CHAPTER 4

ANDEAN ANALOGUE FOR LATE CARBONIFEROUS VOLCANIC ARC AND ARC FLANK ENVIRONMENTS OF THE WESTERN NEW ENGLAND OROGEN, NEW SOUTH WALES

INTRODUCTION

The palaeogeographic context of the New England Orogen during the Late Carboniferous (post-Visean) consisted of a volcanic terrain in the west, a central continental and marine shelf, and an eastern open ocean (Leitch, 1974; Day *et al.*, 1978; Roberts and Engel, 1980). The corresponding tectonic configuration comprised a magmatic arc, forearc basin and subduction complex (Leitch, 1975; Crook, 1980a,b; Fergusson, 1984a,b). The southern portion of the volcanic terrain was located at the margin of the Australian part of Gondwanaland, represented now by the northeastern Lachlan Fold Belt. Younger strata of the Sydney, Gunnedah and Bowen Basins cover the junction of the Lachlan Fold Belt and the New England Orogen, including much of the former volcanic terrain (Fig. 4.1). Parts of the adjacent shelf sequence have been removed by erosion. The Late Carboniferous sedimentary and volcanic rocks of the Currabubula Formation and its equivalents north of the Liverpool Range (Fig. 2.1) provide a record of processes and environments along a section of the eastern flank of the volcanic terrain. Comparison with an analogous modern setting in the Central Andes of South America has enabled reconstruction of the presently concealed or eroded components of the ancient context, and identification of the lithofacies assemblage typical of continental margin arc flank sequences.

CURRABUBULA SECTION OF THE EASTERN FLANK OF THE VOLCANIC ARC

Definition

The Currabubula (volcanic) arc flank section refers to a Late Carboniferous (326 to 290 Ma; Roberts and Engel, 1980) palaeogeographic entity represented by the Currabubula Formation (Voisey and Williams, 1964) and equivalents (White, 1965; McKelvey, 1974) north of the Liverpool Range (Fig. 4.1). The north-south extent so defined is

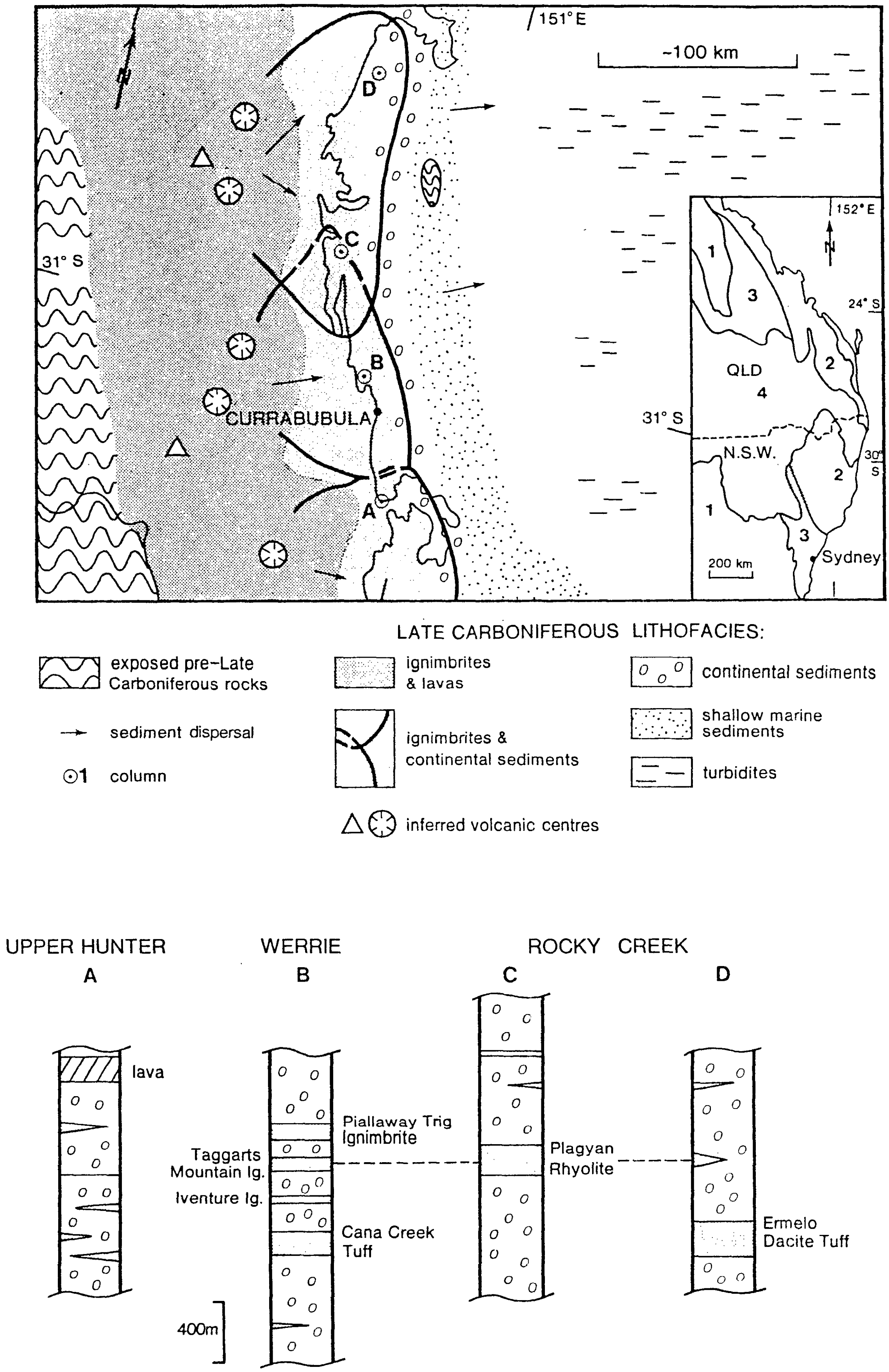


Figure 4.1: Reconstructed lithofacies map of the Currabubula arc and environs at about 310 to 300 Ma. Simplified stratigraphic columns are given for the pyroclastic fields outlined on the arc flank. Inset shows the regional context (Leitch, 1974): (1) Lachlan Fold Belt. (2) New England Orogen. (3) Sydney, Gunnedah and Bowen Basins. (4) Great Artesian Basin.

approximately 300 km. Original east-west dimensions are unknown but exceeded 30 to 40 km.

Lithofacies

Epiclastic rocktypes are substantially more voluminous than pyroclastic rocktypes. Of the former, orthoconglomerate (Fig. 2.6f) is dominant but accompanied by cross-bedded sandstone, laminated mudstone and paraconglomerate. Welded rhyolitic ignimbrite sheets are predominant over non-welded ignimbrite, less silicic (dacitic) ignimbrite, and ash-fall tuff. One silicic (dacitic?) lava flow of local extent has been found (Fig. 5.3).

Lithofacies distribution and palaeoslope

Pyroclastic rock units are most numerous and thickest in sections along the western side of present exposures (Fig. 2.2). Thicknesses of epiclastic rocks present between successive ignimbrites are greater in eastern than in western sections. However, the overall preponderance of conglomerate in the epiclastic component does not appear to alter from west to east. The relationships suggest westerly derivation of both the pyroclastic and epiclastic facies, and are consistent with palaeo-current indicators of sediment transport directions (White, 1968). The depths of incision and orientation of palaeovalleys outlined by ignimbrite sheets are in accord with an easterly dipping palaeoslope. Sources of the ignimbrite sheets were probably at least a few and in some cases tens of kilometres, to the west of present exposures.

Epiclastic processes and products

Conglomerate and coarse sandstone were deposited in an extensive fluvial braidplain environment (*cf.* Miall, 1977,1978; Rust, 1978,1979). Minor laminated mudstone possibly accumulated in restricted lakes and/ or abandoned braidplain channels. Some paraconglomerate has been interpreted as tillite which originated as ablation moraine (White, 1968). Uncommon intercalations of laminated pebbly mudstone containing ice-rafted dropstones also attest to the intermittently severely cold climate (e.g. Rosedale Member, Whetten, 1965). Regional studies of the distribution of late Palaeozoic glacigene facies imply the proximity of the Currabubula arc flank to alpine glaciers (e.g. Crowell and Frakes, 1971a,b; Martin, 1981). Such a setting is consistent with the high latitudes determined by analysis of the palaeomagnetism of Late Carboniferous sequences of New England, including the Currabubula Formation and equivalents (Irving, 1966).

Erosion surfaces evident at outcrop scale are common. Conglomerate beds are lenticular over distances of several hundred metres to kilometres parallel to strike, in part due to erosional truncations. Ignimbrite sheets delineate some major disconformities within the sequence, interpreted as the outlines of palaeovalleys up to a couple of kilometres in width and incised to several hundred metres (Fig. 2.7).

Pyroclastic processes and products

Pyroclastic flows periodically spread across the braidplain, leaving ignimbrite tens of metres thick in topographic depressions and channels, and covering low-lying interfluves. Most reached the eastern limits of present exposures, and presumably beyond, indicating minimum outflow distances of 25 to 30 km. All were partly excavated by later incision of drainage channels and thinned by removal of the non-welded topmost layers. Some have been reduced to a small fraction of their original extent. Airfall ash was probably generated with the pyroclastic flows (*cf.* Sparks and Walker, 1977) but was reworked soon after deposition and dispersed within the epiclastic facies, or else preserved in waterlain laminated tuffaceous mudstone intervals. Co-ignimbrite ash clouds and vent plumes from the most powerful eruptions no doubt bypassed the arc flank altogether, transporting fine ash to distal parts of the forearc and offshore to remote, open marine settings.

At least one large magnitude, explosive, hydrovolcanic eruption occurred, generating debris flows and floods which accompanied emplacement of particularly widespread primary pyroclastic deposits across the arc flank (Cana Creek Tuff Member; Chapter 3). A similar unit was deposited over much of the northern part of the arc flank at about the same time (Ermelo Dacite Tuff) but the details of its mode of eruption and emplacement have not yet been documented.

Pyroclastic fields of the eastern flank of the volcanic arc

Along the length of the Currabubula arc flank, the remnants of three pyroclastic fields have been recognised (Fig. 4.1). These differ in the number, arrangement and character of the included pyroclastic units (*cf.* Smith, 1960a, p.813). The Werrie pyroclastic field is centrally located and characterised by five ignimbrites, four of which have been defined as members of the Currabubula Formation (Chapter 2). The Rocky Creek pyroclastic field has a stratigraphy comprising three major ignimbrite sheets accompanied by numerous locally mappable units (McKelvey and White, 1964; White, 1965; McKelvey, 1968,1974). The Upper Hunter pyroclastic field begins 20 km north of the Liverpool Range and includes non-welded, comparatively thin (less than 15 m), biotite-bearing, quartz-poor ignimbrites of limited extent. This field contains the only known silicic(?) lava flow of the Currabubula arc flank sequence (Fig. 5.3).

The existence of pyroclastic fields is interpreted as a reflection of variation in both the source volcanic terrain and in the processes controlling the site of emplacement of its products. With regard to the former, spatial and temporal changes in frequency, composition, magnitude, style and proximity of volcanic eruptions can reasonably be inferred. Thus, different parts of the arc flank were supplied with distinctive contributions from the source terrain. In addition, topographically-controlled pyroclastic flows may have followed separate and largely static routes from source areas to the arc flank. Physical barriers within the braidplain and the comparatively narrow outflow range of most ignimbrites (a few to several tens of kilometres) would further enhance definition of areas with a shared pyroclastic record. Lateral variations also occur in the epiclastic facies but in general at a scale too large to discern mappable patterns. Particularly in the Currabubula Formation, ignimbrites provide the only means of internal differentiation.

Epiclastic provenance

The epiclastic rocktypes contain the following detrital assemblage:

silicic and intermediate porphyritic lavas; silicic welded ignimbrite; devitrified pumice and shards; quartz, feldspar, biotite; quartz-rich sandstone, one clast of which contains Devonian or older, spiriferid brachiopod and indeterminate bivalve mollusc fossils (Chapter 2), lithic sandstone and polymictic conglomerate; texturally diverse granitoids; micaceous schist, gneiss, slate, quartzite; fine grained hornfelses.

This collection reflects contributions from two major sources: an active volcanic terrain; and a source with exposure of Devonian and/or older, deformed, quartzose and lithic sedimentary sequences, regional metamorphic rocks, and granitoid plutons accompanied by contact aureole hornfelses.

Precursors

The Late Carboniferous conglomeratic sequence overlies similarly extensive volcanolithic, fluvial and paralic, Visean sandstones (Merlewood Formation, Voisey and Williams, 1964; Moore and Roberts, 1976; Caroda Formation, McKelvey and White, 1964; Mory, 1981,1982). Although primary volcanic units have been confirmed as a component of only the upper parts of the unit, intermediate volcanic detritus is ubiquitous. In addition to porphyritic lava clasts, conglomerate intervals contain essentially the same detrital assemblage as those in the Late Carboniferous but silicic volcanics are less abundant.

Outer (eastern) fringe of the Currabubula section of the volcanic arc flank

Shallow marine sedimentary rocks of possible Late Carboniferous age are isolated from the continental facies, and occur 50 km farther to the east (Price, 1973). No primary volcanic units have been reported but the detrital constituents are the same as those of the continental facies (Price, 1973; Mory, 1981). The youngest pre-Permian formation (Crow Mountain Creek beds, Price, 1973) includes minor easterly-derived conglomerate interpreted as evidence of uplifted early Palaeozoic rocks to the east in the outer fringe of the forearc (Price, 1973; Leitch, 1974). No continuous barriers existed because Late Carboniferous (?) flysch sequences to the east (e.g. Coffs Harbour beds, Texas beds) have been linked by provenance studies to the Late Carboniferous volcanic terrain (Korsch, 1978, 1984; Fergusson, 1982). The Currabubula volcanic arc flank may have bordered a Late Carboniferous marine shelf tens of kilometres wide which had emergent areas of limited extent on its eastern extremity, and a deeper marine setting beyond (Fig. 4.1).

LATE CARBONIFEROUS CONTINENTAL MARGIN: THE NORTHERN LACHLAN FOLD BELT

Studies of the history of the New England Orogen propose physical connection of the southern part of the Orogen and the Lachlan Fold Belt from at least the end of the Early Carboniferous (Leitch, 1974; Day *et al.*, 1978; Leitch and Willis, 1982). The present erosion surface of the northern Lachlan Fold Belt coincides closely with the unconformity at the base of Permian basin strata, and the level of exposure during the Late Carboniferous was probably only slightly less deep. The surface geology of the northern Lachlan Fold Belt is thus considered to be representative of its Late Carboniferous constitution.

The sedimentary rocks of the Lachlan Fold Belt are Late Devonian and older quartz-rich fluviatile and paralic sandstone, quartz-rich and quartz-poor flysch (Packham, 1969; Crook and Powell, 1976). Limestone and volcanogenic sequences are locally dominant. Foliated and massive, Carboniferous and older granitoid plutons cover large areas of the Fold Belt, the former being accompanied in some instances by metasedimentary schist and gneiss envelopes (e.g. Nymagee Igneous Complex, Pogson and Scheibner, 1976).

Thus, the provenance indicated by non-volcanic detritus in the Late Carboniferous conglomerates of the Currabubula arc flank section is consistent with derivation from the northern Lachlan Fold Belt. The necessary inference that at least part of the Lachlan Fold Belt was elevated and being eroded cannot be independently established, but is supported by the absence of sedimentary rocks younger than earliest Carboniferous, the early Carboniferous age of regional deformation (Powell *et al.*, 1976; Powell and Edgecombe, 1978) and evidence for pre-Permian excavation of much of the northern part of the Fold Belt.

LATE CAINOZOIC AND MODERN TECTONIC SETTING OF THE ANDES

The western margin of the South American continent meets oceanic crust of the Nazca plate in the vicinity of the Peru-Chile trench (Scholl *et al.*, 1970; James, 1971). The length and trend of the trench is mirrored onshore by the Andean cordillera, the spine of which is located 100 to 300 km to the east and coincident with a modern and late Cainozoic volcanic arc. Ignimbrites and lavas form a near-continuous veneer over older, deformed, lithologically diverse sedimentary rocks which are extensively exposed farther east behind the cordillera. Beyond, there are low-lying, fluvial sedimentary basins draining eastward to the Atlantic coast. The modern setting has antecedents which date back to major uplift and volcanism beginning in the Miocene (about 20 Ma), achieving a configuration similar to the present day in the Pliocene although the precise chronology of events varies regionally (Scholl *et al.*, 1970; James, 1971; Noble *et al.*, 1974; Baker, 1977; Kussmaul *et al.*, 1977; Lahsen, 1982; Thorpe *et al.*, 1982; Jordan *et al.*, 1983).

The principal elements of the South American continental margin, the volcanic arc and the active ocean-continent convergence to which it is related, have parallels with the reconstructed Late Carboniferous palaeogeography and tectonic regime of eastern Australia, parts of which are preserved in the New England Orogen and the Lachlan Fold Belt.

Approximately 280 km of the central Andean arc (Fig. 4.2) has been selected as an analogue for the Currabubula arc section in the Late Carboniferous. Other parts of the Andean arc lack active volcanoes (e.g. between latitudes 28°S and 33°S; Mortimer, 1973; Jordan *et al.*, 1983) or are dominated by andesitic and basaltic lavas of stratovolcanoes (Guest, 1969; Pichler and Zeil, 1972; Paskoff, 1977; Thorpe *et al.*, 1982). The informally-named Calama arc section extends from latitude 23°S northward to 21°S (Fig. 4.2). It is described in terms of three components which have existed since early Miocene times : the central volcanic chain (the Main Cordillera), the eastern zone, and the western flank of the arc.

WESTERN FLANK OF THE CALAMA VOLCANIC ARC SECTION

Modern physiography

Slope and drainage are overall westerly from the 3000 to 4000 m

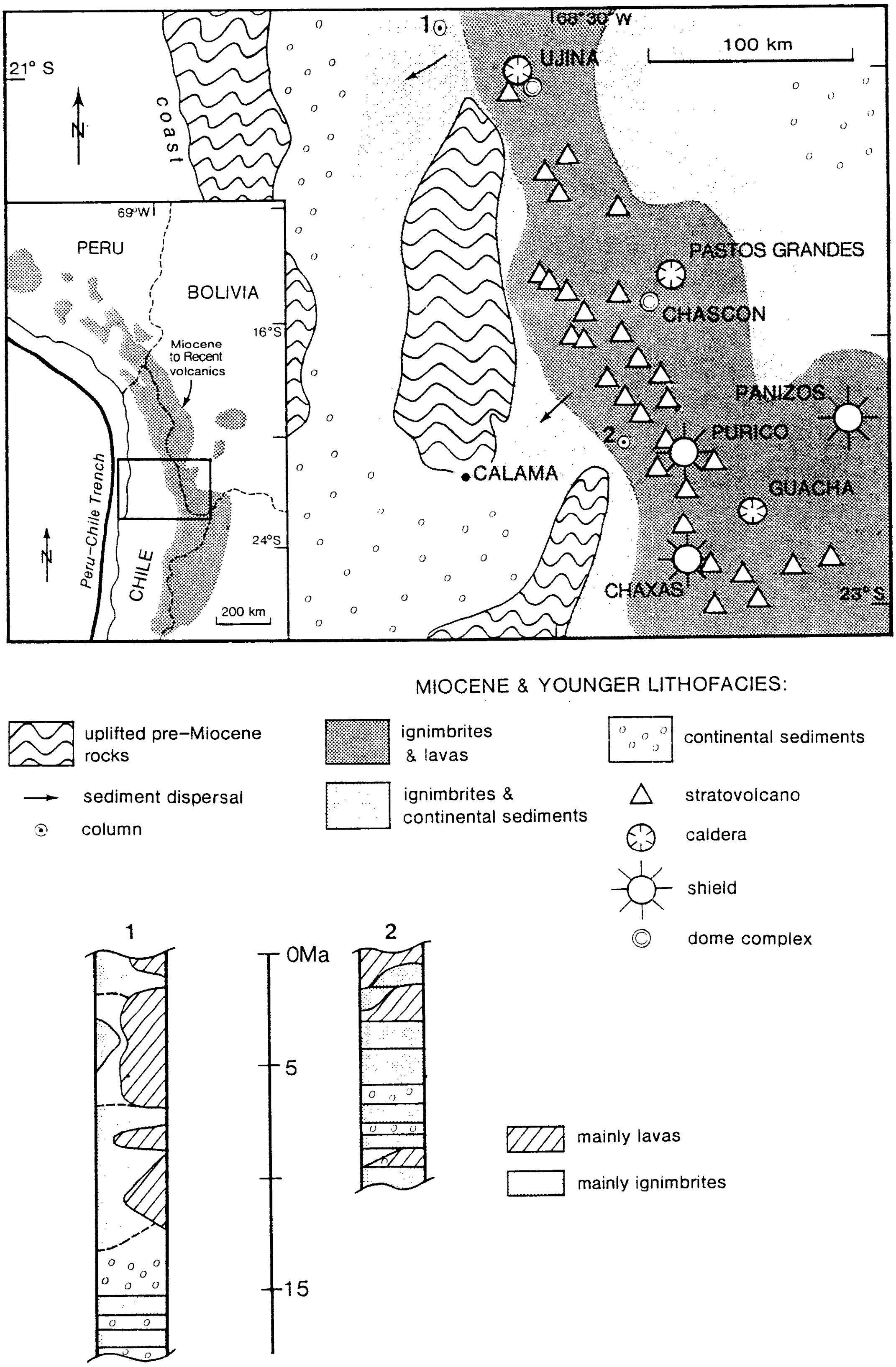


Figure 4.2: Simplified present-day distribution of lithofacies of the Calama section of the Andean volcanic arc. Inset shows the regional setting in relation to the Peru-Chile trench, after James (1971).

elevations of the Main Cordillera. There are meridional ranges (e.g. Cordillera de Domeyko) which interfere with this trend and define interconnected arc-parallel basins (e.g. Pampa del Tamarugal). The principal river channels are incised up to several tens of metres into Cainozoic deposits and fed by numerous less deeply incised tributaries and gullies.

Lithofacies

Regionally extensive continental sedimentary rocks intercalated with ignimbrites constitute the Miocene, Pliocene and Quaternary depositional record west of the Main Cordillera (Dingman, 1965; Hollingworth and Rutland, 1968; Guest, 1969; Pichler and Zeil, 1972; Baker, 1977; Baker and Francis, 1978; Lahsen, 1982). Stratigraphic sections differ in the thickness (400 m to more than 1000 m) and character of the sedimentary and volcanic components in adjacent parts of the western arc flank (e.g. Baker, 1977; Paskoff, 1977; Baker and Francis, 1978; Lahsen, 1982). Typical sections include up to six ignimbrite sheets separated by alluvial conglomerates, collectively 700 m to 1000 m in thickness (Hollingworth and Rutland, 1968; Guest, 1969).

Close to the Main Cordillera some of the youngest sediments are mudflow deposits and glacial moraine gravels, the latter being remnants of lowered snowlines during the Pleistocene (Hollingworth and Guest, 1967; Francis *et al.*, 1974; Paskoff, 1977). Otherwise,volcanogenic alluvial and fluvial conglomerate and sandstone are the dominant sedimentary facies. Modern lakes are common within young basins but are largely saline due to the very arid climate (Stoertz and Ericksen, 1974).

Rhyolitic and dacitic ignimbrite sheets are the most voluminous volcanic units of the arc flank (Francis *et al.*, 1974; Francis and Rundle, 1976; Lahsen, 1982). Welded and non-welded ignimbrites and 'sillar' have been described (Fenner, 1948; Hollingworth and Rutland, 1968; Guest, 1969; Francis *et al.*, 1974). Airfall ash and pumice layers are widespread (e.g. Francis *et al.*, 1974) but their volume and thickness contributions have not been evaluated. Andesitic lavas are restricted to within about 10 km of vents along the western edge of the Main Cordillera (Francis *et al.*, 1974) and silicic lavas are even more closely confined to the volcanic front. Block and ash flows of intermediate composition are similarly restricted in extent, although some debris flows originating on the slopes of volcanoes have reached several tens of kilometres from the source (Francis *et al.*, 1974).

Aggradation and erosion

Being arid, poorly vegetated and adjacent to a major mountain range, the Calama arc flank receives epiclastic detritus transported by infrequent floods and mass-flows, and deposited on alluvial fringes to the elevated region and in topographic depressions (Hollingworth and Rutland, 1968; Guest, 1969; Paskoff, 1977). Glacial advances westward from the Main Cordillera are correlated with deposition of outwash gravel on the flank due to accelerated highland erosion (Hollingworth and Guest, 1967; Scholl *et al.*, 1970; Paskoff, 1977). Pyroclastic flows and avalanches have intermittently inundated the arc flank, the more voluminous of which buried and smoothed the pre-existing topography (Hollingworth and Rutland, 1968; Guest, 1969; Francis *et al.*, 1974).

Erosion is most effective in areas covered by non-welded ignimbrite and unconsolidated airfall ash close to the low ranges formed since the Pliocene (Chong, 1977). At some localities drainage has been gradually altered by repeated pyroclastic influxes (e.g. Hollingworth and Rutland, 1968; Mortimer and Saric, 1975). Elsewhere channels incised in relatively resistant strata have remained fixed, with pyroclastic blockages being excavated instead (Hollingworth and Rutland, 1968).

Tertiary deformation

Deformation accompanying uplift of the Main Cordillera (post 25 to 20 Ma) has produced gentle warps and monoclines related to steeply dipping faults in coëval sequences on the western flank (Hollingworth and Rutland, 1968; Guest, 1969; Scholl *et al.*, 1970; Katz, 1971). Substantial incision of drainage since emplacement of the most recent ignimbrites is presumably a response to another increment of uplift (Hollingworth and Rutland, 1968; Mortimer, 1973; Mortimer and Saric, 1975). Farther from the ranges, thick deposits of principal basins are less affected and essentially conformable although erosional disconformities occur (Hollingworth and Rutland, 1968).

EASTERN ZONE OF THE CALAMA ARC SECTION

East of the Main Cordillera, there is a high plateau (the Altiplano-Puna, 4000 m) and another high range (the Eastern Cordillera, 6000 m). These elevated regions decline eastward to extensive, low-lying riverine plains (James, 1971).

The high plateau is constructed of Miocene and younger, volcanic and continental sedimentary rocks that cover deformed Mesozoic and Palaeozoic basement (James, 1971; Kussmaul *et al.*, 1977). Basement rocks are more extensively exposed in the ranges adjacent to the east (Jordan *et al.*, 1983).

CALAMA SECTION OF THE ANDEAN VOLCANIC ARC

Miocene and younger volcanic centres are clustered within a belt 60 to 100 km across, extending the length of the arc section (Fig. 4.2; Deruelle, 1978,1982; Baker, 1981; Thorpe *et al.*, 1982). The highest peaks are andesitic stratovolcanoes reaching altitudes of 6000 m. Accumulated primary volcanic and volcaniclastic deposits form a bench at 4000 m which declines westward to the ignimbrite fields of the arc flank (Guest, 1969; Francis *et al.*, 1974). Drainage is complicated by the constructional relief of the volcanoes and their products.

Volcanic centres

Stratovolcanoes with major central and peripheral minor vents form spectacular conical mountains built from their products (e.g. San Pedro-San Pablo volcanic centre; Francis *et al.*, 1974). Extinct, dormant and active examples are present and their existence as a component of the Andean arc since its Miocene inception has been established (Pichler and Zeil, 1972; Mortimer and Saric, 1975; Francis and Rundle, 1976; Baker, 1977).

The sources of voluminous and extensive ignimbrite sheets are, in general, topographically inconspicuous and have only been identified relatively recently so none have been mapped in detail (Kussmaul *et al.*, 1977; Baker and Francis, 1978; Baker, 1981; Francis *et al.*, 1983). Some of these centres are calderas with central depressions, the largest of which show signs of resurgent stages (e.g. Cerro Guacha, Fig. 4.2; Baker and Francis, 1978; Baker, 1981). Ignimbrites extend from the caldera rims in sheets and elongate fans controlled by topography. Lava domes occur within and at the margins of the calderas. Other ignimbrite sources lack collapse structures and are termed "ignimbrite shields" (e.g. Cerro Panizos, Fig. 4.2; Kussmaul *et al.*, 1977; Baker, 1981; Thorpe *et al.*, 1982). These are low relief mounds of ignimbrite covering hundreds of square kilometres. Central clusters of lava domes may mark the sites of vents for the apron of ignimbrite. For both the calderas and the shields, constructional relief is only locally developed and confined to the vicinity of silicic lava bodies. None of the ignimbrite centres shows signs of activity but there are Pleistocene ignimbrites in the arc flank sequence (e.g. 0.77 Ma ignimbrite reported by Baker, 1977).

Lithofacies and distribution

The stratovolcanoes produce intermediate composition lava flows, block and ash flows, small-volume ignimbrites, and airfall ash, as well as contributing volcaniclastic materials to mudflows, collapse breccias and epiclastic deposits (Hollingworth and Rutland, 1968; Guest, 1969; Francis *et al.*, 1974). Primary volcanic units from these volcanoes are voluminous but not extensive. Because lavas and block and ash flows rarely reach more than 10 to 20 km from the source (Guest, 1969; Walker, 1973b; Francis *et al.*, 1974), they are not a component of arc flank sequencesexcept very close to the volcanic front. Redeposited and reworked intermediate volcaniclastic accumulations are also relatively closely confined to the slopes and margins of the cones, but intermittent mudflows contribute significantly to the arc flank (Francis *et al.*, 1974).

Ignimbrite from shields and calderas is in places as voluminous as the intermediate volcanics within the arc but is far more extensive on its flanks (Francis *et al.*, 1974; Francis and Rundle, 1976; Baker and Francis, 1978; Baker, 1981; Thorpe *et al.*, 1982). Ignimbrite is found several tens of kilometres from source, having followed drainage channels and then spread upon reaching areas of gentle relief (Hollingworth and Rutland, 1968; Guest, 1969; Francis and Baker, 1978). Though intermediate lavas dominate the arc volcanic pile, silicic ignimbrites are the main arc flank representatives of its activity.

The unconsolidated pyroclastic deposits of the arc contribute pumice, ash and crystals to the arc flank-directed erosion systems, and to intermontane depressions. Co-ignimbrite and vent-produced airfall ash has probably been deposited principally outside the arc (e.g. in gravel sequences of the Pampa del Tamarulgal; Guest, 1969) but no studies have yet provided detailed descriptions.

Spatial and temporal variation in geology of the Calama arc section

Volcanism, uplift and glaciation have shaped the modern Calama arc section and operated at varying rates in different locations during its late Cainozoic development.

Volcanism

Ages of the oldest ignimbrites of the Calama arc section are younger in the south (latitude 21° to 23°S; 10 Ma) than in the north (20° to 21°S; 16 Ma), a pattern which continues into northern Chile (21 Ma north of 19°30'S) and southern Peru (Noble *et al.*, 1974; Baker, 1977; Baker and Francis, 1978; Lahsen, 1982). The distribution of ignimbrite centres is uneven, being clustered toward the south and sparsely scattered toward the north (Fig. 4.2). Stratovolcanoes have outnumbered the ignimbrite centres and been almost continuously active since the mid-Miocene, becoming dominant in the Pliocene and Quaternary (Francis and Rundle, 1976; Baker, 1977; Baker and Francis, 1978; Lahsen, 1982; Thorpe *et al.*, 1982).

The meridional changes in arc volcanism are imprinted on the depositional record of the adjacent proximal arc flank. However, major shifts in the lava-ignimbrite balance have little effect on more distal sites because all but the most extensive lava flows are filtered from the sequence beyond about 10 km from source. These areas preserve pyroclastic stratigraphies with subtle differences which reflect the uncoordinated evolution of the source ignimbrite centres.

Uplift

Volcanism and uplift have proceeded in concert during the late

Cainozoic although the rate of each has varied in time and along the length of the arc (e.g. Clark *et al.*, 1967; Paskoff, 1977; Baker and Francis, 1978; Jordan *et al.*, 1983). Distal and young parts of the Calama arc flank have not been affected by deformation attendant on uplift to the same degree as proximal and older parts, but include low-angle unconformities and are incised by episodically rejuvenated streams (Hollingworth and Rutland, 1968; Guest, 1969; Mortimer and Saric, 1975; Chong, 1977).

Glaciation

Alpine glaciers occur on three of the highest peaks of the Main Cordillera immediately south of the Calama arc section (Hollingworth and Guest, 1967). There were more glaciers during the Pleistocene, as indicated by the glacial sculptures of many of the older stratovolcanoes including those of the Calama arc (Francis *et al.*, 1974; Paskoff, 1977). These glaciers left locally extensive banks of moraine gravels but were in most cases less than 10 km in length (Paskoff, 1977). Beyond the proximal portions of the arc flank the only representatives of glaciation are fluvioglacial outwash deposits, and thick alluvial gravels in the Pampa del Tamarulgal (Hollingworth and Guest, 1967; Paskoff, 1977). At least two and possibly five glaciations have affected the Main Cordillera in the Pleistocene, each correlated with acceleration in erosion rates (Paskoff, 1977).

COMPARISON OF THE CURRABUBULA AND CALAMA ARC FLANKS AND CONTINENTAL MARGINS

The Late Carboniferous Currabubula and late Cainozoic Calama arc flanks match well in terms of environments, lithofacies and processes, and are both dominated by arc volcanism (Fig. 4.3). The outer fringe of the arc flanks and the continental margins in each case are broadly comparable in physiography and lithofacies. However, in contrast to the Calama arc section, there is no evidence for the existence of a Late Carboniferous foreland basin to the Currabubula arc section. Neither were there subaerial ridge and basin features in the outer fringe of the Currabubula section of the arc flank similar in scale to those west of Calama. Smaller, submarine equivalents may have existed (Price, 1973; Leitch, 1974), and elsewhere in the Late Carboniferous forearc, thick

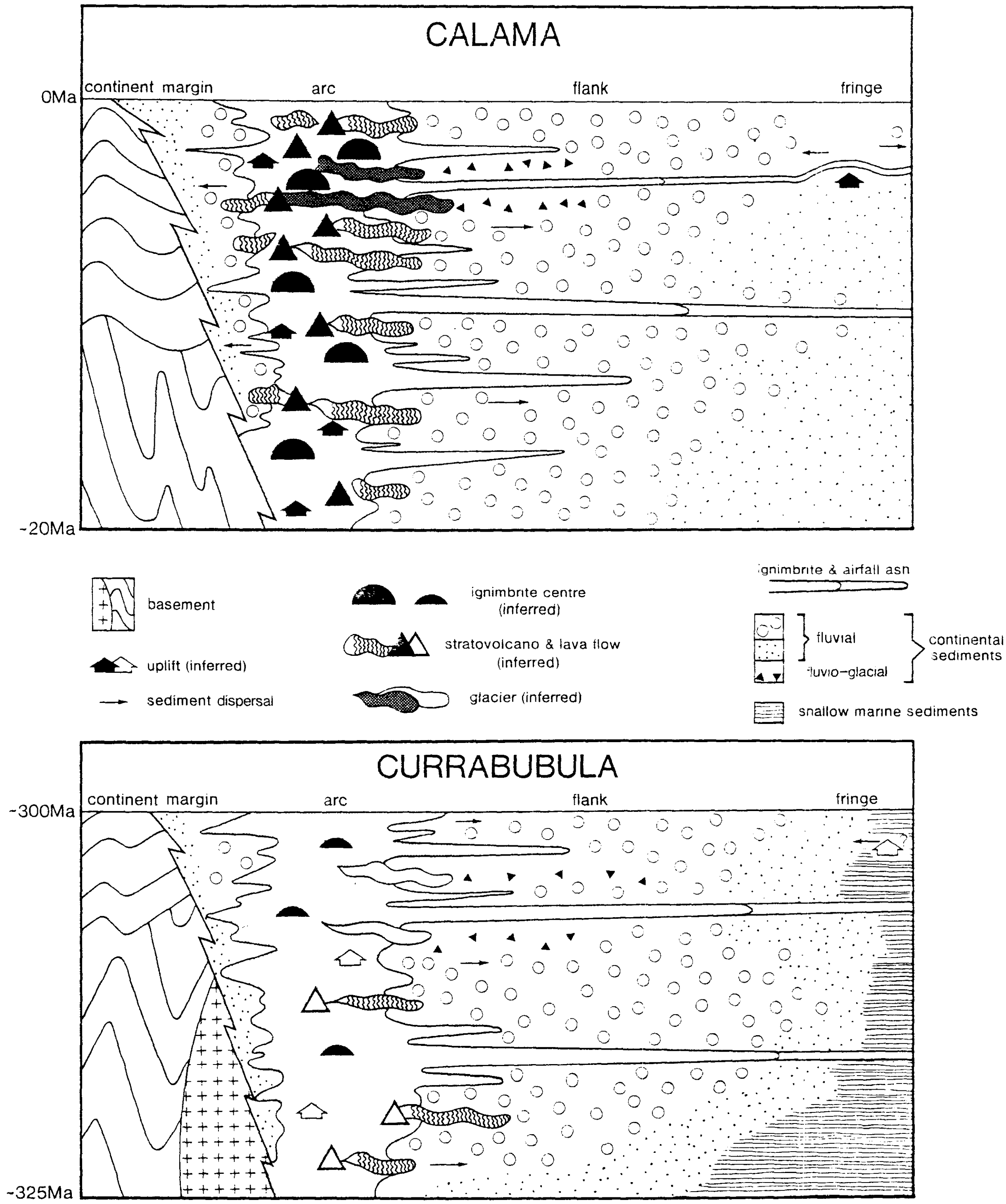


Figure 4.3: Time-space diagram showing the development of the late Cainozoic Calama (E to W) and Late Carboniferous Currabubula (W to E) continental margin volcanic arcs and environs.

shallow marine and continental sedimentary sequences accumulated in depressions interpreted as subsidiary rift basins (Myall Trough; Skilbeck, 1982).

Exposures of the Mesozoic and Cainozoic Andean Batholith (James, 1971) occur west of the Calama arc, and possibly form part of the basement offshore but east of the Chile trench. There were apparently no parallels in similar sites of the Currabubula forearc. Plutons coeval with the Currabubula Formation exist to the west in the northern part of the Lachlan Fold Belt (Facer, 1978) and immediately east of the Tamworth Belt where they intrude rocks of the late Palaeozoic subduction complex to which the Currabubula arc was related (Shaw and Flood, 1981; Flood and Fergusson, 1982).

These discrepancies in the analogy proposed detract little from the generally satisfactory correspondence of tectonic framework and palaeogeography in view of the diversity in structure and constitution accommodated by models for convergent plate margins where subduction operates (e.g. Karig and Sharman, 1975; Dickinson and Seely, 1979).

THE CURRABUBULA VOLCANIC ARC

The similarity of the ancient Currabubula and modern Calama arc flanks provides the basis for reconstruction of the character of the Late Carboniferous volcanic arc (Figs. 4.3,4.4). Volcanism, uplift and glaciation have been the dominant influences on the Cainozoic history of the Calama arc and are inferred to have been important in the evolution of the Currabubula arc.

Silicic, explosive eruptions from ignimbrite centres of the Currabubula arc section at different localities occurred intermittently for 20 to 30 Ma, initially accompanied by restricted, small volume, intermediate pyroclastic flows possibly contributed by stratovolcanoes. Continued activity of stratovolcanoes cannot be confirmed. Whether alive or extinct, they shed porphyritic lava clasts eastward to the braidplain alluvium of the arc flank. As at Calama, limited portions of the flank of the Currabubula arc were supplied with a distinctive set of pyroclastic deposits, ensuring the identity of the three pyroclastic fields delineated so far. Partial overlap of adjacent fields resulted

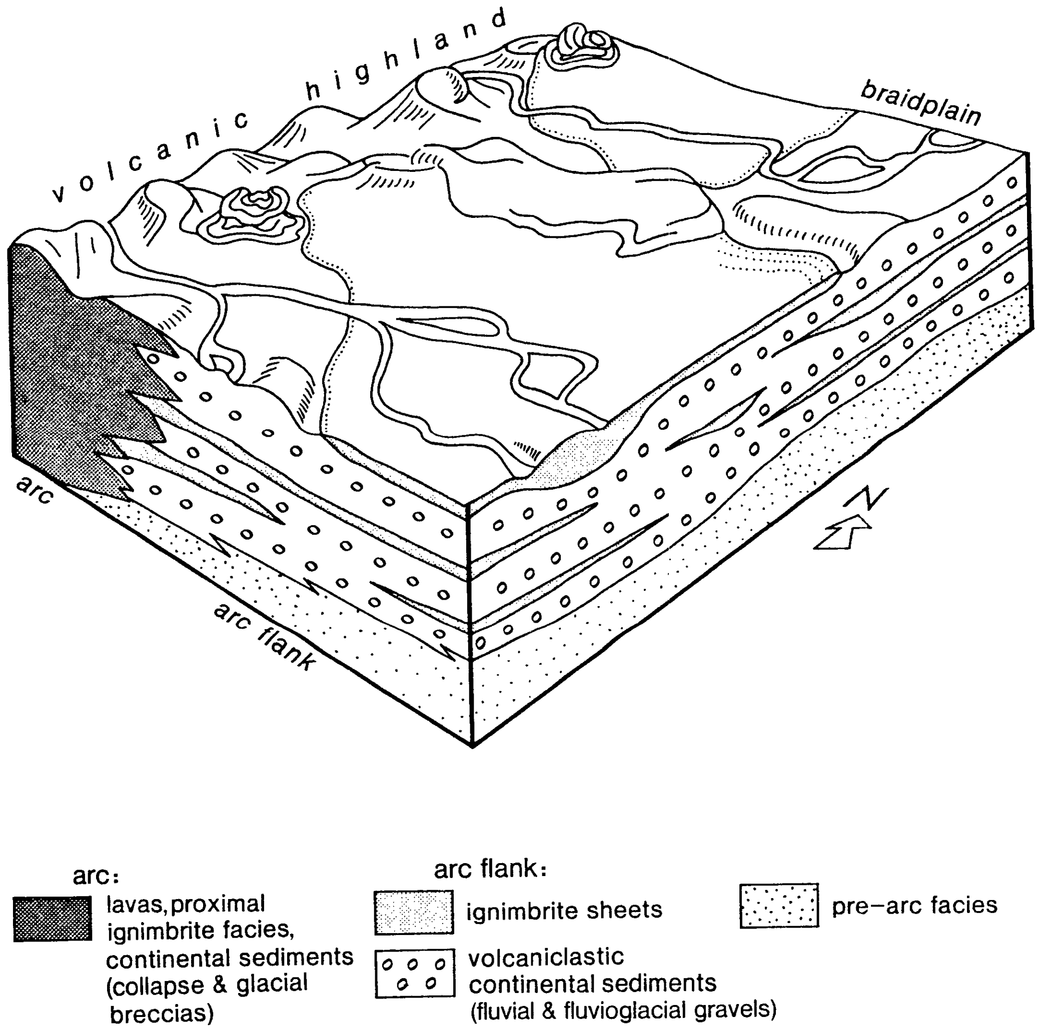


Figure 4.4: Schematic reconstruction of environments and lithofacies of the Late Carboniferous Currabubula continental margin volcanic arc. The surface configuration is intended to match that in existence at about 310 to 300 Ma.

where topographic barriers and drainage had minimal influence on distribution at the site of deposition.

The conglomerates and ignimbrites of the Currabubula Formation may represent a coordinated response to a distinct episode of source area elevation, as displayed by the Calama analogue which has had a history of accelerated uplift and deformation since the inception of Miocene volcanism on the Main Cordillera. The Late Carboniferous uplift was probably initiated by the Visean regional deformation of the northern Lachlan Fold Belt, known to constitute the continent juxtaposed with the Currabubula volcanic chain. At a finer scale, the common disconformities in the Currabubula Formation and equivalents could be symptoms of incremental adjustments of the topography of the mountainous source, as exemplified by sequences of the Calama arc flank.

The highest peaks of the Currabubula volcanic chain were evidently above the Late Carboniferous snowline and some supported glaciers that penetrated beyond the eastern volcanic front for brief periods. In the Calama setting, glaciation has been similarly restricted in extent and frequency, though the effects have been locally significant on landforms and sedimentary facies. Paskoff (1977) related the growth of glaciers to geographical and temporal variations in precipitation and other atmospheric circumstances pertaining to the Andes.

Development of the Calama arc has been dominated by pyroclastic volcanism, particularly the production of landscape-smoothing ignimbrites, and by the mountainous relief of the volcanic spine. Glaciation has had a somewhat subordinate influence, being confined to small areas within or very close to the volcanic chain and to few, brief time intervals. Perhaps a similar balance between these formative processes prevailed for the Late Carboniferous situation in the vicinity of Currabubula.

GEOCHEMICAL CHARACTERISTICS OF THE CALAMA AND CURRABUBULA ARC VOLCANOES

Exhaustive studies of the geochemistry of late Cainozoic volcanics along the Andean arc have demonstrated the presence of voluminous, typically calc-alkaline and high-K calc-alkaline silicic lavas and ignimbrites, bordered to the east by shoshonitic and minor alkaline lavas north of

25°S, or by an alkaline suite farther south (Pichler and Zeil, 1972; Kussmaul *et al.*, 1977; Deruelle, 1978,1982; Lahsen, 1982; Thorpe *et al.*, 1982). Within these major zones, there are meridional variations in the petrology and geochemistry of the late Cainozoic products of the Andean arc (e.g. Francis *et al.*, 1977).

The geochemical character of ignimbrites from the Calama arc section is summarised in Table 4.1 and on Figure 4.5. Compilations of chemical data for rhyolites from the entire South American arc (both lavas and ignimbrites, greater than 69% SiO_2), give similar average values for the tabulated components (Ewart, 1979; Ewart and Le Maitre, 1980). In general, the Late Carboniferous ignimbrites of the Currabubula Formation, especially the welded devitrified samples, are comparable to the ignimbrites of northern Chile (Table 4.1). A more rigorous evaluation of similarities is unwarranted. The data on the Carboniferous ignimbrites are whole rock analyses (Tables 4.2,4.3), rather than pumice analyses. The originally non-welded ignimbrites show the widest range in alkali values (those of the Cana Creek Tuff Member, analyses 1 to 5, Table 4.2). The initially high porosities of these ignimbrites most probably promoted the post-emplacement alteration of glassy pyroclasts. Crystal fragments in the welded, vitrophyric ignimbrite samples show minimal alteration. However, the glassy matrices of these ignimbrites have been hydrated. The K_2O values are consistently lower in the vitrophyric samples relative to devitrified representatives of the same unit. All the analysed ignimbrites are oxidised; norms were calculated with Fe_2O_3:FeO fixed at 0.3 (*cf.* Francis *et al.*, 1974).

Chemical classification based on percentages of SiO_2 and K_2O for analyses recalculated H_2O- and CO_2-free, indicates the calc-alkaline and high-K calc-alkaline character of the Currabubula Formation rhyolitic ignimbrites (Fig. 4.5a). However, those analyses in the calc-alkaline field refer to either vitrophyric or to non-welded ignimbrites. The K_2O values of these samples are considered to be less reliable than those of the devitrified ignimbrites. The latter all plot within the high-K calc-alkaline field. The co-incidence of the welded devitrified samples with the high-K series on the Rb versus K/Rb diagram (Fig. 4.5b) also suggests that the ignimbrites were originally high-K rhyolites. Vitrophyric

TABLE 4.1: Comparison of selected major element oxides and trace elements for Cainozoic ignimbrites of northern Chile and Late Carboniferous ignimbrites of the Currabubula Formation, New South Wales.

	Ignimbrites, northern Chile				Andes[1]	Ignimbrites, Currabubula Formation	
	Thorpe *et al.* (1979)	Deruelle (1982)	Thorpe *et al.* (1982)	Palacios (1984)	Ewart (1979)	14 welded devitrified ignimbrites	29 ignimbrites
major element oxides, wt%							
SiO_2 range	64-76	58-74	64-75				72 - 80
rhyolites:							
SiO_2 average		71		69	73.6		76
CaO	<1				1.59	0.09 - 1.86	0.04 - 3.92
Na_2O	~3				3.70	2.75 - 4.33	0.13 - 4.54
K_2O	~5	5		3.9	4.26	3.61 - 5.31	1.13 -12.10
K_2O/Na_2O	1.6	2		1.15	1.15	0.9 - 1.9	0.3 -90
trace elements, μg/g							
V		40			33	<2 - 25	<2 - 26
Cr		10			n.d.[2]	7 - 12	<2 - 12
Ni		0			12	<2 - 2	<2 - 3
Cu		<10			15	<2 - 4	<2 - 8
Rb	98-224	100-200	high	249	155	109 -201	88 -301
Sr	29-282	250-350	high	150	160	34 -295	5 -672
Ba				640	641	546 -1365	168 -1699
Rb/Sr	0.3-7				1.0	0.4 - 3.5	0.2 - 60

[1]Data compilation encompasses Western South America, and includes lavas.

[2]n.d., insufficient data.

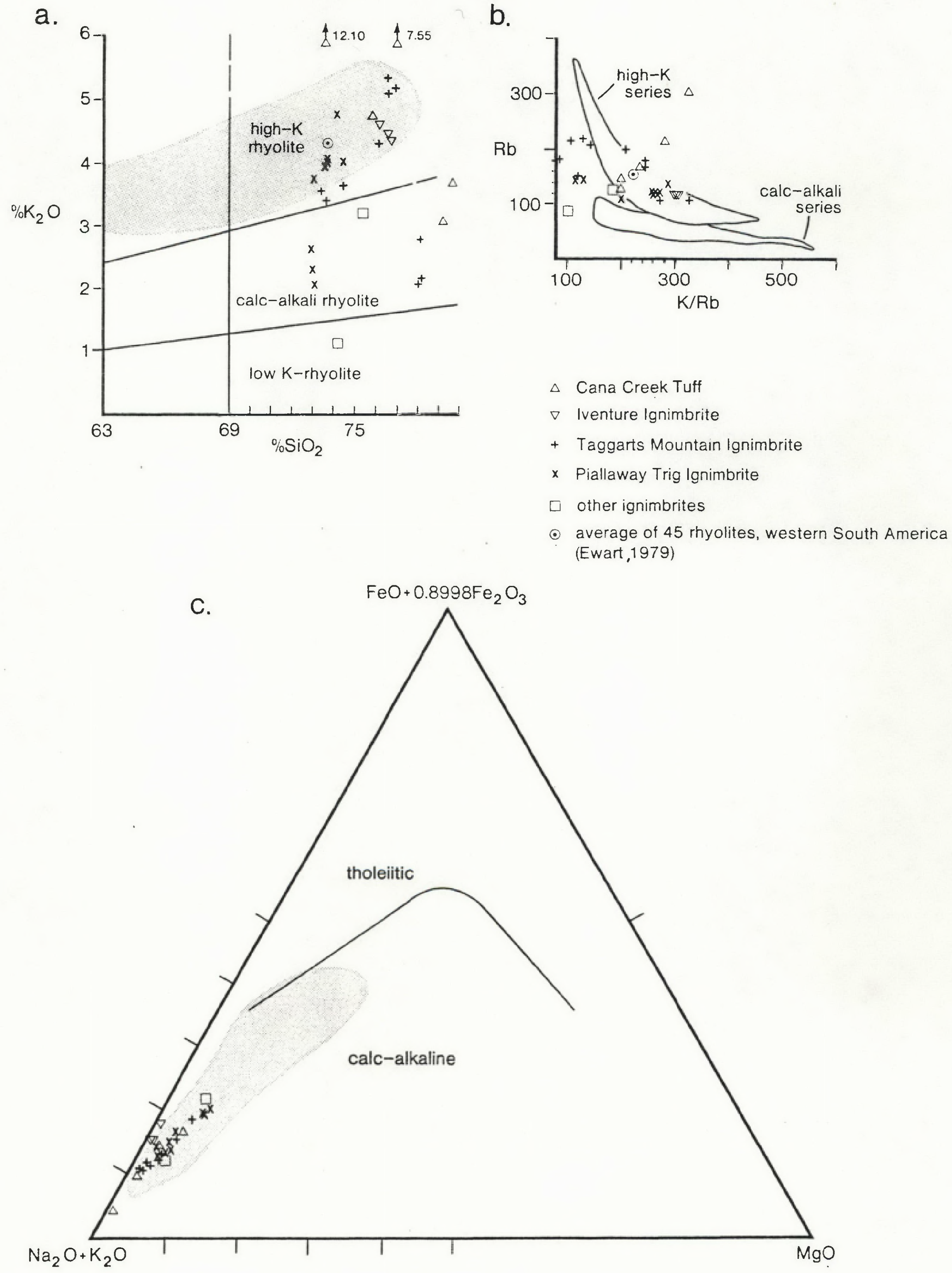
Figure 4.5: a. Anhydrous K_2O versus SiO_2 values of ignimbrites of the Currabubula Formation in relation to the subdivisions of Ewart (1979).

b. µg/g Rb versus K/Rb values of ignimbrites of the Currabubula Formation in relation to the high-K and calc-alkali series identified by Ewart (1979 , Fig. 4).

c. AFM diagram showing the ignimbrites of the Currabubula Formation in relation to tholeiitic and calc-alkaline fields of Irvine and Baragar (1971, Fig. 2).

a,c. Shaded field corresponds to north Chilean ignimbrites after Francis *et al.* (1974, Figs. 15,12).

a,b. Average of 45 rhyolites (more than 69% SiO_2) from western South America, after Ewart (1979, Table 4A).


a.
12.10
7.55
high-K
rhyolite
%K2O
calc-alkali rhyolite
low K-rhyolite
6
5
4
3
2
1
63
69
75
%SiO2
b.
high-K
series
calc-alkali
series
Rb
300
100
100
300
500
K/Rb
Cana Creek Tuff
Iventure Ignimbrite
Taggarts Mountain Ignimbrite
Piallaway Trig Ignimbrite
other ignimbrites
average of 45 rhyolites, western South America
(Ewart,1979)
c.
FeO+0.8998Fe2O3
tholeiitic
calc-alkaline
Na2O+K2O
MgO

samples are scattered on the low K/Rb side of the high-K series and do not fall within the calc-alkaline series identified by Ewart (1979). On the AFM diagram (Fig. 4.5c), the Currabubula Formation ignimbrites display a calc-alkaline rather than tholeiitic trend (*cf.* Irvine and Baragar, 1971).

The predominance of silicic ignimbrites with comparatively high K_2O, Rb, Ba, Sr, low V, Cr, Ni, and the absence of basalts and basaltic detritus in the Currabubula Formation are consistent with location of the volcanic centres in a setting underlain by thick continental or 'mature' (rather than oceanic) crust, as exists beneath the Calama arc (Dickinson, 1970; James, 1971; Ewart, 1979,1982; Ewart and Le Maitre, 1980; Green, 1980; Coulon and Thorpe, 1981; Leeman, 1983).

The overall compositional and petrographic similarity of the products of the ancient Currabubula and modern Calama arcs is in accord with the proposed palaeogeographic analogy.

DISCUSSION

Although the formation and evolution of the Andes is primarily a consequence of subduction of oceanic lithosphere at the Peru-Chile trench, the thickness of the continental crust and the dip of the down-going oceanic plate (indicated by the Benioff Zone) are two subsidiary causes of regional, arc-parallel differences in geology and recent history (James, 1971; Barazangi and Isacks, 1976; Megard and Philip, 1976; Francis *et al.*, 1977; Deruelle, 1982; Jordan *et al.*, 1983). Even those parts of the arc overlying crust of uniform thickness with consistent Benioff Zone inclination display variations in their volcanic and tectonic history: for example, in the central segment of thick crust where the Benioff Zone dips at 25° to 30° (between 15°S and 24°S; Jordan *et al.*., 1983), eruptive records differ in date of onset, frequency, style, location and rates of production, while the timing and intensity of episodes of uplift (and attendant peripheral deformation) have locally been uncoordinated.

The Currabubula arc section was a fraction of the 1600 km volcanic chain bordering the eastern Australian continent in the Carboniferous and related to subduction of the palaeo-Pacific oceanic plate (Leitch, 1974,1975; Day *et al.*, 1978). Though less than half the length of the Andean chain, the ancient Australian system was likely to have displayed comparable variability in constitution and evolution if the processes working to diversify the modern arc also operated in the past.

Differences in detail of the history of the Late Carboniferous volcanic arc, such as has been inferred for the Currabubula section, indirectly attest to at least minor arc-parallel variation. Regional changes in the age of onset and duration of volcanic activity also occur. In the Lower Hunter Valley, Visean ignimbrites are voluminous in the arc flank sequence (Gilmore and Nerong Volcanics; Nasher *et al.*, 1976; Crane and Hunt, 1980; Roberts and Engel, 1980) whereas north of the Liverpool Range, the earliest major ignimbrites are post-Visean. This age trend is not continuous into the northern part of the New England Orogen where there is a Late Devonian record of the inception of the volcanic arc and signs of its extinction by mid-Carboniferous times (Day *et al.*, 1978). Over much of the southern part of the Orogen, volcanism persisted through the Late Carboniferous and into the Early Permian (Boggabri and Gunnedah Volcanics; Seaham Formation; Alum Mountain Volcanics; Leitch, 1969, 1974,1975; Crane and Hunt, 1980).

There are also differences in the products of the arc volcanoes recognisable at the same broad scale. The southern (Hunter Valley) and northern (Connors arc) extremities include substantial thicknesses of andesitic lavas whereas central portions are dominated by silicic ignimbrites (Auburn arc; Currabubula arc section). Similar north-south contrasts in the Andean volcanic chain (and other settings) have been correlated with variations in crust thickness (Francis *et al.*, 1977; Green, 1980; Coulon and Thorpe, 1981). Such an explanation may be appropriate for the Carboniferous volcanic chain of eastern Australia, although proximity to source vents has an important control on the presence or absence of lavas in arc flank sequences.

CONCLUSIONS

The late Cainozoic history and present configuration of the Andean volcanic arc reflect the operation of subduction of oceanic lithosphere at a convergent plate boundary (James, 1971; Jordan *et al.*, 1983).

Regional variation in the circumstances of subduction (Benioff Zone inclination and continental crust thickness) has produced considerable diversity in the constitution of the arc and its environs. The geology of part of the arc flank is comparable in detail to the Late Carboniferous Currabubula Formation, inferred to have formed ocean-ward of a continental margin volcanic arc analogous to the corresponding section of the Andean arc. In both the ancient and modern contexts, the arc flank sequences have developed in response to the interplay of volcanism, uplift and deformation, and glaciation episodically afflicting the highland source over a period of some 20 Ma.

These two cases exemplify the facies expected to permanently accumulate during the active life of a continental margin volcanic arc, summarised by the reconstruction in Figure 4.4. Pyroclastic deposits (ignimbrites, ash-fall tuffs) dominate the arc flank regardless of their proportion relative to lavas within the volcanic chain. Conglomerate-ignimbrite stacks, interrupted by erosion surfaces at a range of scales, are characteristic of the arc flank, passing into volcanogenic sedimentary facies at more distal localities. Coincidence of ice and/or snow with active ignimbrite centres in the volcanic arc provides favourable circumstances for large magnitude hydrovolcanic eruptions, although arc flank sequences are in general typified by 'normal' outflow ignimbrite sheets.

TABLE 4.2: Major element oxide analyses in weight percent and C.I.P.W. norms of ignimbrite[1] members of the Currabubula Formation

Sample	Cana Creek Tuff Member[2] △					Iventure Ignimbrite Member ▽		
	1	2	3	4	5	6	7	8
Oxide	R55119	R55097	R55100	R55104	R55045	R55156	R55157	R55154
SiO_2	73.21	74.02	75.21	76.89	75.20	76.11	75.64	75.44
TiO_2	0.07	0.13	0.09	0.07	0.08	0.15	0.16	0.15
Al_2O_3	13.16	13.11	12.65	11.96	11.92	12.35	12.50	12.35
Fe_2O_3	0.43	1.25	0.76	0.96	0.91	1.66	2.00	1.65
FeO	0.17	0.26	0.28	0.31	0.16	0.08	0.09	0.09
MnO	0.02	0.01	n.d.	0.02	0.01	0.07	0.07	0.06
MgO	0.11	0.20	0.13	0.30	0.23	0.04	0.08	0.10
CaO	0.04	0.69	0.09	0.52	0.45	0.10	0.16	0.09
Na_2O	0.13	3.20	1.14	1.81	2.98	4.29	3.97	4.14
K_2O	12.02	4.59	7.38	3.53	2.90	4.28	4.54	4.36
P_2O_5	n.d.[4]	n.d.	0.01	n.d.	0.02	n.d.	0.02	n.d.
H_2O^+	0.45	2.19	1.70	3.18	3.72	0.37	0.46	0.19
H_2O^-	0.23	0.53	0.38	0.43	1.23	0.21	0.36	0.35
CO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	100.04	100.18	99.82	99.98	99.81	99.71	100.05	98.97
ΣFe as FeO	0.56	1.38	0.96	1.17	0.98	1.57	1.89	1.57
K_2O/Na_2O	90.9	1.43	6.49	1.95	0.97	1.00	1.14	1.05
C.I.P.W. Norms[3]								
Q	26.18	36.32	40.43	52.61	47.49	34.00	34.07	34.08
C		1.67	2.71	4.38	3.28	0.48	0.82	0.67
or	71.51	27.86	44.64	21.66	18.08	25.55	27.08	26.21
ab	0.78	27.81	9.87	15.90	26.60	36.67	33.91	35.64
an		3.52	0.39	2.68	2.22	0.50	0.67	0.45
hy	0.89	2.09	1.41	2.22	1.77	1.97	2.45	2.11
mt	0.05	0.49	0.34	0.42	0.35	0.54	0.65	0.55
il	0.13	0.25	0.17	0.14	0.16	0.29	0.31	0.29
ap			0.02		0.05		0.05	
100 *an*/*ab*+*an*		11.2	3.8	14.4	7.7	1.3	1.9	1.3

Sample	Taggarts Mountain Ignimbrite Member +									
	9	10	11	12	13	14	15	16	17	18
Oxide	R55241	R55240	R55244	R55245	R55242	R55213	R55221	R55214	R55218	R55198
SiO_2	70.28	70.79	72.45	73.11	73.71	74.07	75.25	75.71	76.35	76.17
TiO_2	0.21	0.19	0.12	0.13	0.09	0.26	0.21	0.17	0.14	0.16
Al_2O_3	14.01	13.97	11.86	12.07	12.11	13.53	12.59	12.85	12.59	13.22
Fe_2O_3	0.96	1.11	0.79	0.72	0.77	1.83	1.20	1.12	1.05	1.17
FeO	0.58	0.44	0.12	0.22	0.09	0.21	0.15	0.09	0.07	0.09
MnO	0.05	0.07	0.05	0.03	0.04	0.09	0.03	0.02	0.07	n.d.
MgO	0.36	0.35	0.15	0.18	0.12	0.47	0.28	0.17	0.12	0.30
CaO	1.99	1.89	1.39	1.06	1.34	1.38	0.68	0.61	0.50	0.99
Na_2O	3.77	3.92	3.71	4.20	3.41	4.02	2.76	2.71	3.13	3.59
K_2O	3.38	3.25	1.99	1.93	2.61	3.59	4.96	5.24	5.10	4.27
P_2O_5	0.05	0.04	0.02	0.02	0.01	0.04	0.04	0.04	0.02	0.02
H_2O^+	3.25	3.31	5.54	5.51	4.80	0.62	0.67	0.49	0.36	0.38
H_2O^-	0.82	0.75	1.13	0.98	0.86	0.40	0.44	0.31	0.31	0.33
CO_2	n.d.	n.d.	n.d.	n.d.	n.d.	0.24	n.d.	n.d.	n.d.	n.d.
Total	99.71	100.08	99.32	100.16	99.96	100.75	99.26	99.53	99.81	100.69
ΣFe as FeO	1.44	1.44	0.83	0.87	0.78	1.86	1.23	1.10	1.01	1.14
K_2O/Na_2O	0.90	0.83	0.54	0.46	0.77	0.89	1.80	1.93	1.63	1.19
C.I.P.W. Norms[3]										
Q	31.44	31.64	42.86	41.01	42.93	32.68	38.64	38.46	37.25	35.95
C	0.68	0.69	1.21	1.27	1.34	0.62	1.57	1.73	1.07	0.94
or	20.90	20.02	12.70	12.18	16.37	21.35	29.89	31.39	30.42	25.26
ab	33.38	34.57	33.90	37.96	30.62	34.24	23.82	23.25	26.74	30.41
an	9.99	9.50	7.31	5.48	6.98	6.63	3.17	2.80	2.37	4.79
hy	2.56	2.59	1.41	1.47	1.28	3.25	1.98	1.57	1.48	1.92
mt	0.52	0.51	0.31	0.32	0.28	0.64	0.43	0.38	0.35	0.39
il	0.42	0.38	0.25	0.26	0.18	0.50	0.41	0.33	0.27	0.30
ap	0.12	0.10	0.05	0.05	0.02	0.09	0.09	0.09	0.05	0.05
100 *an*/*ab*+*an*	23.0	21.6	17.7	12.6	18.6	16.2	11.8	10.8	8.1	13.6

Sample	Piallaway Trig Ignimbrite Member X									Other ignimbrites □	
	19	20	21	22	23	24	25	26	27	28	29
Oxide	R55289	R55287	R55288	R55278	R55274	R55271	R55269	R55283	R55273	R55303	R55304
SiO_2	68.77	68.89	68.98	71.90	72.75	73.01	73.22	73.43	74.30	68.47	71.92
TiO_2	0.37	0.36	0.36	0.31	0.33	0.29	0.30	0.30	0.29	0.43	0.26
Al_2O_3	14.00	14.19	13.97	14.22	14.75	14.50	14.15	14.42	14.17	13.41	13.23
Fe_2O_3	1.63	1.63	1.68	1.34	1.73	1.35	1.45	1.27	1.40	1.05	0.91
FeO	0.32	0.29	0.32	0.08	0.09	0.08	0.17	0.11	0.08	0.48	0.21
MnO	0.04	0.06	0.04	0.05	0.06	0.05	0.05	0.05	0.04	0.05	0.01
MgO	0.52	0.48	0.53	0.37	0.33	0.33	0.32	0.25	0.18	0.31	0.31
CaO	2.17	2.15	2.19	1.81	1.57	1.32	1.59	0.92	1.40	3.62	1.79
Na_2O	3.92	4.09	4.29	3.66	4.19	3.98	4.00	3.42	3.85	3.39	3.63
K_2O	2.47	2.17	1.95	3.82	3.71	3.99	3.93	4.68	3.99	1.04	3.04
P_2O_5	0.10	0.09	0.11	0.06	0.05	0.05	0.04	0.04	0.03	0.06	0.04
H_2O^+	4.30	3.99	4.19	0.82	0.52	0.68	0.41	0.87	0.49	6.93	3.01
H_2O^-	0.87	0.85	0.83	0.79	0.49	0.67	0.39	0.38	0.42	1.04	1.08
CO_2	n.d.	n.d.	n.d.	0.52	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	99.48	99.24	99.44	99.75	100.57	100.30	100.02	100.14	100.64	100.28	99.45
ΣFe as FeO	1.79	1.76	1.83	1.29	1.65	1.29	1.47	1.25	1.34	1.42	1.02
K_2O/Na_2O	0.63	0.53	0.45	1.04	0.89	1.00	0.98	1.37	1.04	0.31	0.84
C.I.P.W. Norms[3]											
Q	32.49	32.83	32.44	31.99	29.90	31.21	30.76	33.27	33.00	39.24	36.27
C	1.24	1.51	1.15	0.94	1.11	1.37	0.53	2.18	1.05	0.29	0.85
or	15.50	13.60	12.22	23.15	22.05	23.85	23.43	27.99	23.67	6.66	18.85
ab	35.22	36.71	38.50	31.76	35.66	34.07	34.15	29.29	32.70	31.10	32.23
an	10.74	10.69	10.76	8.81	7.51	6.29	7.70	4.36	6.77	19.04	9.04
hy	3.18	3.09	3.28	2.17	2.47	2.08	2.25	1.81	1.73	2.10	1.77
mt	0.65	0.64	0.67	0.45	0.57	0.45	0.51	0.43	0.46	0.53	0.37
il	0.75	0.73	0.73	0.60	0.63	0.56	0.57	0.58	0.55	0.89	0.52
ap	0.25	0.22	0.27	0.14	0.12	0.12	0.09	0.09	0.07	0.15	0.10
100 *an*/*ab*+*an*	23.4	22.6	21.8	21.7	17.4	15.6	18.4	12.9	17.2	38.0	21.9

[1]Samples for analyses 9,10,11,12,13,19,20,21,28,29 are all black vitrophyric representatives of the ignimbrite members.

[2]Analysis 1, fine ash tuff; analyses 2,3,4,5, ignimbrite.

[3]Norms have been calculated with Fe_2O_3/FeO = 0.3. Analysis 1 has 0.16% *di* and 0.29% *ac*.

[4]n.d. = not detected.

CHAPTER 5

PERMO-CARBONIFEROUS SILICIC VOLCANISM AND PALAEOGEOGRAPHY ON THE WESTERN EDGE OF THE NEW ENGLAND OROGEN, NORTHEASTERN NEW SOUTH WALES

INTRODUCTION

The southern part of the New England Orogen (Day *et al.*, 1978) in northeastern New South Wales is bordered to the west and south by Permian and Mesozoic strata of the Sydney-Bowen and Great Artesian (Australian) Basins (Bembrick *et al.*, 1980). Along much of the western margin of the Orogen meridionally folded Late Carboniferous formations of the Tamworth Belt (Korsch, 1977) are separated by the Mooki Thrust from near horizontal Permian and younger formations farther west. The character of the transition from the Orogen to the Basin as revealed by the Late Carboniferous to Early Permian succession, and correlation of sequences separated by the Thrust, have been difficult to establish, especially since outcrop on its western side is mediocre and age control is imprecise. However, there are some sites where remnants of the Early Permian sequence are preserved east of the Thrust and are well exposed on the western side. This chapter includes descriptions of three such localities between the Liverpool Range in the south and the Nandewar Range in the north (Fig. 5.1), and a reconstruction of the latest Carboniferous-earliest Permian palaeogeography.

At each of two sites east of the Mooki Thrust, the earliest Permian strata are volcanogenic fluviatile sandstones which conformably overlie grossly similar facies of Late Carboniferous age. The Late Carboniferous part also includes undisturbed primary volcanic units which have a westerly provenance, the same as the sedimentary layers, but no surface exposures of source volcanic centres of established Carboniferous age are known. Silicic and intermediate proximal (near source) volcanics exposed west of the Thrust at Boggabri are in part Early Permian, and may be a record of the final activity of the volcanic terrain which had dominated the depositional history of the adjacent part of the Tamworth Belt throughout the Carboniferous. After extinction, younger Permian epiclastic facies gradually encroached on and eventually covered the eroded relics of the western volcanic centres.

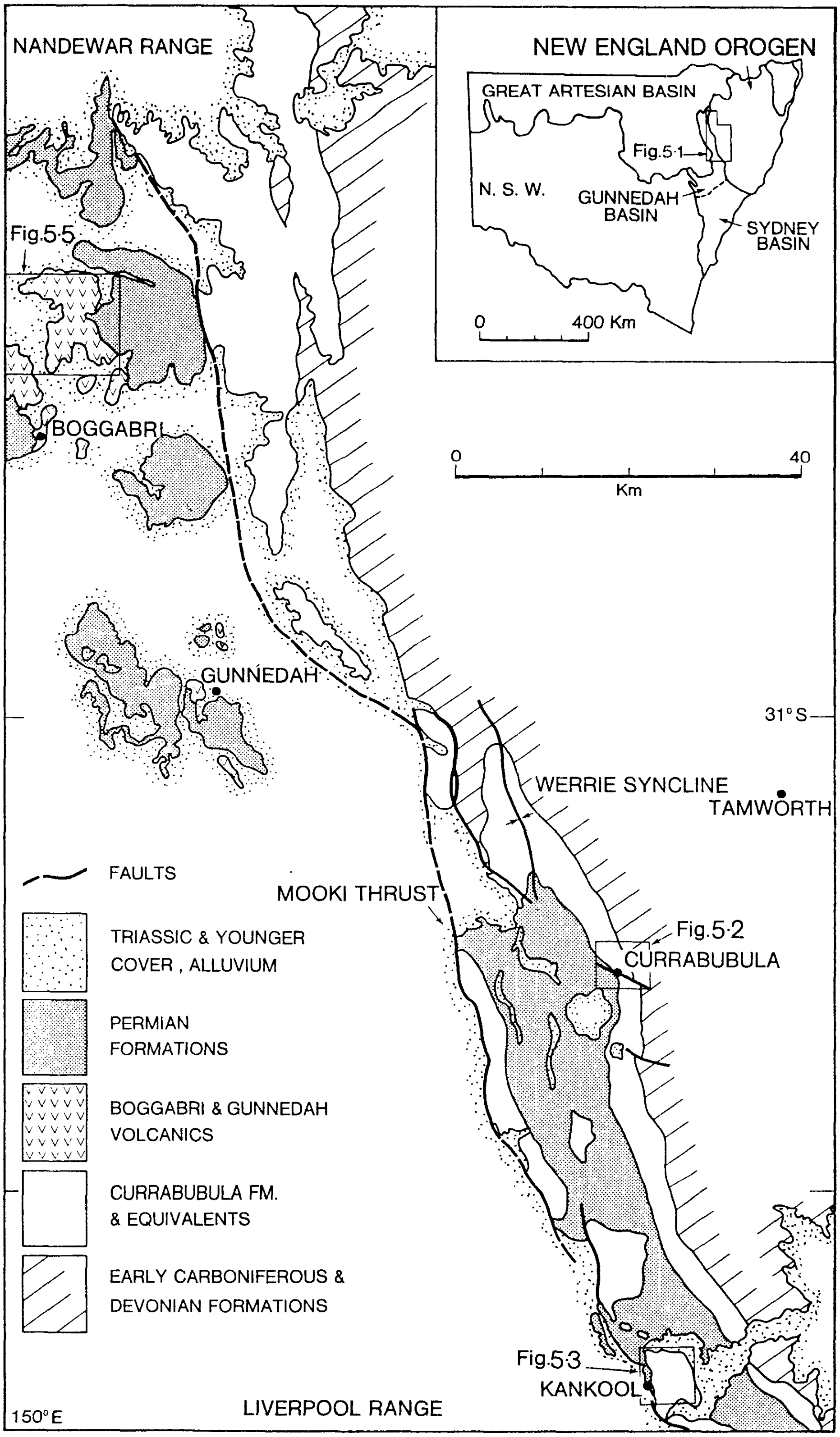


Figure 5.1: Regional geological setting of Permian and Carboniferous formations along the western margin of the New England Orogen in New South Wales.

GEOLOGICAL SETTING

The Tamworth Belt (Korsch, 1977) consists of folded and mildly metamorphosed early Palaeozoic to Early Permian volcanogenic sedimentary rocks (Leitch, 1974) which reflect a protracted regression culminating in continental and paralic sedimentation during the Late Carboniferous and Early Permian. A western provenance prevailed at least from the Early Devonian, and only disconformities of local extent interrupt the succession (Leitch, 1974; Mory, 1982).

The Late Carboniferous Currabubula Formation (Carey, 1934,1937; Voisey and Williams, 1964) and equivalents north of the Liverpool Range include widespread silicic outflow ignimbrite sheets (McPhie, 1983; Chapter 2), interleaved with fluviatile conglomerates and sandstones, and less abundant paraconglomerates and pebbly mudstones of glacial origin (Whetten, 1965), generally amounting to about 2000 m in aggregate thickness. The western source of the ignimbrites and volcaniclastics probably bordered an elevated region of Late Devonian and older Lachlan Fold Belt rocks (Carey and Browne, 1938; Chapters 2,4).

East of the Mooki Thrust at Currabubula and Kankool (Fig. 5.1, Table 5.1) the Currabubula Formation is overlain by fluviatile sedimentary strata (Temi Formation) and mafic lava-dominated volcanics (Werrie Basalt) of Early Permian age. Younger Permian rocks at Werris Creek are coal measures correlated with the Greta Coal Measures of the Hunter Valley (Loughnan, 1975). West of the Mooki Thrust at Boggabri, coal measures (Leard and Maules Creek Formations; Brownlow, 1981) of similar age overlie silicic and intermediate volcanic rocks (Table 5.1). Permian plant fossils have been identified from mudstones near the top of the volcanics but no fossils have been recovered from the underlying section of ignimbrites and lavas. The Permian units are part of the Gunnedah Basin, a subdivision of the Sydney-Bowen Basin (Bembrick *et al.*, 1980). Although outcrop of Permian strata west of the Mooki Thrust is limited, subsurface data are available from drill core recovered for coal exploration in the Gunnedah Basin (e.g. Brownlow, 1981; Beckett *et al.*, 1983).

PERMO-CARBONIFEROUS RELATIONSHIPS IN THE WERRIE SYNCLINE AT CURRABUBULA

The Early Permian Temi Formation and the overlying Werrie Basalt

TABLE 5.1: Stratigraphic nomenclature relevant to the three areas close to the Mooki Thrust which have been studied in detail.

Age \ Locality	East of the Mooki Thrust KANKOOL	CURRABUBULA	West of the Mooki Thrust BOGGABRI
Greta Coal Measures equivalents	Willow Tree Formation	Werris Creek Formation	Maules Creek Formation Leard Formation
	Werrie Basalt	Werrie Basalt	
EARLY PERMIAN (*Glossopteris*-bearing)	Temi Formation	Temi Formation	Boggabri Volcanics
LATE CARBONIFEROUS (*Rhacopteris*-bearing)	Currabubula Formation equivalent	Currabubula Formation	

occupy the core of the Werrie Syncline and extend along its east limb as far south as the Liverpool Range (Fig. 5.1). These units overlie the Late Carboniferous Currabubula Formation which forms a prominent range of strike-parallel ridges.

Previous work

Published descriptions of the Permo-Carboniferous relationships at Currabubula on the east limb of the Werrie Syncline differ: Carey (1934, 1935,1937) disputed a fault proposed by Benson (1920) and recognised a Permo-Carboniferous hiatus which Whetten (1965) interpreted to be a major disconformity.

Carey (1935) reported a thin Permian clastic sequence beneath the Werrie Basalt consisting mainly of fluviatile coarse sandstones and conglomerates containing minor lenses of coal. The relationship of the Permian sedimentary rocks with the underlying Carboniferous "Upper Kuttung Series" was stated to be "... an important non-sequence without angular divergence, .." (Carey, 1937, p.361). Carey's description of the contact in the Woodlands section, 5 km south of Currabubula, reflects the subtlety of the distinction between the Permian and Carboniferous clastic sequences: "These are the highest beds of the Kuttung or the lowest of the Greta Series, ..." (1937, p.345).

Whetten (1965) following Carey, showed locally mappable units in the Currabubula Formation to be continuous across Currabubula Creek. The juxtaposition of the Currabubula Formation north of Currabubula Creek with Permian strata along strike to the south, and the difference in thickness of the Formation either side of Currabubula Creek were explained by means of elaboration of Carey's suggested "non-sequence" at the top of the Carboniferous. Whetten (1965, p.52) maintained that "The contact between the Permian and Carboniferous rocks is a disconformity with about 3000 feet of relief..". Not surprisingly, several workers concerned with the regional tectonic history of the New England Orogen have made reference to, and in some cases, endeavoured to account for, this substantial disconformity (Leitch, 1969, p.31; 1974, p.143; Runnegar, 1970, p.700; 1974, p.15; Herbert, 1980, p.279; Korsch and Harrington, 1981, p.210, p.233).

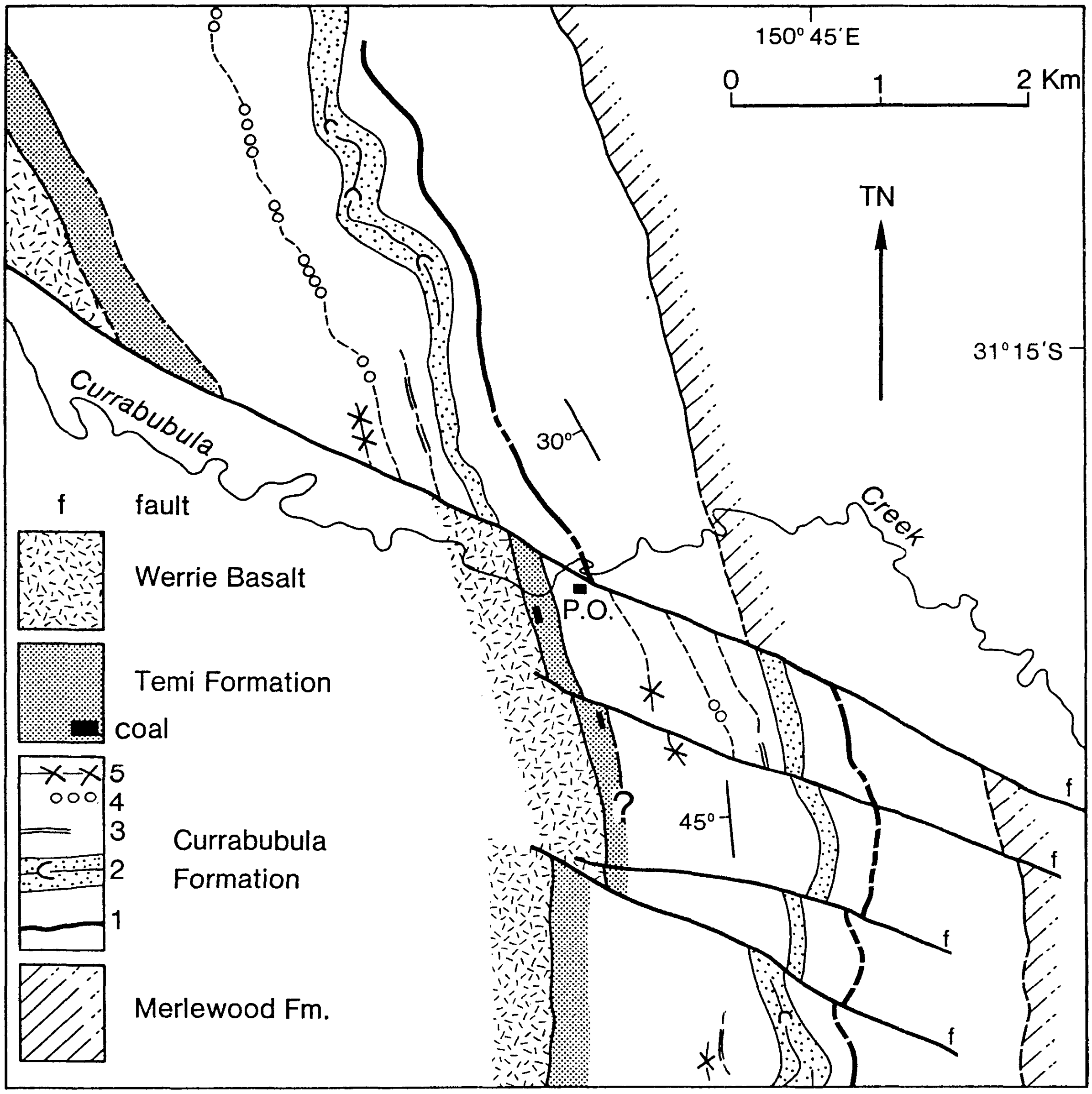


Figure 5.2: Geological map of the Permo-Carboniferous sequence on the east limb of the Werrie Syncline at Currabubula.

(1) Rosedale Member

(2) Cana Creek Tuff Member

(3) mudstone
(4) conglomerate } locally mappable units
(5) tuff

Ignimbrites and disconformities of the Currabubula Formation

Although the Currabubula Formation is predominantly conglomeratic, interbedded silicic ignimbrites of regional extent define an internal pyroclastic stratigraphy, mappable throughout the Werrie Syncline and nearby fault blocks (Chapter 2). The distribution of four ignimbrite members established by outcrop mapping at 1:31680 scale, together with more detailed mapping at Currabubula, have helped clarify the local Permo-Carboniferous relationships.

The member most widespread in the Werrie Syncline is the Cana Creek Tuff (Chapters 2,3). This member cannot be traced directly along strike from where it outcrops north of Currabubula Creek to its position south of the creek (Fig. 5.2). The base of the Currabubula Formation and locally mappable horizons within it show the same disparity in position across Currabubula Creek, and the attitudes of bedding also differ on either side of the Creek. This evidence, in combination with the obvious break in, and displacement of, the strike-parallel ridges at Currabubula strongly supports the existence of a steep, east-southeast trending fault passing through the township, as originally suggested by Benson (1920).

Because the Permian-Carboniferous contact is offset across Currabubula Creek by a fault, the scale of the pre-Permian disconformity was markedly overestimated by Whetten (1965). Furthermore, the distribution of ignimbrites indicates that erosional surfaces are common throughout the Currabubula Formation. One of the best examples is up to 4 km across and is incised through a stratigraphic thickness of several hundred metres (Chapter 2). The relief of Carey's (1937) "non-sequence" at the Permian-Carboniferous boundary is dwarfed by the scale of disconformities within the Late Carboniferous section. The similarity of facies across the Permian-Carboniferous boundary in the Currabubula district renders it harder to detect and no more mappable than the several other disconformity surfaces occurring stratigraphically below it.

Local onlap

Along the east limb of the Werrie Syncline from Currabubula south to the Liverpool Range, the Temi Formation is exposed above the Currabubula

Formation at several localities and is probably continuous as mapped by Carey (1934). Equivalent Permian sedimentary rocks outcrop below the Werrie Basalt on the north side of the fault at Currabubula (Fig. 5.2), but their extent farther northward along the east limb has not been determined. However, in the closure of the Werrie Syncline, the Currabubula Formation is evidently overlain directly by the Werrie Basalt. Although conformable, the Permian units locally onlap northwestward across the underlying Carboniferous sequence.

PERMO-CARBONIFEROUS RELATIONSHIPS AT KANKOOL

East of Kankool on the northern fall of the Liverpool Range (Figs. 5.1,5.3), Late Carboniferous conglomerates and volcanic units equivalent to the Currabubula Formation form a broad anticline plunging gently to the north. The Temi Formation is preserved on the northern part of the west limb beneath the Werrie Basalt but relationships on the east limb are partly hidden by Tertiary basalt and have not been examined. Farther east again, the Temi Formation is present above the Late Carboniferous of the eastern limb of the Werrie Syncline.

The Late Carboniferous sequence at Kankool is comparable with that of the Werrie Syncline but the base is not exposed, and the internal stratigraphy of volcanic units is different and distinctive. Over most of the district three volcanic units are mappable (Fig. 5.3), the youngest being a flow banded silicic lava at least 100 m thick. In contrast, the southwestern quadrant includes several thin (10 to 15 m) ignimbrites of limited extent, the vertical and lateral relationships of which are poorly known in spite of detailed mapping. Though some of this complexity may be primary, bedding data from this sector and the lack of correspondence with the rest of the area possibly indicate disruption by faults. Furthermore, dips of bedding on the west limb of the anticline steepen markedly in all units in westernmost exposures. The anomalous internal stratigraphy of the Late Carboniferous sequence, westward steepening dips, and probably also the apparent southward wedging of the Permian strata may all be a consequence of proximity to a southern extension of the Mooki Thrust, the precise location of which is obscured by alluvium and Tertiary basalt.

Lowe (1971) attributed these relationships to an angular unconformity

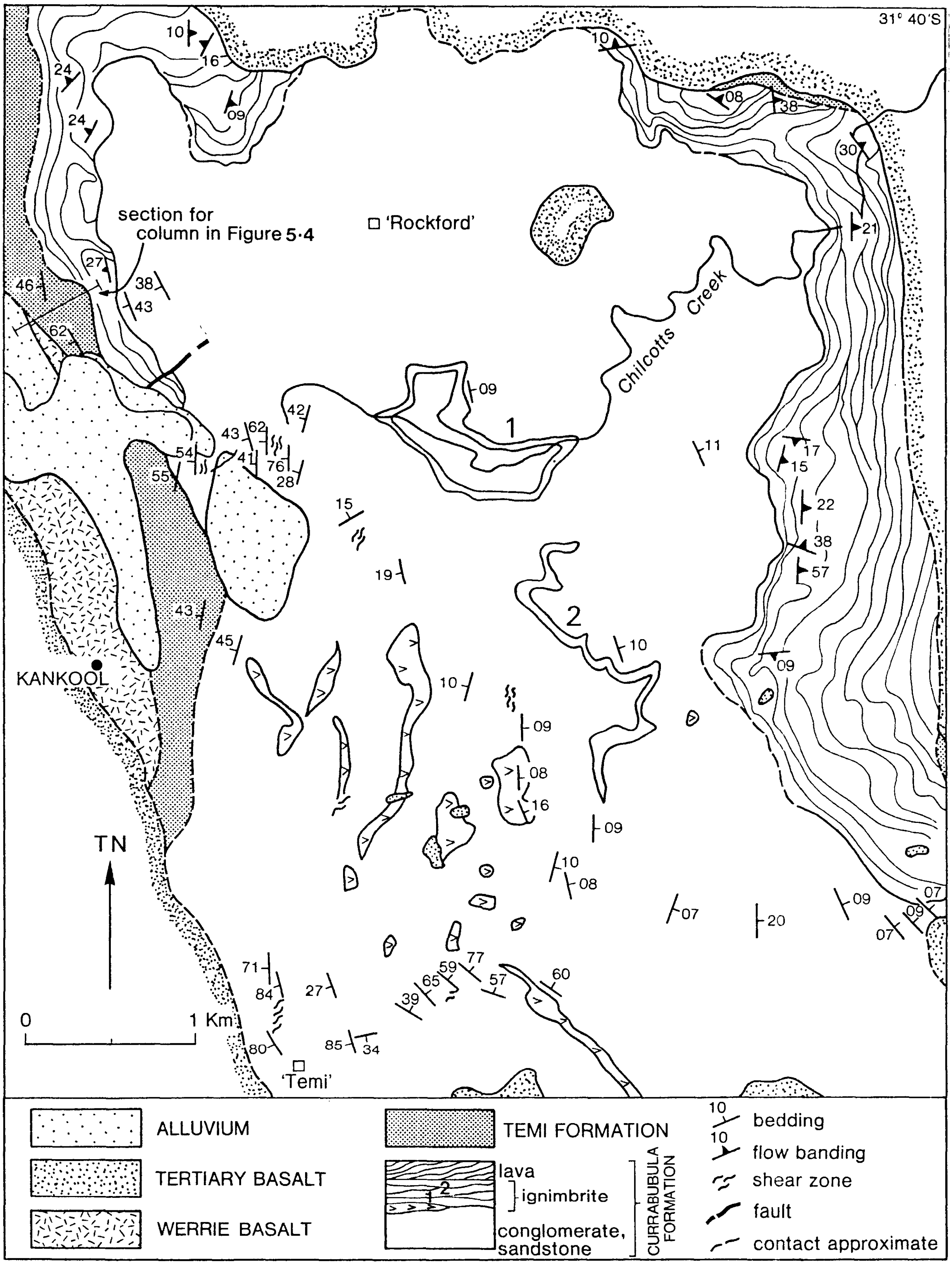


Figure 5.3: Geological map of the Permo-Carboniferous sequence immediately north of the Liverpool Range at Kankool.

and disconformity involving a few hundred metres of relief at the Permian-Carboniferous boundary. This interpretation has been adopted by later workers (e.g. Leitch, 1974, p.143; McClung, 1980, p.65; Korsch and Harrington, 1981, p.210), but is at variance with the conformable, or at most, slightly disconformable relationships reported immediately to the north at Willow Tree (Hanlon, 1947a,b; Runnegar, 1970) and to the south at Murrurundi (Hanlon, 1947c; Warner, 1972; but see also Manser, 1968). The oldest Permian rocks at Kankool may lie on an erosional surface, the significance of which must be evaluated in the light of the common occurrence of disconformities in the underlying Carboniferous sequence (Chapter 2) and their probable existence within the overlying fluviatile Temi Formation.

"Flint clay" sedimentary rocks in the Temi Formation

"Flint clay" sedimentary rocks are widespread at the base of the Early Permian (Greta equivalent) coal measures from Wingen 20 km south of the Liverpool Range, northward to Boggabri (Loughnan, 1973,1975; Brownlow, 1981). Loughnan (1973,1975) has suggested that the "flint clay" clasts were derived from the erosion of an ancient soil developed on the older and in places underlying, deeply weathered Werrie Basalt. This interpretation in part depended on the presence of clasts with relic textures typical of mafic volcanic rocktypes (Loughnan, 1973, Fig. 4, p.333; 1975, p.247,249) and the evidently consistent occurrence of the "flint clay" intervals above the Werrie Basalt.

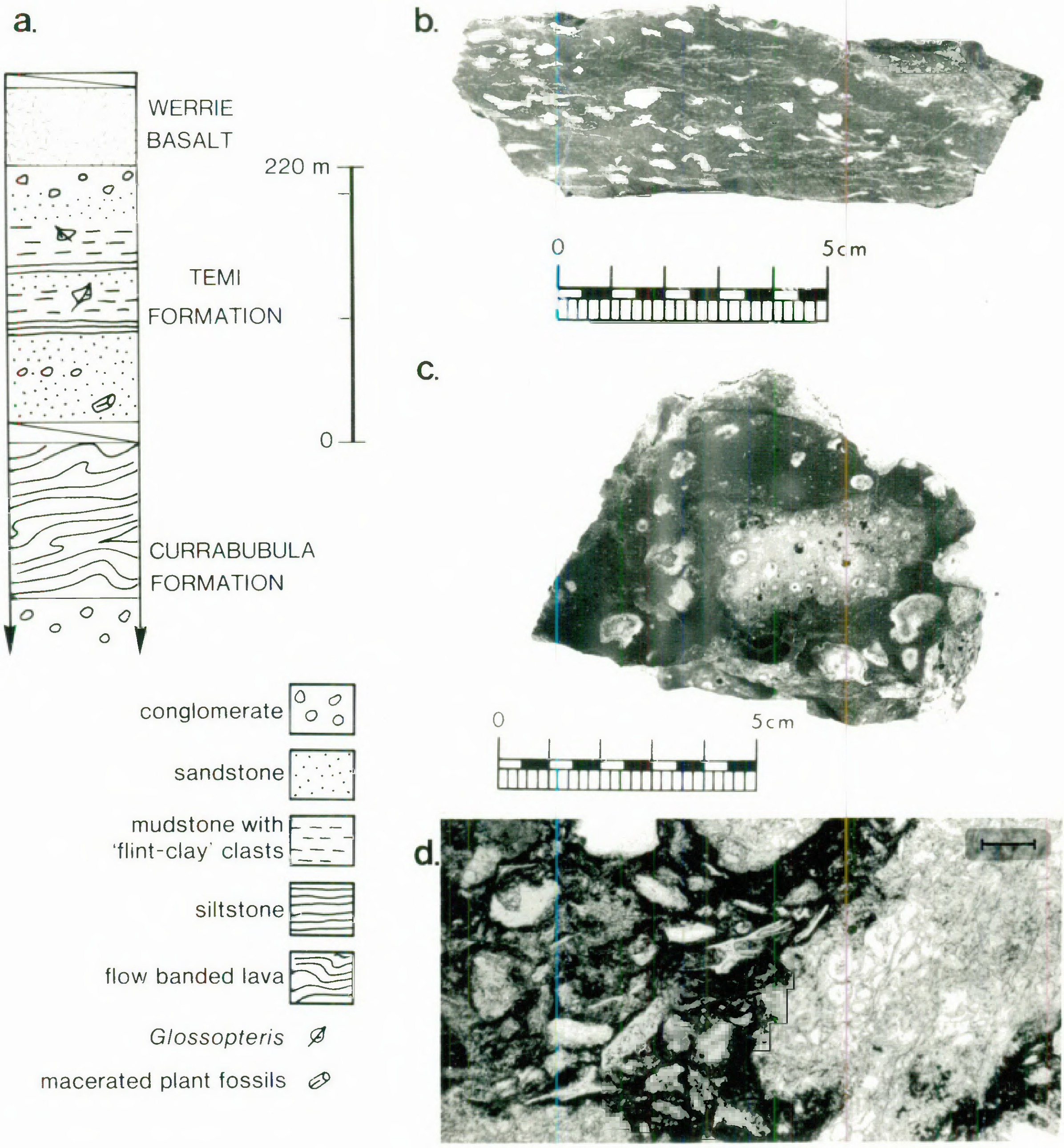
In the northwestern sector of the Kankool area the Late Carboniferous sequence is overlain by 220 m of mudstone, sandstone and pebble conglomerate of the Temi Formation (Fig. 5.4a). *Glossopteris* fossils are abundant a few tens of metres stratigraphically above the silicic lava at the top of the Late Carboniferous. The Temi Formation is overlain in places by Werrie Basalt and elsewhere outcrop is hidden beneath alluvium. Approximately 90 m above the base of the Temi Formation, "flint clay" clast conglomerate (Fig. 5.4b), in places fossiliferous, occurs in a 20 m thick interval with mudstone. At a similar level, clasts of black vesicular mafic volcanic rocktypes are locally conspicuous in the conglomerates (Fig. 5.4c). This interval is entirely within the Temi Formation and well

Figure 5.4: a. Stratigraphic column for the Temi Formation based on the section line shown on Figure 5.3.

b. "Flint clay" conglomerate from the Temi Formation at Kankool. R55334 collected from GR764811 Quirindi-D.

c. Clasts of vesicular basalt in conglomerate of the Temi Formation at Kankool. R55335 collected from GR771798 Quirindi-D.

d. Relic pumice and shards in the "flint clay" interval of the Temi Formation at Kankool. Plane polarised light; bar length approximately 0.1 mm. R55336 collected from GR764812 Quirindi-D.


a.
WERRIE
BASALT
220 m
TEMI
FORMATION
0
CURRABUBULA
FORMATION
conglomerate
sandstone
mudstone with
'flint-clay' clasts
siltstone
flow banded lava
Glossopteris
macerated plant fossils
b.
0
5cm
c.
0
5cm
d.

below the Werrie Basalt. Thus the occurrence of "flint clay" and vesicular mafic volcanic clasts in the Temi Formation at Kankool implies that no close genetic relationship need exist with the Werrie Basalt, and furthermore that these "flint clay" intervals may not be synchronous on a basin-wide scale (*cf.* Loughnan, 1975; Brownlow, 1981; Beckett *et al.*, 1983).

Provenance of the Temi Formation

Lithic detritus in sandstone and conglomerate of the Temi Formation is predominantly volcanic (Whetten, 1965; Lowe, 1971; this study). In addition, shreds of non-porphyritic pumice and shards have been identified in the "flint clay" conglomerates (Fig. 5.4d), providing evidence that pyroclastic volcanism accompanied deposition of the Temi Formation. The pyroclastic components were probably produced by the final eruptions of silicic volcanic centres located to the west near others which had been repeatedly active during the Late Carboniferous. Other rounded "flint clay" clasts that lack a clearly vesicular structure, and the volcanogenic detritus in sandstone and conglomerate of the Temi Formation, may also have come from the western source, resulting from weathering and denudation of the volcanic pile as eruptions became infrequent and finally ceased.

Thus, the evidence suggests that the Temi Formation "flint clay" horizons at Kankool formed while the western volcanic source was still contributing volcaniclastic sediment and minor juvenile pyroclastic debris eastward. The provenance and depositional setting which prevailed during the Late Carboniferous appear to have persisted uninterrupted into the Early Permian at this locality.

RELATIONSHIPS AT BOGGABRI AND GUNNEDAH

Coal exploration drill core data (Russell and Middleton, 1981; Tadros, 1982; Beckett *et al.*, 1983) show that the Permian coal measures of the Gunnedah Basin (Bembrick *et al.*, 1980) overlie undeformed volcanics (Table 5.1). The basement volcanics are exposed north of Boggabri and at Gunnedah (Fig. 5.1) on the crest of the "Boggabri Ridge" (Russell, in Tadros, 1982), a near-meridional, 60 km long basement high. Both these localities are west of the Mooki Thrust and bedding varies from near horizontal to gently dipping.

The Boggabri Volcanics

About 600 m of dominantly silicic volcanic rocks are exposed north and west of Boggabri (Fig. 5.5). Principal components are flow banded lavas and ignimbrites (Fig. 5.6a). At least three distinct lava units, separated by ignimbrites, are present in sections near the Namoi River. Lavas are extensive farther east in the Leard State Forest but because no ignimbrites occur, correlations of these flows with the lavas of the measured sections are uncertain.

The lavas have well developed flow banding though flow folds (Fig. 5.6b) and autobrecciated zones are restricted to the two uppermost units. The irregular occurrence of thin (2 to 3 m) black glassy "vitrophyre" and benches in ridges composed of the youngest lava and of the Leard State Forest lavas, suggest the presence of several stacked flow units. Dark grey porphyritic (feldspar + relic pyroxene) lavas(?) less than a few metres in thickness are a minor component of the sequence, and have not been differentiated on either the map (Fig. 5.5) or the stratigraphic column (Fig. 5.6a) from the common more silicic flows. Examples outcrop at the base of Robertsons Mount (GR014020 Boggabri) and of Gins Leap (GR993049 Boggabri).

Polygonal bedding-normal jointing is displayed by the central non-brecciated part of the youngest lava unit (GR014107 Therribri) and at the top of the oldest unit (GR006076 Boggabri). The structural significance of sporadically developed horizontal polygonal joints (e.g. GR017108 Therribri) is not always evident. However in at least one instance (GR988080 Boggabri; Fig. 5.7) this feature is accompanied by concentric strikes of flow foliations in the lava, pervasively (hydrothermally?) altered lava breccia, and a thin interval of bedded volcaniclastics. The structure is interpreted to be a lava dome, probably occupying its source eruptive vent, being similar in lithological association, arrangement, scale and relief to Cainozoic examples of dome complexes (e.g. Loney, 1968; Smith, 1973; Fink, 1983; Fink and Pollard, 1983). Some of the several other isolated hills in the Boggabri district made of flow banded silicic lava may also prove to be lava domes in view of the common clustering of such small-scale silicic volcanic centres in modern settings (e.g. Mono Craters, California, Loney 1968; Smith, 1973; Medicine Lake,

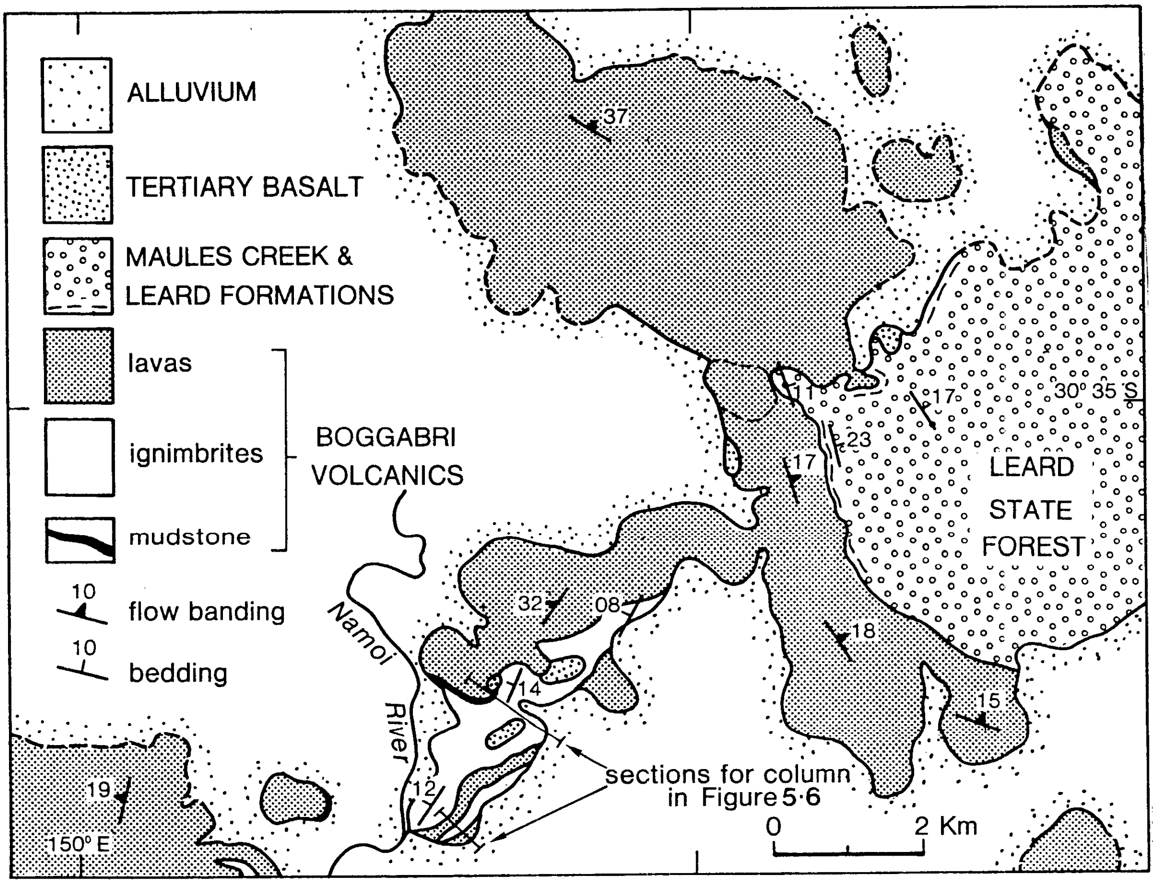


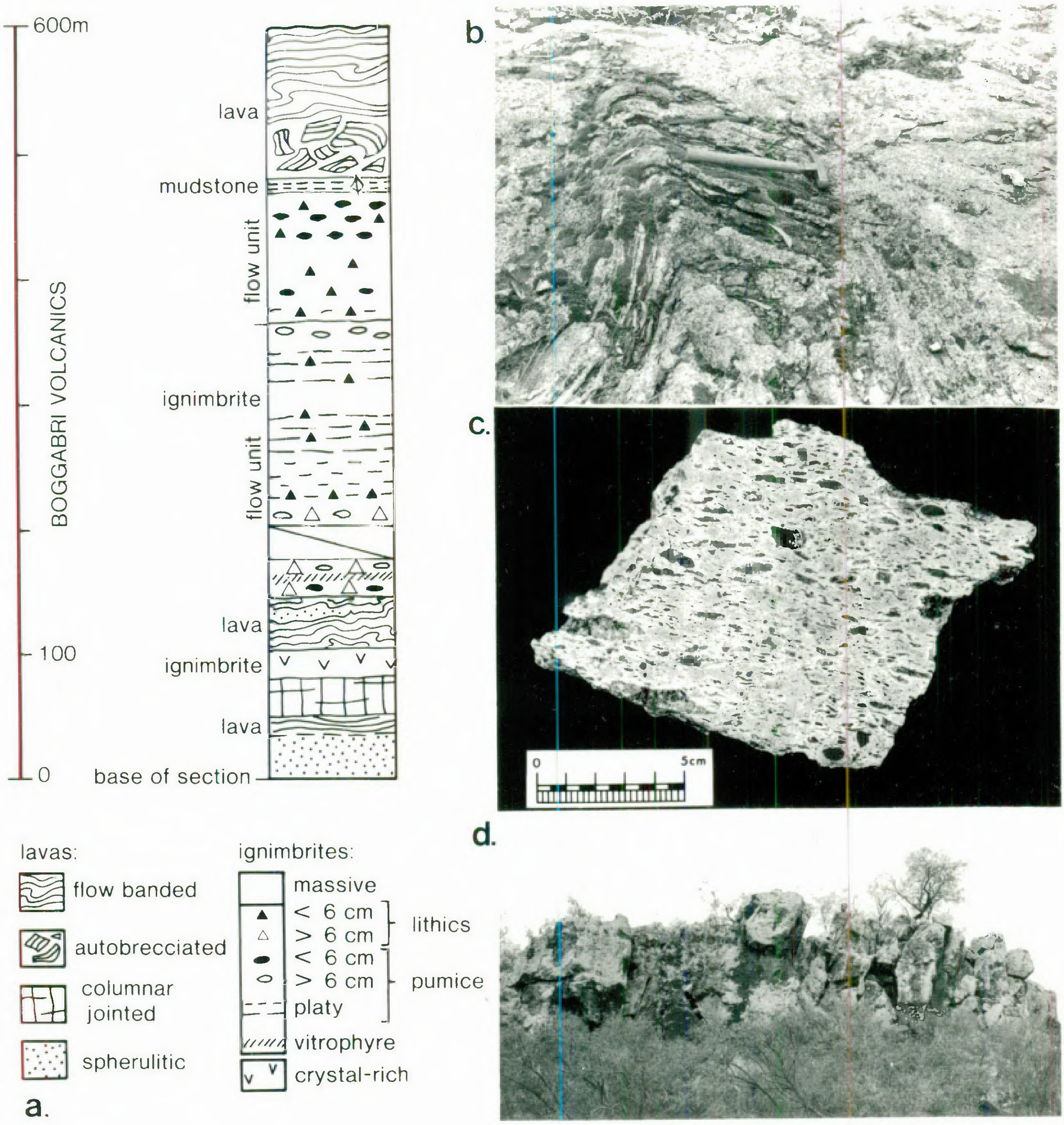
Figure 5.5: Geological map of the Boggabri Volcanics and Permian formations north of Boggabri.

Figure 5.6: a. Stratigraphic column for the Boggabri Volcanics based on the section line shown in Figure 5.5.

b. Flow banded silicic lava of the Boggabri Volcanics (GR971056 Boggabri). Hammer 33 cm.

c. Ignimbrite of the Boggabri Volcanics. R55342 collected from GR028105 Therribri.

d. Columnar-jointed lithic-rich ignimbrite of the Gunnedah Volcanics. View to the southeast, GR226745 Gunnedah.


600m
100
0
BOGGABRI VOLCANICS
lava
mudstone
flow unit
ignimbrite
flow unit
lava
ignimbrite
lava
base of section
lavas:
flow banded
autobrecciated
columnar jointed
spherulitic
ignimbrites:
massive
< 6 cm
> 6 cm
lithics
< 6 cm
> 6 cm
pumice
platy
vitrophyre
crystal-rich
a.
b.
c.
0
5cm
d.

California, Fink and Pollard, 1983).

The younger, better exposed ignimbrite (Fig. 5.6c) is more than 150 m thick. Two flow units have been identified from a measured section (Fig. 5.6a) though there may be more. Each is composed of varying proportions of pumice, crystals (plagioclase + K-feldspar ± quartz ± altered ferromagnesian phase), lithic fragments, microscopic shards and unresolvable ash(?) matrix. Near the base of each flow unit the pumice is platy, shards are deformed and flattened, lithic fragments up to 6 cm or larger are conspicuous, and outcrops have prismatic to columnar jointing. The base of the lower unit in places is black, glassy and densely welded (e.g. GR003078 Boggabri). In uppermost parts, pumices are blocky or lensoid, shards are intact, lithic fragments are less than 3 cm across and outcrops lack regular jointing. These changes are similar to vertical textural variations in Layer 2b of single flow units of modern ignimbrites (Sparks *et al.*, 1973). The compaction/welding profile of the whole section matches the complexity of a compound cooling unit (*cf.* Smith, 1960b).

The older ignimbrite is comparatively thin (20 m) and exposed for less than 2 km along strike. It is crystal-rich, with sparse lithic and pumiceous fragments. Outcrops lack conspicuous vertical textural zonation and have widely-spaced blocky jointing. These limited observations suggest the presence of a single flow unit.

One thin (less than 10 m) epiclastic interval has been found near the top of the volcanics directly above the younger ignimbrite (Fig. 5.6a). It is a laminated tuffaceous mudstone with sparse granule- and pebble-sized "dropped" clasts and minor sandstone interbeds. Plant fossils (*Gangamopteris* and *Glossopteris*) from this unit are Early Permian and considered to be equivalent to or older than floras of the Greta Coal Measures (J. Rigby, written communication, 1983). West of the Namoi River (GR978069 Boggabri) a thin lense of accretionary lapilli tuff occurs at a similar stratigraphic level to the tuffaceous mudstone. The association in each instance with underlying ignimbrite is consistent with these intervals being co-ignimbrite ash-fall deposits (*cf.* Sparks and Walker, 1977), evidently waterlaid and perhaps reworked in the case of the mudstone.

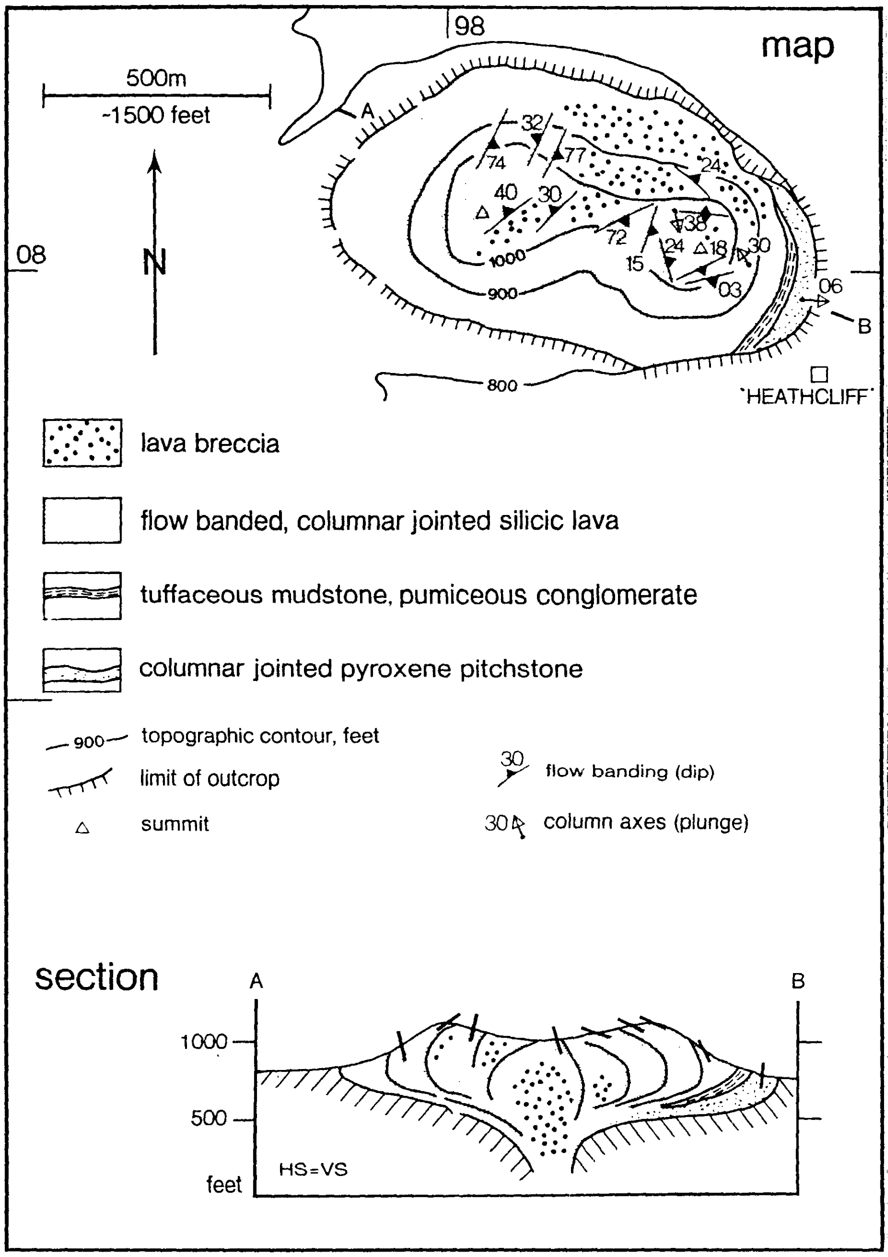


Figure 5.7: Map and interpretive cross-section of a silicic lava dome of the Boggabri Volcanics, north of Boggabri. The numbered grid on the map is the 1000 yard grid on the Boggabri 2 inch to 1 mile topographic map.

The Gunnedah Volcanics

The Gunnedah Volcanics (Manser, 1965a,b) are exposed either as remnants isolated within areas of alluvium, or beneath younger Permian strata. In both cases, poor outcrop and very gentle dips of bedding preclude clear-cut establishment of local vertical and lateral relationships. However the assemblage of volcanic lithologies is identical with that at Boggabri: lithic-rich poorly welded, columnar jointed ignimbrites and flow banded lavas. Residuals of a columnar-jointed ignimbrite sheet form hills north of Gunnedah (Fig. 5.6d; GR227745 and GR223756 Gunnedah). The sheet has a gentle southwesterly dip, implying that it underlies the lavas occurring farther south at and beyond the township.

Hanlon (1948a) reported that *Glossopteris*-bearing sandstones and conglomerates were interbedded with the volcanics but the accompanying map, and also the maps of Manser (1965a,b), show them as a mappable unit overlying the volcanics. Two formations (Gunnible and Condadilly Formations; Manser, 1965a,b) directly overlying the volcanics are considered equivalent to the Leard and Maules Creek Formations, and also to the Greta Coal Measures (Loughnan, 1975; Brownlow, 1981).

Hanlon (1948a) described mafic igneous rocks interbedded with and intruding the more silicic volcanics at Gunnedah. Manser (1965a) proposed that these rocks were in part overlying the Gunnedah Volcanics. In spite of the uncertainties of field relationships, and poor exposures, both these authors supported correlation with the Werrie Basalt, the nearest outcrops of which are east of the Mooki Thrust in the Werrie Syncline (Carey, 1934). Correlation of mafic igneous rocks in drill core from the Gunnedah Basin with the Werrie Basalt (e.g. Loughnan, 1975; Tadros, 1982; Beckett *et al.*, 1983) perhaps should also be regarded as tentative, as younger texturally similar intrusions occur throughout this region (Carey, 1934; Ramsay and Stanley, 1976; Beckett *et al.*, 1983), and also mafic-intermediate lavas are a minor component of the Boggabri Volcanics (see above). Hence the existence and extent of the Werrie Basalt in the Gunnedah-Boggabri district have yet to be established.

Regional significance of the Boggabri and Gunnedah Volcanics

The assemblage of volcanic lithologies preserved in the Boggabri and Gunnedah Volcanics is typical of the proximal (near vent) environments of silicic volcanic terrains: thick silicic lavas, coarse lithic-rich ignimbrites, minor tuffaceous epiclastic intervals (*cf.* Wright *et al.*, 1981; Clough *et al.*, 1982; Druitt and Sparks, 1982). The sequence of eruption styles is similar to that of some modern silicic volcanic centres such as the Okataina Volcanic Centre, New Zealand (Nairn, 1982) and the Cerro Panizos ignimbrite centre, Bolivia (Baker, 1981), in which explosive ignimbrite-producing eruptions were followed by effusion of lava flows and domes prior to extinction. The Boggabri and Gunnedah Volcanics may be the remnant of a multiple-vent volcanic centre which became extinct early in the Permian (pre-Greta Coal Measures), and hence could be the proximal facies equivalent of the Temi Formation, now known to contain signs of contemporaneous, though remote, silicic volcanic activity. The depositional record from the Late Carboniferous Currabubula Formation to the Early Permian Temi Formation is essentially uninterrupted east of the Mooki Thrust. If the Boggabri Volcanics and Temi Formation are coeval, then the western volcanic terrain which had supplied ignimbrites and volcanogenic detritus eastward during the Late Carboniferous evidently still had some active centres at the time when the *Rhacopteris* flora had been replaced by *Glossopteris*, that is, in the Early Permian. Limited isotopic data on the Carboniferous sequence are consistent with this suggestion: two ignimbrites from near the top of the Currabubula Formation in the Quirindi Dome have yielded K/Ar ages of 293 ± 4 Ma (hornblende) and 302 ± 4 Ma (plagioclase; J. Roberts, written communication, 1983; Table 5.2). Problems in dating the Carboniferous-Permian boundary (e.g. Kemp *et al.*, 1977; Waterhouse, 1978; Roberts and Engel, 1980; Archbold, 1982; Murray, 1983) are by no means remarkable in view of the continuity of the rock record representing that time interval, at least in exposures near the western margin of the New England Orogen.

Local onlap

In the Leard State Forest northeast of Boggabri, lava flows of the Boggabri Volcanics are overlain by "flint clay" conglomerate and *Glossopteris*-bearing mudstone of the Early Permian Leard Formation. Bedding in the volcanics and the overlying sedimentary units is gently dipping so the

TABLE 5.2: AMDEL K-Ar analyses of minerals from ignimbrites of the Currabubula Formation.

Sample	% K	$^{40}Ar^*$ (x10 ^{10}moles/g)	$^{40}Ar^*/^{40}Ar_{Total}$	Age (x10^{6}y)
153-9 Hornblende GR 531317 Werris Creek	0.481 0.480	2.6549	0.838	293 ± 4
155-4(R55303) Plagioclase GR 581149 Quipolly	1.48 1.48	8.4430	0.921	302 ± 4

*Denotes radiogenic Argon

Constants used: $^{40}K/K = 1.167 \times 10^{-4}$ mol/mol

$$\lambda_\beta = 4.962 \times 10^{-10}\ y^{-1}$$

$$\lambda_\varepsilon = 0.581 \times 10^{-10}\ y^{-1}$$

contact is irregular (Fig. 5.5) and further complicated because the contact surface is not planar. The westernmost Leard Formation outcrops are isolated patches resting on shallowly buried lava. Farther east, the Leard Formation becomes more continuous and it, plus the Maules Creek Formation, eventually cover the lava completely. This relationship of westward onlap of the Leard Formation onto the Boggabri Volcanics is mirrored by eastward onlap on the west flank of the "Boggabri Ridge" (Tadros, 1982). Drill core data indicate that the Permian section thickens away from the "Boggabri Ridge" to the west (Tadros, 1982; Beckett *et al.*, 1983) and to the east (Runnegar, 1970; Brownlow, 1981; Tadros, 1982) where it is abruptly truncated at the Mooki Thrust (Carey and Browne, 1938).

The relief on the Boggabri Volcanics at the time of deposition of the Early Permian coal measures was in places as much as 100 m (e.g. near GR057136 Therribri). However, a significant component of this was probably constructional rather than erosional, since these subaerial silicic lavas would have been high-aspect ratio bodies (thickness compared with lateral extent; Walker, 1973b; Hulme, 1974; Walker *et al.*, 1980a) with steep flow fronts, as are modern silicic lavas (*cf.* Guest and Sanchez, 1969; Fink, 1980a,b,1983; Clough *et al.*, 1982). Weathering and erosion would have commenced immediately after extrusion, providing detritus to flanking low-lying areas so that progressively younger formations onlap onto the Early Permian volcanic edifice.

"Flint clay" conglomerates of the Leard Formation

The paucity of Werrie Basalt in the Boggabri region has proven difficult to reconcile with the suggested relationship between it and the "flint clay" conglomerates of the Leard Formation (Loughnan, 1975). Brownlow (1981) speculated that the Werrie Basalt was originally more extensive, especially to the east, and uplifted by movement on the Mooki Thrust, thereby providing the "flint clay" clasts of the Leard Formation. In addition, erosion of the Currabubula Formation at a later stage was considered to have supplied the lithic detritus of the Maules Creek Formation. However, the Mooki Thrust truncates and is younger than both the Leard and Maules Creek Formations. The Thrust could not have influenced sedimentation of these formations: rather, these units probably originally extended across the Currabubula Formation and its equivalents east of the present trace of the Mooki Thrust.

Because "flint clay" conglomerates at Boggabri rest directly on silicic(?) lava (e.g. GR066143 Therribri), support for a close relationship with the weathering of Werrie Basalt is further weakened. Clasts in the Leard Formation are massive pale pellets and irregular fragments made of clays, and bleached or relatively fresh fine feldspar porphyries some of which are essentially identical in texture to the underlying lava. Neither pumiceous clasts nor shards have been detected. A genetic link with the local palaeo-high comprising freshly emplaced lavas of the Boggabri Volcanics is implied.

There is a marked similarity between the setting of "flint clay" conglomerates in the Leard Formation and the Temi Formation at Kankool: in both cases, the "flint clay" intervals are in close proximity to an older silicic lava. However, deposition of the Temi Formation was accompanied by eruptions of the waning western volcanic source whereas the Leard Formation accumulated at a stage when the source volcanic centres, both local and remote, were quiet. Lithic clasts of conglomerates of both the Temi and Maules Creek Formations have in common a contribution from the continuous denudation of the same regionally extensive silicic volcanic pile which was constructed to the west during the Late Carboniferous, and active into the Early Permian.

SUMMARY AND DISCUSSION

The Late Carboniferous and Early Permian rocks in exposures east of the Mooki Thrust north of the Liverpool Range are locally disconformable, but no angular discordance exists. At Kankool, the present distribution of the Temi Formation is affected by younger faults but the formation is conformable with the underlying Late Carboniferous sequence, and both steepen in dip westward toward the Mooki Thrust. At Currabubula, strike separation along a minor fault has offset the Permo-Carboniferous transition, and there is no evidence for an intervening disconformity any more remarkable than those which occur throughout the underlying fluviatile volcanogenic sequence. The Early Permian proximal volcanics west of the

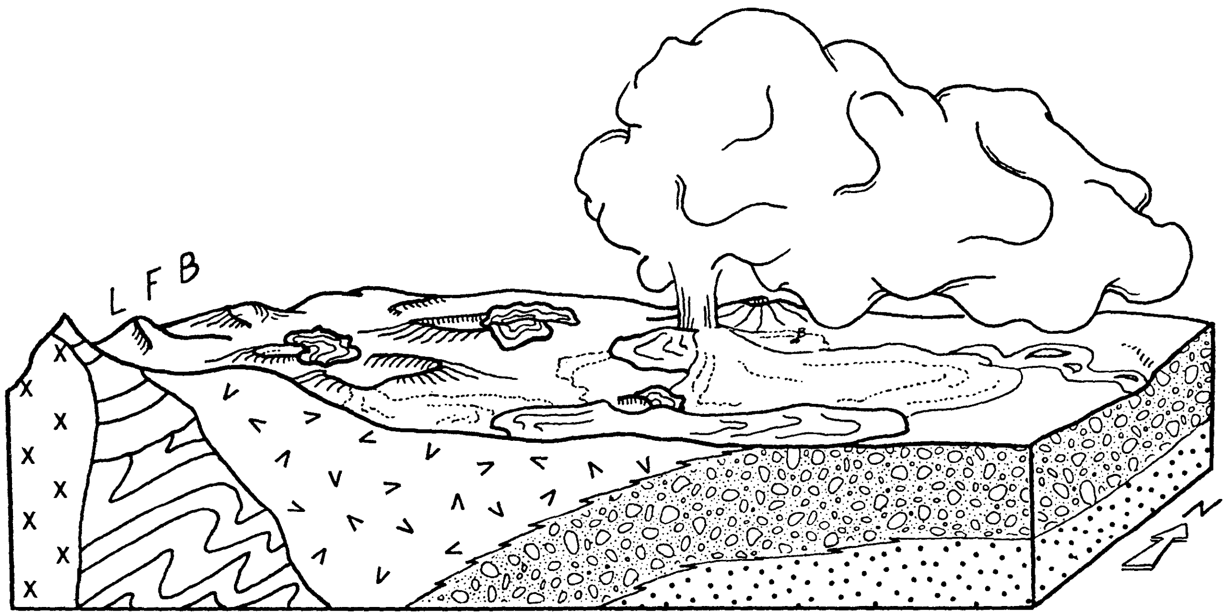


Figure 5.8: A schematic reconstruction of the palaeogeography on the western margin of the New England Orogen during the earliest Permian.
B, Boggabri.
LFB, eastern margin of the Lachlan Fold Belt.
'v' pattern, proximal silicic volcanic pile; denuded Late Carboniferous ignimbrite shields and calderas in the west; active, explosive and effusive centres to the east.
Stippled patterns, volcanogenic fluviatile sands and gravels, and distal pyroclastic deposits (Temi, Currabubula and Merlewood Formations).

Mooki Thrust at Boggabri and Gunnedah may be the youngest representatives of volcanic centres which had been continuously active during the Late Carboniferous. Original constructional relief on the volcanic pile is buried by onlapping but otherwise conformable younger Permian formations.

Late Carboniferous-Early Permian provenance and depositional environments

During the Late Carboniferous the region between the Liverpool and Nandewar Ranges was periodically inundated by silicic ignimbrites and blanketed by airfall pyroclastics delivered from a multiple-vent caldera terrain situated west of present exposures (Chapters 2 and 4). Ambient sedimentation comprised fluvial braidplain conglomerate and sandstone, in which the dominant volcanogenic detritus was diluted by varying amounts of westerly derived, older sedimentary, regional metamorphic and plutonic clasts, probably contributed from the exposed Lachlan Fold Belt (Carey and Browne, 1938; Whetten, 1965; Chapter 2). The pyroclastic influxes temporarily choked the drainage and spread across interfluves, but only a few major ignimbrites survived redistribution by contemporaneous erosion.

The dominant western provenance persisted into the Early Permian during deposition of the Temi Formation. Contemporaneous eruptions changed in style and waned in scale, though input of volcanogenic detritus from denudation of the Late Carboniferous volcanic terrain continued (Fig. 5.8). The eruptions of the Early Permian Boggabri multiple-vent volcanic centre were mainly effusive, producing silicic lavas of limited mobility in the westernmost areas of present exposures. Such activity is considered to be a continuation of the Late Carboniferous ignimbrite-forming volcanism, matching patterns of modern silicic centres closely (Sparks *et al.*, 1973; Sheridan, 1979).

The age of the deformation which produced the Mooki Thrust and the folds east of it is poorly constrained. Marine strata above the Maules Creek Formation in a drill hole adjacent and to the west of the Mooki Thrust near Boggabri, have been dated as early Late Permian (Upper Stage 4; Butel *et al.*, 1983). These rocks are truncated at the Thrust and likely to be equivalents of the Porcupine Formation reported from west of the "Boggabri Ridge" (Tadros, 1982). Thrust movement is thus no older than the early Late Permian, and may be significantly younger, because the most reliable upper limit on the age of movement is Early Jurassic. North of the Nandewar Range, Palaeozoic formations east and west of the extrapolation of the fault trace are covered by Early Jurassic strata of the Great Artesian (Australian) Basin (Exon, 1974). These constraints conflict with suggestions that "flint clay" and lithic detritus in the Leard and Maules Creek Formations was eroded from the area immediately east of the Mooki Thrust (*cf.* Runnegar, 1974; Brownlow, 1981; Hunt *et al.*, 1983). All the Permian formations that are truncated by and to the west of the Mooki Thrust probably extended well to the east, conformably covering the Late Carboniferous and older formations presently exposed east of the Thrust. At this stage in the Boggabri-Gunnedah district, the Early Permian volcanic centres were extinct but still had constructional relief and continued to contribute debris to the coal measure sediments accumulating in flanking low-lying areas (central Gunnedah Basin, Maules Creek). The occurrence of "flint clay" conglomerates near the base of the coal measures at Boggabri, and at the same level farther south (Loughnan, 1975; Brownlow, 1981), perhaps indicates a genetic link with the Late Carboniferous-Early Permian silicic volcanic terrain that was exposed to the west.

Continental coal forming environments existed east of the present day position of the Peel Fault in the Early Permian (Harrington, 1982), since there are coal measures of the same age as the Maules Creek coal measures at Ashford (Stage 4; Evans in McKelvey and Gutsche, 1969). Other remnants of the Early Permian sequences of New England are shallow marine (e.g. Halls Peak, Drake, Texas-Inverell, Silverwood), or deeper marine (e.g. Nambucca, Rockvale, upper Manning district). There are large areas of the New England Tableland for which the Early Permian record remains undocumented. However, the extent of emergence of the region east of the Peel Fault in the Early Permian was probably quite limited. Any easterly component in the provenance of the Maules Creek Formation and equivalents must thus be a distal and early contribution from the New England area which did not become a major source of detritus shedding westward until the Late Permian (Brownlow, 1979; Conaghan *et al.*, 1982).

Thus the sequence from the Late Carboniferous at least to the end of deposition of the Early Permian coal measures is essentially uninterrupted, taking into account its entirely fluviatile character, and dominated by the activity, demise and denudation of the western volcanic terrain, proximal Early Permian remnants of which are exposed at Boggabri and Gunnedah.

CHAPTER 6

VOLCANIC HISTORY OF A LATE PERMIAN CAULDRON : COOMBADJHA VOLCANIC COMPLEX, NORTHEASTERN NEW SOUTH WALES

INTRODUCTION

The Coombadjha Volcanic Complex is the remnant of a Late Permian silicic volcanic cauldron occupying an area of 15 km by 24 km (360 km^2) within which a thickness of 2.3 km of ignimbrites and lavas is preserved. The internal stratigraphy and structure of the Complex provide evidence that cauldron collapse occurred in response to eruption of the youngest ignimbrite, the lowermost parts of which are exposed in the core of the resulting structural basin. The magnitude of subsidence suggests that the total volume of ignimbrite produced during this stage was considerably greater than that preserved. After eruptions had ceased and the cauldron floor had subsided, an adamellite ring pluton was emplaced along the fractured perimeter of the central cauldron block.

The Coombadjha Volcanic Complex is only part of a more extensive sequence of silicic volcanic rocks of Late Permian age which are widespread (greater than 20,000 km^2) throughout the New England Tableland of north-eastern New South Wales. Within much of this area the youngest volcanic unit is crystal-rich ignimbrite which constitutes the only regionally mappable lithological subdivision recognised to date (the Dundee Rhyodacite, Flood *et al.*, 1977). The existence of volcanic cauldrons has long been suspected (e.g. Harrington, 1974) although prior to this study of the Coombadjha Volcanic Complex, only one other has been established by field data (Cuddy, 1978). The Coombadjha Volcanic Complex, herein described, is thus important not only for its exhibition of cauldron-ignimbrite relation-ships but also for its role in the Late Permian volcanology of the region.

REGIONAL GEOLOGY

The eastern portion of the New England Orogen (Day *et al.*, 1978) comprises in part middle and late Palaeozoic, largely marine sedimentary sequences, incorporated in an east-facing subduction complex (Crook, 1980a,b;

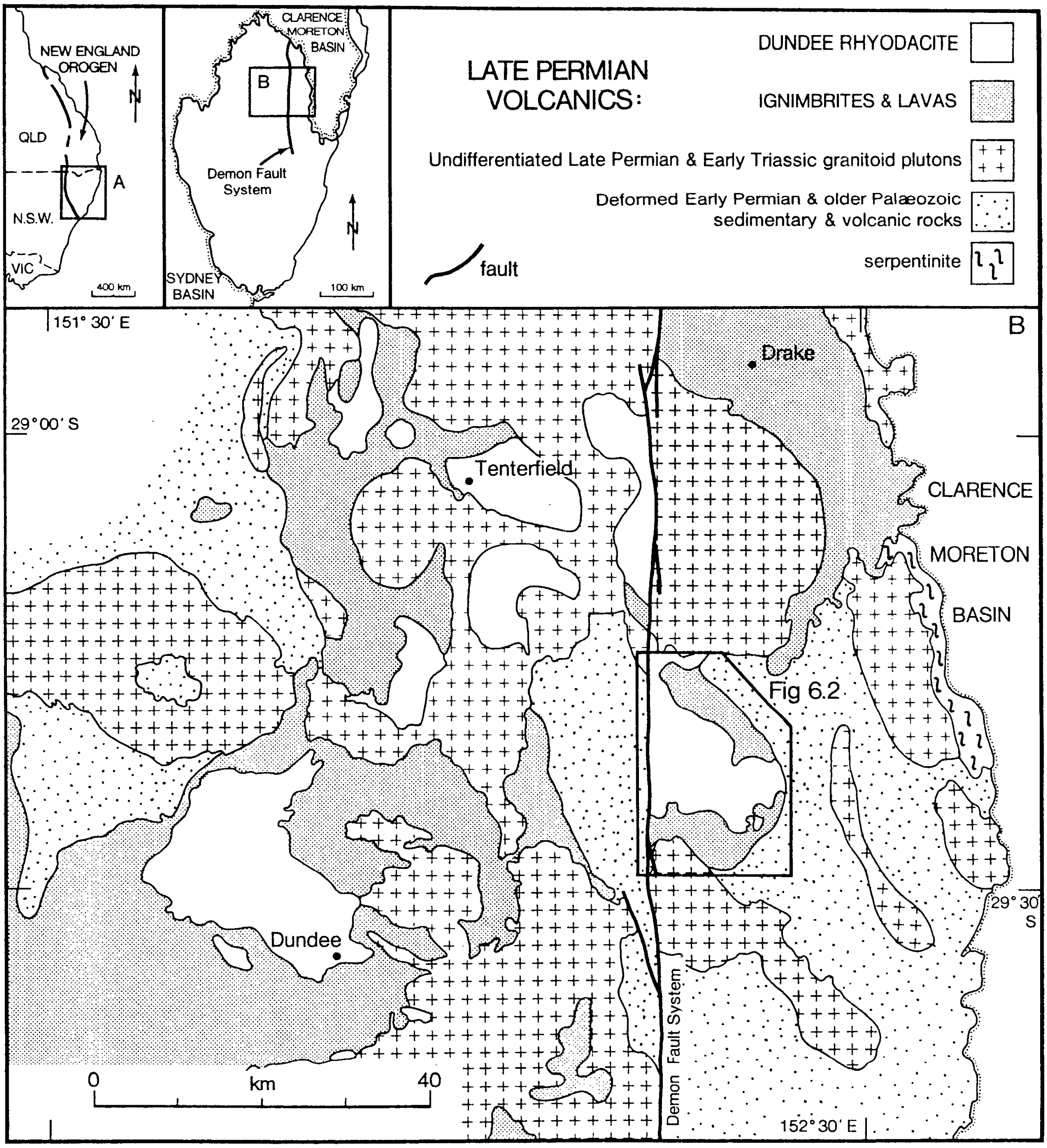


Figure 6.1: Regional geological setting of the Coombadjha Volcanic Complex and Late Permian volcanics of northeastern New South Wales. Modified from Pogson and Hitchins (1973).

Cawood, 1982; Fergusson, 1984a,b). An incomplete cover of flat lying to gently dipping Late Permian volcanic and sedimentary rocks (Korsch, 1977) unconformably overlies the subduction complex, and both these components are intruded by Permian and Triassic plutonic suites of the New England Batholith (Shaw and Flood, 1981). The Late Permian sequence is dominated by widespread ignimbrites accompanied by thin intervals of *Glossopteris*-bearing strata (McKelvey and Gutsche, 1969; Wood, 1982) indicating the preponderance of continental environments, though areas of shallow marine volcaniclastic deposition are also known (e.g. at Drake, Olgers *et al.*, 1974; Thomson, 1976). For much of the region the youngest unit of the Late Permian volcanic pile is distinctive, crystal-rich, blue ignimbrite named the Dundee Rhyodacite (Flood *et al.*, 1977), dated at 247 Ma (biotite K/Ar; recalculated by Shaw and Flood, 1981, from Evernden and Richards, 1962). The Dundee Rhyodacite now outcrops in several discrete areas or 'masses' separated by younger plutons or by older volcanics (Fig. 6.1). Flood *et al.* (1977) have suggested that these masses may be remnants of a once continuous ignimbrite sheet which covered at least 720 km^2. Mapping of one such mass and associated volcanics in deeply dissected country on the eastern edge of the New England Tableland led to the definition of the Coombadjha Volcanic Complex (McPhie, 1982).

COOMBADJHA VOLCANIC COMPLEX

The Coombadjha Volcanic Complex comprises texturally diverse Late Permian continental silicic pyroclastics (mainly ignimbrites), lavas and breccias at least 2.3 km thick, partly encompassed by a ring pluton of biotite-hornblende adamellite (Figs. 6.2,6.3). Limited geochemical data (Tables 6.1,6.2, Fig. 6.4) indicate that the major units are high K_2O calc-alkaline dacites and rhyolites (*cf.* Irvine and Baragar, 1971; Peccerillo and Taylor, 1976; Ewart, 1979,1982; Ewart and Le Maitre, 1980). Bedding in volcanics encircled by the ring pluton dips radially inward, being steepest near the intrusion and decreasing toward the centre. In plan, the Complex is elliptical, with its longer dimension (24 km) oriented northwest-southeast, though the western side is truncated at the Demon Fault System (Shaw, 1969; Korsch *et al.*, 1978). The remainder of

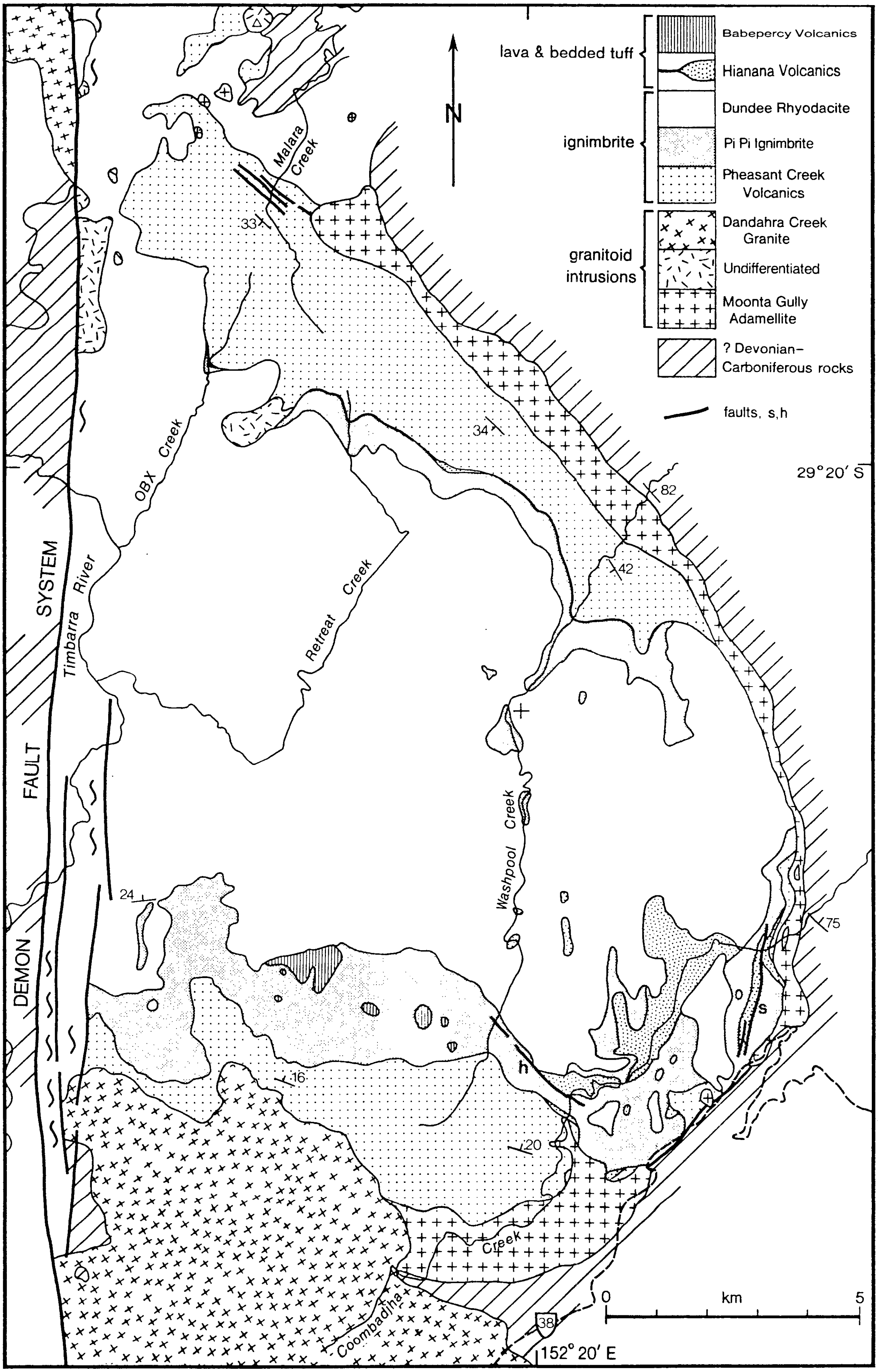


Figure 6.2: Geological map of the Coombadjha Volcanic Complex. Bedding shown is representative of the structure; more comprehensive data are presented in Figure 6.10. s, Sugarloaf faults; h, Hianana fault.

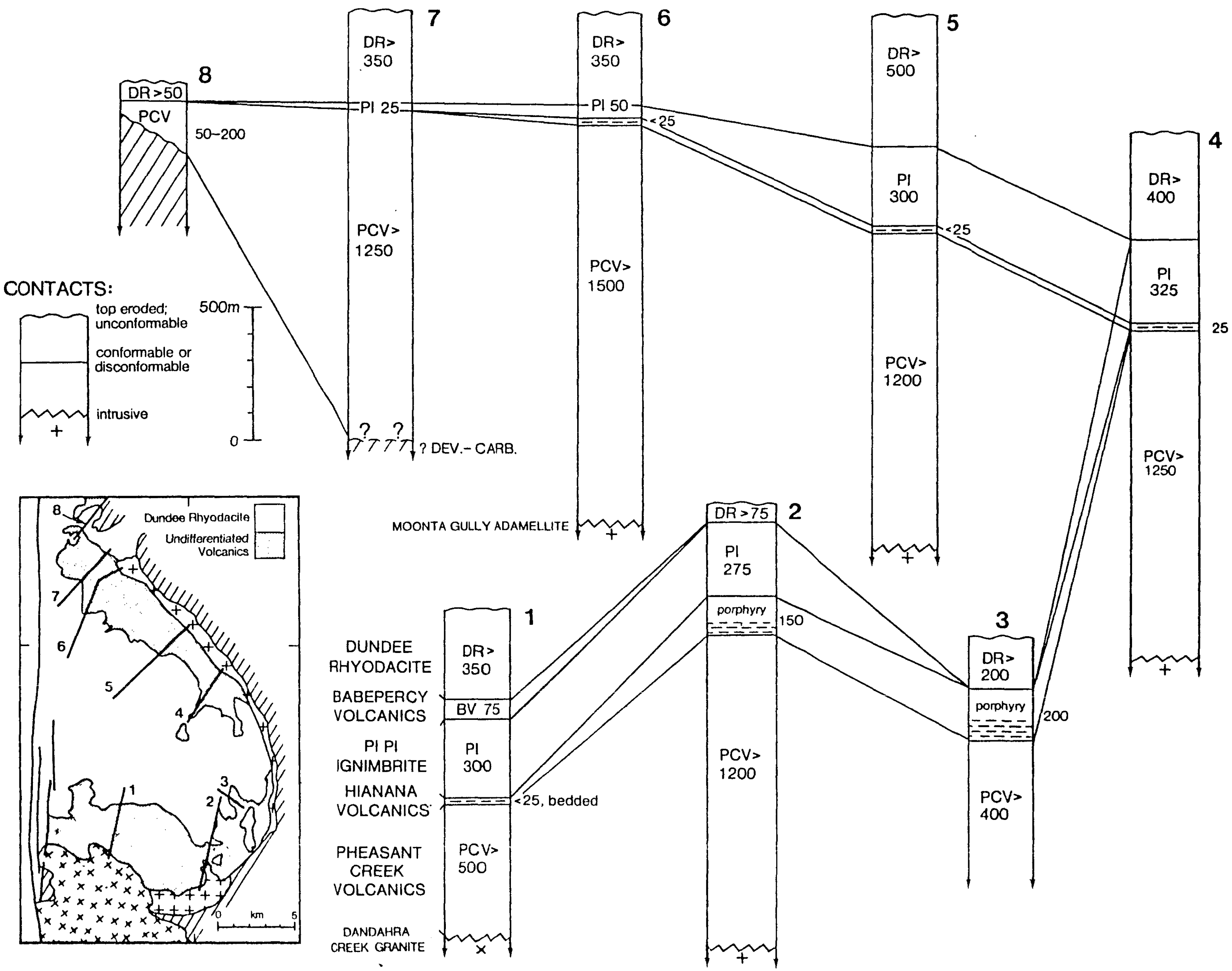


Figure 6.3: Stratigraphy and correlation of volcanic units of the Coombadjha Volcanic Complex. Numbers give thickness in metres.

TABLE 6.1: Major element oxide analyses in weight percent and C.I.P.W. norms of volcanics and the ring pluton of the Coombadjha Volcanic Complex

Sample	Pheasant Creek Volcanics			Hianana Volcanics		Pi Pi Ignimbrite			Dundee Rhyodacite					Moonta Gully	
	ignimbrites			volcani-clastic	porph-yry				Normal facies ignimbrite					Adamellite	
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Oxide	R55379	R55376	R55384	R55416	R55425	R55437	R55434	R55444	R55478[1]	R55480	R55456	R55465	R55481	R55557	R55559
SiO_2	66.05	69.40	72.21	63.50	67.81	66.88	67.53	69.67	62.35	64.79	65.24	65.27	65.43	63.57	64.60
TiO_2	0.50	0.38	0.25	0.75	0.58	0.59	0.59	0.43	0.61	0.60	0.56	0.58	0.58	0.70	0.62
Al_2O_3	16.22	14.87	14.40	15.50	16.24	15.67	15.82	14.47	16.79	15.76	15.62	15.56	15.66	16.03	15.53
Fe_2O_3	0.55	0.82	0.27	0.64	1.76	1.60	1.79	1.35	1.54	1.57	1.44	1.33	1.36	1.74	0.83
FeO	2.75	2.30	1.30	3.75	2.10	2.30	2.49	2.06	3.25	2.68	2.70	2.81	2.77	3.18	3.48
MnO	0.05	0.04	0.05	0.11	0.07	0.08	0.10	0.07	0.11	0.09	0.08	0.10	0.08	0.10	0.10
MgO	1.09	0.52	0.31	1.52	0.85	0.70	0.70	0.59	2.24	1.94	1.89	1.81	1.79	2.20	2.07
CaO	2.35	2.28	1.41	4.01	4.45	3.31	3.06	2.21	5.09	3.86	3.72	3.91	3.94	4.27	3.82
Na_2O	3.09	4.01	3.66	2.40	2.55	3.76	3.32	3.80	3.84	3.43	3.50	3.70	3.40	3.45	3.42
K_2O	6.31	4.14	5.56	3.26	2.36	3.81	3.94	4.44	2.89	3.68	3.97	3.78	3.69	3.52	3.94
P_2O_5	0.11	0.08	0.04	0.19	0.16	0.14	0.13	0.09	0.14	0.10	0.12	0.15	0.09	0.12	0.17
H_2O^+	0.96	0.61	0.42	2.32	1.27	0.48	0.64	0.57	0.71	0.83	0.44	0.53	0.59	0.97	0.97
H_2O^-	0.20	0.31	0.16	0.20	0.22	0.15	0.22	0.11	0.15	0.11	0.23	0.16	0.13	0.13	0.24
CO_2	n.d.[2]	n.d.	n.d.	1.75	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	100.23	99.76	100.04	99.90	100.42	99.47	100.33	99.86	99.71	99.44	99.51	99.69	99.51	99.98	99.79
ΣFe as FeO	3.24	3.04	1.54	4.33	3.68	3.74	4.10	3.27	4.64	4.09	4.00	4.01	3.99	4.75	4.23
Fe_2O_3/FeO	0.20	0.36	0.21	0.17	0.84	0.70	0.72	0.66	0.47	0.59	0.53	0.47	0.49	0.55	0.24
K_2O/Na_2O	2.04	1.03	1.52	1.36	0.93	1.01	1.19	1.17	0.75	1.07	1.13	1.02	1.09	1.02	1.15
						C.I.P.W. Norm									
Q	15.97	23.81	25.64	25.43	33.24	22.19	24.99	24.70	14.43	19.27	18.63	18.12	20.00	17.12	17.40
C	0.30			1.24	1.80		0.85								
or	37.64	24.75	33.03	20.15	14.10	22.78	23.41	26.45	17.28	22.08	23.74	22.56	22.07	21.04	23.62
ab	26.39	34.33	31.14	21.24	21.81	32.19	28.24	32.42	32.87	29.46	29.96	31.62	29.12	29.52	29.35
an	11.04	10.47	6.48	19.50	21.26	14.80	14.41	9.39	20.27	16.99	15.36	14.83	16.77	18.06	15.61
di		0.38	0.25			0.74		0.90	3.54	1.46	2.05	3.08	1.98	2.09	2.05
hy	6.64	4.14	2.50	9.53	3.73	3.50	4.07	3.14	7.84	7.04	6.79	6.38	6.74	7.99	9.15
mt	0.80	1.20	0.39	0.97	2.58	2.35	2.61	1.97	2.26	2.31	2.11	1.95	2.00	2.55	1.22
il	0.96	0.73	0.48	1.49	1.11	1.13	1.13	0.82	1.17	1.16	1.08	1.11	1.12	1.34	1.19
ap	0.26	0.19	0.09	0.46	0.37	0.33	0.30	0.21	0.33	0.24	0.28	0.35	0.21	0.28	0.40
100 an/ab+an	29.5	23.4	17.2	47.9	49.4	31.5	33.8	22.5	38.2	36.6	33.9	31.9	36.5	38.0	34.7

[1] Sample from the crystal-rich basal layer of the Normal facies ignimbrite.

[2] n.d. = not detected.

TABLE 6.2: Trace element analyses in μg/g of volcanics and the ring pluton of the Coombadjha Volcanic Complex

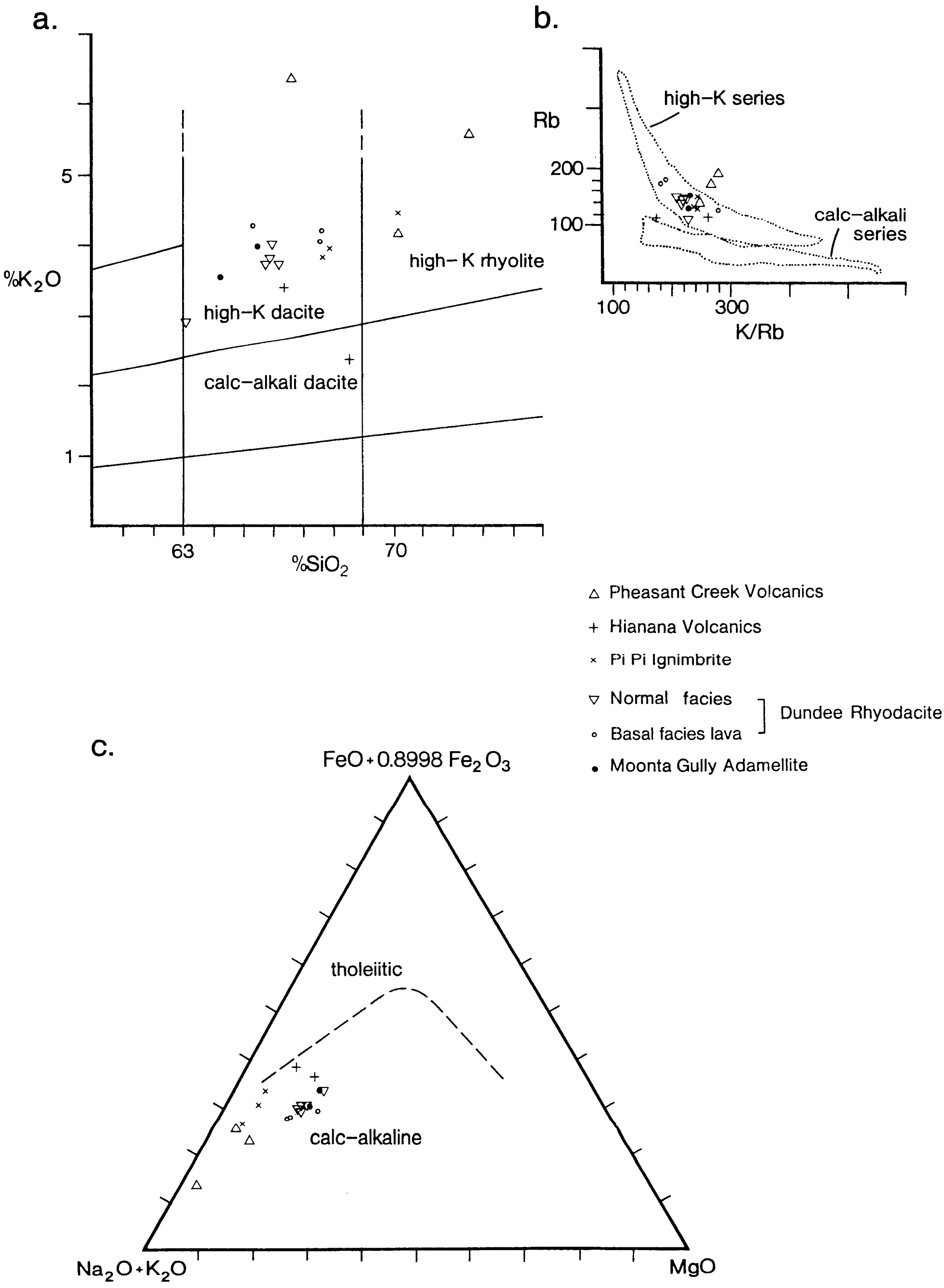
Sample	Pheasant Creek Volcanics			Hianana Volcanics		Pi Pi Ignimbrite			Dundee Rhyodacite					Moonta Gully Adamellite	
	Ignimbrites			volcani-clastic	porph-yry				Normal facies ignimbrite						
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Element	R55379	R55376	R55384	R55416	R55425	R55437	R55434	R55444	R55478[1]	R55480	R55456	R55465	R55481	R55557	R55559
Ti	3250	2500	1830	4650	3550	3970	3800	2870	4300	4110	3690	3660	4160	4430	3980
V	45	25	7	128	43	34	34	20	100	88	75	74	84	95	86
Cr	15	3	3	32	4	4	2	5	39	29	29	28	31	36	31
Mn	452	501	399	918	644	660	816	606	907	813	760	764	837	821	813
Ni	7	3	3	6	8	4	4	3	9	8	7	8	9	8	7
Cu	10	5	5	19	8	6	12	11	11	12	11	14	11	7	18
Zn	65	53	45	58	85	55	79	60	75	69	66	65	66	60	59
Ga	18	18	17	17	16	18	18	19	20	20	19	19	21	20	18
Rb	187	139	173	110	114	132	136	152	106	146	146	142	141	127	144
Sr	281	223	140	319	333	305	305	217	401	308	312	312	316	348	315
Y	35	40	37	30	38	39	44	43	31	32	31	32	28	29	30
Zr	253	249	261	177	207	235	243	277	188	194	192	193	186	200	185
Nb	10	8	9	8	8	10	10	9	7	8	9	8	8	8	9
Ba	1215	775	889	841	551	966	974	1034	689	706	730	713	718	797	734
Pb	75	59	34	30	57	61	64	39	19	26	64	27	24	13	59
Th	18	18	23	13	11	15	15	25	13	17	17	16	18	13	17
U	5	3	5	3	4	4	4	5	4	2	4	4	6	5	·2
K/Rb	283	250	268	257	174	242	242	245	229	213	229	223	220	233	231
Ba/Rb	6.5	5.6	5.1	7.7	4.8	7.3	7.2	6.8	6.5	4.8	5.0	5.0	5.1	6.3	5.1
Th/U	3.6	6.0	4.6	4.3	2.8	3.8	3.8	5.0	3.3	8.5	4.3	4.0	3.0	2.6	

[1] Sample from the crystal-rich basal layer of the Normal facies ignimbrite.

Figure 6.4: a. K_2O versus SiO_2 values (recalculated H_2O- and CO_2- free) for the major units of the Coombadjha Volcanic Complex (Table 6.1) in relation to the subdivisions of Peccerillo and Taylor (1976) and Ewart (1979).

b. µg/g Rb versus K/Rb values of major units of the Coombadjha Volcanic Complex (Table 6.2) in relation to the high-K and calc-alkali series of Ewart (1979, Fig. 4).

c. AFM diagram showing the major units of the Coombadjha Volcanic Complex in relation to the tholeiitic and calc-alkaline fields of Irvine and Baragar (1971, Fig. 2).


a.
5
1
%K$_2$O
63
70
%SiO$_2$
high-K dacite
calc-alkali dacite
high- K rhyolite
b.
Rb
200
100
100
300
K/Rb
high-K series
calc-alkali series
c.
FeO+0.8998 Fe$_2$O$_3$
Na$_2$O+K$_2$O
MgO
tholeiitic
calc-alkaline
△ Pheasant Creek Volcanics
+ Hianana Volcanics
× Pi Pi Ignimbrite
▽ Normal facies
∘ Basal facies lava
Dundee Rhyodacite
• Moonta Gully Adamellite

the Complex west of the Fault has been offset 23 km to the north as a result of dextral strike slip since the Triassic (McPhie and Fergusson, 1983; Appendix A). Parts of the northern and southwestern margin are intruded by Early Triassic granitoid plutons. Deformed ?Devonian-Carboniferous rocks (the Willowie Creek beds, Gundahl Complex, Fergusson, 1984b), including argillite, greywacke, limestone breccia, chert and greenstone, surround the Complex from the north around to the south. Bedding in these units is steeply dipping and strikes northwest-southeast as a result of deformation that predated the formation of the Coombadjha Volcanic Complex (Fergusson, 1984b). The ring pluton occupies the site of a previously faulted contact between the older deformed rocks and the volcanics. The unconformity between the volcanics and the older deformed rocks is exposed in a very limited area at the extreme north (Fig. 6.2).

STRATIGRAPHY OF THE COOMBADJHA VOLCANIC COMPLEX

The Coombadjha Volcanic Complex has an internal stratigraphy made up of five volcanic units and one intrusive unit that has been mapped at 1:25000 scale. The youngest volcanic unit, correlated with the Dundee Rhyodacite of Flood *et al.* (1977), covers much of the central area of the Complex (Fig. 6.2). The older volcanic units are exposed concentrically within the inner margin of the ring pluton and in windows through the Dundee Rhyodacite in the central area. Exposure is excellent and primary textures and structures are generally well preserved in outcrop. All originally glassy fragments are devitrified; feldspars and ferromagnesian minerals in the volcanics older than the Dundee Rhyodacite are at least slightly altered and in some cases, only pseudomorphs remain.

Representative sections through the Complex occur along Coombadjha Creek and its tributaries (downstream from GR355400 Coombadjha) and also along Washpool Creek (especially downstream from GR350503 Washpool). The mappable subdivisions of the Complex, in order from oldest to youngest, are: the Pheasant Creek Volcanics, the Hianana Volcanics, Pi Pi Ignimbrite, the Babepercy Volcanics, the Dundee Rhyodacite and the Moonta Gully Adamellite. These are described in turn below.

Pheasant Creek Volcanics

The predominant rocktype of the Pheasant Creek Volcanics is ignimbrite. Other subordinate components are breccia, crystal tuff, and silicic lava. This unit forms the highest ridges of the area and is exposed almost without interruption concentrically within the margin of the Complex (Fig. 6.2). The thickest sections amount to approximately 1300 m, though this value is a minimum because the base is not exposed (Fig. 6.3). Lowermost levels are truncated by the ring pluton or are in fault contact with the surrounding ?Devonian-Carboniferous rocks, with the exception of the extreme north where a much thinner interval of the Pheasant Creek Volcanics (less than 200 m) unconformably overlies the older deformed sequence. Although the association of ignimbrites and breccias has proven reliably mappable, the internal facies variations and stratigraphy of this unit are the least well known because it lacks extensive marker beds, and lithological changesoccur over short distances. However, there are no contrasts in facies apparent across the Complex at the scale mapped with the exception of the coherent silicic lava which is confined to a single southern locality (Fig. 6.5).

Ignimbrite

There are three main textural varieties of ignimbrite in the Pheasant Creek Volcanics (Figs. 6.5,6.6,6.7a,b,c,d):

(1) Crystal-rich ignimbrite; typically dark blue, relatively coarse-grained (crystal fragments up to 4 mm), originally welded ignimbrite with chloritic relic pumice lapilli (or "lenticles") aligned parallel to bedding Crystal fragments are plagioclase, K-feldspar, quartz, biotite and altered ?hornblende, and amount to between 25 and 45 modal percent.

(2) Crystal-poor ignimbrite, generally pale grey and fine grained, with inconspicuous relic pumice lapilli. The crystal fragments do not exceed 10 to 20 modal percent and K-feldspar is characteristic, accompanied by quartz and plagioclase.

(3) Strongly foliated ignimbrite, characterised by essentially two-dimensional relic pumice which defines a prominant and regular bedding-parallel 'streaky' fabric. This originally densely welded ignimbrite is dark grey-black and crystal-poor (less than 20 modal percent plagioclase and an altered ferromagnesian mineral). Most outcrops display regular

prismatic joints.

Most sections of the Pheasant Creek Volcanics include two discrete intervals of the grey-black strongly foliated ignimbrite (3) and one interval of the pale crystal-poor ignimbrite (2), intercalated with the dominant crystal-rich ignimbrite (1), and minor thicknesses of other rocktypes (breccia, crystal tuff, lava) (Fig. 6.5). The crystal-rich ignimbrite (1) is the most voluminous, and in some sections it amounts to more than 250 m in thickness within which there are subtle changes in proportions and sizes of feldspar grains and of chloritic relic pumice. However, internal flow or welding subdivisions in this ignimbrite type are inconspicuous and no basal layers (2a; Sparks *et al.*, 1973) have been discerned.

Recognition of separate flow units within the other two ignimbrite types (2,3) is precluded by their lack of internal textural variations, though one unit of the dark grey strongly foliated ignimbrite (3) exposed in Washpool Creek has a basal concentration of volcanic lithic fragments which are up to 10 cm across (GR365523 Washpool).

Breccia

Breccia comprising poorly sorted, angular volcanic rock fragments occurs in massive beds a few to several metres thick between the ignimbrite layers, but no exposures of contacts have been observed. Breccia is widespread and evidently not restricted to a single level in the pile but is a volumetrically minor rocktype (less than 10 percent) of the Pheasant Creek Volcanics. Although volcanic clasts predominate, one example has a small proportion of subrounded argillite cobbles (GR276592 Washpool).

In two instances, breccia occurs near the base of crystal-rich ignimbrite (GR305422, GR354418 Coombadjha) and includes dense lithic clasts up to 20 to 30 cm across. These occurrences could be related to the ignimbrite, as dense clasts of this size and abundance have been reported in several Cainozoic pyroclastic flow deposits (e.g. the Ito pyroclastic flow; Yokoyama, 1974; the Taupo ignimbrite; Walker, Self and Froggart, 1981). Pods and lenses of breccia in ignimbrite are attributed to local and transient enhancement of the fluidisation state of the pyroclastic flow (Wilson, 1980,1984; Walker, 1983). More laterally continuous breccia at or near the base of ignimbrite is probably the result of a more sustained but systematically decaying fluidisation condition in the head and body of the pyroclastic flow, typical of comparatively proximal settings, and largely inherited from formation of the flow by eruption column collapse (Wright and Walker, 1977,1981; Wilson, 1980; Druitt and Sparks, 1982; Wilson and Walker, 1982). The breccias near the bases of intervals of the crystal-rich ignimbrite are possibly examples of the latter case (that is, ground breccias and/or coarse 2bL layers) but their lateral continuity has not been established.

A variety of epiclastic processes could also have been responsible for some of the breccias, especially rockfall adjacent to minor fault or erosional scarps, and redeposition from more remote sites by subaerial debris flows.

Crystal tuff

Intervals of diffusely thin bedded, feldspathic crystal tuff, no more than 2 to 3 m thick, are a very minor component of the Pheasant Creek Volcanics, and occur mainly along the northeastern side of the Complex. One exposure in lower Washpool Creek underlies extremely crystal and lithic rich, massive ignimbrite (1) (GR369530 Washpool) and may be a primary pyroclastic deposit (an ignimbrite-related Layer 1 pyroclastic ground surge?; *cf.* Sparks *et al.*, 1973; Wright *et al.*, 1980). Elsewhere an epiclastic origin cannot be discounted although the angularity of crystal fragments suggests minimal reworking.

Lava and lava breccia

Coherent flow banded silicic lava and lava breccia outcrop in a small area on the southern side of the Complex (GR343403 Coombadjha). Too little is exposed to determine the arrangement of facies; whether the body is a flow or dome with an autobrecciated carapace, or a partly intrusive feeder has not been resolved. Breccia made of flow banded silicic lava fragments also occurs on the northern side of the Complex (GR285620 Washpool), without any associated coherent lava. The presence

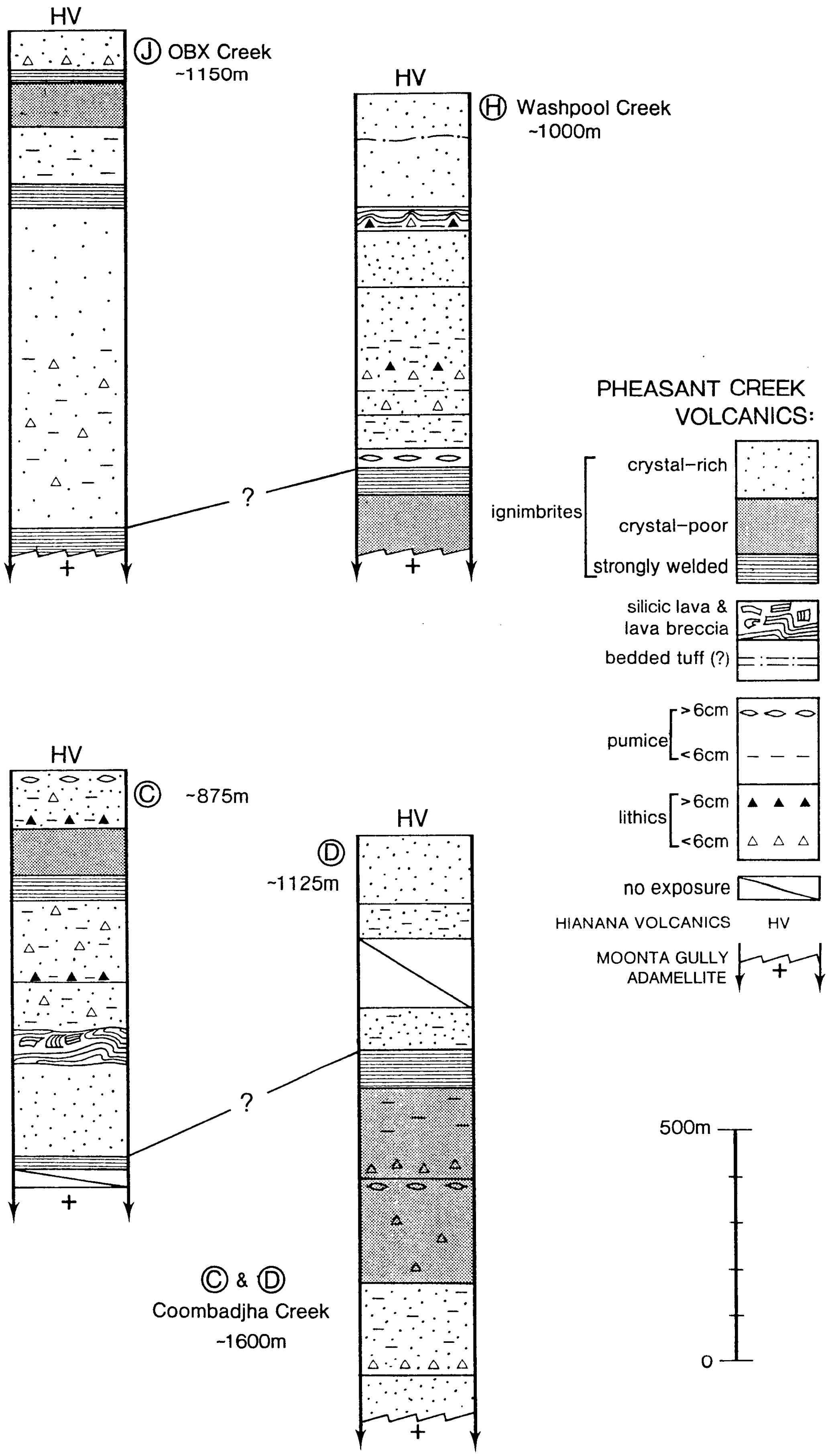


Figure 6.5: Stratigraphic columns showing the internal lithological diversity within the Pheasant Creek Volcanics. The two southern sections (C,D) are about 1 km apart. The two northern sections (H,J) are about 6 km apart (see Figure 6.12).

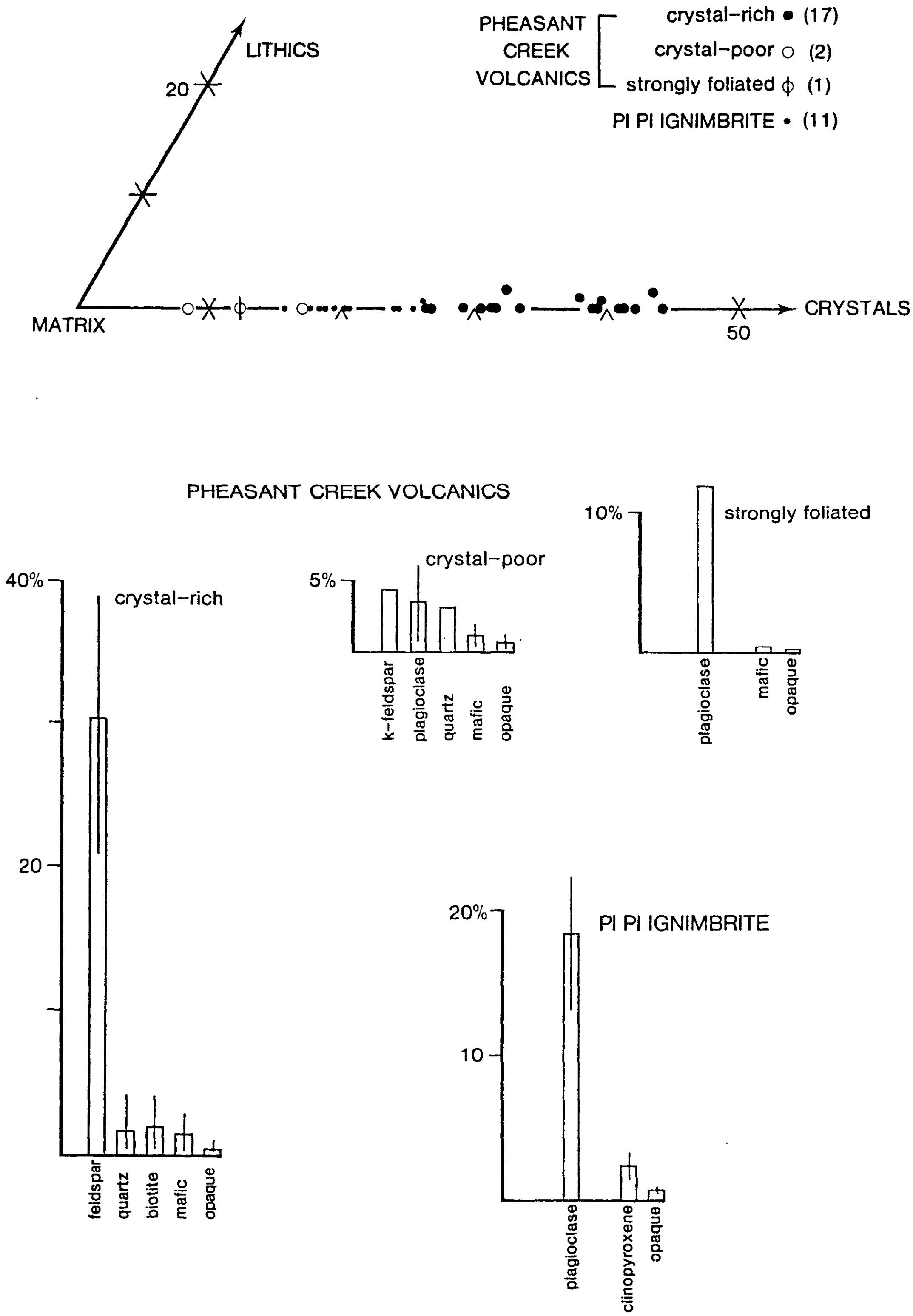


Figure 6.6: Modal data for ignimbrites of the Pheasant Creek Volcanics and the Pi Pi Ignimbrite.'Matrix' includes devitrified pumice, shards and fine ash. Histograms give the average (wide bar) and range (line) in modal percent of the different crystal fragment types. 'Mafic' includes altered ferromagnesian minerals. See also Appendix B.

Figure 6.7: a. Crystal-rich ignimbrite (1) of the Pheasant Creek Volcanics. Note feldspar crystal fragments and small recessive pits indicating former pumice lapilli. GR35304162 Coombadjha. Hammer head about 16 cm.

b. Photomicrograph of the crystal-rich ignimbrite (1) of the Pheasant Creek Volcanics, showing crystal fragments (f, plagioclase; q, quartz) and poorly-preserved matrix shards. R55390, plane polarised light.

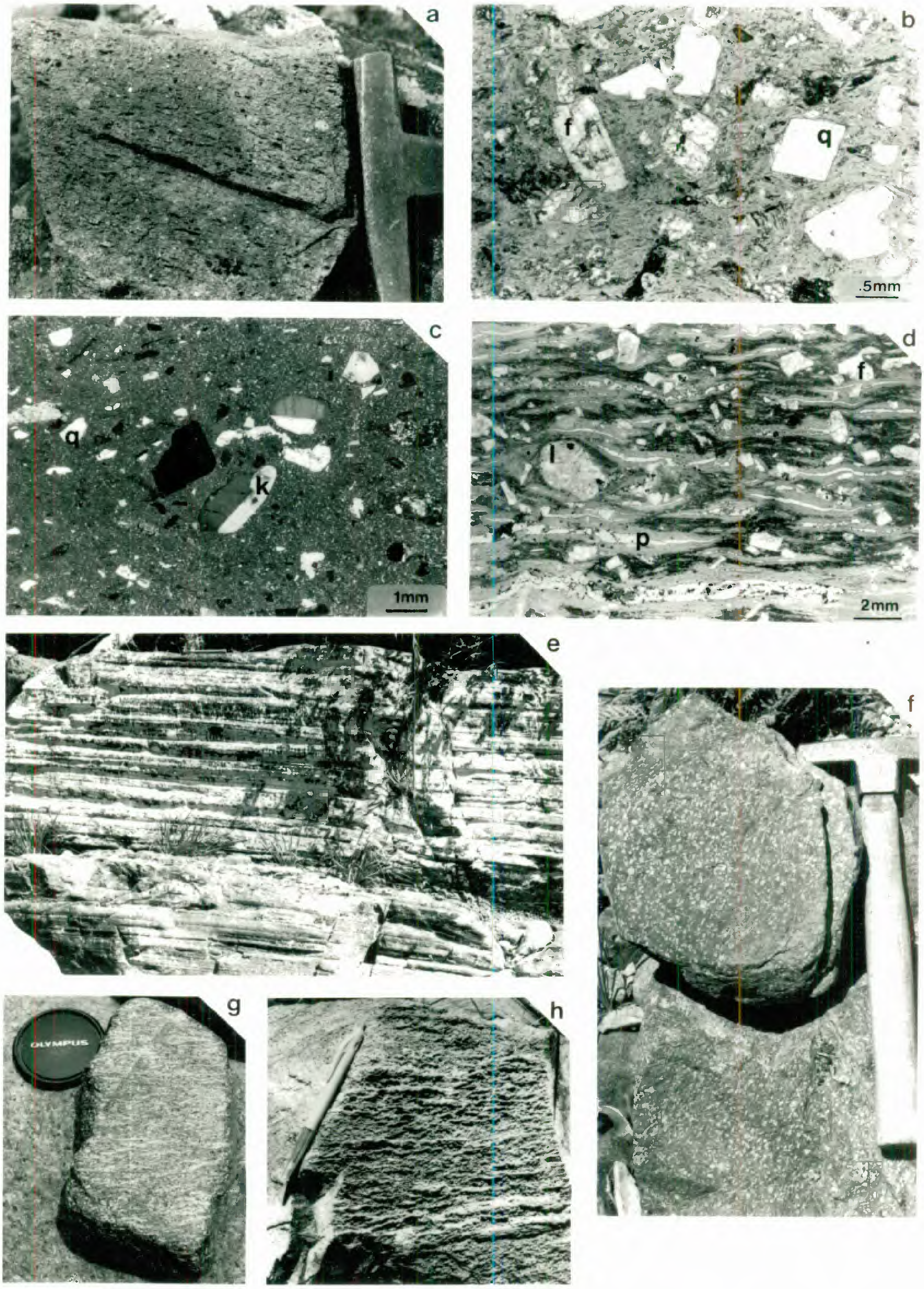
c. Photomicrograph of the crystal-poor ignimbrite (2) of the Pheasant Creek Volcanics, showing crystal fragments of sanidine (k) and quartz (q) in a structureless matrix. R55392, crossed nicols.

d. Photomicrograph of the strongly foliated ignimbrite (3) of the Pheasant Creek Volcanics, showing abundant relic pumice (p), crystal fragments (predominantly plagioclase, f) and lithic fragments (l). R55402, plane polarised light.

e. Outcrop of the bedded volcaniclastic facies of the Hianana Volcanics. Note the crest of long wavelength, very low wave height dune about 1 m below the hammer. GR38454500 Coombadjha. Hammer 33 cm.

f. Massive plagioclase porphyry of the Hianana Volcanics. Boulders in Coombadjha Creek. Hammer 33 cm.

g,h. Handspecimen and outcrop of the Pi Pi Ignimbrite, showing the typically well developed alignment of flattened relic pumice lapilli. Outcrop (h) at GR35355055 Washpool. Lens cap 5.5 cm. Pen 13.5 cm.


a
b
f
q
.5mm
c
q
k
1mm
d
f
l
p
2mm
e
f
g
OLYMPUS
h

of significant amounts of finer grained matrix and of a variety of volcanic clast types suggest that this breccia may be a block and ash flow deposit (*cf.* Wright *et al.*, 1980), resulting from explosive disruption or gravitational collapse of a lava flow or dome, rather than autobrecciated lava debris. In either case, its source has been destroyed by later events involved with the formation of the Complex.

Hianana Volcanics

The Hianana Volcanics consist principally of two facies: dark grey-green, typically thinly bedded volcaniclastic rocks, and massive, dark green-black dacite porphyry. Both facies are present in the southeastern sector of the Complex where they constitute two largely discrete units, the bedded volcaniclastics being conformably overlain by the massive porphyry (Figs. 6.2,6.3). In other parts of the Complex the Hianana Volcanics are represented by the bedded volcaniclastic facies.

Bedded volcaniclastic facies

Rocks of this facies are composed of angular volcanic lithic and crystal fragments (mainly plagioclase), typically in the coarse ash size range (0.63 to 2 mm), accompanied by unresolvable presumably finer grained matrix. Relic chloritic pumice, quartz grains and an altered ferromagnesian mineral occur in trace amounts in thin-sections of some beds.

Most outcrops display uniform thickness, laterally continuous, thin (3 cm to 10 cm) bedding (Fig. 6.7e). In the southeastern part of the Complex where the thickest continuous sequence of the bedded volcaniclastic facies occurs (more than 70 m), there are also long wavelength, low wave height dune bed forms, cross-bedding and wavy bedding (Fig. 7.5). Mantle bedding, bed thickness irregularities, pinch and swell, lensing and low-angle erosional truncations are also evident on close examination of outcrops in this area. Although the bedded volcaniclastic facies extends to the west (near Pheasant Creek) and to the north (OBX Creek), it is much thinner (less than 20 m) and the predominant bed forms are plane parallel very thin beds (1 cm to 3 cm) and laminae (less than 1 cm).

Massive porphyry facies

Phenocrysts in the massive porphyry are complete, euhedral plagioclase