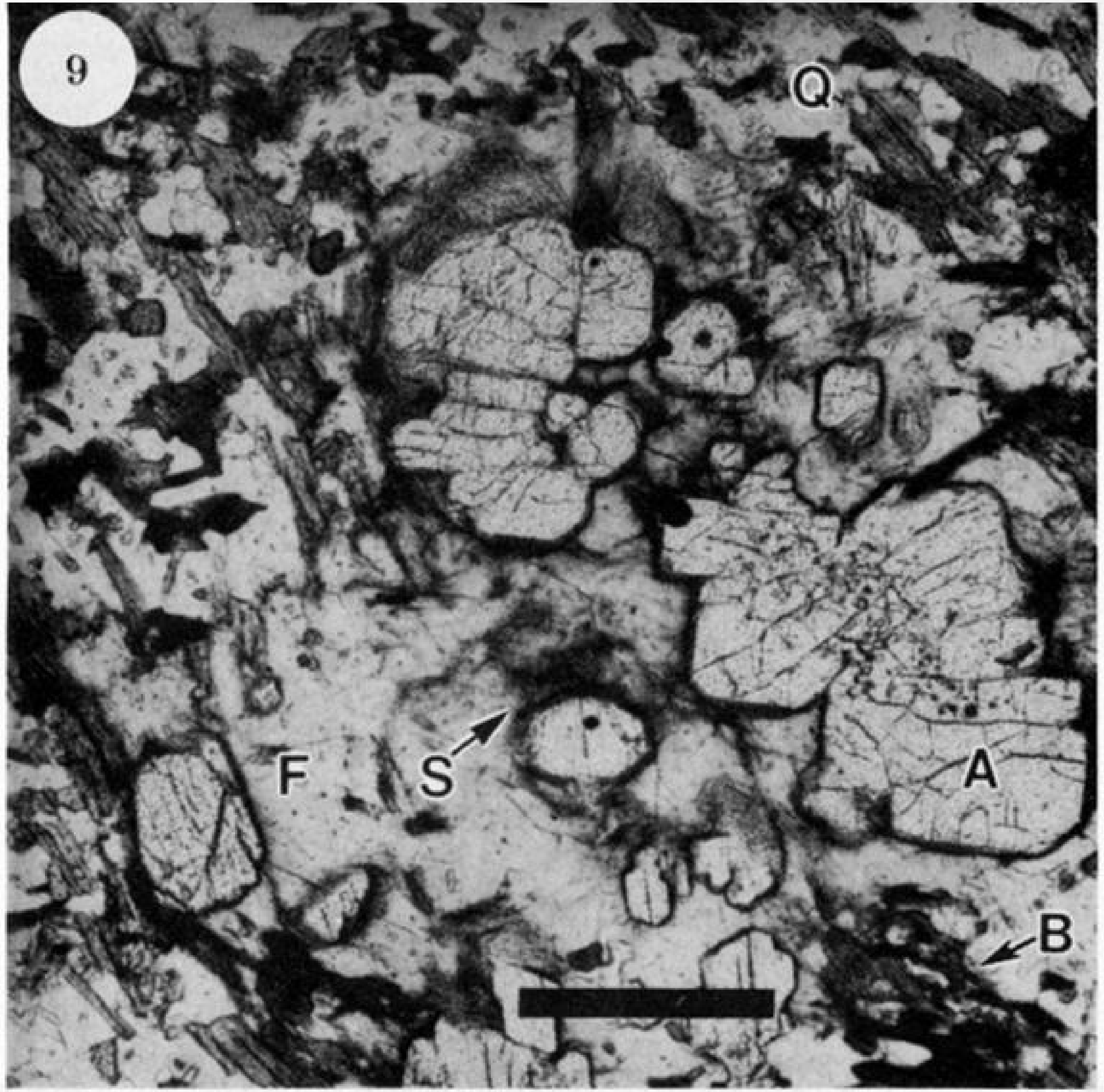
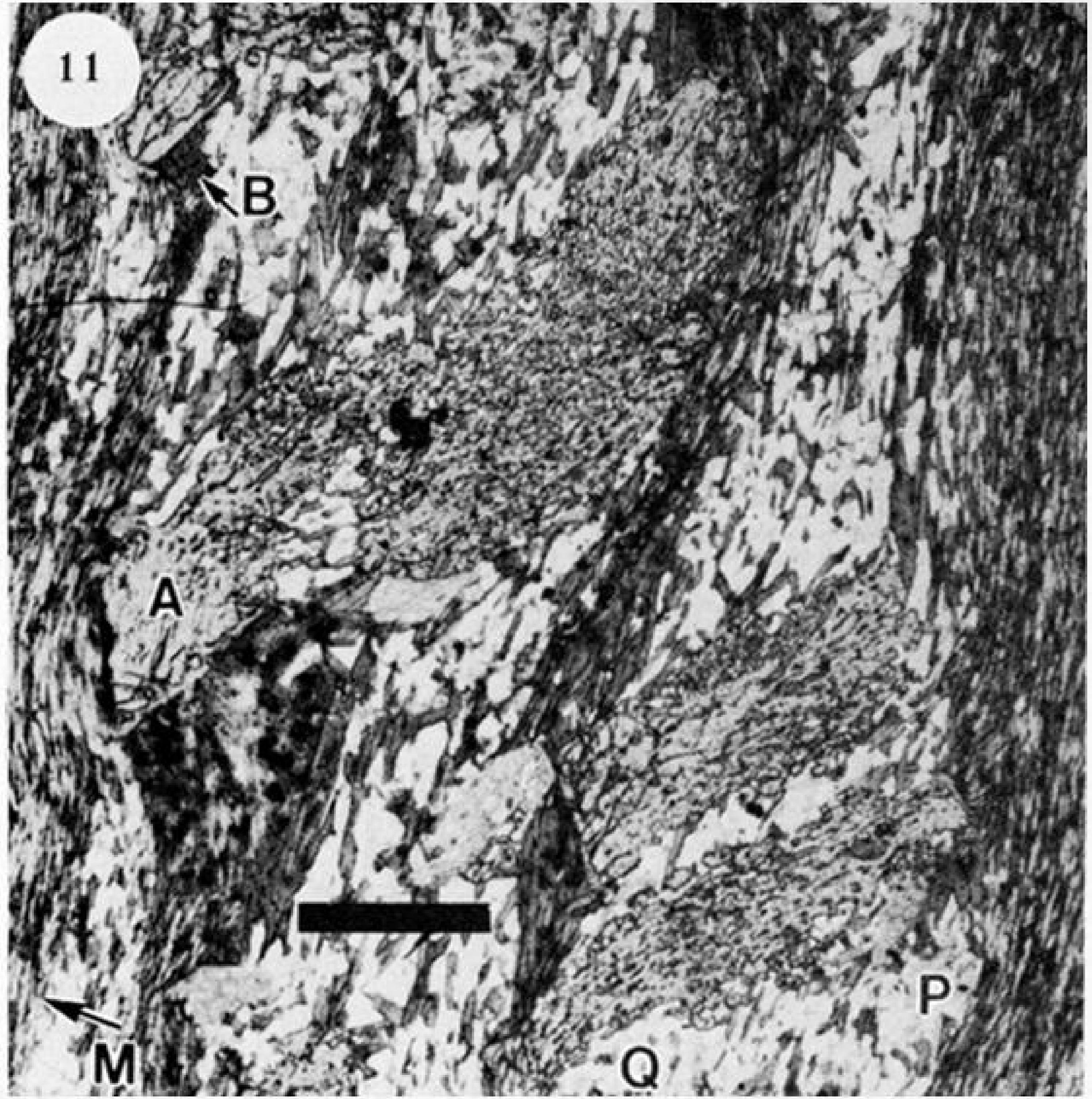


GURE 9. Photomicrograph of relict andalusite in Kfeldspar–sillimanite gneiss. Abbreviations: A, andalusite; B, biotite; F, Kfeldspar; Q, quartz; S, sillimanite. Scale bar represents 0.25 mm.



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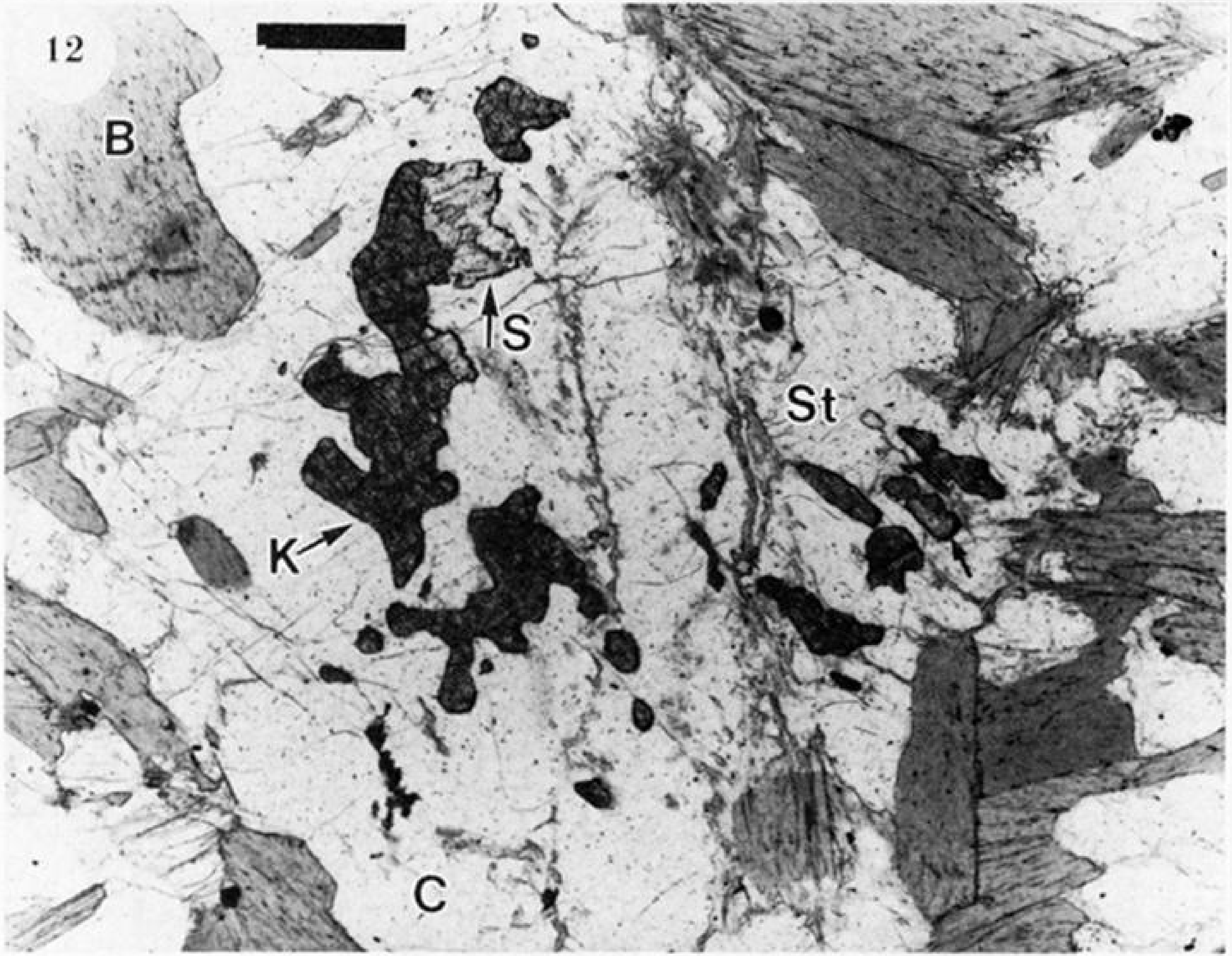
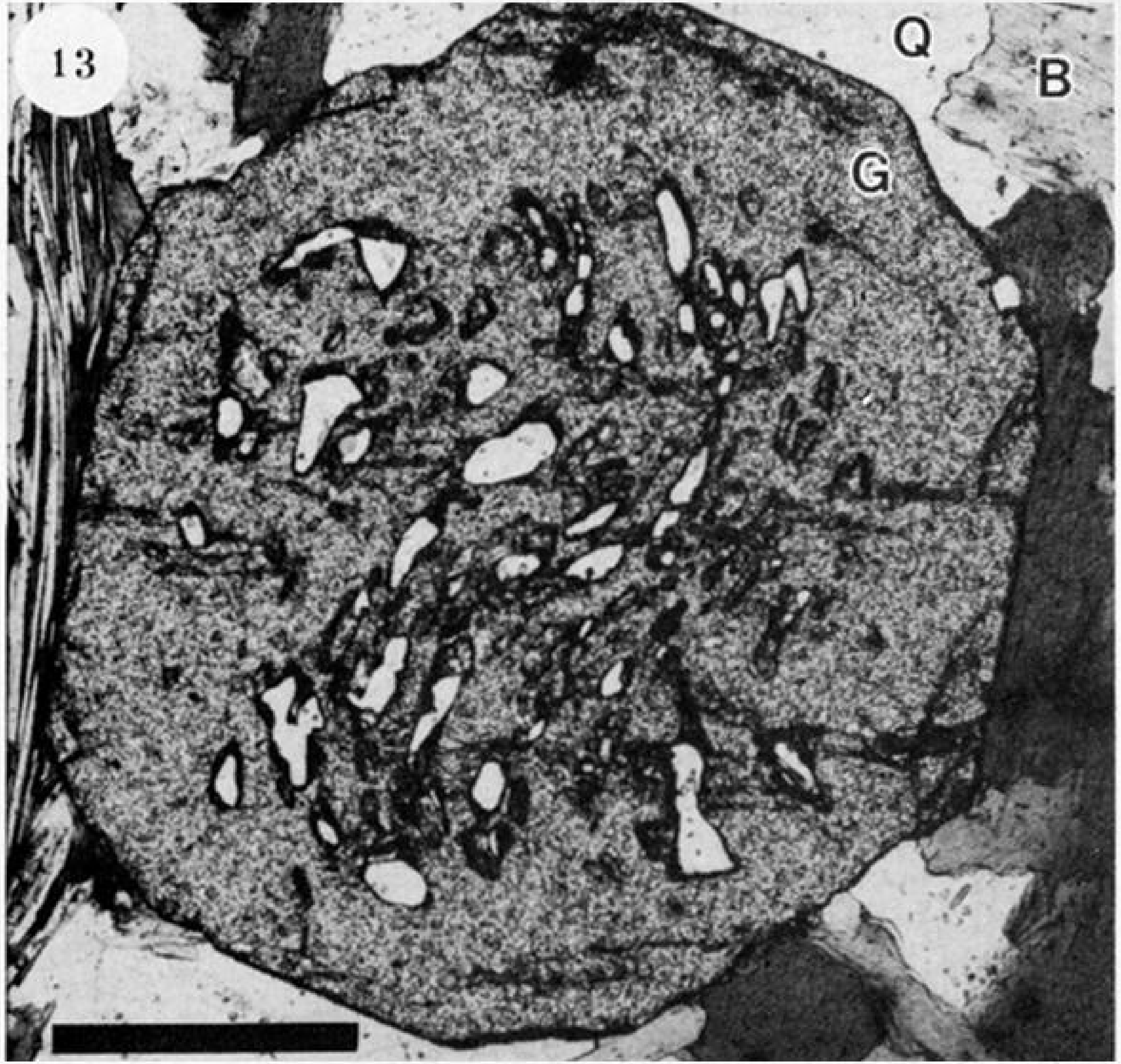
GURE 11. Syn-tectonic andalusite with sigmoidal inclusion trails continuous with exterior (Calderian axial planar) schistosity from high structural levels (H–H′, figure 5). Scale bar represents 0.7 mm. Abbreviations: A, andalusite; B, biotite; M, muscovite; P, plagioclase; Q, quartz.



GURE 12. Photomicrograph of relict high-pressure mineral assemblage, mantled by late cordierite, in autochthonous Proterozoic pelites. Scale bar represents 0.3 mm. Abbreviations: B, biotite; C, cordierite; K, kyanite; S, sillimanite; St, staurolite.



IGURE 13. Photomicrograph of garnet from Calderian allochthon showing a syn-tectonic core, with sigmoidal inclusion trails, mantled by a post-tectonic massive rim. Scale bar represents 0.25 mm. Abbreviations: B, biotite; G, garnet; Q, quartz.

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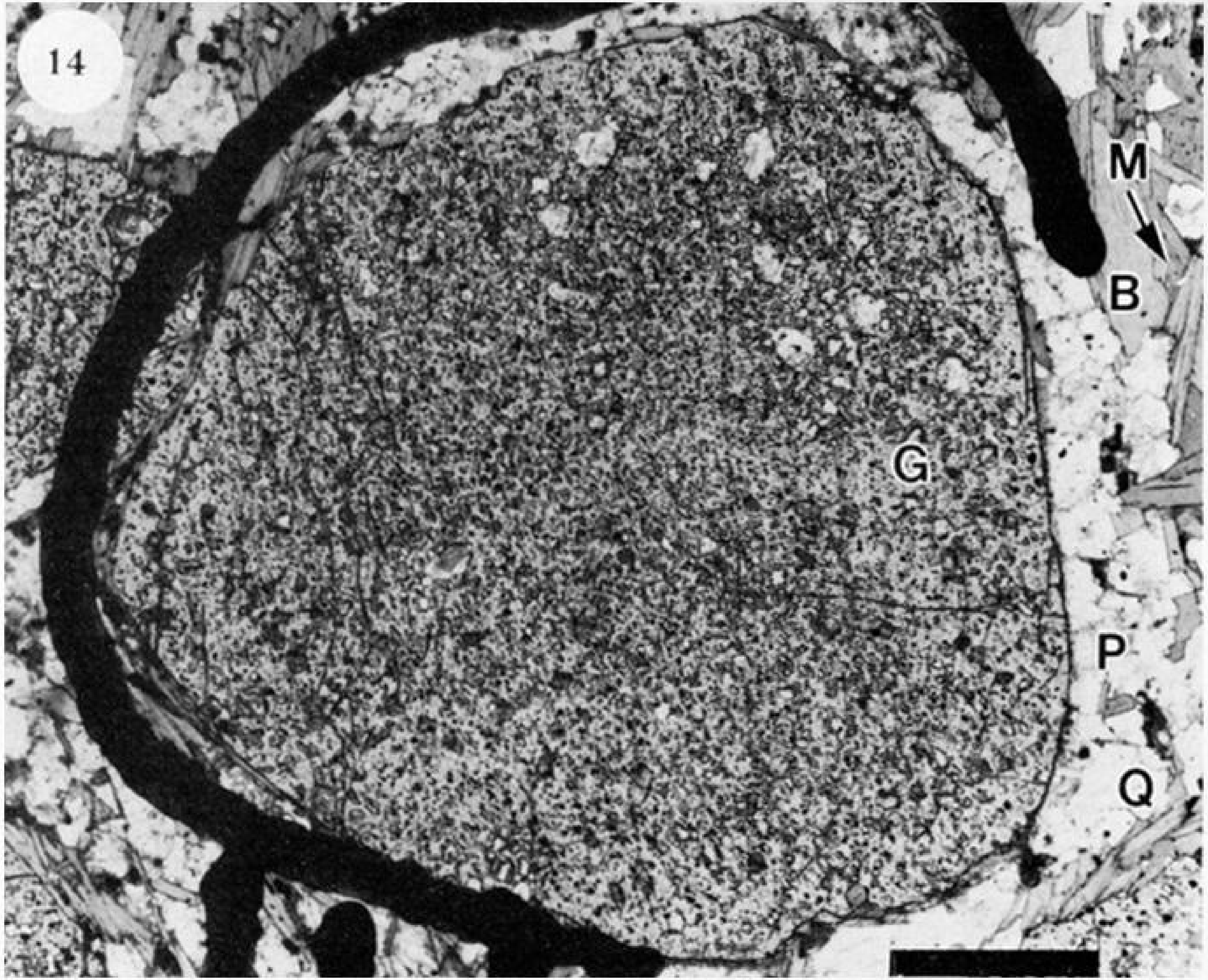

FIGURE 14. Photomicrograph of post-tectonic poikiloblastic garnet in metamorphic sample 3 from the Calderian allochthon (sample location shown in figures 4 and 5). Scale bar represents 1 mm. Abbreviations: B, biotite; G, garnet; M, muscovite; P, plagioclase; Q, quartz.

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