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Practical Volcanology

Lecture Notes for Understanding Volcanic Rocks from Field Based Studies

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Front page of cover: Overview of the Ruapehu volcano in the Central North Island of New Zealand. Tama maar lake is in the foreground.
Back page of cover: *Upper left and right:* Explosion craters along the Tarawera–Rotomahana volcanic fissure system, Taupo Volcanic Zone, New Zealand. *Middle left:* Hot spring in the Tarawera–Rotomahana volcanic fissure system, Taupo Volcanic Zone, New Zealand. *Middle right:* The Rainbow Springs in the Whakarewarewa thermal areas in the Taupo Volcanic Zone, New Zealand. *Lower:* Overview of the Mt Ngauruhoe volcano from the north, Taupo Volcanic Zone, New Zealand.

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Preface

Volcanic rocks are important in compiling geological records because of their characteristic chemistry, relatively fast accumulation and great variety; however, recognizable facies diversity may be useful for reconstructing not only the volcanic processes but also the eruptive environment where the volcanism take place. Volcanic rocks that are significantly fragmented are important from a stratigraphic point of view and they can be used to study palaeoenvironments where these volcanic deposits formed. The increasing importance of fragmental volcanic rocks in geological research is clearly demonstrated by the increasing number of publications that have appeared over recent decades dealing with volcaniclastic deposits and rocks. Different volcanological schools and associated textbooks have been published since the 1980s. Among the many that have become available four are of particular significance These are Fisher and Schmincke (1984): Pyroclastic Rocks; Cas and Wright (1987) Volcanic Successions; McPhie et al. (1993) Volcanic Textures; and Sigurdsson et al (2000) Encyclopedia of Volcanoes. The aforementioned are among the many textbooks that are widely accepted and used in volcanology courses at different levels. The volume Practical Volcanology, as a textbook, does not intend to substitute any of the above books; rather, it tries to deal with volcanic geology from a slightly different aspect from those already cited. Practical Volcanology is a direct result of a series of short courses offered for first time in 2001 at the Geological Institute of Hungary, Budapest, primarily for geologists working in ancient volcanic terrains, and their main aim is general mapping. In addition, these short courses also intended to draw the attention of undergraduate students, postgraduates and research students who came across volcanic rocks during their research. The basic idea of Practical Volcanology is included in a study guide and lecture notes which could be used as a self-standing guide for interpreting volcanic processes and the resulting deposits and rocks. To take full advantage of this book a preliminary geological background is necessary for the user, especially in the field of classic sedimentology, petrology and geochemistry. However, a limited background of geological knowledge would enough to get a basic idea of field-based volcanology in its simplest aspects.

The book's main aim is to introduce basic field volcanology research from a theoretical point of view right through very practical elements. The basic philosophy of the book is that, especially in ancient terrains, the volcanologist's basic data is found through fieldwork, and they are looking for volcanic rocks, especially fragmented ones. This book intends to demonstrate the link between the field subject, a volcanic rock and the volcanic process that may have formed that rock. Such textbooks or study guides are relatively rare these days and often they are too detailed or complicated for undergraduate students or interested amateurs.

This book consists of 8 chapters. Each chapter is fully referenced in order to give a very detailed guide to any user and it clear where the individual citaitons/statements come from. This allows the user to go deeper into the scientific problems such processes, deposits, or the relevant terminology itself. Each chapter is accompanied with figures widely used and referred to in the international literature and there are full colour plates of textures, volcanic activity and the 3D architecture of volcanic deposits. The figures and colour plates are fully explained and referenced. In addition, each chapter has a locality map allowing the user to identify the site locations for future references. At the end of the book there is a detailed glossary along with a collection of terms from widely accepted textbooks, articles, and web resources. The book also contains a detailed index for quick search through the chapters for key volcanological terms.

The 8 chapters set a logical path from an introduction, a key of terminological issues right through to different volcanic processes. The first chapter deals with a short summary and referenced description of major volcanic terminological systems. This chapter also gives a detailed insight of the usage of different terminologies and their potential for future research documentation. The second chapter is a detailed summary of active volcanism and its relationship to volcanic deposits. This chapter intends to make clear the connection between active volcanism and the volcanic rocks that most mapping geologists deal with in the field. The third chapter focuses on fragmented volcanic rocks. Beside its classification scheme and a presentation of the common features of fragmented volcanic rocks this chapter provides a clear guide about the information which can be obtained from fragmented volcanic deposits and rocks. This chapter also gives indications of the limitation the information with respect to its use for inferring volcanic processes and eruptive environments. The fourth chapter gives an introduction to volcanic facies analysis which one of the main goals of studying volcanic rocks in the field. Volcanic facies analysis is the basic tool for broad making interpretations and can be connected to palaeoenvironmental reconstructions. The fifth and sixth chapters concentrate on summarising volcanic processes and the resulting volcanic deposits and rocks which are associated with the two major types of volcanism on Earth: i.e. monogenetic and polygenetic volcanism. In these two chapters not only field examples are given but also a large collection of young deposits and volcanic processes are examined to demonstrate clearly the connection between volcanic processes and the resulting deposits and rocks. The seventh chapter deals with processes which act on volcanic terrains and which can significantly modify the original primary volcanic landforms. Also in this chapter a basic concept - derived from those few studies dealing with the topic - of the erosion of volcanic terrains is introduced. The eighth chapter gives a concise summary of the potentially most widespread, but less known type of volcanism which occurs in subaqueous environments. Probably in ancient terrains the majority of volcanic rocks represent deposits that may have formed in some sort of subaqueous environment. In addition this type of volcanism has the potential to generate volcanic deposits that can host valuable ore minerals.

The book is based on the expertise of two authors gathered over the past 15 years of their work in the field of volcanic geology. The authors have primarily used their own research data to demonstrate key features but where useful these have been collated with other field information from other researchers. The majority of the field and textual data has been provided by the authors. The figure collection is based on published and usually well-accepted research papers or textbooks in order to facilitate the user's ability to connect their own work to individual researchers and their publications.

Practical Volcanology is a study guide which it is hoped will provide a good basis for developing short courses.

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Chapter 1

Terminology of fragmental volcanic rocks

Introduction

Volcanic eruptions can produce large volumes of coherent rocks and/or clastic debris. Volcanic rocks hence consist of coherent ones simply solidified from a melt, and clastic ones that form through a wide range and combinations of different style of fragmentation, transportation and deposition processes. This diversity of processes maybe involved in the formation of a volcanic rock naturally makes difficult to describe and interpret them. Since the late sixties a dramatic advance has taken place in understanding volcanic rocks, and therefore a great variety of description and classification schemes have been formulated. The classification of clastic (fragmental) volcanic rocks, generally named volcaniclastic rocks, became a subject of debates and source of conflicting ways of dealing with their description and classification. The basic problem in describing the volcaniclastic rocks is the need to find a balance between purely descriptive documentation of the rock/deposit itself, while concisely and consistently reflecting their volcanic origin. In past decades many attempts have been made to find a middle ground. The other important problem in classification of volcaniclastic rocks is to express their relationship with the primary volcanic processes and/or to distinguish clearly whether the rock/deposit is of primary or secondary origin. This issue is complicated by the fact that many traditional rock names carry genetic connotations for most workers. In this way, a "lapilli tuff" may be considered suggestive of a primary origin, and "mud" or "sand" would be suggestive of a "normal" sedimentary origin; this despite the fact that terminologies in common use define these terms almost purely in terms of the grain-size characteristics of a rock/deposit. Volcaniclastic rock terms in general apply to rocks consisting of volcanic fragments of any origin, and having any fragmentation, transportation or depositional history (e.g. Fisher 1961). Pyroclastic rocks are understood to be those consisting of pyroclasts. There are at least two commonly used definitions of pyroclasts, however, which is a major issue in consistent description of volcaniclastic deposits. Pyroclasts are defined by Fisher and Schmincke (1984) as fragments entered to the transporting media and to the depositional site through a volcanic vent during volcanic eruption-fed processes, e.g. fragments that originate from volcanic eruptions or as a direct consequence of an eruption (Fisher and Schmincke 1984). Regardless of this definition, many workers define pyroclasts more specifically as particles formed only in explosive eruptions driven by magmatic gas expansion (Fisher and Schmincke 1984). This use of the term, however, is problematic because many important volcanic processes that produce large volume fragmented volcanic rocks would not produce "pyroclasts" of this sort, including pyroclastic flow deposits formed during gravitational collapse of a lava dome (Yamamoto et al. 1993, Carrasco-Nunez 1999, Kelfoun et al. 2000, Robertson et al. 2000, Hooper and Mattioli 2001, Elsworth et al. 2004).

Presently, there are four major line of genetic classification of fragmented volcanic rocks. One of the oldest and still widely used terminological systems was introduced in the early sixties (Fisher 1961, 1966, Fisher and Schmincke 1984, 1994). This is the terminology used in the book titled "Pyroclastic Rocks" (Fisher and Schmincke 1984). In the late eighties another significant work compiled new volcanological data and extended the usage of various terms largely applying classification methods based on facies analysis schemes of the sort used in normal sedimentary environment (Cas and Wright 1987). This work is summarized in the book titled "Volcanic Successions" (Cas and Wright 1987). In the early nineties, urgent need generated a more logical genetic classification of fragmental volcanic rocks, based on the transportation and depositional processes formed the volcanic fragments (McPhie et al. 1993). This work culminated in a book titled "Volcanic Textures" (McPhie et al. 1993). Now therefore there are at least 3 different terminological systems widely used in the volcanology literature, causing confusion. Recent research, especially on explosive subaqueous volcanism (White et al. 2003), magma–water interaction driven phreatomagmatic explosive volcanism (Ross et al.

2005, Ross and White 2005b, 2005a, Martin et al. 2007), laharic systems (Manville et al. 2002, Segschneider et al. 2002, Manville and White 2003) and research on volcanic mass-flow deposits (Calder et al. 2000, Legros and Marti 2001, Roche et al. 2002, Freundt 2003, Hakonardottir et al. 2003, Felix and Thomas 2004, Lube et al. 2004, Schwarzkopf et al. 2005, Lube et al. 2007) highlighted the urgent need to unify the existing genetic classification of fragmental volcanic rocks. Very recently a new terminology system has been suggested, which combines elements of the previous classification systems into a simple and user-friendly terminological system (White and Houghton 2006) that will be discussed further below.

Two types of name definitions now exist for volcaniclastic rocks directly resulting from volcanic eruptions as primary volcaniclastic deposits, both predominantly based on the grain size characteristics of the rock/deposit. One uses terms initially reserved only for pyroclastic rocks (Fisher and Schmincke 1984), the other applies to all volcaniclastic rocks initially a clastic sedimentological terminology (Cas and Wright 1987, McPhie et al. 1993).

Before the existing and new terminological system are considered further, we outline basic textural characteristics of fragmented volcanic rocks. These basic textural characteristics, alongside the grain-size distribution of the fragmental volcanic rock/deposit are the main classification parameters used in classification of fragmental (clastic) volcanic rocks.

General components of volcaniclastic rocks

A fragmental rock is a mixture of different origin of clasts that came to rest together and form a deposit, and after diagenesis, a rock (Figure 1.1, Figure1.2). These fragments can be in various proportions in a single fragmental volcanic rock/deposit. Their proportion, composition and distribution patterns will make a specific rock texture, which is characteristic for the fragmentation, transportation, deposition, and alteration history of the fragmental volcanic rocks. The major fragment types of a fragmental volcanic rocks are juvenile fragments, accidental lithic fragments, and accessory lithic fragments (Fisher and Schmincke 1984, Cas and Wright 1987, McPhie et al. 1993). This classification scheme is widely used, though a new componentry classification suggested very recently divides fragments in a fragmental volcanic rock into juvenile, lithic and composite clasts (White and Houghton 2006). In the next section we describe the

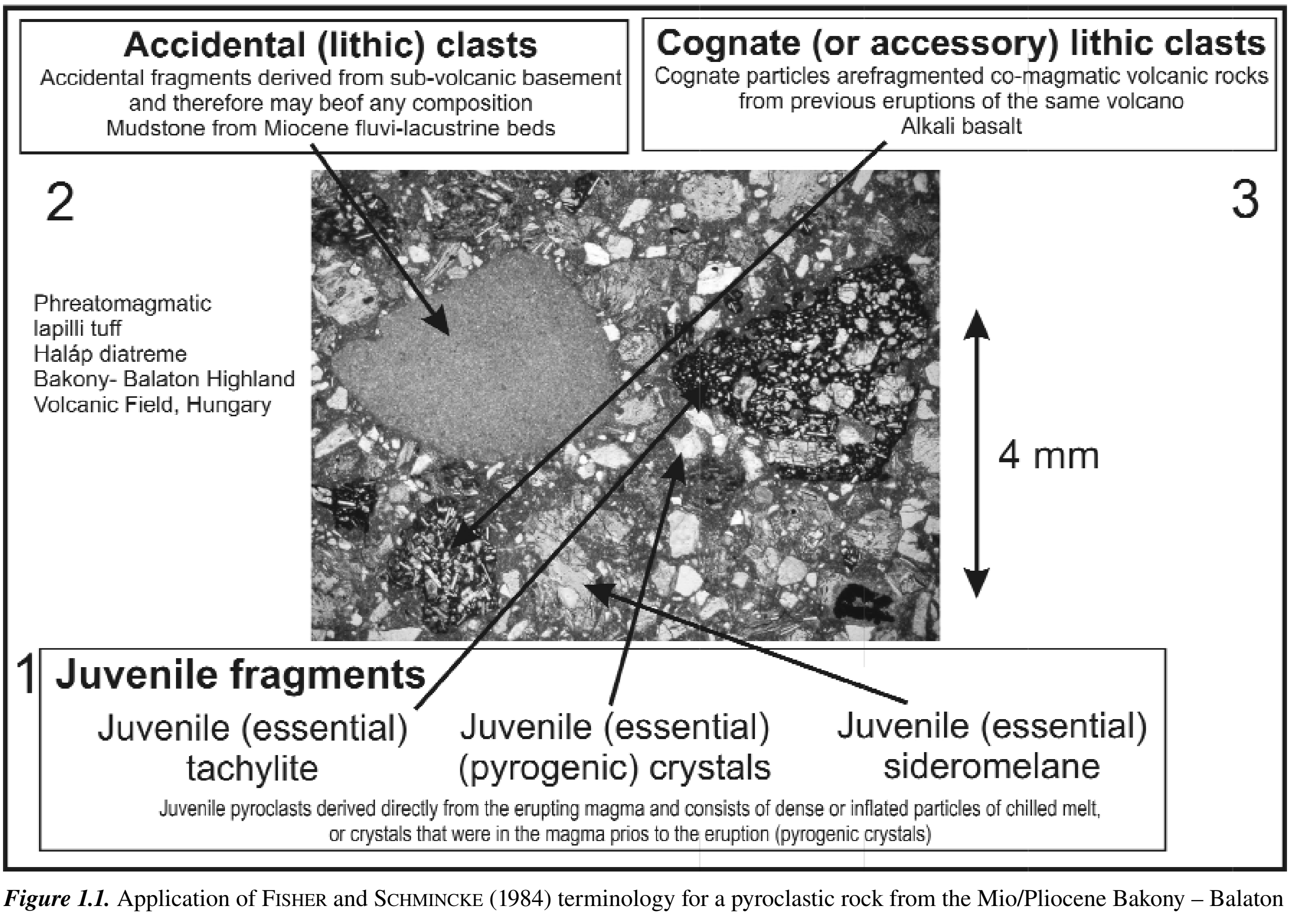

Figure 1.1. Application of Fisher and Schmincke (1984) terminology for a pyroclastic rock from the Mio/Pliocene Bakony – Balaton Highland Volcanic Field maar/diatreme remnant (same sample as on Figure 1.2)

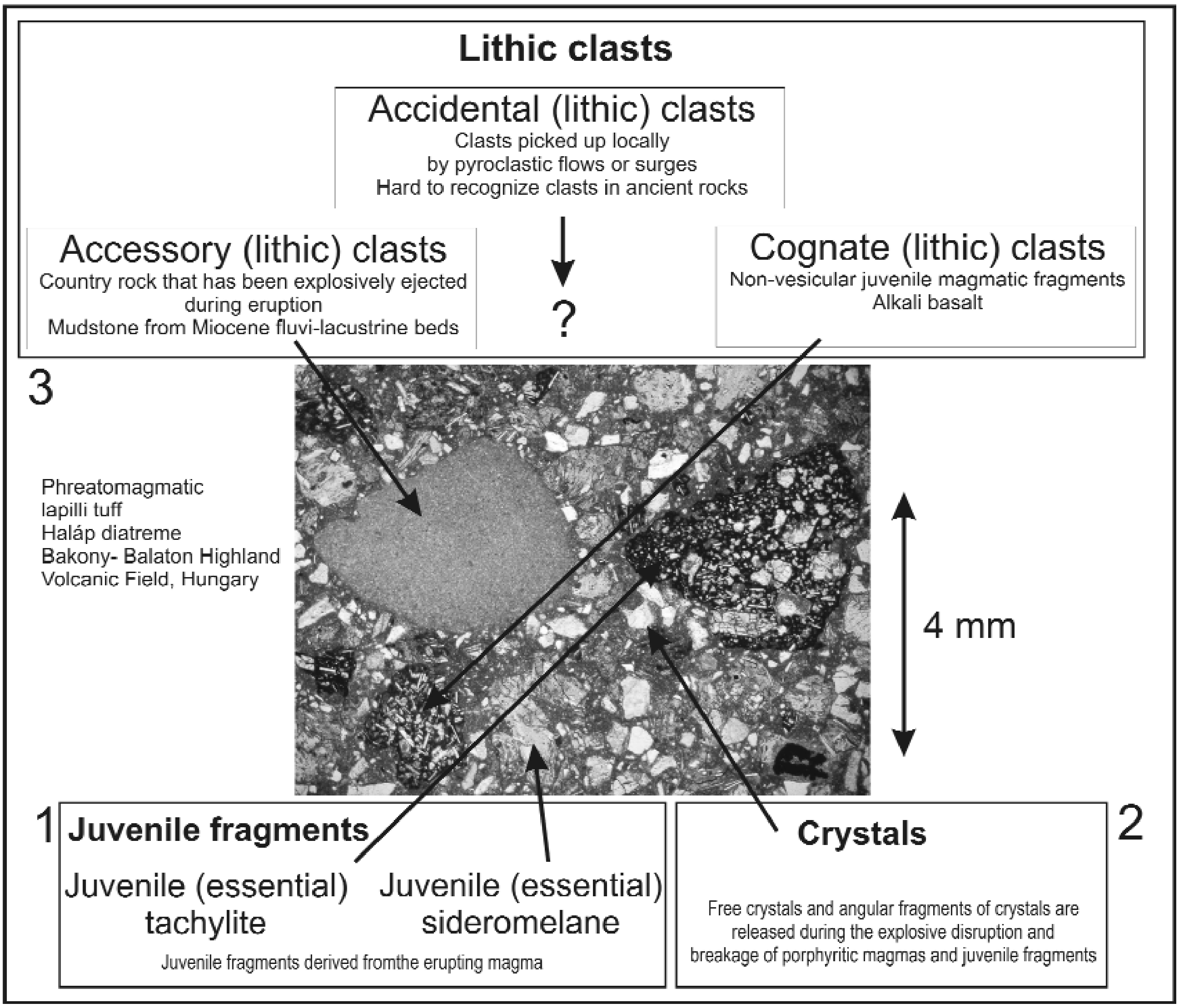

Figure 1.2. Application of CAS and WRIGHT (1987) terminology for a pyroclastic rock from the Mio/Pliocene Bakony – Balaton Highland Volcanic Field maar/diatreme remnant (same sample as on Figure 1.1)

two major traditionally used classification system of the componentry of fragmented volcanic rocks (FISHER and SCHMINCKE 1984, CAS and WRIGHT 1987, MCPHIE et al. 1993). Later, under a separate section we describe the recently suggested componentry classification scheme (WHITE and HOUGHTON 2006).

Juvenile fragments

Juvenile fragments (Plate I, 1) are considered to be derived directly from the erupting magma, and consist of dense or inflated particles of chilled melt, or crystals that were in the magma prior to the eruption (FISHER and SCHMINCKE 1984). The juvenile fragments are commonly distinguished in accordance to their appearance. Such distinction in mafic volcanism separates juvenile fragments into tachylite, sideromelane and crystals (FISHER and SCHMINCKE 1984). Tachylite is a dark volcanic glass, charged with opaque minerals (Plate I, 2). Generally, the presence of tachylite indicates slow cooling of the melt after fragmentation (e.g. aerial transportation system) (FISHER and SCHMINCKE 1984). Sideromelane (Plate I, 3) is a chilled mafic melt, and therefore glassy, transparent (FISHER and SCHMINCKE 1984). Their presence indicates rapid cooling, chilling, such as magma cooled in vigorous fire fountains (e.g. Pele's hair) or/and contact with water (FISHER and SCHMINCKE 1984). Depending on the timing of magma vesiculation, the sideromelane glass shards can be vesicle free, or charged with various shapes and sizes of vesicles. Sudden cooling could also been reflected in collapsed shape vesicles (TADDEUCCI et al. 2004) (Plate I, 4). Given its low viscosity, bubbles in basaltic melt collapse shortly after fragmentation, and the presence of well-developed, round vesicles with thin septa in the sideromelane ash is more readily explained if bubble expansion and coalescence was still in progress when the clast quenched (TADDEUCCI et al. 2004).

Conversely, the lack of well-developed vesicles in the tachylite suggests that gas bubbles had already escaped from the melt or collapsed when the particles quenched (TADDEUCCI et al. 2004). If partial crystallisation of the melt occurs more or less at the same time as the magma is chilled, sideromelane can contain various amounts of microlites (small crystals, typically lath-shaped plagioclase). The relative proportion of tachylite and sideromelane in clasts of a single rock sample can reflect the timing of magma fragmentation in relationship to the time of magma–water interaction, vesiculation, and crystallisation (HOUGHTON and HACKETT 1984, HOUGHTON and SCHMINCKE 1986, WHITE 1991, 1996a, 1996b, HOUGHTON et al. 1999, NÉMETH et al. 2001) (Plate I, 5). Juvenile clasts of more evolved magma compositions can be less informative in regard of their cooling history from simple microscopic observation. Pyrogenic crystals (Plate I, 6) are considered to be juvenile fragments in FISHER and SCHMINCKE (1984) and represent the crystal fraction of the crystallizing melt prior to fragmentation. In the classification of CAS and WRIGHT (1987) the crystals are defined as free crystals and angular fragments of crystals that were released during the explosive disruption and breakage of porphyritic magmas and juvenile fragments (Figures 1.1 and 1.2).

Accidental and accessory lithic fragments

Accidental (lithic) fragments defined by FISHER and SCHMINCKE (1984) are derived from the sub-volcanic basement (Plate II, 1) and therefore may be of any composition (Figure 1.1). In the same classification scheme cognate (or accessory) lithic fragments are defined to be fragmented co-magmatic volcanic rocks from previous eruptions of the same volcano (Plate II, 2). In the CAS and WRIGHT (1987) terminology accessory lithic fragments (Figure 1.2) are defined to be country rocks that have been explosively ejected during eruption. Accidental lithic fragments according to CAS and WRIGHT (1987) (Figure 1.2) are clasts picked up locally by horizontal moving currents such as pyroclastic flows and/or surges. CAS and WRIGHT (1987) define cognate lithic fragments as non-vesicular juvenile magmatic fragments.

Bedding characteristics

The most important bedding characteristic is the bed thickness. Widely used bed thickness categories in volcaniclastic sedimentology are the same as for normal sedimentary deposits/rocks (INGRAM 1954) (Figure 1.3). Thinly laminated to thinly bedded deposits are commonly associated with distal tephra successions or deposits formed from pyroclastic density currents (Plate II, 3 and 4). Important classification categories commonly used in volcaniclastic sedimentology refer to the internal texture of the bed such as massive (e.g. no internal lamination, or other characteristic features such as grading, dish structures, etc.) (Plate II, 5) or weakly to moderately defined beds (Plate III, 1). These textural features carry important information about the transport agent, such as physical aspects of flow including rheology and particle concentration. Bed continuity also an important classification parameter in volcaniclastic sediments, which can be parallel bedded (Plate III, 2), strongly undulating (Figure 1.4), or dune-bedded (Figure 1.5). These types of bedding charac-

Name	Thickness
Very thickly bedded	> 1 m
Thickly bedded	30–100 cm
Medium bedded	10–30 cm
Thinly bedded	3–10 cm
Very thinly bedded	1–3 cm
Thickly laminated	0.3–1 cm
Thinly laminated	<0.3 cm

Figure 1.3. Bed thickness categories widely used in sedimentology after INGRAM (1954) (in FISHER and SCHMINCKE 1984: p. 108, table 5-5)

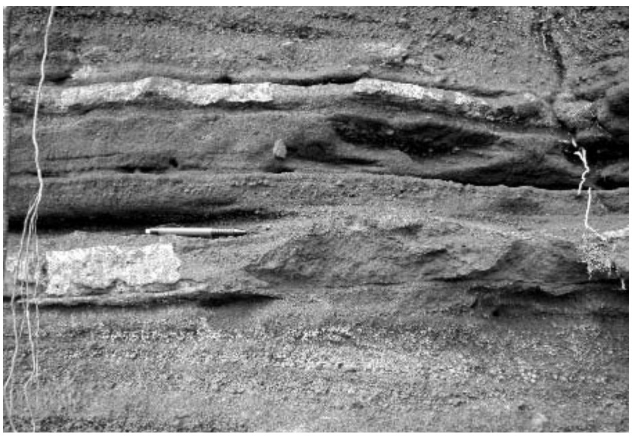

Figure 1.4. Undulating tuff bed (see under the pen) from a tuff ring erupted in 1913 in West Ambrym, Vanuatu. Pen is 15 cm long

Figure 1.5. Dune bedded lapilli tuff succession (upper part of the section) of a tuff ring erupted in 1913 in West Ambrym, Vanuatu. Hammer is 30 cm long

teristics are also indicative of the physical properties of the depositing agents. The particles' vertical distribution pattern, expressed as grading in the deposit, can be very complex (Figure 1.6). The most common grading types are normal grading (Figure 1.7), reverse-to-normal grading and "pure" reverse grading (Plate III, 3). Because of the commonly complex componentry of a volcaniclastic bed, grading can be complex in comparison to that in normal clastic sedimentary beds, and is commonly reflected in density grading instead of strictly size-class-defined grading. This is especially common

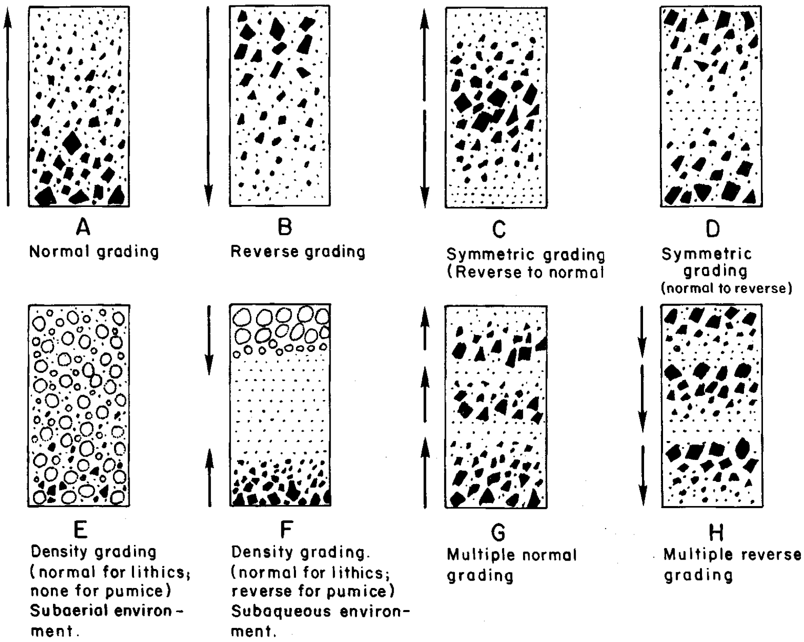

Figure 1.6. Typical grading types after FISHER and SCHMINCKE 1984: p. 109, fig. 5-19

Figure 1.7. Normal graded pumiceous lapilli beds from the Taupo Volcanic Zone, New Zealand. Individual bed starts below the peak of the hammerhead

in pumiceous pyroclastic density current deposits. Cross-bedding is especially important in interpretation of pyroclastic density current deposits' physical properties, and their types can be associated with the current flow regimes (Figure 1.8). Cross-bedding is an important feature in pyroclastic density current deposits of many types (e.g. not important in block-and-ash flow deposits, but very important in pyroclastic surge deposits of any type), but deposition from traction during strong current movement by wind or aqueous currents can also generate cross-bedding (Plate III, 4). Sorting is a description of the size distribution pattern of the deposit/rock (Plate IV, 1), of the unimodal to complex distribution of various grain-size classes, in a single deposit/rock. Well-sorted deposits/rocks have a well defined

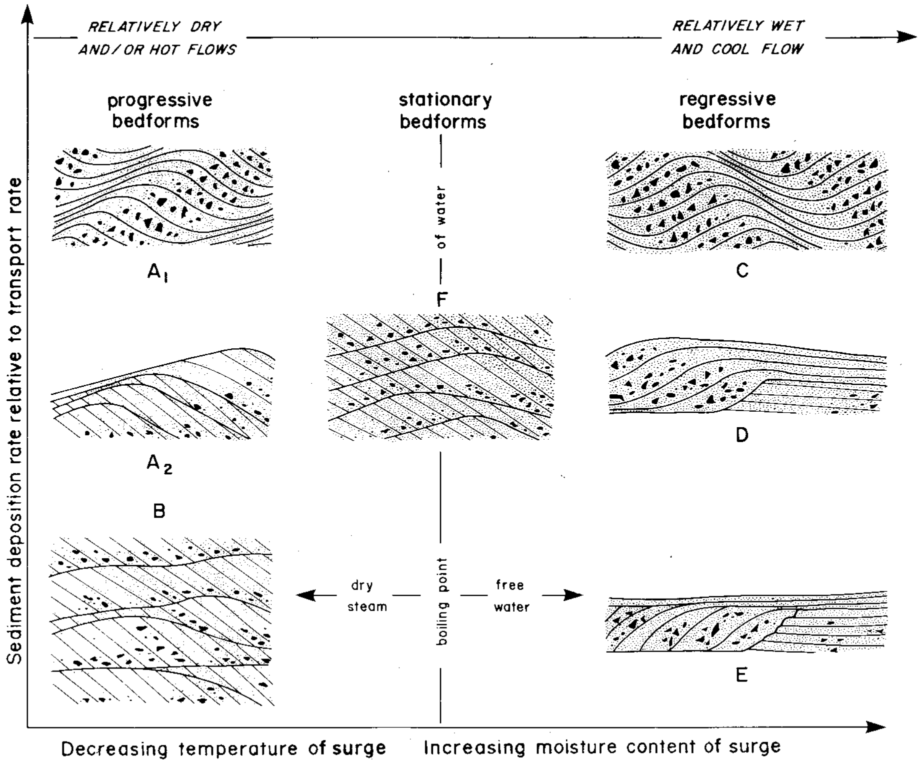

Figure 1.8. Types of pyroclastic surges bedforms and internal cross-stratification (after ALLEN 1982 in CAS and WRIGHT 1987: p. 215, fig. 7.42) as function of depositional rate and surge temperature and moisture content

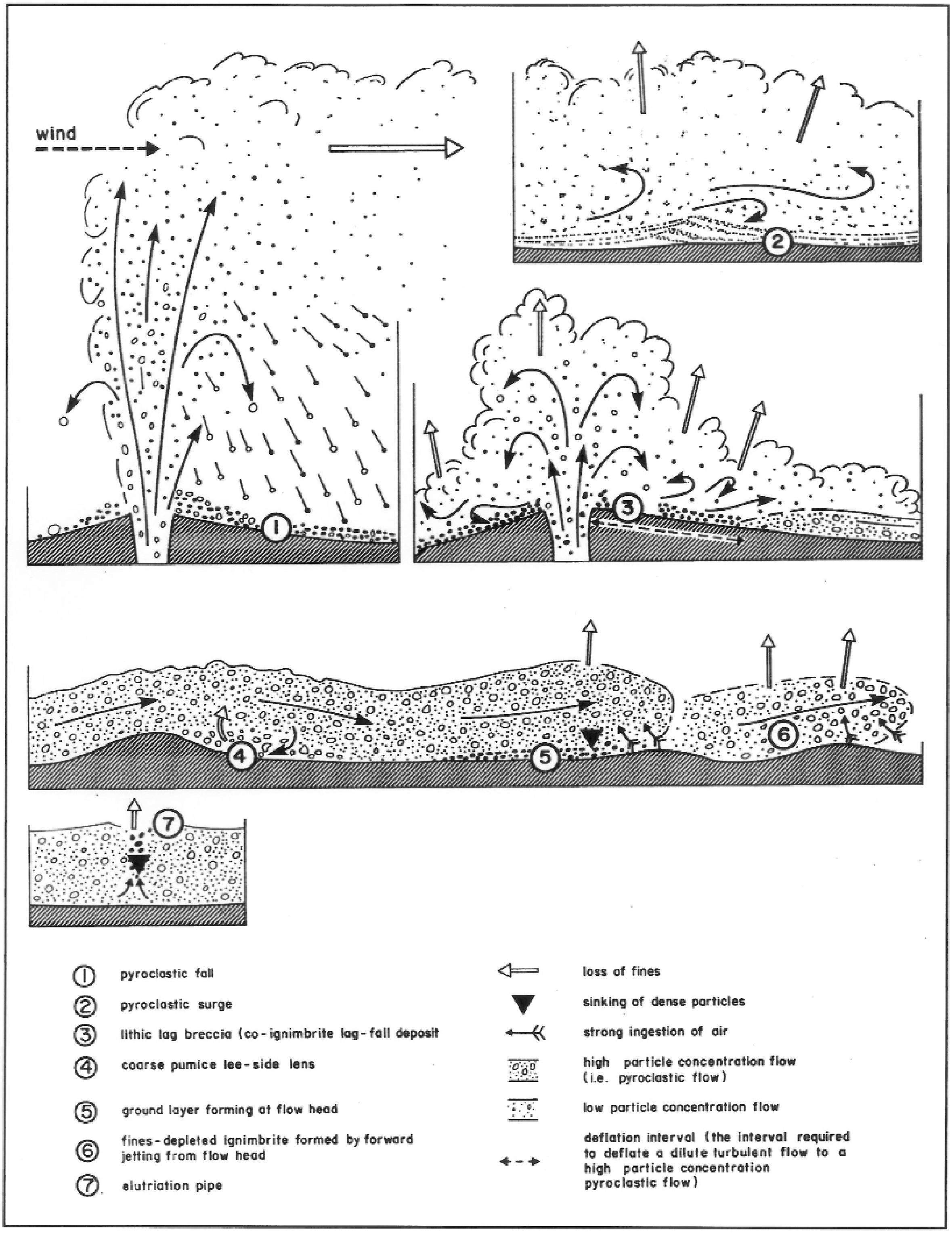

Figure 1.9. Theoretical models of development of good sorting (after WALKER 1983 in CAS and WRIGHT 1987: p. 220, fig. 7.46)

clast population forming the major volume of the deposit/rock (Plate IV, 2). Poorly sorted deposits/rocks are those which have a wide size range (Plate IV, 3). In volcanic deposits good sorting can develop in many way (CAS and WRIGHT 1987), but well-sorted deposits are uncommon, and very poorly sorted deposits, particularly if only grain-size is considered, are common (Figure 1.9).

U.S. Standard Sieve Mesh	Phi	mm	Wentworth (1922)	National Research Council[a]
	−12	4096		
			Boulder gravel	VL boulders
	−11	2048		
				L boulders
	−10	1024		
				M boulders
	−9	512		
				S boulders
	−8	256		
			Cobble gravel	L cobbles
	−7	128		
				S cobbles
	−6	64		
			Pebble gravel	VC gravel
	−5	32		
				C gravel
	−4	16		
				M gravel
5/16	−3	8		
				F gravel
5	−2	4		
			Granule gravel	VF gravel
10	−1	2		
			VC sand	VC sand
18	0	1		
			C sand	C sand
35	1	1/2		
			M sand	M sand
60	2	1/4		
			F sand	F sand
120	3	1/8		
			VF sand	VF sand
230	4	1/16		
			Silt	C silt
	5	1/32		
				M silt
	6	1/64		
				F silt
	7	1/128		
				VF silt
	8	1/256		
			Clay	C clay-size
	9	1/512		
				M clay-size
	10	1/1024		
				F clay-size
	11	1/2048		
				VF clay-size
	12	1/4096		

[a] VL = very large, L = large, M = medium, S = small, VC = very coarse, C = coarse, F = fine, VF = very fine

Figure 1.10. Terminology and grain size terms after WENTWORTH (1922) (from FISHER and SCHMINCKE 1984: p. 119, table 5-7)

Classification of volcaniclastic rocks by grain size characteristics FISHER and SCHMINCKE (1984)

For primary volcaniclastic deposits the FISHER and SCMINCKE (1984) terminology uses grain sizes as core terms, such as ash and lapilli tuff. The basic core terms loosely follow the major size divisions applied in normal clastic sedimentology (WENTWORTH 1922) (Figure 1.10). Pyroclasts in the FISHER and SCHMINCKE (1984) terminology are defined as clasts formed in connection to volcanic eruptions, i.e. clasts expelled through a volcanic vent, without reference of the cause of eruption or origin of fragments (SCHMID 1981) (Figure 1.11). FISHER and SCHMINCKE (1984) distinguish hydroclastic fragments from pyroclastic fragments (the more-specific, use of "pyroclastic" – see previous text). Hydroclastic fragments are those formed by fragmentation due to magma–water interaction. Volcanic fragments formed by weathering of existing rocks are defined as epiclastic fragments (FISHER and SCHMINCKE 1984), though CAS and WRIGHT (1987; see below) use erosion, rather than weathering, to define epiclastic (see below). Fragments formed during mechanical fragmentation of effusive rocks are termed autoclastic fragments (FISHER and SCHMINCKE 1984). Alloclastic fragments in FISHER and SCHMINCKE (1984) are those formed by disruption of pre-existing volcanic rocks by igneous processes with or without direct involvement of magma. Differences between pyroclastic and epiclas-

Clast size	Pyroclast	Pyroclastic deposit	
		Mainly unconsolidated: tephra	Mainly consolidated: pyroclastic rock
	Block, bomb	Agglomerate, bed of blocks or bomb, block tephra	Agglomerate, pyroclastic breccia
64 mm			
	Lapillus	Layer, bed of lapilli or lapilli tephra	Lapillistone
2 mm			
	Coarse ash grain	Coarse ash	Coarse (ash) tuff
1/16 mm			
	Fine ash grain (dust grain)	Fine ash (dust)	Fine (ash) tuff (dust tuff)

Figure 1.11. Granulometric classification of pyroclasts and of unimodal, well-sorted pyroclastic deposit after (SCHMID 1981) (from FISHER and SCHMINCKE 1984: p. 90, table 5-1)

Pyroclastic[a]	Tuffites (mixed pyroclastic-epiclastic)	Epiclastic (volcanic and/or nonvolcanic)	Average clast size (mm)
Agglomerate, agglutinate pyroclastic breccia Lapillistone	Tuffaceous conglomerate, tuffaceous breccia	Conglomerate, breccia	64
(Ash) tuff coarse	Tuffaceous sandstone	Sandstone	2
fine	Tuffaceous siltstone	Siltstone	1/16
	Tuffaceous mudstone, shale	Mudstone, shale	1/256
100% 75%		25%	0% by volume

(increase) Pyroclasts

(increase) Volcanic + nonvolcanic epiclasts (+ minor amounts of biogenic, chemical sedimentary and authigenic constituents)

Figure 1.12. Terms for mixed pyroclastic-epiclastic rocks after SCHMID (1981) (from FISHER and SCHMINCKE 1984: p. 91, table 5-2). "a" – pyroclastic terms according to Figure 1.11

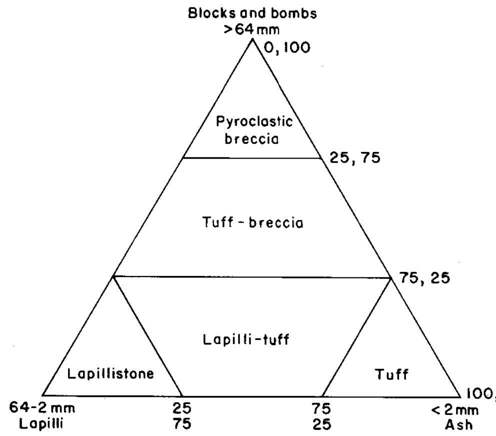

Figure 1.13. Ternary diagram represents mixture terms and end-member rock terms for pyroclastic fragments after FISHER (1966) (from FISHER and SCHMINCKE 1984: p. 92, fig. 5-1). In SCHMID (1981) classification lapillistone is replaced by lapilli tuff

Grainsize (mm)		Pyroclastic fragments: round and fluidally shaped	Pyroclastic fragments: angular	Name of unconsolidated aggregate	Lithified equivalent
	coarse			agglomerate (bombs)	agglomerate (bombs)
256		bombs	blocks	or	or
	fine			pyroclastic breccia	pyroclastic breccia
64					
		lapilli		lapilli deposit	lapillistone
2					
	coarse				
1/16		ash		ash deposit	tuff
	fine				

Figure 1.14. Grain-size limits for proven pyroclastic fragments and pyroclastic aggregates after FISHER (1966) (from CAS and WRIGHT 1987: p. 354, table 12.5). Compare diagram with diagram on Figure 1.11

tic deposits/rocks are also introduced in FISHER and SCHMINCKE (1984) (SCHMID 1981) (Figure 1.12). Pyroclastic rocks are fundamentally defined by their grain size classes (SCHMID 1981). Terms developed for mixed pyroclastic deposits/rocks as well (SCHMID 1981) (Figure 1.13) and various names used to distinguish the loose (deposit) and consolidated (rock) formations (FISHER 1966) (Figure 1.14).

Classification of volcanic rocks by CAS and WRIGHT (1987)

In the CAS and WRIGHT (1987) terminology ash and tuff are the core terms for pyroclastic deposits and rocks. In addition granular-hyaloclastite, granular-autobreccia or hyaloclastic or autoclastic sandstone for autoclastic deposits are used (CAS and WRIGHT 1987). According to CAS and WRIGHT (1987) fragments in volcaniclastic rocks can be produced by primary volcanic processes (in processes contemporaneous with volcanic eruptions) and secondary surface processes (weathering, mass-wasting, erosion). These two main types of processes can produce similar textural types. In CAS and WRIGHT (1987) the

Eruptive mechanism	*Pyroclastic flow*	*Deposit*	*Essential fragment*	
eruption column (fountain) collapse	pumice flow, ash-flow	ignimbrite, pumice flow deposit, ash-flow tuff*	pumice	
	scoria flow	scoria flow deposit	scoria	*decreasing density of juvenile clasts*
lava, dome collapse (explosive and gravitational)	block and ash flow (nuée ardente)	block and ash flow deposit	dense lava	
explosive cryptodome release	hot, dry volcaniclastic debris flow	volcaniclastic debris flow	accessory lithics (± juvenile fragments)	

* In a strict definition, 'ash-flow tuff' should only refer to deposits with >50 wt% finer than 2 mm. 'Ignimbrite' and 'pumice-flow deposit' can be used more loosely irrespective of grainsize, although 'pumice-flow deposit' is sometimes used to emphasise those flow deposits with a large proportion of bomb-sized pumice fragments.

Figure 1.15. Genetic classification of pyroclastic flows and their deposits after CAS and WRIGHT 1987: p. 352, table 12.2)

following types of fragment-forming processes are distinguished: 1) magmatic explosive (e.g. volcanic volatile-driven fragmentation), 2) phreatic or steam explosive (e.g. magmatic heat-driven fragmentation with no direct magma involvement), 3) phreatomagmatic explosive (e.g. magma–water interaction-driven fragmentation). Generally these three types of processes are considered to be pyroclastic eruptions by CAS and WRIGHT (1987). Quench or chill fragmentation (e.g. hyaloclastite formation) and flow fragmentation (e.g. autobrecciation) are additional processes to the previous three classes, all together considered to be result of primary volcanic processes (CAS and WRIGHT 1987). Epiclastic fragments in CAS and WRIGHT (1987) are considered to be those released or remobilised by surface processes (not necessarily weathered from existing rocks). Reworked or redeposited (or both) pyroclastic and autoclastic material become epiclastic upon reworking or redeposition.

Eruptive mechanism	Pyroclastic surge type	Temperature, water content	Essential fragment
phreatomagmatic (outward moving radial collar and column collapse)	base surge	cold, wet (hot, dry)	juvenile (vesiculated to non-vesiculated); accessory lithics
accompanying pyroclastic flows	ground surge	hot, dry	juvenile (vesiculated to non-vesiculated)
	ash-cloud surge	hot, dry	juvenile (vesiculated to non-vesiculated)
accompanying pyroclastic fall eruptions (but without generation of a pyroclastic flow)	ground surge	hot, dry	juvenile (vesiculated)

Figure 1.16. Genetic classification of pyroclastic surges and their deposits after CAS and WRIGHT 1987: p. 353, table 12.4

CAS and WRIGHT (1987) distinguish pyroclastic and epiclastic rocks on the basis of their modes of transportation and deposition (e.g. all the particles in laharic deposits for CAS and WRIGHT 1987 are epiclastic, but pyroclastic for FISHER and SCHMINCKE 1984, as they are particles formed in the eruption, but then get moved again). CAS and WRIGHT (1987) therefore state that epiclastic deposits are the result of normal surface processes, deposited by such processes regardless of whether the fragments are those formed in an eruption or are new particles formed by weathering of older rocks.

For Cas and Wright, volcaniclastic rocks are classified in two ways; 1) genetically and 2) lithologically. The genetic classification identifies their origin such as pyroclastic falls, pyroclastic flows (Figure 1.15) and pyroclastic surges (Figure 1.16). In the lithological classification the grain size limits and overall size distribution of the deposits, the constituent fragments of the deposits and the degree and type of welding are the main classification lines. CAS and WRIGHT (1987) uses same grain size categories as those introduced by SCHMID (1981).

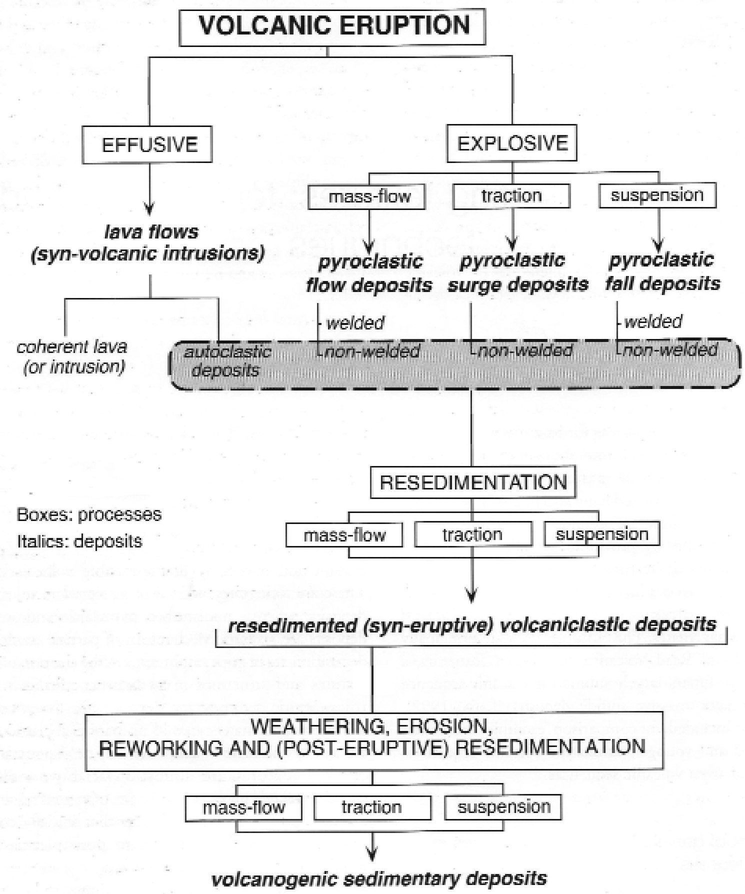

Figure 1.17. Genetic classification of volcanic deposits after MCPHIE et al. 1993: p. 2, fig. 1). Depositional processes are the same (mass flow, traction, suspension) in primary, resedimented and reworking volcaniclastic deposits. Non-welded primary volcaniclastic deposits marked in grey field with dashed line

Classification of volcanic deposits by MCPHIE et al. (1993)

A more-recent classification scheme developed on the basis of field textural characteristics of fragmental and coherent volcanic rocks partly amalgamated the previous terminologies (MCPHIE et al. 1993). MCPHIE et al. (1993) distinguish two major types of volcanic rocks; 1) coherent and 2) volcaniclastic (Figure 1.17). Fragmental rocks with volcaniclastic textures can result from autoclastic fragmentation of a lava flow, or through various types of explosive eruptions. In the MCPHIE et al. (1993) classification the resulting fragmental rocks are classified according to the transportation and depositional processes that generated the deposits/rocks.

Three major types of transportation mechanisms (mass-flow, traction, suspension) are considered to be responsible for the majority of the textural features that form. These three transportation styles are closely linked to traditionally recognised volcanic processes such as mass-flow transportation in pyroclastic flows, traction transportation in pyroclastic surges, and suspension deposition as pyroclastic falls. MCPHIE et al. (1993) also distinguish three levels of processes that form fragmented volcanic rocks. Categories of volcanic rocks in MCPHIE et al. (1993) are 1) coherent lavas and intrusions (Figure 1.18), 2) primary pyroclastic deposits (Figure 1.19), 3) deposits from resedimentation processes (Figure

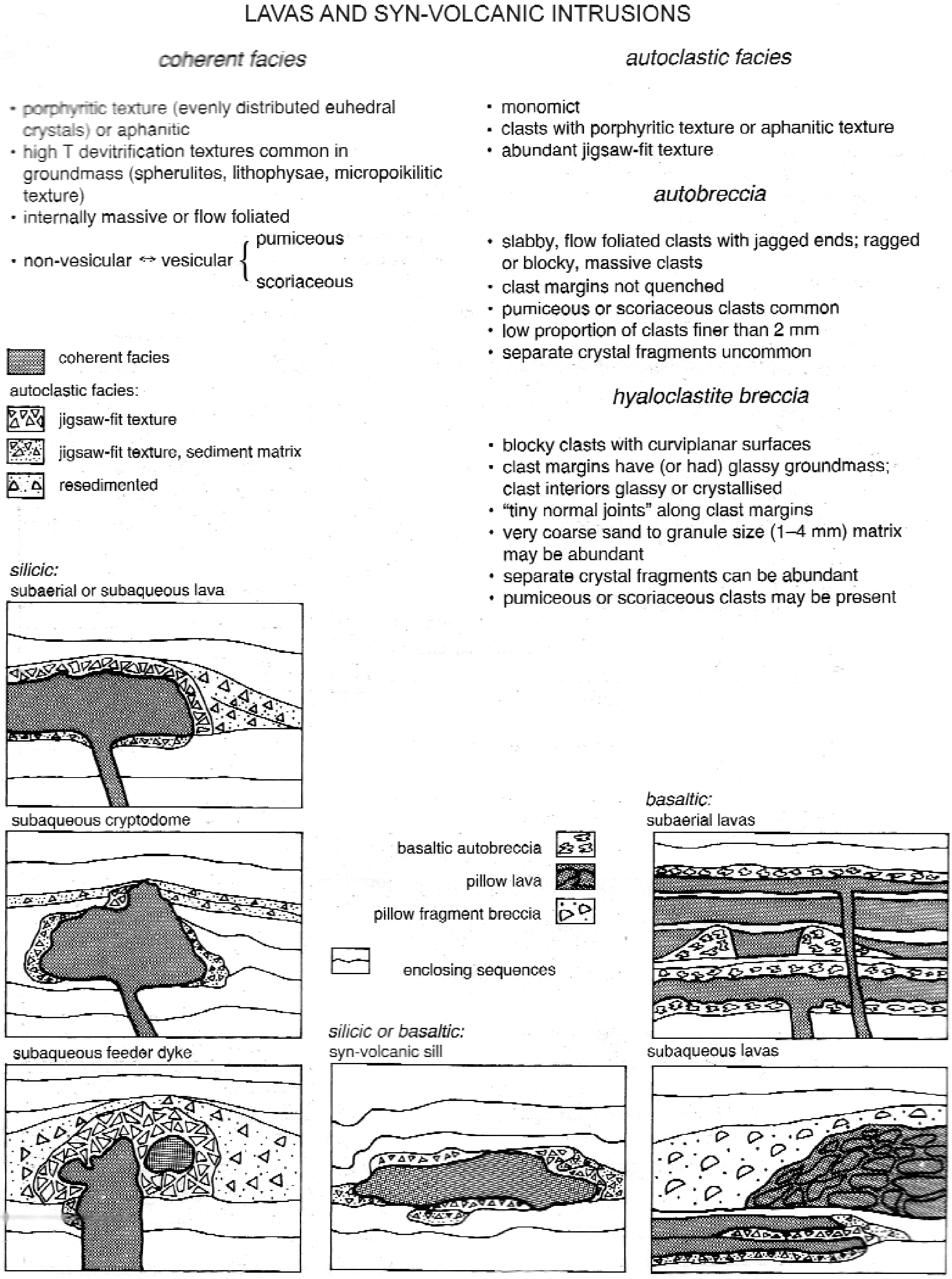

Figure 1.18. Genetic classification and basic characteristics of lavas and syn-volcanic intrusions after MCPHIE et al. 1993: p. 4, fig. 2

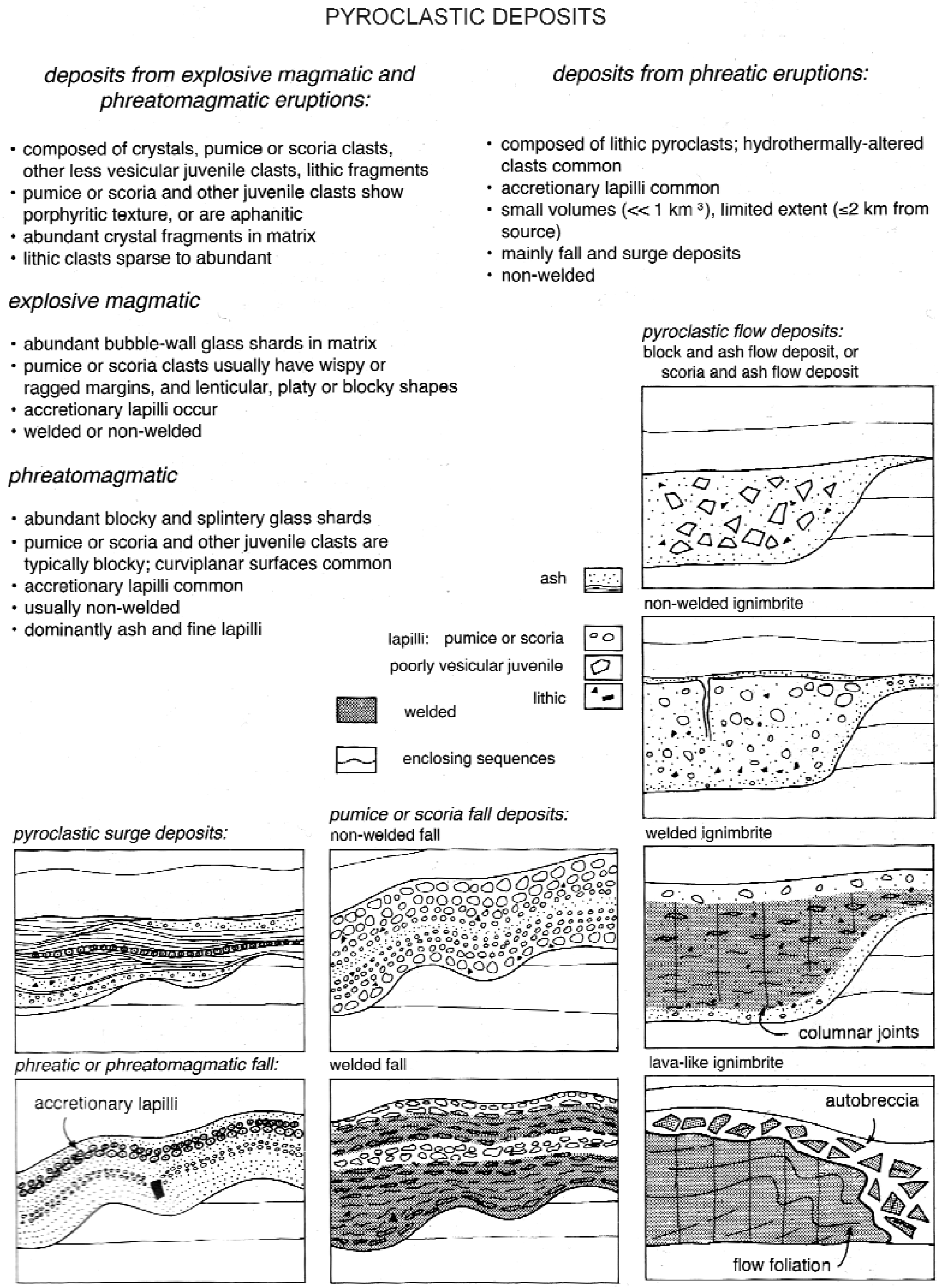

Figure 1.19. Genetic classification and basic characteristics of (primary) pyroclastic deposits after McPhie et al. (1993, p. 5, fig. 3)

1.20), and 4) reworking and post-eruptive resedimentation processes producing volcanogenic sediments (Figure 1.21). This classification perhaps faces major problem to distinguish syn- and post-eruptive processes may be responsible for resedimentation and therefore class 3 and 4 can be indistinguishable, especially on the basis of studies of ancient fragmental volcanic rocks.

McPhie et al. (1993) also emphasize three major types of classification of volcanic rocks ranging from the purely

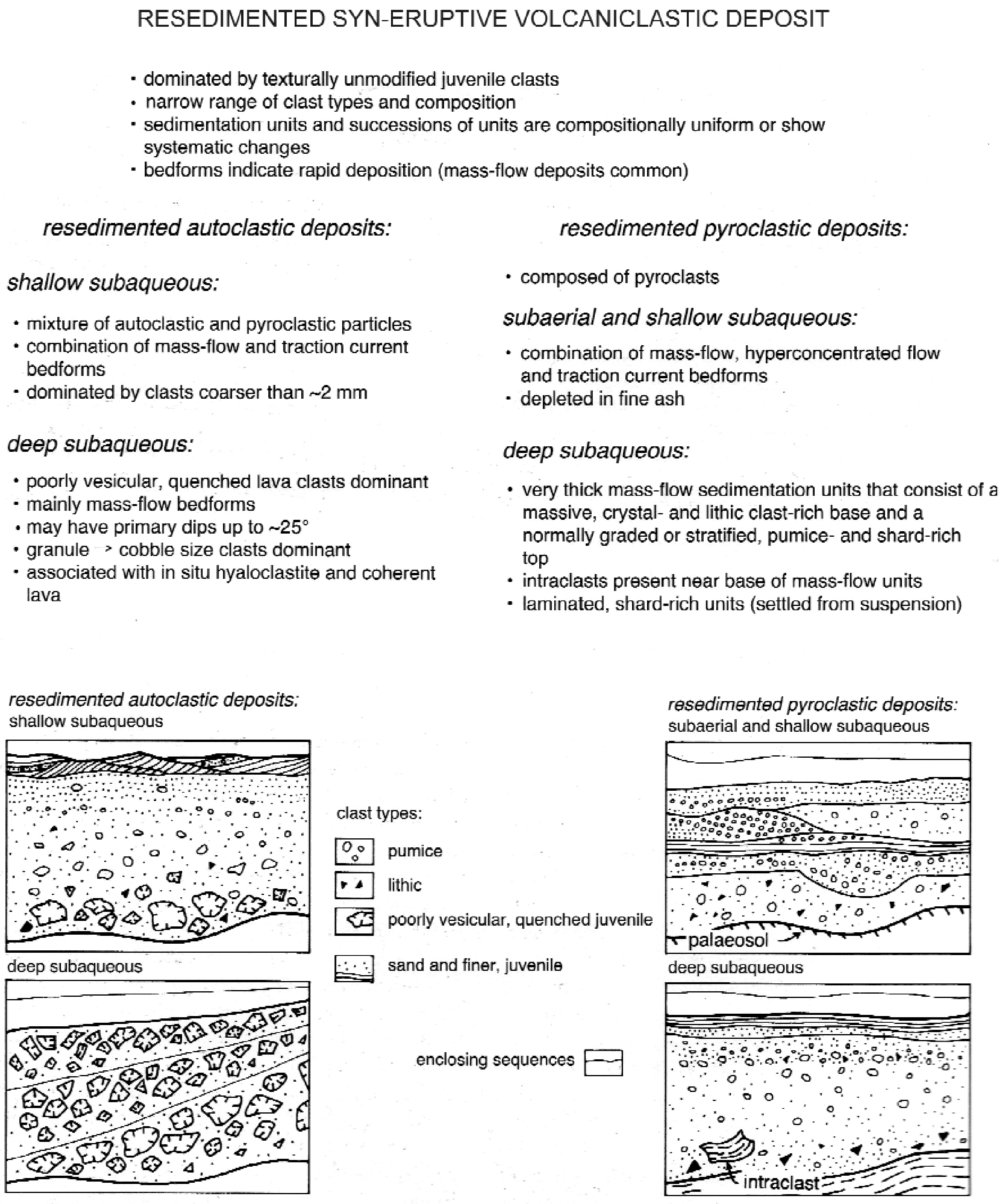

Figure 1.20. Genetic classification and basic characteristics of resedimented syn-eruptive volcaniclastic deposits after McPhie et al. 1993: p. 6, fig. 4)

descriptive to the purely genetic. McPhie et al. (1993) distinguish three major terminology types; 1) lithological terminology, 2) lithofacies terminology, and 3) genetic terminology.

McPhie et al. (1993) lithological terminology provides information on composition, components and grain size.

Lithofacies terminology provides information on facies characteristics evident at outcrop scale in the field, such as structures (bedding, stratification etc.), internal organisation and geometry.

Genetic terminology provides information on eruption and emplacement processes for primary volcanic and volcani-

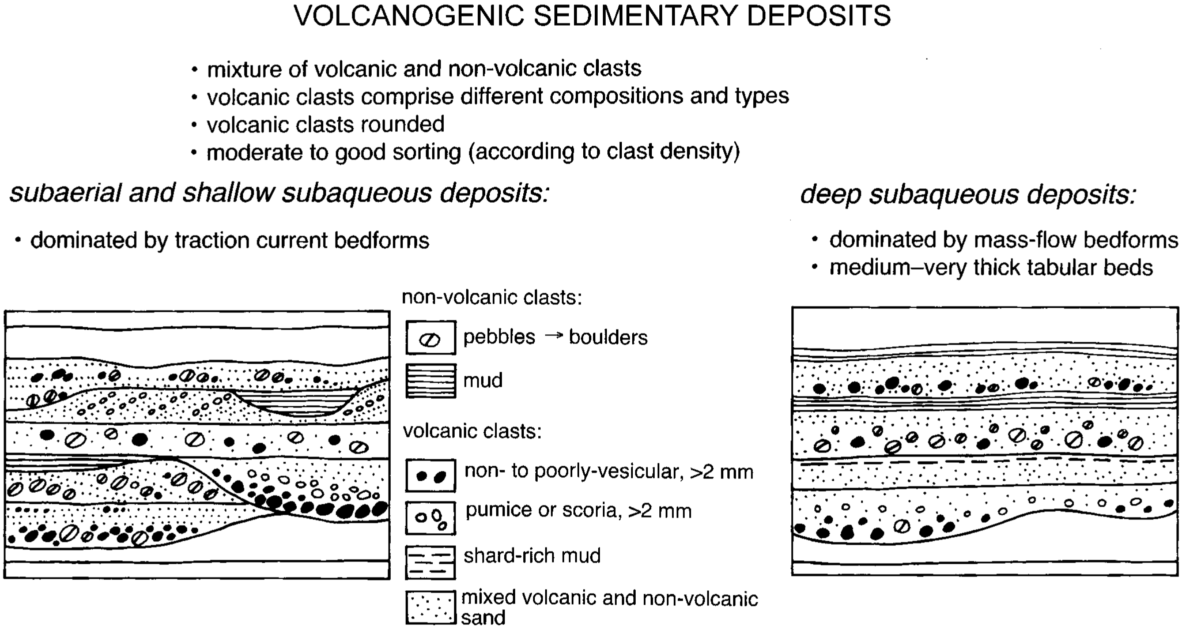

Figure 1.21. Genetic classification and basic characteristics of volcanogenic sedimentary deposits after McPhie et al. 1993: p. 7, fig. 5

clastic deposits, and on subsequent redeposition, erosion, transport and depositional processes for resedimented and volcanogenic sedimentary deposits.

In field volcanology, especially in ancient settings the separation of descriptive from genetic terminology is important in the view of McPhie et al. (1993). McPhie et al. (1993) suggest a useful, field-based terminological algorithm to name identified volcanic rocks using their lithological and lithofacies terminology (Figures 1.22 and 1.23). For McPhie et al. (1993) the lithological terminology for fragmental volcanic rocks the grain size categories represent the core terms, which are then supplemented by the components, lithofacies terms, and then alteration state of the deposit/rock. This terminology is entirely based on clastic sedimentological grain-size terms, with the aim of avoiding any genetic implications in the pure description of the deposit/rock. McPhie et al. (1993) stress that only after detailed interpretation of identified textural features, can a genetic term be applied to a deposit/rock. This approach in fact might be fine from physical and descriptive points of view, but can end up producing very confusing names for some rock types, which most volcanol-

Descriptive names for coherent lavas and intrusions

Ideal combination: ④ + ③ + ② + ①

alteration **texture** **lithofacies term** **composition**

e.g. sericitic, highly quartz-phyric, coarse, flow-banded rhyolite
moderately vesicular, poorly olivine-phyric, fine, columnar jointed basalt

Minimum: ② + ① e.g. blocky jointed rhyolite; massive basalt
③ + ① e.g. hornblende-phyric andesite; aphanitic dacite(?)
④ + ① e.g. sericite-silica rhyolite(?); chlorite-epidote andesite(?)

① COMPOSITION

a. estimate based on phenocryst assemblage:
- *rhyolite*: K-feldspar ± quartz (± Ca-poor plagioclase ± ferromagnesian phase: biotite, amphibole, pyroxene, fayalite)
- *dacite*: plagioclase ± ferromagnesian phase: biotite, amphibole, pyroxene ± quartz (± K-feldspar)
- *andesite*: plagioclase + ferromagnesian phase: biotite, amphibole, pyroxene (± olivine)
- *basalt*: pyroxene + Ca-rich plagioclase ± olivine

b. for aphanitic samples, estimate based on colour:
- *rhyolite(?), dacite(?)* : pale grey, pink, cream, pale green
- *andesite(?), basalt(?)* : dark grey, dark blue, dark green, dark purple

② LITHOFACIES

- massive or flow-foliated, flow-banded, flow-laminated
- jointing: columnar, radial columnar, concentric, tortoise shell, blocky, prismatic, platy
- pillows or pseudo-pillows

③ TEXTURE

- porphyritic: a. phenocrysts – type *(quartz-phyric …, pyroxene-phyric …, etc.)*
  – abundance *(poorly …, moderately …, highly …)*
  – size *(fine ≤1 mm, medium 1–5 mm, coarse ≥5 mm)*
  b. groundmass – glassy, cryptocrystalline, microcrystalline, very fine grained
- aphanitic: uniformly microcrystalline
- aphyric: no phenocrysts present
- glassy: composed of volcanic glass
- non-vesicular or vesicular (or amygdaloidal): sparsely …, moderately …, highly …, pumiceous …, scoriaceous …
- spherulitic, microspherulitic, lithophysae-bearing

④ ALTERATION

- mineralogy: chlorite, sericite, silica, pyrite, carbonate, feldspar, hematite …
- distribution: disseminated, nodular, spotted, pervasive, patchy …

Figure 1.22. Descriptive names for coherent lavas and intrusions as suggested by McPhie et al. 1993: p. 9, table 1)

Descriptive names for volcaniclastic deposits

Ideal combination: ④ **alteration** + ③ **lithofacies term** + ② **components** + ① **grain size**

e.g. chloritic-pyritic, very thickly bedded, volcanic lithic breccia
thinly interbedded, shard-rich mudstone and crystal-rich sandstone

Minimum: ② + ① e.g. crystal-rich sandstone; pumice granule breccia
③ + ① e.g. laminated mudstone; poorly sorted, massive breccia
④ + ① e.g. pyritic sandstone; chloritic breccia

① GRAIN SIZE

mud/mudstone		< 1/16 mm
sand/sandstone		1/16–2 mm
gravel/conglomerate or breccia:	granule	2–4 mm
	pebble	4–64 mm
	cobble	64–256 mm
	boulder	>256 mm

② COMPONENTS

- crystals, crystal fragments: *crystal-rich ...*
- lithic fragments: *lithic-rich ...*
  - volcanic or non-volcanic, polymict or monomict
- pumice or scoria: *pumiceous ..., scoriaceous ...*
- shards: *shard-rich ...*
- accretionary lapilli: *accretionary lapilli-rich*
- vitriclasts: *vitriclast-bearing ...*
- fiamme: *fiamme-bearing ...*
- cement: *siliceous ..., carbonate ..., zeolite ...*

③ LITHOFACIES

- massive (non-bedded) or stratified (bedded)
- bedding: laminated < 1 cm
  very thinly bedded 1–3 cm
  thinly bedded 3–10 cm
  medium bedded 10–30 cm
  thickly bedded 30–100 cm
  very thickly bedded > 100 cm
- equal or unequal thickness
- laterally even or uneven thickness
- laterally continuous or discontinuous
- cross-bedded, cross-laminated
- massive (non-graded) or graded: normal ↑, reverse ↓
  normal–reverse, reverse–normal
- fabric: clast-supported or matrix-supported
  poorly sorted, moderately sorted, well sorted
- jointing: blocky, prismatic, columnar, platy

④ ALTERATION

- mineralogy: chlorite, sericite, silica, pyrite, carbonate, feldspar, hematite ...
- distribution: disseminated, nodular, spotted, pervasive, patchy ...

Figure 1.23. Descriptive names for volcaniclastic deposits as suggested by McPhie et al. 1993: p. 10, table 2

Grain size		Primary volcaniclastic deposit		Sedimentary deposit (rock name)	
(phi)	(mm)	Unconsolidated	Lithified	Unconsolidated	Lithified
>4	<1/16	Extremely fine ash	Extremely fine tuff	Clay	Mudrock, shale
3–4	1/16–1/8	Very fine ash	Very fine tuff	Very fine sand	Very fine sandstone
2–3	1/8–1/4	Fine ash	Fine tuff	Fine sand	Fine sandstone
1–2	1/4–1/2	Medium ash	Medium tuff	Medium sand	Medium sandstone
0–1	1/2–1	Coarse ash	Coarse tuff	Coarse sand	Coarse sandstone
−1 to 0	1–2	Very coarse ash	Very coarse tuff	Coarse sand	Coarse sandstone
−2 to −1	2–4	Fine lapilli	Fine lapilli-tuff	Granule	Grit, granule congl.
−4 to −2	4–16	Medium lapilli	Medium lapilli-tuff	Pebble	Pebble conglomerate
−6 to −4	16–64	Coarse lapilli	Coarse lapilli-tuff	Cobble	Cobble conglomerate
<−6	>64	Block/bomb	Breccia	Boulder	Boulder congl.

Note: The ash and lapilli grain-size ranges have been modified from that given by Fisher (1961) and derivative classifications to match and include the subdivisions within the sand and gravel ranges given by Wentworth (1922). "Extremely fine" ash replaced "fine ash" for particles finer than 4 phi (1/16 mm). Lithified sedimentary deposits with angular grains coarser than 2 mm are commonly termed "breccia."

Figure 1.24. Grain-size terms for primary volcaniclastic rocks after White and Houghton 2006: p. 678, table 1

ogists would not really associate to volcanic processes, such as thickly bedded, accidental lithic-rich, crystal-rich sandstone. In part to deal with such inconsistencies, a new amalgamated terminology is suggested by White and Houghton (2006). This terminology seems to be able to make a clear descriptive and genetic categorisation of primary volcaniclastic rocks.

Primary volcaniclastic deposits after White and Houghton (2006)

Primary volcaniclastic deposits are considered to be primary deposits which do not involve interim storage, regardless of the style of transport of the clasts (White and Houghton 2006). To account for uncertainties in initial treatment of some deposits (McPhie et al.'s "syneruptive… whatever), deposits that appear directly related to an eruption are named with primary-deposit names, rather than sedimentary ones applied to "epiclastic" deposits. This is reflected in White and Houghton's (2006) terminology by restriction of "volcaniclastic" so that it is no longer used as a general term for all deposits containing clasts with a volcanic heritage (including those from weathering of older volcanic or volcaniclastic rocks). Instead, primary volcaniclastic names, which use the same "pyroclastic" core terms from Fisher and Schmincke (1984), are applied to every type of deposit resulting directly from a volcanic eruption (White and Houghton 2006). Deposits of clasts resulting from weathering are no longer named as volcaniclastic, instead being named with normal sedimentary terms such as basaltic sandstone, etc. All primary volcaniclastic rocks are therefore described using grain size terms introduced already in the Fisher and Schmincke (1984) terminology such as ash, tuff, tuff breccia etc. In the White and Houghton (2006) classification scheme, therefore, all deposits considered to be primary volcaniclastic deposits should carry primary volcaniclastic names.

White and Houghton's (2006) description of primary volcaniclastic rocks uses grain-size nomenclature based on classification categories and classes similar to those established during the sixties, seventies and eighties (Fisher 1961, 1966, Schmid 1981). In White and Houghton (2006) the grain size classes are additionally subdivided to follow similar class limits to those used in normal clastic sedimentology (Figure 1.24). This classification follows the earlier Fisher and Schmincke (1984) grain-size classes with slight modification. For mixed primary volcaniclastic rocks (e.g. complex grain size distribution) a similar ternary diagram is suggested based on similar categories to those suggested by Fisher (1961) (Figure 1.25). In this new diagram lapillistone is no longer used, instead the lapilli tuff field is extended (following Schmid 1981).

In the new White and Houghton terminology, 3 major descriptive modifiers are advocated; 1) componentry, 2) sorting, and 3) clast morphology. For primary volcaniclastic deposits, only 3 component classes, one newly identified, are distinguished; 1) juvenile, 2) lithic and 3) composite (Figure 1.26). Juvenile clasts are defined as clasts derived from the newly erupted magma (White and Houghton 2006). Lithic clasts are considered to be any type of country rock-derived fragments (White and Houghton 2006). Composite clasts are defined as mechanical mixtures of juvenile and lithic clasts (and/or recycled (Houghton and Smith 1993) juvenile clasts) (White and Houghton 2006) such as fragments of peperite (White et al. 2000, Skilling et al. 2002) (Plate IV, 4) or cored bombs (Houghton and Smith 1993, Rosseel et al. 2006) (Plate IV, 5).

For the sorting characteristics of primary volcaniclastic rocks, similar classification groups can be applied as in normal clastic sedimentology (e.g.well-sorted, poorly sorted) (White and Houghton 2006). Clast morphology can also be added as a descriptive modifier in naming the primary volcaniclastic deposits (White and Houghton 2006). The term "rounded" is suggested to be used to express clast abrasion, and "fluidal" or "amoeboid" to describe aerodynamically shaped or surface-tension reshaped clasts that have ~round, smoothly curved, shapes not formed by abrasion.

White and Houghton (2006) recognize 4 basic genetic end-member varieties of primary volcaniclastic rocks, based on their manner of deposition or emplacement (Figure 1.27); 1) pyroclastic – sedimentation from pyroclastic plumes and currents, 2) autoclastic – direct deposition of fragments from lava, formed via air cooling, 3) hyaloclastic – direct deposition of fragments from lava, formed via water chilling , and 4) peperitic – deposits emplaced during mingling of magma with wet sediments (in situ deposition).

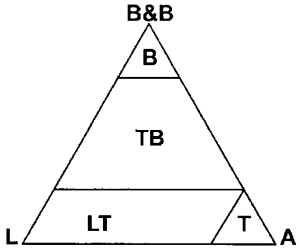

Figure 1.25. Grain-size ternary diagram for naming primary volcaniclastic rocks based on Fisher's (1961) classification (from White and Houghton 2006: p. 679, fig. 1). B&B – blocks and bombs, L – lapilli, A – ash, B – breccia, TB – tuff breccia, LT – lapilli tuff, T – tuff. Blocks are defined as angular shape large pyroclasts and bombs are their fluidal equivalent. Unconsolidated deposits are named after minor-major constituents, e.g. lapilli-ash, bomb-ash etc. Field limits are marked in 25% and 75%

Component	Key criteria	Components within deposits (*example*)
Juvenile	Primary juvenile: derived directly from erupting magma; particle contributes heat to thermal budget of transport and/or fragmentation processes. Recycled juvenile: juvenile clast recycled during the eruption that formed it; not a significant thermal contributor to depositing plume or current.	Dense to inflated fragments of chilled magma (*pumice, scoria, dense juvenile*); may be recycled. Aggregate of relatively finer-grained clasts (*accretionary lapilli, armored lapilli*). Crystals derived directly from the erupting magma (e.g., *juvenile feldspar*); may be recycled.
Lithic	Clast formed by fragmentation of pre-existing rock or incorporated from unconsolidated sediment. These contribute negligible heat energy to transport, depositional, or fragmentation processes.	Fragments derived from wall rock (e.g., *sandstone lithic*). Fragments of solidified magma from conduit walls, blocks of lava or dike rock (e.g., *basalt lithic*). Block of pyroclastic rock (e.g., *tuff block*).
Composite	Clast formed by mingling of magma with a clastic host, or incorporation of lithic debris into magma.	Fragments of peperite (*composite clasts*). Bomb with lithic core (*cored bomb*).

Note: Though "juvenile" is subdivided to distinguish primary from recycled clasts, it is recognized that this significant behavioral distinction can only rarely be made from ancient deposits. Composite clasts are unique in combining lithic and juvenile material.

Figure 1.26. Component classes (juvenile, lithic and composite) for volcaniclastic deposits as suggested by White and Houghton 2006: p. 679, table 2

Process	Deposit adjective (noun)
Sedimentation from pyroclastic plumes and currents	Pyroclastic (various)
Deposition of fragments from lava, formed via air cooling	Autoclastic (autobreccia)
Deposition of fragments from lava, formed via water chilling	Hyaloclastic (hyaloclastite)
Mingling of magma with wet sediment, ~in situ deposition	Peperitic (peperite)

Note: All should be given grain-size names based on grain size and degree of lithification

Figure 1.27. Variety of primary volcaniclastic rocks after White and Houghton 2006: p. 679, table 3

Usage of terminology in classification of fragmental volcanic rocks is very important if researchers are to be clear and unambiguous in their descriptions of volcaniclastic deposits and rocks. Here we suggest following the recent classification scheme for primary volcaniclastic rocks suggested by WHITE and HOUGHTON (2006) because this terminology best fits with real volcanic processes, including those deposits resulting from explosive subaqueous volcanism, and is simple enough to be able to classify a deposit/rock by the key textural characteristics.

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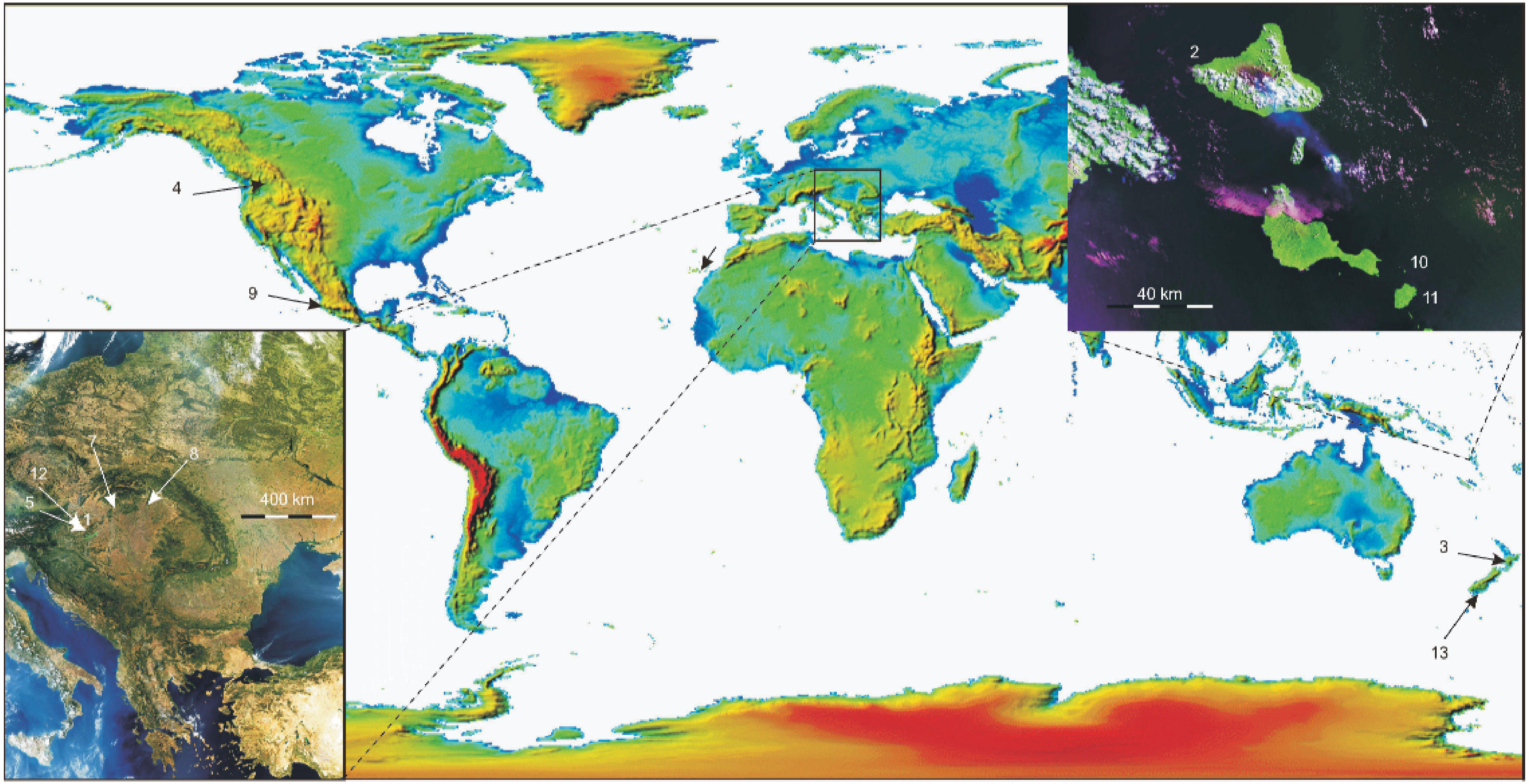
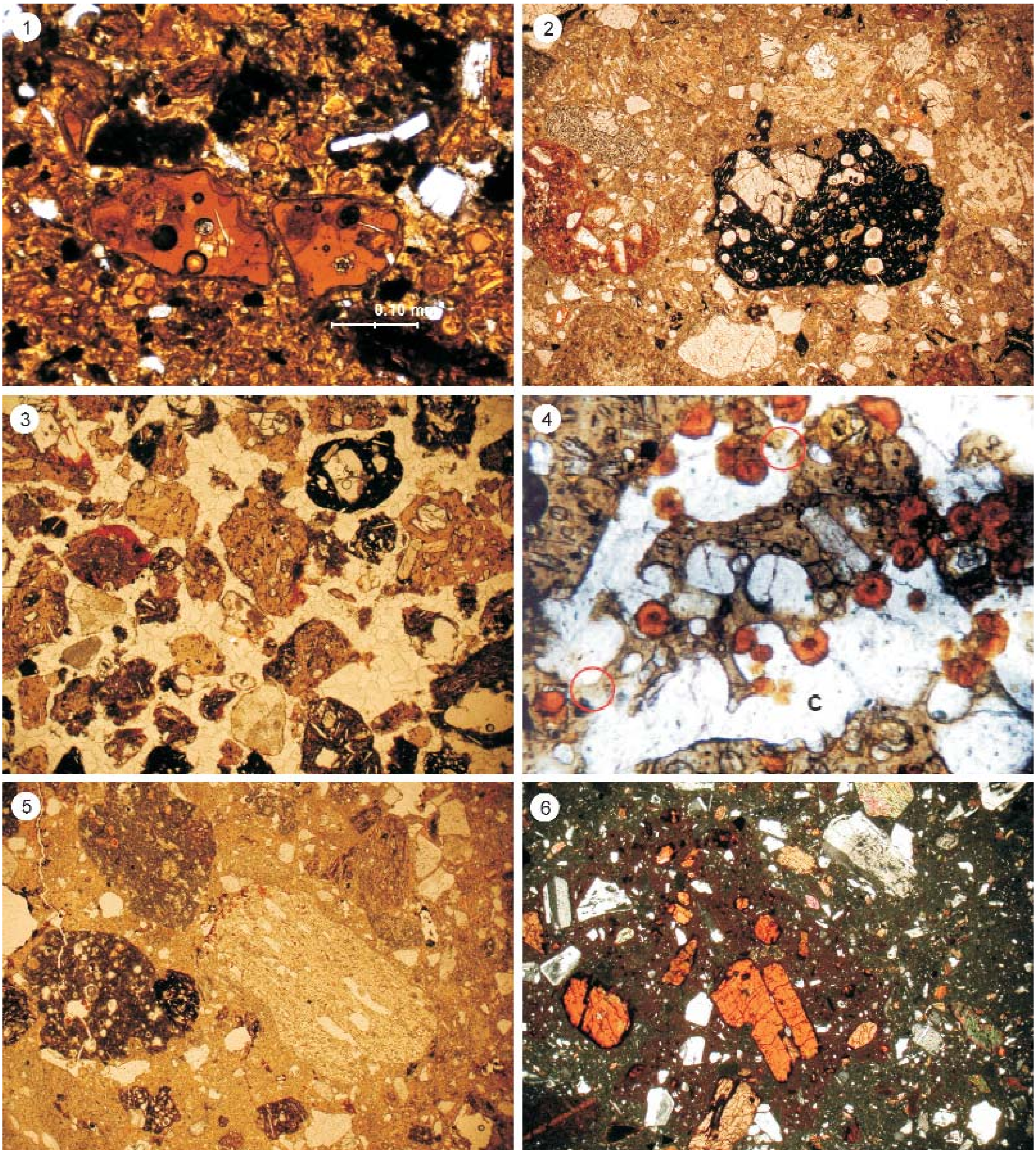
Location map



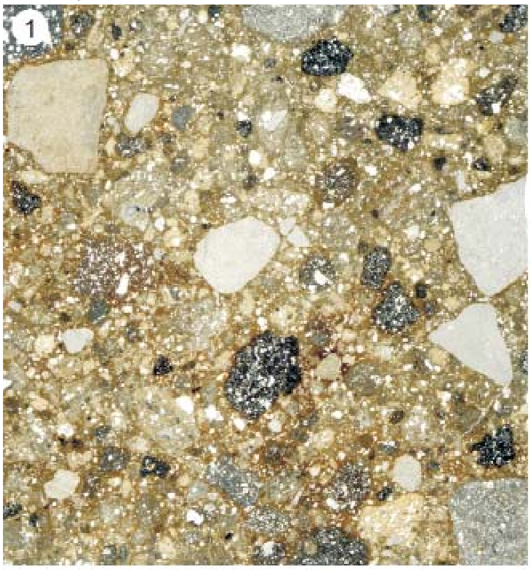
1 — Bakony – Balaton Highland Volcanic Field, Hungary
2 — West Ambrym, Vanuatu
3 — Taupo Volcanic Zone, New Zealand
4 — Sinker Butte, Idaho, USA
5 — Hegyesd diatreme, Hungary
6 — Waipiata Volcanic Field, New Zealand
7 — Pilis Mts, Hungary
8 — Tokaj Mts, Hungary
9 — Ixtlan del Rio, Mexico
10 — Laika Island, Vanuatu
11 — Tongoa, Vanuatu
12 — Hajagos tuff ring, Hungary
13 — Pigroot Hill, New Zealand

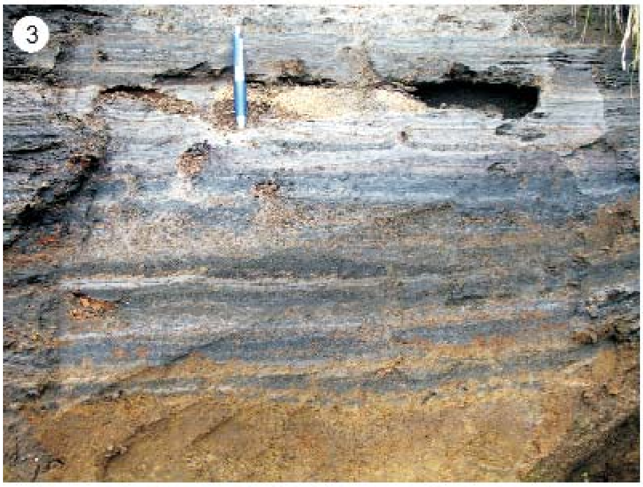
Plate I

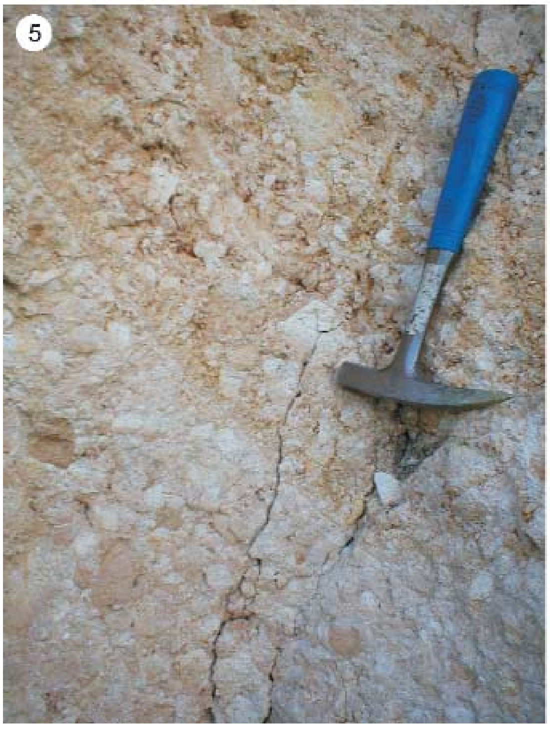


1. Blocky sideromelane glass shards (brown fragment) as juvenile fragments from a phreatomagmatic lapilli tuff of the Sinker Butte, Western Snake River Volcanic Field, Idaho, USA.
2. Moderately vesicular tachylite lapilli in a phreatomagmatic lapilli tuff from the Pliocene Bakony – Balaton Highland Volcanic Field
3. Blocky sideromelane glass shard (yellowish to brown colour fragments) in calcite cemented lapilli tuff from the Hegyesd diatreme, western Hungary.
4. Rectangular contoured, partially collapsed vesicles of an irregular shape sideromelane glass shard from a diatreme filling lapilli tuff of the Miocene Waipiata Volcanic Field, South Island, New Zealand. Red circle marks calcite cement.
5. Mixed textured juvenile clasts (sideromelane and tachylite; vesucular and non-vesicular; crystalline and non-crystalline) from lapilli tuff of the Kishegyestű diatreme remnant of the Bakony – Balaton Highland Volcanic Field, Hungary. The presence of diverse textured juvenile fragments in the same pyroclastic rocks indicates variable vesiculation, crystallisation conditions, and fragmentation styles. Shorter side of the picture is about 5 mm long.
6. Pyroxene, amphibole and plagioclase pyrogenic crystals in a lapilli tuff deposited from Miocene pumiceous pyroclastic flows resulted from dome collapses (Pilis Mts, Northern Hungary).

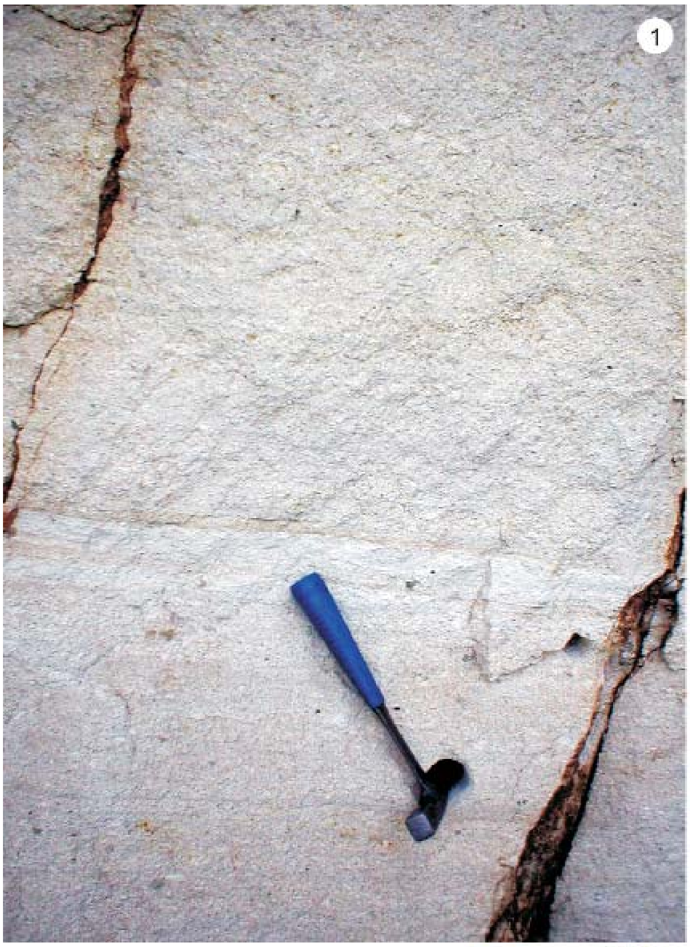
Plate II







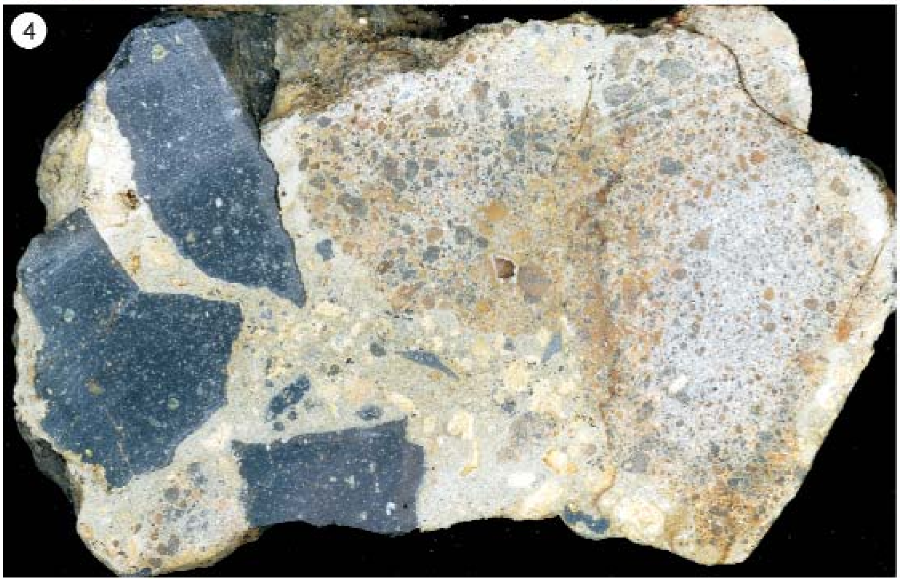
1. Accidental lithic-rich lapilli tuff from the Véndeg-hegy diatreme. In this sample the majority of the accidental lithics are Triassic limestones (pale-grey angular fragments) excavated by the phreatomagmatic explosions responsible for the formation of the diatreme. The side of the picture is 4 cm long.
2. Cognate lithic fragments in Fisher and Schmincke (1987) terminology represent particles of fragmented co-magmatic volcanic rocks from previous eruptions of the same volcano. In the picture the crater complex of the Marum volcanic cone complex on Ambrym Island (Vanuatu, New Hebrides) is shown. In the crater wall thick former lava lake cross sections are exposed (white columnar jointed layers). In the front of the view large angular blocks in a coarse ash matrix qualify to be cognate lithic fragments, since they are inferred to been derived from the former, solidified lava lakes disrupted by subsequent explosive eruptions.
3. Thinly laminated ash beds of the Mangatawai Formation near to the Ruapehu – Tongariro Volcanoes. The origin of the unit is unclear (deposits from fine fall out or horizontal moving pyroclastic density current) and it is under current research.
4. Thickly bedded well-sorted, pumiceous lapilli bed from Plinian fall-out succession (Taupo Volcanic Zone, New Zealand). Hammer is 30 cm long.
5. Massive bed with no internal structures from a rhyodacitic block-and-ash deposit of the Tokaj Mts NE-Hungary. Hammer is 30 cm long.



1. Stratified pumiceous lapilli tuff succession from the Tokaj Mts, NE-Hungary. Hammer is 30 cm long.
2. Parallel bedded fine grained lapilli tuff and tuff of a scoria cone near Ixtlan del Rio, Nayatir, Mexico.
3. Reverse graded lithic-rich lapilli tuff beds from Laika Island, near Tongoa, Vanuatu. The black blocks on the top of each beds are glassy, angular volcanic lithics.
4. A large block of cross bedded tuff from a young tuff cone from West Ambrym, Vanuatu (hammer is 30 cm long).

Plate IV

1. Variations in sorting characteristics of pyroclastic lapilli tuff successions (Ambrym, Vanuatu). Note the coarse lapilli lenses in the ash matrix. The view represents about 2.5 m thick section.
2. Well sorted pumiceous fallout bed from the Taupo Volcanic Zone, New Zealand (hammer is 30 cm long).
3. Poorly sorted lapilli tuff of a pyroclastic density current deposited succession inferred to be accumulated in shallow marine environement (Laika Island near Tongoa, Vanuatu). Pen is about 15 cm long.
4. Fragments of blocky peperite in a lapilli tuff from the Hajagos tuff ring, Bakony – Balaton Highland Volcanic Field, Hungary. The shorter side of the view is about 15 cm.
5. Cored bombs from a vent filling phreatomagmatic tuff breccia of the Pigroot Hill volcanic complex, Waipiata Volcanic Field, New Zealand.



Chapter 2

Actual geological view in volcanology

Rocks associated with volcanoes and volcanic systems play a significant role in the geological record. Volcanic rocks in old and eroded terrains record information not only about the volcanic process that generated them but also of their eruptive environment. Detailed studies of the volcanic rocks preserved in the geological record may give important information about the physical environment in which they are deposited; palaeoclimatology, eruptive environment (e.g. subaerial vs. subaqueous), the sedimentary basin structure and hydrological conditions as well as the tectonic evolution of the region in which they are located. Studies of volcanic rocks therefore are important, and their studies could help to establish the detailed geological evolutional history of the larger region in which they occur. Since volcanoes are either grouped into clusters (CONDIT and CONNOR 1996; CONWAY et al. 1998; CONNOR and CONWAY 2000), alignments (CONNOR et al. 2000) and commonly form individual volcanic fields up to few millions of years duration the information we may able to obtain from volcanic regions could give information of the evolution of a region over time scale of millions of years. Complex volcanoes that are more than an order magnitude in volume to those forming common volcanic fields are also active over thousands to millions of years, with recurrent construction and destruction phases, all leaving a significant mark in the depositional environment where those volcanoes developed. Since composite volcanoes are volumetrically large (km^3 volume range) the preservation potential of their eruptive products is great, and then influence commonly extends one large (hundreds of km^2) areas, where the volcaniclastic sedimentation commonly interferes with normal siliclastic or carbonatic sedimentary processes (FISHER and SMITH 1991). In either system (monogenetic volcanic fields or polygenetic systems) the volcanic sediments and rocks interact with normal background sedimentation to produce a complex sedimentary record (FISHER and SMITH 1991). In this sedimentary record the diagenesis (and consequently the subsequent metamorphic processes) especially in ancient volcanic rocks can overprint many of the characteristic features used to identify the eruptive processes of the volcanic system. The study of ancient volcanic rocks should therefore always compare, and link the identified textural features from the preserved volcanic rocks to the volcanic sediments and coherent magmatic facies of young and active volcanism. In general coherent lava flows and intrusive rocks do not change significantly over long time periods (millions of years), unless the volcanism took place in a setting where diagenesis and subsequent metamorphic processes could be intense and/or intensified over much shorter periods of time. However, coherent volcanic rocks are relatively easy to associate with either effusive or intrusive events. Establishing the link between volcaniclastic rocks and tephra or volcanogenic sediments deposited from young volcanoes is more problematic since diagenesis may cause significant textural changes in the preserved volcanic material. In spite of this, volcaniclastic rocks still held significant amounts of information about their magma evolution, magma fragmentation, pyroclast transportation, deposition and subsequent remobilisation, reworking or redeposition. In the remainder of this Chapter we introduce few basic terms describing types of volcanic eruptions and the volcaniclastic depositional systems commonly encountered in active volcanic field. Using the volcanic rock record as our primary data, we will then interpret the ancient rock record in accordance with those processes observed in active volcanoes.

Volcano types and their relationships to sedimentation

Volcanic landforms are very diverse by volume, shape, relative location, size and composition (COTTON 1944). The size of volcanoes can range from a lava spatter cone few tens of metres across to ocean island shields over 100 km wide, and silicic calderas, that produced thick (tens of metres) pyroclastic deposits covering areas greater than 1000 km^2 (Figure 2.1).

Scoria cones (Plate I, 1) are the most common volcanic landforms on Earth (Cas and Wright 1988). They are a few hundreds of metres across, rarely taller than few hundred metres, and their pyroclastic deposits are commonly restricted to their volcanic cone (Cas and Wright 1988; Vespermann and Schmincke 2000; Schmincke 2004). Large scoria cones however, may have deposits that are more widespread, and may represent sub-Plinian mafic explosive events (Thordarson and Self 1993; Martin and Németh 2006). They are also often hard to distinguish from composite volcanoes (McKnight and Williams 1997). Scoria cones are short lived volcanoes, their eruptive phase usually taking of

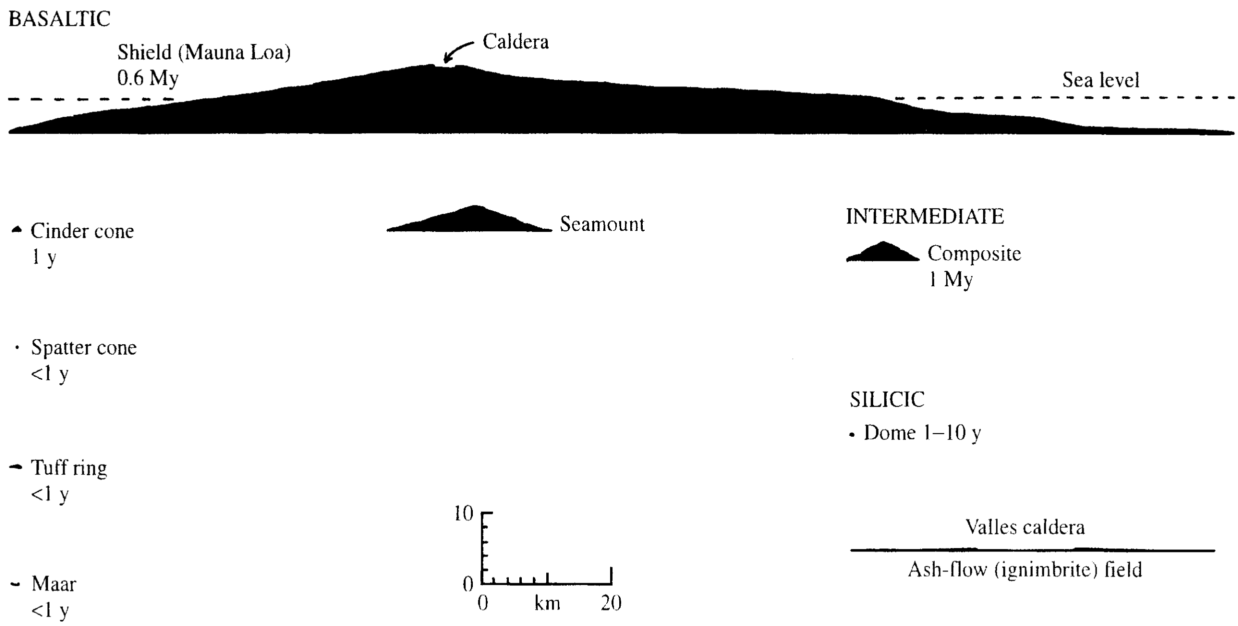


Figure 2.1. Comparative diagram of different volcanic landforms (after Best 2003: p. 246, fig. 10.6)

only a few days duration (Thordarson and Self 1993; Houghton et al. 1999; Vespermann and Schmincke 2000; Calvari and Pinkerton 2004). However, long lived scoria cone eruptions, such as Paricutin (Plate I, 2) (Foshag and Gonzalez 1956; Scandone 1979; Luhr and Simkin 1993) and complex scoria cones (Houghton and Schmincke 1989) are also known. Maars and tuff rings (Plate I, 3) are commonly referred to as a "wet" counterpart of scoria cones (Lorenz 1973, 1986, 2000, 2003), and they considered to be the second most common volcanic landforms on Earth (Cas and Wright 1988). They are produced by magma-water interaction triggered phreatomagmatic explosive eruptions (Lorenz 1973, 1975; Wohletz 1986; Heiken and Wohletz 1991; Zimanowski et al. 1997a; Zimanowski et al. 1997b), that may form a "hole-in-the-ground" (Lorenz 1970) crater-like morphology, commonly filled with water, such as those known from the Eifel, Germany (Schmincke 1977), or southern Patagonia, Argentina (Corbella 2002). Maars are also short lived volcanoes, rarely active more then few days, e.g. Ukinrek in Alaska (Self et al. 1980; Büchel and Lorenz 1993).

Sedimentation around these small volume, commonly mafic volcanoes, also described as monogenetic (indicating they formed in only one eruption event), can be very complex regardless of the small size of the volcano itself (White 1991). Erosion of a scoria cone can produce a broad, reworked scoriaceous sediment "halo" around the cone (Plate I, 4). The crater of the scoria cone may be completely filled up by reworked material, commonly mixed with aeolian sediments (Plate II, 1). The fluvial network around the cone field could be chooked by scoriaceous lapilli and ash, however, such systems have very small preservation potential in ancient settings (Ufnar et al. 1995). Maar volcanoes are more complex systems. The maar lake is often the only place where a detailed sedimentary record is preserved, such as Messel or other Eifel maars in Germany (Pirrung et al. 2001, 2003). Similar continental sedimentary records are also important in Central Europe such as Hajnacka in Slovakia (Vass et al. 2000), and Pula in Hungary (Hably and Kvacek 1998; Willis et al. 1999). Maar lakes are also sites where post-maar eruptions can build intra-crater scoria cones (Plate II, 2). Maars also form small-volume, complex sedimentary environments that are often difficult to recognized and interpret in ancient rock sequences.

Composite volcanoes differ from scoria cones, maars and tuff rings in their size and diversity of volcanic processes that have been involved in their evolution. Composite volcanoes are long lived and over their lifetime multiple and varied effusive and explosive eruptive products accumulate around them (DAVIDSON and DE SILVA 2000) forming diverse volcanic facies. Strato-volcanoes (Plate II, 3) are at least a magnitude larger in volume than scoria cones and are commonly active over millions of years, going through significant aggradational and degradational periods interspersed with periods of increased effusive and/or explosive activity or cone sector collapses, and intermittent long lasting inter-eruptive periods (FISHER and SMITH 1991). Strato-volcanoes are important in volcanic arc settings and over the few hundreds of thousands to millions of years of activity, significantly influence the sedimentation of the region in which they occur (FISHER and SMITH 1991). The ring plain around the strato-volcano is a complex playground (Plate II, 4), where proximal to distal volcanic facies can accumulate by diverse, primary to secondary processes (PALMER 1991; CRONIN et al. 1996; CRONIN et al. 1997; LECOINTRE et al. 1998; KARÁTSON and NÉMETH 2001; GIORDANO et al. 2002).

Volumetrically the largest volcanic systems are those associated with silicic caldera formation (LIPMAN 2000). Mega-calderas may reach sizes up to hundred km across (LIPMAN 1997, 2000). Such depressions are commonly filled by lakes hundreds of metres deep (Plate II, 5) that are long lived, in which thick successions of volcaniclastic sediments accumulate that are derived from the easily to remobilize of tephra from the caldera margin. The primary volcaniclastic sediment surrounding the caldera is also easy to remobilize (MANVILLE and WILSON 2003) and strong erosion carves deep valleys that quickly fill with volcaniclastic sediments (SEGSCHNEIDER et al. 2002). Caldera lake break outs generate large volume floods that could potentially affect thousands of km^2 of surrounding areas, e.g. Taupo, New Zealand (MANVILLE et al. 1999). Caldera forming eruptions may also strongly influence the pre-caldera fluvial network, as has been documented from the Central North Island of New Zealand after the AD 181 Taupo eruption (MANVILLE 2002). In caldera-lakes, post-caldera volcanism may form intra-caldera subaqueous volcanism with associated coherent and fragmented volcanic facies (LARSEN and CROSSEY 1996). Volcanic rocks in ancient settings commonly document similar history (RIGGS and BUSBYSPERA 1991). Caldera systems in which resurgence has occurred (KRUPP 1984; LUONGO et al. 1991; TIBALDI and VEZZOLI 1998; ACOCELLA and FUNICIELLO 1999; ACOCELLA et al. 2000; MORAN-ZENTENO et al. 2004) may uplift the caldera floor and expose successions accumulated in the caldera lake. In ancient settings caldera successions surrounding the caldera are represented by widespread volcanic rocks. Mapping over hundreds of kms across and 3D facies analyses are therefore essential for correct reconstruction.

Large volume, low slope angle volcanic edifices include oceanic island shield volcanoes, such as Hawaii. Significant parts of oceanic island shields are submarine with only the upper cone constructed subaerially (YANG et al. 1994). Large oceanic islands commonly became deeply dissected after the constructional phase of shield volcano development. Truncated morphology with large edifice collapse scars that can be associated with giant volcanic landslide fans on the sea floor are characteristic features of such stage of evolution of oceanic island volcanoes as has been documented in many active shield volcanoes such as Gran Canaria (FUNCK and SCHMINCKE 1998; CARRACEDO 1999), Tenerife (WATTS and MASSON 1995), Hawaii (CLAGUE and MOORE 2002; LIPMAN and COOMBS 2006), and Piton la Fournaise (OLLIER et al. 1998; OEHLER et al. 2004). These lava flow-dominated volcanic shields can be well preserved in ancient settings. The identification of specific lava flow fields associated with fissure vents, scoria cones as well as normal subaqueous clastic (dominantly volcaniclastic) sediments are critical to establishing the eruptive environment in which those rocks formed.

Pyroclast transportation and deposition in primary processes

Pyroclasts are the primary product of a volcanic eruption, i.e. a particle of any origin erupted through a volcanic vent (FISHER and SCHMINCKE 1984). After leaving the volcanic vent the pyroclasts are transported to a site of deposition. The main physical parameters controlling the transportation and deposition of the pyroclasts are (WILSON and HOUGHTON 2000); (1) the particle cohesion i.e. the stickiness (wetness of the pyroclasts) or the plasticity (hotness of the pyroclasts) of the pyroclasts, (2) particle trajectory, the pathway the pyroclasts traverse to their depositional point, (3) solid concentration or density of the eruption clouds. Major transportation agents of pyroclasts are; (1) pyroclastic density currents that are gravity controlled laterally moving mixtures of pyroclasts and gas, (2) pyroclastic falls in which pyroclasts free fall through the atmosphere from an eruption plume caused by explosive eruption, (3) pyroclastic flows that are inferred to be a type of pyroclastic density current, where the pyroclasts and therefore the majority of the momentum of the current is concentrated in the basal, particulate zone of the current, (4) pyroclastic surges that are a type of pyroclastic density current where the pyroclasts and therefore the main momentum of the current are widely distributed in dilute, highly turbulent particulate suspension.

The two major types of transportation agent of pyroclasts could be vertical or horizontal with respect to the ground surface. In vertical transportation the pyroclasts are transported upward by the gas thrust from the vent, and form an eruption cloud (umbrella) from where the pyroclasts fall under influence of gravity and wind direction. Laterally moving currents are driven by direct blasts or gravity collapse and pyroclast deposition is controlled by the speed, and the particle concentration of the current itself. These currents could also be affected by low level wind however the momentum of

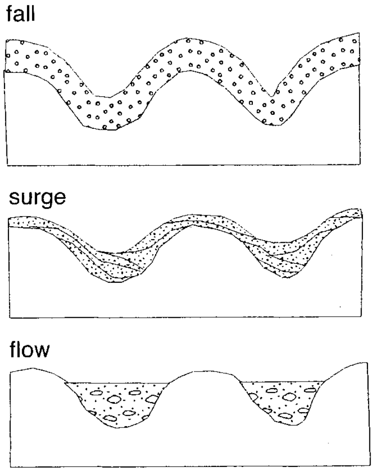


Figure 2.2. Schematic diagram of typical depositional features of pyroclastic fall, surge, and flow after WILSON and HOUGHTON 2000: p. 550, fig. 5

the current is usually large enough to overcome the wind factor. In addition, larger blocks erupt along ballistic trajectories.

The deposition of the pyroclasts by these transportation systems are controlled by the particle trajectory, particle concentration, particle cohesion and also the temporal fluctuations of the pyroclasts added to the system as eruption progresses. In a simplistic way, during pyroclastic fall (Figure 2.2), the vertical trajectory, and the low solid particle concentration are the main controlling parameters. These parameters lead to development of the mantle bedded and relatively well sorted characteristics of the pyroclast fall deposits (Plate III, 1). In pyroclastic surges, the horizontal trajectory and low particle concentration within a turbulent transportation system lead to a morphological depression infill, accompanied with occasional beds deposited on topographic highs (Figure 2.2). The deposits are rich in cross to dune bedded moderately sorted, matrix rich, commonly thinly bedded pyroclastic beds (Plate III, 2). In pyroclastic flows, the horizontal trajectory, the high particle concentration and a predominantly laminar flow structure of the transporting agent leads to deposition of pyroclastic deposits in topographical lows (Figure 2.2). Pyroclastic flow deposits tend to be massive, unsorted, weakly stratified and matrix supported (Plate III, 3). During the eruption transition between flows and surges may occur, predominantly due to particle concentration fluctuations during the eruption.

Classification of eruption types based on fall deposit characteristics and mode of magma fragmentation

Classification of volcanic eruptions from modern volcanic systems is commonly based on the pyroclastic fall deposits generated by a volcanic eruption (WALKER 1973). In this classification scheme the dispersal and the degree of the fragmentation of the magma is taken in account (WALKER 1973). The degree of the fragmentation of the magma is closely related to the grain size of the eruptive product (e.g. finer the grain size the higher the fragmentation). A casual relationship also exists between the height of the eruption cloud and the degree of fragmentation (WALKER 1973); the higher the eruption column the greater the fragmentation. On the other hand the higher eruption column, the more dispersed the pyroclastic fall deposits. This relationship is expressed in a range of similar empirical diagrams (Figure 2.3). On these empirical diagrams, eruption types have been identified, most of them refer to some historical eruption, commonly the type locality where the particular eruption style was first described. These styles of eruption were described from Stromboli (Strombolian) in Aeolian (Lipari) Islands in Italy, Hawaii (Hawaiian), Vesuvius (Vesuvian), and Mt Peleé (Peleean). The Plinian eruption style was named after Pliny the Younger, who gave a detailed description of the AD 79 eruption of Vesuvius, Italy (LIRER et al. 1997). Vesuvian (less energetic) and Plinian (more energetic)-type eruptions are end members of styles of eruption, both based on the AD 79 Vesuvius event.

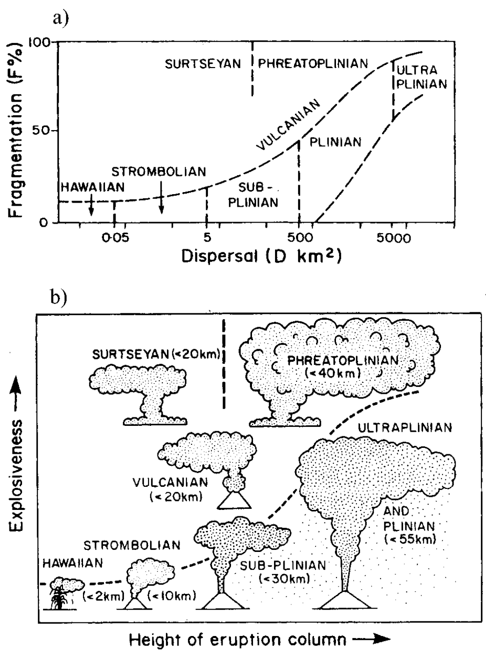


Figure 2.3. a) Dispersal (D) and Fragmentation (F) diagram showing different types of eruptions after WALKER 1973 classification scheme. b) a summary diagram explaining the differences between various types of volcanic eruptions. Diagram redrawn after CAS and WRIGHT 1988: p. 130

This chategorization of eruption styles generates major problems in interpretation of ancient deposits. In older pyroclastic rock records, especially in those representing distal facies accumulated in sedimentary basins, pyroclastic beds often represent condensed sections, comprising a number of different styles of eruption during a relatively short time period in the eruptive history of the same volcano. In addition, eruptive styles are time-dependent phenomena. Observations of active volcanoes demonstrate that eruptive styles quickly can change over time scales of minutes to hours. The resulting tephra successions reflect diverse eruption styles. In young volcanic systems, detailed tephra stratigraphy may allow each tephra layer to be assigned to a single eruptive style (Plate III, 4). In such young systems, even subtle changes in fragmentation style and in associated tephra dispersal pattern within a single event can be related to changes in eruption dynamics. In ancient settings, where only part of the tephra sequence may be preserved each pyroclastic layer, could be interpreted as results of individual eruptions with different style of individual eruption styles either representing a composite record of a complete eruptive cycle or only a temporal variation in the eruptive cycle of one single volcanic event. Interpretation of ancient pyroclastic deposits therefore needs a very careful sedimentological facies analysis in order to establish the time frame in which the succession was emplaced. In the identification of certain eruption style also could be misleading. Type volcanoes such as Hawaii or Stromboli produce far more complex eruptive events than just simple Hawaiian or Strombolian eruptions. In Hawaii, from time to time magma-water interaction triggers violent and explosive phreatomagmatic eruptions producing phreatomagmatic tephra such as the Keanakakoi (McPhie et al. 1990) or Uwekahuna (Dzurisin et al. 1995) Ash Members. The eruption type producing this tephra unit is significantly different from those characterised by *sensu stricto* Hawaiian-style eruptions. In Hawaii lava fountaining produced by normal Hawaiian eruptions may also switches to the more gas bubble disruption-driven Strombolian-style eruptions with its higher eruption clouds and more widespread tephra dispersal. Stromboli, also often produces sub-Plinian to Plinian eruptions (Rosi et al. 2000; Aiuppa and Federico 2004; Francalanci et al. 2004) and the typical Strombolian-style eruption is usually accompanied by Hawaiian-style lava fountaining. The fragmentary nature of the preservation of pyroclastic succession in ancient rocks precludes this level of detailed interpretation.

In young tephra deposits accurate mapping of a single tephra fall unit and detailed granulometry analysis of the tephra gives measurable parameters that can be used to determine the eruption styles that produced the tephra bed (Figure 2.4). The empirical dispersal value (D) is determined by area surrounded by the 0.01 T_{max} isopach where T_{max} is the maximum thickness of the studied single tephra fall bed. The empirical fragmentation (F) value can be determined by the use of multiple sieve analyses of the tephra deposits collected from the area along the main tephra dispersal axis where the 0.1 T_{max} isopach cross the axis line of the tephra dispersal (Figure 2.4). The F value is the average percentage of the finer than 1 mm fraction of the samples collected and sieved from the 0.1 T_{max} isopach axis (Figure 2.4). In this classification scheme major fields are hawaiian, strombolian, sub-Plinian, Plinian, ultra-Plinian, Vulcanian, Surtseyan and phreato-Plinian. The D-F diagram of Walker (1973), while quantitative, restricted in use to young, well exposed and well-preserved tephras that can be readily disaggregated. Moreover observations of recent volcanic eruption suggest that the F estimates not necessary directly related to the level of fragmentation. Many events such as high winds, "vapour rich eruption cloud", and rain flush could alter the grain size distribution pattern of a single tephra unit and therefore the assigned F value may have nothing to do with either the fragmentation of the magma or the style of the volcanic eruption. Such uncertainty in the identification of certain styles of eruptions in young volcanic systems highlights the difficulty of interpreting older pyroclastic deposits, where the field data comprises dissected outcrops, fragmentary sequences and lithified beds.

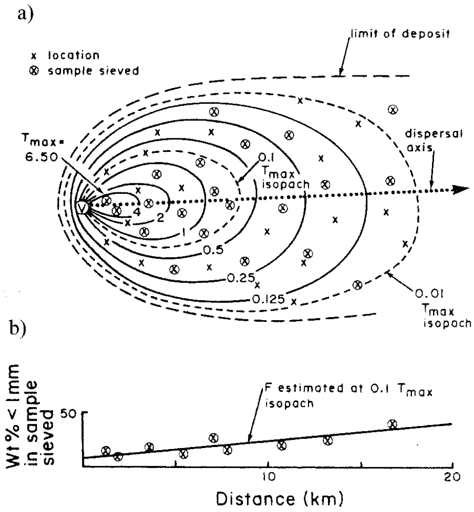


Figure 2.4. Representative diagram used to obtain Dispersion (D) and Fragmentation (F) data
a) map view of distribution of a single pyroclastic fall unit, b) diagram generated along the axis of the dispersal of the pyroclastic fall unit based on the sieved tephra samples (Cas and Wright 1988: p. 130, fig. 6.1)

Characteristic features of eruption types

Hawaiian-, Strombolian-style eruptions

This type of volcanic eruption is typical of basaltic magmas, where magma fragmentation is predominantly driven by the volatile content of the magma. This type of eruption is commonly inferred to be mild explosive, however, many authors suggest, that during the magma fragmentation in Hawaiian–Strombolian eruptions, no explosive process take place (e.g. shock wave generation). Instead, due to sudden decompression of the magmatic feeding system, the magma is torn apart by the constant, and rapid degassing (Head and Wilson 1989; Zimanowski 1998; Wolff and Sumner 2000). Hawaiian-style eruptions are characteristic of low viscosity mafic magmas that produce lava (fire) fountaining (Head and Wilson 1989; Wilson, L. et al. 1995; Wolff and Sumner 2000). During magma rise, large gas bubbles form due to the pressure drop near the vent (Head and Wilson 1989; Wilson, L. et al. 1995). These large gas bubbles are able to propel fragments from the volcanic conduits and form few tens (Plate IV, 1) to hundreds of metres high lava fountains. The lava fountains form a few tens of metres high lava spatter cones composed of welded agglutinates or lava spatter (Head and Wilson 1989; Thordarson and Self 1993). In the centre of the lava spatter cone, fall back lava can retain heat long enough to keep large lenses of magma melted and form small (few tens of metres wide) lava lakes. Lava spatter eruptions are a common form of activity during major inter-eruptive times of stratovolcanoes, e.g. Villarica in southern Chile (Plate IV, 1). Lava fountains commonly switch between Hawaiian and more fragmented Strombolian-styles of eruptions.

Strombolian-style eruptions (Plate IV, 2) are considered to be more explosive and result in the formation of monogenetic scoria (cinder) cones. This type of eruptive activity is driven by the volatile content of the magma and involves bursting of large gas bubbles as much as few tens of metres across in an open volcanic conduit (BLACKBURN and SPARKS 1976; JAUPART and VERGNIOLLE 1988; PARFITT and WILSON 1995; VERGNIOLLE and BRANDEIS 1996; VERGNIOLLE et al. 1996; HOUGHTON et al. 1999; SEYFRIED and FREUNDT 2000; SLEZIN 2003). Most of the ejecta are lapilli to coarse ash size (MCGETCHIN et al. 1974; SELF et al. 1974; RIEDEL et al. 2003). However, recent studies have shown that many scoria cones and their tephra deposits indicate fine grained tephra formation, and the potential transition from Strombolian-style to mafic sub-Plinian-style eruptions in the course of the activity of a larger volume scoria cone (MARTIN and NÉMETH 2006).

Vulcanian-style eruptions

The vulcanian eruption style is named after the type of eruption from Vulcano in the Aeolian (Lipari) Islands, Italy (Plate IV, 3). Typical Vulcanian-style eruptions start with a cannon-like blast, which clears the choked volcanic conduit (SELF et al. 1979; WOODS 1995; CLARKE et al. 2002). Subsequently steam blast-like eruptions take place at regular time intervals. These eruptions occur when the volcanic conduit is partially blocked by fall back ejecta and/or collapse of the wall rock over the magma-filled conduit. When the pressure builds up from the continuous degassing of the melt becomes large enough to lift up the lid, a new blast occurs. Similar rapid decompressions in laboratory experiments replicate the discrete, cannon-like vulcanian explosions and are able to produce two-phase flows through the experimental conduits (CAGNOLI et al. 2002). Large blocks are ejected ballistically (Plate IV, 4) in contrast to finer ash that is carried away in low eruption clouds (YAMAGISHI and FEEBREY 1994). When the conduit-blocking lid is disrupted, vitric ash and lapilli as well as breadcrusted bombs are ejected (FORMENTI et al. 2003). This type of eruption is commonly accompanied by pyroclastic surges and minor pyroclastic flows as has been documented from Asama in Japan (YASUI and KOYAGUCHI 2004) and Colima in Mexico (SAUCEDO et al. 2005). After the pyroclastic eruptions, effusion of lava commonly takes place. Vulcanian-style eruptions are commonly associated with Hawaiian, Strombolian, or sub-Plinian eruptive styles, as it has been documented from Ngauruhoe cone, in southern Taupo Volcanic Zone, New Zealand (Plate IV, 5). The Ngauruhoe stratocone has grown rapidly over the last 2,500 years in an alternation of effusive, strombolian, vulcanian, and sub-Plinian eruptions of andesitic magma (SELF 1975; NAIRN 1976; NAIRN and SELF 1978; HOBDEN et al. 2002).

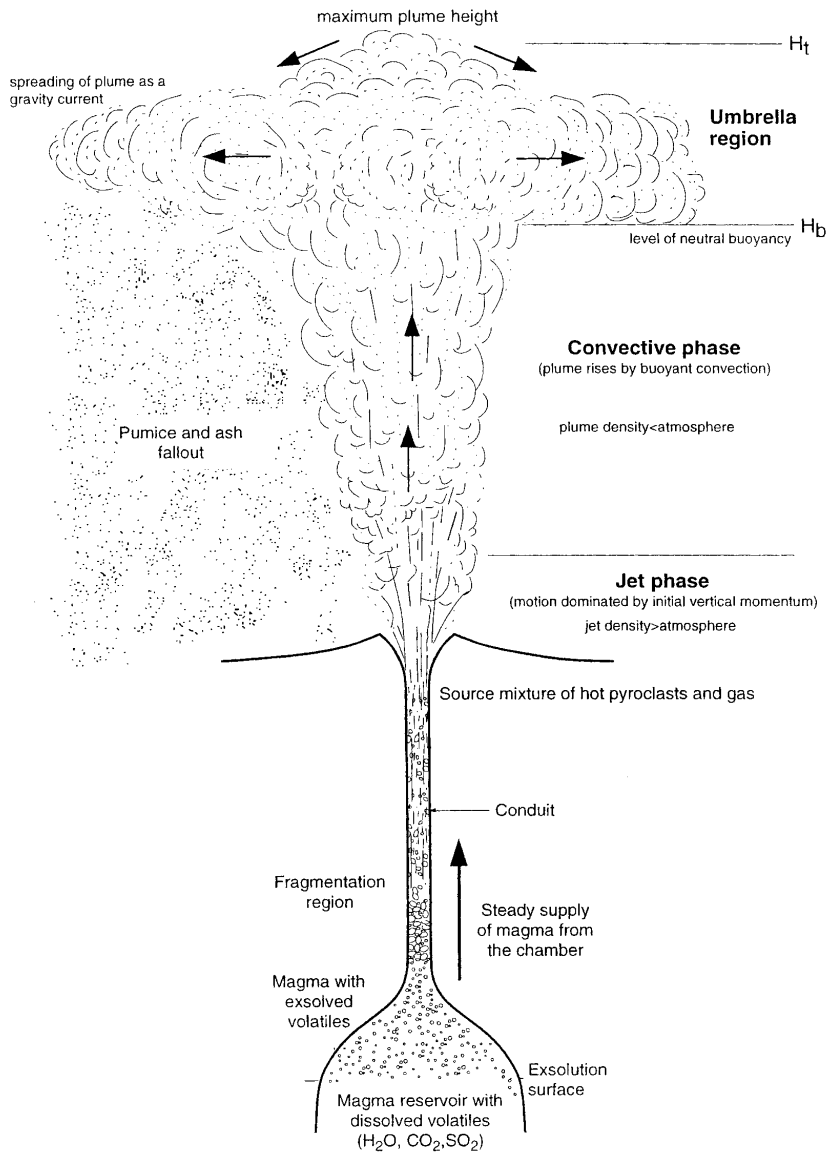


Figure 2.5. Parts of a volcanic plume generated by high silica magma eruption (after CAREY and BURSIK 2000: p. 529, fig. 1)

Plinian-style eruptions

Plinian eruptions are named after the AD 79 eruptions of Vesuvius. It is considered to be a highly explosive eruption where high (tens of kms) eruption clouds (Plinian plumes) form and tephra fall is widespread (Figure 2.5). The tephra dispersal profile is highly dependent on the main wind direction and the fallout pattern can be very asymmetrical (Figure 2.6). Plinian eruptions characterise the eruption of volatile-rich, highly viscous magmas such as dacite – rhyolite, however andesitic or even basaltic eruptions are known to have produced Plinian-type eruptions. S*ub-Plinian* eruptions are those, usually produced by mafic explosive volcanism, and the dispersal of fall out tephra is less than that of Plinian-type eruptions, however, the fragmentation level of the magma maybe as large that of other silicic magmas.

Eruptive velocities are hundreds of metres per second and eruptions could last for days or weeks. During the eruption however, intermittent periods may produce other, often less widespread tephra deposits. Interbeds of immediately reworked volcaniclastic successions (e.g. rain erosion) are common associates of thick (metres thick in proximal sections) Plinian fall out tephras (Plate V, 1). Many of the well known historic eruptions (e.g. AD 79 Vesuvius, 1883 Krakatoa) very destructive and started with formation of thick air fall tephra beds (Plate V, 2). Plinian deposits are rich in block to lapilli size silicic pumice and vitric ash. Plinian fall deposits are sheet-like and moderately to well sorted. One of the largest known historic Plinian eruption of the AD 180 Taupo pumice eruption in New Zealand, produced Plinian fall deposits over 10 m thick about 200 km from the source from an eruption cloud estimated to be about 50 km high (Walker 1980; Walker and Wilson 1983; Wilson, C. J. N. et al. 1995). Sedimentological features of the Taupo pumice eruptive products suggested that at least part of the eruption took place when the vesiculating volatile-rich magma encountered lake water (Wilson and Walker 1985), and magma-water interaction is inferred to be the reason of the formation of the high eruption cloud and very great dispersal of the tephra. Such eruptions are termed *phreato-Plinian*. Phreato-Plinian tephras are rich in very fine ash that are highly to non-vesicular. Accretionary lapilli are common in the tephra beds. Blocky shaped fine ash is also common.

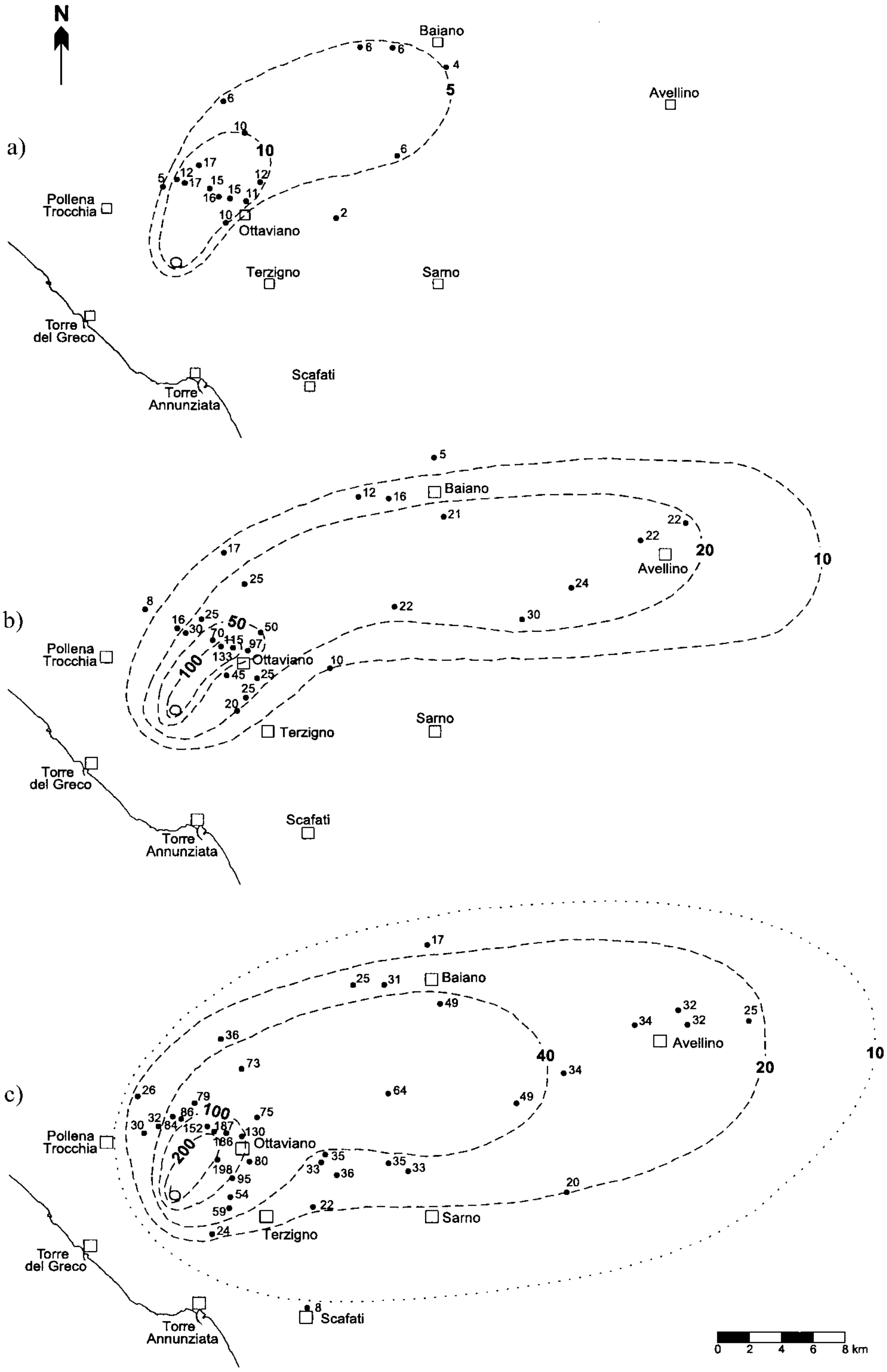


Figure 2.6. Plinian tephra dispersal map of the AD 472 eruption of the Somma volcano, Vesuvio, Italy after Rolandi et al. 2004: p. 304, fig. 8

Individual maps (a, b, and c) show the different distribution pattern of 3 individual fall unit associated with this eruption caused by slight differences/changes in the eruption and the prevailing wind direction

Surtseyan and phreatomagmatic eruptions

The Surtseyan eruption style was described from the 1963 eruption of Surtsey Island near the SW coast of Iceland (THORARINSSON 1967). In a broad sense, Surtseyan-style eruption refers to any eruption where basaltic magma fragmentation take place due to magma and water interaction driven steam explosions (WALKER 1973). The initial definition over the years highlighted many problems in usage of this eruption type as term. Many authors pointed out that there are significantly different types of pyroclastic deposits that could be formed in association with eruptions through a standing water body (lake, sea) or ground water (LORENZ 1974; KOKELAAR 1983; KOKELAAR and DURANT 1983; VERWOERD and CHEVALLIER 1987; WHITE 1996). The most striking difference is that where the external water is a standing water body, the pyroclastic deposits will be dominated by juvenile pyroclasts (WOHLETZ and SHERIDAN 1983; VERWOERD and CHEVALLIER 1987; SOHN 1995). In contrast, when the external water is ground water, the pyroclastic deposits will be rich in accidental lithic fragments (LORENZ 1985, 1986). It is also inferred recently where eruptions took place in standing water mass is involved such as the Surtsey eruption in where the characteristics of the resulting deepest deposits could very greatly depending on the water depth where the eruption occurred. This uncertainty also highlights the difficulty applying a simple term for an eruption. Consequently, the interpretation of ancient pyroclastic rocks could be extremely difficult and needs detailed study before proper interpretations can be made.

Volcanic Explosivity Index and Eruption Magnitude

Volcanic eruptions also can be classified on the basis of their explosivity index. The most widely used index of volcanic size is the 'volcanic explosivity index' (or VEI) of NEWHALL and SELF (1982). The VEI is a semi-quantitative logarithmic scale of eruption size, based on a combination of erupted tephra volume and eruption plume height. On this scale, the largest events (VEI 8) are defined as eruptions with bulk tephra volumes >1,000 km^3. For eruptions of this scale, much of the erupted tephra is in the form of ignimbrites, with a lesser component of ash fallout. One significant practical problem with the VEI scale, and indeed with all scales based on volume, is that it is based on estimated 'bulk volume' and takes no account of the deposit density. Since the density of freshly emplaced tephra deposits may vary by at least a factor of 3, this is potentially a significant problem (MASON et al. 2004). 'Volume' scales also suffer the perennial problem of interpretation when trying to distinguish between 'dense rock equivalent' (DRE) and 'bulk tephra' volumes quoted in the published literature. Assessments of erupted volumes are also prone to a number of potentially significant sources of error or omission. Such problem especially critical in case of caldera forming eruption, which are indeed potentially the largest explosive volcanic events. Large caldera-forming eruptions are associated with three main types of deposit: 1) intracaldera ignimbrite fill; 2) outflow ignimbrite sheets; and 3) tephra fall-out (from Plinian and/or co-ignimbrite ash clouds). It is very rare that the volume of these three type of deposits can be estimated, and therefore large under and overestimates can be made. Given the uncertainties in parameters other than direct estimates of the amount of erupted material, and the equivocal nature of 'volume'-based estimates alone, a logarithmic eruption magnitude scale of eruption size is preferred, that is continuous and based on erupted mass (PYLE 1995). The magnitude scale, M, is defined by:

$$M = \log_{10}(m) - 7.0$$

where m is the erupted mass in kg.

The magnitude scale is defined in this manner so that the scale is close to the original definition of the VEI (Figure 2.7). On this scale, M8 eruptions have masses in the range 1015 kg –

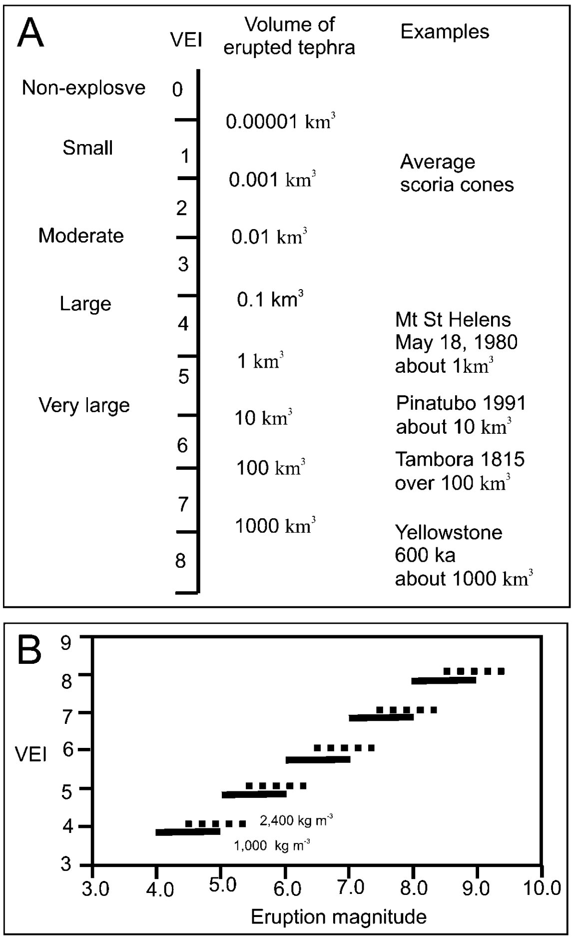


Figure 2.7. A) Relationship between VEI and erupted volume of tephra after NEWHALL and SELF (1982). B) The relationship between eruption magnitude, M, and Volcanic Explosivity Index, VEI, for deposits of bulk density 1,000 and 2,400 kg/m^3 (redrawn after MASON et al. 2004: pp. 735–748, p. 736, fig. 1)

1016 kg, and M9 eruptions have eruptive masses 1016 kg – 1017 kg. In terms of volume, M8 eruptions may have bulk tephra volumes of ~400–10,000 km^3, depending on the bulk density of the deposit (which may range from 1,000–2,500 kg/m^3), and dense rock equivalent (DRE) volumes, for typical rhyolitic compositions, of ~4004,000 km^3. The general relationship between the magnitude scale and the volcanic explosivity index is shown in Figure 2.7, for two representative deposit densities (1,000 and 2,400 kg/m^3).

Extrusion of magma, subsurface magmatic bodies

For the magma to reach the surface basically two requirements need to be met. Firstly a relatively open fracture or permeable conduit needs to exist from the deep magma source to the surface. The magma also needs a driving force which is able to transport the liquid magma through these pathways. The magma buoyancy force may able to open up fractures in favourable tectonic conditions, and create a pathway to the surface. In short, a magmatic overpressure needs to exceed the roof lithostatic pressure to be able to drive the magma to the surface. Various mechanical considerations of fluid-filled crack propagation (LISTER and KERR 1991) through the lithosphere that is under tectonic stress conclude that either extension (e.g. ascent of melt along open fractures) or tectonic inversion (magma ponding at the Moho and other density and/or rheology contrast zones in the lithosphere then expel magma by tectonic forces) seem to be viable mechanisms. During pure extension vertical dyke propagation is favoured. Magma can reach the surface and deep rooted, mostly monogenetic volcanic fields and mantle sourced volcanoes form (WATANABE et al. 1999). When the maximum compressional stress vector is horizontal and the minimum is in vertical position, melt can only reach the surface when the configuration temporarily switches into either pure extension or to the period when the maximum and minimum compressional stress vector change places (WATANABE et al. 1999). If the switching period is short, magma can be trapped in subsurface magmatic bodies and form sills and laccolithes (WATANABE et al. 1999). When the maximum compressional stress vector is in vertical and the minimum stress vector is horizontal position, but their differential stress is small (e.g. strike slip tectonic systems) magma can gradually reach the surface as multiple small batches. In this situation a large magma supply rate is necessary in general for the magma to be able to reach the surface (WATANABE et al. 1999).

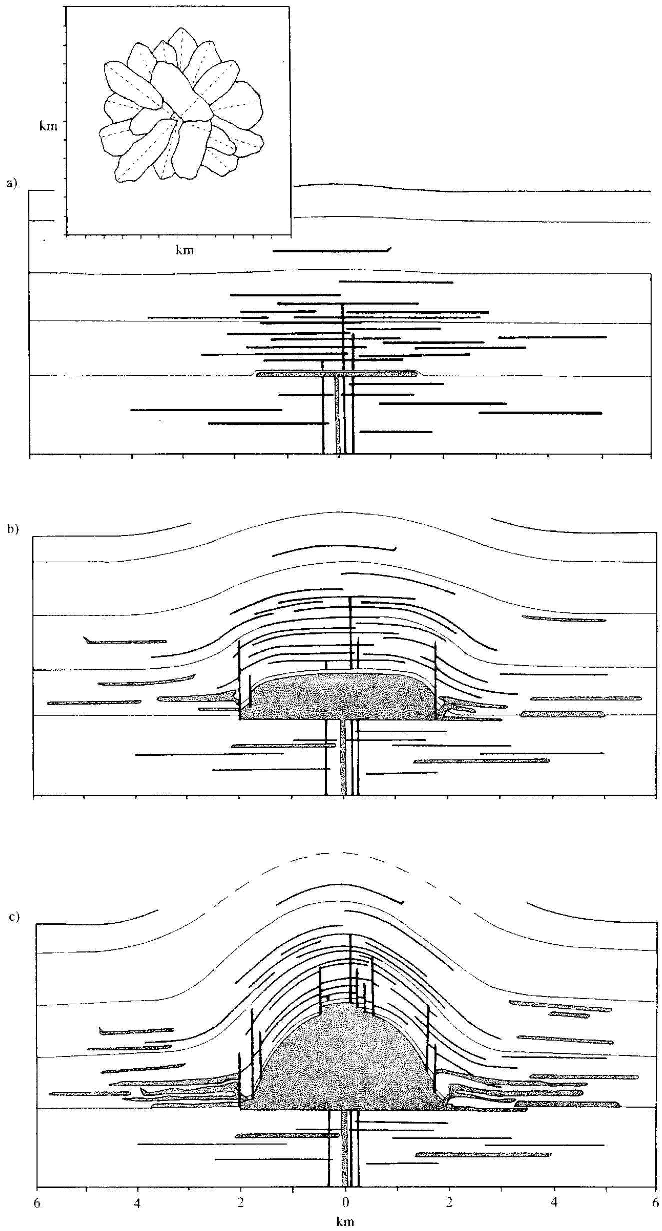


Figure 2.8. Laccolith architecture and growth modelled on the basis of laccolithes of the Henry Mts, Utah (after JACKSON and POLLARD 1988)

The three separate figures (a, b and c) demonstrate the gradual evolution of sills from which a laccolith

The basic physics of magma rise is controlled by the exsolving and expanding volatiles (due to pressure drop en route) that create buoyancy of the melt. During the buoyant (due to the density difference of the newly produced hot mafic magma and the host rock) magma ascends through dykes (WALKER 1989) and from time to time is stored in shallow crustal magma chambers where geochemical evolution (e.g. fractionation, crystallisation, volatile exsolution) can take place. From the shallow magma chambers the melt is able to continue its way up only when the volatile fluid pressure and the buoyant force exceed the tensile strength of the roof rock. Such processes can be;1) magma chamber crystallisation overpressuring the residual melt by volatiles, concentrating and increased buoyancy due to the gradual density drop of the system, 2) sudden decompression of the evolving shallow magma chamber due to volcano collapse, or other tectonic events (LIPMAN and MULLINEAUX 1981); 3) new hot, mafic, magma intruded into the bottom of the magma chamber (SPARKS et al. 1977; PALLISTER et al. 1992). The sudden heat transfer may create additional buoyancy causing the melt to erupt. In addition, the sudden cooling of the new mafic magma can create fast volatile exsolution that leads to volatile overcharge and can generate buoyant force, 4) external water entering into the magmatic feeding system, creating phreatomagmatic explosive disruption and unroofing, leading decompression and magma propagation upward. Magma that ponds in subsurface storages can form great variety of sill-like or laccolith bodies (HYNDMAN and ALT 1987; ZENZRI and KEER 2001). Laccoliths may form a complex system (Figure 2.8), and could be an important component of the total magmatic volume in the volcanic field, such as Elba, Italy (ROCCHI et al. 2002) or Henry Mt, Utah (HUNT et al. 1953; JACKSON and POLLARD 1988; FRIEDMAN and HUFFMAN 1998). Laccoliths may form a complex network of shallow subsurface volcanic features that have limited surface expression such as those in Texas (HENRY et al. 1997). Laccoliths are common in ancient settings and complex architecture of coherent body to the host rocks is more available for study as exposure extend to deeper levels (MOCK et al. 2005). In the Carpatho–Pannon region, Miocene laccoliths of dacitic to rhyolitic composition are known from the Visegrád (KORPÁS 1999) and Tokaj Mts (KULCSÁR and BARTA 1971).

Subaerial lava flows

Volcanic eruptions are commonly associated with wide range of lava flow effusions. Lava flows also can dominate the volcanic eruptions and form extended flood lava fields that cover areas over thousands of km^2. Such volcanic fields are called as large igneous provinces (LIP) and they played a significant role in the global volcanism of Earth (KENT et al. 1992; KING and ANDERSON 1995). Lava flow effusions and the resulting flow morphologies are predominantly controlled by the physical properties of the erupting magma such as temperature, viscosity and volatile content. Most of these physical properties are directly related to the composition of the melt that reaches the surface. The most common magma type is basalt. Approximately half of the total volume of volcanic rocks erupted are basaltic. Basaltic lava flows are characteristically low in viscosity, and therefore commonly form widespread sheet like lava bodies. Effusion of basaltic lavas sourced from fissure vents or central, point like vents or vent complexes. Large igneous provinces (ELLIOT et al. 1999) such as the Karoo, Parana-Paraguay, Columbia River or the Siberian fields (Figure 2.9) are extreme events in the Earth geological history and by many authors suggest that they directly influenced the evolution of life on Earth (PÁLFY and SMITH 2000; WIGNALL 2001; PÁLFY et al. 2002). In spite of the generally effusive eruption style associated with such large igneous provinces, there is growing evidence, that these fields may also have had phases of phreatomagmatic explosive activity in which vast amounts of pyroclastic rocks were produced as has been documented from southern Africa and from the Antarctica (WHITE and MCCLINTOCK 2001; ROSS et al. 2005; ROSS and WHITE 2005; MCCLINTOCK and WHITE 2006). Extrusion rates of such flood lava fields are thought to be huge in order to be able to produce tens of km^3 lava fields over very short period of times (thousands of years). Relatively high effusion rate driven fissure eruptions such as the Laki fissure eruption in 1738 in Iceland were able to produce about 13 km^3 lava flow field just in 28 days and emit vast amount of sulphur and fluor (THORDARSON and SELF 1993; FIACCO et al. 1994; THORDARSON et al. 1996; GRATTAN et al. 2005).

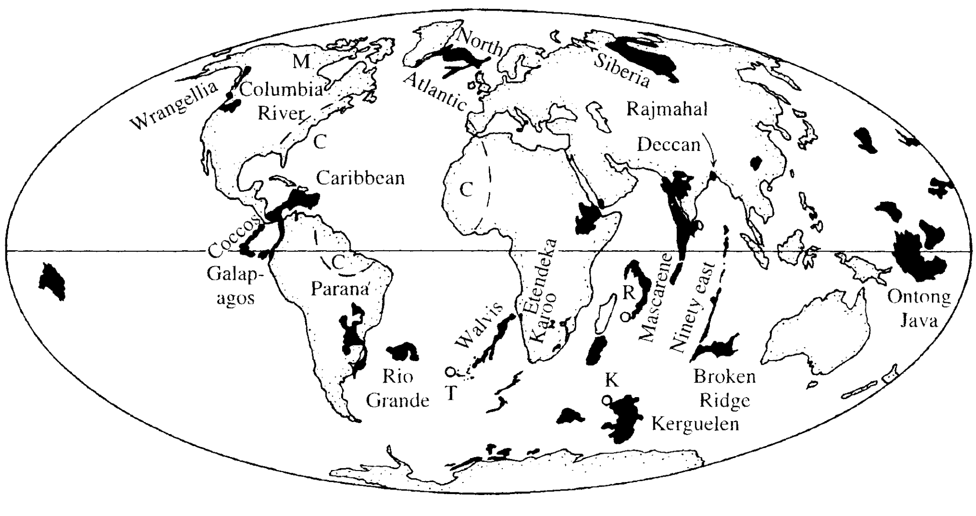


Figure 2.9. World map of location of major Large Igneous Provinces (after COFFIN and ELDHOLM 1994)

Basalt lava fields have flow surface features characteristic of eruption styles. During low eruption rate events pahoehoe lava (Plate V, 3) while during higher effusion rate events aa lava textures form. Pahoehoe lava form during low viscosity basalt effusion which consists of thin, glassy sheets, lava tounges and lobes, commonly in a complicated and overlapping manner (WALKER 1991). The pahoehoe lava fields are rich in lava tubes, where the melt is still hot and fast moving under a chilled tube wall (BYRNES and CROWN 2001). In this type of flow field lava is transported through master lava tube systems, and the flow fields gradually inflate, producing whale back structures, skylights, hornitos, and tumuli (SELF et al. 1998). There are two major types of pahoehoe lava flows (WILMOTH and WALKER 1993);1) S-type (spongy) has abundant ellipsoid vesicles inherited from the time the lava erupted 2) P-type (pipe vesicle-bearing) pahoehoe, that is lower in porosity, reflecting greated gas loss before its cooling. Tumuli are a common feature in pahoehoe lava fields. The magmatic pressure gradually uplifts the solidifying vesicular crust of the flow field. In bulbous pressure ridges tension cracks open up, and melt extruding, leaves behind star like clefts. Tumuli in pahoehoe lava fields occur as (1) lava-coated tumuli, (2) upper-slope tumuli and (3) flow-lobe tumuli (Plate V, 4) (ROSSI and GUDMUNDSSON 1996).

Aa lava flows are more common in more silicic magma effusion, however, aa lava is known in every composition of magma. The aa lava flows generally thicker than pahoehoe flows. The surface of the flow is blocky, clinker-like and commonly has steep flow fronts up to tens of metres in thickness (Plate V, 5). Because the more irregular surface morphology of the aa lavas in comparison to pahoehoe lava flows, the heat loss of the flow is generally greater, therefore the crystallisation can be more advanced in down slope of the flow and highly irregular shaped vesicles may form along the flow. Aa lavas are commonly channelised and along the lava channels levee formation is prominent (Plate VI, 1). Typical aa lava fields are common on the flank of Mt Etna, in Italy (Plate VI, 2). Although tumuli formation is common in pahoehoe flow fields, observation on the Etna lava fields confirm that tumuli also can form in aa lava fields. Detailed textural analyses of the tumuli on aa lava fields distinguish 3 types of tumuli (DUNCAN et al. 2004); (1) focal tumuli, which are formed from the break-up and uplift of 'old', thick lava crust and themselves become sustained sites for the distribution of lava both as flows and within distributary tubes. These focal tumuli are significant centres associated with major tubes. (2) Satellite tumuli, which are typically elongate, whale-back shaped features that branch out from focal tumuli. These satellite tumuli were initially lava flows erupted from a focal tumulus. The crust of the flow slowed or came to a halt and the rigid crust became uplifted and fractured, forming a dome-shaped ridge feature. These satellite tumuli continued to be fed from the focal tumulus and became sites of lava emission with numerous break-outs. (3) Distributary tumuli are formed on the fan associated with short-lived break-outs from tubes and are relatively simple structures formed from limited effusion of lobes and pahoehoe lava.

A continuous change from pahoehoe to aa lava surface morphology is commonly observed, suggest that greater apparent viscosity promotes aa lava formation as a result of temperature loss in down flow, which increases the viscosity of the lava flow. Transitional basaltic lava flow type between pahoehoe and aa lava is the toothpaste lava that form in case of advanced degassing and cooling of the flow. Another type of lava flow texturally similar to aa lava is the block lava flow. This consists of polyhedral chunks of lava blocks, many of them highly vesicular, and clinker-like. This type of lava flow exists from basaltic to highly silicic composition. Original lava surfaces are rarely preserved in ancient volcanic settings, especially in rock units sandwiched between other lithofacies. In the Cainozoic volcanic regions in Central Europe, only few lava surfaces are inferred to be original, and they belong to the younger then Miocene volcanic fields. Pahoehoe-like lava surface is known from the Mio/Pliocene western Hungarian, and southern Slovakian volcanic fields. Lava fields in the arc-related Miocene volcanic regions along the Carpathians are inferred to comprise aa lavas, as well as lava dome-related, preserved coherent lava bodies.

Lava flows commonly develop distinctive joint patterns after solidification of the melt called columnar jointure. Upon cooling the thermal contraction causes tensile stress that exceeds the brittle strength of the rigid magmatic body (SPRY 1962; BUDKEWITSCH and ROBIN 1994). The resulting extensional cracks initiate from point-like sources of the upper and lower margin of the flow (where the largest the heat lost) (DEGRAFF and AYDIN 1987). The resulting cracks are regularly sized polygons of four, five, six or seven sides. As the cooling advances, the cracks migrate toward the centre of the magmatic body, forming usually hexagonal joint pattern. The cooling cracks develop perpendicular to the isothermal cooling surfaces therefore the distribution pattern of the joints is useful to determine the position of the margin of the lava flow body regardless the erosion may remove the valley margins into the lava erupted (LYLE 2000). The columnar jointing has a typical distributional pattern (Figure 2.10) having a basal vertical colonnade, a central curving entablature and a top vertical colonnade layer (SPRY 1962). Jointing pattern analysis is in general useful in determining individual cooling units of lava flow bodies in ancient settings. Columnar jointing is common in association with lava flow remnants as well as volcanic conduit filling plugs or lava lakes in the western Pannonian Basin Mio/Pliocene monogenetic volcanic fields (Plate VI, 3). A complex joint-

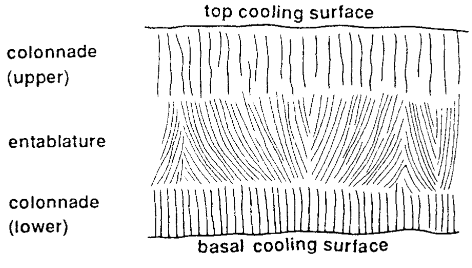


Figure 2.10. Columnar joint patterns of a basaltic lava flow according to SPRY (1962)

ing pattern is known from shallow subsurface coherent basaltic bodies of the western margin of the Mio/Pliocene monogenetic volcanic fields in Western Hungary (NÉMETH and MARTIN 2007) (Plate VII, 1).

Basaltic lava flows can exhibit a wide range of forms from a small, hundreds of metres long individual lava flows to extensive thousands of km^2 areas covered flood lava fields. Individual tongue-like basaltic lava flows are commonly associated with monogenetic, scoria cone fields. They are generally thin (m-scale) and not more then few km long (Plate VII, 2). In long lived volcanic fields however, the total volume of the accumulated lava could reach a few km^3 in volume, and could cover extensive areas such as the Pali Aike Volcanic Field in southern Patagonia (Plate VII, 3). In the Mio/Pliocene monogenetic volcanic fields in the western Pannonian Basin, small to medium sized shield volcanoes host the largest volume of basaltic eruptive products within the volcanic fields. Larger volume mafic lava fields are commonly associated with lava shields that are tens of kilometres across. In extreme situations, such lava shields could form a coalescent network of low-sloped lava dominated volcanoes such as are formed the Miocene to Pliocene volcanic fields along the Snake River in Idaho (Plate VII, 4). This volcanic field with the coalescent type of lava shield were described as a new type of continental volcanism (GREELEY 1982). The Idaho volcanic field no doubt is unusual, however, similar lava shield-dominated volcanoes are known from northern Africa and elsewhere, and probably form a transition between normal monogenetic volcanic fields and true flood lava fields.

Small to medium sized shield volcanoes are common in Iceland. Large lava shields are commonly associated with ocean island volcanism, such as Hawaii. Such ocean island shield volcanoes can be constructed over hundreds of thousands of years, gradually building up a low slope angle, lava flow dominated island.

The largest volume of mafic lava accumulation remains the continental flood lavas, such as the Columbia River Plateau (north-western US), Parana – Etendeka (South America and south-western Africa) or Karoo (southern Africa and Antarctica). According to many model calculations, flood lava fields accumulated in relatively short period of time. The Roza Member of the Columbia River Plateau has been estimated to have accumulated in between 6–14 years, with individual flow units emplaced over 5 to 50 months (THORDARSON and SELF 1998).

Volumes of silicic magma erupted in subaerial settings is volumetrically much less than mafic lava flow fields. Due to the higher viscosity of silicic magmas, the lava flow morphology has much larger aspect ratios (heigh to width ratio) than the mafic flow fields. Silicic lava extrusions can produce lava domes that are mushroom-like, partially solidified lava bodies commonly developed composite volcanoes. Lava domes are common in island arc volcanoes around the Pacific Rim, e.g. Japan or Mexico (Plate VIII, 1). Their size varies significantly from few tens of metres to km across. Such lava domes can be potentially hazardous, especially if they over-steepened. Such process can lead to initiation of deadly dome collapses similar to those the Unzen eruption in the early 1990s in Kyushu, Japan. Large, complex lensoid shape lava domes are also known along the Andean volcanic arc (Plate VIII, 2). These lava domes are individual effusions of silicic, mostly dacitic lavas, and form step-like lava dome volcanic complexes. One of the largest known Quaternary silicic lava body in the world is Cerro Chao in north Chile, a 14-km-long dacitic lava dome – coulee complex with a volume of at least 26 km^3 (DE SILVA et al. 1994) (Plate VIII, 3). Lava domes are complex features, and beside their central coherent body, they are surrounded by carapace-like marginal, autoclastic rock facies (Figure 2.11). Identification of lava domes in the rock record can be difficult. The identification of facies relationships with autoclastic carapace breccias, as well as potential association with lava dome collapse-originated block and ash flow deposits may help to distinguish lava dome bodies from other coherent lava masses (e.g. lava flows, sills, laccoliths). During lava dome growth, viscous silicic melt can be injected below the growing lava dome and make the dome growth endogenous. Other lava domes grow rather exogenously, by adding new lava to the growing lava surface.

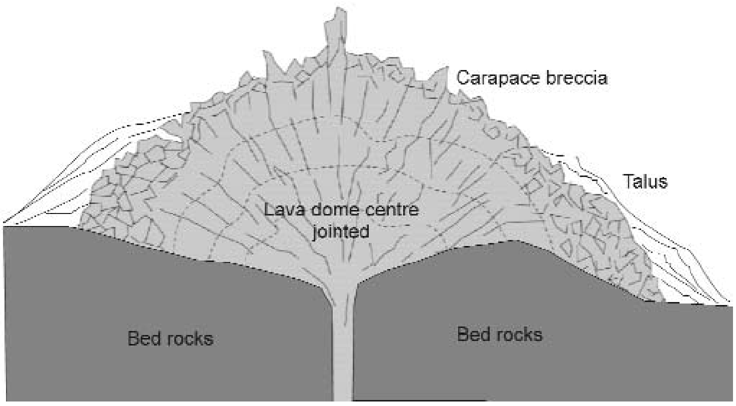


Figure 2.11. Dome cross section diagram. The central part of the lava dome is massive and commonly rosette-like jointed. Strong hydrothermal activity occurs in this part of the dome. The margin of the dome composed of angularly fractured coherent lava fragments, commonly altered and/or mineralized due to strong hydrothermal activity. The flank of the dome forms a steep sloped talus, commonly shows grain flow deposition

Andesitic lava flows are very common in arc-related settings. They are characterised by their steep flow fronts composed of highly vesicular to chilled margined autoclastic breccia and talus. The flow top is very irregular, with sharp edged, commonly scoriaceous blocky breccia. Flow foot is similar to the flow top. Jointing pattern of the coherent flow body could be platy or columnar jointed. Platy joints are generally steepened upward toward the flow fronts and margins.

Rhyolitic lava flows (and domes) have characteristic internal structure given by the variable level of vesiculation, fragmentation and devitrification of the flow. In ancient rhyolitic lava flows such zonation could be useful in identification of location within the lava flow (proximal to distal, base to top). The lava flow tops and bottoms are autoclastically fragmented and are highly variable vesicle content. The autoclastic to coherent body of the flow commonly contains entrapped blocks of lava. The internal part of the lava flow is flow banded. Lava flow interiors of young rhyolitic lava flows (less then a millions of years old) commonly contain obsidian (Plate VIII, 4). Obsidian glass in time gradually hydrates and devitrifies developing perlitic and crystalline textures (Figure 2.12). Devitrification in an early stage produces spherulites along the flow bands. In thick rhyolitic flows devitrification could take place straight after effusion, leaving behind completely devitrified central zones of the flow surrounded by a chilled glassy margin. Over long time periods, the entire coherent flow body devitrifies and the lava flow becomes a mass of aphanitic feldspar and quartz. Such flows can be challenging to identify in old settings. During crystallisation the continuous release of volatiles could lead to the formation of inflated bubble-like zones, lithophysal zones, which could host subsequently complex secondary mineral groups. Silicic lava flows are commonly flow banded where the banded texture given by the oriented feldspars (trachytic texture). Thick rhyolitic subaerial lava flow cross sections from the Little Glass Mountain, California (FINK 1980; FINK and MANLEY 1987) demonstrate typical zonation of rhyolitic lava flows. The flow is surrounded by carapace breccias directly attached to a finely vesicular zone of obsidian. Obsidian forms a relatively narrow belt below the previous zones. The central part of the flow consists of crystallized rhyolite surrounded by spherulitic obsidian. The central part of the flow is completely devitrified and crystallised, and only the margin stayed glassy (obsidian). Silicic lava flows are common in the Tokaj Mts in Northern Hungary, however, original flow surfaces are either not exposed, or not preserved.

Figure 2.12. Perlitic rhyolitic lava flow from the Newberry Volcano, Oregon (photo by S. J. Cronin)

Subaqueous lava flows

Subaqueous lava flows are as diverse as their subaerial counterparts. The most common subaqueous lava flows are basaltic, and form pillow lavas (Plate VIII, 5). Along mid-ocean ridges basaltic magma extrude on the sea floor. The quantity of the magma has been estimated by some to be large enough to cause sufficient temperature increase of the seawater to effect major circulation patterns in the ocean, leading to El Nino events (SHAW and MOORE 1988). On the sea floor massive tabular flow fields, thin sheet flows, block lavas as well as pahoehoe-like lava fields may develop. Surface features described from suabaerial lava flows such as pressure ridges, tumuli, or other inflation features (HONNOREZ and KIRST 1975; BATIZA et al. 1984; CAS 1992) are apparently similar in subaqueous settings. Basaltic lava flows immediately on emplacement could generate quench fragmented large volumes of hyaloclastite. Basaltic lava flows in subaqueous settings are commonly associated with lava spatter cones similar to those described from subaerial settings (HEAD and WILSON 2003). Identification of subaqueous lava flows can be difficult in ancient settings. Common association of the coherent lava bodies with hyaloclastite or lava deltas could be useful criteria to distinguish the two types of flows. Perhaps the most sure way of making the distinction is to establish the depositional environment of the coherent lava's underlaying and overlying sedimentary successions.

Subaqueous silicic lava flows have also recently been recognized as important volcanic features. The physical difference in subaqueous settings from subaerial ones, especially during the emplacement of silicic lavas is the sudden cooling, quenching of the melt, that may lead to the formation of large volumes of quench-fragmented lava clasts, termed hyaloclastite. Due to the high viscosity of the silicic lava flows, in subaqueous settings a lava flow could be dominated by quench-fragmented hyaloclastite, having only domains of coherent bodies preserved in the flow central parts. In the subaqueous environment, silicic lavas commonly form lava domes (Figure 2.13). The internal structure of the young silicic subaqeous domes are similar to those erupted subaerially. The shape of the dome could be flattened due to the higher hydrostatic pressure the water column places on the erupting lava dome. During the growth of the lava dome, in situ hyaloclastite continuously forms along the margin of the lava dome, and could completely hide the coherent body of the lava dome itself. In the case of pulsed intrusion and effusion of the silicic dome, a lava dome complex with thick and

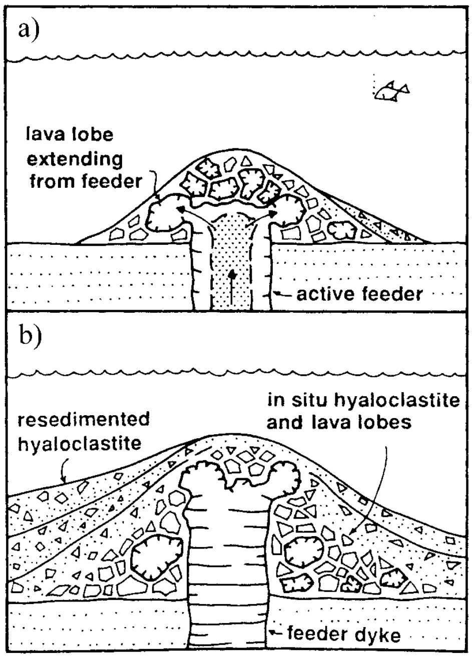


Figure 2.13. Subaqueous lava dome facies after YAMAGISHI (1987)

a) lava extruded into the sea/lake floor. Lava advance only short distance before it is quenched and fragmented, b) the growing hyaloclastite pile is intruded by further feeder dykes producing more in situe hyaloclastite. Gravitational instability may lead to collapse the hyaloclastite pile, leading to accumulate resedimented hyaloclastite nearby

complex in situ hyaloclastite margins could build up over relatively short period of time. Such a succession is well known from Ponza, Italy (SCUTTER et al. 1998; DERITA et al. 2001), and recently identified from the Tokaj Mts., Hungary. Identification of subaqueous silicic lava flows in ancient setting could also be difficult, especially if the quench fragmentation was strong, leaving behind a fragmented "pyroclastic rock-like" texture of the lava flow. Detailed mapping of the succession and the recognition of the facies association with hyaloclastite as well as the stratigraphic relationship with marine sediments, would confirm the subaqueous eruptive environment of such lava flows. Identification of subaqueous lava domes also need a similar 3D understanding of the coherent and fragmented volcanic rock facies before establishing the interpretation of the eruptive environment. In subaqueous environment, the ongoing non-volcanic sedimentation may produce a thick succession of fresh mud and silt that is easily invaded by extruding silicic magma. In such cases, the silicic magma intrudes into the syn-eruptive sedimentary succession and truncates and lifts the succession up. The coherent silicic magma body may stay hidden under the uplifted sea (lake) floor, and a cryptodome may form. Contact between the coherent magma body and the sediments could be complex (e.g. peperite). Intrusive hyaloclastite could also form large volumes of fragmented rocks that are hard to identify in ancient settings.

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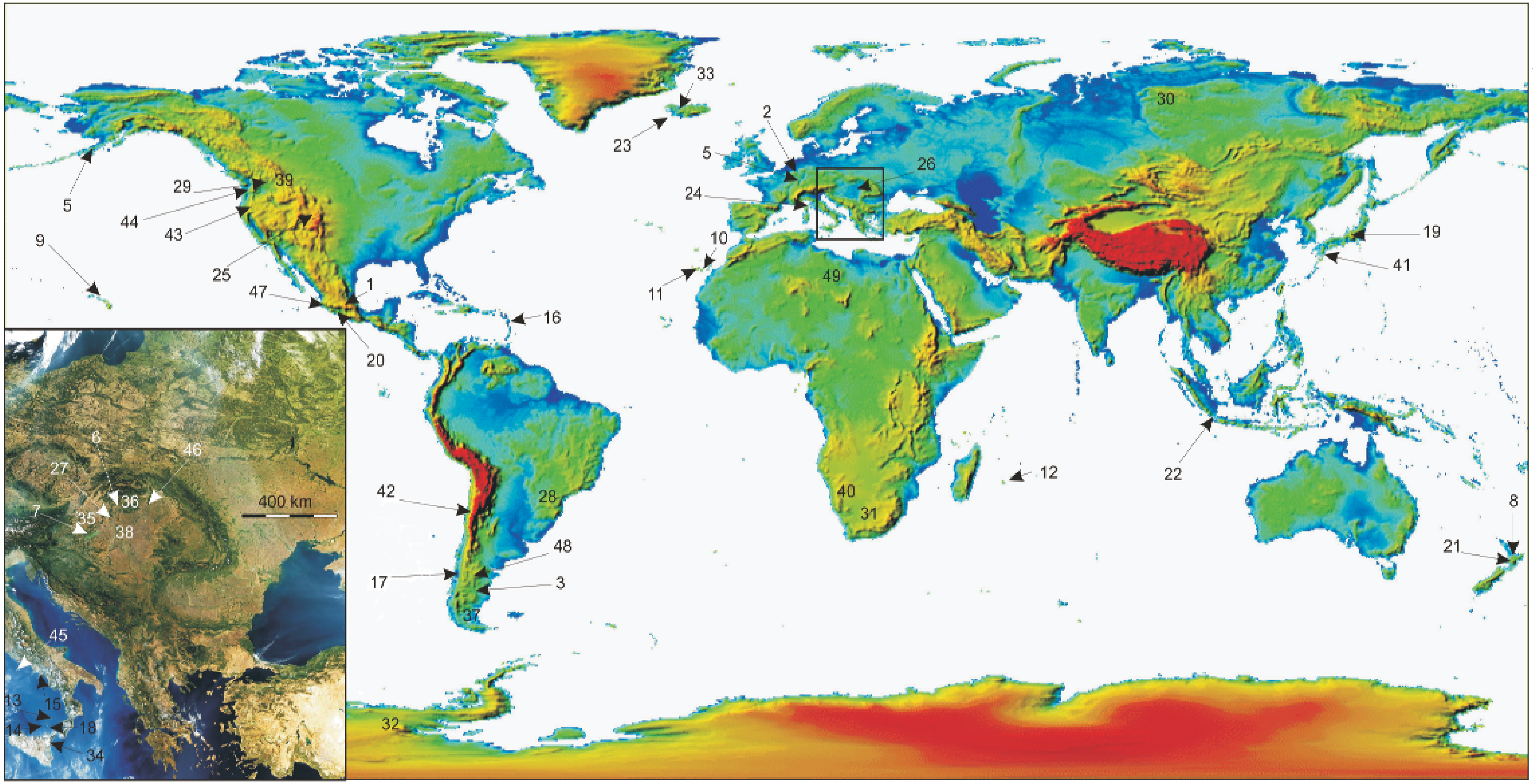
Location map



1 — Paricutin, Mexico
2 — Eifel, Germany
3 — Patagonia, Argentina
4 — Ukrinek, Alaska, USA
5 — Messel maar, Germany
6 — Hajnacka maar, Hungary
7 — Pula maar, Hungary
8 — Taupo Volcanic Zone, New Zealand
9 — Hawaii, USA
10 — Gran Canaria, Spain
11 — Tenerife, Spain
12 — Piton la Fournaise, Reunion France
13 — Stromboli, Italy
14 — Aeolian (Lipari) Islands, Italy
15 — Vesuvius, Italy
16 — Mt Pelee, Martinique
17 — Villarica, Chile
18 — Vulcano, Italy
19 — Asama, Japan
20 — Colima, Mexico
21 — Ngauruhoe, New Zealand
22 — Krakatoa, Indonesia
23 — Surtsey, Iceland
24 — Elba, Italy
25 — Henry Mts, Utah, USA
26 — Carpatho–Pannon Region
27 — Visegrád Mts, Hungary
28 — Parana – Paraguay LIPs
29 — Columbia River Basalt, USA
30 — Siberian LIPs, Russia
31 — Karoo LIPs, South Africa
32 — Transatlantic Mts
33 — Laki Fissure, Iceland
34 — Mt Etna, Italy
35 — Western Hungary
36 — Southern Slovakia
37 — Pali Aike, Argentina
38 — Pannonian Basin
39 — Snake River, USA
40 — Etendeka, SW Africa
41 — Unzen, Japan
42 — Cerro Chao, Chile
43 — Little Glass Mts, USA
44 — Newbery Volcano, USA
45 — Ponza, Italy
46 — Tokaj Mts, Hungary
47 — Volcan Ceboruco, Mexico
48 — Carapacho, Argentina
49 — Al Haruj, Libya

1. Medium sized scoria cone in the NW side of Volcan Ceboruco, Nayarit, Mexico.
2. One of the largest scoria cones on Earth, Paricutin is located in Central Mexico.
3. Carapacho tuff ring in Mendoza, a typical tuff ring with constructional edifice with wide crater, Argentina.
4. Strongly eroded Neogene scoria cones from the Al Haruj Volcanic Field of Libya. Note the reworked volcaniclastic halo surrounds the erosion remnant of the scoria cones.

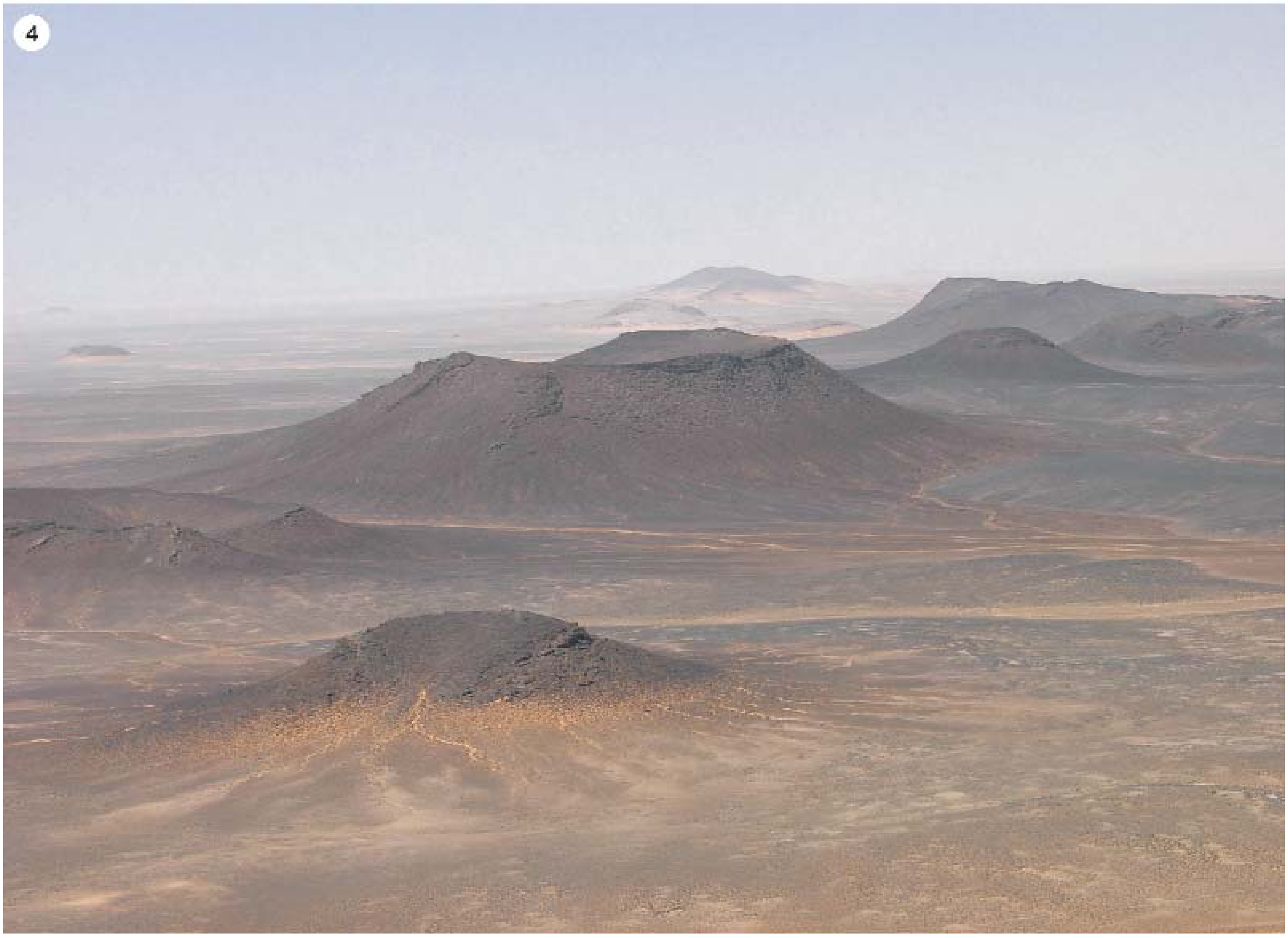


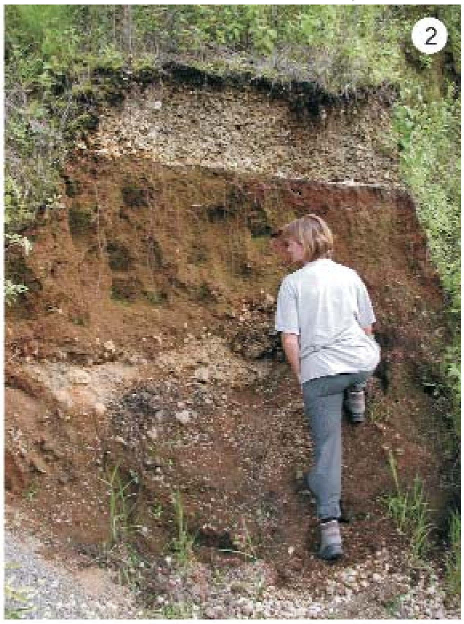
Plate II

1. Aeolian sediments filling eroded scoria cone craters from the Al Haruj Volcanic Field of Libya.
2. Dry maar crater of the La Breńa volcano in Durango, Mexico. The maar crater is filled with post-maar lava flows and a scoria cone.
3. Typical stratovolcano from the Southern Andean Volcanic Zone, Chile. Osorno volcano has a snow-capped peak and very even morphology.
4. Wide ring plain surrounding a stratovolcano such as the plain around the Volcan Ollagüe in Chile, is a playground of primary, secondary volcanic and normal non-volcanic sedimentation.
5. Toyo caldera in Hokkaido, Japan. In the centre of the caldera lake is a group of lava domes as a result of caldera resurgence.

1. Mantle bedded pyroclastic fall beds in SW Hokkaido, Japan. Note the even bed thicknesses of tephra.
2. Pyroclastic surge beds with large dune structures close to the vent of Laacher Sea maar in the Eifel, Germany. Arrow represents inferred flow direction.
3. Pyroclastic flow deposit from the Calbuco volcano, southern Chile. The deposit contains charcoaled, standing tree trunks as a result of high temperature pyroclastic density current activity.
4. Complex tephra succession from the NE ring plain of Ruapehu Volcano, North Island.

Plate IV

1. Lava spatter eruption from the central vent of the Villarica Volcano from Central Chile.
2. Strombolian explosion in night time at Stromboli, Southern Italy (photo by S. J. Cronin).
3. Wide, dish-like crater of Vulcano, Southern Italy. Note the steep inner crater wall as well as the radial erosional gullies.
4. Few metres across, bread crusted, jointed blocks are common around the crater rim of Vulcano, as a result of sudden and catastrophic depressurisation of solidified lava lid during Vulcanian eruptions.
5. Ngauruhoe cone, part of the Tongariro – Ruapehu volcanic chain in the Central North Island, New Zealand.



1. Typical ring plain succession of a major strato-volcano such as the Taranaki (Mt Egmont) in New Zealand. The ring plain succession is a complex depositional sequence of primary distal Plinian fall, volcanic debris flow, and debris avalanche deposits.
2. Plinian air fall bed (white) forms the youngest unit over scoria cone successions near Ceboruco Volcano, western Mexico.
3. Pahoehoe lava formation on an active vent of Kilauela (photo by S. J. Cronin).
4. Flow-lobe tumuli on the extensive lava fields of Al Haruj, Libya.
5. 'Aa' lava flow from the Tromen Volcano, Argentina.

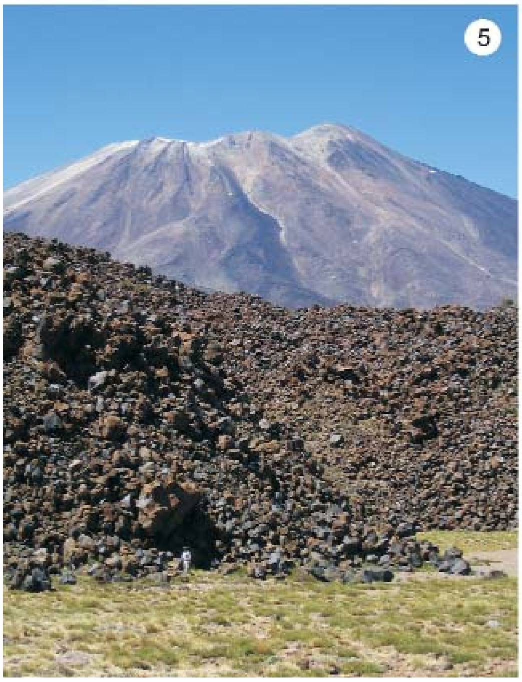


Plate VI

1. ‘Aa’ lava channels on the flank of Tromen Volcano, Argentina.
2. Active ‘aa’ lava flow during the Etna 1993 eruption (A). On a close up view the collapse of hot lava block is clearly visible (B).
3. Well-developed columnar joints in the basanite plug of Hegyestű, Hungary.

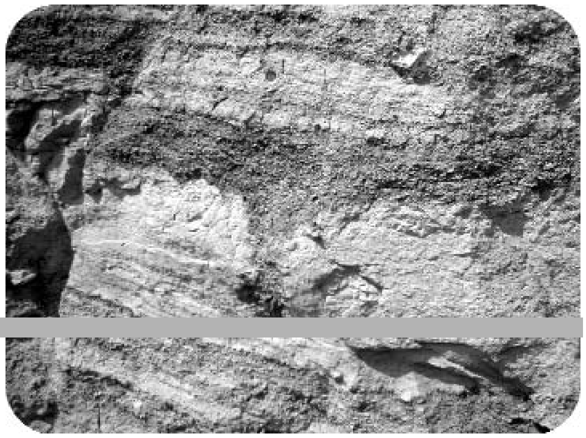
1. Radially jointed basanite as part of a shallow subsurface architecture of a complex small volume basaltic volcano system at Uzsa, Western Hungary.
2. Long lava flow initiated from the Santa Maria scoria cone in Payunia, Argentina.
3. Complex lava fields initiated from one of the youngest scoria cone, Laguna Azul of the Pali Aike volcanic Field, Santa Cruz, Argentina.
4. Low slope angled lava shields of the Snake River plain in Idaho, Pacific Northwest, USA.

Plate VIII

1. Active lava dome of the Showa Shinzan complex in Hokkaido, Japan.
2. Large (tens of kilometres across) lava dome complex in the Altiplano, Chile.
3. Large volume Chao Chao dacitic lava coule in the Altiplano, Chile. Note the steep slope of the coule.
4. Banded strongly spherulitic rhyolite lava flow remnant from the North Island, New Zealand.
5. Miocene pillow lava from associated with an emergent tuff cone at the Boatman Harbour Oamaru, New Zealand (photo by S. J. Cronin).

Chapter 3

Pyroclastic (volcaniclastic) rocks as a key to interpreting volcanic explosive processes and environments



Introduction

Pyroclastic rocks are those that are produced directly by explosive volcanic eruptions. Similarly, "pyroclasts" are erupted clasts of any origin, propelled out from a volcanic vent by any eruptive processes (FISHER and SCHMINCKE 1984, 1994). Pyroclastic rocks may form after diagenesis of freshly deposited sediment. The related term but broader term "volcaniclastic" rock, refers to any type of clastic rock that contains fragments of volcanic origin. This latter term hence carries little information about the formation of the rocks, although it often implies reworked pyroclastic sediments. It is often used to describe rocks or deposits with unknown or suspected secondary origin. In general, autoclastic rock types refer to fragmental rocks that comprise clasts generated by *in situ* fragmentation of coherent magmatic bodies (FISHER and SCHMINCKE 1984, MCPHIE et al. 1993). Autobreccias commonly form on the base and top of lava flows (FISHER and SCHMINCKE 1984, CAS and WRIGHT 1988, MCPHIE et al. 1993).

Clastic volcanic rocks are widespread and important volcanic rock types in the geological record. Detailed study of such rock types can provide important information for the magma fragmentation and vesiculation, both through studies of the component volcanic particles and their overall characteristics (sedimentary structure, grainsize distribution, contacts, geometry, etc) (Figure 3.1). To generate any type of pyroclastic rock, rising and erupting magma must first be fragmented. "Magmatic" fragmentation is the typical fragmentation process, in which coherent melt bodies rising to the surface are disrupted by the exsolution and expansion of gas (primarily H_2O and CO_2) formerly dissolved in the magma (CASHMAN et al. 2000). By contrast, "phreatomagmatic" fragmentation also (and in some cases, primarily) involves contact between magma and external water — ground or surface waters (MORRISSEY et al. 2000). During magmatic fragmentation the volatile content of the magma and melt viscosity play the most important roles (CASHMAN et al. 2000), whereas during phreatomagmatic fragmentation, it is the external cooling media: water (ZIMANOWSKI 1998, MORRISSEY et al. 2000, ZIMANOWSKI and BUETTNER 2003), water-saturated sediment (WHITE 1996a), or warm hydrothermal systems (BERTAGNINI et al. 1991). Any resulting pyroclasts can become remobilized, redeposited and reworked (Figure 3.1).

Studying the primary juvenile fragments of any type of fragmented volcanic rocks at the microscopic level (Plate I, 1) can give vital information about the chemical evolution of the melt. These studies are used to establish the process of how the magma formed and travelled to the surface. The principal questions that can be solved in these studies include; how the magma formed, how quickly and under what conditions it was transported to the surface and under what conditions it emerged (e.g., open versus partially closed vents) (MARSH 2000, RUTHERFORD and GARDNER 2000). Here we do not intend to give detailed explanations of volcanic petrology, but rather concentrate on how juvenile fragments can be used to establish how magma fragmentation took place.

Along with microscopic studies of juvenile fragments, sub-microscopic (SEM – Scanning Electron Microscopy) views of the shape and crack structure of the juvenile clasts can be used to determine the fragmentation and vesiculation history of the magma (HEIKEN 1972, 1974, HEIKEN and WOHLETZ 1986, 1991, DELLINO and LIOTINO 2002, ZIMANOWSKI et al. 2003). Magmatic volatiles come out of solution in rising magma due to its decompression near the surface (CASHMAN and MANGAN 1994a, TORAMARU 1995, CASHMAN et al. 2000). Vesiculation can take place prior, during or after the fragmentation of the

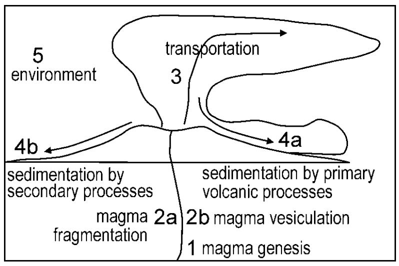


Figure 3.1. Diagram about volcanic processes from magma generation to tephra redeposition. The resulting volcanic rocks will carry characteristic textural features indicative to the volcanic history of the rock from magma to the rock

magma, although it mostly occurs beforehand, hence driving the disruption of the melt (CASHMAN and KLUG 1995). The timing of vesiculation in comparison to fragmentation can give vital information on the style of magma rise (MANGAN et al. 1993, NAVON and LYAKHOVSKY 1998). Also in microscopic and/or hand-specimen level, detailed textural analysis of the fragmented volcanic rock can help to understand the level of fragmentation, e.g. whether it took place in open vent/conduit system, or well below the syn-eruptive palaeosurface (CIONI et al. 1992, HOUGHTON et al. 1999, MASTROLORENZO et al. 2001). Detailed microscopic/macro studies of the accidental lithic fragment population (fragments that were disrupted from the pre-volcanic country rock column) also helps to characterise the depth of magma fragmentation, especially during phreatomagmatic explosions (NÉMETH 2003). Many active volcanic eruption observations suggest that both magmatic and phreatomagmatic fragmentation commonly take place during a single volcanic eruption, and styles of fragmentation may quickly alternate (SELF et al. 1980, ORT et al. 2000).

After magma fragmentation, pyroclasts can be transported by many ways (Figure 3.1). They are commonly transported upward by "gas thrust" of buoyant hot gases to form an eruption column, which if sustained forms a stable eruption plume (CAREY and BURSIK 2000). Pyroclasts can stay in the eruption cloud as long as the eruption plume has enough upward momentum to keeping them lofted (CAREY and BURSIK 2000). When the eruption cloud is not sustained any longer, pyroclasts are drifted away by wind in an eruption cloud from where tephra can fall under gravity (CAREY and BURSIK 2000). In this way pyroclasts are transported to the depositional site via gradual fall-out from a suspension cloud. This can occur in both, air (subaerial) (CAREY and BURSIK 2000, WILSON and HOUGHTON 2000) and water (subaqueous) (FISKE and MATSUDA 1964, CASHMAN and FISKE 1991a, 1991b, CAS 1992, COUSINEAU 1994, KANO et al. 1996, FISKE et al. 1998, ALLEN and MCPHIE 2000, MUELLER et al. 2000, ALLEN and CAS 2001). When an eruption column is overcharged with particles, it may collapse quickly to form high particle concentration pyroclastic flows which hug the ground and are channelled into the valley network surrounding a volcano (CAS and WRIGHT 1988). Pyoclasts are also transported laterally within "direct blasts" (CRANDELL and HOBLITT 1986), or ground-parallel moving, low particle concentration turbulent gas-ash currents, called pyroclastic surges (FISHER 1979, WOHLETZ and SHERIDAN 1979, CAREY 1991, COLE 1991, COLELLA and HISCOTT 1997, VALENTINE and FISHER 2000). To establish the transportation and consequently the depositional mechanism of pyroclasts, microscopic or hand-specimen evidence often needs to be supplemented by outcrop- or landscape-scale field evidence (Plate I, 2). In ancient volcanic terrains, especially in heavily vegetated areas, only hand specimens may be available (Plate I, 3). In this case, there is little chance to establish more of the origin of the pyroclastic rock other than its vesiculation and fragmentation history. To interpret transportation and depositional environment at least meter-scale outcrops are necessary (Plate I, 4). At this scale the pyroclast supporting agents (gas, water, or mixed media; traction, saltation, suspension, ballistic ejecta) can be established (WILSON and HOUGHTON 2000), along with the depositional environment, e.g. subaqueous versus subaerial (Figure 3.1). For the latter, usually good outcrops and potentially a larger scale 3-D understanding of the facies-architecture of the rock units are essential. 3-D volcanological mapping of volcanic rocks over large areas can be used to interpret the eruptive environment and draw a full picture where and how the volcanic eruption took place (Plate I, 5). Volcanic facies analyses are similar to those commonly used in any other sedimentary environment. In this respect, studies of the entire sedimentary succession, including pre and post-volcanic successions are needed for reconstruction of the volcanism in a certain geological time frame.

Features characteristics of different styles of magma fragmentation

The style of magma fragmentation can be studied by normal petrological microscopy or using scanning electron microscopy (SEM) (Figure 3.2) (WOHLETZ and KRINSLEY 1978, BUTTNER et al. 2002, DELLINO and LIOTINO 2002). During magmatic fragmentation, exsolution and expansion of gas causes bubbles to form in the melt, these grow to disrupt the magma. Hence the pyrcolasts have, shapes either defined by curved bubble-walls or contain large numbers of well-developed vesicles with oval, elongate or pipe shapes (Plate I, 6) (HEIKEN and WOHLETZ 1986, 1991). For low viscosity melts (e.g. lava spatter), dark (limited transparency under petrological microscope) droplet-like fragments may

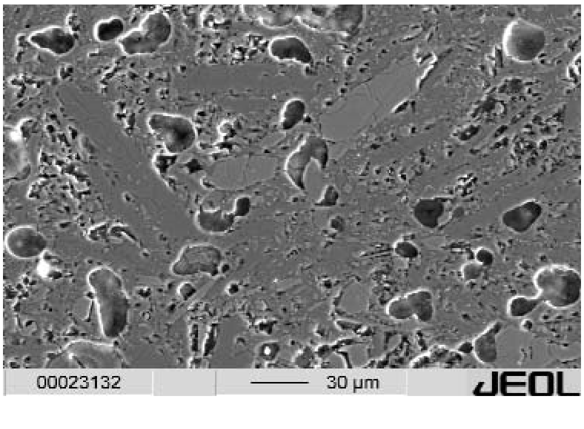


Figure 3.2. SEM studies of volcanic glass shard can help to establish the fragmentation history of the melt. Blocky, fractured glass shards are characteristic for magma water interaction driven explosive fragmentation. An SEM picture of volcanic glass shard from the Volcan Cerro Colorado tuff cone in Sonora, Mexico, shows fractured texture, low vesicularity and angular contour of vesicles with thin palagonite rim, all features characteristic for phreatomagmatic fragmentation

form as well as those that have characteristic features of fracture and bond surfaces characteristic for a high temperature state of the pyroclasts upon formation, transportation and deposition (HEIKEN and WOHLETZ 1986, 1991). During ongoing fragmentation and vesiculation, a high gas thrust in the volcanic conduit can lead to highly sheered margins of the magma against conduit walls and a larger bubble-rich central zone (HEIKEN and WOHLETZ 1986, 1991). As a consequence, elongated pyroclasts with stretched vesicles are formed from magma of these sheared regions (HOUGHTON and WILSON 1989, THOMAS et al. 1994, KAMINSKI and JAUPART 1997, MASTROLORENZO et al. 2001). Scoria (cinder) or pumice is the most common clast types resulting from vesiculating magma. Scoria (Plate II, 1) is generally considered to be of mafic composition, being dark and having a high density), with moderate to high vesicle content (VESPERMANN and SCHMINCKE 2000). Scoria typically ranges from grey to black in colour although it is often oxidised to red. Under petrographic microscope it is made up of dark translucent glass. Less-gassy magmas generate lava fountains that may produce chilled, transparent reddish, weakly to non-vesicular scoriaceous pyroclasts (VESPERMANN and SCHMINCKE 2000). Vesicles in scoria are generally smooth-walled and bubble-like. An extremally vesicular and very low density form of scoria is called reticulate. Pumice the term used for highly vesicular, generally silicic pyroclasts (Plate II, 2). Vesicles can be symmetric to extremely elongated, often forming a "woody" pumice texture with tube vesicles, occasionally associated with subaqueous eruptions (KATO 1987). Pumice of intermediate compositions is commonly banded with layers of variable colours (Plate II, 3). This may either represent mingled magmas of different compositions or may develop via mingling of parts of a single magma body or in a single conduit by locally variable cooling and degassing histories (DONOGHUE et al. 1995, WADA 1995, DERUELLE et al. 1996, MANDEVILLE et al. 1996, PAULICK and FRANZ 1997, SCHMITT et al. 2001, PLATZ et al. 2007). Due to the low density of silicic magma and the high vesicle content of pumice, these pyroclasts can often float on water. Consequently, large pumice rafts may blanket the ocean surface after subaqueous pumiceous eruptions and slowly travel over hundreds of kilometres via ocean currents to be deposited in completely foreign environments (BRYAN 1968, 1971, 1972, RAPP et al. 1973, CAREY et al. 2001, BRYAN et al. 2004). After long periods floating on water, vesicles may fill with water and the pumice will sink (MANVILLE et al. 1998, 2002). This physical difference between floating pumice and scoria (that rapidly sinks) makes these pyroclasts behave very differently during transportation and deposition.

During phreatomagmatic eruptions, magma–water interaction is the cause of the magma fragmentation (MORRISSEY et al. 2000). Explosion occurs when superheated water flashes to the vapour phase, thus rapidly expanding and generating shock waves through the magma. These shock waves force water deeper into cracks in the magma, creating an effective chain-reaction that is highly effective in fragmenting the melt body (ZIMANOWSKI et al. 1986, ZIMANOWSKI 1998). Phreatomagatic explosive fragmentation is very common, due to the almost ubiquitous presence of ground and/or surface water met by rising magma and extruding lava (LORENZ 1985). Actual interaction s usually between magma and water-saturated sediments (WHITE 1996a). Therefore, the physical process causing fragmentation is better known as (molten) fuel-coolant interaction (M)FCI, where the fuel is the magma and the coolant is any type of significantly cooler media (e.g., muddy slurry to water). Superheated liquid (e.g. the water that comes in contact with the magma) is in a metastable thermodynamic state, resulting from extremely rapid heating to a temperature well above the boiling point (WOHLETZ and MCQUEEN 1984a, WOHLETZ 1986). In this way vapour film can form on the magma/water interface and interfacial fluid instabilities are developed by the relative motion of the two immiscible fluids (WOHLETZ and MCQUEEN 1984a, WOHLETZ 1986, BUTTNER et al. 2002). There are two basic types of physical instabilities that may operate between the magma and water (or water saturated sediments); (1) the Kelvin-Helmholtz instability that is induced by shear stress along the interface, and (2) the Rayleigh-Taylor instability that is induced by the density contrast between two independently moving fluids (WOHLETZ and MCQUEEN 1984a, WOHLETZ 1986). In both cases fragmentation occurs when surface tension forces are exceeded. When water comes into contact with magma, it will either transform to steam (vapor) or a two-phase fluid depending of the relative masses of water and magma interacting. The style of activity ranges from passive quenching and granulation of melt to large scale thermohydraulic explosion. The phreatomagmatic fragmentation can be demonstrated by 4 major stages that occur in a sub-second time frame (WOHLETZ 1986): Stage 1: initial contact and coarse mixing of fuel and coolant under stable vapor film boiling (Leidenfrost effect); Stage 2: complete vapor film collapse; Stage 3: episodic increase of heat transfer from fuel to coolant and fine fragmentation leading to superheated and pressurized water. As the coolant is heated, it expands, leading to rapid increase in load stress on the melt. Relaxation of load stress in the brittle mode causes explosive release of seismic energy; Stage 4: volumetric expansion of the fuel-coolant mixture from the transformation of the superheated water to superheated steam. In general it is considered that all (M)FCI, explosive and non-explosive, are initiated by stage 1 (WOHLETZ 1986). Non-explosive (M)FCI terminates in stage 1 or 2, such as in the example of pepperite formation (see later) (WOHLETZ 2002). It is also generally agreed that the wetness of an explosive (M)FCI reflects the degree of fragmentation and the rate of heat transfer during stage 2 and 3. Characteristic clast populations produced by (M)FCI experiments are: blocky/equant; moss-like/convolute; spheres/drop-like; plate-like (WOHLETZ 1986, BUTTNER et al. 1999, 2002). Systematic clast shape analyses of real phreatomagmatic tephra (Plate II, 4) confirmed the similarity between natural and experimental pyroclast shapes (BUTTNER et al. 1999, DELLINO et al. 2001, BUTTNER et al. 2002, DELLINO and LIOTINO 2002). Controlling parameters of phreatomagmatic fragmentation are magma viscosity, temperature/pressure and the water/magma contact mode (supply rate of magma and external water, which is determined by the hydrology of and around the vent and conduit). In general the following parameters may facilitate efficient fragmentation;

— higher temperature melts have greater thermal energy available for conversion to mechanical energy;

— high viscosity retards the mixing of magma and water so that fragmentation is more effective in lower viscosity magmas;

— the amount of water relative to that of magma involved in fragmentation determines the intermediate to final thermodynamic states of water for phreatomagmatic eruptions;

— magma and water temperatures prior to interaction influence the heat-transfer rates and equilibrum temperature approached during mixing (higher the equilibrum T, the more E is available for mechanical work) (WOHLETZ 1986).

In summary, if very little water is available, it can be heated to high temperatures and pressures, but fragments only a relatively small volume of magma (SHERIDAN and WOHLETZ 1983). In cases of abundant water involved with the magma, the system may never get enough thermal energy for fragmentation. As consequence, in intermediate mass ratios, conditions approach an optimal compromise, thus resulting in the most efficient or complete magma fragmentation (SHERIDAN and WOHLETZ 1983, WOHLETZ and HEIKEN 1992). The graphic representation of this relationship is shown (Figure 3.3). In general the highest efficiency of fragmentation can be achieved if the water to magma ratio is 3 to 10 (WOHLETZ and HEIKEN 1992). However recent studies and theoretical considerations demonstrated that the water to magma ratio should be viewed as magma to coolant (Figure 3.4), e.g. sediment laden water (muddy slurry) in real world (WHITE 1996a). The implication of this view is that the syn-eruptive conditions of the near-

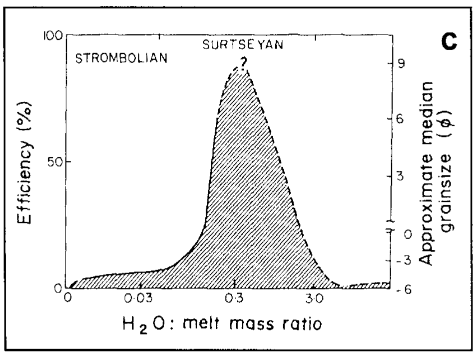


Figure 3.3. Diagram of the relationship between magma-water mass ratios and the fragmentation efficiency after WOHLETZ and SHERIDAN 1983: pp. 385–413

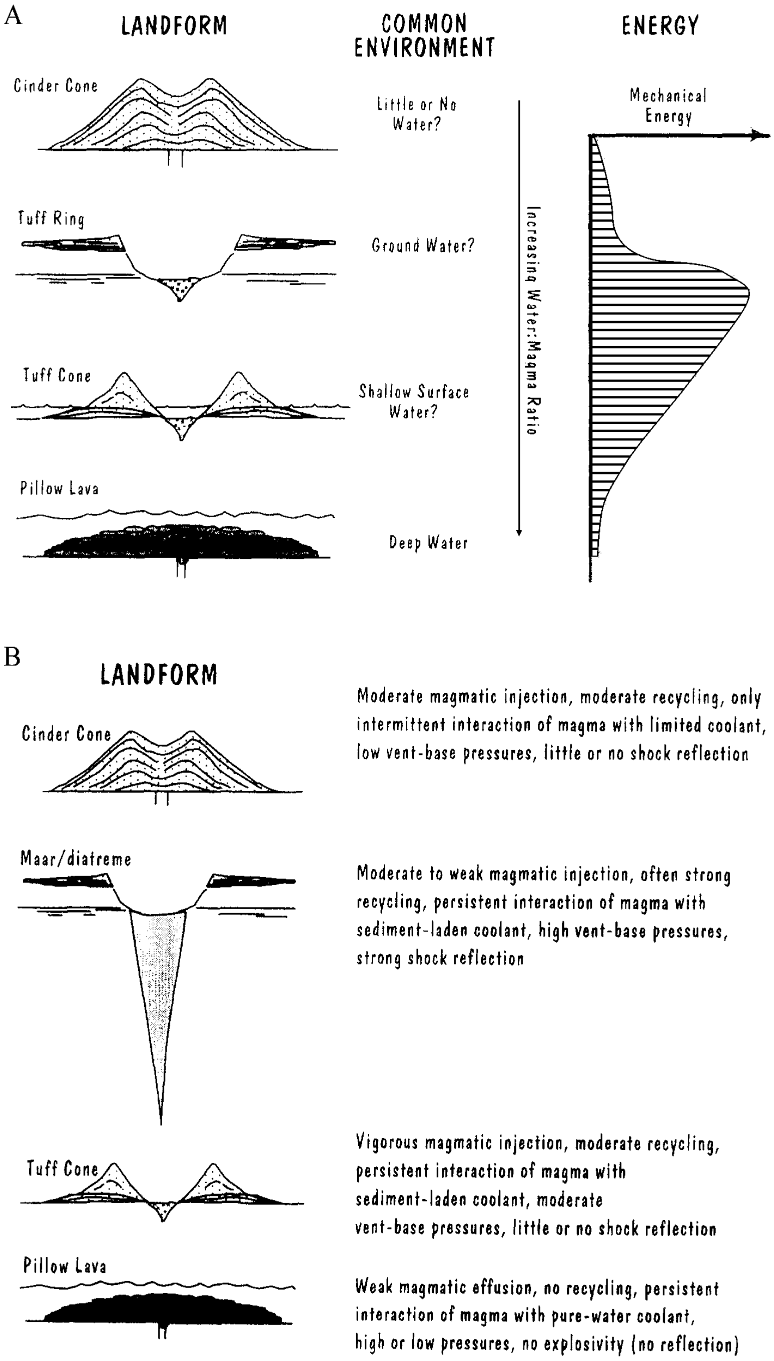


Figure 3.4. Comparative diagram between magma to water ratio mechanical energy and resulting volcanic landforms using pure coolant (water) and impure coolant (muddy slurry) as an interactive media with magma. After WHITE 1996: p. 158, fig. 1 and p. 166, fig. 8

surface plumbing system probably plays a more important role in fragmentation than the pure magma to water ratio (e.g. during the eruption, the water saturation of the vent filling pyroclastic slurry and the vent opening state) (White 1996a).

Micro-textures in relation to magma fragmentation, vesiculation and depth

When magma becomes slightly supersaturated with volatiles, nucleation of bubbles will occur (Sparks 1978). Growth of a fluid bubble is controlled by 1) the diffusion of volatiles dissolved in the magma into the bubbles, and 2) by the rate at which the confining pressure falls as the bubble or the magma, or both, rise (Sparks 1978). Bubble growth rate due to diffusion is controlled by the composition, solubility, concentration, and the degree of supersaturation of the volatiles (Sparks 1978). Bubble growth rate due to decompression (near the vent) is controlled by the rise velocity of magma, the rate at which the magma is disrupted and removed at the free surface in the vent, and by the rise of the bubbles within the magma body (Sparks 1978). Bubbles perhaps cannot grow infinitely. The volatile exsolution cause rapid increase in the viscous resistance to growth the bubbles (McBirney 1973, Sparks 1978, Wilson et al. 1980). Bubbles will not burst because there is no significant pressure gradient across the closely placed bubble walls (McBirney 1973, Sparks 1978). In this condition volatiles continue to diffuse into the bubbles until equilibrium is reached between the fluid pressure in the bubbles and the vapour pressure of the volatile still dissolved in the magma (Sparks 1978). It is generally accepted that magma start to fragment at its free surface in the vent where high pressure gradient exist between the vesiculating magma and the atmosphere (Sparks 1978). Magmatic fragmentation according to Sparks (1978) can be modelled in four steps; 1) at early stages nucleation occurs and bubbles grow freely, 2) bubble growth continue, but larger bubbles start to interfere with newly formed ones, 3) a magma froth form in the top of the magmatic column in the conduit, and further bubble growth may slows down or completely stops, 4) in top of the magma froth rapid fragmentation starts, and fragmentation front migrate downward causing bubble bursts (Sparks 1978).

Magmatic fragmentation in many cases is triggered by magma mixing caused by newly intruded hot (commonly more mafic) magma to a shallow subsurface chamber (Sparks et al. 1977, Woerner and Wright 1984, Koyaguchi and Blake 1989). Magma mixing can trigger magmatic fragmentation because 1) the newly intruded hot melt add volume to the magma chamber, and increase the fluid pressure in the chamber, 2) hot mafic magma can trigger rapid convective uprise in the magma chamber, 3) hot mafic magma usually rich in volatiles, which exsolve during uprise and are transferred by convection, diffusion and mixing into the low volatile acidic magma, leading to fluid pressure build up and explosive fragmentation, 4) the contact between hot mafic and colder acidic magmas can cause rapid crystallisation of the mafic magma, and therefore residual basic fluid can build up the total magma chamber fluid pressure.

In general magmatic fragmentation, pyroclasts are highly vesicular, and in mafic composition non-transparent under microscope (Plate II, 5). Silicic ash from subaerial explosive eruptions can range from fine ash to large pumice clasts. Large pumice clasts can be abraded, and rounded, however, woody shape blocky pumices also common, especially in association with intermediate (andesitic to dacitic) medium volume composite volcanoes in arc settings. Rhyolitic ash is commonly platy shaped as a result of fragmentation of gentle thin walled pumice ash. Cuspate glass shards are Y-shaped fragments commonly associated with blade-like glass shards (Figure 3.5).

In phreatomagmatic tephras, transparent volcanic glass shards, commonly referred as sideromelane are dominant (Plate II, 6). This type of glass is commonly blocky, angular, and moderately or non-vesicular (Plate II, 6). The glass shards size range from fine ash to fine lapilli. Especially in phreatomagmatic tephras, the accidental lithic fragment ratio is to the total volume of the sample is characteristic of the level of fragmentation; e.g. how deep the explosion took place. During shallow-level phreatomagmatic fragmentation and/or relatively open conduit conditions, the tephra will be dominated by blocky, non-to-moderately vesicular volcanic glass shards (Plate III, 1). If the fragmentation took place well below the surface, the shock waves generated by the magma–water interaction fragments the surrounding regions of the explosion chamber and excavate various amounts of country rock fragments (Plate III, 2), that can be related to the local stratigraphy and hence estimate depth (Németh 2003). In similar way the vesicularity study of the resulting pyroclasts will bear information to the volatile content, state of exsolution and pressurisation of the magma prior to fragmentation (Houghton and Wilson 1989). If the degree of vesiculation was more or less uniform throughout the magma body the pyroclasts of a single pyroclastic deposit bed will be similar. By contrast, if the pyroclast population is very diverse, it indicates very changeable conditions during the vesiculation, including variable shearing conditions, changes in rise rates, levels of crystalisation and degree of gas saturation in the melt (Houghton and Schmincke 1989, Houghton et al. 1996, 1999).

The texture of the pyroclasts especially their state of crystalinity give insights into the cooling history of the melt prior to fragmentation (which is often complex, with periods of rapid and slower cooling associated with variable pressure

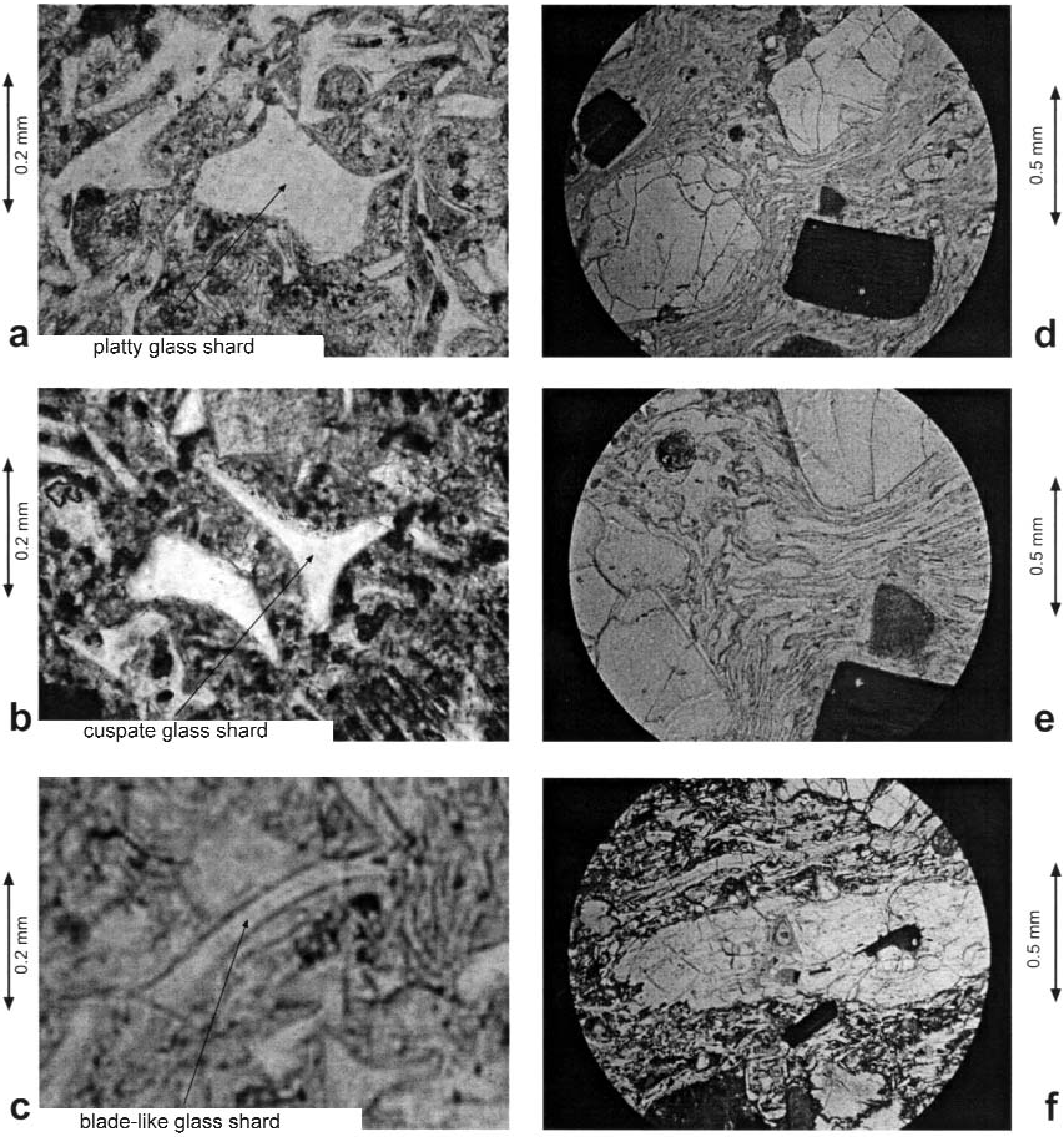


Figure 3.5. Silicic glass shards shards from pumiceous pyroclastic flow deposits. The view is about 4 mm across

conditions). During simple rapid cooling of deep hot melts, the melt chills quickly and the pyroclasts will contain small microlites (showing initial crystal growth), or small microphenocrysts, along with a variable cargo of phenocrysts formed in sub-crustal conditions. Such a scenario of final rapid cooling is expected during magma–water interaction in phreatomagmatic pyroclasts. During slow cooling the fragmented melt droplets will be charged with increasing densities of small microcrystals. In this way, especially in mafic to intermediate melts, when already fragmented melt droplets undergo further magmatic volatile-driven fragmentation, and the pyroclasts travel though air, the pyroclasts will be charged in magnetite and will form, dark, tachylite glass (Plate III, 3). Dark tachylite and light sideromelane pyroclasts may accumulate in repeating alternating layers in the same pyroclastic deposit, which indicates alternating "dry" and "wet" fragmentation (Martin 2002). This can result from incomplete wetting of the entire erupting surface of the magma body, or periods when water is driven off parts of the magma (Houghton and Smith 1993, Houghton et al. 1999). Hence the ratio of tachylite to sideromelane may provide insights into the nature of fragmentation conditions operating at different phases of an eruption.

Characteristics of pyroclastic deposits formed by phreatomagmatic fragmentation

In a single rock fragment collected during field mapping, its shape, vesicularity, and any quenching features (glass versus microlite content) may establish a phreatomagmatic fragmentation history of the melt (HEIKEN and WOHLETZ 1986), although outcrop-scale studies are generally more conclusive. During phreatomagmatic interaction energetic shockwaves often propel pyroclasts away from their source. Such transporting agents commonly form relatively low particle concentration, horizontal moving pyroclastic density currents such as base surges (MOORE 1967, FISHER and WATERS 1970, WATERS and FISHER 1971). Base surge bedforms are relatively easy to recognize in outcrop (Plate III, 4) (FISHER and WATERS 1970), being finely cross and/or dune bedded, and poorly sorted (FISHER and WATERS 1970). They contain a high component of ash-rich matrix-and in consolidated state form lapilli tuff and tuff beds (Plate III, 4). Dune beds are also characteristic (Plate III, 5). Bedforms often indicate deposition from high-energy flows, including structures such as antidunes, with beds of coarse ash and lapilli accumulated in steep bedding surfaces facing their source (SCHMINCKE et al. 1973). Base surge deposits in near-vent positions often show scour and fill contacts to the basal surface. In deep-seated explosions the resulting pyroclastic deposits are charged with accidental lithic fragments disrupted from the country rock column (Plate IV, 1). Shock wave fragmentation causes highly angular clasts over a wide size range, depending also on the strength and jointing properties of the rock. In "soft" substrates, such as fluvial or marine sands and muds the resulting pyroclastic deposits often contain these fragments in the finest particle size range (Plate IV, 2). In extreme situations, such deposits can contain over 90 vol% accidental components in all size ranges (SOHN and PARK 2005, AUER et al. 2006). Such deposits are common in the western Hungarian Mio/Pliocene phreatomagmatic volcanic field, and it may often be a challenge to distinguish them from other siliciclastic sediments.

Accretionary lapilli

Accretionary lapilli (pisolith, chalazoidite) are spherically shaped, lapilli-sized particles made up of aggregated fine ash (SCHUMACHER and SCHMINCKE 1991, 1995, GILBERT and LANE 1994) (Plate IV, 3). Empirical studies of many accretionary lapilli identified 2 major types (SCHUMACHER and SCHMINCKE 1991, 1995); (1) rim-type (Plate IV, 4), that are cored by a coarser ash, lapilli or aggregates of variable sized clasts and rimmed by a fine ash layer and (2) core-type (Plate IV, 5) that have no rims, and usually composed of coarser grained ash. Related particles include armoured or cored lapilli (LORENZ and ZIMANOWSKI 1984), in which lapilli of any type are covered by a homogeneous fine-grained rim that can sometimes reach cm in thickness (Plate IV, 6). Accretionary features such as mud clots or armoured mud balls are also common in fine ash deposits from phreatomagmatic eruptions (Plate V, 1).

The formation of accretionary lapilli results from aggregation of fine ash particles in moisture-rich eruption clouds, and thus they can also be entrained in pyroclastic density currents (ROSI 1992, GILBERT and LANE 1994, SCHUMACHER and SCHMINCKE 1995). Very fine ash particles aggregate due to electrostatic attraction (JAMES et al. 2003). After initial aggregation moisture of the eruption cloud helps to form repeated film-like layers over the aggregates. Accretionary lapilli are commonly used to indicate magma–water interaction (WOHLETZ and MCQUEEN 1984b). However, they can also form in any type of fine particle charged eruption cloud that may travel through a moist region in the atmosphere and/or flushed by rain during the eruption (CAREY and SIGURDSSON 1982, VEITCH and WOODS 2001, SCOLAMACCHIA et al. 2005, TEXTOR et al. 2006). In this latter case, accretionary lapilli beds are often discontinuous, and the accretionary lapilli form clusters in an otherwise pyroclast fall-like bed. They are also commonly associated with Plinian-style eruptions, where the eruption cloud is large enough to be able to get contact with clouds (CAREY and SIGURDSSON 1982), particularly in tropical climates where large rain clouds quickly form due to high humidity. Accretionary lapilli are often ubiquitous in single pyroclastic beds associated with phreatomagmatic eruptions of tuff rings, cones or maars. Large examples occur from eruption columns forming through caldera lakes (Plate V, 2). In such cases, accretionary lapilli beds can be thick (dm-scale) and traceable over large distances (hundreds of kilometres). In another situation, accretionary lapilli are also commonly associated with clastic dykes or segregation pipes (BOULTER 1986). Accretionary lapilli bearing beds show a characteristic distributional pattern around small volume phreatomagmatic mafic volcanoes. The eruption cloud is usually c. 100 °C in near vent position, but this rapidly drops as the cloud moves outward, meaning water droplets condense from the cloud. These cause agglutination of ash around 500–700 metres from the source, and hence there is often a sudden appearance of accretionary lapilli beds at this point. Fine mud clots and aggregates may also form by the simple disruption of entire mud chunks from the vent (Plate V, 3). This process could take place when Plinian eruptions are initiated through a caldera lake. Such fragments can travel together as mud clots over large distances, and mimic accretionary textures. These particles are also common during eruption of mafic small volume volcanoes in fluvio-lacustrine basins (WHITE 1996b, 2001). For this reason such mud clots are common in association with phreatomagmatic successions in the western Hungarian Mio/Pliocene phreatomagmatic volcanic fields (MARTIN and NÉMETH 2004).

Reworking of accretionary lapilli is not well understood. In general, identification of continuous accretionary lapilli beds is used to interpret the bed as primary and deposited subaerially. However, discontinuous and abraded accretionary lapilli are known from reworked volcaniclastic successions associated with large volume plinian eruptions and in subaqueous settings (SELF and SPARKS 1979, BOULTER 1987, JONES and ANHAEUSSER 1993). Despite these examples, accretionary lapilli hardly survive long distance travel, unless the motion is gentle enough and subsequent deposition rapid enough to preserve them against mechanical shear.

Volcanic glass

Volcanic glass reflects rapid chilling of magma and can be thought of as a super-cooled fluid. Glass fragments therefore can be more or less directly related to the original composition of the melt fragmented by any processes. The abundance of volcanic glass shards, especially in mafic pyroclastic deposits, is a good indication of the presence or absence of magma–water interaction. Glass shard morphology is commonly used to indicate fragmentation and deposition history of pyroclasts (MCPHIE et al. 1993). In magmatic explosive fragmentation, especially in silicic magma, three types of glass shards have been documented (MCPHIE et al. 1993); (1) cuspate, that are X- or Y-shaped and represent junctions between vesicles of larger glassy fragments; (2) platy shards, that are curviplanar in shape and usually low-vesicularity; and (3) and highly vesiclular pumiceous shards. These three major types of glass shards can accumulate together and their relative ratio can provide constraints on the conditions during fragmentation, transportation and deposition. The texture of pyroclastic (e.g. ash) form during magmatic fragmentation dependent upon magma composition, temperature, and volatile content (HEIKEN and WOHLETZ 1986, 1991, CASHMAN and BERGANTZ 1991, CASHMAN and MANGAN 1994b, CASHMAN et al. 2000). These factors perhaps control the viscosity and surface tension, which are responsible for the shape the pyroclast may get.

In Hawaiian-style eruptions, during lava fountaning of low-viscosity magma, small, smooth-surfaced glass droplets (e.g. spheres, teardrop-shapes, dumbbells, ovoids), long glass threads (Pele's hair) or irregular shape clots of scoria and glassy pyroclast can form (HEIKEN and WOHLETZ 1986, MANGAN et al. 1993, MANGAN and CASHMAN 1996). The sphere-like glassy pyroclasts are smooth surfaced, having rounded to elongate vesicles and thin skins commonly fractured due to clast collision. Pele's hair are commonly long (few cm), thin hair-like glassy pyroclasts. They are commonly connected to a droplet-like head to where a hair-like tail attached.

During Strombolian-style eruptions low-viscosity magma produce tephra rich in glassy ash consists of pyroclasts from irregular shaped sideromelane droplets to blocky tachylite. The gradation from clear sideromelane droplets into microcrystalline tachylite grains commonly reflects the degree of chilling and crystallization of the melt droplets (HEIKEN and WOHLETZ 1986).

Texture of glassy pyroclasts from more viscous magmas (e.g. andesite to rhyolite) is primarily controlled by their higher viscosities and the higher volatile contents of the source melt. In higher viscosity melt droplets cannot form and therefore the shape of the glassy pyroclasts will be controlled by the vesicle shape and content of the fragments break apart during fragmentation (HEIKEN and WOHLETZ 1986). Fragmentation of magma during a plinian eruption based on assumptions about the timing and mechanisms of fragmentation as key parameters in all existing eruption models. Most models assume that fragmentation occurs at a critical vesicularity (volume percent vesicles) of 75-83% of the melt (HEIKEN and WOHLETZ 1986). However recent evidences indicate that the degree to which magma is fragmented is determined by factors controlling bubble coalescence such as magma viscosity, temperature, bubble size distribution, bubble shapes, and time (KLUG and CASHMAN 1996, KLUG et al. 2002). Bubble coalescence in vesiculating magmas creates permeability which serves to connect the dispersed gas phase. When sufficiently developed, permeability allows subsequent exsolved and expanded gas to escape, thus preserving a sufficiently interconnected region of vesicular magma as a pumice clast, rather than fully fragmenting it to ash. For this reason pumice is likely to preserve information about (a) how permeability develops and (b) the critical permeability needed to insure clast preservation. Both the development of permeability by bubble wall thinning and rupture and the loss of gas through a permeable network of bubbles require time, consistent with the observation that degree of fragmentation (i.e., amount of ash) increases with increasing eruption rate (KLUG and CASHMAN 1996, KLUG et al. 2002). Pumice fragments commonly contain various amounts of microlites. The presence of microlites not only can increase the magma viscosity and effective vesicularity, but appears to be able to aid bubble nucleation and therefore hinder bubble expansion and coalescence (KLUG and CASHMAN 1994). Thus, magmas with microlites may fragment at lower bulk vesicularity than those without microlites. Fragmented microlite-bearing clasts are also likely to expand less after fragmentation and therefore more closely preserve the bubble distribution and structure at the time of magma fragmentation (KLUG and CASHMAN 1994).

As it has been pointed out earlier, phreatomagmatic volcanic glass shards are blocky, non-to-weakly vesicular, and usually bear only a minor content of microlites. With unstable vent and conduit conditions, rapid alterations of phreatomagmatic and magmatic fragmentation can occur and the resulting tephra beds will contain glass shards recording this. Silicic glass shards are more diverse, and due to their lighter colour; hence establishing quench fragmentation (e.g. magma–water interaction) on the basis glass appearance is difficult. In rare cases, large silicic glass fragments can be obsidian clasts derived directly from chilled silicic lava (Plate V, 4). These clasts are black, brown, or red, and commonly have flow-banded textures. Obsidian fragments are common in deposits formed by silicic dome collapse where they could dominate autoclastic deposits (McPhie et al. 1993).

One of the most important glass-rich deposits of this type is hyaloclastite, which forms by the non-explosive quench fragmentation of subaqueous lava flows (Pichler 1965, Furnes and Fridleifsson 1974, Furnes and Sturt 1976, Honnorez and Kirst 1976, Batiza et al. 1984, Smith and Batiza 1989, McPhie et al. 1993, Schmincke et al. 1997, Scutter et al. 1998, DeRita et al. 2001, Martin 2002). There are also theories that infer deep submarine hyaloclastite formation associated with suppressed, low energy explosive fragmentation of lava (Maicher 1999) forming limu shells (Maicher et al. 2000, Maicher and White 2001). Hyaloclastite is rich in blocky, diversely shaped glass fragments that are weakly to non vesicular and usually angular in shape with textural evidence of fracturing (Plate V, 5). Hyaloclastite formation also can be associated with magma and ice interaction and is the major component of "table" mountains in these environments (Smellie et al. 1993, Skilling 1994, Werner et al. 1996, Gudmundsson et al. 1997, Werner and Schmincke 1999, Helgason and Duncan 2001, Stevenson et al. 2006).

Alteration and textural changes of volcanic glass

Because volcanic glass is metastable, it rapidly undergoes alteration upon deposition and at times already during transportation. Basaltic glass shards of phreatomagmatic eruptions often develop thin rims of palagonite during the transportation (Plate VI, 1), and undergo significant palagonitisation soon after deposition, due to low-temperature hydration and alteration (Plate VI, 2). Palagonite is a yellow to brown clay mineral that forms from water, iron, magnesium and alkalia in the glass (Peacock and Fuller 1928, Farrand and Singer 1992, Schiffman and Southard 1996, Techer et al. 2001, Stroncik and Schmincke 2002, Drief and Schiffman 2004). Its name derives from Palagonia, Sicily, where a thick succession of palagonitic pyroclastic deposits crop out in association with the shallow subaqueous to emergent volcanism of the Pliocene Hyblean Mountains (Schmincke et al. 1997). Palagonite is common in tuff cones, where the pyroclastic deposits are dominated by glassy lapilli and ash (Wohletz and Sheridan 1983, Verwoerd and Chevallier 1987, Farrand and Singer 1991, Sohn and Chough 1992, Sohn 1995, Martin 2002). Strong palagonitisation is also known from table mountains where large volume hyaloclastite formation is common (Skilling 1994). Palagonite can also transform quickly to smectites, ferric oxides, zeolites and clorites (Figure 3.6). At thin section level, such changes can be traced from individual clasts, can be bed specific (e.g. as a result groundwater level changes, or saturation), or patchy (Plate VI, 3). There is no direct link know between formation of palagonite and the position of the former water table (e.g. the position of the syn-eruptive water level). Hydrothermal processes and rheological differences between certain pyroclastic beds can cause development of efficient pathways for diverting warm hydrothermal fluids that may facilitate varying degrees of palagonitisation within certain places of a deposit stack.

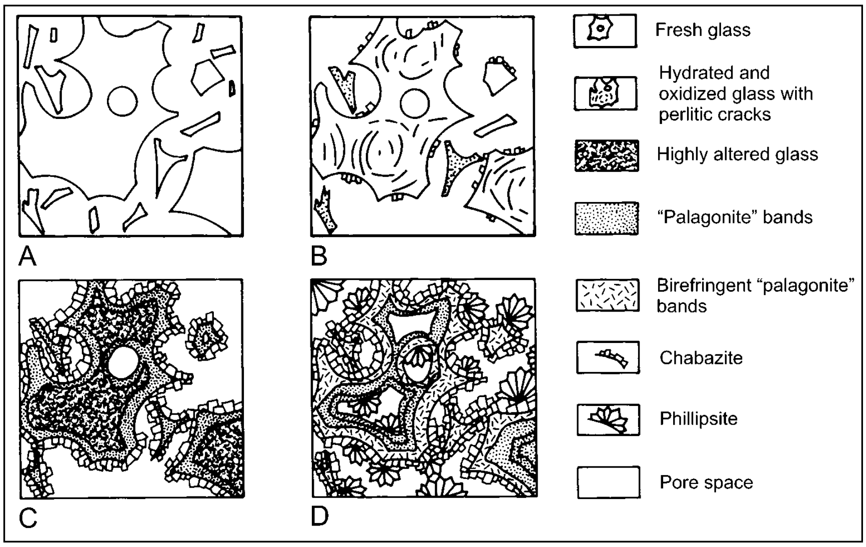


Figure 3.6. Step of palagonitisation of a volcanic glass shard after Fisher and Schmincke (1984)

High-silica glasses can also transform quickly due to devitrification (Lofgren 1971). During this process, the original glass is initially gradually replaced by minerals such as zeolites, phyllosilicate and palagonite (Mandeville 1970, Scott 1971, Stimac et al. 1996). Later stages involve the nucleation and ongoing growth of crystals at subsolidus temperature (e.g. straight after deposition of the pyroclasts). Spherulites (Figure 3.7) and lithophysae (Plate VI, 4) are the most common textures that form during relatively high temperature devitrification of volcanic glass (McPhie et al. 1993). Since the process is gradual, during intitial stages, patchy sphreulites appear, wheras later it becomes spherulate-dominated (Davis and McPhie 1996, Smith et al. 2001, Orth and McPhie 2003). The speed of the process and the advancement of spherulitic texture formation is facilitated by the maintaining elevated temperatures (e.g. welded tuffs that retain heat long enough), and the presence of high temperature alkalia rich fluids. The resulting spherulites are diverse in size, and can reach dm-wide patches of microlite-rich, fibrous, star-like zones in the glass. Lithophysae have a central void that can be later filled by other secondary mineral phases. Lithophysae formation is initiated by spherulite growth during as the melt is still hot and exsolving volatiles. During volatile exsolution, voids are inflated and with cooling of the escaping mineral solution, charged volatiles, quickly become the place for phenocryst (usually quartz) development. Devitrification commonly leads to a micropoicilitic texture, in which single mineral phases such as quartz, enclose small microlites of other mineral phases in a snowflake-like pattern. Spherulites of mm-to-cm in diameter are common in coherent rhyolitic lavas and dome collapse breccias from the Tokaj Mts in NE Hungary (Figure 3.8). Micropoikilitic texture is very common in coherent rhyolitic as well as welded rhyolitic pyroclastic rocks.

In silicic glass-bearing volcanic rocks, perlitic texture is also common (Plate VI, 5). Perlitic cracks are hair-like microcracks that cross-cut the glassy body (Ross and Smith 1955, Friedman et al. 1966). The perlitic cracks develop as a direct response to glass hydration (Friedman and Long 1976). During hydration, glass expands and being brittle it also cracks. Crack development can be random, or follow bands of an original banded texture (common in lava flows). Perlitic rhyolite is widespread in the Tokaj Mts in NE Hungary and associated with subaqueous lava dome and associated hyaloclastite sequences, where both, regular, and flow-banded perlite is common.

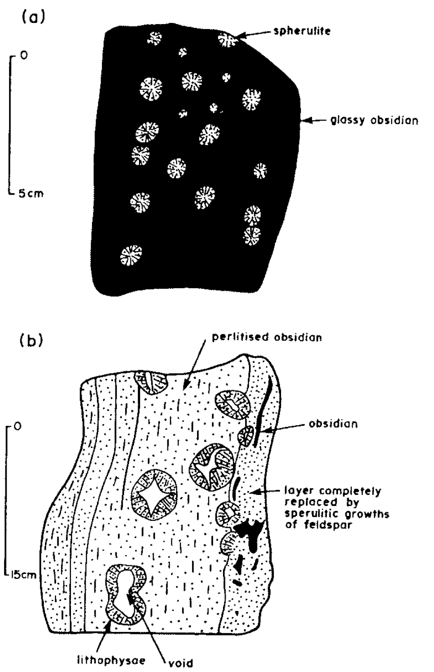


Figure 3.7. Textural development of obsidian lava flows from spherulite to lithophysae development according to Cas and Wright 1988

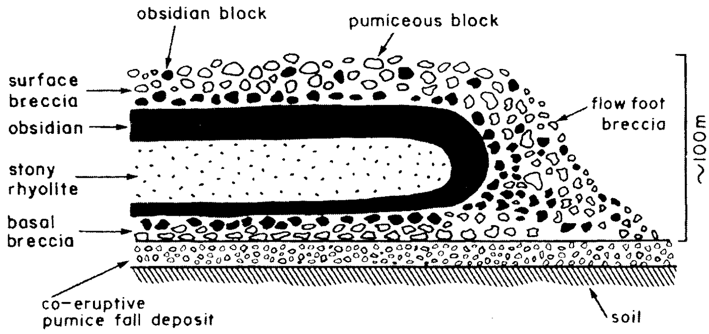


Figure 3.8. Theoretical rhyolite lava flow cross section according to Cas and Wright 1988

Transportation of pyroclasts and depositional styles

Transportation of pyroclasts are governed by the same physical rules regardless of the fragmentation process that may have formed them (McPhie et al. 1993). Outcrop-scale study of the volcanic rock record is necessary for identification of the transport and depositional environment. Particles can be carried more or less directly from their source to their depositional sites, such as during an eruption, or, the pyroclasts may stop and start motion through repeated reworking processes that take them to their final destination (McPhie et al. 1993). In the second case, the transport processes are common to any non-volcanic sedimentary environment.

The main processes involved include (McPhie et al. 1993); (1) mass flow where groups of any type of clast as well as the interstitial fluid is transported and moved together. The flow may vary greatly in particle concentration and rheology; (2) traction-dominated transport, where particles are entrained in an interstitial fluid of any type and they are able to move freely within the fluid; (3) suspension-dominated transport where the transported particles are fully suspended in any type of interstitial fluid. Another approach to distinguish transport mechanisms is to estimate particle concentration, particle trajectory (vertical versus horizontal), cohesion (fines/clay content), and flow/current stability (Figure 3.9) (Wilson and Houghton 2000). In this classification scheme pyroclastic falls would be characterised by low particle concentrations and particles with a vertical trajectory. Pyroclastic surges also have low particle concentrations but with particles travelling in a horizontal trajectory (BURGISSER and BERGANTZ 2002). Pyroclastic flows are transported horizontally but have high particle concentrations (BURGISSER and BERGANTZ 2002). Differences between flow and surge transportation and deposition can also be demonstrated by the particle (density) concentration profile of the currents (Figure 3.10) (WILSON and HOUGHTON 2000). Pyroclastic flows have concentrated flow bases with an abrupt concentration drop above the flow base (Figure 3.10). Pyroclastic surges have lower particle concentrations in the current, but a more gradual concentration drop upward in the current (Figure 3.10). Particle cohesion is responsible for controlling the particle accumulation style, e.g. with low cohesion, particles slump easily forming grain avalanches, whereas with high cohesion

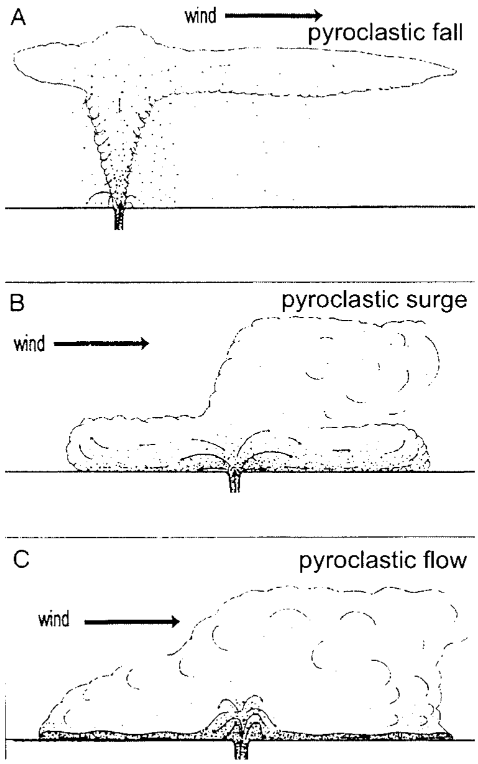


Figure 3.9. Pyroclastic fall, flow and surge formation and the associated eruption clouds according to WILSON and HOUGHTON 2000: p. 547, Figure 1

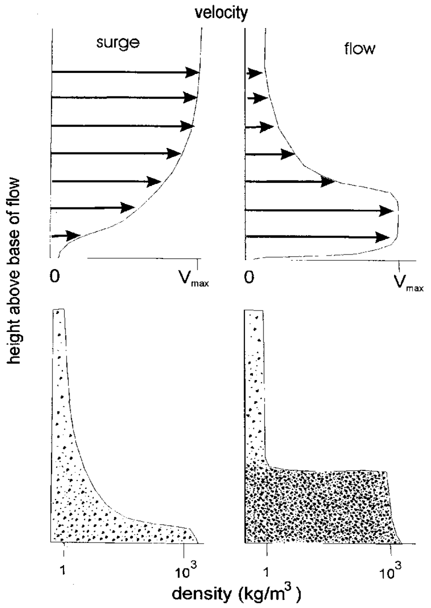


Figure 3.10. Density profile of dilute (surge) and concentrated (flow) pyroclastic density currents according to WILSON and HOUGHTON 2000: p. 548, fig. 2

steep bedforms are created through particles adhering to one another (WILSON and HOUGHTON 2000). Sustained pyroclastic flow eruptions may accumulate thick beds that show gradual particle concentration changes. If their characterised by non sustained pulses of individual phases, the resulting pyroclastic succession will have well developed bedding with sharp contacts (WILSON and HOUGHTON 2000).

Mass flow deposits can be both related directly to volcanic eruptions and resedimentation of them. Volcaniclastic deposits can be emplaced by a range of mass flow transportation including: turbidity currents, debris flows, mud flows, grain flows, density modified grain flows, rock falls or debris avalanches.

Traction-dominated transportation takes place during pyroclastic surges. The same type of transporting currents occur in fluvial or subaqueous systems. Suspension-related primary pyroclastic deposits include pyroclastic falls that can be air or water settled, whereas the reworked equivalent can be any type of suspension deposition such as hemipelagic deposition.

Bedding characteristics from horizontal transport

Horizontal transport of pyroclasts can be identified from outcrop-scale observations of volcaniclastic successions. The full diversity of bedding characteristics viewed in the field are not always indicative of the transport agent. An example is the current lack of understanding of the link between physical characteristics of deposit and flow in the pyroclastic flow system. (Figure 3.11) (BRANNEY and KOKELAAR 1992a, DRUITT 1998). Two major ideas are competing; (1) progressive aggradation by sedimentation from the base of an active flow over its entire length in a similar way to high density turbidity currents (FISHER 1966, BRANNEY and KOKELAAR 1992b, KNELLER and BRANNEY 1995); or (2) "en masse freezing" of the entire flow at once or of its margins (WRIGHT and WALKER 1981). Much field evidence and many experiments have been interpreted to suggest the former mechanism (KOKELAAR and BRANNEY 1996, SUMNER and BRANNEY 2002, BROWN and BRANNEY 2004a, 2004b, CARRASCO-NUNEZ and BRANNEY 2005). However, there are still not complete acceptance of this progressive aggradation hypothesis (BRANNEY and KOKELAAR 1992a, 1994).

Pyroclastic density currents are multi-phase flows where volume, mass flux, grain-size, particle concentration and bulk density can vary over several orders of magnitude (DRUITT 1998, BRANNEY and KOKELAAR 2002). Physical descriptions of the state of flow range from dense, granular flows, where gas plays a subsidiary role (CALDER et al. 2000) and motion is dominated by particle interactions, through gas-fluidised flows, where gas plays a significant role (SPARKS 1976, SPARKS et al. 1978, WILSON 1980, 1984, WILSON and HOUGHTON 2000, ROCHE et al. 2004, 2005), to highly diluted, turbulent systems where gas is the dominant phase and transports particles in turbulent suspensions (DRUITT 1998, FREUNDT and BURSIK 1998, HUPPERT 1998, BRANNEY and KOKELAAR 2002).

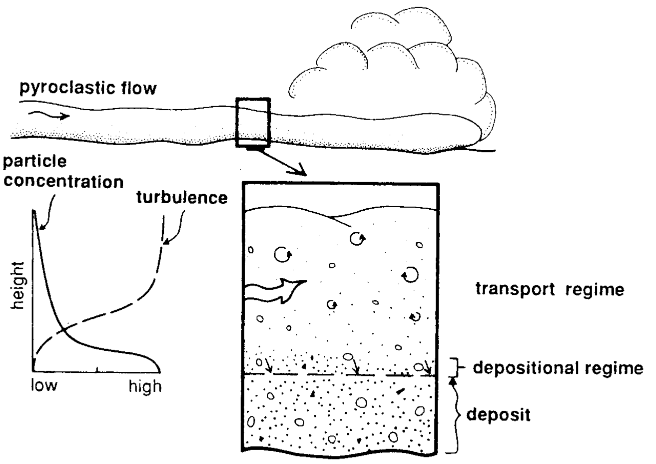


Figure 3.11. Progressive aggradation model for pyroclastic flow evolution according to BRANNEY and KOKELAAR (1992) and DRUITT (1992) ideas

Field textures of pyroclastic flow deposits have been used to interpret variations of the parent flow steadiness (FREUNDT and SCHMINCKE 1985, 1986, 1992), particle concentration gradients (e.g. stratified flow theory), and vertical or horizontal variations in the flow regime (VALENTINE 1987, PALLADINO and VALENTINE 1995, BAER et al. 1997, FREUNDT 1998, VALENTINE 1998, FREUNDT 1999). The most frequently occurring type of pyroclastic flows are those represent the low-energy, dense, granular flow end-member of pyroclastic density currents such as block-and-ash flows (SCHWARZKOPF et al. 2005), typically generated by gravitational collapses from active lava-domes (SAUCEDO et al. 2002, 2004), Vulcanian-style eruptions (NAIRN and SELF 1978, LUBE et al. 2007) or unstable agglutinates and lava autobreccias (RODRIGUEZ-ELIZARRARAS et al. 1991).

Pyroclastic surges including base surges also have horizontal transport regimes (VALENTINE and FISHER 2000, WHITE and HOUGHTON 2000). A base surge is a turbulent pyroclastic density current that radiates from the site of a phreatomagmatic eruption centre (MOORE 1967, FISHER and WATERS 1970, SWANSON and CHRISTIANSEN 1973, VALENTINE and FISHER 2000, NARANJO and HALLER 2002). The resulting beds contain features such as ripples, dunes, antidunes or tabular forms. These deposits commonly build up the majority of small-volume phreatomagmatic volcanoes (SCHMINCKE et al. 1973, WOHLETZ and SHERIDAN 1983, SOHN 1996, STOPPA 1996, VALENTINE and FISHER 2000, VESPERMANN and SCHMINCKE 2000). Identification of base surge deposits can be difficult since they may mimic textures of fluvial deposits (BULL and CAS 2000). Each base surge deposited pyroclastic succession has a typical set of beds that are distinguishable by their colour, structure, texture, componentry and their relative position to each other (VAZQUEZ and ORT 2006). Base surges can be "wet" or "dry" in respect to their free water content (DELLINO et al. 1990, CAPACCIONI and CONIGLIO 1995, ALLEN et al. 1996). Higher moisture contents in the eruption cloud controls the cohesion of the particles during the transportation and deposition. Wet surges are also generally low temperature, hence allowing free water and three phase flow. Dry surges are higher temperature and commonly viewed as gas-supported two phase systems.

Pyroclastic surge deposits are generally identified by their commonly developed cross-bedded texture, including very finely bedded, low-angle cross beds, along with an overall undulating bed thicknesses. These characteristics are however variable, an extreme end member of the deposition spectrum for these flows can produce very thin planar beds that are difficult to distinguish from pyroclastic fall units. In this case, the poor sorting characteristics usually distinguish them from falls. Different types of cross-beds occur and are characteristic for a range in flow regimes within the flow (Figure 3.12). Ripples and small (low amplitude, long wave length) dunes are characteristic for "lower" flow regimes in distal locations or for low-energy pyroclastic surges (Figure 3.12). Preservation of individual bedforms requires rapid deposition, otherwise erosion, especially by high-energy "upper" flow regime currents destroys any earlier deposited bed sets. Transitions from lower to upper flow regimes during a succession of eruptions can result in continuously eroded bed surfaces at any one location that would create parallel-bedded assemblages (VALENTINE and FISHER 2000). Many authors have identified distinct facies variations among different type of bed sets, as a result of flow regime changes from the proximal to distal regions (WOHLETZ and SHERIDAN 1983). The most common facies variation from an initial high-concentration current toward a more dilute, and less energetic flow is inferred to create deposition facies changes from units with sandwave bedforms, through massive units and out to planar bedded associations (WOHLETZ and SHERIDAN 1979, 1983, LAJOIE et al. 1992). From Jeju Island, Korea, lateral facies transformations have also been identified (CHOUGH and SOHN 1990, 1996). In proximal areas, massive and disorganized beds occur, which grade laterally to sand-wave beds and eventually planar and low-amplitude sand-wave beds in the most distal sites (SOHN and CHOUGH 1989, CHOUGH and SOHN 1990). Proximal high-particle concentration and highly turbulent flow regimes are inferred to exist in near-vent positions. As the surges travel outward they become more dilute, suspended-load fall out decreases, giving way to tractional-transport processes. Vertical facies variations can also be identified in deposits of pyroclastic surges or flows, which are usually related to variable rates of energy release in an ongoing eruption (SOHN and CHOUGH 1989).

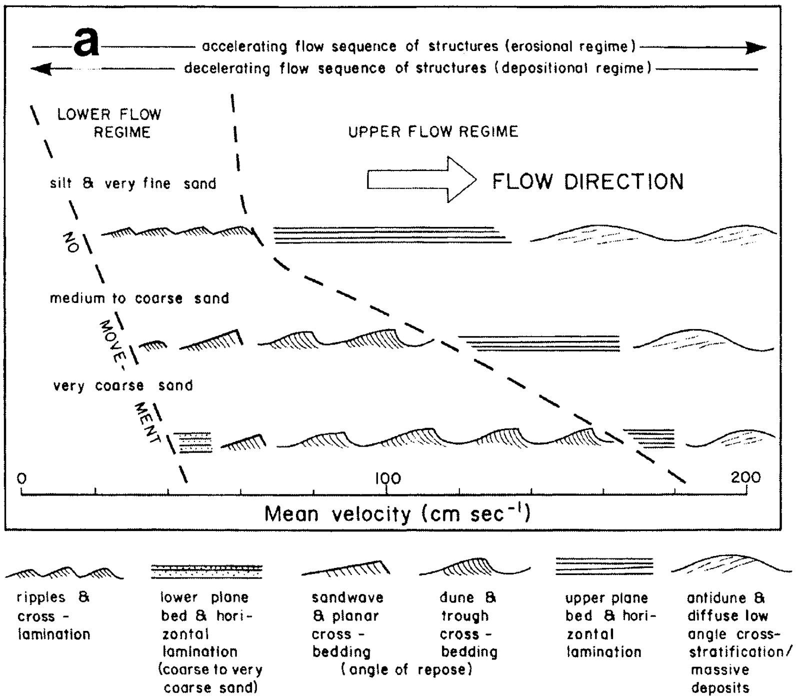


Figure 3.12. Relationship between flow regime and resulting bed forms in horizontal moving sedimentary currents (after CAS and WRIGHT 1988)

Facies variation during any horizontal transport mechanism can also be controlled by the topography, to produce highly contrasting zones of valley filling facies together with either lateral, overbank facies, or so-called ignimbrite veener facies (FISHER et al. 1983, BOGAARD and SCHMINCKE 1984, FREUNDT and SCHMINCKE 1985, NÉMETH and MARTIN 1999).

Pyroclastic surges are commonly associated with pyroclastic flows, either as a ground surges that are produced in front of the pyroclastic flow, as well as ash-cloud surges generated from elutriation of fines with hot gases rising above a flow. In small-scale pyroclastic flows, such as block-and-ash flows from domes, topographic barriers, and hydraulic jumps, may generated ash-cloud surges (EDGAR et al. 2002), in addition when pyroclastic flows enter sea water, secondary explosions may generate dilute surges (EDMONDS and HERD 2005).

Ballistic transportation of clasts

Ballistic clasts are directly propelled from a vent and follow a trajectory similar to cannon fire (hence "ballistic") (MCGETCHIN et al. 1972, CHOUET et al. 1973, 1974). This process is important during Vulcanian- and Strombolian-styles of eruption (YAMAGISHI and FEEBREY 1994, WOODS 1995). Ballistic bombs and blocks are commonly (but not exclusively) associated with phreatomagmatic volcanism, especially when the magma–water interaction takes place in the pre-volcanic country rock succession (Plate VII, 1). Ballistic blocks can travel far from their source (up to 10 km) in extreme conditions of plinian eruptions, however, they are commonly restricted to within 1-3 km of a vent (PFEIFFER 2001). Due to their typically high density they can be very locally destructive to buildings that happen to be located close to the source (Plate VII, 2) (ARTUNDUAGA and JIMENEZ 1997). During phreatic eruptions (Plate VII, 3), ballistic clasts can make up the bulk of erupted material (e.g. hydrothermal explosions), although their distribution is normally very restricted to within tens to hundreds of metres radius (MARINI et al. 1993). Ballistic bombs and blocks often cause impact craters on the immediate underlying bed surface. Mapping the orientation of the impact craters (impact sags) can help to determine the source location in geologic exposures where surface geomorphology is not available (BOGAARD and SCHMINCKE 1984). The depth and shape of the impact sags also indicates the degree of saturation and plasticity of the underlying pyroclastic beds (Plate VII, 4). Recognition of ballistic clasts in a volcaniclastic sequence indicates that the succession is very likely directly related to a volcanic eruption. However, impact sag geometry, in case they very symmetric, can also be confused with other sedimentary processes such as drop stones (Plate VII, 5). For correct interpretation, the 3D outcrop-scale facies analysis is necessary, along with compositional analysis.

Textural features characteristics of soft, unconsolidated sediments

Deformation of freshly deposited tephra beds are only common when the tephra is water saturated, such as when it is of phreatomagmatic origin. Their structure and formation is the same as in other clastic sedimentary environments (Moss and Howells 1996, Massari et al. 2001, Surlyk and Noe-Nygaard 2001, Jolly and Lonergan 2002). Many of them are induced by seismicity, (Mohindra and Bagati 1996, Mohindra and Thakur 1998, Kotlia and Rawat 2004) a common process during eruptions. High density ballistic bombs and blocks impacts can also plastically deform beds (Plate VIII, 1). Also, freshly deposited high density clasts can behave like drop stones, and sink into deeper positions in the bed once deposited. Water-loss of low density tephra can lead to develop hard ground layers with boudinage-like lateral thickness variation (Plate VIII, 2). Such water loss is commonly associated with accumulation of mineral-charged water rich zones, which may enhance alteration of specific beds of the succession (Rosi 1992). Water saturation and escape also can cause flame and dish structures, as well as small sand/mud volcanoes, especially when heavy stacks of tephra caps low-density and saturated sediment. This type of volcaniclastic sediments is common to many phreatomagmatic volcaniclastic successions in submarine basins (Plate VIII, 3) or intra-crater lacustrine successions (e.g. maar lake or caldera lake deposits) (Plate VIII, 5). Soft-sediment deformation structures are also common below high density pyroclastic flow deposits or volcanic debris avalanches, where long flame structures of fine mud can intrude into the overlying volcaniclastic succession. In subaqueous settings, small flame structures are associated with reworked hyaloclastite beds. Large clastic dykes are common in association with subaqueous cryptodome and lava dome complexes , where soft syn-eruptive mud can be squeezed over many tens of metres into the growing dome structure, similar to those described from NE Hungary (Plate VIII, 5) in the Tokaj Mts (Németh et al. 2005).

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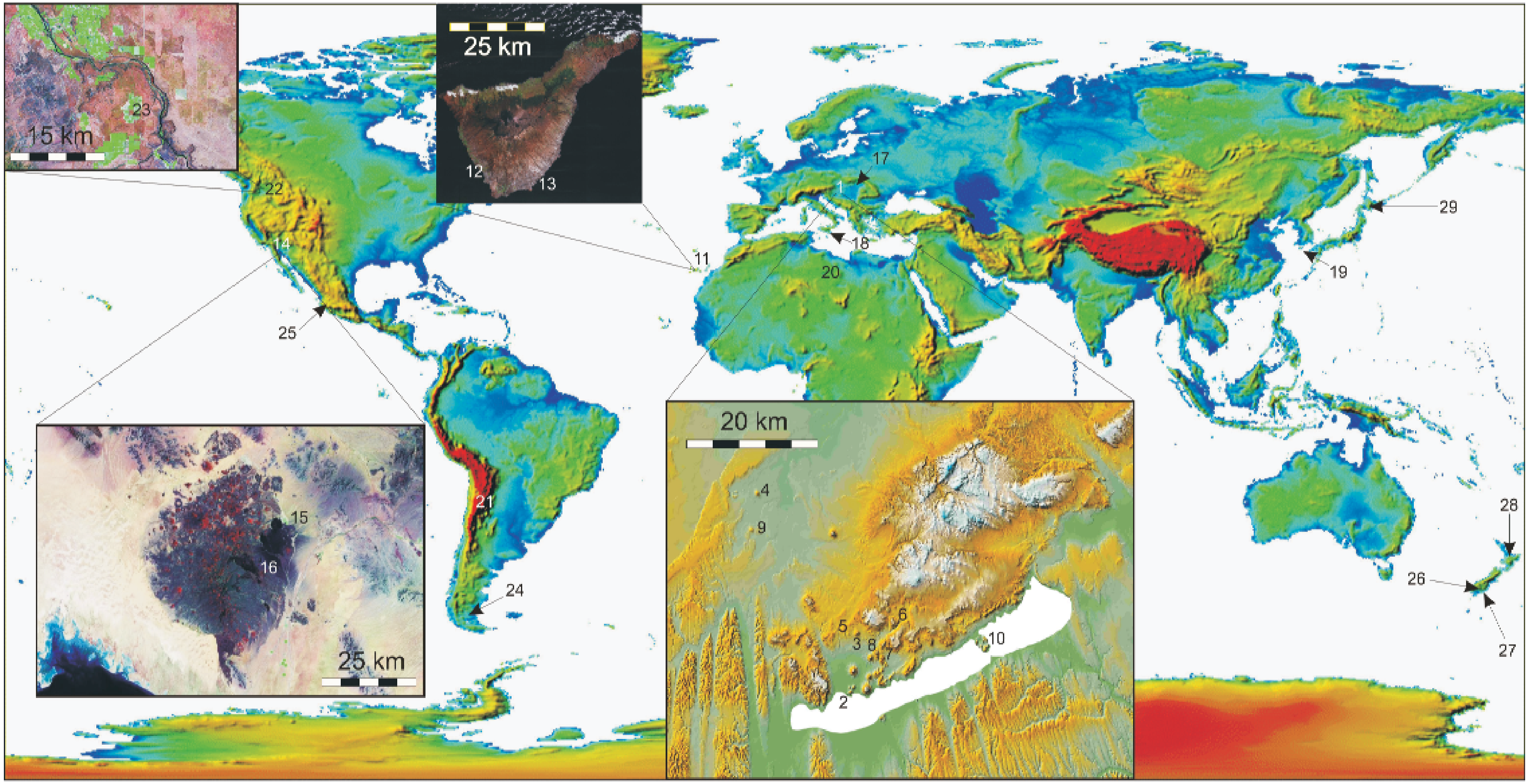
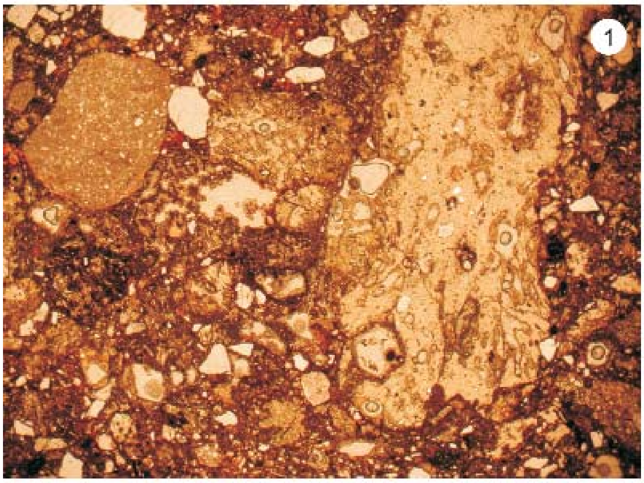
Location map

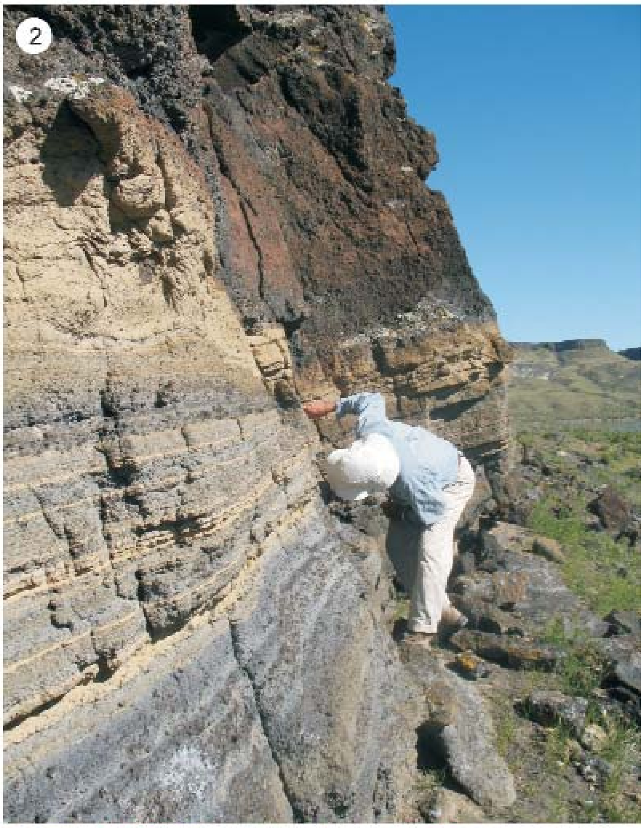


1 — Western Hungary
2 — Szigliget, Hungary
3 — Haláp, Hungary
4 — Ság-hegy, Hungary
5 — Véndeg-hegy, Hungary
6 — Pula, Hungary
7 — Szentbékkálla, Hungary
8 — Hajagos, Hungary
9 — Kissomlyó, Hungary
10 — Tihany, Hungary
11 — Tenerife, Spain
12 — Caldera del Rey maar, Spain
13 — Montana Pelada, Spain
14 — Sonora, Mexico
15 — Cerro Colorado, Mexico
16 — Crater Elegante, Mexico
17 — Tokaj/Pálháza, Hungary
18 — Palagonia/Hyblean Mts, Italy
19 — Jeju Island, Korea
20 — Al Haruj, Libya
21 — Altiplano, Chile
22 — Snake River, Idaho, USA
23 — Snake Butte, Idaho, USA
24 — Pali Aike Volcanic Field, Argentina
25 — Ceboruco volcano, Mexico
26 — Waipiata Volcanic Field, New Zealand
27 — Mt. Charles, Otago peninsula, New Zealand
28 — Taupo, New Zealand
29 — Usu, Japan

Plate I

1. Photomicrograph of juvenile fragment rich lapilli tuff. Blocky glass shard indicates phreatomagmatic fragmentation of melt. Sample from phreatomagmatic lapilli tuff bed of the Szigliget diatreme in Hungary. The shorter side of the view is about 4 mm.
2. Outcrop-scale observations could give hint about the transportation and depositional processes of the tephra formation. Phreatomagmatic tephra layers interbedded with reworked volcaniclastic beds forming extensive sheets along the Snake River in Idaho.
3. Commonly hand specimen can help to establish the fragmentation history and give indication for the transportation/deposition regime the volcaniclastic succession formed. This photo was taken about a graded volcaniclastic lapilli tuff from the Triassic Pietre Verde succession from Hungary.
4. Phreatomagmatic tuff ring succession exposes fall and surge units in the Pali Aike Volcanic Field in Argentina.
5. Volcanic field wide scale investigations of each eruptive centers are especially important in analysing the eruptive history of volcanic fields such as the Pali Aike Volcanic Field, Santa Cruz, Argentina.
6. Oval shape vesicles in volcanic glass shards are characteristic for magmatic vesiculation such as in this sample from Al Haruj, Libya. The shorter side of the view is about 4 mm.





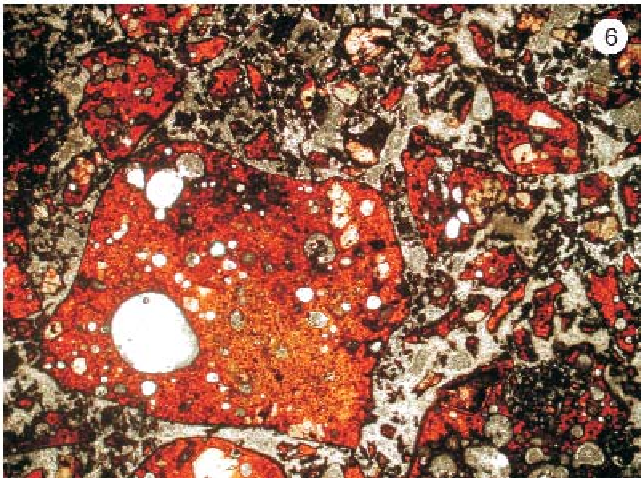
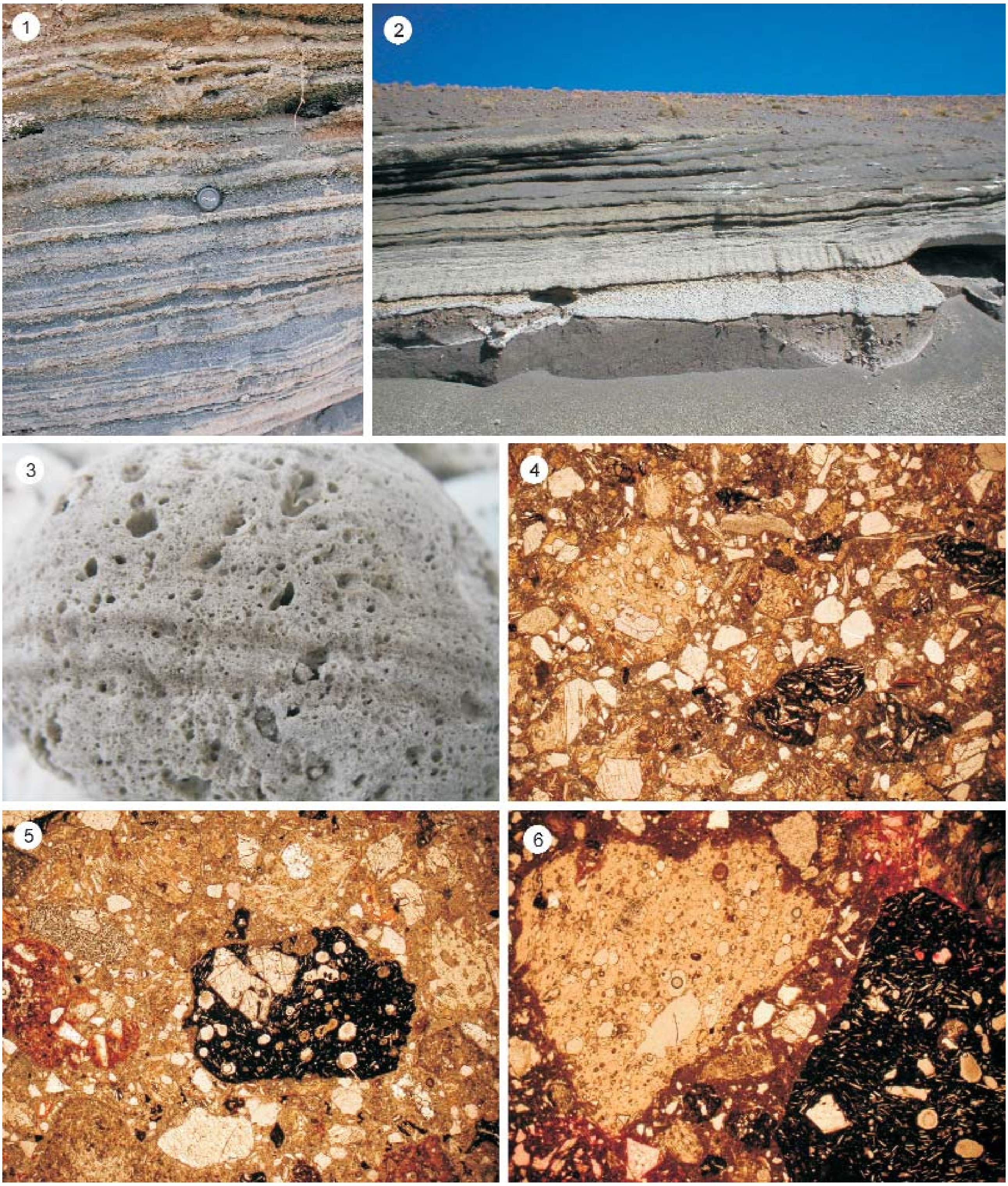
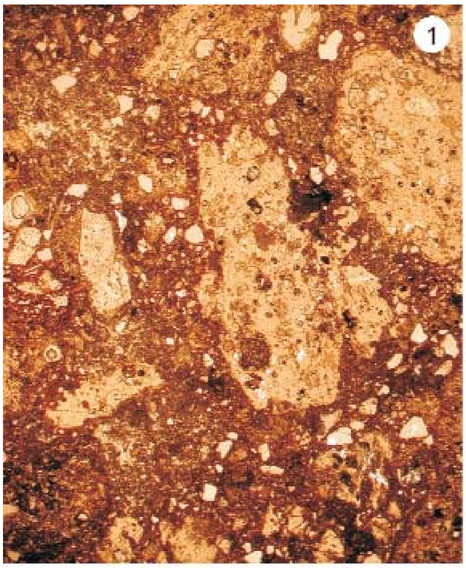


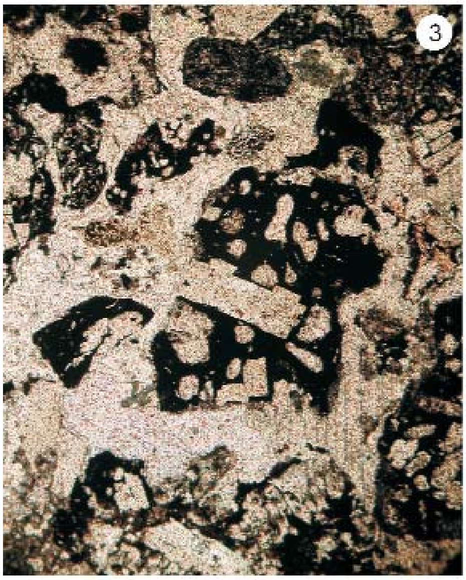
Plate II



1. Black scoriaceous lapilli beds from a scoria cone near Ceboruco volcano in Western Mexico.
2. Pumiceous Plinian air fall beds forming a great succession in the Altiplano, Northern Chile.
3. Block of banded andesite-dacite pumice erupted on May 22, 1915. This initially large block of hot pumice fell on the eastern slopes of Lassen Peak and broke along a set of polyhedral joints that formed as it cooled in place. The striking colour banding reflects mingling of two compositional varieties of erupted magma, contorted by flowage of the incompletely mixed viscous liquids. Photo by R. L. Christiansen, September, 1987.
4. Blocky, angular shape volcanic glass shards from phreatomagmatic lapilli tuff succession of the Haláp tuff ring, Hungary.
5. Dark, vesicular tachylite glass from a lapilli tuff of Ság-hegy indicating magmatic fragmentation. The shorter side of the picture is about 4 mm.
6. Blocky, transparent sideromelane glass shard from the Véndeg-hegy diatreme, Hungary.

Plate III

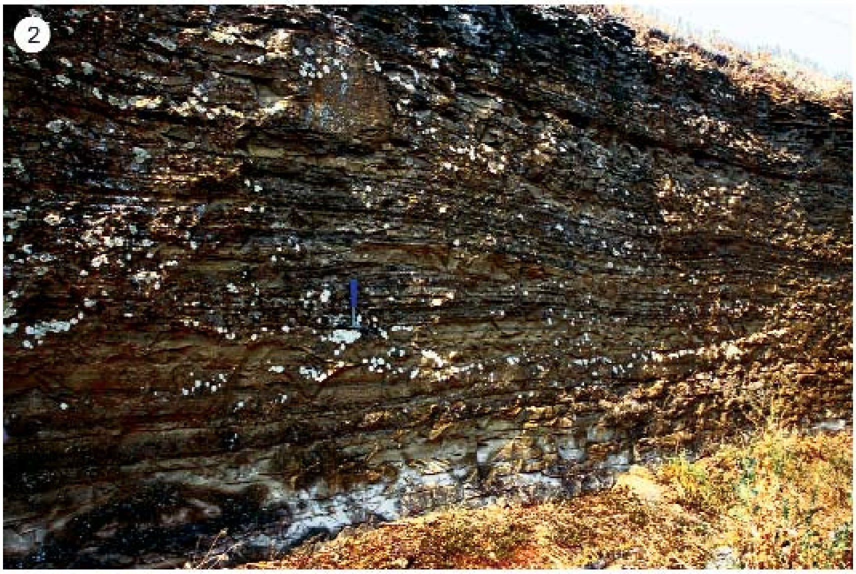


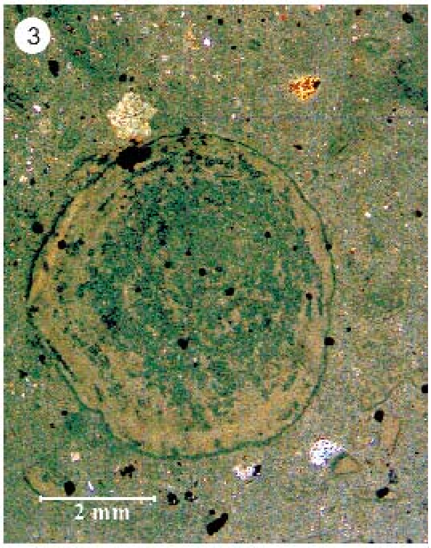


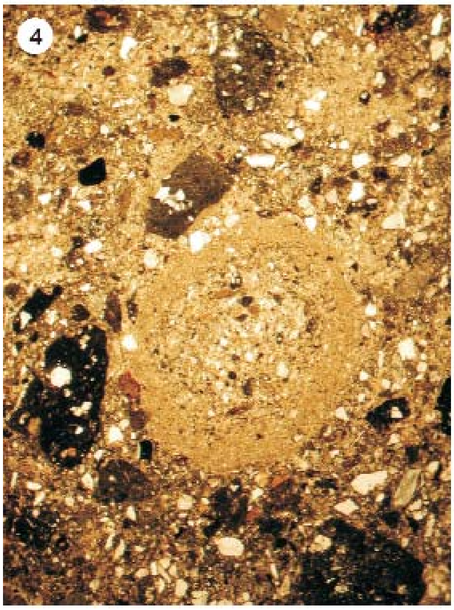
1. Predominantly sideromelane glass shards dominate tephras formed by shallow level fragmentation and/or fragmentation through open vent of melt by magma-water interaction such as in this picture taken from a lapilli tuff of the Szigliget diatreme, Hungary. The shorter side of the view is appx. 4 mm.
2. In fragmentation through deep subsurface level results large number of disrupted accidental lithic fragments in the resulting tephra as it is visible in the sample from Pula, Hungary.
3. Tachylite shard-dominated lapilli tuff indicates travel of pyroclasts through air, and therefore their existence could be used to infer near surface and/or open vent fragmentation such as in this picture of a lapilli tuff from Northern Chile. The short side of the view is about 4 mm.
4. Base surge succession from the Waipiata Volcanic Field, New Zealand.
5. Base surge dune from a maar in southern Tenerife.

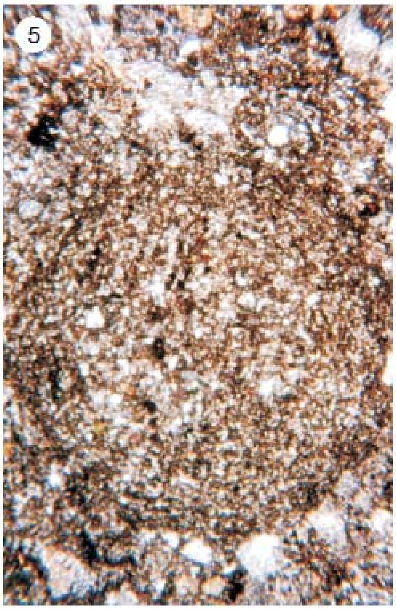
Plate IV

1. Accidental lithic rich surge bed succession from near vent volcanic facies of the Crater Elegante maar from Sonora, Mexico.
2. Base surge beds developed around a tuff ring erupted through a soft substrate (mud and sand) country rock succession in the Little Hungarian Plain.
3. Accretionary lapilli from a maar succession of southern Tenerife.
4. Rim type accretionary lapilli from the Szentbékkalla phreatomagmatic succession, Hungary. The short side of the view is about 4 mm.
5. Core type accretionary lapilli from an emergent tuff cone of Mt Charles in the Otago Peninsula, New Zealand. The short side of the view is about 4 mm.
6. Cored lapilli from the Pula phreatomagmatic succession, Hungary. The short side of the picture is about 4 mm.









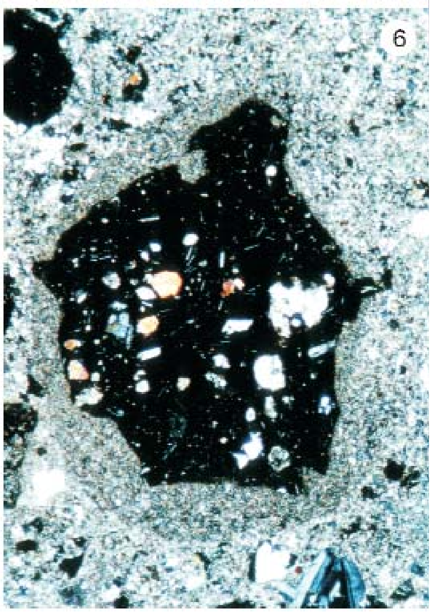
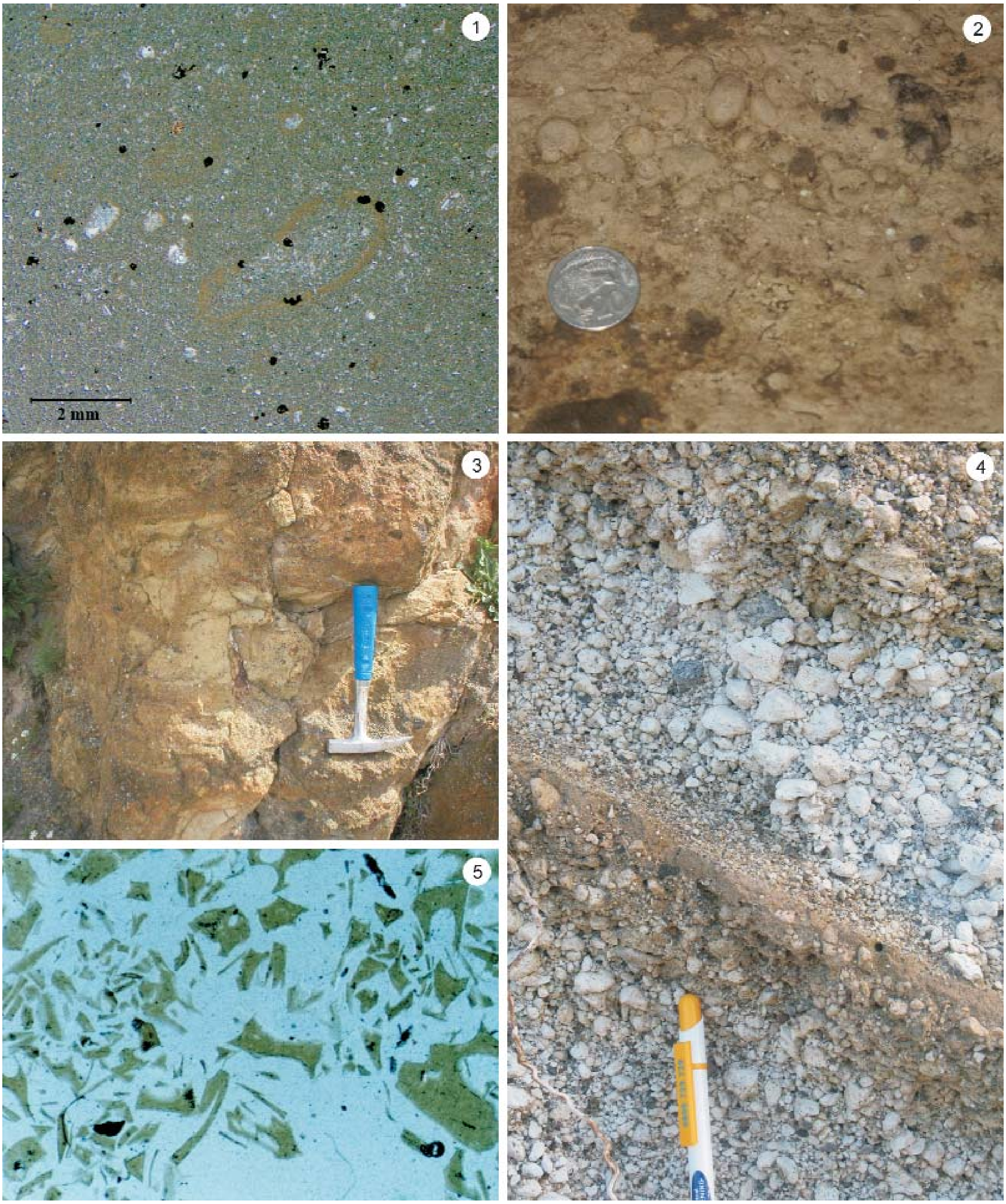
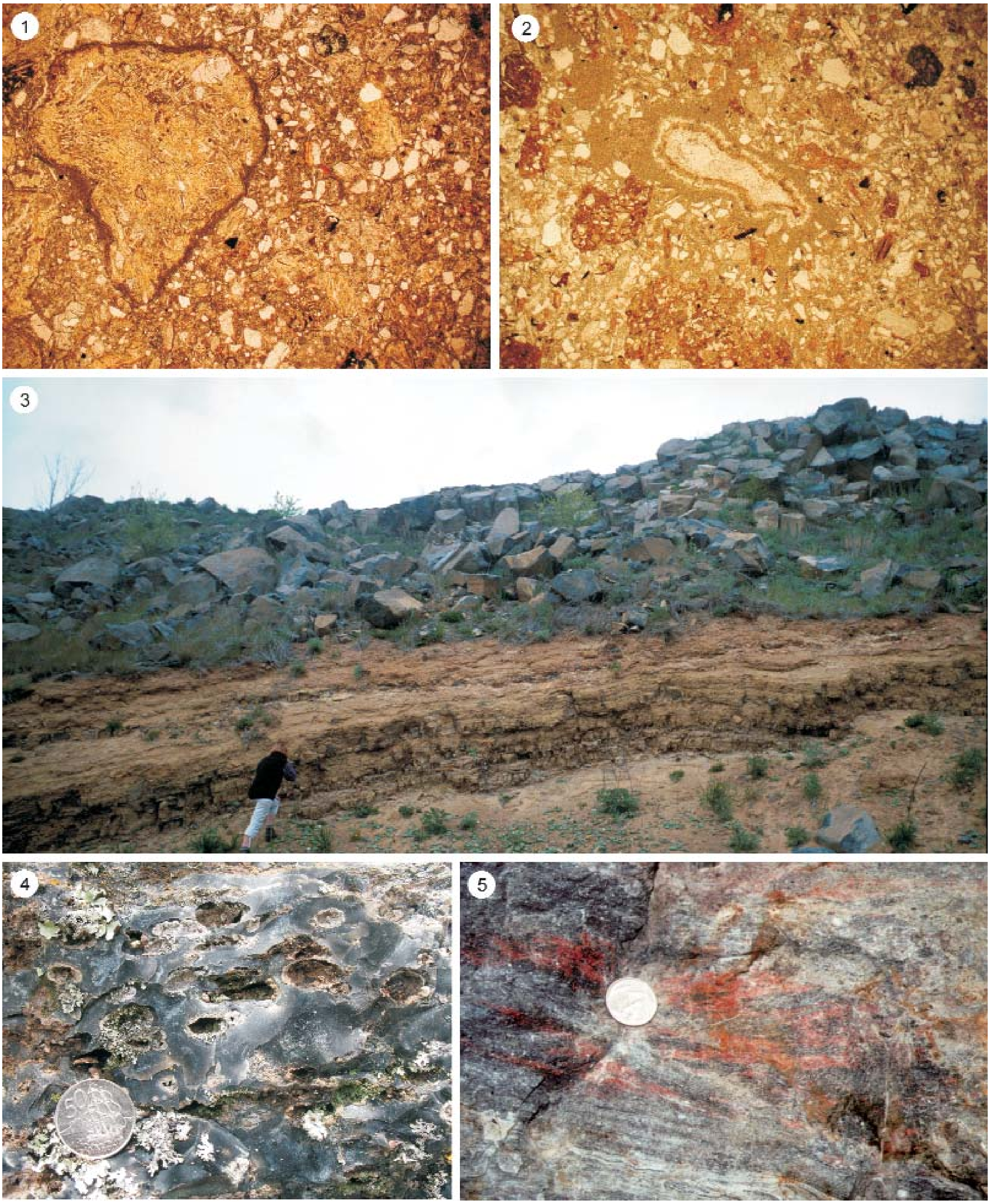


Plate V



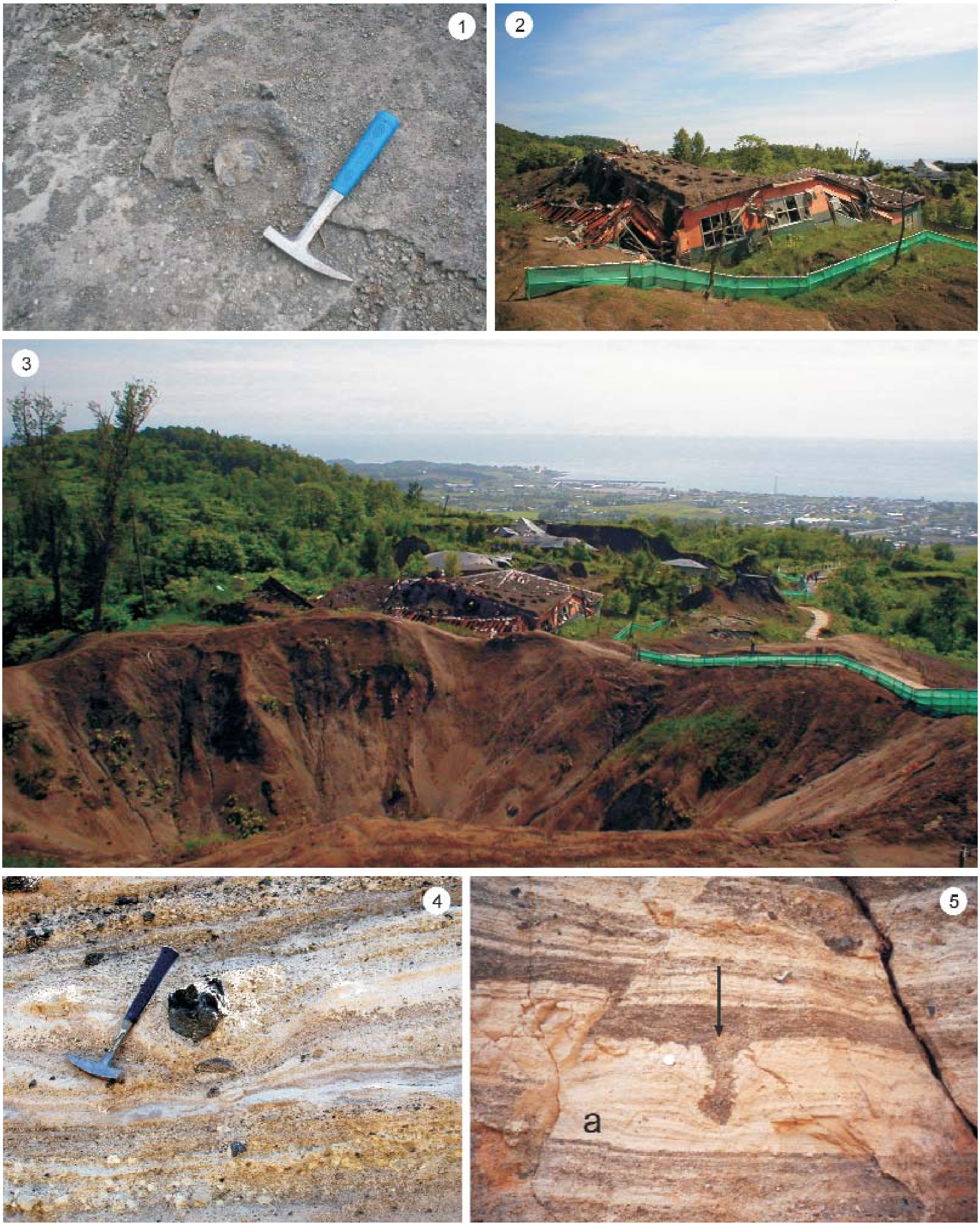
1. Mud clots from tuff bed of the Montana Pelada tuff ring in southern Tenerife. The fine mud aggregates surrounded by thin limonite-rich film layer.
2. Large accretionary lapilli from the phreatoplinian succession of the AD 180 Taupo eruption, New Zealand.
3. Mud chunks from the phratomagmatic succession of Szigliget, Hungary.
4. Black obsidian in Plinian fall deposit.
5. Glass shards (limu shells) from Seamount 9 from the Pacific forming hyaloclastite succession. Shorter side of the picture is about 2 cm.

Plate VI



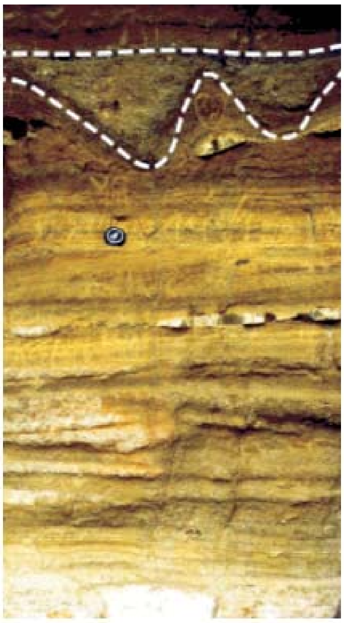
1. Palagonite rim probably developed during transportation of the pyroclasts in the base surge currents formed the pyroclastic succession of Haláp, Hungary. Shorter side of the picture is about 4mm.
2. Advanced palagonitisation of a sideromelane glass shard rich bed resulting red discolorisation of the glass shards as it can be seen from this sample from Hajagos, Hungary. Shorter side of the view is about 4 mm.
3. Palagonite bed of the pyroclastic succession of the Haláp tuff ring, Hungary.
4. Spherulite and lithophysae in a silicic lava flow.
5. Perlitic lava flow from the Tokaj Mtns, NE Hungary.

Plate VII



1. Ballistic bomb in phreatomagmatic tephra bed surface in the Kissomlyó tuff ring, Hungary.
2. Ballistic bombs damaged buildings during the Usu 2000 eruption, Hokkaido, Japan.
3. Phreatic explosion crater in Usu, Hokkaido, Japan. Note the destroyed house showed on Plate VI, 2.
4. Impact sags in a phreatomagmatic succession of a maar from southern Tenerife.
5. Drop stone-like feature (arrow) caused by a ballistic bomb impacted on a soft and water-saturated accretionary lapilli-rich freshly deposited tephra (a) of the Ság-hegy tuff ring, western Hungary.

Plate VIII

1. Plastically deformed impact sag forming twisted and chaotic bed surface of a base surge succession of a maar in southern Tenerife.
2. Hard, strongly diagenised tuff layer (middle in the phreatomagmatic succession of Tihany. Note the V-shaped erosional channels (dashed lines) in cross-section just above the lense cap.
3. Dish structures of a volcaniclastic succession near Sinker Butte, Idaho.
4. Sand volcano in cross section from the maar lacustrine units of the Pula maar, Hungary.
5. Clastic dykes intruded into growing lava dome and hyaloclastite unit of Pálháza, Hungary.