Chapter three

The geomorphic development of volcanic landforms

SYNOPSIS: Terrestrial volcanism alters the landscape through a number of processes, including lava and tephra accretion, regional scale landscape alteration (through magma emplacement), local erosion, local damming of drainage, changed depositional patterns, vegetation destruction and the formation of new substrates. All of these induced palaeogeographic changes are potentially recognisable in the stratigraphic and geomorphic record. Thus, volcanoes enable researchers to place a time-scale on landscape evolution, both on the volcano itself and on surrounding features. This chapter briefly outlines the mechanisms of volcanism and types of volcanic activity. It concentrates on the physical expressions of activity, describing the morphology of constructional topography and the effects that geomorphic agents have on these landforms.

3.1 Volcanic landforms: introduction

As the Earth is tectonically dynamic, it is not unexpected that there is a rich diversity of volcanic landforms, and that volcanism itself has had a significant influence on the nature of the landscape. Variations in volcanic landforms depend largely on the bulk chemistry of the related parent magmas, the nature of the feeders that convey the magmas to the surface, and the modifying presence or absence of the atmosphere and hydrosphere (Cattermole 1989). More than 500 volcanoes have been active in historical time (for example, Mauna Loa, Mt. Fuji, Mt. St. Helens, to name but a few) and thousands of volcanoes may be recognised by their form or structure. Because the preservation potential of these features is so great, they have an ability to yield information that can resolve a range of geological and geomorphological problems. It is possible to determine absolute rates of landscape modification, to model pre- and post-volcanic environments and to interpret landscape (and lithosphere) evolution through the integration of volcanigenic data (for example, pyroclastic deposits and lava flows) with geomorphological evidence (for example, the stratigraphic record). This information can be used to place volcanic episodes within a chronological framework that encompasses geological and environmental events which allow us to determine the mechanisms of regional landscape evolution. However, major problems still exist in New South Wales palaeovolcanology, including the determination of stratigraphic relationships, both internal and external, the detailed relationship between tectonism and volcanism (Chapter 2), petrology and chemistry of many suites, recognition of the type of activity, and the study of specific volcanic features (Branagan 1969). The number of detailed studies published on physical and geomorphic volcanology of Cenozoic intraplate volcanic centres and deposits in eastern Australia is increasing, but many come from the Newer Volcanics province (Victoria), largely because it is a young, well-preserved volcanic area.

While the origin of modern volcanic features may be evident, it becomes more difficult as time progresses to reconstruct the nature of volcanic events and the way in which volcanic materials were related to the geological setting into which ancient volcanoes were erupted. This is generally because most continental volcanoes form constructional features that are reasonably quickly destroyed by erosion. However, the topography displayed by a volcanic complex as a result of erosion depends on its composition and the effects of climate. Thus, even after a volcano has been deeply dissected, distinct landforms may remain, especially if there is preferential preservation of lithology and denudation rates are highly localised over space and time. In this study, it is the analysis of remnant volcanic landforms of central-type volcanism (Section 2.2.2) that is of primary concern, rather than aspects of petrogenesis and physical volcanology, although these aspects are considered within the framework established in Chapter 1.

Typically, the extent to which a regional landscape is controlled by volcanism depends on:

1. *the nature of extruded material*. In general, large dimension, low profile (1-7°) volcanoes and their associated landforms are built by mafic lavas. In contrast, volcanoes associated with more silic magmas are generally steeper (20-35°), less extensive and display landscape features produced by explosively generated materials;
2. *the duration of volcanism*. A volcano may experience a single or several phases of rapid construction during its active lifetime. In this period, the rate of construction exceeds

the rate of erosion. Once eruptive activity decreases or ceases, erosion becomes the dominant process;

3. *the distribution of vents and fissures*;
4. *the volume of outpouring*;
5. *the age(s) of volcanic activity relative to the present and to associated stratigraphic units; and*
6. *the intensity and stage of subsequent erosional activity* (Short and Blair 1986).

3.2 Classification of volcanic landforms

As alluded to in Section 1.6, there are many inconsistent classifications of volcanic landforms in the literature, with most concentrating on petrology and mechanisms of eruption. Increasingly, classification of volcanic phenomena has included the identification of volcanic landforms, emphasising geomorphic processes. Cotton's (1944) *Volcanoes as landscape forms* focused on the development of volcanic landforms by petrogenic mechanisms and modifying geomorphic processes. In general, classification will refer to taxonomy that includes elements of structure, topography and processes accompanying eruption, such as that of Bloom (1979) in which volcanic landforms were categorised around two defining parameters; *the viscosity (quality) of magma* and the q*uantity of magma* (size of landform edifice). Within these parameters Bloom's resultant geomorphic classification included:

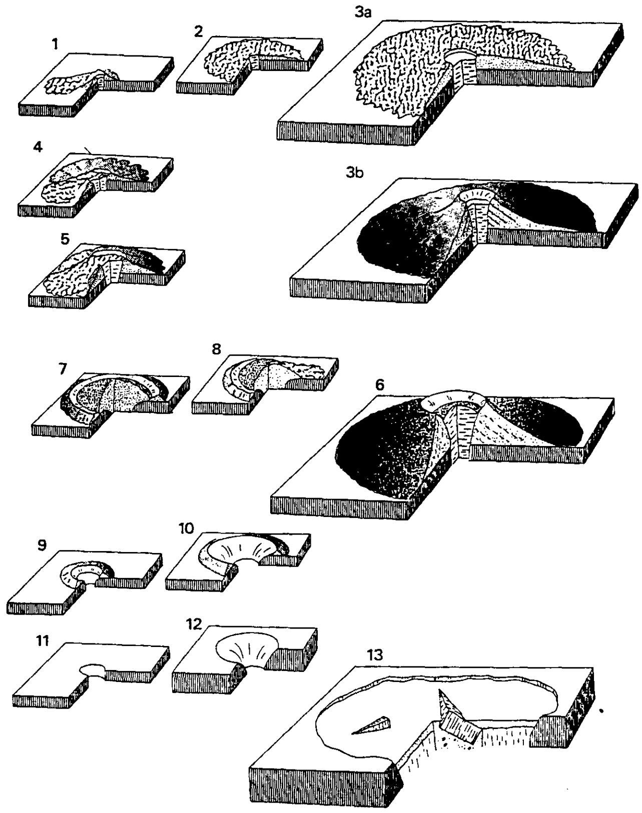
1. *chemical composition of volcanic effluence*;
2. *state of ejecta released*;
3. *history of the volcanic field*;
4. *shape/locations of the vents/fissures*;
5. *nature of the volcanic activity; and*
6. *characteristic landforms.*

Many of the larger landform features arising from volcanism can effectively be classified (Table 3.1), although volcanoes are highly variable in their morphology. Figure 3.1(a) shows how volcanic activity can produce features such as the Warrumbungle Complex. A classification of smaller volcanic landforms appears in Figure 3.1(b). The contents of these Tables and Figures are discussed in the following sections.

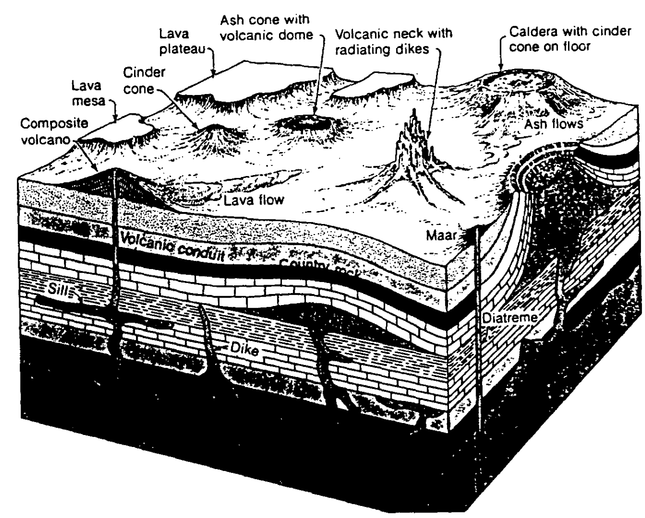
Table 3.1: Simplified classification of volcanoes and related landforms. The numbers refer to the diagrams in Figure 3.1a.

TYPE OF MAGMA	TYPE OF ACTIVITY	QUANTITY OF MAGMA PRODUCED			
		Small ⟵			⟶ Large
Fluid, very hot, basic in composition	Effusive	Lava flows[1]	Exogenous[2] domes	Basalt domes and shield volcanoes	Icelandic[3A] Hawaiian[3B]
Increasing viscosity, gas content, and acidity (high proportion of silica) ↓	↓ Mixed ↓	Scoria cones with flows[4] Loose tephra cones with thick flows[5]	Composite or strato-volcanoes[6]		Volcanic fields with multiple domes
		Endogenous domes (plug domes, tholoids)[7]	Ruptured endogenous domes with thick lava flows[8]		
Viscous, relatively cool, acidic	Explosive	Maars of tephra[9]	Maars with ramparts[10]	Collapse and explosion calderas[13]	Ignimbrite sheets
Extremely viscous, abundant crystals	Explosive, mostly gas	Gas maars[11]	Explosion craters[12]		

Source: Adapted from Rittman 1962.



(a)



(b)

Figure 3.1(a): Schematic representation of simple central-type volcanoes. The numbers refer to the volcano type classified in Table 3.1. The forms of actual volcanoes are normally more complex (as in the case of the Warrumbungle volcanism), since they usually result from a complicated history of numerous eruptive episodes. S*ource:* Summerfield 1991. (b) Schematic diagram showing the characteristic landforms resulting from volcanic action at the surface and their relationship to shallow intrusive activity in the crust. *Source:* Short and Blair 1986.

Typically, volcanoes may take one of several forms. *Monogenetic volcanoes* are the product of a single eruptive episode. Such episodes may occur over a few hours or be as long as several years. Once eruption ceases, the vent is separated from the magma source as the fissures connecting them are cooled. Landforms may take the form of scoria cones, maars, tuff cones, ring dykes and diatremes. *Polygenetic volcanoes* are those that experience more than one eruptive episode in their history. Landforms include plugs, dykes, sills, domes, multiple overlapping lavas and interbedded pyroclastic deposits (Figure 3.1b). *Volcanic complexes* such as the Warrumbungle Complex, may be defined as extensive assemblages of spatially, temporally and genetically related major and minor centres with their associated lava flows and pyroclastic rocks (Francis 1993; but see Section 1.1).

3.3 Shield volcanoes

The magma that erupts from volcanoes originates deep in the Earth and reaches the surface through fissures in the lithosphere (Section 2.2.1). When molten igneous material reaches the surface, the size, shape and nature of the volcano that is formed is largely dependent on the viscosity of the resultant magma. Viscosity is determined by several fundamental properties including temperature, composition and the amount of dissolved gas in the magma. The viscosity of magma is the major factor which determines the type of eruptive activity, the products that are erupted and the landforms that are produced.

The easiest volcano to visualise is one built up of successive flows of fluid (low viscosity) lava and/or pyroclastic material. These flows are capable of travelling great distances down gentle slopes, and of forming thin sheets of lava of nearly uniform thickness (Skinner and Porter 1987). Eventually, this material builds up a pile of lava that resembles a shield, convex side up. Such volcanoes are known as *shield volcanoes*. Terrestrial shields show a range of morphologies from the small, low profile shields of Iceland (for example, Skjaldbreid, Kollota Dyngja), through to the steeper shields of the Galapagos Islands, to the very large shield volcanoes of the Hawaiian-Emperor Seamount Chain. More massive shields are evident on Mars. On this planet, the shield of Olympus Mons is some 27 km high with a basal diameter of 500 km (Hodges and Moore 1994). Icelandic-type shields, like those of the Snake River Plains in the USA, are typically situated within regions of crustal extension. The substantially larger shields of the Galapagos and Hawaiian Islands are in intraplate settings associated with mantle hotspots (Figure 2.3).

3.3.1 Australian shield volcanoes

Intraplate shield volcanoes in eastern Australia are made up of multiple, overlapping, thin lavas and pyroclastic material from long-lived single or polygenetic (multiple) centres of eruption (Johnson 1989), whose life spans may be measured in terms of hundreds of thousands to millions of years (Table 3.2). Middlemost (1985) likened the New South Wales shield volcanoes to the Plio-Pleistocene volcanoes of the south Turkana region in the northern part of the Kenya Rift. These Turkana-type volcanoes were distinguished by Webb and Weaver (1975) as a new volcanic landform. They form low-angle (5°) trachytic shields up to 50 km in diameter, and they characteristically contain structurally complex central zones into which are concentrated the eruptive vents, usually represented by plugs, dykes and small cones. Calderas and craters are minor features (Middlemost 1985). The Focal Peak, Nandewar and Canobolas Complexes conform with this general description (Cas 1989); as does the Warrumbungle Complex, with the exception that this complex contains seven features that may be interpreted as crater remnants or crater-fills (Hockley undated; MacKellar 1980; Ollier 1967; Section 3.6.2; 4.4.3.1 and 6.4.2).

The main Miocene shield volcanoes in eastern Australia show considerable variation in size, volume, structure and types of landforms (Figure 1.1; Table 3.2). Nevertheless, there are enough morphological similarities to make useful comparisons. In general, this type of volcanism produced low-angle shields (<5°) with complex centralised vent systems. These vents are invariably characterised by clusters of domes, plugs and radial dykes which intrude into, or protrude from, a range of lithologically diverse lavas, pyroclastic flows and tephras. Typically, the flanks of the shield are well preserved interbedded lavas and pyroclasts, retaining evidence of original shield morphology. By contrast, the proximal areas were probably originally dominated by trachyte pyroclastic materials covered by a thin veneer of shield-forming lavas. The subsequent removal of this veneer, along with the bulk of the

Table 3.2: A summary of the geomorphic and petrological characteristics of the main central-type (shield) volcanic provinces of southeastern Queensland and New South Wales. See also Figure 1.1.

Volcano	Age (Ma)	General geomorphology	Rock type and distribution
Main Range (Queensland)	26-24, 24-22	This feature forms a continuous belt extending north-northwest from the New South Wales-Queensland border for 22 km parallel to, and close to, the divide. The southern Main Range is heavily dissected on the western slopes. Despite this, the extent of the Main Range shield, like Focal Peak, has been defined by radial drainage patterns (Ollier and Haworth 1994). The northern Main Range is largely plateau country with mesas and, like the southern area, is bounded in the east by the Great Escarpment. The lava pile exceeds 600 m (Stevens *et al.* 1989a). Felsic lavas and mafic intrusions dominate the southwest of the shield.	Volcanic rocks cover an area of 4900 km^2, with a present volume of about 1000 km^3. The oldest series of lavas are predominantly hawaiitic with transitional basalts to mugearite. Later rocks have an alkaline character and include nepheline, basanite, leucite basanite, hawaiite (predominant), *ne* hawaiite and *ne* benmoreite (Ewart and Grenfell 1985).
Focal Peak (Queensland)	23	This volcano is located 60 km west-northwest of the centre of the Tweed Volcano and approximately 20 km east of the Main Range. Mount Barney central complex is the eruptive centre of the Focal Peak shield volcano, within which a central caldera is thought to exist at Focal Peak. Mount Barney also corresponds to a zone of basement uplift. The shield dip is <5°. Mount Barney is characterised by a caldera marked by ring intrusions. The outer zone of Mount Barney includes sill-like or laccolith intrusions of rhyolite (Ross 1989).	The lavas and fine-grained intrusions are mostly alkaline. Hawaiite predominates over alkali basalt, mugearite, basanite, *ol*-tholeiitic basalt and icelandite (Ross 1989).
Tweed (New South Wales)	23.5-20.5	This volcano is characterised by a single distinct igneous complex (Mount Warning) which marks the centre of the volcano. This peak is isolated from the remaining shield by a deep erosion caldera which exposes the underlying Palaeozoic and Mesozoic rocks. Shield slope angles are in the order of 1-3° but the eastern side of the volcano has been removed by erosion. Eruption distances are up to 50 km and form a series of distinct formations. The lava pile reaches a maximum thickness of 900m and individual flows range from 10-150m thick (Stevens *et al.* 1989b).	Generalised volcanic distribution of a lower mafic sequence, followed by rhyolitic units, overlain by a younger mafic sequence. A separate peralkaline rhyolite dyke phase is also recognised (Stevens *et al.* 1989b).
Ebor-Dorrigo (New South Wales)	19-18	Associated lava flows are perched on the edge of an east-to-south facing portion of the coastal escarpment (Ollier 1982a). Flows estimated to range in thickness from an average of 50 m to 300 m (Duggan 1989a). Erosion is estimated to have removed 90% of volume and 60% of the original surface area (Ashley *et al.* 1995). A prominent radial drainage pattern is evident.	Four petrographic groups are present: a dominant tholeiitic basaltic lava suite, alkaline and transitional basaltic lavas, felsic masses, related fragmental and local sedimentary rock as well as the intrusive Crescent Complex (Ashley *et al.* 1995).

Table 3.2 continued: A summary of the geomorphic and petrological characteristics of the main central (shield) volcanic provinces of southeastern Queensland and New South Wales. See also Figure 1.1.

Volcano	Age (Ma)	General geomorphology	Rock type and distribution
Nandewar (New South Wales)	21-17	An older volcanic phase was emplaced onto an irregular basement of Palaeozoic sedimentary and volcanic rocks on the eastern side and Mesozoic sedimentary rocks on the western side. Younger rocks were emplaced during the main shield building phase and were restricted to 18-17 Ma (Stolz 1989). Constructional material dips at 5-10°. Ring dykes, dykes, plugs, lava flows and domes are prominent. Mafic flows are generally 2-10 m thick, trachyte flow units are thicker at 10-20 m. Massive flows outcrop continuously for 6-8 km in the south (Stolz 1989). Localised deposits of lacustrine sediments are evident on the flanks of the shield. Shield slope is 1-3°.	There are two distinct episodes of rhyolite volcanism, with older extrusions along a linear north-northwest fracture in the central area (Stolz 1983).
Warrumbungle (New South Wales)	17.4-13.7	The central portion of the shield is skeletal, characterised by plugs, domes, dykes and poorly bedded pyroclastics. The shield dips at 1-3° and gently dipping constructional lavas dip at 3-12°. Lavas and tephras form sequences up to 806 m thick (Section 6.5.2.1). Localised deposits of lacustrine sediments containing diatomite are evident on the shield flanks. Lavas flowed distances of up to 30 km from their sources.	Hawaiite and mugearite dominate the outer flanks. Xenoliths and megacrysts present. Trachyte dominates the central and western areas of the volcano (Duggan 1989b; Section 4.4.2.1).
Comboyne (New South Wales)	16	Most igneous rocks occur on the Comboyne plateau and unconformably overlie the Triassic Camden Haven Group. Alkaline plugs grouped in the Lansdowne Valley and western rim of the Lorne Basin suggest a tectonic emplacement control. Retreat of the Great Escarpment has removed much of the volcanic material (Pain and Ollier 1986). Several vents were responsible for the bulk of the material since areal extent is large and thickness is limited (Knutson 1989b).	Lavas and high-level intrusions are transitional, ranging from mildly alkaline to subalkaline. Hawaiite, tholeiitic basalt, mugearite, icelandite, benmoreite, dactite, *qz* trachyte and peralkaline rhyolite (Knutson 1989b).
Canobolas Complex (New South Wales)	12-11	Smaller than most central complexes, the central shield area is characterised by numerous felsic domes, dykes and small cones and lava flows surrounded by a thin apron of low-angle mafic lavas overlying Ordovician limestones and sediments of the Lachlan fold belt. Flows are 2-3 m thick. Diatomite is present (Middlemost 1989).	Mainly hawaiite, *qz*-tholeiitic basalt, mugearite, benmoreite, trachyte, peralkaline, rhyolite and peralkaline rhyolite. The core is dominated by trachyte, with hawaiite dominant on flanks (Middlemost 1981).

softer pyroclastic material, has yielded central zones characterised by the more resistant intrusive or extrusive lithologies preserved as domes, plugs and dykes. However, it is acknowledged that the much larger Tweed Volcano may be an exception to this, being dominated by the massive Mount Warning central intrusive complex and erosion caldera. This volcano is eastern Australia's largest intraplate shield, the centre of which is marked by the prominent intrusive complex of Mount Warning. This intrusion is isolated from the shield flanks by a deep erosion caldera. The Warrumbungle Complex differs greatly in morphology to the Tweed in that the centre of the shield and, to a lesser extent the intermediate flanks, are characterised by many intrusions marked by domes, plugs and dykes. The Warrumbungle Complex also lacks an erosion caldera. While the depth of erosion of the Warrumbungle Complex core is comparable to that of Mount Warning (Tweed Volcano) and the Crescent Complex (Ebor-Dorrigo Volcano), Solomon's (1964) surmise that Mount Warning represents a stage of volcano dissection not shown in comparable structures only appears valid in the sense that the Tweed Volcano is a much larger structure. The factors contributing to morphological variations across the Miocene shields are discussed in Chapter 8 in light of the processes which have operated on the development of the present-day Warrumbungle landscape.

3.4 Mechanisms of volcanism

This section briefly discusses the various mechanisms of volcanism as they affect the construction and modification of volcanic landscapes. The properties of molten lavas will be discussed, as well as the products of the fragmentation of magma. Obviously, the physical properties and the chemical constituency of the magma will influence the various types and phases of volcanic activity, which give rise to the considerable variety of volcanic landforms exhibited world-wide.

3.4.1 Types of volcanic activity

There are several types of volcanic activity. The classification of volcanism can be based on a number of parameters, including their activity (effusive *vs* explosive; conventional *vs* hydrovolcanic), their relationships through time and space (location and frequency), eruption size or their violence. While the classification of eruptions can vary, activity can be broadly categorised as either lava eruptions or pyroclastic eruptions. These are generally believed to depend on variations in the chemical composition of the parent magma, although this may not always be the case. Further variations in eruption type may occur through *phreatic* eruptions (the result of contact of magma with groundwater) or fissure eruptions. Tall, central vent volcanoes like Fuji (Japan) appear to have erupted solely through their summit craters over their lifetime. Other large volcanoes are more complex, with eruptions also taking place through points on their flanks (for example, Mount Etna). Such eruptions are termed flank, parasitic or satellite vents (Francis 1993).

3.4.2 Lava eruptions

Lava eruptions may include *hawaiian*, *icelandic* and *viscous* type lava flows. Hawaiian eruptions are generally non-explosive and lava-effusive. Lava fountains may spray at times and build small scoria mounds, but pyroclastics are very much subordinate to lava flows (Ollier 1988). Typical examples of current hawaiian eruptions include Kilauea, Mauna Loa and Mauna Kea in the Hawaiian Islands. By contrast, icelandic activity occurs through persistent fissure eruptions, unlike the more pronounced central activity in hawaiian eruptions, building up horizontal lava plains. Fissure eruptions result when magma-filled dykes intersect the surface. Large scale fissure eruptions occur where the crust is undergoing extension, such as the Laki Fissure in southern Iceland where 14 km^3 of basalt flooded from its 25 km length between June 1783 and February 1784 (Francis 1993). By the cessation of eruption, a line of small cones may mark the location of the fissure. Repetitive icelandic eruptions have been responsible for producing some of the largest basalt plateaus on Earth including the Deccan and Columbia Plateaus, and the Snake River Basalts. On the other hand, viscous lava flows include eruptions of rhyolitic to andesitic composition which are too viscous to flow large distances from their source. These usually build up around the vent to produce small domes and flows. Viscous flows, and more effusive lava eruptions are both evident in the Warrumbungle Complex (Section 6.5).

However, not all fissure eruptions are confined to regional extension zones. Often, the commonest eruptive episodes in large volcanoes (such as Etna) are satellite eruptions low on the flanks. These are fed by dykes radiating out from the core of the volcano. In intraplate or hotspot situations, like the Warrumbungle Complex, dyke propagation and crustal extension are related to the local tectonic regime, which is an expression of the size and shape of the volcano, and not its plate tectonic setting. An example of this type of activity is demonstrated by the rift zones which are the surface expressions of dyke swarms on Mauna Loa and Kilauea volcanoes (Ollier 1988).

3.4.2.1 High viscosity lava eruptions: emplacements (plugs, dykes, sills and domes)

Highly viscous igneous rocks give rise to distinct landforms (provided they do not explode) due to their limited flow. Igneous rocks that cool in the feeding channels or other weaknesses in crustal rock are called intrusive rocks. The formation of intrusions can affect the land surface in two ways. During emplacement, the overlying strata may be significantly uplifted and deformed. Secondly, when erosion has removed the upper portions of volcanoes and adjacent country rock, igneous intrusions give rise to distinct landform features as a result of differential erosion. The different types of rock associated with various forms of intrusive activity, along with the different types of magma from which they originate, play an important role in influencing the ultimate landforms created. Those relevant to the Warrumbungle Complex (plugs, dykes and domes) are discussed below.

Volcanic plugs (necks) are the cylindrical feeders of volcanoes that are filled with solidified lava, or tuff-breccia and agglomerate, depending on the nature of the original eruption and the type of magma. Volcanic plugs often rise dramatically above the surrounding landscape following the erosion of the upper part of the volcano and adjacent material. Australian examples, among others, include Crater Bluff and Tonduron Spire in the Warrumbungle Complex (Plate 3.1) and Mount Warning in the Tweed Shield, New South Wales and the Glasshouse Mountains in Queensland (Plate 2.1). Volcanic plugs often rise dramatically above the surrounding landscape following the erosion of the upper part of the volcano and adjacent material. It is possible for plugs to weather faster than the enclosing rock, creating a trench or depression. Such weathering leaves basins below the level of regional topography, giving rise to amphitheatres bounded by cliffs that form circular walls, breached only at the lowest point on its rim where streams drain the basin (Ollier 1988; Section 6.4.2).

Most volcanoes are fed through vertical fissures which give rise to vertical sheets of igneous rock. These features, known as dykes, may be long and thin (such as those radiating from Ship Rock in New Mexico, or the Breadknife in the Warrumbungle Complex, Plate 3.1), or range in thickness from a few decimetres to hundreds of metres. Dykes may occur as radiating patterns around a volcanic centre or as parallel swarms. While dykes such as the Breadknife give rise to wall-like ridges, more commonly dykes create subdued ridges. Sills, as opposed to dykes, are tabular formations which lie parallel to bedding planes in the country rock into which they are intruded. They range in thickness from 1-2 cm to >100 m. The Palisades Sill in the Hudson Valley, New York is up to 300 m thick. Sills can be differentiated from lavas because the heat of the intruding sill alters the rocks above and below it whereas lavas will only bake the underlying unit. When eroded, sills commonly give rise to landscapes of plateaus and cliffs since they behave the same way as any other sequence of layered rock. Other intrusions of a similar nature to dykes and sills include cone sheets and ring dykes.

Plate 3.1: View showing the structural relationships between the volcanic plugs of Tooraweenah Spire (Tonduron Spire) and Crater Bluff in the Warrumbungle Complex. The circular feature around Crater Bluff is a crater rim. Belougery Spire is thought to be a parasitic vent that formed on the flank of the Crater Bluff volcano (Hockley undated). The Breadknife is a 600 m long, 90 m high, 1-2 m wide radial dyke extends north-northeast from Crater Bluff. Lughs Wall, another dyke, may be observed extending from the head of the Breadknife. The Western Plains form the background. *Source*: Murray (undated).

When viscous lava is extruded, it sags and spreads into convex dome-like bodies. Thus, domes are steep sided protrusions of lava that are too viscous to flow away from a vent. Eventually, the vent is blocked by the lava. Formation of a dome takes place slowly, over periods of months or years. Commonly, extrusion of dactite lava is the last event in an explosive eruptive cycle. The extrusions are so thick that the term lava dome is used, rather than lava flow, for many silicic extrusions (Plate 3.2). However, not all domes are dactitic. Compositions may range from basaltic andesite to rhyolite. Many of the volcanic landforms in Australian shields are domes, coulees or spires, formed by lavas that do not flow great distances from the vent. Instead, they tend to sit astride the vent because of high viscosity and low magma-discharge rates. Many felsic lavas and domes appear to spread laterally as more lava is erupted from the vent into the interior of the flow. New lava rises along sub-vertical flow paths, and is succeeded by more lava that ramps up against it as the flow slowly spreads laterally (Cas and Wright 1987). Many of these domes, coulees or spires are steep sided, whereas the tops are locally irregular or hummocky. Both the margins and tops may be highly autobrecciated (see, for example, Duggan and Knutson 1993).

Plate 3.2: The distinctive lava dome of Bluff Mountain, Warrumbungle Complex. Note the high cliff-bound sides and prominent talus slopes forming at the base of the structure as it weathers along jointing planes. *Photo*: A. Timmers.

3.4.2.2 Low viscosity lava eruptions: lava flows

Lavas have a wide range of compositions from carbonatites through basalts to rhyolites. Excluding composition, the physical properties of lava flows are also influenced by their volatile contents, crystal contents and cooling histories. Thus, flows of marginally different compositions may behave quite differently. Only recently has the rheology of lava flows been closely scrutinised, so it remains poorly understood (Francis 1993).

The velocity of a lava flow is dependent on several factors, including rates of effusion, viscosity, density, volume and the slope and nature of the channel or surface that confines the flow. Velocity diminishes with distance from the source, but also varies across any given cross section of a flow at a given point in time since velocity varies from the middle of the flow to the top, bottom and sides (Williams and McBirney 1979). The rate at which lava is discharged from a source depends primarily on the composition and fluidity of the magma, and the size of the eruptive conduit. The rate of discharge of basaltic lavas may vary over time, from high initial rates through fluctuations over long periods, or they may simply cease abruptly. By contrast, changes in the discharge rates of silic lavas may be barely detectable (Williams and McBirney 1979).

Walker (1973) has suggested that the length of a lava flow is determined primarily by its rate of effusion. If it is slow, lavas of a limited range will pile up on one another, but if the rate of effusion is high, the lava will spread rapidly from its vent as a single flow. Generally, there are two principal types of basalt lavas- *a'a* and *pahoehoe* (Macdonald 1972). A'a is the most common type, with surfaces a jumble of loose, irregularly shaped cindery blocks. In cross section, a'a flows are generally a few metres thick and consist of two distinct zones. The outermost of these is a rubbly skin that rides on the massive, solid interior. If the lava directly overlies the pre-existing soil, this will often be baked, and be separated from the massive base by a thin layer of rubble. On the other hand, pahoehoe lavas are the least viscous of all common lavas and are therefore able to form a wide range of fluidal surface structures and are able to travel far from their source since they are relatively fast moving. Cooling forms a very thin skin that may be dragged into folds by the movement of the still-mobile lava underneath. This produces surfaces known as sharkskin, filamented, corded, ropy, entrail, festooned, and elephant hide lava, among others (Ollier 1988). In Australia, pahoehoe ropy lava surfaces are uncommon, even in the Late Tertiary and Quaternary provinces of Queensland and Victoria, although examples have been recorded in the Toomba Flow, northern Queensland (Stephenson and Griffin 1976) and in the Bridgewater volcanic centre, Victoria (Cas 1989). Pahoehoe and a'a lavas often erupt from the same vent. While visual differences are striking, the differences within the lavas are subtle. While the two may have identical chemistry, the difference between them lies in the physical structure of the silicate melt and the way that it is polymerised, largely as a function of lower flow temperatures. While pahoehoe may experience transition to a'a, a'a lava never reverts to pahoehoe (Peterson and Tilling 1980).

While flows are governed by factors listed above, flows that spread on a broad front will not be as long as those confined to channels or lava tubes. Stephenson and Griffin (1976) reported an hawaiite flow in northern Queensland 160 km long that had flowed down an average slope of 0.3°. The confined channels provided by dry stream beds and lava tubes, combined with the high rate of discharge allowed the lava to flow for great distances despite the fact that its viscosity was higher than that of many basalts. In the Warrumbungle Complex, lavas have been measured that have travelled up to 50 km from their source (Section 6.4.2.1). By contrast, trachyte and andesite lava flows are much shorter and thicker than basaltic lavas. In general, trachyte lavas generally form stumpy flows or pile up over their vent as steep sided domes (Section 3.4.2.1), while basalt flows are the most abundant and voluminous flows, ultrabasic lavas, along with andesites, dactites, rhyolites, trachytes and phenolites, produce much smaller flow which tend to decrease in volume as the content of silica and alkalis increase (Williams and McBirney 1979). Andesitic flows exhibit higher yield strengths and viscosities than basalts so that, in general, andesitic flows are shorter, thicker and travel more slowly than basalt flows. Block lavas are typical of andesites. The flow fronts of blocky lavas are steep, reaching heights of 100 m, and consist of huge angular blocks perched on top of one another. Autobrecciated textures occur when blocks cascade down the front of the moving flow and are later consumed by it, producing a mass of angular fragments solidly welded together (Francis 1993). Such lavas are distinct from true a'a lavas which, as such, have not been recognised from eastern Australian fields, but have been documented on Dunedin Volcano, New Zealand (Sewell and Weaver 1989). Dactite lavas are even more viscous than andesites and form thick, short, steep sided extrusions.

3.5 Other morphological features of basalt lava flows

The topographical expression of a lava flow is often a plateau or flattish plain (Plate 3.3). This is because lava, whether it flows from a crustal fissure or crater, will spread outward from its source approximately parallel to the surface over which it flows (McKnight 1996).

Plate 3.3: The flat-topped lava surfaces of Wallumburrawang Ridge, Warrumbungle Complex. Several flows may be delineated here, with sapping of underlying pyroclastic falls (arrow) causing slow retreat of the overlying lava flow. *Photo*: C. Duthie.

The propagation of low viscosity lava flows is promoted largely by the development of an interior magma feed system of lava tubes and channels. In accordance with the Bingham characteristics (for example, Hulme 1974), the margins of the lava flow cool and may freeze, forming a natural levee (Table 3.3). This acts to confine and insulate the flow, extending downstream as the flow advances, so defining an interior channel through which the lava is fed to the front of the flow (Cas 1989). Marginal cooling of the top of the lava surface may form a solid roof, enclosing the flow, forming a lava tube or tunnel when the lava drains away (Table 3.3). On level topography, lava may spread widely to form extensive veneers of basalt on pre-existing plains. Lava flows of this type occur on the volcanic plains of western Victoria, covering an area of 15 000 km^2 (Ollier 1988). On undulating topography, basalts tend to flow down valleys, partially or completely filling them and spilling over interfluves. Such flows disrupt the pre-existing drainage, displacing rivers and sometimes completely altering drainage patterns. The influence of volcanic activity on drainage is dealt with in Section 3.8.

Other notable surface features of lava flows include *pressure ridges*, *convex lava surfaces*, *lava levees* and *columnar jointing*. The morphological characteristics of these features are described in Table 3.3. While other small-scale features of lava surfaces exist (for example, hornitos and tumuli), those listed here are of more direct relevance to the morphology of the Warrumbungle Complex.

3.6 Pyroclastic eruptions

Pyroclastic eruptions may take several forms depending on their level of explosive activity. This is because explosive volcanic activity involves different modes of fragmentation, different styles of eruption and a diversity of products. These will be reviewed briefly before concentrating on pyroclastic landforms and landscape development. Useful reviews on pyroclastic rocks, terminology, eruption styles and deposits are presented by Wright *et al.* (1980), Fisher and Schmincke (1984) and McPhie *et al.* 1993.

3.6.1 Types of pyroclastic eruptions

Pyroclasts are derived from magmatic explosions driven by the exsolution and explosive expansion of magmatic volatiles, especially water and carbon dioxide. Cas (1989) suggested that magmatic explosions take place in two distinct situations. In the first situation, the vent

Table 3.3: Some morphological features of lava flows common to the Warrumbungle Complex.

Surface feature	Description	Reference
Pressure ridges	These develop perpendicular to the direction of flow and form by updoming of parts of the plastic surface, much like pahoehoe, but on a larger scale.	Cas 1989
Convex lava surfaces	In cross-section many flows have a convex surface, and the more viscous the lava, the greater the convexity. Because lava cools fastest on the edges, this presents an obstacle to spreading and a convexity develops at the edge of the flow. This is important in developing lateral streams.	Cotton 1944; Ollier 1988
Lava levees	Lava levees form when solidified fragments are heaped to the sides of the flow, framing a central depression. Levees aid the propagation of flows and may form in both pahoehoe and a'a flows.	Ollier 1988
Lava channels/canals	These are formed when the supply of magma ceases and lava drains from the interior lava surface causing it to fall or deflate, leaving well defined levees.	Cas and Wright 1987
Columnar jointing	Columnar jointing is the product of progressive cooling of lavas. Cooling at the margin of flows produces polygonal contraction cracks which then propagate perpendicular to the cooling front as the front advances, producing polygonal cooling, such as the columns of the Giants Causeway in northern Ireland.	Cas 1989; Summerfield 1991; Section 6.5.

is open, and vesiculating magma interfaces directly with the atmosphere. Fragmentation of the vesiculated magma is largely caused by high exit velocities during *strombolian, sub-plinian* and *plinian* eruptions. In the second situation, the vent is blocked, and the release of confining pressure may be rapid enough to initiate rapid exsolution and bubble growth and explosive eruption (Sparks 1978). In these cases, explosive activity is marked by *vulcanian* eruptions characterised by short-lived violent vent clearing eruptions.

Of these types of eruption, the two most common types are strombolian and vulcanian eruptions. Strombolian activity has a higher proportion of pyroclastic ejecta than hawaiian activity, although the magma is still basaltic. However, spatter deposits predominate where the eruptions are hawaiian rather than strombolian in style. Strombolian eruptions are commonly characterised by a white cloud of steam issuing from the crater. Eruptions may be continuous or intermittent, ranging from a few minutes to a few hours. Pyroclastic deposits may alternate with lava outpourings such that volcanic deposits may consist of alternating lava and pyroclastics. Scoria fall deposits are diagnostic of strombolian-type eruptions. These falls are generally massive and structureless, can be tens to hundreds of metres thick near the vent and form sheet-like stratified scoria deposits farther away from the source. They result from continuous, maintained strombolian-type eruptions. Scoria cones are common in basaltic fields in intraplate settings in eastern Australia (Cas 1989).

Vulcanian eruptions are characterised by dark cauliflower-shaped clouds caused by the unblocking of a vent. Eruptions are violent and disrupt pre-existing volcanic landforms due to their explosive nature. If ejecta consists of old rock fragments and there is no lava discharged during the eruption, the activity is described as ultra-vulcanian type. Activity of similar appearance at the end of a complex eruption may be termed pseudo-vulcanian. Vulcanian eruptions may contain vesiculated juvenile clasts, but most are dominated by dense blocks of lava derived from the violent, explosive disintegration of vent-blocking lava caps.

The more violent pyroclastic activity are termed *plinian* and *peléan* eruptions. In plinian eruptions, the dispersal of ash is widespread and eruptions may last from several hours to several days. The column formed in plinian eruptions has two parts: the lower gas thrust and the upper convective thrust. When the density of the gas thrust is less than that of the atmosphere, convectional rise takes over. The smaller the pyroclast fragments, the greater the heat exchange, and the higher the ash column. The convectional part commonly makes up 90% of the height (Ollier 1988). *Ultraplinian* eruptions represent the next increment in plinian type eruptions. Such eruptions are characterised by column heights in excess of 45 km. Walker (1980) and Wilson and Walker (1985) have described an ultraplinian eruption in the Taupo area of the north island of New Zealand in which the erupted material covered an area of more than 15 000 km^2 in which tephra deposits 100 km from the source are more than 10 cm thick. The eruption column may have been as much as 50 km high in order to generate such dispersal. To generate a column so high, the rate of eruption must have been exceptional, of the order of 100 000 $m^3\ s^{-1}$. Pumiceous deposits are commonly associated with more violent explosive eruptions such as sub-plinian, plinian and ultraplinian eruptions. While not common in Australia, pumiceous deposits in intraplate settings have been recorded from Canobolas, Focal Peak and the Warrumbungle Complexes.

Unlike plinian activity, peléan activity is more difficult to characterise, and is often linked to vulcanian or plinian activity. Peléan is a general name for very violent eruptions and explosions of highly viscous magma. The magma is usually intermediate to acidic but basalt eruptions have been recorded (Cas 1989). One feature however, characterises peléan activity: the eruption of *nuées ardentes*, or glowing clouds. A nuée consists of solid fragments (huge boulders mixed unsorted with fine dust) which avalanche downslope under the influence of gravity. Hot escaping gases rise upwards, carrying large amounts of dust with them, forming a turbulent wall of ash-laden cloud kilometres high, while the denser part, containing most of the more solid material, rolls rapidly over the ground at great speed (Francis 1993).

A further type of activity may be associated with the presence of groundwater. *Phreatic* activity refers to the violent explosions that occur when ascending magma or hot rock meets groundwater and rapidly produces large amounts of steam. Water can interact with hot volcanic material in a variety of ways: when a vent opens up under the water of a lake or sea, when a volcanic vent on dry land intersects water contained in an aquifer, or when lava or pyroclastic material flows over water-saturated sediment such as river gravel or a swamp (Cas and Wright 1987). Ejected solids may also be country-rock fragments, which may either be volcanic or non-volcanic depending on the history of the area. Hydrothermally altered clay products will also be a significant component where the subsurface rock has been hydrothermally altered. Deposits consist of thick, near-vent sheets of poorly sorted, clay-supported breccia (McPhie *et al.* 1993). The breccia usually contains a range of lithologies that have been explosively excavated from the subsurface succession (Cas 1989). In the Warrumbungle Complex, evidence of phreatic interaction occurs at Bress Peak (see Figure 6.2) where the rock is highly altered, and bleached clay-rich rock indicates that trachytic rock has been affected by hot volcanic water or steam (Duggan and Knutson 1993). Similar highly altered materials are a common feature in present day active volcanic regions such as Yellowstone National Park (USA) and Rotorua in New Zealand.

When large amounts of surface or phreatic water interact with magma rather than hot rock, violently explosive *phreatomagmatic* eruptions occur (Francis 1993). The fundamental characteristic of phreatomagmatic eruptions is a high yield of fine-grained, highly fragmented ash. This is caused by the thermal energy of the magma heating water and turning it to steam. The steam fragments the magma as it expands explosively. Phreatomagmatic deposits are widely recorded in eastern Australia and are frequently used to explain the formation of maars (circular rings of pyroclastics, steep on the inside and shallow (3-4°) on the outside) which, in many areas, are found in places with abundant groundwater or near sea level (Ollier 1988). In the Newer Volcanics, Victoria, many of the maars contain lakes such as Gnotuk (30 m deep) and Bullenmerri (80 m). Some, such as Cobrico are swamps, and others, such as Terang and Wangoom, are now dry (Ollier 1967).

While it is relatively easy to classify types of eruptions from active volcanoes and recognisable phases of activity such as well preserved volcanic deposits, labels such as strombolian or plinian are more difficult to apply to heavily eroded volcanoes (such as the Warrumbungle Complex), that lack eruptive evidence, especially when many eruptions exhibit several different phases of different duration (Williams and McBirney 1979).

3.6.1.1 Pyroclastic ejecta: pyroclastic flows

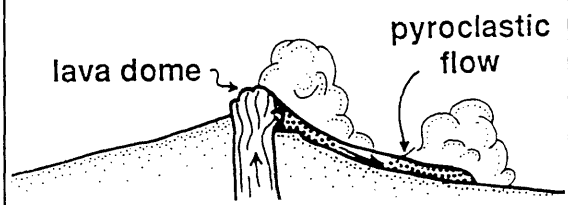
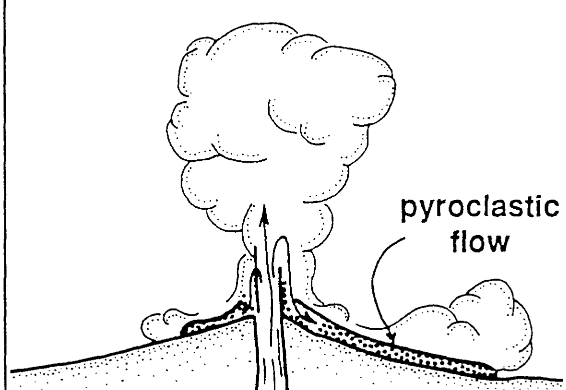
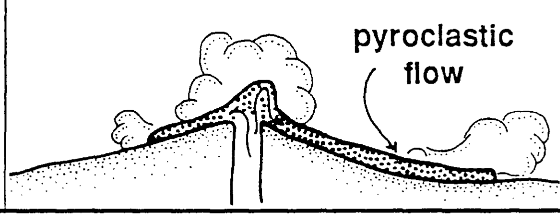
Almost all volcanism involves some form of violent eruption ranging from minor explosive activity to cataclysmic explosions. Generally, pyroclastic volcanism may produce *flows* (Plate 3.4a) or *fall* deposits (Plate 3.4b). Pyroclastic flows result from the association of large amounts of gas and massive and sudden eruptions of volcaniclastic or hydroclastic material. They are hot, highly concentrated, ground-hugging and highly mobile (Wright and Walker 1981). The mobility of such flows are closely related to exsolution of gas, release of gas from broken fragments, and the heating of engulfed air (Ollier 1988). Pyroclastic flows (Table 3.4) are commonly associated with intermediate to felsic Plinian eruptions and, on a small scale, with dense-column, intermediate, vulcanian-strombolian eruptions (Cas 1989). Cas (1989) also reported that there are few pyroclastic flow deposits in eastern Australian intraplate volcanic terrains, and pyroclastic flows are not recognised as significant transporting agents in basaltic terrains. In most cases, pyroclastic flow deposits pond in topographic depressions and are much thinner or completely absent from topographic highs. They have the effect of smoothing out and in-filling topography. The top surface of ponded

(a) (b)

Plate 3.4: (a) Pyroclastic ash billows from the Soufriere Hills volcano in the form of flows at the base of the vertical eruption column. (b) The partially subdued relief of Salem, Monserrat, blanketed by pyroclastic airfall deposits (see also Figure 3.2). *Source*: The Australian 1997.

deposits are flat, although there may be levees at the margins and a gentle sag in the centre were compaction has been greatest (McPhie *et al.* 1993).

Table 3.4: Pyroclastic flow genesis and landforms.

Flow production	Eruption style	Type of flow deposit
Gravitational or explosion-triggered lava dome collapse in association with the extrusion of lava domes.		Nuées ardentes; block and ash flow deposits comprising poorly to moderately vesicular lapilli and ash which are commonly blocky and angular. Ash pyroclasts consist of angular glass shards and some crystal fragments.
Collapse may follow immediately after a single, or series of, explosions as occurs in some vulcanian eruptions. Activity produces scoria and ash flow deposits. Alternatively, collapse of plinian-style eruption columns may cause voluminous pumiceous pyroclastic flows which produce ignimbrites.		Scoria, ash, ignimbrites.
Flows may result directly from the vents by upwelling and overflow, or low fountaining of pyroclast-gas mixtures. Some pumiceous pyroclastic flows and scoria and ash flows are formed in this way.		Pumiceous, scoria, ash.

Source: Adapted from McPhie *et al.* 1993.

3.6.1.2 Pyroclastic ejecta: pyroclastic falls

Pyroclastic fall deposits or tephra (Table 1.1; Table 3.5) on the other hand, are composed of fragments that have fallen through the air after an eruption and may include ballistic ejecta such as bombs, or airfall deposits. They may be generated by the entire range of explosive eruptions as described in Section 3.6.1, and by magmas of any composition. Felsic tephra deposits have not been documented in detail from east Australian intraplate centres, although their presence has been noted or implied in several centres (for example, Focal Peak, Nandewar and Canobolas; Cas 1989). Tephra deposits systematically decrease in grain size and thickness with increasing distance from the source vent. At any one location, they are characterised by even thickness, laterally continuous mantle bedding, and good sorting (Plate 3.4b; Figure 3.2).

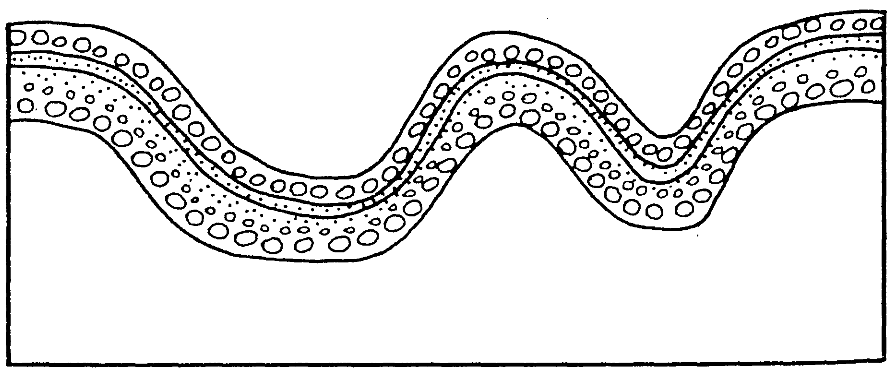


Figure 3.2: The geometry of tephra deposits derived from eruption fallout. The deposit mantles the topography and is well-bedded and sorted. By contrast, pyroclastic flows tend to pond in surface depressions and are relatively poorly sorted. *Source*: McPhie *et al.* 1993.

Table 3.5: Pyroclastic airfall (tephra) deposits.

Fall deposit	Description
Agglutinate	Comprises spatter (poorly vesicular, fluidal, juvenile pyroclasts) and bombs that accumulate near vents in explosive eruptions of low viscosity magma.
Agglomerate	Coarse grained (>64 mm) deposits containing bombs and blocks confined to proximal settings.
Welded fall deposits	Juvenile pyroclasts are sintered together and flattened, forming a coherent rock. These deposits result from very rapid accumulation of pyroclasts that have low viscosity.

Source: Wright *et al.* 1980; McPhie *et al.* 1993.

3.6.2 Landforms of explosive eruption centres

Basaltic explosive-eruption centres may be superficially divided into magmatic and phreatomagmatic explosive centres (Cas 1989). The former typically build scoria cones whereas the latter are characteristically responsible for more diverse features, including maars, tuff rings and tuff cones (Section 3.6.1). Of particular interest in this study is the development and preservation of the crater-like features that may be associated with these types of activity, particularly since the actual landforms (maars, tuff cones) have not been preserved in the Warrumbungle Complex.

Craters, as morphological features, have been defined in Table 1.1. When eruptions cease, the central part of the cone falls into the vent to form a crater. In the case of erosion, craters will vary in size and degree of preservation. Retreat may be rapid on steep crater walls, exposing layers of ash. The crater may thus become enlarged to several times the vent diameter. Bullard (1976) reported that craters rarely exceed "... 0.75 to 1 mile in diameter ...", concurring with Ollier (1988), who suggested that crater diameters may be less than 1 km but rarely greater than 2 km in diameter. Typically, a depression larger than 5 km in diameter is defined as a *caldera*, of which the diameter is many times greater than that of the included vent or vents (Bullard 1976). Calderas are thought to arise through collapse following eruption and partial emptying of the magma chamber several kilometres below the surface. The formation of *erosion calderas*, such as that of the Tweed Volcano, northern New South Wales, are distinct from those features caused by structural collapse. In such cases, there is no evidence of ring dyke structures and associated features such as cone sheets, which would otherwise be exposed by erosion following the removal of all traces of the surface (collapse) caldera. Such features (ring dykes, cone sheets) are interpreted by Reynolds (1956) as integral to the evolution of (collapse) calderas and the subsequent geomorphic development of adjacent areas. Further evidence differentiating craters from calderas is derived from the uniformity of basement rock, which, for crater development, will show evidence of near-horizontal bedding, indicating that subsidence processes have not operated, and the feature is likely to be a crater.

While the distinction between craters and calderas may be made on morphological and structural grounds, it may be noted that craters are active constructional features, related to cone building, while caldera development is related to explosion or collapse and are a negative, passive form. Craters may therefore be distinguished from calderas by constructional morphology, as well as their size.

3.7 Erosion and weathering of volcanic landforms

In volcanic landscapes, all surface deposits and landforms (such as those discussed above), consolidated or unconsolidated, are subject to the same physical and chemical erosional and weathering processes as all other geological materials. The main effects of surface processes following terrestrial volcanism is the degradation of individual features and the regional downcutting of the volcanic landscape. Initially, downcutting will be controlled by the regional base-level. It will also be controlled within the margins of the volcanic terrain by absolute relief. The relief in well-developed continental intraplate central-type provinces is low because of the isolated and low-relief nature of the shields. However, the elevation of the province above regional base-level, and the relief of individual features within the shield, may be significant, depending on pre-existing relief and tectonics (Cas 1989). Walker (1984), for example, calculated downcutting rates of 58 m Ma^{-1} in the cold elevated highlands of Iceland where the average elevation is 400 m ASL. By contrast, Cas (1989) suggested that rates in eastern Australia are likely to be significantly less than this, despite the milder climate, because of the relatively low elevations of volcanic shields above base level. Bishop (1985) and Wellman (1979b) reported post-volcanic denudation rates in the southern highlands of southeastern Australia of 8 m Ma^{-1}; and Timmers (1992) calculated a post-basaltic denudation rate of 0.95 m Ma^{-1} for the New England Tablelands. These rates suggest that Cenozoic erosion rates have been relatively low in these parts of eastern Australia.

In volcanic areas Cas (1989) reported that denudation at the local scale on individual centres may be significant when scoria cones collapse penecontemporaneously with volcanic activity but post-eruptive degradation is surprisingly slow even for loosely packed scoria piles (for example, Wood 1980). However, long-term denudation rates for volcanic areas may not be truly representative of the actual rate of processes because erosion may be rapid and severe in unconsolidated pyroclastic material. Selby (1985), for example, reported gullies 400 m long, and 20 m wide and deep that formed within 24 hours in pumice in the North Island of New Zealand during severe storms. Obviously, rates of denudation will vary from region to region depending on climate, regional relief and lithology.

Even after a volcano has been deeply dissected, anatomically distinct landforms may remain. These may reflect the internal flow structures of the volcano (for example, feeder vents through which lava reached the surface are often preserved long after the rest of the volcano has been eroded away because the material of which they are composed is often more resistant to erosion than adjacent lithologies), or may mark the location and/or extent of erupted material. For example, the erosion of volcanic areas, principally through water erosion, will control succeeding lava-flow geometries, although the volume of the lava flow, its rate of supply and propagation, and the way in which the volume of the flow is accommodated by the topography, will also be principal controls (Cas and Wright 1987).

Erosion rates in unconsolidated material are high, not only because it is unconsolidated, but because stabilising vegetation is often destroyed by continuing eruptions. New particles created solely by surface weathering and erosion are termed *epiclasts*. Epiclastic processes are important in all volcanic landscapes (Cas and Wright 1987) because they operate contemporaneously with, as well as, after volcanism. However, weathering and erosion of pre-existing, poorly consolidated aerially deposited volcanic material simply releases the original pyroclastic or autoclastic material and rapidly provides large volumes of *recycled* particles. These reworked particles form *volcanigenic* sedimentary, rather than epiclastic deposits (McPhie *et al.* 1993). Cas (1989) suggested that epiclastic processes are less dynamic in intraplate settings when compared to (high-profile) stratovolcanoes because intraplate settings are characterised by low-profile centres (Section 3.3), by relatively low proportions of volcaniclastic deposits, and by the topographic smoothing effect of fluidal basaltic lavas. Indeed, little evidence of these deposits and the processes that were responsible for them is preserved in the study area due to high levels of denudation (see Plate 4.1a, b) and the reasons listed below.

In non-volcanic sedimentary environments, facies and facies variation are dependent on several interrelated controls (for example, Gale and Hoare 1992). In active volcanic landscapes, there are additional controls on sedimentation, for example:

1. eruptions strongly influence sedimentary processes and sediment supply;
2. steep slopes and earthquakes are common, and slope failure events make large contributions to sediment load;
3. volcano-tectonism (faults, rapid uplift and subsidence) cause frequent and rapid changes to sedimentation; and
4. some volcanic processes are constructional and quickly create and modify topography and drainage (Cas 1989).

These volcanic controls have a significant effect on sedimentary structures, as well as grain-size distribution and facies architecture. Of interest in this study are sedimentary structures, in association with diatomite and volcano-lacustrine deposits preserved on the outer flanks of the shield (Section 4.4.3.5). The sedimentary structures preserved in this deposit are useful for environmental interpretation because rapid aggradation may preserve, inhibit or modify the development of sedimentary structures. Conversely, erosion of sedimentary structures give an indication of processes that operate during or after eruptions. Thus, different environments are characterised by differences in the kinds of sedimentary processes which operated.

McPhie *et al.* (1993) has suggested that studies of the response of fluvial systems to volcanic eruptions which generate voluminous pyroclastic materials (such as in the Warrumbungle Complex; Section 6.6) identify two fundamental conditions of landscape and sedimentation. The *syneruption* period of eruption is characterised by the geologically instantaneous production of large volumes of pyroclastic material which results in enhanced and variable runoff. This leads to sedimentation by high sediment load flood- or debris-flow processes. The dominance of coarse silt to fine gravel size particles in stream transport has an important effect on river morphology, especially in areas with mountainous headwaters. In areas normally characterised by gravel bedload regimes, the presence of pyroclastic sediments may cause channel widening and decreases in sinuosity (Kuenzi *et al.* 1979). The periods of high volume pyroclastic eruptions are short and separated by longer *inter-eruption* phases where volcanic activity has little or no impact on the fluvial system (McPhie *et al.* 1993). This inter-eruption period is characterised by normal streamflow processes and decreased sediment delivery.

3.7.1 Denudation of volcanic features

One of the most important factors contributing to the survival of landforms in the order of 10^5-10^7 years old is the low rate of denudation that the Australian continent has experienced during the Mesozoic and Cenozoic. In most areas, rates well below 10 m Ma^{-1} have prevailed (Gale 1992; Section 3.7). This means that the Australian continent has been one of the world's most tectonically and geomorphically stable continents. This is because the continent is orogenically stable with the last episode of orogenesis confined to the New England-Yarrol Orogen in the Permo-Triassic. Thus, potential energy is low. Coupled with the progressively arid evolution of Australia's climate, denudation is slow because surface processes operate slowly. However, aridity cannot account for past low rates of denudation throughout the Mesozoic and Cenozoic, during which a more humid climate prevailed (Appendix A). Furthermore, Quaternary glaciations have not significantly influenced mainland Australian denudation rates, because there was little glaciation. Galloway (1963) reported no more than 19 km^2 of ice covering mainland Australia during the last glacial episode.

Once created, denudational processes can rapidly modify the original constructional form of a volcano. The rate of modification is a function of the resistance to weathering and erosion of the erupted material, the initial relief created by the volcano and by the prevailing climatic environment. However, there is little reference in the literature to erosional processes and products from intraplate volcanic settings in eastern Australia. While low rates of denudation have contributed to the maintenance of regional and some individual local Australian landscape features, this alone does not account for the preservation of some ancient landforms, such as those found in this study (Section 6.4.2). The preservation of such landforms means that denudation rates must be highly localised in the landscape in both time and space for long periods of time, as well as being very slow.

3.7.2 Denudation as an indicator of age in volcanic landscapes

In addition to the importance of the constructive nature of volcanism, volcanic activity is also of geomorphic significance for other reasons. Episodic volcanic activity can be dated by radiometric techniques and can therefore provide maximum ages for activity and minimum ages for the landsurface over which it lies. Since shield volcanoes have a symmetrical form when erupted, their original form can be reconstructed with some confidence despite intervening erosion. When these two elements are combined, reconstructions enable estimates to be made of the amount of material eroded over a known period of time. Therefore, volcanoes can be used to produce reliable estimates of denudation. In addition, since volcanoes of similar composition are often found in a range of climatic environments, they provide an opportunity to compare the effects of climate on the rate and nature of denudation (Summerfield 1991), *ceteris paribus.*

The use of geomorphological parameters as indicators of shield ages has been widely applied, especially to Quaternary cinder cones. These parameters include the classification of stages of erosion of cones and their associated lava flows (Colton 1937), the maximum cone slope angle (Scott and Trask 1971), the ratio of cone height to basal diameter (Scott and Trask 1971, Wood 1980), the tangent of the cone slope and the change in surface features of lava flows associated with cinder cones (Bloomfield 1975). Hasenaka and Carmichael (1985) also used gully density and the surface morphology of lava flows to estimate relative ages of cinder cones of the Michoacán-Guanajuato Volcano in Mexico. These authors found that degraded cinder cones, lower slope angles, smaller numbers of gullies (which were larger and deeper), greater soil development, and more weathered and oxidised ejecta differentiated the older cones from the younger cones. Hasenaka and Carmichael (1985) also found that lava flow morphology changed with age. Younger lava flows displayed well preserved original surface features such as flow margins, boundaries of individual flow units, pressure ridges and levees. Flows that were older than Holocene in age tended to lose these characteristics with time.

One problem associated with the application of morphometric parameters to relative age determination in older volcanoes is the variation of lithology over small areas. Erosion of less resistant lithologies causes preferential preservation of landforms. For example, Fried and Smith (1990) reported that lithological control was the main landform determinant in the Glen Innes-Inverell volcanics of northern New South Wales. Further, due to the composite nature of interbedded lavas and pyroclastics, many areas of complex volcanoes yield little information on morphological age because they are buried by other ejecta, colluvium or alluvium or have been highly weathered and eroded, making age-morphology comparisons difficult.

3.8 Drainage development and alteration in volcanic areas

3.8.1 Drainage alteration

Volcanoes can cause immense topographic alteration to their surrounding landscape, with the effect of volcanism ranging from minor to catastrophic on pre-existing drainage. On volcanic cones, radial streams originate below the rim and flow with little sinuosity down the flanks. On the largest scale, the topographical swell above a mantle hotspot may influence large areas of regional topography (Section 2.4.1). Radial drainage patterns produced as a result of hotspot uplifts have been mapped on several continents. Mount Etna is a classic example of this (Chester and Duncan 1982). In eastern Australia, evidence of drainage diversion caused by central-type volcanism has been reported by Ollier (1985) with other examples in the Snowy River associated with the Monaro Volcano in New South Wales (Ollier and Taylor 1988) and the Wannon River in Victoria, which has been diverted around a lava plain (Ollier 1991).

When several volcanic cones erupt close together their drainage patterns interact. This results in several sectors of radial drainage with another line of drainage along the gutter between adjacent cones; such as the drainage of southeast Gough Island in the South Atlantic (Ollier 1984b) and the Focal Peak-Tweed Volcano in northern New South Wales (Ollier and Haworth 1994). On a local scale, volcanic activity may alter drainage by blocking valleys and impounding lakes and rivers behind lava or pyroclastic dams. Additionally, volcanism may contribute to the damming of water bodies that may become important depocentres for fossil biota.

Volcanic eruptions have been known to completely alter regional drainage patterns as pre-existing landscapes are buried by lava. Lava flows may form dams which divert rivers and initiate processes of degradation and aggradation controlled by new base-levels. There are numerous Australian examples of this, including the western slopes of the Tweed Shield, the Comboyne Plateau and around the Ebor-Dorrigo Volcano. If only small cones or flows are produced, streams may deviate around them with little variation to their old courses, but if enough lava is produced or there is significant change to local topography, all pre-existing drainage may be entirely altered to produce a new drainage pattern initiated on the volcanic cones and flows. The various types of drainage diversion that can be caused by lavas are shown in Figure 3.3. The distribution of lava and stream patterns enables quite detailed reconstruction of pre-volcanic drainage. Pyroclastic eruptions also have a considerable impact on drainage. For example, the 1980 eruption of Mount St. Helens (USA) sent a massive pyroclastic/debris flow down the north Toutle Valley, raising its floor by 180 m, damming tributary valleys and creating new lakes and ponds (Findley 1981). Figure 3.4 demonstrates the geomorphic alteration of streams caused by this eruption.

The reconstruction of pre-volcanic, penecontemporaneous and post-volcanic drainage usually requires the determination of both flow direction(s) and the time(s) when the flow occurred. Such a geological approach may include determination of flow directions from gravel fabrics or crossbedding. Alternatively, a morphological direction may be taken whereby valley flow lavas are used to reconstruct former drainage. This method uses the morphology of river valley planforms to determine their ancient flow directions and ages. Bishop (1982), in his study of river systems in eastern New South Wales, favoured the geological approach, which is generally more conservative, concluding that palaeodrainage approximately parallels its modern equivalents. By contrast, the morphological approach

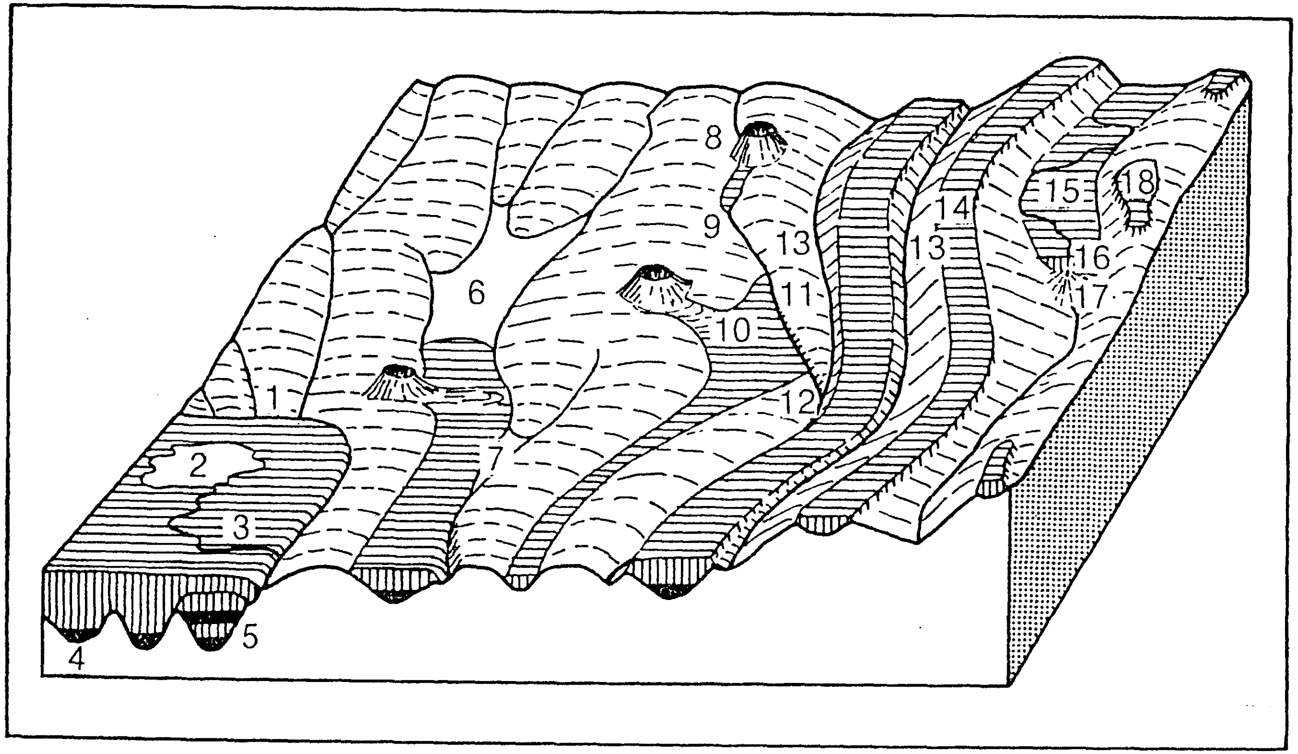


Figure 3.3: Drainage and landforms associated with lava flows. **1**. River diverted by lava plain. **2**. Insequent lake on lava surface. **3**. Insequent streams on lava surface. **4**. Deep lead. **5**. Multiple lava flows and deep leads. **6**. Lava dammed lake (main valley). **7**. Lava dammed lake. **8**. Stream flowing beneath cone and flow. **9**. Spring. **10**. Flow diverting river into next valley. **11**. Lava diverted stream. **12**. Probable site of gorge **13**. Twin lateral streams. **14**. Inversion of relief. **15** Insequent stream on valley flow. **16**. Waterfall. **17**. Alluvial fan. **18**. Residuals of old lava flow. *Source:* Ollier 1988.

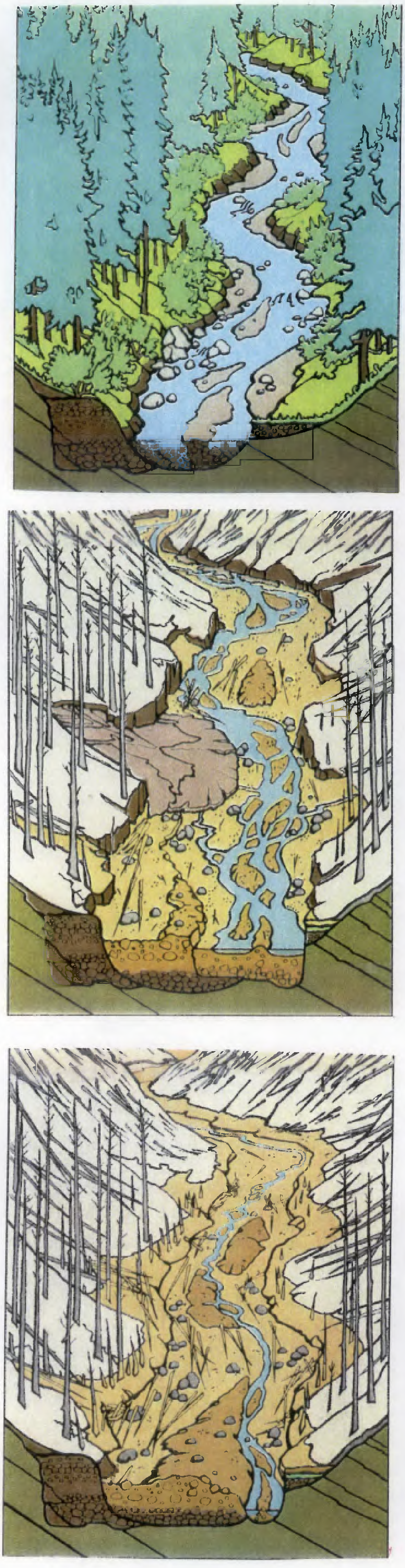


Figure 3.4: The geomorphic effects of the Mount St. Helens eruption on regional drainage. (a) Before the 1980 eruption, the forested catchment retained maximum amounts of rainfall, vegetation stabilised the banks and lag gravels protected the meandering stream bed.

(b) After the 1980 eruption, debris from the volcano ripped up the lag gravel and destroyed the vegetation, causing the hill slopes to destabilise. The stream channel clogged with sediment, straightening the stream course and undercutting and destabilising the banks.

(c) One year after the eruption, altered debris loads caused the stream to braid as a result of changed sediment loads, transport agents and depositional patterns. Debris and alluvial fans formed at tributary junctions and undercut banks collapsed, further widening the channel. *Source*: Modified from Findley 1981.

indicated that there has been major drainage disruption in the Late Cretaceous-Early Tertiary, with concomitant westward migration of the divide (Haworth and Ollier 1992). Drainage alteration in volcanic areas may also be identified by the investigation of associated sedimentary deposits. Such deposits may yield data on flow direction which can be used to discover whether stream courses have been altered in any way by igneous intrusion or by any of the other forms of drainage alteration listed in the preceding paragraph through processes of capture, reversal, entrenchment or inversion. Due to the lack of available geological evidence, largely removed by erosion, a morphological approach is preferred for the investigation of the Warrumbungle Complex.

3.8.2 The erosion of volcanic cones: planeze development

At the end of eruption, many volcanoes have a conical shape. When drainage is initiated on a volcanic cone (Figure 3.5a), there is little erosion near the crater rim because the catchment area is small (for example, Horton 1945). Channelised drainage usually starts mid-slope and reaches its greatest depth there. Channels at the base of the cone are again smaller since channel slope is smaller. Generally, drainage is not regular, even on a symmetrical cone (Selby 1985). Irregularity may further increase if flows are channelised down sites of landslides, follow lava flows, or flow through areas of stratified geology that have differing resistance to erosion. Pyroclastic deposits such as ash, pumice or scoria which have not been welded by heat, consolidated or weathered are extremely permeable are not subject to the rapid effects of erosion. Initially, rainfall will permeate through the tephra deposits. Once water concentrates, erosion may be rapid and severe. However, Wood (1980) suggested that scoria cones are degraded largely through weathering of tephra to clay. Once weathered the development of radial gullies leads to incision, lateral transport, slope lowering and concomitant growth of the width of the cone and apron. The implication of this is that cone degradation does not begin significantly until the scoria pile becomes impermeable. Flat-topped interfluves (or planeze surfaces) will remain on the lower slopes, but will be separated by interlinking gullies which have been carved into the deposit. The lower slopes have the best chance of being preserved as erosional remnants of the original surface. *Planeze* surfaces (Figure 3.5b) are generally preserved between incised channels and are useful tools in the reconstruction of volcanic shield surfaces. For example, Solomon (1964) used planeze surfaces to reconstruct the original height of the Tweed Volcano, concluding that the original height of the shield peaked at 1920 ± 150 m. Planeze development is associated with the evolution of radial drainage on volcanic surfaces, since radial drainage occurs when stream heads are in close proximity to the crater and stream tails are widely distributed at the basal circumference of the volcano. Additional drainage modification of volcanic surfaces may arise along landslide and avalanche gashes that develop over a cone surface once activity ceases. These gashes provide a channel in which drainage may be concentrated and will be rapidly deepened by water erosion, producing a residual volcano (Figure 3.5c).

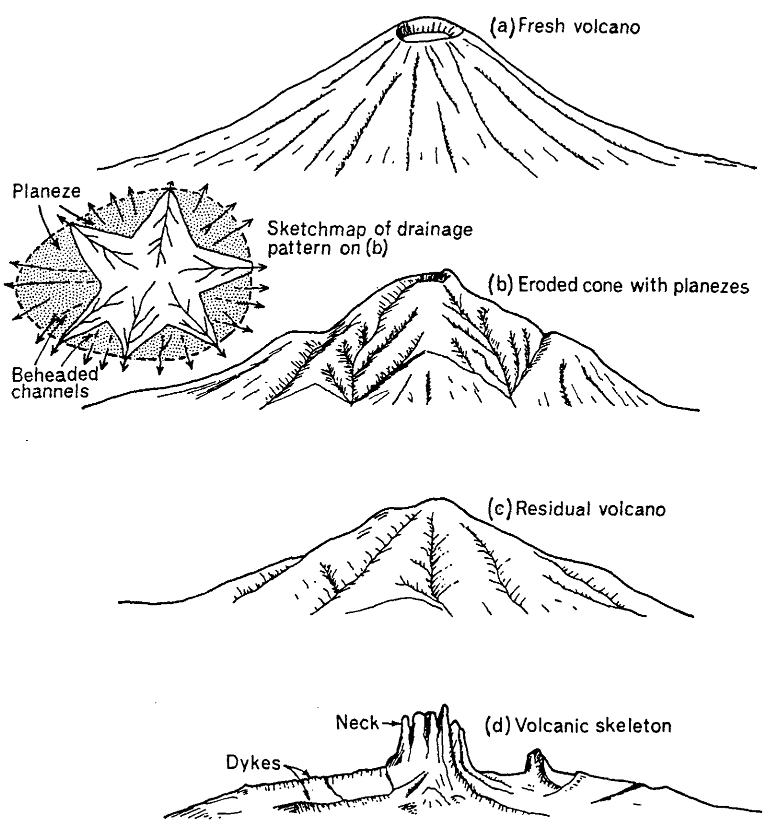


Figure 3.5: The successive stages in the erosion of a large stratovolcano to leave a volcanic neck. The inset shows the drainage pattern where planeze surfaces are present. *Source*: Selby 1985.

When eruption activity is intermittent, ravines or gullies may be entrenched into the flanks while the volcano is dormant. Commonly, these result from landslides (for example, Plate 3.5). Landslides occur when the angle of repose of regolith material is reduced to below that which it is required to support the regolith. There are three main ways in which landslides in volcanic areas, particularly in basalt flows, occur:

1. the internal strength of the material may diminish by gradual loss of particle cohesion;
2. the strength of the material may be weakened by vegetation loss and subsequent exposure to weathering and erosion; and
3. loss of effective strength caused by groundwater pressure building up within the regolith material where ground water rising within the regolith leads to a loss of effective strength (Short and Blair 1986).

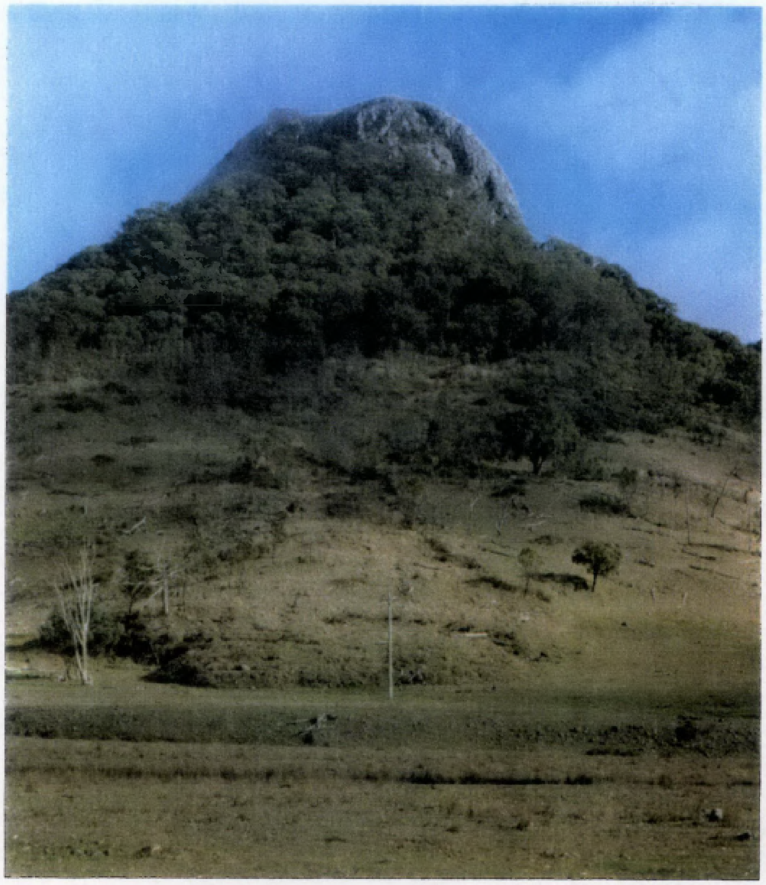


Plate 3.5: One of several small landslides at the base of Mopra Rock, a volcanic plug preserved on the eastern flank of the Warrumbungle Complex, New South Wales. The power pole at the toe of the slide provides a scale. The lack of established vegetation on the slump suggests that it occurred in the recent past. *Photo*: A. Timmers.

In susceptible locations, these factors may cause the loss of sufficient competence to cause sliding. Most landslides associated with volcanic features occur on the slopes and benches of the flanks of the shield which tend to be terraced because they are the truncated ends of multiple thin flows. These flows usually consist of individual stacked lavas, with zones of weathering, alluvial deposition, or pyroclastic deposits in between. In such cases the interbedded soft material is a line of weakness exploited by erosion, spring sapping and undercutting (for example, Plate 3.3). The many joints in basalt lavas render it permeable, allowing rainfall to seep through so that springs may emerge at the base where it overlies impermeable strata (Ollier 1988). These springs erode headward, giving rise to vertical-sided, round-ended side valleys (for example, the alcoves and springs of the Snake River, Idaho, USA).

On a smaller scale to landslides, tephra deposited on steep-sided cones tends to be re-deposited fairly quickly by individual particle free-fall and tumbling, by creep, or by grain-flow (Cas 1989). When the activity is renewed, radial gullies may be filled with tongues of lava or pyroclastic material. These tongues will then become divided and new valleys may be excavated along their sides on the sites of earlier ridges, causing relief inversion (Pain and Ollier 1995) of the volcanic flanks. The infilling of a valley by a lava flow causes, at a larger scale, reorganisation and relocation of the drainage system of that valley (for example, Ollier 1969, 1988).

Continued erosion of a volcano or volcanic complex involves drainage adjustment, with streams cutting through the various lithologies until only the very resistant geological components remain. These components usually take the form of constructional remnants such as erosional necks, plugs and domes which may stand in association with ring or radial dykes (Figure 3.5d; Plate 3.1).

Kear (1957) used planeze to skeletal erosion stages as a relative dating technique in New Zealand, where he proposed that the stage of erosion corresponds to an age. In his study, planeze development occurred in upper Miocene time, a residual stage was reached by Plio-Pleistocene time and the skeletal stage, where plugs and necks are preserved, developed in the mid-Pleistocene to Holocene. However, while erosion following the cessation of activity produces a regular sequence of form through to a planeze stage, a residual stage and a final skeletal stage, different stages may be preserved in different parts of the complex at the same time (for example, planeze surfaces may still be preserved even though skeletal plugs are exposed). Thus, older volcanoes may fail to fit in with age-stage and morphometric relationships because they are highly eroded or simply do not fit the patterns of 'typical' volcanic denudation. For example, the Miocene volcanoes of Uganda do not exhibit planeze development because they are built by a succession of agglomerate deposits (Table 3.5) separated by ash layers. Radial drainage is present, but instead of planeze surfaces the flanks consist of a series of steps. In addition, relative age determination is also hampered in complex volcanoes because of repeated eruption on the same centres in which the time between eruptions may be up to millions of years. However, landforms may still be classified on the basis of morphology. Indeed, Twidale (1976) defined the relative ages of four physiographic provinces in northern Queensland on the basis of the degree of dissection of the margins of lava sheets, the abundance of recognisable eruptive centres and the development of *in situ* soils.

3.8.2.2 Drainage development in craters and on lava plains

By contrast to the radial drainage of volcanic cones, the drainage of craters or calderas is centripetal. Streams run from below the crater rim towards the centre of the vent. If a crater lake or associated centripetal drainage breaches its walls, then the original centripetal drainage simply becomes an appendage at the head of the main stream that drains the breach (Ollier 1988). However, if the crater is breached by the headward erosion of the crater by radial drainage rather than by a lake, then the radial stream will capture the crater and the drainage it contains. By contrast, the drainage of lava plains is largely due to drainage displacement, and the drainage of ignimbrite plateaus is largely parallel due to long, uniform slopes on homogenous material. Drainage density of volcanic areas is low on lava plains, due to the porous nature of the lavas, but is quite high on ignimbrite plateaus largely due to the impermeability of welded ignimbrites and the mechanisms of valley formation on these plains. A detailed discussion of these is unwarranted here since the study area lacks these features. Unlike the straightforward nature of drainage density on lava plains and ignimbrite plateaus, the drainage density of shield volcanoes depends on a range of factors including variations in topography, distribution of lithology, erosion resistance, permeability of rock and the distribution of rainfall. Because drainage evolves differently in different volcanic settings, the analysis of drainage development in volcanic areas can therefore be a useful tool in elucidating sequences of landscape evolution. For this reason, the investigation of drainage development in the Warrumbungle Complex forms an important component of this study.

3.9 Climate as a geomorphic agent

The relationship of landforms to climate is a topic that has generated considerable discussion from geomorphologists in past decades, although it is no longer fashionable. Climate factors are known to influence the nature and rate of geomorphic process, leading some geomorphologists to suggest that different climates are associated with characteristic landform assemblages (for example, Tricart and Cailleux 1972). It has been argued that climatic influences can have an effect on geomorphic processes sufficient to outweigh the influence of tectonic setting, rock type and relief on landform development. However, most elements of the landscape are likely to be, to a greater or lesser extent, out of equilibrium with prevailing climatic conditions because of the magnitude and rapidity of global climate change (for example, relict landforms; Section 1.5). Indeed, reviews of processes and landforms associated with fluvial, glacial and aeolian environments demonstrate that it is only major differences in climatic regime that have marked effects on landscape development. When landscapes are considered as a whole, other factors such as tectonics and lithology often predominate (Summerfield 1991). This is because geomorphic processes tend to be intensified in tectonically active regions. Tectonic activity may reach such a magnitude that a qualitative change occurs, such as in the case of large-scale mass movement triggered by earthquakes and rapid constructional alteration of topography as a result of volcanism.

Despite this dominance by tectonic control, the reconstruction of past climates from the sedimentary record has proved useful over recent decades in determining components of landscape evolution. The application of palaeoclimatic interpretations should not be confused with traditional climatic geomorphology in which climate is thought to be the dominant factor in landscape evolution. Rather, palaeoclimatic interpretations enable researchers to define climatic parameters at the time of deposition and use them, in conjunction with other evidence, to produce a clearer picture of landscape evolution that considers a range of processes and forms (for example, tectonics, climate, biogeography, relict landforms). With this in mind for later discussion, a review of the Miocene climate in Australia is presented in Appendix A. An assessment of climate is relevant to three aspects of volcanology:

1. weathering of the tops of volcanic units before they are covered by the next unit;
2. deep weathering of the volcanic pile; and
3. the rate of denudation, which is dependent on rainfall, lithology and vegetation cover, and which controls the proportion of volcanic material remaining at the present day (Wellman 1989).

3.9.1 Palaeoclimatic indicators and their interaction with volcanism

Climate is a controlling factor in the development of volcanic stratigraphy, especially in pyroclastic materials. However, as a controlling factor, it is difficult to resolve, particularly in a palaeovolcanic setting, unless environmental indicators, such as contemporaneous fossiliferous deposits, can be utilised. Traditionally, the study of volcanism and the study of fossil biotas have been considered to be two very different disciplines. This is especially the case when igneous rocks are generally believed to be unfossiliferous. Despite this, fossils have been found to occur in association with volcanigenic deposits from Precambrian to Recent age. World-wide, volcanism has been found to contribute directly, through rapid burial (for example, the A.D. 79 eruption of Italy's Mount Vesuvius buried Pompeii under ash, pumice and nuées ardente, entombing Pompeii's citizens) and indirectly, through diagenic processes, to the preservation of both floral and faunal remains. Thus, volcanism significantly affects the biostratigraphic record by preserving fossil biotas, and by modifying the palaeoenvironments in which they evolved.

Buesch (1988) has noted that the damming of local drainage by materials of volcanigenic origin can create new lakes and subaerial environments. Such areas are focal points for the development of new biotic communities. Because these points are depocentres, they are likely to be conducive to the accumulation of representative biotic remains in the fossil record. The most obvious influence of localised volcanic activity which facilitates fossil preservation is the abundance of volcanic ash produced by volcanic centres. The quantity of volcanigenic sediments supplements the normal supply of clastic sediments and provides a wide range of fluvial deposits, including deltaic and varved deposits in lakes and the volcanic equivalents of sandstones and siltstones. These deposits are ideal for the preservation of megafossils. While ash falls contribute directly to depositional environments in this way, they also contribute to fossil assemblages indirectly. The drainage of water over ash-covered landscapes and its percolation through pyroclastic deposits dissolves silicon and other elements that are not otherwise available in such concentrations, fostering blooms of siliceous diatoms (Taggart and Cross 1990). Such blooms may result in the preservation of plant and animal materials in diatomite.

3.9.1.1 The application of diatoms in reconstructing environmental and geomorphic change

Diatoms (single-celled siliceous algae, class Bacillariophyacea, Appendix B) have proved singularly successful in reconstructing environmental change. Since diatoms can be identified to species level, it is possible to use their known environmental tolerances to solve a wide range of palaeoecological problems. While most diatom research targets Quaternary sediments or present day communities, it is possible to apply Quaternary analysis and diatom methods to the sediment/diatomite of earlier epochs, since frustules are generally well preserved. Traditionally, palynology has been by far the most widely used analytical method for the reconstruction of past environments, with over one hundred Australian late-Quaternary palaeoecological records constructed, largely from palynology studies (Bleys *et al.* 1991). There are relatively few other examples of the use of palaeoenvironmetal indicators in Australia. By contrast, studies of environmental change in Europe and North America have a long history of the application of floral and faunal indicators to evaluate environmental change (Smol 1990). The use of indicators such as diatoms, ostracods, chyrsophytes and chironomids has, in recent years, shown the versatility of these groups in answering a wide range of questions regarding environmental change.

The integration of diatomological, palynological, sedimentological and macrofossil evidence allows the detailed reconstruction of the palaeoecology of an area. Given that volcanic activity at any specific site will be episodic, the preservation of fossil assemblages will be closely linked to periods of activity in temporal terms. Typically, any activity of a given volcanic centre will involve spurts of eruptive activity punctuated by periods of quiescence. Therefore the pattern of volcaniclastic deposits, including fossils, will reflect the temporal nature of eruptions. Thus there may be exposures of little or no stratigraphic continuity. Continuity is further hampered by the extent to which local volcanism diverts drainage, creates new water bodies and breaches existing bodies. Further discontinuity may be inherent in the stratigraphical record.

3.9.1.2 The value of diatoms for palaeoenvironmental reconstruction

In volcanic areas, denudation can rapidly alter the original constructional form of a volcano. The rate of modification will be a function of the resistance to weathering and erosion of volcanic material, the initial relief created by eruption processes, and the prevailing climatic environment. While Appendix A has outlined the prevailing Miocene climate in Australia, there is no clear account of the true (arid or humid) nature of the climate in the mid-Miocene. One aspect of climate interpretation that could yield light on this is the use of palaeolimnology, more particularly, the use of lacustrine deposits for palaeoenvironmental interpretation. In the Warrumbungle Complex, there are several significant lacustrine deposits that are contemporaneous with eruption (Chalk Mountain, Paddy McCullochs Mountain and Wandiallabah Creek; Figure 1.2). These deposits contain diatomite sequences interbedded with volcanic material that can be used to elucidate the nature of the environmental conditions prevailing over the period of eruption. This palaeoenvironmental data can then be used to extrapolate the role of climate in the geomorphic development of the Warrumbungle Complex. Diatoms are useful palaeoclimatic indicators because they have defined ecological tolerances. As such, they provide considerable information about water quality, pH, alkalinity, salinity and nutrient status and can be used to interpret past environmental conditions (Appendix B). Diatom deposits are intimately associated with the Warrumbungle Complex. They therefore help in developing the geomorphic history of the Complex because they provide information about past conditions, including the periodicity of volcanic activity and, more importantly, yield information on geomorphic parameters operating at the time of deposition. The nature of diatomite deposits in the Warrumbungle Complex is discussed in Section 4.4.3.4.

3.10 Conclusion

Volcanic eruptions are difficult to classify. Some are violent and of short duration, but many others may be active for protracted periods and vary in character in rapid and unpredictable ways. Because eruptions are complex in nature, especially in polygenetic or coeval centres, the style of volcanism is of fundamental importance to subsequent landscape development. The type and style of activity will determine the constructional topography of source areas, and the volume of pyroclastics and lavas will dictate the extent of eruption influence on drainage and sedimentation and the rate of subsequent landscape denudation. Establishing the dynamics of volcanic landscapes (eruption mechanisms, landforms, drainage) through geomorphic analysis provides a useful tool for the elucidation of processes that have operated in palaeovolcanic settings. Interpretation of such processes allows an understanding of pre-volcanic and post-volcanic landscape development within a regional setting, as well as having a wider application of refining the morphotectonic development of intraplate volcanic settings. With this in mind, the nature of the present Warrumbungle landscape is discussed in Chapter 4 in order to establish the nature of the current topography before the geomorphic history of the area is interpreted through palaeovolcanic reconstructions.