CHAPTER 1

INTRODUCTION

The primary aim of this study has been to reconstruct the late Palaeozoic history of explosive silicic volcanism in the New England Orogen of northeastern New South Wales. In pursuing this aim, I have relied on two fundamental principles which have emerged from field investigations of Quaternary volcanoes and the products of their eruptions.

i. Primary pyroclastic layers have distinctive grain size populations, textures and geometries which are inherited from processes operating during eruption and emplacement.

ii. There are systematic variations in lithological assemblages and in the textures of particular pyroclastic layers which reflect proximity of the depositional site to source vents.

The pyroclastic deposits and lavas from the late Palaeozoic eruptions of New England are rocks which have been altered to varying degrees and mildly deformed. However, the quality and continuity of exposure in the areas selected are comparable to those available in many Cainozoic volcanic terrains. Furthermore, most of the New England sequences described herein are dominated by voluminous silicic ignimbrites which were generated on a scale not yet witnessed. The physical volcanology of large magnitude ignimbrite eruptions is currently approached indirectly by extrapolation, employing data from smaller scale events that have generated pyroclastic flows (e.g. Rowley *et al.*, 1981), and by experimental (e.g. Wilson, 1980) or theoretical (e.g. Sparks *et al.*, 1978) modelling. In this regard, the interpretations presented here have the same preliminary status as those drawn from investigation of younger ignimbrites.

METHODS

Volcanic sequences of Late Carboniferous, Early Permian and Late Permian age have been studied at a range of scales. The areas chosen are those with the best exposures and least prior work. Investigation of each sequence concentrated on two aspects.

i. Field mapping to establish the character, distribution, geometry and stratigraphy of each volcanic pile.

ii. Comparison of these data with Cainozoic sequences of similar composition and scale in order to clarify and reconstruct the style and setting of the New England eruptive events.

The former has been accomplished in the course of 40 weeks field mapping between 1979 and 1983. The latter has relied upon recently published descriptions of the lithofacies and environments of Cainozoic volcanic terrains, complemented by 4 weeks of reconnaissance field experience in the Central Volcanic Zone of the North Island of New Zealand.

Where relevant, the petrography and geochemistry of the volcanic rocks are reported although in general these aspects have not been explored in any detail.

THESIS FORMAT

The order of the chapters relates to the age of the rocks concerned, and also to the proximality of the sequences. Thus, Chapters 2,3 and 4 deal with Late Carboniferous outflow ignimbrites that are interbedded with fluvial conglomerates and are remote from their source eruptive centres by at least several kilometres (medial to distal settings). The source volcanic terrain is not exposed but has been reconstructed using clues provided by the ignimbrites and epiclastic rocks. Chapters 6,7 and 8 concentrate on a Late Permian volcanic cauldron which has a dominantly proximal character. Chapter 5 describes an Early Permian volcanic centre that includes prominent silicic lavas. The final chapter is a synthesis and summary of the principal conclusions presented in foregoing chapters. The maps which are the data base for this thesis are attached, with reduced versions within the text.

GEOLOGICAL SETTING

Each chapter includes a resume of the relevant geological setting. A brief but more comprehensive outline follows.

The regional geology of the areas studied is indicated on Figure 1.1. The Late Carboniferous and Early Permian sequences are

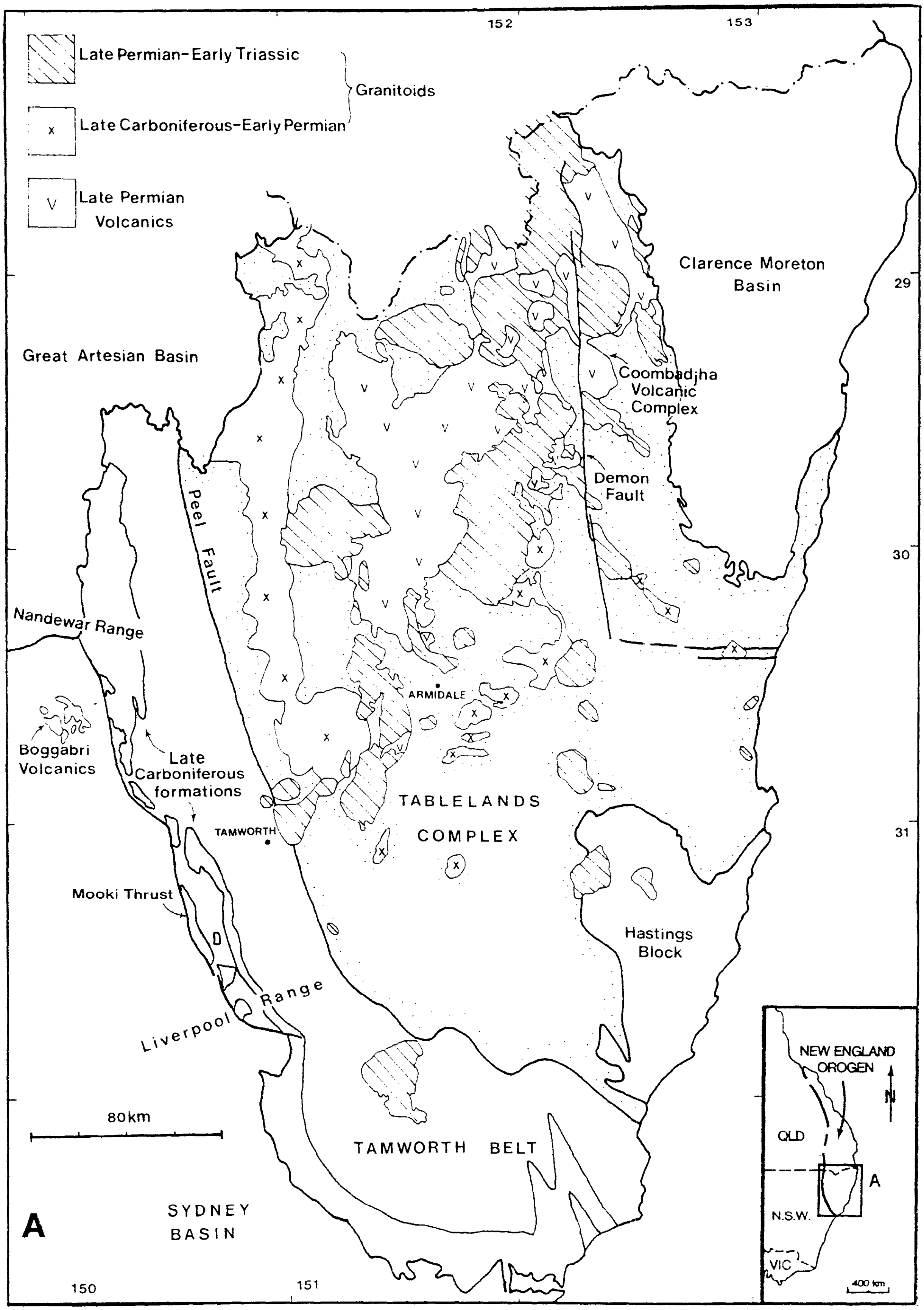


Figure 1.1: Locations of areas of late Palaeozoic volcanic sequences mapped for this study, in relation to the major features of the southern part of the New England Orogen.

within the youngest and westernmost parts of the Tamworth Belt or Zone A of the southern portion of the New England Orogen (Leitch, 1974; Korsch, 1977; Day *et al.*, 1978). The Tamworth Belt consists of folded, early Palaeozoic to Early Permian, predominantly volcanogenic sedimentary formations. The older formations are marine whereas late Early Carboniferous (Visean) and younger formations are mainly continental. Primary volcanic units are common in the Late Carboniferous and Early Permian succession close to the western and southwestern edge of the Belt. The Late Carboniferous source regions are not exposed but were located to the west of present outcrops of the derivative sedimentary and volcanic formations. The Tamworth Belt is considered to represent the fill of a forearc basin eastward of an evolving magmatic arc that was in existence throughout much of the Palaeozoic (Leitch, 1975).

Late Permian ignimbrites, lavas and subordinate, predominantly continental sedimentary sequences form a widespread though incomplete veneer to the Tablelands Complex or Zone B of the New England Orogen (Leitch, 1974; Korsch, 1977). They overlie or are in fault contact with complexly deformed older Palaeozoic, largely marine sedimentary and volcanic rocks, and they are intruded by Permian and Triassic plutonic suites of the New England Batholith (Leitch, 1974; Shaw and Flood, 1981). The deformed marine sequence constitutes a Palaeozoic subduction complex which developed in association with the magmatic arc and forearc system of the Tamworth Belt (Fergusson, 1982,1984a,b; Flood and Fergusson, 1984). The Late Permian volcanic and intrusive episode post-dates this regime although it may be genetically related to it and represent the culmination of regional orogenesis (Leitch, 1974,1975; Crook, 1980a). Recognition of a dextral transcurrent tectonic configuration that affected the entire New England Orogen in the Permian (Murray and Whitaker, 1982; Flood and Fergusson, 1984) poses an alternative context for the Late Permian silicic igneous activity.

PREVIOUS MAPPING

Werrie Syncline

Benson (1920) and Carey (1934,1935,1937) have provided the stratigraphic framework for the Werrie Syncline, later formalised by Voisey and Williams (1964). The late Visean and Late Carboniferous continental sequence comprises the older Merlewood and the younger Currabubula Formations, and conformably overlies marine mudstone (Namoi Formation). Both of the continental formations are volcanogenic, although the primary volcanic units which were the focus of this study are principally confined to the Currabubula Formation. These volcanic units have not previously been separately mapped nor investigated in any detail. Whetten (1965) gave an account of the palaeogeographic significance of a glacial interval in the Currabubula Formation and published a detailed map of the geology close to Currabubula township. Moore and Roberts (1976) presented a map of the northernmost closure of the Werrie Syncline, and concentrated on the lithofacies of a late Visean shallow marine member (Kyndalyn Member) of the Merlewood Formation.

Boggabri and Gunnedah

Hanlon (1948a,b,1949) published maps of the Boggabri and Gunnedah districts which were subsequently revised by Manser (1965a,b). The oldest rocks exposed in these areas are Early Permian ignimbrites and lavas of the Boggabri and Gunnedah Volcanics. These are overlain by Early Permian and younger coal measures and sedimentary formations of the Gunnedah Basin (Brownlow, 1981; Bembrick *et al.*, 1980). There have been no prior field studies concerned with the volcanological and palaeogeographic significance of the Boggabri Volcanics. The Gunnedah Volcanics are very poorly exposed and were examined in reconnaissance fashion only.

Coombadjha Volcanic Complex

This area is shown on regional geological maps as undifferentiated Permian granitoids and volcanics (Pogson and Hitchins, 1973; Brunker and Chesnut, 1976). No other geological data on the locality were available when mapping for this study commenced.

NOMENCLATURE

The terminology adopted for texturally distinct, internal subdivisions of ignimbrite flow units is that of Sparks *et al.* (1973), Sparks (1976), complemented by the additional refinements noted by

Wright *et al.* (1980). As far as possible, the grain size nomenclature for pyroclastic rocks proposed by Fisher (1961; Table 1.1) has been followed.

With the exceptions listed below, potentially contentious terms have been defined in the text where appropriate.

Pyroclastic rocks are composed of fragments generated by explosive volcanism and emplaced by pyroclastic flow, pyroclastic surge or pyroclastic fall mechanisms.

Epiclastic rocks are composed of fragments generated by surface weathering processes affecting pre-existing rocks.

Volcaniclastic is a general term used for rocks composed of volcanic particles (volcanic rock fragments, shards, ash, pumice, crystal fragments). This term has been applied in two situations:

a. where neither an epiclastic nor a pyroclastic origin could be confidently confirmed;

b. where it was necessary to indicate an origin involving significant epiclastic reworking or redeposition of an unconsolidated, initially pyroclastic accumulation.

Tuff refers to a pyroclastic rock of ash grade (finer than 2 mm; Table 1.1) for which the details of the eruption and emplacement processes have not been resolved.

Porphyry is used for an igneous rock with a porphyritic texture; the term does not carry compositional or genetic connotations.

Lava and lava flow have dual senses, catering both for a moving body of magma extruded on the surface, and for the solid rock formed upon cooling.

Continental sedimentary rocks are those which were formed on land, for example in fluvial, lacustrine or glacial environments. In this context, 'continental' is a term complementary to 'marine' (offshore oceanic settings).

GRID REFERENCES

Within the text, some localities are specified by grid references followed by the name of the relevant topographic map. The first half of each grid reference gives 'eastings', and the second half gives 'northings'. In most instances, six-figure grid references apply to Standard 2 inches to 1 mile Topographic Map sheets, whereas eight-figure grid references apply to 1:25000 Series Topographic Maps. Both of these map categories have been produced by the New South Wales Department of Lands, Sydney. A complete list of the map sheets covering the areas studied is given in Appendix D.

TABLE 1.1: Terminology and grain size limits for pyroclastic fragments and rocktypes, after Fisher (1961).

fragment size: (mm)	fragment:	rock name:
	blocks and bombs	(volcanic) breccia
64		
	lapilli	lapilli tuff, microbreccia
2		
	coarse ash	coarse (ash) tuff
0.063		
	fine ash	fine (ash) tuff

TABLE 1.2: Terminology for bed thicknesses.

thickness (cm)	term
>100	very thick
30-100	thick
10-30	medium
3-10	thin
1-3	very thin
<1	laminated

CHAPTER 2

OUTFLOW IGNIMBRITE SHEETS FROM LATE CARBONIFEROUS CALDERAS, CURRABUBULA FORMATION, NEW SOUTH WALES

INTRODUCTION

Field studies of modern centres of pyroclastic volcanism have provided criteria for recognition of undisturbed primary deposits, formed by different eruption and emplacement processes (e.g. Sparks *et al.*, 1973; Wright *et al.*, 1980), and for distinguishing proximal (near vent) from distal settings (e.g. Wright *et al.*, 1981). Application of these criteria to ancient volcanic rock sequences must be prefaced by an appreciation of the low preservation potential of unconsolidated pyroclastic debris in active subaerial environments, the importance of geometry, thickness and distribution in elucidating the origin of a given deposit, and the changes attendant on alteration of vitric components. These qualifications aside, field study of the primary pyroclastic units in ancient sequences may provide constraints on the location, size and style of eruption of centres of volcanism now denuded or not exposed, and yield valuable insights into the processes operating at the site of emplacement, particularly epiclastic processes in medial-distal settings.

Such an approach has been applied to outflow ignimbrites interbedded with fluvial volcanogenic conglomerates of the Late Carboniferous Currabubula Formation in northeastern New South Wales. No remnants of the source volcanic centres have been confidently identified in surface exposures within or adjacent to the areas throughout which the Currabubula Formation extends. The pyroclastic components of this sequence have long been recognised (e.g. Carey, 1934) but no previous attempts have been made to evaluate their significance in the light of recent advances in the understanding of pyroclastic volcanism.

The term "ignimbrite" is used here in the sense of Sparks *et al.* (1973) and Wright *et al.* (1980) and refers to a rock body composed of the deposits from pumiceous pyroclastic flow(s) produced by one eruption. "Welded" ignimbrites are those with collapsed, plastically deformed and moulded matrix shards evident in thin-section. In ancient ignimbrites, the presence of pumice lapilli foliation in outcrop is not necessarily a reliable indicator of welding, because originally spherical pumice fragments in non-welded ignimbrite may collapse to lenticular shapes in response to load accompanying lithification or to later deformation. Also, in some modern ignimbrites pumice fragments have lenticular shapes irrespective of the welding zonation of the unit. For example, lenticular pumice occurs throughout the Ahuroa Ignimbrite (Fransen and Briggs, 1981) and in the least welded zones of the Whakamaru Ignimbrite (Briggs, 1976). Those ignimbrites here regarded as "non-welded" may well include examples of "sillar", that is, pyroclastic flow deposits rendered coherent by vapour-phase crystallisation rather than by welding (Fenner, 1948; Sparks, 1975). Such deposits have a preservation potential which is greater than strictly non-welded ignimbrites, and comparable to that of welded varieties.

GEOLOGICAL SETTING

The western portion of the New England Orogen (Day *et al.*, 1978) of northeastern New South Wales consists of folded Devonian and Carboniferous strata. North of the Liverpool Range, this sequence forms a north-trending belt between two major faults (Fig. 2.1). Carboniferous strata outcrop in the central and western parts of the belt, principally within two gently north and south plunging, coaxial synclines: the Werrie Syncline (Carey, 1934) in the south, and the Rocky Creek Syncline (McKelvey and White, 1964) in the north. Late Visean and younger formations lining these structures are largely continental and volcanogenic, and conformably overlie an earlier Carboniferous and Devonian marine sedimentary sequence. For both synclines the Late Carboniferous record is marked by the prominence of silicic ignimbrites interbedded with fluvial conglomerate and sandstone. In the Werrie Syncline, the mappable pyroclastic stratigraphy of the Late Carboniferous is represented by ignimbrite members of the Currabubula Formation (Voisey and Williams, 1964), described below.

PYROCLASTIC FACIES OF THE CURRABUBULA FORMATION

Character

Rhyolitic (Chapter 4, Table 4.2) welded ignimbrites are the most

Figure 2.1: Distribution of Late Carboniferous strata on the western edge of the New England Orogen, between the Nandewar (north) and Liverpool (south) Ranges.

(1) Cover comprising Permian strata, Tertiary volcanic rocks, or alluvium.

(2) Late Carboniferous Currabubula Formation and equivalents.

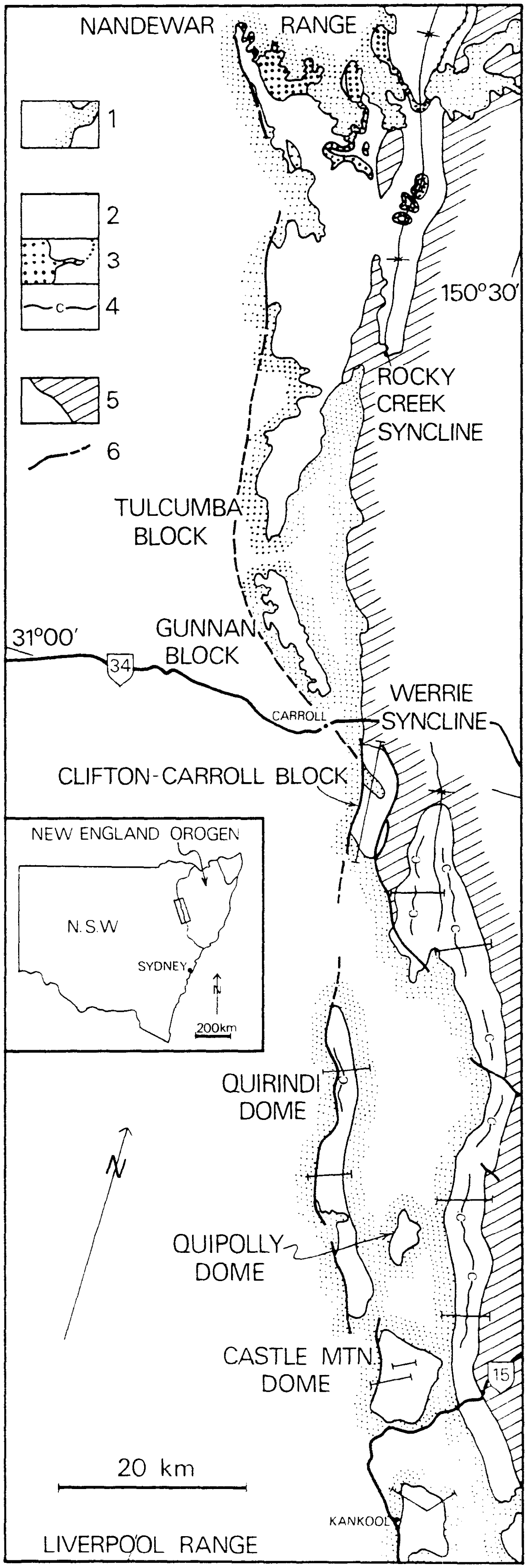
(3) Plagyan Rhyolite (after White, 1965).

(4) Cana Creek Tuff Member of the Currabubula Formation.

(5) Early Carboniferous and Devonian strata.

(6) Mooki Thrust and other fault traces.

Bars indicate positions of sections used for stratigraphic columns in Figure 2.2.


NANDEWAR RANGE
1
2
3
4
5
6
150°30'
ROCKY CREEK SYNCLINE
TULCUMBA BLOCK
GUNNAN BLOCK
31°00'
34
CARROLL
WERRIE SYNCLINE
CLIFTON-CARROLL BLOCK
NEW ENGLAND OROGEN
N.S.W
SYDNEY
N
200km
QUIRINDI DOME
N
QUIPOLLY DOME
CASTLE MTN DOME
15
20 km
KANKOOL
LIVERPOOL RANGE

widespread and voluminous pyroclastic horizons preserved in the Currabubula Formation. These occur as single flow units up to 60 m thick mapped throughout areas of 1200 km^2, compound (multiple flow units, Wright, 1981) ignimbrites, and as composite sheets (Smith, 1960a,b) comprising packets of petrographically similar ignimbrites reaching thicknesses of more than 150 m and presently extending over 2400 km^2. Because basal contacts of the ignimbrites are commonly covered, most exposures display only the homogeneous character of Layer 2b (Sparks *et al.*, 1973). A zone of concentration of lithic lapilli is present in the lowest exposed parts of some flow units (e.g. GR590486 Winton). Vertical zonation in welding, devitrification and lithophysae development is conspicuous. The ignimbrites are capped in places by finer grained and less welded horizons, or separated from overlying, locally-derived conglomerates by erosion surfaces.

Thin primary(?) ash-fall tuffs and non-welded ignimbrites also occur throughout the Currabubula Formation but are volumetrically subordinate to the welded ignimbrites. Discrete pyroclastic units are interbedded with conglomerates and sandstones, and comprise between 50 percent (west) and 10 percent (east) of sections through the Currabubula Formation (Fig. 2.2). However, an additional pyroclastic component is dispersed through the sequence as shards, crystal fragments and pumice incorporated in epiclastic lithologies.

In general, the original pyroclastic textures are well preserved though in most cases formerly glassy components (shards, pumice) have devitrified, or been replaced by zeolites probably developed in response to zeolite facies burial metamorphism (Wilkinson and Whetten, 1964). Though proportions and sizes of crystal fragments vary, the quartz + plagioclase ± K-feldspar ± biotite assemblage is ubiquitous (Table 2.1, Fig. 2.3). Only in a few instances can feldspars and ferromagnesian grains be considered fresh.

Distribution

Outcrop mapping of the Currabubula Formation at 1:31680 shows that it contains a pyroclastic stratigraphy comprising four members in the northern Werrie Syncline (Fig. 2.4). The regional extent of this internal pyroclastic record has been established and the area encompassed is

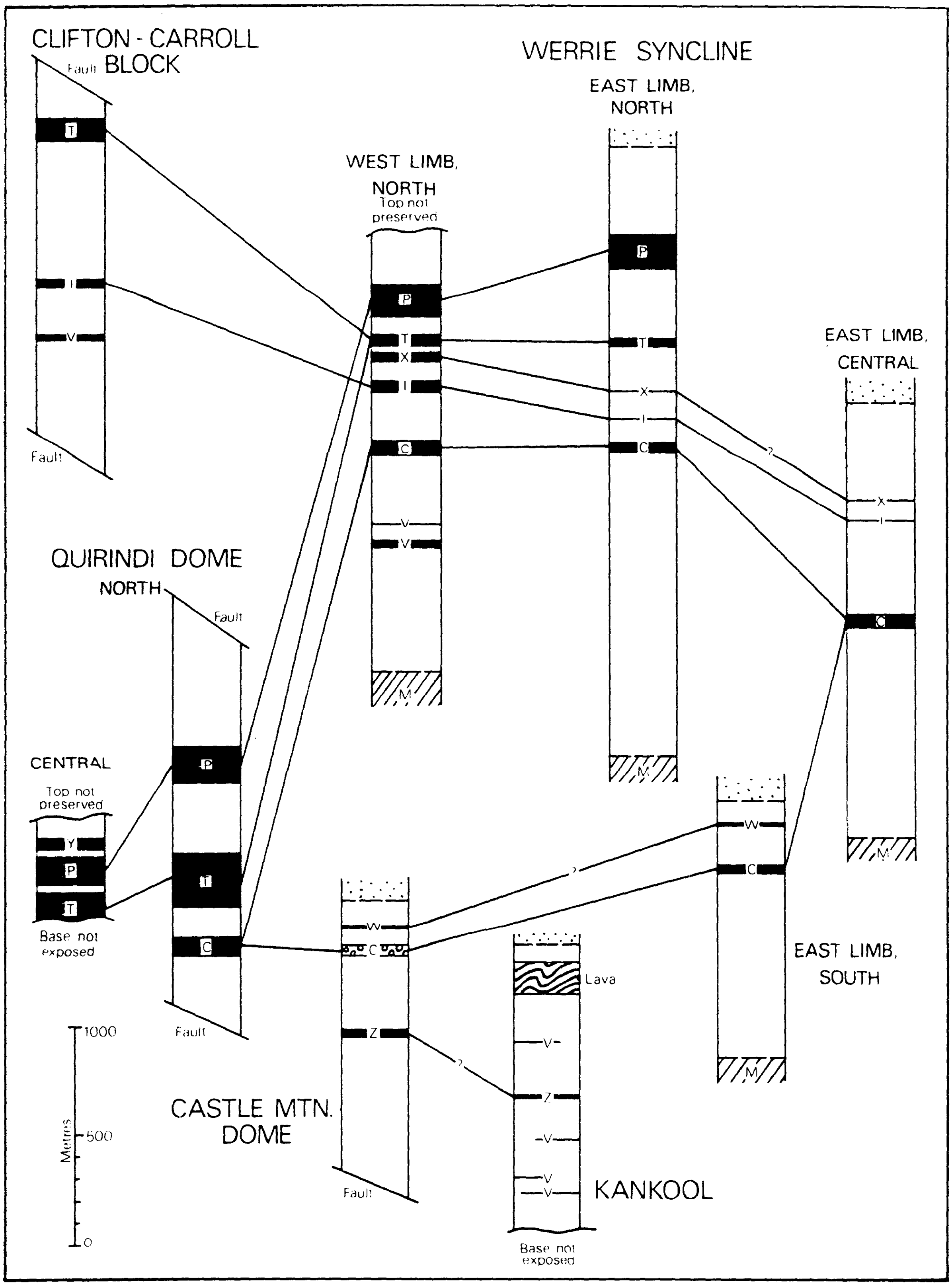


Figure 2.2: Correlations of pyroclastic members of the Currabubula Formation for areas between Carroll (north) and Kankool (south). C, Cana Creek Tuff Member; I, Iventure Ignimbrite Member; T, Taggarts Mountain Ignimbrite Member; P, Piallaway Trig Ignimbrite Member. V,W,X,Y,Z, locally mappable ignimbrite units. M, Merlewood Formation. Stipple, Permian strata.

TABLE 2.1: Averages of modes[1] of ignimbrites from the Currabubula Formation

	Crystal Fragments (%)					Lithics	Matrix[2]
	Quartz	Feldspar	Biotite	Opaques	Total	(%)	(%)
PTIM	3	22	2	1	28	Trace	72
TMIM	17	27	3	Trace	47	Trace	53
IIM	6	13	-	Trace	19	Trace	81
CCTM	3	4	-	Trace	7	Trace	93

[1]See also Table B.1, Appendix B.

[2]Includes relic pumice, shards and unresolvable felsic matrix.

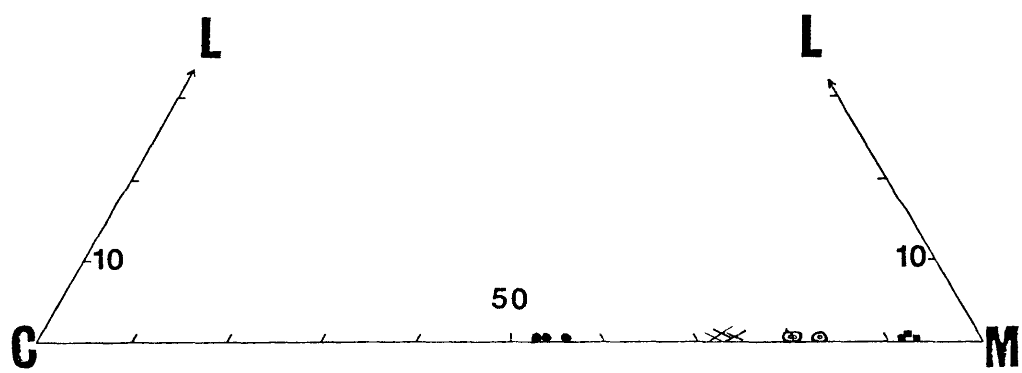


Figure 2.3: Components of representative samples from ignimbrite members of the Currabubula Formation. L, lithic fragments. C, crystal fragments. M, matrix including devitrified pumice, shards and fine ash.

- ■ Cana Creek Tuff Member
- ⊙ Iventure Ignimbrite Member
- x Piallaway Trig Ignimbrite Member
- ● Taggarts Mountain Ignimbrite Member

interpreted to be the remnant of an ancient ash-flow or pyroclastic field (Smith, 1960a, p.813), distinguishable from contemporaneous fields on its northern and southern flanks. The expanse so defined includes the following major structures: the Quirindi and Quipolly Domes; the eastern limb and closure of the Werrie Syncline; the Clifton-Carroll, Gunnan and Tulcumba Blocks; the southernmost Rocky Creek Syncline (Figs. 2.1,2.4, 2.5). Beyond these structures, coeval but independent volcanic stratigraphies are interleaved with the sedimentary facies of the Currabubula Formation and its equivalents. Twenty kilometres north of the Liverpool Range, the pyroclastic members of the Currabubula Formation terminate abruptly and primary volcanic horizons in general are an inconspicuous component of the sequence. Close to the Range at Kankool, pyroclastic units are again common but are typically fine grained, crystal-poor, feldspar- and biotite-bearing, non-welded, thin (less than 10 m) ignimbrites of limited extent (less than 3 km); a flow banded silicic lava also occurs in this area (Fig. 5.3). Hence, the southern limit to the distribution of pyroclastic members of the Currabubula Formation was static and sharply defined during the Late Carboniferous.

North of the Nandewar Range (Fig. 2.1) the Late Carboniferous formations also have an independent pyroclastic record, characterised by abundant but thin (less than 10 m), fine grained, quartz + feldspar ± biotite ± hornblende-bearing ignimbrites and ash-fall tuffs, only in few cases extending more than 5 km. One member of the Currabubula Formation pyroclastic stratigraphy (the Taggarts Mountain Ignimbrite Member) is present in exposures of the sequence between Carroll and the Nandewar Range. However, in this area there are also many ignimbrites having affinities with the northern Rocky Creek sequence, and not belonging to the pyroclastic stratigraphy recognised in the Werrie Syncline area. Away from this area to the north and to the south, the pyroclastic records become increasingly different and eventually independent. The transitional nature of the northern limit of the pyroclastic stratigraphy of the Currabubula Formation reflects fluctuation in the northward distribution of the members during the Late Carboniferous, and their unhindered access to the fringes of the adjacent pyroclastic field.

Members

Each of the pyroclastic members herein defined has distinctive and

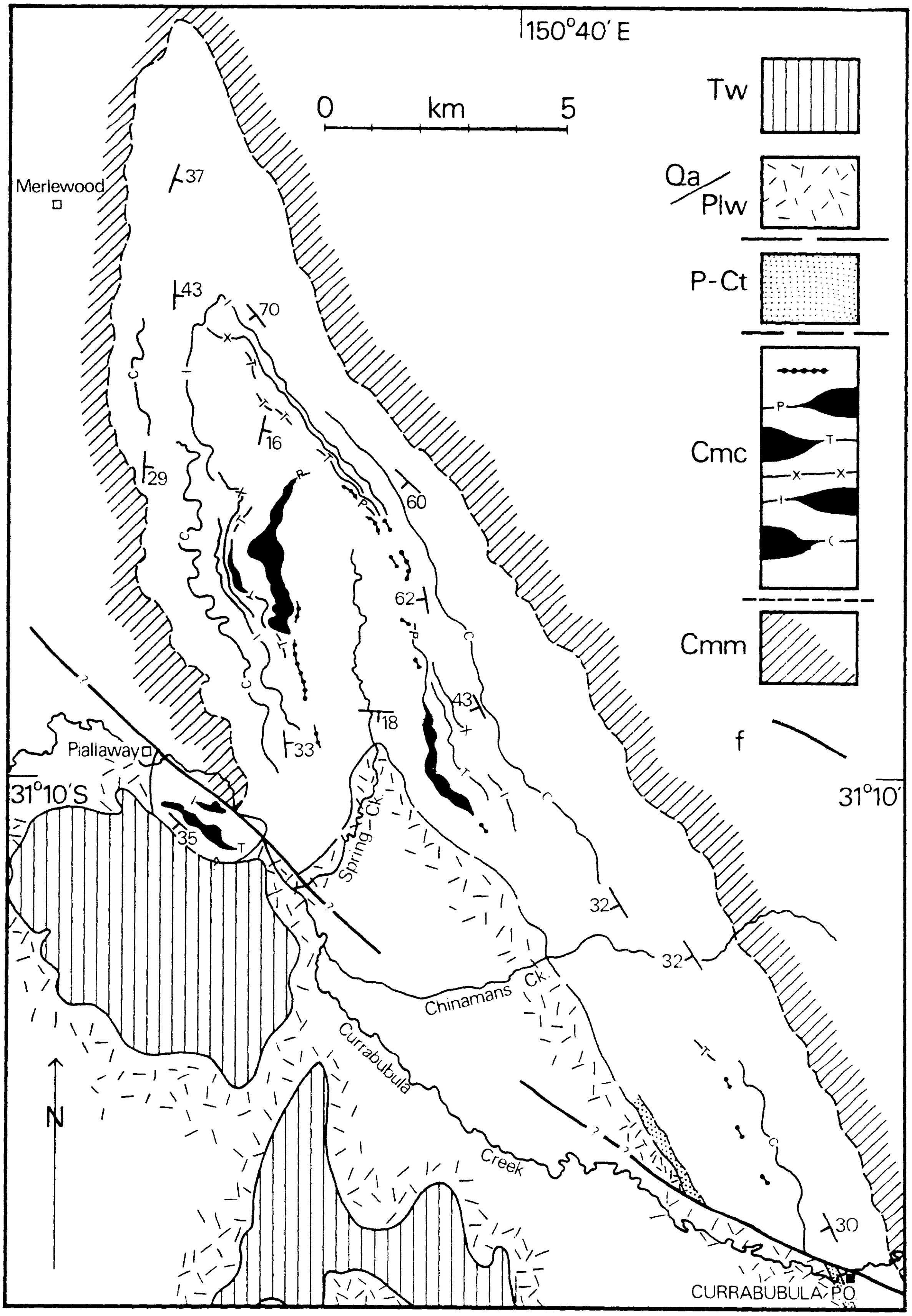


Figure 2.4: Outcrop map of pyroclastic members of the Currabubula Formation of the Werrie Syncline closure. Tw, Tertiary(?) Warrigundi Intrusions. Qa/Plw, Werrie Basalt and alluvium. P-Ct, Temi Formation. Cmc, Currabubula Formation; C,I,X,T,P, as for Figure 2.2; dots, locally mappable ignimbrite-clast conglomerate. Cmm, Merlewood Formation f, fault.

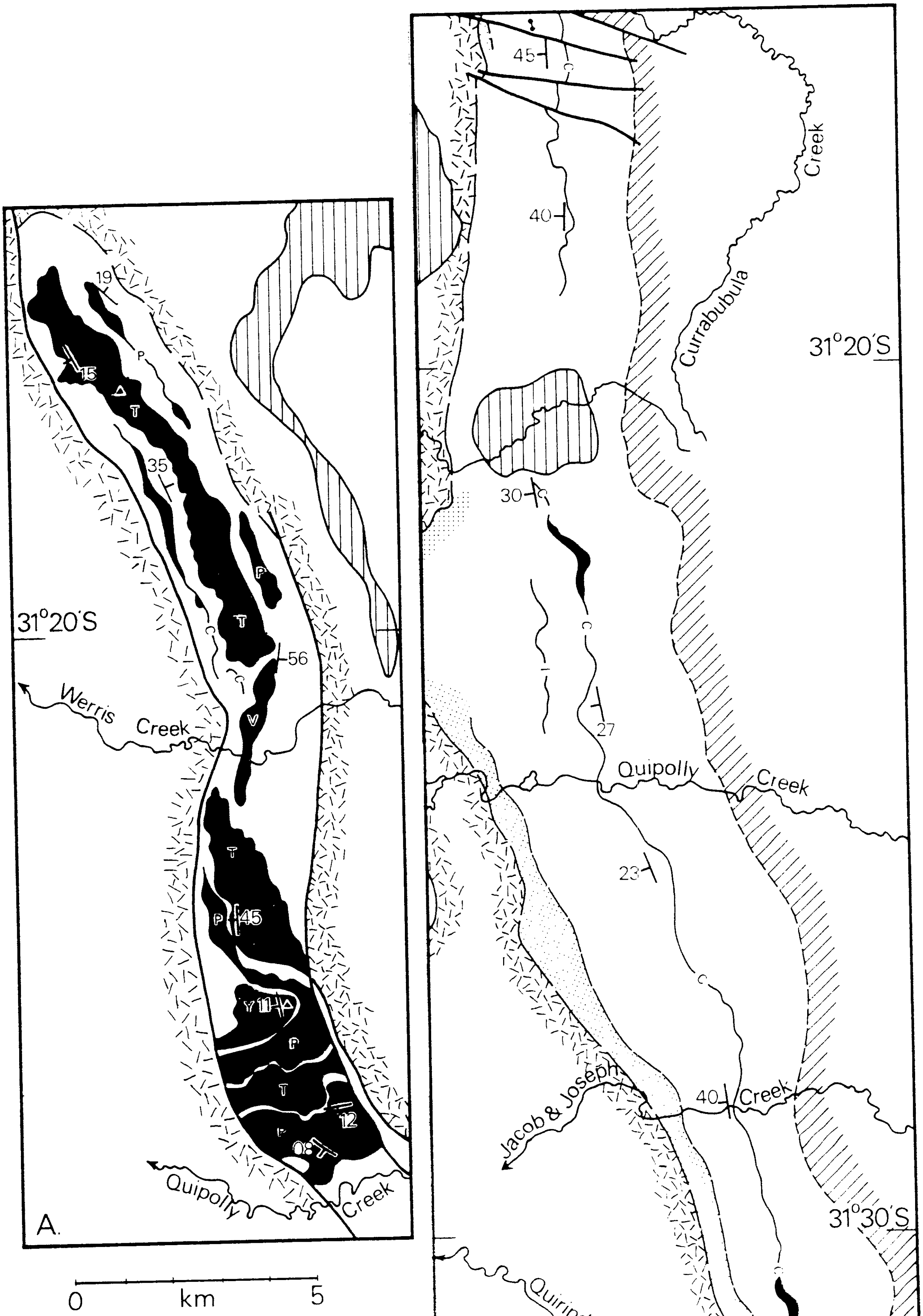


Figure 2.5: Outcrop map of pyroclastic members of the Currabubula Formation. (A) Northern Quirindi Dome. (B) East limb of the Werrie Syncline. Symbols as for Figures 2.2,2.4.

remarkably consistent outcrop features, enabling confident identification even where outcrop is interrupted. A traverse beginning approximately 2 km north of Piallaway homestead and proceeding northeast to Piallaway Trig (GR592486 Winton) includes adequate exposures of all the pyroclastic members.

The Cana Creek Tuff Member (CCTM) comprises non-welded ignimbrite interbedded with ash-fall tuff and compositionally-similar volcaniclastic deposits, in a discrete unit 30 to 100 m thick. Cliff exposures are typically stratified (Fig. 2.6a). The member is characterised by relic pumice lenticles (lapilli) which are now dark green, and by abundant accretionary lapilli. Other components are quartz, K-feldspar and plagioclase crystal fragments, angular lapilli of volcanic rocks, and fine grained, pale green ash(?) matrix. In thin-section, shards are undeformed and circular shapes of slices through spherical bubbles are common.

Although this member is laterally persistent, the arrangement of internal facies is variable. In general, the primary pyroclastic intervals occur lowermost, or are sandwiched between crystal-rich (quartzofeldspathic) pumiceous sandstone and granule conglomerate. The reworked component predominates in easternmost sections whereas primary pyroclastic rocks are thickest farther west. Some of the eastern (more distal) exposures of the pumiceous granule conglomerate of the CCTM contain dispersed angular and subrounded lithic fragments up to 20 cm across in very thick, massive beds. Neither gas escape structures nor basal layers (Layer 2a of the normal sequence in ignimbrite flow units, Sparks *et al.*, 1973) have been identified in these outcrops. The size and lack of grading of clasts suggest that these intervals are debris flow deposits, although it is possible that they are ignimbrites formed by poorly fluidised pyroclastic flows (*cf.* Wilson, 1980, type 1). Other pumiceous and crystal-rich sandstones and granule conglomerates which are planar bedded and internally massive, may have been generated by sheetwash flooding events.

The CCTM has features normally found in the deposits from eruptions affected by external water, and is very similar in character, composition and volume to the phreatomagmatic Wairakei Formation of Pleistocene age in New Zealand (Self and Sparks, 1978; Self, 1983). A hydrovolcanic origin for the CCTM is suggested by the widespread accretionary lapilli, the association of primary with redeposited facies, the lack of welding of the ignimbrite (*cf.* Self, 1983; Walker, 1983) and the fine grain size of ignimbrite and ash-fall tuff (*cf.* Self and Sparks, 1978). The eruption sequence and emplacement processes interpreted from logged sections of the CCTM are discussed in detail in Chapter 3.

The Iventure Ignimbrite Member (IIM), some 60 m thick, forms a continuous cliff line around the closure of the Werrie Syncline, and also in the adjacent Clifton-Carroll Block, though remnants occur over a much broader north-south span (Figs. 2.4,2.5B). Outcrops have a mottled appearance due to the pronounced development of spherulites and lithophysae, and show blocky bedding-normal jointing (Fig. 2.6b,c). Original components are crystal fragments (quartz + feldspar ± biotite), relic welded shards, and pumice lapilli flattened parallel to bedding. Spherulites, 1 to 2 cm diameter, and lithophysae, up to several centimetres across, are concentrated near the bases of most exposures in a zone at least 5 m thick. These devitrification effects, and the size and proportion of crystal fragments decline upwards to a fine grained pumice-flecked horizon. Contacts with overlying conglomerates are erosional, so the original full sequence and thickness of the ignimbrite are not known. The distinctive devitrification effects probably developed soon after emplacement and welding of the single cooling unit preserved and may imply that relatively high residual volatiles still remained in the essential components at this stage (Sparks *et al.*, 1978). Steam vapourised from groundwater trapped beneath the hot ignimbrite could have generated, or at least contributed to, the same effects (*cf.* Mahood, 1980).

The Taggarts Mountain Ignimbrite Member (TMIM) consists of coarse grained, relatively crystal-rich welded flow units. Crystal fragments of quartz, feldspar and biotite comprise approximately 40 to 50 modal percent (Table 2.1, Fig. 2.6d). The thickest and lowermost flow units commonly have a black, glassy, densely welded base 2 to 3 m thick. In the succeeding main devitrified portion, subtle variations in crystal content, grain size and welding have been detected but not systematically mapped. Flattened relic pumice and angular lithic lapilli are only locally conspicuous.

Westernmost exposures of the member in the Quirindi Dome comprise at least three flow units, separated by thin (less than 2 m) pockets of lithic sandstone and granule conglomerate, amounting to approximately 200 m total thickness. In more easterly exposures (Werrie Syncline closure, Clifton-Carroll Block), the member is represented by a single flow unit 50 m thick. The member has been traced northwestward through the structurally isolated Gunnan and Tulcumba Blocks (Fig. 2.1). However, in these areas steeply-dipping, strike-parallel, reverse faults disrupt the sequence so the number, thickness and internal stratigraphy of flow units is imprecisely known. Nevertheless, this member clearly extends to the southernmost Rocky Creek Syncline (Fig. 2.1) where it has been mapped by White (1965) and named the Plagyan Rhyolite.

The Taggarts Mountain Ignimbrite Member is considered to be the remnant of a composite sheet (*cf.* Smith, 1960a,b; Christiansen, 1979) and records a major period of eruption during which the production of widespread voluminous pyroclastic flows was punctuated by quiescent intervals allowing limited erosion, reworking and sedimentation, at least in areas remote from source.

The <u>Piallaway Trig Ignimbrite Member</u> (PTIM) is a compound ignimbrite sheet, up to 180 m thick and typically densely welded throughout. Relic pumice is recognisable as extremely flattened, essentially two-dimensional plates, resulting in pervasive bedding-parallel parting (Fig. 2.6e). Disrupted and warped foliation occurs locally where the unit is thickest (Piallaway Trig exposures) and may be a rheomorphic effect (*cf.* Wolff and Wright, 1981). In thin-section, welded devitrified shards show marked deformation around crystal fragments (quartz + feldspar + biotite) and sparse angular lithic lapilli.

The best exposures of this member occur in the closure of the Werrie Syncline, though it outcrops throughout the Quirindi and Quipolly Domes as well. At least three flow units are present, delineated by breaks in the bedding-normal jointing, concentrations of angular lithic lapilli, thin finer-grained crystal-poor horizons, and/or black glassy (basal?) zones. Lateral fluctuations in thickness are the result of the topographic relief of the depositional site at the time of emplacement and later erosion of the originally more extensive sheet (Fig. 2.7), as has been reported for modern ignimbrites with marked thickness variations (e.g. the upper member of the Bandelier Tuff, Smith and Bailey, 1966; the Ito pyroclastic flow, Yokoyama, 1974; the Whakamaru Ignimbrite, Briggs, 1976).

Other Ignimbrites: In each of the areas mapped, the regionally extensive members described above are accompanied by thin (about 5 m) typically fine grained, crystal-poor (quartz + feldspar) ignimbrites comprising single flow units, either welded or non-welded. Some of these are of very limited extent (less than 1 km along strike). Others have been traced within the structurally isolated blocks but are not sufficiently distinctive to be confidently linked with units in nearby blocks. Such nondescript ignimbrites are much more common in the western and north-western areas (Quirindi Dome, Clifton-Carroll, Gunnan and Tulcumba Blocks). Glassy black horizons, associated with the thickest examples of these minor ignimbrites are similarly restricted to the westerly exposures. Collectively the minor ignimbrites account for less than about a fifth of the total volume of ignimbrites preserved in the Currabubula Formation.

SETTING FOR IGNIMBRITE EMPLACEMENT

Sedimentary facies

Cobble orthoconglomerates are dominant in the Currabubula Formation, and accompanied by lesser coarse to medium sandstones and minor mudstones. Beds of conglomerate range up to about 5 m in thickness and may be massive but more typically display a crude internal stratification (Fig. 2.6f). Most sandstone is confined to lensoid beds less than a metre thick within conglomerate, although there are sandstone or pebbly sandstone intervals several metres thick but of limited lateral extent. Cross-bedding in sets a metre thick is a common feature of sandstone. Mudstone forms relatively thin (less than 10 m), discrete, laterally discontinuous layers which are either laminated (Fig. 2.6g) or massive. Laminated pebbly mudstone and poorly sorted, polymictic, massive paraconglomerate have been interpreted as glacial varves and tillite respectively (Carey, 1937; Whetten, 1965). Plant fossils of the *Rhacopteris* flora, and the invertebrate trace fossil *Isopodichnus* sp. are abundant in some mudstone units. A continental depositional setting controlled mainly by gravel-dominated fluvial processes,

Figure 2.6: a. The cliff-forming Cana Creek Tuff Member of the Currabubula Formation on the western limb of the Werrie Syncline; view to the south from GR581471 Piallaway. Cliff height approximately 35 metres.

b. Typical blocky-jointed cliff of the Iventure Ignimbrite Member, western limb of the Werrie Syncline (GR583482 Piallaway). Hammer 33 cm.

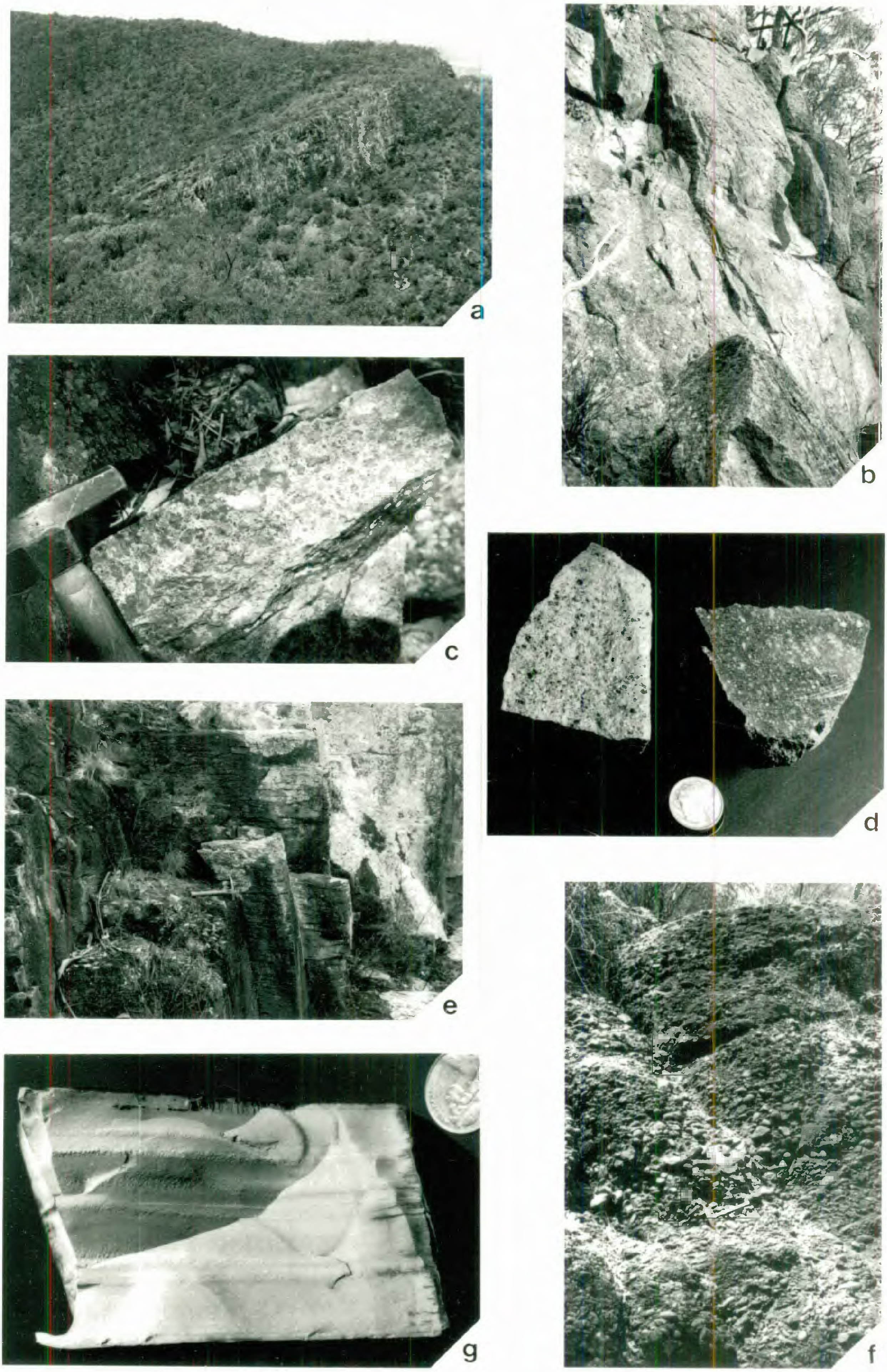
c. Spherulitic devitrification effects in the Iventure Ignimbrite Member. Same locality as b.

d. Hand specimens of the crystal-rich Taggarts Mountain Ignimbrite Member. Left, R55210, right, R55216. Coin diameter 27 mm.

e. Well-developed platy pumice foliation in blocky-jointed Piallaway Trig Ignimbrite Member, western limb of the Werrie Syncline (GR594488 Winton). Hammer 33 cm.

f. Typical crudely stratified, cobble orthoconglomerate of the Currabubula Formation (GR579468 Piallaway). Hammer 33 cm.

g. Hand specimen of laminated mudstone of the Currabubula Formation (R55324, from GR763059 Emblem). Coin diameter 27 mm.


a
b
c
d
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g

and intermittently affected by glacial processes, is evident (Whetten, 1965).

Palaeoslope

Thickness trends revealed by the correlation of pyroclastic stratigraphies (Fig. 2.2) indicate a western source and eastward tilted palaeoslope for both the pyroclastic and sedimentary components of the Currabubula Formation. The proportion of sedimentary rock intercalated with the pyroclastic members increases eastward, whereas the members themselves thin gradually in the same direction. Ignimbrites other than the mappable members are most common and in thicker beds in the westernmost exposures. In addition clast imbrication in conglomerates and cross-bedding in sandstones suggest that alluvial debris was fed to the fluvial system from the west (White, 1968).

Palaeovalleys and disconformities

The ignimbrites of the Currabubula Formation provide extensive time planes with which the balance and scale of aggradation and erosion can be monitored. Conglomerates derived from local erosion of adjacent or underlying ignimbrites are common (e.g. GR584494 Winton). Furthermore, outcrop mapping of the members has revealed much larger scale disconformities and palaeovalleys within the sequence. Such features are best displayed on the east limb of the Werrie Syncline south of the closure, where each of three mappable ignimbrites is successively truncated by the PTIM, or by distinctive ignimbrite-clast-bearing conglomerates (Fig. 2.4). Higher in the section the PTIM is also replaced along strike and overlain by conglomerates derived from its erosion. Figure 2.7 shows the development of the stratigraphic pile incorporating the successive superimposed erosional disconformities so delineated. The site was evidently a persistent discrete palaeovalley, periodically choked by ignimbrites and cleared by fluvial erosion, restoring the pre-existing drainage pattern. Such repeated infilling and excavation of the same valley has been identified in modern ignimbrite outflow sheets, for example, the Bandelier Tuffs of New Mexico (Smith and Bailey, 1966), and is reported to have been common during emplacement of the Quaternary ignimbrites erupted from Vulsini

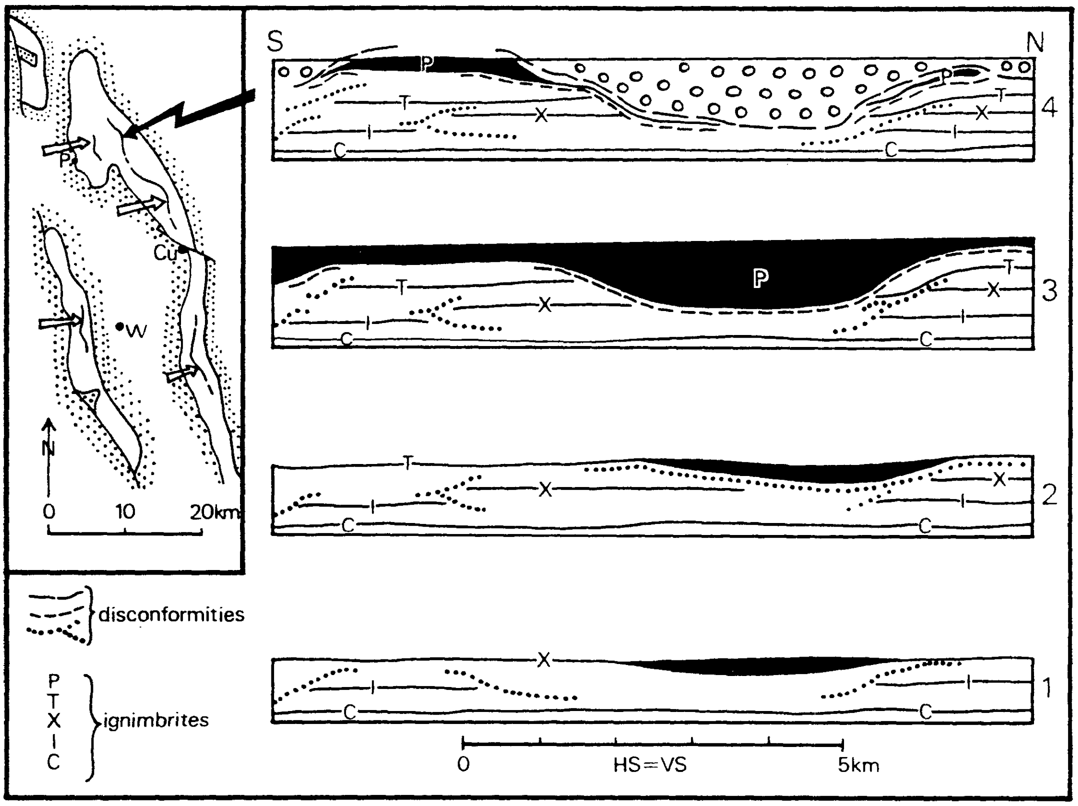


Figure 2.7: Schematic sequential cross-sections showing the superimposed disconformities of a palaeovalley site on the east limb of the Werrie Syncline south of the closure. Ignimbrite members are denoted by letters as for Figure 2.2. The sequence begins (1) after emplacement of ignimbrite X, with at least one older disconformity responsible for partial removal of the Iventure Ignimbrite. The palaeovalley was cleared then filled by the Taggarts Mountain Ignimbrite (2). These events were repeated (3) with a sheet of the Piallaway Trig Ignimbrite lining the restored post-Taggarts Mountain Ignimbrite valley. At this stage another distributary is apparent to the south, but it could be as old as the better defined one shown. The palaeovalley was reinstated (4), on removal of the Piallaway Trig Ignimbrite, and is now occupied by conglomerates (shown by circles), some with abundant clasts of this member. The map inset indicates the sites of this and other palaeovalleys within the Currabubula Formation, as mentioned in the text. Cu, Currabubula. P, Piallaway. W, Werris Creek.

Volcano in Italy (Sparks, 1975).

To the west-southwest on the western limb of the Werrie Syncline near Piallaway, the up-palaeoslope, less-deeply incised extension of this palaeovalley has been detected with the pre- and post- PTIM erosional surfaces recognisable (Fig. 2.4). A disconformity in the northern Quirindi Dome is probably the same age as the post-PTIM surface found at the two Werrie Syncline sites (Fig. 2.5A).

Perhaps the most substantial and persistent area of contemporaneous erosion existed easternmost, and is now preserved along the east limb of the Werrie Syncline from near Chinamans Creek for 35 km south to Quirindi Creek (Figs. 2.4,2.5B). The oldest member, the Cana Creek Tuff Member, is the only continuous representative of the pyroclastic stratigraphy, though in places it, too, is interrupted for some 4 km. Since the full pyroclastic record is exposed due west in the Quirindi Dome, and also to the north, more distal representatives of the ignimbrites were undoubtedly original components of the sequence, but obliterated as coherent units by erosion. The distal and thinner reaches of these Carboniferous ignimbrites would have been non-welded, as are the margins of modern ignimbrite sheets (Smith,1960b; Smith and Bailey, 1966), and less resistant to reworking by the active fluvial processes of the environment of emplacement than welded portions farther west. Without pyroclastic markers, the detailed development of this area cannot be resolved. However, using relationships demonstrated to the north, it is likely that several superimposed disconformities exist in the sequence, rather than a single, major, post-PTIM surface.

PROVENANCE OF THE CURRABUBULA FORMATION

A source area dominated by eruptions from silicic volcanic centres is clearly indicated by both the pyroclastic and sedimentary facies of the Currabubula Formation. The predominant components of sandstones and of the matrices of conglomerates are felsic volcanic rock fragments, grains of volcanic quartz and feldspar (plagioclase and K-feldspar, including sanidine), with traces of biotite and amphibole in some cases. Devitrified shards are in general minor in sandstones but more abundant in the mudstones. Pumice is also a locally conspicuous but subordinate component of sandstones. Clasts in conglomerates are dominantly quartz ± feldspar-bearing volcanics, notably welded ignimbrites, flow banded lava, and fine porphyries of uncertain affinity. Ignimbrite clasts in some conglomerates are petrographically and texturally identical to ignimbrite units of the Currabubula Formation. Clasts of pyroxene- or amphibole-bearing feldspar porphyries (probably lavas) are more common in conglomerates of the lowermost Currabubula Formation and the underlying Merlewood Formation, than in younger horizons. Though the volcanogenic nature of the sedimentary facies primarily reflects contemporaneous silicic volcanism, a contribution from erosion of older inactive volcanic piles is probable.

Non-volcanic detritus consists of silicic plutonic and regional metamorphic (slate, phyllite, gneiss) rocktypes. These are conspicuous in paraconglomerates, though they occur in orthoconglomerates as well. A radiometric K-Ar age of 356 ± 6 Ma has been obtained from hornblende in a granitoid clast (J. Roberts, written comm., 1983;GR748156 Emblem). The nearest exposures of granitoids of this age occur more than 110 km to the southwest in the Lachlan Fold Belt, though intervening strata of the Permo-Triassic Sydney Basin possibly obscure closer plutons. Clasts of indurated quartz sandstones and quartzites are ubiquitous. One such clast found contains Devonian or older spiriferid brachiopods (R55326; identified by J. Roberts, University of New South Wales; pers. comm.). The geological context of the active volcanic edifice was evidently a silicic plutonic and regional metamorphic terrain of probable late Devonian and older age.

CHARACTER OF SOURCE VOLCANOES

Modes of emplacement, volumes, distribution and compositional affinities of the pyroclastic units of the Currabubula Formation collectively provide constraints on the character, scale and eruptive style of the source volcanic centres (Table 2.2). These widespread, welded silicic ignimbrites are outflow sheets produced by pyroclastic eruptions probably associated with episodic foundering and subsidence of large-scale volcanic calderas (*cf*. Ross and Smith, 1961; Smith, 1979). Recent models of pyroclastic eruptions account for such voluminous deposits by continuous collapse of a sustained vertical eruption column (Sparks and Wilson, 1976; Sparks *et al.*, 1978). Ash clouds associated with ignimbrite-forming eruptions and with the moving pyroclastic flows (*cf.* Sparks and Walker, 1977; Fisher, 1979), probably contributed the shards and pumice found dispersed throughout the sedimentary facies of the Currabubula Formation. These components would also have been derived from reworking of non-welded ignimbrites and the non-welded tops and margins of ignimbrites. Less commonly, airfall ash is preserved in discrete units of lacustrine tuffaceous mudstone.

While active, the source for the pyroclastics of the Currabubula Formation was a terrain perhaps comprising overlapping and nested calderas amid coalescing fans of pyroclastic debris, the details of which are now lost. By analogy with modern major centres of silicic pyroclastic volcanism in the western United States (e.g. Valles; Smith and Bailey, 1966), Mexico (e.g. La Primavera; Mahood, 1980; Walker, Wright *et al.*, 1981), the North Island of New Zealand (e.g. Taupo; Walker, 1980), Italy (e.g. Vulsini; Sparks, 1975) and South America (e.g. Cerro Galan; Francis *et al.*, 1978), these Carboniferous calderas were probably multiple-vent systems with low profiles ringed by broad, topographically subdued flanks composed of flat-lying to gently inclined pyroclastic aprons.

The actual distance of the present exposures in the Currabubula Formation from the source calderas can only be roughly approximated. The absence of coherent rhyolitic lava flows or domes, and of proximal facies such as co-ignimbrite lag breccias (Wright *et al.*, 1981; Druitt and Sparks, 1982) in the ignimbrite members, and their intercalation within a regionally extensive sedimentary sequence, imply that the preserved record was emplaced in a setting remote by at least several kilometres from eruption sites. The persistent well-developed welding and marked westward thickening of some members are consistent with a setting in the medial portions of the outflow ring plain. In general lithic clasts in the ignimbrites are uncommon, and less than 2 cm across in eastern exposures and less than 4 cm in western exposures. The grain size versus distance from source graphs for ignimbrites of Japan (Kuno *et al.*, 1964) and Italy (Sparks, 1975) indicate that clasts of this size can be transported at least 40 km from vent. In addition to the 25 to 30 km travelled by all the major ignimbrites in spanning the preserved east-west exposure of the Currabubula Formation, they may have flowed several kilometres and probably as much as a few tens of kilometres from source calderas before reaching the western edge of their present extent.

Accretionary lapilli help constrain vent position only if produced at vent and undisturbed after deposition by airfall (*cf.* Moore and Peck, 1962). Even so, accretionary lapilli of such an origin may be so widely distributed as to offer little further refinement of source vent remoteness (e.g. those in phreatoplinian ashes; Self and Sparks, 1978; Walker, 1981a; Self, 1983). Neither are other models of formation of accretionary lapilli sensitive to primary vent location. Accretionary lapilli formed by rain flushing of co-ignimbrite ash clouds (Sparks, 1976; Sparks and Walker, 1977), by gas streaming in fumarolic pipes within the upper portions of ignimbrites (Walker, 1971), or by phreatic eruptions at secondary vents hold no direct clues as to the site of eruption of the associated pyroclastic flows. Since the accretionary lapilli of the CCTM involve at least two of the above processes their presence provides at best only very broad limits on the distance travelled from source volcanoes.

Although the emplacement of each ignimbrite member was effectively instantaneous, they occur intermittently throughout the post-Visean sequence during a time interval of 25 to 30 Ma. Each major ignimbrite is considered symptomatic of a separate "caldera cycle" (Christiansen, 1979, p.38), involving evolution of a single magma body at shallow crustal levels over a period of the order of a million years. The most complicated of these cycles was that responsible for the Taggarts Mountain Ignimbrite Member, which records several eruption pulses punctuated by recognisable quiescent spells. Christiansen (1979, p.39) relates such protracted cycles to large-volume magma chambers which typically involve ring-fracture vent systems. Furthermore, on a regional scale, the pyroclastic stratigraphy of the Currabubula Formation preserves the remnants of a major pyroclastic field which is distinguishable from those which occur to the north and south, reflecting the existence of coeval but independent magma sources.

The voluminous silicic ignimbrites of the Currabubula Formation constrain the tectonic regime of the source calderas. This style and scale of volcanism is typical of areas underlain by comparatively thick crust (at least 25 km, usually more than 35 km) of continental rather than oceanic character (Coulon and Thorpe, 1981); that is, active continental margins, epicontinental settings (Smith, 1979), "mature" island arcs (Green, 1980), or sites where oceanic arcs impinge on continental crust (Coulon and Thorpe, 1981; Leeman, 1983). Geochemical data for the Currabubula Formation ignimbrites (Chapter 4) and for other Carboniferous volcanic rocks of the Tamworth Belt (Wilkinson, 1971) indicate their overall calc-alkaline character. Provenance considerations (above) require proximity of the source calderas to a region composed of Devonian, and possibly older, plutonic/metamorphic/sedimentary rocks. The caldera terrain evidently bordered the northern part of the Lachlan Fold Belt which was largely stabilised by the Early Carboniferous (Packham, 1969), and is now hidden by Permian and younger basin strata. During the Late Carboniferous, Devonian-Carboniferous plutons, quartz-rich sedimentary rocks and older deformed sequences of the northern Lachlan Fold Belt must have been exposed and sufficiently elevated to shed debris eastward to the conglomerate-ignimbrite alluvial apron. Such an arrangement is consistent with published reconstructions of the Late Carboniferous tectonic framework (Leitch, 1974,1975; Day *et al.*, 1978) and implies support for an active continental margin setting over the other alternatives cited above. A modern analogue for this configuration is described in Chapter 4.

CONCLUSIONS

The Currabubula Formation of northeastern New South Wales includes the record of activity of a Late Carboniferous silicic volcanic terrain spatially associated with an older, partly glaciated (Whetten, 1965) highland. The ignimbrite members are outflow sheets produced by explosive pyroclastic eruptions, emanating from calderas originally situated at least several and possibly tens of kilometres to the west of present exposures, and monitor at least 4 major caldera cycles (Christiansen, 1979). Pyroclastic flows periodically inundated and partly buried the alluvial lowland beneath blankets of ignimbrite and airfall ash, much of which was eroded and redistributed locally. Disconformities and palaeovalleys revealed by pyroclastic units are common throughout the Currabubula

Formation and attest to the prevailing complex interplay between epiclastic sedimentation, the emplacement of ignimbrites and contemporaneous erosion.

Coeval sequences north and south of the Currabubula Formation have independent pyroclastic records, and were part of largely separate pyroclastic fields. The calderas supplying pyroclastic debris to the Currabubula Formation can thus be distinguished as belonging to one of at least three major source regions within a meridional volcanic belt which dominated the Late Carboniferous palaeogeography.

TABLE 2.2: Dimensions of ignimbrites of the Currabubula Formation[1]

Unit:	PTIM	TMIM	Plagyan Rhyolite[3]	IIM	CCTM
Average thickness (m)	160	180	180	50	70
Estimated preserved area (km^2)	806	1480	1238	1375	1375
Estimated volume[2] (km^3)	129	266	223	69	96
Source character:	Volcano-tectonic depressions, collapse calderas, ring-complexes. (Smith, 1960a, 1979).				

[1] Figures given are based on preserved extents and are minima.

[2] For the welded units (all but CCTM), these values are approximately equal to dense rock equivalent volumes.

[3] After White (1965).

CHAPTER 3

PRIMARY AND REDEPOSITED FACIES FROM LARGE MAGNITUDE, EXPLOSIVE, RHYOLITIC HYDROVOLCANISM : CANA CREEK TUFF, LATE CARBONIFEROUS, NORTHEASTERN NEW SOUTH WALES

INTRODUCTION

The Cana Creek Tuff (McPhie, 1983; Chapter 2) is the most widespread and internally complex pyroclastic member of the Late Carboniferous Currabubula Formation in northeastern New South Wales (Fig. 3.1). It is one of a series of rhyolitic outflow ignimbrite sheets interbedded with conglomerate, but differs from the other ignimbrites in the following respects : it comprises ignimbrite and ash-fall tuff which are intimately associated with crystal-rich sandstone and pumiceous granule conglomerate; the primary pyroclastic rocktypes are exceptionally fine grained and accretionary lapilli-bearing; the ignimbrites were entirely non-welded and made of several, thin flow units. The Cana Creek Tuff Member is similar in lithofacies, composition and volume to the 20,000 year old Wairakei Formation of New Zealand, considered to be the product of a large-scale phreatomagmatic eruption (Self and Sparks, 1978; Self, 1983). The influence of water on the eruption and emplacement of the Cana Creek Tuff Member was recognised during initial investigation of the ignimbrites of the Currabubula Formation (McPhie, 1983). That this eruption was hydrovolcanic, involving the interaction of magma with external water (Sheridan and Wohletz, 1981), has proven an effective explanation of many aspects of the geology of the Cana Creek Tuff and its marked contrasts with the other ignimbrite members in the Late Carboniferous sequence.

The account of the eruption and emplacement of the Cana Creek Tuff presented here is inevitably qualitative and handicapped by the lack of systematic granulometric data obtainable from unconsolidated aggregates. Neither is there any precise control on source vent proximity, and the accessible parts of the unit constitute only an incomplete and nct necessarily representative sample of the original distribution. The existing stratigraphy reflects not only the volcanic events at the source but also those processes operating en route and at the site of emplacement, which diversify the final record of the eruption. To a lesser extent these complications hamper studies of Quaternary and even modern day silicic volcanoes, especially those of the 'inverse' (Walker, 1981e) or caldera type, so there is some justification for attempting the exercise in this instance.

REGIONAL GEOLOGICAL CONTEXT

The Late Carboniferous Currabubula Formation is principally composed of volcanogenic braidplain conglomerates. It is between 1500 to 3000 m thick where exposed in the north-northwest trending Werrie Syncline (Carey, 1934) and related fault-bounded blocks along the western margin of the New England Orogen (Day *et al.*, 1978; Fig. 3.1). The outcrop area of the Currabubula Formation is a narrow belt 25 km across and 120 km long but other formations equivalent in age and facies continue for another 130 km to the limits of exposure in a northerly direction, and 160 km southeastward to the coast near Newcastle. Sequences of fluvial sedimentary rocks interbedded with volumetrically subordinate, thin and widespread silicic ignimbrite sheets are typical of the entire belt but in places there is considerable complexity and variation in facies (e.g. Roberts and Engel, 1980).

There are five mappable ignimbrite units in the Currabubula Formation in the closure of the Werrie Syncline, some of which also occur in nearby but structurally isolated areas and along the east limb of the syncline (Chapter 2, Figs. 2.2,2.4,2.5). The Cana Creek Tuff Member is exposed with only minor interruptions for 55 km north to south and at each of the eastern and western extremities of the Currabubula Formation (Fig. 3.1). The sites of eruption of the pyroclastic units are not exposed but were probably located a few tens of kilometres westward from existing outcrops (Chapter 2). The volcanic source terrain included large-scale ignimbrite shields and calderas which were part of a glaciated highland (Whetten, 1965) that provided detritus for the epiclastic facies of the Currabubula Formation. Intermittent glacial expansion is recorded by thin intervals of glaciolacustrine mudstone with ice-rafted dropstones (Whetten, 1965), and possible ablation tillite (White, 1968; Lindsay, 1969). Disconformities

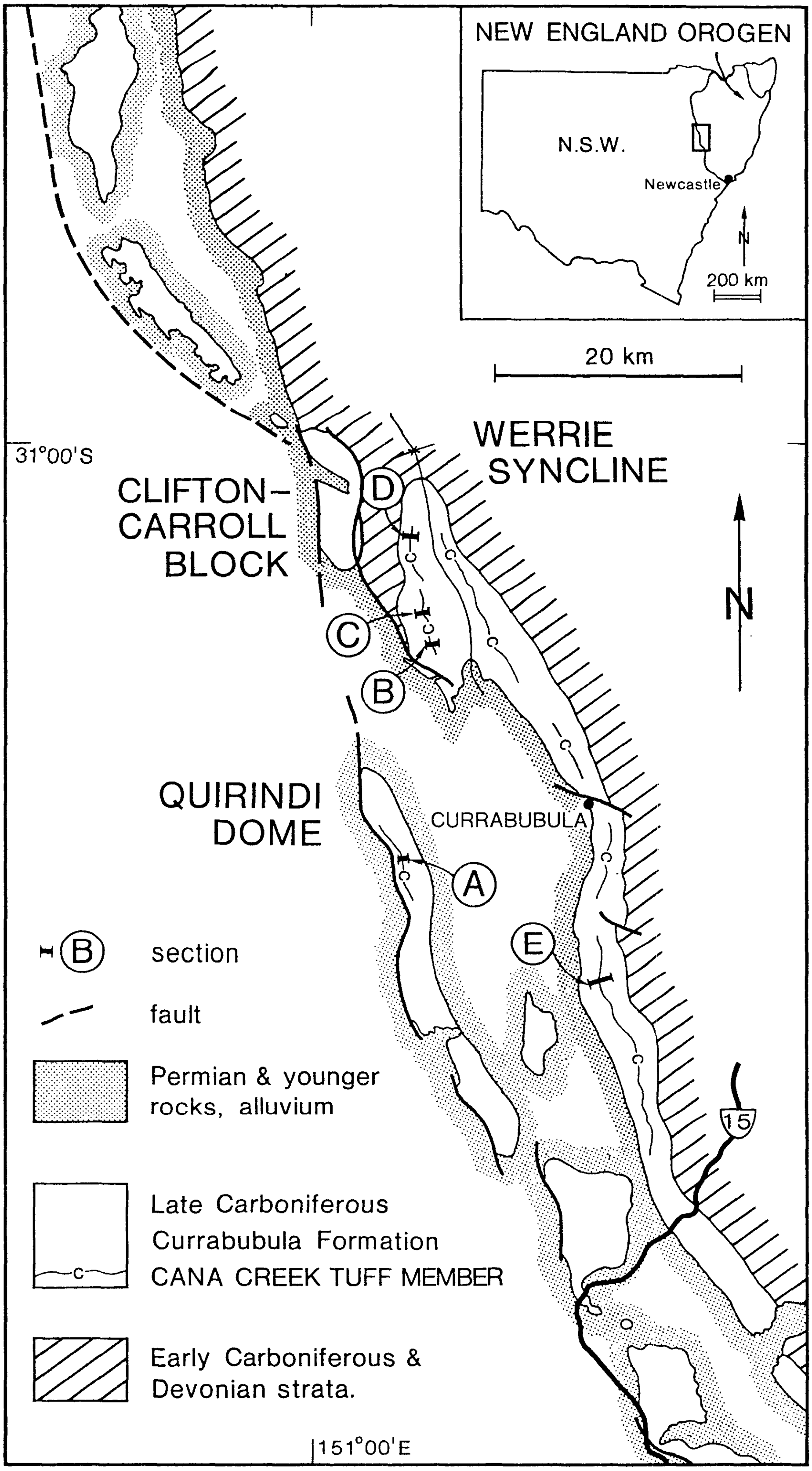


Figure 3.1: Geological setting of the Cana Creek Tuff Member of the Late Carboniferous Currabubula Formation, northeastern New South Wales, following Pogson and Hitchins (1973).

at a range of scales are common within the Currabubula Formation and none of the ignimbrite members still exhibits original thickness or extent.

GENERAL GEOLOGY OF THE CANA CREEK TUFF MEMBER

The Cana Creek Tuff Member is made up of primary pyroclastic rocktypes (ignimbrite and ash-fall tuff), enclosed by an envelope of pumiceous and crystal-rich sandstone and granule conglomerate. For the purposes of this discussion, the former will be referred to as the pyroclastic facies, and the latter as the volcaniclastic facies, of the Cana Creek Tuff. The following description is based primarily on data from stratigraphic sections at five sites chosen for their superior exposure (Figs. 3.1,3.2), supplemented by observations recorded in the course of mapping the member.

Thickness, distribution and volume

The aggregate thickness of the Cana Creek Tuff Member ranges from as little as 30 m to more than 100 m. Relative proportions of the two facies changes most markedly from west to east : thick sections along the east limb of the Werrie Syncline are composed mainly of the volcaniclastic facies whereas in the western thick section of the Quirindi Dome, the pyroclastic facies predominates. Particularly near the top of the member, definition of its limits is in places imprecise, as contacts of the volcaniclastic facies with other sedimentary rocks of the Currabubula Formation are typically gradational over thicknesses up to several metres.

The area encompassed by the present limits of outcrop of the Cana Creek Tuff Member in the Werrie Syncline and Quirindi Dome is approximately 1375 km² (Chapter 2, Table 2.2). This estimate is a minimum because neither the proximal nor the distal parts of the original deposit are represented in existing exposures. A reconstruction based on outflow from source of a few tens of kilometres increases this figure to about 1800 km² (Fig. 3.3a). Such an approximation is realistic for the area actually covered by ignimbrites of the member, but distal and remote ash layers from co-ignimbrite and vent plumes (Sparks and Walker, 1977; Walker, 1981c) were probably far more extensive.

Using the expanded area (Fig. 3.3a) and the average thickness of

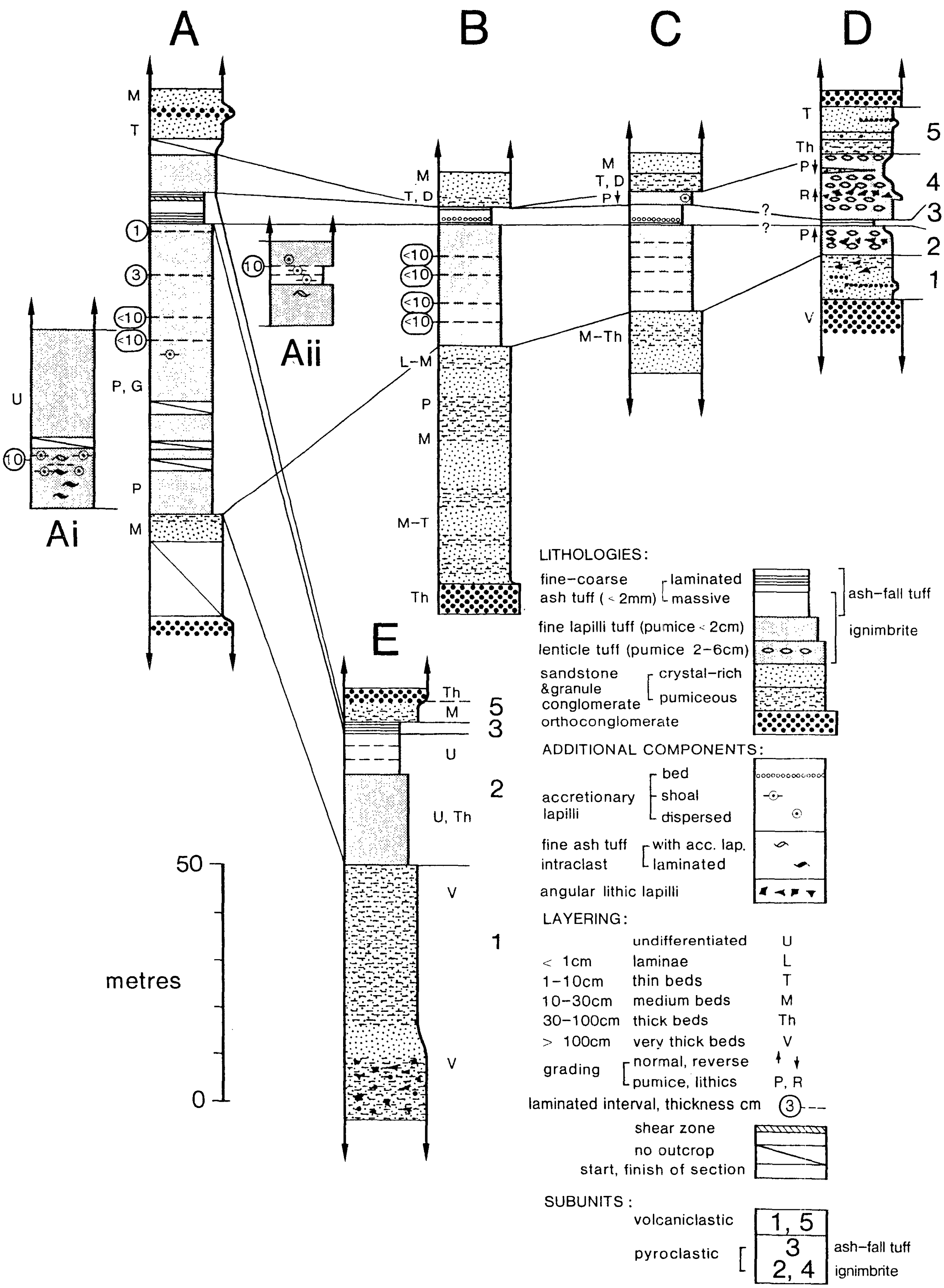


Figure 3.2: Stratigraphic sections of the Cana Creek Tuff Member of the Currabubula Formation, showing lithologies, subunits and correlations. Grid references for sections indicated on Figure 3.1 are: A-GR557261 Werris Creek; B-GR584470, C-GR578481, Piallaway; D-GR567529 Winton; E-GR738149 Quipolly.

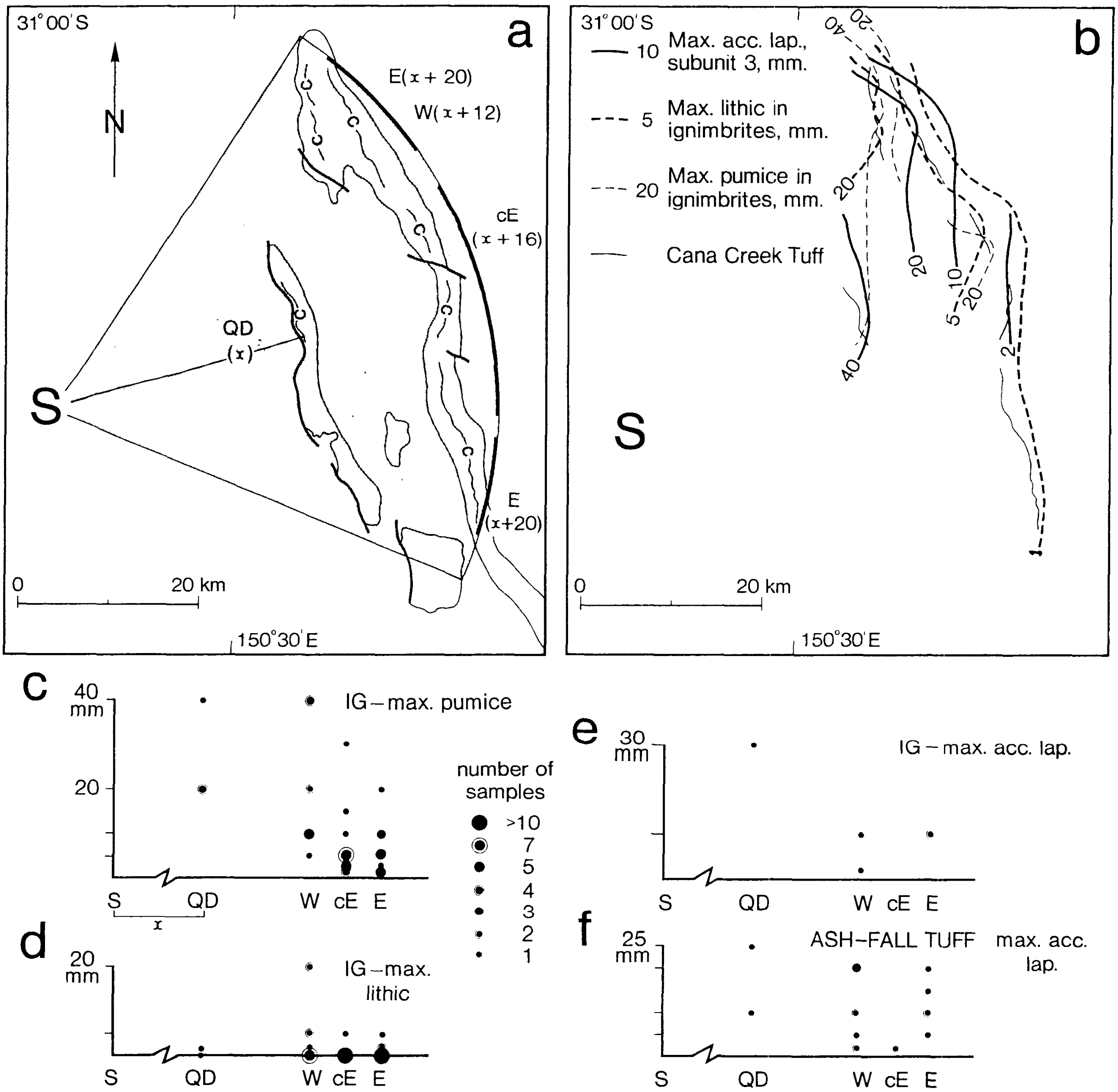


Figure 3.3: Distribution and grain size of ignimbrite and ash-fall tuff of the pyroclastic facies of the Cana Creek Tuff Member.

a. Hypothetical original distribution of the Cana Creek Tuff Member for an eruptive source at S (area ~1800 km²). Distances in kilometres to the Cana Creek Tuff outcrops are given relative to the Quirindi Dome (QD, x km from source), the west limb (W, x +12 km), central east limb (cE, x +16 km) and east limb (E, x +20 km) of the Werrie Syncline.

b. Approximate isopleths in millimetres for maximum dimensions of pumice and lithic clasts in ignimbrites, and accretionary lapilli in ash-fall tuff.

c-f. Maximum dimensions in millimetres of pumice, lithics, and accretionary lapilli in ignimbrite (c,d,e), and accretionary lapilli in ash-fall tuff (f) compared with estimated distance in kilometres from source S. The number of samples with components of the designated size is indicated by symbols. Ignimbrite samples without lithic clasts are plotted along the 0mm axis in d. Distance scale is the same as on a. and b.

TABLE 3.1: Estimated volume of the Cana Creek Tuff Member for the distribution indicated on Figure 3.3a

(present preserved area $\approx$ 1400 km^2)

proposed original extent $\approx$ 1800 km^2 (Figure 3.3a)

average thickness $\approx$ 0.07 km

$\therefore$ original compacted volume = 126 km^3

original compacted density $\approx$ 2000 kg/m^3

dense rock (2500 kg/m^3) equivalent volume $= \dfrac{\text{compacted density}}{2500} \times$ compacted volume

$\approx$ 100 km^3

$= 10^{11}$ m^3

eruption magnitude (Newhall and Self, 1982): very large, VIII

70 m, the bulk volume of the Cana Creek Tuff Member is found to be about 126 km³ of which almost half is ignimbrite (Table 3.1). The initial density of the Cana Creek Tuff Member would have been comparable to the average tephra density of similar uncompacted modern deposits (e.g. 1300 kg/m³ for the Wairakei Formation, New Zealand; Self, 1983). Reduction of porosity by mechanical compaction on burial would have increased the density, perhaps to about 2000 kg/m³, similar to the values reported for partially welded, young rhyolitic ignimbrite (e.g. Bishop Tuff, California; Ragan and Sheridan, 1972). The total dense rock equivalent (DRE) volume represented by the reconstructed Cana Creek Tuff Member is of the order of 100 km³ (10^{11} m³) but this value is conservative in view of the exclusion of distal deposits. These calculations are admittedly poorly constrained but at least demonstrate the large magnitude (Walker, 1980; Newhall and Self, 1982) of the eruption responsible for the Cana Creek Tuff Member.

Components

Primary textures and structures in outcrop, hand specimens and thin-sections are in general well preserved. Originally glassy components are completely devitrified and extensively replaced by chlorite, zeolites and analcime but in most instances, particle outlines are unaffected. Wilkinson and Whetten (1964) attributed the replacement mineral assemblage to low-grade burial metamorphism.

Diverse textures arise from combinations in differing proportions of three fragment types (Fig. 3.4): pumice and pumiceous clasts, including shards, unresolvable fine ash and dust, and accretionary lapilli; mineral grains (quartz, pink K-feldspar, white plagioclase); dense (that is, non-vesicular) lithic fragments. The largest pumice lapilli observed are sparsely porphyritic, having less than 2 modal percent phenocrysts of quartz, K-feldspar and plagioclase. Vesicle size, shape and abundance in pumice vary widely even in a single thin-section but two sorts predominate (Fig. 3.5a): one has round, large (up to 1 mm across), thick-walled (0.1 to 0.2 mm) vesicles whereas the other has aligned, elongate, thinner walled vesicles, imparting a fibrous texture. The latter variety was tube pumice and is not an artefact of primary welding nor of subsequent deformation of vesicles: the directions of vesicle alignment are neither consistent on thin-section scale, nor always parallel to the long dimension of lensoid fragments, and some of these pumices have internally-folded vesicles. There is a corresponding range in the cross-section shapes of the larger shards (0.1 to 0.5 mm) which includes equant blocks, rods and delicate pronged and cuspate spines (Fig. 3.5b). Ash finer than 0.1 mm occurs between shards and commonly partially infills pumice vesicles. Accretionary lapilli are similar to those described from both ancient and modern ash-fall deposits (e.g. Moore and Peck, 1962; Sparks, 1975; Self and Sparks, 1978; Hoblitt *et al.*, 1981; Self, 1983).

Internal stratigraphy

The arrangement of pyroclastic and volcaniclastic facies in each section through the Cana Creek Tuff Member is grossly similar, enabling division into five subunits (Fig. 3.2). Volcaniclastic facies are principally confined to the oldest and youngest subunits (1 and 5). Subunit 3 is ash-fall tuff separating ignimbrite subunits 2 and 4. Sections have been correlated by assuming that the main ash-fall tuff interval in each was part of the same extensive layer. The quality of exposure precludes tracing subunits between sections. All subunits are present in sections A and C but one or more is missing from the other three sections. The most northerly section (D) shows the greatest departure from the standard sequence (discussed below).

VOLCANICLASTIC FACIES

Poorly sorted, pumiceous granule conglomerate and quartzofeldspathic sandstone are typical of the volcaniclastic facies. These two rocktypes are end members of a textural spectrum controlled by the abundance of pumice granules and mineral fragments. Granule conglomerate (Fig. 3.5c) tends to be bimodal, having angular to subangular mineral fragments (about 20 percent) 1 to 2 mm across and greater percentages of larger (up to 20 mm) pumice clasts, presumably reflecting hydraulic sorting according to density rather than size (*cf.* Hume *et al.*, 1975). Quartzofeldspathic sandstone has only minor amounts of pumice and abundant angular grains of quartz, K-feldspar and plagioclase (Fig. 3.5d). Undeformed shards are common in the matrices of both varieties and little more abraded

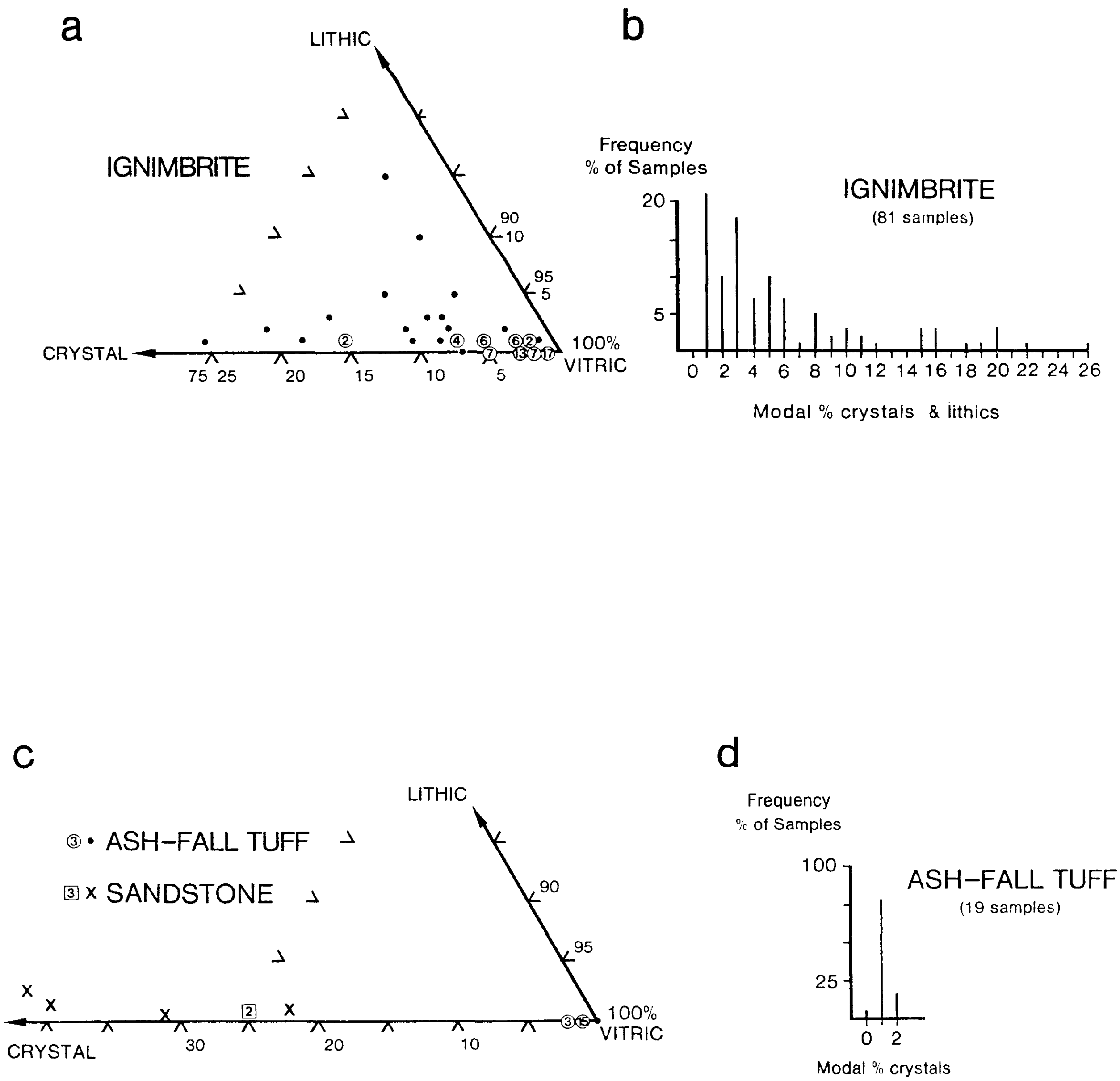


Figure 3.4: Modal data for ignimbrite, ash-fall tuff and sandstone of the Cana Creek Tuff Member. Thin-sections of the sandstones and of 3 ignimbrite samples were point counted (see Table B.2). Proportions of components in all other samples were estimated from hand specimens and thin-sections by comparison with percentage charts and the point count results.

a,c. Crystal-lithic-vitric proportions in ignimbrite, ash-fall tuff and sandstone. The vitric component includes pumice, shards and very fine ash(?), all of which are now devitrified. Figures within circles or squares give the number of samples having the designated proportions.

b,d. Frequency histograms of the modal proportions of crystals ± lithics in ignimbrite and ash-fall tuff.

Figure 3.5: a. Photomicrograph of non-welded ignimbrite containing pumice with round vesicles (v) and tube pumice (t). R55116, plane polarised light.

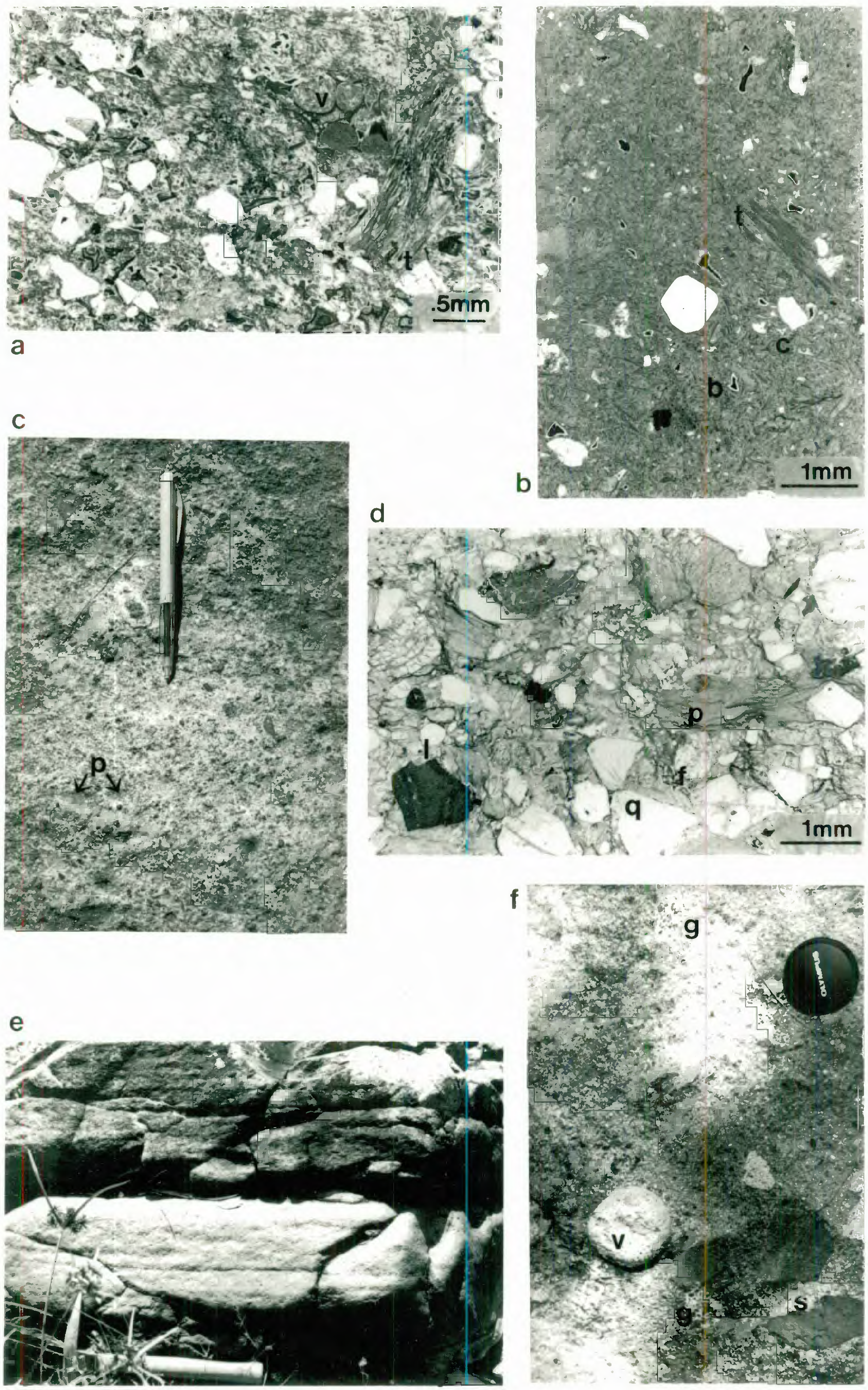
b. Photomicrograph showing blocky (b) and cuspate (c) shards in the matrix of accretionary lapilli-bearing tuff. Note also tube pumice (t). R55056, plane polarised light.

c. Pumiceous and crystal-rich granule conglomerate in internally massive, very thick bed. Recessive, pale, smooth areas are pumice fragments (p) (GR739149 Quipolly). Pen 13.5 cm.

d. Crystal-rich sandstone comprising angular quartz (q), plagioclase (f) and lithic (l) grains, and pumice fragments (p). R55128, plane polarised light.

e. Plane parallel, medium thickness bedding in crystal-rich sandstone (GR557261 Werris Creek). Hammer 33 cm.

f. Slate (s), granitoid (g) and volcanic (v) cobbles in massive, pumiceous, quartzofeldspathic granule conglomerate (GR739149 Quipolly). Lens cap 5.5 cm.


a
v
t
.5mm
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g
s

in comparison with those in the pyroclastic lithologies. Unresolvable matrix may originally have been very fine ash. The proportions of dense volcanic rock fragments varies but most have the same size and angularity as the mineral grains.

Medium (10 to 30 cm) and thick (30 to 100 cm) bedding is well developed. Beds are parallel laminated or, uncommonly, internally massive; they are continuous and even on outcrop scale, and have planar, sharp contact surfaces though more diffuse, gradational boundaries also occur (Fig. 3.5e). Cross-bedding has not been observed in any of the sections.

The grain size and sedimentary structures of the main rocktypes (granule conglomerate, sandstone) of the volcaniclastic facies match the parallel laminated sand lithofacies (Sh) of Miall (1977,1978) and Rust (1978), recognised as the deposits of very high energy stream or sheet flow (upper flow regime; McKee *et al.*, 1967; Miall, 1977; Collinson, 1978a; Gloppen and Steel, 1981; Tunbridge, 1981,1983; Nilsen, 1982). Facies assemblages dominated by this lithofacies are produced by catastrophic ephemeral flooding supplied exclusively by sand grade sediment (McKee *et al.*, 1967; Miall, 1977,1978; Collinson, 1978a; Rust, 1978; Tunbridge, 1981; Cant, 1982).

In some eastern exposures, pumiceous and quartzofeldspathic granule conglomerate locally contains dispersed, rounded and subangular cobbles of slate, gneiss, and granitic and volcanic rocktypes (e.g. section E, Figs. 3.2,3.5f). The proportion of mud grade material is variable but is estimated to be less than 20 percent. Continuous exposures indicate bed thicknesses in excess of several metres, possibly up to tens of metres, and lacking in signs of internal organisation. Lateral extent of these cobble-bearing, massive intervals in the volcaniclastic facies is not known precisely, but does not exceed several hundred metres. These layers are inferred to be subaerial mass-flow deposits. The context suggested by the association with parallel laminated sandstone is that of hyper-concentrated floods at peak discharge, but a laharic origin is equally plausible (*cf.* Crandell, 1971). Each of these regimes may have operated at different places and times during deposition of the volcaniclastic facies. In both cases, flows laden with pumice and crystal fragments which swept along existing drainage would have collected fluvially-rounded, exotic clasts (*cf.* Crandell, 1971). Superficial resemblance to crystal-rich, massive, valley-filling ignimbrite is evident and such an origin cannot be conclusively discounted on the basis of present data.

The rocktypes of the volcaniclastic facies of the Cana Creek Tuff Member differ in composition, depositional structures and grain size from the enclosing sedimentary rocks of the Currabubula Formation. Most of the Currabubula Formation is pebble to cobble grade, very thickly (greater than 1m) bedded conglomerate (massive gravel lithofacies Gm of Miall, 1977; Fig. 2.6f), interpreted to be longitudinal bar and channel lag deposits from flood events in a gravel-dominated braided river system. Sandstone is uncommon but typically cross-bedded and confined to lenses within thicker conglomerate beds (lithofacies St and Sp, Miall, 1977,1978; Rust, 1978,1979). Vertical sequences are similar to those of proximal braided rivers and alluvial plains (*cf.* Scott type, Miall, 1977; facies assemblage G_{II}, Rust, 1978). In both the conglomerates and the sandstones, mineral grains are subordinate in abundance to volcanic lithic fragments and only minor amounts of pumice are present.

Deposition of the Currabubula Formation conglomerates and of the Cana Creek Tuff sandstones are each attributable to episodes of flooding. However, the volcaniclastic facies of the Cana Creek Tuff is dominantly sand grade but not necessarily because floods were less competent : the finer grain size probably reflects the overwhelming input of almost exclusively sand-sized particles indicating that sediment supply and grade were temporarily independent of control by erosion rates in the source area (*cf.* McKee *et al.*, 1967; Davies *et al.*, 1978; Tunbridge, 1981). The background sedimentation of braidplain gravels was evidently abruptly interrupted during the emplacement of the Cana Creek Tuff Member.

Further explanation of the significance of the sandstones and granule conglomerates of the volcaniclastic facies rests on the recognition that they are made of the same constituents as the primary pyroclastic rocktypes : broken crystals of quartz, K-feldspar and plagioclase; pieces of pumice; shards; and sparse volcanic rock fragments (Fig. 3.5d). It is concluded that components of the volcaniclastic facies were initially generated by explosive fragmentation of porphyritic pumice during the same eruption that produced the pyroclastic facies. These components lack indications of having undergone weathering or abrasion by epiclastic reworking, and the aggregates represented by the sandstones and granule conglomerates are largely undiluted by detritus from non-pyroclastic sources. Accumulations of pyroclastic debris at proximal sites were apparently water-saturated and unstable, as the volcaniclastic facies provides evidence of rapid redeposition by sheetfloods which inundated pre-existing drainage. Although strictly sedimentary rocks, the sandstones and granule conglomerates of the volcaniclastic facies were a by-product of an explosive volcanic eruption and in medial settings, constitute the earliest signs of its onset.

PYROCLASTIC FACIES : IGNIMBRITE

One interval of non-welded ignimbrite comprising several flow units occurs in each section (Fig. 3.2). A second, markedly thinner interval is present in two sections (A,C, Fig. 3.2).

Composition

Analyses of samples of ignimbrite from five sites within or close to the stratigraphic sections show they are rhyolites with calc-alkaline or high-K calc-alkaline affinities (Fig. 3.6), as are other ignimbrite members of the Currabubula Formation (Chapter 4, Table 4.2, Fig. 4.5).

Components and grain size

Ignimbrites of the Cana Creek Tuff Member are composed of fragments of pumice, crystals (quartz, pink K-feldspar, white plagioclase), shards and unresolvable material taken to have been fine ash and dust (less than 0.063 mm; Figs. 3.5a,3.7a,b,c). Accretionary lapilli are an uncommon but widespread component of Cana Creek Tuff ignimbrites. Accidental and/or accessory lithic fragments are abundant only in section D ignimbrite, where they occur both as near-base lithic-rich zones and dispersed more sparsely through flow units. Modes show Cana Creek Tuff ignimbrites to be the most pumiceous and crystal-poor of the Currabubula Formation members (Figs. 2.3,3.4). They are also the finest in grain size, comprising pumice

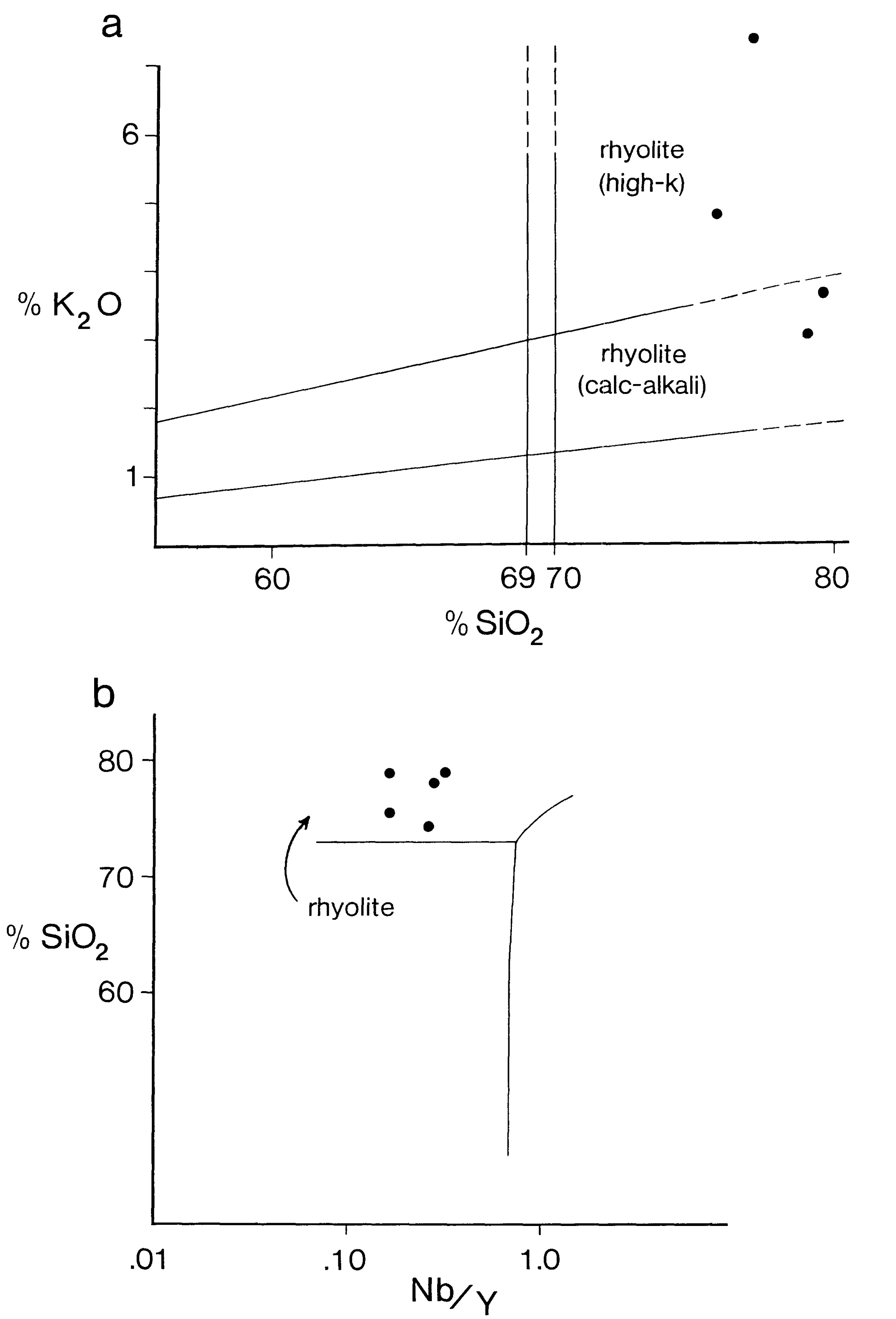


Figure 3.6: Chemical classifications of ignimbrites of the Cana Creek Tuff Member. Analyses 1 to 5, Tables 4.2,4.3, Chapter 4.

a. Anhydrous $\%K_2O$ vs. $\%SiO_2$ classification of Peccerillo and Taylor (1976; rhyolite > 70% SiO_2) and Ewart (1979; rhyolite > 69% SiO_2). Analysis 1 not plotted.

b. Anhydrous $\%SiO_2$ vs. $^{Nb}/Y$ classification (Winchester and Floyd, 1977).

Figure 3.7: Ignimbrites of the Cana Creek Tuff Member.

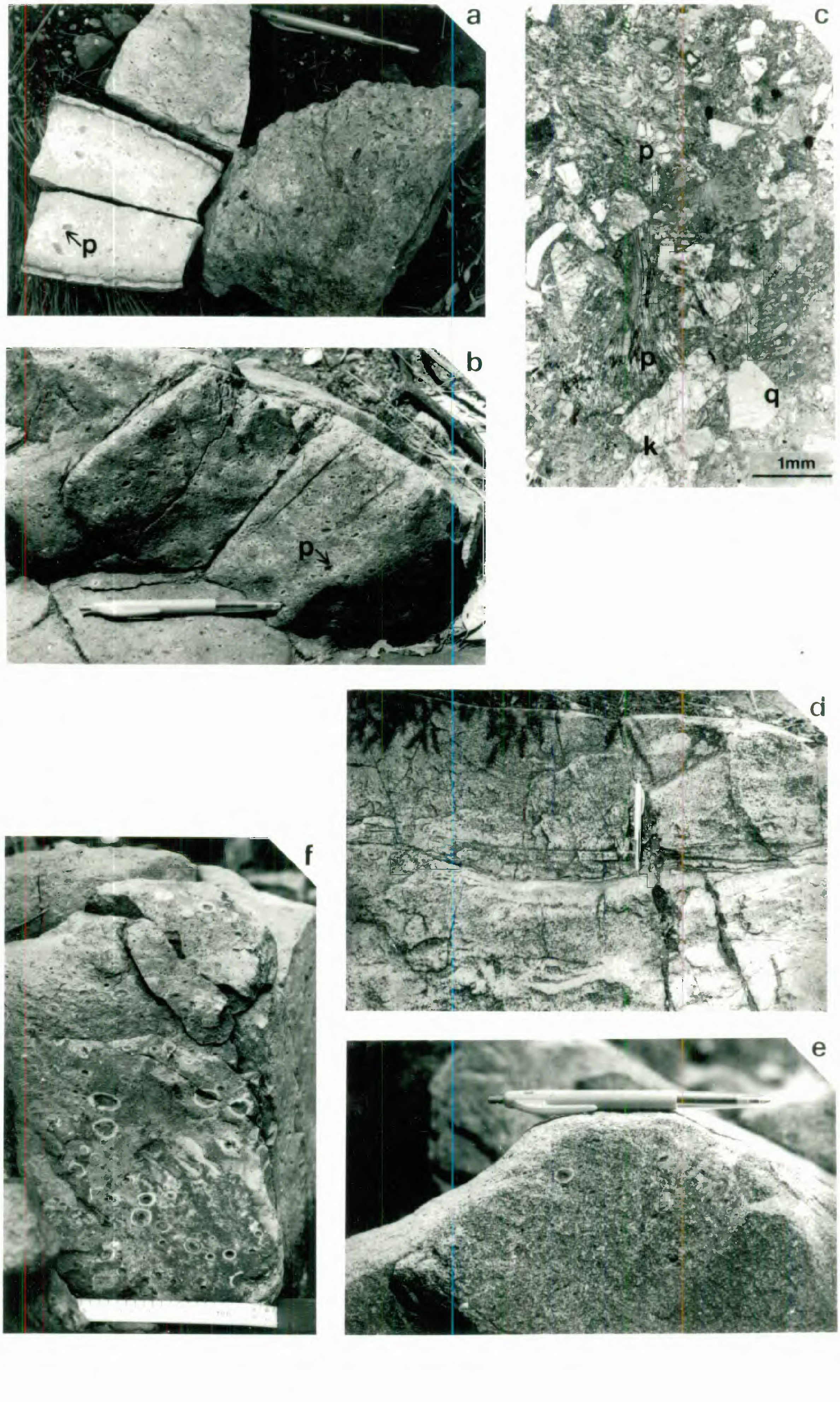
a,b. Typical hand specimens and outcrop of ignimbrite with randomly oriented, equant centimetre-sized pumice fragments (p) and fine grained matrix. Pen 13.5 cm. Samples from GR584470 Piallaway. Outcrop, section A (GR557261 Werris Creek).

c. Photomicrograph of non-welded ignimbrite comprising pumice (p), crystal fragments (q - quartz, k - K-feldspar) and finer matrix of poorly preserved shards and unresolvable ash. R55024, plane polarised light.

d. Lense of laminated tuff between flow units of ignimbrite, section A. Pen 13.5 cm.

e. Accretionary lapilli in ignimbrite, section Ai. Pen 13.5 cm.

f. Shoal of accretionary lapilli in ignimbrite, section Ai. Scale in centimetres.


a
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fragments with few exceptions less than 60 mm and typically less than 30 mm (maximum exposed dimension), crystal fragments smaller than 2 mm, and less than a few modal percent of dense lithics, the largest of which are 5 mm across (Figs. 3.3,3.4). The coarsest ignimbrites are pumiceous and the finest are shard-rich.

Grade

That the ignimbrites of the Cana Creek Tuff Member were originally non-welded is clear from the ubiquitous totally undeformed pumice and matrix shard shapes, even in the thickest parts (Figs. 3.5a,3.7c). Other ignimbrites of similar composition higher in the Currabubula Formation have collapsed pumice and moulded shards indicative of primary welding. Pumice fragments in Cana Creek Tuff ignimbrites vary from lensoid to equant (Fig. 3.7a,b) and pumice foliation is weakly developed or lacking altogether, in contrast to the consistently well-foliated welded ignimbrites. Induration by incipient sintering of shards or vapour phase crystallisation (producing 'sillar', Fenner, 1948) is difficult to distinguish texturally in ancient ignimbrite rocks. However, induration of the Cana Creek Tuff Member ignimbrites is most unlikely because they are not present as clasts in overlying sedimentary rocks. By comparison, the younger welded ignimbrites of the Currabubula Formation commonly constitute a conspicuous component of the detrital assemblage in the succeeding conglomerates.

Being non-welded but of substantial thickness, the Cana Creek Tuff ignimbrites are appropriately described as low-grade (Wright *et al.*, 1980; Walker, 1983).

Flow units and correlations

The main ignimbrite interval (subunit 2) in all sections is characterised by stacked flow units varying from 60 m total thickness in the most proximal section to between 20 and 40 m in sections farther east (Fig. 3.2). At least six flow units are recognisable in section A, the lower ones being more than 10 m each in thickness whereas upper units are thinner than 10 m. As many and possibly more flow units are present in two easterly sections (B,C) but range from little thicker than a metre to

3 or 4 metres each. The poorer quality of exposure of ignimbrite in section E prevented subdivision into flow units. Subdivision of section D was similarly hampered but there may be four units present, some of which exceed 6 m in thickness.

Flow unit boundaries are indicated by thin (10 cm or less) laminated intervals of crystal-rich coarse ash tuff and crystal-poor fine ash tuff. Pumiceous components and crystal fragments are petrographically similar to those occurring in the associated ignimbrites. In section A the laminated intervals lense out along strike (Fig. 3.7d) whereas in sections farther east they are laterally continuous at outcrop scale. Flow base textural variants are uncommon with the exception of lithic-rich layers (2bL, Sparks *et al.*, 1973) in ignimbrite of section D and examples of centimetre scale, planar divisions between thin flow units in section A which could be basal layers (2a; Fig. 3.8; *cf.* Sparks *et al.*, 1973; Sparks, 1976).

There are several possible primary pyroclastic origins for laminated intervals between ignimbrite flow units, some of which imply a genetic link with overlying or underlying ignimbrite: fallout from ash clouds (co-ignimbrite ashes, Sparks, 1976; Sparks and Walker, 1977; Walker, 1981c,1983; vent produced, distal plinian or phreatoplinian deposits; Walker, 1981d), deposition from dilute pyroclastic flows or surges (pyroclastic or ground surge deposits, Sparks *et al.*, 1973; Sparks and Walker, 1973; Sparks, 1976; Fisher, 1979; ash cloud surge deposits, Fisher, 1979), ignimbrite veneer deposits (Walker *et al.*, 1980a,b; Walker, Wilson and Froggatt, 1981) or concentrations of fine ash due to segregation processes in strongly fluidized pyroclastic flows (Wilson, 1980). Secondary explosions at rootless vents in ignimbrite deposited on wet ground or inundated by water may also locally produce laminated pyroclastics (e.g. Rowley *et al.*, 1981; Hildreth, 1983; Walker, 1983).

Compositional similarity of the laminated intervals of the Cana Creek Tuff with the enclosing ignimbrite flow units suggests that the two deposit types are genetically related and not accidentally associated. Small scale lateral thickness variations and the presence of crystal-rich laminae favour emplacement by pyroclastic surges accompanying ignimbrite. Specific correlation of the laminated intervals with presently enclosing ignimbrites cannot be demonstrated and only a general shared origin in the same eruption episode is inferred. Closer study of these laminated intervals may reveal conclusive signs of additional processes; in particular, local fluvial and/or aeolian reworking is plausible. The duration of breaks between emplacement of successive flow units is unknown, but there is no positive evidence for geologically significant interruptions, and spells as brief as hours would be sufficient to account for observed intercalations.

Internal grain size zonation in the ignimbrites is expressed mainly by gradual variation in the size and proportion of pumice although relatively crystal-rich intervals are noticeable in places. Such variation is apparently contained within the 2b portion of the flow units (*cf.* Sparks *et al.*, 1973) but exposure is not good enough to confidently relate grain size changes to flow unit boundaries. Entirely ungraded flow units are at least as common, which is not unexpected given their fine grained nature. Experimental studies (Sparks, 1976,1978a) show that only the larger pumice and lithic clasts are significantly affected by grading processes in fluidised pyroclastic flows. Sizes of clasts which are graded depend on gas velocities and particle densities, and for the conditions studied by Sparks (1976,1978a), ranged from 2 mm to 32 mm. The majority of clasts in Cana Creek Tuff ignimbrite as represented in these medial exposures would have been too small to be graded by gas flow processes.

Section D on the northern margin of the Member includes ignimbrite which is richer in dense volcanic lithics and arranged in thinner flow units than in other sections. This lithic rich variety gives way to ignimbrite more typical of the Cana Creek Tuff over a short lateral distance. The arrangement is similar to narrow lobes with coarser dense lithics in the isopleth maps of Wairakei Formation ignimbrite, thought to be locations of flow routes (Self, 1983, Fig. 12). The same explanation may account for Section D ignimbrite.

Accretionary lapilli-bearing ignimbrite

The best examples of ignimbrite containing accretionary lapilli are in section A. These are described below because the contexts are clearly displayed, but ignimbrite-hosted accretionary lapilli have been found at widespread localities. Accretionary lapilli occur in three circumstances in ignimbrite: as individuals, in 'shoals' (*cf.* "depositional lenses" in ignimbrites of the Wairakei Formation; Self, 1983, p.447), and in intraclasts of ash-fall tuff. Examples of accretionary lapilli-bearing intraclasts have not been observed other than in section A.

Single, whole or partly abraded accretionary lapilli (10 mm diameter) are present in at least four flow units of subunit 2 in section A (Figs. 3.2, 3.7e). Shoals of accretionary lapilli, each a metre or less in thickness, also occur in three of these flow units. In the shoals, accretionary lapilli range up to 40 mm across, are poorly sorted and accompanied by abundant fine grained rind fragments (Figs. 3.7f,3.8). The accretionary lapilli typically have relatively coarse grained cores enclosed by finer grained rinds, but many have additional pairs of coarse and fine shells. Some of the coarser shells are concentrically graded and there is considerable complexity of grain size in detail. Accretionary lapilli are finer grained than the ignimbrite host, and composed exclusively of shards.

None of the rind fragments in shoals can be related to adjacent accretionary lapilli and none of the accretionary lapilli are fragmented *in situ*. Most have an elliptical cross-section shape but some are markedly more elongate than their neighbours. The long dimensions of separate accretionary lapilli and of rind fragments are commonly aligned and slightly oblique to bedding, defining an apparent imbricate structure but many in the same shoal have long axes parallel to bedding, and a few "stand" at right angles to bedding (Fig. 3.8). The elliptical shape was thus acquired independently by each accretionary lapilli and is an inherited primary feature, perhaps due to flattening on impact, and not a response to compaction or tectonic deformation.

The shoals are not laterally continuous and grade into 'barren' fine grained, shard-rich ignimbrite over a few tens of centimetres without any detectable discontinuity between. Vertical limits to the shoals low in section A are similarly diffuse. The higher occurrence in section A is more complicated (Fig. 3.8), consisting of two thin flow units each with shoals, above a planar, sharp contact with massive, accretionary

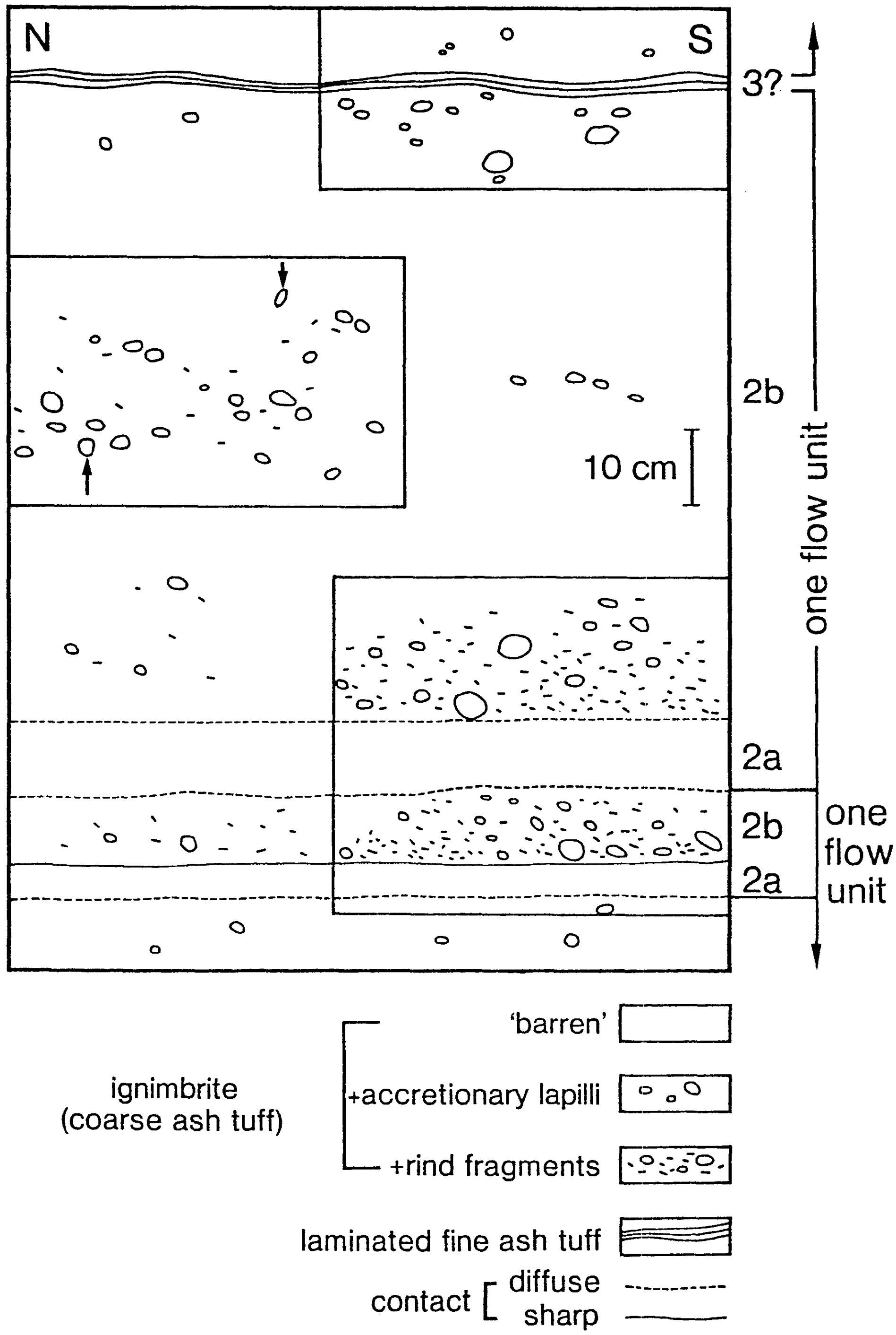


Figure 3.8: Line drawing from photographs of shoals of accretionary lapilli in ignimbrite of section A (GR557261 Werris Creek). The shape, arrangement and number of accretionary lapilli within rectangles have been drawn accurately. Arrows indicate accretionary lapilli which have their long dimensions subvertical, and the interpreted flow unit subdivisions are labelled according to Sparks *et al.* (1973).

lapilli-bearing ignimbrite. These shoals are separated from the overlying ignimbrite by 10 cm of laminated tuff (layer 3 deposit? *cf.* Sparks *et al.*, 1973) and do not continue along strike to the nearest exposures of ignimbrite (tens of metres away).

Intraclasts. Two ignimbrite flow units in section A contain blocks of accretionary lapilli-bearing ash-fall tuff (Fig. 3.9). The examples low in section A (e.g. Fig. 3.9a) are accompanied by contorted ribbons of fine grained laminated tuff which are also interpreted to be ash-fall tuff. Accretionary lapilli within the intraclasts are abundant, whole, unabraded, of uniform shape and alignment, and of much the same size or else crudely graded. Some are composed of multiple alternations of coarse and fine shells although most consist of a simple coarse core and fine rind. Both fragment types are petrographically and texturally identical to *in situ* ash-fall tuff of the pyroclastic facies of the Cana Creek Tuff (described below) and are thus considered to be intraclasts. These blocks of laminated and accretionary lapilli-bearing tuff were evidently soft but competent when incorporated in the host ignimbrite. The accretionary lapilli-bearing intraclast higher in section A has a streamlined shape suggesting that it has deformed to accommodate conditions in the moving ignimbrite (Fig. 3.9b).

The accretionary lapilli intraclasts must have been picked up by the host ignimbrites en route from a more proximal location and indicate their passage across unconsolidated fine grained ash-fall deposits from an earlier phase of the same eruption.

Accretionary lapilli occurring in shoals and singly have also clearly been transported by the enclosing ignimbrite. However, it is unlikely that they were collected from the substrate as they are set directly in ignimbrite matrix without any accompanying fragments of massive or laminated ash-fall tuff. Further to this, the range in sizes of accretionary lapilli in shoals is much wider than that displayed by undisturbed accretionary lapilli populations observed in ash-fall units. If the accretionary lapilli had been collected from the substrate by the enclosing ignimbrite during flow, very erratic but selective sampling by the moving pyroclastic flow would have to be invoked in order to account for the range in their sizes.

Some Quaternary ignimbrites have concentrations of accretionary lapilli and other coarse and dense components in pipe-shaped fumaroles (e.g. Walker, 1971; Yokoyama, 1974; Wright *et al.*, 1980; Self, 1983). It is not clear whether the accretionary lapilli actually formed in the pipes, or were scavenged from elsewhere in the ignimbrite or from underlying ash-fall deposits. Wilson (1980,1984) has demonstrated that degassing pipes can exist in moving pyroclastic flows and that once formed, such segregation structures are likely to persist after the flow comes to rest. Hence, accretionary lapilli are unlikely to be released to the body of a pyroclastic flow by disintegration of pipes which had concentrated them. Shoals of accretionary lapilli in the Cana Creek Tuff ignimbrites are not associated with local increases in proportions of dense components in the host ignimbrites which would be expected if they represent dissociated pipes.

The remaining and favoured origin of the accretionary lapilli occurring in shoals and separately is incorporation in moving pyroclastic flows by fallout from fine grained co-ignimbrite ash clouds.

Accretionary lapilli are evidence for the coexistence of suspended fine ash and water or steam (Moore and Peck, 1962). Ignimbrite eruptions generate clouds of fine ash at the vent, as an accompaniment to moving pyroclastic flows (Sparks and Walker, 1977; Sparks and Huang, 1980; Walker, 1981c), and perhaps also at the sites of secondary phreatic explosions (Walker, 1979,1981c,e,1983; Self, 1983). Accretionary lapilli formed in the first two of these situations could be expected to rain onto and be incorporated into moving pyroclastic flows, with or without also forming discrete ash-fall deposits. Wilson's (1984) experimental study of the behaviour of fluidised beds suggests that accretionary lapilli would sink or float in the pyroclastic flows depending on their density relative to that of the local matrix. Perhaps the uniformity in density of the accretionary lapilli promoted their initial concentration in shoals. Even if accretionary lapilli incorporated in pyroclastic flows close to source did not survive being transported the full outflow distance, contributions from ash clouds generated by the moving flows could have continuously replenished the supply.

The water required to make accretionary lapilli may be provided by an external source, such as a rain storm, and/or condense from steam-rich phreatomagmatic eruptions, or perhaps be ejected directly as water droplets by eruptions from submersed vents (e.g. Walker, 1981a). Distinguishing between these sources on the basis of deposits alone requires detailed data on distribution in relation to vent location. Flushing by rain storms can be demonstrated where thickness trends of undisturbed accretionary lapilli-bearing ash-fall deposits correlate with storm paths rather than vent position (Self and Sparks, 1978; Walker, 1981e). Moving pyroclastic flows which received such accretionary lapilli would interfere with the relationship but possibly preserve the localised pattern of distribution. The extensive distribution of ignimbrite-hosted accretionary lapilli in the Cana Creek Tuff invites speculation that flushing of the co-ignimbrite ash clouds was widespread because the eruption column was inherently steam-rich.

Flow type

Features of the Cana Creek Tuff ignimbrite units are not easily interpreted in terms of fluidisation state (*cf.* Sparks, 1978a; Wilson, 1980). An attempt to quantify a general impression that crystal enrichment is slight is summarised in Table 3.2a. Most pumice clasts are not sufficiently large to include a representative population of phenocrysts and the calculations depend on only three modes of porphyritic pumice. Attention is drawn to the biases in sampling and vagaries in approximations of pumice densities. Results show that most of the ignimbrites are slightly enhanced in crystals (7 to 14 weight percent) relative to the pumice clasts (less than 10 weight percent). Self (1983, p.447) reported similarly moderate crystal enrichment in ignimbrites of the Wairakei Formation (19 weight percent in ignimbrite compared with 14 to 15 weight percent in pumice) which nevertheless are very extensive and have a distribution indicative of deposition from exceptionally mobile pyroclastic flows. Segregation structures (Wilson, 1980) which reflect a high degree of fluidisation during flow have not been identified in Cana Creek Tuff ignimbrites.

There are several reasons for avoiding the conclusion that the pyroclastic flows which deposited the ignimbrites were poorly fluidised or immobile. To some extent the lack of segregation structures is an artefact

TABLE 3.2a: Estimates of crystal enrichment factors for ignimbrites of the Cana Creek Tuff Member

% crystal content of PUMICE[1]		% crystal content of IGNIMBRITE[2]		frequency, % samples	enrichment factor[3]
modal	weight	modal	weight	(Fig. 3.4)	
⩽2	10	⩽3	7	49	0.7
		4-6	10-14	24	1 to 1.5
		7-12	16-26	14	1.7 to 3.1
		15-20	31-39	12	4.0 to 5.8

[1] Crystal density 2600 kg/m³. Original density of pumice 500 kg/m³ (for void fraction 0.77; *cf.* Wilson *et al.*, 1980, p.126).

[2] Original density of ignimbrite matrix 1000 kg/m³ (*cf.* Wright and Walker, 1981, p.122).

[3] Enrichment factor calculation follows Walker (1972, p.137):-

(weight ratio crystals:pumice in ignimbrite) x (weight ratio pumice:crystals in artificially crushed pumice).

Note that components coarser than 2 mm are minor in Cana Creek Tuff ignimbrite.

TABLE 3.2b: Estimated aspect ratio of subunit 2 ignimbrite for the distribution indicated on Figure 3.3a

Aspect ratio (Walker *et al.*, 1980a) $= \dfrac{\text{average thickness}}{\text{diameter of circle with the same area}}$

$$= \frac{20}{48000}$$

$$\approx 1:2400$$

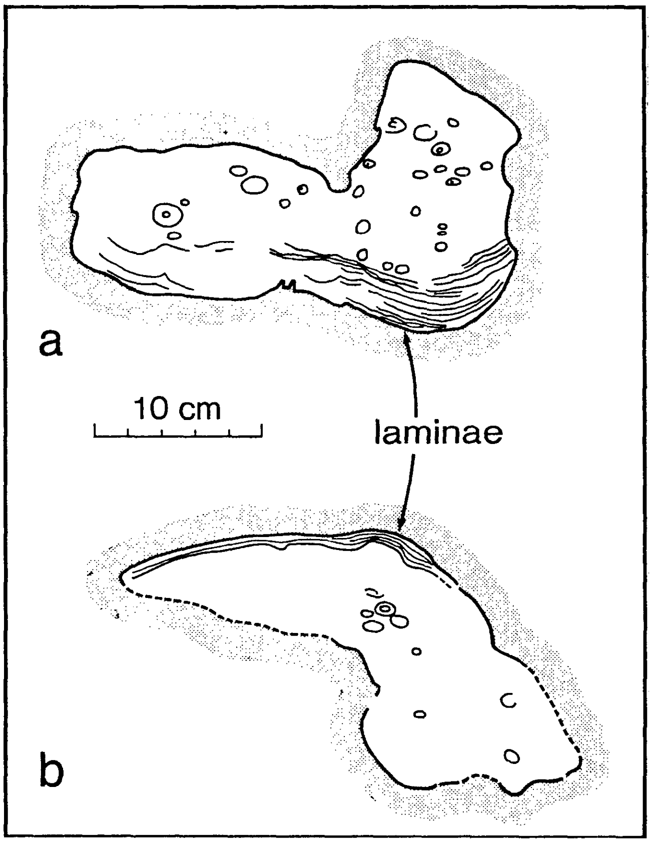


Figure 3.9: Accurate line drawings from photographs of intraclasts of accretionary lapilli-bearing and laminated fine ash tuff in ignimbrite of section A (GR557261 Werris Creek).

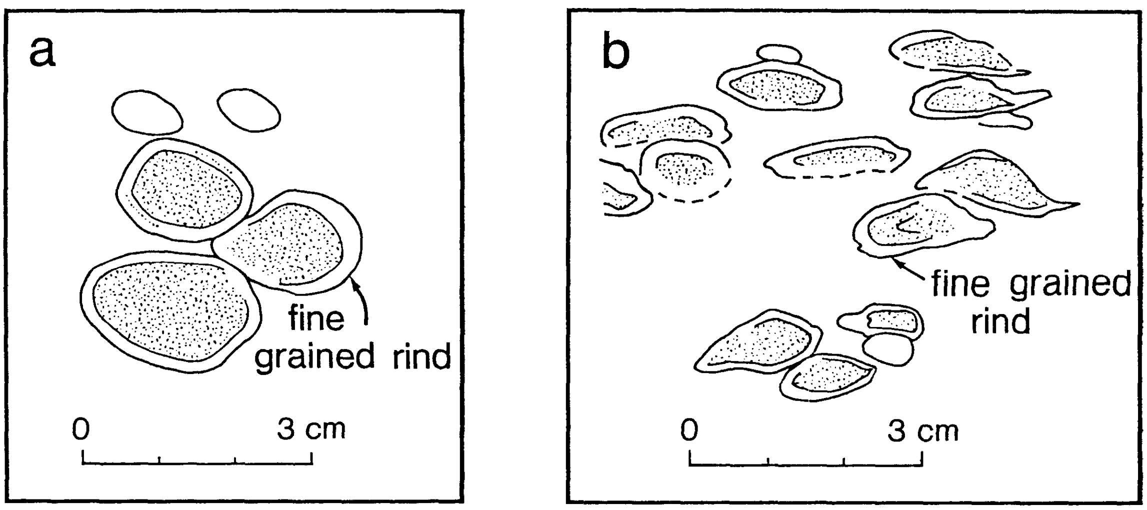


Figure 3.11: Accurate line drawings of deformed accretionary lapilli in ash-fall tuff.

a. R55046. b. R55033.

of the paucity of large and dense clasts. The uncommon intraclasts of ash-fall tuff in section A ignimbrites are large but originally would have been of comparable density (of the order of 1000 to 1300 kg/m^3; *cf.* Walker, 1981a; Self, 1983) to the host pyroclastic flow (*cf.* 1000 kg/m^3 estimated by Sparks, 1978a). Their presence is thus not considered to be a sign of high yield strength in the pyroclastic flows which transported them. Co-ignimbrite ash represented by accretionary lapilli indicates that the Cana Creek Tuff flows were sufficiently fluidised to elutriate fine pyroclasts. Also, flows with high proportions of fines will be fluidised at relatively low gas flow rates and rapidly deposit any available dense clasts near the source (Sparks, 1976). Multiple flow unit ignimbrites, such as those of the Cana Creek Tuff, are possibly the result of splitting of parent pyroclastic flows in which upward increases of gas flow rates and fluidisation state promoted detachment of more mobile, upper portions as thinner, separate flow units (Wilson, 1980). Original morphology of the Cana Creek Tuff pyroclastic flow deposits cannot be evaluated in detail and their sheet-like character may well be the result of efficient topographic smoothing by preceding volcaniclastic units. Even so, ignimbrite of subunit 2 has a moderate to low aspect ratio (1:2400, Table 3.2b; *cf.* Walker *et al.*, 1980a,b; Walker, Wilson and Froggatt, 1981) usually ascribed to flows with low yield strength and adequate fluidisation.

PYROCLASTIC FACIES: ASH-FALL TUFF

Each section measured includes a discrete subunit (3) of ash-fall tuff at the top of the lower ignimbrite subunit (Fig. 3.2). Thickness range is narrow, from approximately 7 m in the proximal section (A) to 4 m in three distal sections (B,C,E) and less than 1 m in section D. Ash-fall tuff is volumetrically minor as preserved, and accounts for 8 to 16 percent of the pyroclastic facies of sections other than D in which it comprises less than 4 percent. Laminated fine tuff separating ignimbrite flow units may in some instances be of ash-fall origin and ash-fall deposits are also represented within the ignimbrite by disseminated accretionary lapilli.

Character

Outcrops identified as ash-fall tuff are characterised by even, laterally continuous lamination (thinner than 1 cm) or very thin bedding (1 to 3 cm) in association with medium (10 to 30 cm) beds of accretionary lapilli. *In situ* beds of accretionary lapilli occur with laminated ash-fall tuff in sections B and C but have not been found in sections A,D or E. Loose blocks of accretionary lapilli-bearing ash-fall tuff are ubiquitous (Figs. 3.10a,b) and their occurrence was used in initial mapping of the Cana Creek Tuff Member.

Bedded intervals of ash-fall tuff comprise multiple centimetre scale layers of equal thickness (Fig. 3.10c). Contacts between beds are commonly diffuse though planar. Laminae in parts of the outcrop of ash-fall tuff in section A are exceptional and appear to be delicately corrugated in detail. The thinner laminae (0.1 mm) in this exposure are composed of very fine grained, shard-rich tuff and alternate with slightly thicker and coarser grained laminae. The thinner laminations are only continuous for several centimetres, being interrupted by gaps of a few millimetres. A thickness of 1 to 2 m is affected by this pattern, the remainder of the ash-fall tuff being plane laminated. There is a strong resemblance to dish structure, a dewatering feature found in rapidly deposited, fine grained sediments (Lowe and Lo Piccolo, 1974; Collinson and Thompson, 1982, p.145). Microbedding (Walker, 1971,1981a; Walker and Croasdale, 1972; Self, 1983) produced by rain splatter onto loose ash, forms a similar layering but the laminae are reported to be depleted in fine ash (Walker, 1981a), unlike those in section A ash-fall tuff.

Maximum grain size of components in the ash-fall tuff in most instances does not exceed a millimetre, so in outcrop separate layers appear to be well sorted. Accretionary lapilli differ from those hosted by ignimbrite in being whole or fragmented *in situ*, unabraded, and the same shape with long axes aligned parallel to bedding (Fig. 3.10d). They form laterally continuous zones within fine grained tuff which may be entirely massive or laminated. Sorting of the accretionary lapilli within zones is variable but moderate sorting and grading are common. Many accretionary lapilli are in contact and examples of moulding at margins indicate that in some instances they were soft when deposited and have not been reworked (Fig. 3.11).

An additional feature attributed to ash-fall deposition is limited to

Figure 3.10: Ash-fall tuff of the Cana Creek Tuff Member.

a,b. Hand specimens and boulder of accretionary lapilli-bearing ash-fall tuff. Coin diameter 27 mm. Boulder, section C. Pen 13.5 cm.

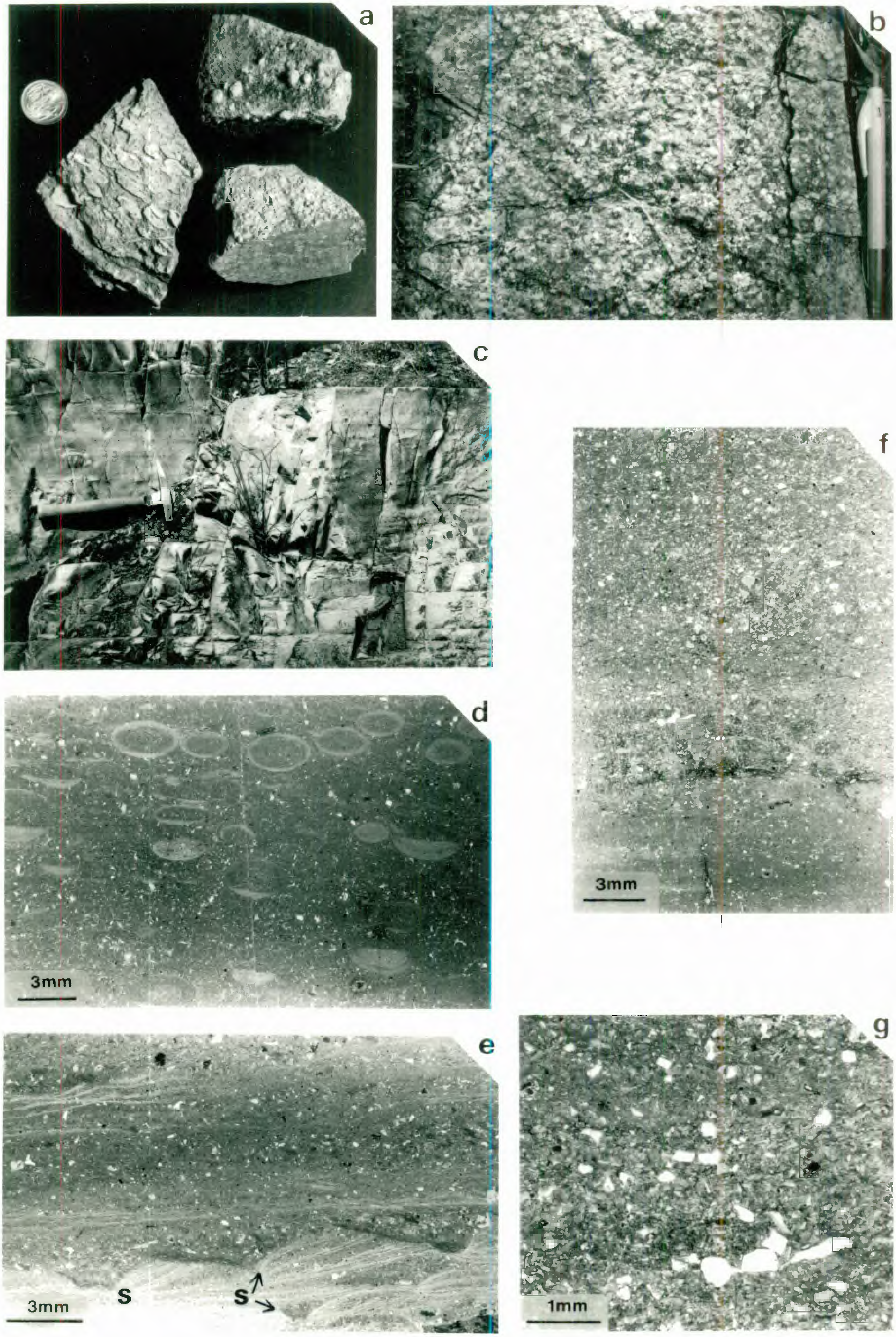
c. Laminated fine ash tuff, section A. Hammer 33 cm.

d. Photomicrograph of accretionary lapilli-bearing ash-fall tuff, section B. R55064, plane polarised light.

e. Photomicrograph of micro-impact sag(s) in laminated tuff, section D. R55038, plane polarised light.

f. Photomicrograph of laminated fine ash tuff, section B. R55065, plane polarised light.

g. Photomicrograph of detail of laminated fine ash tuff, section B, shown in Fig. 3.10f. R55065, plane polarised light.


a
b
c
d
3mm
e
S
S
3mm
f
3mm
g
1mm

section D where laminae in fine grained ash-fall tuff are disrupted by micro-impact sags (Fig. 3.10e). The "bombs" are crystals and pumice fragments (1 mm or less) which puncture several laminae in an asymmetrical fashion indicating oblique impact. Such exceedingly delicate structures would not be expected to survive unless rapidly covered by other non-erosive layers. Finding them is equally remarkable as their preservation, since they account for a tiny proportion of a single section.

Components and grain size

Ash-fall tuff of the Cana Creek Tuff Member is composed exclusively of juvenile pyroclasts, predominantly undeformed shards and unresolvable fine ash (?) with sparse angular crystal fragments and millimetre-size pumice flecks (Figs. 3.4, 3.10). Approximately 90 percent of most samples consist of sub-millimetre particles. Clearly discernible shards (0.1 to 0.2 mm) are angular, and equant or rectangular in cross-section; curved spines and prongs are also present. These relatively large shards are probably fragments of the walls of vesicles in pumice. The abundant, much finer pyroclasts are not separately resolvable and the details of their shapes are unknown.

No marked grain size trend has been detected between proximal and more distal sections. The maximum diameters of accretionary lapilli decline very slightly eastward (Figs. 3.3b, f). The conclusion that the ash-fall tuffs are uniformly fine grained is tentative because the difficulties of distinguishing bedded pumice-fall layers from reworked pumice deposits in ancient rocks may have introduced sampling bias to this study.

Classification and origin

The most widely used classification of ash-fall deposits (Walker, 1973a) compares their dispersal with the degree of fragmentation of pyroclasts. The scheme requires an isopach map of deposits from one eruption and their systematic granulometric analysis, or isopleth maps of maximum lithic, maximum pumice or median deposit grain size (Walker, 1981d). Ash-fall deposits characterised by fine grain size and wide dispersal are typical of large magnitude phreatoplinian eruptions involving silicic magma and water (Self and Sparks, 1978). Phreatoplinian airfall ash is also distinguished by comparatively poor sorting and weak fractionation with distance from source (Sparks *et al.*, 1981; Walker, 1981a,d). Other common features are small-scale thickness variations, abundant accretionary lapilli (Self, 1983), microbedding and vesicles (Walker, 1971,1981a,d,e). Deposits may be significantly affected by shortlived but intense fluvial erosion because water (either as steam or water droplets) apparently constitutes a substantial proportion of the ejecta (Walker, 1981a; Self, 1983).

Subunit 3 ash-fall tuff matches phreatoplinian ash on the grounds of abundant accretionary lapilli, fine grain size, and association with redeposited volcaniclastics. The presence of possible local dewatering features is consistent with rapid deposition of fine, saturated ash. However, the occurrence of co-eruptive ignimbrite introduces another possibility of derivation from ash clouds generated from pyroclastic flows by primary (e.g. elutriation of fine particles; Curtis, 1968; Sparks and Walker, 1977; Walker, 1981c) and/or secondary processes (e.g. rootless vent explosions; Walker, 1979). Consideration of volumes is inconclusive: subunit 2 ignimbrite is sufficiently voluminous (36 km^3 DREV) to have generated the relatively small volume (4 km^3 DREV) of ash-fall tuff preserved in subunit 3 (*cf.* Walker, 1971; Sparks and Walker, 1977), and phreatoplinian ash deposits have widely ranging volumes (78 km^3 DREV for the Wairakei Formation, Self, 1983; 1.02 km^3 DREV for Hatepe ash, Walker, 1981a). Without necessary dispersal data, the relative importance of genuine phreatoplinian activity as opposed to co-ignimbrite ash-fall in the genesis of subunit 3 cannot be confidently evaluated.

THE VOLCANOLOGY OF THE CANA CREEK TUFF

The Cana Creek Tuff Member is the product of a large magnitude explosive eruption of vesiculated rhyolitic magma. Furthermore, it preserves evidence of the involvement of external water in its eruption and emplacement. Thus the Cana Creek Tuff eruption is considered to have been hydrovolcanic. The basis for this conclusion is explained below.

Models for hydrovolcanic eruptions and the character of their deposits

Recent advances have been made in understanding the influence of water-magma interaction on the explosivity of an eruption and the character of its deposits (e.g. Colgate and Sigurgeirsson, 1973; Self and Sparks, 1978; Sheridan and Wohletz, 1981,1983a; Sheridan *et al.*, 1981; Self, 1983; Wohletz, 1983). The model for hydrovolcanic eruptions developed by these workers is supported by observations of small magnitude eruptions which are predominantly basaltic in composition (e.g. Moore, 1967; Kienle *et al.*, 1980). Examples of deposits of similar magnitude though more silicic in composition (e.g. Schmincke *et al.*, 1973; Sheridan and Updike, 1975; Yokoyama, 1981) may indicate that magma chemistry is not critical; however, wider validity has yet to be rigorously demonstrated.

Sheridan and Wohletz (1981,1983a) and Wohletz (1983) argue that the behaviour of hydrovolcanic eruptions is fundamentally controlled by the efficiency of transfer of magmatic thermal energy to mechanical energy of the fluid-particulate mixture when water and magma initially meet. The efficiency of energy conversion is a function of the water:magma mass ratio, the level of superheating of the water and the degree of fragmentation of magma. Patterns of hydrovolcanic activity are explained in terms of fluctuations in the efficiency of energy transfer as the water:magma mass ratio changes (Sheridan and Wohletz, 1981, Fig.1; 1983a, Figs. 5,6; Wohletz, 1983, Fig. 1). The spectrum ranges from activity essentially magmatic in character in which water:magma mass ratios are low and little of the explosive energy is derived from vapourisation of water, through a peak when steam-rich pyroclastic flows, ash clouds and surges are produced. At this stage superheated steam separates from the pyroclasts during transport so flows are hot and deposits are dry. Further elevation of the water:magma mass ratio reduces explosivity and promotes the early condensation of steam in the particulate mixture. Wet, cold flows (mudflows, mud hurricanes, wet surges) produce saturated deposits prone to redeposition.

Reconstruction of large magnitude, rhyolitic hydrovolcanic eruptions is presently dependent on analogy with these smaller scale phenomena, on constraints provided by theoretical approximation of physical parameters and their relationships (e.g. Bennett, 1972; Colgate and Sigurgeirsson, 1973; Peckover *et al.*, 1973; Settle, 1978; Wilson *et al.*, 1978,1980; Wohletz, 1983), and on interpretation of detailed granulometric data obtained from their products (e.g. Self and Sparks, 1978; Walker, 1981a; Self, 1983).

Regardless of magnitude or composition, typical deposits from hydrovolcanic eruptions have several features in common. The association of primary pyroclastic units with compositionally similar redeposited epiclastic units is part of the diagnosis of hydrovolcanism (Sheridan and Wohletz, 1981,1983a; Self, 1983). Primary units are commonly water-saturated when emplaced and are readily remobilised by mass-flows and sheetwash floods. Hydrovolcanic ash is especially fine grained (Walker, 1973a,1981a,c,d; Self and Sparks, 1978). Rapid vapourisation of water by magma increases the explosivity of eruptions and enhances fragmentation of the magma (Colgate and Sigurgeirsson, 1973; Honnorez and Kirst, 1976; Self and Sparks, 1978; Sheridan and Wohletz, 1983a; Wohletz, 1983). Another sign of hydrovolcanism is the presence of abundant accretionary lapilli in both flow and fall deposits, attributable to the coincidence of fine grained suspended ash and steam (Moore and Peck, 1962; Self and Sparks, 1978; Brazier *et al.*, 1983; Self, 1983).

Additional features displayed principally by the primary pyroclastic deposits of large magnitude rhyolitic hydrovolcanic eruptions are the exceptionally wide dispersal of airfall ash, the low grade of associated ignimbrites, and the bimodal grain size and distinctive shape of ash pyroclasts. Explanation of these features requires reference to current models for explosive ignimbrite-forming eruptions.

Eruptions of large magnitude at dry vents produce columns in which the lower part is dominated by gas thrust driven by decompression of exsolved magmatic gas and the upper part is driven by convective thrust sustained by buoyancy (Sparks and Wilson, 1976; Wilson, 1976; Sparks *et al.*, 1978; Settle, 1978; Wilson *et al.*, 1978). For a given volume eruption rate, maintenance of the convecting plume and the height it reaches depend on efficient transfer of the magmatic heat of the pyroclasts to air entrained in the lower portions of the column so that its density remains less than that of the surrounding atmosphere (Sparks and Wilson, 1976; Wilson, 1976). Heat transfer efficiency in turn is primarily controlled by the size and surface area of pyroclasts, as thermal energy is more readily available from small particles (Sparks and Wilson, 1976; Settle, 1978). In explosive silicic eruptions, disruption of vesiculating magma by bursting of gas bubbles produces gas and hot rigid shards in abundance at the fragmentation surface in the conduit (Sparks, 1978b; Wilson *et al.*, 1980). If shards mix with water in submersed vents, they are subjected to a second stage of fragmentation due to rapid vapourisation of water and steam expansion before explosive ejection (Self and Sparks, 1978; Wohletz, 1983). Scanning electron microscope studies reveal that plate-like, pyramidal and blocky, angular pyroclasts are characteristic of silicic hydrovolcanic ash (Heiken, 1972; Self and Sparks, 1978; Self, 1983; Wohletz, 1983). Such ash may in detail be bimodal: the coarse particles (larger than 63 μm, Wohletz, 1983; 100 to 200 μm, Self, 1983) are shards from bursting of vesicles in pumice; the finer mode (less than 50 μm, Wohletz, 1983; less than 30 μm, Self, 1983) consists of fragments probably derived from shattering of these bubble-wall shards by hydrovolcanic processes.

Hydrovolcanic eruptions involving vesiculated magma supply highly fragmented pyroclasts to the column and convective plume. However, part of the thermal energy of the magma is used in the conversion of water to steam at the site of eruption so there is less magmatic heat available from these pyroclasts to promote convection than for equivalent volume discharge rates from dry vents (Wilson *et al.*, 1978). Wilson *et al.* (1978) predict that the ability of the eruption to sustain a convective plume will be further retarded by condensation of steam in the column, one effect of which is to increase its density. Lower eruption column heights and greater propensity to collapse to form pyroclastic flows are thus to be expected of eruptions where sufficient water has access to the magma than for dry vent eruptions of comparable intensity (or emission rate: Walker, 1980). Dispersal of airfall ash from the convective plume need not be impaired however, since the lower eruption column may be adequately compensated by the fine grain size of pyroclasts (Self and Sparks, 1978). Pyroclastic flows so generated deposit low-grade ignimbrite probably because the eruptions were water-cooled (Self, 1983; Walker, 1983). Thus, pyroclastic flows of hydro-volcanic origin do not necessarily involve column collapse from great heights inferred to be responsible for cooling of pyroclastic flows producted at dry vents (Sparks and Wilson, 1976; Sparks *et al.*, 1978). The superior fluidisation predictable for fines-rich pyroclastic flows (Sparks, 1976; Wilson, 1980) would enhance outflow, in spite of the normally inhibiting effects of collapse from eruption columns of low heights (Sparks *et al.*, 1978). Perhaps under such circumstances, superheated steam entrapped at the vent provides the gas for fluidisation, compensating for the reduced supply of residual magmatic gas available from the relatively cool pyroclasts.

The Cana Creek Tuff: evidence for its hydrovolcanic origin

In most modern and Cainozoic volcanic terrains where inundation of vent areas can be demonstrated with confidence, the operation of water-magma interaction is the starting point for volcanological analysis of eruptions on the basis of their deposits. For ancient pyroclastic rocks water-magma interaction can only be inferred from the association, distribution and character of the lithofacies preserved. In spite of the limitations imposed by age and incomplete exposure, the evidence in favour of a hydrovolcanic origin for the Cana Creek Tuff Member is convincing (Table 3.3). Water-cooling of vesiculated pyroclasts at the vent accounts for the low grade of Cana Creek Tuff ignimbrites and is possibly also responsible for their fine grain size. The flow deposits lack any signs of being hot when emplaced but were fluidised sufficiently to elutriate fine ash. Superheated steam evidently separated efficiently from the pyroclastic flows but condensed in co-ignimbrite ash clouds from which accretionary lapilli rains fell (*cf.* Sheridan and Wohletz, 1981). The association of low-grade ignimbrite with water-flushed, fine grained ash-fall tuff and volcaniclastic rocks made of pumice and crystals redeposited by floods, further supports the contention that the Cana Creek Tuff was generated by a hydrovolcanic eruption.

The water reservoir for the Cana Creek Tuff hydrovolcanic eruption

For ancient hydrovolcanic deposits, the requirement of a water reservoir at the site of eruption with adequate capacity has to be compatible with the known geological context. The existence of calderas in the source area of the Currabubula Formation has been inferred from consideration of the ignimbrite sheets it contains (Chapter 2). Lakes are a characteristic feature of modern calderas (e.g. Lake Rotorua, Lake Taupo, Yellowstone Lake) and a Late Carboniferous caldera lake may have provided the reservoir required by the hydrovolcanic eruption of the Cana Creek Tuff. However, another potential source should be considered in view of the broader

TABLE 3.3: Comparison of the Cana Creek Tuff and other ignimbrite members of the Currabubula Formation with common features of the deposits from large magnitude, rhyolitic hydrovolcanic eruptions

Characteristics of deposits from HYDROVOLCANIC ERUPTIONS:	CANA CREEK TUFF:	Other ignimbrite members of the Currabubula Formation:
Association of primary pyroclastic facies with compositionally equivalent redeposited volcaniclastic facies	PRESENT	absent
Fine grain size of ignimbrite and airfall ash - substantial sub-millimetre and fine (<63 μm) ash	√	ignimbrites 'normal' (ash-fall tuff absent)
Shard shapes - coarse mode: vesicle walls	√	√ but flattened
- fine mode: fragments of vesicle walls	fines present, shape not established	fines very minor, shape not established
Accretionary lapilli	ABUNDANT	absent
Low-grade ignimbrite (i.e. non-welded even though of substantial thickness)	√	all almost entirely welded
Wide dispersal of airfall ash	?	?
Erupted from submersed or 'wet' vent	ice and snow, and/or caldera lake inferred at source	vents 'dry'

palaeogeographic context of the region in the Late Carboniferous. Epiclastic facies influenced by source area alpine glaciation are widespread in Late Carboniferous sequences of northeastern New South Wales (Crowell and Frakes, 1971a,b). The speculation that volcanic facies in the same sequence might also be so influenced is warranted. In particular there are two documented occurrences of glacial sedimentary rocks in the immediate vicinity: the Rosedale Member of the Currabubula Formation at Currabubula (Whetten, 1965) and the Spion Kop Conglomerate 60 km to the north (White, 1968; disputed by Lindsay, 1969). Both occurrences are at stratigraphic levels close to but below the Cana Creek Tuff Member. That this Member was generated by a hydrovolcanic eruption for which the water supply was melted from glacier ice and/or snow, perhaps ponded in a caldera depression, is considered feasible. Although smaller in magnitude, the 1912 eruption of Novarupta, Alaska, took place in an ice- and snow-bound setting and produced an assemblage of primary and redeposited facies (Curtis, 1968; Hildreth, 1983) similar to that of the Cana Creek Tuff.

The reconstruction of the course of the Cana Creek Tuff hydrovolcanic eruption attempted below, and illustrated in Figure 3.12, incorporates the possibility of ice and/or snow impoundment of the source caldera.

The Cana Creek Tuff eruption

The volcaniclastic facies of the lowest subunit in each section registers the onset of eruptions and implies the prior existence of voluminous unconsolidated pyroclastic debris (pumice, shards, crystals) at more proximal localities. The volcaniclastic facies is interpreted to comprise sheetflood and debris flow deposits (Fig. 3.12a) and provides evidence that the proximal pyroclastic accumulations were water-saturated. It is possible that from the outset, the vent for the eruption was submersed and that water was ejected with pyroclasts (*cf.* Walker, 1981a). Even if the vent was dry to begin with, the initial primary pyroclastic deposits may have been dumped onto ice or snow, or into water, and subject to redeposition as a result of slumping (*cf.* Curtis, 1968; Crandell, 1971; Janda *et al.*, 1981; Hildreth, 1983) and/or secondary phreatic explosions (*cf.* Walker, 1979,1981c,e; Rowley *et al.*, 1981; Hildreth, 1983; Self, 1983).

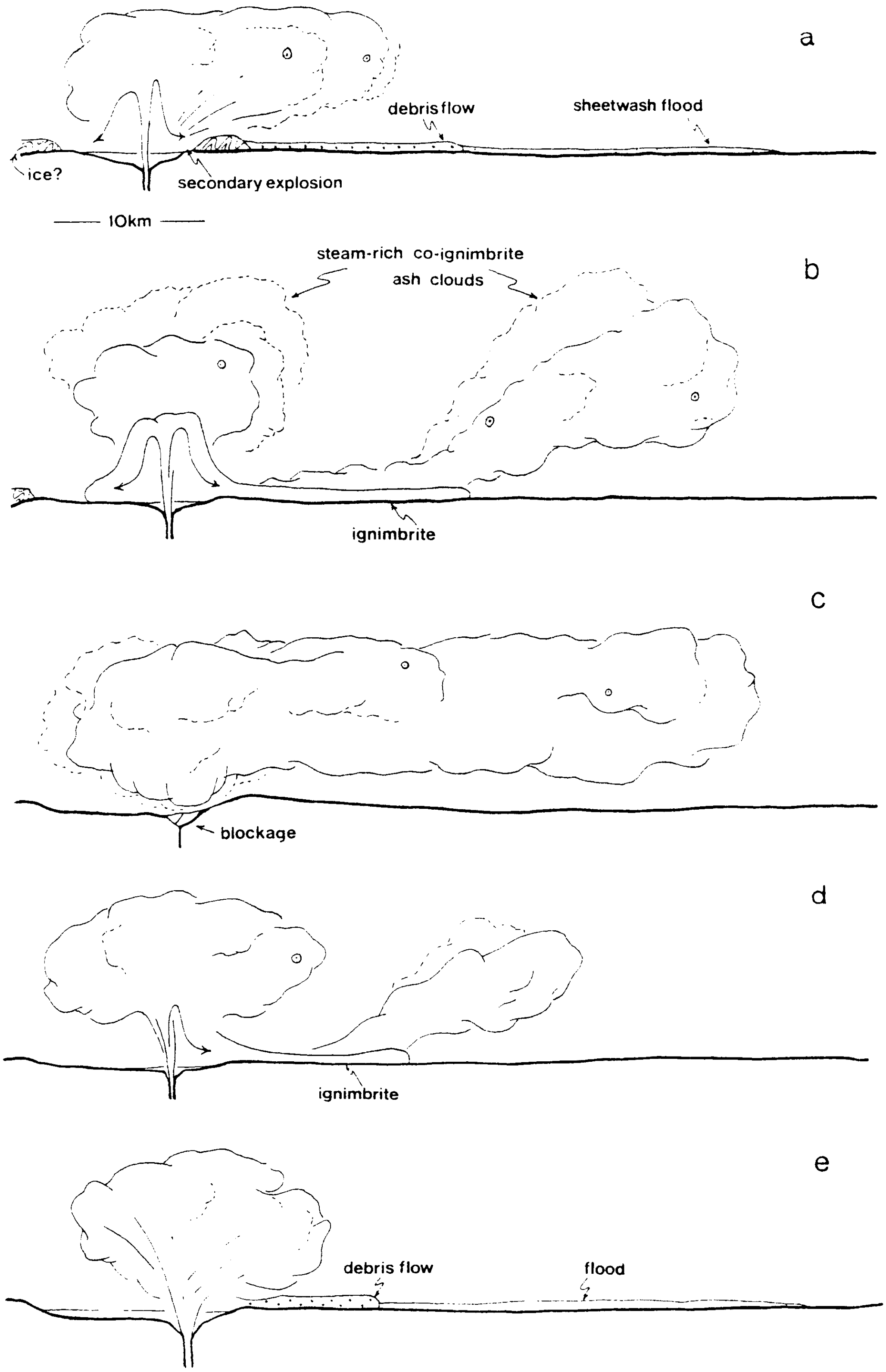


Figure 3.12: Schematic reconstruction of the eruption responsible for the Cana Creek Tuff Member (see text for further explanation).

There is a second clue indicating that water was interacting with the erupting vesiculated magma at an early stage: the lowest ignimbrites of subunit 2 contain intraclasts of fine grained accretionary lapilli-bearing ash-fall tuff. Fine wet ash could have been supplied by an opening phreatoplinian style eruption column, analogous to the plinian phase of ignimbrite-forming eruptions at dry vents (Sparks and Wilson, 1976; Sparks *et al.*, 1978; Wilson *et al.*, 1978; Wilson *et al.*, 1980). Water-flushing by condensation in such steam-rich ash clouds is known to result in the premature deposition of fine airfall ash close to the source (Self and Sparks, 1978; Sparks *et al.*, 1981; Walker, 1981a,c,e,f; Self, 1983).

The details of events at the source are lost, but the medial record is consistent with rapid elevation of the water:magma ratio at the start of the eruption and chaotic interaction of water with hot pyroclasts. Large volumes of pyroclastic material were redeposited from proximal sites, reaching more distal localities without undergoing modification by weathering or abrasion in fluvial systems. Saturated mass-flows and floods travel at velocities ranging between about 6 m/s and 40 m/s (e.g. Bijou Creek flood, Colorado, 6 m/s, McKee *et al.*, 1967; Mt St Helens debris flows and mud-flows, Washington, up to 40 m/s, Janda *et al.*, 1981). The first deluges produced by the Cana Creek Tuff eruption may have taken as little as half an hour to flow from the source eastward to Currabubula, followed by flood peaks several hours later.

Fine grained, low-grade, multiple flow unit ignimbrite of subunit 2 records the escape of water-cooled but primary pyroclastic flows from the source, signifying a shift in the water:magma mass ratio from that which prevailed during flooding responsible for subunit 1 (Fig. 3.12b). Either the vicinity of the vent was sufficiently drained by the earlier events, or the volume discharge rate of vesiculating magma increased.

Whether a full-scale phreatoplinian convecting eruption plume was established and maintained prior to sustained eruption column collapse is not discernible from the available data. Consideration of the special effects of the proposed involvement of ice and/or snow in the eruption suggests that any such stage was likely to have been shortlived. Additional expenditure of magmatic heat energy in converting ice to water prior to vapourisation might further retard the capacity of the system to maintain a high convective plume for those cases where the discharge rate would normally be adequate. Consequently early column collapse and production of steam-rich pyroclastic flows would be favoured instead of progression to an eruption of strictly phreatoplinian character.

The separation of steam from moving pyroclastic flows was evidently effective and deposits were dry and stable (*cf.* hot dry surges, Sheridan and Wohletz, 1981,1983a) but too cool to weld. Fine ash was elutriated by escaping steam and exsolving residual magmatic gas, and some of it was returned to the flows by accretionary lapilli rains. Estimates of velocities of pyroclastic flows range from less than 10 m/s to more than 200 m/s (e.g. Sparks, 1976; Francis and Baker, 1977; Wilson and Walker, 1982). Moderately fluidised pyroclastic flows move at 30 to 80 m/s (Wilson and Walker, 1982). Travelling at 50 m/s, Cana Creek Tuff pyroclastic flows erupted from the postulated source (Fig. 3.3a) would reach Currabubula in about 15 to 20 minutes. The entire stack of medial flow units of subunit 2 could have been emplaced within a couple of hours if, as the record suggests, there were no pauses in this stage of the eruption.

A change in the course of the eruption is reflected by the deposition of a thin layer of fine ash-fall tuff (subunit 3; Figs. 3.2,3.12c). Settling of airborne ash accompanied the emplacement of subunit 2 ignimbrite but this did not survive as discrete units. The formation of permanent beds was delayed until the supply of ignimbrite had been terminated, presumably in response to events at the source. If eruptions temporarily ceased altogether, a regional ash cloud no doubt lingered as has been inferred for a break in the Valley of Ten Thousand Smokes eruption (Hildreth, 1983). The atmosphere would have been charged with steam and fine ash produced at the source vent and as a by-product of prior pyroclastic flows, and ripe for wholesale flushing upon cooling and condensation of steam. Stratification in parts of the ash-fall tuff unit could be accounted for by deposition from a complexly-layered, mixed-generation ash cloud, enhanced by primary fluctuations in discharge from the vent (*cf.* Bond and Sparks, 1976; Walker, 1980) and/or by wind interference (*cf.* Curtis, 1968).

Even if subunit 3 originated from a full-scale phreatoplinian convecting plume, a spell in the eruption following discharge of subunit 2 pyroclastic flows is still implied : once a collapsing eruption fountain has been established at a dry vent, reinstatement of a plinian column requires reduction in vent radius and mass discharge rate or an increase in gas content (Sparks *et al.*, 1978; Wilson *et al.*, 1980). If the same constraints apply in hydrovolcanic systems, a depositional record that includes an ignimbrite to vent-produced ash-fall transition is also a record of a fundamental alteration in the course of the eruption.

The cause of the discontinuity in the course of the Cana Creek Tuff eruption, correlated with deposition of the ash-fall tuff subunit, is a matter of conjecture. Wilson *et al.* (1980, p.145,146) recount causes of temporary interruptions to eruptive styles, and conduit blockages due to collapse of vent walls are considered likely. In view of the common association of ignimbrite source areas with collapse features of a range of scales, such a process is indeed plausible in this case. Furthermore the subsequent behaviour of the Cana Creek Tuff eruption episode is consistent with the predicted effects of vent blockages. Transition from a collapsing eruption fountain (subunit 2 ignimbrite) to a convective plume (subunit 3 ash-fall tuff) requires conduit restriction and reduced eruption intensity (Wilson *et al.*, 1980), both of which would be expected if vent walls caved in. If the eruption fountain ceased altogether, the constriction of the vent by collapse in the interim would have produced a convective plume at least briefly, on resumption of the eruption (Sparks *et al.*, 1978; Wilson *et al.*, 1980). In each of these circumstances, foundering of the vent might have temporarily impeded water-magma access, thereby producing water:magma mass ratios initially conducive to peak explosive efficiency and a phreatoplinian plume. This would be shortlived however, because ponding of water in the vicinity of the subsided vent would shift the balance in favour of a collapsing pyroclastic flow-forming fountain (subunit 4 ignimbrite).

Eventual reinstatement of an eruption column, whether or not persistently phreatoplinian, is indicated by the succeeding ignimbrite of subunit 4 (Fig. 3.12d). The deposits of this episode are much less voluminous and missing from some sections (B,E) although otherwise similar to subunit 2 ignimbrite in being low-grade and containing sparse accretionary lapilli. Inferred general reduction in the magnitude and intensity of the eruption must be qualified by the possibility of significant removal of the original record of this phase by subsequent erosion. However, such an inference simplifies the overall pattern of the eruption to one of gradual waning following peak discharge during the earlier ignimbrite stage, with one brief but significant interruption (prior to or during emplacement of subunit 3).

It is perhaps this decline in magnitude that led to production of the volcaniclastic facies of subunit 5 similar to deposits of the opening stages of the eruption (Fig. 3.12a,e). Subunit 4 ignimbrite was the final primary deposit to reach medial areas. Weaker eruptions probably continued, temporarily piling up saturated pyroclastic debris at the source, destined to be redeposited. It is also likely that earlier, more proximal primary deposits from the eruption were eroded at this stage. There would have been little topographic impediment or control of the sediment-laden floods because of the landscape-smoothing effects of prior ignimbrite. Widespread large-scale inundation of remnants of the braidplain fluvial network bordering the pyroclastic apron presumably followed cessation of eruptions, as happened when the Wairakei Formation eruptions terminated (Self, 1983).

FACIES VARIATIONS IN DEPOSITS FROM A LARGE MAGNITUDE HYDROVOLCANIC ERUPTION

Figure 3.13 illustrates a model for the correlation of the facies of the Cana Creek Tuff Member with the inferred events at the site of the eruption, and with facies likely to occur at more distal localities.

Eruptions began from a vent with excess water available (submersed?) and subsequent events only temporarily redressed the situation. Undisturbed pyroclastic units would be unlikely to survive close to the source when water is significant in the ejecta because the early record of a voluminous hydrovolcanic eruption is marked by erosive breaks. Deposition of each of the three main ash-fall units of the Wairakei Formation was followed by a shortlived erosive event which partly or completely removed the eruption record within 30 km from the vent (Self, 1983). One can reasonably infer that the lost proximal pyroclastic facies were transformed into medial and distal volcaniclastic facies, such as are preserved at the base and

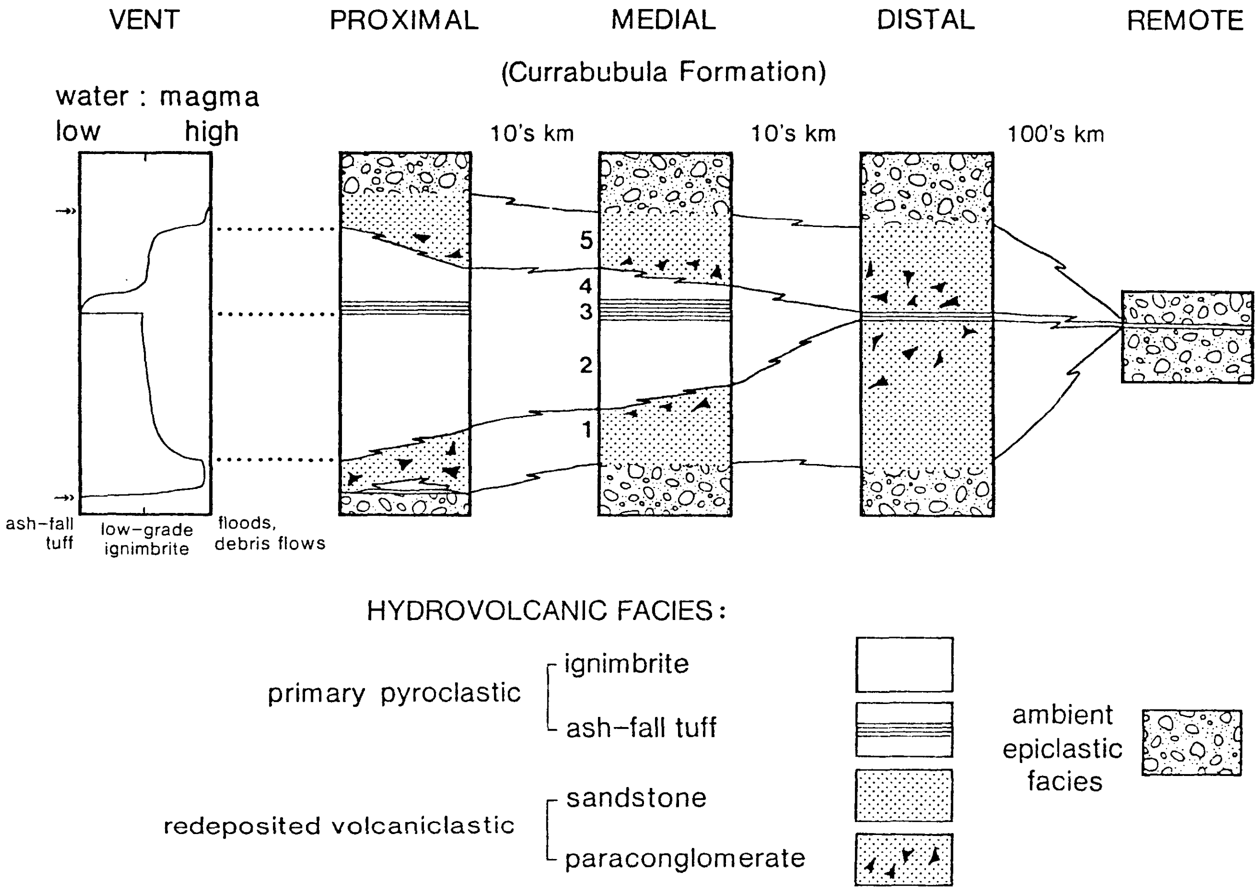


Figure 3.13: Proximal, distal and remote facies associations produced by a large magnitude, explosive, hydrovolcanic eruption. Medial facies are exemplified by the Cana Creek Tuff Member of the Currabubula Formation. The internal stratigraphy of the hydrovolcanic facies is correlated with the postulated temporal variation in water:magma mass ratio at the vent.

top of the Cana Creek Tuff Member. The most remote symptoms of the onset and decline of this type of eruption would be flood deposits in fluvial sediments in which the volcaniclastic components were conspicuous but modified by epiclastic reworking and diluted by contributions from other sources.

The model of Sheridan and Wohletz (1981) implies that the medial ignimbrites would be represented by pumiceous debris flow deposits at more distal locations as a result of the cooling and condensation of entrapped steam with increasing distance from the source. Continued sedimentation and separation of water-rich flows might be expected to produce a gradation into volcaniclastic fluvial facies:debris flows dominating on the outer reaches of the pyroclastic apron would give way to sheetwash floods at more distal localities (*cf.* Collinson, 1978a; Miall, 1978; Rust, 1979; Gloppen and Steel, 1981; Nilsen, 1982). The vent eruption and the pyroclastic flows probably both generated large volumes of fine ash. Its remote dispersal (that is, hundreds of kilometres from source) may have been enhanced by the fine grain size, as has been proposed for the widespread Wairakei Formation ash (Self, 1983).

Following the eruption, new drainage systems established on the transformed landscape gradually encroached on the great expanse of devastation, leading to vent-ward transgression of volcaniclastic facies and eventual reinstatement of gravel-dominated sedimentation typical of the ambient braidplain system.

A PLEISTOCENE ANALOGUE FOR THE CANA CREEK TUFF

Recognition of the hydrovolcanic character of the Cana Creek Tuff was prompted by its remarkable lithological similarity to the 20,000 year old Wairakei Formation (Vucetich and Pullar, 1969; Self and Sparks, 1978; Self, 1983), regarded as typical of the deposits of large-scale silicic phreatomagmatic volcanism (Sheridan and Wohletz, 1983a, p.7,17; Walker, 1981d, p.236, 1983, p.76). Parallels between the Cana Creek Tuff and Wairakei Formation in the sequence and distribution of facies are as striking as those in lithologies (Table 3.4). The Wairakei Formation has two ignimbrite members each underlain by ash-fall units primarily of phreatoplinian origin. Self (1983) identified the change from the lower ignimbrite to plinian and phreatoplinian airfall ash above as a sign of an alteration of the eruption sequence and interpreted it as symptomatic of reduced water-magma interaction and/or vent blockage due to subsidence. Much of the primary record from the Wairakei eruption was reworked and deposited by fluvial systems encroaching on the pyroclastic fan, and now constitutes part of the pumiceous, quartzofeldspathic Hinuera Formation (Hume *et al.*, 1975).

Regardless of the causes, the match betweeen the Cana Creek Tuff and the Wairakei Formation suggests the operation of common processes peculiar to those eruptions which combine large magnitude and hydrovolcanic exchange. The main differences are those of scale and the role of a genuine phreatoplinian convecting plume in producing ash-fall tuff. Scale discrepancies are however difficult to assess as the Late Carboniferous unit is plainly only a fragment of the original deposit and considerable vagaries pervade volume calculations because the vent location is unknown. At least the order of magnitude is comparable but correspondence in dispersal is impossible to confirm. There is only scant evidence for a sustained phreatoplinian convecting plume prior to formation of the Cana Creek Tuff ignimbrites. Perhaps ice and/or snow involvement in the Cana Creek Tuff eruption promoted early column collapse to form pyroclastic flows rather than maintenance of a convecting phreatoplinian plume.

ANOTHER EXAMPLE OF LATE CARBONIFEROUS SILICIC HYDROVOLCANISM?

The Cana Creek Tuff Member is unusual but not unique in the Late Carboniferous sequence of New England. The Ermelo Dacite Tuff (McKelvey and White, 1964; McKelvey, 1968) is an extensive unit in the northern portion of the same belt of Late Carboniferous volcanogenic sedimentary rocks and intercalated ignimbrites. Reconnaissance inspection (McPhie, unpublished data) and published descriptions reveal diverse facies (non-welded ignimbrite; accretionary lapilli-bearing ash-fall tuff, crystal-rich sandstone, boulder grade paraconglomerate) and a complicated internal stratigraphy, contrasting with other welded ignimbrite sheets in the rest of the sequence. A hydrovolcanic origin for this unit is consistent with its presently known characteristics. In addition, correlations of the

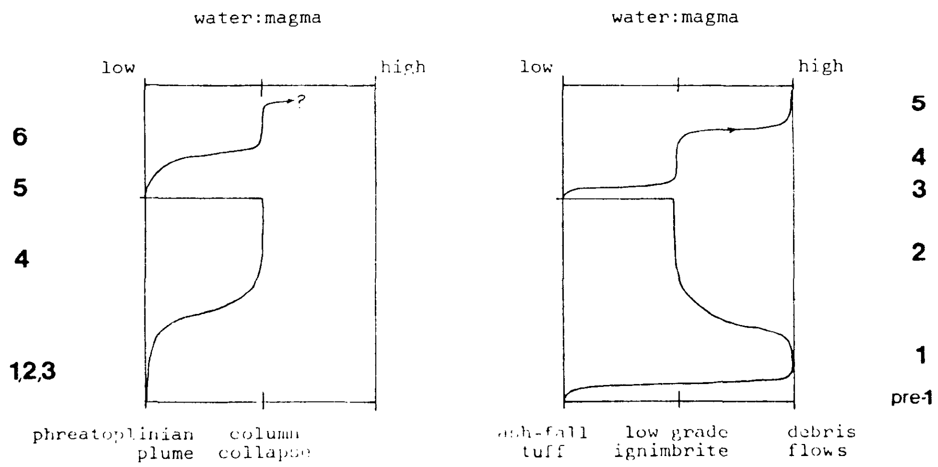
TABLE 3.4: Comparison of two large magnitude rhyolitic hydrovolcanic eruptions. Pleistocene example is based on Hume et al. (1975), Self and Sparks (1978) and Self (1983).

	20 000 Y B.P., TAUPO, N.Z. Wairakei & Hinuera Fms, Mokai Sand	LATE CARBONIFEROUS, N.S.W. Cana Creek Tuff Member
FACIES	**Fine grained** rhyolitic airfall ash + **low-grade** ignimbrite. Fluvial, lacustrine & aeolian **volcaniclastic** sediments. **Accretionary lapilli** present in all facies.	Fine grained rhyolitic ash-fall tuff + **low-grade** ignimbrite. **Volcaniclastic** sandstone & paraconglomerate. **Accretionary lapilli** present in pyroclastic facies.
FACIES DISTRIBUTION	Pyroclastic facies dominant near source; erosion surfaces between members. Redeposited & reworked volcaniclastics below (?), above & at distal margins of pyroclastic facies. Widespread remote fine airfall ash (>800km).	(no data) Redeposited volcaniclastic facies below & above pyroclastic facies in medial exposures. (no data)
MAGNITUDE (DREV)	~155 km³	~100 km³
ENVIRONMENT OF DEPOSITION	Terrestrial, cool, **periglacial** tundra, poorly vegetated.	Terrestrial, proximity to alpine **glaciers**, poorly vegetated.
RESERVOIR	Caldera lake (Lake Taupo).	Meltwater? and/or caldera lake.

INFERRED ERUPTION SEQUENCE

	MEMBERS:	INTERPRETATION (Taupo)	INTERPRETATION (Cana Creek)	SUBUNITS:	
post-6	(Mokai Sand, Hinuera Fm.)	epiclastic redeposition & reworking		volcaniclastics	5
6	airfall ash / ignimbrite	secondary explosions / column collapse		ignimbrite	4
5	airfall ash	phreatoplinian convective plume	co-ignimbrite ± phreatoplinian plume	ash-fall tuff	3
	hiatus	vent blockage (subsidence?)		hiatus	
4	ignimbrite + airfall ash	column collapse		ignimbrite + accretionary lapilli	2
	proximal ignimbrite?	minor column collapse?	secondary explosions / column collapse?		
1,2,3	airfall ash	phreatoplinian convective plume	phreatoplinian plume?	volcaniclastics	1

INTERPRETED WATER:MAGMA MASS RATIO VARIATION



northern Late Carboniferous section with the Currabubula Formation (Chapter 4) indicate that the Ermelo Dacite Tuff and the Cana Creek Tuff are approximate time equivalents. The expanse which separates the northern limit of the Cana Creek Tuff from the southern limit of the Ermelo Dacite Tuff includes the area of outcrop of tillite described by White (1968). These preliminary data suggest a repetition of the association of widespread hydrovolcanic ignimbrite and ash-fall tuff with a Late Carboniferous glacial interlude.

CONCLUSIONS

The Cana Creek Tuff Member of the Currabubula Formation is the product of a large magnitude, explosive eruption involving vesiculating rhyolitic magma and external water. The features which collectively indicate its hydrovolcanic origin are the combination of primary pyroclastic facies with compositionally similar volcaniclastic facies which were emplaced by saturated debris flows and sheetwash floods; the abundance of accretionary lapilli in both ignimbrite and ash-fall tuff of the primary pyroclastic facies; ignimbrite which is uniformly low-grade; the fine grain size of pyroclasts in the ignimbrite and ash-fall tuff. Furthermore, the composition, facies and stratigraphy of the Cana Creek Tuff are very similar to the medial and distal deposits from a Pleistocene phreatomagmatic eruption of comparable magnitude at Taupo, New Zealand (Wairakei Formation, Self, 1983; Hinuera Formation, Hume *et al.*, 1975).

The internal stratigraphy of the Cana Creek Tuff can be correlated with fluctuations in the eruptive style which are interpreted to reflect alterations in the balance between water and vesiculating magma. The emplacement of relatively cool pyroclastic flows was preceded and followed by floods and debris flows which remobilised more proximal accumulations of water-saturated pyroclastic debris. Thus, most of the medial and distal records of such large magnitude, hydrovolcanic eruptions are likely to comprise redeposited volcaniclastics at the base and top, with primary pyroclastic units in between. Ideally, the deposits may show a lateral transition in facies from proximal primary pyroclastics to more distal redeposited volcaniclastics. Significant deviations from these simple vertical and lateral facies patterns will be produced by more complicated eruption sequences, and also result from the interference of local, contemporaneous erosion. The water to magma mass ratio at the eruption site is considered to be a major control on the character of the hydrovolcanic facies so generated.