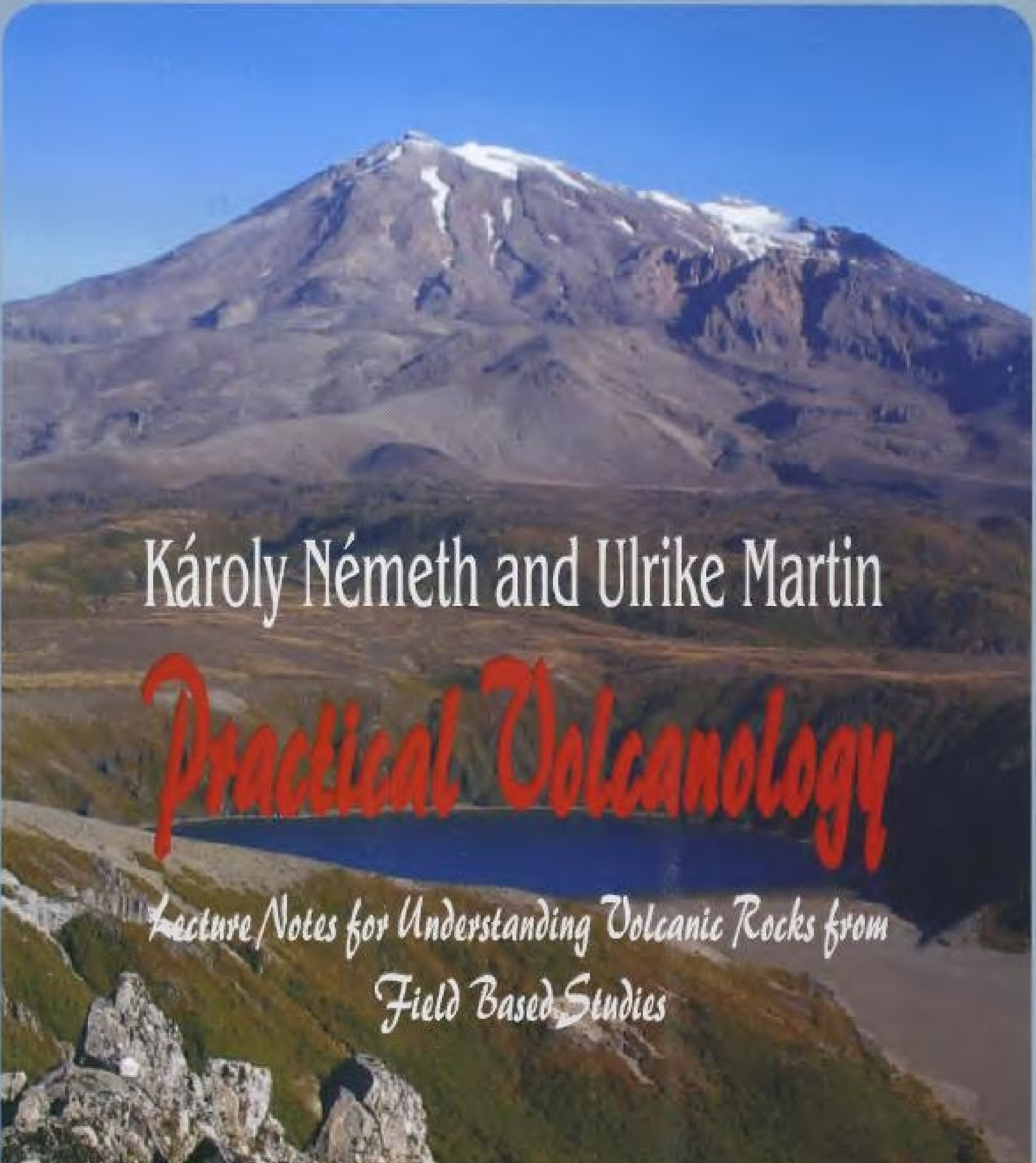

Károly Németh and Ulrike Martin

# Practical Volcanology

## Lecture Notes for Understanding Volcanic Rocks from Field Based Studies

Budapest, 2007

Occasional Papers of the Geological Institute of Hungary,
volume 207

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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.

Published by the Geological Institute of Hungary
*Responsible editor:*
LÁSZLÓ KORDOS
director

**ISBN 978-963-671-259-4**

# Contents

# 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 which can take place at the Geological Institute of Hungary, Budapest in the future.

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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

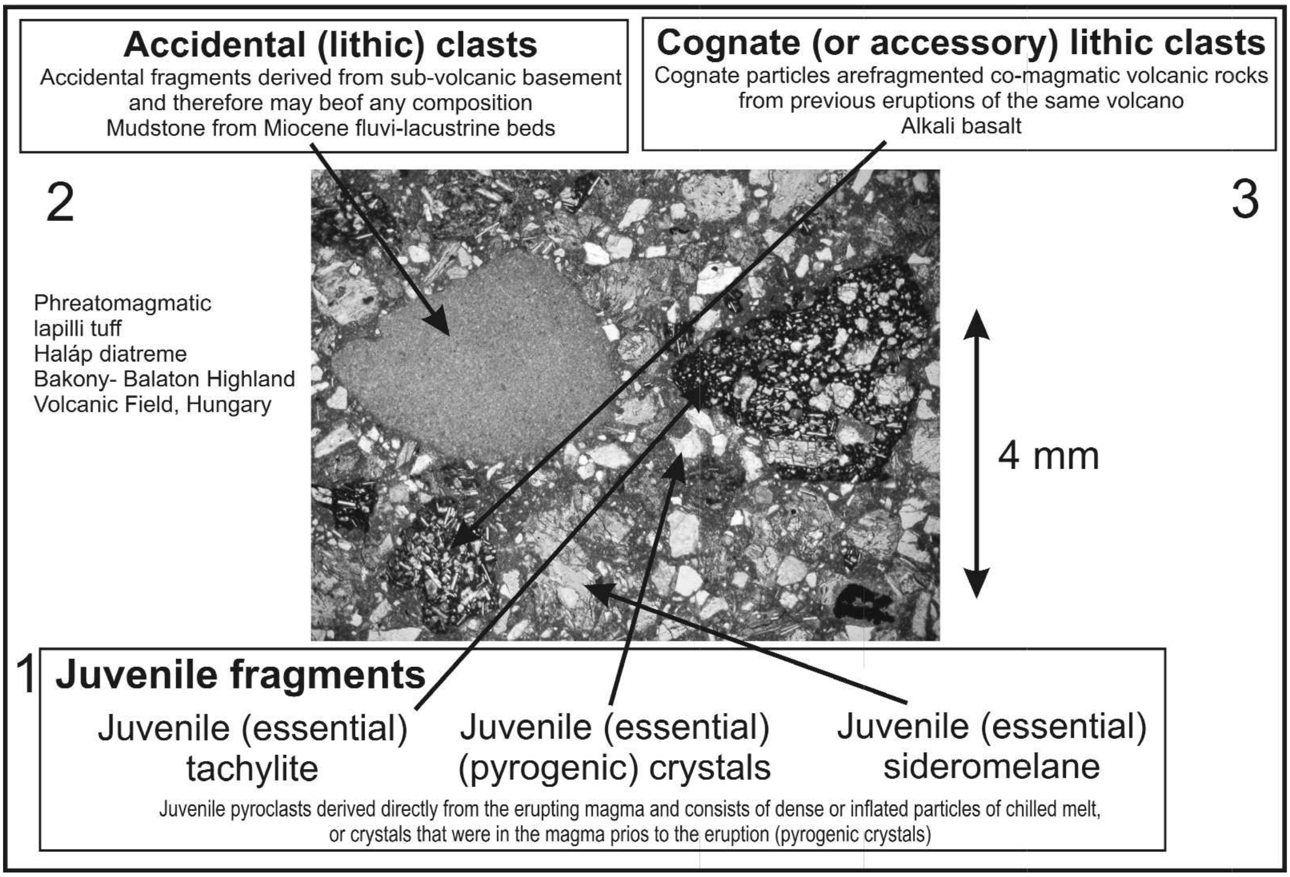


*Figure 1.1.* Application of Fisher and Schmincke (1984) terminology for a pyroclastic rock from the Mio/Pliocene Bakony – Balaton Highland Volcanic Field maar/diatreme remnant (same sample as on Figure 1.2)

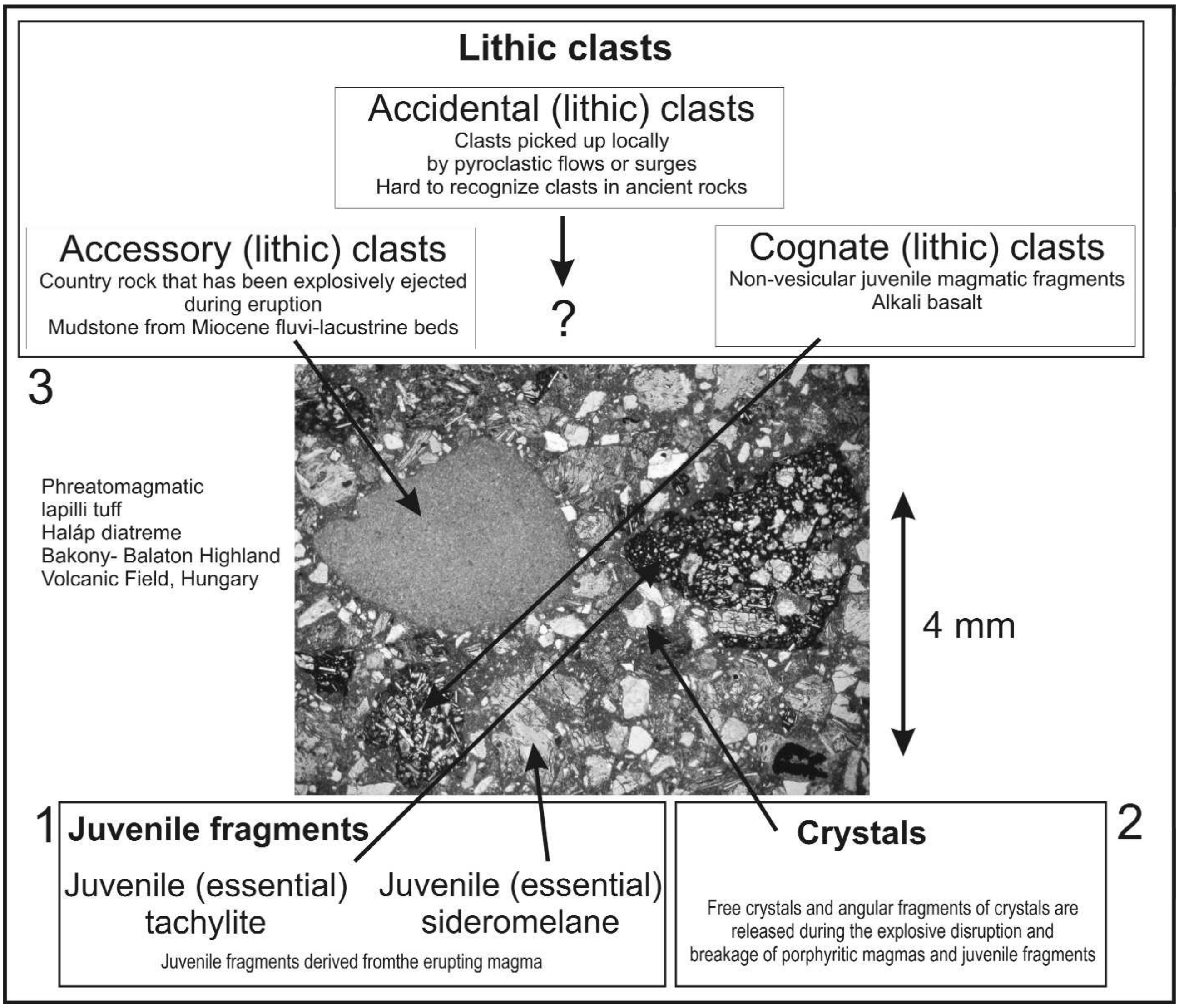


***Figure 1.2.*** Application of Cas and Wright (1987) terminology for a pyroclastic rock from the Mio/Pliocene Bakony – Balaton Highland Volcanic Field maar/diatreme remnant (same sample as on Figure 1.1)

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)

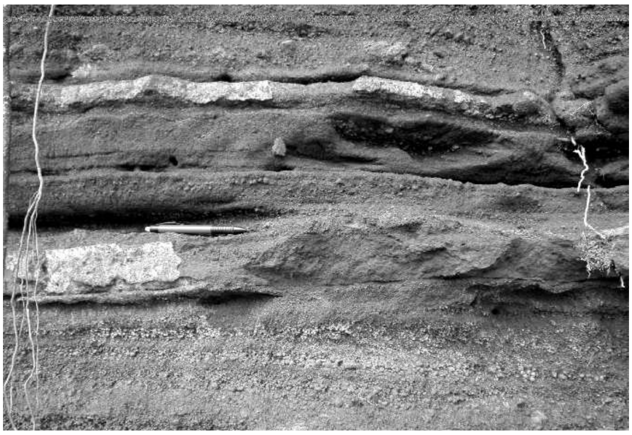


***Figure 1.4.*** Undulating tuff bed (see under the pen) from a tuff ring erupted in 1913 in West Ambrym, Vanuatu. Pen is 15 cm long

***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 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

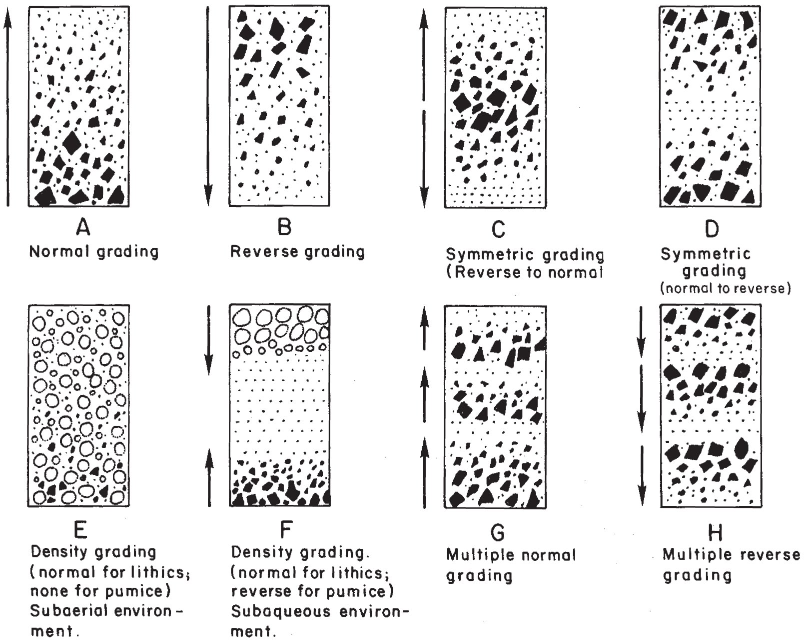


***Figure 1.6.*** Typical grading types after Fisher and Schmincke 1984: p. 109, fig. 5-19

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

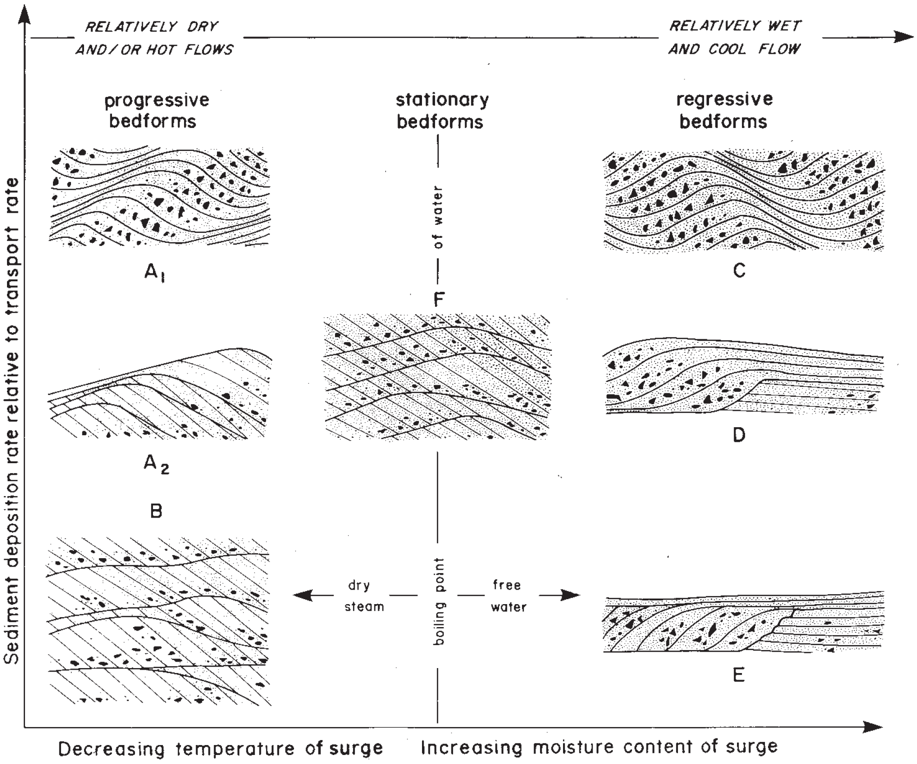


***Figure 1.8.*** Types of pyroclastic surges bedforms and internal cross-stratification (after Allen 1982 in Cas and Wright 1987: p. 215, fig. 7.42) as function of depositional rate and surge temperature and moisture content

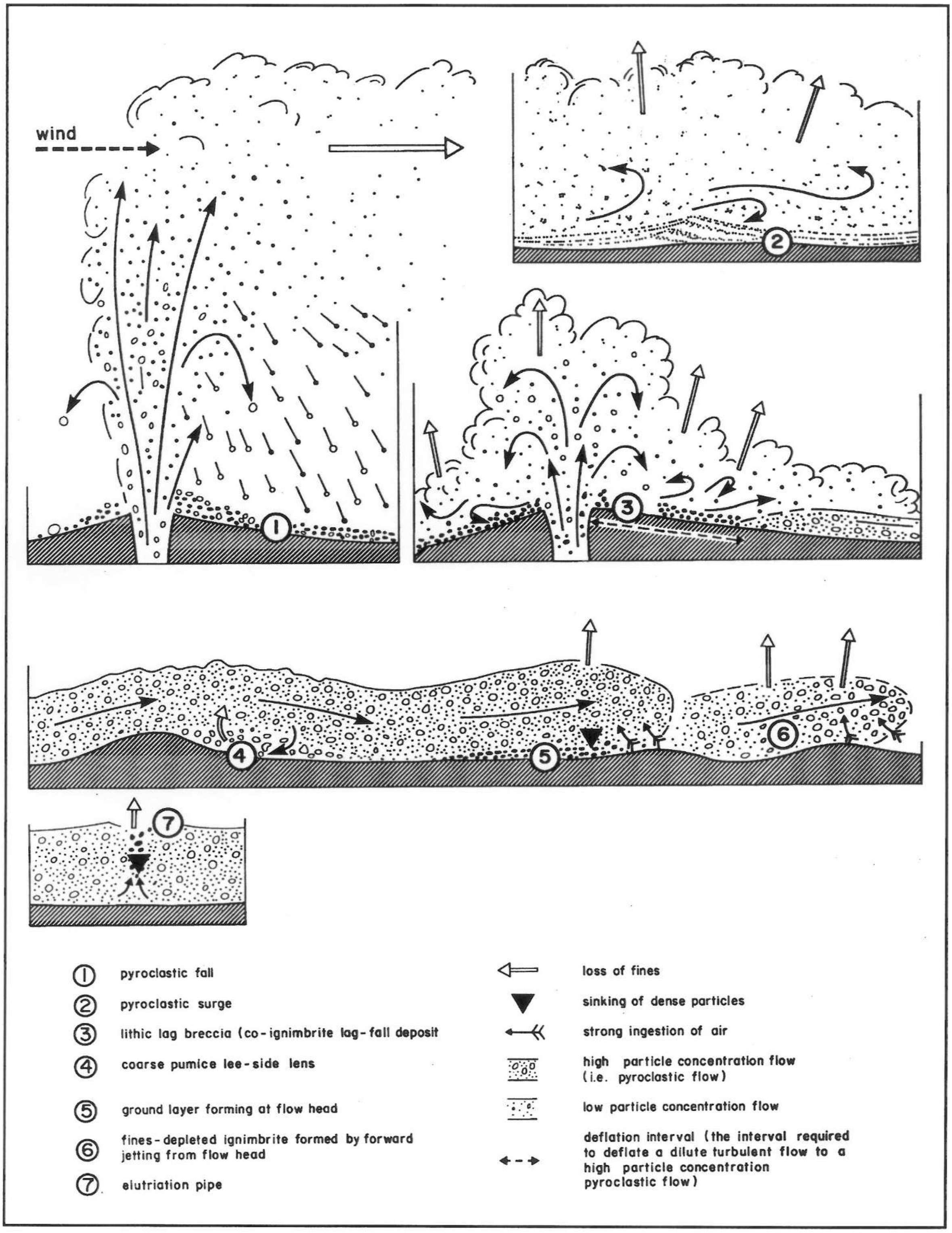


***Figure 1.9.*** Theoretical models of development of good sorting (after Walker 1983 in Cas and Wright 1987: p. 220, fig. 7.46)

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).

## 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-

| 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)

| Clast size | Pyroclast | Pyroclastic deposit | |
|---|---|---|---|
| | | Mainly unconsolidated: tephra | Mainly consolidated: pyroclastic rock |
| | Block, bomb | Agglomerate, bed of blocks<br>or<br>bomb, block tephra | Agglomerate, pyroclastic breccia |
| 64 mm | | | |
| | Lapillus | Layer, bed of lapilli<br>or<br>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)

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).

| Pyroclastic[a] | Tuffites (mixed pyroclastic-epiclastic) | Epiclastic (volcanic and/or nonvolcanic) | Average clast size (mm) |
|---|---|---|---|
| Agglomerate, agglutinate<br>pyroclastic breccia<br>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

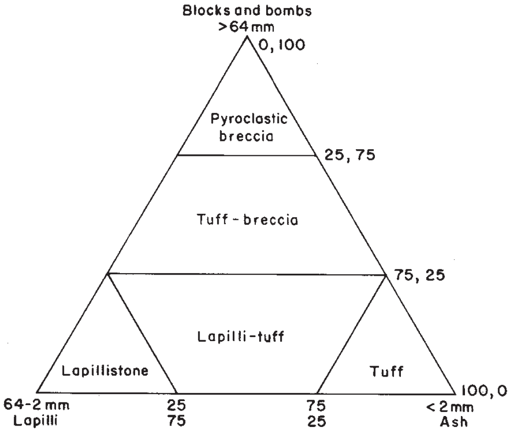


*Figure 1.13.* Ternary diagram represents mixture terms and end-member rock terms for pyroclastic fragments after FISHER (1966) (from FISHER and SCHMINCKE 1984: p. 92, fig. 5-1). In SCHMID (1981) classification lapillistone is replaced by lapilli tuff

| Grainsize (mm) | Pyroclastic fragments: round and fluidally shaped | Pyroclastic fragments: angular | Name of unconsolidated aggregate | Lithified equivalent |
|---|---|---|---|---|
| coarse | | | | |
| 256 | bombs | blocks | agglomerate (bombs) or pyroclastic breccia | agglomerate (bombs) or pyroclastic breccia |
| fine | | | | |
| 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

## 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.

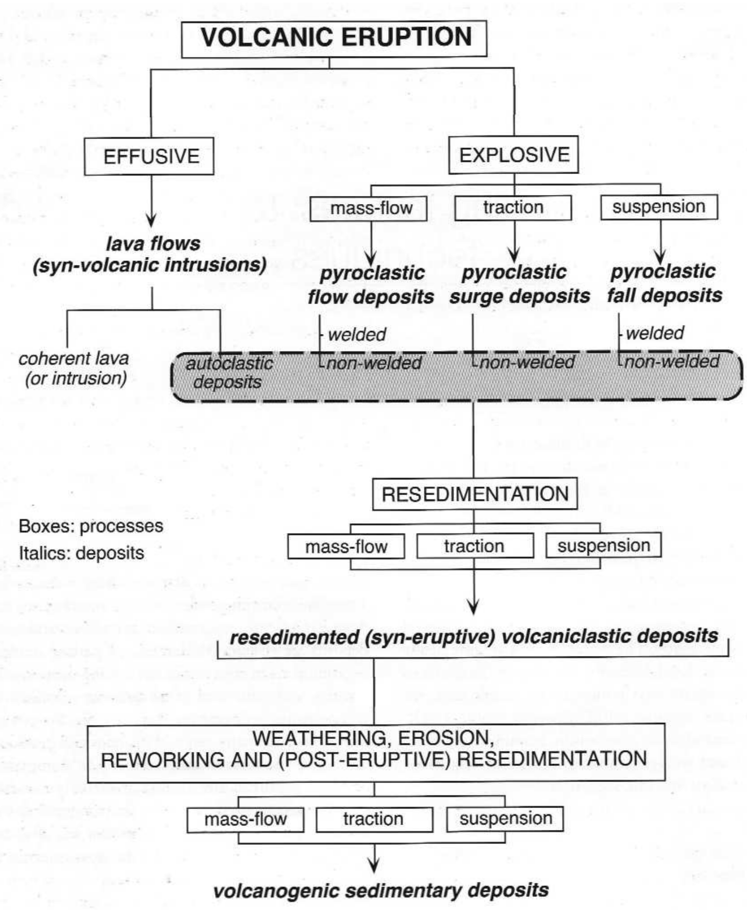


*Figure 1.17.* Genetic classification of volcanic deposits after McPhie et al. 1993: p. 2, fig. 1). Depositional processes are the same (mass flow, traction, suspension) in primary, resedimented and reworking volcaniclastic deposits. Non-welded primary volcaniclastic deposits marked in grey field with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

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).

### 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

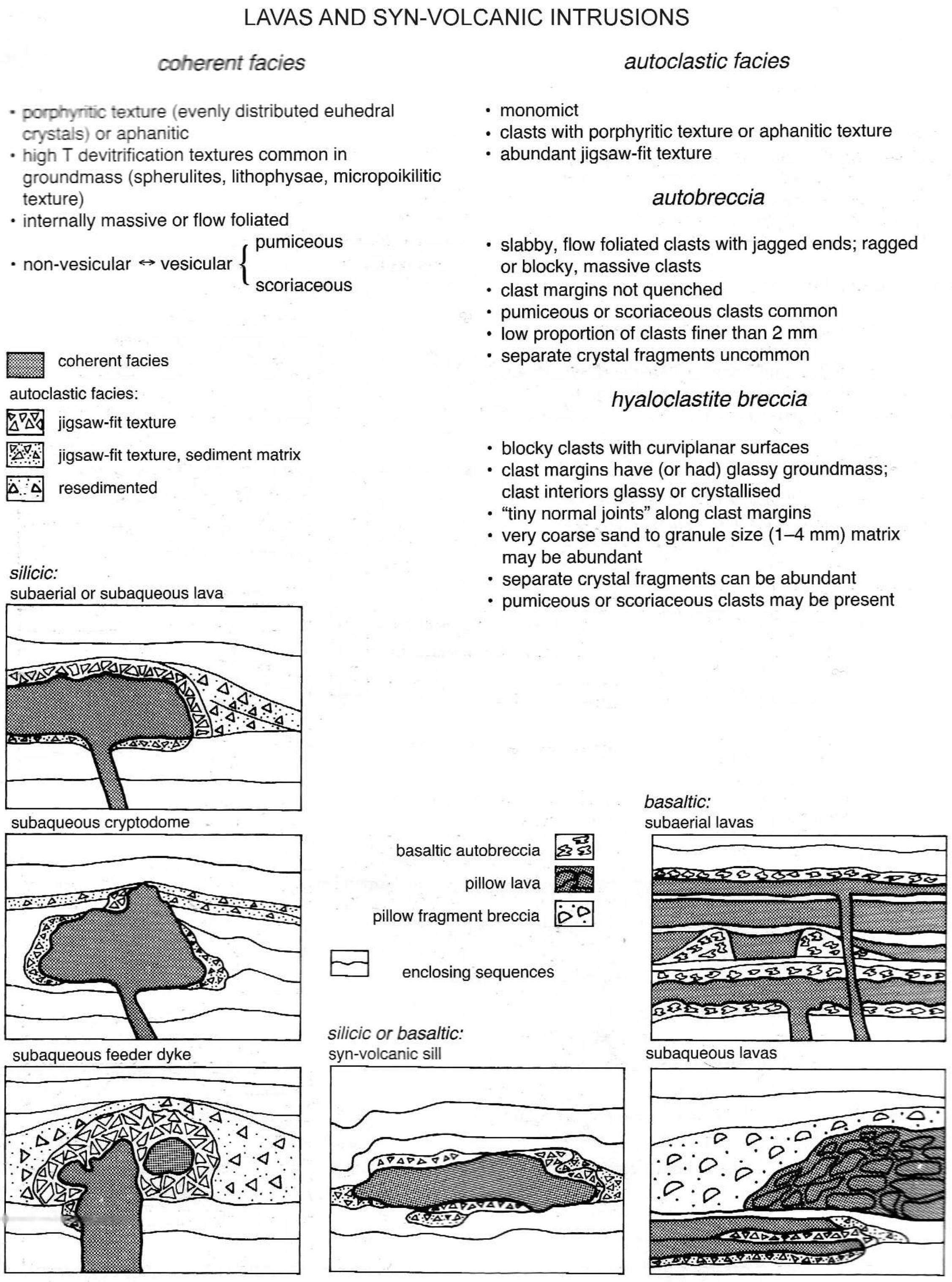
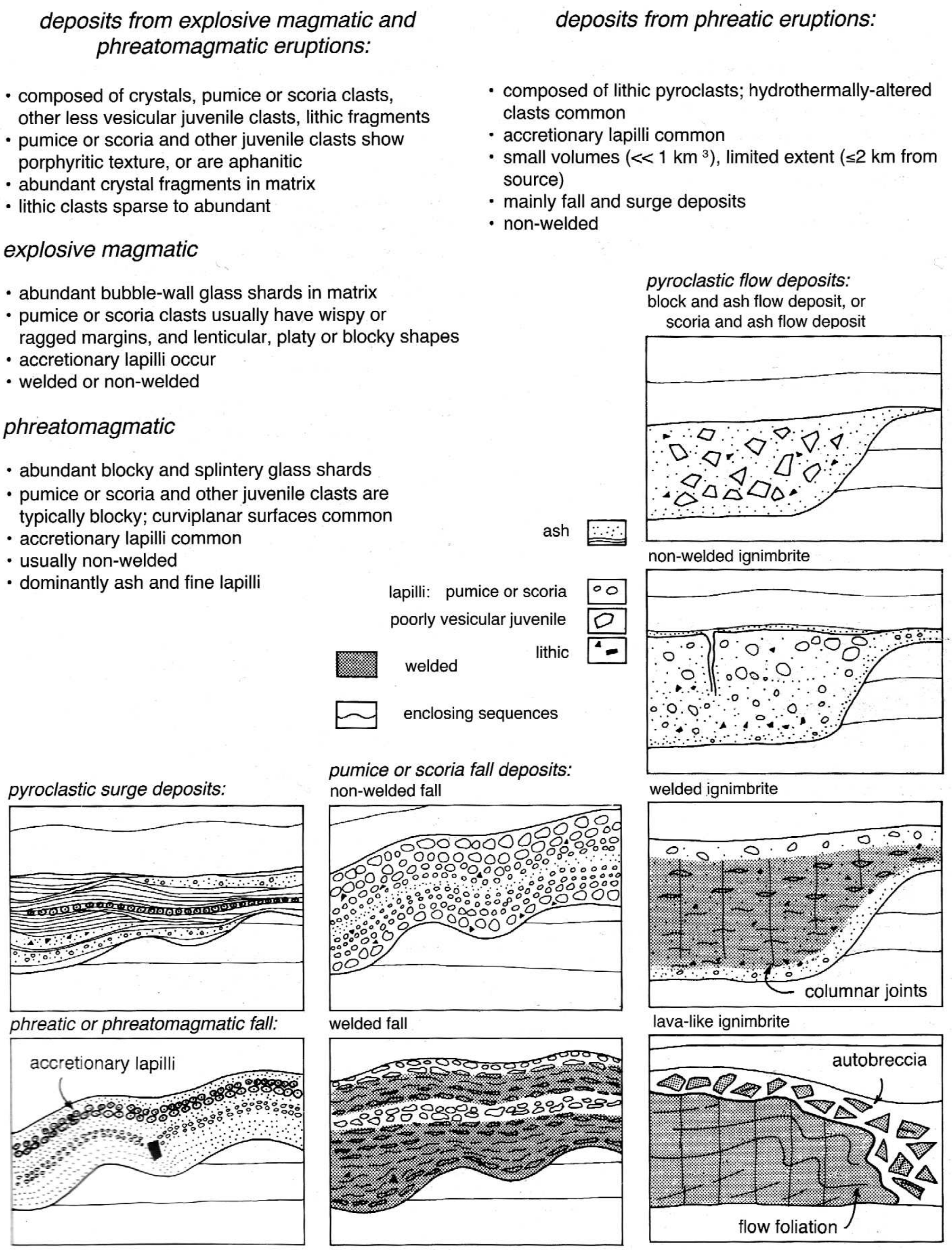


*Figure 1.18.* Genetic classification and basic characteristics of lavas and syn-volcanic intrusions after McPhie et al. 1993: p. 4, fig. 2



*deposits from explosive magmatic and phreatomagmatic eruptions:*

- composed of crystals, pumice or scoria clasts, other less vesicular juvenile clasts, lithic fragments
- pumice or scoria and other juvenile clasts show porphyritic texture, or are aphanitic
- abundant crystal fragments in matrix
- lithic clasts sparse to abundant

*explosive magmatic*

- abundant bubble-wall glass shards in matrix
- pumice or scoria clasts usually have wispy or ragged margins, and lenticular, platy or blocky shapes
- accretionary lapilli occur
- welded or non-welded

*phreatomagmatic*

- abundant blocky and splintery glass shards
- pumice or scoria and other juvenile clasts are typically blocky; curviplanar surfaces common
- accretionary lapilli common
- usually non-welded
- dominantly ash and fine lapilli

*deposits from phreatic eruptions:*

- composed of lithic pyroclasts; hydrothermally-altered clasts common
- accretionary lapilli common
- small volumes (<< 1 km $^3$), limited extent (≤2 km from source)
- mainly fall and surge deposits
- non-welded

***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

## RESEDIMENTED SYN-ERUPTIVE VOLCANICLASTIC DEPOSIT

- dominated by texturally unmodified juvenile clasts
- narrow range of clast types and composition
- sedimentation units and successions of units are compositionally uniform or show systematic changes
- bedforms indicate rapid deposition (mass-flow deposits common)

### *resedimented autoclastic deposits:*

*shallow subaqueous:*

- mixture of autoclastic and pyroclastic particles
- combination of mass-flow and traction current bedforms
- dominated by clasts coarser than ~2 mm

*deep subaqueous:*

- poorly vesicular, quenched lava clasts dominant
- mainly mass-flow bedforms
- may have primary dips up to ~25°
- granule → cobble size clasts dominant
- associated with in situ hyaloclastite and coherent lava

### *resedimented pyroclastic deposits:*

- composed of pyroclasts

*subaerial and shallow subaqueous:*

- combination of mass-flow, hyperconcentrated flow and traction current bedforms
- depleted in fine ash

*deep subaqueous:*

- very thick mass-flow sedimentation units that consist of a massive, crystal- and lithic clast-rich base and a normally graded or stratified, pumice- and shard-rich top
- intraclasts present near base of mass-flow units
- laminated, shard-rich units (settled from suspension)

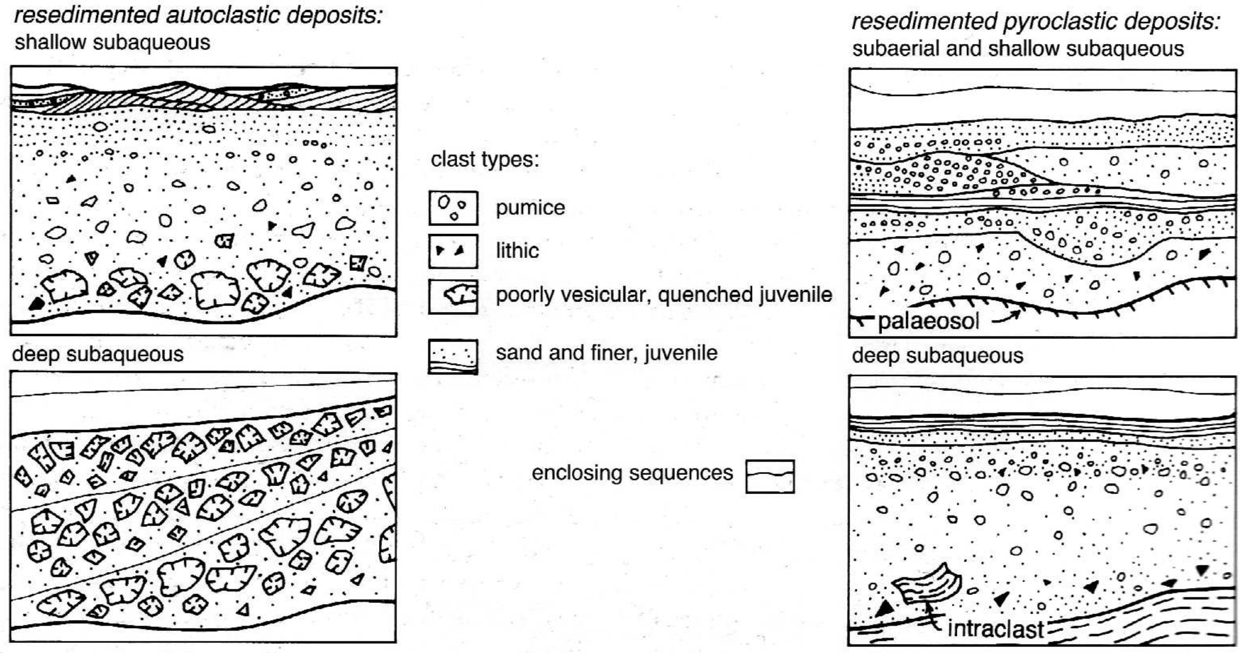


***Figure 1.20.*** Genetic classification and basic characteristics of resedimented syn-eruptive volcaniclastic deposits after McPhie et al. 1993: p. 6, fig. 4)

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-

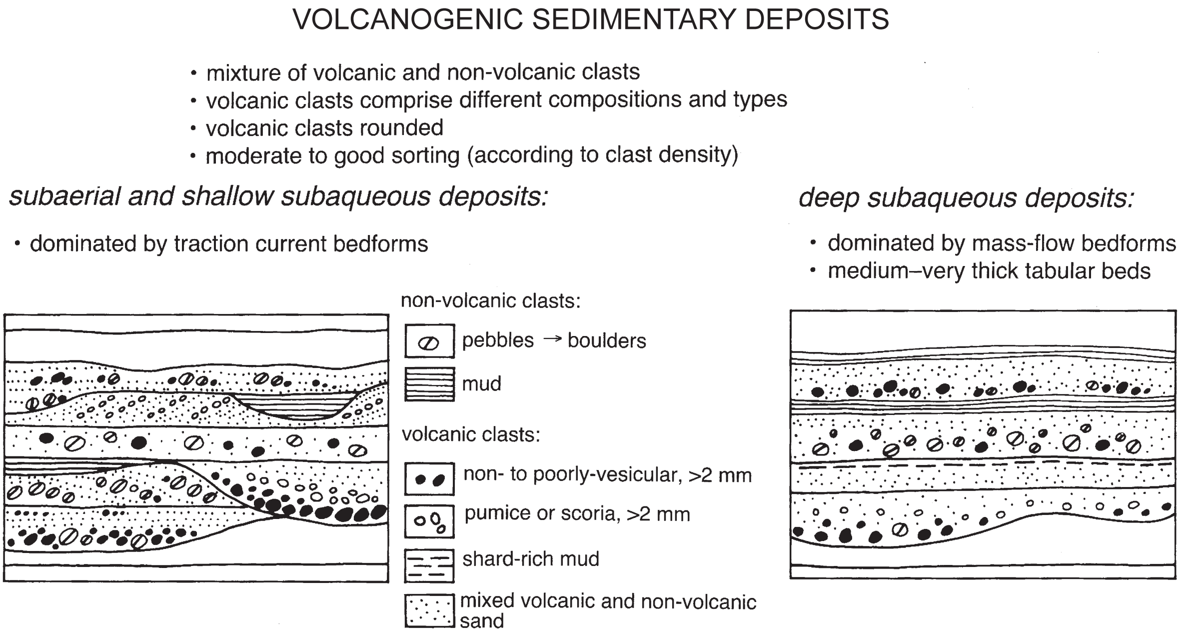
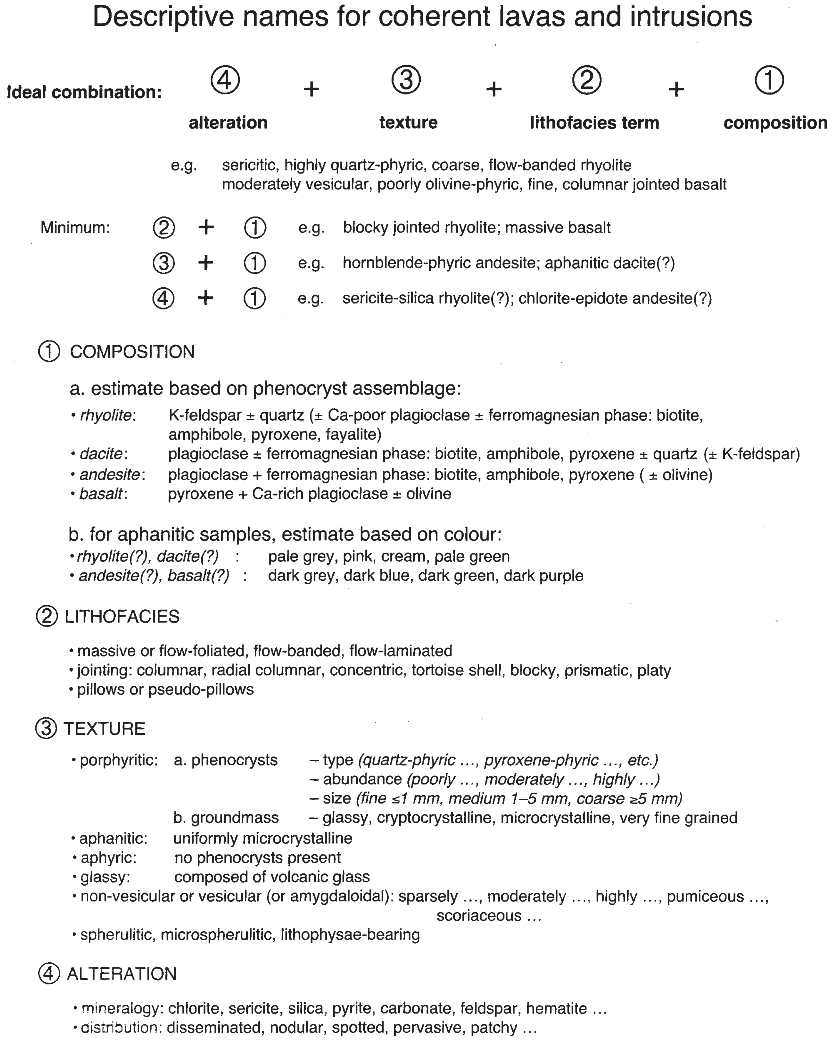


***Figure 1.21.*** Genetic classification and basic characteristics of volcanogenic sedimentary deposits after McPhie et al. 1993: p. 7, fig. 5

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 alteration + ③ texture + ② lithofacies term + ① 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

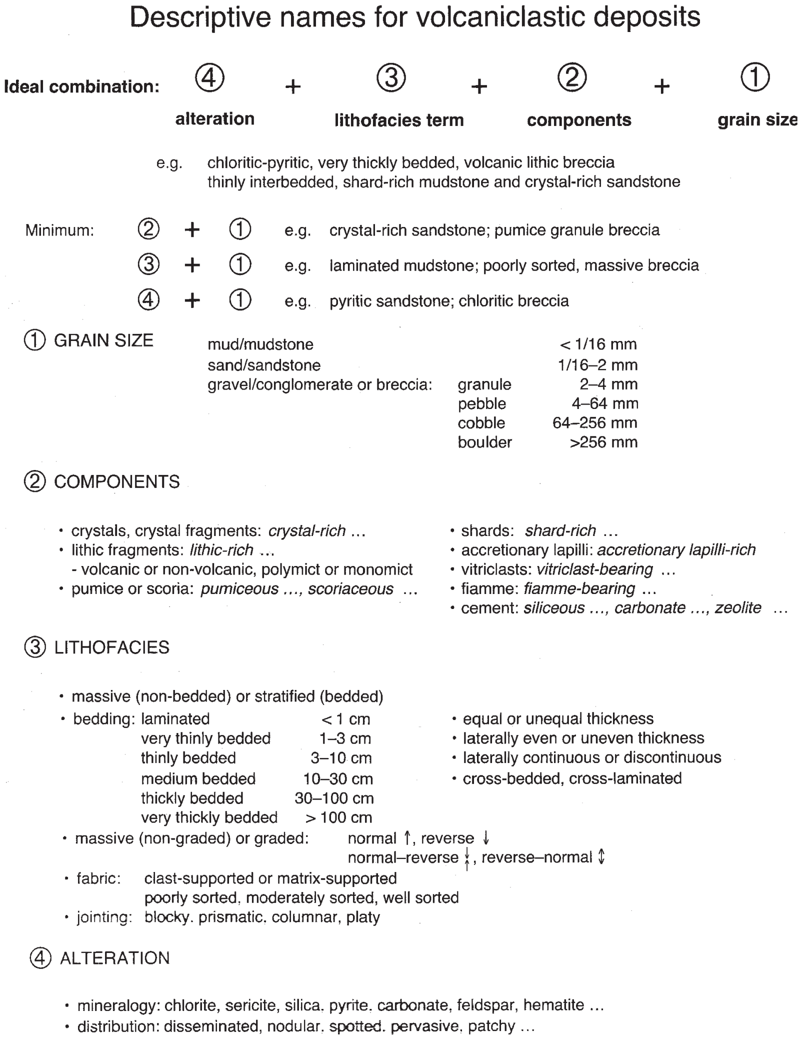
- mineralogy: chlorite, sericite, silica, pyrite, carbonate, feldspar, hematite ...
- distribution: disseminated, nodular, spotted, pervasive, patchy ...

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

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.



Descriptive names for volcaniclastic deposits

**Ideal combination:** ④ + ③ + ② + ①

**alteration** **lithofacies term** **components** **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:

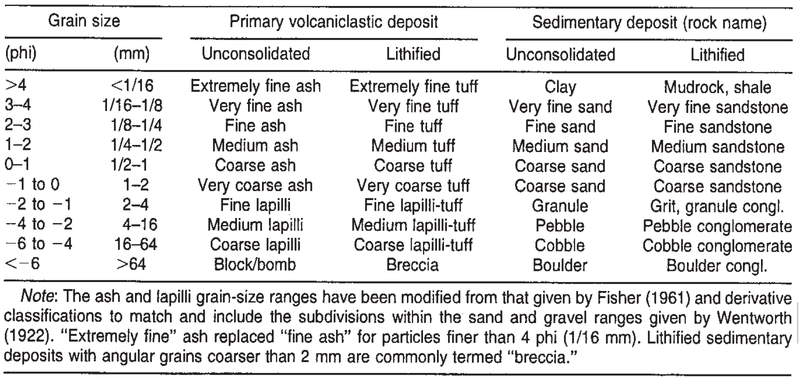
| | |
|---|---|
| laminated | < 1 cm |
| very thinly bedded | 1–3 cm |
| thinly bedded | 3–10 cm |
| medium bedded | 10–30 cm |
| thickly bedded | 30–100 cm |
| very thickly bedded | > 100 cm |

- 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

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).

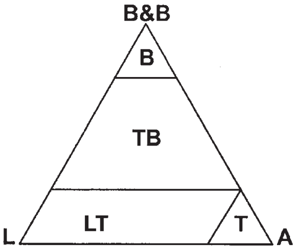


***Figure 1.25.*** Grain-size ternary diagram for naming primary volcaniclastic rocks based on Fisher's (1961) classification (from White and Houghton 2006: p. 679, fig. 1). B&B – blocks and bombs, L – lapilli, A – ash, B – breccia, TB – tuff breccia, LT – lapilli tuff, T – tuff. Blocks are defined as 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%

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).

| 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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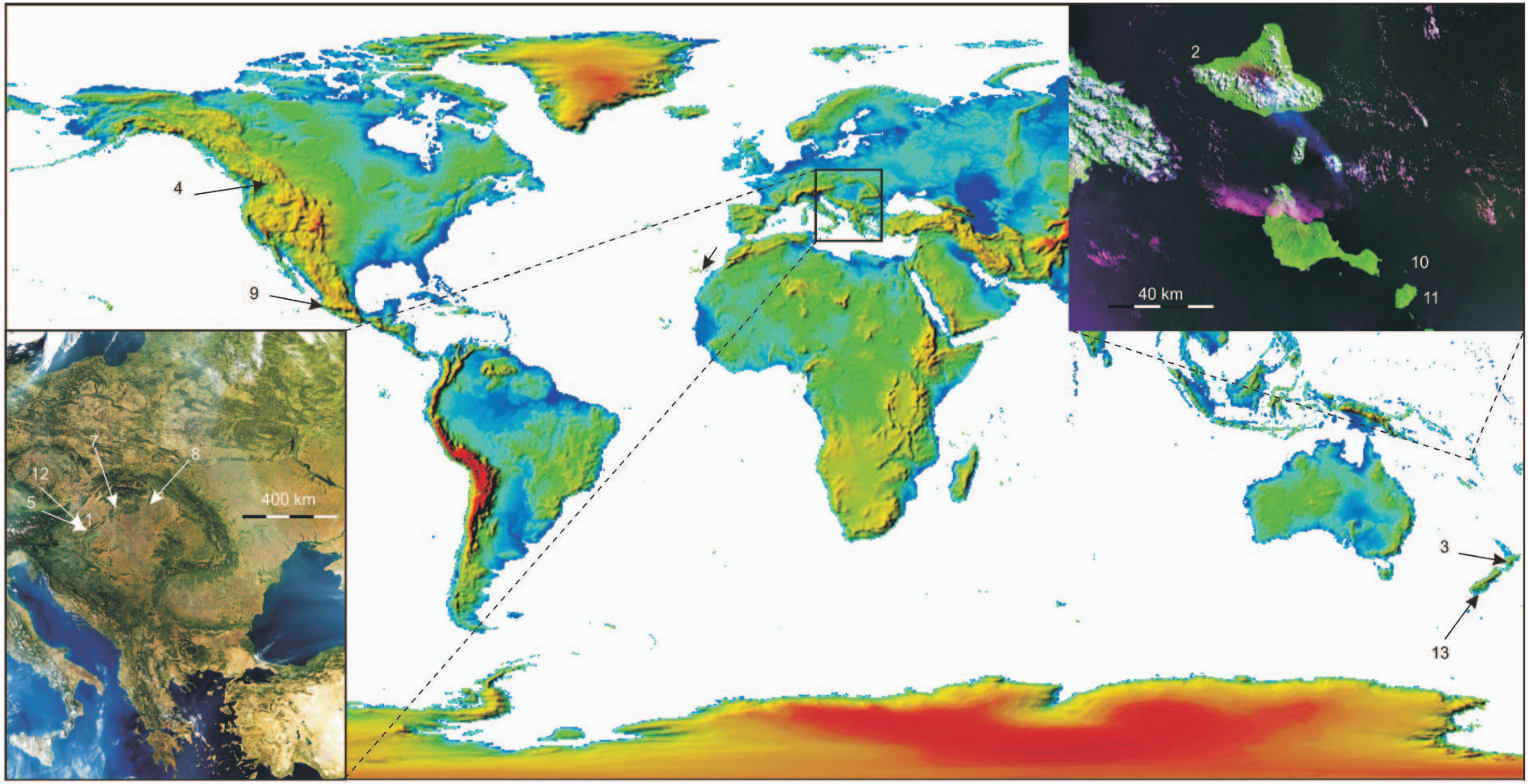
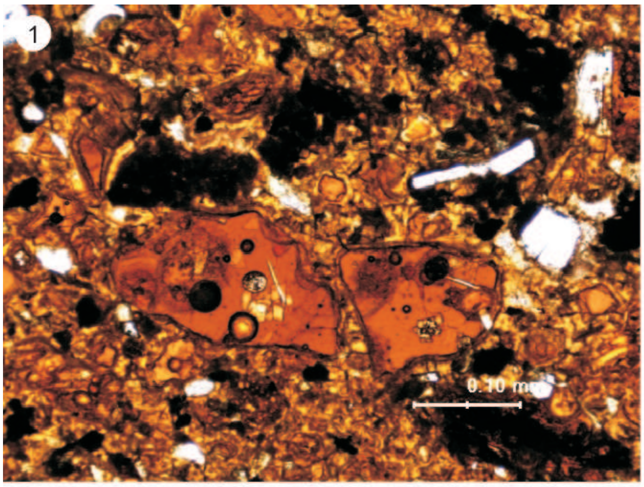
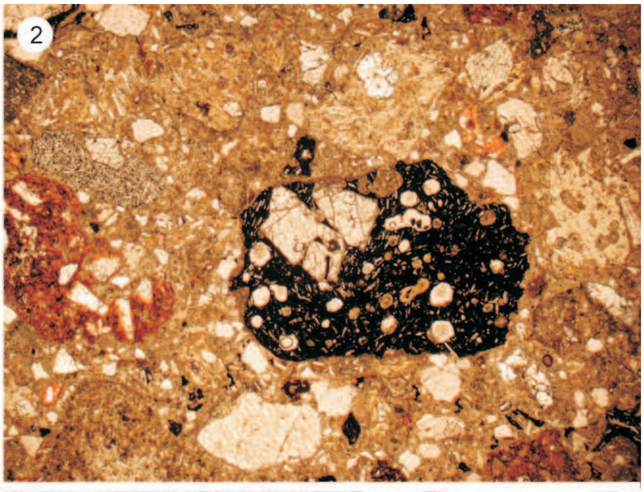
## Location map

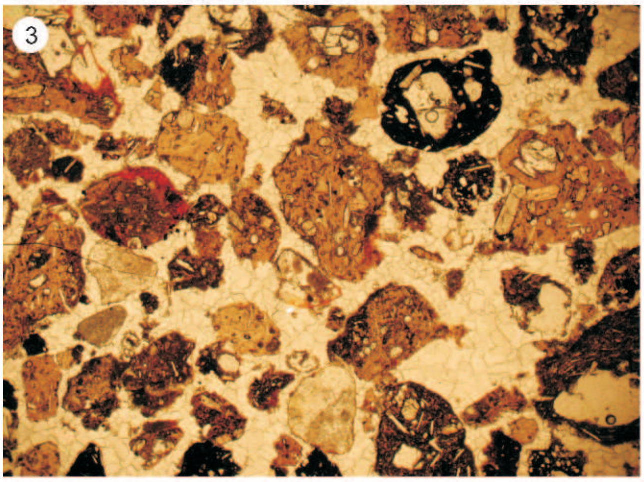


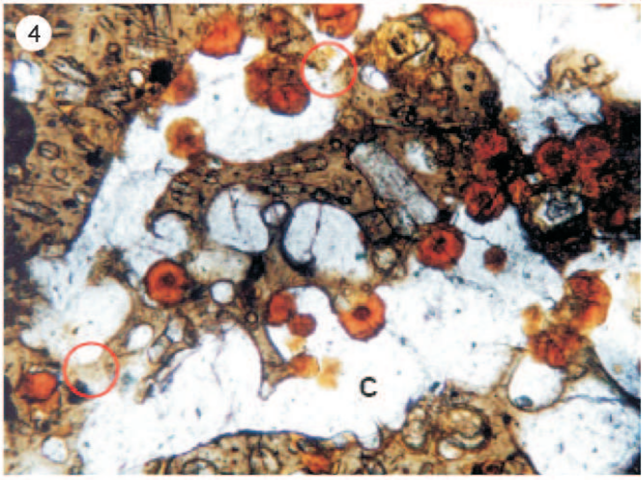
1 — Bakony – Balaton Highland Volcanic Field, Hungary
2 — West Ambrym, Vanuatu
3 — Taupo Volcanic Zone, New Zealand
4 — Sinker Butte, Idaho, USA
5 — Hegyesd diatreme, Hungary
6 — Waipiata Volcanic Field, New Zealand
7 — Pilis Mts, Hungary
8 — Tokaj Mts, Hungary
9 — Ixtlan del Rio, Mexico
10 — Laika Island,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

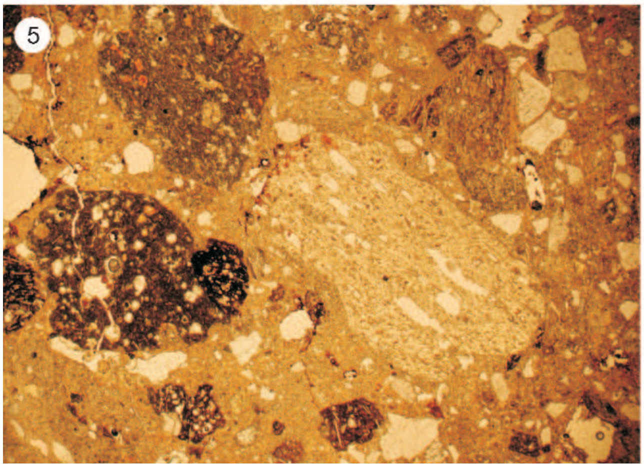
# Plate I

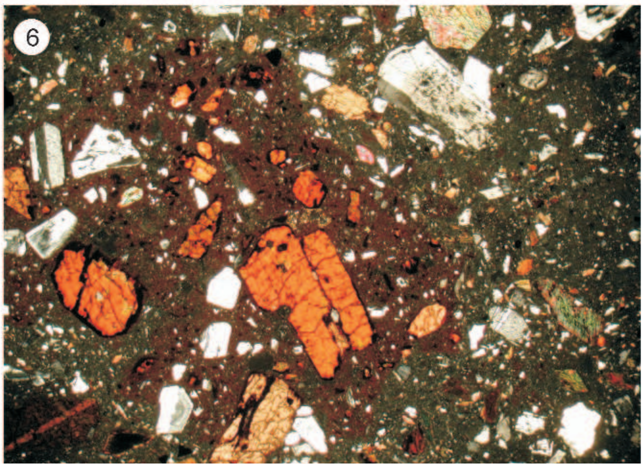






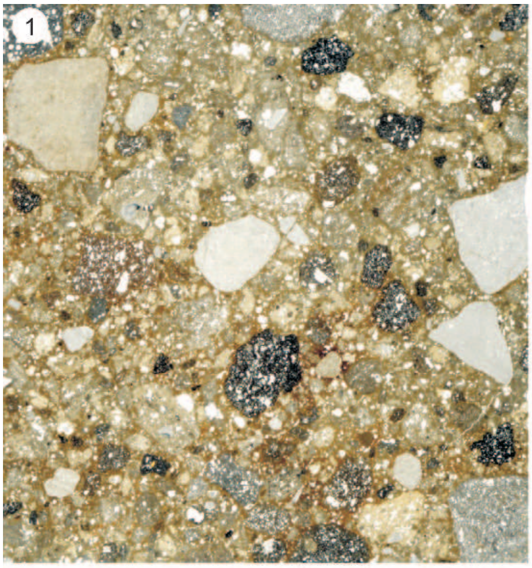


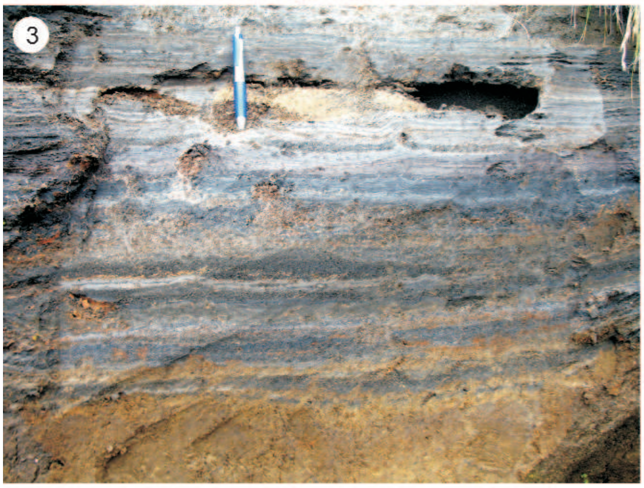




1. Blocky sideromelane glass shards (brown fragment) as juvenile fragments from a phreatomagmatic lapilli tuff of the Sinker Butte, Western Snake River Volcanic Field, Idaho, USA.
2. Moderately vesicular tachylite lapilli in a phreatomagmatic lapilli tuff from the Pliocene Bakony – Balaton Highland Volcanic 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





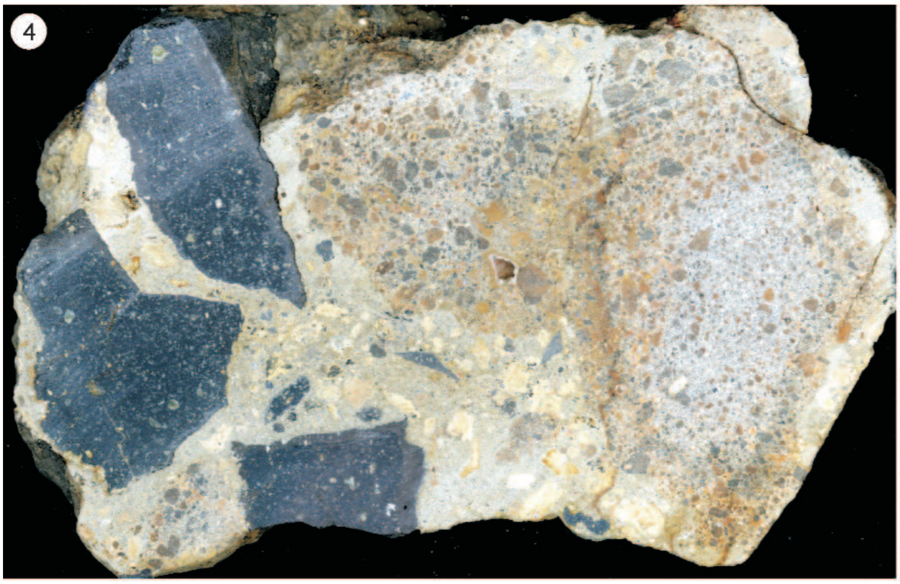
1. Accidental lithic-rich lapilli tuff from the Véndeg-hegy diatreme. In this sample the majority of the accidental lithics are Triassic limestones (pale-grey angular fragments) excavated by the phreatomagmatic explosions responsible for the formation of the diatreme. The side of the picture is 4 cm 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.

# Plate III

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 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).

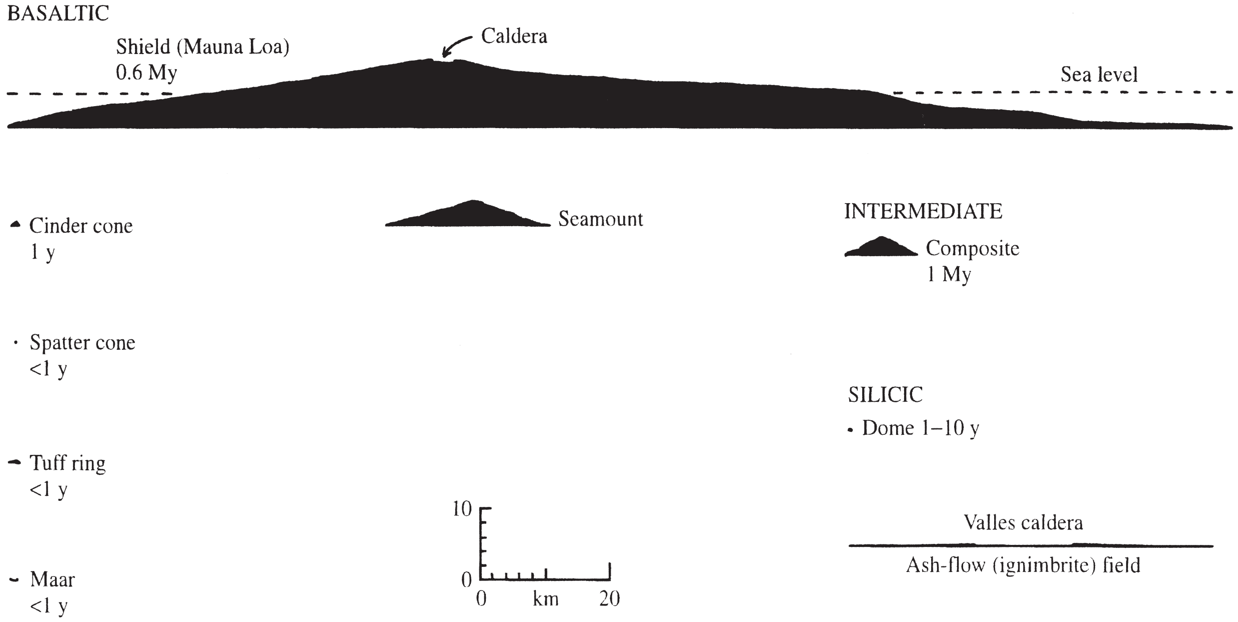


***Figure 2.1.*** Comparative diagram of different volcanic landforms (after Best 2003: p. 246, fig. 10.6)

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 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.

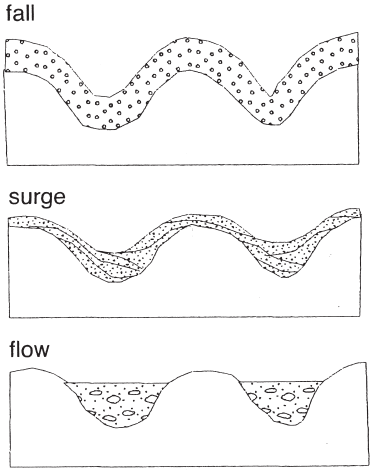


*Figure 2.2.* Schematic diagram of typical depositional features of pyroclastic fall, surge, and flow after WILSON and HOUGHTON 2000: p. 550, fig. 5

## 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.

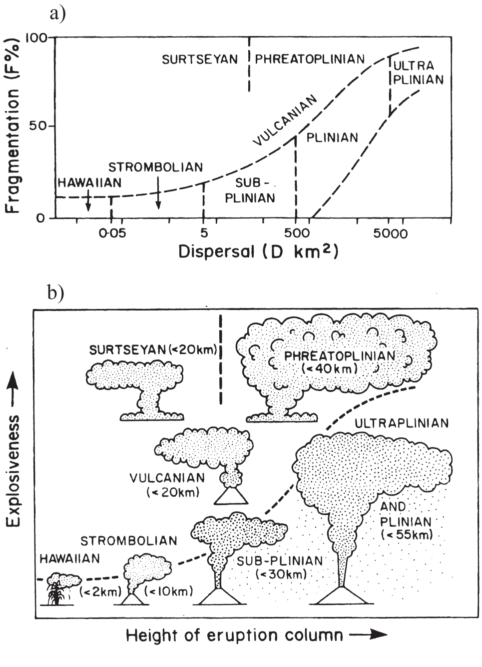


*Figure 2.3.* a) Dispersal (D) and Fragmentation (F) diagram showing different types of eruptions after WALKER 1973 classification scheme. b) a summary diagram explaining the differences between various types of volcanic eruptions. Diagram redrawn after CAS and WRIGHT 1988: p. 130

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.

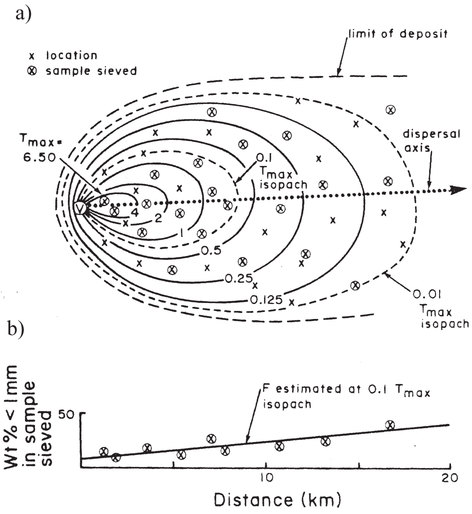


***Figure 2.4.*** Representative diagram used to obtain Dispersion (D) and Fragmentation (F) data
a) map view of distribution of a single pyroclastic fall unit, b) diagram generated along the axis of the dispersal of the pyroclastic fall unit based on the sieved tephra samples (Cas and Wright 1988: p.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

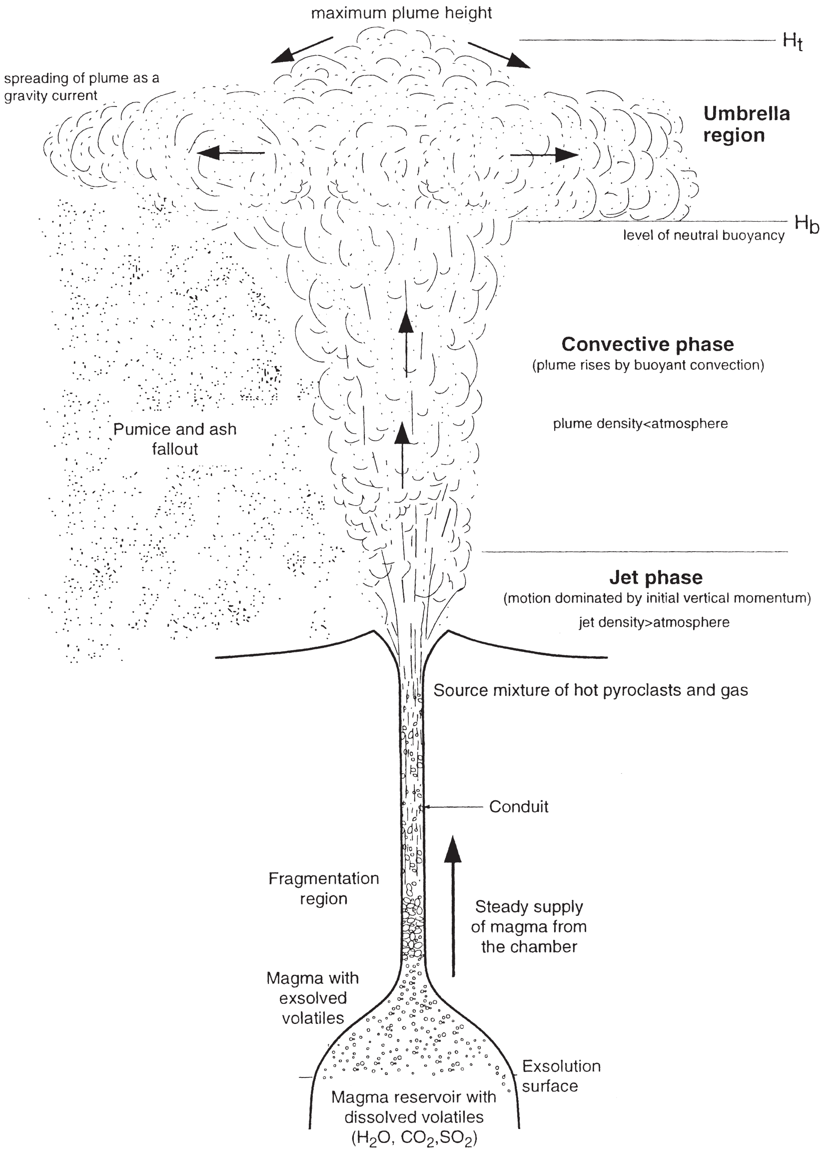


***Figure 2.5.*** Parts of a volcanic plume generated by high silica magma eruption (after Carey and Bursik 2000: p. 529, fig. 1)

### *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.

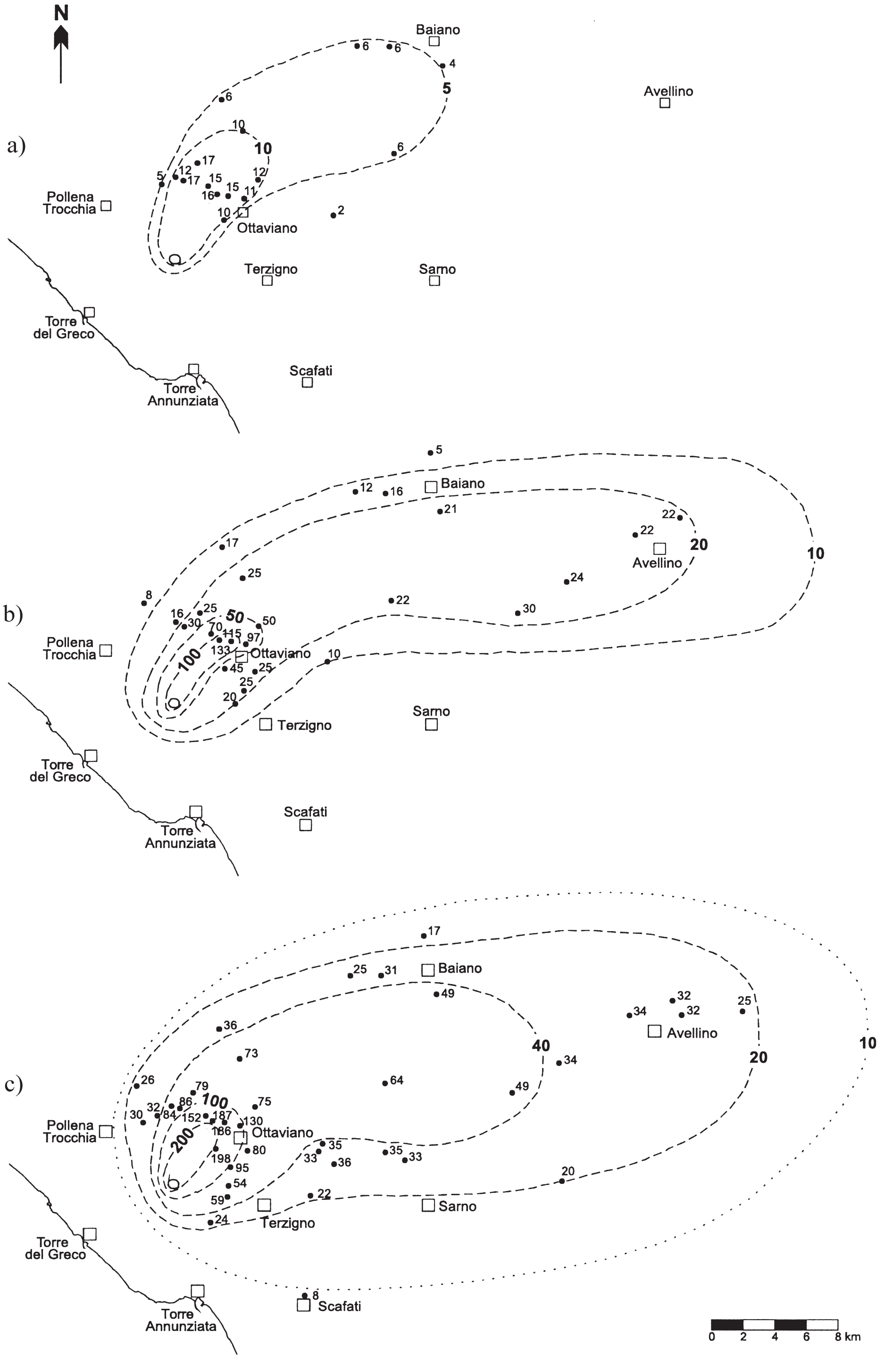


***Figure 2.6.*** Plinian tephra dispersal map of the AD 472 eruption of the Somma volcano, Vesuvio, Italy after Rolandi et al. 2004: p. 304, fig. 8
Individual maps (a, b, and c) show the different distribution pattern of 3 individual fall unit associated with this eruption caused by slight differences/changes in the eruption and the prevailing wind direction

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.

### *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.

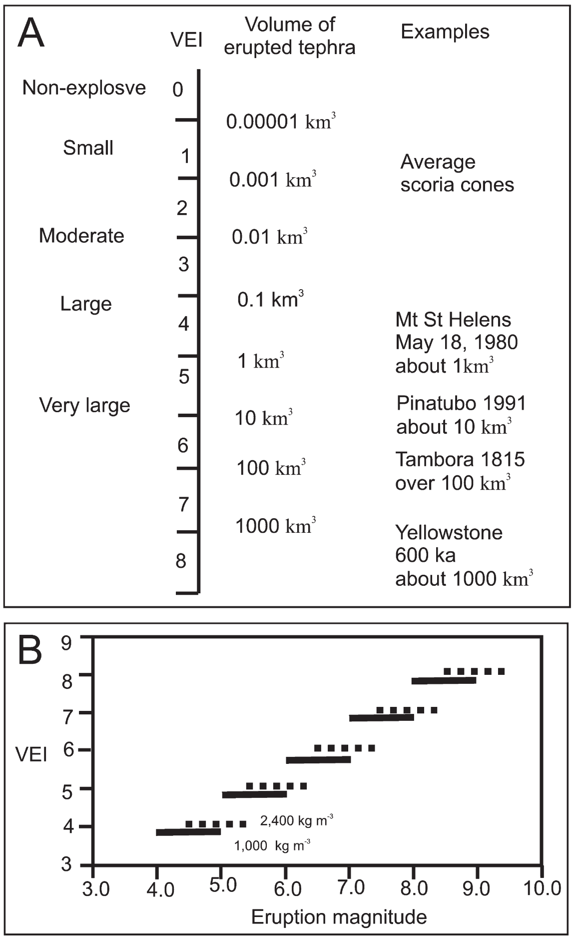


***Figure 2.7.*** A) Relationship between VEI and erupted volume of tephra after NEWHALL and SELF (1982). B) The relationship between eruption magnitude, M, and Volcanic Explosivity Index, VEI, for deposits of bulk density 1,000 and 2,400 kg/m$^3$ (redrawn after MASON et al. 2004: pp. 735–748, p. 736,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

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 –

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).

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

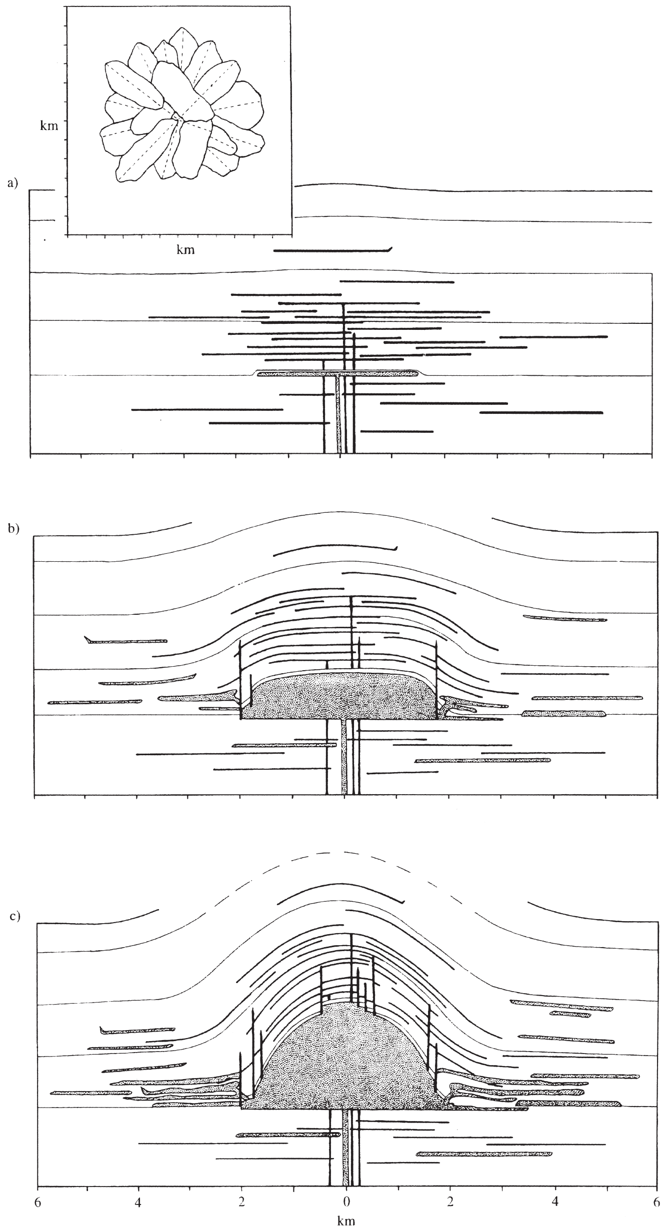


***Figure 2.8.*** Laccolith architecture and growth modelled on the basis of laccolithes of the Henry Mts, Utah (after Jackson and Pollard 1988)

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

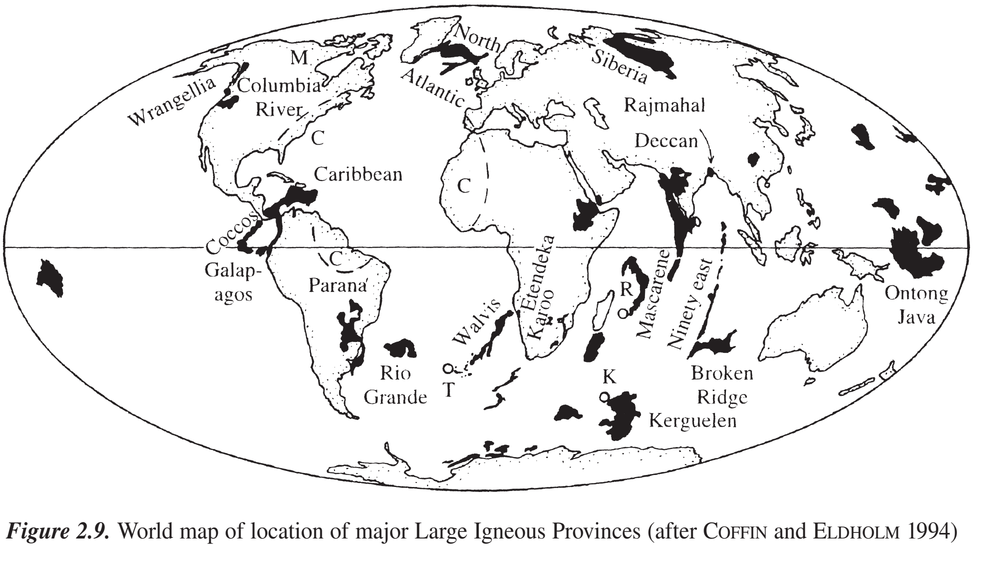


***Figure 2.9.*** World map of location of major Large Igneous Provinces (after Coffin and Eldholm 1994)

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-

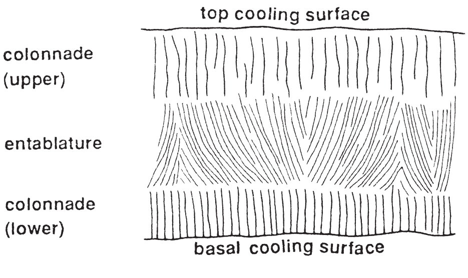


***Figure 2.10.*** Columnar joint patterns of a basaltic lava flow according to SPRY (1962)

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.

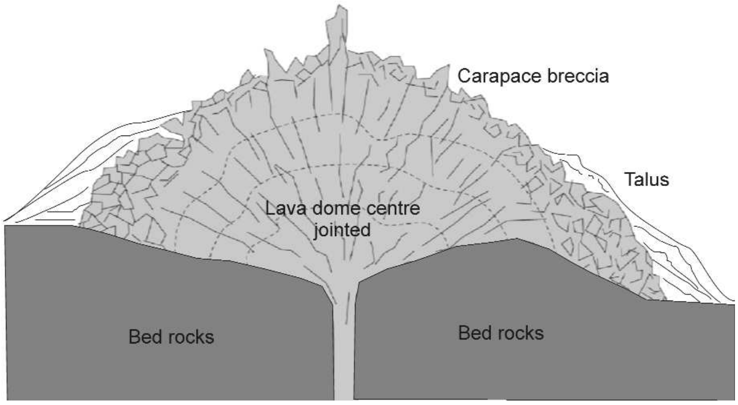


*Figure 2.11.* Dome cross section diagram. The central part of the lava dome is massive and commonly rosette-like jointed. Strong hydrothermal activity occurs in this part of the dome. The margin of the dome composed of angularly fractured coherent lava fragments, commonly altered and/or mineralized 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

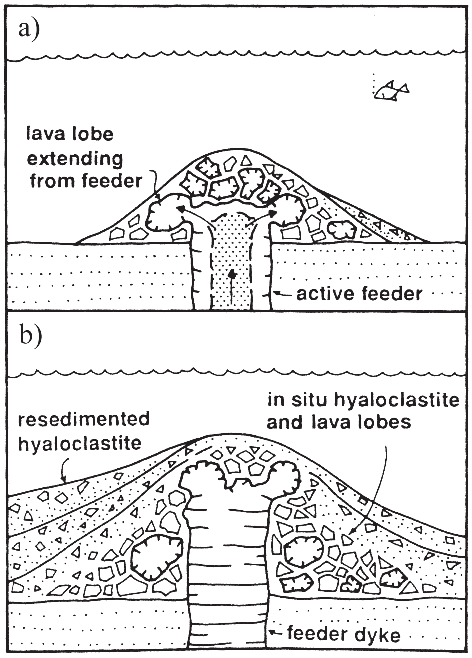


***Figure 2.13.*** Subaqueous lava dome facies after Yamagishi (1987)
a) lava extruded into the sea/lake floor. Lava advance only short distance before it is quenched and fragmented, b) the growing hyaloclastite pile is intruded by further feeder dykes producing more in situe hyaloclastite. Gravitational 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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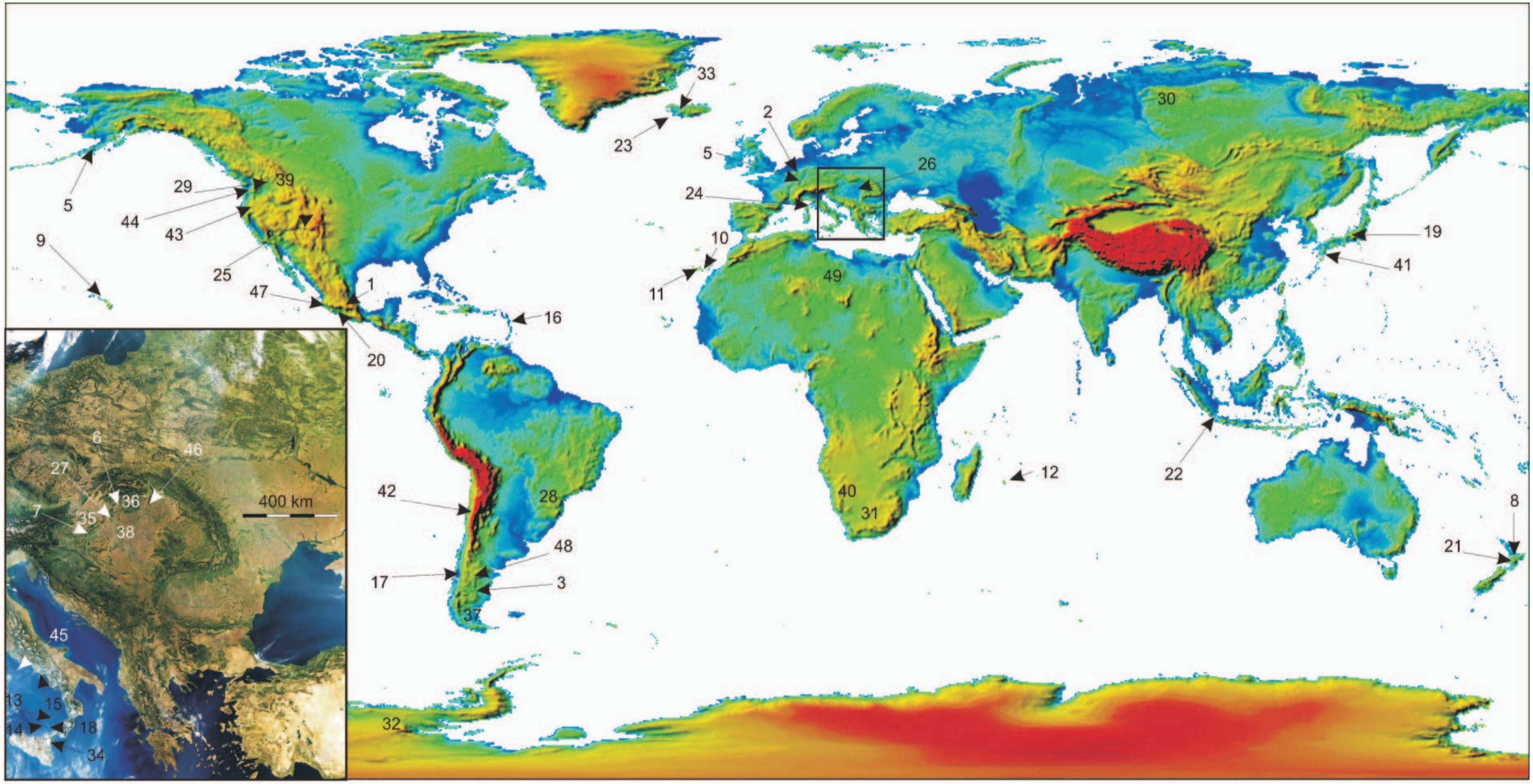
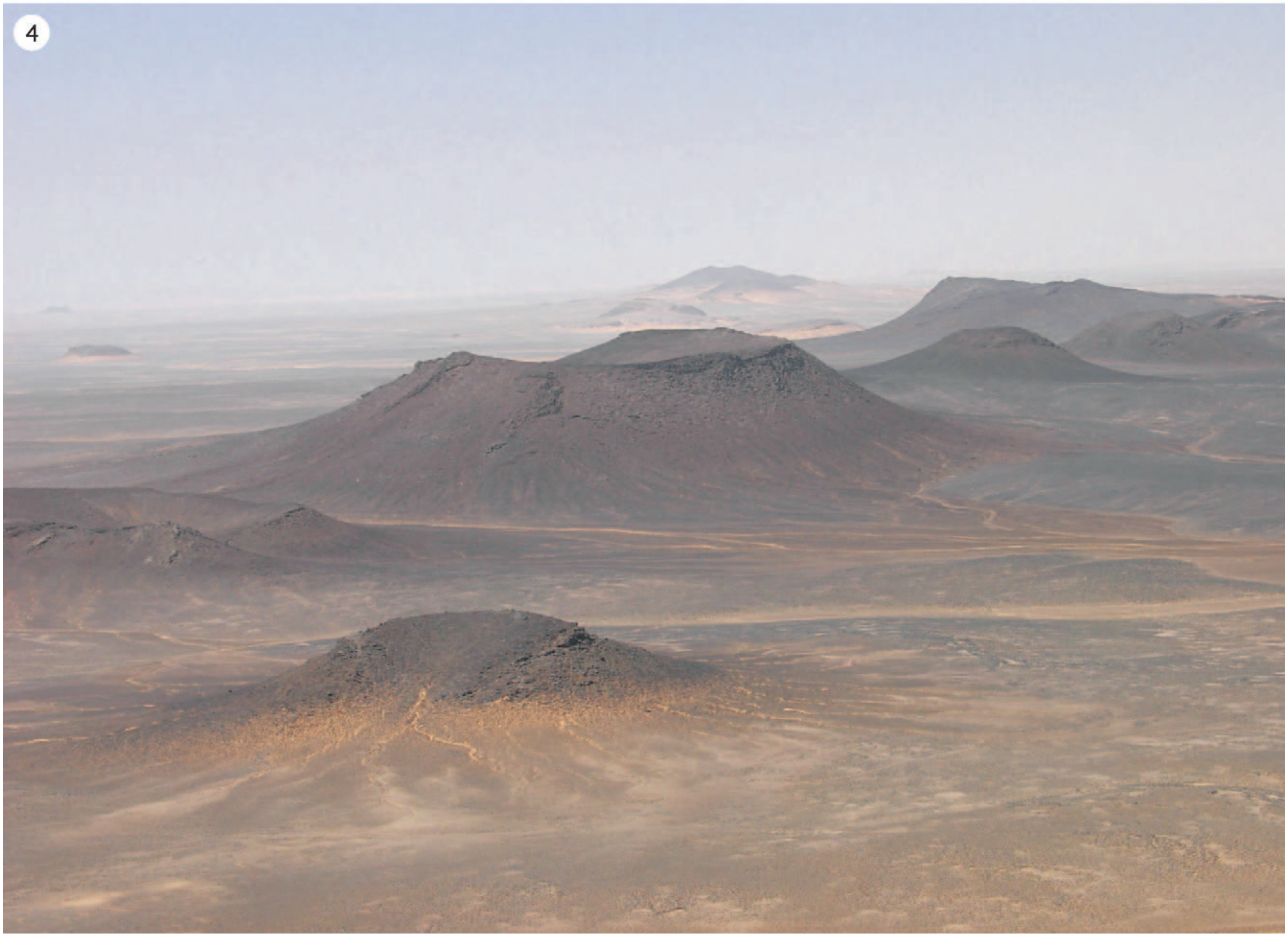
## Location map



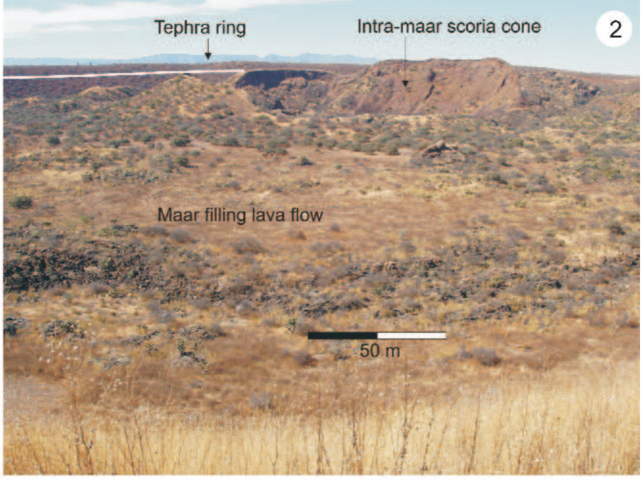
1 — Paricutin, Mexico
2 — Eifel, Germany
3 — Patagonia, Argentina
4 — Ukrinek, Alaska, USA
5 — Messel maar, Germany
6 — Hajnacka maar, Hungary
7 — Pula maar, Hungary
8 — Taupo Volcanic Zone, New Zealand
9 — Hawaii, USA
10 — Gran Canaria, Spain
11 — Tenerife, Spain
12 — Piton la Fournaise, Reunion France
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

# Plate I

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.

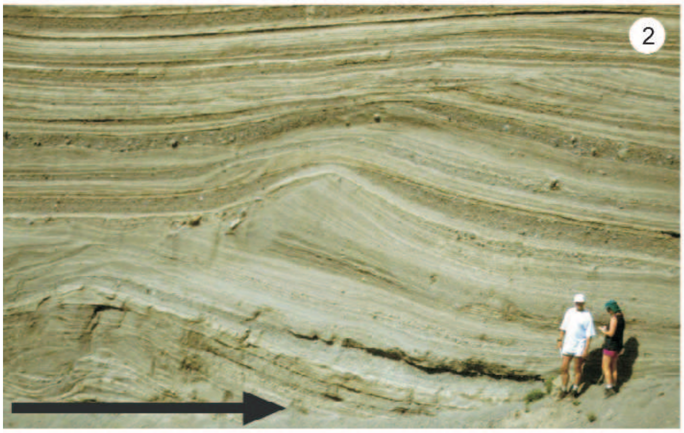


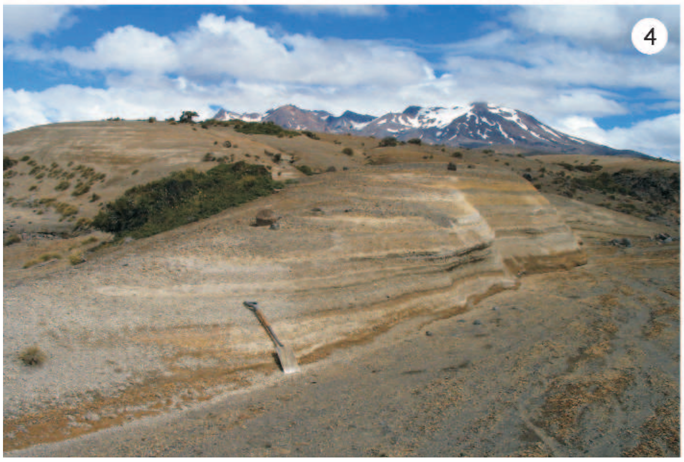
# Plate II



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.

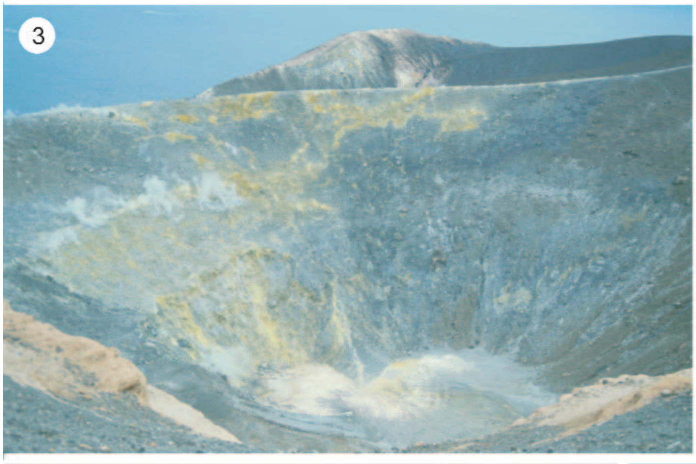
# Plate III





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

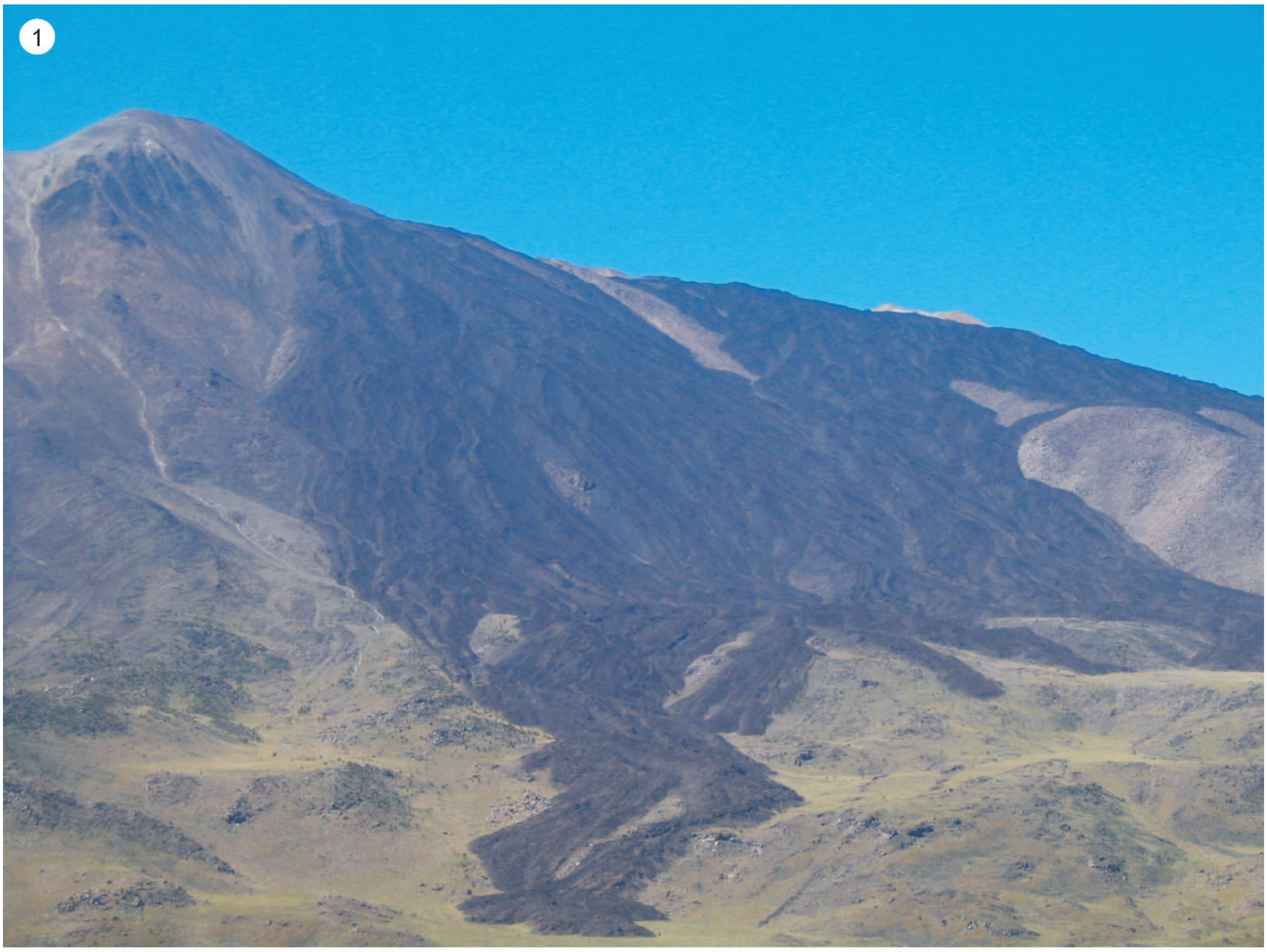


1. Lava spatter eruption from the central vent of the Villarica Volcano from Central Chile.
2. Strombolian explosion in night time at Stromboli, Southern Italy (photo by S. J. Cronin).
3. Wide, dish-like crater of Vulcano, Southern Italy. Note the steep inner crater wall as well as the radial erosional 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.

# Plate V

1. Typical ring plain succession of a major stratovolcano 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.

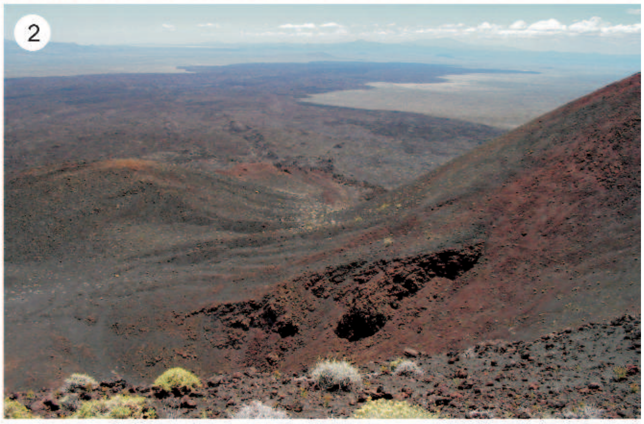
## Plate VI



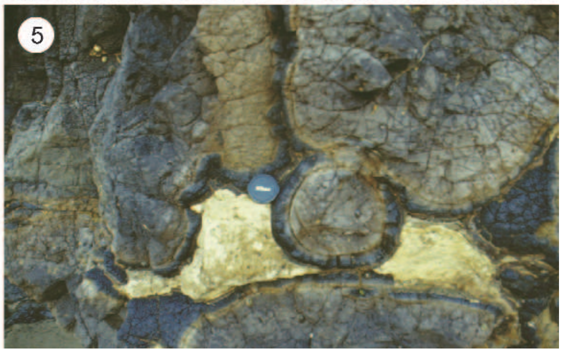
1. ‘Aa’ lava channels on the flank of Tromen Volcano, Argentina.
2. Active ‘aa’ lava flow during the Etna 1993 eruption (A). On a close up view the collapse of hot lava block is clearly visible (B).
3. Well-developed columnar joints in the basanite plug of Hegyestű, Hungary.

# Plate VII

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.



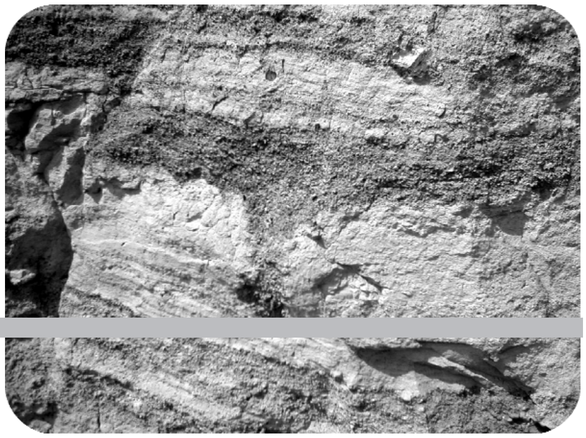
## Plate VIII



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.

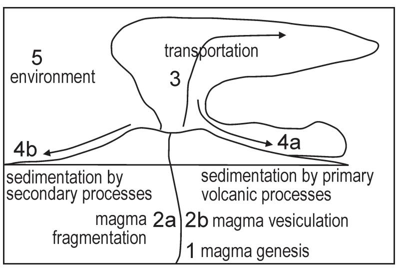


*Figure 3.1.* Diagram about volcanic processes from magma generation to tephra redeposition. The resulting volcanic rocks will carry characteristic textural features indicative to the volcanic history of the rock from magma to the rock

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 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

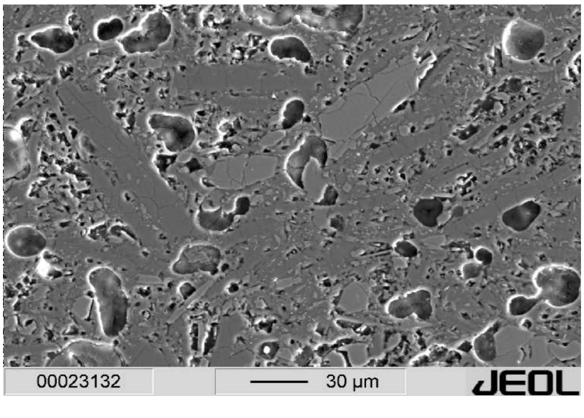


*Figure 3.2.* SEM studies of volcanic glass shard can help to establish the fragmentation history of the melt. Blocky, fractured glass shards are characteristic for magma water interaction driven explosive fragmentation. An SEM picture of volcanic glass shard from the Volcan Cerro Colorado tuff cone 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-

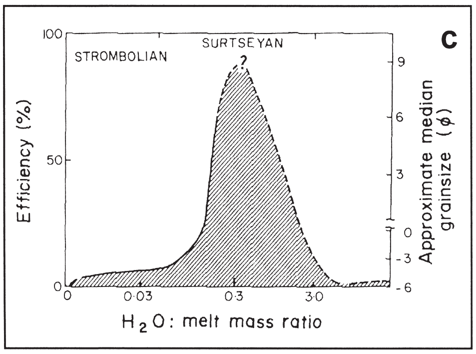


*Figure 3.3.* Diagram of the relationship between magma-water mass ratios and the fragmentation efficiency after Wohletz and Sheridan 1983: pp. 385–413

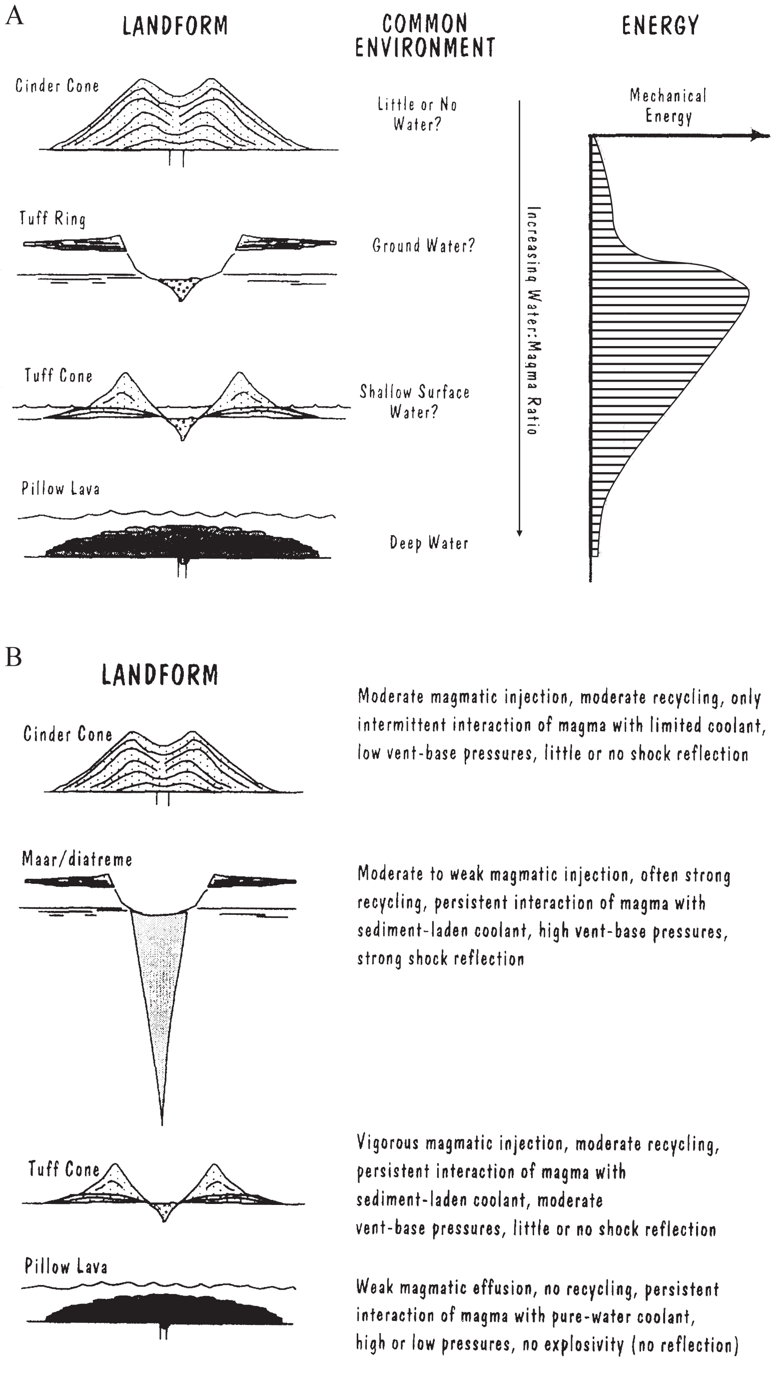


*Figure 3.4.* Comparative diagram between magma to water ratio mechanical energy and resulting volcanic landforms using pure coolant (water) and impure coolant (muddy slurry) as an interactive media with magma. After White 1996: p. 158, fig. 1 and p. 166, fig. 8

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

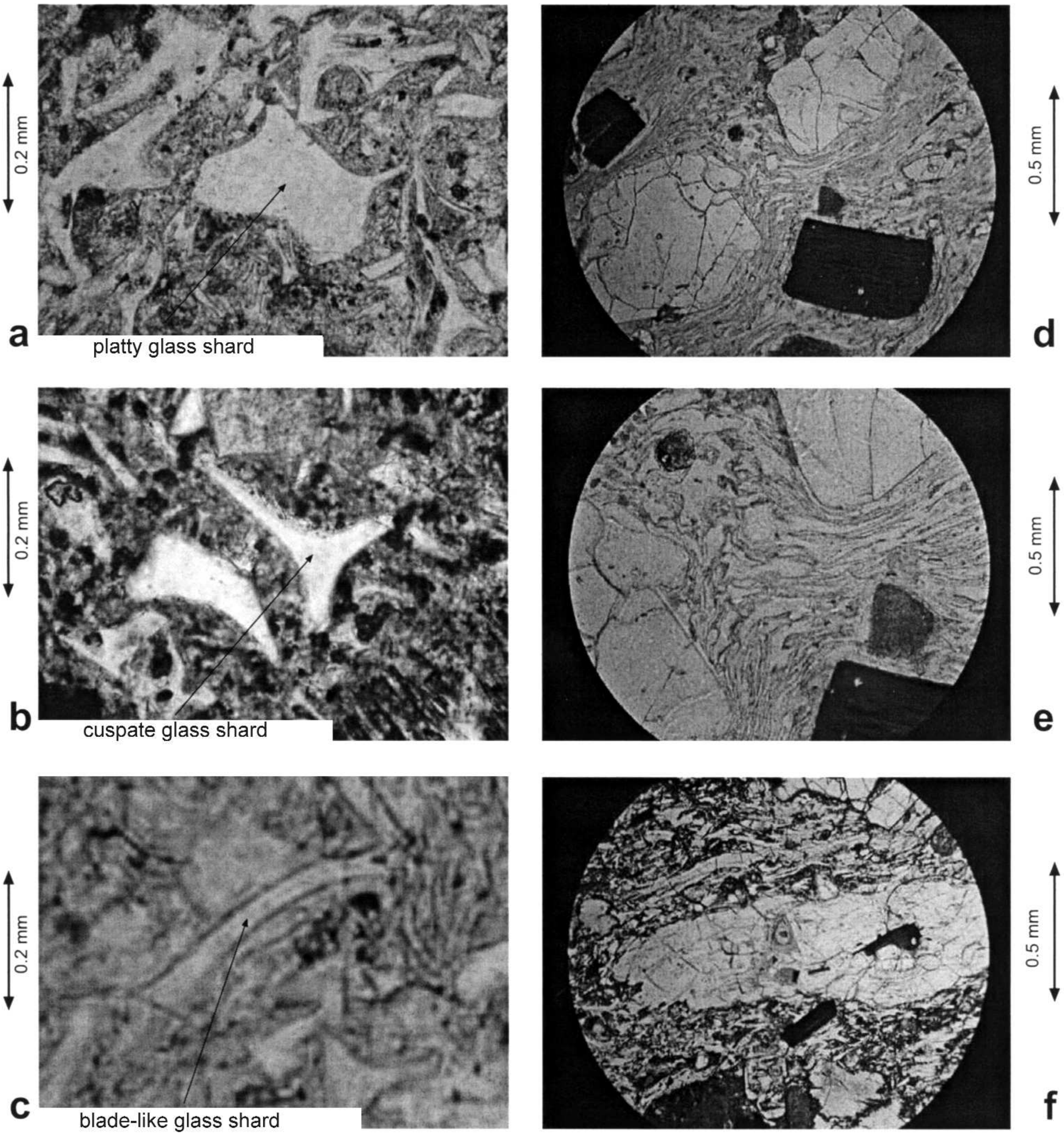


***Figure 3.5.*** Silicic glass shards shards from pumiceous pyroclastic flow deposits. The view is about 4 mm across

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.

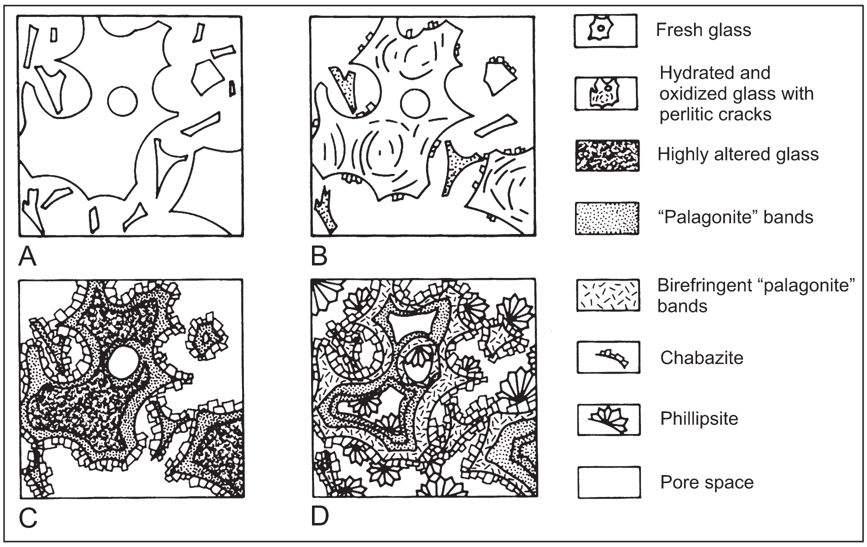


*Figure 3.6.* Step of palagonitisation of a volcanic glass shard after Fisher and Schmincke (1984)

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.

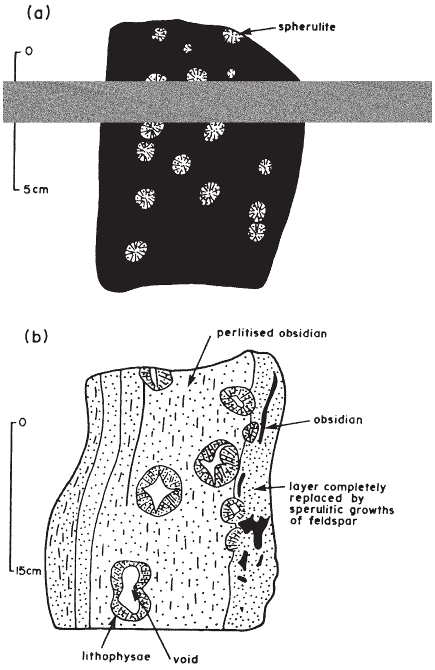


*Figure 3.7.* Textural development of obsidian lava flows from spherulite to lithophysae development according to Cas and Wright 1988

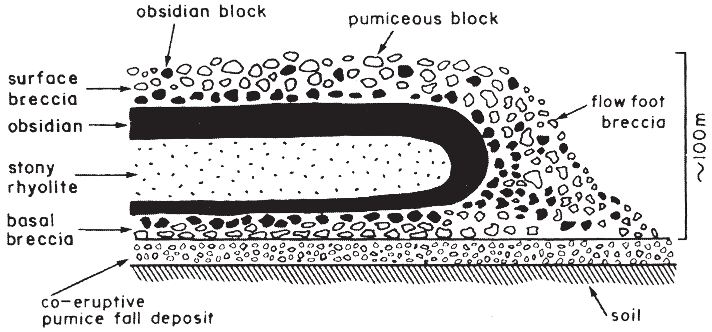


*Figure 3.8.* Theoretical rhyolite lava flow cross section according to Cas and Wright 1988

### 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 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).

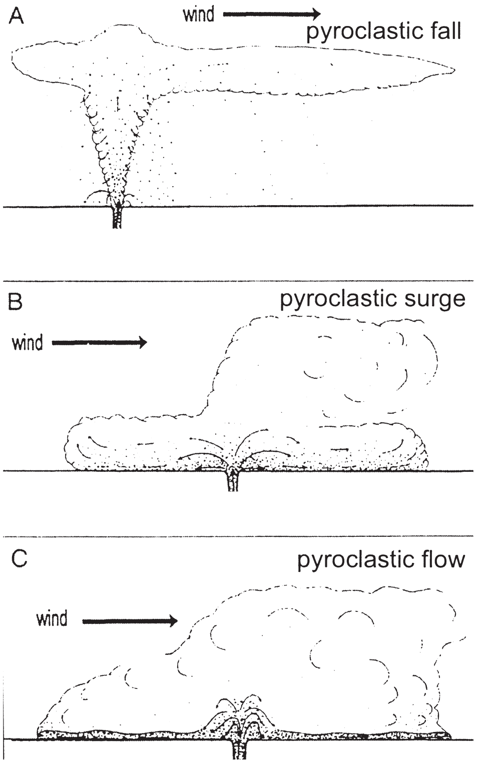


***Figure 3.9.*** Pyroclastic fall, flow and surge formation and the associated eruption clouds according to Wilson and Houghton 2000: p. 547, Figure 1

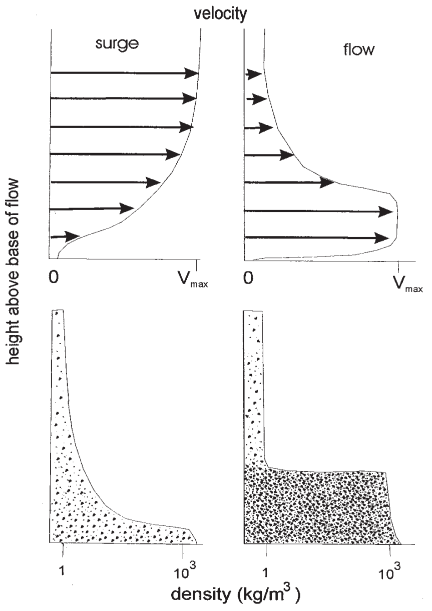


***Figure 3.10.*** Density profile of dilute (surge) and concentrated (flow) pyroclastic density currents according to Wilson and Houghton 2000: p. 548, fig. 2

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).

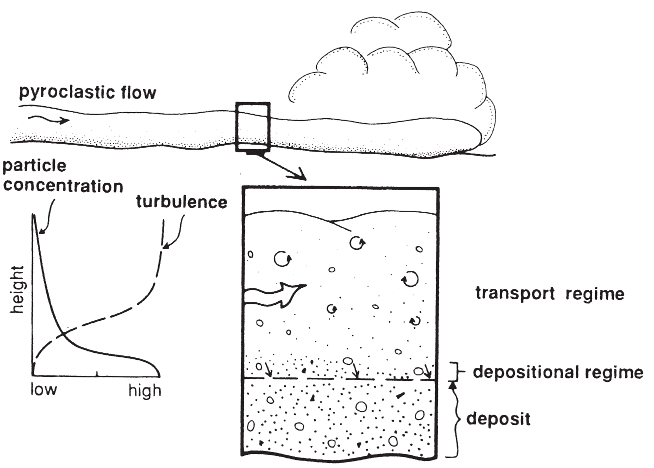


***Figure 3.11.*** Progressive aggradation model for pyroclastic flow evolution according to Branney and Kokelaar (1992) and Druitt (1992) ideas

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).

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 veneer facies (Fisher et al. 1983, Bogaard and Schmincke 1984, Freundt and Schmincke 1985, Németh and Martin 1999).

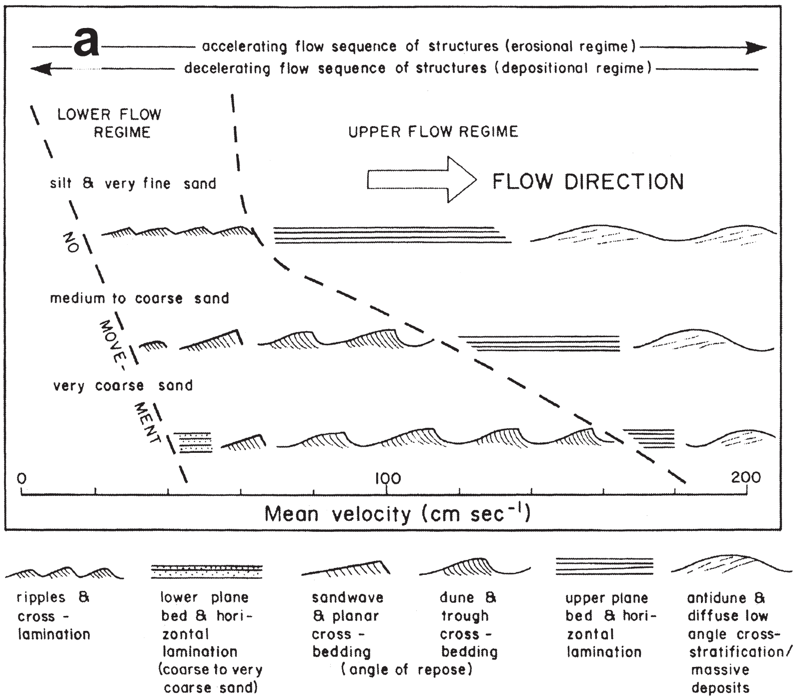


*Figure 3.12.* Relationship between flow regime and resulting bed forms in horizontal moving sedimentary currents (after Cas and Wright 1988)

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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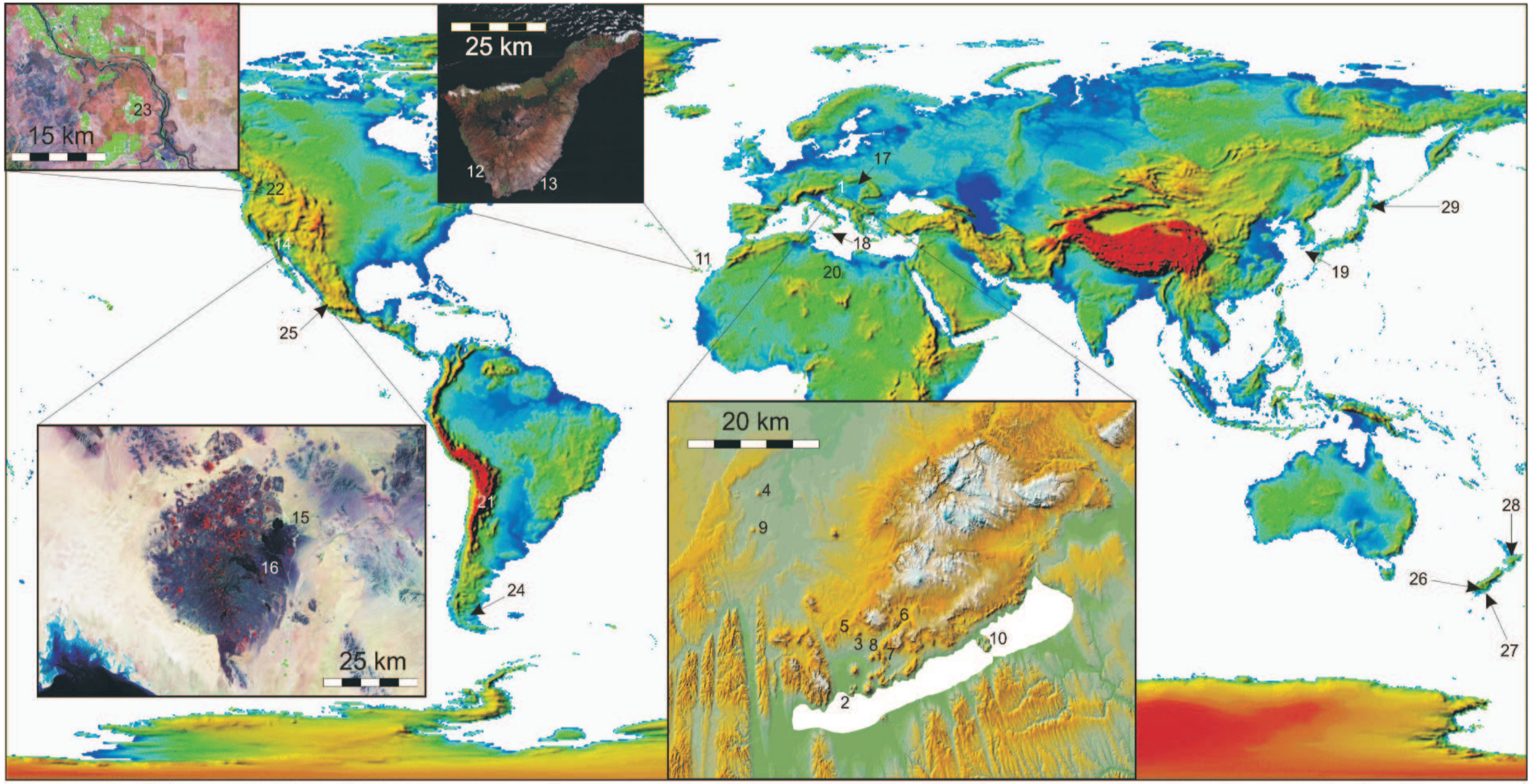
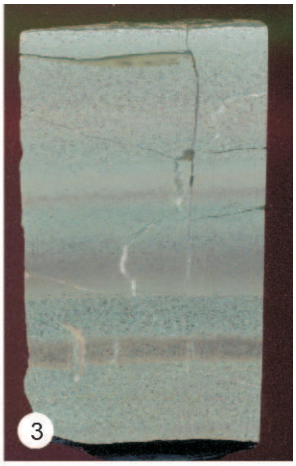
## Location map

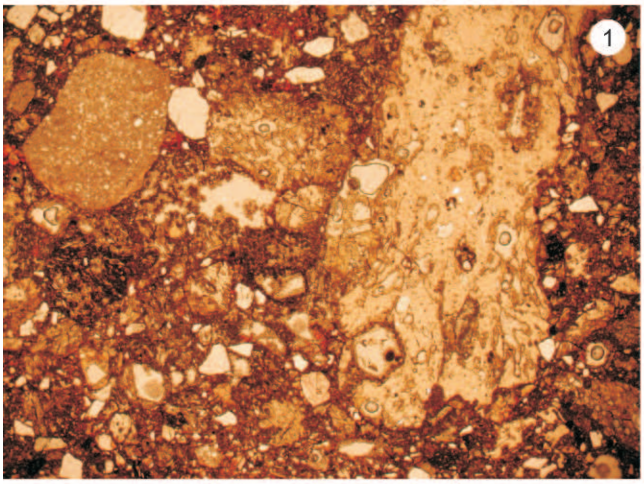


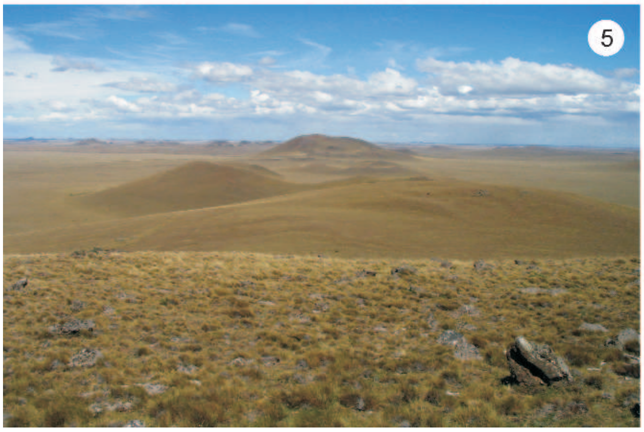
1 — Western Hungary
2 — Szigliget, Hungary
3 — Haláp, Hungary
4 — Ság-hegy, Hungary
5 — Véndeg-hegy, Hungary
6 — Pula, Hungary
7 — Szentbékkálla, Hungary
8 — Hajagos, Hungary
9 — Kissomlyó, Hungary
10 — Tihany, Hungary
11 — Tenerife, Spain
12 — Caldera del Rey maar, Spain
13 — Montana Pelada, Spain
14 — 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

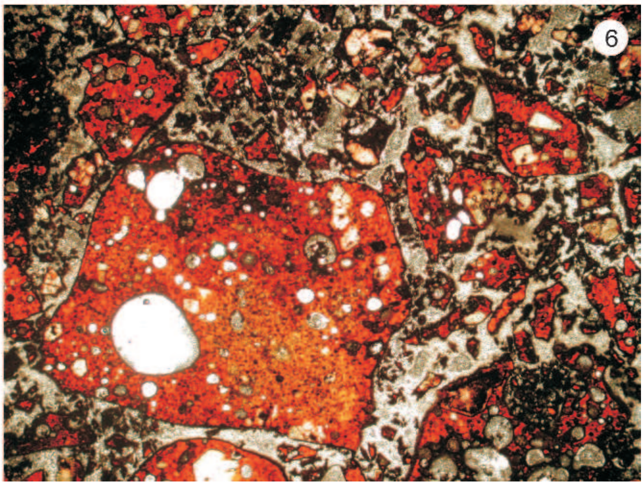
# Plate I

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.

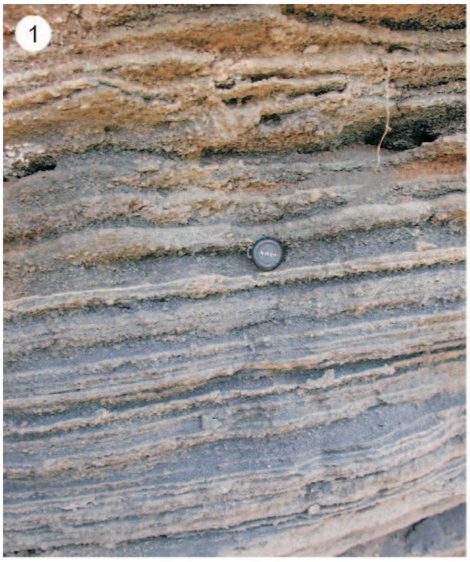


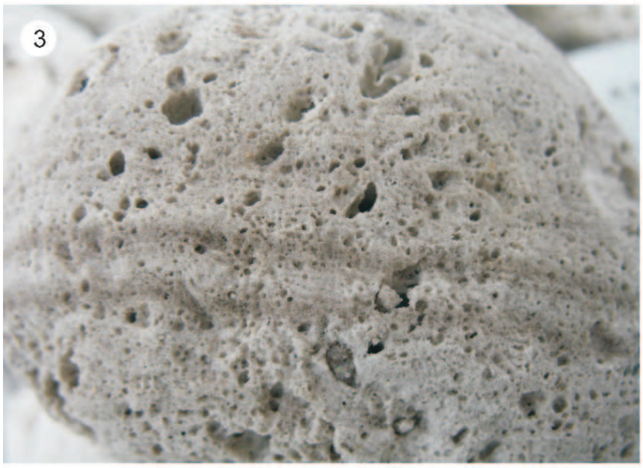


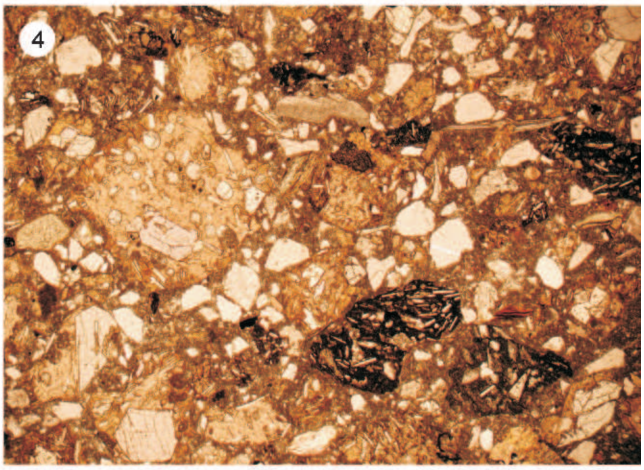


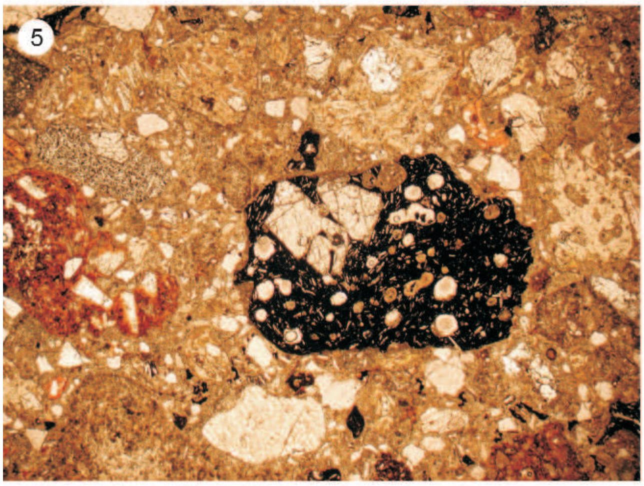


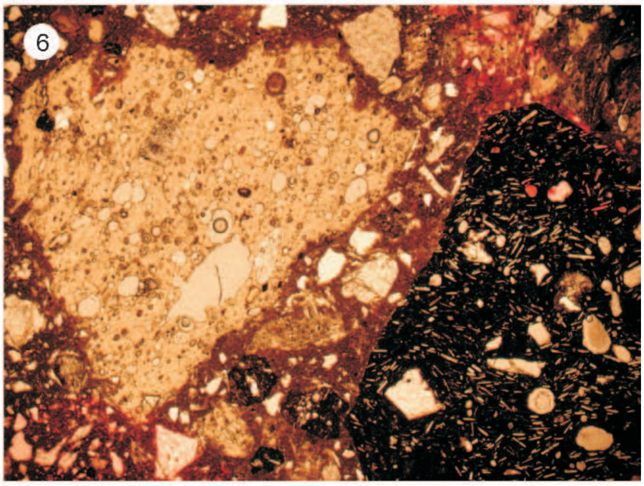
# Plate II





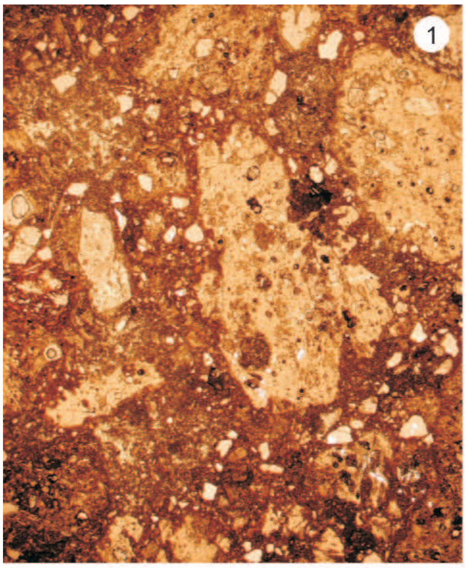


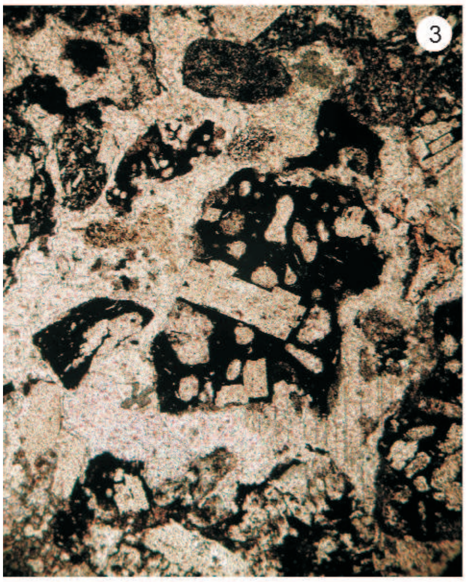


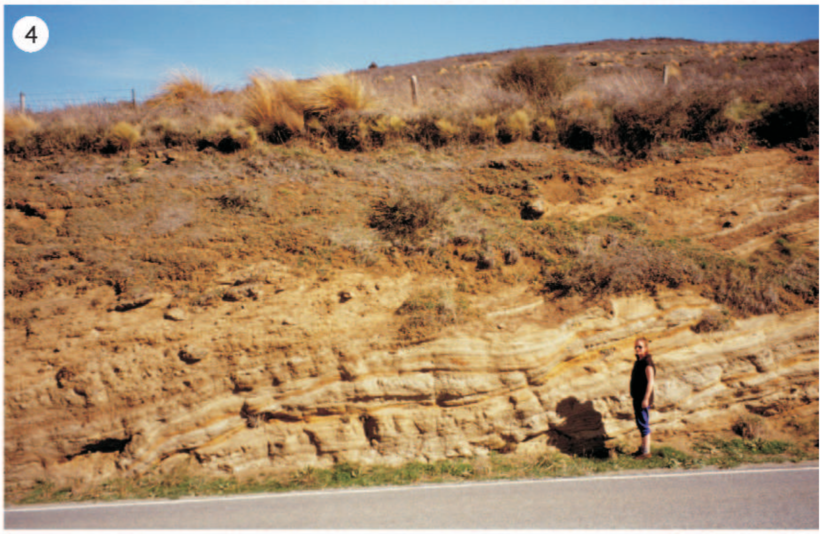


1. Black scoriaceous lapilli beds from a scoria cone near Ceboruco volcano in Western Mexico.
2. Pumiceous Plinian air fall beds forming a great succession in the Altiplano, Northern Chile.
3. Block of banded andesite-dacite pumice erupted on May 22, 1915. This initially large block of hot pumice fell 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

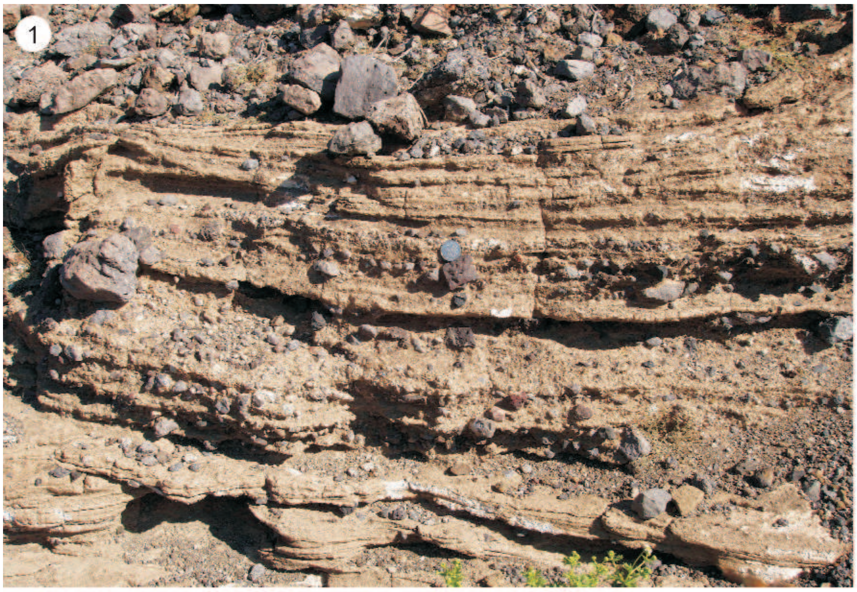




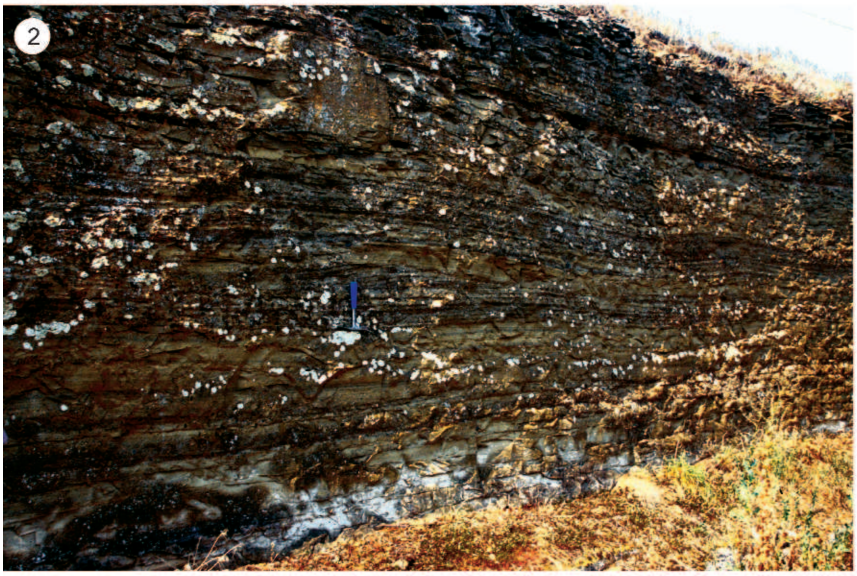


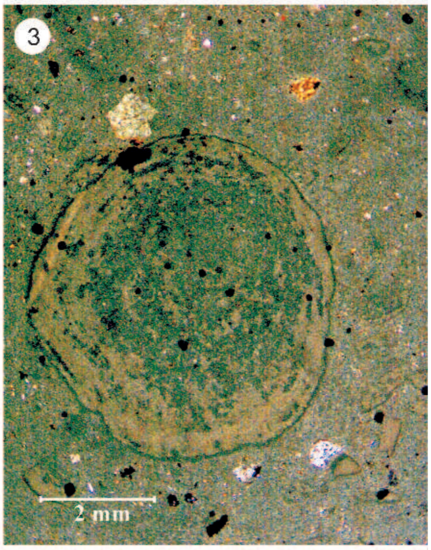
1. Predominantly sideromelane glass shards dominate tephras formed by shallow level fragmentation and/or fragmentation through open vent of melt by magma-water interaction such as in this picture taken from a lapilli tuff of the Szigliget diatreme, Hungary. The shorter side of the view is appx. 4 mm.
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.

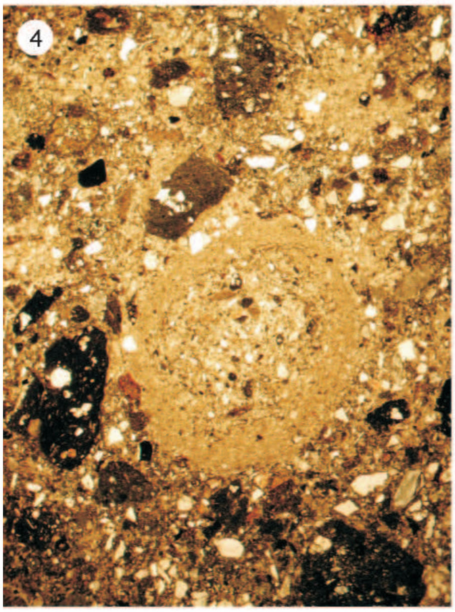
# Plate IV

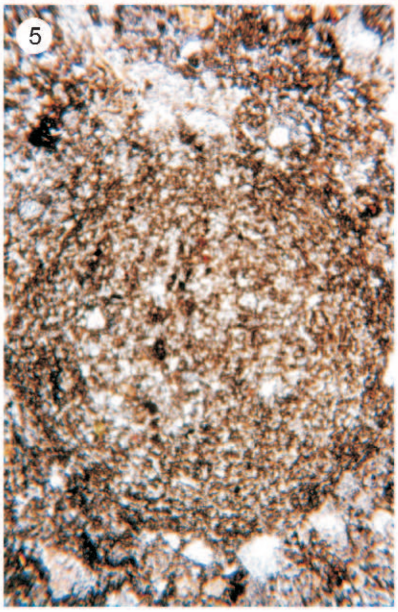


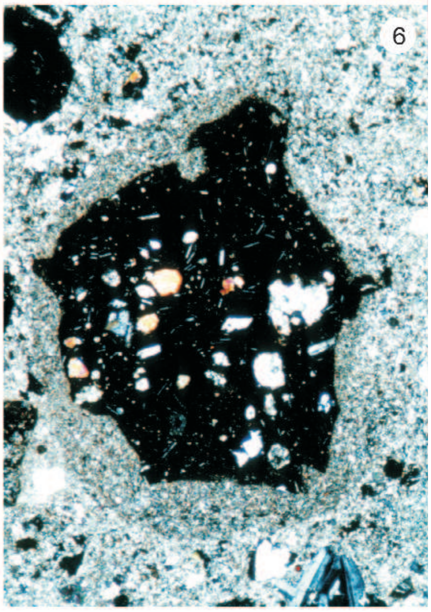
1. Accidental lithic rich surge bed succession from near vent volcanic facies of the Crater Elegante maar from Sonora, Mexico.
2. Base surge beds developed around a tuff ring erupted through a soft substrate (mud and sand) country rock succession in the Little Hungarian Plain.
3. Accretionary lapilli 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.



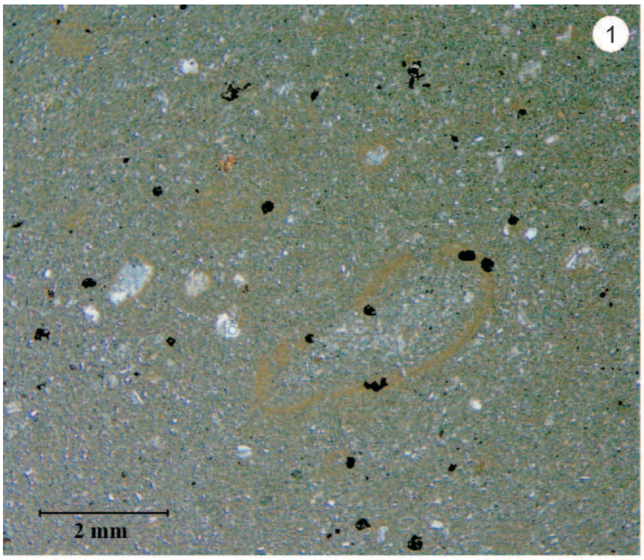


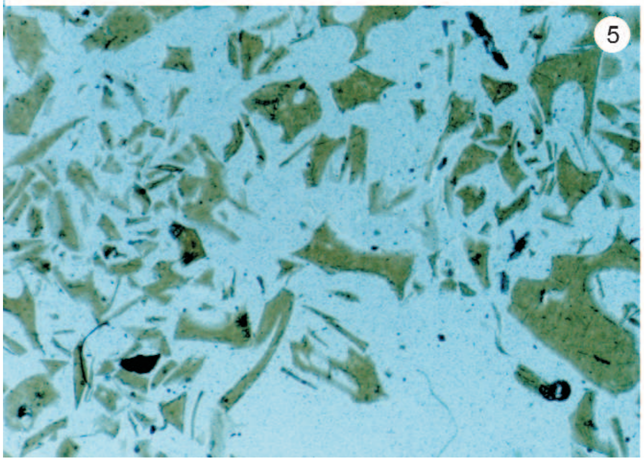






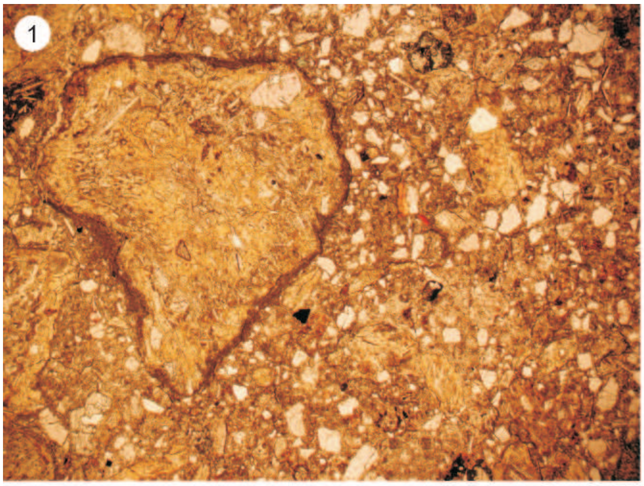
## Plate V

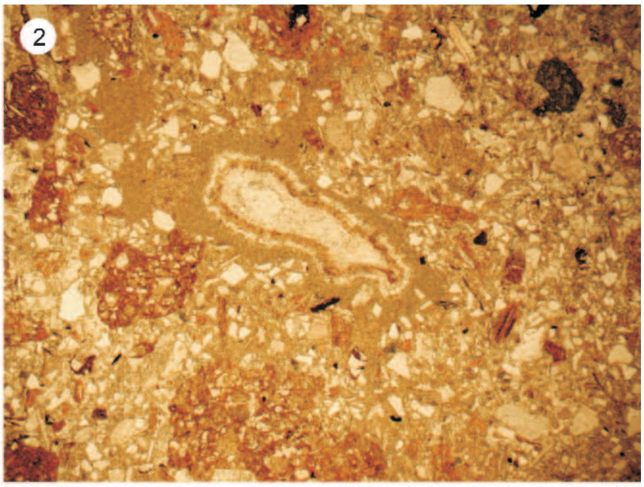


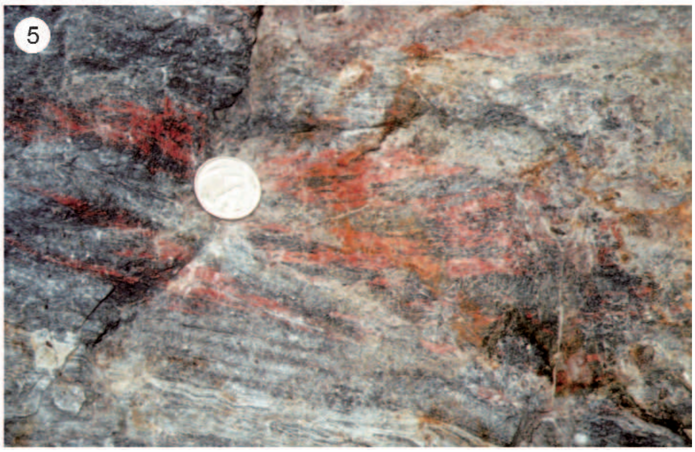


1. Mud clots from tuff bed of the Montana Pelada tuff ring in southern Tenerife. The fine mud aggregates surrounded by thin limonite-rich film layer.
2. Large accretionary lapilli from the phreatoplinian succession of the AD 180 Taupo eruption, New Zealand.
3. Mud chunks from the phratomagmatic succession 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

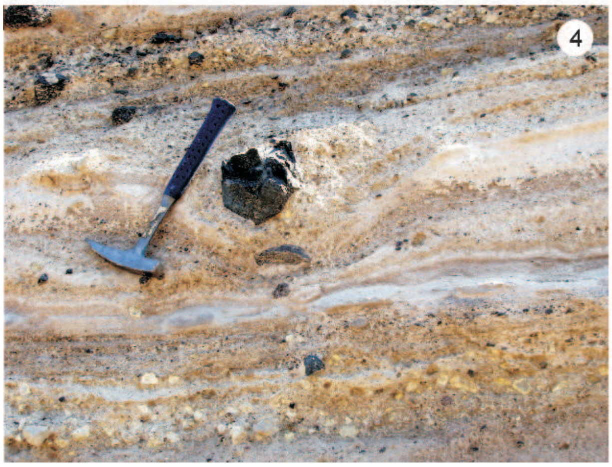


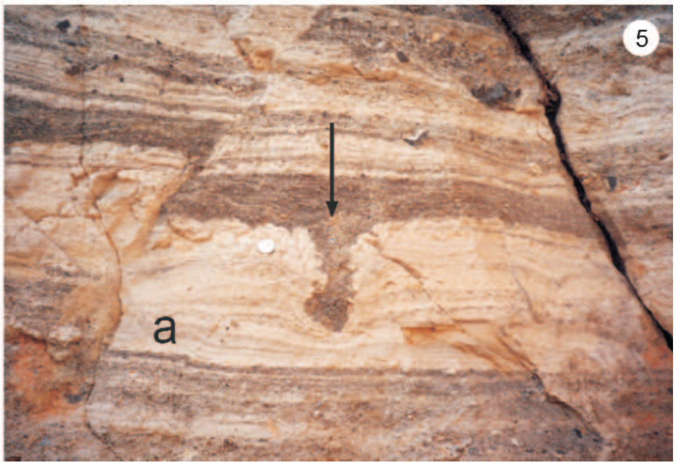




1. Palagonite rim probably developed during transportation of the pyroclasts in the base surge currents formed the pyroclastic succession of Haláp, Hungary. Shorter side of the picture is about 4mm.
2. Advanced palagonitisation of a sideromelane glass shard rich bed resulting red discolorisation of the 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.

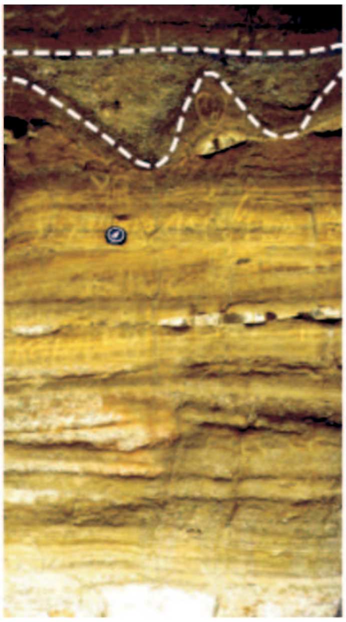
# Plate VII



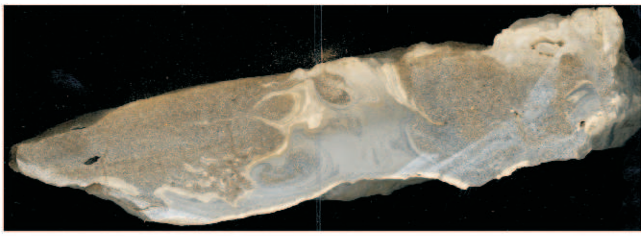


1. Ballistic bomb in phreatomagmatic tephra bed surface in the Kissomlyó tuff ring, Hungary.
2. Ballistic bombs damaged buildings during the Usu 2000 eruption, Hokkaido, Japan.
3. Phreatic explosion crater in Usu, Hokkaido, Japan. Note the destroyed house showed on Plate VI, 2.
4. Impact sags in a phreatomagmatic 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.

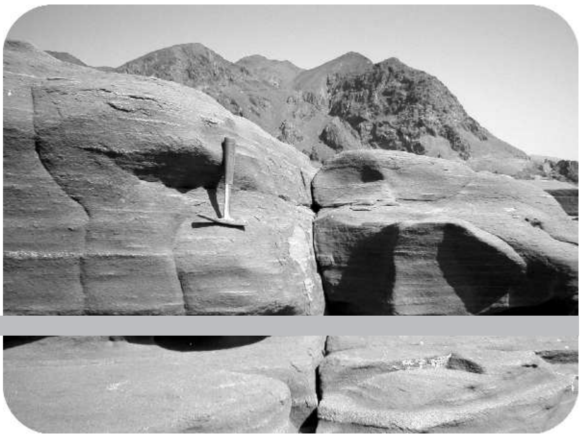
## Plate VIII



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.



# Chapter 4



# Volcanic facies analysis

Volcanic processes are providers of large volumes of volcanic sediment to the surrounding areas of an active volcanic system. They can be classified as either effusive or explosive. During effusive eruptions, various volumes of magma may reach the surface, forming lava flows. Lava flows can provide significant amounts of volcanic debris for secondary, surface erosional processes. Such volcanic material can build up large volumes of clastic sediments that may accumulate in the surrounding effusive eruption dominated terrains. Explosive volcanic eruptions, either triggered by magmatic gas exsolution of the emerging magmatic column in a volcanic conduit and/or triggered by magma and water interaction can deposit large volumes of tephra that is ready to be remobilised subsequently by non-volcanic surface processes. In either way, the interpretation of volcanic rocks (both clastic and coherent) need a broader understanding of the relationship between non-volcanic and volcanic successions. Such studies can be performed during detailed investigation of the volcanic successions, and their relationships to the immediate pre-, syn- and post-volcanic sedimentary successions. This approach is commonly referred to as volcanic facies analysis, and applies similar methods as facies analysis commonly done in non-volcanic sedimentary environments. In this chapter we primarily focus on volcanic sediments, and their importance in understanding the evolution of a volcanic terrain.

## Volcanic facies

Facies is in general a collection of features of a geological unit that define the processes of origin and source as well as the environment of deposition. Facies is in this respect a broad term widely used in geology. Any volcanic facies, perhaps any volcanic rock unit, may define the above basic characteristics. Facies in sedimentology has been defined as the physical, chemical and biologic variations of three-dimensional rock bodies deposited within a specific interval of geological time (Pettijohn et al. 1972, 1975).This definition is valid for any sedimentary rock. The definition itself determines that a facies (or volcanic facies) is a well-defined, usually mapable rock unit that bear common textural, compositional, and internal structural features indicative for a well-defined source, transportation and depositional mechanism. Volcanic facies differ from non-volcanic facies in many ways. In order to understand better their formation, the identification of a single fragmentation history of the juvenile volcanic fragments (in case of primary pyroclastic successions) or multiple fragmentation histories for individual preserved volcanic clasts (in case of reworked or resedimented volcaniclastic rocks) accumulated in a rock unit is essential.

Especially in clastic sedimentology a useful subdivision of certain facies is used to separate proximal, medial and distal facies (Figure 4.1).

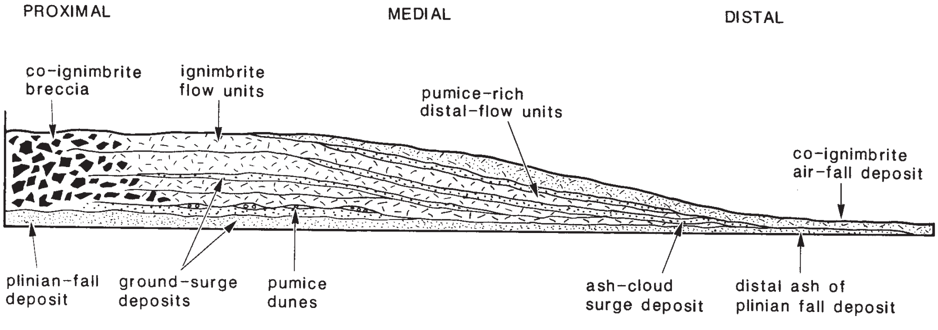


***Figure 4.1.*** Proximal to distal facies variations along an ignimbrite unit (after Wrigth et al 1981)

This subdivision is meaningful since many types of geophysical mass flows (Wright et al. 1981) significantly change their transport and deposition behaviour “en route”. This subdivision is also very useful among volcanic facies, especially in association with pyroclastic density current, laharic or volcanic debris avalanche deposits as well as any secondary, surface erosion resulted volcanogenic mass flow deposit (Mathisen and McPherson 1991) (Figure 4.2). This type of subdivision in volcanic mass flow deposits is very relative. In base surge deposits accumulated around tuff rings or maars the distance between proximal (Plate I, 1) and distal facies (Plate I, 2) is about 1 km (Waters and Fisher 1971, Chough and Sohn 1990, Vazquez and Ort 2006). Similar short distance facies variations have been observed at small volume emergent or subaqueous volcanoes such as Surtseyan-style tuff cones (Sohn and Chough 1992, Mueller et al. 2000, Martin 2002). Often particular bedding features (Plate I, 3) are characteristic and restricted to specific distances from source. This has been well documented for the Laacher See Tephra in Germany (Fisher and Waters 1970, Schmincke et al. 1973). Facies variations may occur over much larger distances in the case of large volume, pyroclastic flow forming eruptions. The 1980 eruption of Mount St. Helens reveals facies changes in a distance of a few kilometres from source (Carey and Sigurdsson 1982, Rowley et al. 1985, Fisher et al. 1987, Druitt 1992, Alidibirov 1995). Facies changes occur even in longer distances in case of ignimbrite forming events, such as the Campanian Ignimbrite in Italy (Fisher et al. 1993). Proximal to distal facies changes can take place over even larger, tens of kilometres in case of large volume ignimbrite eruptions such as those formed in the Andes (de Silva 1989, Ort 1993, Lindsay et al. 2001). The recognition of proximal to distal facies variations as well as the establishment of a correct stratigraphy of extensive ignimbrite sheets can be problematic. Here geochemical studies can prove useful to demonstrate facies changes and to identify facies relationships. This problem is especially pronounced in ancient settings, where the three-dimensional architecture of the otherwise extensive volcanic rock unit is often poorly constrained. This is the case in the Miocene “Rhyolite Tuff Formations” in the Pannonian Basin (Capaccioni et al. 1995, Póka et al. 1998, Szakács et al. 1998, Lukacs et al. 2004).

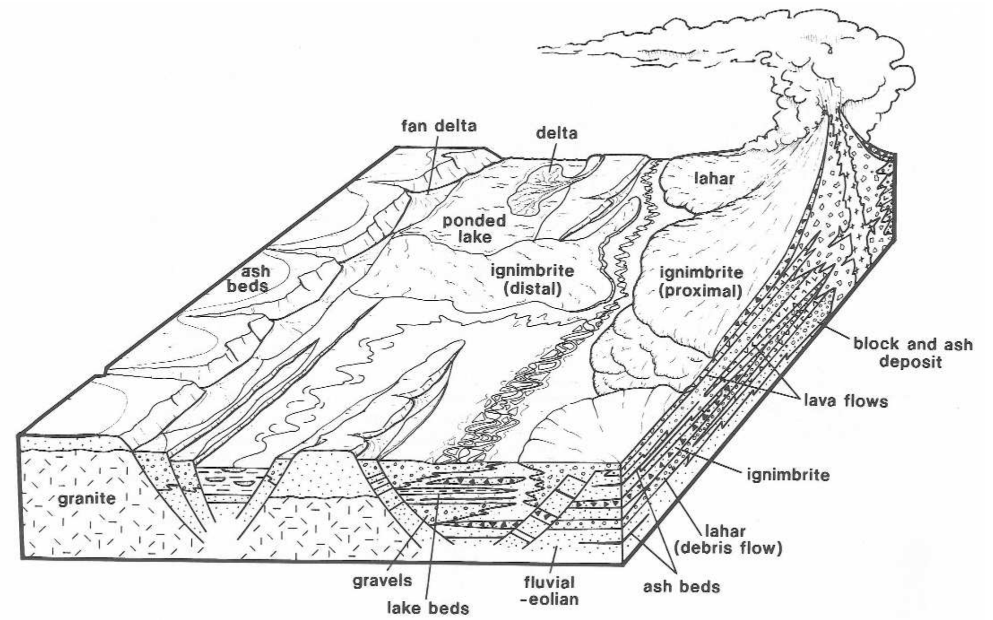


***Figure 4.2.*** Volcanic facies relationships along an active rift zone, complicated by intra-rift faulting and sub-basin development (after Mathisen and McPherson 1991: p. 30, fig. 5)

In general, in volcanic environments proximal volcanic facies usually include the source volcano (e.g. the volcanic rock units accumulated and formed near to the source vent). In young volcanic terrains, where the original volcanic landform is more or less still intact, the association of proximal to distal volcanic facies to the geometrical position of the preserved rocks units in comparison to the source volcano is generally straight forward. In old volcanic successions however, the source volcano may be completely vanished, or displaced by tectonic processes into far position, hindering the identification of proximal to distal facies relationships (Plate I, 4). One of the most important facts in studying old volcanic successions is, that in most of the cases volcanic rocks accumulated and preserved in basinal settings. Basin fills can be intercalated with distal facies portions of primary to secondary volcaniclastic successions, and only inference of the texture, extension, and geometry of the proximal volcanic deposits is possible (Smith et al. 1988, Besly and Collinson 1991, Montanari et al. 1994, Dostal et al. 2003, Kataoka 2005). The study of basin filling volcaniclastic successions can tell us important details of the deposition environment (e.g. subaqueous versus subaerial) (Barcat et al. 1989, Bull and Cas 1989, Turbeville 1991, Yogodzinski et al. 1996, Schmincke et al. 1997, Jerram et al. 1999, Schneider et al. 2001, Breitkreuz et al. 2002, Martin and Németh 2005). In the geological record volcanic sandstones often form the only remnants (Plate I, 5) preserved in the continental sedimentary record (Ufnar et al. 1995, Millward et al. 2000, Zimmermann and Bahlburg 2003).

There is also significant facies change from primary and secondary volcanic successions. In distal regions, primary pyroclastic successions commonly transform laterally into secondary volcaniclastic successions. Such changes often occur in continental to marine deposit transitions, as many primary pyroclastic density currents enter to lakes or the sea (Plate I, 6) (Bull and Cas 1989, Mueller 1991, de Rita et al. 2002). Therefore primary pyroclastic successions, commonly inter-finger with marine deltaic, turbidity current or deep marine successions (Cole and Stanley 1994, Cousineau 1994). The recognition of such facies changes and the establishment of a three-dimensional relationship between individual rock units may help to establish a basin-wide environmental reconstruction (Kano 1991, McPhie and Allen 1992, Macdonald and Barr 1993, Fergusson et al. 1994, Mangano and Buatois 1997, Nagy et al. 1999, Hathway and Kelley 2000, Dostal et al. 2003, Franzese et al. 2003).

Volcanic eruptions, especially major arc related strato-volcanic eruptions, immediately influence the surrounding sedimentary environment, and can provide significant volume of volcaniclastic material into the sedimentary basin. Such composite and long-lived volcanic systems simultaneously provide primary pyroclastic deposits in syn-eruptive time and their reworked derivates during inter-eruptive periods. In this way, a complex primary and secondary volcanic sedimentary record can accumulate, commonly inter-fingered with non-volcanic deposits. It may be further complicated by the fact that horizontally moving volcanic gravity currents (either primary or secondary) obey to physical laws to deposit the material they transport. The transformations of the physical properties of volcanic gravity currents are the major cause of formation of changes in the facies of the accumulating volcanic deposits (Fisher 1983). Such changes can be caused by (1) density separations within the passing particle-laden gravity current due to gravitational settling of particles (gravity transformation), (2) velocity variations caused by slope angle changes without significant variations of interstitial fluid content of the body of the gravity current (body transformation), and (3) separations of particles by the turbulent mixing within the boundary between the ambient fluid and the flow surface (surface transformation) (Fisher 1983) (Plate II, 1).

## Volcanic facies analysis

With volcanic facies analyses volcanic deposits (both coherent and fragmented) are studied in an aim to establish horizontal and vertical facies relationships that could help to understand the volcanic eruptions, the subsequent reworking processes and the relationship as well as the interaction between volcanic and non-volcanic sedimentary environments. During volcanic facies analysis we can identify general trends of the texture, composition, lateral distribution, magma fragmentation and eruption style generated the studied volcanic successions.

In general, lithofacies studies focus on distinguishable volcanic units, that can be mapped and therefore limit in the field (e.g. in outcrop scale). To define a lithofacies, common, and ready to recognized features such as structure, texture, internal organisation, geometry, and componentry should be identified in outcrop scale. A volcanic lithofacies can contain more than a single bed, and is not necessary the result of a well distinguished volcanic event (e.g. a single pyroclastic density current deposited succession). An identified lithofacies often reveals common vertical and lateral facies variations that can be interpreted as changes in the depositional behaviour of the parental flow.

To describe a volcanic lithofacies, similar methods are used as in clastic sedimentology. To distinguish certain lithofacies, commonly the grain size characteristics of the deposit (rock) unit is one of the key parameters, since grain size pattern of a certain deposit (rock) has direct link to the physical properties of the transporting and depositing currents (Figure 4.3). The size grading (graded, non-graded, normal or reverse graded, density graded) texture of the unit is an additional useful parameter to distinguish certain lithofacies. In addition, the internal stratification and clast orientation (e.g. imbrication) are important parameters to define a certain lithofacies (Sohn and Chough 1989, Chough and Sohn 1990). On the basis of these three major parameters, a table-like matrix easily can be created, and every well-defined rock unit (e.g. bed or group of beds) can be assigned to one of the defined category. The categories may vary (Figure 4.4) in accordance of the usual lithofacies categories used in previous studies and perhaps for practical reasons (e.g. not to be lost in a too detailed descriptive study) (Sohn and Chough 1989, Chough and Sohn 1990, Sohn and Chough 1992, Sohn 1995, Németh et al. 1999, Németh et al. 2001, Németh and White 2003). In a well-defined matrix of rock unit parameters may give certain categories cannot be identified in the field.

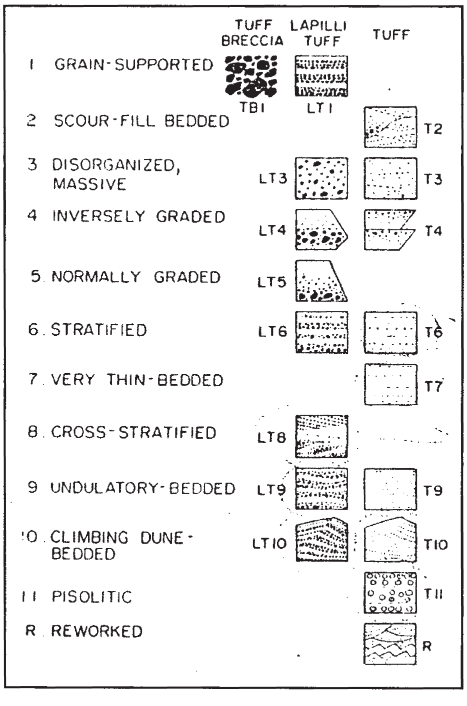
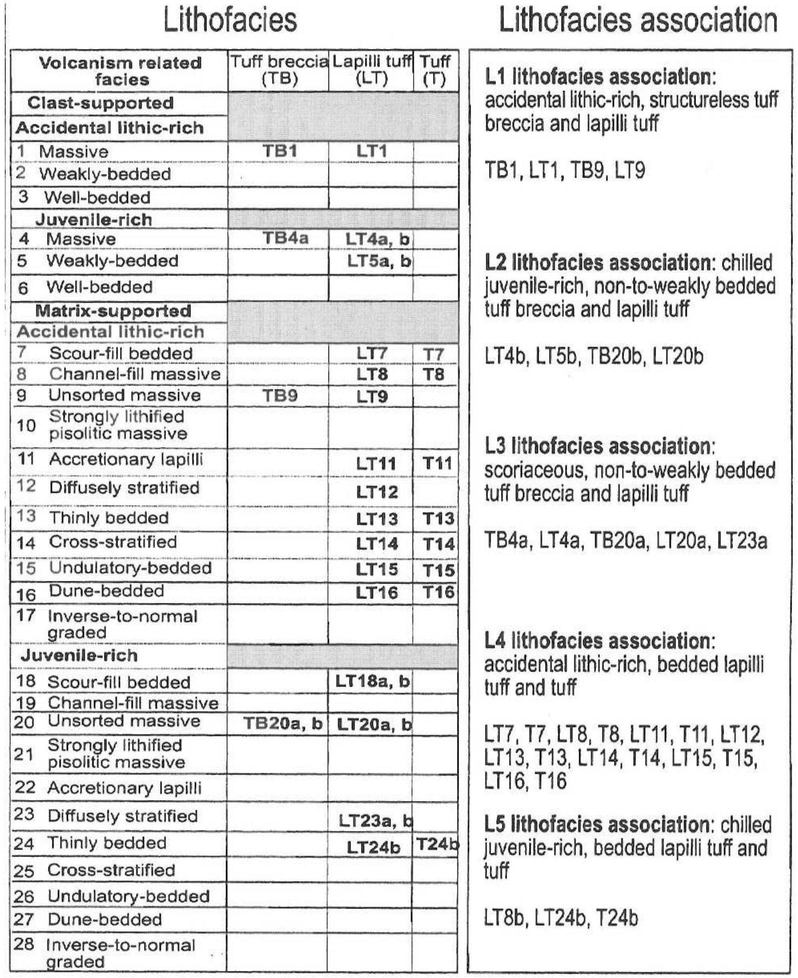


*Figure 4.3.* Lithofacies classification diagram used in Sohn and Chough (1989) and Chough and Sohn (1990) to describe volcanic units from the phreatomagmatic volcanoes of the Jeju Island, Korea. In the horizontal "axis" the grain size (from tuff breccia to tuff) and in the vertical "axis" the main bedding features have been used to identify key lithofacies categories



Lithofacies

| Volcanism related facies | Tuff breccia (TB) | Lapilli tuff (LT) | Tuff (T) |
|---|---|---|---|
| **Clast-supported** | | | |
| **Accidental lithic-rich** | | | |
| 1 Massive | TB1 | LT1 | |
| 2 Weakly-bedded | | | |
| 3 Well-bedded | | | |
| **Juvenile-rich** | | | |
| 4 Massive | TB4a | LT4a, b | |
| 5 Weakly-bedded | | LT5a, b | |
| 6 Well-bedded | | | |
| **Matrix-supported** | | | |
| **Accidental lithic-rich** | | | |
| 7 Scour-fill bedded | | LT7 | T7 |
| 8 Channel-fill massive | | LT8 | T8 |
| 9 Unsorted massive | TB9 | LT9 | |
| 10 Strongly lithified pisolitic massive | | | |
| 11 Accretionary lapilli | | LT11 | T11 |
| 12 Diffusely stratified | | LT12 | |
| 13 Thinly bedded | | LT13 | T13 |
| 14 Cross-stratified | | LT14 | T14 |
| 15 Undulatory-bedded | | LT15 | T15 |
| 16 Dune-bedded | | LT16 | T16 |
| 17 Inverse-to-normal graded | | | |
| **Juvenile-rich** | | | |
| 18 Scour-fill bedded | | LT18a, b | |
| 19 Channel-fill massive | | | |
| 20 Unsorted massive | TB20a, b | LT20a, b | |
| 21 Strongly lithified pisolitic massive | | | |
| 22 Accretionary lapilli | | | |
| 23 Diffusely stratified | | LT23a, b | |
| 24 Thinly bedded | | LT24b | T24b |
| 25 Cross-stratified | | | |
| 26 Undulatory-bedded | | | |
| 27 Dune-bedded | | | |
| 28 Inverse-to-normal graded | | | |

Lithofacies association

**L1 lithofacies association**: accidental lithic-rich, structureless tuff breccia and lapilli tuff

TB1, LT1, TB9, LT9

**L2 lithofacies association**: chilled juvenile-rich, non-to-weakly bedded tuff breccia and lapilli tuff

LT4b, LT5b, TB20b, LT20b

**L3 lithofacies association**: scoriaceous, non-to-weakly bedded tuff breccia and lapilli tuff

TB4a, LT4a, TB20a, LT20a, LT23a

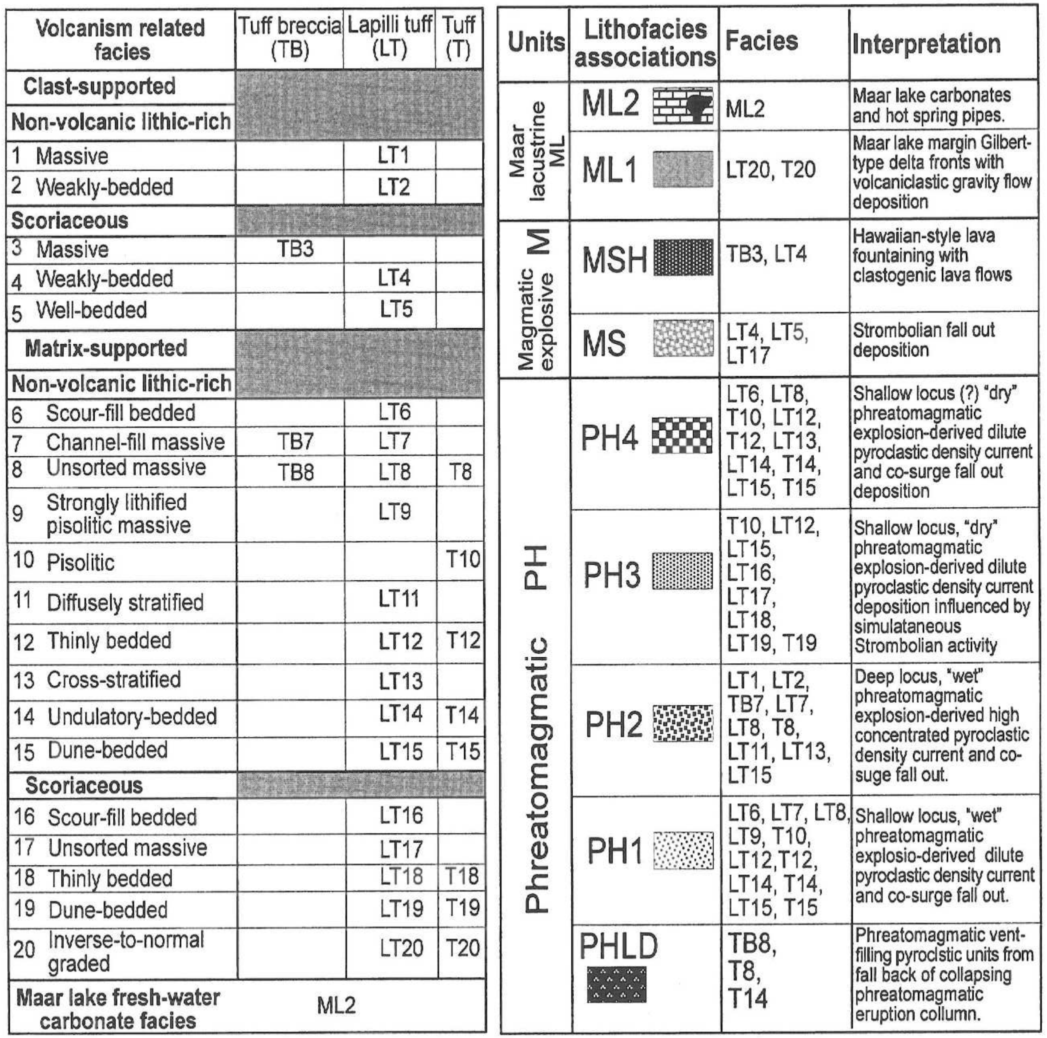
**L4 lithofacies association**: accidental lithic-rich, bedded lapilli tuff and tuff

LT7, T7, LT8, T8, LT11, T11, LT12, LT13, T13, LT14, T14, LT15, T15, LT16, T16

**L5 lithofacies association**: chilled juvenile-rich, bedded lapilli tuff and tuff

LT8b, LT24b, T24b

*Figure 4.5.* Lithofacies classification diagram used in NÉMETH and WHITE (2003) to describe volcanic units from the phreatomagmatic volcanoes of the Waipiata Volcanic Field, Otago, New Zealand. The main categories to identify certain lithofacies were similar to the SOHN and CHOUGH (1989) and CHOUGH and SOHN (1990) lithofacies scheme. In this classification scheme the volume of accidental lithic fragments versus juvenile fragments in the deposits has been used to refine the classification scheme



| Volcanism related facies | Tuff breccia (TB) | Lapilli tuff (LT) | Tuff (T) |
|---|---|---|---|
| **Clast-supported** | | | |
| **Non-volcanic lithic-rich** | | | |
| 1 Massive | | LT1 | |
| 2 Weakly-bedded | | LT2 | |
| **Scoriaceous** | | | |
| 3 Massive | TB3 | | |
| 4 Weakly-bedded | | LT4 | |
| 5 Well-bedded | | LT5 | |
| **Matrix-supported** | | | |
| **Non-volcanic lithic-rich** | | | |
| 6 Scour-fill bedded | | LT6 | |
| 7 Channel-fill massive | TB7 | LT7 | |
| 8 Unsorted massive | TB8 | LT8 | T8 |
| 9 Strongly lithified pisolitic massive | | LT9 | |
| 10 Pisolitic | | | T10 |
| 11 Diffusely stratified | | LT11 | |
| 12 Thinly bedded | | LT12 | T12 |
| 13 Cross-stratified | | LT13 | |
| 14 Undulatory-bedded | | LT14 | T14 |
| 15 Dune-bedded | | LT15 | T15 |
| **Scoriaceous** | | | |
| 16 Scour-fill bedded | | LT16 | |
| 17 Unsorted massive | | LT17 | |
| 18 Thinly bedded | | LT18 | T18 |
| 19 Dune-bedded | | LT19 | T19 |
| 20 Inverse-to-normal graded | | LT20 | T20 |
| **Maar lake fresh-water carbonate facies** | ML2 | | |

| Units | Lithofacies associations | Facies | Interpretation |
|---|---|---|---|
| Maar lacustrine ML | ML2 ▨ | ML2 | Maar lake carbonates and hot spring pipes. |
| | ML1 ▨ | LT20, T20 | Maar lake margin Gilbert-type delta fronts with volcaniclastic gravity flow deposition |
| Magmatic explosive M | MSH ▨ | TB3, LT4 | Hawaiian-style lava fountaining with clastogenic lava flows |
| | MS ▨ | LT4, LT5, LT17 | Strombolian fall out deposition |
| Phreatomagmatic PH | PH4 ▨ | LT6, LT8, T10, LT12, T12, LT13, LT14, T14, LT15, T15 | Shallow locus (?) "dry" phreatomagmatic explosion-derived dilute pyroclastic density current and co-surge fall out deposition |
| | PH3 ▨ | T10, LT12, LT15, LT16, LT17, LT18, LT19, T19 | Shallow locus, "dry" phreatomagmatic explosion-derived dilute pyroclastic density current deposition influenced by simulataneous Strombolian activity |
| | PH2 ▨ | LT1, LT2, TB7, LT7, LT8, T8, LT11, LT13, LT15 | Deep locus, "wet" phreatomagmatic explosion-derived high concentrated pyroclastic density current and co-suge fall out. |
| | PH1 ▨ | LT6, LT7, LT8, LT9, T10, LT12,T12, LT14, T14, LT15, T15 | Shallow locus, "wet" phreatomagmatic explosio-derived dilute pyroclastic density current and co-surge fall out. |
| | PHLD ▨ | TB8, T8, T14 | Phreatomagmatic vent-filling pyroclstic units from fall back of collapsing phreatomagmatic eruption collumn. |

*Figure 4.4.* Lithofacies classification diagram used in NÉMETH et al. (2001) to describe volcanic units from the phreatomagmatic volcanoes of the Tihany Peninsula, Hungary. The main categories to identify certain lithofacies were similar to the SOHN and CHOUGH (1989) and CHOUGH and SOHN (1990) lithofacies scheme. Next to bedding characteristics, the dominance of accidental lithic fragments as well as scoriaceous particles in the sediment has been used to distinguish between phreatomagmatic and magmatic explosive eruption related deposits

After describing and interpreting each rock units in this way, a very detailed lithofacies group can be established. In the field perhaps certain cycles in vertical sections and certain vertical facies relationships can be traced. Such cycles and/or common vertical facies relationships can give important information to the transportation and depositing gravity currents, and their physical properties. Certain vertical facies relationships could be identified in specific geometrical distributions, and therefore a further horizontal facies relationship can be established. Such horizontal facies relationships are characteristic for the proximal, medial or distal section of the preserved volcanic succession. As an outcome of such a detailed study, a three-dimensional reconstruction of volcanic processes, their environmental variations and the identification of source regions can be done.

In volcanic successions, fragmentation history of juvenile clasts or the relative ratio between juvenile and accidental lithic fragments can also be added as a dividing category (Figure 4.5). In this respect, magmatic and phreatomagmatic fragmentation history of the juvenile clast population of an otherwise similarly transported pyroclastic system can result in two distinct lithofacies (in case other sedimentological parameters are the same).

Using assigned facies to describe outcrop-scale volcanic successions can be illustrated in a stratigraphic column

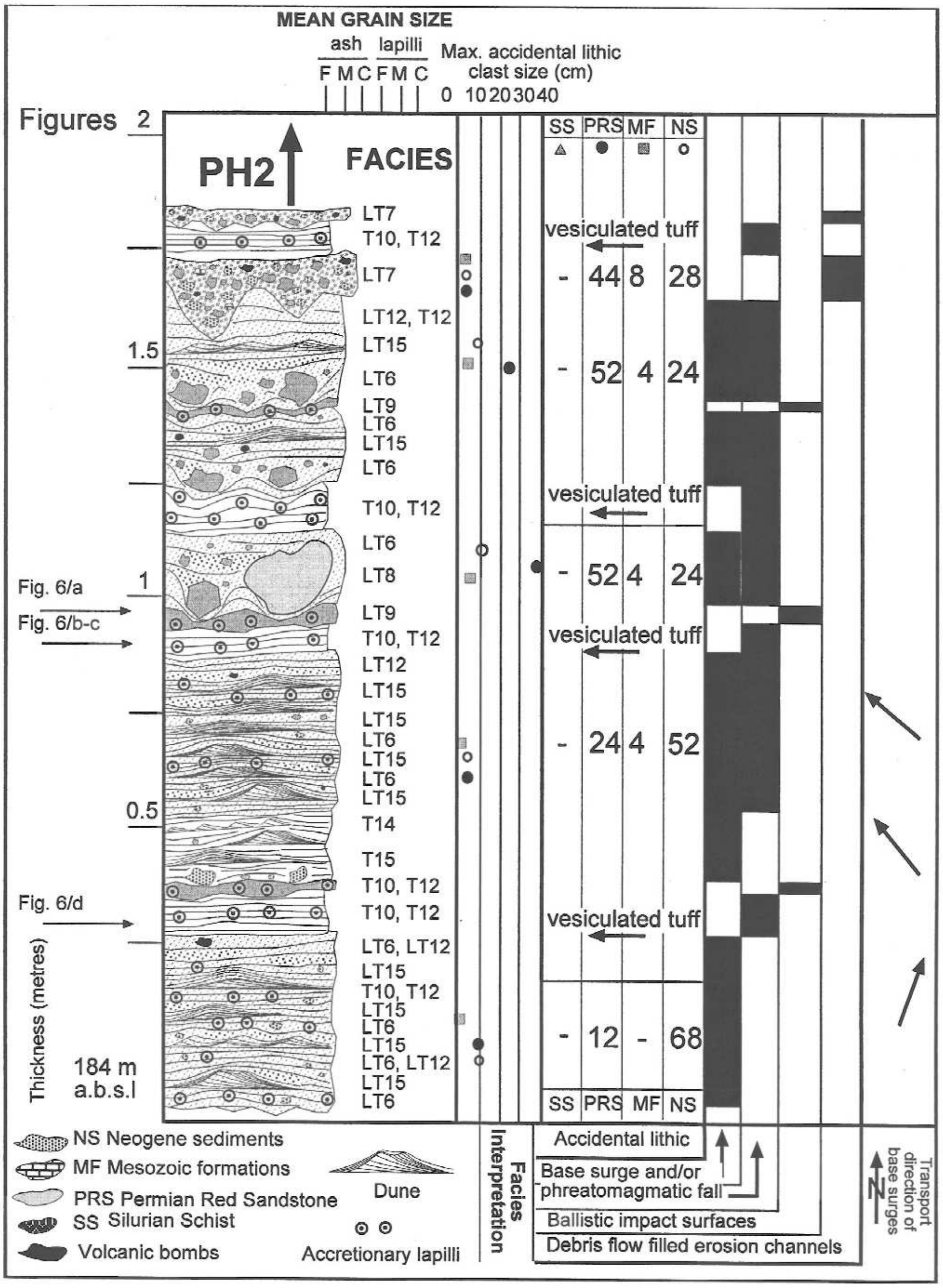


*Figure 4.6.* A graphic stratigraphic log of the phreatomagmatic successions of Tihany, Hungary, detailed with the identified lithofacies categories (NÉMETH et al. 2001)

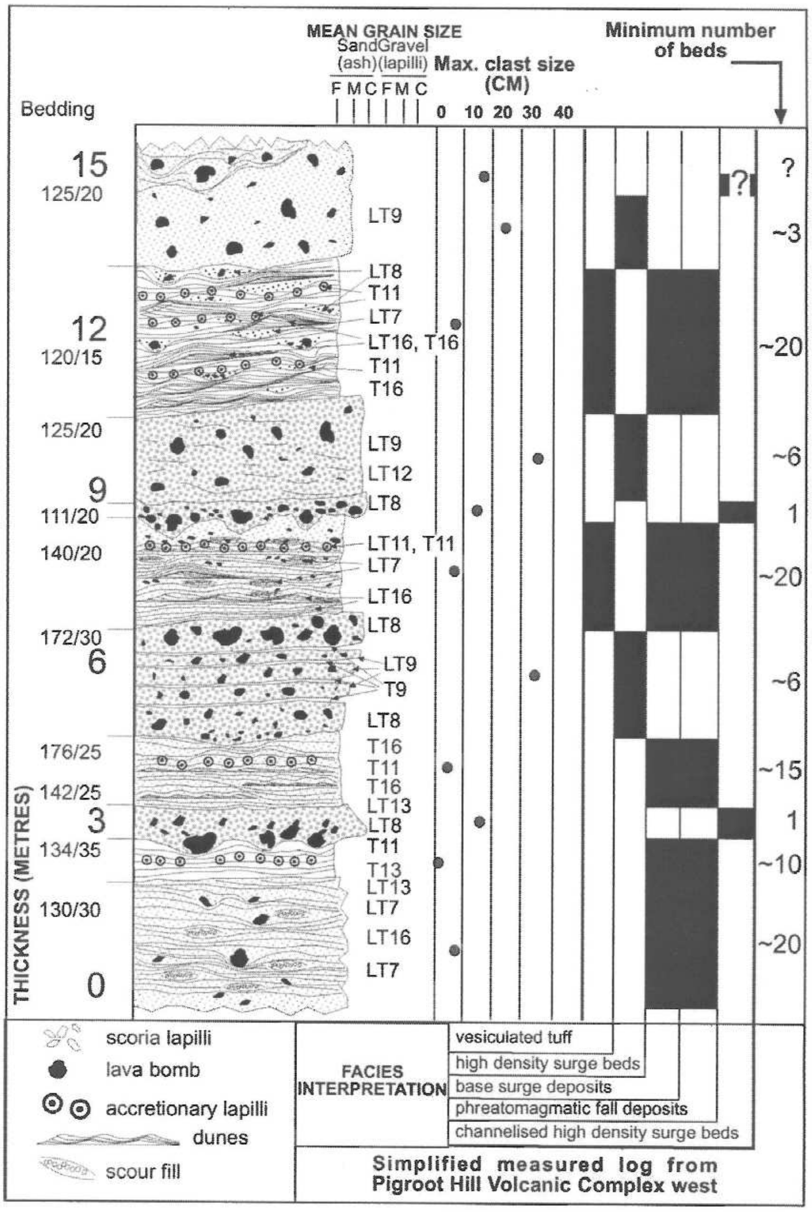


*Figure 4.7.* A graphic stratigraphic log of the phreatomagmatic successions of one of thephreatomagmatic volcano of the Waipiata Volcanic Field in Otago, New Zealand (NÉMETH and WHITE 2003)

(Figures 4.6 and 4.7). In volcanic field-wide mapping, such columns can help to establish vertical and lateral facies relationships that may be used to reconstruct the volcanic eruptive environment.

### Differences of volcaniclastic sediments from non-volcanic sediments

From a methodological point of view studies and interpretations of volcaniclastic rocks/sediments do not differ from those performed on clastic non-volcanic sediments. However, there are characteristic differences that made volcaniclastic studies commonly complicated. Such differences can be taken in account during defining our lithofacies descriptive categories. The specific features of volcanic processes should be treated with care. Descriptive methods should not be mechanically applied during the identification of certain lithofacies.

#### *Time-scale of deposition*

One of the most difficult issues in studies of volcaniclastic sediments/rocks is to understand the time-scale of deposition. Usually a bedding plane in normal sedimentary environments clearly demonstrates a time break in sedimentation that allows forming a layer physically different to those located below that surface. Such differences may be generated by slight changes of physical, chemical or compositional changes of the depositing sediments subsequently accumulating over a freshly developed bedding plane. But in fact, the time break is in most of the cases significant. In accumulating tephra however, characteristic bedding surfaces can develop without significant time breaks between certain depositional events. Such changes commonly manifest as bed planes and instead of giving evidence of time breaks during the sedimentation giving signs of changes of eruption conditions. In this respect a primary volcaniclastic succession may consists of many well-defined beds but none of them reflect breaks, rejuvenation or erosional events may affect the accumulating tephra. Based on direct observation of volcanic eruptions it is well known that characteristic changes could take place in the development of the eruption cloud over the duration of the eruption (Plate II, 2). In Plinian eruptions, the eruption continuously forms a large umbrella like eruption cloud. This eruption plume is initially supported by the uprising gas thrust and can reach few tens of kilometres above the volcano. However, there is a fine balance, when the eruption column will be charged by expelled pyroclasts that become heavy and the gravitational forces eventually overcome the upward directed forces. In such situations, the eruption cloud will start to partially collapse and the particles fall back to the surface and initiate horizontally moving pyroclastic density currents. In this simple eruption scenario, the accumulating tephra will reflect these changes upsection its stratigraphy (Plate II, 3). In spite of the dramatic changes of the transportation and therefore the depositional sequences of the accumulating tephra, the entire pyroclastic succession produced in such events can be volumeneous, and may reflect not more then few hours history, with no significant time breaks. However, the sequence will be strikingly complex, with many well-defined beds in the commonly few tens of metres thick successions.

The depositional time of primary volcaniclastic deposits are generally also significantly shorter than that of any other non-volcanic deposits. In extreme case, during magma–water interaction triggered phreatomagmatic explosive eruptions or during scoria cone forming events, few tens of metres thick pyroclastic sequence can accumulate in just few hours (Plate II, 4). In spite of the very short accumulation time frame of such events, the resulting deposits, again, can show complex bedding features, but almost none of them reflecting time breaks of the eruption itself. The complexity of such quickly accumulated sequences clearly associated with the complexity of the eruption dynamics of phreatomagmatic explosive eruptions.

Even volcanic events, that are able to mobilise few km$^3$ volume of material in a single event, such as a formation of volcanic debris avalanches due to volcano collapses, are believed to form within a few days time. The identified deposits still can be very complex, and their studies need a very different approach to studies of other non-volcanic sediments.

### *Volume and aerial extension of deposits*

The accumulated volume of erupted pyroclastic successions can also differ significantly to other non-volcanic sediments. Extensive ignimbrite sheets for instance can be traced over hundreds of kilometres distances, and their thickness may change insignificantly over large distances. Especially large caldera systems can be associated with such large-volume, commonly very monotoneous (at least from textural point of view) ignimbrite successions. The erupted volume is even more dramatic in sheet-like ignimbrite fields like those formed in South America, which lack apparent known sources. The commonly large extension and volume of volcaniclastic deposits perhaps bear some internal textural variations. Such variations are commonly not obvious in short distances, but can be very pronounced over large (tens of kilometres) distances. Textural changes therefore can be very gradual, and only detailed facies analysis can reveal them.

In other hand pyroclastic deposits can also be very limited in extension and volume, and could be interbedded with deposits an order larger in volume and lateral extension. Especially phreatomagmatic pyroclastic deposits related to tuff rings and maars are this type of deposits. Lateral textural changes in this type of deposits can be dramatic in very short distances (few hundreds of metres).

### *Cross sectional variations*

Many volcanic eruptions produce horizontally moving pyroclastic density currents. Such currents can move away from an erupting vent, and gradually depositing tephra “en route” following the changing physical properties of the currents. However, the physical properties of pyroclastic density currents do not exclusively vary in downstream direction but also in lateral extent. Therefore different depositional behaviour characterises their marginal and central zones, which can be identified by significant textural, bedding, and even compositional differences along the cross section of pyroclastic flow deposits. Pyroclastic density currents often shift their main depositional axis over time. Such shifts can accumulate valley and overbank facies over each other creating characteristic vertical facies relationships. In a similar way, other volcaniclastic mass flow deposits (e.g. any laharic deposits) can also be associated with characteristic marginal and central facies. In this respect identification of cross sectional variations of volcaniclastic mass flows is important to distinguish from other textural variations caused by longitudinal changes of deposition behaviour.

### *Variations in composition*

Eruption conditions can change dramatically over short times. A sudden entry of water to an erupting vent can switch magmatic gas exsolution driven magma fragmentation to more magma–water interaction driven phreatomagmatic fragmentation (Barberi et al. 1989). As a consequence the resulting eruption column will be charged in excavated country rocks that may deposit in well-distinguished beds of fall or pyroclastic density currents. Such changes can produce a volcaniclastic sequence which maybe deposited from the same eruption cloud or pyroclastic density current but transported different type of clasts. This type of changes can also be gradual, with no apparent sharp marks may result distinguishable, or easy to recognize beds. Also, the near surface magmatic feeding system of a volcano can change vastly during the eruption. The volcanic conduit can collapse or become wider, changing the style of an ongoing eruption. Typical accidental lithic clast horizons may reflect vent clearing events (Plate II, 5). The level of vesiculation of erupting magma can change resulting pyroclasts with different vesicularity. A result of this process is commonly recognised in scoria cone sequences (Houghton et al. 1999).

Chemical zonation of a magma chamber also can result in gradual changes of the composition of juvenile fragments in an otherwise texturally uniform pyroclastic succession. Such changes commonly have no obvious manifestation in the accumulated pyroclastic successions, and only detailed geochemical study of the juvenile fragments can reveal its existence (Briggs et al. 1993, Wilson et al. 1995, Beresford et al. 2000, Edgar et al. 2002, Maughan et al. 2002, Sumner and Branney 2002). In other cases chemical zonation of a pyroclastic succession can also be abrupt as the result of a gradual emptying of a layered magma chamber.

### *Thermal properties of pyroclasts*

Juvenile pyroclastic particles erupted from volcanic eruptions are commonly hot, and their hot temperature can be retained long during transport. Gradual cooling of the depositing pyroclastic system can effect the behaviour of the freshly deposited tephra. In extreme case, ignimbrites in near vent position tend to retain their heat long enough to form welded zones. In more distal position, ignimbrite units can be still hot enough to vapourise entrapped water which may result in secondary explosion craters on the deposit surface (Moyer and Swanson 1987). Such active thermal circulation can create textural features such as hard layers, degassing pipes, or hydrothermal mineralization. To distinguish these features from primary depositional features can be challenging, and detailed facies analyses of the preserved volcanic succession is necessary.

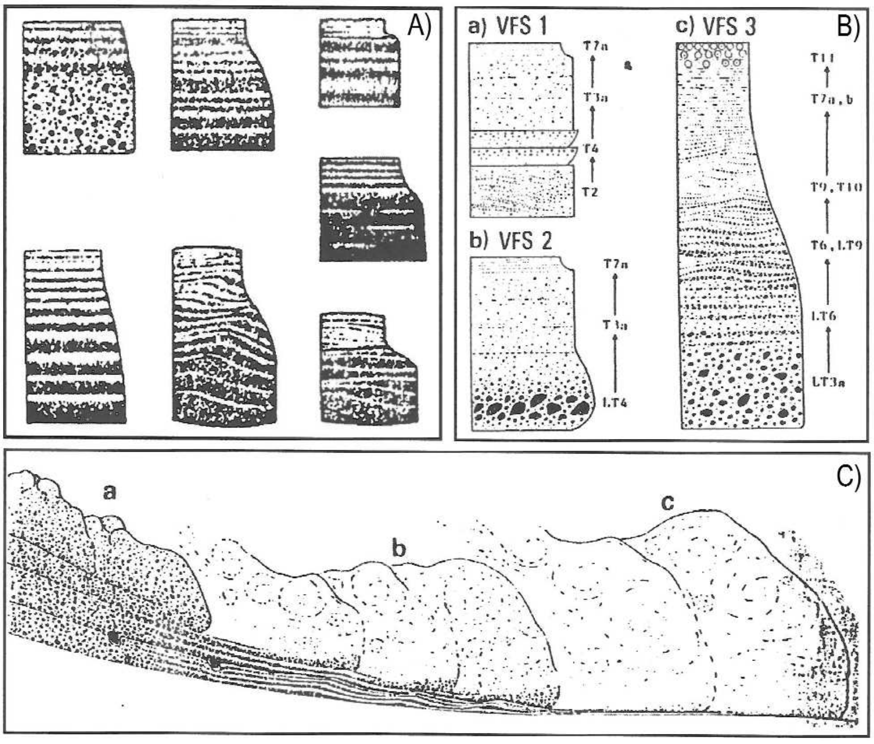


*Figure 4.8.* Identified vertical facies variations in relationship to base surge evolution (Sohn and Chough 1989; Chough and Sohn 1990; Sohn 1996). A) Graphic representations of dominant types of vertical facies relationships identified from phreatomagmatic volcanoes of the Jeju Island, Korea. B) 3 main vertical facies relationship from phreatomagmatic volcanoes of the Jeju Island, Korea. Compare the facies codes to the facies table on Figure 4.3. C) A model to associate the identified vertical and horizontal facies variations of the phreatomagmatic volcanoes of the Jeju Island, Korea with the evolution of a base surge current, which gradually become more turbulent due to loss of particle concentration "en route"

### Vertical and lateral facies relationships

During volcanic facies analysis, certain trends can be outlined both vertical and longitudinal manner. Vertical facies relationships are important to understand the timescale, apparent shifts of the vent location, or changes in eruption style during the course of the eruption (Figure 4.8). Meanwhile the longitudinal facies relationships can tell us important facts about the evolution of a volcanic gravity current from the proximal to the distal areas (Figure 4.9). Such changes are com-

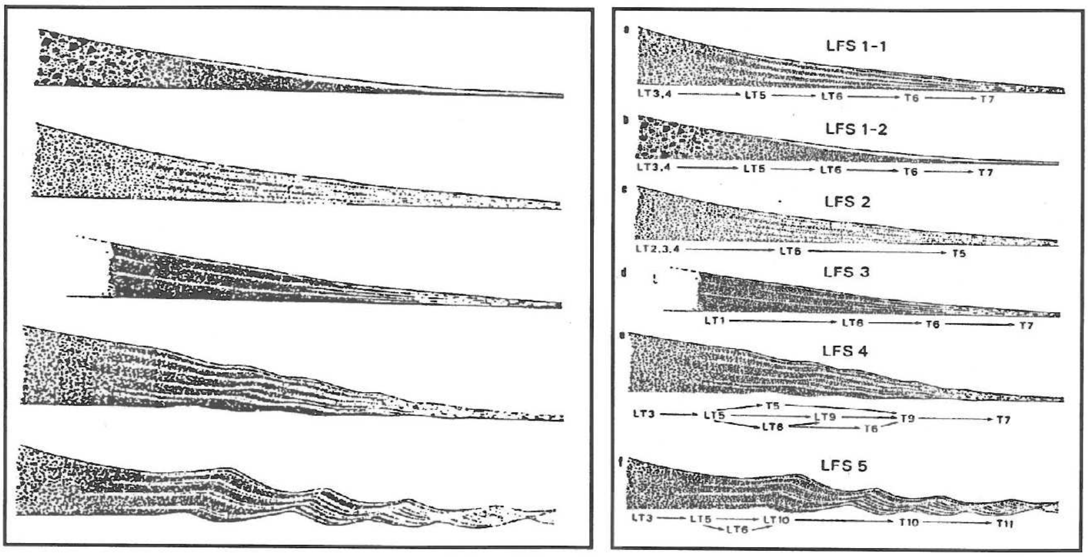


*Figure 4.9.* Identified lateral (horizontal) facies relationships in base surge deposits of phreatomagmatic volcanoes of the Jeju Island (Sohn and Chough 1989; Chough and Sohn 1990). Compare the facies codes with the facies table on Figure 4.3

monly related to the flow transformation of certain volcanic gravity currents. During careful facies analyses, very fine changes and relationships can be established as it has been demonstrated from small-volume mafic explosive volcanoes from Jeju Island, Korea (Sohn and Chough 1989, Chough and Sohn 1990, Sohn and Chough 1992, 1993, Sohn 1995, 1996), Sonora, Mexico (Wohletz and Sheridan 1979, 1983b) or from Linosa, Italy (Lajoie et al. 1992). However, recognizing facies relationships and connect to them to real natural processes is not easy and ongoing subject of recurring researches of similar fields. Lateral facies variations of tuff ring sequences has been established on the basis of tuff rings and maars of Sonora (Wohletz and Sheridan 1979, Wohletz and Sheridan 1983a). Similar studies from Linosa, Italy (Lajoie et al. 1992) revealed a facies change trend from massive to parallel bedded and then dune bedded facies groups larger distances from the source (Figure 4.10). An even more detailed study from Korea (Sohn and Chough 1989, Chough and Sohn 1990), suggested far more complex facies relationships (Figure 4.11) highlighting the role of the level of details how a facies study can be performed in the resulting interpretations. Similar problems are even more evident in larger volume volcanic systems, where outcrop availability, preservation and the large lateral extension can hinder the identification of key facies relationships.

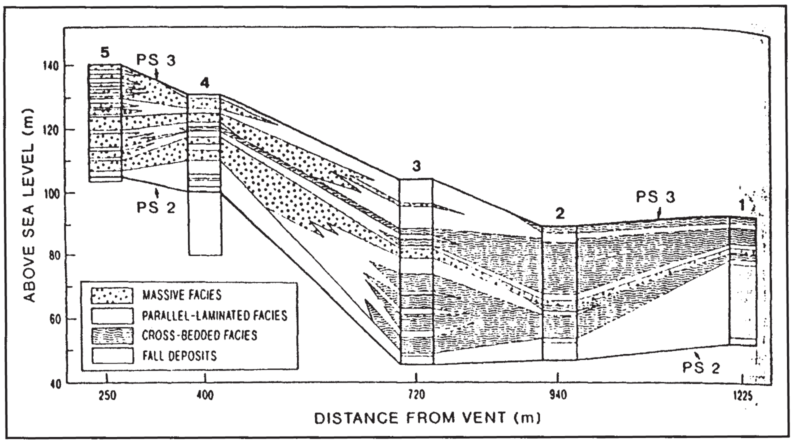


*Figure 4.10.* Horizontal facies relationships of identified facies in relationship with increasing distance from the source in a tuff ring sequence of Linosa Island, Italy (Lajoie et al. 1992). Massive facies gradually give places to cross-bedded facies in increasing distance from the source

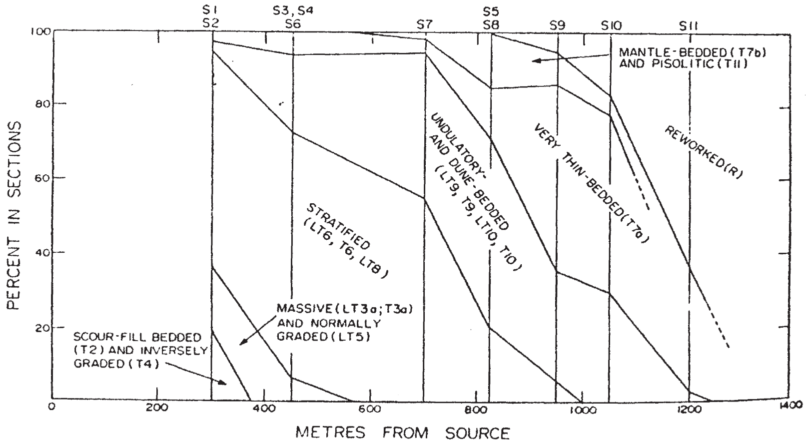


*Figure 4.11.* Horizontal facies relationships of identified facies in relationship with increasing distance from the source in a tuff ring sequence of Jeju Island, South Korea (Sohn and Chough 1989; Chough and Sohn 1990). Stratified facies gradually give places to undulatory and dune-bedded facies in increasing distance from the source

### Volcaniclastic aprons

Large-scale explosive volcanic eruptions spread large volumes of volcaniclastic materials to volcanic fan, alluvial plain, lacustrine and marine settings. The volcaniclastic sediments may become redistributed shortly after an eruption by various reworking processes, such as debris flows, hyperconcentrated flows, dilute fluvial flood flows, turbidity and storm currents (Smith 1987, Scott 1988). Resedimentation of volcaniclastic debris in terrestrial settings is commonly associated with major and catastrophic environmental changes caused by the supply of high sediment load into the basin (Palmer and Walton 1990, Mathisen and McPherson 1991, Smith and Lowe 1991). Large-scale eruptions can potentially result in catastrophic impacts over very wide areas and in far-distant places. However, resedimentation induced by large-volume (i.e. hundreds of km$^3$) ignimbrite-emplacing eruptions has not been observed directly and remains poorly understood.

Volcanism in subaerial and submarine settings can accumulate large volume of apron-like deposits. Pyroclastic fans are common around large frequently active volcanoes, and could build up thousands of metres thick units. Volcano Mayon in the Phillippines (Rodolfo and Arguden 1991), is a well-known example for extensive volcaniclastic fans with interfingered primary and reworked depositional units (Figure 4.12). These types of volcaniclastic fans are commonly named block and ash fans (Siebe et al. 1993) and can reveal a huge complexity. Large volcaniclastic fans can also develop in alluvial settings, especially where active tectonism and basin subsidence can enhance the inflow of large volumes of volcaniclastic sediments from nearby stratocones. Such settings are very common in the geological record and their potential for hydrocarbon exploration is still not fully explored (Mathisen and McPherson 1991).

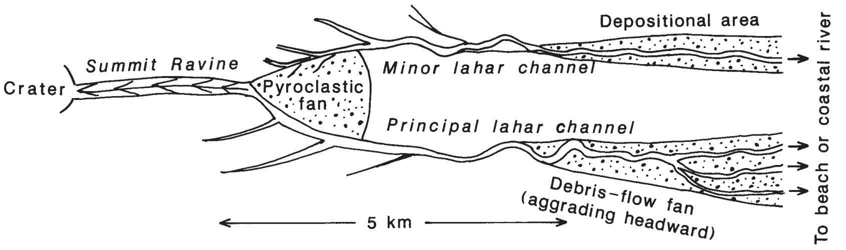


***Figure 4.12.*** Faices model of the Mayon lahar system in Philippines (after Rodolfo and Arguden 1991). Repeated volcanic activity can build up extensive volcaniclastic aprons and channel filling successions

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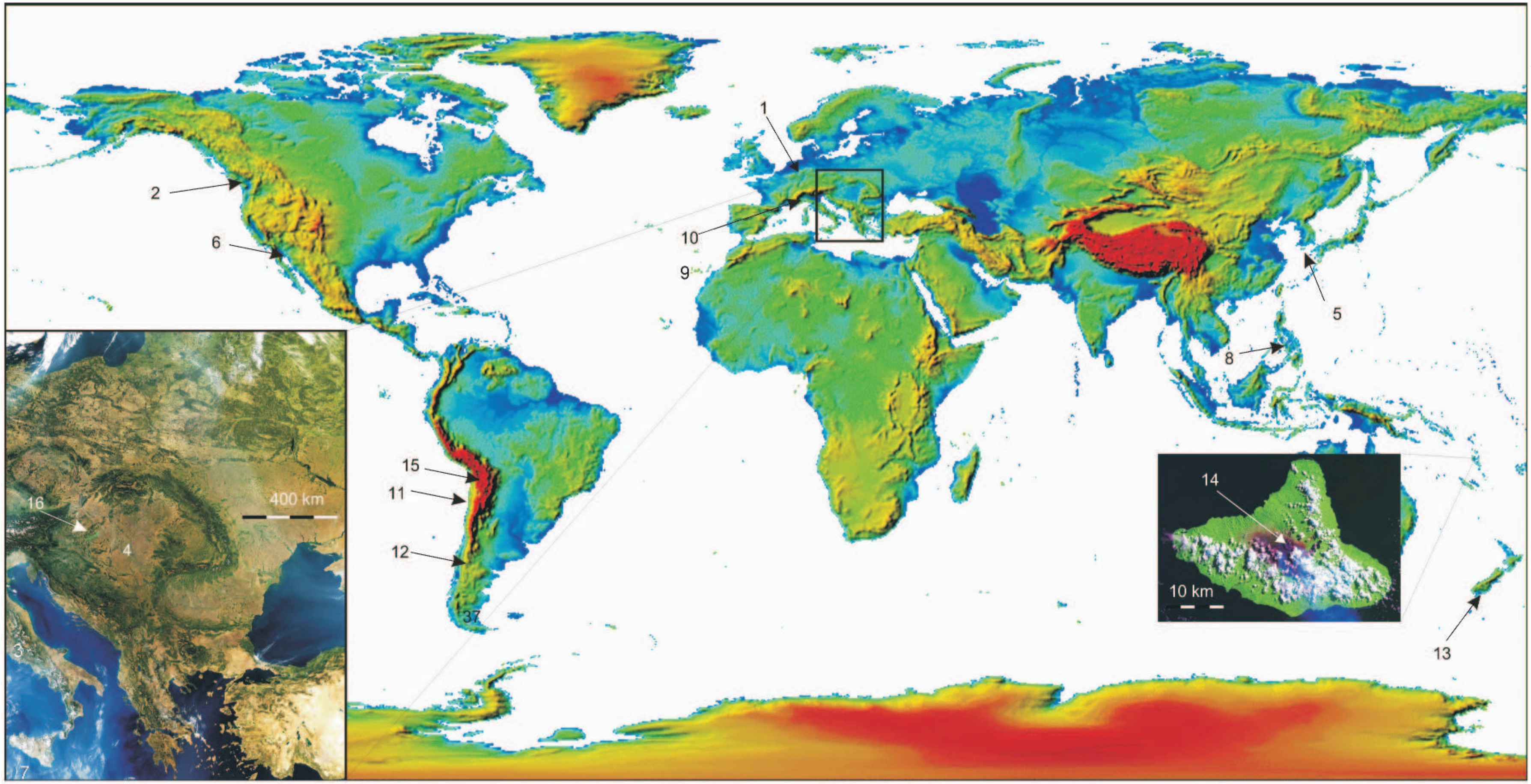
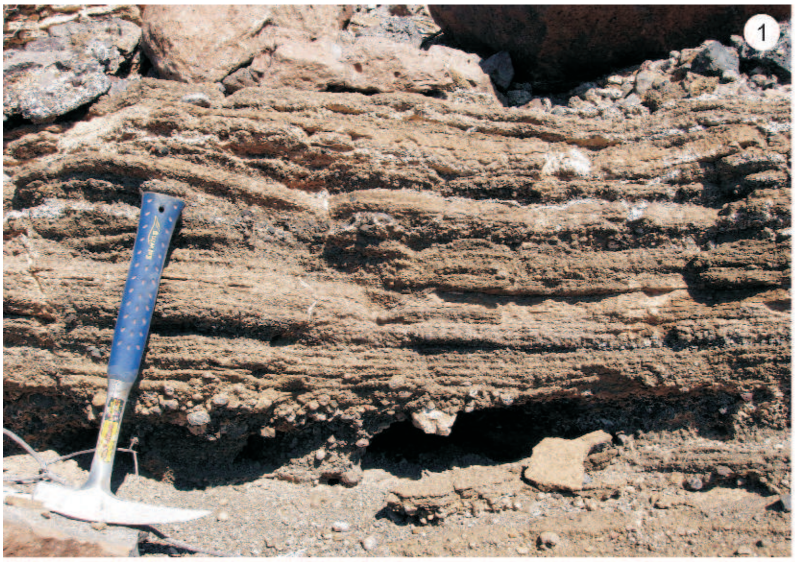
## Location map



1 — Laacher See, Germany
2 — Mount St. Helens, Washington, USA
3 — Campania, Italy
4 — Pannonian Basin, Central Europe
5 — Jeju Island, Korea
6 — Sonora, Mexico
7 — Linosa Island, Italy
8 — Volcan Mayon, Phillippines
9 — Caldera del rey, Tenerife, Spain
10 — Dolomites, Italy
11 — Coastal Coordillera, Chile
12 — Volcan Villarica, Chile
13 — Waipiata Volcanic Field, New Zealand
14 — Ambrym Island (Marum volcano) Vanuatu
15 — Altiplano, Chile
16 — Tihany Peninsula, Hungary

# Plate I



1. Proximal base surge facies from the crater rim of Crater Elegante maar in Sonora, Mexico.
2. Distal base surge facies about 1500 m away of the crater rim of Planchon, Mendoza, Argentina-Chile border.
3. Dune-bedded base surge beds from the distal section of the phreatomagmatic succession along the Snake River, Idaho, USA.
4. Triassic volcaniclastic siltstone (Pietre Verde) bed (greenish bed) interbedded with normal siliciclastic turbidite beds from the Dolomites, northern Italy.
5. Volcaniclastic sandstone successions from the Jurassic La Negra Formation, Northern Chile.
6. Pumiceous deposits from beach setting in southern Tenerife. A) Dewatering pipes in the pumiceous deposits formed when ignimbrite entered to the shallow sea, beach environment, and entrapped water that find escape path to the surface after deposition. B) Cauliflower shape pumiceous lapilli as evidence of sudden contact with cold sea water upon the ignimbrite entered to the sea.

# Plate II

1. Transitional deposits of a block and ash fan of the Calbuco volcano, southern Chile. A) Overview of the thickly bedded, coarse grained volcaniclastic succession, deposited by debris flows from lahars B) Charcoal fragments in pyroclastic flow deposits initiated laharic succession of the Calbuco volcano, southern Chile.
2. Sub-Plinian to Vulcanian eruption cloud over Marum volcanic cone complex in Ambrym Island, Vanuatu. Note the large drifted eruption cloud with well-defined zones of slightly higher cloud. These zones formed in previous Vulcanian explosive eruptions, but the cloud was drifted quickly away. Such pulsating style of eruption can produce characteristic cyclic deposits.
3. Plinian-type of eruptions are associated with initial Plinian-type fall deposits, that can accumulate clast supported, pumice-rich, relatively well-sorted plinian fall deposits such the case in this section near to Rotorua, North Island, New Zealand.
4. Small volume monogenetic volcanoes such as the scoria cones and tuff rings of the Auckland Volcanic Field in New Zealand can form in few hours, and can accumulate tens of metres thick pyroclastic successions.
5. Vent and volcanic conduit dynamics can be closely followed in any pyroclastic succession such as the phreatomagmatic units of the Pula maar in Hungary. Beds richer in accidental lithic fragments represent volcanic conduit collapse and/or vent clearing events in the course of the relatively short-lived volcanic eruption.

# Chapter 5

## Monogenetic volcanism and related features

Small-volume volcanic eruptions are commonly referred to be monogenetic form tephra cones, rings, or mounds consisting of bedded pyroclastic deposits that are composed by fallout, density currents and/or down slope remobilisation of tephra (Connor and Conway 2000, Vespermann and Schmincke 2000).

Monogenetic volcanic fields are commonly related to explosive eruptions (phreatomagmatism) driven by magma–water interaction associated with magma interacting with shallow or deep ground water and/or surface water sources (White 1991a). In many cases seasonal climatic changes as well as the ratio of available surface and ground water play an important role in the formation of different types of volcanic landforms (Aranda-Gomez and Luhr 1996, Carn 2000, Németh et al. 2001, Siebe et al. 2005). From these reasons a great variety of volcanic landforms are expected to develop especially in low lands where the hydrogeology of the country rocks maybe very complex (White 1991a, 1991b).

The resulting volcanic landform especially in such settings is strongly depending on the nature of the pre-eruptive surface (e.g. depositional environment), lithology and mechanical properties of volcanic conduit wall rocks, vent geometry, type and availability of external water (Lorenz 1987). Most volcanological studies on monogenetic volcanic landforms are based on young volcanoes and focus on their short term morphological changes, tephra transport and depositional processes in syn- and post-eruptive times. Due to the erosion the exposed inner architecture of monogenetic volcanoes reveal volcanic litho-facies that bear important information on the eruptive mechanism of the volcano (Lorenz and Kurszlaukis 2007).

Monogenetic volcanic fields commonly consist of large numbers of volcanic clusters and/or alignments that may reach the amount of hundreds of volcanoes in a volcanic field (Connor 1987, 1990, Connor et al. 1992, Condit and Connor 1996, Convay et al. 1998, Connor and Convay 2000, Connor et al. 2000, Valentine et al. 2006). Such large volcano numbers in a volcanic field may evolve over millions of years. Over such long time periods (thousands to millions of years) individual volcanoes may erode significantly, and a volcanic field maybe a group of variable eroded volcanic landforms preserved on a gradually eroding landscape (White 1991b, Konecny et al. 1999, Konecny and Lexa 2000, Németh and White 2003b, 2003a, Lorenz and Haneke 2004). Over longer time periods, a relatively uniform landscape could be dissected, lowered and commonly inverted having preserved clusters of formerly low-lying zones of the syn-eruptive landscapes in elevated positions (Németh and Martin 1999).

In this point of view, monogenetic volcanic fields are very useful to characterize the course of erosion of the surrounding syn-eruptive landscape over the time length during which the volcanic field was active and after the cessation of the last eruption. Especially volcanic erosion remnants of volcanoes in a same age cluster may help to draw a geomorphic horizon for the same time horizon (Németh and Martin 1999). With substantial enough age clusters of volcanic erosion remnants and good enough age data for such clusters a refined erosion history of an often large (hundreds of $km^2$) area can be reconstructed. In this geomorphic reconstruction a combination of general geomorphic consideration of erosion trends of certain volcanic landform such as scoria cones, tuff rings, tuff cones, maars are critical. There are many relatively mechanical ways to re-establish the erosional stage of a certain monogenetic landform and therefore estimate the erosion stage of the syn-eruptive landscape on which such volcano erupted. Especially in older (millions of years) volcanic fields, erosional remnants of phreatomagmatic volcanoes bear important information on the syn-eruptive country rock stratigraphy.

## Monogenetic volcanic fields

Volcanic activity in terrestrial settings often results in the formation of volcanic fields rather than single volcanic edifices (Connor and Convay 2000, Walker 2000). Volcanic fields, especially basaltic ones, are common volcanic systems on Earth (Walker 1993). They can develop as a cluster of small-volume volcanoes such as the Hopi Butte (Plate I, 1) in Arizona (White 1991b), or around a central volcano, such as the volcanic field around Lamongan (Figure 5.1), in Java (Carn 2000). Monogenetic volcanic fields are those in which individual volcanoes (mainly basaltic) commonly form during single episodes of volcanic activity, without subsequent eruptions, while the volcanic field as a whole may be active for millions of years (Walker 1993). Intracontinental volcanic fields commonly are characterized by low magma supply rates over relatively long periods of time (millions of years) (Takada 1994, Connor et al. 2000). They typically consist of scattered volcanic vents that are often considered to be monogenetic as they apparently never made it into the construction stage of stratovolcanism (Walker 1993). In fact, these volcanoes are generally small in size and in volume of accumulated eruptive products, but on closer inspection they do often show signs of multiple eruption histories, and therefore their architecture can be complex regardless of their small size. It is also notable, that volcanic fields in continental settings are often associated with large shield volcanoes and lava flow fields (Greeley 1982, Walker 1993, Hasenaka 1994, Németh 2004).

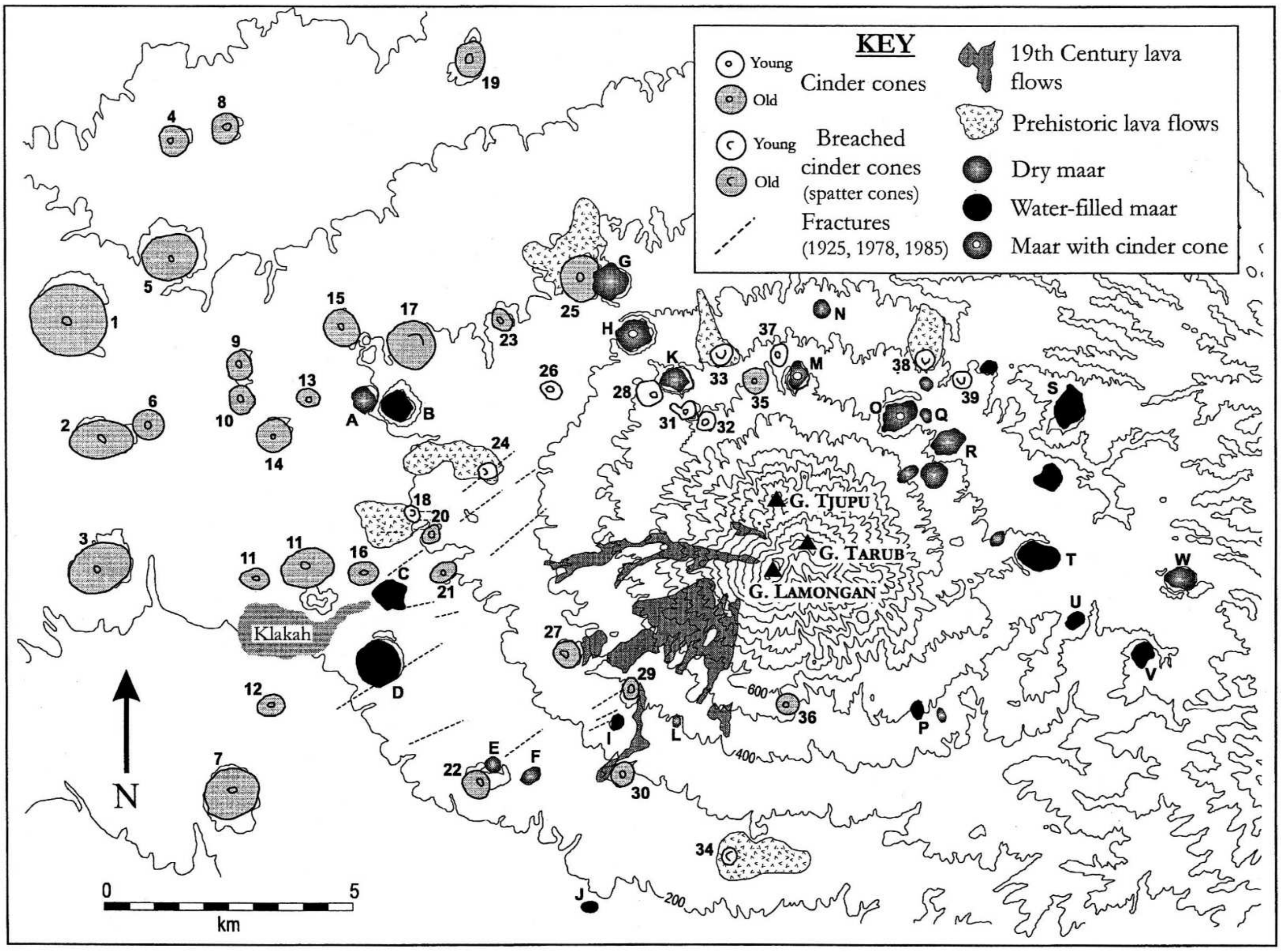


*Figure 5.1.* Volcanic field around the Lamongan stratovolcano in East-Java (after Carn 2000: p. 84)

Phreatomagmatic volcanoes in a volcanic field are commonly associated with low-lands or valleys (Lorenz 1973, 1986, Lorenz and Büchel 1980). Magmatic explosive eruptive centres and extensive lava fields are commonly located in elevated lands or areas with limited water availability. Lava flows are commonly confined to valleys or stopped behind syn-volcanic geomorphic barriers. Lava flows may not leave their source vent zone, forming lava lakes, or filling wide craters of phreatomagmatic volcanoes. Distribution of different styles of vents gives vital information of the syn-volcanic landscape drainage system as well as its physiography (Lorenz and Büchel 1980). Identification of widespread phreatomagmatism in many fields suggests extensive surface and ground water availability of the region in syn-volcanic times. Volcanic fields in terrestrial settings, especially those developed in a fluvio-lacustrine basin such as the western Pannonian Basin Mio/Pliocene fields (Martin and Németh 2004) or the Snake River Plain (Godchaux et al. 1992, Wood and Clemens 2004, Németh and White 2007) volcanic fields are great volcanological interest both from volcanological and palaeogeomorphological point of view. The relatively long volcanic history of such volcanic fields and the adjacent lake\fluviatile environment makes them an ideal area for studying the sublacustrine, peri-lacustrine and post-lacustrine volcanism, the palaeogeomorphological evolution. Such fields have a great opportunity for developing our knowledge about eruption mechanisms resulting from magma–water interactions across the entire magma\water-ratio spectrum with the special relations to the palaeoenvironment, and the related palaeohidrology and physical characteristics of the pre-volcanic units.

Fundamental physical characteristics of volcanic fields that are the focus of current research include 1) the number, type and eruption history of individual vents (Connor 1990, Siebe et al. 2005, Valentine et al. 2006); 2) the timing and recurrence rates of the volcanic eruptions in a given volcanic field (Tanaka et al. 1986, Condit and Connor 1996, Convay et al. 1998), 3) the distribution of vents and volcanic complexes (Connor 1987), and 4) the relationship of volcanic fields and the volcanoes within them to tectonic features such as basins, faults and rift zones (Connor et al. 1992, Stamatakos et al. 1997, Connor et al. 2000). In general there are three major elements to be considered in the ascent and emplacement of magma either on Earth or other planets, and each strongly depends on the physical properties and structure of the lithosphere encountered by the magma. The three factors are (Walker 1989): 1) magma generation and buoyancy 2) rheological boundaries in the lithosphere and 3) density boundaries in the lithosphere. In addition to these factors, the stress field (local and regional) plays an important role in controlling magma ascent which is generally related to the structural features of the lithosphere encountered by the magma.

Monogenetic volcanoes are traditionally referred as those volcanoes that erupt only once during their eruptive history. They are small and occur as scoria cones (Plate I, 2), tuff cones (Plate I, 3) and rings (Plate I, 4), and maars (Plate I, 5). They form from typically short-lived single and brief eruptions. Their characteristic feature is that the duration of the eruption is usually shorter than the solidification time required for the feeding system to provide the melt for the eruption. This definition maybe useful to classify volcanoes that erupt mafic magmas and produce small-volume cinder cones associated with variable long lava flows. Since phreatomagmatic volcanoes such as maars and tuff rings are considered to be the wet equivalent of scoria cones (Lorenz 1986), a general preconception of the short eruption duration of phreatomagmatic volcanoes is accepted. Eyewitness of a few historic maar volcanic eruptions such as Ukinrek (Alaska) (Kienle et al. 1980, Self et al. 1980, Ort et al. 2000) and or Rininahue (Chile) (Müller and Veyl 1956) support this working hypothesis that such eruptions usually last a short time. Calculation of the necessary melt involvement into such eruptions commonly gave a very low proportion of primary magma that needs to create such volcanoes. This conclusion is supported by the componentry analyses of phreatomagmatic tephra in many places in various volcanic settings. To solidify such small volume of melt in the feeding system it may need less time than the a few days or weeks. Conversely, there are reports from scoria cones which show gradual transition toward composite volcanoes, and they are hard to classify in term of monogenetic and polygenetic systems (McKnight and Williams 1997). On the basis of historic eruption it is clear, that most of these volcanoes are large in volume and the deposited tephra commonly consists of a great diversity of eruptive products from strikingly different fragmentation history of the melt (e.g. magmatic vs. phreatomagmatic). Especially phreatomagmatic volcanoes are often associated with scoria cones, spatter cones and lava flows and they form together a volcanic complex. Such volcanic complexes are closely spaced individual volcanoes that individually may fulfill the requirement of the sensu stricto term monogenetic, however, to identify these features in ancient setting maybe problematic if not impossible.

Monogenetic volcanoes may form in any type of geological environment, however, their volcanic landform strongly depends on the water availability and therefore the environment where they erupt. In fully subaqueous environments either in sub-lacustrine or sub-marine settings lensoidally shaped pyroclastic mounds form (White 1996). These mounds consist of flat lying pyroclastic density current deposits (White 2000). In shallow subaqueous environment after the construction of a pyroclastic mound, eruption clouds and directed pyroclast jets breach the water surface and form steep flanked tephra cones over a basal tephra mounds (Belousov and Belousova 2001, White 2001). In fully subaerial settings (Figure 5.2) when magma interact with near surface water tables and/or very shallow lake, sea or river water, gently dipping broad tephra rings develop (Heiken 1971). The tephra ring edifice consists of alternating base surge and phreatomagmatic fall (Plate I, 6) tephras (Vespermann and Schmincke 2000). When magma interact with ground water, hole-in-the-ground, maar volcano forms (Plate I, 7) that is surrounded by flat lying tephra beds (Vespermann and Schmincke 2000). The crater floor of maars undercut the syn-eruptive surface. The maars have four distinct part (Figure 5.3) in cross sectional view (Lorenz 1986, White 1991b, Lorenz and Kurszlaukis 2007); 1) root zone with intrusions that are often mixed with matrix of conduit filling volcaniclastic debris and collapsed country rock blocks; 2) a lower diatreme which is the deep subsurface zone of the conduit filling primary and intra-vent mixing of different origin volcaniclastic debris; 3) an upper diatreme that consists of near surface primary pyroclastic deposits and 4) the crater lake setting that is an accumulation of different origin sediments in a small sedimentary basin with variable influence from background sedimentation. The above mentioned zones of a maar volcano maybe exposed in accordance to the level of erosion of the surrounding landscape (Plate II, 1). Therefore the clear identification and the geometrical consideration of the maar volcanic architecture bear vital information of the estimation of the level of erosion. Using eroded monogenetic volcanic fields to calculate long term erosion and to estimate possible palaeogeography of a region are powerful tools to reconstruct syn-volcanic landscapes. Therefore such works could give vital information of a region‘s landscape evolution. Maar/diatreme volcanoes are often the only sites where already eroded pre-volcanic sediments may have been preserved, thus their study may give information to understand the stratigraphy. Phreatomagmatic explosive phases occur in almost every small-volume volcanoes especially in valley settings with good water availability. Subsequent magmatic explosive eruptions, however, are commonly accompanied with such volcanoes in the waning phase of the eruption history of such volcanoes (Aranda-Gomez et al. 1992). This phase commonly produces large scoria cones accompanied by lava effusion commonly erupting inside the crater (Plate II, 2) (Aranda-Gomez et al. 1992). Phreatomagmatic volcanic fields commonly erupted into fluvio-lacustrine basins, therefore the recognition of great diversity of phreatomagmatic volcanoes in ancient settings almost certainly means that the syn-volcanic landscape is rather basin-like, and/or valley settings. Such an interpretation may alter the extrapolation of the estimated landscape erosion data. It also recently has been highlighted that volcanic activity may rejuvenate from time to time in an otherwise small-volume volcano that has been considered to be monogenetic. This may challenge the „monogenetic“ characteristics of such volcanism. Recognition of such processes in the preserved geological records of such small-volume volcanoes must be taken seriously in the estimation of the landscape erosion of the syn-eruptive settings. In the next paragraphs we list a few basic considerations that may alter the estimation of the certain syn-volcanic landscape erosion level.

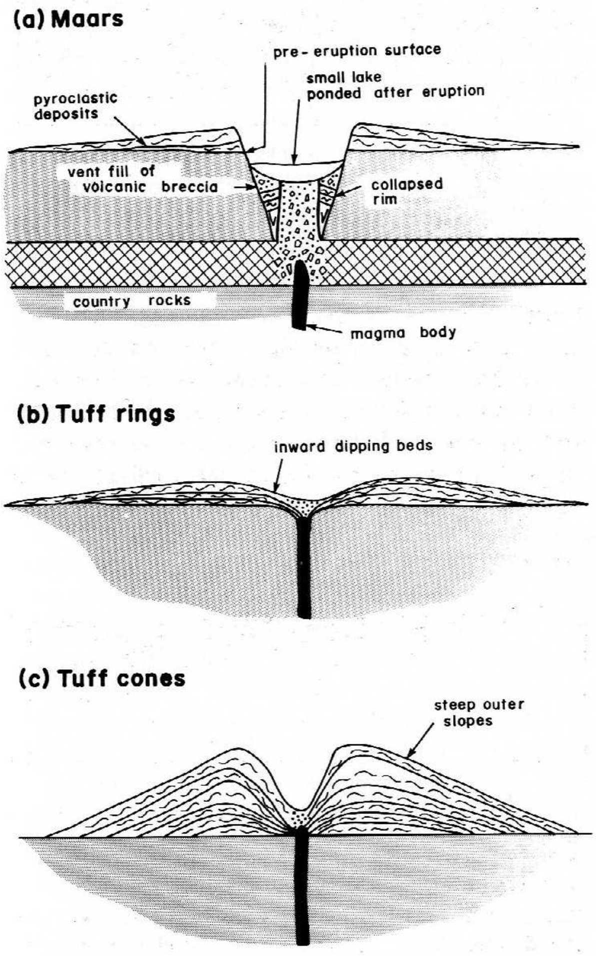


*Figure 5.2.* Schematic cross sections of the three major types of monogenetic volcanoes (after Cas and Wright 1988: p. 377, fig. 13.17)

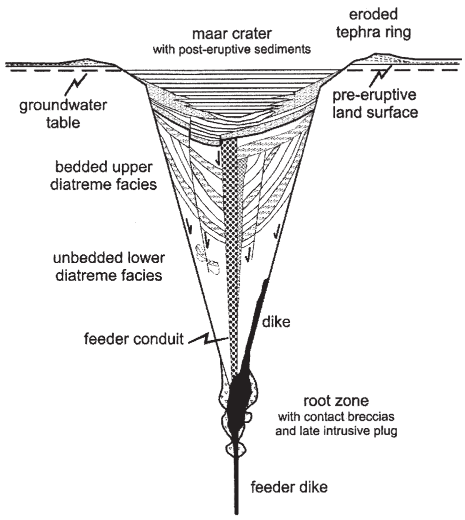


*Figure 5.3.* Cross section of a typical maar (after Lorenz 2007: p. 290, fig. 1)

### Lava fields, spatter cones, and scoria cones

Hawaiian to Strombolian-type eruptions build up spatter, scoria and/or cinder cones ((Plate II, 3). Eruption rate, volatile content, magma composition and temperature are the most obvious controlling factors during their eruption (Houghton and Schmincke 1989, Houghton et al. 1999, Vespermann and Schmincke 2000). 95% of observed cinder-cone eruptions lasted less than a year in contrast to composite volcanoes formed from multiple eruptions over thousands of years — an important notion in view of the hazard assessment. Comparative morphology of scoria cones is a useful dating tool, however, new researches suggest that their erosion could be more complex. Rare basaltic Plinian eruptions are poorly known but dangerous volcanic phenomena. Shield volcanoes (Plate II, 4) are common and give the major sources of lavas in intraplate provinces (WALKER 1993). Eruptions of large Hawaiian-type volcanic centers usually related fissure-vent systems (Plate II, 5), but in small plains-basalt province eruptions related to central vent systems (JOHNSON 1989), but there are several examples where shield volcanoes developed along basement fissure systems (JOHNSON 1989). The individual lavaflows associated with intracontinental volcanic fields tend to be around 5 to 10 km long however exceptionally long lava flows are also known (CONNOR and CONVAY 2000, KILBURN 2000). The total thickness of the lava cover could reach several tens of meters and could cover significant portions of a volcanic field. Adjacent to the lava field Strombolian scoria cones and Hawaiian spatter cones are common features in a continental volcanic field. These volcanoes are commonly point sources of lava flows (Figure 5.4). Basaltic volcanic fields commonly accompanied by extensive lava fields ranging from aa' to pahoehoe types of lava flows (KILBURN 2000). Lava flow fields bear characteristic surface morphological features such as tumuli (Plate II, 6), sky rise, whale back humps, lava tubes, and pressure ridges (KILBURN 2000). Such surface morphological features are commonly large in dimension (tens of metres) and characteristic for eruption rate, composition, preflow morphological features and composition of the flow (KILBURN 2000). Among these features, tumuli, are whale-back-shaped uplifts are common in most of the pahoehoe lava flow fields, such as the Deccan, India (DURAISWAMI et al. 2001), Hawaii, USA (WALKER 1991), Etna, Italy (CALVARI et al. 2003), Iceland (MATTSSON and HOSKULDSSON 2005), or in Eastern Australia (OLLIER 1964, WILMOTH and WALKER 1993). Tumuli are positive topographic features with wide range of slope angles and surface morphological features that are common on pahoehoe lava flow fields. Commonly three types of tumuli are distinguished on shield volcanic systems, such as (1) lava-coated tumuli, (2) upper-slope tumuli and (3) flow-lobe tumuli in accordance to their distance from their source (ROSSI and GUDMUNDSSON 1996). Flow-lobe tumuli are common in the medial and distal parts of pahoehoe flow fields, whereas the other two tumulus types are more frequent in the proximal parts of the flow fields (ROSSI and GUDMUNDSSON 1996). The flow- lobe tumuli are significantly larger, have shallower flanks, and do not have extensive outflows from the cracks in comparison to more proximal types of tumulus (ROSSI and GUDMUNDSSON 1996). They commonly group into cluster in gentle slopes of shield volcanoes, such Hawaii (WALKER 1991). Large tumuli are comparable in size to small lava spatter cones, and therefore their recognition from morphological point of view is important to establish the eruption history of volcanic fields,. It has been demonstrated that recognition of tumuli features and their characteristic surface textures may give hint to quantify eruption duration (MATTSSON and HOSKULDSSON 2005). There are thin lava foot breccias between the individual lava beds.

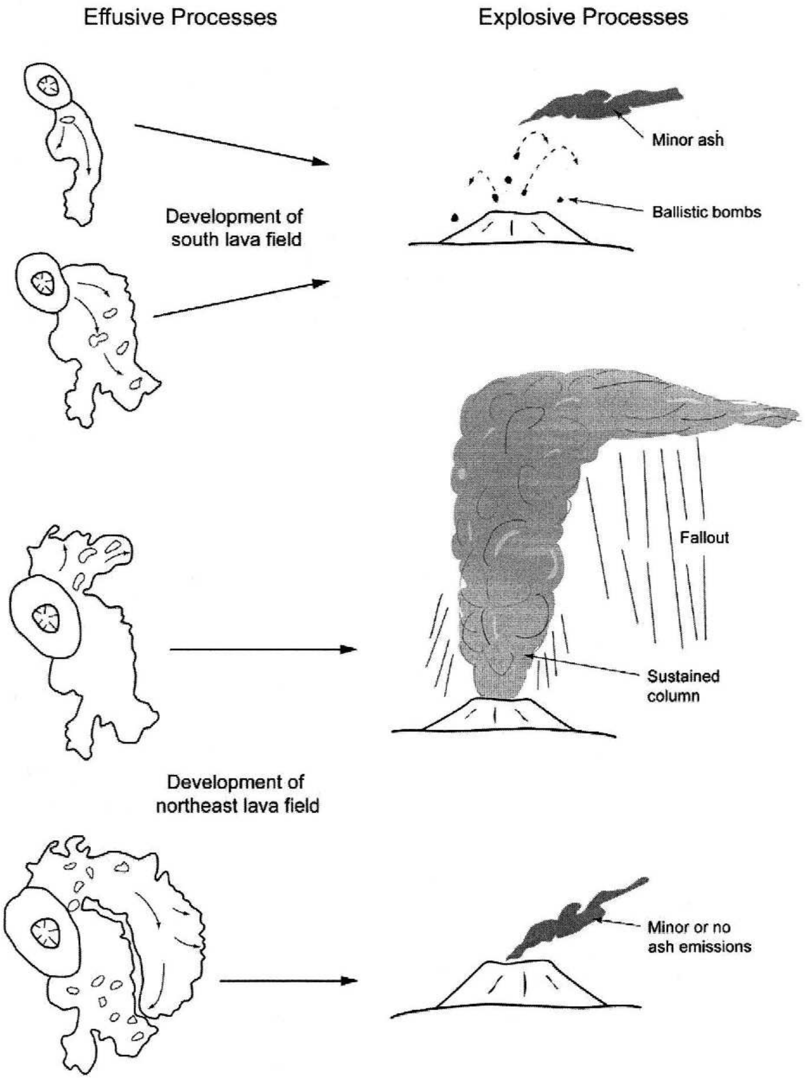


***Figure 5.4.*** Interpretation of effusive and explosive eruptive processes responsible for the formation of the Lathrop Wells volcano in the Yucca Mountain, Nevada (VALENTINE et al. 2006: fig. 17)

Spatter cones consist of a near vent strongly baked, red, slightly bedded sequences with large spindle or highly vesiculated fluidal bombs (Plate III, 1). These deposits usually reflect strong reworking of volcaniclastics in near vent position. Spatter cones and scoria cones can build up steep spatter and agglutinate piles, that can collapse gravitationally (Figure 5.5), or driven away by moving lava flow initiated from the flank of the cone (Plate III, 2).

Strombolian scoria and spatter deposits are common in relation with maar volcanic centers. Even maar volcanic centers may produce phases of Hawaiian and Strombolian-style eruptive activity from several distinguished eruption points, leading to agglutinates or even clastogenic lavas. There are examples in the Tihany Volcano maar volcanic

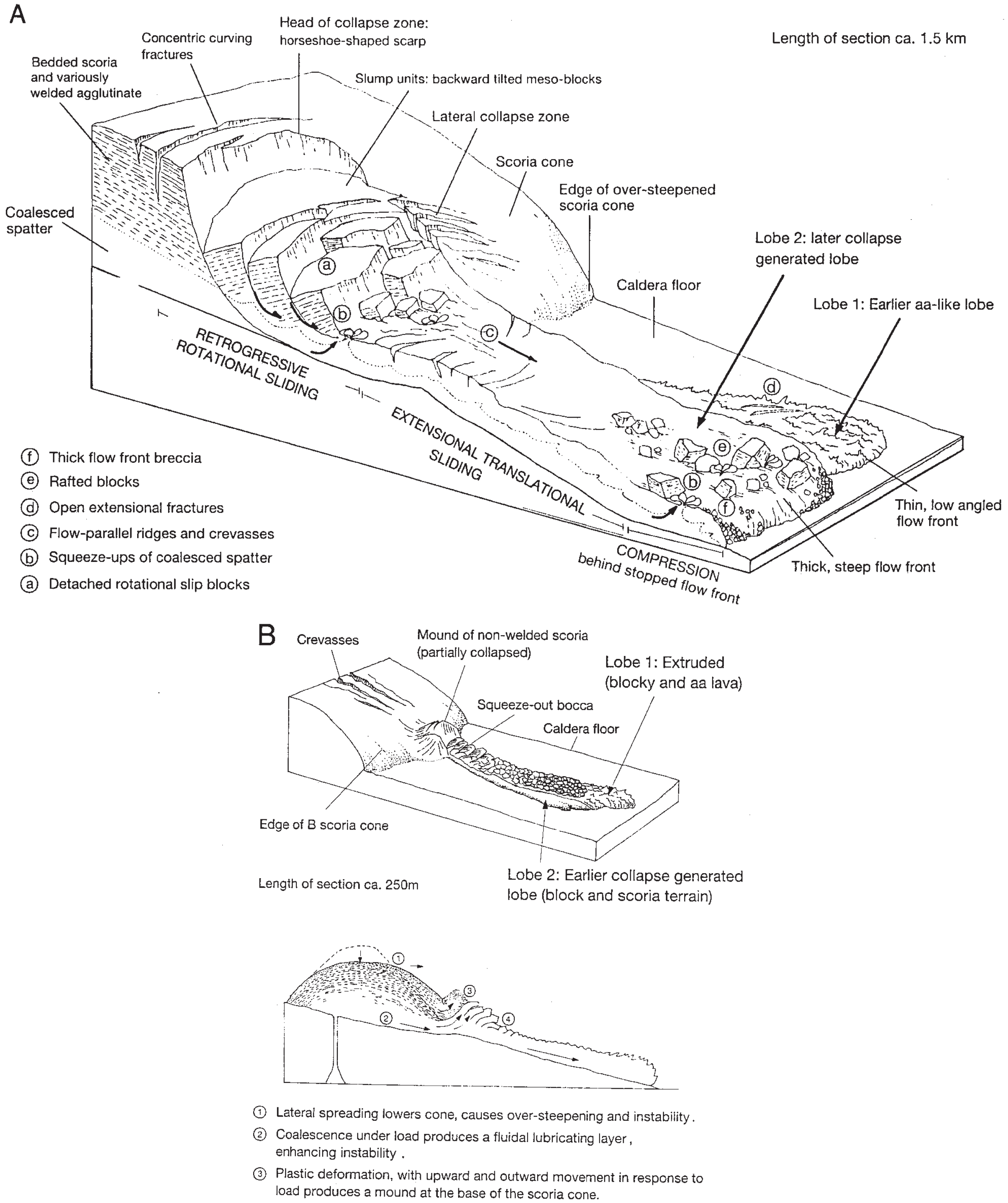


***Figure 5.5.*** Model of spatter rampart collapse after Sumner 1998

complex for this kind of deposits in the northern part of the maar complex (Gödrös–Diós) (Németh et al. 1999). A remnant of Strombolian scoria cone in the Füzes-tó region in the Bakony–Balaton Highland Volcanic Field, Hungary shows near vent scoriaceous volcaniclastic breccia in peperitic matrix representing water saturated slurry in the vent during the Strombolian activity (Németh and Szabó 1998).

Magmatic explosive and/or degassing processes as result of the fragmentation of the uprising mafic magma leading the formation of scoria cones with common welded core zones (Vespermann and Schmincke 2000). The textural characteristics of the pyroclasts, such as high vesicularity, fluidal shape, and the dark, often red colour (Plate III, 3) indicate a magmatic degassing and fragmentation history due to Strombolian-style explosive eruptions (Jaupart and Vergniolle 1988, Vergniolle and Brandeis 1996, Vergniolle et al. 1996, Sumner 1998, Vespermann and Schmincke 2000). The closely packed, slightly oriented texture of lava ash and lapilli rich pyroclastic rocks are interpreted to be the result of Hawaiian-style lava fountaining (Thordarson and Self 1993, Vespermann and Schmincke 2000, Wolff and Sumner 2000, Sumner et al. 2005). The common intercalation of scoria beds of scoria cones with welded fall out deposits and/or clastogenic lava flows indicate a sudden and common change in eruption style from Strombolian to Hawaiian and vice versa (Plate III, 4) (Parfitt and Wilson 1995, Parfitt et al. 1995, Wilson et al. 1995). The presence of pyroclastic breccias, lapilli tuff and tuff interbeds in scoria cones are common signs of a phreatomagmatic influence on the eruptions (Houghton and Hackett 1984, Houghton et al. 1991, Doubik and Hill 1999, Houghton et al. 1999).

Scoria cones are the most common subaerial volcanic landforms on Earth and are generally considered to be a result of mild explosive eruption of mafic magmas in a short period of time (days, weeks) (Vespermann and Schmincke 2000), however long-lived scoria cone eruption such as Parícutin in Mexico was active for 9 years (Luhr and Simkin 1993). In spite of the numerous scoria cones associated with volcanic fields and central volcanoes (e.g. along rift zones) there are only a few detailed studies that have been carried out on their architecture (McGetchin et al. 1972, Chouet et al. 1974, McGetchin et al. 1974, McGetchin and Settle 1975, Head and Wilson 1989, Riedel et al. 2003). Scoria cones inner part usually consists of welded agglutinate (Figure 5.6). Such welded part usually more resistant against erosion and can be preserved long time (Plate III, 5). Detailed analyses of deposits preserved on scoria cones lead to the clarification of the role of the shallow seated magmatic system in the control of the explosive eruptions of such volcanoes (Houghton et al. 1999). Among the identified parameters the variations in degassing patterns, magma ascent rates and degrees of interaction with external water are thought to be responsible for sudden changes in the eruption sequence from deposits representative for "wet" and "dry" eruption conditions (Houghton et al. 1999). In general scoria cone-forming eruptions are linked to Strombolian-type activity driven by magmatic fragmentation occurring in the near surface region of the open volcanic conduit (Blackburn and Sparks 1976, Houghton et al. 1999). Among scoria cones a great variety has been observed and described, which show gradual transitions between Hawaiian lava fountaining (Figure 5.7) to moderate Strombolian-type eruptions. It has been suggested that magma ascent speed is the most important factor causing such transitions, with gas content and viscosity also influencing the ascent speed at which the transition occurs (Parfitt and Wilson 1995, Parfitt et al. 1995). A decrease in gas content does not cause a transition from Hawaiian to Strombolian activity, but instead causes a transition to passive effusion of vesicular lava (Parfitt and Wilson 1995). Some authors suggest that a change from Hawaiian to Strombolian-style requires a significant reduction in magma ascent speed (Parfitt and Wilson 1995). Among many of the recently identified deposits related to basaltic explosive volcanism Tarawera (New Zealand) 1886 eruption is a classical example of characteristic mafic (sub)-Plinian-style (Houghton et al. 2004). "Violent Strombolian" eruptions are explosive eruptions of mafic magma characterised by eruption column heights <10 km, voluminous ash production, and simultaneous lava effusion (Houghton

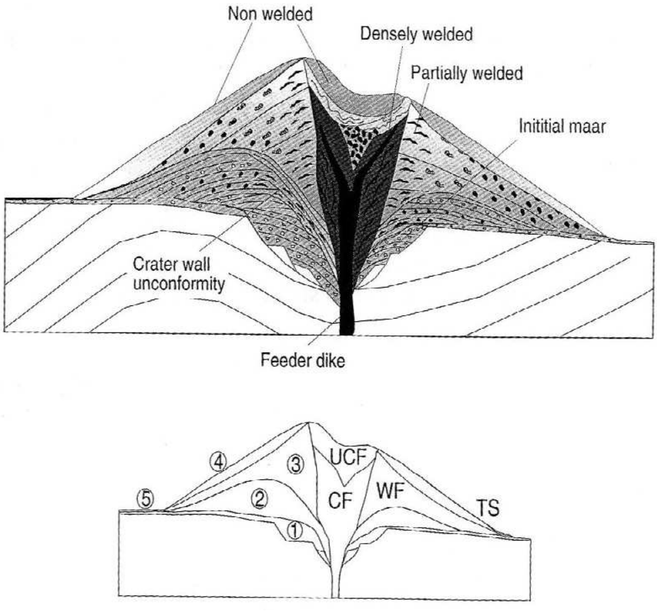


***Figure 5.6.*** Scoria cone structure (after Vesperman and Schmincke 2000)
1 — initial phreatomagmatic units, 2 — strombolian units with intercalated phreatomagmatic beds, 3 — strombolian eruption formed cone facies, 4 — post-strombolian talus, 5 — distal fallout tephra, CF = crater facies, UCF = upper crater facies, WF = wall facies, TS = talus slope

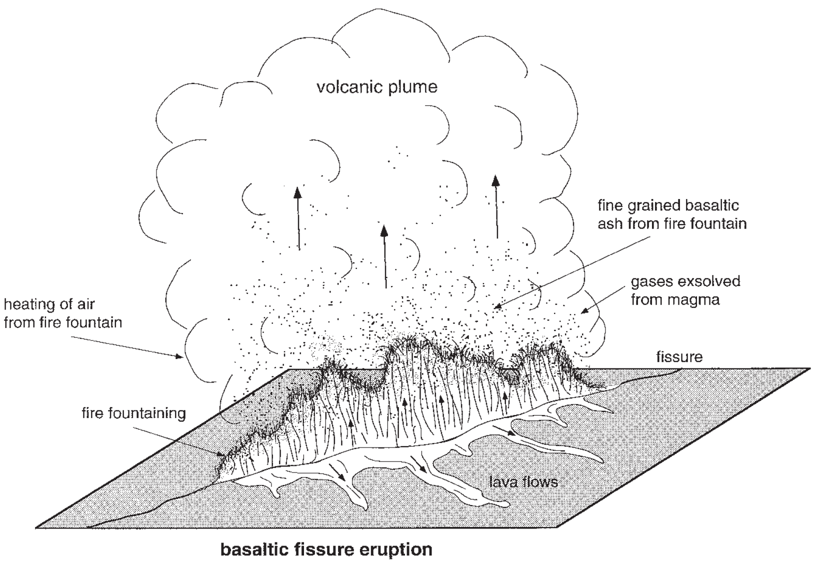


***Figure 5.7.*** Hawaiian lava fountaing model (after Carey and Bursik 2000)

et al. 2004). The mechanism of generation, fragmentation, transportation and deposition of ash in these eruptions is poorly understood, however such eruption was just recently identified to have occurred not only during eruption of composite volcanoes but also in single scoria cones. The best example for such eruptions is Volcán Parícutin, Mexico (1943–1953). Eruptive products of Parícutin eruption include thick tephra deposits of alternating ash and lapilli beds few kilometres from the cone building pyroclastic units (Luhr and Simkin 1993, Newton et al. 2005). Comparative studies indicated that other young monogenetic cones in the Michoacán, Mexico region may have erupted in a similar style as Parícutin (Hasenaka and Carmichael 1985). Further field evidences from the western Transmexican Volcanic Belt (Plate III, 6) indicate that mafic violent Strombolian-style eruptions among scoria cones may be more common than expected (Martin and Németh 2006).

## Maar – diatreme volcanoes

Maars are small-volume volcanoes that are the second most common volcanic landforms on Earth (Lorenz 1985 1986). Maar volcanoes have characteristically wide and deep craters, commonly referred as ‘hole-in-the-ground’ features (Lorenz 1985). Magma and water explosive interaction is considered to be the main driving force that causes their creation (Lorenz 1986, Zimanowski et al. 1991, Zimanowski et al. 1997, Buttner et al. 2002). The sudden formation of steam during magma and water interaction produces a steam explosion that disrupts the country rocks and creates a mass deficit in and around the explosion centre, which leads to the formation of a collapse crater. The explosion locus gradually migrates downward (Figure 5.8) due to the gradual exhaust of water sources to fuel magma–water interaction in the explosion sites (Lorenz 1986). Maar volcanoes are not a strictly or well characterised group of volcanoes. There is an almost continuous transition to scoria (cinder) cones, which are considered to be the most common volcanic landforms on Earth. Scoria cones have frequently had only a short-lived eruptive history, when magma– water interaction took place. In this respect, maars may be viewed as ‘wet’ equivalents of scoria cones. Maar volcanoes are complicated, but small volume volcanoes if we consider that they commonly form groups or clusters, or have structurally-controlled alignments. In volcanic fields, the volcanoes commonly fall between the two end-members such as maars and scoria cones and therefore the study of such landforms should include both.

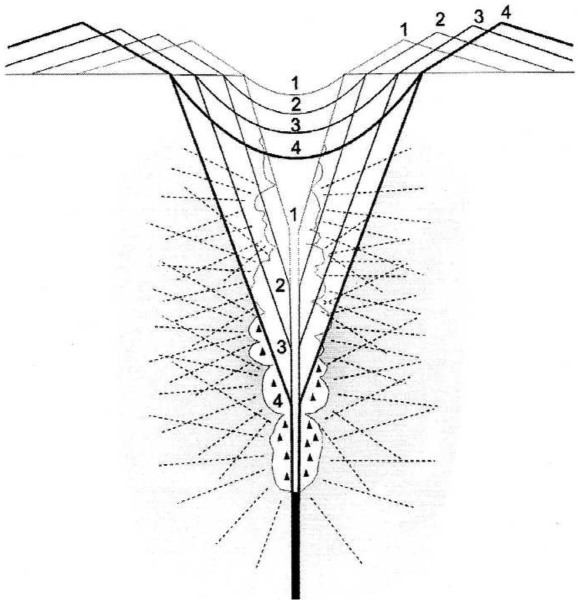


***Figure 5.8.*** Theoretical model of the evolution of a maar due to downward migration of explosion locus in time (after Lorenz 2007: fig. 2)

There exist two models on the formation of maar-diatreme volcanoes: the magmatic model and the phreatomagmatic model. The magmatic model is especially concerned with ultrabasic, ultramafic and carbonatitic magmas. It invokes volatile rich fluid magmas which, close to the Earth´s surface, fragment the country rocks thus forming progressively from deeper levels to almost the surface the irregular shaped root zone. Explosive breakthrough to the surface is supposed to result in the formation of the maar crater and then via downward propagating fluidization of the root zone contents to shaping of the cone-shaped diatreme, and mixing of the diatreme clasts (WOOLSEY et al. 1975, MCCALLUM et al. 1977, WOLFE 1980, MITCHELL 1986).

The phreatomagmatic model, in contrast, invokes explosive interaction of rising magma with ground water, originally close to the surface and then downward penetration of the sites of explosion (LORENZ 1985). The various individual explosion sites or chambers jointly form the root zone. Ejection of explosively fragmented country rocks leads to a mass deficiency and consequently to collapse of the overlying rocks. Via these processes the diatreme forms and, in principle, it represents a collapse feature like a sink hole. Downward explosive penetration of the root zone on its own feeder dyke and consequent collapse phases of the diatreme leads to a growing diatreme and a growing maar crater above (LORENZ 1986). Phreatomagmatic explosions and eruptions represent the key mechanism for the formation of maar-diatreme volcanoes, irrespective of magma composition or host rock type. However, it is environmental conditions (including magma and host-rock characteristics) that apparently cause quite a variability in size and shape of maars and the characteristics of their deposits. Furthermore, phreatomagmatism in maar volcanoes can occur with eruptive stages of purely magmatic explosivity, mixed magmatic/phreatomagmatic eruptions, or even final stages of lava lake effusion. Because the formation of maars is driven by the interaction of ground water and uprising melt, after the collapse of the crater, ground water inflow quickly fills the volcanic depression and creates a deep crater lake (BÜCHEL and LORENZ 1993). Maar lakes are steep sided, often surrounded by unstable steep tephra cliffs that can erode into the lake quickly (Plate IV, 1). Maars produce complex deposits both inside and outside of their craters. The record of the eruption is complicated by complex interactions between magmatic and phreatomagmatic processes in the conduit, transport processes in the vertical and lateral currents, and depositional processes.

A diatreme is the substructure of a maar crater and its tephra ring (LORENZ 1986, WHITE 1991b, LORENZ and KURSZLAUKIS 2007). In maar-diatreme volcanoes a large amount of fragmented country rocks and commonly juvenile lapilli and bombs are ejected. Formation of these lapilli and bombs is attributed to thermohydraulic explosions in the root zones of the diatremes. Sections of fragmented country rocks may collapse or slide into the partially evacuated root zone and thus form a mass-flow deposit subsequent to each explosion (LORENZ et al. 2002, LORENZ and KURSZLAUKIS 2007). In Patagonia, Argentina (Plate IV, 2) for instance in the center of the Patagonian mafic Cenozoic plateau lava fields newly discovered diatremes stand about 100m above the surrounding plane exposing the lower diatremes of former phreatomagmatic volcanoes and their feeding dyke systems (MARTIN et al. 2005). Diatremes themselves are cone-shaped volcanic structures cut into pre-eruptive rocks. They are up to 2.5 km deep and up to 1-2 km in upper diameter. They are filled by clastic debris, subsided larger blocks and frequently intrusive rocks. The volume of the diatreme fill is about the same as that of the thinly bedded tephra ring and distal ash deposits. Thus, diatremes form an important part of the maar-diatreme volcano. The rather regular cone-shaped diatremes continue at depth into a root zone. This rootzone is irregular in shape and overlies the magmatic feeder dyke of the volcano (LORENZ and KURSZLAUKIS 2007). Uprising magma from the underlying feeder dyke into the diatreme root zone intrudes the clastic debris in the diatremes, inflates them and mingles with the debris to commonly form subterranean peperite (LORENZ et al. 2002, LORENZ and KURSZLAUKIS 2007).

In "soft substrate" environment (Figure 5.9) maar volcanoes are broad and underlain by "champagne glass" shape diatremes (LORENZ 2003, SOHN and PARK 2005, AUER et al. 2007). In contrast the crater wall of maar volcanoes that erupted through "hard rock" environment (Figure 5.9) will be steep, fulfilled with volcaniclastic delta deposits and underlain by deep diatremes (LORENZ 2003, SOHN and PARK 2005, AUER et al. 2007). Maar-diatreme volcanoes are associated with any magma type involved in volcanism.

The West Eifel Volcanic Field, Germany, represents a classic example of how pre-existing country rock structures influenced and controlled the position and emplacement behaviour of rising magma in the uppermost crust (SCHMINCKE 1977, LORENZ and BÜCHEL 1980, LORENZ 1984, BÜCHEL 1993). The West Eifel maar-diatreme volcanoes were formed by phreatomagmatic activity on the intersections of mostly basaltic dykes with local water bearing faults or joints especially, but not exclusively, underneath fault-controlled valley floors (Figure 5.10) (LORENZ 1984). If meteoric water was not available, scoria cones were formed on dykes by magmatic activity, without the formation of diatremes (LORENZ 1984).

Kimberlites and carbonatites also seem to follow an emplacement pattern that appears to be controlled by country rock structures (DOWNES et al. 2007). While their final emplacement mechanism is still under discussion (the major diatreme-forming process is thought to be related either to the expansion of a juvenile gas phase or to the interaction of the rising magma with ground water), work published mainly over the last two decades has shown that the position of many of these economically important pipes is to at least some extent controlled by crustal discontinuities. Deep-seated shear zones, faults, mobile belts, transform faults, arch-style uplifts of pre-existing basement structures and the general stress field at the time of emplacement of these magma types have been quoted as being responsible, or at least as having influenced, the emplacement of pipes and dykes (JAQUES and MILLIGAN 2004, JONES and CRAVEN 2004, GRIFFIN et al. 2005). Of interest, particularly for exploration purposes, is the extent to which the

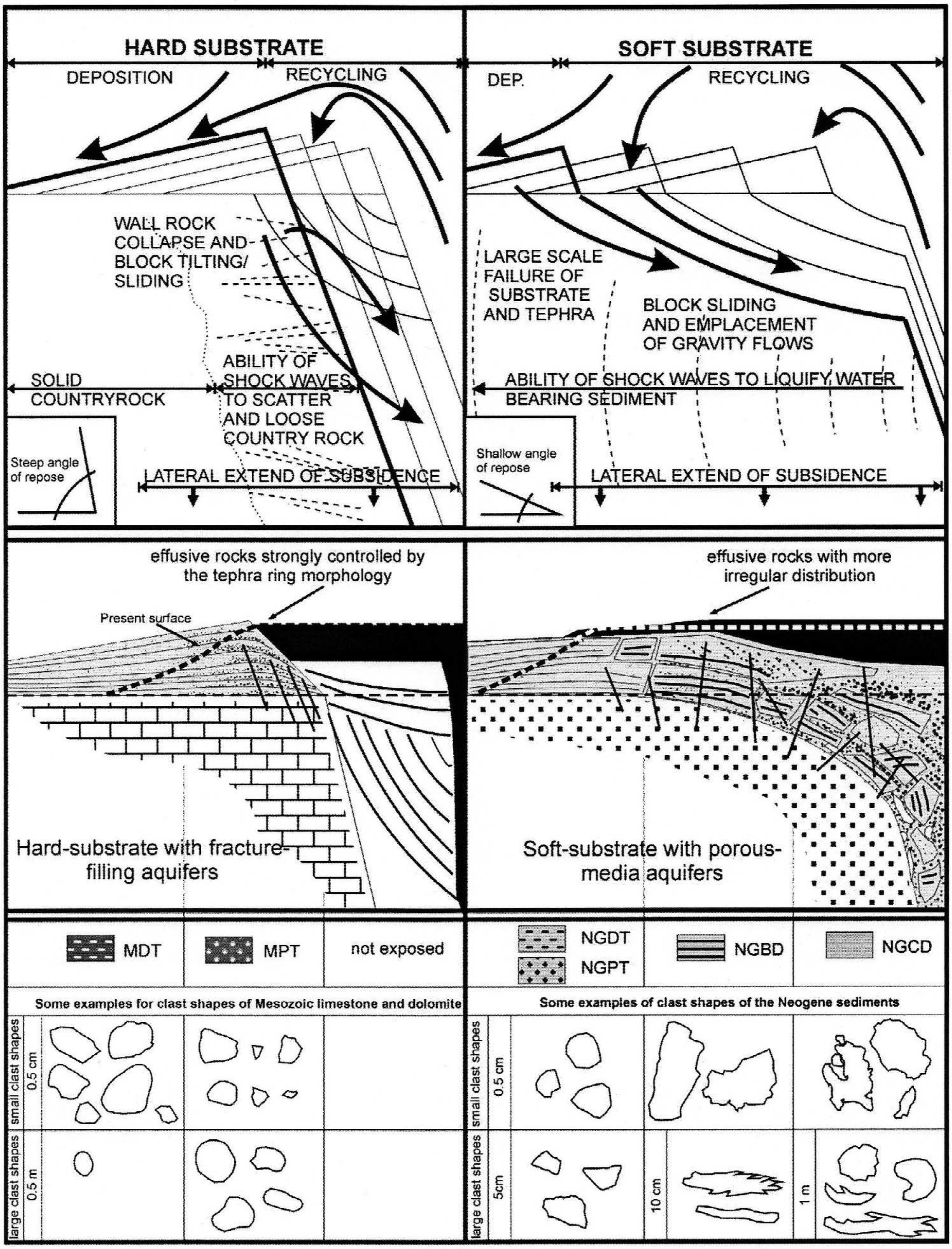


*Figure 5.9.* Theoretical model of maar crater evolution in hard and soft substrate environment based on study of the Fekete-hegy maar volcanic complex in western Hungary (after Auer et al. 2007)

position and emplacement of a kimberlite pipe is controlled by country rock structures. In addition, it is important for mining purposes to define the interaction of the pipe shapes with inhomogeneities in the country rock. The morphology of root zones of maar-diatreme volcanoes seems to be particularly susceptible to country rock faults and joint patterns. Depending on magma type involved, locality and state of erosion, maar-diatreme volcanoes may be of

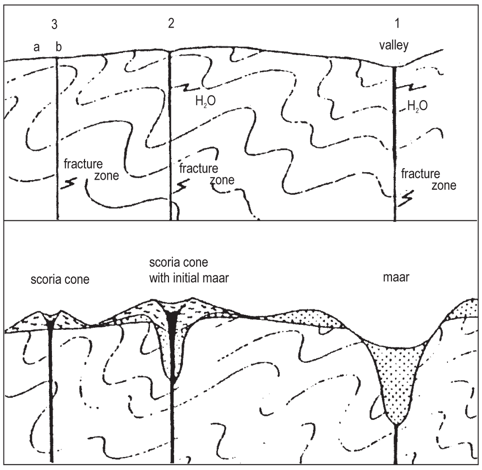


***Figure 5.10.*** Theoretical model of the formation of maar volcanoes along hydrologically active zones in the Eifel, Germany (after Lorenz 1984)

economic relevance. The economically most relevant maar-diatreme volcanoes are diamondiferous kimberlite and lamproite pipes which occur on all major cratons. In the West Eifel, Germany, some diatremes underlying maar craters channel $CO_2$ towards the surface which may be used in fizzy water or soft drinks, or for purely industrial purposes. The pyroclastic rocks of diatremes and maar tephra rings may represent material suitable for use as road metal and similar purposes.

## Tephra ring deposits

The tephra ring successions are commonly interpreted to be deposited predominantly from pyroclastic density currents such as base surges (Plate IV, 3) and deposited through a gradual loss of energy as well as input of fall material into the passing base surge currents (Dellino et al. 1990, 2004a, 2004b, Dellino 2000, Dellino and La Volpe 2000, The interbedded scoriaceous fall deposits are common in tephra ring successions (Plate IV, 4) and are interpreted to be a result from an intermittent change in eruptive style, where water availability to fuel magma–water interaction was suppressed (Houghton et al. 1999). In the distal sections, the base surge dominated beds are gradually replaced by reworked dm-thick beds of fine ash deposited from syn-volcanic reworking by debris flows and/or hyperconcentrated mass flows (Plate IV, 5) (Lajoie et al. 1992, Sohn 1996, Vazquez and Ort 2006). Facies variation in tuff ring sequences have been recognized characteristic changes. From the Crater Elegante in Mexico proximal sandwave bed facies replaced by massive facies in the medial regions, and the by planar bed facies in the most distal regions (Figure 5.11) (Wohletz and Sheridan 1979). A facies variation of a single surge unit from the Hopi Butte recognized different trend; having disorganized beds in the rim section, stratifified beds and sandwave beds in the medial sections,and plane parallel beds in distal regions (Figure 5.12) (Vazquez and Ort 2006). In the distal areas the bedding tends to be subhorizontal, with no dramatic undulations of bed contacts. The number of impact sags or accretionary lapilli-bearing beds in the distal sections of the tephra ring succession become less common. The beds are laterally continuous over tens of metres. The common inter-fingering of phreatomagmatic deposits with syn-volcanic reworked volcaniclastic sediments in the tephra ring successions generally indicates an ongoing remobilisation of freshly deposited tephra during the eruption. This remobilisation of freshly deposited tephra is best to be inferred to be similar to sheet erosion on the flank of the volcano. Mud flows and debris flows in the distal areas form tabular beds mantled by predominantly base surge generated tephras. The proportion of reworked tephra in the volcaniclastic succession is increasing with distance from the vent.

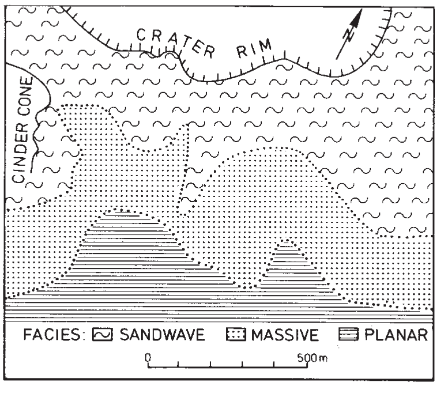


***Figure 5.11.*** Facies distribution around the Crater Elegante, Mexico (after Wohletz and Sheridan 1979)

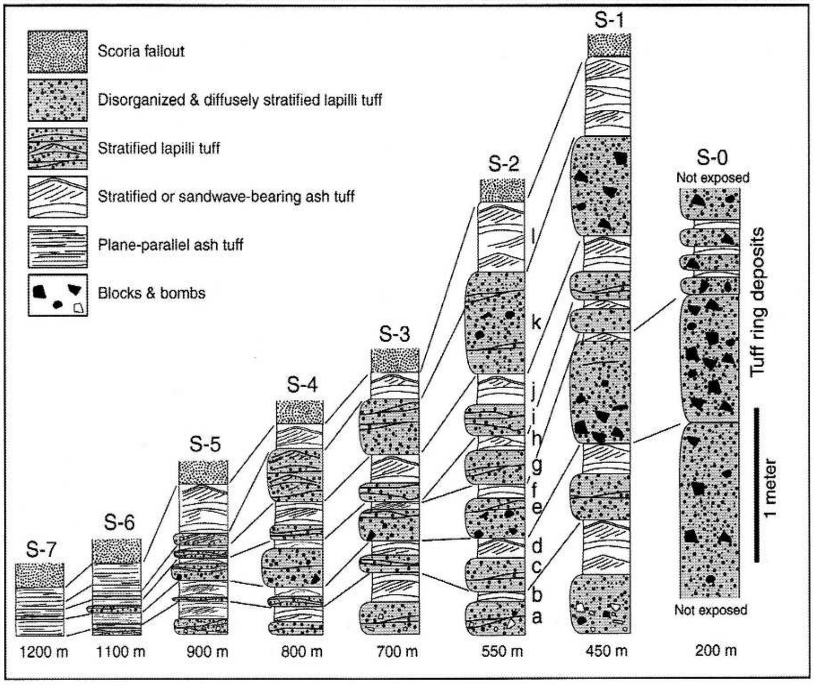


***Figure 5.12.*** Facies distribution in measured stratigraphic logs from the Hopi Butte, Arizona (after Vazquez and Ort 2006: fig. 3)

## Maar lakes

When the phreatomagmatic explosions of a maar-diatreme volcano finally come to an end, the crater fills up with water. The resulting maar lakes are deep in relation to their diameter and damned from the surroundings by the ejected material (crater wall)(Plate IV, 6). This special architecture effects the lake and its sediments. These lakes trap the material of an extreme small catchment area (Wilde and Frankenhauser 1998, Zolitschka et al. 2000, Scharf et al. 2001, Mingram et al. 2004a, 2004b, Kotthoff and Schmid 2005, Sabel et al. 2005). Allochthonous clastic material reaches the crater mainly as turbitity currents originating from the crater rim (ejected material) (Drohmann and Negendank 1993). The autochthonous sediment in maar lakes is often dominated by algal material. Algal bloom layers alternate with the background sediment layers creating laminated deposits (Mingram 1998, Belis et al. 1999, Brukner-Wein et al. 2000, Schabetsberger et al. 2004). And maar lakes often develop a meromiktic division of their water column providing in this, way exceptional conditions for the preservation of sedimentary structures as well as fossils. During the post-eruptive history of a maar volcano, landslides, crater-wall collapses and the initiated volcaniclastic debris flows and turbidity currents form a typical lacustrine succession in the maar lake (Plate IV, 7). This destructive period is often accompanied by a lengthy period of quiet, during which the deposition of suspended material may produce laminated sequences. These laminites are often characteristic of the palaeoenvironment, and therefore maar volcanoes are commonly considered to be excellent sites in which continental depositional records can be well preserved (Vazquez et al. 2004, Crausbay et al. 2006, Garcin et al. 2006).

Large tuff rings and maars may host significant thickness of lacustrine sediments accumulate over long time (thousands of years) in their crater. In the western Pannonian Basin for instance at least 3 maar crater are known where few tens of metres thick lacustrine succession is preserved (Martin and Németh 2004). Such lacustrine sediments are wet, and having significant pore space, and over time the thickness of the succession could drop significantly, that may cause significant subsidence of the crater fillings (Suhr et al. 2004). Study of such lacustrine successions bear many evidences of such processes (Suhr et al. 2004), in a form of convolute and distorted beddings, water escape structures or unusual blocks upon each other (Németh et al. 2002, Csillag et al. 2004). After erosion and exhumation of the lacustrine succession the position of the lacustrine succession maybe used to estimate the original position of the syn-eruptive landscape on what the volcano developed. However, the gradual subsidence and compaction of the crater filling succession must be taken in account for correct estimate of the syn-volcanic surface estimates (Figure 5.13). In any case when subsidence and/or truncation of the lacustrine succession are recognised, the position of the syn-eruptive surface could be significantly higher than it maybe expected from the pure basis of the size and sedimentary features preserved in the succession.

In case a phreatomagmatic (e.g. maar) volcano forms, its crater may reach up to a few km in diametre. Such a size of crater could be filled with a significant amount of water (tens of metres range). In such crater lakes, in case of rejuvenation of volcanic activity, eruption may take place purely subaqueous to emergent (Németh et al. 2007). After erosion volcaniclastic sediments could accumulate in the crater lake forming complex sedimentary facies associations similar to those forming in normal subaqueous sedimentary basins (Németh et al. 2007). To distinguish between sedimentary environments purely related to the syn-eruptive basin settings from sedimentary environments created by volcanic eruptions is important to reconstruct the syn-volcanic landscape evolution of the region. Only the correct identification of 3D relationships between volcanic facies associated with the volcanic complexes allow correct reconstruction of the position of the syn-volcanic landscape.

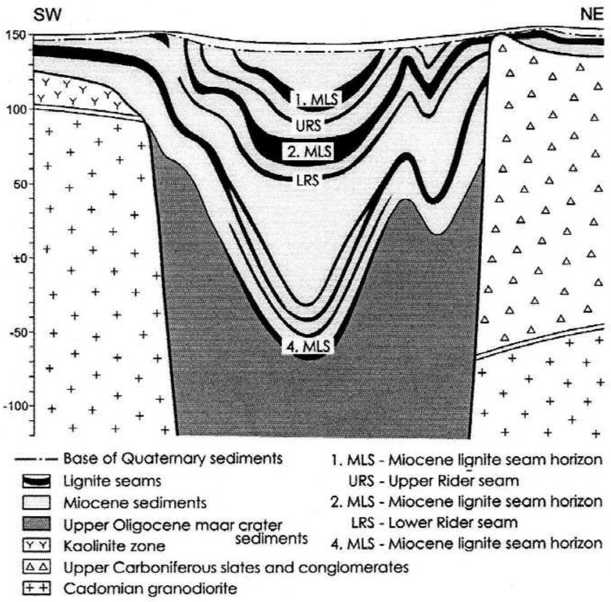


***Figure 5.13.*** Posteruptive subsidence and deformation of Miocene lignite bearing sediments above the Oligocene Kleinsaubernitz maar-diatreme volcano, eastern Saxony, Germany (after Lorenz 2007: fig. 4)

Geopysical anomalies over maars and diatremes vary in their character and do not provide definitive evidence for a phreatomagmatic origin (Büchel 1987, Schulz et al. 2005, Cassidy et al. 2007). Many known maars and diatremes are buried by post-genetic sediments. Geophysical methods resulted in their initial discovery and subsequent drilling provided geologic samples, which confirmed their phreatomagmatic origin. Interpretation of a single geophysical data set over a suspected maar or diatreme structure can be ambiguous. When combined, however, with complementary geophysical methods and the existing database over other known maar or diatreme structures, a more definite assessment can be made. The most notable geophysical signature associated with maars or diatremes is a negative gravity anomaly (Schulz et al. 2005). These gravity lows are generally circular and cover the whole structures. They are due to lithological and physical changes associated with the preatomagmatic explosion. In well-preserved maar structures, low-density sedimentary infill of the topographic depression of the crater contributes to the gravity low. In general, magnetic anomalies associated with maar or diatreme structures are more complex than gravity anomalies (Schulz et al. 2005). Their reasons are very complex intrusion processes in the diatreme or in the sedimentary infilling of the maars.

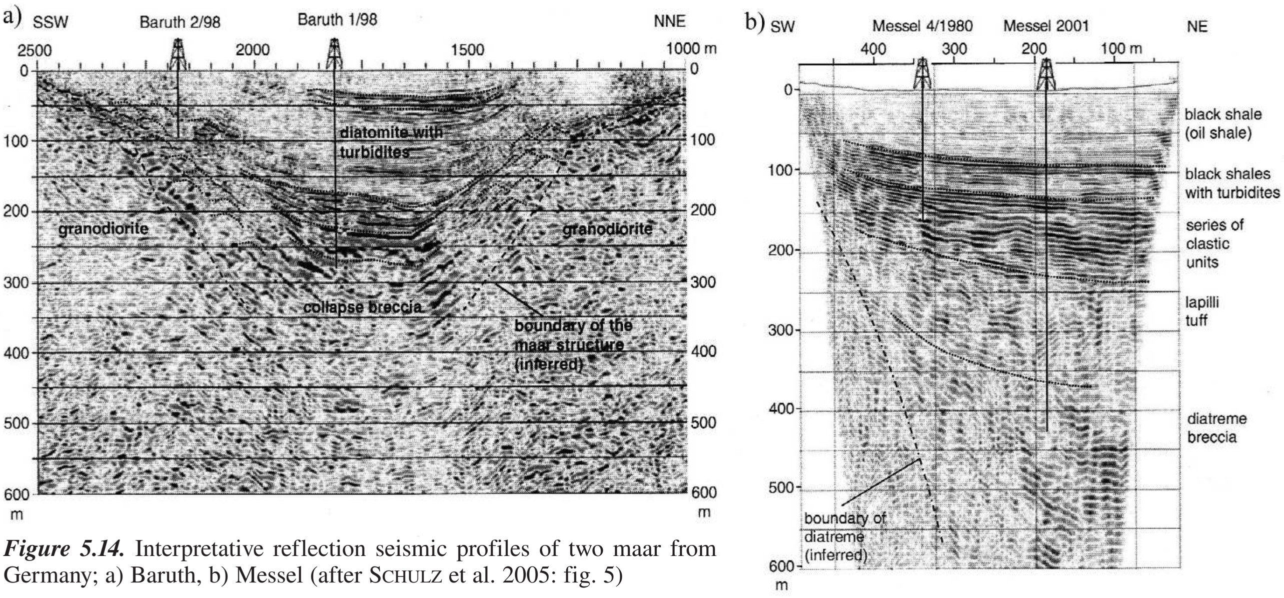


*Figure 5.14.* Interpretative reflection seismic profiles of two maar from Germany; a) Baruth, b) Messel (after Schulz et al. 2005: fig. 5)

Also the development of spatter cones into the maar crater can cause a magnetic anomaly.The presence of fluids in explosion-induced fractures and pore spaces of the maar and diatreme rocks leads to decreased resistivity levels that can be mapped effectively by various electrical methods (Brunner et al. 1999). Reflection seismic surveys allow for detailed imaging of maar structure morphology (Figure 5.14). Well logging methods are very useful for the detailed investigation of drill holes in maar and diatreme structures.

### Peperite associated with small volume phreatomagmatic volcanoes

Peperite results from interaction between magma and wet sediment and exhibits a range of complex textures (White et al. 2000, Skilling et al. 2002, White and Houghton 2006). Peperite is common in many geological settings where magma comes in contact with wet sediment (Skilling et al. 2002). The presence and recognition of peperite indicate contemporaneous volcanism and sedimentation, and they provide insight into the nature of subsurface magma transport and host-sediment properties at the time of eruption (Skilling et al. 2002). Peperite is common along the margin and adjacent to intrusive bodies and where lava flows travelled through wet sediments (Skilling et al. 2002). Recognition of peperite-forming processes in a sub-volcanic region of a phreatomagmatic volcano is given from several sites (Hooten and Ort 2002, McClintock and White 2002).

Peperites commonly form when magma intrudes wet unconsolidated sediment. The host can be any clastic sediment, and interactive phenomena occur in great variety under a wide range of physical conditions. In general, peperite are considered to be very common in arc-related and other volcano-sedimentary sequences associated with composite volcanic systems. In such settings peperite and associated volcaniclastic sediments may form large volume of deposits of a very diverse textural and compositional range from rhyolite to basic rock types. Peperites and their associated volcaniclastic environment are important in reconstructing palaeoenvironments. It has been recently recognized that peperite and their associated pyroclastic environment are as common in small volume phreatomagmatic vent-filling deposits and/or along the contacts between sediment and mafic, generally small-volume intrusions and lavas as in more complex stratovolcanes and or caldera volcanic systems (Martin and Németh 2007, Németh and Martin 2007). In the Pannonian Basin, a wide range of peperite has been described lately in phreatomagmatic, small-volume mafic volcanic systems (Plate V, 1). The recognition of the existence of peperite in small-volume, phreatomagmatic volcanoes and their associated intrusive/effusive products highlights the importance of identification of such textures and the interpretation of their meaning in these types of volcanoes (Martin and Németh 2007). Peperites are more common in association with small-volume terrestrial phreatomagmatic volcanoes than it has been considered previously, especially in volcanic fields, which developed in a region where surface and sub-surface aquifers provided substantial water and/or water saturated sediments to sustain phreatomagmatism. Peperite is often described on the basis of the juvenile clast morphology as blocky or fluidal, but there is a big range in shapes that have to be taken in to account as well (Busby-Spera and White 1987). The shape differences are often explained by the simple textural differences of the host deposit as fine and/or coarse grained texture that may produce globular (Plate V, 2) and/or blocky peperite respectively (Busby-Spera and White 1987). However, examples from the phreatomagmatic volcanic fields in the Pannonian Basin show that this simple classification is not the only controlling parameter (Martin and Németh 2007). Peperite in association with phreatomagmatic volcanoes can develop when feeder dykes intruded water-saturated sediment infilling 1) maar or tuff ring basins and/or 2) vent zones and/or 3) lava flow outpouring from such volcanoes (Figure 5.15). However, these three types of peperite may suggest strikingly different environmental reconstruction and therefore their interpretation may imply different geomorphological reconstruction. In the western Pannonian Basin (Hungary) lava capped buttes, e.g. Kissomlyó (Martin and Németh 2005) are commonly accompanied with a great variety of peperite that indicate wet environment into which the lava flow erupted. However, the reconstruction of the position of the lava flows need to be established before any geomorphological conclusion. In the western Pannonian Basin, most of the lava flows erupted inside of a crater of a phreatomagmatic volcano, and therefore the presence of peperite only means that the crater zone and not necessarily the surrounding syn-volcanic landscape was "wet" and water saturated. The identification of peperite in such settings therefore does not guarantee the reconstruction of the eruptive environment nor the establishment of the syn-volcanic landscape position onto the volcano erupted.

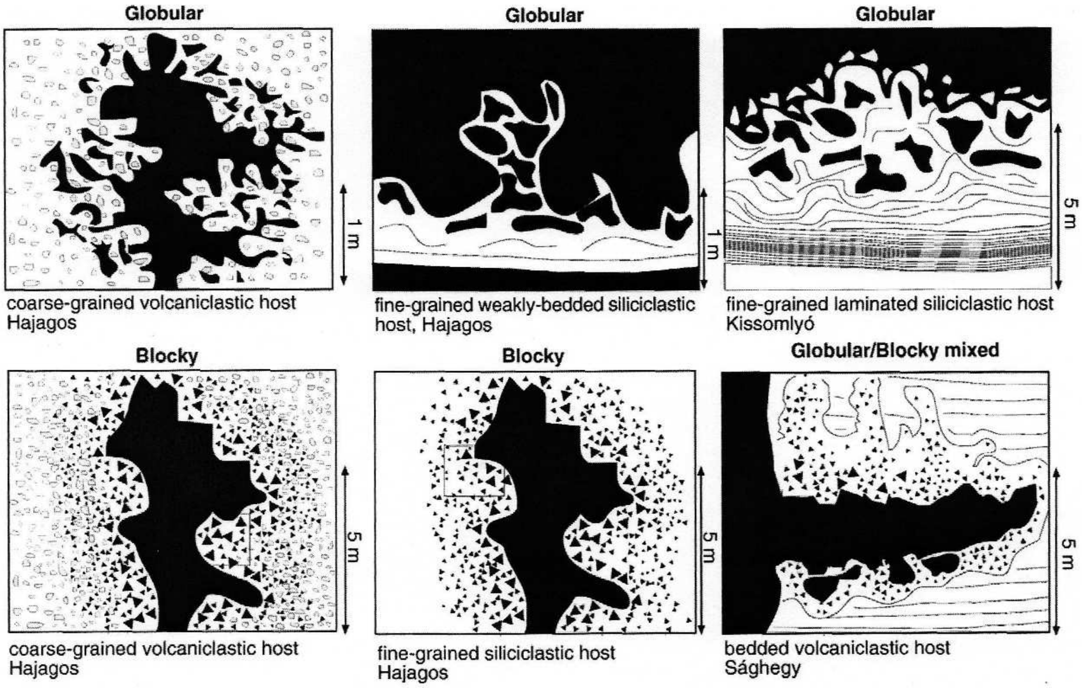


*Figure 5.15.* Schematic representation of peperite types identified from the Neogene phreatomagmatic volcanic fields of Hungary (after Martin and Németh 2007: fig. 13)

### "Dry"magmatic processes associated with phreatomagmatic volcanism

Maar-diatreme volcanoes are created by phreatomagmatic eruptions arising from a contact of ascending magma with ground water in available aquifers. If this contact is eliminated magma continues its ascent towards the surface giving rise to a wide range of volcanic forms and products associated closely with maar-diatreme volcanoes. Usually this happens in the advanced or closing stage in evolution of the volcano. However, in arid regions the maar-forming eruptions were often preceded by effusive and/or Strombolian activity (Plate V, 3) (Gutmann 2002). Various scenarios are possible and factors controlling changes in the eruption style are understood poorly. Phreatomagmatism might be inhibited unless the magma flux is low relative to the rate of water supply and unless the top of the magma column has subsided, probably below the water table. A general evolutionary sequence from hydromagmatic eruptions during formation of the maars, through Strombolian eruptions of the post-maar scoriae and ashes, and finally to the post-maar lavas (Plate V, 4) appears to reflect the declining influence of magma – ground water interactions with time, which process widely documented in many volcanic fields such as Hopi Buttes (Arizona) (White 1991b). The general time sequence of eruption events (from wet to dry explosive processes) and the distribution of volcanic vent type (dry and wet) in regard to the geomorphology of the pre-volcanic landscape is well established, having phreatomagmatic volcanoes in low lying areas and scoria cones in highlands (White 1991a). Transient hydromagmatic events e.g. occurred relatively late in the 1975 eruption sequence of the Tolbachik volcano in Kamchatka (Doubik and Hill 1999). Maar rims are thin and composed of easily erodable tephra. Intermittent phreatomagmatic acitivity due to sudden ground water access to the volcanic conduit in the vaning stage (low magma supply rate) of the eruption have also been described in large ocean islands such as Hawaii Kilauea (Dzurisin et al. 1995). The most widespread of these deposits at Kilauea (Uwekahuna Ash Member) is a few metres thick basaltic surge and fall deposit (Dzurisin et al. 1995). These few metres thick deposits inferred to be a result of two major pyroclastic surges, each preceded by unusually vigorous lava fountaining from a vent near the volcano's summit (Dzurisin et al. 1995).

Simultaneous magmatic and phreatomagmatic explosive events in the same volcano have been recognized from tuff ring sequences of many volcanoes. For instance, a series of alternating phreatomagmatic ("wet") and magmatic ("dry") basaltic pyroclastic deposits forming tuff rings, such as the Crater Hill tuff ring in New Zealand, contains unit(s) which can only be interpreted as the products of mixing of ejecta from simultaneous wet and dry explosions at different portions of a multiple vent system (Houghton et al. 1999). In other type of mixed deposit has been recognized from the Eifel, where documented mixed (wet and dry) basaltic pyroclastic deposits interpreted to represent mixing from two point sources (e.g. vents) of quite different but stable character (Houghton and Schmincke 1986, 1989). These two well-documented site highlights the complexity of eruption dynamics of a predominantly "wet", mafic volcano.

Magma, that has not reached the surface, appears as dykes, sills and/or plugs in diatreme/maar filling. Some of them may represent feeders to surficial activity. Outpourings of lava feed up lava flows and/or lava lakes in maar depressions, often in fully subaqueous environment. In such case pillow lavas, hyaloclastite breccias and/or peperite breccias are present. Phreatomagmatic eruptions of the Surtseyan-type due to interaction of ascending magma with water in the maar lake give rise to palagonite tuff cones. Eruption rate and water depth are factors controlling Surtseyan-type eruptions and transition towards Strombolian-type eruptions. Lava flows and lava lakes provide an excellent opportunity to study evolution of jointing and its relationship to the form of the lava body. They are also good objects to calibrate and compare Quaternary dating methods, remote sensing methods, and rates of geomorphic processes.

### Surtseyan volcanism

Surtseyan eruptions (Figure 5.16) are characterized by interaction of a fluid erupting magma with abundant external water (Kokelaar 1983, White and Houghton 2000). Commonly they start in shallow subaqueous environments (Figure 5.17) where accumulating tephra forms a mound-shaped volcano prior to emergence (White 1996, White and Houghton 2000). During this subaqueous stage density currents play a significant role in shaping the edifice and imparting characteristic bedding features (White 1996, Smellie and Hole 1997, Mueller et al. 2000, White 2000, Martin 2002). Emergent to subaerial Surtseyan eruptions generally produce cone-shaped edifices by accumulation of wet fall deposits with subordinate density currents (Verwoerd and Chevallier 1987, Sohn and Chough 1993, Sohn 1995, Cole et al. 2001). Surtseyan deposits typically consist almost entirely of glassy fragments formed by fragmentation of the erupting magma, and lack a significant country-rock component (Kokelaar 1983, White and Houghton 2000). This is an important difference from deposits of maar-forming eruptions, and it indicates that in Surtseyan eruptions fragmentation occurs at very shallow levels in the edifice or/and as the magma emerges from it. If or when the erupting magma no longer encounters water (e.g. by enclosure of an emergent vent, or isolation from/depletion of ground water), both Surtseyan and maar-forming eruptions may transform to Strombolian or Hawaiian ones. The subaqueous phases of the eruption of an emergent volcano, such as the well-documented Pahvant Butte eruption initiated few tens of metres (85 m in the case of

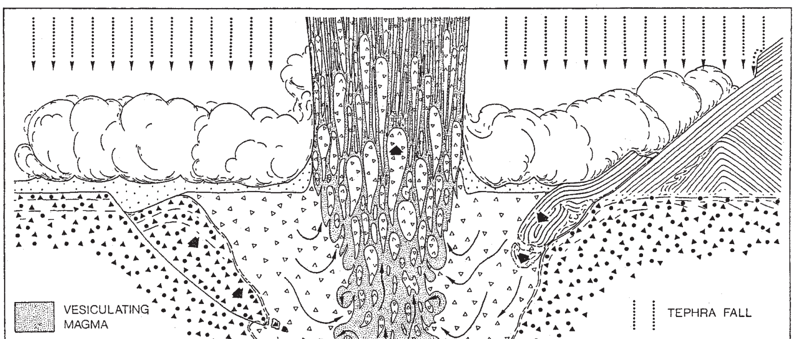


*Figure 5.16.* Eruption model of a Surtseyan eruption (after Kokelaar 1983)

Pahvant Butte) beneath the water surface produced a broad mound of tephra (Figure 5.18)(White 1996). A variety of distinctive lithofacies identified from Pahvant Butte allows reconstruction of the eruptive and depositional processes active prior to emergence of the volcano above lake level (White 1996). Early in the eruption subaqueous tephra jetting from phreatomagmatic explosions discontinuously fed inhomogeneous, unsteady, dilute density currents (White 1996). Dunes and crossbeds which are better developed upward in the section of Pahvant Butte resulted from interaction between sediment gravity flows and surface waves triggered as the explosion-generated pressure waves and eruption jets impinged upon and occasionally breached the surface (White 1996). Intermingling of (a) tephra emplaced after brief transport by tephra jets within a gaseous milieu and (b) laterally flowing tephra along vent margins during parts of the eruption in which episodes of continuous uprush produced localized water-exclusion zones above a vent. Mass flow deposits formed by disruption and remobilization of mound tephra. Intermittent, explosive magma– water interactions occurred from the outset of the Pahvant eruption, with condensation, entrainment of water and lateral flow marking the transformation from eruptive to "sedimentary" processes leading to deposition of the mound lithofacies (White 1996). Surtseyan-style eruptions have been observed in marine environment (Thorarinsson et al. 1964, 1967) and crater lakes (Németh et al. 2006). The observations confirmed that two type of explosive activity accompanied with the eruption (Kokelaar 1983, White and Houghton 2000); 1) repeated continuous particle uprush, repeated in about up to 20 minutes intervals, and 2) explosive bursts forming dark coloured, ash charged cock's tail jets repeated in few minutes intervals.

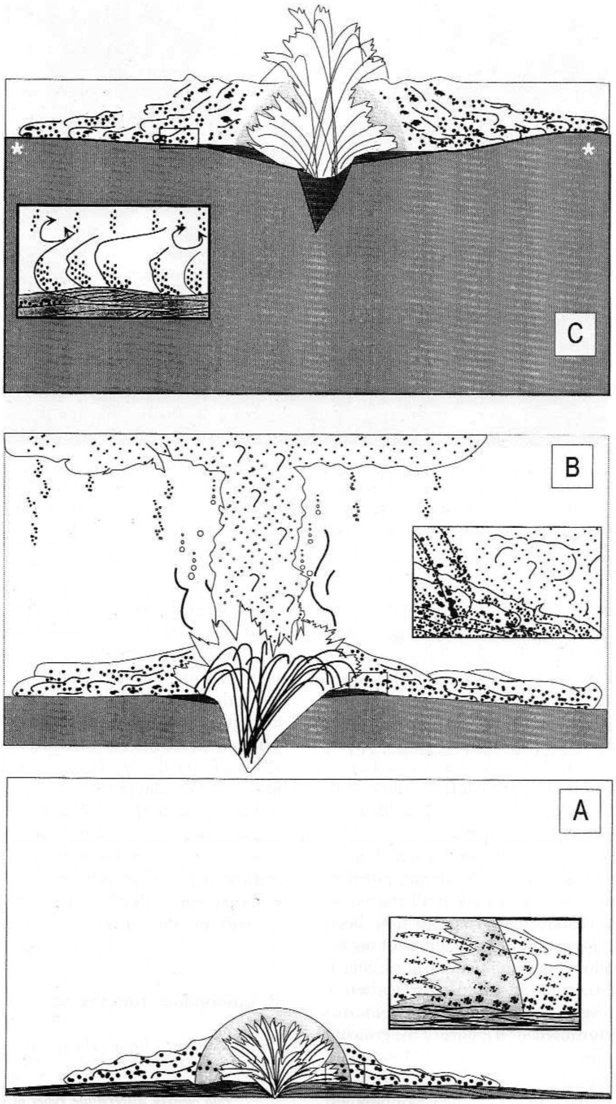
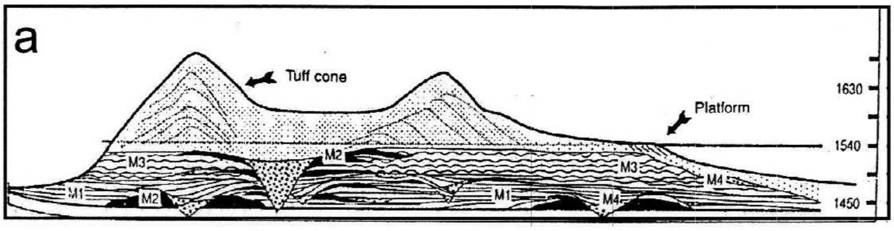
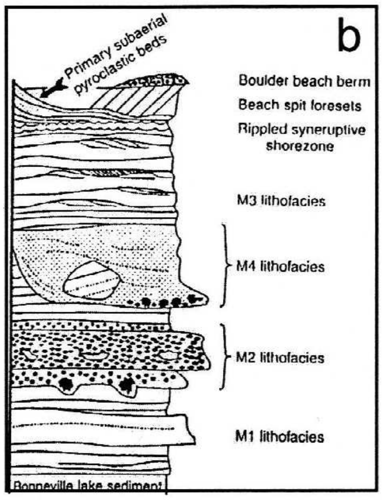


*Figure 5.17.* Eruption model of the Pahvant Butte. A) initial subaqueous phase, B) subaqueous fall formation, C) emergent phase,eruption cloud breaches the lake surface (after White 1996)

*Figure 5.18.* Cross section (a) of Pahvant Butte with identified lithofacies (b) (M1-4). The structure and morphology of the butte probably adoptable to other subaqueous phreatomagmatic volcanoes' architecture (after White 1996)





Surtseyan-style volcanoes are commonly develop in the rift edges of ocean islands. Along the rift axis scoria cones form, and in the low-lands, near the sea level, phreatomagmatic volcanoes as well as Surtseyan-style volcanoes form such as the 1957-58 eruption of Capelinhos in Azores (Figure 5.19) (Machado et al. 1962, Waters and Fisher 1971, Cole et al. 2001), or the 1913 eruption sites on Ambrym, Vanuatu (Németh and Cronin 2006). Surtseyan-style eruptions are also common in island arc settings such as the Izu-Bonin arc in Japan. In such settings, volcanic islands (tuff cones) can form

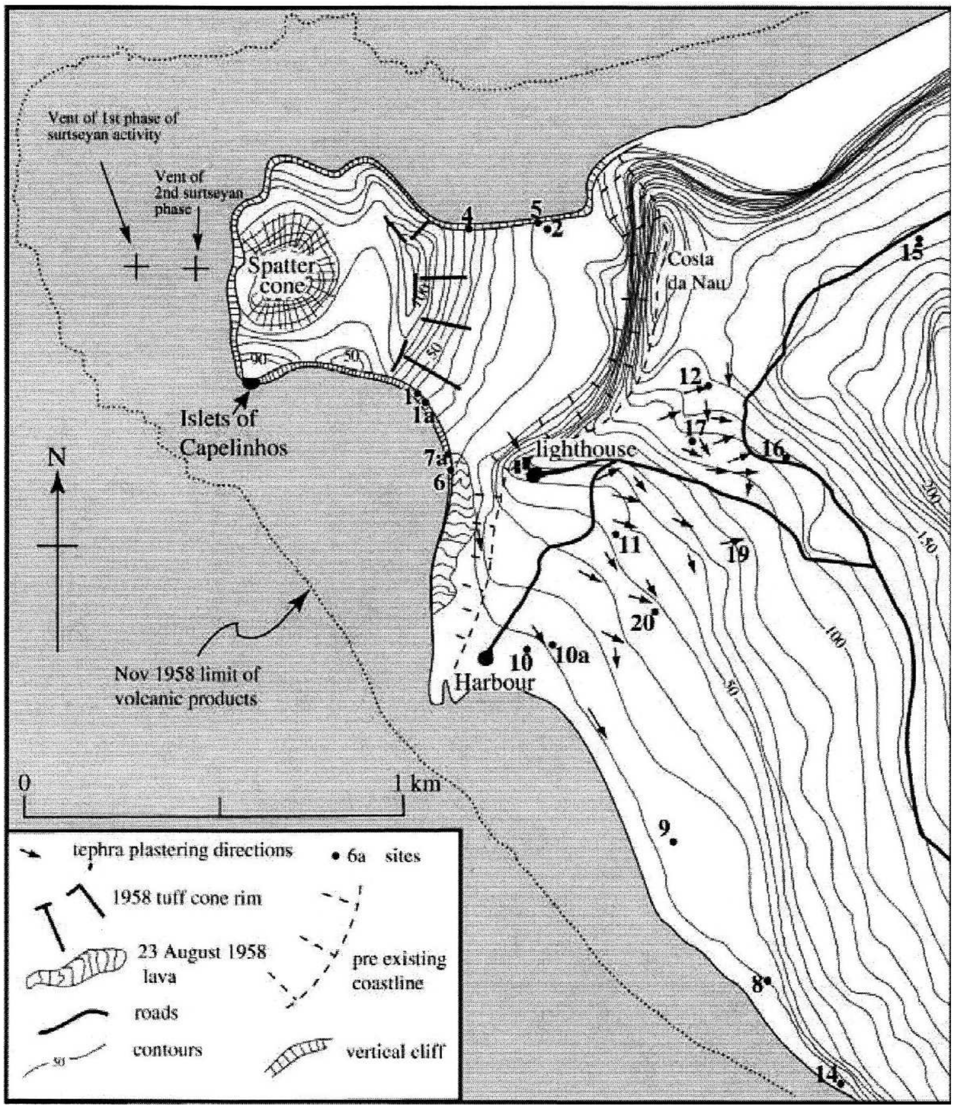


*Figure 5.19.* Map of the Capelinhos volcano developed in the edge of a volcanic island, Azores (after COLE et al 2001)

quickly but they commonly completely vanish by wave action, such as the Graham Island (Ferdinandia) in the Channel of Sicily in 1831 (PlateV, 5). Surtseyan eruption also can take place in caldera lakes such as the 2005 Ambae (Plate V, 6) eruption in Vanuatu (NÉMETH et al. 2006), where the 50 m high tephra cone erodes quickly after the eruption ceased.

### Volcanic hazards of monogenetic volcanoes

Hazards associated with maar eruptions are: volcanic earthquakes (up to c. M: 4-5), possibly several 1000 individual eruptions, eruption clouds rising to maximum heights of economic air travel, ejection velocities of tephra clasts of up to 400 m/s, ejection distances of ballistic clasts up to 4 km; size of ejected clasts up to 8 m, base surges travelling up to several km and with time building a tephra ring of a height up to 100 m and of a radius of up to 4 km (measured from centre of crater), thin distal tephra falls extending to more than 100 km, syn- and post-eruptive slumps and lahars inside and in part also outside the crater, destruction of buildings and transport lines within a radius of up to 5-6 km (LORENZ 2007). Associated formation of the maar crater floor and underlying diatreme results in subsidence of country rocks, tephra, and buildings to depths of possibly 1000-2000 m (SUHR et al. 2004). In addition, recent studies have shown that there are hazards associated to recurrence of activity within volcanic fields but also in single maars. Volcanic hazard studies of a volcanic field commonly target to understand the recurrence rate, the eruption frequency and style of eruptions may occur in the future (HO 1992, CONNOR and HILL 1995, HO and SMITH 1998, CONNOR et al. 2000, CRONIN et al. 2001, CRONIN and NEALL 2001, EDBROOKE et al. 2003, HURST and SMITH 2004, MAGILL and BLONG 2005a, 2005b, MAGILL et al. 2006).

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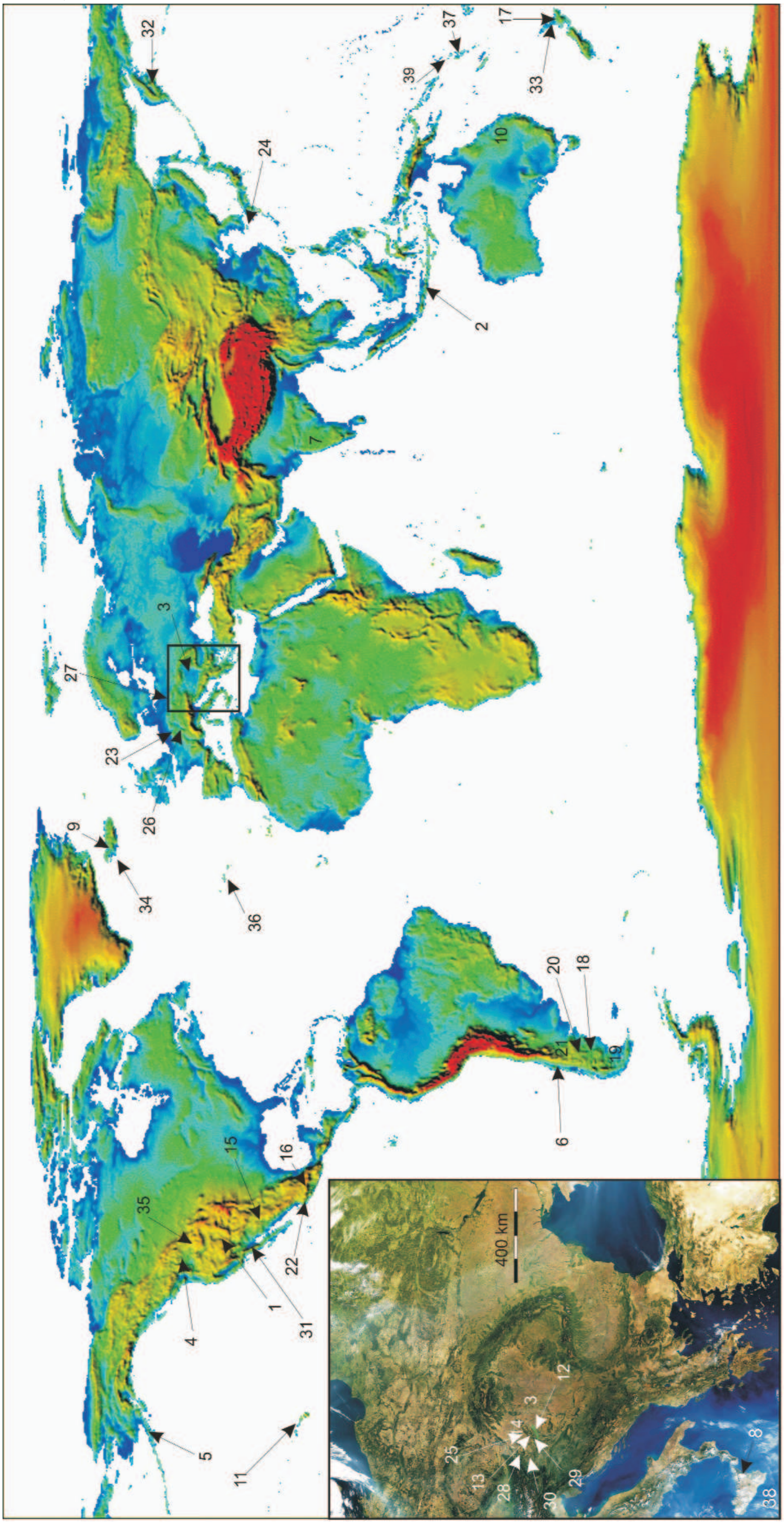
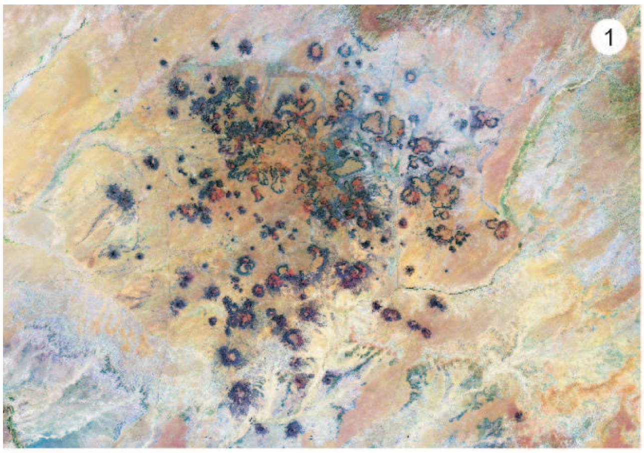
## Location map



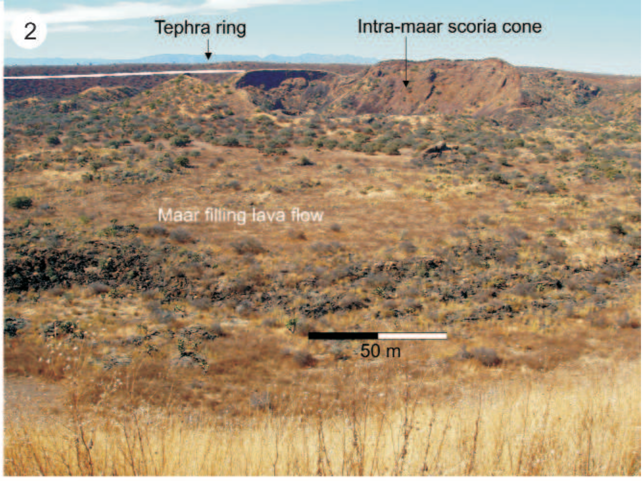
1 — Hopi Butte, Arizona, USA
2 — Lamongan, Java, Indonesia
3 — Pannonian Basin, Central Europe
4 — Snake River, Idaho, USA
5 — Ukinrek maar Alaska, USA
6 — Rininahue, Chile
7 — Deccan, India
8 — Etna, Italy
9 — Iceland
10 — Eastern Australia
11 — Hawaii
12 — Tihany, Hungary
13 — Füzes-tó, Hungary
14 — BBHVF, Hungary
15 — La Brena maar, Mexico
16 — Paricutin, Mexico
17 — Tarawera, New Zealand
18 — Patagonia, Argentina
19 — Pali Aike, Argentina
20 — Chubut, Argentina
21 — Payunia, Argentina
22 — Ceboruco, Mexico
23 — West Eifel, Germany
24 — Jeju Island, Korea
25 — Pula maar, Hungary
26 — Messel maar, Germany
27 — Baruth maar, Germany
28 — Ság-hegy, Hungary
29 — Hajagos, Hungary
30 — Kissomlyó, Hungary
31 — Crater Elegante, Mexico
32 — Tolbachik, Russia
33 — Crater Hill, New Zealand
34 — Surtsey, Iceland
35 — Pahvant Butte, Utah, USA
36 — Capelinhos, Azores, Portugal
37 — Ambrym Island, Vanuatu
38 — Graham Island, taly
39 — Ambae Island, Vanuatu

# Plate I

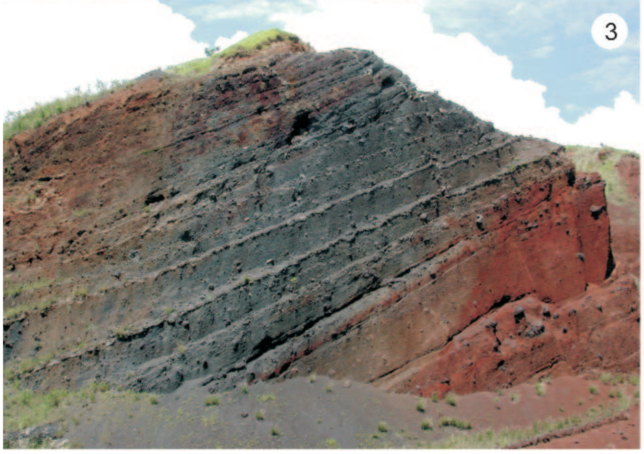


1. MrSID satellite image of the Hopi Butte volcanic field where about 300 monogenetic volcanic vents have been identified. The map view is about 80 km wide.
2. Typical scoria cone from Payunia, Mendoza, Argentina.
3. A well developed tuff cone in the western edge of Ambrym, Vanuatu.
4. Low tuff ring from Payunia, Mendoza, Argentina.
5. A maar with lake from the Pali Aike volcanic field, Patagonia, Argentina.
6. Typical tephra ring sequence consists of alternating base surge and phreatomagmatic fall beds from a maar rim in the Pali Aike Volcanic Field, Argentina.
7. Deep and wide crater of the Crater Elegante maar in Sonora, Mexico.

## Plate II

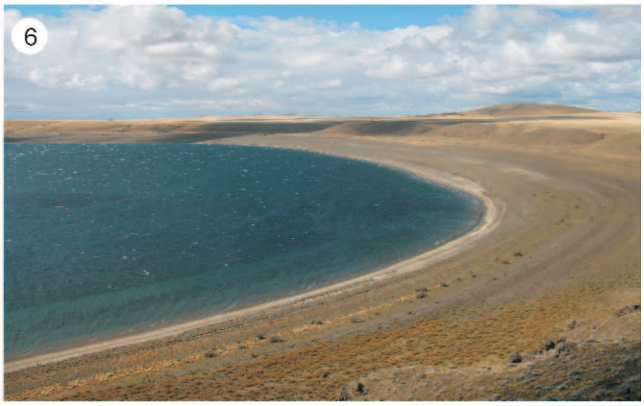


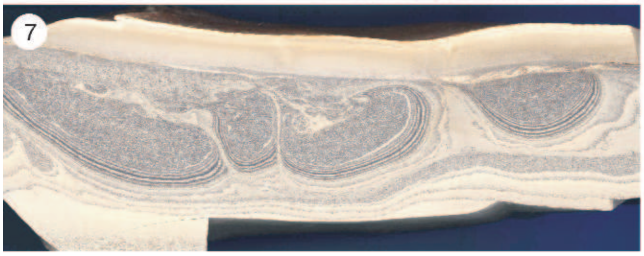
1. Exposed diatreme from Chubut, Argentina.
2. Scoria cone and lava field in the maar crater of the La Brena maar, in Durango, Mexico.
3. Large scoria cone of Paricutin in Michoacan, Mexico.
4. Intracontinental lava shield volcano from Al Haruj, Libya.
5. Lava spatter-dominated scoria cone chains in the Pali Aike Volcanic Field, Argentina.
6. Large tumuli from the Al Haruj, Libya.



1. Lava spatter deposits preserved the steep flank of a volcanic cone of the Al Haruj Volcanic Field, Libya.
2. Collapsed spatter cone fragments carried away on a lava flow as documented from a cone complex of the Crater Basalt Volcanic Field, Chubut, Argentina.
3. Typical scoria cone deposit of a scoria cone from scoria cone nearby Ceboruco, Western Transmexican Volcanic Belt, Mexico.
4. Welded agglutinate succession from Al Haruj, Libya.
5. Completely welded and exhumed interior of a scoria cone from Al Haruj, Libya.
6. Chain of scoria cones near Ceboruco, Nayarit, Mexico.

# Plate IV





1. Maar lake in Rininahue, Chile.
2. Exposed diatrema from southern Slovakia, the sharp plain is the contact between diatreme (left) and host (right) rocks.
3. Typical base surge beds from the 1913 eruption of west Ambrym, Vanuatu.
4. Interbedded scoriaceous fall deposit in the 1913 phreatomagmatic succession of west Ambrym, Vanuatu.
5. Reworked volcaniclastic succession in distal areas of a tephra ring of the 1913 eruption site of west Ambrym, Vanuatu.
6. Large maar lake (Potrok Aike) from the Pali Aike Volcanic Field, Patagonia, Argentina. The tephra ring is almost entirely eroded.
7. Soft sediment deformation in maar lake sediment of the Pula maar, western Hungary.

# Plate V

1. Blocky peperite from the Hajagos maar in western Hungary.
2. Globular peperite along a sill from the Ság-hegy phreatomagmatic volcano.
3. Scoria cone half section exposed in the crater wall of the Crater Elegante maar in Sonora, Mexico.
4. Transition from phreatomagmatic succession (base of hill along the cliff of the Snake River valley) to strombolian scoria cone units (capping topmost part of hill) from the Sinker Butte, Western Snake River Plain, Idaho.
5. Two paintings of the eruption of the Graham Island in 1831 in the Channel of Sicily, Mediterranean Sea; on (A) white eruption cloud probably caused by steam is visible with dark jets closely resembling cocks' tail jets reported from Surtseyan eruptions. On (B), lava fountaining is visible, possible representing the late stage of the eruption of the emergent volcano, where no or just limited magma-water interaction took place.
6. Surtseyan eruption in Lake Voui in Ambae Island in December, 2005.

# Chapter 6

# Polygenetic volcanism and associated features

## Polygenetic volcanism

Polygenetic volcanism (Figure 6.1) is considered to be volcanism that is associated with long lasting and usually complex eruption history of a volcanic system (DAVIDSON and DE SILVA 2000; LIPMAN 2000; WALKER 2000). Such volcanic systems can either consist of a single volcanic edifice, or where slight shifting of position of the active pathway of magma to the surface occurs a nested and multiple edifice complex can result. In general, we view a volcanic system as polygenetic if successive magma batches causes eruptions in more or less in the same place (DAVIDSON and DE SILVA 2000). In this way, the term can be slightly misleading because many relatively small volume eruptions may recur in the same place and may result complex, nested volcanic systems such as those found in many long lived basaltic volcanic fields (HOUGHTON and SCHMINCKE 1989; HOUGHTON et al. 1996; CONNOR and CONWAY 2000; VESPERMANN and SCHMINCKE 2000). Many of relatively small volume volcanoes in a basaltic volcanic field can exhibit very complex volcanic architectures, and in many cases their eruption history shows signs of multiple and recurring activity over long periods of time (AUER et al. 2006). However, it is rare for small volume basaltic volcanoes to stay active for longer then few months (SELF et al. 1980; LUHR and SIMKIN 1993), in contrast to true polygenetic volcanoes commonly forming composite volcanic edifices that are active over thousands of years. The transition between small volume but complex basaltic volcanoes and long-lived but relatively small volume composite volcanoes is rather gradual and it has been described from many volcanoes from Central America (ABRAMS and SIEBE 1994; MCKNIGHT and WILLIAMS 1997).

Polygenetic volcanoes can occur in many different edifice forms as well as being very different in duration of activity. In this way a large, long-lived shield volcano associated with hot spot magmatism is also classified as polygenetic (Plate I, 1), and can result in accumulation of large volumes lava fields, with very diverse geochemical signatures, reflecting slight chemical changes in the rising melt over long periods of time. A good example of these types of volcanoes is the islands of Hawaii where hot spot activity formed four coelescing broad shield volcanoes over the last million years of activity (FREY et al. 1991; MOORE and CLAGUE 1992; MOORE 1992). The activity resulted in the accumulation of diverse composition and texture pahoehoe lava fields (KESZTHELYI and SELF 1998; CROWN and BALOGA 1999; BYRNES and CROWN 2001). Such shield volcanoes can also have a complex architecture and the volcanic edifice can contain a reasonable volume of pyroclastic material resulting from occasional explosive activity (MCPHIE et al. 1990; DZURISIN et al. 1995; MASTIN and WITTER 2000). Their reworked counterpart may also form extensive volcaniclastic aprons around the main lava shield. Due to gradual growth of a shield volcano, and their relatively unsupported seaward edifice constructs (MARTI et al. 1997; MASSON et al. 2002; MORGAN and CLAGUE 2003), such volcanoes may also develop extensive volcaniclastic aprons in the sea floor, commonly as a result of occasional volcano collapse (ANCOCHEA et al. 1994; GARCIA and HULL 1994; MOORE et al. 1994; WATTS

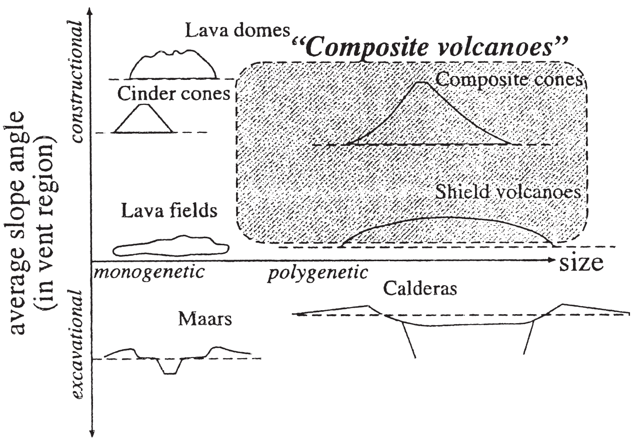


*Figure 6.1.* Variation of volcanic landforms as a function of their size and morphological parameters (from DAVIDSON and DE SILVA 2000: p. 664, fig.1)

and MASSON 1995; MCMURTRY et al. 1999; CERVELLI et al. 2002; ENGELS et al. 2003; COOMBS et al. 2004). Large shield volcanoes can also form large volume of hyaloclastite due to their repeated effusive eruptions in the sea floor (Plate I, 2) (QUANE et al. 2000; WALTON and SCHIFFMAN 2003; COOMBS et al. 2004; SCHIFFMAN et al. 2006) and littoral cones (JURADOCHICHAY et al. 1996) where lava and sea water contact take place. Such hyaloclastite aprons can be drifted away by currents and resedimented around the volcanic edifice (FURNES and STURT 1976; RUIZ et al. 2000). Since a lava shield volcano can be active over hundreds of thousands of years, such sedimentary units can form a significant volume of sediments (GEE et al. 2001; KRASTEL and SCHMINCKE 2002; MITCHELL et al. 2002). Such sediments can be preserved in a sedimentary record. In ancient sedimentary sequences thick piles of resedimented hyaloclastite marks long lasting and polygenetic evolution of a lava shield. Lava shield volcanoes are also common in subaerial settings in intracontinental tectonic settings, for example (Plate I, 3) in northern Africa (ADE-HALL et al. 1974; FRANZ et al. 1997, 1999; WILSON and BIANCHINI 1999; KENEA et al. 2001; NÉMETH 2004; MARTIN and NÉMETH 2006), along the Snake River Plain (Plate I, 4) in Idaho (GREELEY 1982; GEIST and RICHARDS 1993; CUMMINGS et al. 2000; GODCHAUX and BONNICHSEN 2002) or in Eastern Australia (FREY et al. 1978; SUTHERLAND 1983; JOHNSON 1989; OREILLY and ZHANG 1995). Volcanoes in these environments can also be long lived and usually their position marks major magma rise paths in an intracontinental volcanic field. The large accumulation of lava flows in such volcanic fields (STEPHENSON and GRIFFIN 1976; STEPHENSON et al. 1996) may also have associated explosive phases of activity (Hawaiian- or Strombolian-style eruptions) as well as occasional phreatomagmatic explosive activity (NÉMETH et al. 2003; PEREGI 2003; NÉMETH 2004). Lava shields in such settings adjacent to a lacustrine basin may give rise to lava deltas and hyaloclastite successions (Plate I, 5) may develop in the areas where lava flows enter fluvio-lacustrine networks (GODCHAUX and BONNICHSEN 2002; HUGHES et al. 2002; WOOD and CLEMENS 2004). Similar facies may develop where lava enters shallow marine environment (BEHNCKE 2004). Although intracontinental lava shield volcanoes can be volumetrically the same as those forming ocean islands, their size is commonly a magnitude smaller. Intracontinental flood basalt volcanism is also inferred to be long lived (tens of thousands of years of activity), and commonly forms large volumes of phreatomagmatic tephra as described from the Jurassic Mawson Formation in Antarctica (ROSS and WHITE 2005; MCCLINTOCK and WHITE 2006) or Karoo from South Africa (ROSS et al. 2005). Such phreatomagmatic pyroclastic rocks may form as a result of long lasting phreatomagmatic eruptions that may generate so called phreato-cauldrons, which leave large volume of deposits (WHITE and MCCLINTOCK 2001).

Composite or strato-volcanoes are a major type of polygenetic volcanoes. They occur predominantly in subduction zones, however, intraplate volcanoes of similar architecture are also common such as those forming a Cenozoic volcanic chain in Eastern Australia. Composite volcanoes are those that erupt over long period of time (thousands of years), and producing extensive lava flows alternating with major explosive eruptive products as well as their reworked, redeposited and resedimented counterparts (Figure 6.1). Volcanoes that form amalgamation of relatively small volume eruptive vents, commonly termed compound volcanoes, are typical of basaltic volcanic systems. Composite volcanoes are surrounded by a ring plain, a place where distal primary volcaniclastic sediments as well as lava flows and associated reworked material can accumulate and form a basin fill succession, commonly intercalated with normal siliclastic sediments (PALMER and NEALL 1991). Composite volcanoes usually have a central vent (Plate II, 1) surrounded by satellite vents that are individual, short lived, monogenetic volcanoes. However, their feeding systems are commonly coupled with the central vent of the composite volcano (Plate II, 2). Such satellite vents are located in areas that are favoured by local stress fields which allows melt to reach the surface. Commonly they can change their position as the composite volcanic edifice evolves. Composite volcanoes that are active over thousands of years, can reach an equilibrium edifice architecture, that is, a shape that is controlled by equilibrium between the total volume of the erupted products and the erosion that acts on it especially in inter-activity periods (DAVIDSON and DE SILVA 2000). Shape of the composite volcano is therefore strongly controlled not only by the style of eruptions, and the texture and physical properties of the eruptive products, but also by external forces such as rainfall, rainfall distribution, principal wind direction, and characteristics of the substrate that may affect the stability of the growing volcanic cone (DAVIDSON and DE SILVA 2000). In a very simplistic way, the type of volcanoes from morphological point of view can be characterised by their edifice shape (constructional versus excavational) and whether they are monogenetic or polygenetic (Figure 6.1) (DAVIDSON and DE SILVA 2000). Polygenetic volcanoes in this classification can be constructional, such as shield volcanoes and composite cones, or excavational such as calderas (DAVIDSON and DE SILVA 2000). Excavational polygenetic volcanoes, can be extremely complex, and their original landform can be truncated by uplifted resurgent part of the caldera floor (ACOCELLA and FUNICIELLO 1999; LIPMAN 2000). Such resurgence may occur more than once and, especially in ancient settings, can produce a constructional volcanic landform, where geomorphic inversion is significant (ACOCELLA et al. 2000; LINDSAY et al. 2001a).

Composite volcanoes are the most common type in convergent plate margins, where chains of large volume and long lived arc volcanoes can develop, and affect the sedimentation in their surroundings. Spacing of these volcanoes shows a regular pattern (STERN 2004). However, exceptions may occur, such as the Rininahue volcanic field which developed in the space between two major composite volcanoes (STERN 2004). Such exceptions may be caused by plate tectonic and regional stress field distribution anomalies. These types of volcanoes are usually andesitic or dacitic in composition. However, rhyolitic eruptions also occur. In the divergent plate margins composite volcanoes are rare, however, Hekla and Askja in Iceland are two large volume composite volcanoes in such settings (Baldridg et al. 1973; Sparks et al. 1981; Brown et al. 1991; Gudmundsson et al. 1992; Sturkell et al. 2006). In the rifting stage of continental break up, large intraplate composite volcanoes can develop such as those associated with the East African Rift system (Bosworth 1987; Roche et al. 2001). Calderas can form during the evolution of normal composite volcanic cones. Large volumes of volatile rich rhyolitic magma can form through fractionation, that may triggers a major explosive eruption that reduce the former composite volcanic edifice to a caldera such an example is the Crater Lake in Oregon, formed by the Mazama eruption (Bacon 1983; Klug et al. 2002).

Larger caldera forming eruptions are directly related to long-lived magma infiltration and shallow level magma evolution that produced very explosive eruptions, resulting in the formation of calderas tens of kilometres in diameter. Such caldera eruptions commonly have evolutionary stages, and the formation of calderas may be associated with intensive post caldera volcanism, that can form individual, small volume constructional volcanic edifices (Lipman 2000). Mega-calderas that formed few tens to thousands of km$^3$ volume of pyroclastic material may have erupted over long period of time, and their eruption may have taken long time are common in the Andean volcanic arc (Lindsay et al. 2001a). Those volcanoes have only recently been reported due to their deposits and hundreds of kilometres across extension broad, flat morphology (Lindsay et al. 2001a).

Classical cone-shape composite volcanoes have certain morphological characteristics especially dacitic compositions. Lava domes commonly cap the summit of such volcanoes (Plate II, 3), and these may grow over decades. Gravitational instability, partially due to volatile pressure, can trigger dome collapse-induced pyroclastic flows (Sato et al. 1992). Composite volcanoes are commonly truncated by major edifice failures in the form of volcanic debris avalanches (Plate II, 4). In the morphological development of a composite volcano the interplay between aggradation and degradation is important (Hathway and Kelley 2000). When external forces are strong (tropical climate, heavy rainfall) and the eruption frequency is relatively low, the volcanic edifice can be deeply incised, steep, and, in case of volatile rich magma involvement during eruptions, can commonly fail (Plate II, 5). In other end-member, when the degradation is slow, eruptive products can accumulate quickly and lead to gravitational failure of the volcanic edifice. The characteristic conical shape of active composite volcanoes such as Mayon in Philippines, or Taranaki in New Zealand (Plate III, 1), indicates an equilibrium between aggradation and degradation (Davidson and de Silva 2000). The total volume and size of the composite volcanoes is therefore controlled by factors such as magma supply rate, magma composition, climatic forces, and perhaps the physical properties of the crust over which the volcanoes develop. Where a relatively steady and long lived magma supply exist, such as the case of many intraplate volcanoes, the edifice can grow significantly, forming composite volcanoes such as Ararat in Turkey amongst the largest on Earth (Pearce et al. 1990; Yilmaz et al. 1998; Adiyaman et al. 2003). This given tectonic setting composite volcanoes are commonly of the same size, reflecting the common nature of the melt and the tectonic regime in which they developed. In the aggradational stage of the volcano, the total eruption product volume and eruption style (effusive versus explosive) is important in the development of a certain cone shape. Magma ascent is powered by the "hydrostatic head effect", which is an overpressure in the magma reservoir, in turn controlled by the chemistry of the melt (volatile content) and the tectonic stress in the crust (Davidson and de Silva 2000; Murray and Stevens 2000; Rutherford and Gardner 2000; Pinkerton et al. 2002). Final magma extrusion initiates from relatively shallow magma chambers beneath composite volcanoes (Figure 6.2) The maximum height the magma may be able to reach in an "open" conduit is therefore strongly controlled by magmatic overpressure. Higher initial magmatic overpressure drive high eruption rates, and significant volumes of lava can erupt early in the evolution of the composite volcano, forming lava dominated, high density ("heavy") volcanic edifice. Such an edifice can increase the lithostatic load and therefore suppresses the ability of magma to reach higher zones of the growing edifice, and, as final stage, magma cannot reach the summit vent (Ida 1999; Murray and Stevens 2000). During magma ascent, storage of magma in magma chambers can allow differentiation to take place (Hansteen et al. 1998), and over time low density, but higher viscosity magmas can evolve. As a result the magmatic overpressure can be maintained due to the lower density and higher volatile content of the evolved melt, even under the increasing lithostatic load of the growing edifice, but the eruptions gradually could switch from dominantly effusive to dominantly explosive. Such a process can lead to volcanic eruptions that may destabilize the edifice and cause volcano collapse. This in turn can cause decrease in lithostatic load, and returns the volcanic system to near-initial conditions. Such pulsating growth and destruction of composite volcanoes are common in the geologic record, and probably one of the factors most responsi-

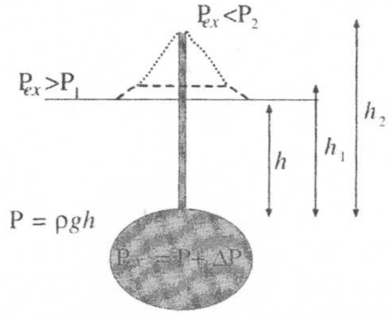


***Figure 6.2.*** Shematic diagram representing the relation between lithostatic pressure (P) and eruption driving overpressure (Pex) in the magmatic system of an active volcano. (after Davidson and De Silva, 2000: p. 671, fig. 5)

ble in the development of the volcanic landforms of composite volcanoes (Ancochea et al. 1990; Palmer and Neall 1991; Beget and Kienle 1992; Cacho et al. 1994). Observation suggests that there are physical limits to edifice heights at around 3000 m above the base of the volcano, suggesting non-evolved compositions for the initial eruptive products. This dynamic relationship between effusion rate and style of eruption are the main controlling parameters of the shape and morphology of volcanic edifice. Equilibrium stage volcanic edifices are those that have reached their maximum height and the further morphological evolution of the volcano can only occur as aresult of major edifice failure that could reset the physical conditions. This way, this is the point when degradational processes come important. In this stage degradation can form deposits that accumulate in a ring plain that may form a stabilizing flank around the volcano to prevent major edifice failure. Edifice failure perhaps can take place in the near summit vent area that can be masked quickly by fresh tephra that has been recognized in many composite volcanoes (Plate III, 2). The shape of the volcanic edifice is largely controlled by slight shifting of the vent location. Volcanic edifices that not have a stable conduit wall can develop closely spaced vent location that can even be active simultaneously. However, the individual active vent activities can be strongly coupled.

Volcanic products of composite volcanoes are very diverse. They can grow relatively quietly, through Strombolian and Hawaiian-style eruptions (Gardeweg et al. 1998) and associated lava effusions through their central vents. Such eruptions could change to Plinian-style eruptions ower a longer time frequency and produce pyroclastic flows, commonly controlled by major eruption column collapses. Many arc volcanoes however, grow lava domes that could collapse due to gravitational instability and/or minor explosive disruption, both leading to form block-and-ash flows such as the Unzen eruption in 1990–1995 in Japan (Nakada et al. 1999). In inter-eruptive phases reworking and non-eruption triggered lahar and/or volcanic debris avalanche formation could take place. Since pyroclastic flow development is one of the most common and dangerous phenomena associated with composite volcanoes, their origin and depositional features will be considered further.

### Pyroclastic flow genesis

Pyroclastic flows are generated byvolcanic eruptions and are considered to be very mobile, hot, and have high particle concentration, move horizontally moving and be dominanted by gas-particle dispersion (Freundt et al. 2000; Wilson and Houghton 2000). The particles in the pyroclastic current are pyroclasts, since they are generated by explosive fragmentation of the magma and the conduit wall and then erupted through a volcanic vent. The particle-support mechanism of the pyroclastic flow is dominated by fluidisation, buoyancy, grain to grain collision, and hindered settling (Sparks et al. 1978; Wilson 1980; Carey 1991; Freundt et al. 2000; Wilson and Houghton 2000). One of the most common ways in which pyroclastic flows form is by collapse of a vertical eruption column (Figure 6.3) (Cas and Wright 1988). Column collapse can take place immediately after a single eruption, or during a series of closely timed explosions, such as those occur in many Vulcanian-style eruptions (Nairn and Self 1978). The resulting pyroclastic currents can be strikingly different. In a case of collapsing Plinian eruption plumes that may reach 30 km in height, large volumes of pyroclasts can collapse into pyroclastic flows that radiate outward from the eruption centre. Since Plinian eruptions are generally involve evolved and volatile-rich magma types, the resulting eruption clouds are charged with pumiceous pyroclasts. Such eruption plumes can generate pumiceous pyroclast-dominated pyroclastic flows or ignimbrites. Small eruption plumes generated by Vulcanian-style eruptions (Gourgaud et al. 2000) or vigorious lava fountain activity (Bertagnini et al. 1991; Mastrolorenzo et al. 1993; Marianelli et al. 1999; Wolff and Sumner 2000) intermittent with ongoing Strombolian-style eruptions commonly generate scoriaceous (more mafic) pyroclasts that can initiate high temperature pyroclastic flows comprising hot scoria and ash (Gardeweg et al. 1998). Such currents termed as scoria-and-ash flows. Scoria-and-ash flows can also be generated by collapse of gradually accumulating scoriaceous deposits around an active vent because of over-steepening. Such an unstable pile of pyroclasts can perhaps collapse initiated a subsequent eruption. One major type of pyroclastic flow is associated with lava dome evolution on composite volcanoes. Lava domes are slow growing lava accumulations in the summit crater that slowly degas, and form a thickening crust. Such lava domes are hydrothermally altered and over time can become gravitationally unstable (Plate III, 3). When the lava dome reaches an unstable state, a relatively small volume, less energetic explosive eruption, commonly triggered by interaction between hot melt and hydrothermal systems, can lead to collapse of the lava dome (Bourdier et al. 1997; Sparks 1997; Nakada et al. 1999; Navarro-Ochoa et al. 2002; Behncke et al. 2003). Such dome-collapses then form a hot, avalanche-like current composed of large blocks derived from the lava dome and finer grained, generally ash-sized pyroclasts derived from the explosive eruption (Zobin et al. 2002; Behncke et al. 2003). These currents are termed as block-and-ash flows. Dome collapse and edifice failure is known to reoccur and produce repeated episodes of block-and-ash flows and volcanic debris avalanches, as has been demonstrated on many volcanoes, e.g. Kamchatka, Russia (Ponomareva et al. 1998; Belousov et al. 1999). Many block-and-ash flows are directly related to significant and energetic explosive eruptions and can be very hot. Such block-and-ash flows are also known as nuée ardentes or hot avalanches (Westercamp 1987; Bourdier et al. 1989; Lajoie et al. 1989; Boudon et al. 1993; Tanguy 1994; Abdurachman et al. 2000; Voight and Davis 2000; Carn et al. 2004; Tanguy 2004). Explosive eruption-triggered block-and-ash flows are commonly associated with laterally directed blasts that open up and destabilise the growing lava dome as occurred in the 1980 eruption of Mt St Helens (Hoblitt et al. 1981; Voight 2000). However, edifice failure also can also provoke violent explosive activity, as was the case in the 1964 eruption of Shiveluch in Kamchatka, Russia (Belousov 1995). A similar directed blast was formed during the 1956 eruption of Bezymianny volcano, Kamchatka, Russia (Belousov 1996). This blast was generated by decompression of an intracrater dome and cryptodome that had formed during the preclimactic stage of the eruption (Belousov 1996).

The particle transportation and depositional processes are very similar for either type of pyroclastic flows. Many authors suggest that the resulting pyroclastic flow deposits do not always preserve key textural characteristics that allow us to interpret the transportation and depositional processes of pyroclastic flows (Wilson and Houghton 2000). The transportation and depositional mechanism of pyroclastic flows are therefore the subjects of ongoing research and debate. Two contrasting depositional mechanisms are inferred (Wilson and Houghton 2000) on the basis of the sedimentary textures of pyroclastic flow deposits;1) the progressive aggradation model (Fisher 1966, 1979, 1983; Fisher and Schmincke 1984; Branney and Kokelaar 1992, 1994; Fisher and Schmincke 1994; Kokelaar and Branney 1996) which postulates continuous deposition of an active flow base in the entire run out length of the flow. In this model the deposits therefore only give information about the base of the flow and the overall pyroclastic flow should be viewed as a high-density turbidity current from a physical point of view (Kneller and Branney 1995); 2) “en masse” freezing model which states that the entire flow freezes in a single moment (Wright and Walker 1981). The textural characteristics would therefore represent the textural characteristics of the flow itself. In this model, pyroclastic flows are viewed as cohesive debris flows (Vallance and Scott 1997; Capra and Macias 2000, 2002; Capra et al. 2002; Lecointre et al. 2002).

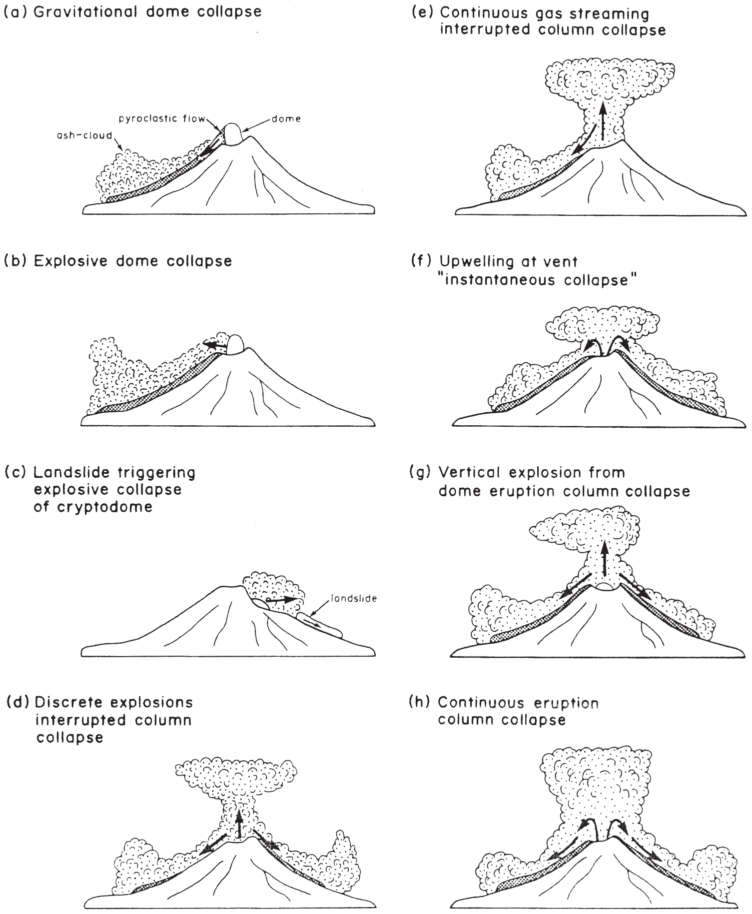


***Figure 6.3.*** Types of pyroclastic flow generation according Cas and Wright 1988: p. 106, fig. 5.11

### Pyroclastic flow deposits

Pyroclastic flow deposits result from all types of pyroclastic flows and predominantly consist of juvenile particles (Fisher and Schmincke 1984; Cas and Wright 1988; Fisher and Schmincke 1994). Accidental lithics that have been picked up from the volcanic conduit and/or en route during deposition are relatively minor component in the resulting deposits. Juvenile particles are predominantly either pumice or scoria, depending on the source magma chemistry. Volcanic lithic fragments are especially common in block-and-as flows, where material is derived from the disrupted and collapsed lava dome (Plate III, 4). Pyroclastic flow deposits can also be associated with accretionary lapilli-bearing beds where the eruption was partially phreatomagmatic.

Pyroclastic flow deposits are generally thick-bedded, matrix-rich and unsorted deposits that are commonly associated with a thin basal ground surge and overlying ash cloud surge beds (Figure 6.4). The three major types of pyrcolastic flow deposits are common in textural point of view. Block-and-ash flow deposits are rich in non-to-moderately vesicular lapilli and ash. Clasts in general are angular, and many of them show high temperature oxidation alteration. Larger lapilli and block size fragments are commonly jointed (Plate III, 5). The deposits are generally non-welded, however, they commonly show evidence of high temperature depositional environment, such as large charcoalized logs, and as uniform magnetic fabric. Scoria-and-ash flow deposits are predominantly composed of scoria, many of them bed-flattened, especially in near vent locations (Plate IV, 1). Ignimbrite deposits (Plate IV, 2) are pumice rich and fine glassy shards comprise the matrix of the deposits. Gas escape pipes are common, especially in thick ignimbrite units (Plate IV, 3). Welding textures (Plate IV, 4) are common features in ignimbrite deposits and reflect of heat retantion of thick accumulations of pumiceous deposits. The base of a pyroclastic flow deposit is commonly marked by thin, cross bedded deposits from ground surges [(Plate IV, 5) immediately preceding the arrival of the main current body of the pyroclastic flow (Sparks and Walker 1973). The base of the pyroclastic flow deposit itself is richer in dense volcanic lithic fragments that may form lenses, or clast strings. The main flow deposit unsorted, and enrichment of pumiceous and/or scoriaceous fragments up-section is common (Plate V, 1). It is commonly covered by fine ash deposits that may be massive or slightly stratified. This unit is deposited from ash-clouds that emanate from passage of the main body of the pyroclastic flow current (Fisher et al. 1980a; Vazquez and Ort 2006).

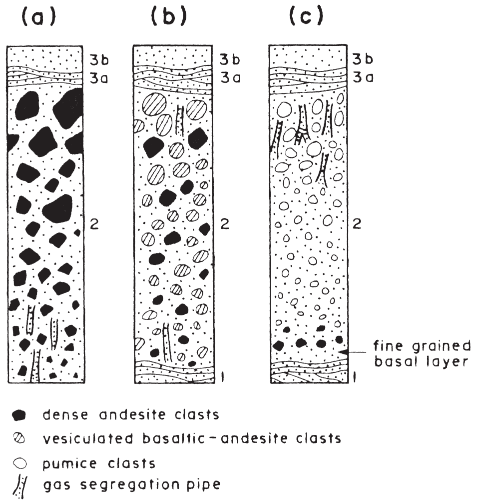


*Figure 6.4.* Ideal, simplistic sections of major pyroclastic flow deposit types (after Cas and Wright 1988). (a) block and ash flow deposit, (b) scoria-flow deposit, (c) pumice-flow deposit or ignimbrite

In pumiceous pyroclastic flow deposits welding can occur (Freundt 1998; Sparks et al. 1999; Beddoe-Stephens and Millward 2000; Sparks et al. 2000). The high temperature of the individual pyroclasts can retain heat long enough that could lead to a partial melting of the fine-walled, low density clasts of evolved composition (Wilson and Hildreth 2003). During the welding process, pumiceous and scoriaceous pyroclasts become bed-flattened, and their vesicularity decreases due to the gradual collapse of vesicles. The more or less bed-oriented flattening of the pyroclasts and the associated plastic deformation of the entire unit lead to the formation of dense juvenile fragments of lenticular shape, referred to as fiamme (Plate V, 2). The glassy shards of the matrix can sinter together to form a eutaxitic texture and produce the coherent, rock-like matrix of the fiamme-bearing succession. The welding process can be very effective in thick pumiceous deposits, and could lead to the development of a single cooling unit over the entire succession. The degree of welding is expressed in the grade of pyroclastic deposits (Freundt 1998; Sumner and Branney 2002). High grade pyroclastic flow deposits are rheomorphic, and can form lava-like units which are especially hard to distinguish from coherent lava bodies, particularly in ancient settings (Branney et al. 1992; Smith and Cole 1997; Freundt 1998; Mukhopadhyay et al. 2001; Sumner and Branney 2002; Allen 2004). Medium grade pyroclastic flow deposits are moderately welded (Duncan et al. 1999). Low grade pyroclastic flow deposits are non-welded (Szakács et al. 1998; Yokoyama 1999; Edgar et al. 2002; Thouret et al. 2005).

The distribution of pyroclastic flow deposits strongly depends on the morphology (Plate V, 3). Pyroclastic flows tend to follow topographic lows. Energetic pyroclastic flows, however, can flow uphill, flow over obstacles (Plate V, 4) (hundreds metres scale) (Woods et al. 1998; Legros and Kelfoun 2000; Gurioli et al. 2002) even up to hundreds of metres high (Taupo). Vegetation is commonly completely destroyed by pyroclastic flows (Plate V, 5). Valley pond deposits are generally thick units deposited from the axis of the pyroclastic flow currents (Wilson 1991; Pittari et al. 2006) whereas veneer deposits are common along the valley margins and/or over obstacles. This lateral facies variation is very common, and the resulting deposits are distinguishable (Brown et al. 2003; Pittari et al. 2006). The veneer deposits are commonly cross bedded, stratified, and finer grained than those deposited in the axis of the valley (Fisher et al. 1980b; Walker et al. 1980, 1981; Wilson 2001; Giordano et al. 2002). From proximal to distal areas pyroclastic flow deposits also show some horizontal facies variations as a result of the loss of momentum by the current. Due to the physical properties of the pyroclastic currents, large but low density clasts can be transported far from their source. In contrast, dense volcanic lithic clasts decrease their grain size away from the vent. In near vent positions, proximal, coarse lithic breccia units are common (Druitt and Sparks 1982; Druitt 1985; Druitt and Bacon 1986; Druitt 1995; Wilson 2001). Compositional zonation has been recognized in some large volume pumiceous pyroclastic flow deposits as a result of the ongoing and rapid emptying of the magma chamber (Edgar et al. 2002; Sumner and Branney 2002). Magma mixing has also been recognized in large volume ignimbrites and may be a triggering mechanism for the eruption (Briggs et al. 1993; Freundt and Schmincke 1995; Hildyard et al. 2000; Troll and Schmincke 2002).

Pyroclastic flow deposits because of the diverse mechanism of formation are very variable in volume. Scoria-and-ash flow deposits in general are less than 1 $km^3$ by volume. Large ignimbrites associated with caldera formation however, can form in excess of 1000 $km^3$ deposits (Plate VI, 1). The generally large volume of pyroclastic flow deposits means that they are well preserved in the geological record. In long lived composite volcanic systems, pyroclastic flows can accumulate deposits to thicknesses of hundreds of metres, and therefore may play important role in the evolution of a sedimentary basin.

### Calderas and sedimentation associated with silicic volcanism

Calderas are large depressions formed by subsidence driving voluminous volcanic eruptions (Lipman 2000). They are in general circular in map view, and can reach of depth of hundreds of metres (Plate VI, 2). They range up to several kilometres across and are commonly filled by caldera lakes (Plate VI, 3–4), and post-caldera eruptive deposits (Plate VI, 5) and associated reworked sediments. Caldera formation is generally associated with Plinian-style eruptions that form extensive Plinian pumiceous deposits and associated pyroclastic flows (ignimbrites) (e.g. Campanian Ignimbrite, Campi Flegrei, Italy) (Plate VII, 1). Pre-caldera successions are exposed in and the caldera walls and are overlain by syn-caldera volcaniclastic deposits and effusive products (Lipman et al. 1984). Very young calderas regardless to their well-preserved morphology do not give vital information about the subsidence processes and the root of the caldera itself. After formation of a caldera, the central zone of the volcanic system becomes structurally unstable. Where the magmatic plumbing system remains active after caldera subsidence, the empty magma chamber being refilled by post-caldera magmas. The central zone of the volcanic structure may be raised (Acocella et al. 2000; Lindsay et al. 2001a; Masturyono et al. 2001). The uplift can be pushing significantly up the central part of the caldera (Masturyono et al. 2001) causing a significant geomorphic inversion over short period of time. Uplift can range in hundreds of metres scale and can cause a significant geomorphic inversion over relatively short period of time (thousands of years) (Hulen and Nielson 1991; Battaglia et al. 1999; Newman et al. 2001). In the central part of the resurgent caldera system, caldera lake sediments can be raised to high stratigraphic positions, and post-caldera small-to-medium volume composite volcanoes can develop (Krupp 1984; Chen et al. 1995; Tibaldi and Vezzoli 1998; Moran-Zenteno et al. 2004). The inversion results in the formation of resurgent dome emplacements. During the post-caldera formation, newly emplaced magma in the shallow feeding system can evolve and accumulate significant volume of volatiles, leading to further caldera forming eruptions. A resurgent caldera can experience repeated caldera subsidence, forming a complex architecture in the caldera volcano (Orsi et al. 1996; Di Vito et al. 1999). Such subsidence and resurgence can be cyclic, and lead to the development of nested, complex calderas (Tibaldi and Vezzoli 2004; de Vita et al. 2006).

Although caldera forming eruptions are common in the geological record, few have been documented in historic times. These rare occurrences are generally small volumes and include the 1991 Pinatubo, Philippines (Rosi et al. 2001), 1968 Fernandia, Galapagos (Rowland 1996), 1912 Katmai, Alaska (Hildreth and Fierstein 2000), 1883 Krakatau, Indonesia (Deplus et al. 1995) or 1750–1790 Kilauea, Hawaii (Swanson and Christiansen 1973) eruptions. For example, the Pinatubo eruption only produced an approximately 2 km diameter caldera and formed up to 5 km$^3$ of dense rock equivalent (DRE) eruptive products. In the Cenozoic, however, there are many large volume caldera structures (Wilson et al. 1995) with deposits of hundreds of km$^3$ in volume. Such eruptions have also been considered to have had climatic effects as well. Mega-calderas from the Cenozoic geologic record indicate eruption with potential devastating effects.

Especially older, and resurgent, and therefore well-exposed volcanic successions of calderas indicate a typical caldera cycle (Figure 6.5) most of the calderas went through during their life (Lipman 1976; Colucci et al. 1991; Lipman et al. 1996; Lipman 1997). The sequence of events leading to caldera formation is marked by pre-caldera volcanism (Colucci et al. 1991). During this stage small magma batches erupt from initial accumulation of melts in shallow magma chambers. The gradual filling of the shallow magma chambers causes general uplift of the central region of the volcanic zone and formation of ring fractures and associated radial fractures, many of them filled with dykes. In this stage of the caldera development, magma may reach the surface and form a radial network of small-to-medium volume composite volcanoes, commonly with associated, long-lived effusive phases. Caldera subsidence takes place after favourable structural and magma chemical conditions in the central magma plumbing system trigger rapid eruption of large volumes of magmas. Although the resulting calderas are diverse in morphology, size, and the chemistry of the formed eruptive products, the generated calderas have a characteristic structure and morphology (Figure 6.6). The caldera rim is marked by a characteristic escarpment facing toward the central part of the caldera. This rim encircles both the structural boundary of the subsidence feature as well as the talus deposits formed by the collapse of the retreating topographic margin of the caldera. The inner topographic wall of a caldera forms a concave architecture and exposes pre-caldera successions, which may be covered by talus deposits formed by mass wasting in the rim. The collapse collar of the caldera is the region located between the structural caldera boundary and the inner topographic wall. In young calderas the collar slope angle (the line from the structural boundary to the topographic rim) is about 45 degrees but this can decrease to10-15 degrees in eroded structures. The structural caldera boundary is marked by bounding faults along which the caldera subsidence took place. These faults are ring-like and only exposed in older calderas where the caldera fill is removed and/or post-caldera

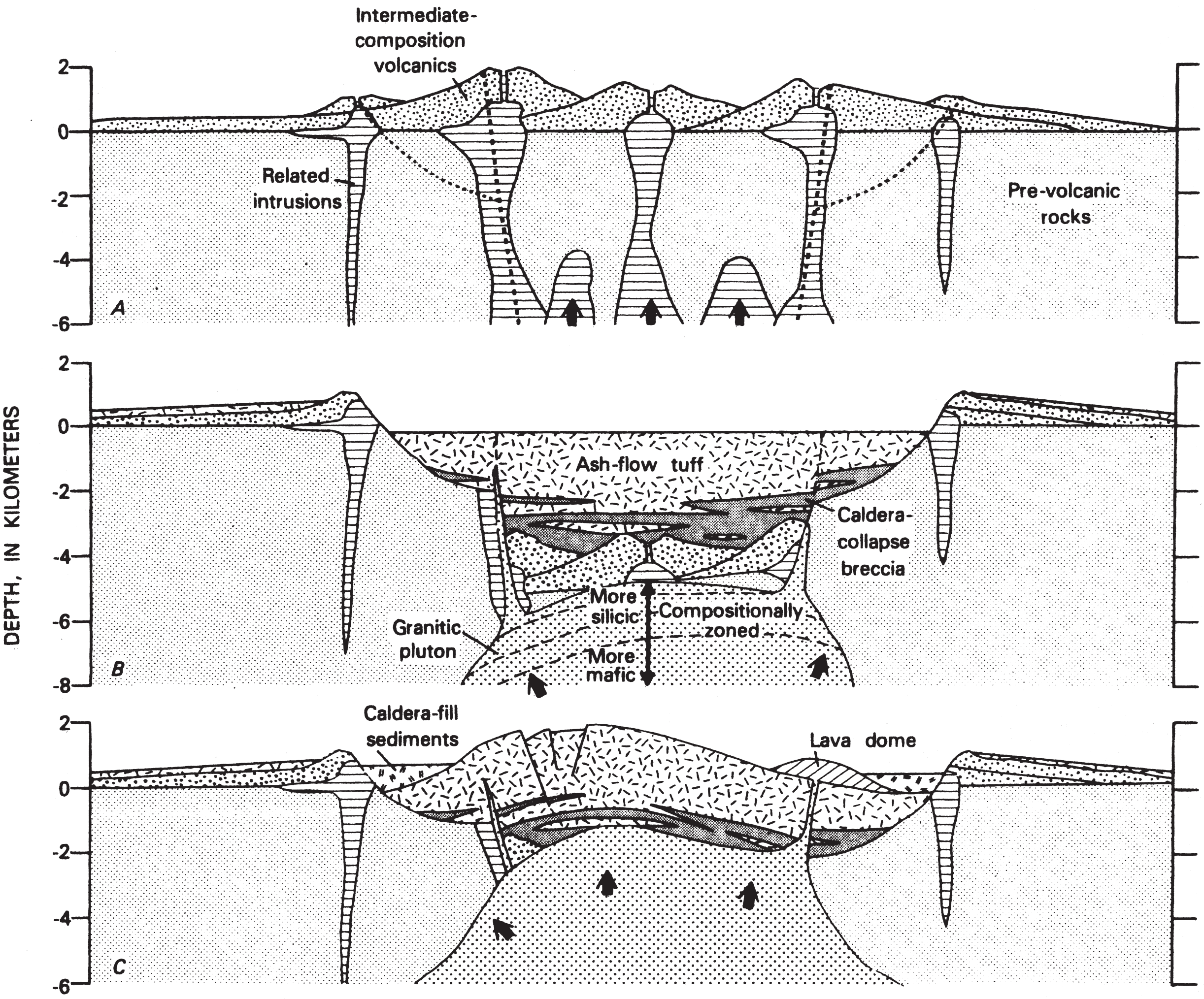


*Figure 6.5.* Caldera cycle after Lipman 2000: p. 648, fig. 2]

resurgent uplift makes the deeper structure accessible. The calderas are generally filled with intra-caldera ash flow deposits and subsequent lacustrine successions. In the caldera topographic boundary, large blocks and caldera margin breccias may be associated with intra-caldera ignimbrites. The latter are commonly welded with multiple cooling units, and may form a strong, cap-like succession. The caldera floor is generally defined by the structural caldera floor, located well below the intra-caldera pyroclastic successions. The sub-caldera magma chamber (or remnant of it) can be found below the structural caldera floor in resurgent calderas, where the deep structures of the caldera can be in uplifted and exposed (Lipman 1984; Johnson et al. 2002). These magma chambers usually consist of plutonic rocks with strong hydrothermal alteration and common mineralization. As a final stage of the caldera cycle, small-to-medium volume volcanic edifices may develop in the central part of the caldera.

The most important evolutionary stage in the development of a caldera is its subsidence. Study the structural architecture of the root zone of calderas, as well as detailed morphological analysis of young calderas aided by scaled model experiments, suggested few basic lines of subsidence styles (Figure 6.7) (Roche et al. 2000; Walter and Troll 2001; Troll et al. 2002; Holohan et al. 2005). The most common subsidence style is the piston or plate style of collapse of the caldera floor (Acocella et al. 2000; Roche and Druitt 2001; Folch and Marti 2004). In this style the subsidence takes place along near vertical faults and the entire floor subsides as a plate. Trapdoor subsidence is an asymmetric type of subsidence, where an initial downsagging of the caldera floor takes place, followed by a plate-like col-

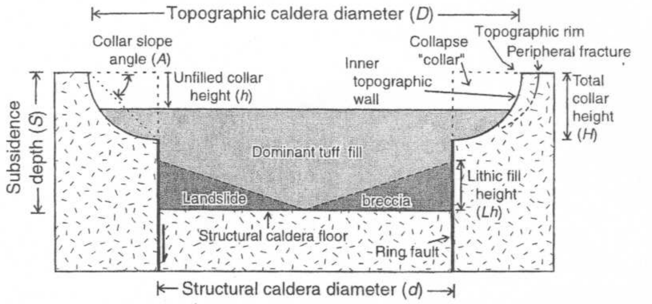


*Figure 6.6.* Morphological and structural elements of a caldera after Lipman 2000: p. 649, fig. 3

lapse (Mahood and Hildreth 1983; Beresford and Cole 2000; Milner et al. 2002; Kennedy et al. 2004; Ramelow et al. 2006). Pure downsagging is apparently rare (Ort 1993). Piecemeal fault controlled subsidence is very common in many calderas, as a result of differential subsidence along many well defined structural weakness zones (Rosi et al. 1996; Moore and Kokelaar 1997, 1998; Milner et al. 2002). Such subsidence can result differently subsided blocks in the central part of the caldera floor. Any of the above types of caldera subsidence can lead to chaotic subsidence which is also probably the most common styles in complex and large mega-calderas. Funnel-shape subsidence is an end member of subsidence types, and probably only associated with small calderas. The internal structure of calderas can be studied when the resurgence of the caldera floor uplift the central part of the caldera as it is the case in the Valles Caldera in Texas, one of the most well-studied caldera in Earth (Plate VII, 2).

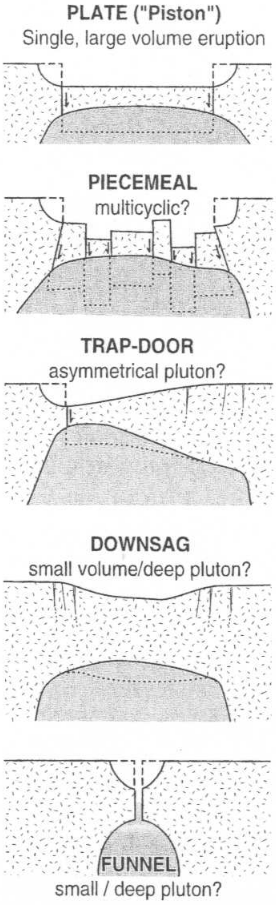


***Figure 6.7.*** Styles of caldera subsidence after Lipman 2000: p. 654, fig. 6

Well-defined calderas and associated ignimbrite sheets are well known and well described in many Cenozoic calderas. In general the identification of large volume and extensive ignimbrite sheets are therefore often used to infer the location of associated calderas. However, in many regions very extensive ignimbrite sheets (Plate VII, 3) had no apparent association with caldera structures, such as the widespread and large volume Cenozoic ignimbrites in the Andes (Lebti et al. 2006) or in Armenia. Such silicic ignimbrites are commonly referred as ignimbrite shields (de Silva 1989). Only in recent times have researchers recognised that these ignimbrite shields are also associated with caldera of structural elements (Ort 1993; Lindsay et al. 2001b; Richards and Villeneuve 2002). Identification of caldera structures, and separating their structural elements from regional tectonic structures is difficult in areas where ignimbrite-forming eruptions are associated with regional extension and volcanic products accumulate in a volcano-tectonic structure, such as the Taupo Volcanic Zone of New Zealand (Wilson et al. 1995). Similar problems hinder the identification of the source of extensive Neogene basin-wide ignimbrite sheets in the Miocene Pannonian Basin (Pantó 1963; Szabó et al. 1992; Capaccioni et al. 1995; Póka et al. 1998; Szakács et al. 1998). Similar volcano-tectonic graben structures are inferred to contour a chain of large calderas in the Tokaj Mountains in Hungary. However, more detailed studies still need to be done confirm this interpretation of the preserved volcanic units (Molnár and Zelenka 1995; Seghedi et al. 1998; Bajnóczi et al. 2000; Pécskay and Molnár 2002; Martin et al. 2003). In the Taupo Volcanic Zone, it also has been recently recognized, that certain ignimbrite sheets can be associated with previously not recognized eruptive centres, and growing number of hard to recognize calderas have been identified (Nairn et al. 1994; Wilson et al. 1995; Beresford and Cole 2000; Wilson 2001; Milner et al. 2002, 2003).

The size of calderas is very diverse, but many evidences in the geological record suggest that even the larger calderas can form in a relatively short period of time. Since we don't have experience of such eruptions in our history, the volcanic hazard implication of caldera eruptions is poorly understood. They are however, potentially the largest and most hazardous eruptions on Earth. Such eruptions commonly are considered to also be climate-forcing and even effect evolution (Fedele et al. 2002; Oppenheimer 2003a, b; Manville and Wilson 2004; Self 2006). However, because of the more evolved composition of magmas known to be involved in caldera formation, the climatic effect of such eruptions could be smaller than those medium-volume, but more mafic composition volcanoes that produce larger volumes of sulphur aerosols, known to have a large impact on the Earth's thermal balance.

## Post volcanic hydrothermal activity

Post-volcanic hydrothermal activity of polygenetic volcanoes, especially silicic ones are is very diverse and play major role in mineralization. In caldera systems, the shallow magma chamber still can provide heat source effective enough, to heat ground water, and produce solfatara, mofetta or geyser activity long time over eruption, and/or inter-eruption periods. Solfatara (Plate VII, 4) is a dominantly sulphur producing vent named after Solfatara, a small post-caldera tuff ring in the Campanian Field in Italy. Mofetta is a $CO_2$ producing vent, commonly associated with solfatara fields. Major solfatara fields in silicic volcanic systems are commonly accompanied with small hydroexplosion sites (Plate VII, 5). Due to the elevated heat of the shallow magma chamber provide to the ground water, overpressurised hydrothermal systems, time to time can outburst into hydroexplosions, and may form few tens of metres wide craters surrounded by dm-to-m thick chaotic breccias (Plate VIII, 1) dominated by hydrothermally altered volcanic lithic fragments. Such hydroexplosion (phreatic explosions) sites are commonly forming extensive (few km$^2$ area) fields (Plate VIII, 2), a perfect sites for mineralization. Explosion craters are commonly filled with thermal water and can form reasonable sized volcanic craters similar in architecture to maars (Plate VIII, 3). Water level of such craters can change dramatically (Plate VIII, 4). Hydrothermal fields commonly have geysers, and associated outflow springs rich in minerals. Such mineral-rich waters may accumulate laminated lacustrine beds in shallow hot water pools in the thermal areas (Plate VIII, 5). In ancient settings similar shallow lacustrine systems maybe preserved in a form of laminated fine grained lacustrine sediments composed of angular volcanic lithics as well asclay minerals such as deposits in the Miocene Tokaj Mts in NE-Hungary (Plate VIII, 6). Hydroexplosions can form over ignimbrite sheets. Ignimbrite can retain heat long enough after deposition and generate overheated steam that may disrupt in an explosion the overlying ignimbrite sheets. Such eruptions are documented from the Mount St Helens. The resulting deposits and crater morphology is similar to those formed by phreatomagmatic eruptions, however, there is no juvenile magma involvement in such hydroexplosions, and therefore the deposits will be only consist of accidental lithics derived from the ignimbrite units.

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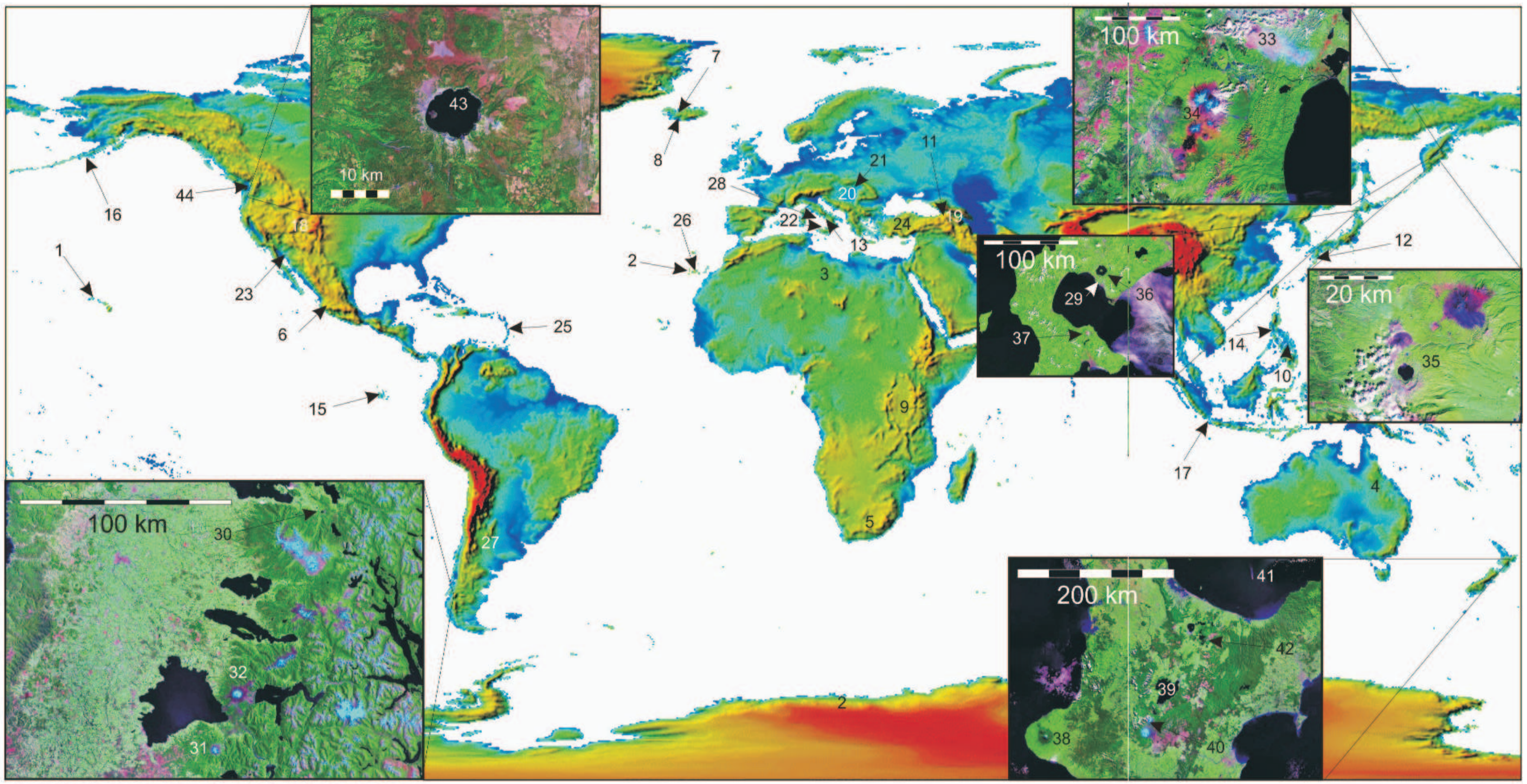
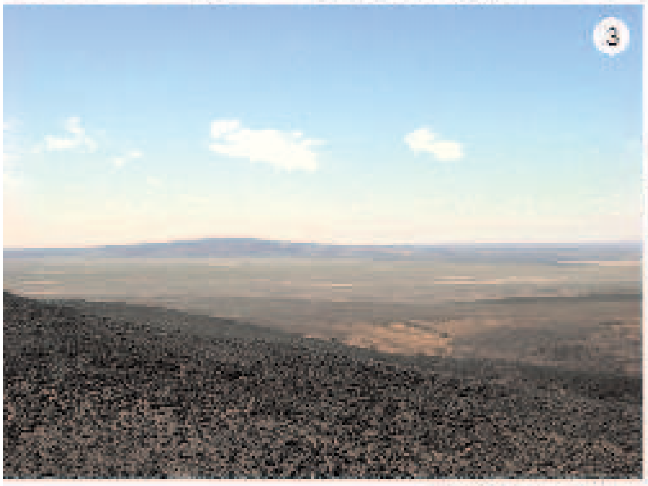
# Location map



1 — Hawaii, Kilauea, USA
2 — Tenerife, Spain
3 — Al Haruj, Libya
4 — Eastern Australia
5 — Karoo, South Africa
6 — Ceboruco, Mexico
7 — Hekla, Iceland
8 — Askja, Iceland
9 — East African, Rift
10 — Mayon Phillippines
11 — Ararat, Turkey
12 — Unzen, Japan
13 — Campi Flegrei, Italy
14 — Pinatubo, Phillippines
15 — Fernandia, Galapagos
16 — Katmai, Alaska, USA
17 — Krakatau, Indonesia
18 — Valles Caldera, Texas, USA
19 — Armenia
20 — Pannonian Basin
21 — Tokaj Mts, Hungary
22 — Vulcano, Italy
23 — Santa Catarina Mts, Sonora, Mexico
24 — Cappadocia, Turkey
25 — Mt. Peleé, Martinique
26 — Gran Canaria, Spain
27 — Mendoza, Argentina
28 — Ischia, Italy
29 — Usu, Japan
30 — Rininahue, Chile
31 — Calbuco, Chile
32 — Osorno, Chile
33 — Shiveluch, Russia
34 — Bezymianny, Russia
35 — Karymsky, Russia
36 — Toyo Caldera, Japan
37 — Hokkaido-Komagatake, Japan
38 — Taranaki Volcano, New Zealand
39 — Lake Taupo, New Zealand
40 — Ngauruhoe Volcano, New Zealand
41 — White Island, New Zealand
42 — Tarawera, New Zealand
43 — Crater Lake, Oregon, USA
44 — Mt St Helens, Washington, USA

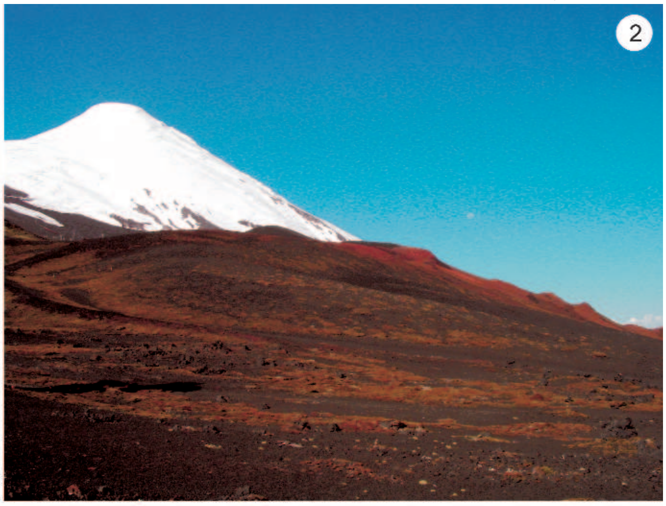
## Plate I



1. Shield volcano of the Tenerife Island (in horizon) having broad base composed of thick succession of lava flows and subordinate pyroclastic rock units similar to the succession in the foreground of Gran Canaria, Spain (photo by S. J. Cronin).
2. Hyaloclastite successions of the old shield edifice of Tenerife Island, Canary Islands, Spain
3. Medium size Pliocene intracontinental shield volcano from the Al Haruj al Abyad volcanic field, Central Libya.
4. Low-relief, Pliocene medium size lava shields forming an amalgamated network of volcanic landforms along the Snake River Plain in Idaho.
5. Lava delta in cross sectional view along the Snake River valley, Idaho.

# Plate II

1. Crater of Volcan Ceboruco in the western Transmexican Volcanic Belt. The crater of Ceboruco is occupied by small volume lava coulees and domes (centre of view).
2. Satellite vents on the flank of the Santa Catarina shield volcano, in the Pinacate Volcanic Field in Sonora, Mexico.
3. Lava dome complex on top of an arc stratovolcano in Northern Chile forming a complex summit morphology of the stratocone.
4. Truncated summit region and morphology of Calbuco Volcano in Southern Chile a result of multiple collapse events.
5. Strongly modified and glaciated summit of a southern Chilean stratovolcano as a result of strong external erosional forces and multiple collapse events.



1. Conical shape of Taranaki stratovolcano, New Zealand.
2. Summit region of Osorno in Southern Chile with a young and fresh scoria dominated cone evolving over a truncated possible collapsed summit region of the volcano.
3. Hydrothermally altered "ready-to-collapse" architecture of the Hokkaido-Komagatake lava dome in Hokkaido, Japan
4. Cross section of block-and-ash flow deposits from the 1973–75 eruption of the Ngauruhoe volcano, New Zealand.
5. Prismatically jointed lava block in block and ash flow deposit from the 1980 eruption of Mt St Helens (photo by S. J. Cronin).

# Plate IV

1. Scoria-and-ash flow deposit from the 1979 eruption of Ngarahue volcano, New Zealand. Note the levees formed by the channelised scoria-and-ash flow deposit.
2. Ignimbrite deposits from the Central Anatolian Ignimbrite Province, Zelve Valley, Cappadocia, Turkey.
3. Gas escape pipes preserved as pillars in ignimbrite deposits of the Mazama eruption, Crater Lake (photo by S. J. Cronin).
4. Welded ignimbrite unit of the Crater Lake, Oregon (photo by S. J. Cronin).
5. Ground surge deposit of Mt Pelee 1904 eruption exhibit dune bedded structure (photo S. J. Cronin).

1. Main pyroclastic flow unit in Miocene non-welded ignimbrite with elutriation pipes from the Tokaj Mtns, NE Hungary.

2. Fiamme in welded ignimbrite from Gran Canaria (photo by S. J. Cronin).

3. Valley filling ignimbrite pinching out in the palaeovalley margin (Mendoza, Argentina).

4. Ignimbrite deposits (white blanket on the flank of the cone) over scoria cones in southern Tenerife (Montana Pelada tuff ring) indicating high inertia of the pyroclastic flow running over tens of metres high obstacles.

5. Fallen trees on small hill side the 1980 pyroclastic surge and flow run over at Mt St. Helens (photo by S. J. Cronin).

# Plate VI

1. Extensive sheet-like ignimbrite plateau in the Altiplano, Northern Chile.
2. Taupo caldera with Lake Taupo, New Zealand. Note the steep escarpment along the caldera lake shore formed by ignimbrite successions.
3. Toyo caldera lake in MrSID satellite immage, Hokkaido, Japan. The islands in the lake (black) are resurgent domes.
4. Karymsky caldera in Kamchatka on oblique satellite image (Google Earth). Note the intra-caldera Surtseyan-style tuff cone in the shoreline of the caldera lake formed in 1996. The tuff cone is eroded by now, and only a pyroclastic mound left on the lake floor.
5. Crater Lake caldera with Wizard Island in its centre, as a post caldera scoria cone (photo by S. J. Cronin)

# Plate VII

1. Oblique view of a satellite image of the Campi Flegrei, a large caldera system produced the Campanian Ignimbrite. The caldera structure is occupied by small post caldera tuff rings, and scoria cones.
2. Resurgent caldera structure of Ischia in Italy on oblique satellite image (GoogleEarth).
3. Extensive ignimbrite shield in Northern Chile. The yellowish landscape on the entire view is an extensive ignimbrite shield formed in the Quaternary.
4. Solfatara field in the inner crater wall of Vulcano, Lipari Island, Italy.
5. Deep hydrothermal explosion crater in New Zealand.

# Plate VIII

1. Hydrothermal explosion breccia from New Zealand.
2. Hydrothermal explosion crater field in New Zealand.
3. Maar-like explosion craters around Tarawera, New Zealand.
4. Water level changes represented in narrow benches in the shoreline of an explosion crater near Tarawera, New Zealand.
5. Thermal water pools in thermal fields in the New Zealand.
6. Miocene hydrothermal deposits associated with a thermal area developed over Miocene silicic pyroclastic successions of the Tokaj Mtns, NE-Hungary.

# Chapter 7

# Depositional processes related to erosion of volcanic terrains

Volcanoes produce large amounts of brecciated materials that contribute to the construction of a stratocone, and accumulate around the volcanic edifice to form a ring plain or volcaniclastic apron. Large volume volcanic edifices (i.e. >100 km$^3$) form over long periods of time (e.g. hundreds of thousands of years). During periods of quiescence, volcanoes can be effected by external forces (e.g. climatic agents) that continuously remove parts of the volcanic edifice and redistribute volcanic debris to the surrounding terrains (MATHISEN and MCPHERSON 1991, RAMPINO 1991, SCHMINCKE 2004). Redistribution of primary volcanic deposits can take place immediately after, or during the course of an eruption, when volcanic processes rapidly initiate secondary depositional mechanisms. Volcano-sedimentary facies evolve almost continuously between primary and secondary volcanic deposits, in almost any type of volcanic system.

Secondary volcanic processes that generate large volumes of sediments, such as lahars, may also be initiated by external forces such as heavy rainfall and storms, a common occurrence in tropical environments (SMITH and LOWE 1991, FISHER and SCHMINCKE 1994). Flank instabilities and sector-collapses have also been recognized (VOIGHT and ELSWORTH 1997, VOIGHT 2000) as key mechanisms leading to the catastrophic emplacement of volcano-sedimentary successions, especially around polygenetic, long-lived strato-volcanoes. Cone collapses generate volcanic debris avalanches that often transform into debris flows in distal reaches of the ring plain (SMITH and LOWE 1991, UI et al. 2000). Long-lived composite volcanoes usually reach an advanced stage of growth before flank collapse occur, resulting in exceptionally large volume deposits (tens of cubic kilometres) made of chaotic volcaniclastic successions accumulated around the central volcanic edifice (SIEBERT 1984, BEGET and KIENLE 1992, NEHLIG et al. 2001). Such catastrophic volcanic collapses are commonly triggered by moderately explosive eruptions (e.g. Shiveluch Volcano, Kamchatka; PONOMAREVA et al. 1998). However, such collapses could also take place during periods of quiescence, when mechanical failure of unstable portions of the edifice are suddenly triggered by hydrothermal circulations, seismo-tectonic events or dyke emplacement (VAN WYK DE VRIES and BORGIA 1996, VAN WYK DE VRIES 1998, VAN WYK DE VRIES and MATELA 1998, VOIGHT 2000, TIBALDI 2001, TIBALDI et al. 2003). In fact, gradual spreading of an unstable volcanic edifice over long periods of time (thousands of years) may also truncate, redistribute and modify original primary volcanic successions with no single catastrophic event involved (VAN WYK DE VRIES 1998, LAGMAY et al. 2000, CECCHI et al. 2004). Studying volcanic successions accumulated around stratovolcanoes may help to identify generating processes of lahars and volcanic debris avalanches that contributed to the construction of the ring plain. In both cases, the key question is, what the triggering mechanisms are, and how closely that triggering mechanisms are linked to eruptive activity (e.g.: was a volcanic eruption involved in the formation of lahars and volcanic debris avalanches or not?).

Volcanic sediment redistribution can also take place without any sudden, catastrophic mass-wasting. Instead, sustained erosion of the original volcanic edifice may lead to the development of specific erosional landforms, and the accumulation of volcaniclastic sediments in the surrounding sedimentary basins. Effects of (or impact of) long term erosion on volcanic landforms are poorly understood, and the link between the original edifice and the resulting accumulated volcaniclastic sediments may be difficult to establish with certainty in older volcanic terrains.

## Volcanic ring plain

Volcanic ring plain was introduced in the literature from studies of central New Zealand volcanoes (Table I, 1) where volcaniclastic sequences have been mapped in detail, and key facies changes observed between the central volcanic edifice and the surrounding low lands have been properly recorded (NEALL 1975, PALMER and NEALL 1991, PALMER et al.

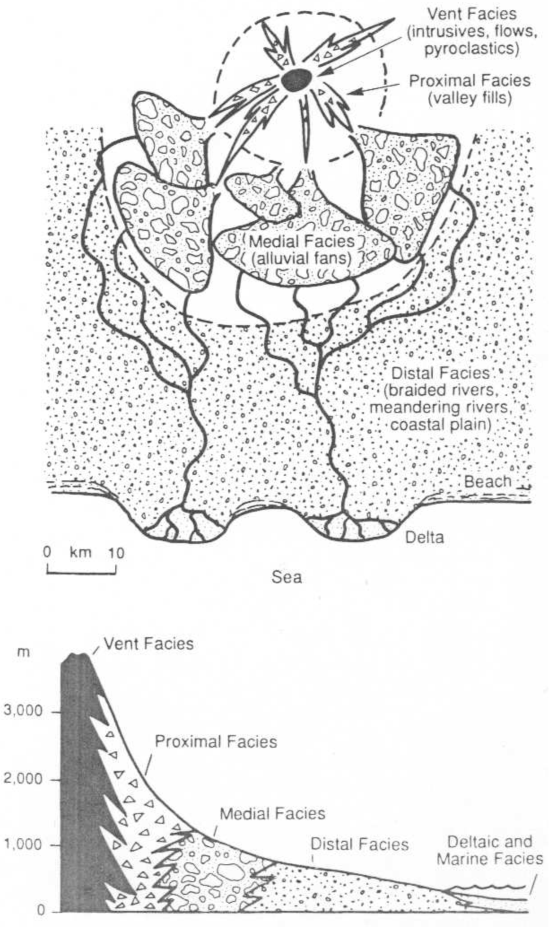


*Figure 7.1.* Diagram shows major facies associations around an active composite volcano (after Mathisen and McPherson 1991: pp. 29, fig. 3)

1993, Cronin and Neall 1997, Lecointre et al. 1998, Donoghue and Neall 2001). A volcanic ring plain is defined as the circular area that surrounds a centrally constructed volcanic edifice (usually a stratovolcano) (Figure 7.1). Ring plains are therefore sedimentary basins where both primary and secondary volcaniclastic products accumulate (Palmer et al. 1993). During periods of volcanic activity, the ring plain is progressively built with the deposition of primary volcanic products (Cronin et al. 1996b, 1996c). Layers of pyroclastic debris (ash, lapilli) mantle the volcaniclastic apron (Table I, 2), preferentially accumulating in sectors of the ring plain where dominant winds control the direction of the eruption cloud (Neall 1975, Palmer and Neall 1991, Alloway et al. 1995). Such fall deposits can be used successfully as marker beds for calibrating key lithostratigraphic sections, with an aim to establish the detailed record of volcanic and mass wasting events that affected a centrally constructed volcanic edifice and its surrounding ring plain. Because the wind direction change over time, each individual fall deposit may accumulate in a slightly different area of the volcanic ring plain, as demonstrated around the main central North Island volcanoes of New Zealand (Froggatt and Lowe 1990, Donoghue and Neall 1996, Lowe et al. 1998, Donoghue et al. 1999, Shane 2000) and elsewhere (Schmincke and van den Bogaard 1991, Andreastuti et al. 2000, Legros 2001). In extreme cases, fluctuating wind directions during successive eruption phases can lead to the accumulation of tephra beds that are not overlying each other (Figure 7.2). If magma sources are geochem-

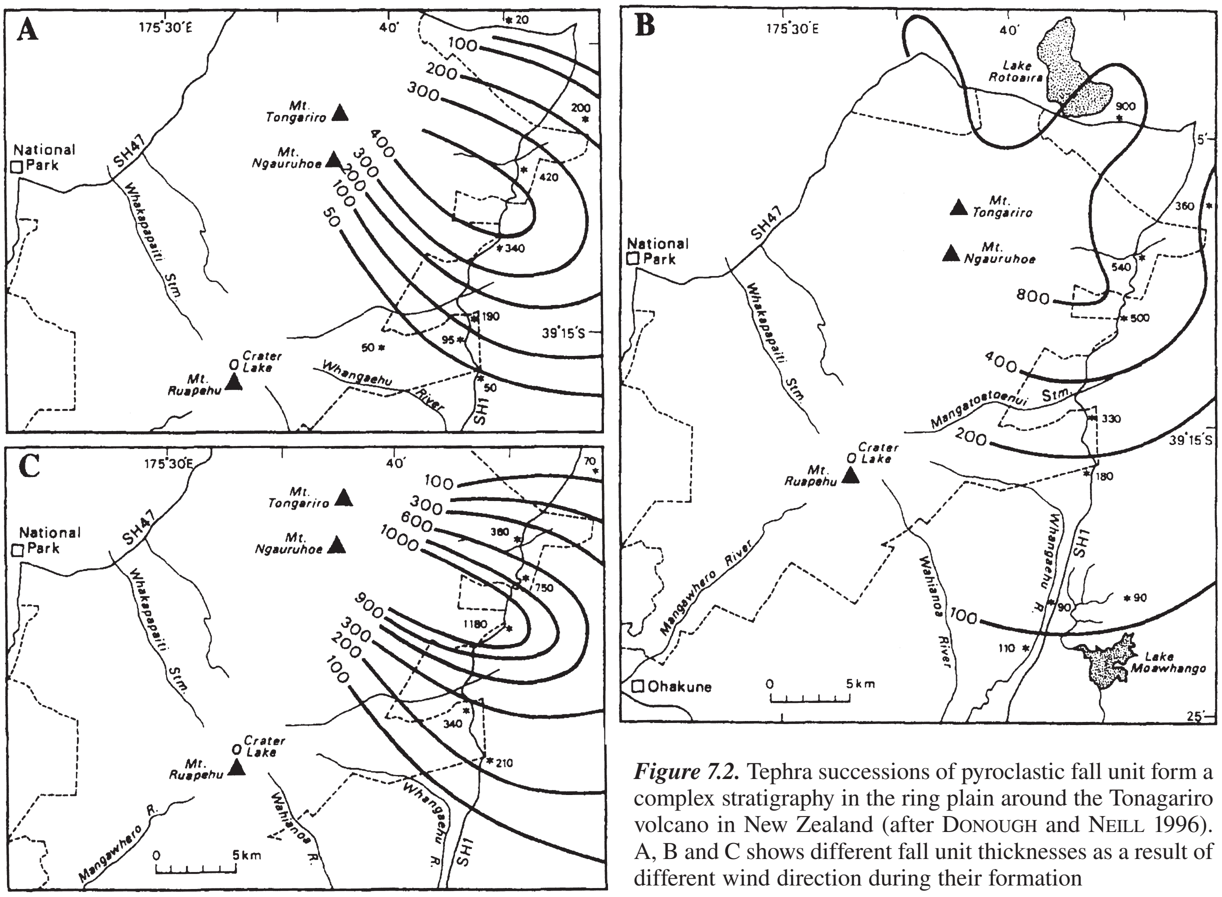


*Figure 7.2.* Tephra successions of pyroclastic fall unit form a complex stratigraphy in the ring plain around the Tonagariro volcano in New Zealand (after Donough and Neill 1996). A, B and C shows different fall unit thicknesses as a result of different wind direction during their formation

ically very similar, the correct identification of specific fall tephra units could be challenging, and thus, leading to incorrect tephrostratigraphic interpretations. Tephra accumulating on the ring plain are commonly intercalated with other volcaniclastic deposits resulting from erosional and/or major destructional processes that affected the volcanic edifice (Palmer et al. 1993, Lecointre et al. 1998, Lecointre et al. 2004). The ring plain is also a place where distal volcanogenic deposits sourced from pyroclastic flows and primary lahars may accumulate and mix (Palmer 1991, Cronin et al. 1997b, Lecointre et al. 1998). In ancient volcanic terrains, volcanic deposits commonly represent lithostratigraphic successions formed in a sedimentary basin around volcanoes, such as ring plain associations (Breitkreuz 1991, Spalletti and DallaSalda 1996, Aydar 1998, Corcoran et al. 1999, Cummings et al. 2000). Because of the incomplete preservation of primary deposits on the slopes of the cone, ring plain volcaniclastic sequences could reveal a more detailed chronostratigraphic record that include discrete volcano-sedimentary events of the past. Therefore, the study of these distal volcanogenic accumulations is an essential step to properly reconstruct the eruptive history of a stratovolcano. To understand the volcanic architecture of a complex, ancient volcanic system, it is thus very important to interpret volcanic terrains in the light of understanding the depositional processes that contributed to the construction of the volcanic ring plain.

## Lahars

Lahar is an Indonesian term defining a whole spectrum of laterally spreading volcanic mass flows such as debris flows, transitional flows or hyperconcentrated streamflows that originated from a volcano by any genetic processes (Smith and Lowe 1991, Fisher and Schmincke 1994, Lavigne and Thouret 2000, Vallance 2000). Lahar therefore is a very broad type of gravity-driven volcaniclastic current that can produce many different types of deposits (Figure 7.3). Lahar as a term should be used to describe the process, not the deposit (Smith and Lowe 1991). Lahars deposits therefore have to be carefully examined in order to determine the physical process(es) responsible for their emplacement. Lahars can be generated directly by a volcanic eruption (Waitt et al. 1983, Mothes et al. 1998) or by sudden rainfall that quickly remobilise volcanic debris on the flanks of an active volcano (Hodgson and Manville 1999, van Westen and Daag 2005). Lahars can also be triggered by catastrophic precipitations induced by hurricanes over a dormant volcano such as the Casita in 1998 in Nicaragua (Scott et al. 2005). Deposits resulting from these two distinct processes could look very similar to each other by their texture, bedding characteristics or composition. Lahars start often as minor debris flows that may transform into debris flows incorporating large volume of volcanic and non-volcanic debris en-route (Smith and Lowe 1991, Vallance 2000). Lahars commonly reach the ring plain (Table I, 3) and form extensive sheet-like deposits intercalated by other primary deposits such as ash layers (Palmer et al. 1993). Lahars represent a very significant hazard in volcanoes located in tropical countries such as Mt Merapi in Indonesia (Lavigne et al. 2000, Lavigne and Thouret 2003), or Mt Pinatubo in the Philippines (Chorowicz et al. 1997, van Westen and Daag 2005, Carranza and Castro 2006). Ancient lahar deposits interbedded with palaeosols and loesses may also provide clues on palaeoclimatic conditions of the region where the volcano erupted. Lahars are also commonly produced by active volcanoes that have craters occupied by a lake, such as Mt Ruapehu in New Zealand (Cronin et al. 1996a, Lecointre et al. 1998, Manville et al. 1998, Cronin et al. 1999). Eruption through a crater lake may suddenly expel large amounts of water on the steep, tephra-covered, upper slopes of the cone. This violent outpouring of hot and

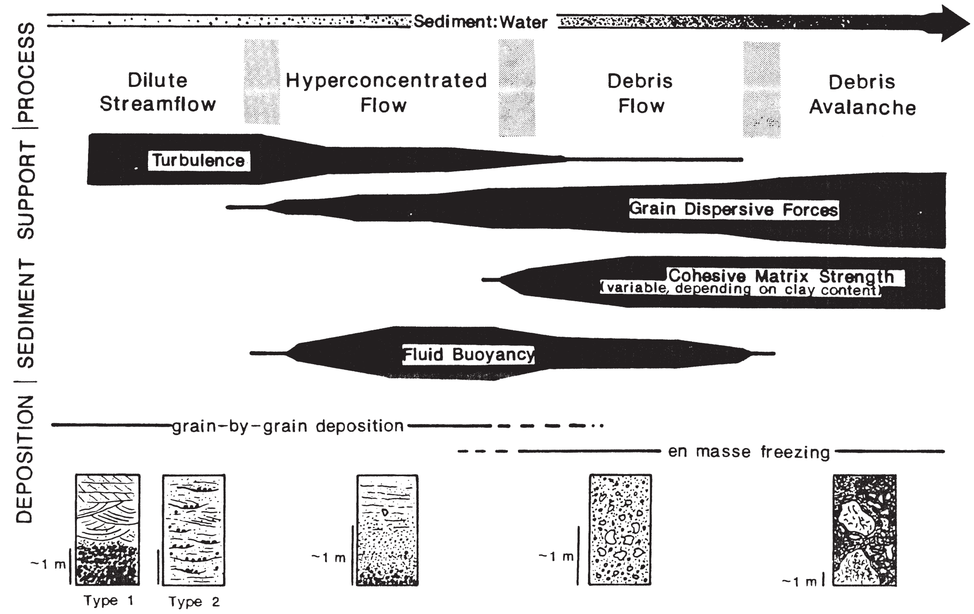


*Figure 7.3.* Relationship between lahar types, their deposits, and resulting theoretical sedimentary structures after Smith and Lowe 1991: p. 60, fig. 1)

often acidic waters initiate debris flows that by bulking process, will incorporate large volume of extra sediment (sand, boulders) on the way down to the ring plain (Mastin and Witter 2000). Eruptions through crater lakes are considered to be dangerous, a common hazardous situation that affect exposed populations on volcanic islands such as Ambae in Vanuatu (Németh et al. 2006a).

Lahars can also be initiated by sudden melting of a glacier, or snow- and ice-cap (Table I, 4 and 5), that often cover the summit region of a high composite volcano (Manville et al. 2000) and may lead to the rapid formation of "snow slurry lahars" (Cronin et al. 1996a). Such processes can be triggered by a growing lava dome beneath the summit of the volcano. An elevated heat gradient, due to hydrothermal and magmatic fluids in the upper flank of the edifice, may lead to the sudden melting of the snow-and-ice cap. Such catastrophic process is unfortunately very common on the highly elevated volcanoes of the Andean volcanic arc, and is responsible for devastating mudflow inundations in surrounding valleys (Branney and Gilbert 1995, Stern 2004). Lahars initiated from the sudden melting of an ice cap are able to carry large ice blocks (tens of m$^3$) that can be deposited at great distances from source, downstream the river catchment. After melting of the ice, voids may form in the sandy deposit that once collapsed, will be filled with chaotic debris (Cronin et al. 1996a). Lahars also carry large volumes of eroded soil, vegetation, and truck- or house-sized debris from infrastructure destroyed on their path.

Lahars are also commonly able to run over small obstacles (tens of meters high), flow uphill over small morphological barriers, and can reach long distances (tens of kilometres) from their source (Carrasconunez et al. 1993, Kohlbeck et al. 1994, Mothes et al. 1998, Hall et al. 1999). Such lahars are therefore a real threat to unaware populations that live or travel in areas generally considered to be safe from a distant — and often dormant — volcano. Large volume deposits resulting from these catastrophic lahars may represent a significant proportion of the sedimentary basin fill, even at far distances from the source volcano. This observation implies that identification of laharic deposits in ancient settings should be conducted very carefully, in order to determine the exact source location of the studied diamictons (Walton and Palmer 1988, Spalletti and DallaSalda 1996, Karátson and Németh 2001, Nehlig et al. 2001, Encinas et al. 2006). Lahars generally flow down the valley network radiating around the volcano, and accumulate deposits in major basinal areas of the ring plain. The sudden input of large volume of volcanic debris in existing fluvial networks may significantly alter the course of the main river and its secondary tributaries (Table II, 1). Therefore, modelling and prediction of lahar inundation areas is an essential but complex task, especially in volcanic fields where lahars can occur frequently without too much warning (e.g. in tropical volcanic islands).

The generation of lahars and the volume of resulting deposits could vary hugely. After the Pinatubo 1991 eruption, tephra covering vast expenses of bare surfaces was quickly remobilized into lahars after heavy rain falls, with no volcanic eruption involved (Gran and Montgomery 2005, van Westen and Daag 2005, Carranza and Castro 2006). Over a 3 years period, 30% of the total tephra volume produced by the 1991 eruption have been remobilized and accumulated in stream valleys and onto the Pinatubo ring plain (Torres et al. 2004). Lahar deposits can be subsequently recolonized by vegetation once soil development occur and form a stable landscape (Oba et al. 2004). Another famous lahar, which occurred on the Nevado de Ruiz in Columbia in 1985, was initiated by medium volume hot pyroclastic flows that interacted with the ice- and snow-cap (Kohlbeck et al. 1994, Pierson and Janda 1994). Initial melt water mixed with the ejected volcanic debris, and produced a small volume lahar that evolved into a series of devastating, widespread lahar waves (0.1 km$^3$), killing c. 23,000 people. The lahars reached zones located 100 km from the source, and modified significantly the drainage network of the area. Nevado del Ruiz-type lahars are commonly triggered by pyroclastic flow-generating eruptions. The resulting mass flows cover the lower flanks of the volcano and the surrounding ring plain, where a complex network of intercalated and inter-fingering primary and secondary volcaniclastic sequences form a fan-like sedimentary cover (Table II, 2). This succession commonly referred as block-and-ash fan, has been described initially at the Pico de Orizaba, Mexico (Siebe et al. 1993).

Laharic diamictons in ancient volcanic settings are common. Many of them correspond to very large volume of deposits produced by a single mass wasting event such as the Osceola Mudflow from Mount Rainer, in the Cascades Range, Washington State, USA (Vallance and Scott 1997). The giant, clay-rich lahar initiated catastrophically from a fluid-saturated debris avalanche, deposited 3.8 km$^3$ of volcaniclastic sediment over a distance >120 kilometres (Vallance and Scott 1997). Similar volume of lahar deposits are also known in South America, for instance around Mt Cotopaxi in Equador (Aguilera et al. 2004). The Chillos Valley lahar has also been triggered by pyroclastic flow-generating eruptions, and reached a distance of 300 km from their source. It filled radiating valleys with sediment accumulations up to 200 meter deep (Mothes et al. 1998).

The preservation potential of lahar deposits –especially for debris flows- is in general good. Lahar are quickly channelled in valleys before reaching the ring plain, and their resulting deposits may contain interbedded layers hardened by dewatering processes, and also, many large lithic blocks that reinforce their broad compaction. Therefore, confined lahars tend to "fossilize" palaeo-valley networks around a volcano.

### From volcanic debris flows to hyperconcentrated (flood) streamflows

Lahar, as a volcaniclastic mass flow initiated on the flank a volcano, can evolve quickly during its course downstream a river channel. As a result, very different types of deposits can form, showing a broad horizontal facies variations along the stream valleys (Lavigne and Suwa 2004). If lahars are initiated on a volcanic terrain covered by abundant, loose, generally fine-grained sediment (e.g. ash), they can quickly transform from hyperconcentrated streamflows (HcFs) to debris flows (DFs), depending on the amount and size of particles the lahar current may pick up en route (Pierson and Scott 1985, Smith and Lowe 1991, Cronin et al. 1997a, Vallance 2000). Further downstream, the lahar may become more diluted, as a water-rich hyperconcentrated flow that could quickly change into normal flood, generating sediment-laden streamflow deposits (Smith and Lowe 1991, Vallance 2000). Volcanic debris avalanches initiated by sudden collapse of a volcanic edifice tend also to transform rapidly into debris flows, then into hyperconcentrated flows and normal floods downstream, as the mass flow hits water stored in lower inundation zones (Pierson and Scott 1985, Smith and Lowe 1991, Cronin et al. 2000). During lahar emplacement, significant erosion can affect the bed rock units, especially when the lahar wave moves over erodable terrains (e.g. loose tephra cover) (Smith and Lowe 1991). Undercutting of channel walls may also cause significant erosion and incorporation of sediment into the active lahar current (Smith and Lowe 1991). This erosional effect can be significant during both debris flow and hyperconcentrated flow phases of the lahar current. As previously mentioned, the lahar current may incorporate downstream large volumes of sediments (bulking), a process that is responsible for the transition from normal flood flow events to hyperconcentrated flows and debris flows (Smith and Lowe 1991). Such mechanism is observed specifically during the waxing and waning phases of the flow. Sediments picked up during the passage of the lahar waves tend to be segregated into distinct beds rich in exotic lithologies. Large volume debris flows can be loaded with volcanic and non-volcanic lithic fragments, colluvium, tree debris as well as glacial drifts. The waxing phase of such lahar currents can be very erosive (Smith and Lowe 1991). The fine sediment-rich tail of the lahar is generally less erosive that the front wave, facilitating the deposition of sediments (Smith and Lowe 1991). The waning stage of the lahar current is in general water-rich, of smaller amplitude than the frontal head, but still rather erosive. At this late stage of evolution, it incises into the previously deposited deposits and can create complex sedimentary features in cross sectional view through the stream channels (Smith and Lowe 1991).

Lahars go through significant particle segregation processes (Figure 7.4) due to density and particle concentration variations throughout the entire current (Pierson and Scott 1985, Smith and Lowe 1991). Light, pumiceous clasts quickly reach the top of the lahar current. In a similar way, low density solids also tend to pop up on the surface of the current due to pore water escape in the body of the flow. Such low density particles commonly form rafts on the surface of the current, and can travel together over long distances (Smith and Lowe 1991). Due to the viscous nature of the lahar current, its bottom part moves more slowly then its top part. This differentiation leads to the development of typical vertical flow profiles (for velocity and particle concentration) (Smith and Lowe 1991). The low density particles therefore migrate upward, then to the front of the lahar wave. Conversely, high density particles tend to settle quicker. This particle segregation process results in a typical normal grading texture for the lahar deposits. However, kinetic sieving also occurs in a vibrating or sheared grain flow, as a result of both percolation and fluid expulsion. This process, induces downward movements of small particles, progressively replacing larger clasts (Smith and Lowe 1991). As a consequence, large particles tend to reach the surface of the flow, and drift progressively towards its lateral margins, depositing lenses of well sorted sediments.

During its migration downstream a river catchment, a lahar will gradually incorporate more water from the stream channel in amounts large enough to change the physical properties of the flow and drop its capacity to carry sediments (Pierson and Scott 1985, Thouret et al. 1998, Cronin et al. 1999, Lavigne et al. 2000, Lavigne and Suwa 2004). This process is important

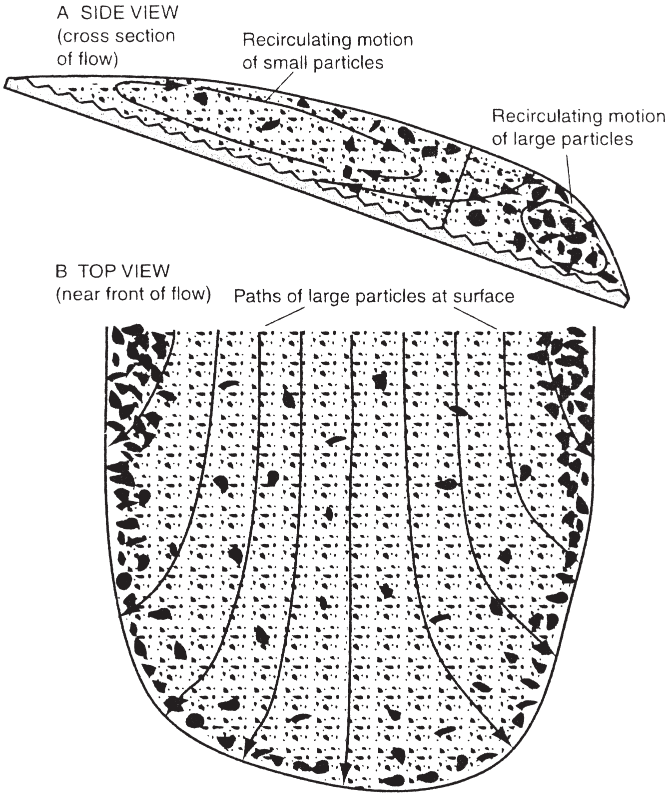


***Figure 7.4.*** Particle-size segregation through a laharic current after Vallace 2000: p. 607, fig. 3

only for those lahars that are small enough in comparison to the water course they enter. The rheology and physical behaviour of large volume lahars are generally not affected significantly by this mechanism (Lavigne and Thouret 2000). Small volume clay-poor lahars (i.e. "non-cohesive" lahars; Scott 1988, Scott et al. 1995) tend to have an open framework, sandy matrix, and can be effected by dilution from water (Mothes et al. 1998). By contrast, clay-rich lahars (or "cohesive" lahars) are harder to dilute due to the buffering effect of clay particles in the matrix. Lahars that reach the ring plain and follow major tributaries push the stream water ahead of the front wave, and then gradually mix. Water ingestion into the lahar body leads to a drop in its capacity to carry large clasts, a process that leads to the formation of lag-like lenses of cobbles and boulders.

The textural characteristics of the resulting deposits depend on the type of lahar flow (Figure 7.5). Debris flows are generally poorly sorted, massive, and non-stratified (Table II, 3). By contrast, hyperconcentrated stream flow deposits are better sorted, and show a faint internal stratification (Table II, 4). Textures of hyperconcentrated streamflow deposits may also reveal clast alignments with compositional variations among clast groups, leading to a graded vertical distribution of coarse particles. Some of these clast groups show evidences of internal abrasion (cataclasis), and tend to be associated with more stratified units.

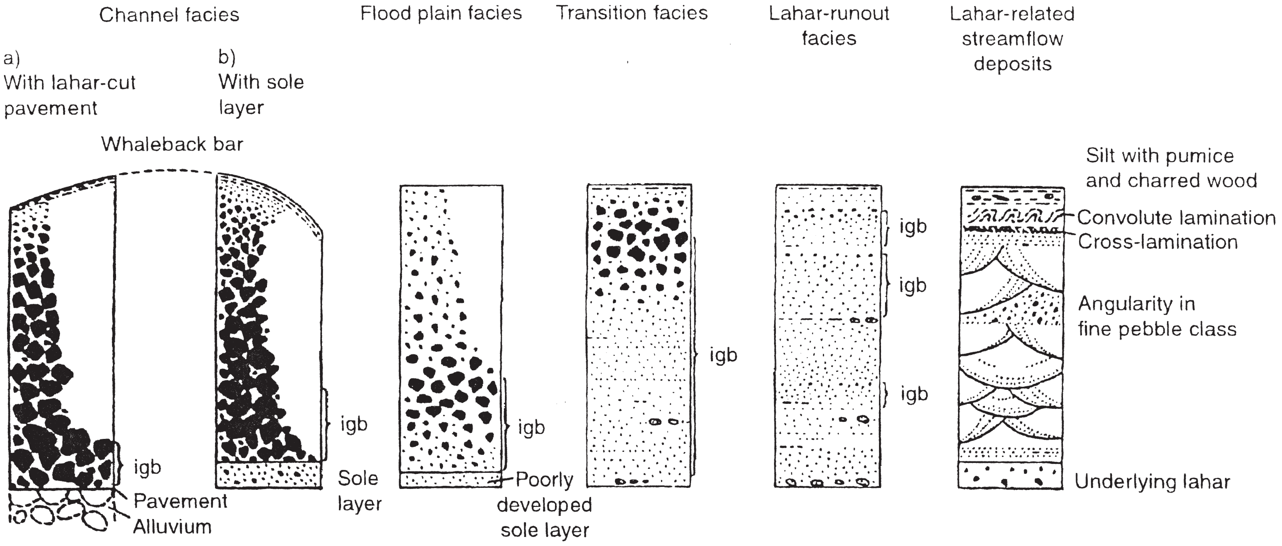


*Figure 7.5.* Facies types of laharic deposits after Vallace 2000: p. 611, fig. 7

Distinguishing lahar-originated deposits (Table II, 5 and 6) from those non-welded fine grained and matrix-rich pyroclastic flow deposits is sometimes difficult (Brantley and Waitt 1988, Waythomas 1999). However, detailed studies of lateral facies variations can be helpful in order to distinguish the textural characteristics of non-welded pyroclastic flow deposits (juvenile particle content, vesicle-free, and non-indurated matrix) from those of lahar deposits. Since pyroclastic flows are hot during emplacement, their deposits commonly contain charcoals as a result of burnt vegetation (Scott and Glasspool 2005), but a distinctively rare occurrence in lahar deposits (Table III, 1). However, pyroclastic flows can also hit major water courses, snow-and-ice caps, and be partially diluted by heavy rain, thus initiating hot lahars (Table III, 2 and 3). When lahars are generated in a valley previously filled by a pyroclastic flow, the mass flow will remobilise the pumiceous material that will be re-deposited in more distal areas of the ring plain (Table III, 4). In this case, lahar deposits can incorporate charcoals that have been originally contained in a pyroclastic flow deposit channelled further upstream in the valley (Table III, 5). Pyroclastic flow deposits are usually hot enough to develop magnetically oriented clasts, and have a distinct magnetic fabric (Porreca et al. 2003, Saito et al. 2003, Tanaka et al. 2004). Lahar deposits usually have no such magnetic fabric due to their low transportation and deposition temperatures (Table III, 6). Pyroclastic flow deposits have usually a loose texture, contain less mud in their matrix, and are less compact than lahar deposits.

When a pyroclastic flow travels through an ice cap, it induces melt water and generates large volume lahars. In this type of scenario, the transition between pyroclastic flows and pumice-rich lahars ) can be gradual along the transportation axis (Crandell 1987, Brantley and Waitt 1988, Scott 1988, Siebe et al. 1993). In the similar way, transition between volcanic debris avalanche deposits and clay-rich lahars is more or less gradual and complex (Calvari et al. 1998, Nehlig et al. 2001, Capra et al. 2002). Typical volcanic debris avalanche deposits are larger in volume, contain hummocks as the result of megaclasts transportation, and display a more chaotic internal architecture than lahar deposits. Till deposits may also look very similar to lahar deposits. However, tills generally do not incorporate vegetation fragments and do not expose lateral facies variations typical of lahar deposits. Finally, landslide deposits (that are sometimes texturally similar to lahar deposits are generally very localised, and can be often associated to scars or other morphological features (headwalls, ridges, etc) pointing towards the source of the unconsolidated material. These small scale, landslide-associated diamictons can be easily distinguished in the field from lahar deposits by detailed surveying, careful mapping supported by aerial photo coverage if necessary.

Lahars are among the most dangerous volcanic hazards. Lahars commonly reach unexpectedly inhabited areas destroying villages and cities. Ironically in many places worldwide lahar inundated areas re-inhabited, even larger scale than before the volcanic disaster stroke (Table IV, 1). Reduction of volcanic hazards by lahars commonly facilitated by huge effort to construct dams that may break the mechanic energy of the current, and hold back larger clast to travel long distances (Table IV, 2).

### Volcanic debris avalanches

Volcanic debris avalanches are products of major sector collapses affecting a volcano under water (or fluids-)-saturated conditions (Siebert 1984, Glicken 1991, Smith and Lowe 1991, Ui et al. 2000). Debris avalanches are very rapid, inertial, granular flows, generally resulting from giant landslides occurring on an unstable portion of a mature volcanic edifice. Sector (or flank) collapses results in horse-shoe shaped amphitheatres from where unstable lava and brecciated pyroclastics are being removed and transported as a chaotic avalanche (Table IV, 3 and 4). Destabilization of the volcano's superstructure by a shallow intrusion of magma high into the edifice is a process that has been linked to some of these collapses (e.g., Mount St. Helens in 1981) (Table IV, 5), while other debris avalanches have occurred without any production of juvenile magmatic materials (e.g., Bandai in 1888). Historic debris avalanches not linked to a magmatic eruption are commonly associated with enhanced fluid circulations and pressurization in shallow hydrothermal systems at the time of the collapse (Reid 2004). Three main types of volcanic debris avalanches are recognized in the literature (Ui et al. 2000), each of them refering to a different eruption style. Bezymianny-type debris avalanches refers to the blast-induced flank collapse of Mt Bezymianny in 1956 (Kamchatka, Russia; Belousov 1996). This type of volcanic debris avalanches is triggered by a very powerful magmatic eruption. Bandai-type volcanic debris avalanches refer to the mass flow generated by a major phreatic eruption that decapitated Bandai volcano in Japan in 1888 (Yamamoto et al. 1999). In this later case, no juvenile material was found in the resulting volcanic diamicton. Finally, Unzen-type volcanic debris avalanches refer to the mechanical destabilization of an old parasitic dome situated on a steep flank of the erupting Mt Unzen in 1792 in Japan (Ui et al. 2000). This type of catastrophic collapse is related to earthquakes rather than volcanic activity.

Deposits from volcanic debris avalanches comprise two major facies types (Ui et al. 2000): a block (or megaclasts-rich) facies, and a matrix facies (Figure 7.6). A debris avalanche block (or megaclast) is a fractured (and sometimes deformed), commonly overturned piece of the source volcano. Such blocks can reach tens to hundreds of metres in diameter, and may form an obscure stratigraphical section appearing in an exotic place (Cacho et al. 1994, Reubi and Hernandez 2000, Collot et al. 2001). Usually these blocks show many evidences of mechanical stress, and internal fracturing can be very significant. Along the milled and strongly fragmented clasts, jig-saw fit textures are common (Figure 7.7). The matrix of debris avalanches is generally chaotic in texture (Table IV, 6), finely grained, and can host various lithologies representative of the sampled parts of the source volcano (Reubi and Hernandez 2000). Areas affected by volcanic debris avalanches commonly show a typical hummocky morphology (Table V, 1 and 2) resulting from the distal transportation of fragmented megaclasts

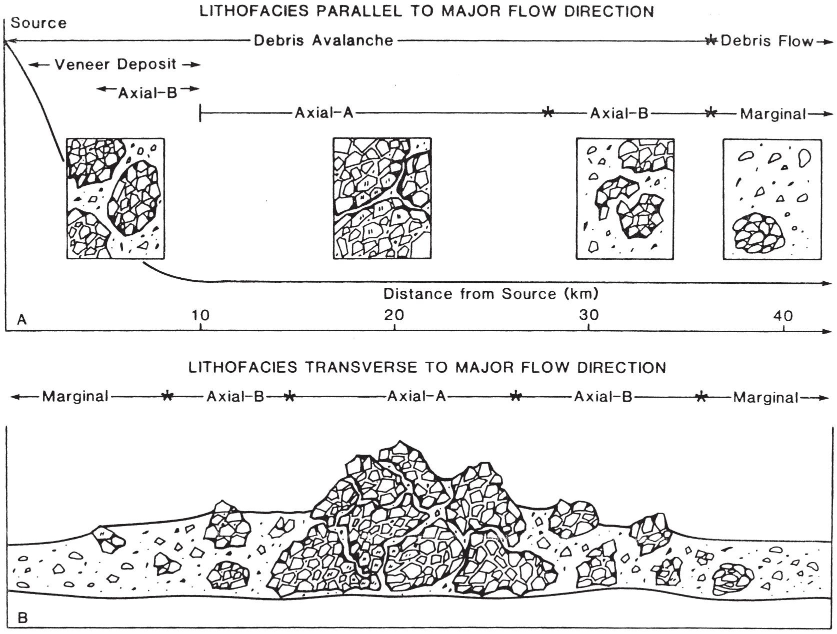


***Figure 7.6.*** Lithofacies types and changes recognized in the volcanic debris avalanche deposits of the Taranaki volcano, New Zealand (Palmer et al. 1991: p. 95, fig. 7). On "A" facies variations from source to distal areas demonstrated. On "B" a theoretical cross section of a volcanic debris avalanche deposit is shown

*Figure 7.7.* Jig-saw fit texture of large volcanic lithic fragments of a volcanic debris avalanche deposit of the Ruapehu volcano, New Zealand

*Figure 7.8.* Milled clast between large volcanic lithic clasts of a volcanicdebris avalanche deposit from the Ruapehu volcano

(Ponomareva et al. 1998, Clavero et al. 2002, Le Friant et al. 2002). Hummocks commonly reach tens of metres across and no apparent distributional pattern has been identified yet (Siebe et al. 1992). One of the main characteristic of volcanic debris avalanche deposits is their common fracture pattern. Jig-saw fit cracks are common on larger, brittle clasts such as volcanic lithics (e.g. fragments from coherent lava flows and domes). Jig-saw cracks of an avalanche block are joint-like and irregular, and usually remain closed. As megaclasts are transported over long distances (i.e. >15 km), joints can open up and be filled progressively by finely grained matrix sediment coming from the milled part of the same fractured blocks (Figure 7.8).

Despite the abundant geological evidence showing that volcanic debris avalanches are a natural consequence of volcano growth and destruction, their actual significance has just been recently emphasized. Volcanologists worldwide started to focus their attention on volcano collapses and the generation of debris avalanches after the catastrophic and well documented collapse of Mt St. Helens in 1980 (Waitt et al. 1983, Donnadieu et al. 2001). Since then, volcanic debris avalanche deposits have been identified around many major volcanic complexes located in many different geological settings. The significance of volcanic debris avalanches in the evolution of a volcano therefore is increasingly recognized.

Volcanic collapses generating debris avalanches leave very typical morphological features in the landscape. The most recognisable features are: (1) a large scar — or amphitheatre — located at the source of the initial avalanche on the upper part of the volcanic edifice, often partially filled by subsequent lava domes, and (2) a widespread hummocky surface in the medial or distal reaches of the deposit. The hummocky region include blocks (megaclasts) of various sizes that are usually larger and more closely spaced in the proximal area of deposition. Recognition of hummocky surfaces however is not a sufficient criteria to confirm the volcanic debris avalanche origin of a diamicton in the field. Cross-sectional views of individual hummocks are usually required in order to identify fractured blocks, exotic volcanic stratigraphy and lithologies (Table V, 1). The recognition of a jig-saw fit texture affecting large coherent lava bodies embedded in altered matrix is necessary to infer the hummocky origin of small hills. Natural levees exceeding a few tens of metres in elevation are commonly reported in volcanic debris avalanche deposits such as the Socompa volcano in northern Chile (Wadge et al. 1995). Such cliffs are usually well developed in distal areas where the debris avalanche reaches flat lying areas of the surrounding ring plain (Siebe et al. 1992).

Slided and tilted blocks (or megaclasts) in debris avalanche deposits are more or less intact parts of the original volcanic edifice. Despite the fact that such blocks preserve the basic texture and lithology of the original portion of the volcanic cone affected by destabilization, they are often truncated and deformed, especially in distal regions, reflecting the impact of transportation over long distances. Very large blocks can significantly erode the substratum during the emplacement of the mass flow, and they are frequently mantled by fractured and crushed clast fragments where hummocks are formed. The magnetic fabric of a single avalanche block (megaclast) is usually fairly uniform. However, the magnetic fabric orientation can be very variable block by block, indicating uniform, en masse displacement of the source rock units into homogeneous megaclasts. The avalanche matrix is a mixture of finer grained sediment, derived from source rock lithologies. The magnetic fabric of the matrix is very randomly oriented, as a reflection of the well fragmented and diversified source rock. Debris avalanche matrix commonly contains colluvial and/or fluvial fragments picked up during travelling. Such fragments can be deformed and squeezed, forming flame-like structures injected between larger clasts. This is a common occurrence with water-saturated clasts of sedimentary rocks. The base of debris avalanche deposits is commonly fine grained, and has many textural features indicating high shear stress on the sole of the flow (Table V, 3). Soft sediment deformation, unidirectional flame structures, and other dewatering structures reflect loading and shearing in the base of the flow (Table V, 3).

An increasing number of large volume volcanic debris avalanche deposits have been identified recently in many different volcanic settings. The largest known volcanic terrestrial debris avalanche deposit has been identified around Mt Shasta in the Cascades (USA), where the deposit reaches at least a run-out distance of 45 km (UI and GLICKEN 1986). Volcanic debris avalanches have been extensively studied in volcanic arc settings, such as Unzen, Hokkaido-Komagatake (both in Japan), or at many Chilean or Mexican volcanoes. As previously indicated, volcanic debris avalanches can transform into clay-rich ("cohesive") lahars, as it has been originally demonstrated for the Osceola Mudflow at Mount Rainer (US) (VALLANCE and SCOTT 1997). After travelling a distance of 2 km, the initial avalanche hit a major river valley and travelled an additional 120 km as a cohesive lahar (VALLANCE and SCOTT 1997). Therefore, volcanic debris avalanches can affect very large areas, and can drastically modify the landscape. Despite numerous studies reporting the occurrence of volcanic debris avalanches deposits worldwide, only the Mt St Helens blast-triggered collapse was directly observed and well documented in recent historic times. The geological record of Quaternary volcanoes in Japan indicates that volcanic debris avalanches occurred at least once in every hundred years (UI et al. 2000).

In ancient terrains, identification of volcanic debris avalanches is more difficult to achieve. Volcanic debris avalanches can displace entire rock units from the source volcano as megaclasts measuring hundreds of metre across. Such very large blocks can be preserved and mimic original "in situ" volcanic successions. In this context, differentiation between "in situ" and allochton units is hindered by superimposed tectonic and metamorphic processes. The high mobility of these fluid-saturated mass flows may also result in the displacement of large sections of a cone tens of kilometres away from its source. In such a situation, stratigraphic and lithological correlations of avalanche block units with source rock units are difficult to establish. Detailed mapping is essential to delineate the avalanche boundaries, and their exact position in an exotic rock environment.

Volcanic cone collapses and associated large volume debris avalanches have been well described in ancient volcanic settings such as the Cantal massif in central France (CANTAGREL 1995, REUBI and HERNANDEZ 2000, NEHLIG et al. 2001). Some large scale cone collapses are also related to major tectonic activity, such as movement of strike-slip faults beneath a volcano (LAGMAY et al. 2000). CECCHI et al. (2004) and ACOCELLA (2005) inferred that volcano spreading and associated collapses (CECCHI et al. 2004, ACOCELLA 2005) can be enhanced by favourable substrate conditions beneath an oversteepened cone, usually dominated by the accumulation of lava domes. Many volcanoes in Northern Chile, where volcanic edifices grown over salar deposits, illustrate this genetic relationship (Table V, 3, 4 and 5).

Amphitheatres open at the head of volcanic valleys such as the Valle del Bove on the eastern, sea-facing flank of Mt Etna, are thought to have been shaped by catastrophic collapses (CALVARI et al. 1998, PARESCHI et al. 2006). Large scale avalanches are generated offshore by these giant landslides, as a result of unsupported parts of the growing volcanic edifice collapsing into the sea (CLOUARD et al. 2001, CLEMENT et al. 2003, MILIA et al. 2003, HURLIMANN et al. 2004, HILDENBRAND et al. 2006). Major volcano collapses have been similarly suggested to explain the formation of large half depression features found around volcanic islands such as Tenerife (Table V, 6) (HURLIMANN et al. 1999), and Gran Canaria in Spain (MEHL and SCHMINCKE 1999). However the general lack of associated volcanic debris avalanche deposits may complicate the interpretation of these depressions. Around volcanic islands however, bathymetric surveys often reveal the existence of an irregular seafloor morphology that could be interpreted as the result of submarine landslides (KRASTEL and SCHMINCKE 2002, MASSON et al. 2002) and/or volcanic debris avalanches feeding submarine turbidity currents in their distal portion (SCHNEIDER et al. 2004). Such deposits have also been identified around most of the Hawaiian volcanic islands and around Reunion, in the Indian Ocean (OEHLER et al. 2004). The collapse of island volcanic edifices usually leaves a subaerial to submarine horseshoe-shaped scar. On the island of Ischia for instance, the southern flank of the volcano was affected by a large scale collapse that generated a debris avalanche incorporating thousands of giant blocks, dispersed as far as 50 km offshore on the sea floor (CHIOCCI and DE ALTERIIS 2006). The transition of volcanic debris avalanche deposits into submarine mass flow deposits has been well documented for many volcaniclastic sequences, such as around Gran Canaria, Spain (SCHNEIDER et al. 2004). However, the exact interpretation of the succession of volcano-sedimentary events (e.g. primary debris avalanche deposits or secondary origin via reworking and redistribution) are in many cases not straight forward.

In the Carpathian arc in Cenrtal Europe, volcanic debris avalanche deposits have recently been recognized. Large horseshoe-shaped morphological features are found in many Miocene to Pliocene erosional remnants of stratovolcanoes. Recent detailed studies suggest these residual morphologies are the result of major ancient volcanic collapses (KARÁTSON 1999, KARÁTSON et al. 1999). Since the initial recognition of the large collapsed crater zones, associated volcanic debris avalanche deposits have been identified. The authors suggest that an exhumed volcanic amphitheatre and the related volcanic debris avalanche deposits may have been a cause for a major curvature on the Danube river channel (Table VI, 1), called Danube Bend (KARÁTSON et al. 2006). In addition, volcanic debris avalanche deposits in the Miocene Inner-Carpathian Volcanic Chain have been also identified in the Visegrád (Table VI, 2), Börzsöny and Mátra Mountains so far (Karátson et al. 2001). Volcano collapse and associated volcanic debris avalanche deposits have been also identified recently in the Outer Carpathians and their potential role to host mineral deposits has been emphasized (Lexa et al. 1999, Szakács and Seghedi 2000, Szakács and Krézsek 2006).

### Relationships between lahars and volcanic debris avalanches

The idea that volcanic debris avalanches transform into lahars is now commonly accepted (Lecointre et al. 2002). When a volcanic edifice collapses, the resulting mass flow is often partially channelled by a major river system, and the avalanche becomes progressively more diluted downstream (Carrasco-Nunez et al. 1993). This is especially the case for "dry" debris avalanches that have the capacity to incorporate large amounts of water during their emplacement in a river catchment. The flow can transform into various type of lahars s (Palmer and Neall 1989, Carrasco-Nunez et al. 1993, Encinas et al. 2006). Such mass flows usually contain megaclasts, representing rock types that are found in the upper volcanic edifice such as blocky fragments of lava domes. Deposits resulting from this type of lahar could be very similar to those left by lahars initiated by volcanic eruptions or meteoric events, especially in the distal zone of the ring plain. In many cases, volcanic debris avalanches and lahar currents are closely related. Debris avalanches form an end member in the spectrum of granular mass flows characterized by a very high sediment-to-water ratio in the current (Figure 7.3) (Smith and Lowe 1991). Under-saturated debris avalanches are able to transform, with increasing water content along a river channel, into debris flows, hyperconcentrated flows and finally, into dilute stream flows. In "dry" debris avalanches (i.e. clay-poor), the main support mechanism for clasts and particles is the grain dispersive force, which is gradually decreasing in strength once transformation into debris flows and hyperconcentrated flows occur. At the end of the spectrum, turbulence dominates in dilute stream flows (Figure 7.3). However, matrix cohesion of the initial mass flow may prevent this evolution. Grain support mechanisms in debris avalanches and debris flows are very strongly related, to the clay content of matrix in the current. Typically, debris avalanches sourced from hydrothermally altered portions of a volcanic cone are fluid-saturated (very high pore pressure), and will transform rapidly into "cohesive" debris flows that will not dilute further downstream (Lecointre et al. 2002). Fluid buoyancy also plays a significant role in particle support for grain size particles, especially in hyperconcentrated flows. This process is less efficient in debris flows, and has now role in debris avalanches, as the relative coarse sediment to water ratio increases. The deposition mechanism of debris avalanches is predominantly en masse freezing, similar to cohesive debris flows. This differs notably from hyperconcentrated flow and dilute stream flow where grain-by-grain settling tend to dominate the deposition process.

### Lahars and fluvio-lacustrine depositional systems

Lahars and catastrophic flood events can be triggered by major plinian eruptions (Manville et al. 2005). Large volume pumiceous deposits can block major water courses and may create temporary barriers such as tephra dams, resulting into the development of extensive lakes (Macias et al. 2004). When the dams break, major volcaniclastic floods are initiated, and the generated lahars may cause significant changes on the landscape and devastation well after the volcanic eruption (Manville et al. 1999, Waythomas 2001). Major pyroclastic events, such as the 12,900 yr BP Laacher Sea eruption in Germany, can modify the drainage system of an area, cause damming of major tributaries, and later trigger major floods once temporary water storages fail (Schmincke et al. 1999). Similar reshaping of a large hydrographic network has been well documented for the 1.8 ka pumice-rich Taupo eruption (Segschneider et al. 2002a), where major re-sedimentation processes were particularly active in distant river catchments (Segschneider et al. 2002b).

Lahars can leave along their route large volume of water-saturated sedimentary valley fills and overbank deposits. On the ring plain, such lahar deposits may create semi-permanent lakes (years to decades) that are gradually filled by fine grained sediment eroded from the barren surface. The gradual reestablishment of the original fluvial network progressively carve valleys into the lahar deposits, generating complex facies associations. In ancient volcanic settings, such fluvio-volcaniclastic successions can be very difficult to decipher. On vertical sections, stratigraphic sequences showing sudden facies variations between debris flow, hyperconcentrated flow and normal stream flow deposits can be intercalated with normal lacustrine beds (Figure 7.9). In horizontal view, lacustrine beds may be found as incorporated lenses into more massive debris flow diamictons-, alternating with sandy hyperconcentrated flow deposits. Because of intense erosion along the valley margins, diamictons resulting from late lahar activity and associated fluvial sequences may display sharp and steep erosional contacts with earlier debris flow and hyperconcentrated flow deposits. Lahars can also be generated by relatively small-volume pyroclastic eruptions, such as phreatomagmatic explosions that trigger ground-hugging density currents. Such pyroclastic flows may enter a maar lake and transform rapidly into a mudflow, a process described for the Peperino Albano phreatomagmatic eruption from Colli Albani volcano, located 30 km to the south-east of Rome, Italy (Giordano et al. 2002). Large scale, basin-wide volcaniclastic successions associated with Large Igneous

Provinces have been previously interpreted to be made of lahar deposits resulting from major destructional phases of quickly grown volcanoes such as the volcanic systems of Middle Jurassic Prebble and Mawson Formations, Antarctica (Elliot 2000, Elliot and Hanson 2001). These deposits have been however recently re-evaluated, and re-interpreted as a direct result of phreatomagmatic explosive events, and being rather primary in their origin (White and McClintock 2001, Ross et al. 2005, Ross and White 2005)

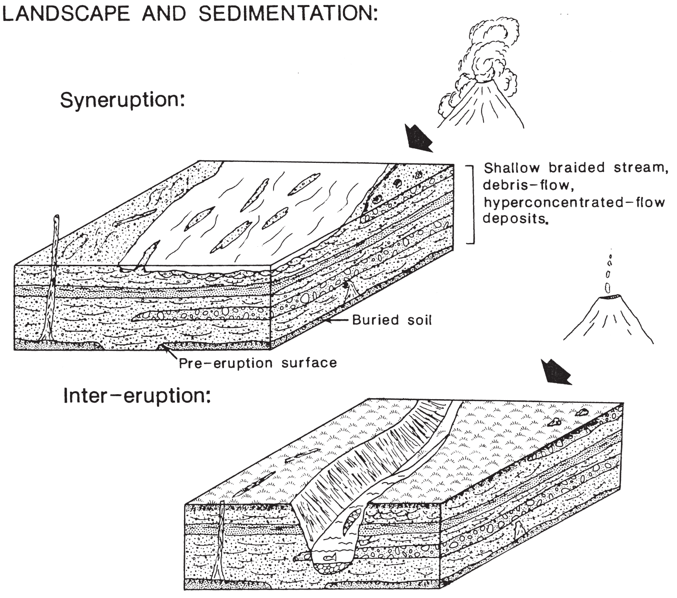


*Figure 7.9.* Diagram of syn and inter-eruption sedimentary facies may develop around a long lived strato-volcano after Smith 1991: p. 113, fig. 3

### Erosion of volcanic landforms

Erosion of volcanic landforms has been studied with many different approaches. Sediment loss and surface modification that lead to the development of an erosional volcanic landforms can be a very efficient process. The quantification of the erosion rate is however quite complex. Monogenetic volcanoes are an appropriate target when the objective is to quantify the erosion of small volume volcanic landforms. The process of erosion on scoria cones has been well documented (Table VI, 3), and emphasizes a specific relationship between the height of the edifice, the opening of the crater, and slope angle changes (Wood 1980b, 1980a). Such morphometric studies usually associate specific residual shapes of scoria cones (Figure 7.10) with a peculiar stage of evolution (Porter 1972, Wood 1980b, Dohrenwend et al. 1986, Hooper and Sheridan 1998). Various authors suggested that scoria cones can be almost totally flattened after a million year of erosion history, low in slope angle and their crater more or less unrecognizable (Wood 1980b, Hooper and Sheridan 1998, Vespermann and Schmincke 2000). Such studies however have been derived from volcanic fields that evolved in arid climate, and where scoria cones are relatively simple in morphology, as well as uniformly composed of coarse-fine lapilli and ash beds (Dohrenwend et al. 1986, Siebe 1986, Inbar et al. 1994, Heizler et al. 1999, Inbar and Risso 2001). More recent studies have demonstrated in addition that erosion stages for such volcanic landforms can be very much predetermined by their original morphology. In some cases, that original morphology may have been hindered by eruptive activity (e.g. phreatomagmatism) that led to the formation of a wide and steep crater in the cone (Németh et al. 2005, Martin and Németh 2006). The transition between cinder cone and a small volume composite volcano is very often gradual, and many erosional models cannot be directly applied in this situation (McKnight and Williams 1997). Also, many scoria cones go through an eruptive phase dominated by vigorous lava fountaining that produce thick welded tephra units (Németh 2004). Such units can also react to erosion as hard layers, and modify the expected erosional path of the corresponding volcanic landform (Table VI, 4). In this scenario, erosion may be driven by undercutting and small collapses on the outer crater rim, rather than chemical dissolution rain-generated debris flows, and/or eolian abrasion (Table VI, 5) (Németh et al. 2005). The situation is more complex for those small volume volcanoes, where the eruptive history included a significant phreatomagmatic phase (Németh 2004). Erosion of maars, tephra rings and tuff cones has not been studied in detail yet, and they seem to follow different rules to those applying to scoria cones. Tephra generated by phreatomagmatic activity can form a "concrete-like" deposit due to a sudden loss of water, and behave as hard layers, resistant to erosion (White 1991a). In such beds, erosion and especially the development of gullies (Table VI, 6) play an important role in the redistribution of volcanic material (White 1991a). Phreatomagmatic activity produces also large volumes of water-saturated tephra. Those tephras can be easily remobilized during the eruption, forming major debris flows that carve a gully network on the flanks of the volcano (Németh and Cronin 2006). As previously mentioned, debris flows tend to transform into hyperconcentrated flows in distal areas, where accumulations of tabular debris flows and hyperconcentrated flows units are intercalated with thin base surge and phreatomagmatic tephra beds (Sohn 1996, Sohn and Park 2005, Németh and Cronin 2006). Thick accumulations of syn-volcanic, reworked tephra units can form a collar of flat lying volcaniclastic succession around a phreatomagmatic volcano (Sohn and Chough 1989, Sohn 1996, Sohn et al. 2003, Sohn and Park 2005). Due to the low slope angle and the broad opening of the vent, reworked volcaniclastics accumulate quickly in their crater (Table VI, 7) (White 1991b). Under favourable conditions, crater lakes can be sedimentary traps of volcanic and non-volcanic debris. In extreme cases, maar-type volcanoes can have a crater hundreds of metres deep that can act as an active sedimentary basin over tens of thousands of years (White 1992). During the initial erosional stage, entire blocks from the maar or tuff ring can collapse into the lake and initiate subaqueous

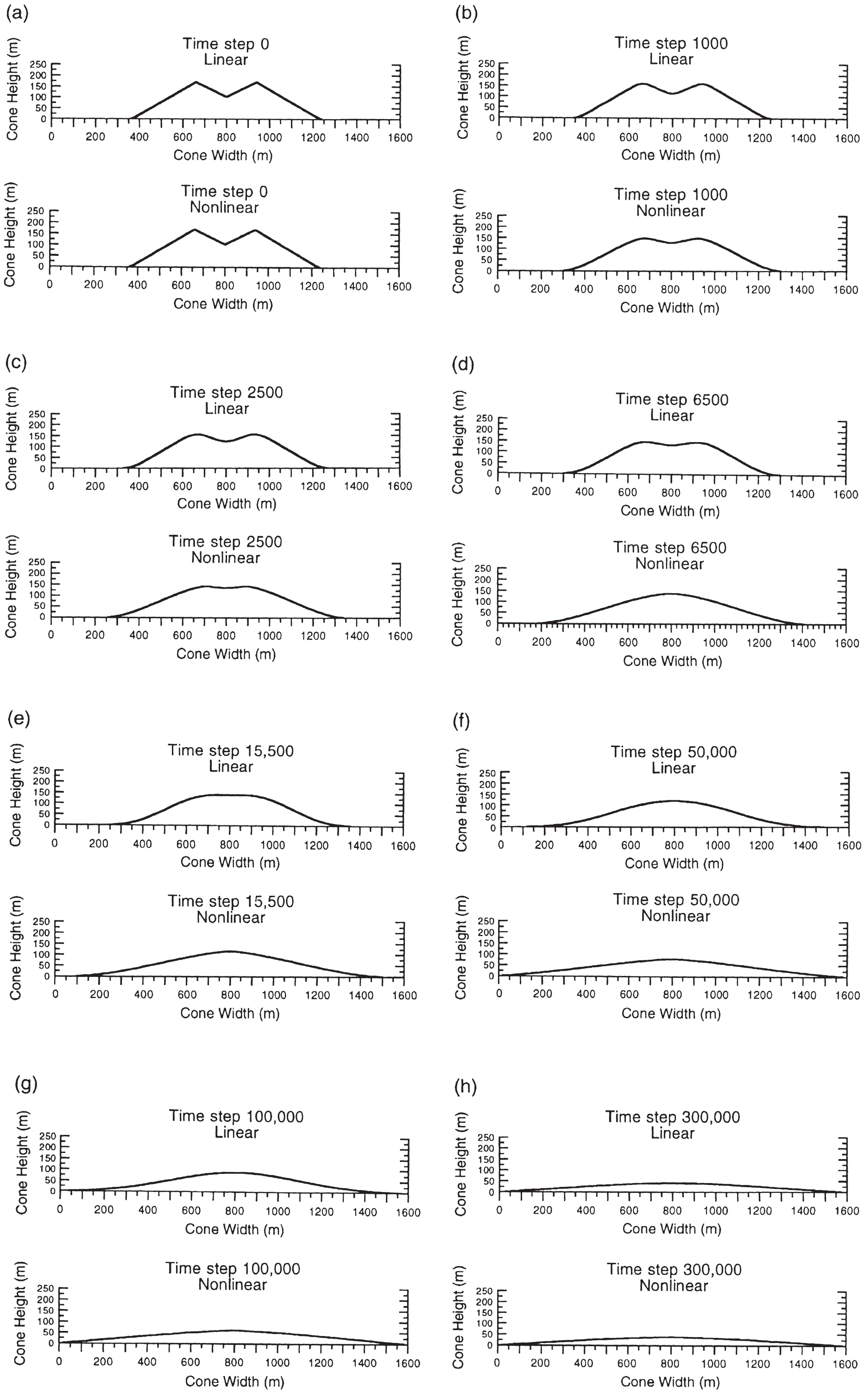


***Figure 7.10.*** Stages of erosion (from "a" to "h") of scoria cones modelled by Hooper and Sheridan (1998)

debris flows and turbidity currents (White 1992, Büchel and Lorenz 1993, Fisher et al. 2000). The tephra rim in many cases can be quickly eroded (thousands of years), and the maar lake acts as a sheltered, deep sediment trap (Drohmann and Negendank 1993, Mingram 1998, Zolitschka et al. 2000). Studies of lithofacies associations from eroded phreatomagmatic volcanic fields can help to reconstruct the erosion history of the landscape where the volcanoes erupted (Németh 2001). This long-term erosion rate studies suggest that lowering of the base level is in the order of a few tens of metres per million year for regions affected by a continental climatea result comparable with those previously published (Németh and Martin 1999, Németh 2001, 2003). However, (Németh et al. 2006b) recently proposed that erosion rate calculations should be done with great care, and that a correct lithofacies study of erosion remnants should be completed before attempting a detailed erosion calculation (Table VII, 1).

In composite volcanoes, erosion is far more complex than in small volume volcanic edifices (Wood 1978). Stratovolcanoes are generally long-lived, and extended periods of quiescence separate eruptive phases (Davidson and de Silva 2000). Periods of eruptive activity can also be very complex in terms of eruption style and intensity. Freshly produced tephra can be quickly remobilized into syn-eruptive lahars, commonly accompanied by primary pyroclastic density currents (Davidson and de Silva 2000). In inter-eruptive periods, and prior to any recolonization of the bare landscape by vegetation, rain-induced lahars can redistribute large volumes of loose volcanic sediment onto the ring plain. As a result, especially on stratovolcanoes with low frequency eruptions and long periods of quiescence, the continuously eroded edifice is progressively surrounded by a growing and thickening succession of volcaniclastic sediments (Lecointre et al. 1998, Davidson and de Silva 2000, Lecointre et al. 2004). Eruptions, coupled with intermittent erosional phases on a volcano that is active over hundreds of thousands of years, are able to generate a ring plain hundreds of metres thick, and mostly dominated by reworked volcaniclastic sediments (Table VII, 2). Numerous long-lived composite volcanoes are surrounded by such a volcanogenic sedimentary pile. The detailed study of the corresponding stratigraphies help to understand the nature of the depositional processes for each volcanic succession found in the ring plain, and may provide vital information about early evolution stages of the central cone. Very often, products on the cone have been already removed by erosion, and only preserved by redeposition in the lowlands of the ring plain. Erosion of composite cones has been studied by applying a range of complementary methods on many different volcanic settings. From the Carpathians, Miocene to Pliocene stratovolcanoes (Table VII, 3) have been selected to quantify the main factors controlling landform modification, and identify long term erosion rates (Karátson 1996, 1999). Erosion of these volcanoes resulted in a significant widening of their central craters, forming a talus bordering depressions similar to calderas (termed erosional calderas). Erosion of composite volcanoes can generate very complex structures, especially when flank or sector collapse on the volcano occurs in its history (Table VII, 4). Thouret (1999) demonstrated that on many lava dome-dominated arc volcanoes, the landscape evolution is predominantly controlled by pulsating growth and successive collapses (Thouret 1999). Such volcanoes can develop largely as an asymmetric cone with truncated flanks (Table VII, 5). In this context, flank and ring plain sections show widespread hummocky surfaces determined by the distribution of major volcanic debris avalanches, such as the Calbuco volcano in southern Chile.

Erosion of volcanic landforms can produce spectacular micro, meso- and macro-features. Strong wind erosion can form micro yardangs in basaltic rock surfaces (Table VII, 6), as well as mega-yardangs on ignimbrites (Table VIII, 1 and 2). Wheatering processes of basaltic lava flows in arid climate can produce fractured lava surface morphologies (Table VIII, 3, 4 and 5) as well as thick varnish cover over individual clasts. Redistribution of tephra in every volcanic region could be significant especially in high altitude, high wind zones, such as the High Andes (Table VIII, 6). Significant volume of tephra accumulates every year in areas very distal from the tephra sources.

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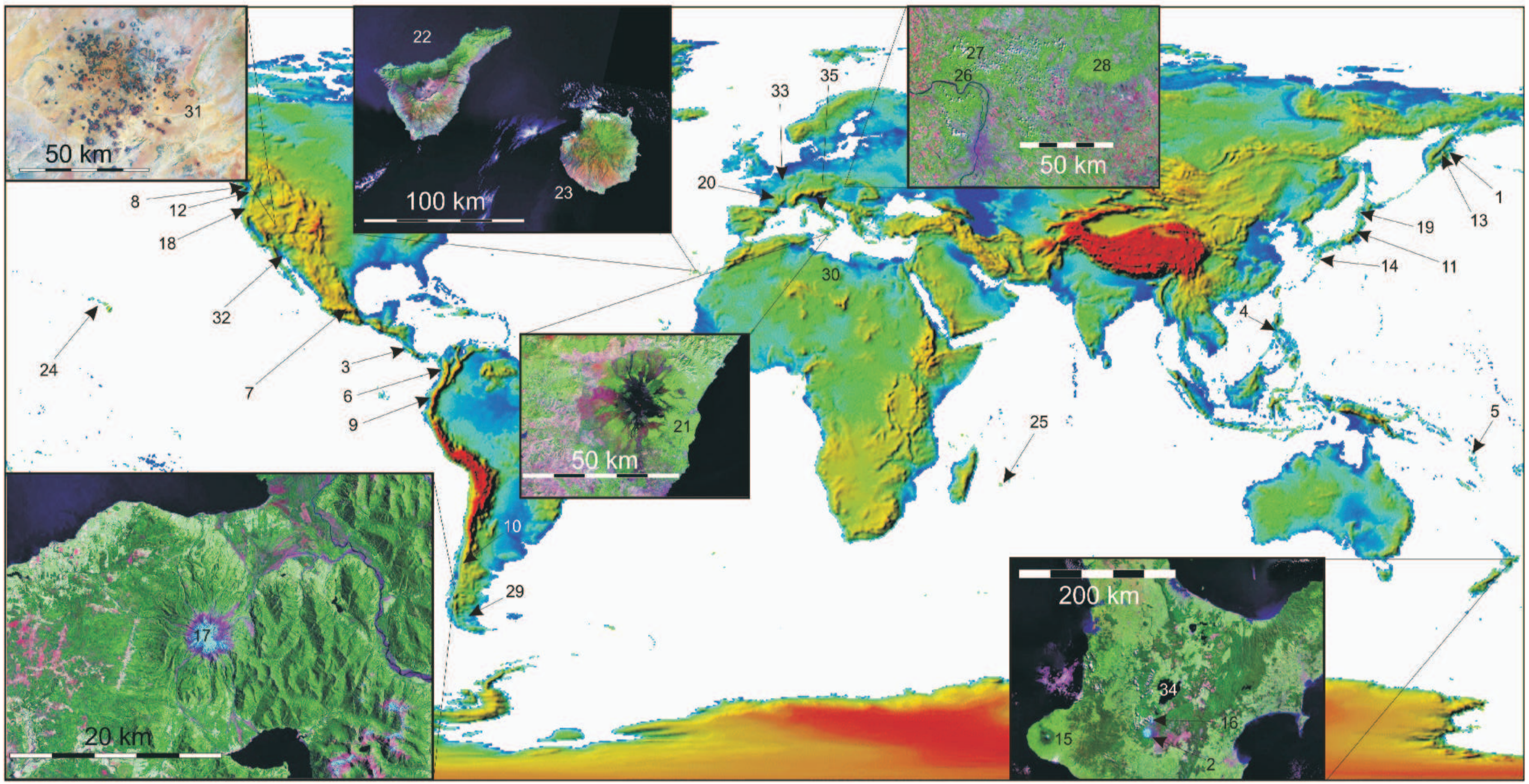
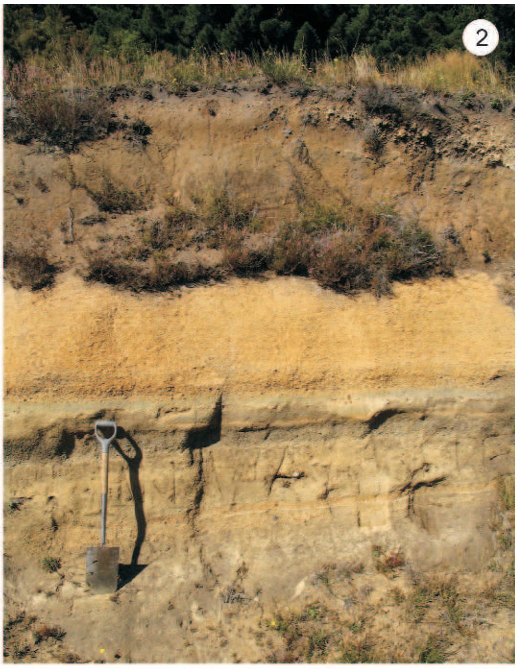
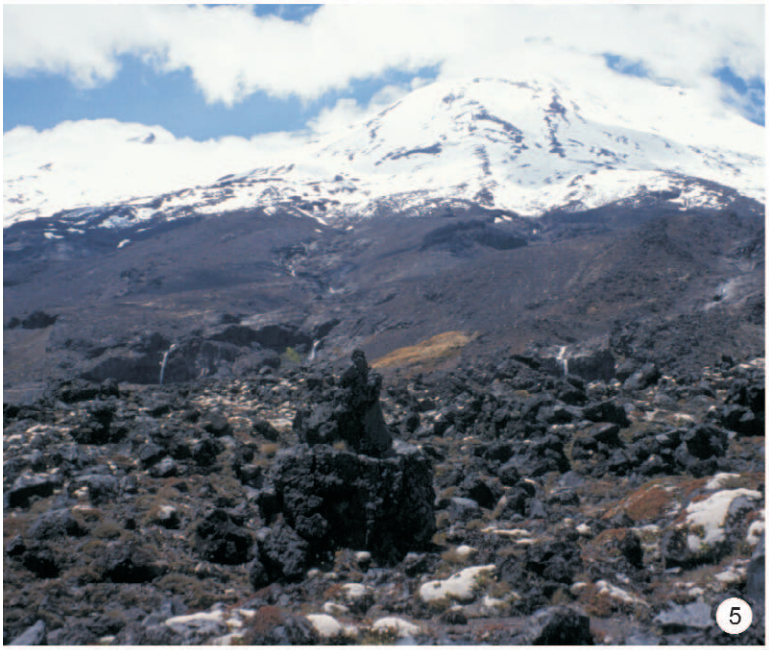
# Location map



1 — Shiveluch, Russia
2 — Ruapehu, New Zealand
3 — Casita, Nicaragua
4 — Pinatubo, Phillippines
5 — Ambae, Vanuatu
6 — Nevadodel Ruiz, Columbia
7 — Pico de Orizaba
8 — Mt Rainer, Washington, USA
9 — Cotopaxi, Ecuador
10 — Villarica, Chile
11 — Bandai, Japan
12 — Mt St Helens, Washington, USA
13 — Bezymianny, Russia
14 — Unzen, Japan
15 — Taranaki, New Zealand
16 — Tongario, New Zealand
17 — Calbuco, Chile
18 — Mt Shasta, California
19 — Hokkaido-Komagatake, Japan
20 — Cantal, France
21 — Valle del Bove, Etna, Italy
22 — Tenerife, Spain
23 — Gran Canaria, Spain
24 — Hawaii, USA
25 — Renuion, France
26 — Danube Bend, Hungary
27 — Börzsöny Mts, Hungary
28 — Mátra Mts, Hungary
29 — Pali Aike Argentina
30 — Al Haruj, Libya
31 — Hopi Butte, Arizona, USA
32 — Cerro Colorado, Sonora, Mexico
33 — Laacher See, Germany
34 — Lake Taupo, New Zealand
35 — Colli Albani, Italy

# Plate I





1. A) Volcanic ring plain around Ruapehu volcano in the central North Island, New Zealand. B) Volcanic ring plain of the Karymsky volcano (Russia) evolving in a caldera.
2. Tephra succession around the Tongariro–Ruapehu volcanoes, New Zealand.
3. Lahar deposits in the Ruapehu ring plain. The deposit consists of large rounded blocks of volcanic lithics in fine-grained matrix. The deposit is located about 15 km from the source.
4. Columnar jointed block in block and ash flow triggered lahar deposit in Ruapehu.
5. Snow capped peak of Ruapehu a perfect site to initiate lahars in case of hot pyroclastic flows would travel through this region. A combination with an existing crater lake in the peak of Ruapehu makes this volcano hazardous of lahar initiations.

## Plate II

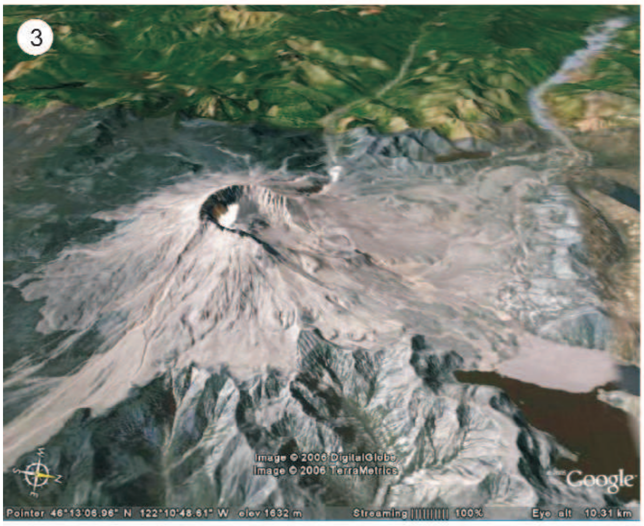
1. Deep valley network filled with ancient lahar deposits in the western flank of Ruapehu.
2. Block and ash fan on the Calbuco volcano in southern Chile dominated by block and ash flow deposits and minor reworked volcaniclastic successions.
3. Debris flow deposit in the Osorno volcano ring plain in southern Chile.
4. Hyperconcentrated mud flow deposits from the 18th of March 2007 lahar from Ruapehu volcano, New Zealand, near the Tangiwai Bridge.
5. Lahar destroyed Aztec city of Cholula in the ring plain of Popocateple in Mexico. Popocateple is visible in the distance.
6. Aztec pyramids under matrix supported distal lahar deposit facies in Cholula, Mexico.

# Plate III

1. Non-charcoaled bed flattened tree trunks in the1971 Turbio Valley lahar deposits near Pucon, Chile.
2. Massive, thick unit of volcaniclastic breccia in the Liucura Valley near Villarica volcano in Chile. The origin of the sequence is still under debate. The textural characteristics is similar to debris flow deposits from lahars, however, the angular and largely monomict volcanic lapilli content suggestive for pyroclastic flow origin. It is a good compromise to interpret this section as a pyroclastic flow triggered lahar deposit.
3. Close up view of a hot lahar deposit (same as in Plate III, 2).
4. Fluvial deposits of the Turbio valley. These valley is a common routes of lahars such as the 1971 lahar initiated from the NE flank of the Villarica volcano, Chile.
5. Turbio valley filled by the 1971 lahar deposits that formed a delta in the Lake Villarica, Chile.
6. Block-and-ash flow deposit texturally similar to lahar deposited debris flow deposit, but have characteristic magnetic fabric as AMS study reviled its magnetic fabric (Mendoza, Argentina).

# Plate IV

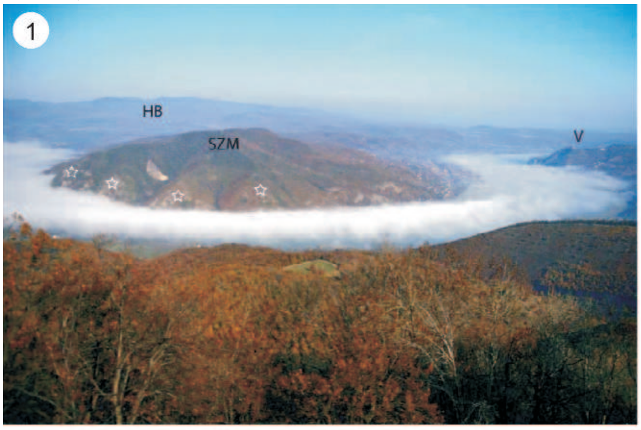
1. Ring plain around Popocateple volcano in Puebla, Mexico. In the historic time devastating lahars inundated large Aztec cities in the region, regardless of the complete destruction of those ancient cities, Cholua and other large cities, including Mexico City of 15 million people, are in potential danger of lahars may initiate from Popocateple.
2. Sebu dams built in many Japanese volcanoes, to reduce the mechanical energy of down flow lahars.
3. Horse-shoe shape escarpment of the Mt St Helens volcano on an oblique satellite image (Google Earth) as a result of sector collapse of the summit of the volcano. Note the flat smooth surface area in the ring plain partially filling a lake (black field).
4. Horse shoe-shaped escarpment of the summit of Mt St Helens, as a result of the 1980 eruption (photo by S. J. Cronin).
5. Mt St Helens collapsed summit area looking from the ring plain (photo by S. J. Cronin).
6. Debris avalanche matrix with chaotic volcanic lithic fragments from a volcanic debris avalanche deposit of the Ruapehu volcano, New Zealand.

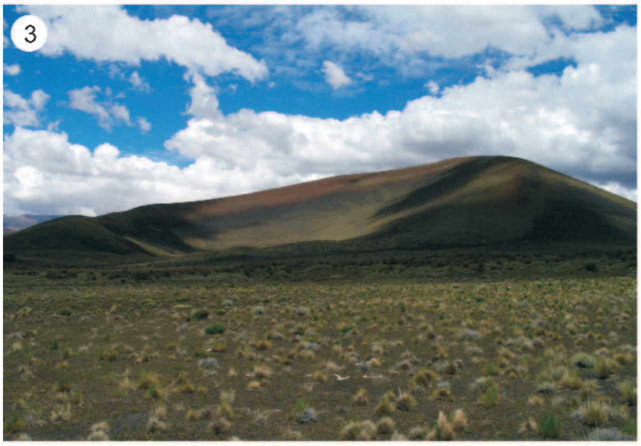


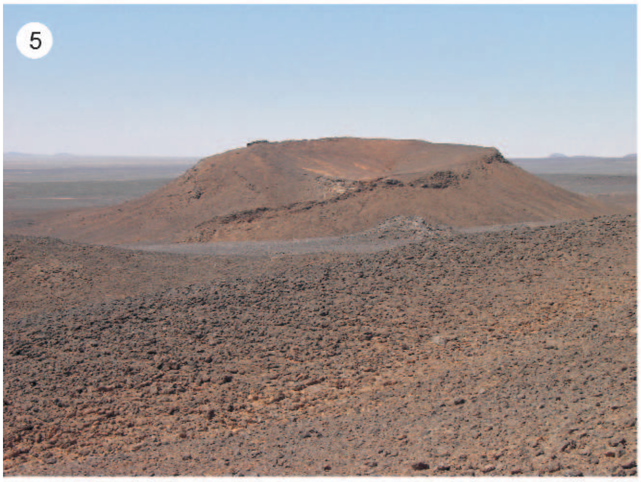
# Plate V

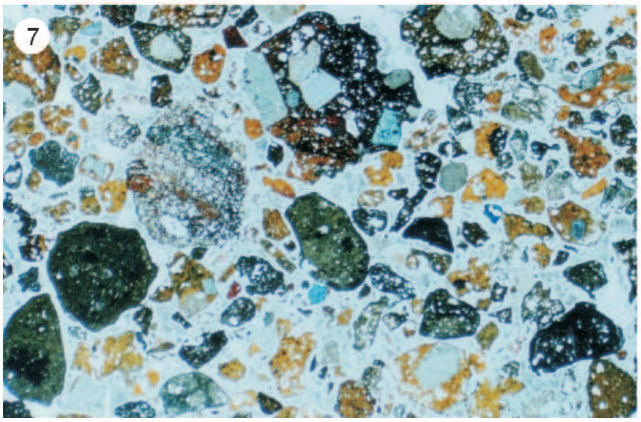
1. Proximal part of a volcanic debris avalanche with large (tens of metres across) hummocks about 15 km from the crater of Popocateple volcano.
2. Distal part (about 50 km away from the source) of hummocky surface with small (metre scale) hummocks around Calbuco volcano, southern Chile, as a result of a major volcanic debris avalanche of this volcano.
3. Sheared base zone of a debris avalanche deposit on Popocateple, Mexico.
4. Large, oversteepened strato cones develop over salar deposits in the Altiplano, ready to collapse due to the lubricated basal rock units.
5. Hummocky surface of a volcanic debris avalanche deposit in the Altiplano developed over salar deposits.
6. Steep northern flank of Tenerife Island inferred to collapsed several times in its history. The present day edifice of Pico del Teide is over 4000 m above sea level and unsupported in the northern side of the island.

# Plate VI





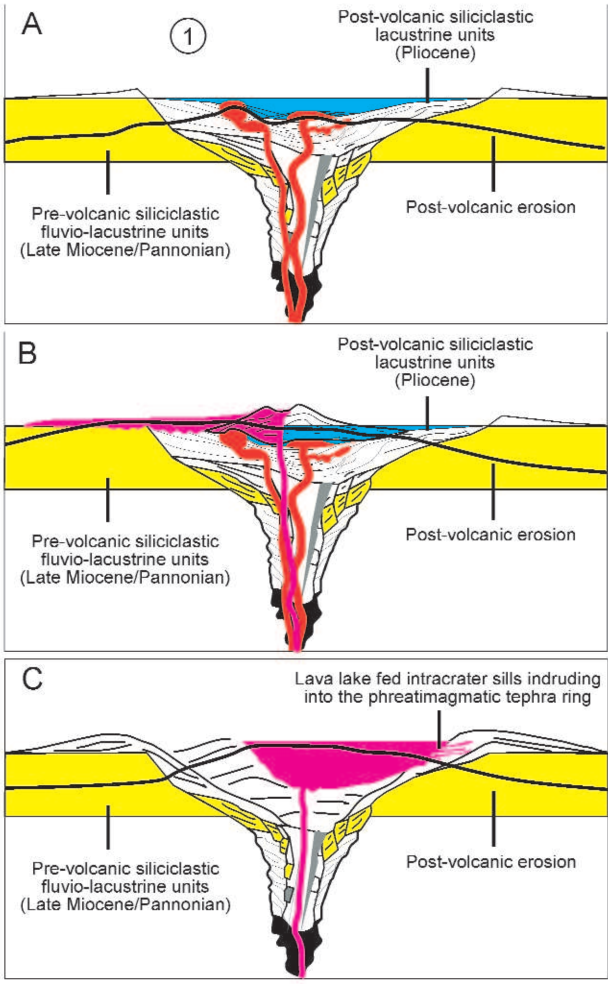


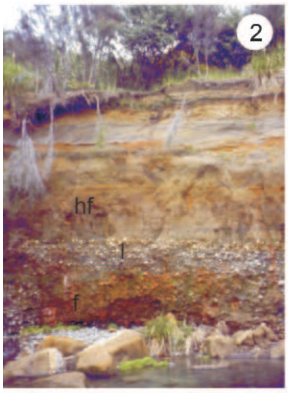


1. Danube bend is inferred to be a former horse-shape amphitheatre structure, a result of a former volcano collapse.
2. Volcanic debris avalanche deposits in the Visegrád Mt, Hungary.
3. Eroded scoria cone from Mendoza, Argentina.
4. Skeleton shape volcanic cones in the Pali Aike Volcanic Field in Argentina. The erosion dominated buy high wind, removed the loose tephra, leaving behind skeleton-like structures dominated by lava spatter layers.
5. Undercutting of scoria cone beds causing sudden collapses of large part of the pyroclastic units of scoria cones of the Al Haruj al Abiyad Volcanic Field in Libya. This way, the cones erode in a different path than it has been predicted from scoria cones dominated by granular lapilli successions.
6. Deep gully network on the outer flank of the Cerro Colorado tuff cone in the Pinacate Volcanic Field, Sonora, Mexico.
7. Volcaniclastic sediment accumulated in a crater lake of a maar of the Hopi Butte, Arizona. Note the diverse texture of individual clasts and their rounded shape.

# Plate VII

1. General model after NÉMETH et al. (2006) of the erosion of maar volcanic complexes in western Hungary. (A) Subaqueous vent(s) in water filled maars. (B) Scoria cone and associated lava flow(s) erupted in a water filled maar. (C) Scoria cone erupted in a dry maar and fed lava lake.
Continuous black lines represent potential erosion level in each case. Note the significantly different meanings of the preserved palaeo-surfaces covered by lava units.
2. Thick volcaniclastic succession in an incised valley 40 km from the Ruapehu volcano, New Zealand.
3. Eroded Miocene volcanic landforms of the Visegrád–Börzsöny Mtn.
4. Collapse scars (horse-shoe-shaped craters) of a lava dome dominated strato volcano in the Altiplano, Chile.
5. Irregular cone morphology of Calbuco volcano in southern Chile is a result of pulsating lava dome growth and collapse. This volcanic evolution and the erosion together produce a truncated cone morphology.
6. Micro-yardangs on vesicular basaltic lava flow surface of a pahoehoe lava field of Al Haruj al Abuyad, in Central Libya.

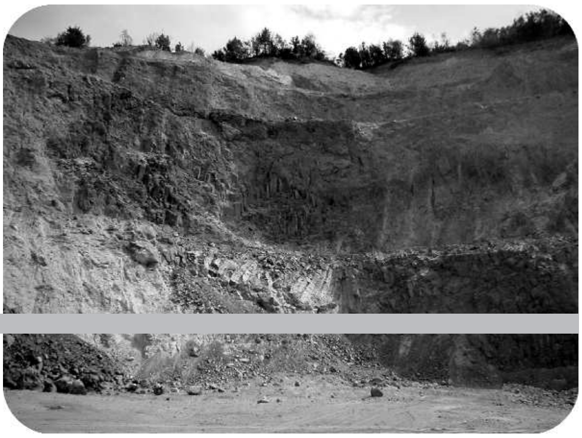




## Plate VIII

1. Elongated mega-yardang developed on an ignimbrite surface in the high wind area behind the Andean mountain chain in Mendoza.
2. Close up view of a mega-yardang surface developed over an ignimbrite surface. Note the micro-yardang texture similar to those developed on vesicular basaltic lava flows in Libya (see Plate VII, 4).
3. Thermal wheatering forming polygonal fractures over basaltic lava flow surface in Central Libya.
4. Displaced fratured blocks of a lava surface due to thermal wheatering of basaltic lava flow in Central Libya.
5. Radial jointing pattern over large coherent lava body due to thermal wheatering of lava flow surface in Central Libya.
6. Large-scale redistribution of pumiceous fall deposits (white zones in the peak) from volcanoes in the high Andes forming dunes in the lee-side of the mountain ranges.

# Chapter 8



# Subaqueous volcanism and associated features

## Introduction

Subaqueous volcanic eruptions are the most common on earth, considering the total volume of magma erupting along mid-oceanic rift axes, hot spot-related volcanoes and volcanoes erupting in shallow marine, lacustrine or subglacial environments. The total volume of effusive and explosive eruption products and their economic significance is huge in comparison to subaerially erupted volcanic products. In spite of the abundance of subqueous volcanism on Earth, the understanding of volcanic processes in such environments is very limited for two reasons; 1) logistical problems associated with collecting samples and studying sites underwater and 2) the long distances to and remoteness of the most suitable study sites.

From a logistical point of view our access to subqueous eruption sites is very limited, and even in cases where good access is possible we are largely limited to observations from the water surface (e.g. direct observations of the eruption clouds that may breach the water surface). Even in the case of submersible techniques, visibility in water is much less then in air and detailed observations are very difficult to obtain. Also, sample collection from the seafloor (e.g. after eruptive events) is largely limited to dredging or coring, which do not allow identification of outcrop (tens of metres scale) observations of key sedimentary features that may be key to interpreting the transportation and depositional regime that formed the volcanic unit. The remoteness of many subaqueous eruptions also limits our ability to reach eruption sites quickly enough to see the main eruptive phases, commonly responsible for the formation of the majority of the volcanic sediments and effusive products. Therefore our understanding of subaqueous volcanism relies mainly on studies of ancient volcanic successions, where we are confidently able to establish the subaqueous environment where the studied volcanic rock unit formed (e.g. identification of subaqueous deposition for under- and overlying sedimentary successions sandwiching volcanic rock units (WHITE et al. 2003). However, although this approach is widely used, potentially serious errors resulting from simple sedimentological problems that arise in the interpretation of contact zones between non-volcanic, marine and volcanic rock successions remain a source of debate in many studies. This is partially the reason why many Precambrian volcanic successions are used for understanding subaqueous volcanic processes, since at that time, the Earth's surface was covered by water (MUELLER 1991, DOUCET et al. 1994, LAFRANCE et al. 2000, MUELLER et al. 2000, 2002). In spite of every effort to use careful, well-supported arguments and reasoning to pursue such studies on older rock formations, we all face major difficulties in understanding the tectonic and palaeoenvironmental development of subaqueous volcanic settings. It is also common to find that although older volcanic rock units may be well-suited to understanding subaqueous volcanism, they are commonly discontinuous or dissected, hindering any chance to see the full 3D facies relationships of different volcanic rock units (WHITE et al. 2003). This problem is one of the major issues that makes it difficult to confidently use older successions for facies modelling of subaqueous volcanic successions, even if they are relatively young (a few millions of years old).

## Effusive subaqueous volcanism

Effusive volcanism in subaqueous settings is the most widespread type of volcanism on Earth, especially along the mid-ocean ridges (BATIZA and WHITE 2000). Along with extensive lava flows, large volumes of non-explosive, quench fragmented rock called hyaloclastite (Plate I, 1) forms due to non-explosive fragmentation of erupting lava (LONSDALE and BATIZA 1980, BATIZA et al. 1984, 1989, SMITH and BATIZA 1989, BATIZA and WHITE 2000). In addition, significant

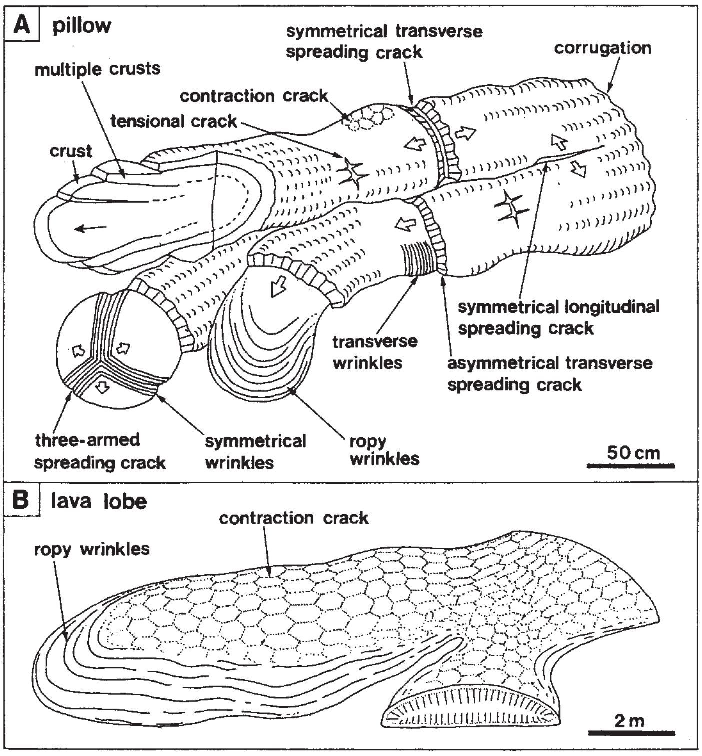


*Figure 8.1.* Theoretical architecture of a pillow lobe (after Yamagishi 1985 in Goto and McPhie 2004: p. 322, fig. 13). Multiple crusts form on the tip of the advancing lava toe. Two individual lava lobe initiates from a single master lobe. The open and closed arrows refer to spreading and flow direction respectively. A) architecture of a pillow lava lobe, B) architecture of a subaquaeous lava lobe

volumes of magma is emplaced subaqueously as intrusive bodies (Plate I, 2) and associated peperite, formed by *in situ* mingling and mixing of fragmented shallow dyke and sill magma with host marine and/or lacustrine sediments (Einsele 1986, White et al. 2000, Skilling et al. 2002, White and Houghton 2006). Magma rise near the sea/lake floor is predominantly controlled by the marine and lacustrine sediment pile. If the loose marine and lacustrine sediment pile is thick, magmatic crack propagation stops functioning and magma can stop rising and spreads laterally over the sea/lake floor, or can invade the marine/lacustrine sediment, forming extensive peprite facies (White and Busby-Spera 1987, McPhie 1993). Effusive and intrusive-dominated subaqueous volcanism produces large volumes of sediment, fragmented during non-explosive or weakly explosive events, that are distributed by currents or gravitational failure events and commonly form sheet-like deposits on the sea/lake floor, inter-bedded with non-volcanic marine and lacustrine sediments (White and Busby-Spera 1987).

Seamounts are a common volcano type associated with mantle up-welling sites, where magma pours out onto the seafloor, and forms submarine volcanic edifices (Staudigel and Schmincke 1984, Maicher 1999, Corcoran 2000, Trua et al. 2002, Clouard et al. 2003, Tarduno et al. 2003, Kopp et al. 2004). Large volcanic plateaus (e.g. Ontong-Java plateau) are also commonly associated with hot spot magmatism, and may form sheet-like flood lava fields and associated hyaloclastite units that cover thousands of square kilometres (Tarduno et al. 1991, Berger et al. 1992, Kerr et al. 2000, Wessel and Kroenke 2000, Mann and Taira 2004).

The most obvious subaqueous lava forms are pillow lavas (Plate I, 3), lobate lava flows and sheet-like lava fields (Batiza and White 2000). Laboratory experiments show that at low effusion rates and on gentle slopes pillow lavas form (Moore 1975, Cas 1992, Batiza and White 2000). Increasing effusion rates and steeper slope angles result in lobate lava flows and sheet-like lava fields (Batiza and White 2000). Pillow lavas have many distinct morphological features (Figure 8.1). Pillows can be elongate, oval shaped to complex interconnected piles and are commonly associated with hyaloclastite facies (Walker 1992). Mega-pillows (Figure 8.2) have also been recognised (Goto and McPhie 2004). The rim of the individual pillows is glassy, sometimes with multiple glassy layers (Kawachi and Pringle 1988); their interior, however, is usually more crystalline due to their slower cooling rate. Pillow lava piles commonly form complex pillow volcanoes. Subaqueous lobate lavas form large lobate-shaped, often sack-like forms that are interconnected. Sheet lava flows are associated with fast spreading mid-ocean ridges and are believed to fed by lava tube networks.

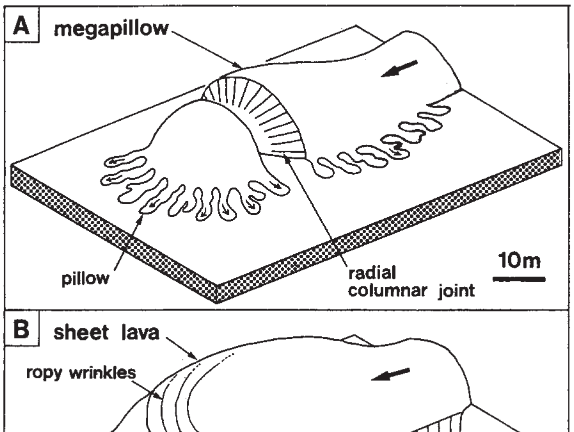


*Figure 8.2.* Diagrammatic representation of megapillow (A) and lobate lava flow (B)(after Goto and McPhie 2004: p. 324, fig. 14)

Hyaloclastite is a glass-fragment rock that results from non-explosive, quench fragmentation of subaqueous lava flows (Silvestri 1963, Batiza et al. 1984). Basaltic hyalo-

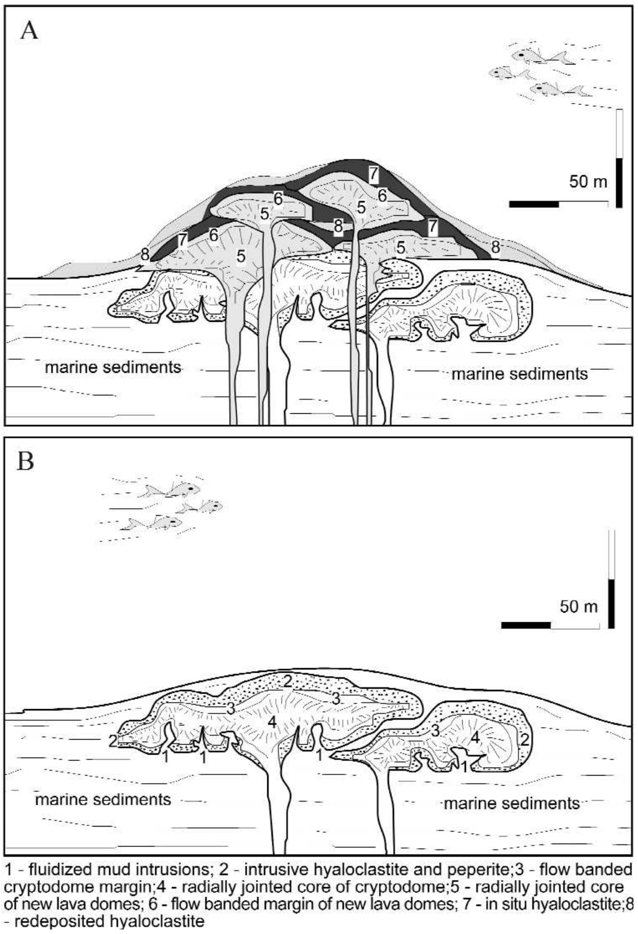


***Figure 8.3.*** Theoretical model of the evolution of the Pálháza cryptodome- and dome-complex (after Németh et al. 2007). In the initial stage rhyolitic magma intrude into the seafloor sediment, forming intrusive hyaloclastite and various types of peperites (A). As new rhyolite magma intrudes, the new melt breaches the seafloor sediments, and pour out to the seafloor. Upon contact with the water, thick pile of hyaloclastite forms around the lava dome. Currents, and gradual increase of the slope angle of the growing dome complex initiates remobilisation, and reworking of the accumulated hyaloclastites

clastite is commonly associated with extensive pillow, lobate and sheet-like lava flows (Batiza et al. 1984). In modern sea floors acidic hyaloclastite piles are relatively rarely identified and are mostly associated with rift settings. However, rhyolitic hyaloclastite is relatively common in the geological record, mostly inter-bedded in marine sediments and associated with lava domes or cryptodomes that grew within and on top of sediments on the former seafloor (Plate I, 4) (Cas et al. 1990, Yamagishi 1991, Scutter et al. 1998, DeRita et al. 2001, Rinaldi and Venuti 2003, Németh et al. 2007). These hyaloclastite-dominated rhyolites, rhyodacites and dacite domes can be very complex volcanoes (Figure 8.3) (McPhie and Allen 1992, McPhie et al. 1993, Gimeno 1994, Goto and McPhie 1998, Doyle and McPhie 2000, Gifkins et al. 2002, Rinaldi and Venuti 2003). Subaqueous volcanic successions and associated rhyolite lava domes are common from a range of tectonic settings (Cas et al. 1990, Gimeno 1994, DeRita et al. 2001). Lava domes, cryptodomes and lava flows (Figure 8.4) easily burrow into and apparently rapidly expand into soft sediments in subaqueous settings (Kano 1989, Hanson and Hargrove 1999, Gifkins et al. 2002), forming various types of peperite and associated hyaloclastite units. The term cryptodome is used to describe coherent bulbous magmatic bodies emplaced at shallow levels into host sediment that never breached the sedimentary cover (Goto and McPhie 1998, Stewart and McPhie 2003). Cryptodomes in subaqueous settings are surrounded by intrusive hyaloclastites (Stewart and McPhie 2003) formed by the quench fragmentation of the margin of the intruding magmatic body (Figure 8.5 and 8.6). Intrusive hyaloclastite is also known as peperitic hyaloclastite (McPhie et al. 1993). Hence, intrusive hyaloclastite and peperite are very closely related terms. Some intrusive hyaloclastite bodies could also be termed peperite, but only if there is evidence of host sediments injected into the intrusive body.

By contrast, while endogenous lava domes are partially grown in soft sediment (McPhie et al. 1993, Goto and Tsuchiya 2004), at their upper margins lava breaches the

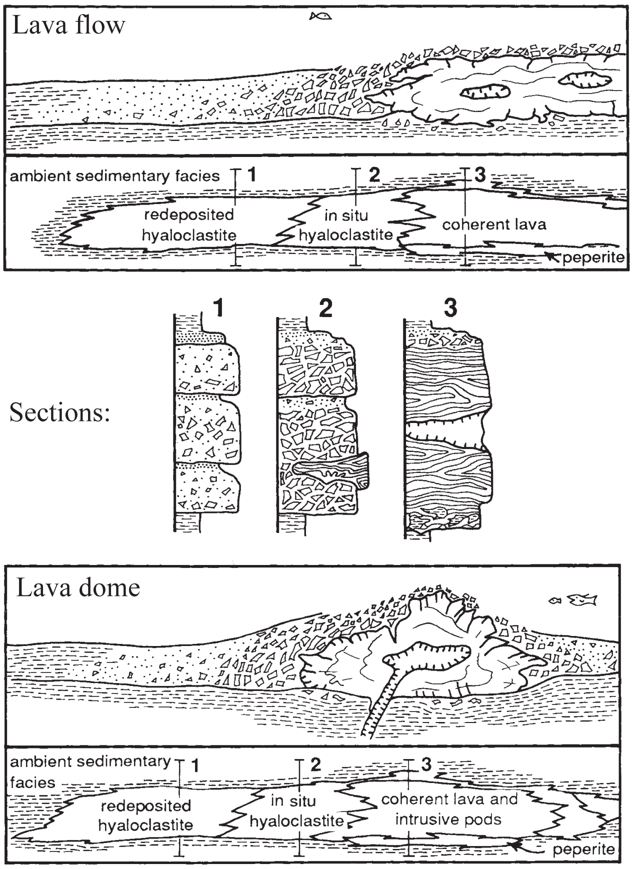


***Figure 8.4.*** Diagrammatic representation of facies relationships of subaqueous lava flows and lava domes (after McPhie et al. 1993). Numbers represent theoretical sections across the lava flows and lava domes. The sections are very similar, and made a challenging work to distinguish subaqueous lava flows from lava domes preserved in the geological record

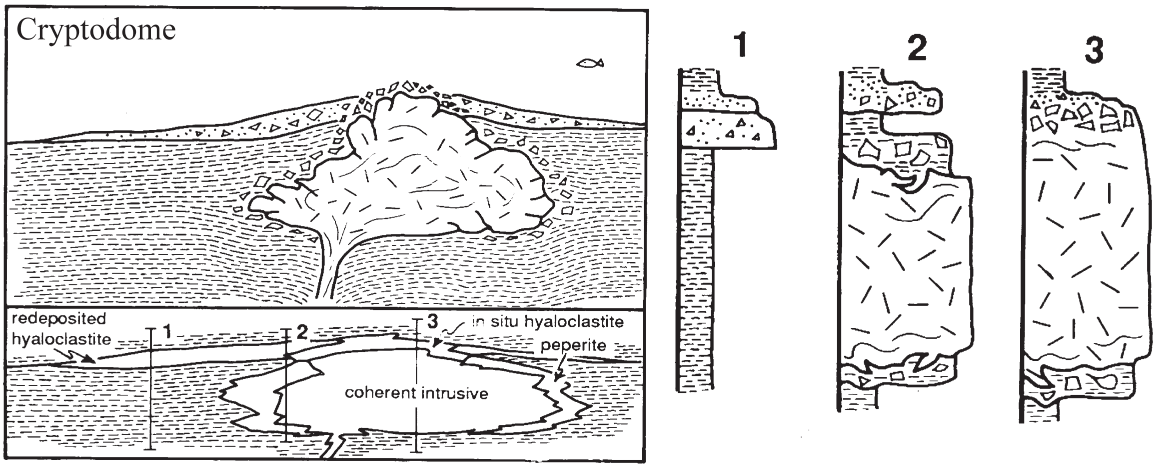


*Figure 8.5.* Theoretical facies distribution associated with a subaqueous cryptodome (after McPhie et al. 1993)

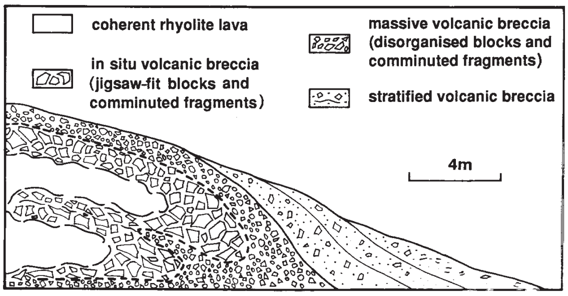


***Figure 8.6.*** Facies association of a subaqueously emplaced rhyolite lava flow from the Miocene, Ushikiri Formation, Japan (after Kano et al. 1991)

sedimentary cover and, upon contact with the sea/lake water, quench fragmentation of the lava surface forms *in situ* hyaloclastite. These hyaloclastite successions may laterally interconnect with redeposited hyaloclastite units formed by volcaniclastic deposits carried away from the *in situ* hyaloclastite piles by currents (Goto and Tsuchiya 2004). Extrusions in very shallow subaqueous environments may generate explosive eruptions and form tephra mounds and cones that overlie domes (Cas et al. 1990, 1992). Distinguishing geologic exposures of subaqueous cryptodomes from endogeneous lava domes may be challenging due to the fact that a cryptodome may expand enough to breach the host sediment. Cryptodome and dome structures can overlap with a highly intricate architecture that can contain a variety of rock textures, as is documented from the Tokaj Mountains, NE Hungary (Németh et al. 2007).

### Eruption dynamics of explosive subaqueous volcanism

The basic difference between subaqueous and subaerial volcanism is that the magma, upon reaching the lake or sea floor, can encounter water (White et al. 2003). Therefore, there are two basic problems we have to face (White et al. 2003); 1) how this magma–water interaction took place, and 2) when, or what stage of the eruption was affected (most)? It is known from direct observations and inferred from ancient settings that many subaqueous volcanic eruptions occurred in shallow subaqueous environments where the eruption intensity was large enough to breach the water surface and produce eruption clouds that emerged above the water surface, or created physical conditions underwater that are similar to subaerial eruption clouds (White et al. 2003).

In summary, we can state that water differs significantly from air as an eruption media (White et al. 2003) with respect to the following factors: 1) the role of steam generated during the eruption, 2) the role of pressure caused by the water column, 3) the role of heat capacity and conductivity of water, and 4) the water rheology. These four fundamental parameters are significantly different from eruption conditions in air.

When magma rises and erupts in dry environments, its explosive fragmentation is driven by the exsolution and expansion of gases that were trapped within it under high pressures at depth. However, in wet environments, magma is dominantly fragmented through a conversion of thermal energy to mechanical energy when the >1000 °C magma meets water. This contact leads to a chain-reaction process often referred to as molten-fuel-coolant-interaction (MFCI) (Zimanowski et al. 1991, 1997a, Zimanowski 1998). The MFCI process not only pulverises and chills of the magma, but the resulting shock waves also permeate and disrupt surrounding rock and sediment (Wohletz 1986, Zimanowski 1998). Water (the coolant) vaporizes above its boiling point upon contact with hot magma (the molten fuel) and its expansion enhances magma fragmentation. Entrapped water can expand quickly and fragment the melt. During the premixing stage between magma and water, the water becomes superheated. Thermohydraulic fragmentation of the melt take place and the fragments dispersed in the superheated water, followed by sudden fragment dispersion when the superheated water flashes to steam (Zimanowski et al. 1997a, b). The resulting steam affects the dispersion of the fragmented magma during the lifetime of the steam (which is usually short due to its buoyancy and low density). Steam also has an insulating effect, which could help to form spatter and locally weld ejecta close to eruption sites, where no direct contact between hot particles and cold water occurs (Mueller and White 1992, Kano et al. 1994, Mueller et al. 2000, Cas et al. 2003). In shallow water, water can boil to steam and expand dramatically. However, it is inferred that the expansion of and the phase change from water to steam is gradually suppressed with increasing water depth, and generally believed that at over 3 km water depth no meaningful phase changes are possible and only limited expansion can take place.

The water pressure also increases with depth, which affects the solubility, expansion, and release of magmatic volatiles, as well as the development of steam (Fisher 1984, Fisher and Schmincke 1984). In practical terms, this means that increasing pressure decreases magma fragmentation since volatile expansion is suppressed. In a simple way this means that much higher concentrations of magmatic volatiles are required in order to reach the same amount of fragmentation at greater depth (e.g. volatile-rich magmas) (Dimroth and Yamagishi 1987). The role of pressure on magma–water interaction issues, however, remain controversial. Experimental studies demonstrate that at fixed magma–water ratios the violence of magma–water interaction increases with pressure (Wohletz 2003) (e.g. deeper water would cause stronger, more powerful MFCI interactions); however, this might be just a reflection of the experimental set up.

Water heat capacity and conductivity is significantly higher in water than in air and therefore rapid cooling of magma and erupted gases are expected. The erupting melt will quench fragments quickly when direct contact between magma and water occurs. Further quench granulation can also take place as large hot particles break apart, leading to formation of smaller and smaller particles as the process continues until the magma has cooled completely. Intrusion into seafloor water-saturated sediments can result in quench fragmentation and production of large volume of spalled glass from the surface of the erupting magma. Already fragmented hot pyroclasts leaving a subaqueous vent can suffer further fragmentation in the margin of the eruption column, where individual clasts come into contact with ambient water. In general fluidal clasts (e.g. spindle bombs) are not expected to form in subaqueous environments due to the brittle fragmentation of the rapidly chilled fragments, unless water is temporarily excluded from eruption sites (Mueller and White 1992). Such situations are expected from time to time, when magmatic volatiles are able to exclude water temporarily or when significant volumes of steam are produced during the contact between the hot magmatic bodies and water. Still, if fluidal particles are hot enough when they come into contact with ambient water, they can brittly fragment and generate blocks and shards from the original fluidal clasts (Mueller and White 1992, Clague et al. 2003, Mueller 2003). On the other hand, large-volume fluidal clasts can generate steam envelopes that insulate the clast from surrounding water and thus prevent further brittle fragmentation

In comparison to air, water is dense and viscous, which strongly affects the eruption plume geometry as well as the dispersal of the ejected material (White et al. 2003). Ballistically transported fragments are not expected to be common due to the high viscosity and density of the water in comparison to air (e.g. the erupted fragments will slow down, stall and fall back through vertical movement, instead of following a ballistic trajectory). Ballistic transport is only likely in near-vent water exclusion zones, or when the eruption takes place in shallow water, and fragments can breach the water surface (Kokelaar 1986, White 1996). Buoyant particles can be transported to the water surface, and then aqueous currents can carry them away. Even basaltic (high density clasts) can be buoyant, at least while they are hot and have entrapped magmatic volatiles in their core. Subaqueously erupted pumice can float until their pores become water saturated and they sink (Manville et al. 1998). Ash size particles can accumulate "en masse" from hot/warm water plumes as they drift away from the eruption sites (Cashman and Fiske 1991, Fiske et al. 1998). The resulting deposits of such aqueous plumes perhaps form different bedding characteristics than deposits settled from subaerial eruption plumes. It has recently been recognized that large volumes of fine particles from subaqueous eruptions, initially transported near the water surface, can form vertical density currents (Manville and Wilson 2004). Once they reach the sea/lake floor these vertical density currents form radially dispersing, now horizontal, density currents (White 1996, 2000). Perhaps in the case of drifted subaqueous ash plumes vertical density currents can form radiating density currents on the sea/lake floor, complicating the reconstruction of their source. Gas-supported subaqueous density currents are considered to be true subaqueous pyroclastic flows (White 2000). To generate such currents, the system must have high particle concentrations and have very low concentrations of low density, buoyant (e.g. pumice) particles in order to keep the current moving on the sea/lake floor. The higher density of the water as the supporting media will make these flows slow-moving phenomena. The few field studies and theoretical considerations available suggest that the resulting subaqueous pyroclastic flow deposits are probably very similar to subaerial pyroclastic flow deposits, even including significant welded successions (Sparks et al. 1980). Subaqueous surges (e.g. turbulent, low particle concentration gaseous density currents), however, are not expected to move along the sea/lake floor due to their low density and therefore high buoyancy (White et al. 2003).

### Shallow subaqueous explosive volcanism

In shallow subaqueous environments where the water depth is less than few hundred metres, Surtseyan-type (predominantly basaltic) explosive subaqueous eruptions can take place (White and Houghton 2000). Surtseyan volcanoes are known from all shallow subaqueous environments (submarine, lacustrine, glacial). The low confining pressure of the water column contributes to the explosiveness of this style of volcanism, and is inferred to occur at water depths of up to about 200 m (White and Houghton 2000). Surtseyan-type volcanoes are generally monogenetic (see chapter 5) and their eruption duration ranges from few days to years (Plate I, 5) (White and Houghton 2000).

Surtseyan-type eruptions share many characteristics with subaerial maar and tuff ring forming eruptions, particularly magma fragmentation as a consequence of magma–water interaction driven by MFCI processes. Magma–water interaction in Surtseyan eruptions mostly takes place at the interface between the top of the conduit and sea/lake floor (Kokelaar 1983, 1986, White and Houghton 2000). This means that during magma–water interaction abundant water and/or water saturated slurry is available to drive explosions (Kokelaar 1983). Surtseyan eruptions build up steep sided tuff cones of phreatomagmatic tephra constructed by fallout and pyroclastic density currents (White and Houghton 2000). The volcanic edifices are surrounded by thick gravity mass flow deposits resulting from mass wasting of the loose and wet tephra (Verwoerd and Chevallier 1987, Sohn 1995, Cole et al. 2001, Martin 2002, Maicher 2003). Volcano collapses are also common and are recorded by large scars draped by younger tephras. The initial eruption of such volcanoes commonly starts with lava effusion, forming thick piles of pillow lavas (Staudigel and Schmincke 1984, McPhie 1995, Andrews 2003). Over these pillow volcanoes a partially subaqueous and partially subaerial tuff cone can develop (Staudigel and Schmincke 1984, Cas et al. 1989, Schmidt and Schmincke 2002). Tuff cones that form over a pile of hyaloclastite and/or pillow mounds in subglacial settings are commonly named tindars (Smellie 2000).

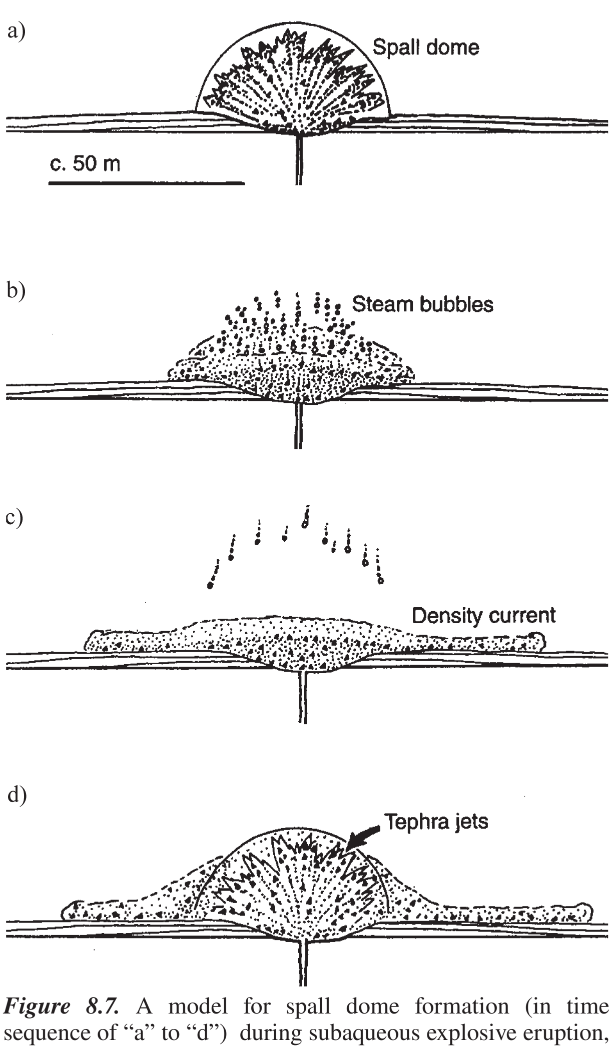


***Figure 8.7.*** A model for spall dome formation (in time sequence of “a” to “d”) during subaqueous explosive eruption, based on the reconstruction of the formation of the Pahvant Butte sublacustrine volcano (after White 1996: p. 259, fig. 9a–d)

Magma fragmentation in shallow subaqueous settings is inferred to be similar to subaerial volcanoes, with decompression triggered by magma rise toward the surface (in this case, the sea- or lake-floor) where volatiles exsolve and form a magma-bubble foam in the upper section of the conduit. The general lack of accidental lithic fragments derived from the conduit wall indicates that fragmentation mostly takes place in the topmost part of the conduit or in the crater. Fragmentation of magma in different stages of vesiculation produces pyroclasts with wide-ranging but generally low vesicularity; this range of vesicularity is characteristic of Surtseyan deposits (Houghton and Wilson 1989). The magma foam or pyroclast-charged slurry will exit the vent if its pressure exceeds the ambient pressure which is controlled by 1) the depth of fragmentation relative to the sea/lake floor (Wilson and Head 1981) and 2) the volcanic conduit radius (Woods 1998) (narrower conduits in hard rocks produce higher pressure). In this respect, in subaqueous environments it is inferred that the fragmented slurry-like material leaves the vent slowly and looks rather like a “boil over” event, especially if the eruption is characterised by low mass flux or occurs in deeper water (Woods 1998). Direct observations confirm that during Surtseyan eruptions a gas + water + ash mixture radially leaves the vent accompanied by a hemispherical shock wave manifested as an expanding spall dome (Figure 8.7) (White 1996). During the onset of a subaqueous eruption a gas-thrust zone of the eruption plume forms over the vent, but this gas-thust region quickly transforms to a buoyant convective eruption column. This buoyant eruption column can reach the water–air interface, but is usually not able to breach it, unless the eruption is initiated in a very shallow subaqueous environment (Kokelaar 1983, Kokelaar and Durant 1983). In deep water eruptions, an extensive mushroom-like cloud may form near the water surface as a result of the rise of the buoyant convective cloud (Cashman and Fiske 1991). These near surface regions can quickly become overcharged in ash particles, and the particle cloud can fall back to the sea floor via vertically descending sediment gravity currents (Carey 1997). Once these currents hit the sea/lake floor, they are inferred to continue as topographically controlled sediment gravity flows. These sediments are frequently reported from the geological record and are inferred to be one of the main parts of the growing volcanic edifice (White 1996, Smellie and Hole 1997). Identifying the link between the eruption and the resulting deposit is critical to interpreting these sediment gravity flow deposits, especially distinguishing them from other, non-volcanic gravity current deposits. Deposits resulting from pyroclastic density currents directly initiated by subaqueous volcanic eruptions are named eruption-fed pyroclastic density current deposits and are grouped according to their eruption column dynamics (Mueller and White 1992, White 2000, Mueller 2003). The behaviour of a subaqueous eruption column inferred to be controlled predominantly by the mass flux of the magma versus the mass fraction of the external water (Koyaguchi and Woods 1996). In the case of a low external water fraction in the eruption column (due to a high magmatic flux), the column will be steam-poor and therefore dense. Such columns are inferred to be low and probably easily collapse shortly after eruption, forming dense pyroclastic gravity flows that are more or less directly initiated from the crater. If particle concentrations are very high, the resulting deposits may even be welded. In the case of a high water fraction (due to a lower magmatic flux) the column can be charged in steam, and therefore could behave more buoyantly. The column will grow until it becomes too heavy for rise further, leading to collapse along its margins and spawning column margin-fed density currents (White 2000). If the column hits the water surface, extensive laterally spreading mushroom-like clouds may form, from which ash gradually falls out to form graded fallout beds (Cashman and Fiske 1991). If the rising column ingests more water it becomes gravitationally unstable and collapses to form low temperature, dense pyroclastic gravity flows (White 2000). Formation of eruption columns during shallow subaqueous eruptions is mainly restricted to the quasi-steady state eruption phases, commonly referred as continuous up-rush activity (Figure 8.8) (Kokelaar 1983, 1986). In vigorous eruptions in subaqueous environments, water exclusion zones can

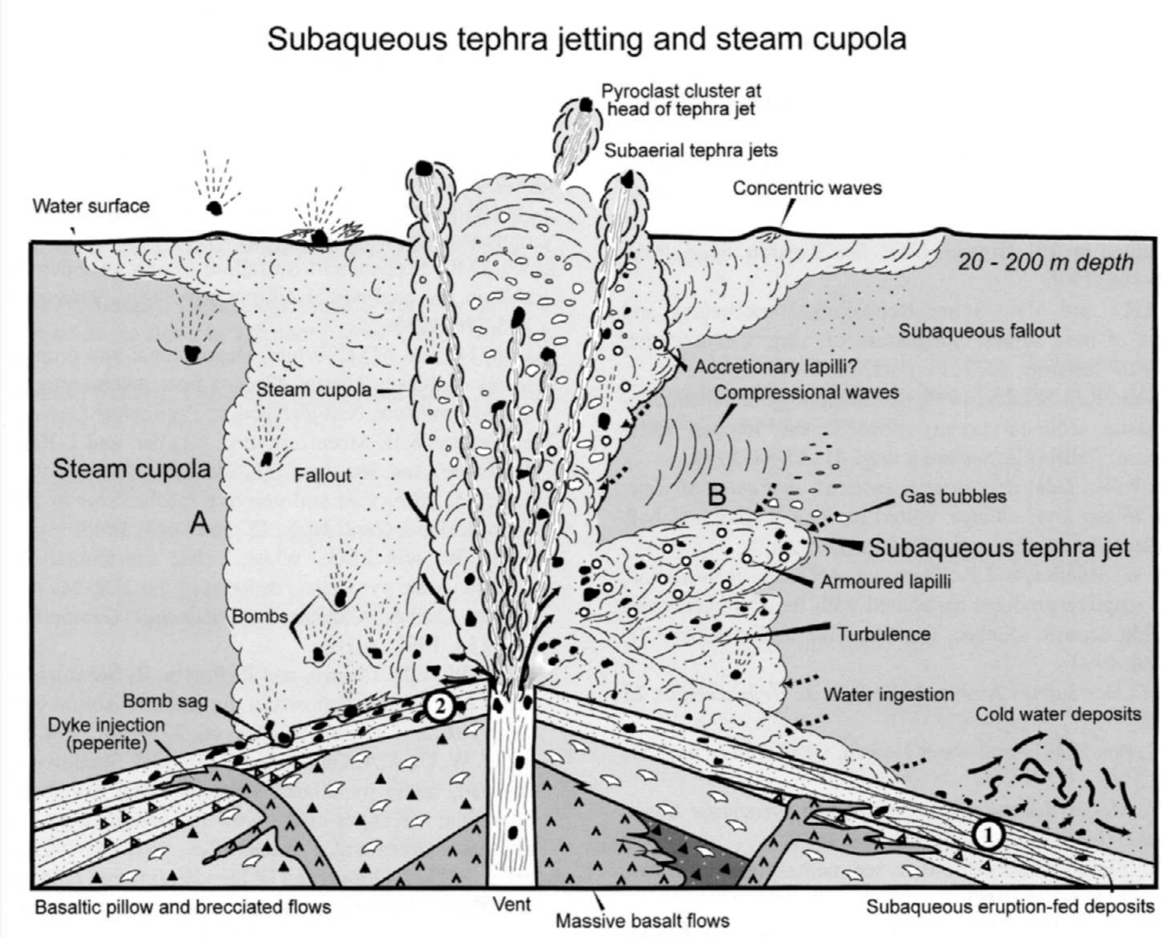


***Figure 8.8.*** A model cartoon of a small-volume shallow-water, Surstseyan-type eruption with continuous uprush conditions (shown on the left hand side of the figure "A") and individual tephra jets (shown on the right hand side of the figure "B"). Volcanic facies of lapilli tuff bed-dominated succession deposited from subaqueous eruption-fed density currents (1) and lapilli tuff breccia-dominated succession rich in ballistically emplaced impact structures formed under a steam cupola (after Mueller 2003: p. 201, plate 4)

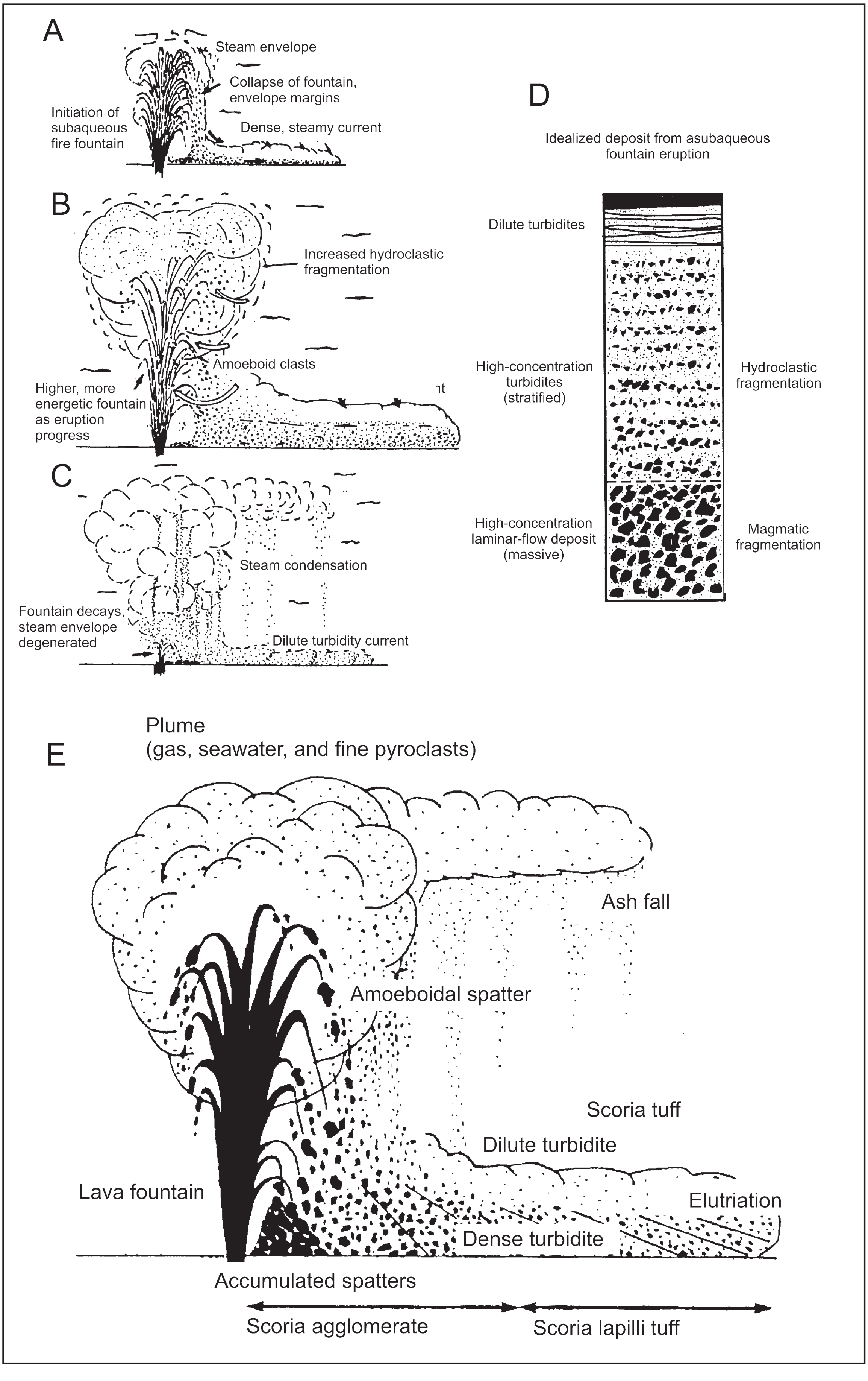

A
Steam envelope
Collapse of fountain, envelope margins
Initiation of subaqueous fire fountain
Dense, steamy current
B
Increased hydroclastic fragmentation
Amoeboid clasts
Higher, more energetic fountain as eruption progress
C
Steam condensation
Fountain decays, steam envelope degenerated
Dilute turbidity current
D
Idealized deposit from asubaqueous fountain eruption
Dilute turbidites
High-concentration turbidites (stratified)
Hydroclastic fragmentation
High-concentration laminar-flow deposit (massive)
Magmatic fragmentation
E
Plume (gas, seawater, and fine pyroclasts)
Ash fall
Amoeboidal spatter
Scoria tuff
Dilute turbidite
Lava fountain
Elutriation
Dense turbidite
Accumulated spatters
Scoria agglomerate
Scoria lapilli tuff

***Figure 8.9.*** Schematic model for subaqueous lava fountain eruption (after MUELLER and WHITE 1992). A) Initiation of subaqueous lava fountain on the sea/lake floor and formation of steamy, dense eruption-fed current. B) Eruption plume fully develops when eruption progress. Eruption-fed pyroclastic density currents form. Along the eruption cloud margin increased hydroclastic fragmentation occurs. In steam envelop, hot, fluidal pyroclasts can form. C) In the final stage of the eruption the steam envelope breaks, and dilute eruption-fed density currents passing the vent. D) In a theoretical section changes from magmatic to hydromagmatic fragmentation are apparent on the basis of the textural changes of the deposits formed (deposition from high particle concentration density currents to dilute ones). E) Theoretical relationships between the location of clasts formed in the column in the steam envelop and by the passing density currents (from high particle concentration to more dilute ones)

develop near the vent zone that allow formation of fluidal shaped bombs and armoured lapilli; armoured lapilli are generally considered to be able to form only in subaerial conditions, so careful work is required to reconstruct eruption conditions to determine whether eruptions took place entirely underwater or were partly emergent (Figure 8.9) (KOKELAAR 1983, KOKELAAR and DURANT 1983, KOKELAAR and BUSBY 1992, WHITE 1996, WHITE and HOUGHTON 2000). During low magma discharge rates, only intermittent explosions take place and activity is dominated by tephra jetting (WHITE 1996, MASTIN and WITTER 2000, NÉMETH et al. 2006). The deposits of tephra jets are poorly understood.

During shallow subaqueous to emergent eruption phases (e.g. Surtla), mostly subaqueous pyroclastic density currents carry fragmented pyroclastic particles away from the vent (WHITE 1996, 2000). However, after explosions breach the water surface, aerial transport begins to dominate; that is, dense eruption columns are formed that generate fallout-dominated deposits as well as the deposits of horizontal base surges (LORENZ 1974a, b). The ratio between fall versus base surge dominated deposits is very variable in the resulting tephra cone, as documented from Surtsey (LORENZ 1974a). As a result of continued deposition a tephra cone grows above water level, commonly forming one or a few crescent-shaped or sub-circular islands (SOHN and CHOUGH 1992, WHITE 2001). These islands are very fragile, being made up of loose tephra deposits, and are strongly at the mercy of wave action. In most cases they are short-lived, as in the case of Graham Island, which formed in 1831 just south of Sicily, sparking a three-way international dispute over its ownership before it disappeared beneath the waves eight months later. In rare cases these islands can be efficiently armoured by solid rock if the eruption becomes "dry" and a lava fountain forms in the last phase of the sequence, leading to the formation of a small lava shield in the crater, as at Surtsey (THORARINSSON 1965, 1967, LORENZ 1974a). It has also been reported that the immediate and ongoing palagonitisation of the volcanic glass shards in the fine tephra could form hard and impermeable beds in the ejecta construct (THORARINSSON 1965, JAKOBSSON 1972).

### Deep subaqueous explosive volcanism

The range of processes responsible for generating explosive eruptions in deep water and transporting and depositing eruption products are poorly understood. The very limited access to sites on active volcanoes and the problem of interpreting stratigraphical relationships between volcanic and non-volcanic rock units inferred to form in deep subaqueous environments hinder our understanding of this type of volcanism (WHITE et al. 2003).

In deep subaqueous environments explosive eruptions driven by magmatic gases can occur (HEAD and WILSON 2003). Hawaiian- and Strombolian-style eruptions are predicted in deep water, and their eruption products are inferred (HEAD and WILSON 2003). The basic idea of such eruptions take place is that magmatic volatiles can accumulate prior an eruption, and then the over-pressurized magma can fragment in deep water in a similar way that it would in subaerial conditions (HEAD and WILSON 2003). Young deposits interpreted to be result of such deep water hawaiian or strombolian eruptions have now been identified in many places in the sea floor in water depths ranging from 1000 to 4300 metres (CLAGUE et al. 2003). Because this type of eruption in deep subaqueous environments produces only low eruption columns, only very fine pyroclasts are expected to be transported and deposited extensively on the sea/lake floor, such as "limu o Pele" (limu shells) (MAICHER and WHITE 2001, CLAGUE et al. 2003). Many authors suggest that these deep water eruptions are dominantly driven by magmatic fragmentation, combined with varying degrees of fragmentation driven by magma–water interaction (BATIZA and WHITE 2000). Because steam can theoretically only be produced at water depths shallower than the critical depth of sea water (about 3 km), it is expected that fragmentation by explosive magma–water interaction will be of importance only at depths of 3 km or less.

In the case of acidic magmas, eruption of highly vesicular pyroclasts forms pumice, which is capable in floating in water until it becomes water-saturated and sinks as its density exceeds that of water. In addition to explosive eruptions, pumice can form in association with subaerial and subaqueous lava flows (MCPHIE and ALLEN 1992, MANLEY 1996, AKAY and ERDOGAN 2001). However, pumice-rich deposits formed in association with subaqueous lava flows can be carried away by currents, and may be deposited far from their source lava flow. In this instance, identification of pumice deposited by sinking of clasts in pumice rafts in subaqueous settings is not unambiguous evidence for nearby explosive subaqueous eruptions (WHITE et al. 2003). Pumice-forming subaqueous eruptions are divided into four categories (Figure 8.10) (KANO 2003); 1) subaqueous Plinian-types, 2) those generating subaqueous flows, 3) those involving explosive

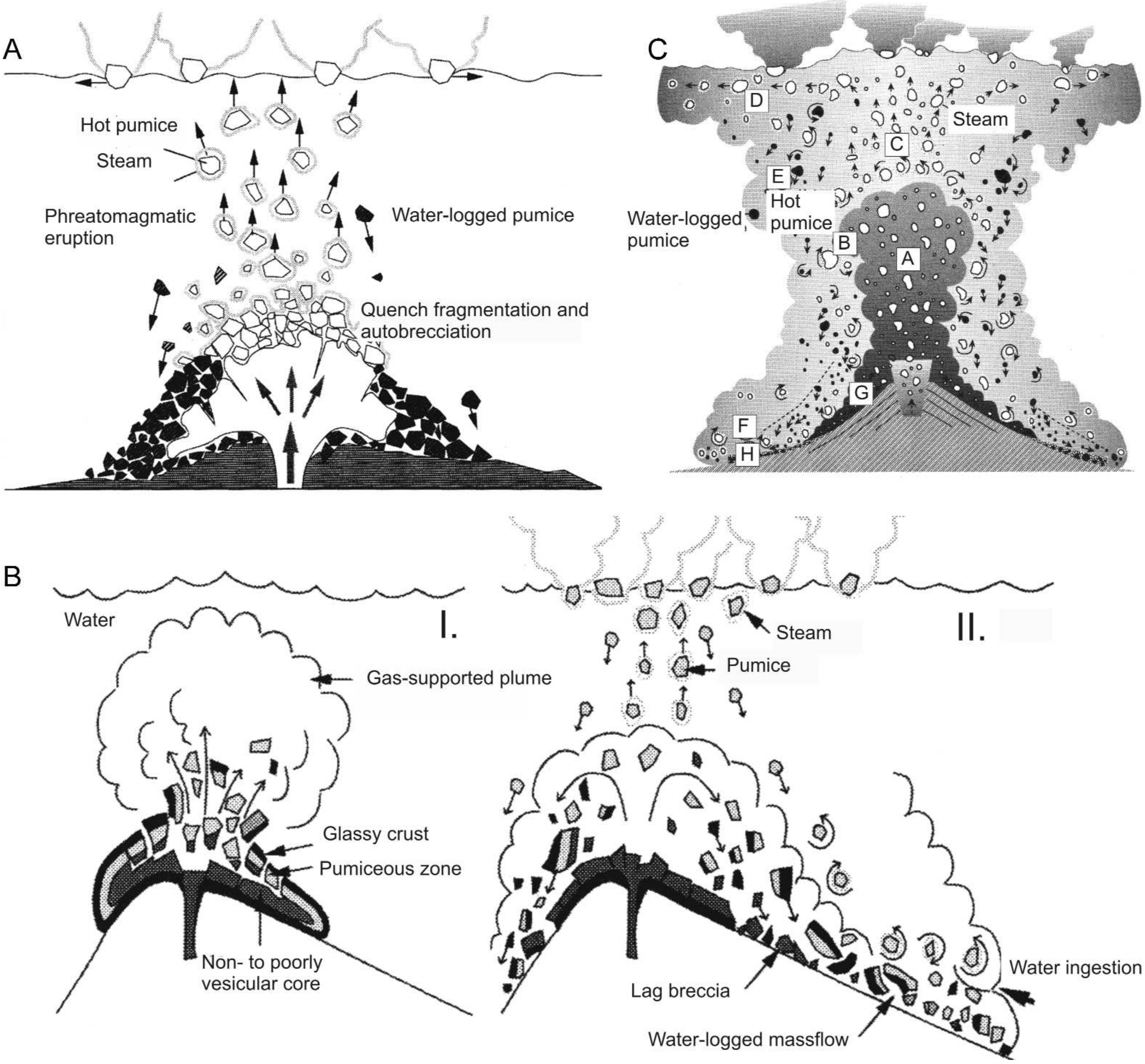


***Figure 8.10.*** Summary cartoons of the possible ways to form pumiceous pyroclasts during subaqueous eruptions (after Kano 2003: A) p. 225, fig. 16, B) p. 224, fig. 14, C) p. 223, fig. 13). A) Quench and mechanical fragmentation in the margin of a subaqueous pumiceous dome. Hot pumice blocks ascend to the surface and after drifting settle on the seafloor. Especially larger pumice blocks can fragment further upon contact with sea water. Fragmentation can be enhanced by phreatomagmatic eruptions from the interior of the pumiceous dome as well. B) Subaqueous Vulcanian-style eruption can disrupt the top of a pumiceous dome (I) and generate coarse pumice rich mass flows on the flank of the dome (II). C) A general model to show subaqueous flow-generating processes. “A” represents the gas-supported eruption plume. “B” is the zone where mixing of eruption plume and water takes place. “C” is the zone where buoyancy carries pumice and ash from the convective plume. “D” is the zone where fine pyroclasts transported in suspension. “E” is the region where water-logged pumice and dense lithics fall back. “F” represents gas- and water-logged density currents. “G” is the region where gas-supported hot subaqueous pyroclastic flows operate. “H” is the accumulation zone of pumice. D) Hot pumices are charged with gases and buoyant. Until they are hot, steam envelop prevents pumice clasts from hydroclastic fragmentation, rapid cooling and water-entrance to the vesicles

bulk interaction of vesiculating magma and water and 4) those ones generating pumice during non-explosive processes. Welding of subaqueous pumice deposits is under debate, and various explanations and conditions are inferred to explain their generation (Sparks et al. 1980, Cas and Wright 1991, Kokelaar and Busby 1992, Kokelaar and Koniger 2000, Simpson and McPhie 2001).

An increasing number of subaqueous calderas have been identified on the seafloor recently on the basis of high resolution bathymetry (Cas 1992, Wright and Gamble 1999, Yuasa and Kano 2003). These features have significant economic value because of their potentially extensive mineralization. Their proximal deposits in particular are potentially of economic significance. Since these identified calderas are large (tens of km across) features, it is expected that extensive distal pyroclastic successions may be associated with them as well.

## Subglacial volcanism

Subglacial eruptions refer to those eruptions where the erupting vent is located under ice cover (Figure 8.11) (Smellie 2000). This could occur in any type of glaciated area, such as high mountains or close to the poles where permanent ice sheets are present (Smellie 2000). Volcanoes formed in englacial settings share many features with volcanoes formed in other subaqueous settings (e.g. in lacustrine environments). This is because the eruption can produce large volumes of melt water that acts as a subaqueous environment below the ice sheet (Smellie and Skilling 1994, Tuffen et al. 2001, Sigvaldason 2002, Gudmundsson 2003, Schopka et al. 2006, Smellie 2006, Stevenson et al. 2006). Eruptions under ice can generate large volumes of hyaloclastite and melt water, which can initiate floods termed jökulhlaup (glacier outburst floods). Jökulhlaup are devastating floods capable of depositing large volumes of sediment and causing significant landscape modification (Carey et al. 2000, Alho 2003, Bjornsson 2003, Carrivick et al. 2004, Roberts 2005, Evatt et al. 2006, Rushmer 2006, Russell et al. 2006, Smith et al. 2006). Their power is well demonstrated in many Icelandic subglacial eruptions, such as the October 1996 eruption beneath the Vatnajökull glacier (Gudmundsson et al. 1997, Maria et al. 2000, Gudmundsson et al. 2004, Stefansdottir and Gislason 2005).

Eruptions beneath glaciers are largely controlled by the thickness of the ice cover, e.g. thin (100–150 m) or thick (over 150 m) (Smellie and Skilling 1994, Smellie 2000). The classification of the resulting volcanic structure is commonly based on the relative proportion of volcanic, volcaniclastic, and sedimentary lithologies present, perhaps combined with their 3D architecture (Figure 8.11) (Smellie 2000).

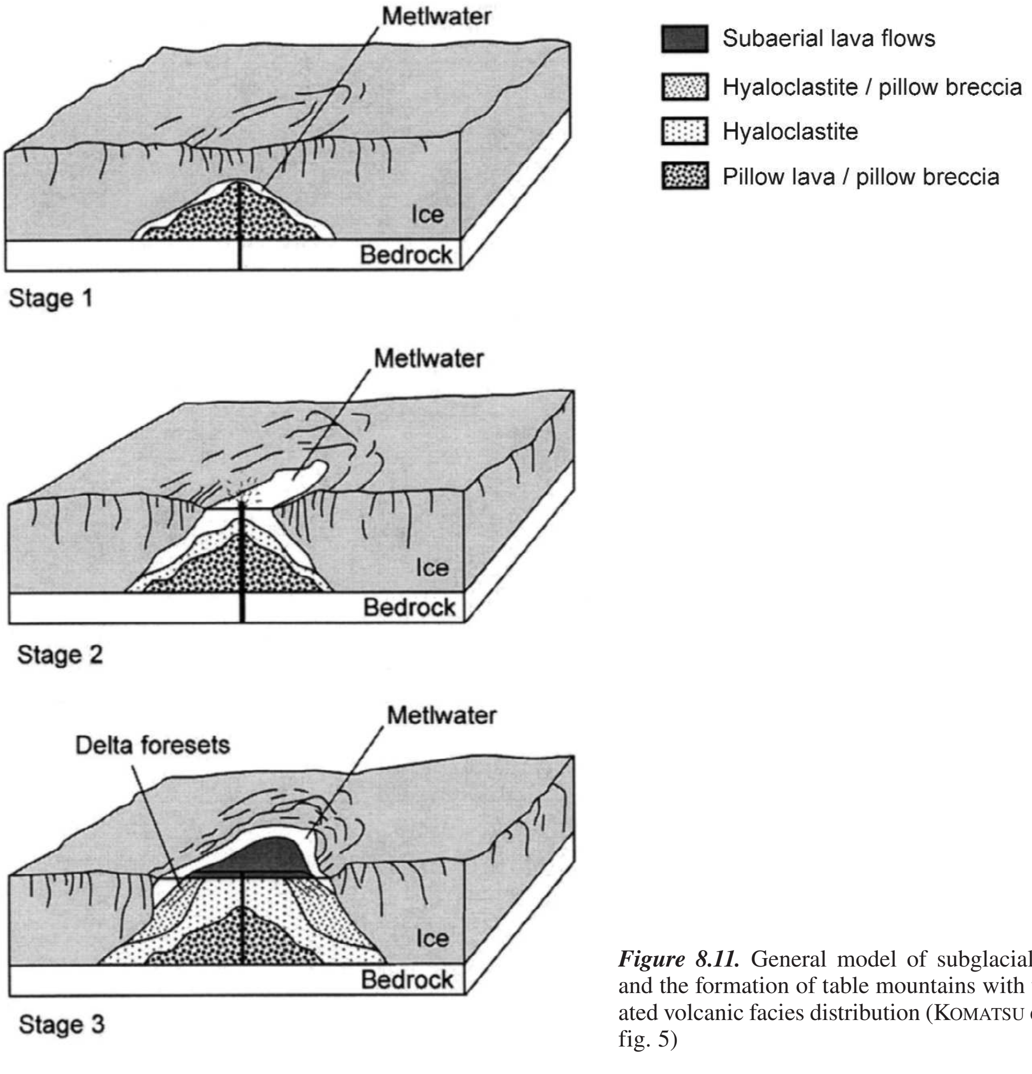


*Figure 8.11.* General model of subglacial eruptions and the formation of table mountains with the associated volcanic facies distribution (Komatsu et al. 2007: fig. 5)

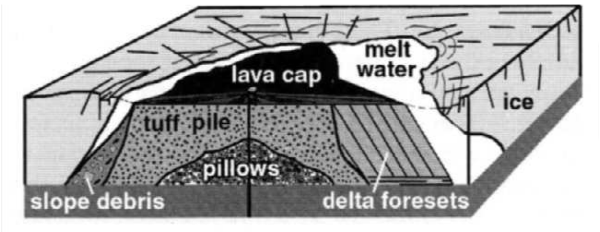


*Figure 8.12.* A model architecture of a tuya (CHAPMAN 2002: p. 278, fig. 5)

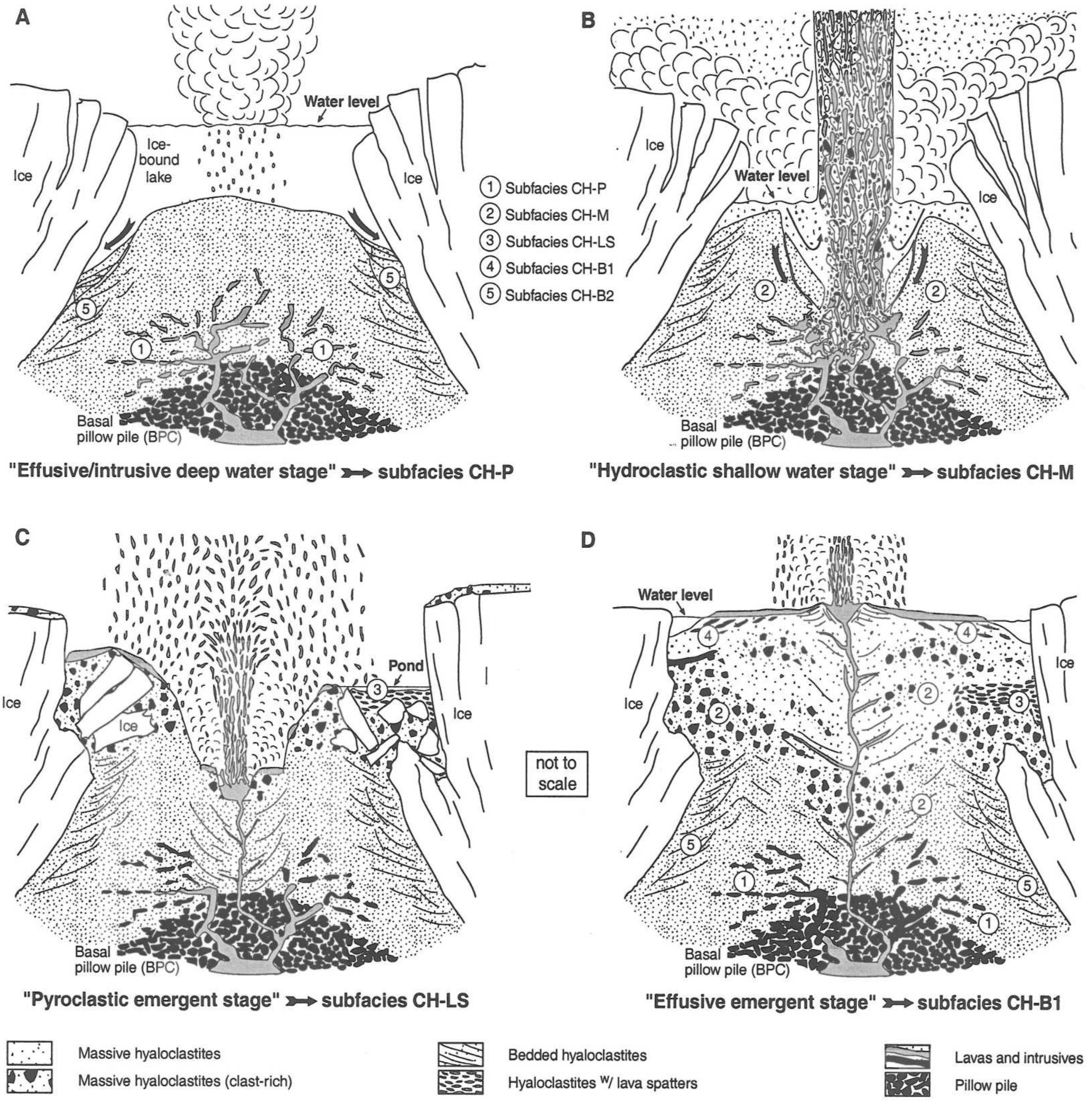


*Figure 8.13.* Theoretical model illustrating the environmentally controlled eruptive processes of an active englacial volcano in a narrow ice-bound lake (after WERNER and SCHMINCKE 1999: p. 350, fig. 10A–D). A) Effusive/intrusive deep water eruption and accumulation of pillow lavas. B) Hydroclastic fragmentation and shallow water, Surtseyan-style eruption. C) Emergent stage, more magmatic fragmentation (Strombolian- and/or Hawaiian-style eruption). D) Effusive emergent stage producing lava flows over the volcanic edifice

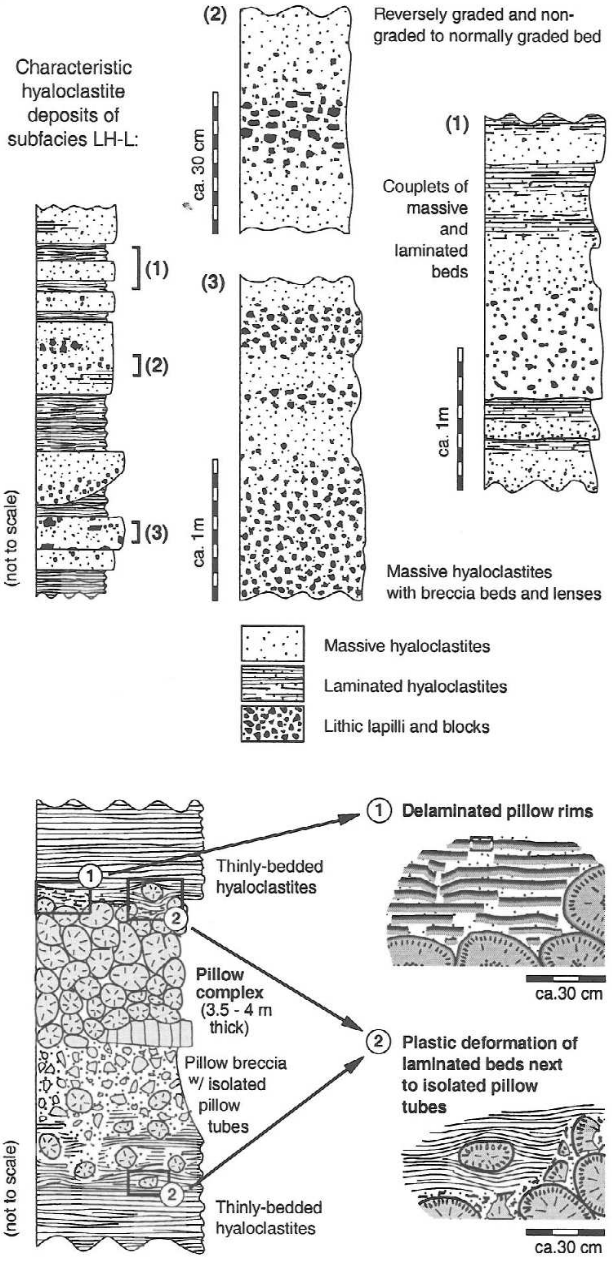


***Figure 8.14.*** Schematic columnar sections of a subglacial volcano described from Iceland showing vertical volcanic facies relationships after Werner and Schmincke 1999: p. 344, fig. 6B–C]

In the case of thin ice cover, the eruption quickly generates a cavity (vault) in the ice and melt water escapes through a complex network of tunnels generated by the heat of the magma (Smellie 2000). The eruption site quickly becomes open to air and phreatomagmatic eruptions triggered by magma–water interaction often forms a tuff cone in the newly formed hole in the glacier. The vent site of this new phreatomagmatic volcano is quickly sealed off from external water, and eruptions turn toward 'drier'lava fountaining hawaiian- or mildly explosive Strombolian-style events. In the final stages the heat of the newly formed volcano may melt more ice at the margins of the ice hole, and melt water can be dammed between the newly formed cone and the ice vault walls. Subsequent eruptions pour lava flows into this meltwater lake, forming lava deltas and associated hyaloclastite units.

During eruptions through thicker ice piles, the initially formed melt water can be sealed and locked into the newly formed cavity over the vent, creating prefect subaqueous conditions for the subsequent eruption(s) (Smellie 2000). This condition can be reached especially in those glaciers where the upper permeable layer of the glacier is thick. In this case the melt water depth is not sufficient to lift up the ice cover by floating, and therefore melt water cannot leave the eruption site other than by slow escape along thermally eroded channels. When lava effusion takes place, lava can enter the englacial lakes and accumulate thick piles of hyaloclastite, preserved as table mountains (Figure 8.12) (Werner et al. 1996, Werner and Schmincke 1999) (*stapi* in Icelandic, or *tuyas* in British Columbia) after the ice cover disappears. Before the growing volcanic pile can emerge and become fully subaerial a palagonite tuff and tuff breccia composed edifice develops, commonly referred to as *tindar* (Smellie 2000). Explosive eruptions that take place in the trapped melt water are typical of those in shallow subaquous environments (Figure 8.13). Their eruption products are also similar to the deposits of Surtseyan eruptions (Figure 8.14).

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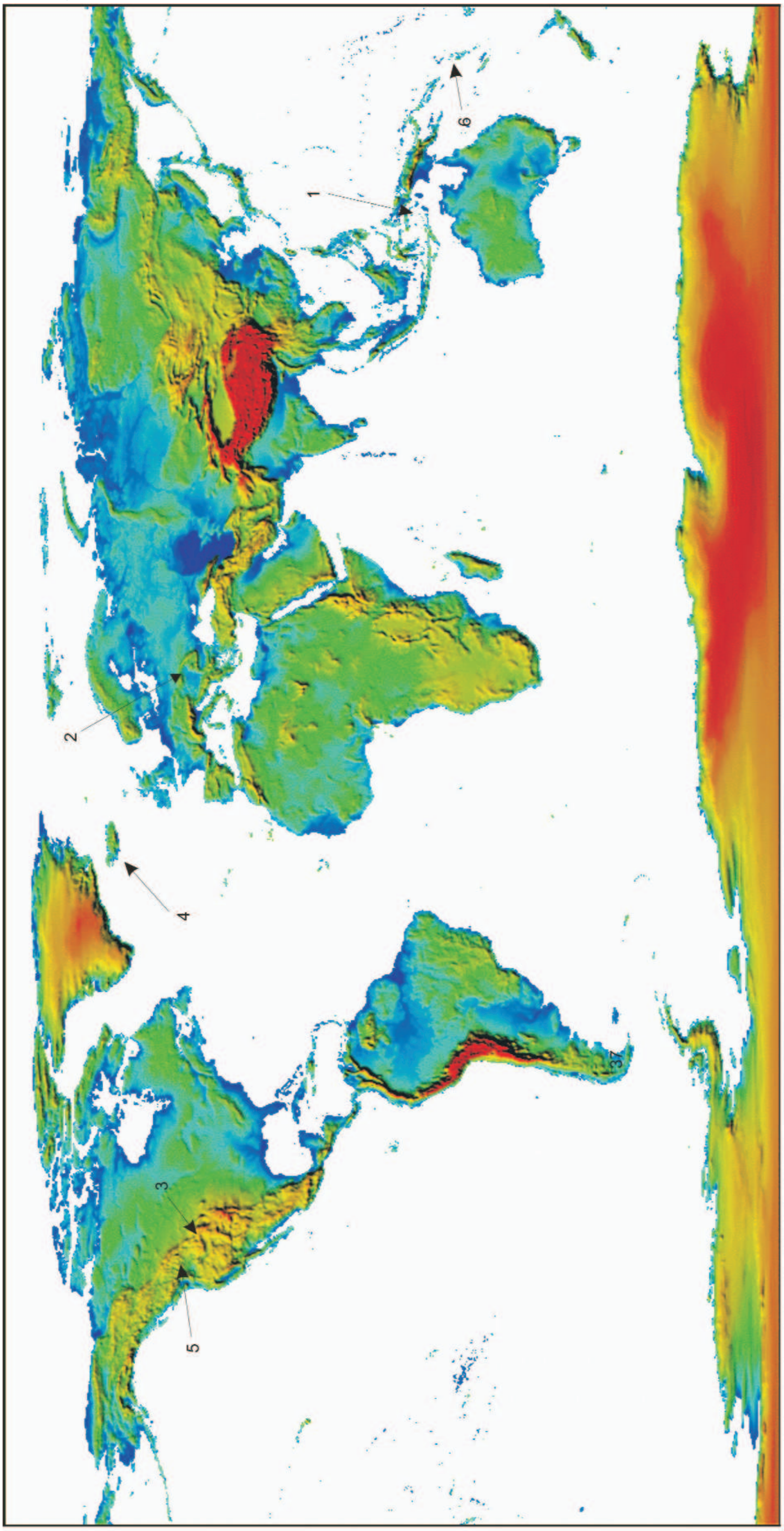
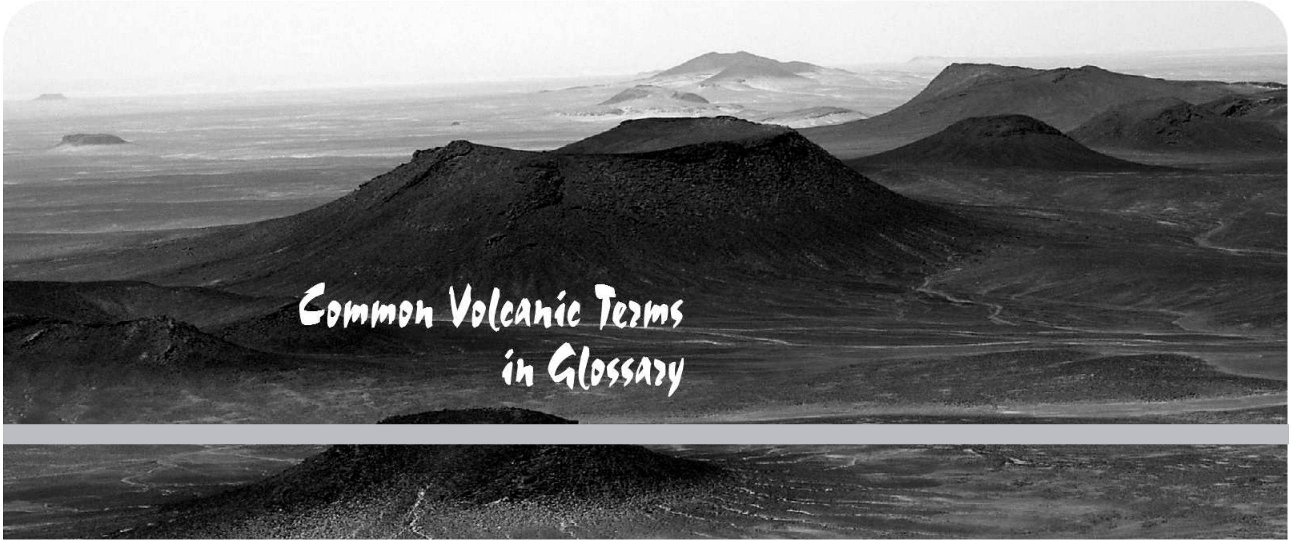
## Location map



1 — Ontong – Java Plateau
2 — Tokaj Mts, Hungary
3 — Pahvant Butte, Utah, USA
4 — Surtsey Island, Iceland
5 — Western Snake River Plain, Idaho, USA
6 — Ambae, Vanuatu

## Plate I

1. Rhyolitic hyaloclastite unit with flow banded, perlitic obsidian clasts from the Pálháza (Tokaj Mts, NE Hungary) Miocene subaqueous cryptodome/dome complex.
2. Overview of a coherent rhyolite body of the Pálháza subaqueous cryptodome/dome complex with complex network of coherent rhyolitic intrusive bodies and associated hyaloclastite successions.
3. Pillow lava cross section from a sublacustrine lava flow from the Western Snake River Plain, Idaho, USA. Note the typical radial joint pattern of the pillow lava and the brownish rim composed of palagonitized glassy lava.
4. Overview of the Pálháza hyaloclastite succession surrounding irregular shape coherent rhyolite intrusive/extrusive bodies.
5. A tuff cone formed in the Lake Voui caldera lake on Ambae Island, Vanuatu in December 2005. Dark tephra jets erupt from the vent.



# Common Volcanic Terms in Glossary

**'A'a:** Hawaiian word used to describe a lava flow whose surface is broken into rough angular fragments.

**Accessory:** A mineral whose presence in a rock is not essential to the proper classification of the rock.

**Accessory fragment:** A lithic fragment composed of country rock that has been explosively ejected during an eruption (Cas and Wright 1987: p. 54). Accessory fragments within pyroclastic deposits may be difficult to distinguish from accidental fragments. In general terms, referred to as a xenolith.

**Accidental:** Pyroclastic rocks that are formed from fragments of non-volcanic rocks or from volcanic rocks not related to the erupting volcano.

**Accidental fragment:** A clast picked up locally by pyroclastic flows and surges (Cas and Wright 1987: p. 54). Accidental fragments may be difficult to distinguish from accessory fragments. In general terms, referred to as a xenolith. Accidental rock fragments in maar and tuffring deposits provide additional important information: (1) their maximum size allows estimation of explosion energy; (2) the type of crustal rocks present permits inferences about explosion depths if crustal stratigraphy is known; and (3) mantle-derived ultramafic xenoliths, are common in some maar deposits.

**Acid:** A descriptive term applied to igneous rocks with more than 60% silica ($SiO_2$).

**Accretionary lapilli:** Accretionary lapilli are mud balls which result from a wet nucleus falling through a volcanic ash cloud. They flatten on striking the ground or may roll on loose ash and grow like a snowball. Spherical aggregates (commonly with a concentric structure) formed by the accretion of moist ash in eruption clouds (White and Houghton 2000: p. 495). Also used for all ash aggregates, including mud lumps (Houghton et al. 2000: p. 513).

**Achnelith:** A type of juvenile fragment characterized by smooth, glassy molded surfaces formed from lava spray from extremely fluid mafic eruptions (Walker and Croasdale 1972).

**Agglomerate:** Agglomerate is a mix of volcanic material which has been solodified into a rock. A course, pyroclastic deposit composed of a large proportion of fluidal-shaped volcanic bombs that are formed, in the strictest sense, by a fall deposit in the immediate vicinity of a volcanic vent. It is best applied to describe bomb and scoria deposits that build strombolian cones, and should never be used as a non-generic term for a "volcanic breccia" (Cas and Wright 1987: p. 359).

**Agglutinate:** A pyroclastic deposit consisting of an accumulation of originally plastic ejecta and formed by the coherence of the fragments upon solidification.

**Agglutination:** Instantaneous flattening of hot, soft pyroclasts upon landing. The resultant deposit is an agglutinate or spatter pile; particle outline is in part retained (Wolff and Sumner 2000: p. 321).

**Airfall:** Volcanic ash that has fallen through the air from an eruption cloud. Airfall deposits are characteristically well-sorted and well-layered, and typically exhibit mantle bedding (Cas and Wright 1987).

**Alert level:** Alert level is a measure of the current status of the volcano.

**Amygdaloidal:** A volcanic texture comprising vesicles (rounded holes resulting when magma cools around gas bubbles) which have been subsequently filled by secondary minerals.

**Amygdule:** An individual vesicle which has been subsequently filled-in by secondary minerals.

**Annular flow:** One of four two-phase flow regimes, in which magma lines the conduit walls and gas flows in a central jet (Jaupart 2000: p. 237).

**Armoured lapilli:** A type of accretionary lapilli composed of a crystal, pumice, or lithic fragment core which is surrounded by a rim of fine to coarse ash (McPhie et al. 1993: p. 29).

**Ash:** A textural term for volcanic fragments less than 2 mm in diameter (Fisher 1966; Schmid 1981). Ash is the typical product of explosive volcanic eruptions. Measuring less than 1/10 inch in diameter, ash may be either solid or molten when first erupted. By far the most common variety is vitric ash (glassy particles formed by gas bubbles bursting through liquid magma).

**Ash cloud:** A cloud of ash produced during pyroclastic eruptions. These clouds can result from rapid rising of the hot, buoyant ash-rich eruptive plume, or can be derived by elutriation at the top of a pyroclastic flow (Cas and Wright 1987).

**Ash flow:** A type of pyroclastic flow comprising dominantly ash-sized particles. Hot ash flows may be called "glowing avalanches" or "nuee ardentes", and if their volume is large enough, may eventually form deposits known as welded tuffs. These types of flows are extremely dangerous and historically have killed hundreds of thousands of people. A turbulent mixture of gas and rock fragments, most of which are ash-sized particles, ejected violently from a crater or fissure. The mass of pyroclastics is normally of very high temperature and moves rapidly down the slopes or even along a level surface.

**Ashfall (Airfall):** Volcanic ash that has fallen through the air from an eruption cloud. A deposit so formed is usually well sorted and layered.

**Atmospheric shock wave:** A strong compressional shock wave caused by a combination of volcanic ejecta and sonic waves.

**Autoclastic volcanic breccias:** Result from internal processes acting during movement of semisolid or solid lava; they include flow breccia and intrusion breccia (FISHER 1960).

**Avalanche:** A large mass of material or mixtures of material falling or sliding rapidly under the force of gravity. Avalanches often are classified by their content, such as snow, ice, soil, or rock avalanches. A mixture of these materials is a debris avalanche.

**Ballistic projectiles:** Ballistic projectiles are pieces of rock thrown from a volcanic vent in an eruption. They are generally confined to less than 3 kilometres radius from the vent because of their size.

**Base surge:** Base surges are ground hugging, outward moving clouds of gas and ash. They result from water–magma interactions. A turbulent, low-density cloud of rock debris, water, and/or steam that moves over the ground surface at extremely high speeds. Base surges are commonly the result of directed volcanic explosions. Base surge deposits are commonly composed of cross-bedded deposits comprising ash and lapilli. During 1965 Taal volcano (Philippines) erupted and base surges travelled 4 km and killed 189 people. Base surges were first identified during ocean nuclear weapons explosions in the Pacific.

**Bedding sags:** Also known as "bomb sags" (WENTWORTH 1926), form by the impact of ballistically-ejected bombs, blocks and lapilli into beds capable of being plastically deformed. They are characteristic of hydroclastic deposits and have been described from the deposits of many maar volcanoes, tuff rings and tuff cones. Beds beneath the fragments may be completely penetrated, dragged down and thinned, fold or show micro-faulting (HEIKE ~1971). Deformation is commonly asymmetrical, showing the angle and direction of impact if three-dimensional exposures are available.

**Bed forms:** The surface configuration of a bed (VALENTINE and FISHER 2000: p. 571).

**Bed forms by base surges:** Bed forms occur as three main kinds — sandwave, massive and planar (plane parallel) beds (SCHMINCKE et al. 1973; SHERIDAN and UPDIKE 1975), and are grouped into three facies types (WOHLETZ and SHERIDAN 1979) related to a fluidization model of transport and deposition. FISHER and WATERS (1970), FISHER and CROWE (1973) and SCHMINCKE et al. (1973) have emphasized bed forms in terms of the flow regime concept These different approaches are treated separately although they are not mutually exclusive.

**Bedset:** A sequence of beds with distinct internal structures, textures, colours or compositions that sets them apart from other sequences, usually bounded by unconformities, or by fallout layers (VALENTINE and FISHER 2000: p. 571).

**Bench:** The unstable, newly-formed front of a lava delta.

**Blister:** A swelling of the crust of a lava flow formed by the puffing-up of gas or vapor beneath the flow. Blisters are about 1 meter in diameter and hollow.

**Block:** Angular chunk of solid rock ejected during an eruption. Fragments of solid rock greater than 64 millimeters in diameter that are ejected during volcanic eruptions. Blocks are commonly composed of accessory fragments made up of crystallized magma associated with the eruption (e.g. pieces of a lava dome).

**Block-and-ash flow deposit:** Small-volume pyroclastic flow deposit characterized by a large fraction of dense to moderately vesicular juvenile blocks in a medium to coarse ash matrix of the same composition (FREUNDT et al. 2000: p. 581).

**Blocky lava:** Lava flows that are characterized by highly fractured surfaces which contain fragments of debris (usually flow fragments) up to several metres in diameter. The size of the surface fragments in blocky lavas is controlled by the rheology of the lava in the interior of the flow (KILBURN 2000: p. 291).

**Bomb:** Fragment of molten or semi-molten rock, 2 1/2 inches to many feet in diameter, which is blown out during an eruption. Because of their plastic condition, bombs are often modified in shape during their flight or upon impact.

**Bubbly flow:** One of four two-phase flow regimes, in which the gas phase appears as bubbles suspended in a continuous magma phase (JAUPART 2000: p. 237).

**Bulking:** The erosion and incorporation of secondary, exotic debris by lahars as they move downstream (VALLANCE 2000: p. 601)

**Caldera:** The Spanish word for cauldron, a basin-shaped volcanic depression; by definition, at least a mile in diameter. Such large depressions are typically formed by the subsidence of volcanoes. Crater Lake occupies the best-known caldera in the Cascades.

**Caldera cycle:** A commonly observed evolutionary sequence recognized in many caldera complexes. From oldest to youngest, the seven stages of the caldera cycle are: 1) regional tumescence and generation of ring fractures; 2) ignimbrite (pyroclastic) eruption(s); 3) caldera collapse; 4) pre-resurgent volcanism and intra-caldera sedimentation; 5) resurgent doming; 6) major ring fracture volcanism; and 7) terminal fumarolic and/or hot spring activity.

**Capping stage:** Refers to a stage in the evolution of a typical Hawaiian volcano during which alkalic, basalt, and related rocks build a steeply, sloping cap on the main shield of the volcano. Eruptions are less frequent, but more explosive. The summit caldera may be buried.

**Cinder cone:** A volcanic cone built entirely of loose fragmented material (pyroclastics).

**Clastogenic lava:** A lava flow formed by the rheomorphic flow of coalesced and agglutinated hot pyroclasts, typically fed by a lava fountain (WOLFF and SUMNER 2000: p. 321).

**Coalescence:** The process by which hot fluidal pyroclasts form a homogeneous liquid in which the particle outlines are obliterated (WOLFF and SUMNER 2000: p. 321).

**Cognate lithic fragment:** Non-vesiculated juvenile magmatic fragments that have silicified from the erupting magma (CAS and WRIGHT 1987: p. 54).

**Column collapse:** Column collapse is caused by an eruption column reaching a critical level then collapsing under its own weight. This forms the most dangerous of volcanic processes — pyroclastic flows and surges.

**Columnar jointing:** A type of fracture pattern resulting from the thermal contraction of hot volcanic rocks after their crystallization which commonly is expressed in elongate, pentagonal or hexagonal columns oriented perpendicular to the cooling surface. Columnar jointing is common in all compositions of lava flows, although it is generally best developed in mafic (basalt) lava flows and in felsic welded tuffs.

**Continuous uprush:** A style of explosive eruption in shoaling volcanoes, which combine into a continuous uprush and are characterized by a tall non-spreading eruption column (SMELLIE 2000: p. 403).

**Cooling unit:** A group of hot pyroclastic deposits (ignimbrites) that cools at more or less the same time. A deposit from a single eruption that shows simple variations in the degree of welding is known as a simple cooling unit. When many ignimbrites occur over an extremely short period of time, each individual ignimbrite may be deposited, and start to weld over a previous deposit or group of deposits that are cooling and undergoing welding. The resulting deposits have several zones of partial and dense welding, and since they more or less cool together, are known as compound cooling units (CAS and WRIGHT 1987: p. 253–255).

**Coulée:** A type of rhyolite lava flow that forms when lava issues from one side of a volcanic vent and produces a lava flow which is elongate in plan view (Cas and Wright 1987: p. 81).

**Composite volcano:** Relatively large, long-lived constructional volcanic edifice, comprising lava and volcaniclastic products erupted from one or more vents, and their recycled equivalents.

**Compound volcano:** Volcanic massif formed from coalesced products of multiple, closely spaced, vents.

**Conduit:** A passage followed by magma in a volcano.

**Country rocks:** The rock intruded by and surrounding an igneous intrusion.

**Crater:** A steep-sided, usually circular depression formed by either explosion or collapse at a volcanic vent.

**Cryptodome:** A cryptodome is a mound caused by the accumulation of viscous magma just beneath the surface.

**Cupola (Water dome):** A water cupola is a dome formation on the surface of the water just before an underwater eruption breaks through. The dome may rise to form a cylinder. Cupolas have been observed with a base of 100-200 m and height of 26 m.

**Curie point:** The temperature at which a body loses (by heating) or preserves (by cooling) its permanent magnetization. As rocks cool, the electromagnetic field aligns magnetic minerals in the magma, and their orientation is preserved as the rocks cool below the Curie point.

**Curtain of fire:** A row of coalescing lava fountains along a fissure; a typical feature of a Hawaiian-type eruption.

**Debris avalanche:** A rapid and unusually sudden sliding or flowage of unsorted masses of rock and other material. Catastrophic landsliding of gravitationally unstable volcano flanks resulting in a widely dispersed deposit at the foot of the edifice, typically characterized by a hummocky surface. As applied to the major avalanche involved in the eruption of Mount St. Helens, a rapid mass movement that included fragmented cold and hot volcanic rock, water, snow, glacier ice, trees, and some hot pyroclastic material. Most of the May 18, 1980 deposits in the upper valley of the North Fork Toutle River and in the vicinity of Spirit Lake are from the debris avalanche. Debris avalanches differ from debris flows in that they are not water saturated and in that the load is entirely supported by particle-particle interaction (Vallance 2000: p. 601).

**Debris flow:** A mixture of water-saturated rock debris that flows downslope under the force of gravity (also called lahar or mudflow). A type of mass flow comprising a dense, cohesive, flowing mixture of sediment (mud through boulder sized materials, generally >50% by volume), water, and commonly, organic debris. Debris flows generally move downslope in laminar fashion due to the force of gravity (Vallance 2000: p. 601; Carey 2000: p. 627). Debris flows generated at volcanoes are commonly referred to as lahars. A uniform mixture of solid and liquid phases in vertical profiles characterizes debris flows and distinguishing them from more water-rich hyperconcentrated flows (Vallance 2000: p. 601).

**Debris fall:** A debris fall is the near free fall of debris from an overhang or vertical face.

**Debulking:** A process in which the lahar selectively deposits certain particles, owing to their size or density, as it moves downstream (Vallance 2000: p. 601).

**Decompressive melting:** Melting that occurs when rocks undergo a decrease in pressure. This commonly occurs in the vicinity of hot spots as mantle rocks rise to shallower levels in the earth due to convective rise and upwelling (Sigurdsson 2000: p. 15). Melting occurs as a result of decreasing pressure, not increasing temperature

**Destructiveness index:** the logarithm of the area covered by lava, pyroclastic flows, and surges, or buried under more than 100 $kg/m^2$ of tephra during an eruption (Pyle 2000: p. 263).

**Devitrification:** The solid-state transformation of volcanic glass into crystalline materials. Devitrification tends to be more prevalent in densely-welded tuffs, but may also occur in less densely-welded or unwelded pyroclastic and/or volcaniclastic deposits. The main products of devitrification are cristobalite ($SiO_2$) and alkali feldspar ($KAlSi_3O_8$) (Cas and Wright 1987: p. 258).

**Diatreme:** A funnel-shaped, pipe-like volcanic conduit, usually filled with volcaniclastic debris, emplaced by the explosive energy of gas-charged magmas. Diatremes are believed to result from hydrovolcanic fragmentation and subsequent wall rock collapse (Vespermann and Schmincke 2000: p. 683), and may reach depths up to 2500 metres. Diamond-bearing diatremes are economically important and are referred to as kimberlite pipes.

**Dike:** A discordant, sheetlike body igneous body formed from the injection of magma into a fracture within the brittle crust of the earth (Carrigan 2000: p. 219; Marsh 2000: p. 191). Generally a tabular igneous body which cross-cuts the planar structures in the adjacent rocks.

**Directed blast:** A hot, low density mixture of gas, rock debris, and ash that is propelled by a volcanic eruption and generally moves along the ground at high speeds.

**Dispersed flow:** One of four two-phase flow regimes, in which magma takes the form of fragments in a continuous gas phase (Jaupart 2000: p. 237).

**Dispersal index (D):** A measure of the area covered by a pyroclastic fall deposit, specifically the area enclosed by an isopach draw at 1/100 of the maximum thickness of the deposit (Houghton et al. 2000: p. 513).

**Dome:** A steep-sided mass of lava that is generally formed immediately above the volcanic vent from which it was extruded. Domes are generally circular in plan and have a relatively small surface area relative to other types of lava flows. Domes may be spiny, rounded, or flat on top, and often have rough, blocky surfaces formed by the fragmentation of the dome's crust during intrusion. Domes may grow by extrusion of lava onto the outer surface of a previously formed dome (exogenous dome) or may be formed by inflation of a pre-existing dome (endogenous dome). Domes are most commonly the result of extrusion of viscous lava (primarily of the composition of rhyolite and dacite, but andesite may occur as well).

**Downsag caldera:** A type of caldera characterized by inward sloping topography, inward tilted wall rocks, and an apparent absence of large displacement caldera bounding faults (Lipman 1997). Downsag calderas are believed to result from small volume eruption from a deep-seated subvolcanic intrusion.

**Edifice (Volcanic):** A volcanic edifice is a constructional feature built from erupted material.

**Ejecta:** Material that is thrown out by a volcano, including pyroclastic material (tephra) and lava bombs.

**Elutriation:** Loss of small particles by an upward flow of gas through a deposit.

**Epiclastic volcanic breccias:** Result from transportation of loose volcanic material by epigenetic geomorphic agents, or by gravity, and include laharic breccia, water-laid volcanic breccia, and volcanic talus breccia (Fisher 1960).

**Episode:** An episode is a volcanic event that is distinguished by its duration or style.

**Eruption:** The process by which solid, liquid, and gaseous materials are ejected into the earth's atmosphere and onto the earth's surface by volcanic activity. Eruptions range from the quiet overflow of liquid rock to the tremendously violent expulsion of pyroclastics.

**Eruption cloud:** The column of gases, ash, and larger rock fragments rising from a crater or other vent. If it is of sufficient volume and velocity, this gaseous column may reach many miles into the stratosphere, where high winds will carry it long distances. A volcanic cloud is a convoluted rolling mass of water vapour and ash that is highly charged with electricity and overhangs a volcano during an eruption. The cloud is produced by a column of gases, ash, and rock emitted from a crater. Eruption clouds may reach great heights. The 1883 eruption of Krakatau produced an eruption column 50 km high. Eruption clouds may take on the shape of pine trees or cauliflowers.

**Eruption column:** An eruption column is the lower vertical part of the eruption cloud, where the ash and gases rise with great speed.

**Eruptive vent:** The opening through which volcanic material is emitted.

**Extrusion:** The emission of magmatic material at the earth's surface. Also, the structure or form produced by the process (e.g. a lava flow, volcanic dome, or certain pyroclastic rocks).

**Facies:** A part of a rock body that can be differentiated from another part of a related rock body by textural or compositional variations. The general appearance or composition of one part of a rock body as contrasted with other parts (AGI 1976: p. 155).

**Facies changes:** The textural and compositional changes that occur laterally and/or vertically within related rock bodies.

**Fall deposit:** Mantling blanket of pyroclastic particles (ash, scoria, pumice, etc) erupted explosively and transported through the atmosphere before falling back to the ground.

**Felsic:** An igneous rock having abundant light-colored minerals.

**Fire fountain:** See also: lava fountain

**Fissures:** Elongated fractures or cracks on the slopes of a volcano. Fissure eruptions typically produce liquid flows, but pyroclastics may also be ejected.

**Flank eruption:** An eruption from the side of a volcano (in contrast to a summit eruption).

**Flood basalt:** voluminous and thick sequences of mafic lavas erupted from fissures over relatively short periods of time (plateau basalts) (PEFIT and DAVIDSON 2000: p. 89).

**Flow banding:** A foliation commonly observed in intermediate and felsic lavas, that results from shearing of the lava during laminar flow (CAS and WRIGHT 1987: p. 78). In rhyolite flows, flow banding is commonly exhibited by alternating bands comprising volcanic glass and spherulites (small, radiating bodies of devitrified glass).

**Flow regime:** Hydraulic conditions of noncohesive flow of sand and silt that develop ripples, dunes, plane parallel beds, and antidunes. Progressive changes in bed forms occur as flow regime increases. Low-flow regime conditions form small scale ripples that progress to dunes; high flow regime conditions form plane beds and then antidunes (VALENTINE and FISHER 2000: p. 571).

**Flow transformation:** Reversable changes in sediment gravity flows between turbulent and steady flow related chiefly to particle concentration, thickness of flow, and flow velocity (VALENTINE and FISHER 2000: p. 571).

**Formation:** A body of rock identified by lithic characteristics and stratigraphic position and is mappable at the earth's surface or traceable in the subsurface.

**Fragmentation:** Fragmentation is a process whereby bursting bubbles tear apart magma as it reaches the surface. At a depth of 10 km gas filled magma is at a pressure of 3000 atmospheres. Therefore a large pressure drop occurs as the magma rises towards the surface through the conduits. Fragmentation also can be defined as a transition from a continuous melt with a dispersed gas phase to disconnected parcels of bubbly melt within a continuous gas phase (CASHMAN et al. 2000: p. 421).

**Fragmentation index:** a parameter measuring the grain size of a pyroclastic fall deposit, specifically the percentage of ash finer than 1 mm at the point on the dispersal axis to1/10 of the maximum thickness of the deposit (HOUGHTON et al. 2000: p. 513).

**Fuel-coolant interaction:** The interaction of magma (fuel) with external water (coolant) that may result in thermal explosions (VESPERMANN and SCHMINKE 2000: p. 683).

**Fumarole:** A vent which releases volcanic gases. These include steam ($H_2O$), carbon dioxide ($CO_2$), sulphur dioxide ($SO_2$), hydrogen sulphide ($H_2S$), as well as other volatile gases emitted from subterranean magmas.

**Geyser:** Most geysers are hot springs that episodically erupt fountains of scalding water and steam. Such eruptions occur as a consequence of groundwater being heated to its boiling temperature in a confined space (for example, a fracture or conduit). A slight decrease in pressure or an increase in temperature will cause some of the water to boil. The resulting steam forces overlying water up through the conduit and onto the ground. This loss of water further reduces pressure within the conduit system, and most of the remaining water suddenly converts to steam and erupts at the surface.

**Guyot:** A type of seamount that has a platform top. Named for a nineteenth-century Swiss-American geologist.

**Hawaiian eruption:** Hawaiian eruptions describe a style of eruption typically seen on shield volcanoes. Hawaiian eruptions involves lava fountains and lava flows. Commonly seen in Hawaii, Iceland, Reunion, Hawaiian eruptions are spectacular and the most photogenic of all eruption types.

**Heat transfer:** Movement of heat from one place to another.

**Horizontal blast:** An explosive eruption in which the resultant cloud of hot ash and other material moves laterally rather than upward.

**Hornito:** A small rootless spatter cone that forms on the surface of a basaltic lava flow (usually pahoehoe) is called a hornito. A hornito develops when lava is forced up through an opening in the cooled surface of a flow and then accumulates around the opening. Typically, hornitos are steep sided and form conspicuous pinnacles or stacks. They are "rootless" because they are fed by lava from the underlying flow instead of from a deeper magma conduit.

**Hot spot:** A volcanic center, 60 to 120 miles (100 to 200 km) across and persistent for at least a few tens of million of years, that is thought to be the surface expression of a persistent rising plume of hot mantle material. Hot spots are not linked to arcs and may not be associated with ocean ridges.

**Hot-spot volcanoes:** Volcanoes related to a persistent heat source in the mantle.

**Hummocks:** These are rounded or conical mounds within a volcanic landslide or debris avalanche deposit. Hummocks contain a wide range of rock debris, reflecting the variation of deposits that previously formed the flanks of the volcano. Some hummocks contain huge intact blocks tens to hundreds of metres in diameter. Some of the original layering of lava flows and other deposits can be seen in these large hummocks, but most of the large rock fragments are thoroughly shattered. In other hummocks the rock debris is thoroughly mixed as if the material had been in a blender and thoroughly mixed together.

**Hyaloclastite:** A deposit comprising small, angular glass fragments formed by nonexplosive shattering of lava or magma flowing into water, ice, or water-saturated sediment (BATIZA and WHITE 2000: p. 361; SCHMIDT and SCHMINCKE 2000: p. 383).

**Hydroclastic rocks:** a general term applied to volcanic rocks formed by fragmentation of magma in the presence of water (Batiza and White 2000: p. 361).

**Hydrovolcanic eruptions:** A general term for eruptions caused by the mixing of magma with water (Vespermann and Schminke 2000: p. 683). Encompasses hydroclastic, hydromagmatic, and phreatomagmatic eruptions.

**Hypabyssal:** A shallow intrusion of magma or the resulting solidified rock.

**Hyperconcentrated flow:** A gravitationally driven, nonuniform mixture of debris and water content larger than that of debris flow but less than that of muddy streamflow (Vallance 2000: p. 601).

**Ignimbrite:** Fiery raincloud rocks. The rock formed by the widespread deposition and consolidation of ash flows and Nuees Ardentes. The term was originally applied only to densely welded deposits but now includes non-welded deposits. Ignimbrite deposits are poorly sorted pyroclastic deposits consisting of glass shards, crystals and lithic fragments. Ignimbrites are formed by the deposition of hot, rapidly expanding, turbulent magmatic gases. Ignimbrites have a volume of 1 cubic km to 2000 cubic km. A term used for pyroclastic flow deposits, that is synonymous with “ash tuff” (Lipman 2000: p. 643). According to Cas and Wright (1987: p. 98), the term should only be used to describe pumiceous pyroclastic flow deposits.

**Isopach:** Line joining points of equal thickness of deposit (Houghton et al. 2000: p. 555).

**Isopleth:** Line joining points where the sizes of the largest clasts are the same (Houghton et al. 2000, p. 555).

**Jökulhlaup:** A Jökulhlaup is a glacial outburst caused by melt water from a subglacial volcano. Examples of Jökulhlaup Grimsvötn (Iceland) 1996, 1937.

**Jigsaw cracks:** These are characteristic joint patterns within a debris-avalanche block (Ui et al. 2000: p. 617).

**Juvenile fragment:** Glassy or partially crystallized fragments which represent samples of an erupting magma. These include fragments such as pumice, scoria, reticulate, achneliths (Pele’s tears, Pele’s hair), and various types of volcanic bombs (Cas and Wright 1987: pp. 47–53).

**Kimberlite:** Kimberlite is a rock formed by explosive eruptions in volcanic pipes. It is formed at depths of 100-200 km and pressures of 40,000 to 50,000 atm. Kimberlite is the main source of diamonds.

**Kipuka:** Is a Hawaiian term for an “island” of land completely surrounded by one or more younger lava flows. A kipuka forms when lava encircles a hill or a slight rise in the ground as it moves downslope or across relatively flat ground. Because they are surrounded by more recent flows, kipukas are often covered with mature vegetation.

**Laccolith:** A body of igneous rocks with a flat bottom and domed top. It is parallel to the layers above and below it.

**Lahar:** The Indonesian term for a debris flow or a mudflow originating on a volcano (Harris 2000: p. 1301). Lahars are generally composed of volcanic materials, but can contain significant amounts of non-volcanic materials derived from erosion during flow. Most volcanologists prefer this term to be used for the process and not the sedimentary deposits that it forms, but unfortunately, this distinction has been largely ignored in the geological literature. Many lahars are composed of sand and coarser materials, and thus, can be distinguished from “mudflows” which predominantly contain silt- or clay-sized grains (Rodolfo 2000: p. 973). A torrential flow of water-saturated volcanic debris down the slope of a volcano in response to gravity. Lahars are also referred to as volcanic mudflows or debris flows. They form in a variety of ways, chiefly by the rapid melting of snow and ice by pyroclastic flows, intense rainfall on loose volcanic rock deposits, breakout of a lake dammed by volcanic deposits, and as a consequence of debris avalanches.

**Laminar flow:** flow regime where viscous effects dominate and flow trajectories are parallel (Jaupart 2000: p. 237).

**Lapilli:** Literally, “little stones.” A textural term for fragments in volcanic rocks and volcanic deposits that range from 2 mm to 64 mm in diameter (Fisher 1966; Schmid 1981).

**Lateral blast:** A volcanic eruption which is directed horizontally instead of vertically. Lateral blasts may be caused by sudden decompression of a shallow magma chamber residing within the flanks of a volcano (for example, the 1980 eruption of Mt St. Helens), or along the base or side of a lava dome (for example, the 1902 eruption of Mt Pelee in Martinique) (Nakada 2000: p. 945).

**Lava:** Magma which has reached the surface through a volcanic eruption. The term is most commonly applied to streams of liquid rock that flow from a crater or fissure. It also refers to cooled and solidified rock. An outpouring of lava from a vent or fissure that spreads along the ground surface, as well as the crystallized rock resulting from solidification of the outpouring (Peterson and Tilling 2000: p. 957).

**Lava breaching:** Breaching is a term to describe a lava flow breakout. The conditions from breaching require a high channel pressure and more fluid lava in the channel compared to the edges. Aa lava is the most prone to breaching. Breaching usually occurs in the upper half of the lava flow. Therefore breaching tends to widen the lava flows rather than extend them length ways. Note: It is rare for breached lava flows to extend more than 50% more than the original flow length. Note: lava flows cool quicker on steep slopes and therefore stops quicker.

**Lava delta:** Lava entering the sea often builds a wide fan-shaped area of new land called a lava delta. Such new land is usually built on sloping layers of loose lava fragments and flows. On steep submarine slopes, these layers of debris are unstable and often lead to the sudden collapse of lava deltas into the sea.

**Lava dome:** Mass of lava, created by many individual flows, that has built a dome-shaped pile of lava.

**Lava flow:** An outpouring of lava onto the land surface from a vent or fissure. Also, a solidified tongue like or sheet-like body formed by outpouring lava.

**Lava fountain:** A rhythmic vertical fountain-like eruption of lava.

**Lava lake (Pond):** A lake of molten lava, usually basaltic, contained in a vent, crater, or broad depression of a shield volcano. Lava takes are large volumes of molten lava, usually basaltic, contained in a vent, crater, or broad depression. Scientists use the term to describe both lava lakes that are molten and those that are partly or completely solidified. Lava lakes can form (1) from one or more vents in a crater that erupts enough lava to partially fill the crater; (2) when lava pours into a crater or broad depression and partially fills the crater; and (3) a top a new vent that erupts lava continuously for a period of several weeks or more and slowly builds a crater higher and higher above the surrounding ground. A region typically within the summit of a shield volcano which contains partially crystallized or molten lava which lies immediately above a volcanic conduit which joins the lava lake to the magma chamber. Strong magma convection within volcanic conduits sustains lava lakes within their respective volcanic vents (Walker 2000: p. 285).

**Lava levee:** A lava levee is a boarder of lava blocks at the edge of a flow.

**Lava shields:** A shield volcano made of basaltic lava.

**Lava tube:** A tunnel formed when the surface of a lava flow cools and solidifies while the still-molten interior flows through and drains away.

**Limu:** Limu, or Limu o Pele (Hawaiian for “seaweed of Pele”), consists of thin flakes of basaltic glass that sometimes form when pahoehoe lava pours into the ocean. As waves wash atop exposed streams of lava, some water may become trapped and boil, resulting in delicate steam-filled bubbles of lava. Abrupt chilling and continued expansion of the delicate bubble walls form thin plates and shattered pieces of brownish-green to nearly-clear glass.

**Littoral cone:** A cone of lava fragments built on the surface of a lava flow pouring into a body of water, usually the sea, is called a littoral cone ("littoral" refers to a shoreline). Lava entering the ocean heats and boils seawater, often generating steam explosions that hurl tephra onto the shore, including spatter, bombs, blocks, ash, lapilli, and, rarely, limu. As the various tephra accumulates on the shoreline, a well-developed cone may be created.

**Lithophysae:** Radial aggregates of fibrous crystals which have formed around an expanding vesicle in a melt which is capable of flowing (Cas and Wright 1987: p. 84). Lithophysae are commonly the result of vapor-phase crystallization within a rhyolitic magma. They should not be confused with spherulites, which are similar-shaped structures formed from devitrification of volcanic glass.

**Lithic:** Fragments of previously-formed rocks or dense fragments that occur within volcaniclastic deposits. Lithic fragments may be accessory fragments, accidental fragments, or juvenile fragments.

**Littoral:** An adjective describing physical features or processes associated with shorelines of oceans, seas, or lakes (Peterson and Tilling 2000: p. 957).

**Lobate lava:** A submarine lava comprising elongate, flattish lobes with smooth, outer glassy skins (Batiza and White 2000: p. 361).

**Maar:** A maar is a low-relief, broad volcanic crater formed by shallow explosive eruptions. The explosions are usually caused by the heating and boiling of groundwater when magma invades the groundwater table. Maars often fill with water to form a lake. A maar is a type of tuff ring which has been affected by sagging so that it lies below the the level of the surrounding surface. A maar is often filled with water and surrounded by a rim of ejected material that was probably formed by explosive interaction of magma and groundwater. A type of monogenetic volcano, generally formed by subterranean phreatic or phreatomagmatic eruptions that occur as magma explosively interacts with ground water or subsurface moisture. Maar craters are cut into the surrounding country rock, vary from 10-500 metres deep, and range from a few hundred metres to 3 km in diameter. Maar volcanoes are generally surrounded by low, shallowly outward-dipping beds of well-bedded volcanic ejecta that rapidly decrease in thickness away from the vent. The volcanic deposits are mainly emplaced by base surges and fallout, and commonly contain very little (or in the case of phreatic eruptions, no) juvenile volcanic materials (Vespermann and Schmincke 2000: p. 685; Cas and Wright 1987: pp. 376–377). Examples of Maars Lake Nyos (Cameroon), Suoh (Sumatra, Indonesia), Karpinsky Group (Kurile Islands), Ukinrek Maars (Alaska).

**Mafic:** A compositional term for igneous rocks which contain 45–55% $SiO_2$ (by weight). Mafic rocks are generally dark coloured, and are characterized by mineralogy including pyroxene and calcium-rich plagioclase, variable amounts of olivine, and accessory minerals such as ilmenite and magnetite. Examples of mafic rocks include basalt and gabbro.

**Mantle bedding:** Pyroclastic deposits generated by ash fall which maintain a uniform thickness and drape over all but the steepest topography (Cas and Wright 1987: p. 96).

**Matrix:** The solid matter in which a fossil or crystal is embedded. Also, a binding substance (e.g. cement in concrete).

**Megabreccia:** Coarse, heterolithic breccia deposits formed during caldera collapse, which contain fragments which are generally greater than one metre in diameter (Lipman 1976). Megabreccia fragments may be so large that individual fragments may not be readily recognizable on the scale of an outcrop.

**Mesobreccia:** Heterolithic breccia deposits formed during caldera collapse which contain fragments that are generally less than 1 metre in diameter (Lipman 1976).

**Moat sediments:** A general term for sedimentary deposits that occur between the tographic walls and the resurgent central cores of the calderas. In felsic caldera systems, moat sediments are commonly intruded by, and associated with, lava domes.

**Monogenetic volcano:** A volcano that erupts only once (Walker 2000: p. 283).

**Mudflow:** A flowage of water-saturated earth material possessing a high degree of fluidity during movement. A less-saturated flowing mass is often called a debris flow. A flowing mixture composed of water and mud (clay- and silt-sized sediments). The term should be used exclusively for mud-dominated mass flows, and should not be used as a substitute for the term "lahar" (Rodolfo 2000: pp. 973–974). Mudflows are common in both volcanic and non-volcanic environments.

**Muddy streamflow:** A flow that essentially transports sediment as normal streams do, with fine-grained sediments in suspension and coarse-grained sediment moving piecemeal along the bed as bedload (Vallance 2000: p. 602).

**Nested caldera:** A type of caldera which is found within a larger, older caldera structure.

**Nuees Ardentes:** A French term applied to a highly heated mass of gas-charged ash which is expelled with explosive force and moves hurricane speed down the mountainside. The term used for a "glowing avalanche" resulting from a small-volume block and ash flow produced by the collapse of an actively growing lava dome (LaCroix 1904). In recent years, the term has unfortunately been more widely used as a synonym for "ignimbrite". Its use should be restricted to the original definition of LaCroix (Cas and Wright 1987: p. 225).

**Outwash:** Sediments deposited by glacial meltwater beyond the active glacial ice. Outwash sediments are commonly characterized by poorly bedded gravels interlayered with well-bedded (and commonly cross-bedded) sands.

**Pahoehoe lava:** A Hawaiian term to describe lava flows with smooth, continuous surfaces (Kilburn 2000: p. 291). Pahoehoe flows may have a variety of surfaces described as smooth, ropy (characterized by rope-like, commonly braided flow folds on the lava flow's surface), or shelly (vesicular and cavernous; Cas and Wright 1987: pp. 66–67). Pahoehoe toes and lobes form when largely degassed mafic magma issues from tubes relatively far from the erupting vent.

**Palagonite:** Conspicuous yellowish alteration product of hydrated basaltic glass, mainly composed of clay minerals (Smellie 2000: p. 403).

**Pali:** Hawaiian word for steep hills or cliffs.

**Particle cohesion:** The sticking together of particles due to either the presence of water (at low temperatures) or to particle plasticity (at high temperatures) (Wilson and Houghton 2000: p. 545).

**Pele:** The mythological Polenesian goddess of volcanoes. In Hawaii, this temperamental goddess makes her home in Kilauea's fiery vent, Halemaumau (Sigurdsson and Lopes-Gautier 2000: p. 1297).

**Pele's hair:** A type of achnelith composed of thin, hair-like strands of volcanic glass. These thin, cylindrical strands of volcanic glass are commonly golden in color, have diameters between 1-500 mm in diameter, and may be up to 1 metre in length. They are formed from stretched magma droplets emitted into the atmosphere during fire fountaining and strombolian eruptions (Vergniolle and Mangan 2000: p. 447).

**Pele's tears:** A type of achnelith composed of small droplets of shiny black volcanic glass that have been ballistically molded and quenched during flight into spherical, dumbbell, or tadpole shapes. These droplets generally range from a few millimetres to a few centimetres in size, are generally dense, but locally may be quite vesicular (Vergniolle and Mangan 2000: p. 447).

**Pelean eruption:** A type of volcanic eruption characterized by a ground hugging glowing avalanche (pyroclastic flow) resulting from a mixture of hot volcanic gases, ash, and incandescent lava fragments. Pelean eruptions may occur when pyroclasts are blown out of a central volcanic vent and then collapse onto the earth's surface to form a pyroclastic flow (TILLING 1985). Pelean eruptions may also occur as a result of the explosive disintegration of a lava dome (as was the case for the lava dome on Mt Pelee, Martinique in 1902).

**Peperite:** A genetic term for a rock formed by in-situ disintigration and mixing of molten magma or lava with wet, poorly consolidated sediment (BATIZA and WHITE 2000: p. 361). A breccia-like deposit formed from the extrusive or intrusive mixture of lava or magma with wet sediment (SCHMIDT and SCHMINKE 2000: p. 383).

**Perlite:** Hydrated obsidian, generally light grey in color, that is commonly characterized by rounded, onion-skin-like fractures (perlitic cracks). Apache's tears are unhydrated clumps of fresh obsidian that are commonly found within regions containing perlite.

**Phreatic eruption:** Phreatic eruptions are steam-driven explosions that occur when water beneath the ground or on the surface is heated by magma, lava, hot rocks, or new volcanic deposits (for example, tephra and pyroclastic-flow deposits). The intense heat of such material (as high as 1,170 °C for basaltic lava) may cause water to boil and flash to steam, thereby generating an explosion of steam, water, ash, blocks, and bombs. A steam eruption, commonly associated with water, mud, and other earth materials, that is caused when ground water, heated by a magma, flashes (and explosively expands) into steam (HARRIS 2000: p. 1301). Phreatic eruptions expel no juvenile (magmatic) material, and are commonly the precursor to magmatic eruptive activity.

**Phreatomagmatic eruption:** A type of explosive volcanic eruption that occurs when water (ground water or surface water) comes in contact with hot magma. The quenching of the magma by the water causes the magma to violently fragment into juvenile (cognate) particles that are bounded by fracture surfaces and by rounded walls of broken vesicles. Due to the moisture present, accretionary lapilli are also common in volcanic deposits resulting from phreatomagmatic eruptions (WILLIAMS and MCBIRNEY 1979: pp. 247–248).

**Piecemeal caldera:** A type of caldera characterized by an internal structure composed of several individual fault-bounded blocks (LIPMAN 1997). Piecemeal calderas may result from non-uniform subsidence of a caldera formed from a single eruption, or may be the result of subsidence following a series of large eruptions (multicyclic; LIPMAN 1997; LIPMAN 2000: pp. 655–656).

**Pillow breccia:** A mixture of coarse, typically glassy fragments and broken to whole pieces of pillow lava formed from the shattering of pillow lava crusts (BATIZA and WHITE 2000: p. 361). Pillow breccias commonly form in areas where pillow lavas are not strong enough to maintain their competence along steep submarine slopes or scarps.

**Pillow lava:** A type of submarine lava flow consisting of interconnected, elongated lava tubes. Cross-sections of individual lava tubes resemble pillows with convex upper surfaces and flat or concave lower surfaces (SCHMIDT and SCHMINCKE 2000: p. 383). Both radial and concentric cooling fractures may be present along the margins of individual pillows, and these fractures are brought on by thermal contraction during cooling. Growth of the pillow tubes takes place as the outer, commonly striated outer glassy surface of the pillow tube fractures, and a new tube "buds" from the fracture in a manner similar to the way that toothpaste is squeezed out of a tube.

**Pipe:** A vertical conduit through the Earth's crust below a volcano, through which magmatic materials have passed. Pillow interspaces are commonly filled with volcanic breccia and fragments of older rock.

**Pit crater:** Pit craters are circular-shaped craters formed by the sinking or collapse of the ground. Fissures may erupt from the walls or base of a pit crater, but pit craters are not constructional features built by eruptions of lava or tephra. Pit craters may also partially fill with lava to form a lava lake. They are common along rift zones of shield volcanoes; for example, Mauna Loa and Kilauea volcanoes in Hawaii. No one has observed the formation of a large pit crater, but they are thought to form as a consequence of the removal of support by withdrawal of underlying magma.

**Plate (piston)-type caldera:** A type of caldera in which the caldera floor subsides more or less evenly as one coherent block. Plate-(piston)-type calderas are believed to result from single, large volume pyroclastic eruptions from relatively shallow depth (hypabyssal) magma chambers.

**Plinian eruption:** Plinian eruptions are large explosive events that form enormous dark columns of tephra and gas high into the stratosphere (>11 km). Such eruptions are named for Pliny the Younger, who carefully described the disastrous eruption of Vesuvius in 79 A.D. This eruption generated a huge column of tephra into the sky, pyroclastic flows and surges, and extensive ash fall. Many thousands of people evacuated areas around the volcano, but about 2,000 were killed, including Pliny the Older.

**Plug:** Solidified lava that fills the conduit of a volcano. It is usually more resistant to erosion than the material making up the surrounding cone, and may remain standing as a solitary pinnacle when the rest of the original structure has eroded away.

**Plug dome:** The steep-sided, rounded mound formed when viscous lava wells up into a crater and is too stiff to flow away. It piles up as a dome-shaped mass, often completely filling the vent from which it emerged.

**Pluton:** A body of rock which has formed beneath the earth from crystallization and consolidation from a magma (AGI 1976: p. 334). Plutons may be considered extinct magma chambers (MARSH 2000: p. 191). Large plutons (>40 square miles in area) are called "batholiths".

**Polygenetic:** Originating in various ways or from various sources.

**Pumice:** Pumice is a light, porous volcanic rock that forms during explosive eruptions. It resembles a sponge because it consists of a network of gas bubbles frozen amidst fragile volcanic glass and minerals. All types of magma (basalt, andesite, dacite, and rhyolite) will form pumice. Pumice is similar to the liquid foam generated when a bottle of pressurized soda is opened — the opening depressurizes the soda and enables dissolved carbon dioxide gas to escape or erupt through the opening. During an explosive eruption, volcanic gases dissolved in the liquid portion of magma also expand rapidly to create a foam or froth; in the case of pumice, the liquid part of the froth quickly solidifies to glass around the glass bubbles. Solidified fragments of quenched, highly vesicular (>60%) silicic magma or lava (CASHMAN et al. 2000: p. 421). The highly vesicular nature of pumice results from large volumes of gas rapidly expanding within a rapidly cooling magma. The low density of pumice commonly permits it to float on water for extended periods of time. Hot pumice, however, has been shown experimentally to sink rapidly upon interacting with water (WHITHAM and SPARKS 1986 ).

**Pyroclastic:** Refers to processes resulting from the explosive fragmentation of a magma or lava. May also be used to describe the deposits formed by explosive volcanic activity and directly deposited by transport processes resulting directly from this activity (CAS and WRIGHT 1987: p. 8). Pyroclastic is a Greek term which means "fire-broken" (HARRIS 2000: p. 1301).

**Pyroclastic breccia:** This is produced by volcanic explosion and includes vulcanian breccia, pyroclastic flow breccia, and hydrovolcanic breccia (FISHER 1960).

**Pyroclastic density current:** A gravity controlled, laterally moving mixture of pyroclasts and gas (WILSON and HOUGHTON 2000: p. 545)

**Pyroclastic fall:** The "rain-out" of pyroclasts through the atmosphere from an eruption jet or eruption plume during an explosive volcanic eruption (WILSON and HOUGHTON 2000: p. 545; HOUGHTON et al 2000: p. 555).

**Pyroclastic fall deposit:** Volcaniclastic (pyroclastic) deposits formed from the rain-out of clasts through the atmosphere from an eruption jet and/or plume during an explosive eruption (Houghton et al. 2000: p. 555). Fall deposits typically exhibit mantle bedding, are well sorted, and commonly show well-developed planar stratification (Cas and Wright 1987: pp. 95–96).

**Pyroclastic flow:** A dense, hot, dry, high particle concentration mixture of gas and hot rock fragments (ash, pumice, blocks, etc.) that travels along the ground surface, typically at high velocity (generally on the order of hundreds of feet or metres per second; Harris 2000: p. 1301) away from a volcano. The high speeds of pyroclastic flows are possible because they flow over a thin layer of hot, commonly expanding and escaping gases. Most of the material within a pyroclastic flow is contained within concentrated particle dispersion located at the flow's base (Wilson and Houghton 2000: p. 545).

**Pyroclastic flow deposit:** Pyroclastic (volcaniclastic) deposits that are left by pyroclastic flows (Cas and Wright 1987: p. 96). The deposits are usually topographically controlled (infilling valleys and topographic depressions), massive, and poorly sorted. Depending upon their thickness and heat retention, pyroclastic flow deposits may coalesce into welded tuffs. Pumice-rich pyroclastic flow deposits are often called "ignimbrites".

**Pyroclastic surge:** A type of turbulent, low density (low particle concentration) pyroclastic cloud or pyroclastic density current. Being more dilute than pyroclastic flows, surges can sweep over ridges, hills, and other topographic boundaries. Two kinds of surges are known: wet surges have temperatures <100 °C and contain steam that condenses into water droplets that surge along the ground surface with gas and pyroclasts; and dry surges, which have temperatures >100 °C, and form by either hydrovolcanic eruptions with low water/magma ratios, or by magmatic eruptions driven solely by expanding magmatic gases (Valentine and Fisher 2000: p. 571).

**Pyroclastic surge deposit:** Pyroclastic deposits that are left by pyroclastic surges. These deposits mantle topographic features but also generally thicken within topographic depressions. These deposits are generally well-sorted, and are enriched in crystals and lithic fragments relatively to pyroclastic flow deposits. Surge deposits commonly exhibit unidirectional sedimentary bedforms, including low angle cross-bedding, dune forms, climbing dune forms, pinch and swell structures, and chute and pool structures (Cas and Wright 1987: p. 98).

**Quenching:** The rapid cooling of magma to form glass (Batiza and White 2000: p. 361). Fuel-coolant interactions commonly lead to quenching. Abrupt quenching may cause a rapid volume decrease which leads to fragmentation of the glass (cooling-contraction granulation).

**Renewed volcanism state:** Refers to a state in the evolution of a typical Hawaiian volcano during which — after a long period of quiescence — lava and tephra erupt intermittently. Erosion and reef building continue.

**Repose:** The interval of time between volcanic eruptions.

**Reticulite:** Reticulite is basaltic pumice in which nearly all cell walls of gas bubbles have burst, leaving a honeycomb-like structure. Even though it is less dense than pumice, reticulite does not float in water because of the open network of bubbles. The delicate glass threads between the bubbles are so fragile that reticulite was first called "thread-lace scoria" by the great American mineralogist, James Dana. An exceptionally porous type of scoria containing porosities ranging from 95-99% (Vergniolle and Mangan 2000: p. 447; McPhie et al. 1993: p. 27). Commonly referred to as "thread-lace" scoria, reticulite is made up of a honeycomb-like network of thin glass fibers.

**Resurgent dome:** The central highland in many large calderas formed by gradual upwarping of the caldera floor after caldera collapse as a result of renewed magma intrusion.

**Reynolds number:** A dimensionless quantity characterizing the relative importance of inertial or momentum related forces to viscous forces in fluid flow. In magmas viscous forces usually dominate, resulting in a Reynolds number that is less than 1 (Carrigan 2000: p. 219)

**Ring fracture/Ring fault:** The arcuate bounding faults upon which caldera (cauldron) subsidence takes place. Ring fractures (faults) define the structural limits of calderas. Most observed ring faults are nearly vertical or dip steeply inward (toward the center of the caldera), and this is thought to be a result of doming of the caldera structure following its initial formation (Lipman 2000: pp. 649–650).

**Ring plain:** Region surrounding a volcano beyond lower topographic flanks, over which tephra and mass-wasting products are radially distributed.

**Ropy pahoehoe:** A type of pahoehoe lava characterized by flexible crusts that are bent into tight folds as lava flows. These tight folds form lava surfaces that appear to be made up of a series of braided ropes (Kilburn 2000: p. 295).

**Scoria:** Scoria is a vesicular (bubbly) glassy lava rock of basaltic to andesitic composition ejected from a vent during explosive eruption. The bubbly nature of scoria is due to the escape of volcanic gases during eruption. Scoria is typically dark grey to black in colour, mostly due to its high iron content. The surface of some scoria may have a blue iridescent colour; oxidation may lead to a deep reddish-brown colour. Solidified fragments of quenched, highly vesicular (>60%) mafic magma or lava (Cashman et al. 2000: p. 421). The highly vesicular nature of scoria results from rapid cooling of gas-rich lava.

**Scoria (cinder) cone:** Small volcanic landforms formed from focused (single-vent) subaerial strombolian eruptions of basalt or basaltic-andesite magma. These features have an inverted cone-shaped profile and are generally circular in plan, although elongate scoria cones can be formed from multiple-vent volcanic eruptions (Cas and Wright 1987: pp. 371–372).

**Seamount:** A submarine volcano.

**Sector collapse:** A destructive volcanic process during the growth history of a volcano. Debris avalanche deposits are the products of sector collapses (Ui et al. 2000: p. 617).

**Shearing:** The motion of surfaces sliding past one another.

**Shelly pahoehoe:** A type of pahoehoe lava characterized by highly vesicular, extremely fragile crusts that form over hollow lava blisters. The surfaces of these blisters break easily when stepped upon, giving the impression of walking on eggshells (Kilburn 2000: p. 295).

**Shield volcano:** A broad, low-relief volcano constructed by flows of relatively fluid lava (e.g. basalt: Spudis 2000: p. 698). Flank slopes on shield volcanoes are typically < 5° (Zimbelman 2000: p. 771).

**Sill:** A tabular body of intrusive igneous rock, parallel to the layering of the rocks into which it intrudes.

**Sinter:** A type of fragile, commonly white or grey rock formed by precipitation of silica from cooling hydrothermal solutions at or near a hydrothermal vent. Precipitation of siliceous sinter (often with associated sulphide minerals and precious metals) commonly occurs in neutral and acid hydrothermal systems under the influence of biogenic agents such as algae and bacteria (Cas and Wright 1987: p. 316).

**Skylight:** An opening formed by a collapse in the roof of a lava tube.

**Slabby pahoehoe:** A type of pahoehoe lava with a surface composed of slabs of broken lava crust that are up to metres across and up to several centimetres thick (Kilburn 2000: p. 295).

**Slug flow:** One of four two-phase flow regimes, in which large gas pockets,which are almost as large as the eruption conduit rise through magma (Jaupart 2000: p. 237).

**Solfatara:** A type of steam vent or dry fumarole that is characterized by quiet discharge (<20 m/s), and that precipitates a significant amount of sulphur (Hochstein and Browne 2000: pp. 850-851).

**Spall fragments:** Formed by shattering of the brittle, quenched, crust of a lava by a combination of thermal contraction and flexure at the margins of a still-flowing lava (Batiza and White 2000: p. 361)

**Spatter bomb:** A glassy pyroclast greater than 64 mm in diameter that takes on a fluidal shape by the force of ejection (Vergniolle and Mangan 2000: p. 447).

**Spatter cone:** A low, steep-sided cone of spatter built up on a fissure or vent. It is usually of basaltic material.

**Spatter rampart:** A ridge of congealed pyroclastic material (usually basaltic) built up on a fissure or vent.

**Spherulite:** Typically rounded, radiating arrays of crystal fibers produced by the high temperature devitrification of volcanic glass. In felsic rocks, the crystal fibers are generally composed of alkali feldspar and a silica polymorph (either quartz or cristobalite), whereas in mafic rocks the fibers commonly consist of plagioclase and/or pyroxene. Spherulites typically have diameters of 0.1-2.0 cm, but can be much larger (commonly up to 20 cm). Isolated spherulites are generally spherical, but adjacent spherulites may impinge upon one another to produce long chains that are often aligned with flow foliation (McPhie et al. 1993: pp. 24–25).

**Spindle bomb:** Volcanic bombs are masses of lava greater than 64 mm diameter ejected from a vent. Bombs are viscous when expelled and deform during flight to assume a characteristic shape on impact.

**Spines:** Horn-like projections formed upon a lava dome.

**Steady-state or equilibrium profile:** Shape of the edifice (cone) once an active volcano has become well established — follows the initial cone building, precedes long-term erosional degradation, and represents a balance between construction through mass addition (eruption) and degradation through erosion.

**Stratovolcano:** A generally steep sided volcano composed of alternating layers of lava flows, pyroclastic deposits, and commonly, volcaniclastic sedimentary deposits (Walker 2000: p. 283). Stratovolcanoes commonly have increasing slopes toward their summits since they generally have mainly lava flows and sedimentary deposits near their base and pyroclastic (tephra) deposits near their summits. Stratovolcanoes also called as “composite volcano”.

**Stony rhyolite:** Very finely crystalline rhyolite lava (Cas and Wright 1987: p. 84).

**Strombolian eruption:** Volcanic eruptions of basaltic magma, slightly more violent than Hawaiian eruptions, that produce large amounts of scoria and ash around a central vent to form a cone. Strombolian eruptions are typically pulsating and have periods of several seconds (Wolf and Sumner 2000: p. 321). The deposits consist of lava spatter, vesicular bombs, scoria lapilli, and mafic ash (Vespermann and Schmincke 2000: p. 683). It named after Stromboli, an Italian volcano.

**Superheated liquid:** A metastable thermodynamic state of a liquid resulting from rapid heating to a temperature well above the boiling point (Morrissey et al. 2000: p. 431).

**Surge:** A ring-shaped cloud of gas and suspended solid debris that moves radially outward at high velocity as a density flow from the base of a vertical eruption column accompanying a volcanic eruption or crater formation. A pyroclastic surge is a turbulent cloud of gas and rock fragments that flows across the ground. A pyroclastic surge is more dilute than a pyroclastic flow. Surges are not constrained by topography but can move over obstacles such as ridges and hills.

**Surtseyan eruptions:** Hydrovolcanic eruptions dominated by jets of wet tephra (scoria and ash) that result in the formation of tuff cones. The term “surtseyan” is generally used for volcanoes erupting through seawater. Named after Surtsey, a volcano which emerged from the sea off the coast of Iceland in 1963 (Vespermann and Schmincke 2000: p. 683).

**Talus:** A slope formed a the base of a steeper slope, made of fallen and disintegrated materials.

**Tephra:** A general term used by volcanologists to describe all fragmental volcanic ejecta produced during explosive volcanic eruptions (Dehn and McNutt 2000: p. 1271). This includes ash (<2 mm diameter fragments), lapilli (2–64 mm diameter fragments and fragements greater than 64 mm in diameter known as bombs (semi-solid or plastic ejecta) or bombs (solid ejecta) (Tilling et al. 1987).

**Tephra (finger) jet:** parcels of bombs, wet tephra, and gas that follow spreading parabolic trajectories and are ejected by discrete, relatively low frequency shallow explosions (Smellie 2000: p. 403)

**Tephrochronology:** The collection, preparation, petrographic description, and approximate dating of tephra.

**Topographic inversion:** Process whereby through time valleys become ridges and vice versa — can occur on volcanoes as volcanogenic products such as lavas are channeled down valleys, focusing subsequent erosion along their edges.

**Trap-door caldera:** A type of caldera formed when one part of the caldera floor subsides to a greater depth than the other side of the caldera floor. In general, trap-door calderas have a partial ring fracture (associated with the side of greatest caldera collapse) and a hinge area (associated with the side of least collapse). Trap-door calderas may represent either calderas that have undergone incomplete collapse, or calderas formed from eruptions from shallow asymmetrical magma chambers (Lipman 1997; Lipman 2000: p. 654).

**Tree mold:** Fluid basaltic lava may preserve the shapes of trees and other objects by solidifying around them. Tree molds are formed when lava surrounds a tree, chills against it, and then drains away. The standing structure left behind is often called a lava tree. Tree trunks engulfed and incinerated by lava leave cylindrical hollows, or tree molds, where lava solidified against them; tree molds often preserve the original surface texture of the tree. Tree molds are found within standing lava trees and on the surfaces of lava flows. They are common in pahoehoe flows and occasionally found in `a`a flows.

**Tremor:** A continuous vibration of the ground around active volcanoes (Vergniolle and Mangan 2000: p. 447). Tremors defined on seismographs may have either a regular sine-wave appearance (harmonic tremor) or an irregular, pulsating appearance (spasmodic tremor) (McNutt 2000: p. 1015).

**Tuff:** A lithified volcaniclastic rock composed primarily of ash, with up to minor volumes of lapilli and/or blocks and bombs (Fisher 1966). Originally used as a non-genetic rock name, common use today typically implies (incorrectly) that the tephra comprising the rock was deposited while hot. Similar deposits that have no indication of being hot while deposited are commonly referred to as “tuffaceous” (McPhie et al. 1993: p. 8).

**Tuff cone:** A type of hydroclastic volcano that is generally higher than (generally >50 m high), and has steeper external flanks (commonly >25°) than tuff rings or maars (Vespermann and Schmincke 2000: p. 684). Craters within tuff cones are generally higher in elevation than the adjacent land surface. Tuff cones are made up primarily of juvenile clasts deposited from lateral surges, airfall, and associated volcaniclastic remobilization processes.

**Tuff ring:** A type of hydroclastic volcano, generally <50 m high, defined by craters with low depth/width ratios that sit at or above the elevation of the adjacent land surface. The rims around tuff rings are composed of juvenile and accidental clasts and are deposited in beds with dips <25° (Vespermann and Schmincke 2000: p. 684).

**Tumescence:** The doming or uprising of a volcano commonly due to inflation of a shallow magma chamber. Regional tumescence commonly occurs prior to a major pyroclastic eruption, but may also occur following an eruption as less volatile magma is emplaced into the shallow crust (Smith and Bailey 1968).

**Tumulus:** The surfaces of pahoehoe flows on flat or gentle slopes often exhibit elliptical, domed structures called tumuli. A tumulus is created when the upward pressure of slow-moving molten lava within a flow swells or pushes the overlying crust upward. Since the solid crust is brittle, it usually breaks to accommodate the "inflating" core of the flow. Such fractures generally extend along the length of a tumulus, and are frequently accompanied by smaller irregular cracks down the sides. Lava commonly squeezes out through these fractures, and sometimes drains from the tumulus to leave a hollow shell.

**Turbidite:** Sediment or rock deposited from a gravity-driven flow of suspended sediment in water (turbidity current), characterized by well-developed sedimentary structures arranged in a regular sequence (Smellie 2000: p. 404)

**Turbulent flow:** flow regime where inertial effects dominate, such that the flow is well mixed by small-scale eddies (Jaupart 2000: p. 237)

**Tuya:** A flat-topped, steep-sided volcano that erupted into a lake thawed into a glacier by volcanic heat (Smellie 2000: p. 403). Tuya commonly referred to as a "table mountain".

**Unconformity:** A surface of erosion that separates younger strata from older rocks (AGI 1976: p. 448).

**U-shaped channels:** These channels in base surge deposits, described by several authors, are symmetrical in cross section with curving bottoms that clearly cut underlying layers. Most range from about 0.3 m to 7 m across and are a few centimetres to 3 m deep, but unusually large channels (30 m across 20 m deep) are also reported. The curving bottoms are best described as U-shaped, not parabolic curves, even though some are very broad in cross section. Beds reflect the shape of the channels, but the curvature of individual beds decreases upward, and the final fill extends uniformly across the channel and is conformable with the sequence outside the channel. Thus, beds thicken toward the centers of channels and therefore do not resemble draped fallout layers.

**Vapor film:** A layer of vapour formed at the interface of water and hot liquid or solid body (Morrissey et al. 2000: p. 431).

**Variolite:** A spherulite-like radiating aggregate composed of feathery, needle-like crystals of plagioclase and pyroxene that occur in mafic volcanic rocks (typically basalt). Variolites may result from devitrification, but are commonly believed to be formed in subaqueous rocks by quench-induced crystallization (Cas and Wright 1987: p. 420).

**VEI:** The Volcanic Explosivity Index, or VEI, was proposed in 1982 as a way to describe the relative size or magnitude of explosive volcanic eruptions. It is a 0-to-8 index of increasing explosivity. Each increase in number represents an increase around a factor of ten. The VEI uses several factors to assign a number, including volume of erupted pyroclastic material (for example, ashfall, pyroclastic flows, and other ejecta), height of eruption column, duration in hours, and qualitative descriptive terms.

**Vent:** A surface opening through which volcanogenic materials are erupted (Davidson and DeSilva 2000: p. 663). Typically thought of as a hole in a planet from which volcanic products (magma, ash, etc.) are erupted (Spudis 2000: p. 697).

**Vesicle:** A frozen bubble in a volcanic rock. Vesicles are formed when magma crystallizes around a gas bubble (Spudis 2000: p. 697).

**Vesicular tuff:** Tuffs containing millimetre to centimetre-sized, irregular to round vesicles which are interpreted to form during trapping of air or vapor in wet ash deposits (Vespermann and Schmincke 2000: p. 683).

**Vesiculation:** Nucleation and growth of gas bubbles in a magma (Cashman et al. 2000: p. 421)

**Vesuvian eruption:** Commonly used as a synonym for a "Plinean" eruption (e.g. Tilling 1985), but also used to describe basaltic eruptions which involve long-sustained gas streaming with little ash being released (as in the 1906 eruption of Vesuvius; Cas and Wright 1987: p. 130).

**Viscosity:** A measurement of the ratio of shear stress to the rate of shear strain in a fluid (Williams and McBirney 1979: p. 20). In common language, how easily a fluid flows. Considered the most important physical property of a magma because it largely determines eruptive style as well as volcano morphology. Magma viscosity generally increases as the silica content of the magma increases (due to silica polymerization) and as the temperature of the magma decreases. Magma viscosity may also be affected by the presence of trace elements (e.g. Ti) or volatiles (e.g. $H_2O$, $CO_2$, $SO_2$, etc.). In general, common magmas increase in viscosity in the following order: komatiite, basalt, andesite, dacite, rhyodacite, rhyolite.

**Vog:** Smog of volcanic origin, composed of volcanic ash and gases

**Volcanic field:** These are clusters of small volcanoes which erupt only once. They usually have a cinder cone and lava flows. A typical field may contain 10 to 100 volcanoes. One of the largest is the Michoacan–Guanajuato field in Mexico which contains 1000 volcanoes.

**Volcanic bomb:** Juvenile fragments of semi-solid or plastic magma ejected during a volcanic eruption. Based on their shapes after they hit the ground and cool, bombs are given various textural names including breadcrust bombs, cow-dung (cow pie) bombs, spindle bombs (fusiform bombs) and ribbon bombs. Type of bombs: Rotational bomb, Tear-shaped bomb, Spheroidal bomb, Spindle-shaped bomb, Pancake-shaped bomb, Slag bomb, Pumiceous bomb, Ribbon bomb, Cored bomb, Olivine bomb, Breadcrust bomb, Cow pat bomb

**Volcanic breccia:** The term volcanic breccia is used as a general term applying to all coarse-grained rocks composed of angular volcanic fragments (Fisher 1960).

**Volcanic complex:** A persistent volcanic vent area that has built a complex combination of volcanic landforms.

**Volcanic cone:** A mound of loose material that was ejected ballistically.

**Volcanic cycle:** A general term used to describe a period of increased volcanic activity.

**Volcanic field:** A region comprising a large number of volcanic edifices. Volcanic fields are usually associated with basaltic volcanism, and commonly comprise a number of small, monogenetic volcanoes (e.g. cinder cones, maars, tuff cones, tuff rings, small shield volcanoes, lava domes). Fields may form in linear trends associated with tectonic structures (such as faults), on the flanks of larger composite or shield volcanoes, or within calderas (Connor and Conway 2000: p. 331).

**Volcanic magnitude:** Volcanic magnitude is the total mass of erupted material. The scale spans from 0 to 9 magnitude = log10 (erupted mass kg) –7. The magnitude will be similar to VEI which also measures mass erupted.

**Volcanic neck:** A massive pillar of rock more resistant to erosion than the lavas and pyroclastic rocks of a volcanic cone.

**Volcaniclastic:** A non-genetic term used to describe any fragmental aggregate of volcanic parentage (Cas and Wright 1987: p. 8). Rocks formed by the fragmentation of volcanic materials (either magma or volcanic rocks) irrespective of the method of fragmentation. Pyroclastic rocks and epiclastic rocks are both considered to be "volcaniclastic".

**Volcaniclastic facies:** These facieses are defined by distance from source, type of transporting agent, environment of deposition, and in some cases, by composition. First-order volcaniclastic facies are generally defined by position of the rock body relative to source within non-marine or marine environments, e.g. proximal, medial and distal facies.

**Vulcan:** Roman god of fire and the forge after whom volcanoes are named.

**Vulcanian eruption:** A vulcanian eruption is a type of explosive eruption that ejects new lava fragments that do not take on a rounded shape during their flight through the air. This may be because the lava is too viscous or already solidified. These moderate-sized explosive eruptions commonly eject a large proportion of volcanic ash and also breadcrust bombs and blocks. Andesitic and dacitic magmas are most often associated with vulcanian eruptions, because their high viscosity (resistance to flow) makes it difficult for the dissolved volcanic gases to escape except under extreme pressure, which leads to explosive behavior. An explosive volcanic eruption generally expelling less than 1km3 of material, but with an eruption column that may reach heights of up to 10-20 km (Nakada 2000: p. 945). These eruptions last on the order of seconds to minutes (Morrissey and Mastin 2000: p. 463).

**Welding:** The sintering together of hot, glassy fragments, irrespective of shape and size, by compactional lithostatic load (Cas and Wright 1987: p. 165). Postdepositional compaction of a hot pyroclastic deposit under its own weight. Note that agglutination, coalescence, and postdepositional welding are part of a process continuum (Wolff and Sumner 2000: p. 321).

**Welded tuff:** A hard pyroclastic rock compacted by internal heat and pressure from overlying pyroclastic deposits.

**Wet/dry eruptions:** Depending on the quantity of available heat energy, when water comes into contact with magma it may become a liquid-rich fluid or a vapour-rich fluid, resulting in a dry eruption or a wet eruption respectively (Morrissey et al. 2000: p. 431).

**Xenocrysts:** A crystal that resembles a phenocryst in igneous rock, but is a foreign to the body of rock in which it occurs.

**Xenoliths:** A foreign inclusion in an igneous rock.

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