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EXCURSION GUIDE

INFLUENCE OF ENVIRONMENT ON CONTRASTING STYLES OF VOLCANISM, VOLCANICLASTICS, AND SEDIMENTATION: THE DEVONIAN BOYD VOLCANIC COMPLEX, SOUTHEASTERN AUSTRALIA

Ray Cas and Stuart Bull

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Influence of Environment on Contrasting Styles of Volcanism, Volcaniclastics, and sedimentation: the Devonian Boyd Volcanic Complex, Southeastern Australia

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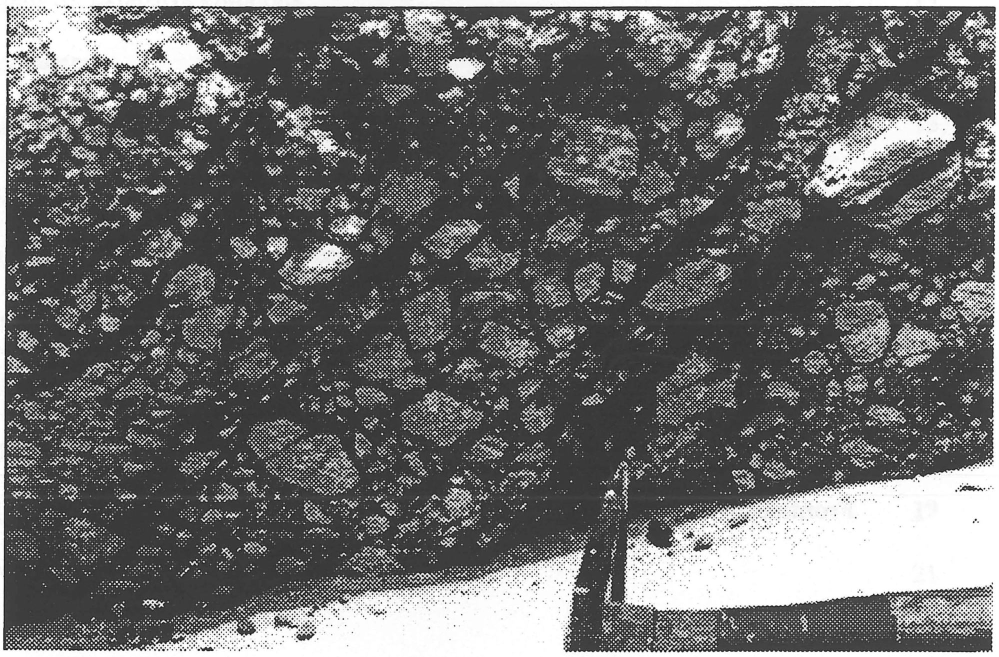

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Fronticepiece: Hyaloclastite with clast rotated textures, Bunga Head.

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1-INTRODUCTION: Regional Setting of Palaeozoic Volcanism in Southeastern Australia

GENERAL GEOLOGY AND HISTORY OF VOLCANISM OF AUSTRALIA

Volcanic rocks are represented in all the rock systems of Australia from the Archean through to the Recent. Komatiitic volcanics are common in the Archean of Australia, with felsic volcanics being of only minor significance. From the Proterozoic onwards, komatiitic volcanics become rare, and conversely, felsic volcanics become more prominent although mafic volcanism remained a significant part of the volcanic record.

As elsewhere in the world, there is on-going debate about the tectonic significance of Australia's ancient volcanic successions, especially for the Archean to mid-Paleozoic volcanics. The common tectonic scenarios that are debated are subduction related arc setting versus rift setting, and plate margin versus intraplate settings. The occurrence of fault bounded mafic-ultramafic complexes ("ophiolitic" complexes) associated with many of these ancient volcanic successions has also fueled the debate on how significant oceanic plate and island arc tectonic settings were compared with continental plate tectonic settings. Lastly, the application of terrane concepts is an on-going process, with considerable work being required to establish the terrane status of crustal blocks of all ages.

In general, the age of basement complexes youngs to the east (Rutland 1976). Archean belts are almost completely confined to the western one third of the continent, Proterozoic successions are confined to the western two thirds whereas the eastern third of the continent consists of Paleozoic basement complexes. This pattern gives the impression of progressive continental accretion in an eastwards direction by plate margin subduction related processes. The rock record is consistent with numerous orogenic events throughout geologic time that would support such an interpretation, but many uncertainties remain to be resolved were such a simple scenario to be adopted. For example, in eastern Australia, one of the great uncertainties is whether or not the pre-Paleozoic basement was ensialic or ensimatic (Cas, 1983).

From the Carboniferous and Permian on, there is the clearest indication of a plate margin arc setting along the eastern margin of Australia. This arc, originally located in northeastern New South Wales (the New England arc), migrated northeastwards into eastern Queensland during the Mesozoic and earliest Tertiary, and appears to represent the precursor to the modern arc-trench system in the southwest Pacific (New Zea-land-Kermadec-Tonga-Vanuatu-Solomons) that represents the present day eastern margin of the Australian-Indian Plate. Nonetheless, the tectonic significance of the Mesozoic to Tertiary volcanics of the Queensland coastal belt is debatable. They could represent subduction related arc volcanics, or volcanics originating from behind-the-arc rifting associated with the eastwards migration of the active plate margin and the formation of the marginal basins that extend from the east coast of Australia to the current plate margin (e.g. Bryan et al., in press), or both.

From the Jurassic on, Australia's western, southern and eastern margins were subject to rifting associated with the breakup of Gondwanaland. Volcanism was associated with many of these rifted margins (Veevers 1984). In eastern Australia, rifting along the southern margin produced the Otway, Bass and Gippsland Basins along the south coast of Victoria, and rifting and back-arc spreading led to the opening of the Tasman Sea and Coral Sea basins along the east coast (Veevers 1984).

Intraplate volcanism, largely basic in character, has been a notable feature of the geological development of the eastern Australian coastal and highlands belt since the Jurassic (Johnson et al. 1989). Volcanics resulting from this intraplate volcanism are widespread along the eastern margin of the Australian mainland, and are also known from offshore oceanic settings (Johnson et al. 1989).

GEOLOGY OF SOUTH-EASTERN AUSTRALIA

The exposed basement in southeastern Australia consists of the deformed remains of the early to mid-Paleozoic Lachlan Fold Belt (LFB; Fig. 1). This orogenic belt consists of deformed metasedimentary, volcanic and granitic rocks. The LFB appears to have evolved through the cumulative effects of four cycles of crustal extension and compression from the beginning of the Cambrian to the Early Carboniferous (Cas 1983). During the Cambrian to end Ordovician the paleogeography was largely of an open marine basin marginal to the paleo-Australian craton. A number of mafic volcanic belts were active during the Cambrian in Victoria. There is debate as to whether or not these represent subduction related island arcs or rift related events, and whether or not the basement was oceanic or subsided continental crust (Crawford and Keays 1986; Cas, 1983). In the Ordovician an andesitic island arc appears to have been active in central New South Wales (Cas et al. 1980). A trench may have been located along and offshore from the coastline, and a marginal sea separated the arc from the continental margin which lay in

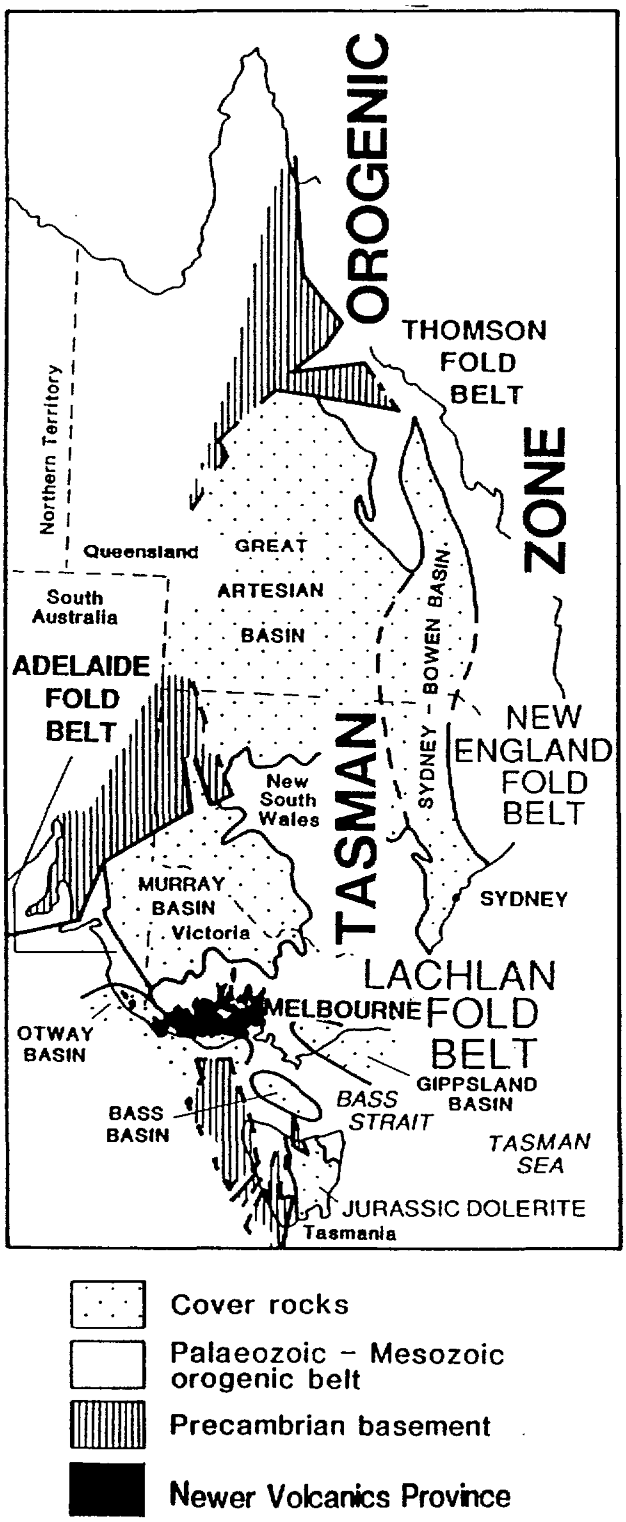

Figure 1: *Principal geological and tectonic elements of eastern Australia (after Cas 1983, Gray 1988).*

western New South Wales (Fig. 2; Cas, 1983, Powell, 1984). This palaeogeographic assembly was terminated by the end-Ordovician Benambran Orogeny.

By the Siluro-Devonian the crust was largely sialic based on the widespread occurrence of Siluro-Devonian granites which occupy 25% of the area of the Lachlan Fold Belt. Based on the geochemistry of these granites, and geochemical differences in regionally extensive suites of granitoids as well as the model ages for their source rock successions, Chappell et al (1988) proposed that the LFB shows no evidence of the existence of supracrustal terranes, but appears to show evidence of the existence of the remains of terranes in the lower crust that seem to have amalgamated by the beginning of the Paleozoic. According to this line of reasoning the Paleozoic LFB evolved on a largely sialic basement (Cas, 1983; Chappell et al., 1988). The Siluro-Devonian paleogeography (Fig. 3) was a mixed marine-subaerial one and the numerous basins of the LFB appear to have originated by extension or transtension, accompanied by voluminous, largely felsic magmatism following the Benambran Orogeny.

As a result of the mid-Devonian Tabberabberan Orogeny, the paleogeography became almost wholly subaerial. By the Late Devonian another cycle of crustal extension producing rift basins and bimodal volcanism, including major caldera complexes, reactivated the tectonics of the LFB (Fig. 4). During the early Carboniferous Kanimblan Orogeny the crust of southeastern Australia was fully cratonised and stabilised into a sialic crustal complex of deformed metasedimentary and meta-igneous rocks ranging from zeolite to greenschist and locally to amphibolite facies metamorphic grades.

From the Devonian to Permian the New England Fold Belt (Fig. 1) evolved as a major plate margin orogenic belt in northeastern New South Wales and Queensland. During the latest Carboniferous, early Permian and Triassic, orogenic uplift of the New England Orogen led to the Formation of the Sydney-Hunter- Bowen Basin system as a foreland basin to the west of the New England Orogen (Herbert 1980; Fig. 5). The basement of the Sydney Basin consists of the deformed succession of the eastern part of the Lachlan Fold Belt. Early highland environments were soon inundated by transgression of the sea during the Permian. Australia lay in circumpolar latitudes at this time, and this reflected by the remains of glacial and iceberg sedimentary deposits.

Following Permo-Triassic orogeny tectonically, southeastern Australia remained stable until the latest Jurassic-Early Cretaceous, when rifting associated with the breakup of Gondwanaland led to the formation of the Otway, Bass and Gippsland Basins along the southern margin of Victoria, the opening of the Southern Ocean between Australia and Antarctica, and the opening of the Tasman Sea along the east coast of the continent (Veevers 1984).

The Otway, Bass and Gippsland Basins (Fig. 1) were initiated in the latest Jurassic to earliest Cretaceous when the then contiguous land masses of Australia, Antartica and New Zealand began to extend, forming a linear continental rift basin system. The Otway rift basin continued to widen, with breakup (i.e. separation of Australia and Antarctica) occurring about 95 million years ago with the development of a seafloor spreading system in the Southern Ocean (Veevers, 1984). The east-west trending Otway Basin thus represents half of the original rift basin and has evolved into a passive continental margin basin

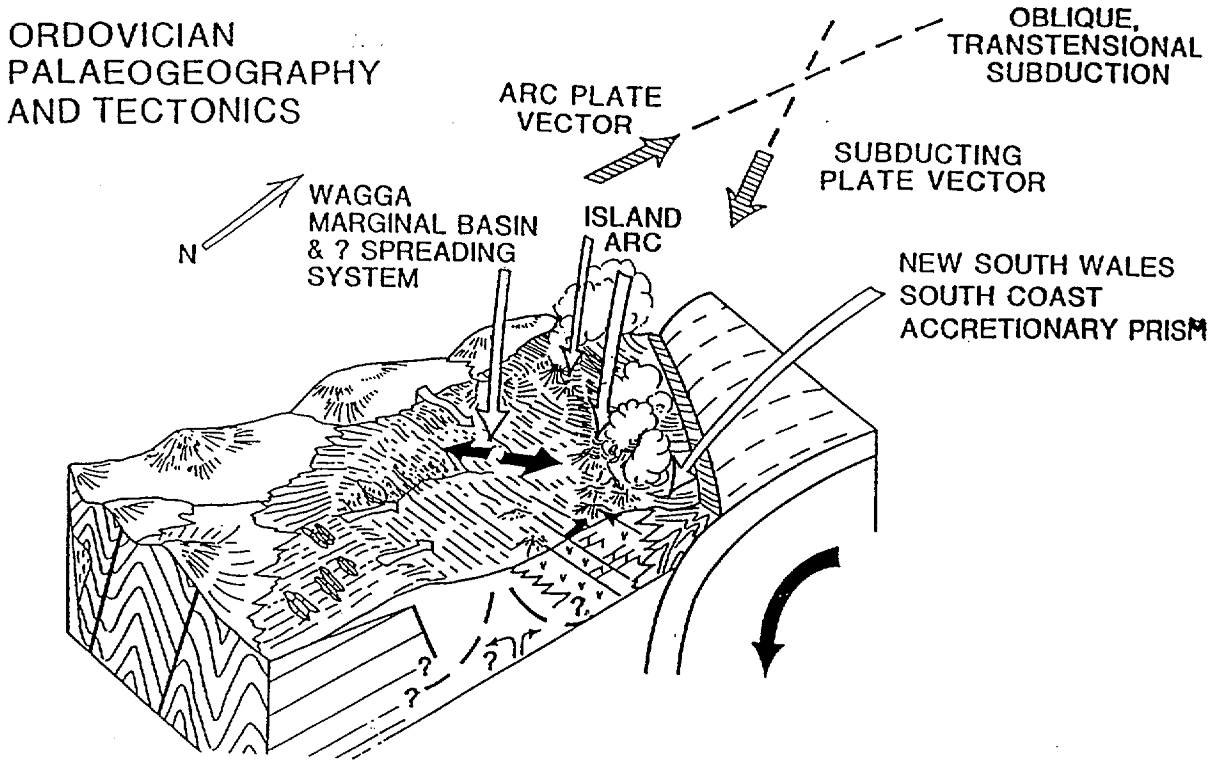

Figure 2: *Paleogeographic reconstruction of the Lachlan Fold Belt of southeastern Australia during the Ordovician (From Cas et al. 1980, Cas 1983)*

that is open to the ocean. Its basement thus consists of deformed sialic crust of the Paleozoic Lachlan Fold Belt onshore, but offshore this passes into younger oceanic crust. The Otway Basin is still an active basin with both active subaerial and marine erosional and depositional systems. The Bass and Gippsland Basins are failed rift basins, are wholly ensialic, and have led only to the limited separation of the mainland from Tasmania. The Bass

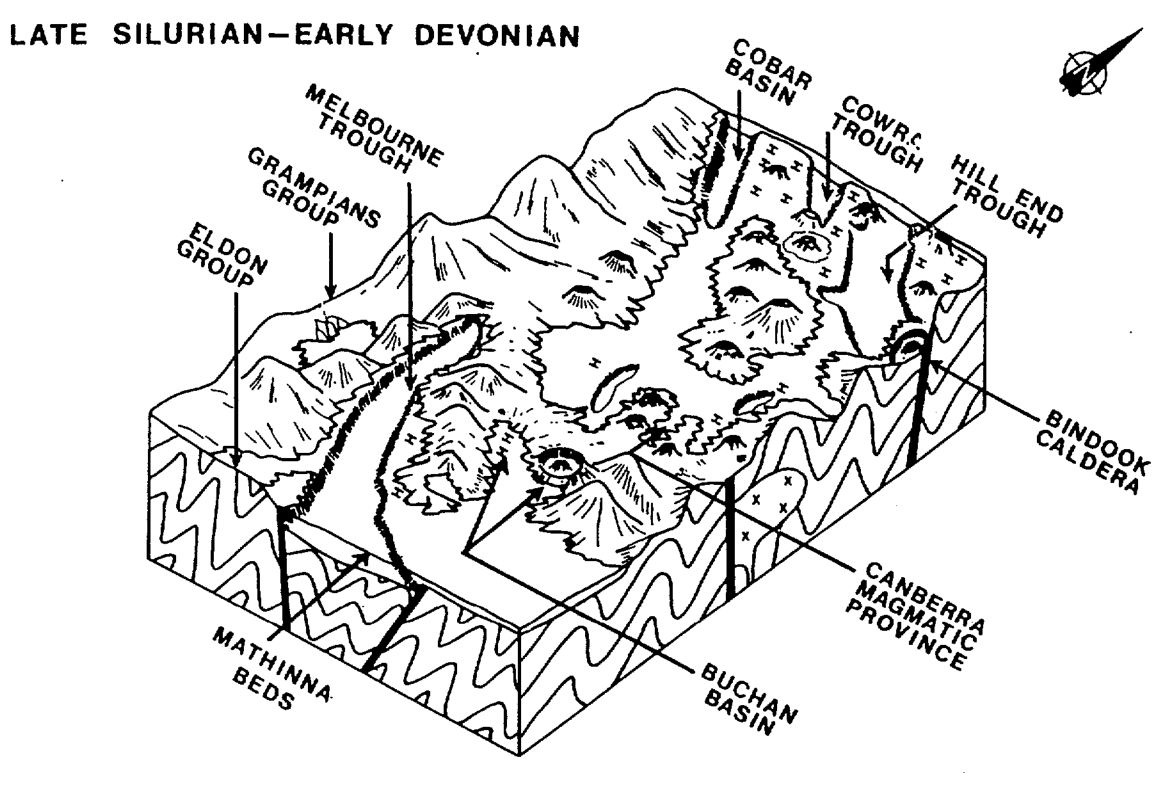

Figure 3: *Paleogeographic reconstruction of the Lachlan Fold Belt of southeastern Australia during the Siluro-Devonian (From Cas 1983)*

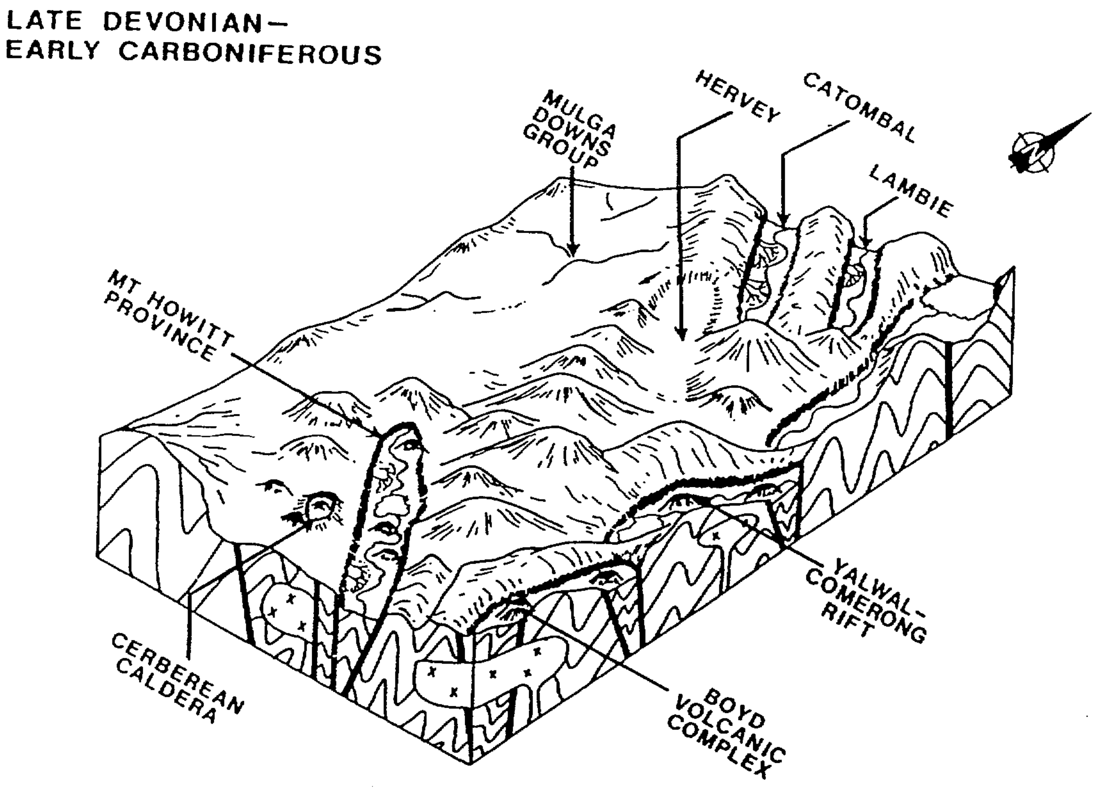

Figure 4: *Paleogeographic reconstruction of the Lachlan Fold Belt of southeastern Australia during the Late Devonian (From Cas 1983)*

Basin is currently wholly submarine, whereas the Gippsland Basin, like the Otway Basin, has both continental and marine components. They are also both still active basins. After the original continental rift forming phase all basins eventually subsided thermally and isostatically and were transgressed by the sea. Until recently, the Gippsland Basin was Australia's major producer of oil and gas. The Otway Basin has only economic gas resources, and the submarine Bass Basin's potential is still being evaluated.

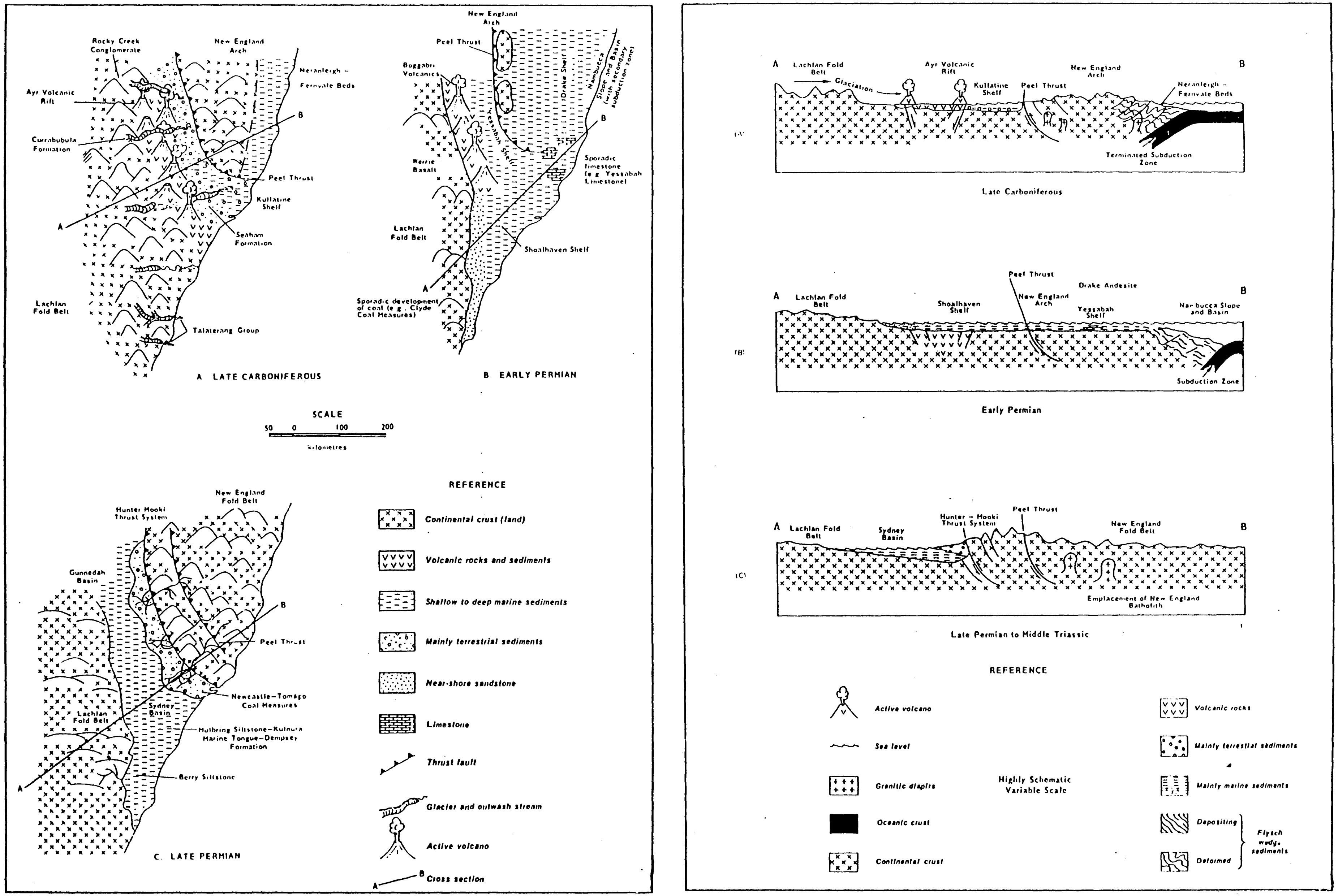

Figure 5: *Paleotectonic reconstruction of the development of the Sydney Basin of southeastern Australia (FromHerbert 1980)*

II - CONCEPTS AND PROBLEMS

Introduction

The significant differences in the products of subaerial versus subaqueous volcanism imply that environment plays a significant role in influencing not only eruptive products, but also the eruptive style. Whereas in subaerial environments lavas, pyroclastic fall, flow and surge deposits are the principal primary volcanic products, in subaqueous environments, lavas, including pillow lavas, hyaloclastites, and limited occurrences of in situ pyroclastics are testimony to the influence that environment plays.

The principal aspects of environment that will affect both eruption style and resultant deposits are

- nature of the substrate through which magmas pass on their way to the earth's surface
- nature of the ambient medium
- nature of the transportation and depositional processes

Effect of substrate on eruptive processes: syn-sedimentary intrusions or lavas?

The nature of the substrate through which magmas pass on their way to the earth's surface may significantly affect the form and level at which magmas come to rest during their ascent. In subaerial environments, nearly all of the substrate through which magmas rise, with the exception of the present soil and sediment profile, will be lithified bedrock. If magma fluid pressure is high enough, this fluid pressure will generate or seek out brittle fractures as conduits and rise to the earth's surface. Thereafter, under the conditions of low atmospheric pressure, the eruption style and resultant products will be the result of the properties of the magma, especially volatile content and viscosity.

In subaqueous environments, the magma will often have to pass upwards through a water saturated sediment pile. The nature of the interaction will depend on the state of the sediment succession, water depth, and the uprise rate of the magma. At shallow water depths explosive interaction may occur due to low confining lithostatic and hydrostatic pressures. At greater depths, where explosive interaction is suppressed, the nature of the interaction will depend on the uprise rate of the magma, the state of consolidation of the sediment and the density contrast between magma and sediment.

If the uprise rate is high, the magma will rise to the seafloor, along nearly planar conduits. If the uprise rate is low, and the enclosing sediment is unconsolidated and has little strength, the magma is likely to spread laterally at or above its density level (McBirney 1963). Due to the lack of strength, the form of the conduit and the intruding magma mass may be highly irregular, especially when dealing with low viscosity magmatic fluids such as basalts. Irregular pillow-like lobes and irregular intrusive apophyses should be common.

Fluidisation of sediments and peperites

The effect on the sediment could be dramatic, as heat is transferred to the interstitial pore water, heating it and initiating fluidisation streaming and convective mixing. The result should include significant loss of structure in the enclosing sediment, leading to homogenisation through fluidisation (Kokelaar 1982). Where quench fragmentation of the magma occurs, hyaloclastite debris may be mixed into the fluidised sediment producing peperite (e.g. Hanson and Schweickert 1982; Hanson and Wilson 1993; Kokelaar 1982; Busby-Spera and White 1987).

In many subaqueous settings significant proportions of coherent igneous masses may thus be expected to be shallow, essentially syndepositional intrusives, including cryptodomes for rhyolitic magmas, and irregular digitate, pod and lobate masses for basalts.

Nature and effects of the ambient medium

Should the magmas reach the earth's surface, the nature of the ambient medium will be a significant influence on eruption style and dynamics. In subaerial environments, with low ambient pressures of 1 bar or 1 atmosphere, lavas will be erupted if the volatile content is low, and pyroclastic eruptions, driven by exsolving magmatic volatiles, will result where magmatic volatile content is high (e.g. Sparks 1978). If the magma rises through groundwater aquifers, phreatic and phreatomagmatic eruptions may occur at shallow levels in the crust (< 500-800m).

In subaqueous environments, the ambient fluid is water, with a density of 1 gm/c.c., and a conductivity and thermal transfer potential that is significantly higher than that of air.

Film boiling, coherent lavas and hyaloclastites

Initial contact between the erupting magma and water will cause superheating of the seawater at the contact, producing a film of steam (e.g. Moore 1975).~If this film of steam remains stable, the magma will be insulated from direct contact with the magma. Coherent lava will result (e.g. pillow lava). If the steam film collapses due to instabilities (e.g. convection induced turbulence), direct contact between water and magma will cause the magma to chill to glass, and then to shatter to glassy debris (hyaloclastite) as contractional fractures propagate through the glass (e.g. Yamagishi 1987, Kokelaar 1986). Therefore at one extreme subaqueous lavas may be almost wholly coherent and at the other they may consist almost wholly of hyaloclastite breccia, thicknesses of tens of metres not being unusual.

Jigsaw fit textures and gradation into coherent lava will be diagnostic of in situ hyaloclastite. If further movement of the lava occurs, for example through upward growth of a dome lava, hyaloclastite may be resedimented by sliding, grain flow, debris flow, producing an apron or enclosing carapace of hyaloclastite with clast rotated textures. Hyaloclast fragments may be angular to irregular in shape, with curviplanar margins, and often a rind with small scale, closely spaced jointing formed due to cooling. Hyaloclastite may be finely granulated down to sand size material, including fractured angular crystal fragments.

Maximum depths for explosive eruptions

Steam however is a compressible fluid, and with increasing water depth the insulating vapour film should become progressively more compressed. At a water depth of 2,120m (=pressure of 212 bars) in fresh water, and 3,150m (=pressure of 315 bars), water vapour or steam is virtually incompressible (McBirney 1963, Kokelaar 1982, Cas and Wright 1987). These limiting pressures are known as the critical point of water. Below these critical water depths, film boiling is unlikely to be able to sufficiently insulate lava surfaces from quenching, and so hyaloclastite should theoretically be pervasive at greater water depths (Clifford, 1992). Although there is thus no theoretical shallow water depth limit to the formation of hyaloclastite, hyaloclastite should theoretically be more abundant with increasing water depth, especially beyond water depths corresponding with the critical point pressure.

At shallow water depths, under conditions of low hydrostatic pressures, superheated steam could expand instantaneously - i.e. explosively. Considerations of specific volume changes of water (steam) at magmatic temperatures by McBirney (1963) suggest that the growth rates of steam bubbles required to produce explosive expansion are unlikely to occur unless hydrostatic pressures are less than 50-100 bars (= 500m - 1000m water depth). This is consistent with the calculations of Colgate and Sigurgeirsson (1973), who suggested that the maximum water depth for steam explosions is about 700m. Clearly, with increasing water depth the intensity of explosions is suppressed by increasing hydrostatic pressure.

If the magma is sufficiently volatile rich, volatiles will begin to exsolve if the vapour pressure exerted by the volatile phase exceeds the ambient pressure. Volatiles will continue to exsolve into vesicles, and vesicles will grow if the vapour pressure of the volatile component in the magma is greater than the fluid/gas pressure in the vesicles (Sparks 1978). The maximum pressures required to cause vesicles to grow, the excess pressure, is about 1 bar for basalts and up to 10 bars for some rhyolites (Sparks 1978). the maximum pressure that will develop in gas bubble in any magma, the equilibrium pressure is defined by the state when the vapour pressure equals the gas pressure in the vesicles (Sparks 1978). This however is not the minimum pressure required to cause explosive fragmentation, defined here as a pressure that produces high enough strain rates in the magma walls of gas bubbles to cause brittle fracturing or failure of the magma. Clearly where gas bubble pressures are less than this minimum pressure for explosive fragmentation, vesicles will grow at a steady (but non-explosive rate) until the magma walls of the vesicles become so thin they run out of matter, and collapse *non-explosively*.

In subaqueous environments, the minimum pressure required to produce explosive fragmentation of a vesiculating magma will be the minimum pressure for subaerial lavas, plus the hydrostatic pressure at the depth of eruption.

From this it is possible to estimate the maximum water depth at which explosive fragmentation of a magma can occur using the relationship

$$P_{b\ equilibrium} = P_{min.\ explos.} + P_{hydrostat.}$$

where
$$P_{hydrostat.} = \frac{H_{max.\ water\ depth}}{10}$$

where $\frac{1}{10}$ is the depth pressure conversion factor (1 bar/ 10m water depth).

This relationship can be written as

$$H_{max} = 10\ (P_{b\ equilib.} - P_{min\ explos.})$$

For rhyolites with 2.5% H_2O and 60% vesiculated Sparks (1978) calculated that the equilibrium pressure is 100 bars. For a rhyolite with 4% H_2O and 60% vesiculated the equilibrium pressure is 200 bars. The value of $P_{min.}$ explosive is now known, but will be greater than the minimum excess pressure for vesicle growth in a rhyolite which could be as high at 10 bars (= 100m water depth). As a result, the maximum water depth for explosive fragmentation of a vesiculating water rich (4%) rhyolite is 2000 metres (200 bars ¥ 10m) minus the ($P_{min.\ explos.}$ ¥ 10m) (= ? 10^{+} bars = 100^{+} m).

Few felsic magmas are more water rich than a 4% bearing rhyolite. Clearly due to the suppressing effects of hydrostatic pressure at 2000m, the explosive intensity would be imperceptible. Significant explosions are probably only likely at depths of <500-1000m. It is of note that all known historical explosive eruptions were initiated in water depths of < 200m (Cas, 1992).

Many pyroclastic eruptions will be driven by both exsolving magmatic volatiles and superheated external water, and are called phreatomagmatic eruptions. In continental settings such eruptions can take place between rising magmas and groundwater aquifers or with a surface body of water such as a lake, including crater lake waters. In marine environments explosive interaction with seawater is common, but is subject to the depth constraints discussed above.

Pumice in deep-water

A consequence of the above is that if the pressure in the vesicles is only greater than the hydrostatic pressure by more than the excess pressure but less than the minimum pressure for explosive fragmentation, then vesicles will grow, and the magma may vesiculate to levels of pumice. If quench fragmentation were to occur, then aggregates of vesiculated pumice and shards of quench fragmentation, not pyroclastic origins, could be formed, even in deep water.

From the preceding, it is clear that pumice found in deep water is unlikely to be of in situ pyroclastic origin. It could be of quench fragmentation origin, or be the result of mass-flow resedimentation of pumice from shallow water or subaerial volcanic centres (e.g. Nishimura *et al.*). Alternatively it could be of water-settled fall origin.

The significance of sedimentary/clastic intervals in volcanic-sedimentary successions

Clearly a major problem in assessing the role of environment in influencing eruptive styles and products is distinguishing the environment in the first place, which will be based largely on the characteristics of interstratified sedimentary/clastic intervals, as well as the primary volcanic facies characteristics.

Clastic intervals in volcanic successions may have the following origins:

(1) they may represent the ambient sedimentary facies, which reflect the normal conditions in the environment of deposition;

(2) they may be primary volcaniclastics, resulting directly from eruption of magma, and including pyroclastics, quench fragmented debris or hyaloclastites, and autobreccias;

(3) they may be the products of the contemporaneous transformation of primary pyroclastic transport phenomena from gas-supported systems to water-supported systems;

(4) they may be the products of post-eruptive reworking and resedimentation of contemporaneous primary volcaniclastics;

(5) they may be the products of the normal weathering and erosion of older volcanic source rocks; or

(6) they may be the packages of sediment that are a response to tectonic activity in the basin in which volcanism occurs.

The influences on sedimentation in volcanic terrains are thus extremely complex, and include volcanism, normal surface processes, tectonics and climate.

Distinguishing hyaloclastite from pyroclastic deposits

The foregoing has highlighted the overlap in characteristics that may develop in hyaloclastite and pyroclastic debris. Both may be non-vesicular, moderately vesicular or highly vesicular. Both may be intensely fragmented or relatively coarse. Explosive processes disperse clasts, as does mass flow resedimentation of both hyaloclastite and pyroclastic debris, leading to clast rotated textures in the final deposits. Clearly the most diagnostic characteristic of hyaloclastite is gradational relationships between coherent lava, in situ jigsaw fit textures and clast rotated textures (Yamagishi 1987, Cas 1992). It is also clear that in subaqueous settings large volumes of hyaloclastite breccias should be produced, and that in the past these may often have mistaken for pyroclastic aggregates.

Distinguishing primary pyroclastics from resedimented pyroclastics

In subaerial settings, the products of the primary pyroclastic transport processes (flow, surge and fall) generally produce distinctive depositional facies and textural characteristics (e.g. Fisher and Schmincke 1984, Cas and Wright 1987). Resedimentation of primary pyroclastic deposits however can commence almost immediately after deposition (e.g. Torres 1993), producing a spectrum of secondary volcaniclastic deposits from those that are texturally indistinguishable from primary deposits (e.g. secondary pyroclastic debris flows; Torres 1993), to those that clearly show the imprint of surface reworking by virtue of improved sorting and rounding of pyroclastic fragments and aggregates compared with primary pyroclastic aggregates, and development of diagnostic tractional sedimentary structures, such as ripples and cross-stratification.

The cross-stratification produced by pyroclastic surges (e.g. Moore 1967, Crowe and Fisher 1973, Schmincke *et al.* 1973, Sheridan and Updike 1975, Sheridan and Wohletz 1981) can usually be distinguished from normal sedimentary cross-stratification by primary surge deposits having: low angle cross-stratification, poor sorting (except for very fine ash surge deposits), cross-strata which are continuous from crest to crest in a dune set, and significant grain size variations between adjacent layers in a cross-bed set. Only high flow regime deposits such as subaqueous storm surge deposits are capable of producing similar features such as hummocky and swaley cross-stratification. In these however, the sorting is better, and the regular asymmetry produced by surges is lacking.

In subaqueous settings, the influence of environment can be significant on the depositional characteristics of pyroclastic deposits. Cas and Wright (1987, 1991) have reviewed the occurrence of pyroclastic flow deposits in subaqueous settings and concluded that there are very few examples of pyroclastic flow deposits (sensu stricto hot gas-supported flows, Cas and Wright 1991), other than at vent or near vent, that appear to have resulted from successful penetration of gas supported flows into and under water. Cas and Wright (1991) point out however that transformation of gas supported flows into water supported flows of pyroclastics debris appears to be common. Such deposits are generally better sorted hydraulically than subaerial pyroclastic flow deposits (e.g. Cas 1979, Cas *et al.* 1992 and in prep).

Pyroclastic surges are unlikely to be able to propagate subaqueously because being low concentration flows, their progress would be kindered by the dense aqueous ambient water mass. Kokelaar (1983) however proposed that in shallow water craters a steam rich cupola may form immediately above the vent, providing a limited low density atmosphere in which vent confined base surges might propagate within the vent. Cas *et al.* (1990) also proposed that subaqueous explosions might produce high energy turbulent aqueous blast flows of water and pyroclastic debris to account for wavy bedded crystal rich horizons in a rhyolitic tuff cone succession in the Bunga Beds of the Boyd Volcanic Complex. These deposits are hydraulically well sorted (crystal rich, matrix poor).

The identification of and the characteristics of water settled pyroclastic fall deposits has received minimal attention, apart from the characteristics of deep-sea distal (co-ignimbrite?) ashes (e.g. Sparks and Huang 1980). More recently Cashman and Fiske (1991) have recognised that water-settled fall deposits will undergo more efficient hydraulic sorting than air settled fall deposits, and have proposed that they could be recognisable by virtue of a 5-10:1 pumice:lithic clast diameter ratio, compared with a 2-3:1 ratio for air fall deposits. Such ratios are only likely to be valid for single explosion fall deposits. In a maintained eruption the finer pumice clasts liberated early in the eruption will be settling with large lithics released later in the eruption, and so the diameter ratios may be much less than proposed by Cashman and Fiske (1991).

It is likely that water settled fall deposits will be planar stratified, especially in deeper water away from vent, will have diffuse boundaries and will show grainsize fluctuations when eruptions proceed as a sequence of explosive bursts. Near vent turbulence generated by subaqueous blasts, or strong shallow water currents may produce scours, undulating bedding, and some cross-stratification.

Fully resedimented and/or reworked pyroclastic debris should exhibit sedimentary structures consistent with the mode of deposition (e.g. graded bedding, Bouma sequences for turbidites, wave ripples, cross-bedding, tidal bed forms etc. for shall water reworked deposits. Reworked deposits should also be more polymictic than primary pyroclastics or contemporaneously redeposited pyroclastic debris. Although there are some general guidelines as outlined above, it is clear that distinguishing primary pyroclastics from resedimented and reworked pyroclastics is not always simple, especially in subaqueous settings.

Determining environments in ancient volcanic sedimentary successions

Given the influence that environment has on eruptive style, products and associated volcanic hosted mineral deposits (e.g. Cas 1992), determining the environment of ancient volcanic successions is important. Two parallel approaches can be used: first using the ambient sedimentary sequence, and secondly using the primary volcanic products.

The use of ambient sedimentary sequences is the same as the approach of sedimentologists - i.e. recognising facies that characterise particular processes, and by association, particular environments. The ambient sedimentary succession may be volcaniclastic, or non-volcaniclastic.

In parallel primary volcanic facies should also assist and palaeoenvironmental recognition, with coherent to autobrecciated lavas, and abundant pyroclastics, including pyroclastic flow, surge and airfall deposits being diagnostic of subaerial environments. Lavas, especially pillow lavas, with associated hyaloclastite are diagnostic of subaqueous environments, as outlined previously.

Determining water depth for ancient successions

One of the most vexing problems in reconstructing ancient subaqueous volcanic successions is determining water depths. The use of sedimentary structures such as wave ripples, planar stratified beach deposits, tidal bedforms and hummocky cross-stratification are useful in constraining approximate water depths for shoreline to shelf water depths (0-200m). For deeper environment (200-10,000m) the available tools are very limited. Turbidites and (hemi-)pelagic sediments reveal nothing about specific water depths of the basin floor except in relative terms. Even the use of fossils is not definitive. An example is the differences in interpretation of water depths for the felsic volcanic host rock succession to the Kuroko massive sulphide deposits of the Miocene green Tuff Belt of Japan, using microfossils. Whereas Guber and Merrill (1983) proposed water depths of 3.4 to 4.1km based on analysis of benthonic and resedimented foraminifera in sedimentary intervals, Hasegawa *et al.* (1989) proposed water depths of 0.7 to 2.8km, and Kitazato (1979) proposed water depths of 0.5-2.0km. These differences are significant and either preclude explosive activity in the case of Guber and Merrill (1983) or allow suppressed explosive activity in the case of Hasegawa *et al.* (1989) and Kitazato (1979).

The use of primary volcanic facies is also only of significance in assessing relative water depths - pillow lavas and hyaloclastite form in any water depth, vesiculated debris can be resedimented into any water depth, and the depth at which magmatically driven pyroclastic eruptions occur, is dependent on magma properties, including composition, viscosity, volatile content, which will vary from magma to magma.

An interesting approach was used by Pisutha-Arnond and Ohmoto (1983), who used fluid inclusion compositions (H_2O-NaCl-CO_2 system) and palaeotemperature assessments to evaluate what the minimum confining pressures must have been during precipation of Kuroko massive sulphide deposits on the sea floor to prevent boiling of the ore forming fluids before reaching the seafloor. Their resultant water depth estimate was 1,750m based on a minimum confining pressure of 175 bars.

Assessing the time significance of primary volcanic successions, associated sedimentary successions, and boundaries

Whereas volcanic successions can often be very thick and associated sedimentary interval relatively thin, this is no indication of the relative durations of the two phases of activity. Clearly in the absence of intervening sedimentary intervals, the time taken for primary volcanic successions to accumulate is very short, especially if evidence exists for preserved erodible units such as non-welded pyroclastic deposits. Deposition of sedimentary successions especially through processes of tractional transport is very time dependent.

Summary of problems

In summary, it is clear that there are significant problems associated with evaluating the influence of environment on eruption style, eruptive products and depositional processes, including

1. Maximum water depths for explosive eruptions

2. Distinguishing pyroclastic and hyaloclastite debris

3. The origin of pumice on the deep-sea floor

4. Interpreting the environments of ancient volcanic successions, especially where ambient sediment sequences are limited in their development

5. Distinguishing primary pyroclastic deposits from resedimented deposits of pyroclastic debris

6. Determining water depths in ancient subaqueous volcanic and sedimentary successions.

7. Assessing the time significance of primary volcanic successions, associated sedimentary successions, and boundaries

III - FIELD GUIDE

DAY 1 : THE PERMIAN "GERRINGONG VOLCANICS": LAVA-SEDIMENT INTERACTION AND VOLCANICLASTIC SEDIMENTATION IN A STORM AND TIDE DOMINATED SHELF SETTING

Directions
The coastal township of Kiama lies approximately 100 km south of Sydney on the Princes Highway (Highway 1).

Introduction
The upper, Late Permian part of the Shoalhaven Group of the Permo-Triassic Sydney Basin is exposed in the region of the coastal township of Kiama, New South Wales, 120 km south of Sydney (Figs 6, 1.1). It consists of two formations, the Berry Siltstone and the Broughton Formation, which together with the overlying coal-bearing Pheasants Nest Formation constitute a conformable regressive sedimentary succession which has a regional dip of 2° to the northwest. Bowman (1970, 1980) suggested that the Berry Siltstone represents marine deltaic sedimentation, the Broughton Formation marine delta-front and shoal-delta sandstone, and the Pheasants Nest Formation fluvial and delta plain deposits. The Broughton Formation and lower part of the Pheasants Nest Formation have been informally termed the Gerringong Volcanic Facies (Bowman 1974; Carr 1982) because they encompass nine shoshonitic basalt to andesite lavas and shallow intrusions (Carr 1985), previously termed latites, as well as three intervening volcaniclastic stratigraphic units of immature volcaniclastic provenance. The four lowermost members of the Gerringong Volcanic Facies, the Westly Park Sandstone Member, the Blow Hole Latite Member, the Kiama Sandstone Member and the Bumbo Latite Member are all well exposed in coastal cliffs in the Kiama area (Fig. 1.1). An examination of this succession indicates that the emplacement of shoshonitic volcanics, in addition to having an obvious local effect on the sediment composition, has also exerted a degree of physical control over sedimentary processes in the host shallow marine shelf environment.

STRATIGRAPHY

Westly Park Sandstone Member (WPM)
The base of the WPM is not exposed in the Kiama area, however, the unit is at least 120 m thick (Fig. 1.2). It consists of light brown to green colored volcaniclastic sandstone (litharenite/felspathic litharenite), pebbly sandstone and minor conglomerate. The major framework grain component is porphyritic or glassy volcanic fragments derived from the intra-basinal volcanic activity associated with the emplacement of the shoshonite members in the Kiama area. As is the case for much of the Permian sedimentary succession of the Sydney Basin, cobble- to boulder-sized "lonestones" occur sporadically within the WPM and the overlying Kiama Sandstone Member (STOPS 1,2 and 4). In the Kiama region these consist almost exclusively of intermediate volcanic lithologies and they have been interpreted by previous workers (eg. Raam 1968; Bowman 1974) as having been rafted and then dropped by floating ice (ie. "dropstones").

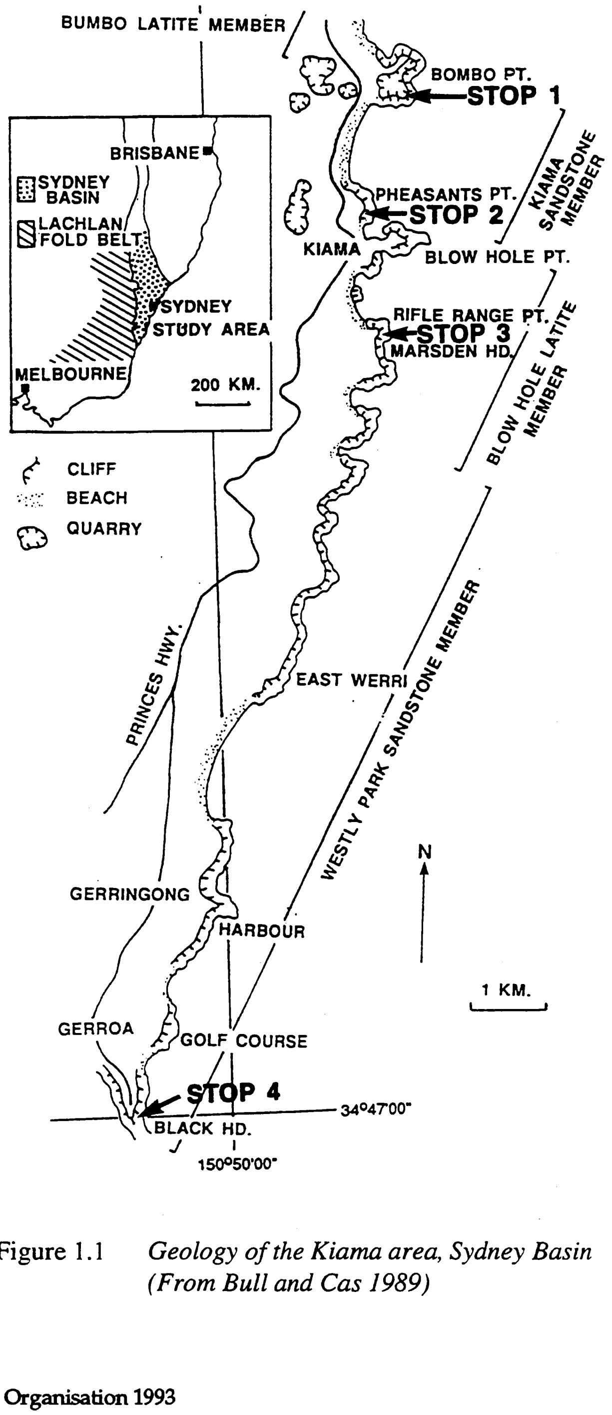

Figure 1.1 *Geology of the Kiama area, Sydney Basin (From Bull and Cas 1989)*

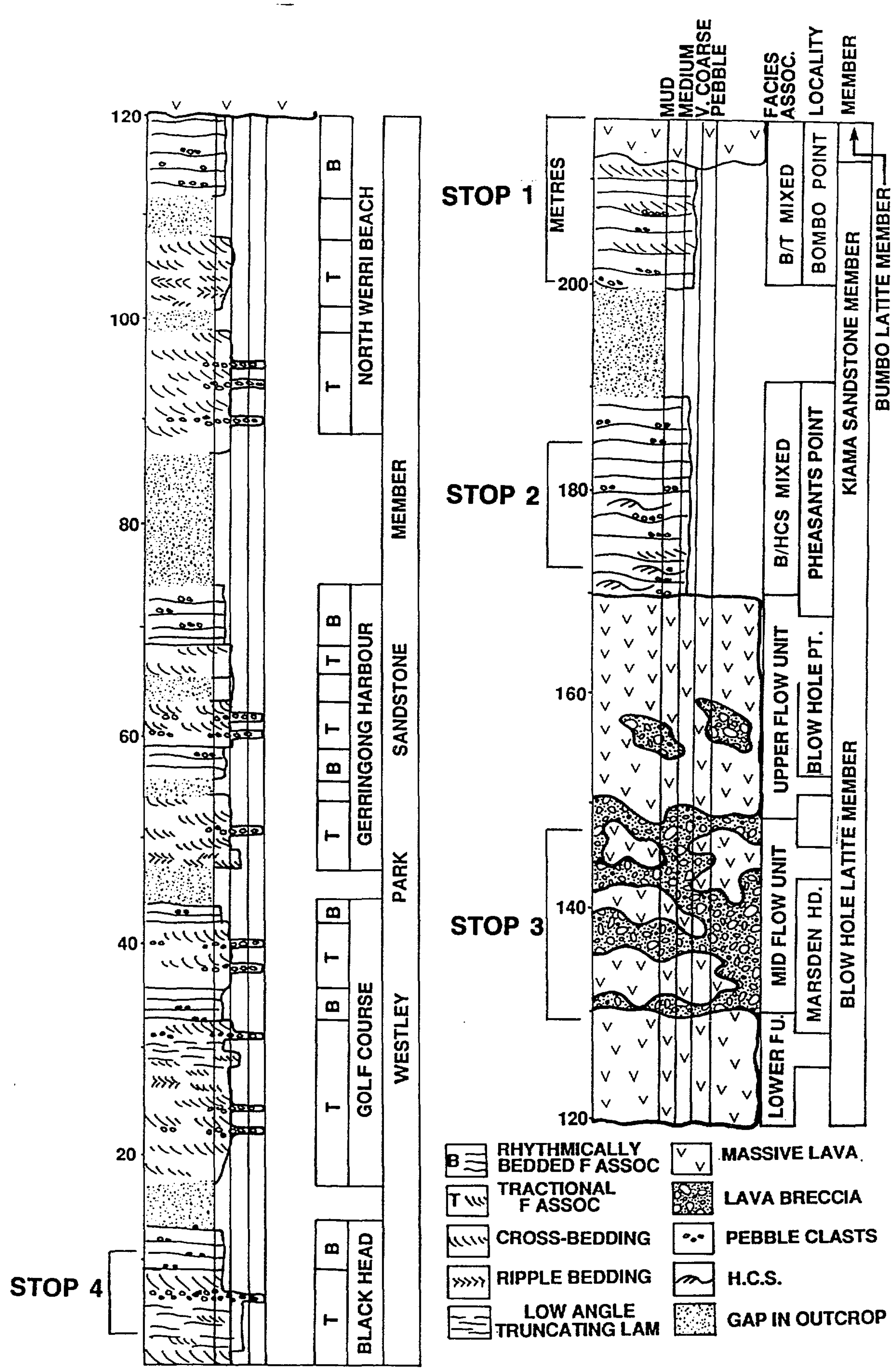

Figure 1.2 *Permian stratigraphy of the Kiama area (From Bull and Cas 1989)*

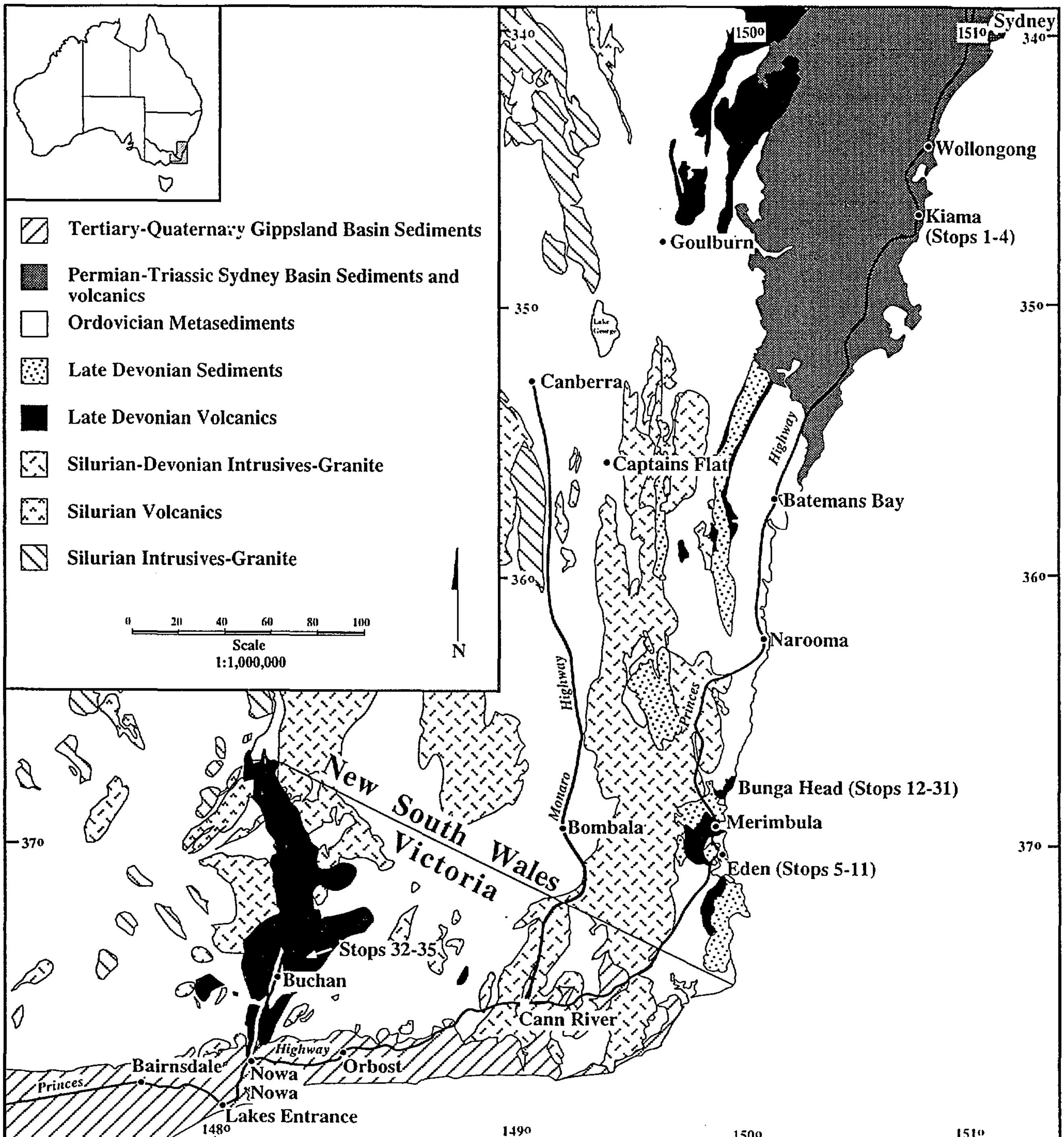

Figure 6 *Route map and generalised geological map of southeastern Australia*

The depositional structure of the WPM will be examined at locality 4. Overall, the unit can be considered in terms of a background facies association of bedded deposits which have been locally reworked by tractional currents (Fig. 1.2). The bedded facies association consists of 50 to 500 mm thick, massive to normally graded sandstone beds which may have basal conglomerate/pebbly sandstone and are separated by mudstone intervals. Individual beds have sharp bases, may exhibit basal sole structures, commonly load casts and less commonly scour marks. These units have the general characteristics of turbidity current deposits, however, they lack the plane laminated and ripple cross-laminated units of the classical Bouma sequence. They are, therefore, interpreted to have been deposited from poorly-expanded, high-density turbidity currents. High degrees of particle interaction within such systems inhibits individual particle freedom, leading to frictional freezing and excessively rapid sedimentation rates, rather than the progressive fallout from suspension deposition necessary for development of tractional sedimentary structures of the Bouma sequence (Lowe 1982).

The bedded depositional structure of the bedded facies association is partially to completely obscured at many localities by bioturbation which consists of diverse, horizontal branched and backfilled burrows (locality ?). This structure is typical of the *Cruziana* facies (the zone of generalized sediment feeders; Seilacher, 1978) which represents regions of relatively low hydraulic energy at the bed surface, where deposit and sediment feeding organisms are prevalent (eg. below wave base). Body fossils are also locally abundant within the WPM. They generally occur as dispersed fragmental, or at least disarticulated shells, however, two species, the bivalve *Myonia elongata* and an unidentified species of the brachiopod *Sulciplica* occur in articulated assemblages which appear to be in growth position. Overall, the fossil assemblage belongs to the Late Permian Ovalis Zone (after *Echinalosia ovalis*; Runnegar 1980b), and it has been interpreted to be indicative of a quiet open marine environment (eg. Raam 1968; Bowman 1974). The bedded facies association of the WPM is, therefore, interpreted to have been deposited in the continental shelf environment below wave base by periodic poorly expanded. These deposits were subsequently bioturbated by benthic deposit feeding organisms.

The tractional facies association of the WPM consists of scattered tractional sedimentary bedforms and structures dispersed within the bedded facies association. The tractional structures range in scale from large scale cross-bed sets (amplitude >1.5 m, wavelength >20 m) to ripples, and occur in intervals ranging in thickness from >1 m to about 8 m. Individual cross-bed sets may have basal pebble lenses and, in rare exposures, medium scale cross-bed sets have their foresets defined by pebble trains. Mudstone is rare and is only present as drapes on individual foresets (STOP 4). Palaeocurrents in the tractional facies association are generally to the north, although with a decrease in the amplitude of the structures present a weak southerly component becomes evident. Biogenic structures are restricted to scattered sub-vertical burrows. These are interpreted to belong to the *Skolithos* and *Glossifungites* facies (the zone of suspension feeders; Seilacher, 1978) which represents regions of relatively high hydraulic energy at the bed surface where most detrital food is kept in suspension. Body fossils in the tractional facies association occur as fragments which are concentrated into lags at the base of individual cross-bed sets. Overall, the tractional facies association of the WPM is interpreted to represent longshore directed, tidally influenced current activity which sporadically reworked the shelf density current deposits of the bedded facies association.

Blow Hole Latite Member (BHL)

The Blow Hole Latite Member is a 50 m-thick shoshonitic basaltic andesite which overlies the Westly Park Sandstone Member (Fig. 1.2). It was initially interpreted as a three-component intrusive (Raam, 1964). However, Bull and Cas (1989) considered that only the middle flow unit can be demonstrated to be at least partially intrusive because of local fluidal, delicately sutured intrusive upper lava/sediment contacts. As far as can be ascertained, the lower and upper flow units of the BHL have planar upper contacts consistent with them representing extrusive lava flows (Fig. 1.2). The partially intrusive middle flow unit of the BHL will be examined at STOP 3.

The middle unit of the BHL also differs from the other lava units in the Kiama area, which are composed dominantly of massive/columnar jointed lava, in that it consists of digitate shoshonite bodies from tens of centimetres to tens of metres in diameter, and large volumes of volcanic breccia typified by a structureless, homogenous tuffaceous sediment matrix (cf. Kokelaar 1982). These features will be examined at STOP 3. In detail, the digitate lava masses may be internally massive or composed of radially oriented columnar joints, and are enclosed in a tuffaceous sediment. At the contact with the enclosing sediment, the margins of the digitate masses may be planar or jagged, in which case there is commonly an intimately associated volcanic breccia zone, or alternatively the margins may be delicately sutured and fluidal. The associated breccia zones consist of angular, blocky shoshonite fragments which may have closed frameworks and jigsaw fit (or nearly so) relationships suggesting *in situ* fragmentation, particularly where adjacent to coherent lava bodies. More commonly, the breccias have an open framework with clasts supported by a matrix which, even well away from the associated lava bodies, consists of structureless tuffaceous sediment.

The middle flow unit of the Blow Hole Latite Member is interpreted as a complex lava flow which was emplaced over an unconsolidated sedimentary substrate. Local fluidization and resultant mobilisation of sediment below the lava, as well as unequal pressure exerted by the propagating digitate lava masses, could have allowed some of these digitate lava bodies to "burrow" into the underlying sediment pile forming intrusive, pillow-like lava pods. Quench fragmentation of the margins of the lava bodies, associated with fluidization-mixing of the resultant hyaloclastite and sediment would have produced the associated lava-sediment breccias, and accounts for the jigsaw-fit to open framework breccias.

Kiama Sandstone Member (KSM)

The KSM overlies the Blow Hole Latite Member and is approximately 40 m thick (Fig. 1.2). As is the case for the Westly Park Sandstone Member, the KSM comprises a background bedded facies association interpreted to represent sediment density currents in a shelf setting. In detail, however, the KSM differs markedly from the Westly Park Sandstone Member in that: (1) The mudstone facies is haematitic indicating oxidising conditions. (2) The tractional facies association is largely replaced by the distinctive hummocky cross-stratified sandstone facies which is interpreted to represent reworking of the substrate by storm waves and currents (eg. Mount, 1982; Swift *et al.* 1983; Walker, 1984). (3) Body fossils are absent, although rare megadesmid pelecypods have been recorded (Runnegar 1980b), and bioturbation is mainly restricted to mudstone intervals and never totally obliterates the original depositional fabric. Overall, these features (which will be examined at STOPS 1 and 2) indicate that, as was the case for the Westly Park Sandstone Member, the KSM was deposited in a marine shelf environment. In this case, however, the oxidised nature of the sediment and the presence of the hummocky cross-stratified sandstone facies indicates that deposition occurred in markedly shallower water above the storm wave base.

Bumbo Latite Member (BL)

The Bumbo Latite Member is a 150 m-thick massive, columnar jointed shoshonitic basalt which occurs above the Kiama Sandstone Member. The contact between these two units will be examined at locality 1 where there is no evidence of the complex lava/sediment interactions which characterise the middle flow unit of the Blow Hole Latite. The BL has been interpreted as a tri-composite extrusion (Raam, 1964; Bowman, 1974). A hyaloclastite breccia facies with a matrix of fluidized sediment is present locally, however, it is restricted to scattered, sub-vertical pipe-like bodies which are volumetrically insignificant compared with the columnar jointed and massive lava facies. These structures are interpreted as reflecting the escape of overpressured, water-saturated fluidized sediment, through steam explosions. This material could have moved upwards as a diapir from the base into the interior of the liquid lava flows, causing local quench brecciation of the lava in contact with the water-rich fluidized sediment, and subsequent mixing of resultant lava clasts with sediment, as well as local steam explosions.

DISCUSSION

Ambient depositional conditions on the host marine shelf during the accumulation of both the Westly Park and Kiama Sandstone Members are represented by the bedded facies association. High-density, poorly-expanded turbidity currents have, therefore, been the major source of sediment throughout. The identification of the hummocky cross-stratified sandstone facies widely considered to be diagnostic of storm associated activity within the Kiama Sandstone Member, provides a mechanism for the generation of these turbidity current deposits. Storm associated activity has been implicated in the generation of turbidity currents in the pericontinental environment where storm surge and/or storm associated current activity entrain sand in the nearshore zone and transport it offshore (Hayes, 1967; Hamblin & Walker 1979; Brenchley & Newall 1982). This process may explain the poorly-expanded, high-density nature of the turbidites which could have been the result of the proximity and relatively low gravitational potential of the sediment source area, in this case the shoreline, implying a short transport distance and low substrate slope so that fully expanded flows did not develop. Alternatively, if fully expanded flows had been able to develop, flow turbulence may have imparted sufficient mobility for them to travel further offshore, leaving poorly expanded flows as a residual deposit.

That sediment accumulated via poorly-expanded, high-density turbidity currents throughout the Westly Park and Kiama Sandstone Members indicates that they represent closely related depositional systems, however, the other depositional processes which were operative differ markedly. During deposition of the Westly Park Sandstone Member, areas of the substrate were reworked by relatively deep, probably partly tidal, northerly directed longshore currents but there is no evidence of storm related reworking. Conversely, there is no evidence of significant tidally associated reworking during deposition of the lower part of the Kiama Sandstone Member sequence, but the seafloor was clearly subject to storm related reworking. This rapid change in depositional conditions between the deposition of the Westly Park and Kiama Sandstone Members is attributed to the emplacement of the Blow Hole Latite Member. In this model, the emplacement of the three lava flow units which comprise the Blow Hole Latite Member could have rapidly raised the depositional surface by as much as 50m, and this elevation was apparently sufficient to put the Kiama Sandstone Member above the storm wave base. In addition, the emplacement of the lava body, presumably roughly perpendicular to the shoreline, appears to have shielded the depositional surface from the effects of northward directed tidally influenced currents evident during the deposition of the Westly Park Sandstone Member. This allowed development and preservation of HCS. in that region of the Kiama Sandstone Member closest to the lava flow. Tidally influenced currents were still operative, because tractional sedimentary structures are present higher in the Kiama Sandstone Member at Bombo quarry (locality ?), indicating that by this stage the sheltering effects of the Blow Hole Latite had been eliminated, as continued sedimentation led to its burial.

STOP 1 - BOMBO POINT: UPPER KIAMA SANDSTONE MEMBER OVERLAIN BY BUMBO LATITE MEMBER

Directions: The turnoff to Bombo Beach goes beneath a railway bridge to the left off Highway 1 approximately 1.2 km north of the Kiama town centre and at the foot of the hill on which the hamlet of Bombo sits. Once under the bridge turn left (north) along the track inland from Bombo Beach and proceed to the carpark in front of the prominent quarry in Bombo Point.

The contact between the Kiama Sandstone Member (KSM) and the Bumbo Latite (BL) is exposed in the cliff which forms Bombo Point. The contact is planar overall, but has local irregularities and breccia zones up to 1 m thick. It appears to represent the relatively passive effusion of lava over the sedimentary substrate. The tabular, massive/graded bedded structure of the KSM is clearly visible at this locality as is the red coloration due to abundant haematite, particularly in the mudstone intervals. Some evidence of northerly directed current activity at this level is also present in the form of scattered medium-scale planar and trough cross-bed sets. Pebble/cobble/boulder dropstones related to floating ice are also present in the KSM. In the quarry at the back of Bombo Point the massive/columnar jointed nature of the internal part of the BL is exposed. Poorly-exposed pipe-like bodies of quench fragmented lava breccia with a sediment matrix are present in the quarry floor and walls (Fig. 1.3). These are the structures interpreted to be the result of steam diapirs rising through the lava flow during or shortly after emplacement.

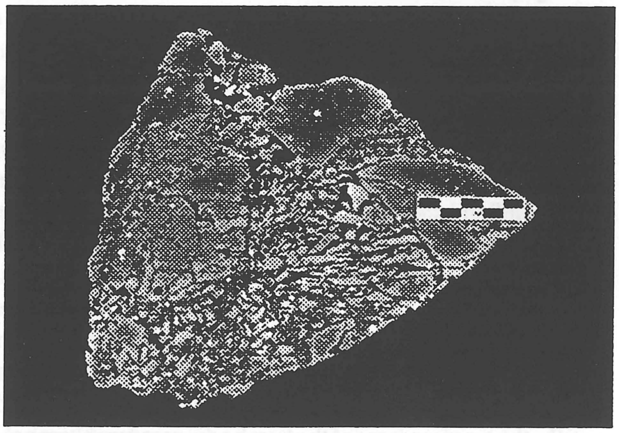

Figure 1.3 *Quench fragmented latite from a pipe like body, Bumbo Quarry*

STOP 2 - PHEASANTS POINT: LOWER KIAMA SANDSTONE MEMBER

Directions: Return along the Bombo Beach track to Highway 1 and turn left (south) towards the Kiama town centre. Half way down the main street of Kiama town centre there is a turn to the left marked Pheasants Point. Take this road which goes beneath a railway bridge to a car park at a tidal swimming pool on the rock platform at Pheasants Point.

The KSM exposed in Pheasants Point has a similar depositional structure to STOP 1 see summary sedimentary section: Fig; 2.1, however, medium-scale cross-bed sets have been replaced by the distinctive hummocky cross-stratified sandstone (HCS) facies. The HCS facies comprises ovate to circular domes and intervening swales defined by low angle truncating laminae. Individual hummocks and swales are variable in scale, with bed thicknesses of up to 400 mm and wavelengths of up to 5 m. These structures are interpreted (eg. Dott & Bourgeois 1982; Swift *et al.*, 1983; Walker 1984) as reflecting storm wave and/or storm surge re-working, usually below the fair weather wave base. They are best developed in fine to medium grained sandstone. Other features to note at this locality are the bioturbated nature of the mudstone intervals and the vertically oriented "escape" burrows through the sandstone beds. Dropstones at this locality reach boulder size.

STOP 3 - MARSDEN HEAD: MIDDLE BLOW HOLE LATITE MEMBER AND RIFLE RANGE TUFF MEMBER

Directions: Return to main street of Kiama town centre (Highway 1) and turn left (south). Approximately 1 km south of the town centre is a turn to the left signposted Kendalls Beach Caravan Park. Take this road down to the beach and park at the southern end of the beach past the caravan park.

A section through the middle flow unit of Blow Hole Latite Member (BHL) is exposed in Marsden Head. All of the features described above associated with the intrusion of pods of lava into unconsolidated sediment can be observed, including both quench brecciated and fluidal lava/sediment contacts at lava pod margins (Fig. 3.1). Where bedding in the intercalated sediment is visible it is generally disturbed, but in most exposures, particularly where the sediment is interstitial to brecciated lava clasts, the sediment pile is completely homogenized and structureless. This is interpreted to be due to extensive fluidization and mixing (Kokelaar, 1982) caused by heating of interstitial sediment pore water by the intruding shoshonite pods, leading to convective circulation (and perhaps even boiling). The discordant intrusive contact between the upper margin of the middle flow unit of the BHL and a relatively small sediment interval termed the Rifle Range Tuff Member (Raam, 1964) is exposed in the upper part of the northeastern face of the bluff which forms Marsden Head.

STOP 4 - BLACK HEAD: WESTLY PARK SANDSTONE MEMBER - VOLCANI-CLASTIC, TIDALLY INFLUENCED SHELF SEDIMENTATION

Directions: Return to Highway 1 and turn left (south). Turn left off Highway 1 approximately 7 km south of Kiama onto the road to Gerringong and Gerroa. Proceed through the township of Gerringong to Gerroa where there is a left hand turn marked Black Head. Take this road to the carpark on Black Head.

The bluff which comprises Black Head exposes typical green/light brown colored Westly Park Sandstone Member (WPM). A prominent, large-scale trough cross-bed set indicating a northerly palaeoflow direction is present at the level of the rock platform. It has a basal pebble lag

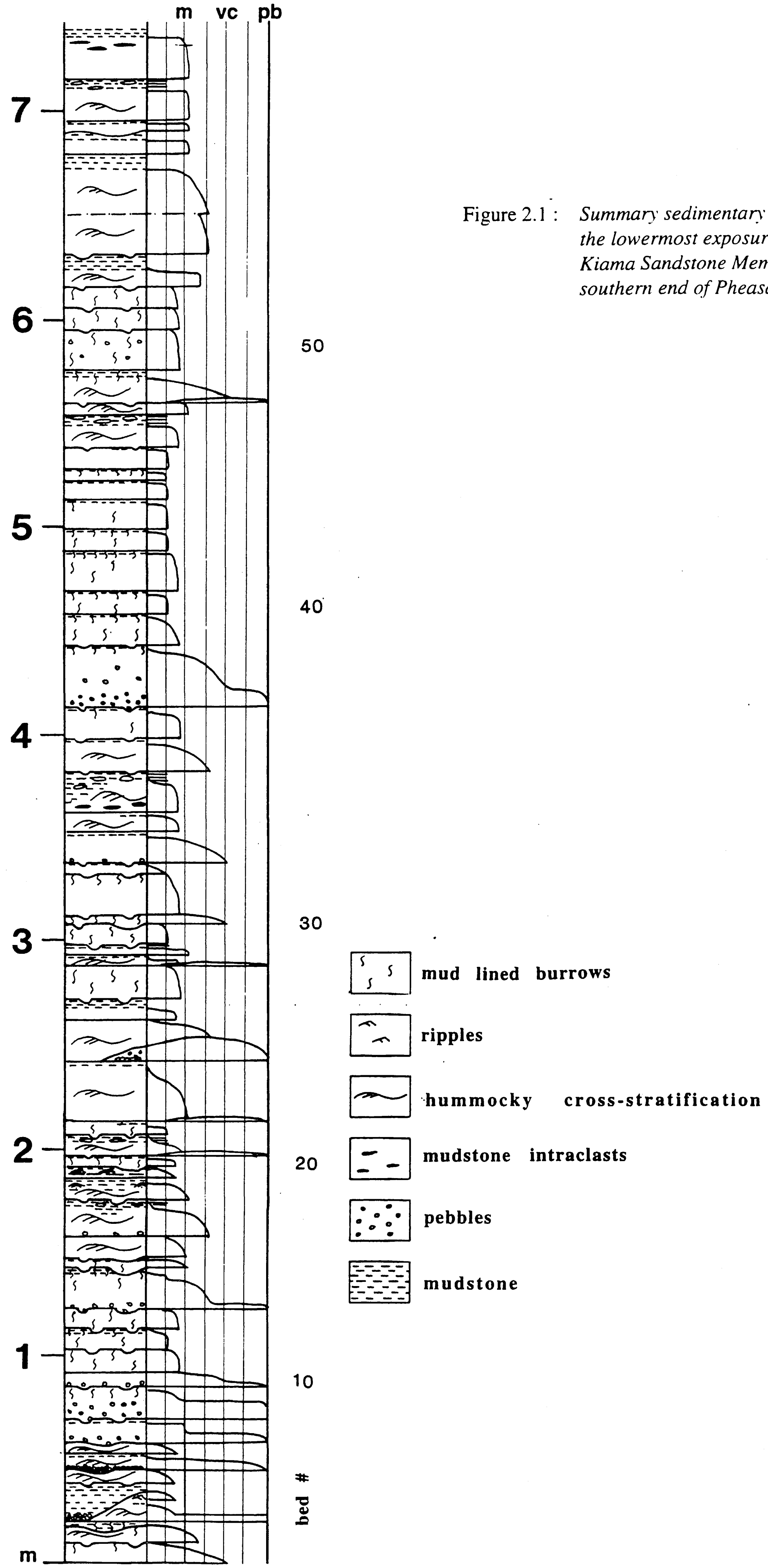

Figure 2.1 : *Summary sedimentary section of the lowermost exposure of the Kiama Sandstone Member at the southern end of Pheasants Point*

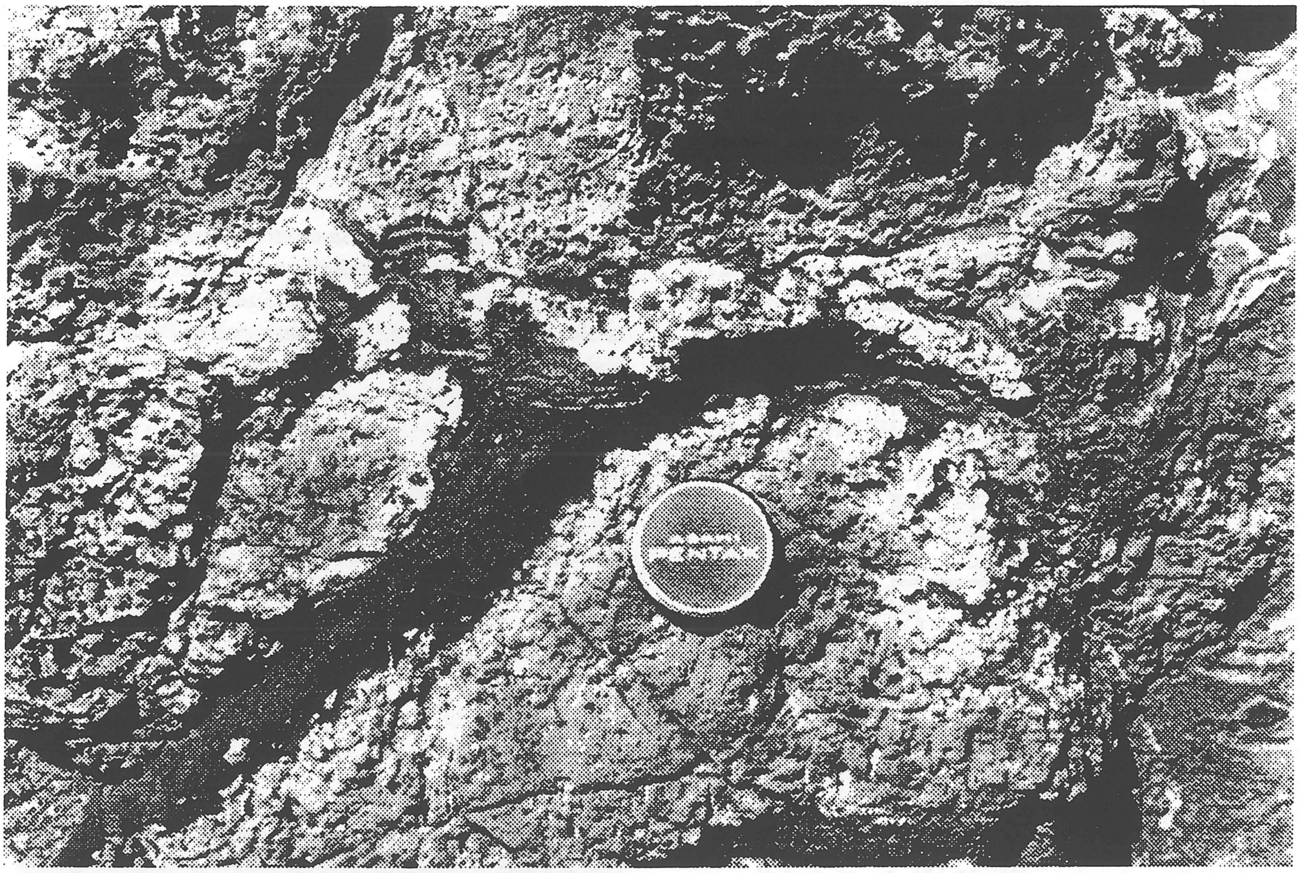

Figure 3.1 *Highly irregular contact between pillow-like intrusive latite lobe and partly fluidised volcaniclastic sand stone.*

deposit and mudstone draped foresets typical of tidally deposited tractional structures. It should be noted that not many of the tractional sedimentary structures which occur throughout the WPM and parts of the Kiama Sandstone Member are this large-scale. Medium-scale trough and planar cross-bed sets such as are exposed half way up the cliff section a short distance to the north of the large-scale structure are actually the most abundant tractional sedimentary structure present. The cliffs in this region also show the typical depositional structure of the WPM in places, which consists of tabular, graded/massive bedded deposits locally obliterated by horizontally oriented bioturbation.

DAY 2: THE LATE DEVONIAN BOYD VOLCANIC COMPLEX : THE INFLUENCE OF ENVIRONMENT ON CONTRASTING STYLES OF VOLCANISM AND SEDIMENTATION

Introduction

The Late Devonian Boyd Volcanic Complex (Fergusson et al. 1979) is a north-south trending belt of bimodal rhyolitic and basaltic volcanics and associated sedimentary rocks which occur along the south coast of New South Wales (Fig. 5.1). In the main outcrop belt around Eden and Pambula, the Boyd Volcanic Complex consists of coherent rhyolite lavas, intrusives, and ignimbrites, basaltic lavas and intrusives, and intercalated red bed fluvial and lacustrine sedimentary rocks (Fergusson et al., 1979). The Boyd Volcanic Complex is in unconformable and faulted contact with Ordovician metasedimentary turbidite successions, and Early Devonian granites of the Bega Batholith. The rhyolite lavas and ignimbrites of the rift zone are relatively alkaline potassic rhyolites (K. Dadd, pers. comm.), whereas the mafic volcanics comprise quartz normative tholeiites and olivine normative tholeiites (Dadd, 1984). The Boyd Volcanic Complex passes upwards into a continental red bed succession with minor marine intercalations, known as the Merimbula Group (Steiner 1972, 1975, Fergusson et al. 1979; Fig. 1).

This bimodal volcanic-sedimentary rock succession is the

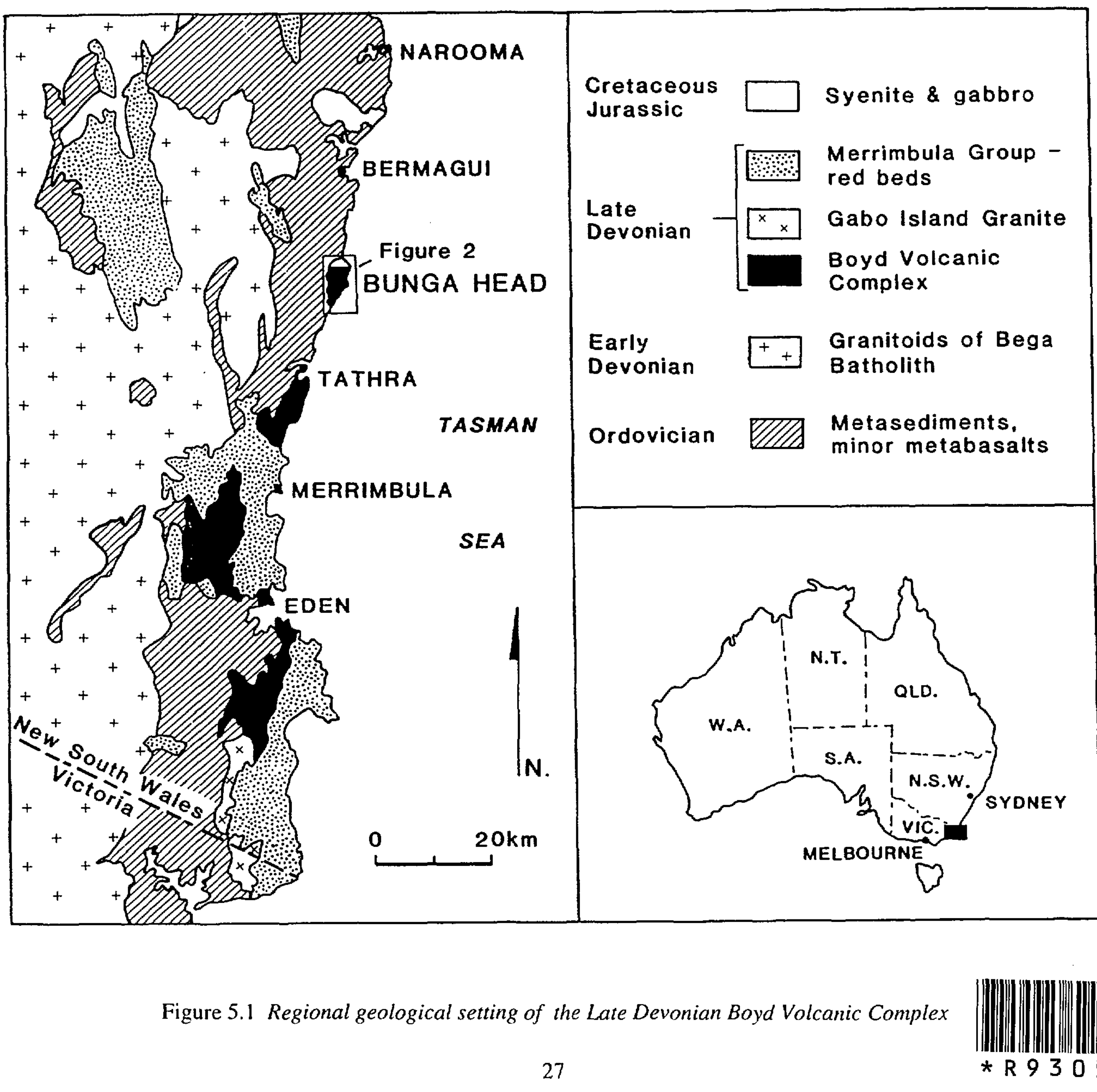

Figure 5.1 *Regional geological setting of the Late Devonian Boyd Volcanic Complex*

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southern most part of the regionally extensive, north-south trending Yalwal-Comerong-Eden Rift Zone (McIlveen 1974), which was part of the more widespread Basin and Range-like continental rift terrain of the Lachlan Fold Belt of southeastern Australia in the Late Devonian (Cas 1983). This extensional rifting and bimodal volcanism immediately succeeded the mid-Devonian Tabberabberan Orogeny, which virtually terminated all marine sedimentation in the Lachlan Fold Belt, excepting a shortlived Frasnian incursion. The Late Devonian rock successions of the Lachlan Fold Belt were deformed by the Early Carboniferous Kanimblan Orogeny whic was the termonal event in the history of the Lachlan Fold Belt.

The nature of the Boyd Volcanics varies enormously on a regional scale (Fig. 5.2). South of Eden in the Wonboyn area, the succession appears to be dominated by a very thick rhyolitic ignimbrite succession which is intruded by the co-magmatic and contemporaneous Gabo Island Granite (Fergusson et al. 1979). This region is considered to represent the remains of a caldera, which may have been the source for ignimbrites in the Eden area.

Around Eden a succession of rhyolitic ignimbrites, lavas, and intrusives and lesser basalts and sedimentary rocks (Fig. 5.2) appear to represent outflow volcanism and sedimentation in a continental graben rift setting. West of Pambula, basalts dominate the succession (Fig. 5.2), representing either a graben/half graben ponded basalt pile or the remains of a basaltic shield volcano.

The northernmost part of the Boyd Volcanic Complex is an outlier known as the Bunga Beds (Scott 1972, Fergusson et al. 1979; Fig. 5.1). The Bunga Beds are surrounded by a basement of deformed Ordovician metasedimentary quartz-rich turbidites and pelites with quartz veins, and Lower Devonian granitic plutons (Fig. 5.1). The only known contact with the Ordovician basement along the southern margin at Picnic Point is faulted, but appears to have been an unconformity (see below). Outcrop of the Bunga Beds is poor except along the coast. The preserved north-south extent of the Bunga Beds is 8 km, but the size of the original basin is unknown.

The Bunga Beds also consist of a bimodal association of volcanics and associated sedimentary rocks but differs from the rest of the Boyd Volcanic Complex in lacking continental red beds and ignimbrites. Rhyolites predominate over basalts, and both are represented by coherent bodies and a diversity of volcaniclastics. The sedimentary rocks contain fossil fish and plant fragments of late Givetian-early Frasnian age (latest Middle-earliest Late Devonian), representing the oldest part of the Boyd Volcanic Complex (Fergusson et al. 1979). The fish fossils indicate a brackish to marine environment (J. Long, pers. comm.). The sedimentary succession varies from shallow basin margin clastics to deep water mass flow hemi-pelagic deposits. Pyroclastics, although present, are localised and represent a minor part of the preserved succession. The stark contrast between the Eden area sucession and the Bunga Beds suggests that environment played a significant role in influencing the nature of volcanism and the products.

THE CONTINENTAL BOYD VOLCANIC COMPLEX, EDEN: SUBAERIAL VOLCANISM AND SEDIMENTATION IN A RIFT GRABEN BASIN

The Boyd Volcanic Complex around Eden consists of extensive rhyolitic ignimbrites, rhyolite lavas, minor basaltic products and sedimentary rocks (Figs 5.3, 5.4). The aim of today's field program is examine the subaerial facies of the Boyd Volcanic Complex (BVC) to assess the relationships between volcanism, tectonics and sedimentation in this Late Devonian continental rift basin setting. Principal aims will be to assess the nature of the depositional environments, tha factors that influenced changes in environment, including volcanism and tectonics, the duration and frequency of volcanism, the eruptive style and eruptive products.

STOP 5: THE BASAL FACIES OF THE BVC, QUARANTINE BAY, SOUTH OF EDEN.

From Merimbula drive south through Eden along the Princes Highway. 5km south of Eden take the left hand turn into Quarantine Bay. Park on the left hand side, and take track to the north along the coastal rock platform to the contact between the deformed Ordovician and the basal facies of the Boyd Volcanic Complex, an interbedded sandstone and conglomerate succession (Fig. 5.4). This limited outcrop is an outlier of the BVC. The contact between the BVC and Ordovician is a fault with reverse movement. The conglomerates consist of quartzite clasts, many with quartz veins, clearly derived from the adjacent Ordovician metasedimentary succession. Similarly the sandstones are quartz rich and derived from the Ordovician.

The conglomerates vary from closed to open framework pebbly sandstones. They are massive, generally lensoidal, and frequently scour into the associated sandstones. The sandstones are massive to planar laminated, with rare cross stratification. These characteristics are consistent with a high flow regime fluvial-alluvial depositional origin. The preferred depositional environment is a basin margin alluvial fan (Fergusson et al. 1979). The reverse fault may be a reactivated original basin margin normal fault scarp against which the alluvial fan succession developed. Similar lensoidal occurrences of basement derived conglomerates and sandstones occur to the north along the western margin of the BVC at its contact with the Ordovician. It is significant that this basal succession lacks volcanic detritus, indicating that crustal extension, and normal faulting and formation of relief preceded the onset of volcanism at least in this part of the basin.

STOP 6 : THE BASAL FACIES OF THE BOYD VOLCANIC COMPLEX, ROTARY LOOKOUT, EDEN

From Quarantine Bay return to Eden, turn right at the roundabout, drive through the shopping centre, down the hill past the wharves, and up the hill alopng Imlay St. to the

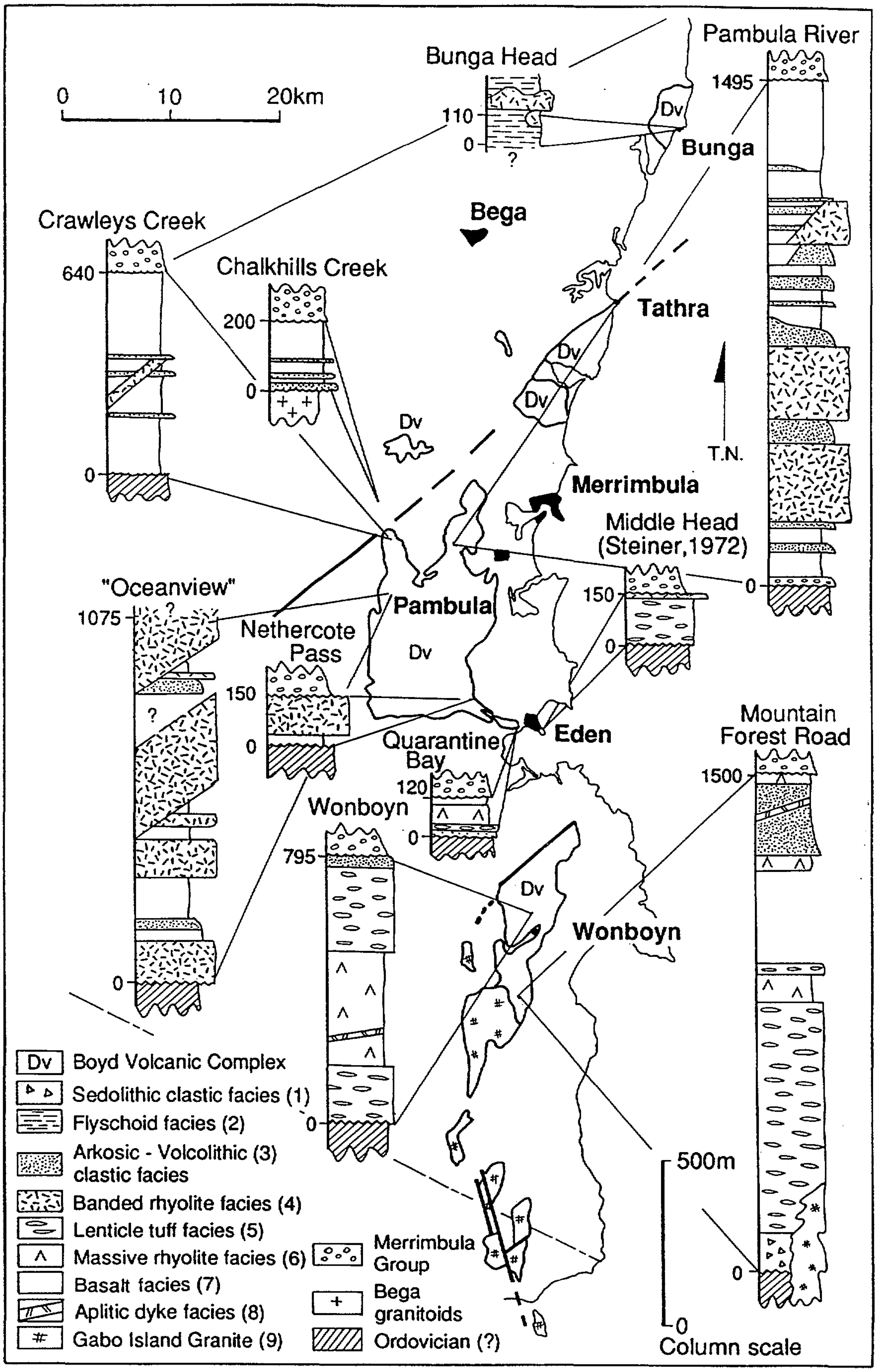


Figure 5.2 *Regional facies distribution pattern for the Boyd Volcanic Complex (after Fergusson et al. 1979)*

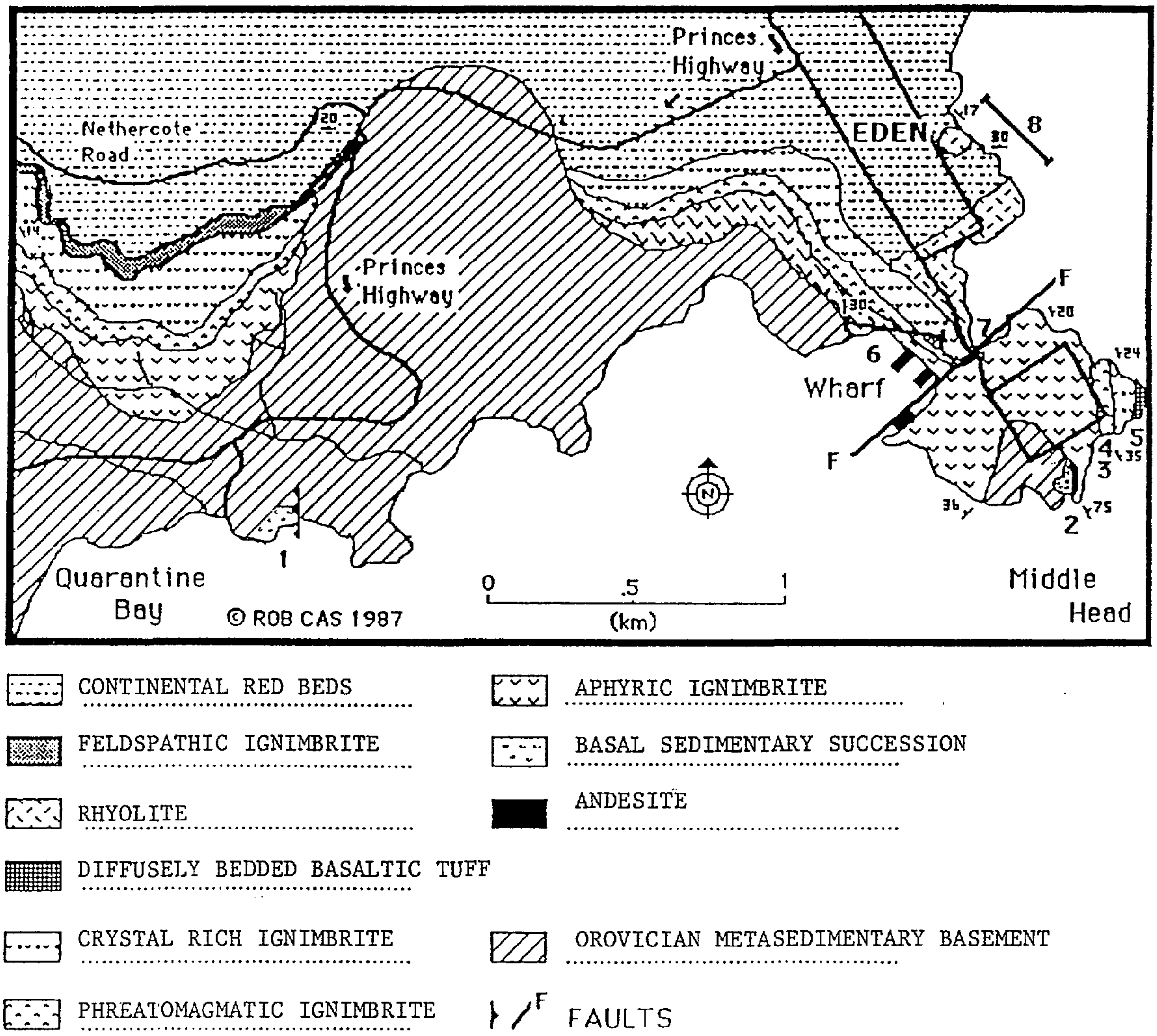


Figure 5.3 *Geological map of the Boyd Volcanic Complex, Eden area.*

Rotary Park lookout. Take the track to the lookout and then veer left through the bush along a narrow steep track down into the bay below. At the lookout note the thick columnar jointed unit making up the cliff to the north.

Locality 1: Down in the bay look first at the headland and platform exposure to the right. The cliff consists of polydeformed Ordovician metasediments that are at phyllite grade. On the platform is the basal unconformity between the BVC and the Ordovician. The unconformity is irregular, with conglomeratic and grit lenses filling in hollows locally. The basal succession consists of tabular massive beds of quartz sandstones interbedded with thin siltstones. Higher in exposed blocks in the bay, the succession consists of tabular thin beds of cryptically graded beds of sitlstone interbeded with grey-black mudstone. This succession is interpreted as lacustrine turbidite and hemipelagic succession deposited in a lake ponded by either faulting (?half graben fault basin) or a high profile volcanic unit that is not exposed here.

Lying at the base of the BVC this facies occurs at the same stratigraphic level as the Quarantine Bay succession, as depiscted schematically in the stratigraphic succession for the day (Fig. 5.4).

Locality 2: Now examine the massive fracture jointed brown-grey rock unit making the cliff exposure behind the bay. The unit is massive and has a weathered sandy, granular texture. Tracing this unit around to the north isolated occurrences of pseudomorphed tabular crystals can be found. This rock unit makes up the base of the cliff to the east.

Carefully clamber up the steep track through the bush and then down again to the boulder beach in the bay. PLEASE TAKE EXTREME CARE NOT TO DISLODGE BOULDERS. Outcrop of the brown grey unit agin reveals pseudomorphed phenocryts. The unit is interpreted as an andesite lava. At its western end where it occurs along strike from the bedded lacustrine succession, the contact

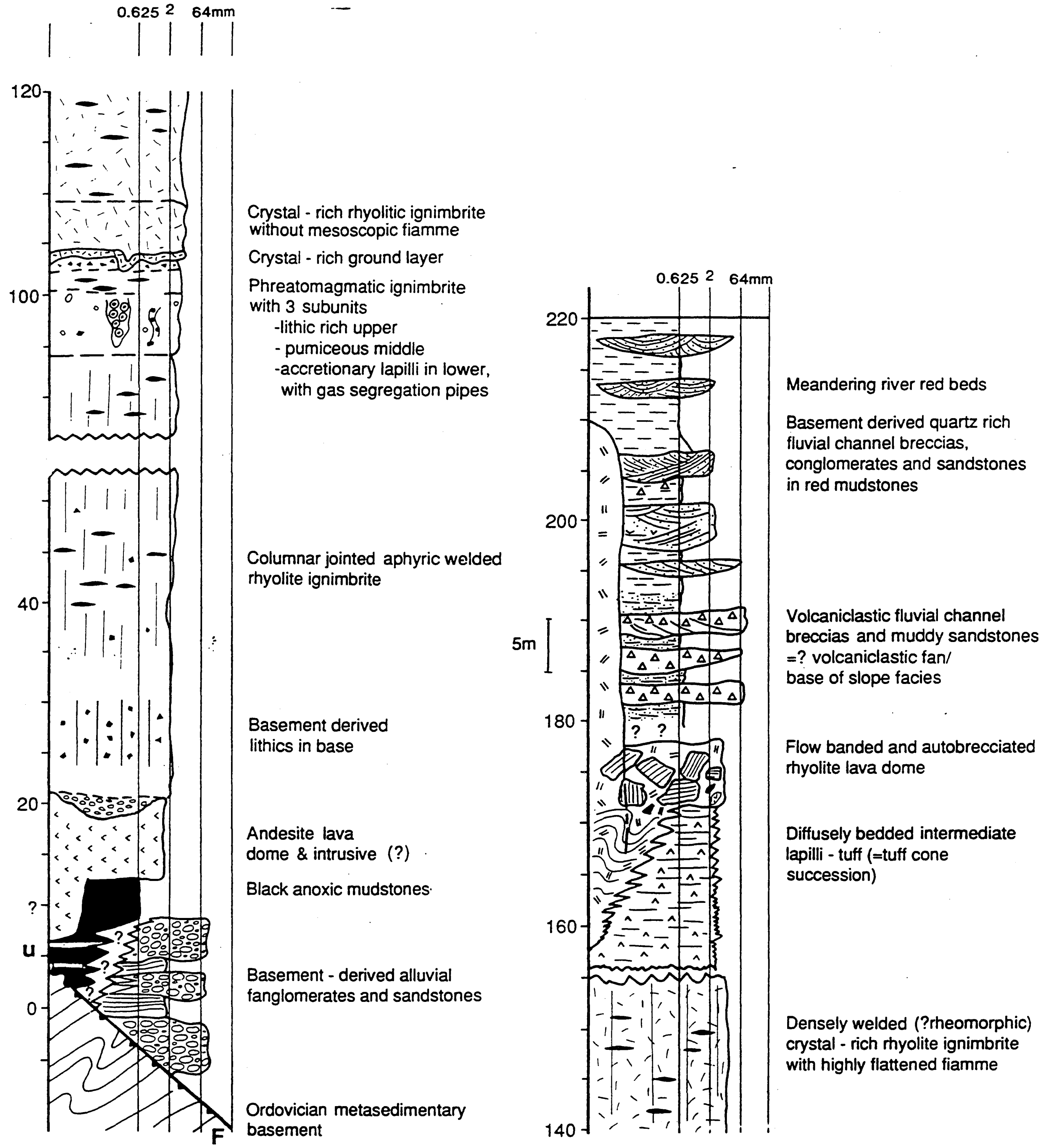


Figure 5.4 *Stratigraphy of the Boyd Volcanic Complex, Eden area.*

is either intrusive or faulted.

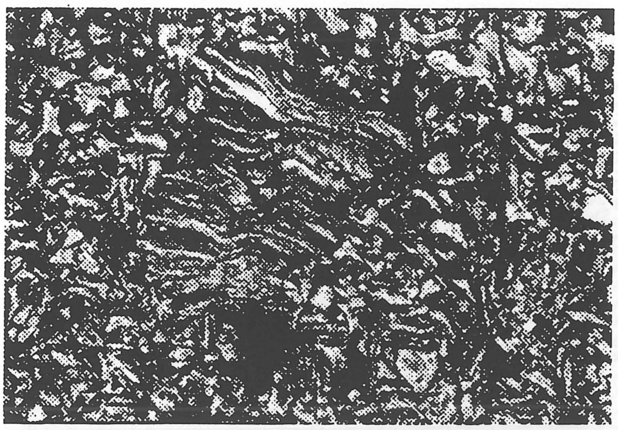
Locality 3: Now climb up to a cave about 5m above sealevel. Above the andesite occurs a quartzite conglomerate succession about 2m thick. When traced laterally, this unit shows significant thickness variations, and is lensoidal in form. It appears to be a fluvial gully deposit, cut into the top of the andesite and deriving debris from basement highland sources.

Overlying the conglomerate in a gradational relationship is a diffusely stratified to massive sandy textured lithology that passes up gradationally into the cliff forming columnar jointed rock unit. In thin section the basal sandy textured interval contains pseudomorphed pumice and shards which show no evidence of welding (Fig. 6.1a). This basal zone is interpreted as the basal non-welded zone and ? precursor plinian fall ash of an ignimbrite eruption succession. This ignimbrite is interesting because it is very fine grained, and essentiall aphyric suggesting it was erupted at about or above its liquidus temperature.

Along the cliff to the east above water level, the base of the ignimbrite contains lithics and and vughy cavities (?lithophysae).

Now return WITH CARE to the lookout car park.

a)



b)

c)

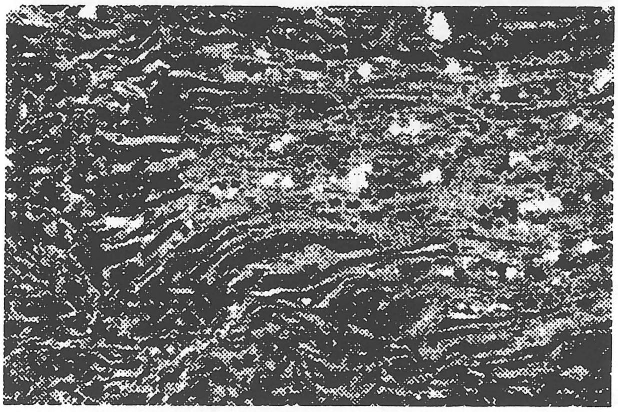


Figure 6.1 *Welding zonation in aphyric ignimbrite, Eden area. (a) Uncollapsed pumice and shards from non welded base. (b) Densely welded zone with extreme welding foliation (?rheomorphicayering). (c) Non-welded top.*

STOP 7 : APHYRIC IGNIMBRITE, COASTAL PARK

Walk to the north along Bramble St. past the fisherman's monument, houses and the start of the park. 200m along enter the park opposite the intersection of Bramble St. and Yule St., and head for the cliff line, finding another steep track down to water level. PLEASE DESCEND WITH GREAT CARE.

The outcrop here is of the same aphyric ignimbrite as seen at STOP 6. It is 70-80m thick. Spectacular columnar jointing (Fig. 7.1) clearly indicate that this outcrop is within the the densely welded zone of this ignimbrite. Perpendicular to the columnar jointing is a closely spaced planar jointing which may be an incipient rheomorphic flow layering. This is suggested by the very intense welding foliation seen in thin section (Fig. 6.1b).

The top of this ignimbrite occurs at the foot of the next headland to the north, and is marked by a non-welded top (Fig. 6.1c).

Carefully return to the road.

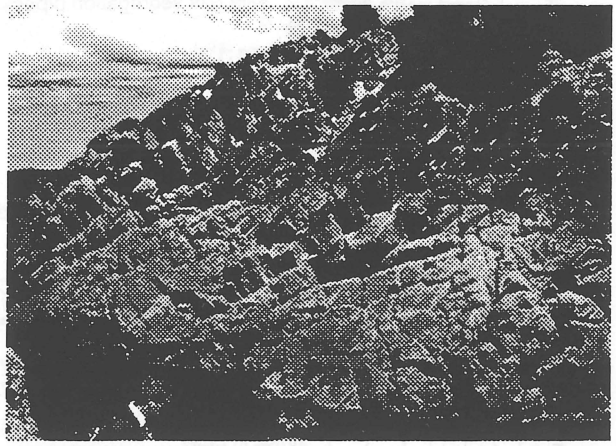


Figure 7.1 *Columnar jointed aphyric welded ignimbrite*

STOP 8 : IGNIMBRITE TYPES, MIDDLE HEAD

From STOP 7 walk northwards along Bramble St., taking the dirt track running below and parallel to it in an easterly direction to the car park bordered by the very large logs. Walk to the cliff edge and look back towards STOP 7, noting the magnificent development of columnar jointing. Follow the track along the cliff edge and through the scrub southwards until you emerge from the scrub. Then carefully clamber down the cliff along the easiest route, veering slightly right as you go. When you have gone as far as you can easily, look at the rock type and its features. Then climb down the final 3-4 metres to the rock platform on the bay to the south.

When on the rock platform look at the rock type it consists of and its features, and then look back at the 3-4 metres rockface you have just climbed down. In it you should see three distinctive facies:

-a lower crystal-poor one (it also makes up the platform)
-a middle, 5-10cm thick, very crystal-rich one (at 1-3m above the platform), and
-an upper crystal-rich one (it makes up the upper cliff).

The lower facies interval can be subdivided into at least three units. The lowest, from water level to the base of the cliff is very fine grained, is massive, and contains accretionary lapilli (Fig. 8.1), often in gas concentrations which are pipe like (Fig. 8.2). This unit, which is about 8m thick

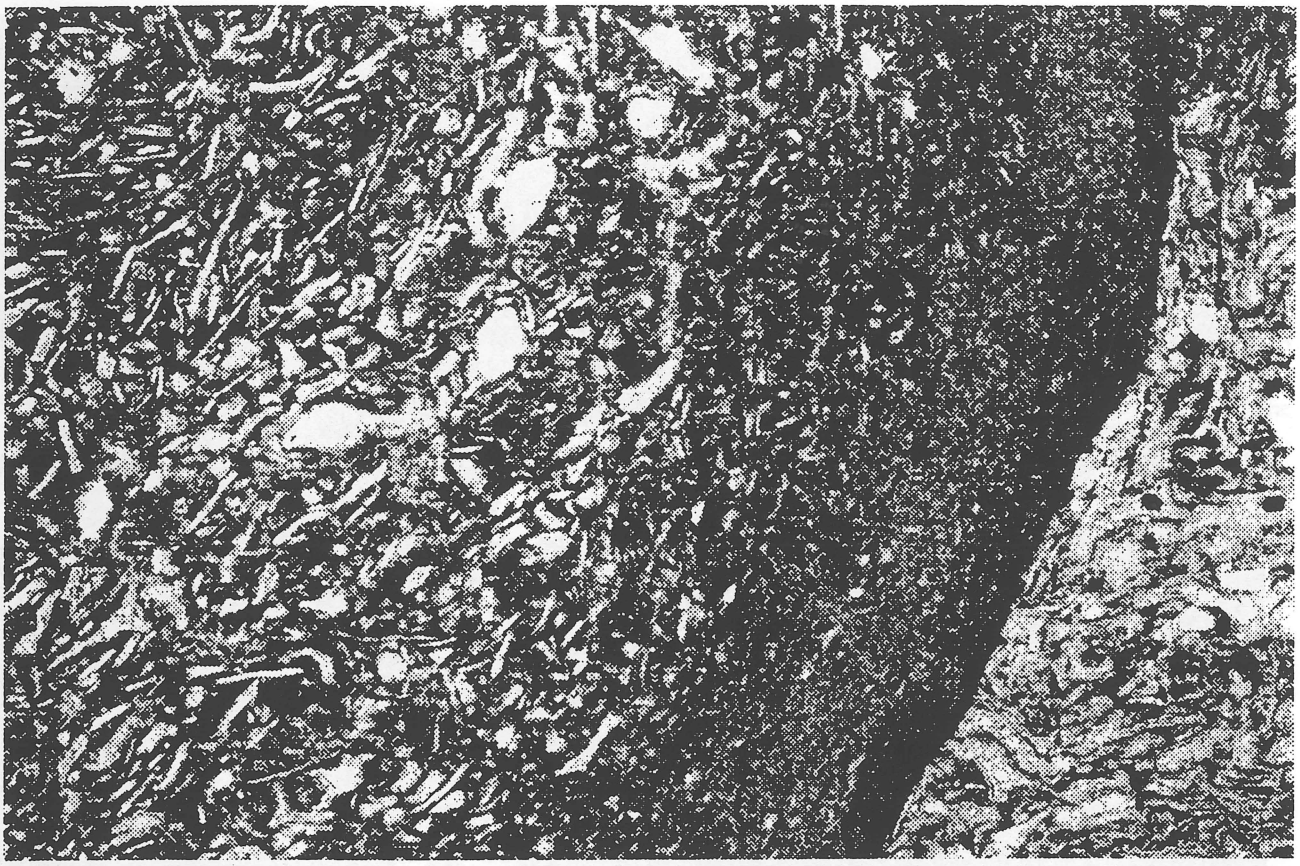


Figure 8.1 *Photomicrograph of shard rich accretionary lapilli, and shard rich matrix of non-welded phreatomagmatic ignimbrite*

is also almost aphyric, and in thin section is full of moderately thick walled glass shards which are non-welded to incipiently welded. The acretionary lapilli also contain abundant shards (Fig. 8.1). Near the base of the cliff surge dune forms in fine tuff are found within the unit (Fig 8.3).

The unit is interpreted to be a phreatomagmatic ignimbrite with preserved gas segregation pipes filled largely with accretionary lapilli. The surge dune structure is interpreted to be an ash cloud surge deposit preserved between two flow units.

The second unit is coarser grained, containing large tube pumice clasts. It is also interpreted as a thin ignimbrite. The top unit, which is also thin (c. 1m), massive and is more lithic rich, is also considered to be a thin ignimbrite.

The second facies, represented by a thin 5-10 cm thick crystal rich layer, has a sharp contact with the underlying phreatomagmatic facies. It is internally massive, Locally thickens and thins, locally scours the underlying unit and contains clasts of it. The upper contact although generally sharp also grades in places into the overlying unit. The coarse crystals are identical to the coarse dispersed crystals of volcanic quartz and feldspar in the 40m thick overlying unit, but the crystal concentration is 5-10 times higher. This unit is clearly genetically related to the overlying facies which is and ignimbrite, and is interpreted as a thin crystal rich ground layer, resulting from elutriation of fine

Figure 8.2 *Pipe-like concentration of accretionary lapilli, interpreted as a gas segregation pipe.*

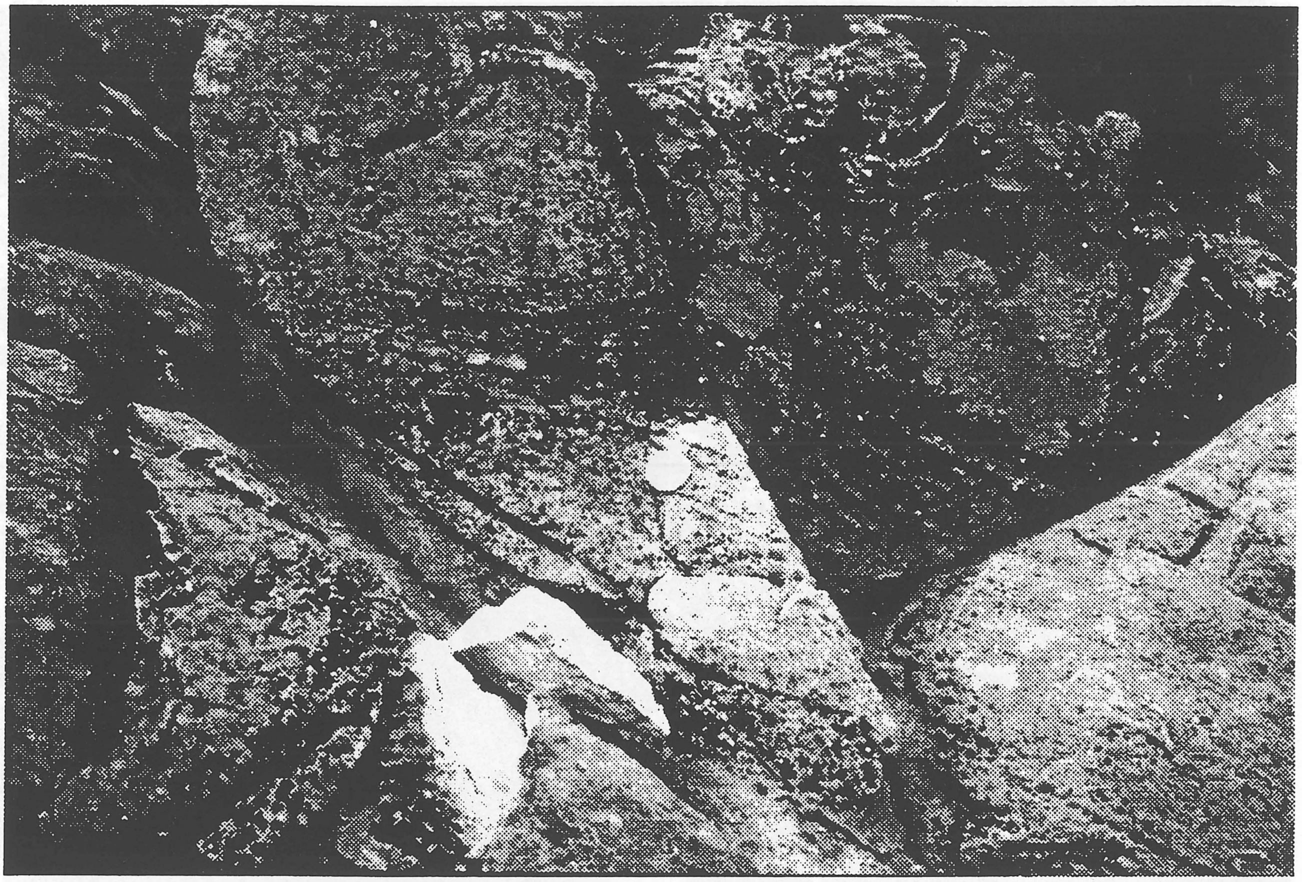


Figure 8.3 *Surge dune form, STOP 8.*

ash from the head of the moving pyroclastic flow, so concentrating dense solids, in this case coarse crystals, in the heaad from which they are deposited, and then overridden by the body of the ignimbrite.

Climb back up to the first ledge where highly attenuated welded pumice clasts are exposed. The attentuation is probably at least ten times. Thin section textures of this ignimbrite are poor, but outcrop characteristics indicate that it is a densely welded ignimbrite. Also exposed are dykes of the phreatomagmatic ignimbrite. Then return to the car park.

STOP 9 : BASALTIC TUFF CONE, MIDDLE HEAD

Walk back into the scrub and follow the cliff edge back to the car park. Then take a path to the eastern side of Middle Head until you can again clamber down the cliff. In the gorge to the east you will see the columnar jointed top of the crystal rich ignimbrite. On the seaward side of the gorge a brown weathering facies makes up the cliff face. Clamber down to the rock platform and assess the characteristics of this facies.

Although the outcrop is generally non-descript, there is a crude seaward dipping stratification through the unit. In addition several fresh enclaves, the fresh cores of spheroidally weathered joint blocks are visible in the middle part of the cliff and further around on the rock platform. Is this facies clastic or coherent?

In places small dark mafic clasts can be seen. Thin sections from the fresh core of joint blocks reveal magnificently preserved poorly vesiculated blocky shards with quench fractures through them (Fig. 9.1). The succession is interpreted as the remains of the distal margins of a phreatomagmatic basaltic tuff cone. Walking further around the platform to the east a vertical weathered basalt dyke is encountered. This basaltic succession, juxtaposed with the rhyolitic succession seen so far is inducative of the bimodal nature of the BVC.

Return to the car park.

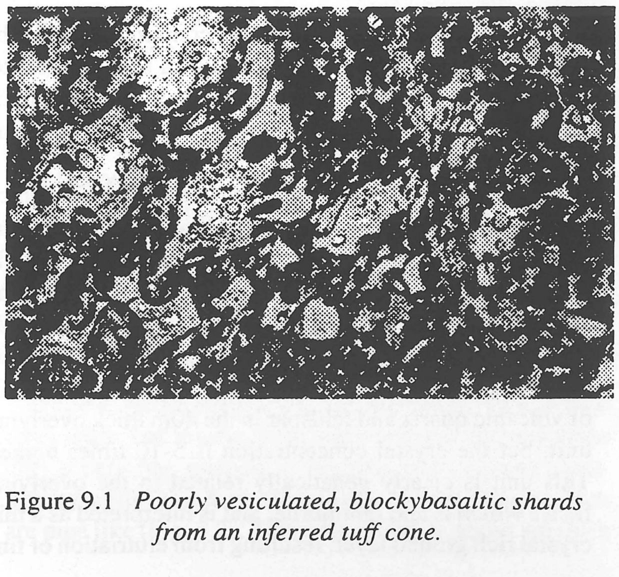


Figure 9.1 *Poorly vesiculated, blockybasaltic shards from an inferred tuff cone.*

STOP 10 : IGNIMBRITE SEQUENCE, FISHERMAN'S WHARVES

Driving back along Bramble and Imlay Sts. towards the shopping centre, turn left past the Fishermen's Co-Op. and park where you can. Walk west along the foreshore, looking briefly at the rock succession as you go. Note the boulders at the end of the road as you walk onto the gravel beach. If the tide permits, walk around to the swimming pool, from where you will be able to see cliffs of the Ordovician basement to the west. Now walk back along the foreshore, noting carefully the stratigraphic succesion as you go. It is the same as you saw at STOPS 7 and 8. Can you find the crystal rich ground layer. Note its gradational relationship to the crystal rich ignimbrite.

STOP 11 : VOLCANIC-RED BED TRANSITION, "CONVENT SECTION"

Drive up the hill towards Eden and turn first right, driving to the corner of Cocora and Calle Calle Streets. Park in the parking area and walk down the hill along the fenceline to the cliff edge. Then clamber down the cliff northwards. First cross over to the small "island" .

Locality 1: This island is made up of flow banded and autobrecciated rhyolite which appears to lie along strike at the same stratigraphic level as the basaltic tuff cone succession. Where first encountered in the bay there is evidence of hydrothermal alteration in the form of vughy outcrop with silicification and manganese alteration. In this area, fine irregular clasts, some vesicular, occur. In thin section these are basaltic, and appear to be cognate fluidal basaltic inclusions suggesting magma mixing.

Further around there are also sedimentary xenoliths, and excellent exposures of vertically orientated coherent flow banded domains and autobreccia occur. A significant part of this rhyolite which is interpreted as late stage rhyolite dome, is autobrecciated.

Locality 2: Cross back to the "mainland" and traverse north along the coast. Although there is poor outcrop, the boulders at the base of the cliff indicate the nature of the facies. The dominant facies is a massive to planar stratified, sometimes cross-stratified volcanic breccia-conglomerate consisting of rhyolite clasts and minor vein quartz clasts. Thin red-grey mudstone horizons also occur. The first outcrop confirms this facies nature. Upsequence the conglomerates become finer grained, the percentage of volcanic clasts decreases, the percentage of vein quartz fragments increases, trough cross bedding becomes more abundant and massive-planar stratified beds less common, and the proportion of mudstone increases (Fig. 5.4). Sandstone bodies higher in the succession often fine upwards, and the mudstones and associated sandstones develop the typical continental red bed character.

This succession is interpreted to represent the degradation of a continental volcanic landscape after the cessation of volcanism, the last volcanic event being represented by the flow banded-autobrecciated rhyolite. The upsequence fining and change in provenance is thought to represent a change from initial distal alluvial fan-braid plain environments to high sinuosity meandering river environments as the gradient of the volcanic source and adjacent depopsitional surface declined through time. The increase in vein quartz, presumably derived from erosion of the deformed basement metasedimentary succession, represents headward erosion of the fluvial system draining the volcanic landscape, beyond the margins of the rift, and back into the adjacent basement highlands. The whole succession thus represents the gradual infilling of the initial extensional rift graben basin system and the lowering of the regional relief (Fig. 11.1).

Locality 3: Towards the end of the day's traverse there is an outcrop of rhyolite. Does this represent late stage volcanic resurgence? The margins of the rhyolite are essential vertical and in abrupt contact with the enclosing sedimentary succession (Fig. 11.2). four possible origins exist for the rhyolite:
-intrusive
-late stage resurgent extrusive
-up faulted
-erosional relict of high aspect ratio rhyolite erupted at the same time as the "island rhyolite".

There is no evidence of contact metamorphism, inclusions of sedimentary clast or faulting. The preferred interpretation is that the rhyolite is an erosional relict of the final rhyolite forming eruptive phase. The enclosing sedimentary bedding laps up and drapes against the margins of the rhyolite (a buttress unconformity) and clasts of rhyolite are found in the sedimentary succession adjacent to the contact, indicating spalling of debris from the rhyolite mass into the accumulating fluvial floddplain succession around.

Summary:

1. Extensional tectonics were responsible for the initiation of the basin and the devolopment of the intial environments (alluvial fan, ponded lacustrine).
2. Voluminous volcanism soon flooded the basin with pyroclastics and lavas. The pyroclastics are dominated by both welded and non-welded rhyolitic ignimbrites which appear to represent an outflow succession, not a caldera fill succession. The location of the eruptive centre is unknown but may have been in the Wonboyn area
3. Both magmatic and phreatomagmatic pyroclastic eruptive phases appear to have been important.
4. Localised basaltic tuff cone forming eruptions testify to the bimodal nature of the volcanism and the influence of mantle processes in the volcanism and tectonics.
5. The last phase of volcanism appears to have been localised non-explosive rhyolite dome forming eruptions which produced an irregular relief.
6. The great thickness of volcanics and the numerous eruptive units represented appear to have been emplaced

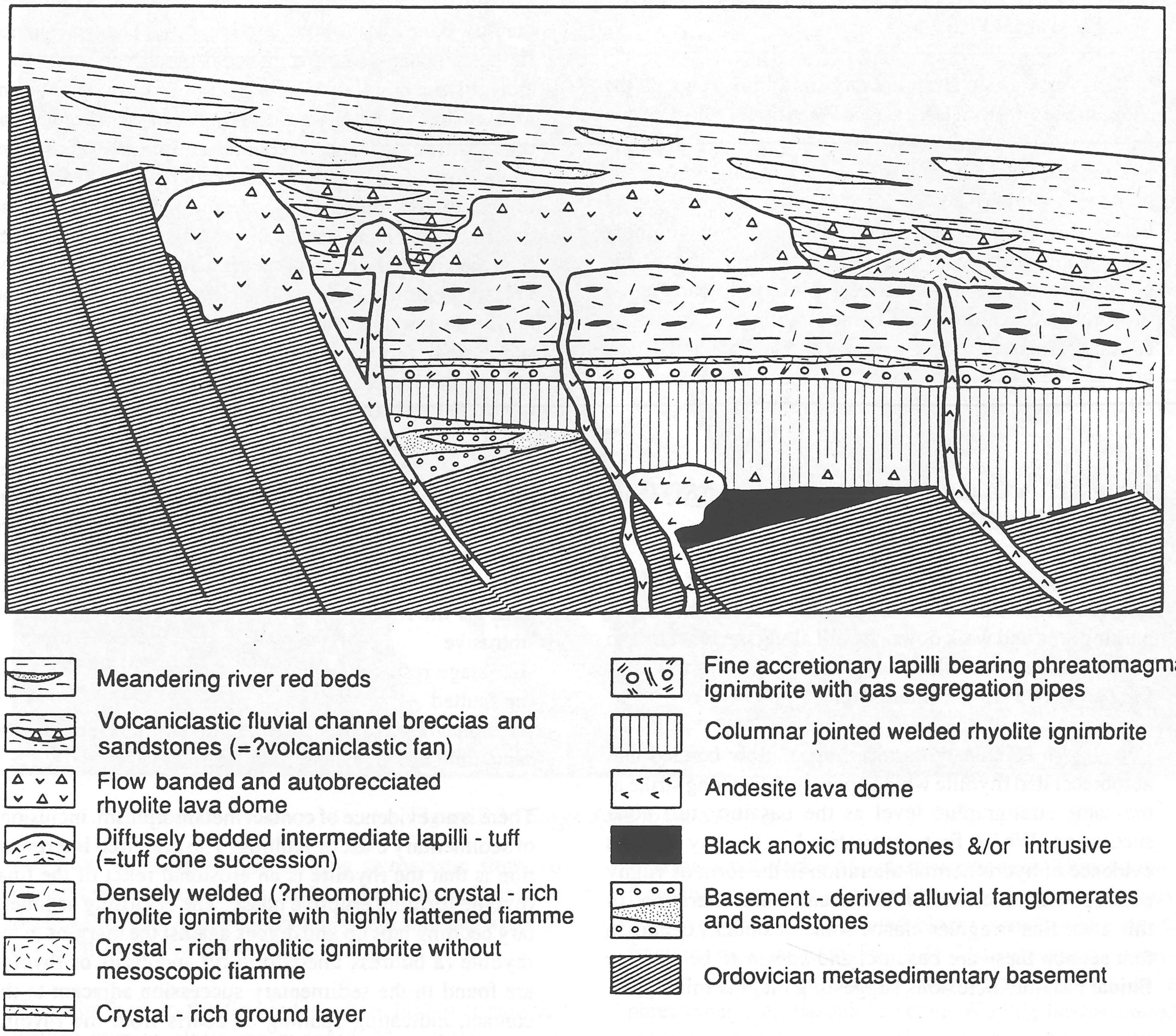


Figure 11.1 *Schematic cross-section of the rift basin margins showing the spatial relationships between basement topogrpahy, sedimentation, and volcanic units, Eden area.*

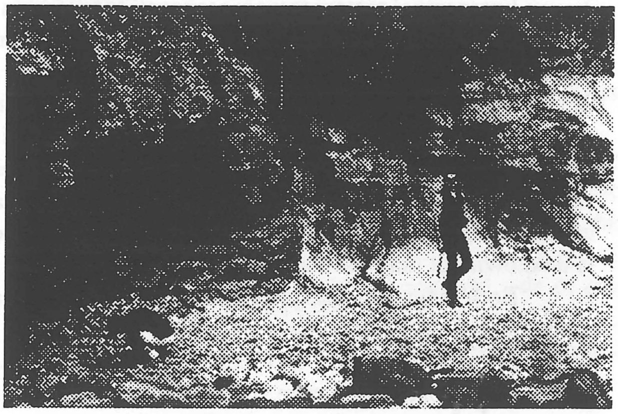


Figure 11.2 *Vertical buttress unconformity between rhyolite and younger sedimentary successions draping the contact.*

in very rapid succession in a short time period, as suggested by the paucity of intervening sedimentary units between volcanic units, the presevation of the non-welded top of at least one ignimbrite, as well several thin non-welded ignimbrites, and the preservation of part of a tuff cone.

7. The high relief volcanic landscape was eroded and buried, and this transition from high relief active volcanic landscape to low relief fluvial-alluvial non-volcanic landscape is reflected in the upsequence provenance changes and changes in the depositional environment. The time involved in this transition is likely to have been very long by comparison with the time required to emplace the primary volcanic succession.

DAY 3: BASIN MARGIN FACIES ASSOCIATIONS OF THE ? SUBAQUEOUS BOYD VOLCANIC COMPLEX, BUNGA BEDS, LATE DEVONIAN BOYD VOLCANIC COMPLEX

INTRODUCTION

The northernmost part of the Boyd Volcanic Complex is an outlier known as the Bunga Beds (Scott 1972, Fergusson et al. 1979; Fig. 12.1). The Bunga Beds are surrounded by a basement of deformed Ordovician metasedimentary quartz-rich turbidites and pelites with quartz veins, and Lower Devonian granitic plutons (Fig. 12.1). The only known contact with the Ordovician basement along the southern margin at Picnic Point is faulted, but appears to have been an unconformity (see below). Outcrop of the Bunga Beds is poor except along the coast. The preserved north-south extent of the Bunga Beds is 8 km, but the size of the original basin is unknown.

The Bunga Beds also consist of a bimodal association of volcanics and associated sedimentary rocks but differ from the rest of the Boyd Volcanic Complex in lacking continental red beds and ignimbrites. Rhyolites predominate over basalts, and both are represented by coherent bodies and a diversity of volcaniclastics. The sedimentary rocks contain fossil fish and plant fragments of late Givetian-early Frasnian age (latest Middle-earliest Late Devonian), representing the oldest part of the Boyd Volcanic Complex (Fergusson et al. 1979). The fish fossils indicate a brackish to marine environment (J. Long, pers. comm.).

The sedimentary succession is also different in consisting not of red beds. Two main facies associations, a basin margin facies association, and a deeper water basin centre facies association, have been recognised. The basin margin facies association occurs at the northern end of the Bunga Beds outlier, extending southward from Goalen Head, where a Jurassic gabbroic diorite outcrops, and at the southern end of the outlier, extending northwards from Picnic Point (Fig. 12.1). The basin centre facies association outcrops between these areas and is the facies association most frequently associated with the rhyolites. It is therefore significant for determining the palaeoenvironmental setting for volcanism.

The basin centre facies association consists of grey to black mudstones, turbidites, debris flow deposits and slide-slump deposits, all indicative of a relatively deep subaqueous environment. The origin of this subaqueous environment is not known, but the most likely origin is crustal extension and subsidence, leading to marine transgression or formation of a deep brackish lake.

The coincidence of difference in environment together with difference in volcanic products is significant, and provides and ideal opportunity to study the effects of environment on eruptive style and products.

The the day will be spent examining the northern and southern basin margin successions along Bunga Beach and at Picnic Point respectively (Fig. 12.1).

Northern Basin Margin Successsion: Bunga Beach

From Merimbula drive north along the Princes Highway. Just north of the main shopping centre turn right into Reid St. which is sign posted for Tathra and Tura Head along the coast road. Continue 17-18 km north to a T-junction with the Snowy Mountains Highway, turn right to Tathra. Take the coast road to Bermagui. Pass through both segments of the Mimosa Rocks National Park as well as Wapengo. Bunga Beach is not signposted, and is on private property off an unidentified road to the right of the main coast road which is dirt.

The northern basin margin facies association consists of interbedded sandstones and grey mudstones, is approximately 100 m thick, is openly folded, and its base is truncated by the Jurassic gabbro at Goalen Head. In addition it is intruded by contemporaneous rhyolites, basalt dykes and a (?) Jurassic microdiorite dyke (Fig. 12.1). The top of the sedimentary succession is buried under beach sand and appears to be conformably overlain by about 30 m of black mudstones of the basin centre facies association at the southern end of Bunga Beach.

The most pervasive facies of the basin margin facies association is a grey, massive to laminated mudstone and siltstone facies, which represents the ambient conditions in the depositional environment. Although intervals of this facies are up to 8 m thick, beds are rarely more than 50 cm thick and are interbedded with other facies (Fig. 12.2). Beds are tabular except in the upper part of the succession where together with other facies of this association, they fill channels. Occasional short sub-vertical to horizontal burrows only several millimetres in length (?Chrondites) and carbonaceous plant fragments, including large logs of Lepidodrendon, are found. Fossil soils, rootlet horizons and mudcracks are lacking, and so this facies was deposited under quiet water conditions, away from the influence of persistent bottom or surface currents. The depositional surface was generally flat except for occasional channels. The grey mudstones suggest moderately oxidising conditions in the basin margin environments compared with the anoxic conditions represented by the black mudstones of the basin centre association .

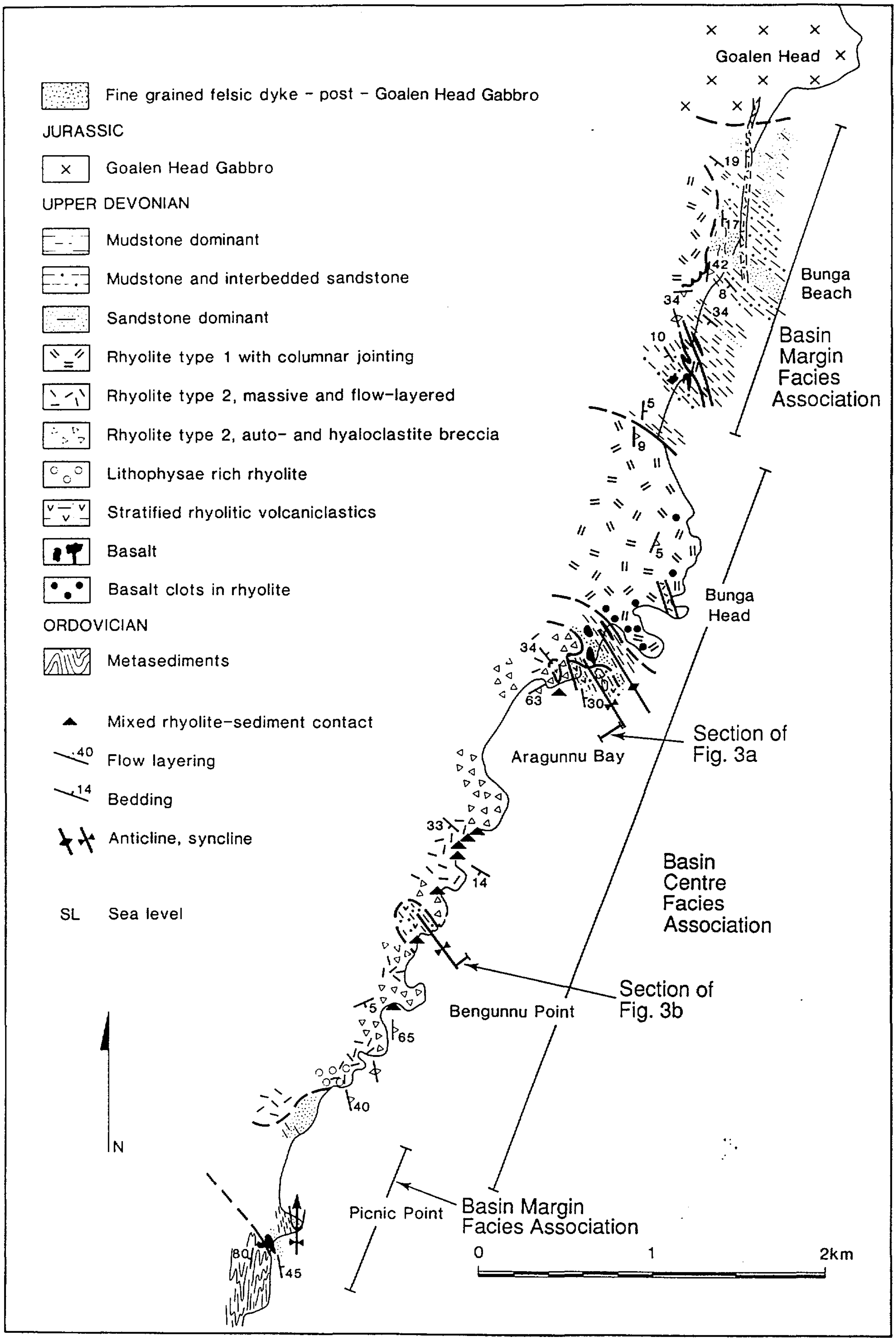


Figure 12.1 *Geology of the Bunga Beds outlier, Late Devonian Boyd Volcanic Complex (From Cas et al. 1990)*

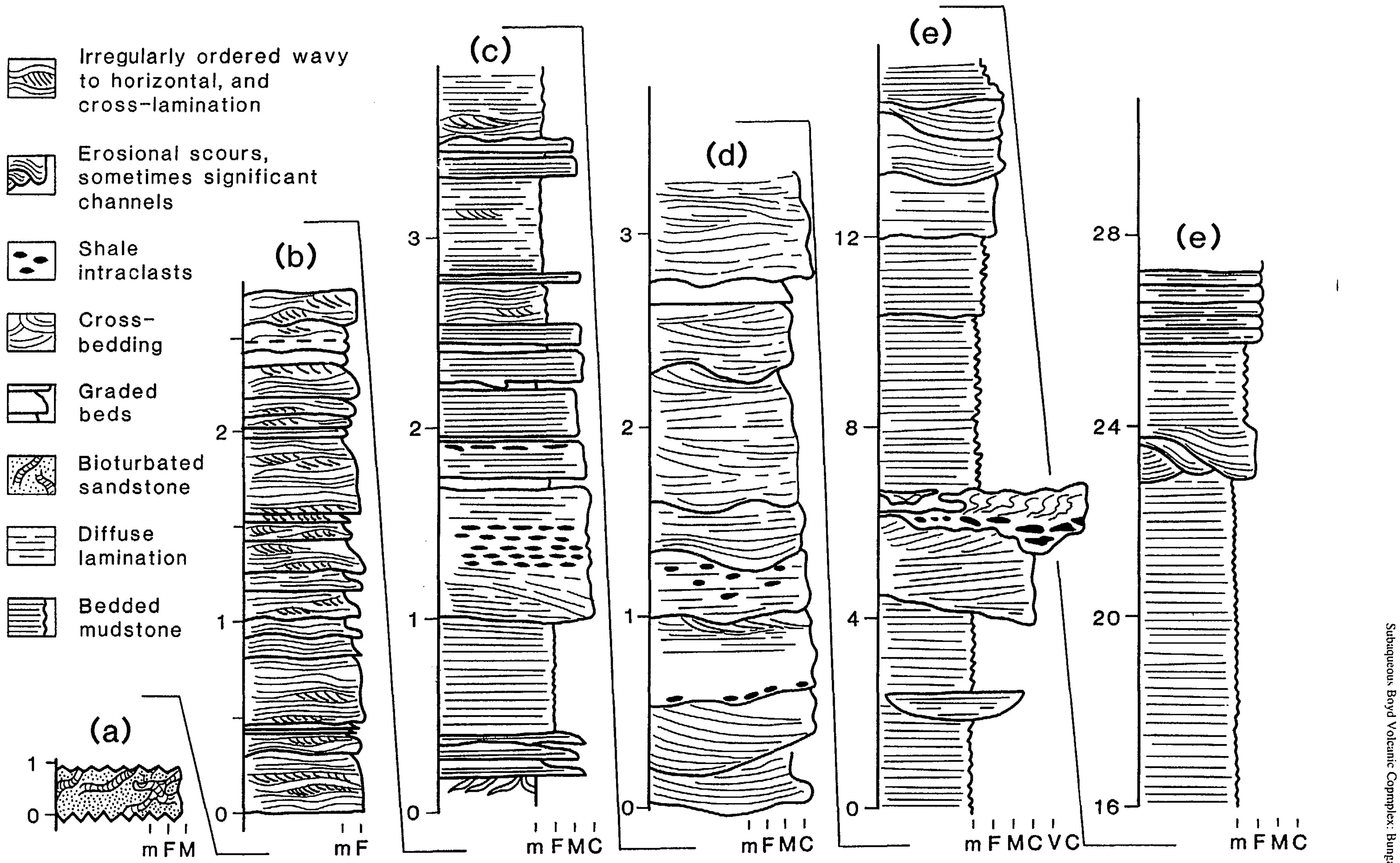


Figure 12.2 *Representative logs of the northern Basin Margin Facies Association, Bunga Beach.*

Tabular, laminated to cross-laminated vitric siltstones and fine sandstones are interbedded with the mudstones and occur in intervals of several or more metres thick. Beds are tabular and laterally continuous, and vary in thickness from several to 40 centimetres. Bases are generally sharp and tops gradational into mudstone. Internally the beds contain random alternations of, and various combinations of planar and wavy lamination and asymmetrical ripple cross laminae. The facies consists of > 90% cuspate and straight walled glass shards, and minor angular volcanic quartz and feldspar fragments.

Although this grain assemblage indicates an unconsolidated ash grade rhyolitic pyroclastic source, the general facies aspect is suggestive of turbidites, indicating resedimentation by epiclastic processes. However, the random order of the internal sedimentary structures is indicative of fluctuating flow velocities rather than steady decelerating flow as occurs in turbidity currents. The depositional agent was a finite flow, not a normal long-lived bottom current, as indicated by interbedded mudstone layers. This facies is interpreted to be the product of continuous feed turbulent underflows sourced by river mouths in flood (e.g. Lambert and Hsu 1979), and may represent slope deposits perhaps associated with a basin margin delta. Cross-laminae indicate palaeo-flow directions to the west.

Also interbedded with the mudstones are tabular, massive to laminated to cross-stratified sandstone beds. Beds vary from several centimetres to over one metre thick. Intervals of beds may be several metres thick, and amalgamation occurs. Beds are tabular and laterally continuous, their bases are sharp and conformable or erosional, and the tops are sharp or gradational into siltstone or mudstone. Individual beds are either internally massive, diffusely layered, horizontally laminated, or graded. Some graded beds pass from a basal massive interval up into cross bedding, then horizontal lamination, then cross lamination and mudstone. In this facies group, concentrations of mudstone intraclasts define layering in places.

This facies consists of both basement derived and volcanic detritus. Subangular to subrounded monocrystalline quartz comprises 44% of the framework, sericite and chlorite altered volcanic lithics 17%, and metasedimentary slate fragments 10%. Minor components include rounded feldspar, chert, polycrystalline quartz and rare tourmaline and detrital mica. Silty matrix constitutes up to 20% of this facies.

This facies like the previous facies represents subaqueous mass-flow events. Beds with simple internal features and sharp tops and bases represent high concentration, rapidly decelerating turbidity currents. The internally more complex beds with multiple structural divisions represent steadily decelerating and in some cases fluctuating velocity flows. These flows could also represent river mouth fed density currents, involving coarser sediment and higher energy flood events. Isolated beds of this facies could represent crevasse splay events from a basin margin river-delta system.

The only facies indicating any substantial tractional reworking in the Bunga Beach section, is a cross bedded sandstone facies interval 8m thick (Fig. 12.2). The base of the facies is slightly erosional and internally it consists of broad, scouring, sets of trough cross-bedded sandstone with cross beds indicating flow to the southwest. This facies is interpreted as a distributary channel fill or channel mouth bar.

The sandstones of this facies are moderately to well sorted and contain angular fragments of basement derived pelite, quartz-rich siltstones, minor vein quartz and chert, together making up 26% of the framework. 74% of the framework consists of texturally immature volcanic quartz, vitric lithic fragments, and minor feldspar and cuspate shards, indicating a probable rhyolitic pyroclastic source. The cross-bedding indicates that, like the previous sandstone and siltstone facies, the volcanic detritus of this facies is resedimented, probably from a subaerial source.

Locally clastic dykes of pebbly sandstone also occur along Bunga Beach close to a 15m wide rhyolite dyke which is interpreted to have caused fluidisation and upward intrusion of unconsolidated, coarse, water-saturated sediment through which it passed.

The total facies association is therefore representative of subaqueous sedimentation in a quiet water environment. The grey mudstones may in part be hemipelagic sediment, but may also be the deposits of overflows or underflows, derived from a density stratified clay-rich water column, originating from the mouths of rivers in flood (e.g. Sturm and Matter 1978). The cross-bedded facies could be either a channel deposit or channel mouth bar and occurrence within a mudstone succession suggests a distributary channel associated with a basin margin delta or fan/delta slope system. The coarser tabular siltstones and sandstones are interpreted as short-lived to continuous feed underflows, and where they occur in isolation, as crevasse splays originating from a breach in a distributary channel margin. The whole succession probably accumulated in a small delta or fan delta at the margin of the basin. Lack of dessication features, rootlet beds, soil horizons and pervasive bioturbation suggest a constantly submerged setting.

There is an overall upward fining trend in the Bunga Beach succession (Fig. 12.2), suggesting transgression and deepening. This is supported by a change from grey oxidised mudstones to the black anoxic mudstones of the basin centre facies association at the southern end of Bunga Beach.

STOP 12 : GOALEN HEAD GABBRO, BUNGA BEACH

From access road where the road first comes closest to the sea, park and walk to the beach. At the northern end is massive outcropping gabbro of Juassic age. It is cut buy a fine grained glassy microdiorite, which also cuts the Bunga Beds sedimentary succession to the south. It is also thought to be Jurassic in age. Jurassic plutonism also occurs further to the north around Tilba Tilba, and is thought to be related to the Cretaceous opening of the Tasman Sea.

STOP 13 : BIOTURBATED VITRIC SANDSTONE, BUNGA BEACH

Walk south along the beach and head away from the shore line towards the foot of the dunes. Amongst the sand infront of the dunes is an outcrop of fine sandstone with bioturbation.

Bioturbated sandstone occurs in only this small outcrop near the preserved base of the succession on Bunga Beach, and consists of a high density of cross cutting, sub-horizontal to sub-vertical burrows (? Rhizocorallium). The intensity and style of burrowing suggests a relatively shallow environment. The facies consists of >85% juvenile volcanic material, predominantly straight-walled and cuspate shards, with minor quartz and feldspar fragments. Clearly the original source of the sediment was pyroclastic although the depositional mode is now uncertain due to the bioturbation. A clue to this lies at the next outcrop.

STOP 14 : BASIN MARGIN CONTINUOUS FLOW VITRIC TURBIDITES, BUNGA BEACH

Continue walking south to a low rock face about 2m high near the back of the beach. Tabular, laminated to cross-laminated vitric siltstones and fine sandstones are interbedded with the mudstones and occur in intervals of several or more metres thick (Fig. 12.2). Beds are tabular and laterally continuous, and vary in thickness from several to 40 centimetres. Bases are generally sharp and tops gradational into mudstone. Internally the beds contain random alternations of, and various combinations of planar and wavy lamination and asymmetrical ripple cross laminae (Fig. 14.1). The facies consists of > 90% cuspate and straight walled glass shards, and minor angular volcanic quartz and feldspar fragments.

Although this grain assemblage indicates an unconsolidated ash grade rhyolitic pyroclastic source, the general facies aspect is suggestive of turbidites, indicating resedimentation by epiclastic processes. However, the random order of the internal sedimentary structures is indicative of fluctuating flow velocities rather than steady decelerating flow as occurs in turbidity currents. The depositional agent was a finite flow, not a normal long-lived bottom current, as indicated by interbedded mudstone layers. This facies is interpreted to be the product of continuous feed turbulent underflows sourced by river mouths in flood (e.g. Lambert and Hsu 1979), and may represent slope deposits perhaps associated with a basin margin delta. Cross-laminae indicate palaeo-flow directions to the west.

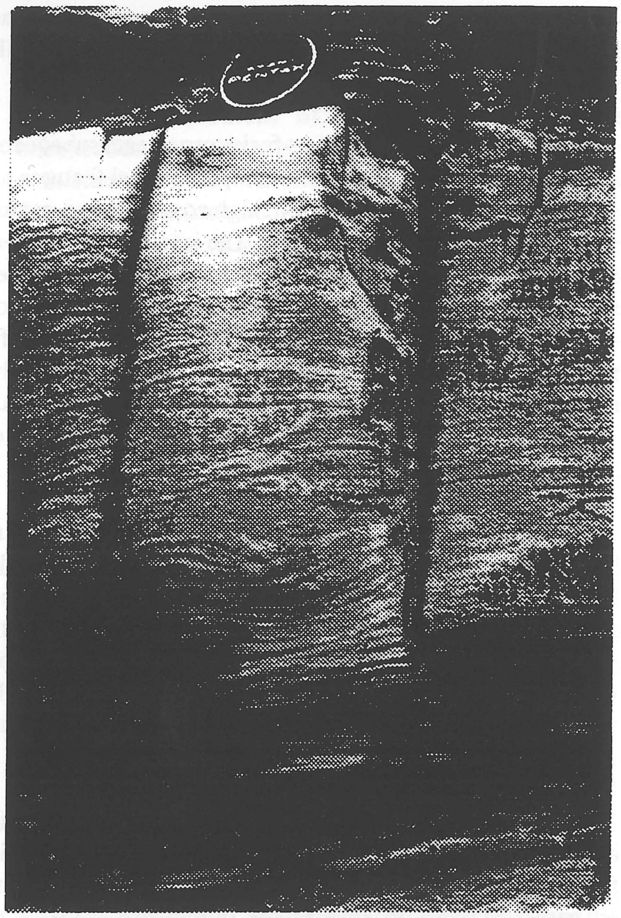


Figure 14.1 *Graded tabular bed of laminated to cross laminated vitric sandstone.*

STOP 15 : BASIN MARGIN SANDSTONES AND GREY MUDSTONES, BUNGA BEACH

Return to the shoreline zone where on the rock platform, the vitric sandstone facies passes into a succession of interbedded grey mudstones and tabular, massive to laminated to cross-stratified sandstone beds (Fig. 12.2). This facies which dips southwards extends down the beach for 400-500m, and for much of this distance is cut by the glassy microdiorite dyke noted at STOP 12.

Beds vary from several centimetres to over one metre thick. Intervals of beds may be several metres thick, and amalgamation occurs. Beds are tabular and laterally continuous, their bases are sharp and conformable or erosional, and the tops are sharp or gradational into siltstone or mudstone. Individual beds are either internally massive, diffusely layered, horizontally laminated, or graded. Some graded beds pass from a basal massive interval up into cross bedding, then horizontal lamination, then cross lamination and mudstone. In this facies group, concentrations of mudstone intraclasts define layering in places.

This facies consists of both basement derived and volcanic detritus. Subangular to subrounded monocrystalline quartz comprises 44% of the framework, sericite and chlorite altered volcanic lithics 17%, and metasedimentary slate fragments 10%. Minor components include rounded feldspar, chert, polycrystalline quartz and rare tourmaline and detrital mica. Silty matrix constitutes up to 20% of this facies.

This facies like the previous facies represents subaqueous mass-flow events. Beds with simple internal features and sharp tops and bases represent high concentration, rapidly decelerating turbidity currents. The internally more complex beds with multiple structural divisions represent steadily decelerating and in some cases fluctuating velocity flows. These flows could also represent river mouth fed density currents, involving coarser sediment and higher energy flood events. Isolated beds of this facies could represent crevasse splay events from a basin margin river-delta system.

The only facies indicating any substantial tractional reworking in the Bunga Beach section, is a cross bedded sandstone facies interval 4m thick (Fig. 12.2). The base of the facies is slightly erosional and internally it consists of broad, scouring, sets of trough cross-bedded sandstone with cross beds indicating flow to the southwest. This facies is interpreted as a distributary channel fill or channel mouth bar.

The sandstones of this facies are moderately to well sorted and contain angular fragments of basement derived pelite, quartz-rich siltstones, minor vein quartz and chert, together making up 26% of the framework. 74% of the framework consists of texturally immature volcanic quartz, vitric lithic fragments, and minor feldspar and cuspate shards, indicating a probable rhyolitic pyroclastic source. The cross-bedding indicates that, like the previous sandstone and siltstone facies, the volcanic detritus of this facies is resedimented, probably from a subaerial source.

STOP 16 : INTRUSIVE RHYOLITES, BUNGA BEACH.

Continue down the beach through the interbedded sandstones and mudstones. At the back of the beach several large closely spaced outcrops of rhyolite will come into view.

Two distinct rhyolite compositions can be recognized in the field in the Bunga Beds based on differences in size and abundance of phenocrysts. Four bodies of coherent rhyolite in the northern half of the area are all highly porphyritic rhyolite characterized by 25-30% coarse 2mm phenocrysts of quartz, plagioclase, K-feldspar and amphibole (rhyolite type 1 of Fig. 12.1). By comparison, rhyolites to the south of the Bunga Head area are only moderately porphyritic, being characterized by 10-20% 1mm quartz, K-feldspar and plagioclase phenocrysts (rhyolite type 2 of Fig. 12.1).

The outcrops at the back of the beach represent are an example of the highly porphyritic, coherent rhyolite. The body outcrops lower in the stratigraphy (Fig. 12.1). At its southern margin, polygonal joints converge radially inwards from a locally preserved smooth convex surface, interpreted to be the original margin of the rhyolite body or a cooling surface parallel to the margin. This surface is discordant to the nearby sedimentary sequence and is interpreted to be intrusive, probably into wet semi-consolidated sediments, as other examples below are. Consequently the rhyolite is inferred to be an intrusive dome (cryptodome) or a partly intrusive lava dome. A similar but smaller crystal rich rhyolite occurs at the northern end of Aragunnu Beach, in close association with moderately porphyritic rhyolite.

STOP 17 : RHYOLITE FEEDER DYKE AND SYN-DEPOSITIONAL BASALTIC AND CLASTIC INTRUSIVES, BUNGA BEACH

Walk to the end of the main beach where a small rocky headland represents the end of the beach. In this location there is a complex association of features including open folding of the sedimentary succession, a prominent vertical rhyolite dyke, irregular basalt intrusives and coarse clastic dykes. The prominent 15m thick dyke forms the fourth highly porphyritic rhyolite body in the area (Fig. 12.1) although it is less porphyritic and finer grained than the other bodies. The dyke is vertical, virtually perpendicular to bedding in the sedimentary sequence, and displays strong flow-layering parallel to the dyke margins. Orientation, thickness, composition, and position in the stratigraphy suggest that this rhyolite could be a feeder dyke for the thick, tabular Bunga Head rhyolite body which it must intersect just off the coast (Fig. 12.1) and/or to the large intrusive body to the north discussed above.

The basaltic intrusives are highly irregular and occur on the platform both to the north and south of the headland. Their irregularity indicates that the enclosing sedimentary successsion had no mechanical strength and the intrusives were thus syn-depositional intrusives. The close spatial relationship between the rhyolite and basalts is again a reflection of the bimodal nature of the Bunga Beds and Boyd Volcanic Complex.

The clastic dykes are coarse and pebbly and were produced by liquefaction of deeper seated paarts of the stratigraphy. Liquefaction may have been caused by loading effects, so causing excess pore pressures, or to heating of interstitial pore waters in proximity to rising magmas. The close association of the clastic dykes with the intrusive rhyolite and basalts lends some support to the latter hypothesis.

STOP 18 : BASIN MARGIN TO BASIN CENTRE TRANSITION, BUNGA BEACH

100m south of the rhyolite dyke there is platform and low cliff exposure of the last of the northern basin margin mudstone facies. The top of this facies association is buried under beach sand and appears to be conformably overlain by about 30 m of black mudstones of the basin centre facies association.

The most pervasive facies of the basin margin facies association is a grey, massive to laminated mudstone and siltstone facies, which represents the ambient conditions in the depositional environment. Although intervals of this facies are up to 8 m thick, beds are rarely more than 50 cm thick and are interbedded with other facies (Fig. 12.2). Beds are tabular except in the upper part of the succession where together with other facies of this association, they fill channels. Occasional short sub-vertical to horizontal burrows only several millimetres in length (?Chrondites) and carbonaceous plant fragments, including large logs of Lepidodrendon, are found. Fossil soils, rootlet horizons and mudcracks are lacking, and so this facies was deposited under quiet water conditions, away from the influence of persistent bottom or surface currents. The depositional surface was generally flat except for occasional channels which suggest periodic extension of perhaps deltaic distributary channels. The grey mudstones suggest moderately oxidising conditions in the basin margin environments compared with the anoxic conditions represented by the black mudstones of the basin centre association. Thin horizons of black mudstone have occurred in the basin margin facies association, indicating fluctuating oxidising-reducing conditions, but these have been minor. To the south black mudstones of the basin centre facies association are prominent.

STOP 19 : BASIN CENTRE SEDIMENTS AND BUNGA HEAD RHYOLITE LAVA CONTACT, BUNGA BEACH

Walk across the smaller beach southwards towards the high headland of Bunga Head. At low tide rock platform exposure reveals prominent exposures of basin centre black mudstones. At the base of the cliff massive sandsones underlie the sharp base of the very thick flow banded Bunga Head rhyolite body.

Of the four highly porphyritic rhyolite bodies in the north, the Bunga Head rhyolite is the thickest and most extensive. The body is a single unit at least 180m thick and 1300m long with subvertical columnar jointing, diffuse subhorizontal flow-layering throughout, and a homogeneous coarsely porphyritic texture. Small fluidal-shaped basaltic inclusions are scattered through the rhyolite in some areas, especially at the southern margin of the unit (STOP 23). The basal contact of the rhyolite body is well exposed at the northern end of Bunga Head (Fig. 12.1). It is subhorizontal, parallel to flow-layering in the rhyolite, and completely concordant with underlying sediments of the deep-water basin centre facies association.

In detail this contact varies from relatively smooth to finely undulating with 2-3cm amplitude load casts of coherent rhyolite and intervening flames of black mudstone. The rhyolite is coherent and columnar jointed right to the base but has a 30cm thick nodular textured contact zone which is moderately vesicular and contains rounded autoclasts set in a matrix of coherent flow-layered lava (Fig. 9a). Sediments beneath the rhyolite are relatively undisturbed except for minor soft sediment distortion and partial mobilization of sandy beds due to the loading effect of the rhyolite. This contact therefore records the remarkably passive emplacement of a thick lava over wet unconsolidated sediments of the basin floor. The very consistent subhorizontal orientation of flow-layering and the vertical orientation of jointing, together with the nature of the basal contact suggest the original morphology of the rhyolite was that of a thick tabular sheet rather than a dome and also suggests high effusion rates. This is in spite of the high crystal content, which suggests a sub-liquidus temperature of emplacement. Low angle discordances in the flow banding suggests localised ramping during flow in this viscous lava.

STOP 20 : SOUTHERN BASAL CONTACT AND BASIN MARGIN ASSOCIATION, PICNIC POINT

Return along the beach to the vehicles. Return to the coast road and turn south. Drive through the northern segment of the Mimosa Rocks National Park. Take the Wapengo Lake turnoff on the left. This is narrow dirt road. Drive to the end and turn left to Picnic Point. From the parking-camping area climb down to the small beach to the south, and walk across it to the major headland.

At Picnic Point the only contact between the Bunga Beds and the deformed Ordovician metasedimentary basement is exposed (Fig. 12.1). At the contact the basal facies of the Bunga Beds is a closed framework conglomerate-breccia consisting of basement derived metasedimentary clasts with a matrix of coarse lithic sandstone Figure 20.1). It is in faulted contact with the Ordovician and is itself sheared, but probably represents a faulted unconformity. The conglomerate is about 5 m thick and although its top is not exposed, it does not appear to be substantially thicker, the next outcrop being massive to laminated grey mudstone (Fig. 20.1). Given that the basal conglomerate is in contact with the fault and contains clasts of metasediments and vein quartz derived from the adjacent Ordovician basement, it is likely that the fault represents a reactivated basin margin normal fault scarp against which the conglomerates were deposited (perhaps in an initial alluvial fan?) and from which they derived much of their debris. Basaltic dykes are associated with this southern bounding fault suggesting it may have acted as a magma conduit.

STOP 21 : BASIN MARGIN SEDIMENTATION, PICNIC POINT

Walk back to the headland that the vehicles are parked on and traverse around it to the north. The southern basin margin succession extends from Picnic Point north (Fig. 12.1), is also openly folded and also contains a similar facies association of massive to laminated grey mudstone with contemporaneous, infilled channel-like downwarps, tabular laminated-cross laminated siltstone fine sandstone, graded tabular siltstones and sandstones, and cross-bedded pebbly sandstone (Fig. 20.1). These are considered to have had the same depositional origins and

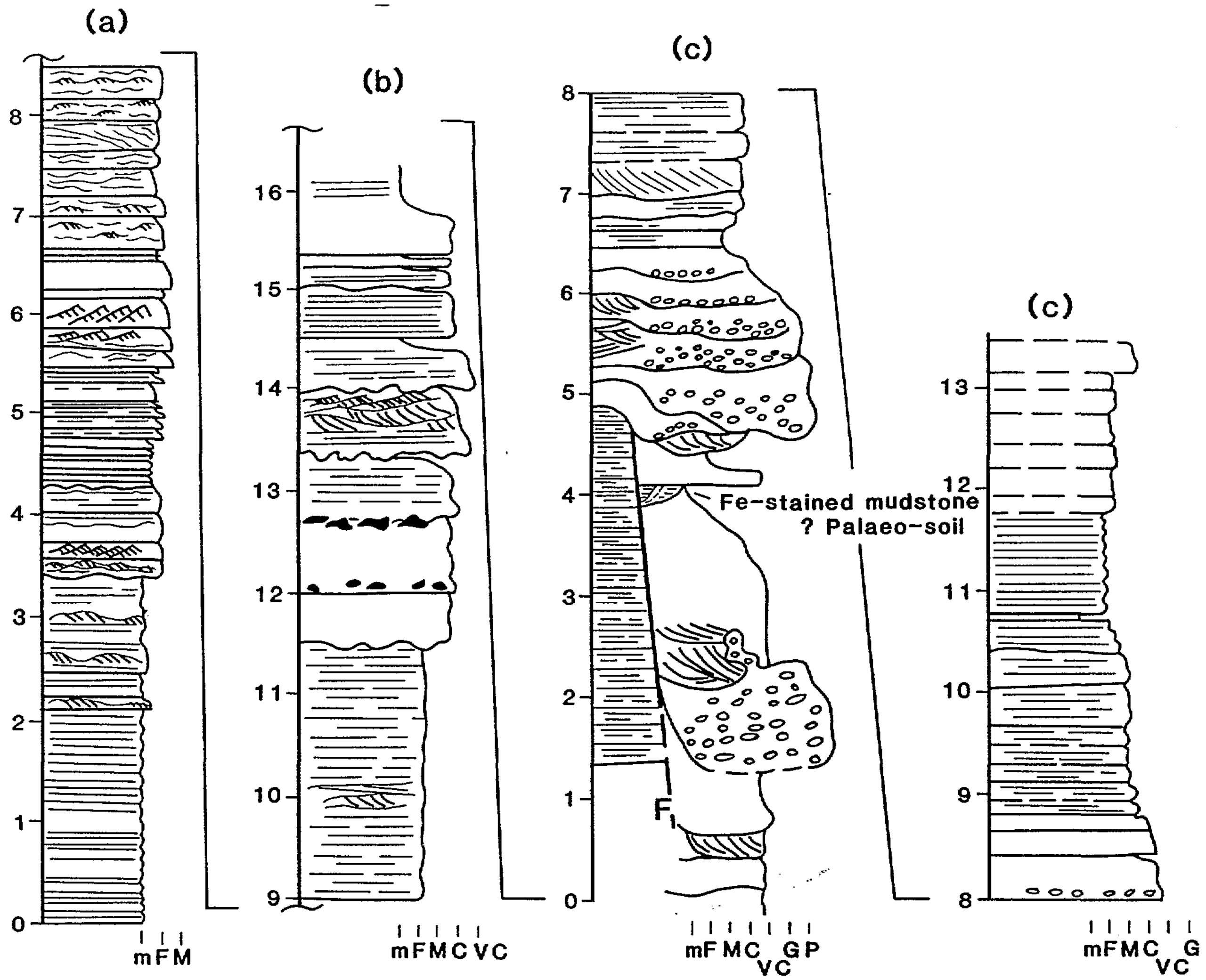


Figure 20.1 *Repesentative measured sections through the southern basin margin facies assocation, extending from Picnic Point north.*

palaeoenvironmental significance as the corresponding facies in the Bunga Beach succession.

Local, thin black mudstone beds and laminae occur in the grey mudstone, as does one horizon of mottled, iron stained mudstone (=? fossil soil), and an interval of several beds of tabular sandy mudstone, thought to represent slurry flow deposits (Fig. 20.1).

The sedimentary succession from Picnic Point north shows an initial upward fining trend, defined by the change from the basal conglomerate to the grey mudstone sequence (Fig. 20.1), and indicating initial rapid transgression. This is followed by an upwards coarsening succession of mudstones, tabular siltstones and sandstones, and following a break in outcrop across a long beach, by an interval of cross-bedded pebbly sandstone 10m thick, marked by abundant vein quartz pebbles (Fig. 20.1). This succession suggests that progradation of a coarse clastic wedge from the basin margin occurred, forming either a coarse delta or fan-delta. Above this is a subtle upward fining succession (Fig. 20.1), again representing deepening. To the north near Bengunnu Point black shales associated with rhyolites suggest rapid lateral transition into the deeper water, basin centre facies association. Measured palaeocurrent directions in the southern basin margin succession indicate flow directions to the east, the opposite to flow directions in the northern preserved basin margin succession along Bunga Beach. This succession could thus represent the opposite basin margin to that seen in the north although the origianl three dimensional orientation of the basin is unknown.

STOP 22 : SPHERULITIC RHYOLITE, NORTH OF PICNIC POINT

If time permits walk further north from the cross bedded conglomeratic sandstones to outcrops of the southern most rhyolite. This rhyolite preserves well developed spherulites, often aligned in trains, presumably superimposed on an original cryptic flow layering.

DAY 4 : BIMODAL BASIN CENTRE MAGMATISM AND SEDIMENTATION: SYN-DEPOSITIONAL INTRUSIVES, PEPERITES AND RHYOLITIC TUFF CONES, BUNGA BEDS, BOYD VOLCANIC COMPLEX

Drive from Merimbula through Tathra along the coast road north as for yesterday. 5km north of the Wapengo Lake turnoff, there is a hard right turnoff into the Mimosa Rocks National Park at the top of a crest. Follow this dirt track to the coast. Park at the first car park overlooking the northern end of Aragunnu Beach or follow the track north, and take the first track to the right to a headland-camping area.

The program today will focus on the characteristics of the basin centre facies association, and by considering some of the associated volcanics the influence that environment may have had on eruptive style. The basin centre facies association outcrops between the northern and southern basin margin facies association areas (Fig. 12.1) and is the facies association most frequently associated with the rhyolites. It is therefore significant for determining the palaeoenvironmental setting for volcanism.

The basin centre facies association is distinctive from the basin margin facies association in that the mudstones are black, the coarser clastics are classical subaqueous mass flow facies (e.g. Walker 1984) and there is no evidence for any tractional reworking. The association is best preserved just south of Bunga Head in an extensive platform exposure. It is openly to tightly folded, faulted, locally cleaved and intruded by domes and dykes of rhyolite and basalt. Elsewhere this association occurs as small areas of exposure between rhyolite bodies. Outcrops are either bedded or structureless and pervasively mixed by fluidisation processes as discussed below.

STOP 23 : BUNGA HEAD RHYOLITE: OUTFLOW AND MAGMA MIXING?

From the small headland find a track to the rock platform and boulder beach to the north. Walk north across platform exposures of basin centre facies association outcrops, across a boulder beach, across a necked headland of rhyolite, across another boulder beach to outcrops of rhyolite. Climb over the next small headland to a point where a vertical rhyolite dyke cuts across the more massive rhyolite. On Figure 12.1 this is shown as the southern side of Bunga Head.

The rhyolite crossed represents the southern end of the Bunga Head rhyolite whose northern margin was seen at STOP 22.

Of the four highly porphyritic rhyolite bodies in the north, the Bunga Head rhyolite is the thickest and most extensive. The body is a single unit at least 180m thick and 1300m long with subvertical columnar jointing, diffuse subhorizontal flow-layering throughout, and a homogeneous coarsely porphyritic texture. Phenocrysts are euhedral and the groundmass consists of a fine grained devitrified mosaic of quartz and feldspar. Small fluidal-shaped basaltic inclusions are scattered through the rhyolite in some areas, especially at the southern margin of the unit. The basal contact of the rhyolite body at this southern margin is not exposed. It is thought to be subhorizontal, parallel to flow-layering in the rhyolite, and probably concordant with underlying sediments of the deep-water basin centre facies association.

The dyke is finer than the enclosing rhyolite and has vertical flow layering. Whether it is genetically related to the main rhyolite or is a later event is not clear. Walk back over the main rhyolite to the south. Enclaves of irregular weathered basalt clasts become more abundant. The highly irregular form of many suggests that they were fluidal when incorporated into the rhyolite, suggesting magma mixing. In thin section quench needle textures are visible. On the south side of the necked headland, the diversity of clasts increases enormously.

The fluidal basalt inclusions may either occur as dispersed clasts or in "swarms" of clasts. The swarms are up to a metre or more in maximum dimension and basalt clasts constitute 50% or more of the rhyolite locally. Individual basalt clasts are generally up to 5cm long but occasionally 10cm. Petrographically the basalt clasts have a needle-like quench texture, with very fine crystallites of a pseudomorhed mafic mineral (?pyroxene) $\pm$ an iron oxide phase and plagioclase needles. The margins of the basalt inclusions are highly irregular, and occasional basalt clasts contain xenocrysts of volcanic quartz identical to the quartz phenocrysts in the rhyolite. All of these characteristics indicate that the basalt inclusions were still fluidal and hot at the time they were incorporated into the rhyolite, presumably in the magma chamber or in the magma conduit to the surface.

Walking back along the boulder beach, at low tide rock stacks in the bay can be seen. the northern ones are rhyolite, but the southern ones are bedded sedimentary. These beds are folded but the northern most appear to dip under the rhyolite.

STOP 24 : BASIN CENTRE SEDIMENTATION, BUNGA COVE

Walk back to the rock platform where the best exposures of the basin centre facies association occur. Work over these towards the south. Several distinctive facies are recognisable (Figs 24.1, 24.2).

The distinctive black mudstones are massive to laminated, occur in intervals 5m or more thick, are variably pyritic and are interpreted as hemipelagic muds deposited in a deep, quiet, anoxic environment. The laminae are pale silty layers, sometimes graded and represent small distal turbidites.

Variably graded, massive to laminated sandstones-pebbly sandstones are also prominent (Fig. 24.2). They are up to 2m thick, are sometimes amalgamated, and have sharp bases with classical turbidite sole structures such as load casts, flame structures and ball and pillow structures. Reverse graded bases also occur. The beds are usually massive, diffusely layered or laminated internally. They are interbedded with black mudstones. This facies represents proximal turbidites deposited from poorly expanded, high concentration turbidity currents, and most beds consist of Bouma a, ae or abe divisions (Bouma 1962, Walker 1984).

The sandstones are moderately to poorly sorted litharenites, consisting of vein quartz (49%), metasedimentary lithics (quartz sandstones, minor slate and chert-25%), and granite rock fragments, derived from the metasedimentary and granitic regional basement. Black mudstone intraclasts also occur and were deformed by soft sediment deformation and subsequent diagentic compaction. The sandstones are texturally immature to submature.

Crudely layered, clast supported intraclast breccias consisting of mudstone and siltstone intraclasts in a matrix of the above sandstone type represent a coarser equivalent of the previous facies with a high intraclast content. A highly concentrated turbidity current or debris avalanche origin is likely.

Much of the sedimentary succession south of Bunga Head has been subjected to soft sediment deformation and down slope slumping forming intervals of chaotic slump facies. This has been caused by slope instability, perhaps in part at least due to contemporaneous intrusion of rhyolitic and basaltic cryptodomes and basaltic dykes and sills (c.f. Kokelaar et al. 1985). Facies include essentially in situ coherent slumps of bedded intervals of the previous facies, and slide deposits consisting of disupted bedded clasts up to several metres long. Intervals of slumps and slides are up to 12m thick. Slump folds and ramp structures indicate a palaeoslope dipping to the east to east-southeast.

The last facies of this association is dominated by basaltic debris and is an open framework pebbly, basaltic sandstone facies (Fig. 24.2) occurring in internally massive, tabular beds up to 6.6m thick. The basalt clasts vary from highly fluidal bomb shapes to scoriaceous fragments with highly irregular delicate margins. Intraclasts of sandstone, mudstone as well as wood fragments also occur. The muddy sandstone matrix consists of scoriaceous basalt fragments, basaltic shards, plagioclase fragments, and minor basement derived grains. The basalt grains are subangular and have cuspate margins which are truncated vesicles. The grains contain aligned plagioclase crystals in a variably chlorite-carbonate altered microlitic vesicular groundmass, vesicles being filled with chlorite±carbonate. The fine part of the matrix includes a very minor component of volcanic quartz and more felsic shards.

This facies is represented by a group of six depositional units in the sediment succession south of Bunga Head, and is the only indication of the eruption of basaltic magmas in the Bunga Beds, since all other occurrences of basalt are intrusive bodies (see below). The scoriaceous clasts and shards indicate magmatic explosive eruptions or phreatomagmatic explosive eruptions of highly vesiculated magma. The six units are interpreted as debris flows that formed contemporaneously with eruption and formation of a small scoria cone or tuff ring within or at the margins of the basin. The concentration of these beds in one part of the stratigraphy may represent collapse of part of the small cone and resedimentation by debris flows into deeper water.

STOP 25 : BIMODAL SYN-DEPOSITIONAL RHYOLITE-BASALT INTRUSIONS, BUNGA COVE

In the southern half of the rock platform and cliff exposures of locality 24 irregular bodies of basalt occur through the basin centre facies succession. Basalt occurs in a number of localities and has a variety of forms and origins in the Bunga Beds. Lavas are not known, the most common occurrence of basalt being as sills, dykes and irregular intrusive pods. These commonly have marginal fragmental zones. In addition basalt is represented by beds of vesicular basaltic debris, and as inclusions in rhyolite. Volumetrically however, basalt is only a very minor component of the Bunga Beds.

They vary in form and include tabular to irregular dykes and sills, irregular branching networks of dykes, and small domal pods. The basaltic intrusives generally occur in close proximity to rhyolites which are themselves intrusive.

Petrographically, the basalts vary texturally from almost aphanitic with tiny cryptic plagioclase(?) laths to a coarse doleritic texture dominated by large (1-2mm) sericite(-quartz) pseudomorphed plagioclase laths. These two textural types can occur in the same intrusive mass in an irregularly intermingled fashion, suggesting perhaps different pulses of magma. Pyroxene crystals are notably lacking suggesting that the true composition may be basaltic andesite.

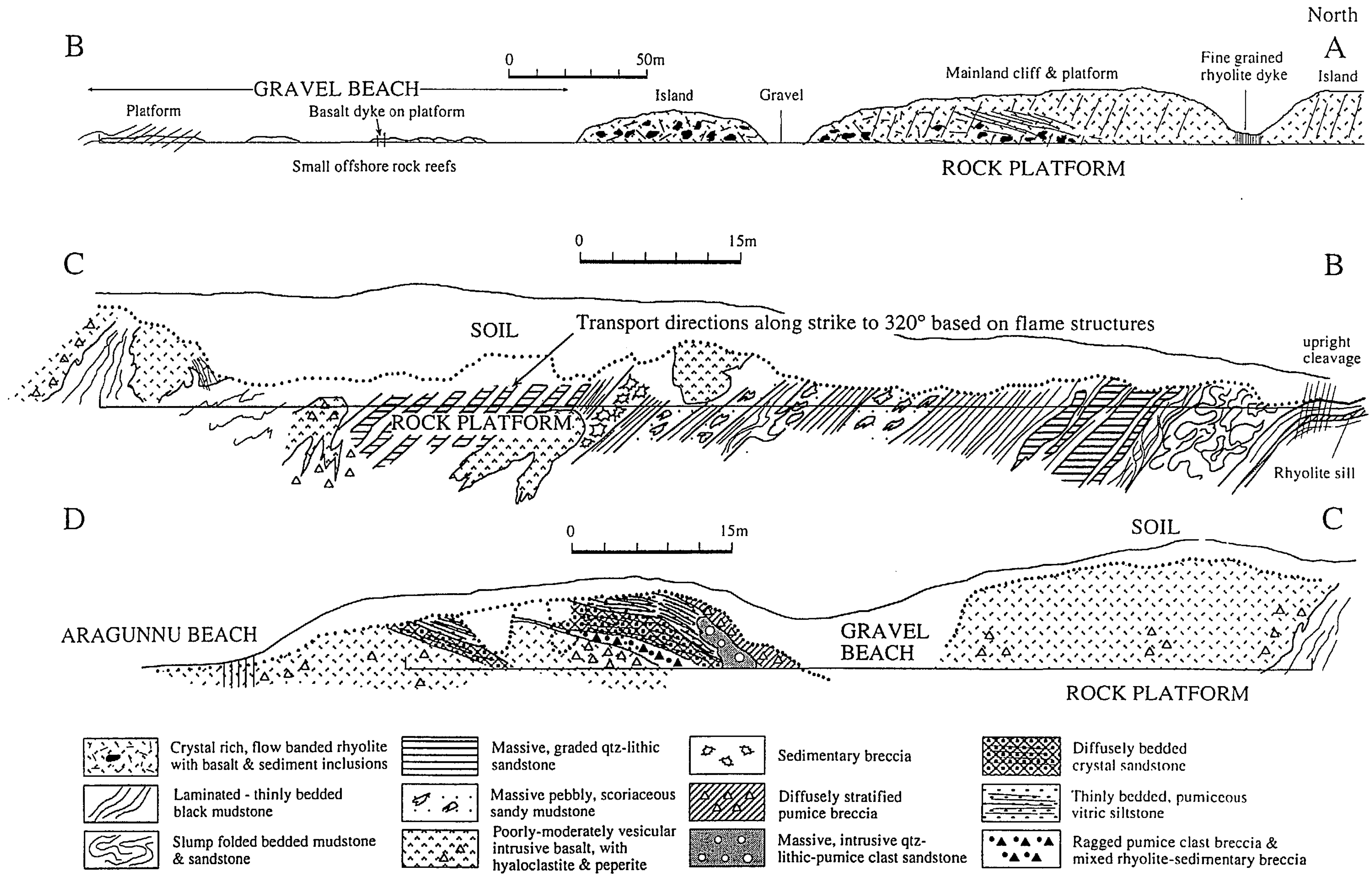


Figure 24.1 *Cross section of the basin centre association, STOP 24.*

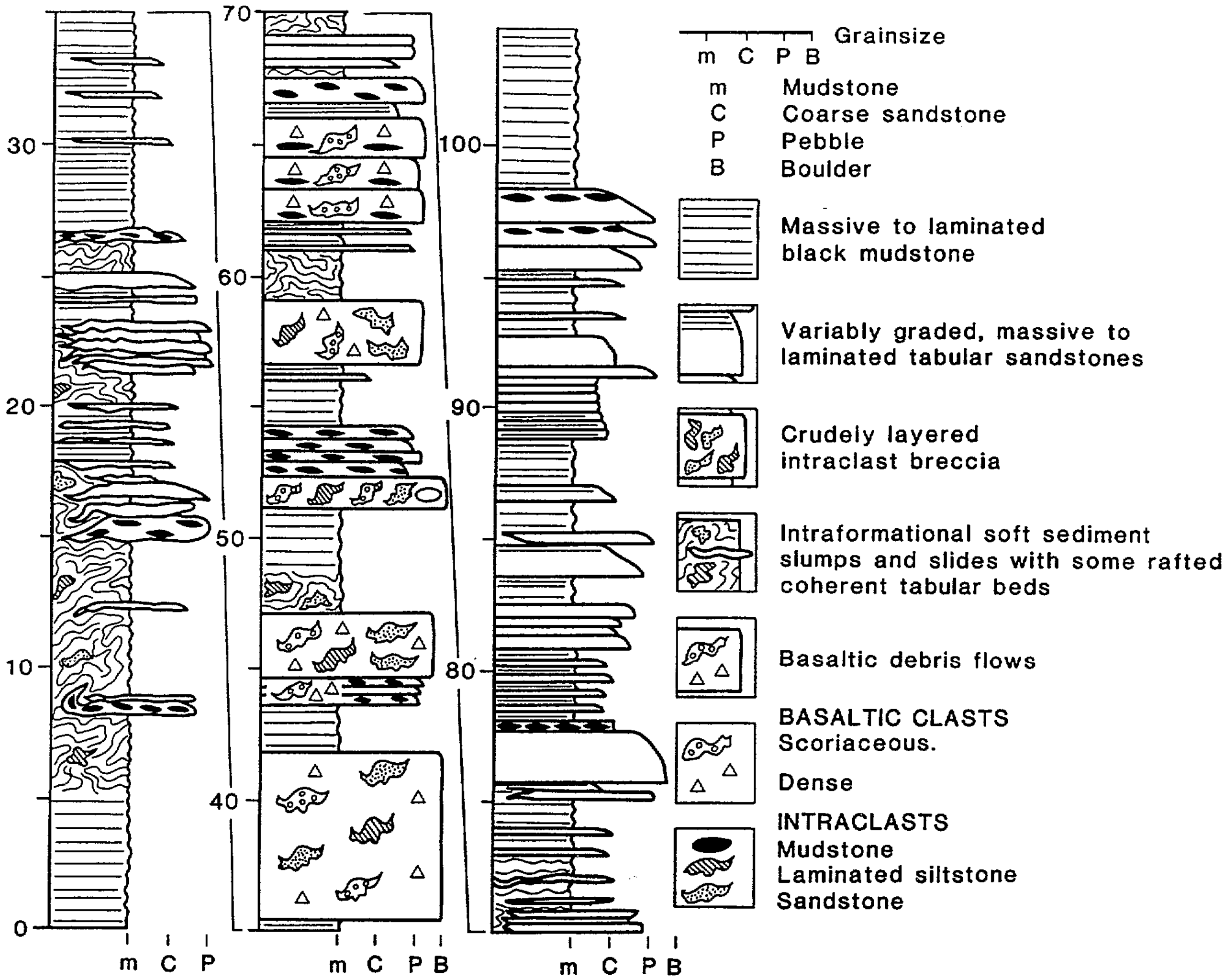


Figure 24.2 *Measured section of basin centre facies association, STOP 24.*

The basalts are commonly vesicular, with 10-15% vesicles which are filled with quartz and chlorite, commonly with a colloform layering and structure. The quartz appears to be a replacement/recrystallisation product after chalcedonic and opaline quartz. The groundmass surrounding the vesicles is very fine grained and indicates local chilling at the time of vesiculation and before the coarser crystallisation of the rest of the groundmass.

Dykes and sills are rarely more than 1m wide, and often thicknesses vary markedly along a single intrusive, reflecting the irregular margins of many of the the basaltic intrusives. This irregularity is often marked by flame-like apophyses of basalt intruding the enclosing sedimentary succession with complementary tongues and lobes of sediment. Irregular pods and cryptodomal masses of basalt are seldom more than 5-10m in diameter, and their margins are discordant with the enclosing bedding.

The margins of some intrusives are brecciated and grade from in situ incipiently fractured basalt to slightly dispersed jig-saw breccia to breccia with rotated clasts of basalt in a sedimentary matrix, similar to some of the margins of the rhyolites. The basalt clasts are blocky to splintery in shape, they have planar to curviplanar margins and are up to 10 centimetres in size.

Where a matrix is present it consists of the enclosing sedimentary succession which appears to have been injected into fractures in the surrounding basalt. Finer clasts of basalt also occur in the matrix. The sedimentary rock in contact with the the basalt is often structureless compared with the bedded nature away from the contact.

These interactive relationships with the enclosing sedimentary succession, suggest that the sedimentary succession was unconsolidated at the time of intrusion. The breccia is interpreted to be hyaloclastite resulting from the quench shattering of basalt intruding into wet, unconsolidated sediment. Fluidisation induced mixing of hyaloclastite clasts and water saturated sediment thus appears to have occurred, forming typical peperite textures.

In the cliff at the southern end of the cross section (Fig. 24.1) relatively simple subvertical discordant contacts between two bodies of flow-layered rhyolite and black mudstone of the deep-water basin centre facies association occur. The contacts are irregular, locally show flow-layering terminating outward against mudstone, and at one point a long flame-like structure of mudstone penetrates 0.5m into a fracture in the rhyolite surface of the northern body. The margin of the rhyolite displays development of fine polygonal cooling joints perpendicular to the contact, is extensively in situ quench fractured, and possibly hydraulically brecciated as well. These contacts are interpreted as part of an original irregular lobate intrusive margin between rhyolite and wet sediments. The southern body appears to be pervasuvely quench and hydraulically fractured.

Evidence that the enclosing sedimentary succession was unconsolidated at the time that both rhyolites and basalts were intruded into the basin sedimentary succession, and that basaltic inclusions in the Bunga Head rhyolite were fluidal indicate that rhyolitic and basaltic magmatism were contemporaneous with each other and with sedimentation. To what degree magma mixing sensu stricto occurred has not been established in this study, but it is clear that contemporaneous bimodal magmatism was a fundamental part of the active history of the Bunga Beds sedimentary basin.

STOP 26 : RHYOLITIC DOME-TOP TUFF CONES: A MODEL, ARAGUNNU HEAD-LAND

From the two intrusive rhyolite bodies walk across the small boulder beach to the next small cliff exposure. This exposure contains a diversity of facies and some interesting relationships (Fig. 26.1). It is the northern end of the headland at the northern end of Aragunnu Beach (Fig. 12.1).

This succession is part of a more complex stratified volcaniclastic succession that dips northeast, and has been interpreted to have been the product of a dome top tuff cone succession (Cas et al. 1990). The base of the succession lies 50m further south.

The base of the succession (Fig. 26.2) consists of a rhyolite with coherent intervals and pods, autobreccias, extensive quench fragmented zones, and at the top, stringers of mudstone and sandstone matrix filling in spaces between in situ rhyolite blocks. This passes up into a 6m thick mixed rhyolite-sediment megabreccia with rotated, flow-banded angular blocks of rhyolite up to 4m in diameter at the base, whereas at the top, rhyolite clasts are generally up to 20cm long, but up to 1m and mudstone intraclasts are up to 2m. The matrix is a finer silt-rhyolite fragment mixture.

This facies passes gradationally, but rapidly, into a massive, non-graded to graded rhyolite pumice breccia with pumice clasts and minor black mudstone clasts up to 5cm long in a crystal rich matrix of volcanic quartz and feldspar as well as finer pumice shreds. Clasts are aligned parallel to bedding and the top of the unit is abruptly graded.

The next facies in the succession is a crudely stratified, coarse crystal-rich sandstone facies that is 10m thick and is conformable on the pumice breccia. Layers are 5-10cm

Figure 26.1 *Sketch of small cliff face showing wet-sediment slides, 1, and mixed rhyolite-sediment intrusion, 2. Dashed line represents trace of slide surface subsequently intruded by the intrusive body. Note large black mudstone raft in the core of the intrusive body, faint flow laminations and alignment of mudstone clasts parallel to the margin, local concentrations of mainly pumice clasts, and lobate to flame-like wet-sediment style intrusive margin, 3 = massive rhyolite pumice breccia, 4 = crudely stratified coarse grained crystal rich tuff, 5 = thinly bedded fine grained crystal poor tuff, 6 = diffusely stratified pumice breccia, 7 = beach cover.*

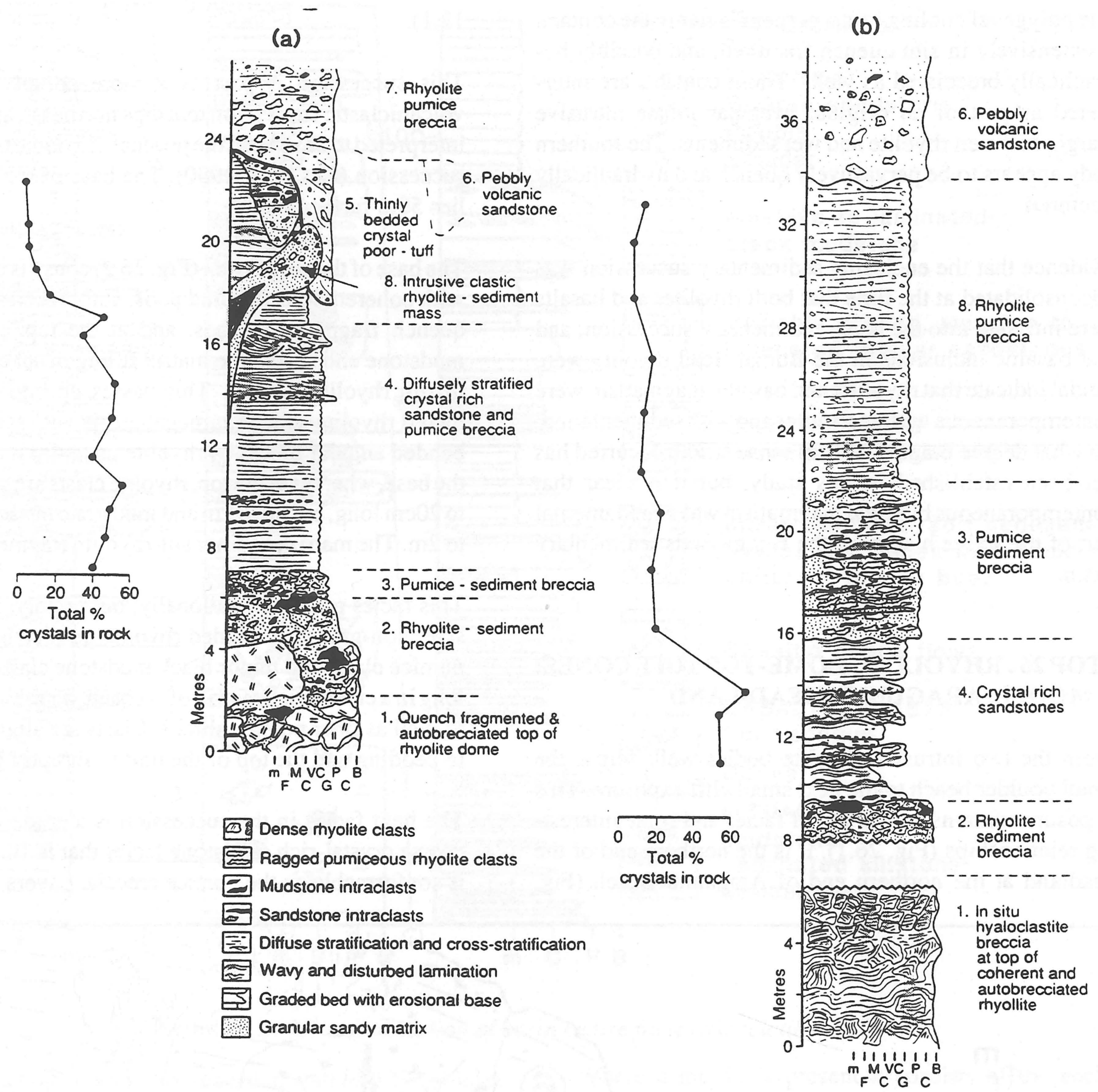


Figure 26.2 *Measured section through two dome top tuff cone successions of the Bunga Beds (a) headland north of Aragunnu Bay, STOP 26. (b) headland north of Bengunnu Point, STOP 27. Note the changes in crystal content upsequence. See text for explanation*

thick and boundaries are gradational both upwards and downwards. This unit has local laterally discontinuous layers of ragged rhyolite pumice clasts, low angle cross-beds, wavy layering and, at the top, one set of large scale cross beds consisting of both tuff and pumice breccia. The facies is composed largely of very coarse sand to granule size broken volcanic quartz and feldspar crystals (Fig. 26.3), lesser ragged pumice fragments, and isolated black silt clasts. The lack of interbedded mudstone horizons suggest rapid deposition.

Conformably above the crystal-rich facies is a 7m thick <u>thinly bedded, fine grained, crystal-poor sandstone and siltstone facies</u> that is dominated by shreds and shards of pumice (Fig. 26.4). Beds are 2-5cm thick and are mostly planar and laterally continuous, although some low angle truncations occur. Bedding varies from poorly defined to sharp, and beds may be internally massive, laminated or cross-laminated. Grainsize varies greatly from bed to bed, and some beds are crudely graded. Deposition of this facies was also rapid as indicated by the lack of interbedded mudstone.

The succession so far is interpreted to be the result of the intrusion of a rhyolite cryptodome into and through wet unconsolidated sediments, and the growth of a phreatomagmatic dome-top tuff cone (Cas et al., 1990; Fig. 26.5). This model is based on observations of the 1953-57 eruption of Tuluman Volcano in the Bismarck Sea (Reynolds and Best 1976; Reynolds et al. 1980). The

Figure 26.3 *Crystal concentration of crystal rich tuff.*

Figure 26.4 *Tube pumice shred in upper crystal poor tuffs*

upward movement of the viscous cryptodome caused autobrecciation, and contact with wet sediments caused quench fragmentation. As the dome approached the sediment surface it caused it to arch up, and as it emerged through the sediment surface, sliding and slumping of sediment and blocks of rhyolite occurred down the steepening slopes of the dome. The pumice breccia appears to represent the onset of explosive activity or spalling of the vesiculating top of the dome and simultaneous resedimentation. Explosive activity would have been initiated as the dome grew to moderately shallow water depths where confining pressures were low enough, and as water gained access to the interior of the dome through its fragmenting margins.

The crystal-rich and crystal-poor facies are interpreted to represent efficient magmatic and probably phreatomagmatic explosive eruptions, with the separation into a lower crystal-rich and an upper crystal-poor zone being effected by sorting in the water column. During active eruption the dense crystal fragments would have settled quickly, whereas the fine pumiceous ash and shards would have remained suspended for some time in the turbulent water column. The characteristics of the crystal-rich facies indicate sedimentation directly from the water column through continuous settling (diffuse stratification) and limited resedimentation through high concentration grainflows (massive layers). In addition the settling crystal aggregates would also have been affected and transported by turbulent surges caused by subaqueous blasts, so depositing the low angle cross-bedded and wavy bedded horizons. The crystal-rich facies is thus interpreted to be a water settled crystal tuff and resedimented equivalents.

The fine, crystal-poor facies appears to represent the waning and cessation of explosive eruptions. As the subaqueous eruption column and associated turbulence waned (cf. Fiske and Matsuda 1964), finer material would have settled and also would have been resedimented by

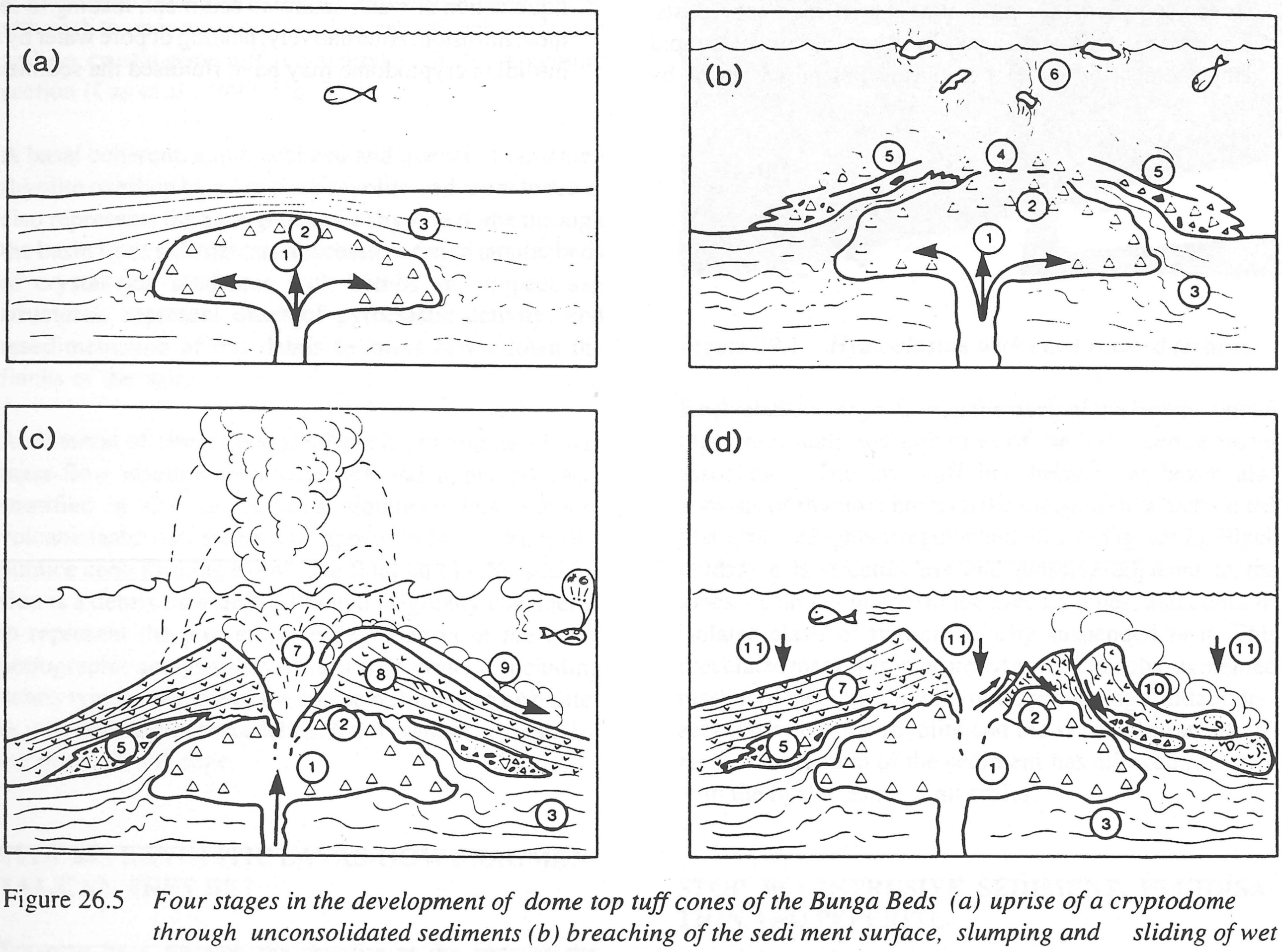


Figure 26.5 *Four stages in the development of dome top tuff cones of the Bunga Beds (a) uprise of a cryptodome through unconsolidated sediments (b) breaching of the sedi ment surface, slumping and sliding of wet sediment and autobreccia and hyaloclastite clasts (c) onset of phreatomagmatic pyroclastic activity, with hydraulic settling of dense crysta fragments, whilst pumice and shards remain suspended un til eruption ceases (d) collapse of cone producing slides (from Cas et al. 1990)*

turbidity currents, producing the succession of thin, variably graded fine beds, which are thus crystal-poor vitric tuffs.
The rhyolite and rhyolite-sediment megabreccia are thus related to the emergence of a rhyolite dome through the basin floor sediment sequence. The pumice breccia, crystal tuffs and crystal poor tuffs are the relics of a small tuff cone that developed on the dome and its margins as the dome grew to relatively shallow water depths.

Above this autoclastic and pyroclastic succession lies a lower channel-form, diffusely stratified and an upper well stratified pebbly volcaniclastic facies. Both facies contain abundant rhyolite and laminated mudstone clasts and together are nearly 4m thick. The base of the lower unit is erosional into the underlying fine crystal-poor tuffs and its top contains a marked concentration of large black mudstone clasts as well as a hummocky surface with up to 30cm relief. It is interpreted as a debris flow deposit. The upper stratified pebbly sandstone appears to have been deposited by small volume mass-flows. Both units are considered to represent the post-eruptive epiclastic degradation of the tuff cone succession.

The uppermost stratigraphic unit in this stratified volcaniclastic succession is a faintly stratified rhyolite pumice breccia that is nearly 10m thick. Pumice clasts are up to 2-3cm large, are ragged in form and make up a close packed framework. The matrix consists of finer pumice shreds, sparse crystals and scattered small mudstone clasts. The homogenous, almost massive character indicate rapid emplacement by either pyroclastic fall processes or by mass-flow. It is likely that this pumice deposit represents renewed pyroclastic activity from this centre or from an overlapping one.

However, the pumice unit is separated from the underlying succession by a sharp locally concordant to discordant surface, interpreted as a slide plane, that cuts down up to 6m into the underlying crystal-rich tuff. Another small discordant surface and normal faults are associated with this major slide surface, as well as soft sediment deformation in the bedding adjacent to the slide surface.

A discordant, intrusive pebbly rhyolite-sediment mass up to 6m by 3m in cross-section, has intruded up into the stratified volcaniclastic sequence along the major slide plane. It consists of pumiceous rhyolitic debris from the underlying stratified succession, as well as grains and soft sediment deformed clasts of sandstone and black mudstone derived from the underlying basin centre, turbidite-black mudstone facies association. The texture of this intrusive mass is similar to the fluidised sediment adjacent to rhyolite cryptodome contacts.

The close relationship between the intrusive mass and the slide system is significant and is interpreted to reflect renewed emplacement of rhyolite cryptodomes in the subsurface, causing updoming and slip failure of a sector of the tuff cone. Sudden movement and rapid loading by the slide blocks may have caused overpressuring and liquefaction of water saturated sediment, leading to sediment intrusion. Alternatively, heating of pore water by the intruding cryptodome may have fluidised the sediment.

DAY 5: SUBMARINE RHYOLITE LAVAS, INTRUSIVES, HYALOCLASTITES AND TUFF CONES, BUNGA BEDS, BOYD VOLCANIC COMPLEX

Drive from Merimbula to the car park overlooking the northern end of Aragunnu Bay, as for yesterday. Descend down stairs and traverse along the coastline to the headland north of Bengunnu with the clearly recognisable syncline. Parts of the coastline are very rugged and considerable care must be taken. It is best to make this traverse only around either side of low tide. Then progressively work back to Aragunnu Bay.

STOP 27 : RHYOLITIC DOME-TOP TUFF CONES: VARIATIONS ON A THEME, HEADLAND NORTH OF BENGUNNU POINT

A similar succession but with some variations to that seen at the northern end of Aragunnu Bay occurs at this locality, and a similar interpretation is applied.

On the headland north of Bengunnu Point (Fig. 12.1) there is a succession generally similar to that described above, although there is more pumice, less crystal-rich sandstone and no crystal-poor tuff, compared with the Aragunnu section (Cas et al., 1990; Fig. 26.2).

A basal coherent, autobrecciated and quench fragmented rhyolite overlain by a 1m thick rhyolite-sediment breccia, also represents the emergence of a rhyolite dome through the basin floor sedimentary succession. 6m of tabular beds of crystal-rich sandstone with bombs and impact sag structures, represent onset of pyroclastic activity, and resedimentation of this debris by mass flows down the flanks of the dome.

An interval of 19m of pumice deposits, including a lower mass-flow resedimented sequence and upper diffusely stratified in situ fall deposits, dominate this southern volcaniclastic succession, and appear to be the relics of a pumice cone forming event. The final unit in this succession is a debris flow unit. Although originally considered to represent the post-eruptive degradation of the cone, petrographic analysis shows that it is polymictic, including debris typical of some of the coarse basin centre turbidites as well as basaltic scoria debris. It thus cannot be related to the rhyolitic tuff cone.

STOP 28 : RHYOLITIC LAVAS-HOW FRAGMENTAL CAN THEY BE?

Traverse back through the rhyolite at the base of the stratified succession and note the proportion of the flow that is fragmental. Initial outcrops appear to be autobrecciated flow banded lava breccias. The rhyolite then consists of varying coherent and finer brecciated zones, the latter being particularly evident at the outcrops at the southern end of the small beach to the north.

STOP 29 : RHYOLITIC HYALOCLASTITE AND PEPERITE

The northern most exposure of the rhyolite that underlies the tuff cone succession is well exposed at the southern end of the small beach to the north and consists of a rhyolite breccia with clast rotated textures interpreted as hyaloclastite (Fig. 29.1).

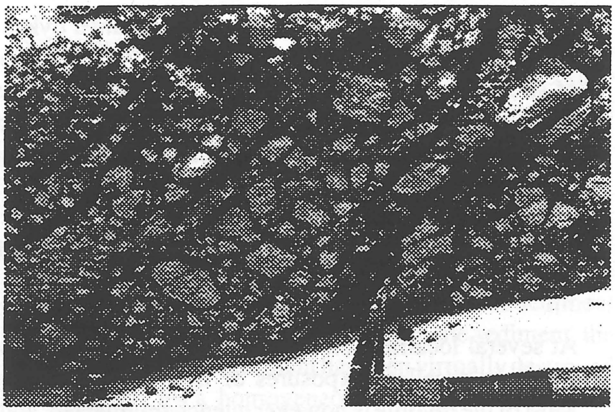


Figure 29.1 *Hyaloclastite with clast rotated textures*

Rock platform exposures at the back of the beach contain black mudstone and turbidites of the basin centre facies association. The low cliff line behind the beach also consists of rhyolitic breccia, the margins of which on the platform are highly irregular and lobate (Fig. 29.2). Black mudstone is structureless and massive adjacent to the lobes, occurs as matrix in the breccia lobes, and contains isolated clasts of rhyolite wholly suspended in it. This brecciated margin is interpreted as the quench fragmented margin of and intruding rhyolite, producing fluidal interaction between the rhyolite and the water saturated sediment. Fluidisation of the sediment has allowed it to mix with the hyaloclastite at all scales.

STOP 30 : INTRUSIVE SEDIMENT, FLUIDISATION AND PEPERITE

Follow this next rhyolite north around the next headland. Again note the relative proportions of coherent and

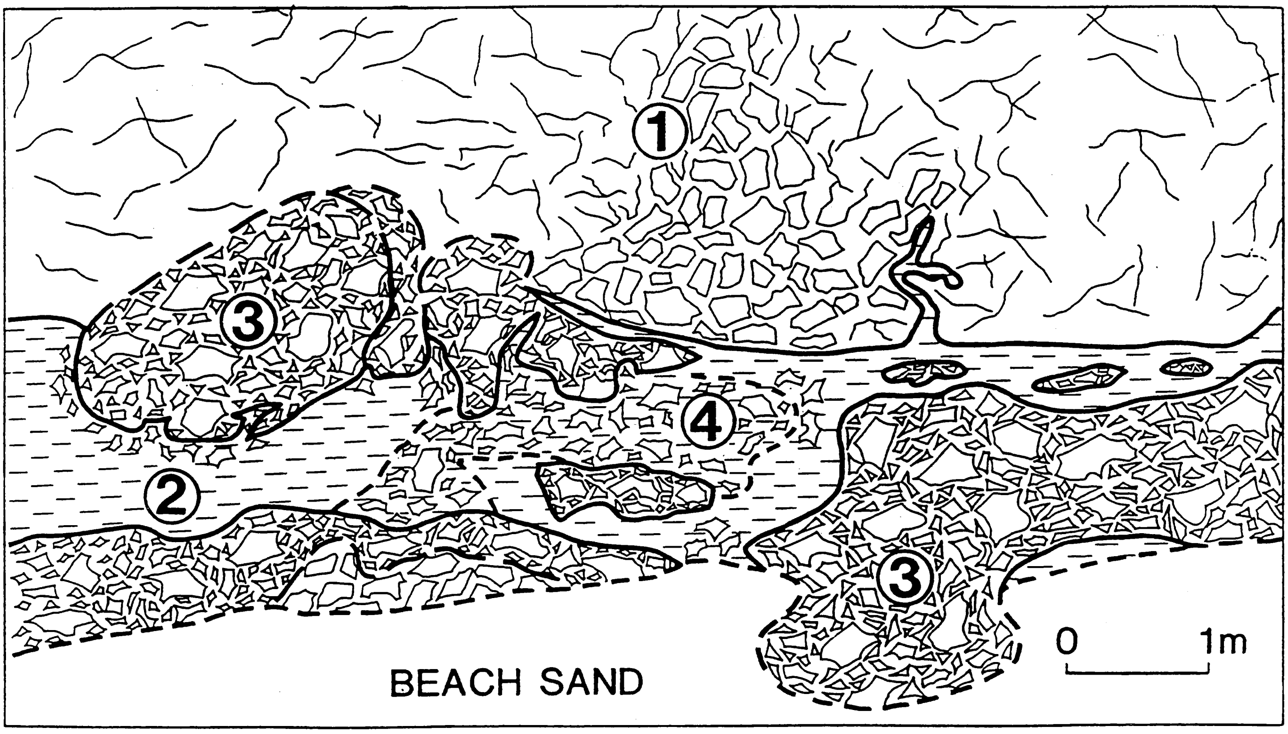


Figure 29.2 *Sketch of relationships between lobate brecciated margin of intrusive rhyolite and fluidised black mudstone*

brecciated rhyolite. En route you will encounter dykes of sediment in the rhyolite, and at one locality ther is a zone of structureless mudstone with dispersed rhyolite clasts through it, again testifying to dynamic interaction between rhyolite and the wet unconsolidated sediment through which it rose.

STOP 31 : FLOW FOLDS AND MORE HYALOCLASTITE

At several localities on the way back to Aragunnu Bay there are excellent exposures of flow banded rhyolite. Most of the rhyolite is however quench fragmented, and in numerous localities has sedimentary inclusions.

Discussion: Rhyolites of the southern half of the Bunga Beds

The moderately porphyritic rhyolite which dominates the southern half of the area invariably forms more complex bodies than the crystal-rich rhyolites to the north. Coherent portions are intimately associated with, and grade into, massive rhyolite breccia of similar composition. Distribution of coherent material within the rhyolite units, and rhyolite-sediment contact relationships discussed below, suggest that the coherent rhyolite mainly forms the cores of complex lava and intrusive bodies.

Coherent moderately porphyritic rhyolite usually has a fine spherulitic flow-layering which may vary widely in orientation and is locally flow folded. An alignment of lenticular quartz-filled vesicles also defines flow-layering in one area and vesicularities of up to 20% have been recorded. Columnar jointing is not widespread. Perlitic fracture textures are common in the groundmass, indicating that much of the rock was originally glassy.

Autobreccias or flow breccias also occur in the massive moderately porphyritic rhyolite bodies. These breccias are clast supported frameworks of irregular, angular rhyolite fragments which vary in size from 10-20 cms up to several metres. Flow layering is common in the fragments and highlights two modes of brecciation. In the first, the continuity of flow layering through the breccia indicates that in situ, mainly extensional fragmentation, with little or no rotation of clasts occurred. In the second, brecciation involving significant rotation of fragments occurred, implying the influence of a flow induced shear stress.

Irregular, soft state deformed, black mudstone clasts are particularly common within autobreccia at the margins of the rhyolite bodies. On the south side of Aragunnu Bay some isolated mudstone clasts also occur in autobreccia, apparently well within a rhyolite unit, indicating incorporation of sediment into the rhyolites as they were emplaced in poorly consolidated wet sediment.

Coherent, autobrecciated, and mildly fractured moderately porphyritic rhyolite in many cases grades outward through a zone of in situ brecciation to an intensely granulated marginal breccia both of which are interpreted to be hyaloclastite breccia. This marginal breccia is in turn either in contact with sediments or is overlain by stratified rhyolitic volcaniclastics (see above). Thickness of the in situ breccia varies widely from a few metres to at least 100m and is generally greater than that of the intensely granulated marginal breccia. The marginal zone is up to 30m thick at one locality at the northern end of Aragunnu Bay, but elsewhere appears to be no more than 10m thick. Variations in this pattern occur and at a few locations coherent, autobrecciated and in situ brecciated rhyolite are in direct contact with sediments with no intervening intensely granulated zone.

The in situ breccias show a general trend of decreasing clast size, and hence increasing intensity of fragmentation, outward. Well inside the margins of rhyolite bodies (10 to 50m) the breccia comprises slightly dislocated, roughly polygonal blocks of rhyolite averaging 15cm in size. Close to the margins the breccia forms a dense in situ mosaic of highly angular, polygonal, generally non-rotated, 2cm sized fragments. These breccias are strongly altered and the fracture system that defines them is commonly highlighted as a fine black network due to abundant fine grained pyrite in the fracture filling.

In situ breccia fragments are composed of dense, poorly to nonvesiculated rhyolite. Fragment boundaries are most commonly planar to broadly arcuate, but a smaller number of finely serated boundaries are also found. Planar to broadly arcuate fragment boundaries is a distinctive feature of this type of breccia. Within the fragments themselves a strong development of finely arcuate perlitic fractures is common and these can locally be detected in cut slabs or even in outcrop.

At the northern end of Aragunnu Bay lobate pods of relatively coherent rhyolite 2-3m in diameter are set within a matrix of strongly fragmented material of identical composition. The pods have brecciated margins which grade outward into the matrix. Continuous outcrop in this area shows that these pods form a 20m thick transition zone between coherent flow-layered rhyolite and more uniformly brecciated material.

The intensely granulated marginal breccias appear as granular, poorly sorted, strongly pyrophyllite-quartz altered, rhyolitic rubble. In detail they comprise scattered irregular shaped patches of strongly in situ-brecciated rhyolite set in a more intensely fragmented and disorganized rhyolitic matrix. Matrix fragments vary from angular blocky to sharply irregular shapes and are mostly less than 2cm in size. Many of these fragments are composed of altered glassy rhyolite. However, some of the irregular shaped ones may be pumiceous.

The in situ breccia and marginal breccia described above are both interpreted to be rhyolitic hyaloclastite. Features which characterize these breccias in the Bunga Beds are : intimate and gradational relationship with coherent or autobrecciated subaqueous rhyolite lava, general trend of increasing intensity of brecciation outward, widespread occurence of intense in situ fracturing with little or no rotation of clasts, pervasive and sharply angular nature of the brecciation, and the abundant evidence for an original glassy composition of the breccia fragments. These features are consistent with an origin of quench fragmentation at the margins of subaqueous lava, and are all common to other well documented rhyolitic hyaloclastites (Pichler 1965; DeRosen-Spence et al. 1980; Furnes et al. 1980; Yamagishi and Dimroth, 1985; Yamagishi 1987, 1991).

The in situ breccia is regarded as hyaloclastite formed by the propagation of a network of thermal contraction fractures deep into the glassy rhyolite. Intense quenching at the surface of the lava has produced the finer granular marginal breccia. This marginal breccia is therefore also essentially formed in situ. However, all the hyaloclastite material, and in particular the marginal breccias, appear to have suffered local rotation of clasts and further granulation due to continued emplacement and expansion of the lava, confirming Pichler's (1965) conclusion that hyaloclastite largely forms by the combination of quench fracturing and autobrecciation. Quenching may also be accompanied by small scale, local steam explosions at sufficiently shallow water depths, although we have no evidence of this here.

The rhyolite pods set in fragmental matrix, described from the northern end of Aragunnu Bay, may simply be isolated relic coherent patches within otherwise strongly fragmented material. Alternatively, they may be broken dislocated lava lobes, extruded into and enveloped by, their own or earlier disintegration products (cf. De Rosen-Spence et al., 1980; Furnes et al. 1980; Yamagishi 1987, 1991).

Rhyolite-sediment contact relation-ships of the moderately porphyritic rhyolite bodies : implications for emplacement.

In contrast to the passive, and simple transgressive contacts of the crystal-rich bodies, contacts of the moderately porphyritic rhyolite in the southern half of the area all record disruptive interaction between lava and sediment.

At STOP 30 contact relationships are more complex. Quench fragmented blocks of rhyolite occur dispersed in sediment adjacent to contacts, and dykes of sediment have intruded several metres into the rhyolite. In the sediment dykes, and where rhyolite clasts occur in the sediment, the original fabric of the sediment has been virtually destroyed and replaced by a homogeneous fabric of sand grains scattered evenly in a black mud matrix. Minor fluidal soft state deformed black mudstone clasts are the only relic of the original lithologies. The bedded turbidites and mudstone of the basin centre facies association have therefore been completely mixed to a homogeneous slurry while in a wet and poorly consolidated state. Furthermore, the sediment mixtures have been injected as dykes into fractures in the rhyolite, and have intruded outward into undisturbed sediment transporting quenched rhyolite fragments in the process. These features are consistent with the fluidization of the wet sediments by the hot rhyolite, as described by Kokelaar (1982) elsewhere.

Emplacement of the hot rhyolite into wet sediments would have heated pore water, adjacent to contacts with the sediments. Turbulent circulation of this heated water (perhaps locally steam), and continued intrusion of the rhyolite into the fluidized carapace probably mobilized the fluidized mixtures of water, steam, sediment, and rhyolite clasts into fractures and away from the rhyolite body. At the fluidized intrusive contacts described here from STOP 30, rhyolite mainly overlies and laterally abuts gently dipping sediments indicating the contacts are at the base or

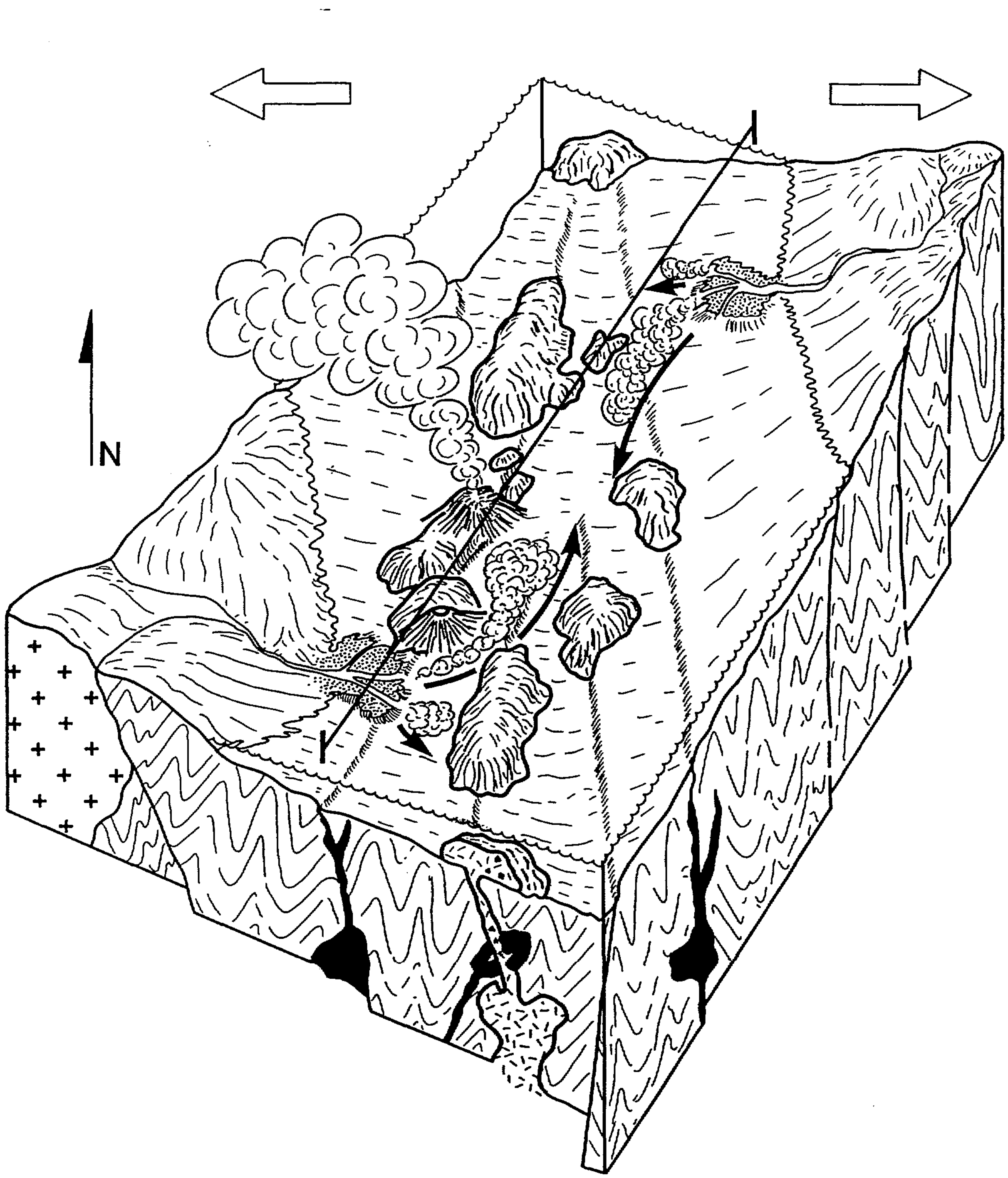


Figure 31.1 *Schematic reconstruction of the Bunga Beds basin and the depositional and volcanic processes.*

margin of an at least partly intrusive rhyolite body.

The most complex mixed contact occurs at STOP 29 between sediments and granular hyaloclastite breccia at the margin of a rhyolite body (Fig. 29.2). Here the contact is a zone several metres wide consisting of numerous irregular patches of hyaloclastite breccia dispersed in reconstituted black silty sediment. In addition, one area of black mudstone hosts a number of isolated pods and blocks of coherent massive rhyolite up to 2m in size. One pod (approximatley circular in cross section and 1m in diameter) has a moderately hackly to cauliflower-like crenulate margin, with some sections of this margin having radial cracks which pass a few centimetres into the pod. The coherent pod therefore appears to have a chilled crust, and may be a blocky, more viscous version of a pillow structure.

In summary, the disruptive contacts described above all document the intrusion of rhyolite lava into wet poorly consolidated sediments. These include contacts where sediment underlies, overlies or laterally abuts rhyolite. Together with the occurence of mixed rhyolite-sediment breccias described below, these contacts indicate that the majority of the rhyolites is intrusive and, therefore, the moderately porphyritic rhyolite bodies are rhyolite cryptodomes.

Summary of the depositional environment in which volcanism occurred

The depositional environment into which magma was erupted or intruded was a quiet, restricted small basin, or an embayment of a larger basin which is no longer preserved. At the margins, which are preserved at the northern and southern extremities of the outcrop belt, the environment was relatively shallow, as shown by the cross-bedded sandstone facies, the grey ?oxidized mudstones and local intense bioturbation. The environments at the basin margins probably consisted of deltaic and foreslope deltaic environments (Fig. 31.1). The deeper middle part of the basin was anoxic, and below the influence of reworking agents as indicated by the black pyritic shales and turbidite facies associations. Although the preserved fish fossils (Fergusson et al., 1979) appear to be marine, no other marine fossils have been found and so a brackish water or even a deep lacustrine basin are possible.

Although the magmatic activity was clearly bimodal the preserved volcanic succession is wholly rhyolitic, dominated by domes and cryptodomes. Basaltic intrusives occur together with voluminous rhyolitic intrusives. Volumetrically rhyolites and autoclastics are considerably more abundant than in situ pyroclastics, which appear to be confined to small intrabasinal dome top tuff cone successions. The only other pyroclastic debris occurs in massflow resedimented facies which indicate that both basaltic and rhyolitic eruptive centres were active around the basin margins.

An interesting aspect of the rhyolites is that the majority appear to have been shallow syn-depositional intrusives. The question that arises is whether the scarcity of in situ, intrabasinal pyroclastics was due to water depths great enough to suppress explosive eruptions or were the magmas simply low in volatile content? A quantitative answer or assessment of this is not possible because of the absence of any quantitative depth indicator. However, even if the magma volatile content was low, superheating of external pore water or basin water should have produced at least phreatic explosions if the confining pressure and water depths were low enough.

Finally, in comparing the continental parts of the Boyd Volcanic Complex with the subaqueous parts, it is clear that there are significant differences in both the volcanic and sedimentary facies that can be attributed to differences in the environment.

DAY 6: EARLY DEVONIAN SNOWY RIVER VOLCANICS: DISTINGUISHING PRIMARY PYROCLASTICS FROM SECONDARY REDEPOSITED SEDIMENTS OF PYROCLASTIC DEBRIS IN AN ANCIENT CALDERA LAKE SUCCESSION

DIRECTIONS

Proceed south along the Princes Highway (Highway 1) from Merimbula through Eden and into Victoria (Fig. 6). Turn off Highway 1 northwards to Buchan approximately 240 km to the southwest at the township of Nowa Nowa (Fig. 6). Turn right at Nowa Nowa onto the Nowa Nowa/ Bruthen Road, and right again 7 km later onto the Buchan/ Gelantipy Road. Buchan is 27 km north of this intersection. The road cutting to be examined is on the Buchan/ Gelantipy road cutting at W-Tree, a further 15 km north of Buchan.

INTRODUCTION

A Lower Devonian sequence comprising the Snowy River Volcanics (SRV) and the overlying Buchan Group carbonates accumulated within a prominent tectonic feature called the Buchan Rift. The Buchan Rift is one of numerous basin (trough) features which originated as fault-bounded extensional rift/graben features in the Lachlan Fold Belt during the Siluro-Devonian (Cas, 1983). The SRV, which form the bulk of the fill of the Buchan Rift, outcrop in an extensive north-south trending belt parallel to, and mainly to the west of, the Snowy River and centred on the township of Buchan in eastern Victoria (Fig. 32.1). The present day margins of the SRV are thought to coincide, at least in part, with original margins of the Buchan Rift (VandenBerg, 1988; Bull, 1992).

The SRV and Buchan Group carbonate succession in the Buchan Rift unconformably overlies Ordovician and Silurian clastic and volcanic sequences and associated mainly Silurian granites (VandenBerg, 1988; Fig. 32.1). The Lower Devonian sequence was deformed during the Middle Devonian Tabberabberan Orogeny and regional metamorphism reached lower greenschist facies (Talent, 1965; Orth, 1982; VandenBerg, 1988). The deformation produced a series of large, open, north-south trending folds which plunge shallowly to both the north and south. The axis of the main synclinal structure, the Murrindal Syncline, is marked by the largest preserved remnant of the Buchan Group carbonates which occur around the Buchan township (Fig. 32.1). The area where we will examine the SRV, Buchan/Gelantipy Road cutting at W-Tree, occurs at the northern end of this carbonate unit in the axis of the Murrindal Syncline.

Overall, the SRV are a largely silicic but bimodal volcanic suite which has been described as a stratigraphically complex sequence of rhyodacite and rhyolite ignimbrites with lesser amounts of rhyolite, rhyodacite and andesite lava, basalt lava and a great variety of epiclastic, pyroclastic and non-volcanic sediments (VandenBerg, 1988). In detail, the SRV record the depositional development of the

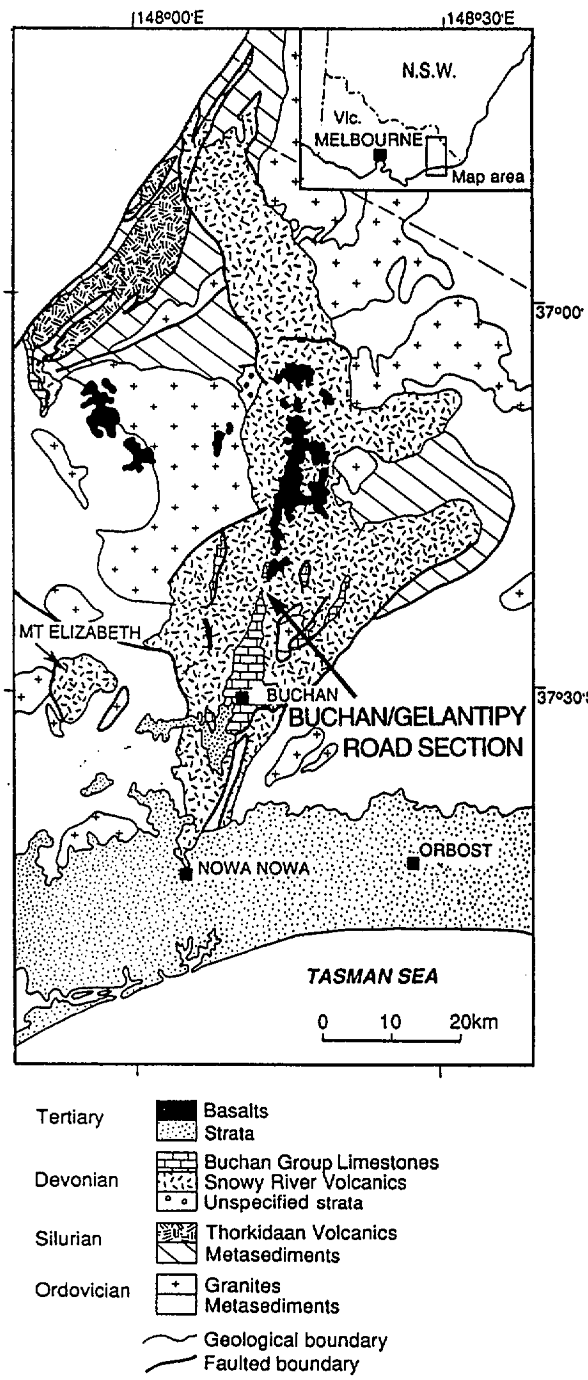


Figure 32.1 *Geological setting of the Lower Devonian Snowy River Volcanics*

deepest (?axial) graben of the Buchan Rift system (Bull, 1992). Basal deposits in the southwestern part of the belt consist of initially non-volcanic, basement-derived alluvial fan deposits which occur within palaeovalleys incised into the hanging wall of intra-rift normal faults. These units are succeeded by volcanic-dominated deposits which accumulated in different micro-environments within the developing graben system. These environments included both subaerial and relatively deep, quiet marine conditions early in the history of the graben (Bull and Cas, 1991; Bull, 1992), however, subaerial conditions prevailed during the accumulation of the upper part of the SRV succession. These upper deposits are dominated overall by welded ignimbrite deposits, and it is part of this succession which will be examined in the Buchan/Gelantipy Road cutting at W-Tree.

The contact between the SRV and the overlying Buchan Group is complex and variable, but is gradational on a regional scale and the Buchan Group is interpreted to have been deposited in an extensive shallow sea following the cessation of volcanism and gradual, regional scale subsidence (Orth, 1982; VandenBerg, 1988). The Buchan Group carbonates are now preserved at three major and at least thirteen minor localities within the volcanic belt, most of which are structural depressions (Gaskin, 1943; VandenBerg, 1988).

The Buchan/Gelantipy road cutting at W-Tree, north of Murrindal, exposes a 1.2 km long section which is part of a volcano-sedimentary succession informally referred to as the W-Tree Rhyolite sequence (WTRS; Bull, 1992). The WTRS is an extensive succession (ie. > 40 km^2) of coherent and fragmental silicic volcanic and volcaniclastic sedimentary rocks which occurs high in the stratigraphy of the SRV, intercalated with the uppermost few ignimbrite units of the SRV succession (Poxon, 1979; Orth, 1982; Orth et al., 1989; Bull, 1992). The road cutting exposes part of a distinctive central zone within the WTRS. This central zone differs markedly from the rest of the stratigraphy at this high level in the SRV, in that coherent rhyolitic porphyries and volcaniclastic sediments are well represented volumetrically relative to pyroclastic deposits. The overall facies association in this area is similar to that occurring in many modern-day silicic volcanic centres (eg. Cas and Wright; 1987), suggesting that the sequence may represent a volcanic centre within the SRV (Orth et al., 1989; Bull, 1992).

STRATIGRAPHY

In addition to the mapping by Bull (1992), the area encompassing the WTRS has also recently been mapped as part of a regional stratigraphic study of the SRV undertaken by the Geological Survey of Victoria to produce a Murrindal 1:100,000 scale geological map (Orth et al., in prep.). In this study the WTRS was not differentiated, however, it occurs within a unit termed the Fairy Sandstone Formation which is part of the uppermost stratigraphic unit defined within the SRV, Little River Sub-Group, (Orth et al., in prep.). In detail, the WTRS has been subdivided into three volcanic units termed the Lower, Middle and Upper W-Tree Ignimbrite, which are separated by intervals of clastic tuffaceous debris of variable affinity which are collectively termed the W-Tree Clastics (Bull, 1992). The Buchan/Gelantipy road cutting at W-Tree exposes the W-Tree Clastics sequence where it occurs between the Upper W-Tree Ignimbrite at the base of the section and the main body of lava present in the area, the Raymond Falls Rhyolite at the top of the section (Fig. 32.2). A detailed examination of this exposure reveals a complex interplay between volcanic processes including pyroclastic flow and base surge and lava effusion, and sedimentary processes including high-energy, ephemeral fluvio-deltaic deposition and quiet water turbidite and pelagic deposition in an intra-caldera environment. In addition, since the sedimentary deposits consist almost entirely of volcaniclastic material, this sequence offers an opportunity to determine criteria for distinguishing primary volcanic and volcaniclastic sediment deposits in ancient successions.

STOP 32

Directions: Proceed north from Buchan along the Buchan/Gelantipy Road. Approximately 15 km north of Buchan there is a bridge over the Murrindal River. Park in the carpark on the eastern side of the road south of the bridge. Walk approximately 500 m north up the Buchan/Gelantipy Road to a quarry on the right hand side of the road which is STOP 32.

The lowest part of the stratigraphy dissected by the Buchan/Gelantipy road cutting at W-Tree is exposed in and around the quarry at STOP 32 (Fig. 32.2). The Upper W-Tree Ignimbrite, which forms the base of the sequence in this area, is poorly-exposed at the southern end of the quarry in a low cutting on the eastern side of the bend in the road. However, the features of this unit are best observed in boulders piled on the slope at the western edge of the road. The Upper W-Tree Ignimbrite is overlain by a number of crudely bedded massive ash units which are poorly exposed in the erosion gully at the northern edge of the quarry (note: these units are better exposed in a road cutting approximately 100 m north of the quarry; Fig. 32.2, 140-150 m). These bedded units are interpreted to represent mass-flow deposits of ash which were the result of gravitational stabilization of the depositional surface immediately after the emplacement of the underlying ignimbrite. They appear to have been subaqueously deposited, possibly in a caldera lake. The crudely bedded mass flow units are successively overlain by an interval of poorly-sorted, well-bedded units interpreted as base surge deposits. These units are exposed in the main bench of the quarry approximately 25 m above the level of the road (Fig. 32.2). Typical base surge features which can be observed are overall poor sorting and the alternation of thin laminated units, some of which display low-angle sigmoidal cross-stratification, with massive accretionary lapilli-rich ashes, which are thought to be co-surge fall deposits (Walker 1984). The surge succession appears to represent the remains of a phreatomagmatic tuff ring or

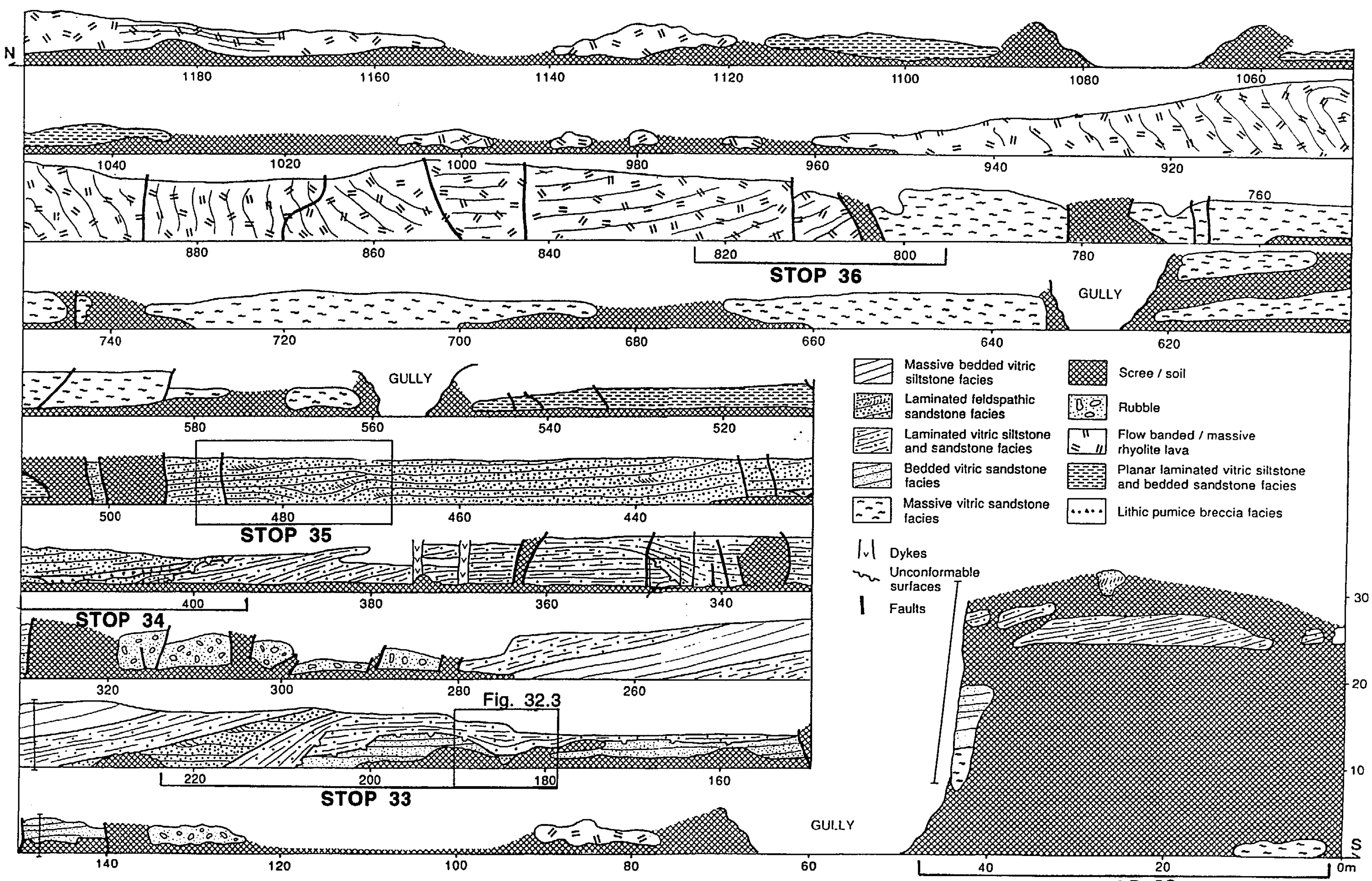


Figure 32.2 *Field sketch of the Buchan Gelantipy Road section at W-Tree*

cone. Rhyolite boulders along strike appear to represent the remains of a small intrusive mass.

STOP 33

Directions: STOP 33 is marked by a gully structure exposed in the cutting on the eastern side of the road approximately 180 m north of STOP32.

Due to the gentle northerly dip of the section, STOP 33 is approximately along strike from the main quarry bench at STOP 32 (Fig. 32.2). At STOP 33 it is clear that the base surge deposits are infilling an undulose erosion surface in the underlying crudely bedded mass flow deposits. The main erosive feature present is a U-shaped gully which has a partially obscured V-shaped basal portion (Fig. 33.1). This feature is interpreted to represent a steep sided erosion gully which was modified into a U-shape due to erosion by base surges (Fisher, 1977). Twenty metres north of the U-shaped gully a planar disconformable surface occurs within the base surge sequence (note: another example of this type of feature is exposed 130 m to the north; Fig. 32.2, 340-350 m). The base surge units deposited across the planar surface show typical mantle bedding characteristics The planar disconformity is interpreted to represent a slide/slump surface resulting from minor sector collapse of part of the rapidly accumulating base surge succession. The presence of this type of feature indicates that the accumulating base surge succession had significant topographic relief, and formed some form of phreatomagmatic cone or rampart. Ten metres north of the slide surface flat-lying stratified volcaniclastic sandstone deposits occur which truncate against the northerly-dipping upper surface of the underlying base surge deposits. These units are well-sorted relative to the base surge deposits, and medium scale cross-bed sets with high-angle foresets are present (note: the features of this facies are better exposed at STOP 34). They are interpreted to represent fluvial reworking of the base surge deposits (see below).

STOP 34

Directions: STOP 34 is marked by an undulose erosion surface exposed in the cutting on the eastern side of the road approximately 200 m north of STOP 33.

At STOP 34 the upper surface of the base surge dominated part of the succession is exposed. It is marked by an undulose, gullied horizon indicative of a degree of erosion of the base surge sequence. The bulk of the material which infills the eroded surface is crudely stratified, relatively well-sorted volcaniclastic sandstone with scattered medium-scale cross-bed sets with high-angle foresets. This is

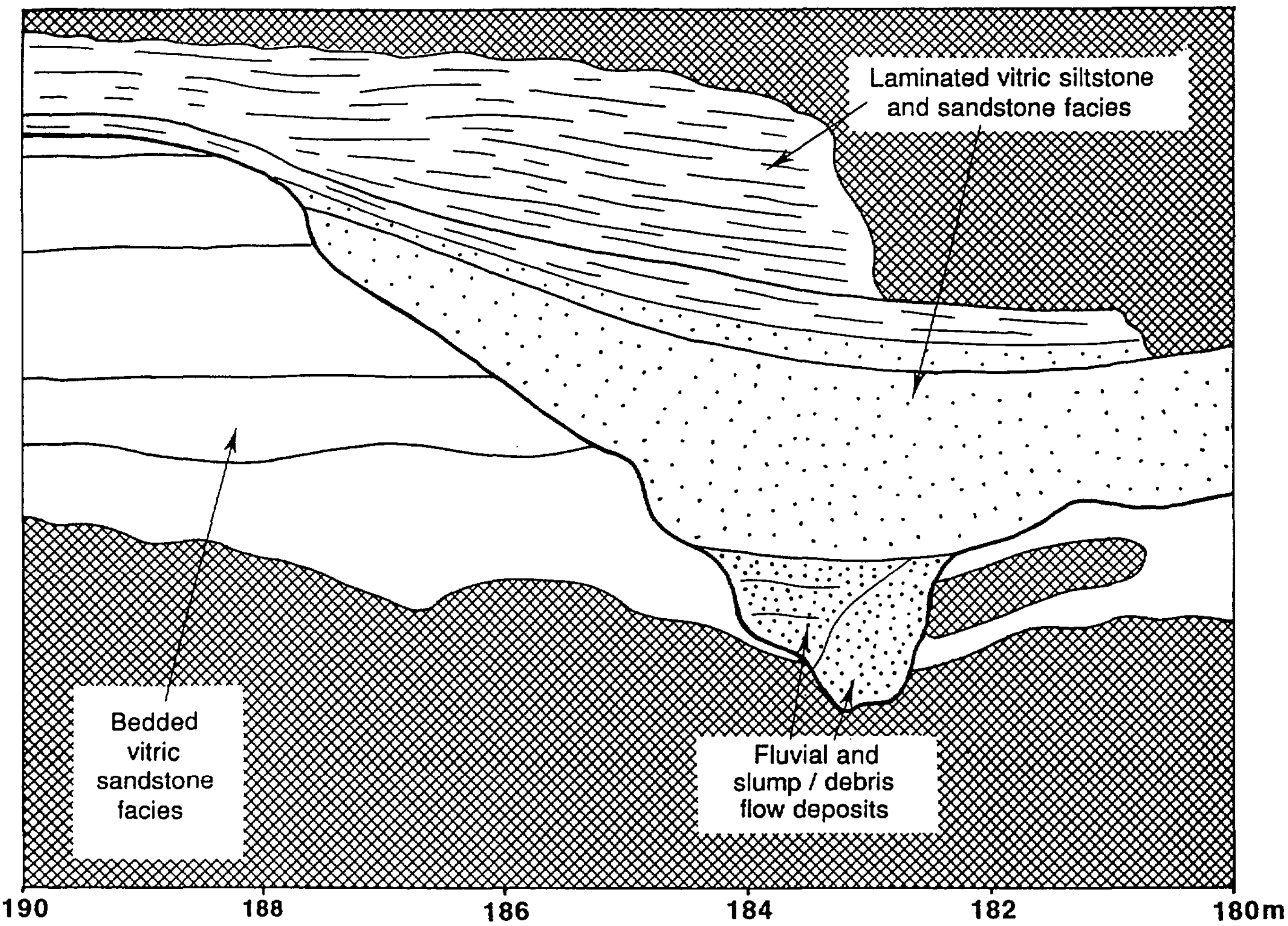


Figure 33.1 *Field sketch of a u-shaped erosion gully in the base surge facies at STOP 33.*

the same facies which occurs associated with the base surge deposits at STOP 33, however, the ubiquitous crude stratification is better exposed. It can be seen to consist of planar/wavy low-angle truncating lamination/low-angle ripples. Taken overall, this assemblage of tractional sedimentary structures is indicative of upper flow regime depositional conditions, probably at the dune to upper plane bed transition (Harms et al., 1982; Dam and Andreasen, 1990). This indicates that the fluvial system in which the base surge deposits were reworked was relatively high-energy, and probably ephemeral in nature. A prominent, metre thick, massive breccia unit low in the fluvial succession has a valley-fill geometry indicating that it is some form of flow deposit. It could represent a thin pyroclastic flow or an epiclastic mass flow of volcaniclastic material.

STOP 35

Directions: STOP 35 is marked by a 10 m wide interval of downwarped bedding exposed in the cutting on the eastern side of the road approximately 70 m north of STOP 34.

At STOP 35 bedding in the fluvial deposits is downwarped into a U-shape so that the resultant feature superficially resembles the U-shaped gully developed in the base surge deposits at STOP 33. However, a medium-scale cross-bed developed within the structure indicates an upslope palaeoflow direction. It is attributed to syn-sedimentary faulting in the underlying succession, possibly further sector collapse of the underlying base surge rampart.

STOP 36

To the north of STOP 35 volcanic sediment succession becomes faulted and passes into a feldspathic ignimbrite with accretionary lapilli especially towards the top. Its northern margin is in fault contact with a flow banded rhyolite (W-Tree Rhyolite), the base of which is in contact with a laminated lacustrine sedimentary succession. The rhyolite is interpreted to represent a caldera rhyolite dome complex. A limited interval of rhyolitic tuffs or tuffaceous sediments lie to the north as an inlier in the rhyolite. Further north, the rhyolite becomes progressively more hydrothermally altered, and outcrops immediately before the W-Tree Falls show intense silica-pyrite alteration.

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