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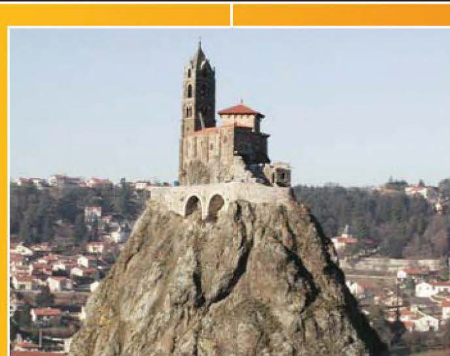
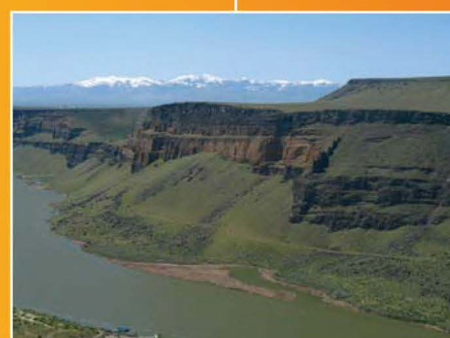
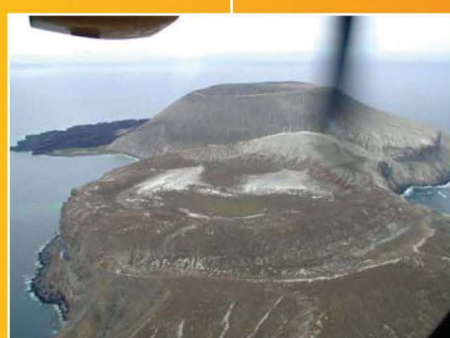
What is a Volcano?


edited by Edgardo Cañón-Tapia and Alexandru Szakács

What Is a Volcano?

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Cover: It is very difficult to capture in a single image the complexity and diversity of features encompassed by the term "volcano," as shown by the various images on the cover, some of which can be considered to be individual volcanoes either on Earth or on other planets or moons of our solar system, whereas some others illustrate essential components of a volcano that nonetheless seldom enter the definition of that term.

Images from left to right: top row: regional dike (central Tibet), volcanic neck (Le Puy en Velay, France)*, stratovolcano (Villarica, Chile)*. Second row: mud volcano (Lusi, Indonesia, from Chapter 3, this volume), sequence of lava flows (Snake River plain, from Chapter 4, this volume). Third row: guttae (Neptune's moon Triton, from Chapter 2, this volume), tephra cone and flattened lava dome (San Benedicto island, Mexico), shield volcano and scoria cone (Mendoza region, Argentina)†. Bottom row: rhyolitic dome with coulée (Socorro island, Mexico), filled lava tube (Pinacate volcanic field, Mexico). Photo credits: *Frida Cañón, †Alexandru Szakács, the rest courtesy Edgardo Cañón-Tapia or from corresponding chapters in this volume as noted.

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Introduction

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Humankind has been aware of volcanic activity in this planet since ancient times. This can be inferred from the remains of human settlements giving evidence of destruction by volcanic activity, and by the many myths around the world describing events that can be interpreted in relation to volcanic eruptions. Furthermore, such occurrences are evidenced by the special words that some human groups created to designate the special cases of "fire mountains," thus distinguishing these from other (nonvolcanic) mountains found in the same region. More recently, films and television shows devoted to exploring volcanoes have become very common, making it easier for the general public to gain access to "firsthand" experiences concerning this type of natural phenomenon. Consequently, it is only fair to say that at present almost everyone has an "intuitive" knowledge of what a volcano is.

To some extent the same intuitive knowledge could be said to exist relative to planets. Nevertheless, as recent events have shown, the definition of a planet is far from being well established, and there is much current debate and controversy surrounding such a definition. Actually, the definition of a planet has not been constant through human history. For instance, there was a time when the four largest moons of Jupiter (Io, Europa, Ganymede, and Callisto) were called "Medicean Planets," and at another time the first known asteroids (Ceres, Pallas, Juno, and Vesta) were also considered planets (Croswell, 1999). As more discoveries of objects within our solar system, and of large objects around other stars, started to accumulate, the astronomical community has been forced to adjust the definition of a planet, and this process seems to have continued until the present. For instance, as a result of the accumulation of new observations made during the last half of the twentieth century, it appears that only eight of the nine planets considered to form our solar system during most of that century will have such a status in the twenty-first century. This is because Pluto is no longer considered to be a planet, at least from a formal scientific point of view (IAU General Assembly, 2006). Noteworthy, the definition of a planet adopted in 2006 spurred debate and controversy among the scientific community (e.g., Britt, 2006; Sarma et al., 2008), showing that despite astronomy being an ancient discipline, the meaning of one of its pivotal concepts has required formal revision as new discoveries have been made.

Evidently, scientists devoted to studying Pluto will not become unemployed suddenly because the new definition does not grant it the status of a planet; Pluto will continue to exist and deserves to be studied in the same detail regardless of whether it is a planet or not. Nevertheless, the status of Pluto as a planetoid might turn to be advantageous, because it can highlight the fact that there are other objects in our solar system that have many similarities to Pluto, and that such similarities put them apart from the other eight objects now considered as the only planets of our solar system. Consequently, the new definition (or actually, the debate associated with the proposed new definition) can help the scientific community to increase the knowledge of our solar system by having opened the door for an in-depth examination of a pivotal term—that of a planet—in one of the most ancient scientific disciplines.

The attitude of astronomers throughout the history of this discipline, concerning one of their fundamental terms, teaches us several important lessons. First, there is an important difference between the "intuitive" and the "formal" definition of an object. Second, no matter how ancient a discipline (or a term) is, the formal definition, even of concepts for which an intuitive definition has been around for a long time, might eventually require adjustment to accommodate more recent observations. Third,

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the process of revising formal definitions may be advantageous, because it provides elements that increase our understanding of relationships between different objects, and of processes related to their origin as well. Fourth, the field of astronomy does not cease to exist, even if some of the pivotal terms are changed over time. All of these lessons might be relevant for volcanology, especially if it is considered that this discipline has undergone a rapid growth during recent decades.

Unfortunately, unlike the case of astronomy, a vast sector of the volcanological community seems to remain resistant to a discussion concerning definitions of fundamental concepts. This attitude is illustrated by remarks such as, "I've been studying volcanoes for over 30 years, and therefore I ought to know very well what a volcano is." Nevertheless, a small group of volcanologists (including both editors of this book) diverges from that perception, feeling rather that we may be experiencing the first stages of an inflection point, similar to that observed in the field of astronomy. In particular, we believe that the intuitive definition of a volcano, given by personal experience in research, may no longer be sufficient for a formal definition of the object of study. This is because development of new technologies has led to discoveries that are commonly beyond the field of expertise of one single person; thus, the exercise of critically revising the formal definition of fundamental terms is not an idle task.

After having met by chance in several scientific meetings, both editors of this book started to identify common concerns about the restrictions and ambiguities imposed by the current definition of a volcano, and other fundamental concepts. Eventually, we decided to probe the volcanological community on these topics by organizing a special session with the title "What's a volcano? New answers to an old question," which was held during the 2007 American Geophysical Union–Joint meeting in Acapulco, Mexico. Although the number of works contributed to that session was much smaller than those contributed to other, more technical (or less philosophical) sessions at the same meeting, it became clear to us that we were not the only two volcanologists with similar concerns. Fortunately for us, the editors of Geological Society of America Books seemed to have had the same impression and invited us to put together those contributions in the form of a Special Paper. We gladly accepted this invitation, and with high optimism we launched ourselves to the task of editing the present volume.

The initial response from colleagues invited to contribute a chapter to this book, even if they had not been present at the 2007 meeting, was most positive. Unfortunately, along the road we found several obstacles, many of which were related to the refractory attitude of other colleagues, almost leading us to cancel the project. Nevertheless, after critically reading the material available at that time we concluded that the term *volcano* was indeed not well defined, and that the whole volcanological community could benefit from a discussion parallel to that exemplified by the recent story of a planet. Consequently, despite the relatively small number of contributed papers, compared to those originally committed, we decided to continue with editing the book. Fortunately for us, we had the continued support of GSA. The result of these efforts is the present book, which we think provides the basis required to initiate an introspective exercise. We hope this exercise will lead to open discussion of fundamental aspects of our discipline, and ultimately lead to a better definition of fundamental terms in volcanology.

Whereas we will not claim to have found a final answer to the question posed by the book's title, we think that enough elements to promote a healthy revision of the term are provided by the several conclusions reached in the chapters of the book. For instance, a review of currently available definitions is made in the first chapter (Borgia et al.), in which it is shown that these definitions only consider morphological and eruptive aspects. The authors state that volcanoes should be defined by considering other aspects, related to various geologic disciplines. Consequently, these authors propose a new definition in which a volcano is considered a geological environment characterized by three elements: magma, eruption, and edifice. In Chapter 2, in contrast, after reviewing the different forms of volcanic activity seen in other worlds of our solar system, Lopes et al. define a volcano as an opening on the planet's (or moon's) surface from which magma and/or magmatic gas is erupted. Although not far from current definitions of a volcano, the definition proposed by Lopes et al. depends strongly on the term *magma,* which also needs to be defined with more precision than is current, because the same material may be considered magmatic on one planet and not so on another.

Actually, some of the problems concerning what can be considered magmatic within a planet are also found when a special type of volcanoes, studied only recently on Earth, is included in the discussion. This is illustrated in Chapter 3 (van Loon), in which "sedimentary volcanoes" are examined in some detail. It is shown that these volcanoes share many characteristics with "igneous volcanoes," including morphological aspects, and even processes that contribute to the genesis of both igneous and sedimentary volcanoes. Such similarities certainly suggest that we need to think again what the real meaning is of the terms *volcano* and *magma.* Yet a different set of problems related to the morphologically based definitions of a volcano are encountered when attention is focused on the distinction of monogenetic and polygenetic volcanism. Some of those problems stem from the relative scale of the structures, as one "vent" within a large composite volcano could be called a volcano if formed within a volcanic field; other problems concern the duration of activity within a region, the cumulative volume of erupted products, and the hazards associated with volcanic activity. Some of these topics are examined in some detail in Chapter 4 by Németh, and revisited from a different perspective in Chapter 5 by Szakács. Actually, Chapter 5 discusses the elements required to create an acceptable definition of any natural phenomenon, and examines several of the available definitions of a volcano (including those proposed in Chapters 1 and 2) in light of those rules. As a result, two different alternative definitions of a volcano are proposed.

Starting with Chapter 6, each of the following chapters addresses fundamental concepts in volcanology besides the term *volcano.* In Chapter 6, Cañón-Tapia calls our attention to the role played by a definition, and the strong influence that any definition might have in the development of a particular field of study. The particular case examined in his chapter is "Large Igneous Provinces," which may be thought of as types of extreme volcanoes that have been observed not only on our planet, but also elsewhere within the solar system. The influence of a definition is also illustrated by Chapter 7, where van Loon shows that there are reasons to consider composite volcanoes as "sedimentary" structures, because a large proportion of the rocks that form such a structure (pyroclastic deposits) share in common their mechanism of emplacement with sedimentary rocks, and are actually considered by the sedimentological community as a special type of sedimentary rock. Another interesting problem dealt with in Chapter 7 concerns the actual lateral dimensions of a volcano, and the form in which the definition of each of its components influences the very definition of the larger object (i.e., the volcano itself). The problem of considering the dimensions of a given volcano is also examined in Chapter 8, but with an extraterrestrial perspective. In that chapter, Borgia and Murray compare some terrestrial and Martian structures, leading to the conclusion that extremely large and flat volcanoes can be found on a planet in somewhat unexpected forms. Finally, in Chapter 9, Szakács and Cañón-Tapia identify some challenges that volcanologists might face in the near future, most of them arising from the rapid accumulation of data related to an ever increasing spectrum of processes associated with volcanic activity in our solar system. As pointed out in that final chapter, perhaps the most difficult challenges would be to foster effective communication between subdisciplines, and to promote effective cooperation in an effort to solve general (yet fundamental) questions, instead of losing ourselves in the jungle of super-specialization.

Viewed in retrospect, it is clear to us that a large part of the volcanological community may not feel a particular need to revise current definitions of a volcano, or other fundamental concepts of the discipline, perhaps because their personal experience and most immediate research are uninfluenced by the exact definition of those terms. Nevertheless, it is also true that attempting a formal definition that includes the most recent advances and discoveries is not trivial. Actually, as shown by all of the chapters of this book, it is possible to advance more than one formal definition of apparently intuitively simple terms, and some degree of discussion would be required to finally adopt a common one. In turn, as shown by other chapters of the book, the adopted definition can influence in subtle form the development of a particular field of science, and this can in turn determine how fast or slow a particular field evolves.

Whether or not volcanologists follow the example of astronomers, and assemble in the near future to vote on what a volcano is (and equally important, what it is not), is not really important. Neither is it important if any of the definitions advanced in this book are accepted as the best definition of a volcano. What is important is that by promoting discussion of this type of problem we start to create a more mature science, one that will be ready to accept the challenges posed by new discoveries that certainly will be made in the future. Paraphrasing Stern and Levinson (2002), we believe that volcanology has reached a time when new facts and new understandings motivate new classification schemes for some of the bodies volcanologists study, and that at any rate, "one reason for developing a new classification scheme is to provide a framework for the evaluation of new ideas." We only hope that this book stimulates such discussions.

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What is a volcano?

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ABSTRACT

The definition of a volcano is discussed, and a new encompassing version is provided. The discussion focuses on the observations that volcanism is a self-similar process that ranges many orders of magnitude in space and time scales, and that all kinds of geologic processes act on volcanoes.

Former definitions of *volcano*, such as that from the *Glossary of Geology* (1997, p. 690)—"a vent in the surface of the Earth through which magma and associated gases and ash erupt" or "the form or structure, usually conical, that is produced by the ejected material" are clearly insufficient. All definitions that we encountered tend to consider volcanoes from the point of view of a single discipline, each of them neglecting relevant aspects belonging to other disciplines. For the two cases mentioned above a volcano is seen only from the point of view of eruptive activity or of morphology.

We attempt to look at *volcano* holistically to provide a more comprehensive definition. We define a volcano as a geologic environment that, at any scale, is characterized by three elements: magma, eruption, and edifice. It is sufficient that only one of these elements is proven, as long as the others can be inferred to exist, to have existed, or to have the potential to exist in the future.

INTRODUCTION

Figure 1 shows what everyone would unquestionably call a volcano: an eruption in course. The meaning of the word *volcano* has, however, slowly changed through time. During Roman times it was the name of the god of fire (cf. Bullard, 1962; Sigurdsson, 1999), the son of Jupiter and Juno, who lived under the island now called Vulcano, in the Aeolian Archipelagos north of Sicily. Accordingly, the blacksmith work of the god Vulcano was considered the energy source for volcanic activity, whereas the eruptions were the smoke, sparks, and scoria produced during his work. The term later became used in a broad sense to indicate a mountain that throws "fire." However, it is not the etymology of the word that we are concerned with but its meaning, that is, the nature of the thing that the word *volcano* indicates. Obviously, whatever we call this thing, its actual nature and reality do not change because of our definition; rather, the definition reflects our perception and understanding of this reality. This is why an appropriate definition is so important. Accordingly, Schmincke (1986) observes that, since we do not know much about volcanoes, their definition depends strongly upon the background of the scientist giving it. Have we, in the past 20 yr, increased our knowledge about volcanoes enough that we are now able to give a more accurate definition? We would like to believe so.

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Borgia, A., Aubert, M., Merle, O., and van Wyk de Vries, B., 2010, What is a volcano?, *in* Cañón-Tapia, E., and Szakács, A., eds., What Is a Volcano?: Geological Society of America Special Paper 470, p. 1–9, doi: 10.1130/2010.2470(01). For permission to copy, contact editing@geosociety.org.

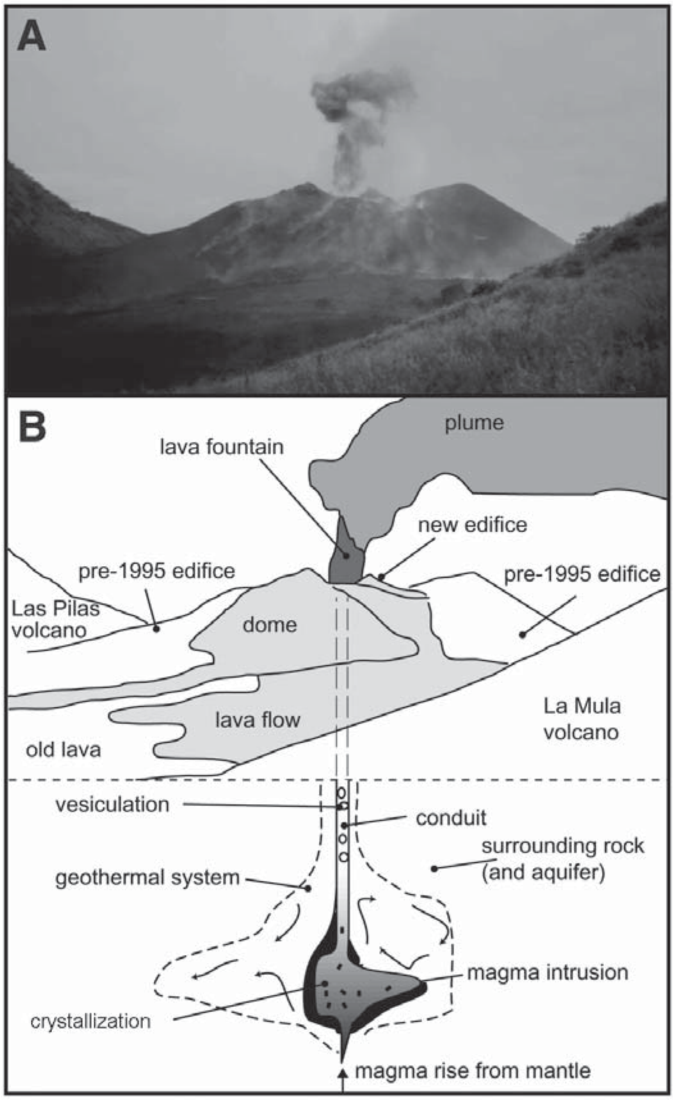

Figure 1. (A) Cerro Negro (Nicaragua) volcano during the November–December 1995 eruption. It is composed of an old cone made of scoria and lava, with a new scoria cone and lava dome growing on it. A hydrothermal system affects the style of eruptions. Volcanic material fell on the city of Leon, some 20 km away, and the gas plume extended over the Pacific Ocean. Faults formed on the cone in response to magma motion, tectonic stresses in the basement, and gravitational loading. The volcano deposits merge with those of the neighboring strato-volcanoes, forming a complex volcanic setting. (B) Sketch of Figure 1A with the "plumbing" system within the basement.

In fact, after a survey of the scientific literature about the definitions that are given to volcanoes (see Table 1), it becomes manifest that the past preference for brief definitions resulted in incomplete, limited, or sometimes contradictory and inconsistent descriptions of the complex phenomenon of volcanism. We conclude that we all probably have quite different perceptions of what volcanoes are—perceptions that, as Schmincke observed in 1986, are invariably influenced by our diverse backgrounds. This disagreement becomes particularly amplified in the corollary definitions of the words *volcanism, volcanology,* and *volcanologist.*

As we stated before, to be meaningful and useful, a definition must reflect our understanding of the nature of volcanoes, and in particular it should not exclude former definitions. Therefore, in this paper, we first analyze former definitions of the word *volcano,* pointing out facts that have become "common knowledge" but that now appear to be at odds with these definitions. Then we propose a new definition that, we think, reflects better our current understanding of volcanoes without contradicting former definitions. Of course we try to stimulate discussion by being provocative and perhaps also polemic. Thus, we expect disagreement.

Editor's note: *In the "References Cited" section at the end of this chapter, many of the alphabetical listings depart from the traditional listings by author(s) by beginning with the volume titles of encyclopedias/dictionaries, in alphabetical order, followed by the name(s) of the editor(s).*

FORMER DEFINITONS OF *VOLCANO*

Perhaps the best place from which to start in describing former definitions is the *Encyclopedia of Volcanoes* (2000; Table 1). In the more than 1400 pages of this volume, by all means a very knowledgeable volcanic tome, we cannot find an explicit proper definition of *volcano,* though in the introduction it is stated that "volcanoes and their eruptions . . . are merely the surface manifestation of the magmatic processes operating at depth in the Earth" (p. 2). On the other hand, the same *Encyclopedia* states that "Volcanology is the study of the origin and ascent of magma through the planet's mantle and crust and its eruption at the surface" (p. 2). Now, the original meaning of the term *volcanology* is the study of volcanoes. Therefore, one could infer that either volcanology has lost its original meaning or that volcanoes are the magma that rises through the planet's mantle and crust to erupt at the surface—which is certainly in contrast with what most scientists would think a volcano is. Some authors, in fact, would prefer the term *magmatology* for what the *Encyclopedia* calls *volcanology*.

What the *Encyclopedia of Volcanoes* states is analogous to the Roman myth of the god Vulcano, which focuses on the source and mechanism of volcanism more than on volcanic products, edifices, or successions; this definition has been (see, e.g., Cotton, 1944), and probably is now, accepted by many petrologists. Instead, we want to point out that the *Encyclopedia*'s definition, by itself, indicates how volcanoes are in general not very well defined and perhaps not well known. As Bullard (1962) states: "to describe what a volcano is not is much easier than to give a concise definition of what it is" (p. 8). Perhaps to avoid this problem, the *Encyclopedia of Earth System Science* (1992), in spite of the impressive erupting volcano on the cover and the numerous definitions of volcanic "things," reports no specific definition for the term *volcano,* nor for the terms volcanism or volcanology.

We have to open a parenthesis: the general understanding is that during an eruption volcanoes emplace silicate rocks. In a few

TABLE 1. COMMON DEFINITIONS OF VOLCANO, VOLCANISM, AND VOLCANOLOGY FOUND IN GEOLOGIC BOOKS, DICTIONARIES, GLOSSARIES, AND ENCYCLOPEDIAS

Year	Title	Editor/Author	Volcano	Volcanism	Volcanology
1845	Kosmos I	Humboldt	—	The essence of all the reactions of a planet against its crust and surface.	—
1858	Lehrbuch der Geognosie	Naumann	—	All the phenomena and the manifestations of forces coming from the interior of the Earth and that originate from the interaction of fluid core-rigid crust.	—
1911	Die Vulkanischen Erscheinungen der Erde	Schneider	—	The phenomena through which juvenile masses coming from deep in the Earth are transported to the interior or the surface of the crust.	—
1911	Recherches sur l'exhalaison volcanique	Brun	A place on the surface of the globe where the temperature can achieve, in a rhythmic or permanent fashion, a temperature much higher than that of the surrounding area. The temperature difference can be over 1000 °C.	—	—
1914	Der Vulkanismus	Wolff	A place at the Earth's surface where the magma and its products have come or are coming out.	All phenomena directly related to the rise of magma.	The study of the rise and the conformation of the magma.
1936 1962	Vulkane und ihre Tätigkeit	Rittmann	—	All phenomena associated with the breaking through of molten material.	The science that studies volcanism.
1944	Volcanoes as Landscape Forms	Cotton	—	The superficial manifestation of the deeper-seated processes of igneous injection or intrusion.	—
1957	Glossary of Geology and Related Sciences	Howell et al. (eds.)	**1**. A vent in the earth's crust from which molten lava, pyroclastic materials, volcanic gases, etc., issue. **2**. A mountain which has been built up by the materials ejected from the interior of the earth through a vent.	Volcanic power or activity; volcanicity. The term ordinarily includes all natural processes resulting in the formation of volcanoes, volcanic rocks, lava flows, etc.	The branch of science treating with volcanic phenomena.
1959	Geological Nomenclature	Schieferdecker (ed.)	A place at the surface of the earth where magmatic material from the depth erupts or has erupted in the past (A. Rittmann), usually forming a mountain, more or less conical in shape with a crater at the top.	All phenomena connected with the rise of magmatic material in a compact state, as injections of magma or effusion of lava, or in a dispersed state, as emanations of ejecta or gases. They are processes in and properties of the hypomagma, mainly produced by physicochemical processes in the magma itself.	The branch of science primarily treating eruptions of magma on the earth's surface, or in levels not far beneath, but also of related features in the tectonical, petrological, seismological, and geophysical fields.
1962	Volcanoes in History, in Theory, in Eruption	Bullard	A vent or chimney which connects a reservoir of molten matter known as "magma," in the depth of the crust of the earth, with the surface of the earth. The material ejected though the vent frequently accumulates around the opening, building up a cone called the "volcanic edifice."	—	The branch of science which deals with the eruption of magma upon the surface of the earth or its rise into levels near the surface.
1963	Geologisches Wörterbuch	Murawski	The construct created from the effusion and eruption of volcanic products, both on land and below the sea.	General concept indicating volcanic and subvolcanic processes.	—
1972 1980 1987 1997	Glossary of Geology	Gary et al. (eds.) Bates and Jackson (eds.) Jackson (ed.)	(**a**) A vent in the surface of the Earth through which magma and associated gases and ash erupt; also, the form or structure, usually conical, that is produced by the ejected material. (**b**) Any eruption of material, e.g., mud that resembles a magmatic volcano.	The processes by which magma and its associated gases rise into the crust and are extruded onto the Earth's surface and into the atmosphere.	The branch of geology that deals with volcanism, its causes and phenomena.
1972	Volcanoes	Macdonald	A volcano is both the place or opening from which molten rock or gas, and generally both, issue from the earth's interior onto the surface, and the hill or mountain built up around the opening by accumulation of the rock materials.	—	The science of volcanoes.

(*Continued*)

TABLE 1. COMMON DEFINITIONS OF VOLCANO, VOLCANISM, AND VOLCANOLOGY FOUND IN GEOLOGIC BOOKS, DICTIONARIES, GLOSSARIES, AND ENCYCLOPEDIAS (*Continued*)

Year	Title	Editor/Author	Volcano	Volcanism	Volcanology
1980	Geological Nomenclature	Nijhoff (ed.)	A hill or mountain built up from the accumulation of volcanic products around a crater, i.e., the accumulation of lavas and/or pyroclastics.	The aggregate of processes associated with the surface phenomena involved in the transfer of materials from the earth's interior to or immediately below its surface.	The branch of geology that deals with volcanism, its causes and its phenomena both at the earth's surface and at deeper levels.
1984 1988 1992 1995	Dictionnaire de Géologie	Foucault and Raoult (eds.)	Place where lavas (molten magma) and hot gases reach the surface of the Earth's crust (or of the Moon, or of the planets) either on the ground or below the water. After cooling, the lavas become volcanic rocks. A volcano generally includes a volcanic cone (formed by the accumulation of the lavas and/or blocks, scoriae, and cinders) around a crater, which is the site of extrusion of the volcanic rocks brought up by the conduit.	Set of volcanic manifestations and associated phenomena.	The study of volcanoes.
1986	Vulkanismus	Schmincke	The definition is different depending on the background of the scientist giving it.	—	—
1987	Grand Larousse		Relief, in general, of conical shape, formed by magmatic products, which reach the surface of the earth in the air or under water.	—	—
1992	Dictionary of Science and Technology	Morris (ed.)	1. A vent or fissure in the earth's surface through which magma and its associated materials are expelled. 2. The generally conical structure formed by the expelled material.	Any of the processes in which magma and its associated gases rise up from the earth's interior and are discharged onto the surface and into the atmosphere.	The study of the causes and phenomena associated with volcanism.
1992	Encyclopedia of Earth System Science	Nierenberg (ed.)	Volcanoes are the landforms that are made when magma (molten rock) erupts onto the surface of the earth.	—	—
1993	The Encyclopedia of the Solid Earth Sciences	Kearey et al. (eds.)	—	—	—
1993	Volcanoes: A Planetary Perspective	Francis	A site at which material reaches the surface of the planet from the interior.	The manifestation at the surface of a planet or satellite of internal thermal processes through the emission at the surface of solid, liquid, or gaseous products.	—
1994	Volcanoes: An Introduction	Scarth	A volcano is usually a cone-shaped hill or mountain composed of materials erupted through an opening in the Earth's crust which extends from the hotter zone below.	—	—
1996	L'Etna et le monde des volcans	Tanguy and Patanè	A volcano is the edifice built by the accumulation of tephra falls and lava flows emplaced by eruptions that have been concentrated in the same place of weakness of the earth's crust.	—	—
1996	Vulcani e Terremoti	Casertano	A fracture on the Earth's crust through which the magma passes from the Earth's interior to its surface. The mountain (usually conical) that forms around and above the fracture owing to the accumulation of the emitted materials.	The set of phenomena and manifestations more or less directly related to the magmatic activity. Volcanism is not equivalent to magmatism.	

(*Continued*)

TABLE 1. COMMON DEFINITIONS OF VOLCANO, VOLCANISM, AND VOLCANOLOGY FOUND IN GEOLOGIC BOOKS, DICTIONARIES, GLOSSARIES, AND ENCYCLOPEDIAS (*Continued*)

	Title	Editor/Author	Volcano	Volcanism	Volcanology
2000	Encyclopedia of Volcanoes	Sigurdsson et al. (eds.)	Volcanoes and their eruptions … are merely the surface manifestation of the magmatic processes operating at depth in the Earth.	—	Volcanology is the study of the origin and ascent of magma through the planet's mantle and crust and its eruption at the surface. Volcanology deals with the physical and chemical evolution of magmas, their transport and eruption, and the formation of volcanic deposits at the planetary surface.
2000 to 2008	This work		A volcano is a geologic environment that, at any scale, is characterized by three linked elements: magma, eruption, and edifice. It is sufficient that only one of these elements is proven, as long as the others are inferred to exist, to have existed, or to have the potential to exist.	The set of processes associated with a volcano.	The study of volcanoes.

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cases these rocks can be carbonatitic (as Oldoynio Lengai in Tanzania; cf. Le Bas, 1977), sulfuric (as on Io; Lopes-Gautier, 2000), icy (as on the icy satellite and Triton; Geissler, 2000), clayey (as in many terrestrial mud volcanoes; Macdonald, 1972), or of just about any kind of rocks (as in many terrestial phreatic or hydrothermal explosions with no direct involvement of magma; cf. Macdonald, 1972). Also in these cases the word *volcano* is used, adding immediately before, the kind of erupted material (e.g., mud volcano).

Surprisingly, an eruption may create no volcanic edifice, in the hypothetical case of the erupted material being completely dispersed in the atmosphere and/or rapidly eroded away. This is why, perhaps, the classic definition of volcano remains general (*Glossary of Geology,* 1997, p. 690): "(a) a vent in the surface of the Earth through which magma and associated gases and ash erupt; also, the form or structure, usually conical, that is produced by the ejected material; (b) any eruption of material that resembles a magmatic *[sic!]* volcano."

In fact, the much older definition from *Geological Nomenclature* (1959) may, by intuition, seem more adequate to many geologists: "a place at the surface of the earth where magmatic material from depth erupts or has erupted in the past, usually forming a mountain, more or less conical in shape with a crater at the top" (p. 223). Note that (in contrast to the definition given just above) here the crater, equivalent to the vent, is the last feature to appear, as if it is not so essential. A more recent edition of the same dictionary reports a definition of *volcano* that makes no explicit mention of magma or eruptions (*Geological Nomenclature,* 1980, p. 106): "a hill or mountain built up from the accumulation of volcanic *[sic!]* products around a crater, i.e. the accumulation of lava and/or pyroclastics."

A definition that is a combination of the former ones is given by the *Dictionnaire de Géologie* (1995): "a place where lavas, molten *[sic!]* magma, and hot gas reach the surface of the Earth's crust (or that of the moon, or of a planet) either in the air or below the water. By cooling, these lavas give volcanic (or effusive) rocks. A volcano generally includes a volcanic cone (the accumulation of lavas and/or blocks, scorias and ash) surrounding a crater, which is the place of exit of the volcanic rocks *[sic!]* that rise through the conduit."

We could spend much longer presenting the quite varied set of definitions published in the scientific literature (Table 1). As can be easily seen, all these definitions maintain some degree of inaccuracy and dependence on the background of their authors. Given the substantial increase in knowledge about volcanoes that has been achieved during the past decades, we believe that these definitions tend to be now inadequate and need some form of integration. Indeed, a volcano is far more complex than what is implied by each of the definitions.

GENERAL OBSERVATIONS ON FORMER DEFINITIONS

The definitions in the scientific literature fall, in general, between three end-member headings: rising magma (cf. Schneider, 1911; Bullard, 1962; *Encyclopedia of Volcanoes,* 2000), eruptions (cf. Wolff, 1914; *Glossary of Geology and Related Sciences,* 1957; *Glossary of Geology,* 1997; Francis, 1993), and volcanic edifice (cf. *Geologisches Wörterbuch,* 1963; *Geological Nomenclature,* 1980; Tanguy and Patanè, 1996). Obviously, petrologists will tend to like the definitions that focus on the magma, geophysicists the ones that focus attention on the eruption, and structural geologists and stratigraphers the ones centering on the edifice or the deposits in general; other specialists will endorse definitions somewhere in between. This is why, in addition to some internal inconsistencies, all definitions appear to be generally unsatisfactory.

Even the simple, early definition of Brun (1911) that a volcano is a *hot spot* obviously misses some major aspects. Indeed, our perception is that mass and momentum transport in a volcano are as important as transport of thermal energy.

For instance, to criticize the most common definition of volcano reported above (*Glossary of Geology,* 1997), one may observe that the first part of the definition should include the feeding conduit in addition to the vent, and possibly the eruptive plume as well. In addition, to be polemic, the use of the words "surface of the Earth" excludes most terrestrial volcanoes that form below sea level or beneath glaciers (making them intrusions) and all planetary volcanoes, which are not terrestrial! Also, a vent formed during a phreatic eruption, which by definition has no magma, could not be considered a volcano. For the second part of the same definition, all "volcanic" intrusive bodies are excluded, such as dikes, sills, magma chambers, and cumulate complexes. In addition, are the volcanic deposits that remain after erosion has taken away parts of the original edifice still to be called a volcano? Also, are the volcanic deposits produced by the various kinds of edifice collapse still part of the volcano?

Another major deficiency of this and all other definitions is that they ignore the hydrothermal systems, which are an integral part of all volcanoes (at least on Earth). In fact, all books on volcanoes include chapters on the hydrothermal system (cf. Macdonald, 1972; *Encyclopedia of Volcanoes,* 2000). In addition, are the non-eruptive volcanic processes, such as volcanic spreading (cf. Borgia et al., 2000a) and the interaction between volcanic edifice stability and basement tectonics (Lagmay et al., 2000; Tibaldi, 2005) part of the dynamics of volcanoes?

Even if we accept standard definitions, more problems arise in defining simple geometric parameters of volcanoes: how is the volcano radius or height measured? Is the radius measured to a basal arbitrary break in slope, to the edge of the lava fields, or to the distance at which volcanic deposits remain continuous? According to the standard definition of *volcano* the radius should be to the edge of the ejected deposits, which usually is time dependent and far beyond the distance of any "reasonable" radius. In the extreme, a volcano that erupts deposits over the whole surface of the Earth has the shape of a spherical shell, not that of a cone!

How is the height of volcanoes measured? Is it the thickness of the volcanic pile, and if so, where is the lower boundary? Is it measured from the level of the basement rocks, which "hopefully" are non-volcanic, or is the feeding conduit included also? What about volcanoes that overlap, such as in Hawaii? Here the height of each single volcano is practically impossible to determine. Not surprisingly, *Volcanoes of the World* (Simkin et al., 1981) reports no dimensions for the listed volcanoes.

A PROPOSED NEW DEFINITION

Before we propose what we think volcanoes are, we would like the reader to ponder two other aspects of volcanoes, which are usually not stressed enough. First, consider the diversity of geologic processes. Clearly, *all processes, properly scaled down, in addition to volcanism, occur on volcanoes:* from chemical and clastic sedimentation to erosion, from dynamo-thermal metamorphism to rock-water interaction, from earthquakes to giant landslides, from volcano-basement interaction to regional geodynamics controlling volcano structural evolution, from death to life. No other geologic environment is so inclusive.

Second, consider the aspect of scale. Large volcanoes, Etna for instance, are commonly made of the superposition of volcanic cones (the "Concazze" and "Il Piano" centers that are the most recent, among the many others; Calvari et al., 1994). These, in turn, have smaller cones on their edifice (like the "cratere di nordest" or the "cratere centrale"; Chester et al., 1985) on which are the actual eruptive boccas. In fact, volcanoes range in size over at least 4–5 orders of magnitude, from meters to hundreds of kilometers, the largest volcanoes always including smaller ones. Their life span may range from hours to millions of years. In addition, as volcanoes grow in size, they tend to change from the generally asymmetric single volcanic deposits, to the radial symmetry of stable scoria cones, to the pseudoradial symmetry of small spreading stratocones like Concepción (Borgia and Van Wyk de Vries, 2003) or Etna (Borgia et al., 1992), to the bilateral symmetry of the large, spreading shield volcanoes like Kilauea (Hill and Zucca, 1987).

In this view, the leap forward in scale to spreading mid-ocean ridges should not be unexpected, and it is intriguing (Borgia and Treves, 1992). Indeed, all of the definitions of *volcano* given in Table 1 apply to mid-ocean ridges as well. They are simply very large volcanoes, so large and massive that gravitational pressure overcomes the rock strength at their base (in the asthenosphere), making them relatively "flat" sectors of spherical shells. Volcanoes of that mass could have only that shape. This fact should not disturb us: There is more difference in size between a scoria cone and Mauna Loa than between Mauna Loa and a mid-ocean ridge. Indeed, the stratigraphy, structure, and processes occurring in the Hawaiian rift zones are frequently compared to those of mid-ocean ridges (cf. Hill and Zucca, 1987; Borgia and Treves, 1992). A similar relationship between volcanism and tectonics at planetary scale may also hold for Mars and Venus (Borgia et al., 2000b; Borgia and Murray, this volume) and for the asteroid volcanism, which generally involves the whole body (Wilson and Keil, 1996).

Indeed, T.A. Jaggar and A. Rittmann considered volcanism ubiquitous within and beneath the Earth's crust (Rittmann, 1936). In a quite figurative fashion, A. von Humboldt (1845) viewed volcanism as "the essence of all the reactions of a planet against its crust and surface," an opinion shared also by Naumann (1858). Similarly, we think that the Earth itself may be considered a self-organizing, self-stratified giant volcano (H. Shaw, 1995, written commun.), so big and so weak that the gravitational pressure collapses it into an orbiting "spherical" planet (Borgia, 1994). We remark that this statement is nonexclusive; other disciplines may have a similar claim without weakening our proposal. It is no surprise, then, that all the geologic processes that occur on a planet are found, properly scaled, at the size of each volcano. Small scoria cones and planets are the volume end members of the same general process: the interaction and feedback between gravitational, thermal, and chemical fields on matter. We observe that, up to now, the cutoff between these two end members has been too drastic. As usual, we may well benefit from dropping our old, textbook ideas.

In view of the above, we are forced to accept that volcanoes are not simple geologic "objects" like crystals, strata, faults, or fossils. Indeed, they are analogous to complex geologic environments, perhaps like sedimentary basins or orogens. Thus, we need a definition that will tend to stress a context much broader than the usual one. Accordingly, we propose that

> Volcanoes are geologic environments where magma, generated at a source within the crust or mantle, flows upward and is subject to varying amounts of physicochemical evolution, intruding and reacting with the encasing rocks and other magma, and originating a geothermal system. Once near the lithosphere top (that is, of a major rigid-fluid, high-low density zone of interface) the magma erupts, piercing the interface. Volcanic deposits are accumulated from eruptions giving rise to a volcanic edifice. In turn, these deposits may become intruded or modified by magma, eruptions, geothermal fluids, tectonics, erosion, landsliding, and all other kinds of geologic processes. The boundaries of this environment (volcano) are frequently time dependent, transitional, ill-defined, or unknown. However, working boundaries can be based on different arguments using factors such as geometry, morphology, and structure.

Of course, we would like a less baroque and more down-to-earth definition. Therefore, we may state in short that

> A volcano is a geologic environment that, at any scale, is characterized by three linked elements: magma, eruptions, and edifice.

It is sufficient that only one of these elements is proven, as long as the others are inferred to exist, to have existed, or to have the potential to exist. Here we use the word *edifice* to indicate all the various geomorphic expressions of volcanoes, including negative features such as calderas and maars.

These definitions may be easily applied to "volcanoes" made by materials different from magma, such as sulfur, ice, mud, and in principle any fluid, only if a direct analogy between them and "magmatic" volcanoes is acceptable. In this case a more general definition can be easily obtained just by substituting, in the short definition above, the word *fluid* to the word *magma*. Consequently, we think that the corollary definitions of *volcanism* as the set of volcanic processes, and *volcanology* as the study of volcanoes, may still be considered quite adequate.

CONCLUSIONS

We realize that the new definition of *volcano* may require adjustment after broader discussion. However, we think that it is now closer to what volcanoes really are: highly complex geologic environments. It stresses the fact that volcanoes belong to a wide range of scales in time and space and that they are dynamic environments—not systems, which have well-defined boundaries—where all kinds of geologic processes act on the rising magmas, the eruptions, the volcanic edifices, and their basements. Therefore, a volcano cannot be limited to the volcanic edifice proper, but must include its basement at least up to where volcanic processes, such as (but not limited to) intrusions, geothermal activity, metamorphism, and edifice-basement tectonic interactions, occur in it. Obviously, our definition does not require unique (rigid) geometric boundaries, which will be better defined on the basis of the studies conducted for specific volcanoes.

This definition has one other important consequence: all geologic disciplines should be applied to and integrated in the study of volcanoes. Therefore, sarcastically, since volcanoes (volcanic environments) may extend to planetary scales, *volcanology,* as it is thought of today, should not exist as an independent discipline because it is only a scaled down *geology*. Inversely, and we are certainly biased toward this view, geology could simply be called *volcanology,* so that volcanologists could continue to exist!

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Beyond Earth: How extra-terrestrial volcanism has changed our definition of a volcano

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ABSTRACT

The discovery of numerous extra-terrestrial volcanoes, including active ones, has stretched our traditional definition of what a volcano is. We now know that the nature of volcanism is highly variable over the solar system, and the traditional definition of a volcano as defined for Earth needs to be modified and expanded to include processes such as cryovolcanism, in which aqueous mixtures are erupted from the interior to the surface. In this chapter, we review past volcanism on the Moon, Mercury, and Mars, active volcanism on Io, and cryovolcanism in the moons of the outer solar system. We suggest the following definition that encompasses the different forms of volcanic activity seen in other worlds: *A volcano is an opening on a planet or moon's surface from which magma, as defined for that planetary body, and/or magmatic gas is erupted*.

INTRODUCTION

Volcanism is a fundamental geologic process that has affected all rocky planets and most moons in the solar system and, presumably, other solar systems as well. As we explore other worlds, we come across signs of active and past volcanism, some in unexpected places. Volcanic materials and eruptions on extraterrestrial worlds can be different from the examples we see on Earth, but the similarities are also striking. An important consequence of this variety is that our understanding of what volcanoes are has changed, and even their currently used definition may no longer be suitable.

The discovery of numerous extra-terrestrial volcanoes, including active ones, has stretched our traditional definition of what a volcano is. Prior to the *Voyager 1* and *2* spacecraft observations during the late 1970s and early 1980s, the Earth was the only planet known to have active volcanism, with the moon, Mars, Venus, and possibly Mercury showing signs of past activity. When *Voyager 1* found active volcanism on Jupiter's moon Io, our understanding of active volcanism, and what causes it, dramatically changed. Io's volcanism is driven by tidal dissipation, fundamentally different from what causes volcanism on Earth. To date, no planet outside the Earth shows evidence of

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plate tectonics. Despite these differences, the eruption styles and products on other planets show great similarity to Earth's.

Voyager 2 went on to observe geysers on Neptune's moon Triton (Kirk et al., 1990), showing the first possible evidence for cryovolcanism, a process that has no direct terrestrial analogue but which appears to be widespread in the outer solar system. The *Galileo* spacecraft showed that relatively recent cryovolcanic activity may have occurred on Europa and Ganymede. In the last few years the *Cassini* spacecraft showed dramatic active plumes on Saturn's moon Enceladus. Features thought to be volcanic have been shown to exist on Titan's young surface, raising the possibility that active or recently active cryovolcanism may have been present there. As we continue our exploration of the solar system, we need to redefine the fundamental geologic processes using a planetary rather than a terrestrial context. Francis and Oppenheimer (2003) defined volcanism rather broadly as "the manifestation at the surface of a planet or satellite of internal thermal processes through the emission at the surface of solid, liquid, or gaseous products." However, we consider this definition to be too broad; for example, geysers on Earth would be considered volcanoes.

In this chapter we review the types of extra-terrestrial volcanism so far found on other planets, and propose a new definition of what a volcano is.

DEFINITION OF A VOLCANO

The word *volcano* has multiple definitions, even if restricted to technical literature (MacDonald, 1972). The traditional definition of a volcano is a place or opening from which molten rock and gas, and generally both, issue from the Earth's interior onto its surface. But a volcano has also been defined as the hill or mountain built up around the opening by accumulation of the rock material poured or thrown out (MacDonald, 1972). Even on Earth, the behavior of erupting volcanoes shows wide variations, depending both on the nature of the material being erupted and on the surrounding environment. Several other alternative definitions of a volcano on Earth can be found in the literature; yet, as discussed by Borgia et al. (this volume), the true essence of what a volcano is seems to be somewhat elusive.

Because the variations in volcanic activity are closely dependent on the kind of molten rock (magma) involved, it is necessary also to define magma types. On Earth, this is generally done by using the name of the rock that results from the consolidation of the magma, such as basalt, andesite, dacite, and rhyolite. Those that are closer to mantle compositions, such as basalts, tend to have higher iron contents and are referred to as mafic or basic, whereas those that have had the time and conditions to evolve and lose their denser components, such as rhyolites, have lower iron contents and are referred to as felsic or acid.

Many mafic magmas appear to be the result of buildup of heat in the mantle from radioisotopic decay, which leads to dynamical instabilities and the escape and rise of buoyant plumes of partial melts. These will ascend until they stall either at a rheological boundary or, owing to neutral buoyancy in the upper few kilometers of the crust, forming magma reservoirs or chambers. Changes in pressure allow magma-filled fractures to form, referred to as dikes, and where those intersect the surface, small quantities of juvenile volatiles brought up from depth such as CO_2 allow the magma to ascend through these dikes and erupt onto the surface. The archetypical example would be the Hawaiian Islands, which are shield volcanoes formed predominantly from effusive flows. When mafic magmas are involved, it is usually, although not exclusively, only through interaction with surface or near-surface bodies of water that powerfully explosive eruptions occur.

A wider range of magma chemistries, including felsic ones, are more commonly found around convergent plate tectonic margins, and appear to be at least partly recycled materials from the surface or near-surface. Remelting of these rocks is the result of any combination of (1) heating by friction as one plate is driven under another, (2) heating by rising bodies of hotter, typically more mafic melts, and, perhaps most importantly, (3) lowering of the melting point due to the addition of water. The mafic magmas in this tectonic setting tend to have higher volatile contents, often several weight percentages of water, which cause more explosive eruptions than in other tectonic settings. Nevertheless, when volatile contents or extrusion rates are low, effusive domes or very thick flows may form.

Other, more exotic volcanic materials—such as carbonatites—exist, but in terms of process they are similar to some sort of felsic or mafic activity. Furthermore, on the Precambrian Earth very low viscosity ultramafic lavas called komatiites erupted, forming large lava flow fields with thermal erosion channels that hosted magmatic nickel-copper-platinum group element ore deposits. In summary, on Earth we see considerable variety in both the composition and genesis of magmas. So what are the commonalities?

On Earth all magmas are molten rock that, at atmospheric temperatures, become solid. The source of their heat is directly or indirectly the interior of the planet itself, and they can be considered primary or secondary expressions of planetary heat loss. While these characteristics coincide with the definition of a volcano made by Francis and Oppenheimer (2003), it is noted that other types of material are erupted onto the surface of the Earth, also typically as an expression of planetary heat loss; yet these other products do not define a volcano. The most obvious example is a geyser, which is not considered to be a volcano. The reason? Water is not magma on Earth, and is instead a volatile material able to exist in solid, liquid, or metastable gaseous form on the surface and in the atmosphere. Although in other respects its origin is similar to some forms of rhyolitic volcanism, in which the rhyolites are remelted and activated by rising pockets of hotter magmas, typically basalts (e.g., Branney et al., 2007), the fact that water in geysers cannot be considered a magma is a key factor for excluding them from the definition of volcanoes on Earth (see discussion below). In a similar vein are sulfur flows, which are volcanic volatile deposits remobilized by heat (e.g., Greeley et al., 1984) and the so-called mud volcanoes that basically are formed when methane is released to the surface, mobilizing the overlying sediments on its way up (van Loon, this volume). Finally, there is a range of other types of extrusive fluid activity, most of which are the result of near-surface chemical processes (e.g., decomposition of volatiles) rather than heating by rising magma.

Various planetary missions have revealed that volcanoes exist on other worlds. On rocky bodies, there is relatively little information about the chemical composition of rocks, as samples have so far been brought back to Earth only from one other body—the Earth's Moon—although most evidence points to more basic silicate compositions, typically basaltic. However, on icy worlds, including Pluto and most of the moons of the outer solar system, volcanic processes are very different. On those worlds, the surface and near-surface temperatures are extremely low. The crusts are composed primarily of water-ice, and the molten near-surfaces are also primarily water, and so, in a very real sense, the "rocks" on these worlds are ices, as they fill the same role as silicate rocks on Earth. On many of these worlds we see strong evidence for both active and past eruptions, which are at times unexpectedly Earth-like in expression, and which have been referred to as cryovolcanic, an expression coined by Croft et al. (1988). In the sections below, we review the different types of volcanic eruptions in the solar system, their commonalities with the Earth, and strive to define what a volcano is in a solar system context.

VOLCANISM ON THE MOON AND TERRESTRIAL PLANETS

The terrestrial planets of our solar system are defined as those that are relatively small in size (compared to the outer giant gas planets) that have rocky surfaces and interiors, including Mercury, Venus, Earth, the Moon (considered a planet for this discussion), and Mars. Jupiter's moon Io also may be considered in this category, but because of its uniqueness it will be discussed separately. All of the terrestrial planets have surface features consistent with volcanic phenomena; however, with the exception of the Earth, there is little evidence of active volcanism, or eruptions in historic times at bodies other than Io. Most of our information on the volcanic histories of the terrestrial planets beyond Earth comes from spacecraft data, particularly images from orbiting spacecraft, along with compositional estimates from a few landers on Venus, the Moon, and Mars, as well as analyses of samples returned from the Moon, and meteorites from the Moon and Mars. In this section we review the evidence for volcanism on the Moon, Mercury, and Mars, discuss the styles and composition of eruptions, and note their similarities to terrestrial volcanism.

Volcanism on the Moon

The Moon is thought to have formed by the collision of a Mars-sized object with the Earth shortly after the Earth's formation, ca. 4.5 Ga (Hartmann and Davis, 1975). The ring of Earth-orbital debris resulting from the collision eventually aggregated to form the Moon, which underwent gravity-induced differentiation to form an anorthositic crust, mafic mantle, and perhaps a small Fe-rich core. Many consider formation of the Moon's feldspathic crust to have resulted from a "magma ocean" that cooled and crystallized after the differentiation event (Solomon and Töksoz, 1973; Taylor and Jakes, 1974). A hypothesized late heavy bombardment of the inner solar system by unaccreted rocky material at ca. 3.9 Ga (Ryder et al., 2000) formed large impact basins on the Moon, thinning the crust on the near side and thickening the crust on the far side.[1] These events enabled eruption of volcanic materials on the Moon. Volcanic features found on the Moon (see Fig. 1) include (1) maria (lava flows); (2) cryptomaria; (3) sinuous rilles; (4) domes, cones, and small shields; (5) dark mantle deposits (DMD); (6) volcanic centers: Aristarchus Plateau–Rima Prinz, Marius Hills; (7) non-mare domes; and (8) light plains.

The lunar *maria* (sing. *mare*, Latin for "seas") cover ~17% of the lunar surface (Head, 1976), and are recognized as massive floodlike lava flow fields that fill impact craters and basins. The maria were mostly emplaced from ca. 3.9 Ga to ca. 3.2 Ga, but with some minor effusive eruptions continuing until ca. 1 Ga (Hiesinger et al., 2000, 2003). Over geologic time the effects of space weathering have destroyed the primary volcanic features such as lava flow boundaries in all but a few cases. Compositional measurements on lava samples returned from the Apollo program, combined with visible-near-infrared (VNIR) spectroscopy from Earth-based telescopes and spacecraft data, indicate that lunar mare lavas are, in general, anhydrous mafic to ultramafic in composition, with higher titanium and iron contents than terrestrial basalts (see Lucey et al., 2006, and references therein), although there is considerable variation in both titanium and iron contents.

Cryptomaria are ancient (>3.8 Ga) mare deposits that have been partially buried or covered by highland impact crater or basin ejecta (Head and Wilson, 1992), and are recognized either by dark-halo craters that excavate darker basalts from underneath brighter crater ejecta (Schultz and Spudis, 1979; Hawke and Bell, 1981), or by spectral unmixing analyses applied to multispectral data (Greeley et al., 1993; Head et al., 1993), or by Apollo orbital geochemistry (Hawke and Spudis, 1980; Hawke et al., 1985). Prominent areas containing cryptomaria include the Schickard (Greeley et al., 1993) and Mendel-Rydberg regions (Williams et al., 1995), and the Lomonosov-Fleming basin and Balmer-Kapteyn region (Giguerre et al., 2003; Hawke et al., 2005). Cryptomaria cover about another 3% of the lunar surface, bringing the total mare cover to ~20% of the Moon.

Lava channels on the Moon are referred to as *sinuous rilles*. They appear as levee-less meandering channels, ~2–300 km in

[1]The Moon, and many of the outer planet satellites, undergo *synchronous rotation*, in which the time required to rotate once on their axes is equivalent to the time required to revolve once around their parent planets. This means, from a point of view on the Earth's surface, that we can see only one side of the Moon in sunlight (the near side); the opposite side is called the far side. The far side can be seen in sunlight only from orbiting spacecraft.

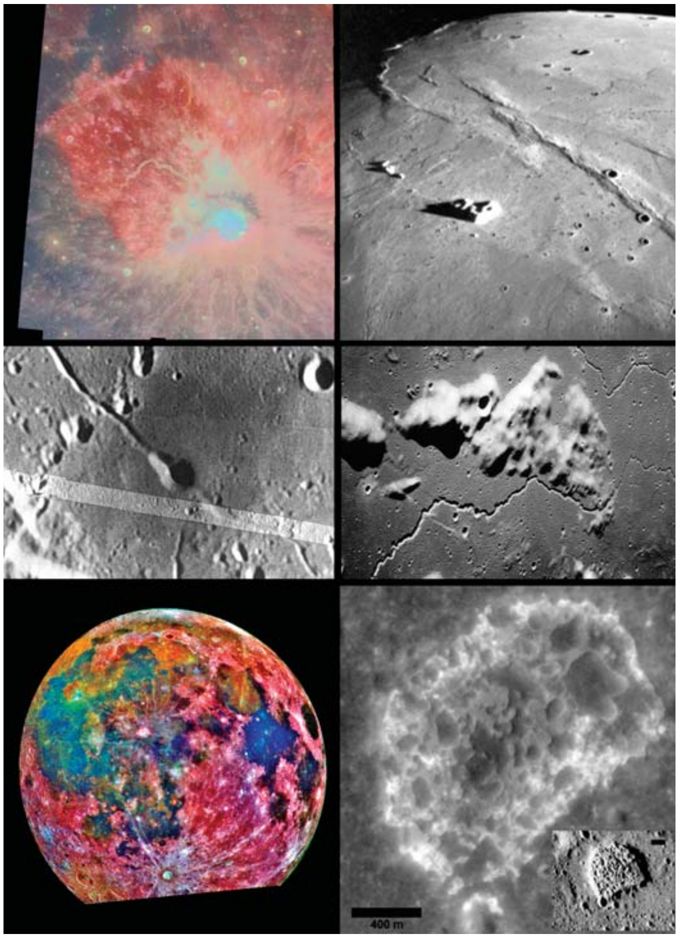

Figure 1. Montage of images of the Moon, showing putative volcanic features. Clockwise from upper left: False color *Clementine* mosaic of the Aristarchus Plateau of the lunar near side, showing the reddish pyroclastic deposit on the plateau, and the sinuous rille (lava channel) known as Schröter's Valley; *Apollo 15* image of well-preserved lava flows in Mare Imbrium; *Apollo* orbiter image of lunar sinuous rilles (lava channels); *Apollo* orbiter images of the Ina structure of the lunar near side, thought to represent a recent gasp of lunar degassing ca. 10 Ma (Schultz et al., 2006); false color *Galileo* mosaic of the lunar nearside, in which orange units represent low-titanium mare deposits, and blue units represent high-titanium mare deposits, derived from ancient lava flows; image of a dark mantle (pyroclastic) deposit in the near side crater Alphonsus.

length, meters to 3 km in width, mean depths of ~100 m, with a craterlike depression at the head of the rille and a fading downslope into the maria at the end of the channel (Greeley, 1971). It is likely that many channels formed by thermo-mechanical erosion of substrate by the very fluid, high-temperature lunar mare lavas (Hulme, 1973; Williams et al., 2000b), although a constructional mechanism cannot be ruled out. Other features observed in the mare resemble collapsed lava tubes on Earth (Greeley, 1971).

There are many volcanic domes, cones, and small shield volcanoes that have been preserved in the maria. Guest and Murray (1976) mapped 80 low domes, which appear as semicircular, broad, convex landforms that range from 2.5 to 24 km in diameter and 100–250 m in height, with slopes of 2°–3°. There is a major concentration of domes in the Marius Hills (Greeley, 1971), which contain some domes with steeper slopes (7°–20°) and summit fissures or craters (Hiesinger and Head, 2006). Cinder cones have been identified in association with linear rilles on the Moon, particularly in crater Alphonsus (Head and Wilson, 1979). These cones are typically 2–3 km wide, and <100 m high, with summit craters <1 km in diameter (Guest and Murray, 1976). There are no shield volcanoes on the Moon larger than ~20 km in diameter, indicating a lack of shallow buoyancy zones, and demonstrating that lava extrusion did not involve low-volume, short-duration eruptions from shallow reservoirs (Hiesinger and Head, 2006). This is a unique aspect of lunar volcanism compared with the other terrestrial planets. Several small shield volcanoes appear in the maria, filling parts of the Orientale impact basin.

Explosive volcanism has occurred on the Moon, as exemplified by deposits of rounded, colored glass beads and black vitrophyric beads found in lunar soil samples. These lunar glasses are thought to result from lava fountaining of gas-rich, low-viscosity mafic magmas erupting in a vacuum (Heiken et al., 1974). These deposits, called *dark mantle deposits* (DMD), include regional deposits (>2500 km^2) on highlands adjacent to major maria, and local deposits that are smaller and widely dispersed (Head, 1976; Gaddis et al., 1985; Coombs et al., 1990). Because the Moon is anhydrous, the volatile gas associated with the formation of the DMD is CO_2 or CO. The youngest proposed volcanic event(s) on the Moon, the outgassing of volatile material at the Ina structure, is estimated to have occurred ca. 10 Ma (Schultz et al., 2006), possibly indicating that the Moon may still be outgassing today.

Two major volcanic centers on the lunar near side include the Aristarchus Plateau–Rima Prinz region and the previously mentioned Marius Hills. These are not individual volcanoes, as they contain multiple vent centers. The Aristarchus Plateau–Rima Prinz region is a positive-relief region ~40,000 km^2 jutting up through the maria of north-central Oceanus Procellarum, built around the Aristarchus and Heroditus impact craters. This region contains the Schröter's Valley sinuous rille and is covered by a regional pyroclastic mantling deposit (Guest and Murray, 1976; Whitford-Stark and Head, 1977). The surrounding Rima Prinz region contains ~36 sinuous rilles within the mare lavas, perhaps indicative of high-volume, high-effusion-rate eruptions that may have filled much of the north-central Oceanus Procellarum. The Marius Hills, in central Oceanus Procellarum, contains >100 mare domes and cones, as well as 20 sinuous rilles (Greeley, 1971). Both regions have multiple types of volcanic deposits, consistent with contemporaneous effusive and explosive eruptions.

Several regions on the lunar near side have features similar to mare domes, but with a much higher albedo, indicative of a non-mare composition. Examples include the Gruithuisen domes, the Marian domes and cones, Hansteen Alpha, and Helmet. These non-mare domes are <20 km in diameter and

>1000 m in relief, with morphologies consistent with viscous lava domes or explosive volcanism. Spectrally, these features have a downturn in the ultraviolet, leading to their reference as "red spots." Although these features have not been sampled, current theories suggest that they may be compositionally similar to terrestrial rhyolites, dacites, or basaltic andesites (Head and Wilson, 1999).

The lunar light plains include regions with a high albedo similar to the lunar highlands, but also with a smooth appearance indicating formation after the end of the late heavy bombardment, suggesting that these regions might represent some form of effusive highlands volcanism. Sampling during the *Apollo 16* mission showed that the Cayley Formation of the light plains is composed of impact breccias and is not related to volcanism. However, other regions of the light plains elsewhere on the Moon may have been produced by volcanism, possibly KREEP (KREEP stands for material rich in potassium [K], rare earth elements [REE], and phosphorus [P], and this material is thought to occur in a layer in the lower lunar crust or upper mantle, exposed by nearside impact basins) volcanism associated with the formation of the Imbrium impact basin (Spudis, 1978; Hawke and Head, 1978).

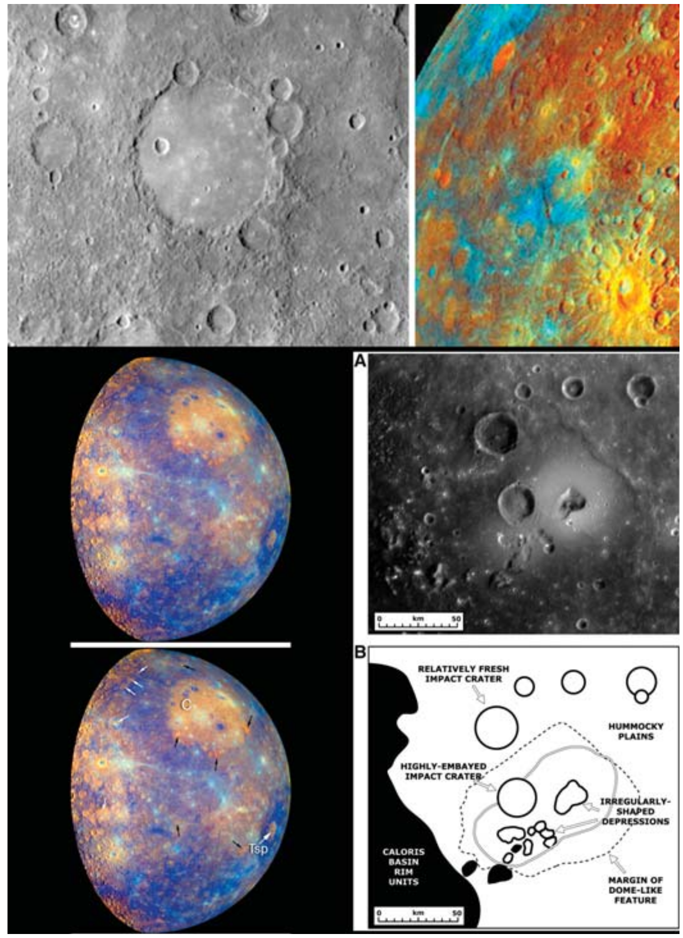

Figure 2. Montage of images from NASA's *Mariner 10* and *MESSENGER* spacecraft of putative volcanic features on Mercury. Clockwise from upper left: gray-scale *Mariner 10* image of larger crater containing smooth plains, inferred to be volcanic flows; false color image in which orange units, thought to be lava flows, filling topographic lows and crater floors (from Robinson and Lucey, 1997); *MESSENGER* image (A) and interpretive sketch map (B) of a putative volcanic caldera and associated deposits (from Head et al., 2008); false color global mosaic of Mercury from the first *MESSENGER* flyby (January 2008), in which white arrows point to inferred ancient lava flows filling depressions (from Robinson et al., 2008). The black arrows indicate red units interpreted to be small volcanic centers. C marks the Caloris Basin, and "Tsp" refers to high-reflectance smooth plains within the Tolstoj basin.

Volcanism on Mercury

Mercury, the innermost planet, is still steeped in mystery because (until recently) over half the planet has remained unobserved up close by spacecraft. NASA's *Mariner 10* spacecraft performed three flybys of Mercury in 1973–1974, and imaged 45% of the surface. In 2004 NASA launched the *MESSENGER* spacecraft, which will begin orbiting Mercury in March 2011. Prior to arriving in Mercury orbit, *MESSENGER* performed three flybys of Mercury, in January 2008, October 2008, September 2009, and all of these flybys were successfully executed. These Mercury flybys enabled another 50% of the surface to be imaged. Because we have incomplete high-resolution imaging and limited spectroscopic and other data of the surface of Mercury, any conclusions about the abundance and nature of volcanism must remain tentative.

Evidence of volcanism on Mercury has previously taken two forms: (1) identification of smooth plains units on the surface, morphologically similar to the lunar maria (Murray et al., 1974a, 1974b); and (2) limited spectroscopic data showing color differences in some plains units, possibly indicative of compositionally distinct lava flows (Robinson and Lucey, 1997). Figure 2 shows examples of these from *Mariner 10* data. New insights are coming from *MESSENGER* data; in particular the color spectral data are showing additional areas of potential volcanic deposits, likely mafic (basaltic) in composition (Robinson et al., 2008). By combining the *Mariner 10* and *MESSENGER* coverage of Mercury (97.7% of the planet), the new data suggest that at least 40% of Mercury's surface is composed of volcanically derived plains (Solomon et al., 2010). New high-resolution images have detected a potential shield volcano in the newly imaged terrain (Head et al., 2008), as well as evidence of widespread explosive pyroclastic deposits in some areas (Solomon et al., 2010). More information on the styles and compositions of Mercurian volcanism can be expected as additional data from the *MESSENGER* orbital mission are obtained.

Volcanism on Mars

Mars is a volcanic planet (Fig. 3). It has the largest shield volcanoes in the solar system; it has vast plains of lava flows; it has many channels, domes, and cones; and it has evidence of

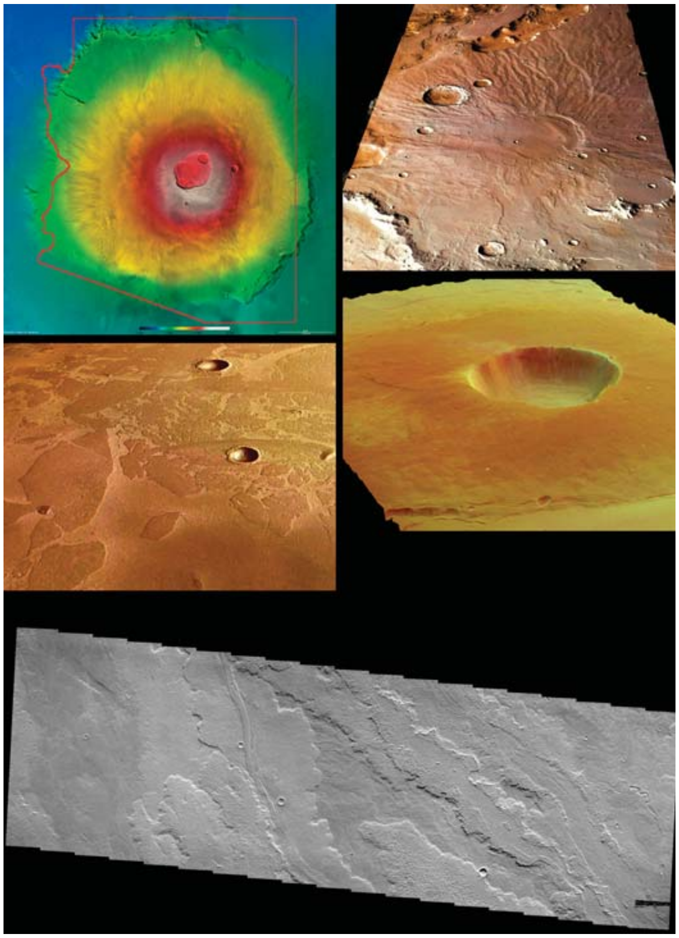

Figure 3. Montage of Mars orbiters images of volcanic features on Mars. Clockwise from upper left: European Space Agency (ESA) *Mars Express* high-resolution stereo camera (HRSC) mosaic of the Olympus Mons volcano in the Tharsis Volcanic Province, color-coded for surface elevation based on HRSC digital terrain model (DTM), with the Arizona border for scale; HRSC color perspective view of Hadriaca Patera, a putative pyroclastic volcano in the Circum-Hellas Volcanic Province; HRSC color perspective view of the Albor Tholus volcano in the Elysium Volcanic Province; NASA *Mars Odyssey* thermal emission imaging system (THEMIS) image of channel-fed lava flows in Alba Patera, north of Tharsis; HRSC color perspective view of platy lava flows in the Elysium Volcanic Province. Image credit for HRSC images: ESA/DLR/FU Berlin (G. Neukum). Image credit for THEMIS image: NASA/JPL/ASU.

explosive volcanism and extensive pyroclastic deposits (Greeley and Spudis, 1981; Greeley et al., 2000; Carr, 2006). Because Mars has been the primary focus of planetary exploration since the mid-1990s, and because Mars has been (successfully) visited by six orbiters and six landers or rovers as of 2008, there is a vast quantity of data available on Martian surface features (at resolutions from 30 cm/pixel to several kilometers/pixel), surface compositions and role of volatiles, atmospheric dynamics and composition, and ionospheric composition and interaction with the solar wind. Only the composition and nature of the Martian interior have not been well investigated by post-*Viking* missions. In this section we will briefly review the diversity of volcanic features on Mars.

Although volcanic features are present ubiquitously across the planet, Martian volcanism was centered in three distinct provinces: Tharsis, Elysium, and Circum-Hellas. The Tharsis volcanic province occurs on the Tharsis bulge, is ~8000 km in diameter, and covers 25% of the Martian surface. This province includes the Olympus Mons shield volcano; three smaller shield volcanoes named Arsia, Pavonis, and Ascraeus Montes; their associated rift aprons and small shield fields (Crumpler and Aubele, 1978; Bleacher et al., 2007a); several smaller (<200 km diameter) features named *tholi* (sing. *tholus*, Latin for "dome") and *paterae* (sing. *patera*, "an irregular crater, or a complex one with scalloped edges," similar to terrestrial calderas); and fields of interconnecting lava plains. Olympus Mons is >500 km across and 25 km high, with maximum slopes of 5°. It is dominated by tube-fed and channel-fed lava flows of probable basaltic composition (Bleacher et al., 2007b), with evidence of glacial ice on top of the volcano at times in the past (Neukum et al., 2004), and a series of nested calderas >60 km across. A unique volcanic structure in north Tharsis is Alba Patera, which covers an area larger than Olympus Mons but has flank slopes of <1°, thus lacking the relief of shield volcanoes. Alba Patera is heavily fractured, indicating the influence of regional stress patterns, and is antipodal to the Hellas impact basin, perhaps indicative of a link in their formation. Alba Patera contains two discrete calderas, channel- and tube-fed flows in some regions, and more sheetlike flows in others, as well as evidence of pyroclastic activity. All of the volcanoes in Tharsis (and Mars as a whole) have been eroded or modified by fluvial, glacial, and/or eolian activity.

The Elysium volcanic province is another central-vent volcanic region on Mars, ~3500 km in diameter and located ~75° of longitude west of Tharsis. This province is dominated by three smaller volcanoes: Elysium Mons, Hecates Tholus, and Albor Tholus, and their surrounding lava flow fields. Elysium Mons is 500 by 700 km wide and ~13 km high, and it contains well-defined channel-fed flows and flank flows (Greeley et al., 2000). In contrast, Hecates Tholus is ~160 by 175 km wide and ~6 km high, in which its flank is heavily dissected by linear, radial shallow valleys thought to be fluvial channels formed from the melting of ice by underground magma sources (Fassett and Head, 2006). The dissected nature of Hecates Tholus suggests that it may be composed of ash or other easily eroded pyroclastic deposits (Mouginis-Mark et al., 1982), as opposed to Elysium Mons, which appears to be composed of effusive flows.

The Circum-Hellas Volcanic Province covers >4.9 million km^2 of the surface and, as its name suggests, is dominated by patches of volcanoes and their associated flow fields surrounding the Hellas impact basin (Williams et al., 2009). These include the well-known, low, shield-like edifices Tyrrhena Patera (and the surrounding, putative flow field of Hesperia Planum), Hadriaca Patera, and Amphitrites Patera, as well as three caldera-like depressions, Peneus, Malea, and Pityusa Paterae, and the putative lava plains that make up Malea Planum and parts of the Hellas rim region. The heavily channeled and dissected shields of Tyrrhena, Hadriaca, and Amphitrites Paterae suggest that these volcanoes are composed of easily erodible, friable pyroclastic deposits rather than lava flows (Greeley and Crown, 1990; Crown and Greeley, 1993). The caldera-like depressions and surroundings of Peneus, Malea, and Pityusa Paterae suggest that they may have been similar to terrestrial "supervolcanoes," and possibly produced ignimbrite deposits that have since been modified by fluvial, eolian, and periglacial processes (Williams et al., 2009). Ages based on crater counts suggest that most of the volcanic activity in the Circum-Hellas Volcanic Province occurred between ca. 3.5 and 4.0 Ga, with little evidence that any activity occurred more recently than 1 Ga.

The compositions of Martian volcanic features have been assessed by Earth-based, orbital, and surface visible and near-infrared spectroscopy, orbital and surface thermal infrared spectroscopy, in situ measurements of volcanic rocks and soils by landed spacecraft, and study of the Martian meteorites. These studies show that most of Mars' volcanic deposits are mafic in composition, consistent with terrestrial basalts (McCord et al., 1982) to basaltic andesites (Bandfield et al., 2000). Measurements of the basalts filling Gusev Crater by the Mars Exploration Rover *Spirit* suggest that these lavas are likely high-magnesium basalts (~11–12 wt% MgO: McSween et al., 2004; Gellert et al., 2004). In contrast, analysis of the rock "Barnacle Bill" at the 1997 NASA *Mars Pathfinder* landing site in Ares Vallis suggested a composition slightly more silicic, perhaps an "icelandite" (McSween and Murchie, 1999). Elsewhere on Mars, orbital measurements by the thermal emission imaging system (THEMIS) on NASA's 2001 *Mars Odyssey* orbiter have identified outcrops of more silicic compositions, likely dacites in Syrtis Major (Christensen et al., 2005).

Despite the widespread occurrence of volcanic features on Mars, there is limited evidence of active or recent volcanism. Crater counts of the youngest lava flows on the Tharsis shield volcanoes yield ages of a few million years to a few tens of millions of years; the central calderas have ages of 100 Ma to <1 Ga (Neukum et al., 2004). Both Earth-based instruments and the planetary fourier spectrometer (PFS) on the European Space Agency *Mars Express* orbiter have detected methane in the Martian atmosphere, with concentrations in discrete locations. These results have been interpreted to signify either volcanic outgassing or perhaps biological activity (Encrenaz, 2008). Nevertheless, despite 10 yr of imaging the surface of Mars with orbiting spacecraft, no evidence of surface changes caused by volcanic activity has been observed (Malin et al., 2006).

Volcanism on Venus

The surface of Venus is covered by volcanic plains and a variety of volcanic features. There are many superposed volcanic plains units that cover hundreds of kilometers across, providing evidence for multiple episodes of vast outpourings of lava (Head et al., 1992). Thousands of volcanoes have been identified on Venus, some of which are the most unusual volcanoes in the solar system (Stofan, 2004). Sizes range from a few kilometers to flood lavas covering hundreds of kilometers across. Most of the data used for the study of volcanic features on Venus comes from the *Magellan* spacecraft, which carried a radar that could penetrate the thick atmosphere of the planet.

Small volcanoes (<10 km in diameter) include shields and cones similar in morphology to terrestrial types. Shields are more common than cones and tend to occur in clusters or groups. Many of the shields and cones have summit pits, indicating that either collapse into a magma chamber occurred or else small explosions took place (Guest et al., 1992). The great atmospheric pressure on Venus (90 bars at the surface), combined with lack of water, inhibits explosive activity, and large explosive eruptions are unlikely.

Intermediate sized volcanoes (10–100 km across) are fewer in number but are still common on the surface (Crumpler et al., 1997). These include shields that are surrounded by long lava flows that have similar radar brightnesses (and likely surface textures) to basaltic lava flows on Earth (Stofan, 2004). Despite higher surface temperatures on Venus (~750 °C), basaltic lavas should form a crust relatively rapidly after the eruption (Bridges, 1997) and therefore should produce surface textures similar to those on Earth.

Volcanoes larger than ~100 km in diameter (Fig. 4) are relatively few in number and are surrounded by extensive flows (Head et al., 1992; Stofan et al., 2001a). These larger volcanoes tend to be found on top of broad topographic rises, perhaps similar to volcanoes on Earth atop hotspots. The hotspots on Venus appear to be randomly distributed, with no alignments that might indicate plate boundaries. The larger Venusian volcanoes have much larger volumes than volcanoes on Earth, probably owing to the lack of plate tectonics on Venus (Stofan et al., 1995), which allows volcanoes to remain stationary over a hotspot.

Some of the larger volcanoes have broad summit calderas, whereas others have sets of radial fractures that could have been caused by diking. Lava flows on Venus can extend for hundreds of kilometers, and lava channels can also be much larger in size than those on Earth, some also hundreds of kilometers long (Baker et al., 1992). Lavas on Venus are thought to have been of very low viscosity to form these features. Several types of compositions have been proposed for these lavas, including carbonate-rich lavas, sulfur-rich lavas, and ultramafic lavas. The unusual length of some channels also could be due to thermo-mechanical erosion of the surface, similar to the mechanism proposed for rilles on the Moon. Vast expanses of lava flows are common on Venus and are thought to have been emplaced as flood lavas (Lancaster et al., 1995).

Venus has several types of volcanic constructs that are morphologically unusual compared with what is seen elsewhere in the solar system. Among these are steep-sided domes, also known as pancake domes, which are flat-topped and have steep margins (Pavri et al., 1992). Some of these domes show collapsed,

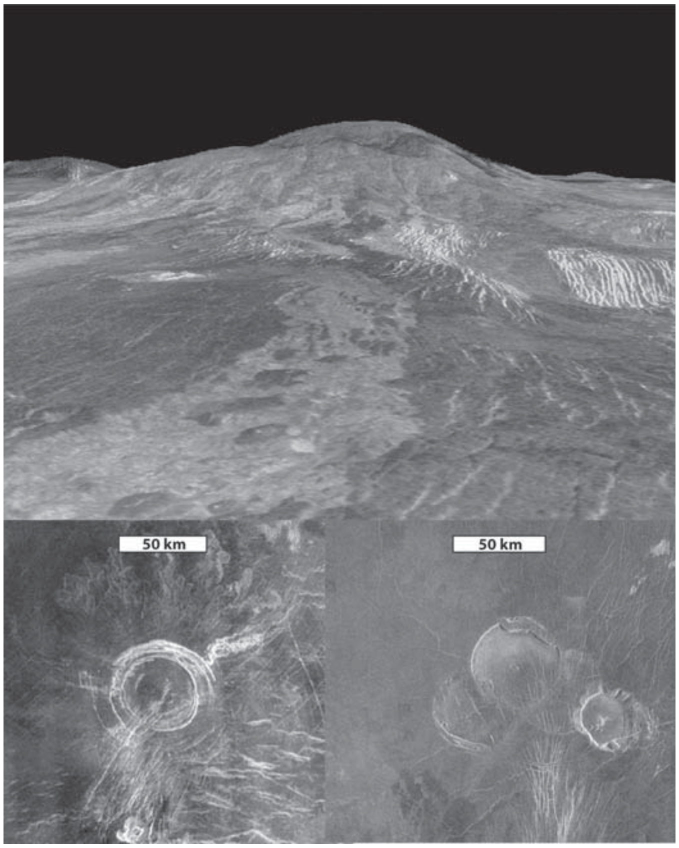

Figure 4. Montage of Venus images from *Magellan*. Top: the volcano Sif Mons on Venus is seen in this computer-generated view of the surface. Synthetic aperture radar (SAR) and altimetry data were combined to produce this 3-D view, with simulated color and vertical exaggeration to show detail. Sif Mons is 350 km in diameter and ~2 km high. Note the lava flows in the foreground. Lower left: Serova Patera has a broad caldera and is surrounded by lava flows. The patera has been fractured by tectonic activity. Lower right: flat-topped volcanoes ("pancake domes") have steep sides and a flat top. The volcano at the lower right has scalloped edges, probably the result of multiple collapses along its margins.

scalloped margins and are known as "ticks" (Stofan et al., 2000; Guest et al., 1992). Steep sided domes may have been formed by slow eruptions of relatively high viscosity lavas.

A relatively common, morphologically unusual feature on Venus is the corona. Coronae are large (>100 km across) circular features that show concentric ridges. Over 400 coronae have been identified on Venus, the largest being 2,500 km across (Stofan et al., 2001b). They are thought to have been formed by rising magma deforming the surface, producing concentric ridges. Many coronae have shields in their interiors and are surrounded by long lava flows. Coronae appear to have been long-lived features, similar to volcanoes formed over hotspots on Earth.

The Venusian surface is very young, as there are relatively few impact craters, and estimates of surface ages range from 300 million to 1 billion yr, similar to the oldest continental areas on Earth (McKinnon et al., 1997). Recent data from Venus Express (Visible and Infrared Thermal Imaging Spectrometer) identified compositional differences in lava flows at three locations, interpreted as a lack of surface weathering (Smrekar et al., 2010). These results suggest that the flows are younger than 2.5 million years and probably much younger, about 250,000 years or less, indicating that Venus is actively resurfacing. Other evidence that active volcanism may still be happening on Venus comes from measurements of declining SO_2 in the atmosphere obtained in 1978 (Esposito, 1984). No surface changes were detected by the *Magellan* radar during the mission (1990–1994), though changes could have happened below the resolution of the instrument (~120 m). The question of current activity on Venus remains open. In conclusion, all indications at present are that Venusian magma is silicic, possibly basaltic, though more exotic compositions such as carbonatites have been suggested.

ACTIVE VOLCANISM ON IO

Volcanic features dominate Io's surface, and Io's volcanoes cover a wide range of sizes and present varying characteristics such as power output, persistency of activity, and association with plumes. Active volcanism on Io was discovered in 1979 when the Voyager spacecraft revealed active plumes (Morabito et al., 1979) and thermal anomalies (Pearl et al., 1979). Results from the *Galileo* mission, which observed Io from 1995 to 2001, substantially advanced our understanding of volcanism on Io (see reviews by Lopes and Williams, 2005; Lopes and Spencer, 2007).

Most of Io's volcanoes manifest themselves as caldera-like depressions (Fig. 5), referred to as paterae (Lopes et al., 2004). Unlike terrestrial volcanoes, Io's volcanoes rarely build large topographic structures such as shields or stratovolcano-like mountains. There are only a few structures, called *tholi*, scattered across Io. Some of these features were suggested to be evidence of possible shield-building basaltic volcanism (Schaber, 1980). Paterae are the most common type of volcanic feature on Io. Although the origin of paterae is still somewhat uncertain, they are thought to be similar to terrestrial volcanic calderas, formed by collapse over shallow magma chambers following partial removal of magma. Some paterae show angular shapes that suggest some structural control, indicating that they may be structural depressions that were later used by magma to travel to the surface. At least 400 Ionian paterae have been mapped. Their average diameter is ~40 km, but Loki, the largest patera known in the solar system, is >200 km in diameter. In contrast, the largest caldera on Earth, Yellowstone, is ~80 km by 50 km in size. The larger sizes of the Ionian features probably reflect the much larger sizes of magma chambers, which are thought to be relatively shallow (Leone and Wilson, 2001).

Io's surface shows some remarkably large lava flows (*fluctii*, sing. *fluctus*, meaning "flow field"). The lava flow field from Amirani is ~300 km long, the largest active flow field known in the solar system. Io's large lava flows are possibly analogues of the continental flood basalt lavas on Earth, such as the Columbia

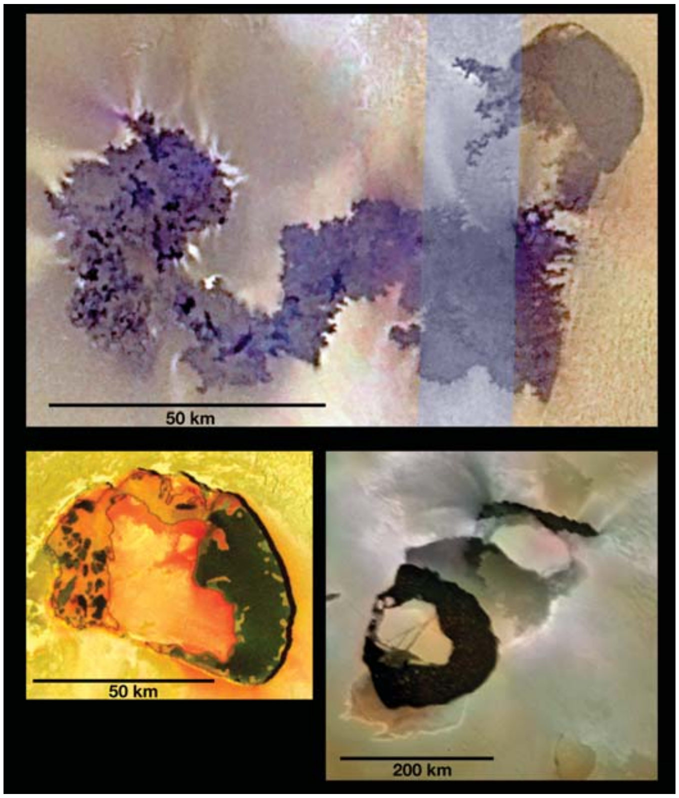

Figure 5. Montage of Ionian paterae. Top: Prometheus Patera and its lava flow field (*Galileo* image). Lower left: Tupan Patera (*Galileo* image). Lower right: Loki Patera (*Voyager* image).

River Basalts in the United States. Repeated imaging of Amirani during the *Galileo* flybys allowed eruption rates to be estimated (50–500 m^3s^{-1}, Keszthelyi et al., 2001). These are considered moderate effusion rates, and the ability of lava to travel large distances at moderate effusion rates suggests not only that lavas had a low viscosity but also that they were emplaced as insulated flows, so that the cooled crust would insulate the hot material underneath. Thermal profiles along the Prometheus and Amirani flows (Lopes et al., 2001, 2004), and high spatial resolution images (Keszthelyi et al., 2001), suggest that these large Ionian flows were similar to terrestrial inflated pahoehoe flows (Hon et al., 1994).

A major question about Ionian volcanism after *Voyager* was the nature of volcanism—whether sulfur or silicates were predominant. Although temperature measurements from *Galileo* clearly showed that many active volcanoes have temperatures far too high for sulfur, the possibility that some sulfur flows occur on the surface cannot be ruled out. The colorful flows around Ra Patera that were argued by Sagan (1979) to be sulfur had been covered over by new eruptions before *Galileo*'s first observations of the area in 1996 (McEwen et al., 1998a). However, other places may have sulfur flows. Although most Ionian flows appear dark, a few show pale yellow or white flows that may well have been molten sulfur. Williams et al. (2001a) proposed that flows radiating from Emakong Patera may be sulfur and that low-temperature liquid sulfur heated to 450 K could explain many of the morphological features seen around Emakong, such as a meandering channel 105 km in length that appears to feed a gray-white flow some 270 km in length. Temperature measurements using the near-infrared mapping spectrometer (NIMS) (Lopes et al., 2001, 2004) indicate temperatures <400 K inside Emakong caldera, and much cooler (below the instrument's detection capabilities) over the flows. Another bright flow field, Tsui Goab Fluctus, is similarly the location for a low-temperature hotspot and could represent effusive sulfur volcanism (Williams et al., 2004). However, the *Galileo* instruments could not distinguish between sulfur flows or cooled silicates coated by bright sulfurous materials after erupting. One possibility suggested by Greeley et al. (1984), based on studies of a yellow sulfur flow at Mauna Loa, Hawaii, is that rising silicate magma may melt sulfur-rich country rock as it nears the surface, producing "secondary" sulfur flows (as opposed to "primary" flows that originate from molten magmas at depth). The widespread occurrence of sulfurous deposits detected around Ionian hotspots as seen by *Galileo* provide more evidence that this may be the case. Therefore, the presence of sulfur flows on Io remains open.

Inferring the composition of Io's lavas has been problematic, as no high-resolution spectroscopic measurements could be taken (Lopes et al., 2001). With no direct measurements of lava composition, temperatures detected at active hotspots provide the best clues to magma composition. Temperatures can be calculated by measurements made from two *Galileo* instruments, the Solid-State Imaging System (SSI) and NIMS. The wavelength range used for the temperature determinations was 1.1–5.1 µm. Temperatures determined by remote sensing data depend on the spatial resolution and wavelength range, and because of the rapid cooling of lava, they have to be considered minimum temperatures (e.g., Lopes et al., 2001; Kargel et al., 2003). One Ionian hotspot, Pillan, showed temperatures above 1800 K during an eruption in 1997, which was detected by both NIMS and SSI (McEwen et al., 1998b; Davies et al., 2001; Williams et al., 2001b), although recent analysis suggests that this number could be as low as ~1600 K (Keszthelyi et al., 2007). This temperature far exceeds terrestrial basaltic lava temperatures, which rarely exceed 1450 K, and mantle temperatures on Io, which may be as high as ~1570 K (Keszthelyi et al., 2007). Temperatures measured at other hotspots could be consistent with either basaltic or ultramafic temperatures (e.g., Kargel et al., 2003).

At present it is not known whether lavas on Io are erupted at very high temperatures or not, since it is difficult to detect sufficiently large areas where very fresh (and therefore very hot) materials are exposed. If lavas are indeed hotter than basalt, it is also not known what is causing the very high temperatures. Two main hypotheses have been put forward: ultramafic volcanism (McEwen et al., 1998b; Williams et al., 2000a) and superheated basaltic volcanism (McEwen et al., 1998b; Kargel et al., 2003; Keszthelyi et al., 2007). Color data on Io's dark volcanic materials obtained from SSI and NIMS (Geissler et al., 1999) indicate the presence of orthopyroxene very rich in Mg, requiring ultramafic bulk compositions (but apparently lacking in olivine). The only occurrence of orthopyroxene phenocryst-bearing komatiites on Earth is in the Commondale greenstone belt, South Africa, which had inferred liquidus (eruption) temperatures of ~1884 K (Williams et al., 2000a). These are the closest likely analogues to the lavas erupted at Pillan and perhaps at other Ionian hot spots identified thus far.

A second hypothesis to explain Io's hottest eruptions is superheating, which was also discussed by McEwen et al. (1998b). Magma can be heated considerably by ascent from a deep source if driving pressures are high. This effect is small on Earth, but the crust of Io is under considerable compression, which leads to significant driving pressures; assuming adiabatic ascent, this can result in 50–100 K heating, possibly greater (Keszthelyi et al., 2007). Melting temperatures of dry silicate rocks increase with pressure; therefore the erupted lava can be significantly hotter than its melting temperature at surface pressure. This can provide a partial explanation for the high temperatures but does not, on its own, account for the ~1800 K observation (with liquidus temperatures of ~1400–1550 K at 1 bar) and must be combined with higher-than-terrestrial mantle temperatures and errors in the original observation (Keszthelyi et al., 2007). Furthermore, if the magma were to contain even small quantities of juvenile volatiles—of an order 0.1 wt%—significant cooling would occur during ascent that could reverse or, if volatile contents were high enough, exceed the superheating effect. Similarly, if ascent were sufficiently slow, loss of heat through wall rocks could also reduce the temperature (Kargel et al., 2003). At present, the question of whether Io's lavas are mafic or ultramafic remains open.

Another possible magma on Io is SO_2. The higher-spatial-resolution NIMS data provided the opportunity to map SO_2 distribution at local scales (e.g., Lopes-Gautier et al., 2000; Lopes et al., 2001; Douté et al., 2002, 2004). A nearly pure SO_2 region (corresponding to a 90%–100% concentration of SO_2, Lopes et al., 2001) was found to be topographically confined within the Baldur caldera (see Fig. 6). Smythe et al. (2000) suggested emplacement by liquid flow of SO_2 rising from the subsurface. Although liquid would normally boil off when exposed to Io's tenuous atmosphere, estimates by Smythe et al. (2000) suggest

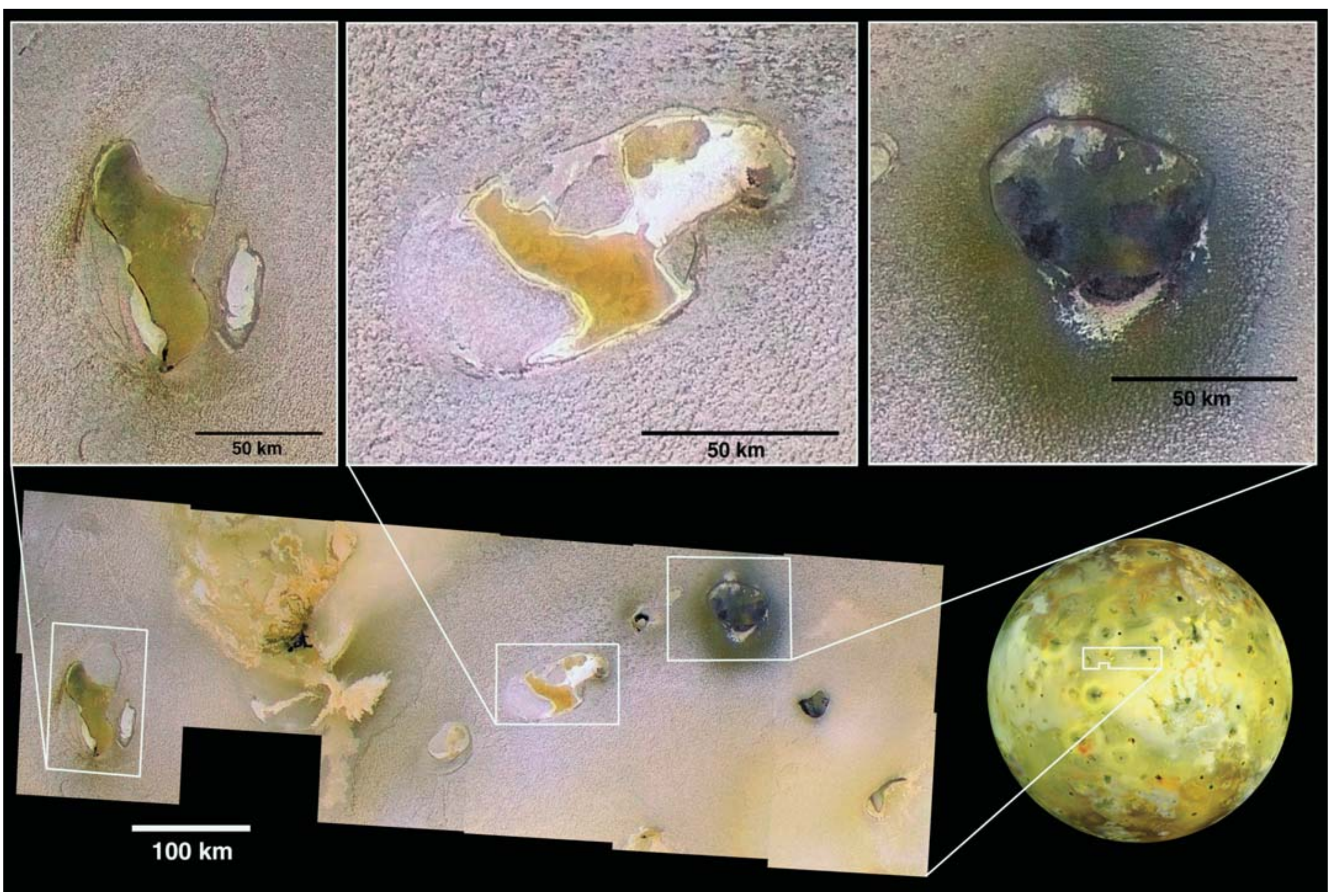

Figure 6. Paterae of the Chaac-Camaxtli region of Io. Upper left: Chaac Patera and neighboring Balder Patera (white floor); upper center: Ababinili Patera, with a complex floor; upper right: Camaxtli Patera, with fresh dark flows and white and dark diffuse material on its margins.

that, given sufficiently large quantities, some of the liquid could freeze to form a layer of sulfur dioxide ice inside the caldera. Another candidate location for an SO_2 flow field, on the southwest part of Tohil Patera, was discussed by Williams et al. (2004).

In conclusion, the primary magma on Io is silicate, possibly basaltic or ultramafic. Although eruptions of sulfur and SO_2 have been suggested and are possible, these can be considered secondary forms of volcanism on Io.

CRYOVOLCANISM ON THE OUTER SOLAR SYSTEM

Several satellites of the outer planets present evidence of eruptions of material (including solids, liquids, and gases) from the interior. The processes that involve eruptions of material in the icy bodies of the outer solar system are closely analogous with volcanic processes of the terrestrial planets, although differences remain; in particular, a buoyant rise of cryomagmas from the interior to the near-surface seems less likely, owing to the high density of liquid water relative to solid water ice (more specifically, ice-I), the main constituent of the ice shells. The differences in the respective geological and eruptive activities of icy satellites are produced by the different conditions as thermal-orbital evolution and internal composition. Because of the analogy with volcanism, we call the eruptive processes of icy satellites *cryovolcanism,* after Croft et al. (1988).

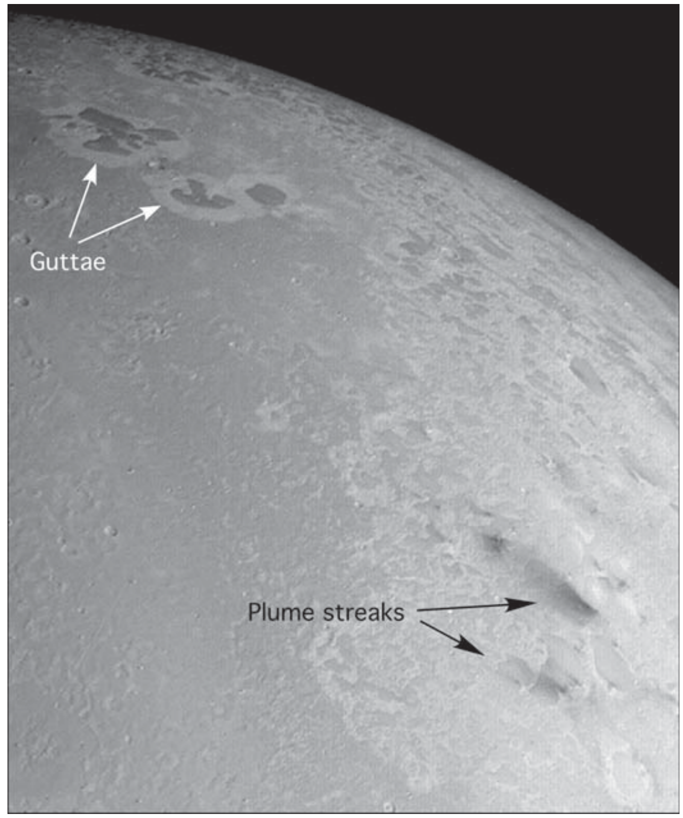

Figure 7. Neptune's moon Triton has several features interpreted to be volcanic, including guttae, which are lobate features with a dark core surrounded by a bright annulus (Prockter, 2004). Dark plume streaks are seen over the bright south polar cap. The image is ~1500 km across.

Neptune's icy satellite Triton (mean radius, 1353 km) presents a complex surface geology (Fig. 7). Eruptive plumes up to 8 km high were observed (Smith et al., 1989; Soderblom et al., 1990). Smith et al. (1989) suggested that the gas venting may be driven by solar heating and the subsequent vaporization of subsurface nitrogen. Other mechanisms for gas venting were proposed as solid state greenhouse (Brown et al., 1990) and convection in the solid nitrogen caps (Duxbury and Brown, 1997). Also Triton shows evidence of resurfacing processes (Smith et al., 1989) possibly by cryovolcanism (Croft, 1990; Schenk, 1992). The resurfaced terrains could be the surface manifestation of high heat production in the interior. A possible explanation was proposed by McKinnon (1984). The retrograde orbit around Neptune suggests that Triton was likely a captured satellite (McKinnon, 1984). McKinnon (1984) showed that as a consequence of the capture of Triton in the Neptune system, Triton may have undergone large tidal heating and melting of the interior ices. On Triton's surface, circular features were observed as in cantaloupe terrains, interpreted as caused by diapirism or upwelling of material from the interior (Schenk, 1992).

Despite its small dimensions, the satellite of Saturn, Enceladus (mean radius 252 km), presents geological activity (Porco et al., 2006) that could result from tectonic resurfacing and cryovolcanism (Squyres et al., 1983; Kargel and Pozio, 1996). The present-day eruptive activity of this small icy moon is concentrated at its south pole (Spencer et al., 2006). The south polar regions are tectonically disrupted (Fig. 8) and present four linear subparallel predominant fractures of the ice shell (Porco et al., 2006). The tectonically disrupted south polar regions are delimited by sinuous escarpments (Porco et al., 2006). The south polar regions present a higher surface temperature (114–157 K) with respect to the surrounding regions (70–80 K) with an estimated thermal emission of 3–7 GW (Spencer et al., 2006). An active venting of water vapor and icy particles (Fig. 8) was observed at the south polar area (Porco et al., 2006; Spahn et al., 2006; Waite et al., 2006). Porco et al. (2006) proposed that the water vapor venting might result from a local reservoir of liquid water in the ice shell, requiring considerable local heat (273 K) in the near surface. However, Kieffer et al. (2006) argued that the high concentrations of N_2 and CH_4 in the plume were beyond what seemed likely in eruption of an aqueous solution, instead preferring decomposition of solid clathrates at lower temperatures, into gas, ice, and vapor, and released by tectonic fracturing. Nimmo and Pappalardo (2006) show that diapiric activity at the south pole can explain the localized geological activity of Enceladus. However, Collins and Goodman (2007) showed that the geological activity of the south pole could result from a partial melting of the ice shell. Tidal internal heating, likely localized in a thermal plume or diapir, could explain the formation of partial melting of the ice shell and the high surface temperature of the south polar regions (Mitri and Showman, 2008a, 2008b; Tobie et al., 2008).

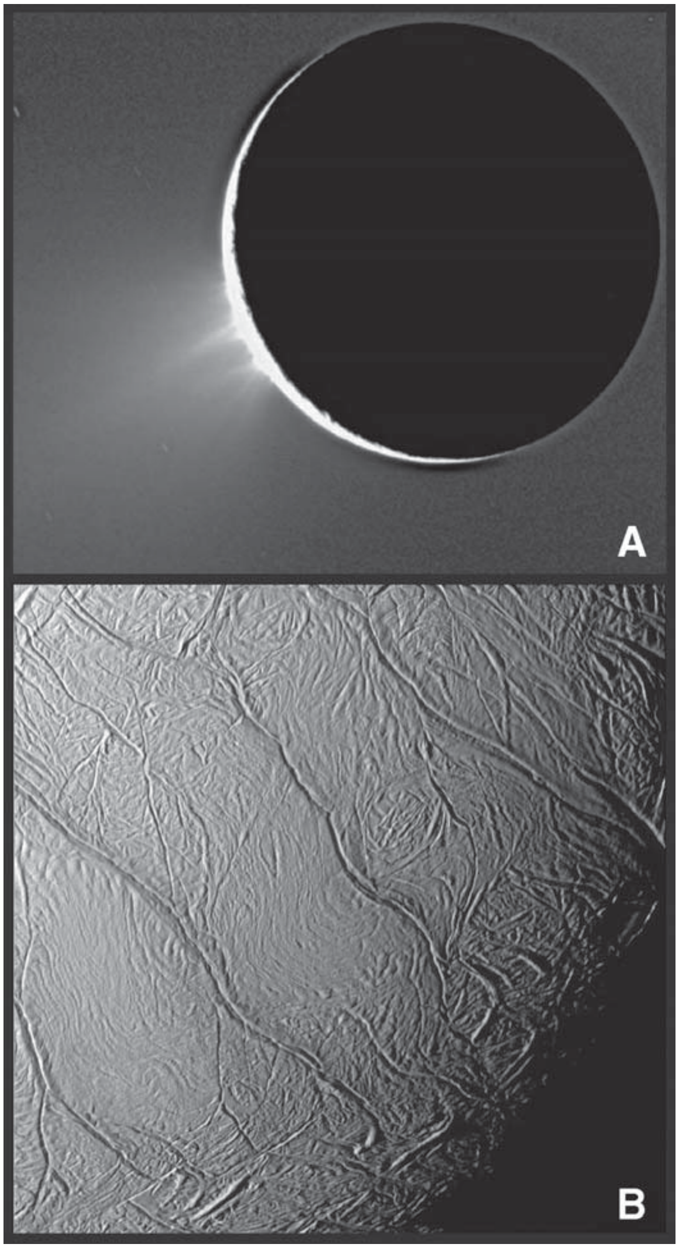

Figure 8. (A) Active venting of water vapor and icy particles observed at the south polar area of Enceladus. From Porco et al. (2006). (B) South polar regions of Enceladus. These south polar regions are tectonically disrupted and present four linear subparallel predominant fractures. An active venting of water vapor and icy particles was observed at the south pole of Enceladus. Image is ~125 km across.

The surface of Jupiter's satellite Europa (mean radius, 1569 km) presents numerous pits, domes, and mottled features (3–30 km in diameter) as well as larger chaotic terrains (Pappalardo et al., 1999; Greenberg et al., 1999). An example is the Conamara Chaos, which presents tectonic disruption of the preexistent terrains surrounded by a not smooth matrix. Thrace Macula (Fig. 9) is a second example of chaotic terrain, which also presents flowlike features that appear to embay ridged terrains with confinement by topography (Miyamoto et al., 2005). It was proposed that the terrain matrix of Thrace Macula is the result of disruption or degradation of preexistent terrain (Collins et al., 2008) or is the result of emplacement of viscous fluid (ice or slush) on the surface (Fagents, 2003; Miyamoto et al., 2005). Fagents (2003) proposed possible mechanisms to form chaotic terrains on Europa that involve eruption of water ice from the interior. However, other geological processes can explain the chaos formation on Europa as melt-through of the ice shell (Greenberg et al., 1999; Thomson and Delaney, 2001; O'Brien et al., 2002), as the surface manifestation of diapirism (Pappalardo et al., 1998; Pappalardo and Barr, 2004), as thermal convection in the ice shell (Pappalardo et al., 1998; Showman and Han, 2005), as spatial localization of tidal heating within a thermal plume in the ice shell (Sotin et al., 2002; Tobie et al., 2003; Mitri and Showman, 2008a), as brine mobilization (Head and Pappalardo, 1999), and as propagation of crevasses at the base of the ice shell to the surface (Crawford and Stevenson, 1988).

Before the *Cassini-Huygens* mission, various workers suggested that Titan, Saturn's largest moon (mean radius of 2575 km) may be cryovolcanically active, on the basis of geochemical and geophysical models, and made possible by substantial quantities of ammonia in the interior, which facilitate a subsurface liquid layer, or subsurface ocean. The dominant paradigm is one of ammonia-water cryomagma (e.g., Croft et al., 1988), and rheological studies of ammonia hydrates (Kargel et al., 1991) suggest that eruptants could be quite viscous, comparable with terrestrial basalts or basaltic andesites, contrasting with the runnier brines proposed for the Jovian icy satellites. Detection of ^{40}Ar (Niemann et al., 2005) in the atmosphere lends strong credence to the case, as it implies outgassing from Titan's interior. Various features observed by *Cassini* on Titan have been interpreted as evidence of cryovolcanic activity. Putative cryovolcanic features detected by radar include lobate flows (e.g., Fig. 10) of varying apparent thicknesses with relatively uniform radar properties (Elachi et al., 2005; Lopes et al., 2007; Wall et al., 2009). Figure 10 shows a putative flowlike feature. The visible and infrared mapping spectrometer (VIMS) observed brightening at two places, which is inconsistent with changes in cloud cover and may be due to ongoing cryovolcanic activity (Nelson et al., 2009a, 2009b). At one of these sites the Hotei Arcus region (26° S, 79° W), high-contrast, flowlike features embay surrounding terrains and channels. Wall et al. (2009) proposed that the observed features are likely cryomagma flows, although fluvial-sedimentary deposits can offer a second explanation. VIMS also provided observations of features interpreted to be cryovolcanic on morphological grounds (Sotin et al., 2005; Barnes et al., 2006). To explain the first observations of putative cryovolcanic activity, Fortes et al. (2007) proposed a model based on the assumption that large amounts of sulfur leached into the internal ocean, differing from the ammonia hydrate paradigm. However, the data from the *Cassini* radiometer (Paganelli et al., 2007) seem inconsistent with large amounts of ammonium-sulfate compounds on the surface

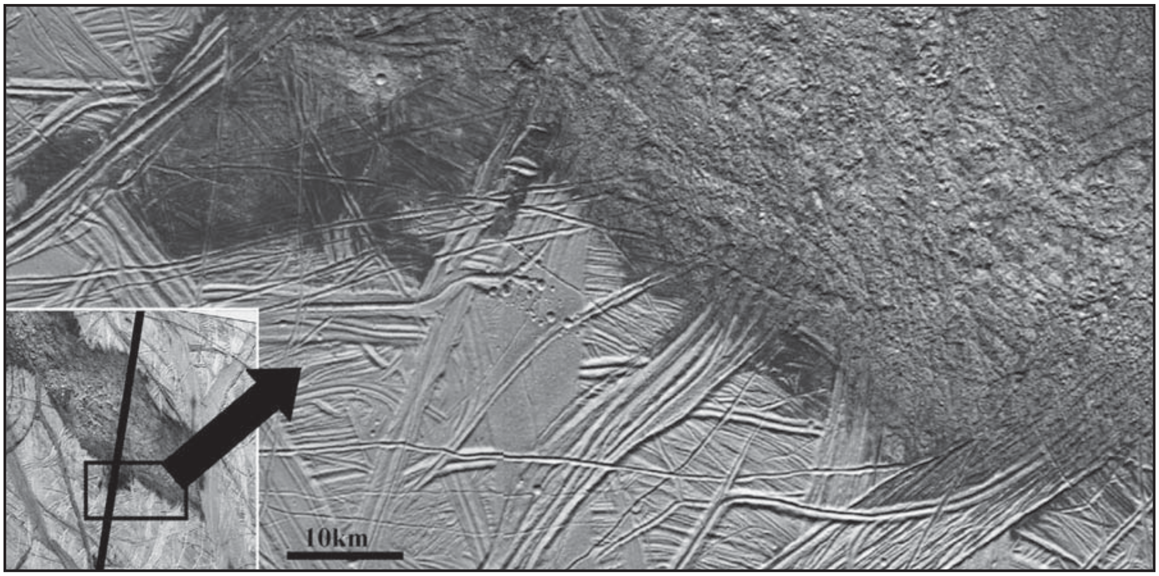

Figure 9. Thrace Macula on Europa (42° S, 172° W). This region presents flowlike features that appear to embay ridged terrains with confinement by topography. From Miyamoto et al. (2005).

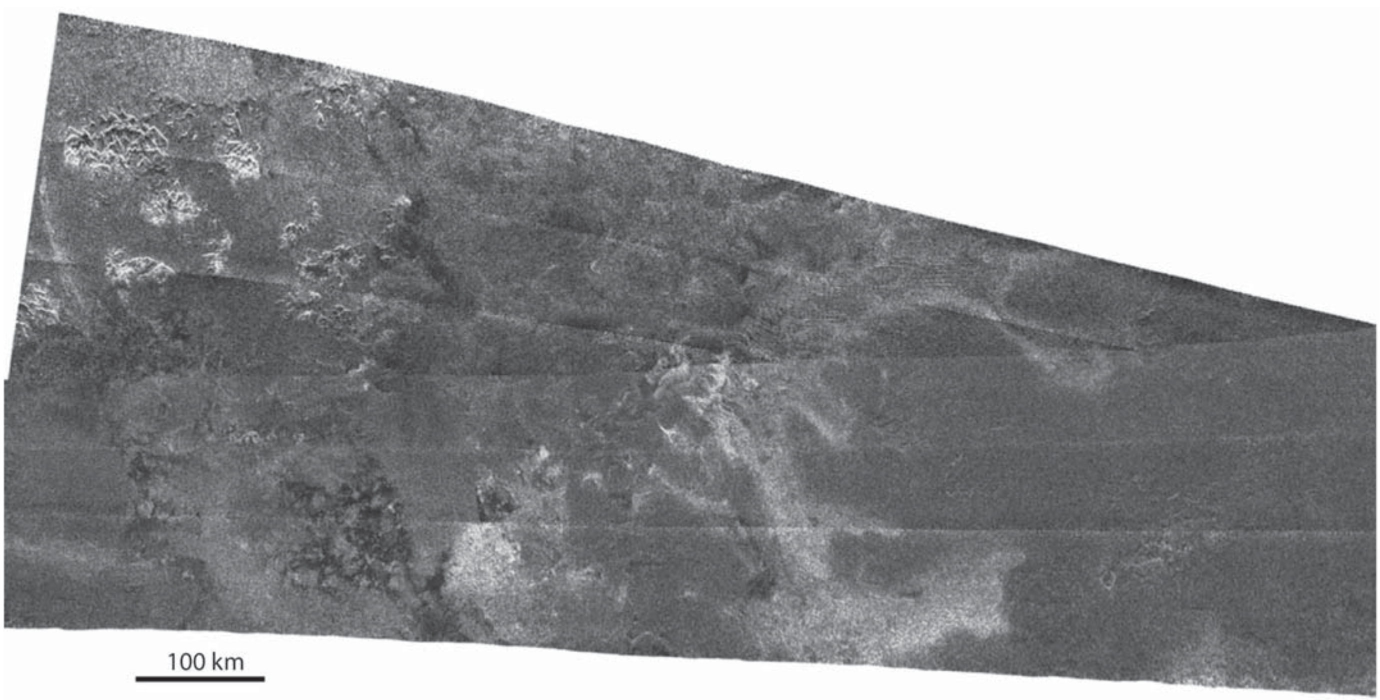

Figure 10. Flowlike feature called Winia Fluctus (45° N, 30° W) imaged by the Titan radar mapper using the SAR (synthetic aperture radar) mode. The radar illumination is from the south. From Mitri et al. (2008).

as requested from the model by Fortes et al. (2007), and so the ammonia-water paradigm continues to dominate. Mitri et al. (2008) showed that cryovolcanic activity on Titan could be related to the presence of a subsurface ammonia-water liquid layer underlying the ice shell (Figs. 11, 12). Accretion models suggest the presence of ammonia within Titan (e.g., Mousis et al., 2002), and that a subsurface ocean of ammonia-water is likely present within Titan (e.g., Tobie et al., 2006; Mitri and Showman, 2008b). Mitri et al. (2008) proposed a model of cryovolcanism that involves cracking at the base of the ice shell and formation of near-neutrally buoyant ammonia-water pockets in the ice. They showed that although the ammonia-water pockets cannot easily become neutrally buoyant and promote effusive eruptions, large-scale tectonic stress (from solid state convection, tides, nonsynchronous rotation, satellite volume changes, and/or topography) may promote the rise of cryovolcanic fluids that could erupt onto the surface. Alternatively, solid state convection in the ice shell, akin to that proposed for Europa, may transport heat to the near-surface, enabling melting or destabilization of surface materials, which may then erupt (e.g., Choukroun et al., 2007).

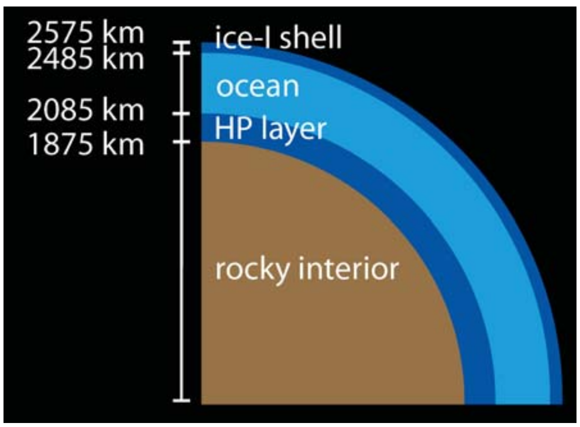

Figure 11. Model of the interior structure of Titan, which is at least partially differentiated and composed of a rocky core, a layer of ice high-pressure (HP) layer, and an ice-I shell. An ammonia-water subsurface ocean could be present. From Mitri and Showman (2008b).

Icy satellites, such as Ganymede, Ariel, Miranda, and Titania, show features that are related to resurfacing and possibly to eruption of material on the surface as the consequence of cryovolcanic activity, much like the grooved terrain on Ganymede (Showman et al., 2004). Also trans-neptunian objects and dwarf planets as Quaoar could present cryovolcanic activity (Jewitt and Luu, 2004).

In summary, the geological activity and surface features on various icy satellites strongly suggest the presence of cryovolcanic activity. The physical processes that are at the origin of eruption of cryomagma from the interior of icy satellites are still not known, and may vary considerably from world to world. Several mechanisms have been proposed (see Mitri et al., 2008, for a review): gas exsolution following depressurization in fluid-filled fractures that propagate upward from the base of the ice shell (Crawford and Stevenson, 1988; Lorenz, 1996); explosive eruptions of sprays (Fagents et al., 2000); pressurization of liquid chambers in an ice-I shell (Fagents, 2003; Showman et al., 2004); resurfacing driven by pressurization of the entire ocean during satellite volume changes (Manga and Wang, 2007); and partial melting of the ice shell by tidal dissipation (Mitri and Showman, 2008a; Tobie et al., 2008). The differences in the respective

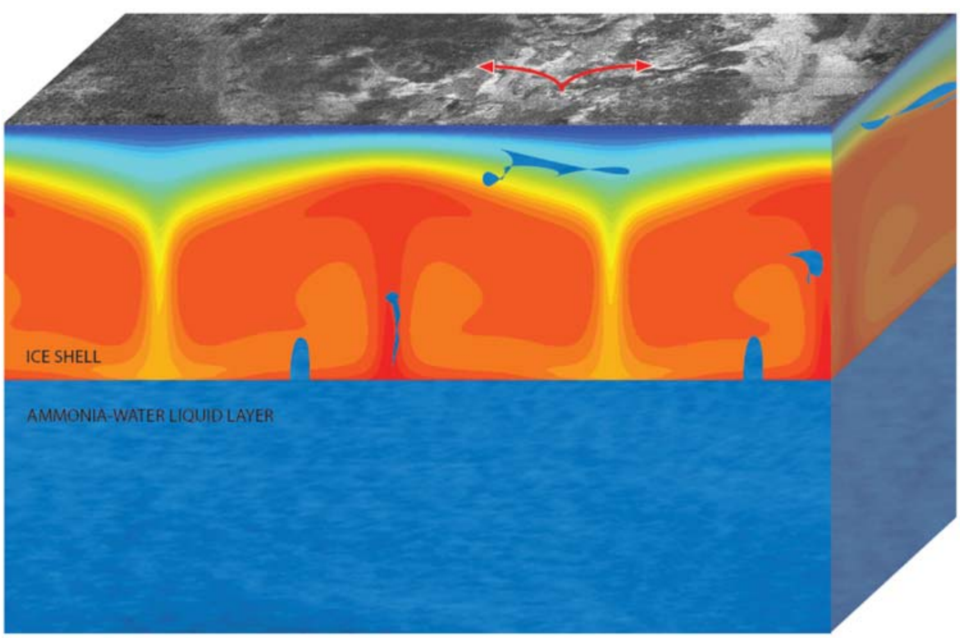

Figure 12. Schematic illustration of the model on Titan proposed by Mitri et al. (2008) for cryovolcanism, which involves cracking at the base of the ice-I shell and formation of ammonia-water pockets in the ice. As these ammonia-water pockets undergo partial freezing in the cold ice shell, the ammonia concentration in the pockets increases, decreasing the negative buoyancy of the ammonia-water mixture. Although the liquid cannot easily become buoyant relative to the surrounding ice, these concentrated ammonia-water pockets are sufficiently close to the neutral buoyancy point that large-scale tectonic stress patterns (tides, nonsynchronous rotation, satellite volume changes, solid state convection, or subsurface pressure gradients associated with topography) would enable the ammonia to erupt effusively onto the surface. Thermal convection in ice might play an important role.

geological and cryovolcanic activities of icy satellites could be produced by the different conditions as thermal-orbital evolution, internal composition, and grade of interior differentiation.

PROPOSED NEW DEFINITION: WHAT IS A VOLCANO?

The sections above reviewed the various types of volcanic processes that occur in the solar system, some of which are exotic for Earth-based standards. If we are to accept the wide variety of volcanic features revealed in the last few decades, it is necessary to come up with a clear and simple definition that encompasses them all. From a geological perspective, if a process fulfills the same role and has a similar expression on one world as on another, despite the different environments, it should have the same geological classification. Thus, fractures on the surface of Europa are still referred to as tectonic, even though they are in an icy crust, and standing bodies of liquid on Titan are still referred to as lakes, even though their likely composition is liquid hydrocarbons. Hence, eruptions of materials from depth onto the surface of an icy satellite can also qualify as volcanic. However, classifying the eruption of water-melts on icy satellites as some type of volcanic activity opens the question as to whether geysers on Earth should be considered to be volcanoes, or if geysers on Earth are not volcanoes why should we call the eruptions of water-melts in other bodies of our solar system as volcanic in origin rather than geysers?

Part of the problem remains, however, in that what we are observing on other planets is usually the remnant of a process, which is open to subjective analysis. Furthermore the processes themselves can be uncertain when a significant fraction of what goes on occurs beneath the surface, and in most cases the precise mode of fluid transport cannot be understood with current data. There are theoretical reasons to suspect that the transport processes may differ in some cases. On Earth, magma transport from the mantle to the near-surface is facilitated by the generally positive buoyancy of silicate melts relative to the crust. This appears to be a dominant mechanism in hotspot volcanism at, e.g., Hawaii, where it is thought that melts rise from the mantle as plumes that form owing to Rayleigh-Taylor instabilities, stalling typically within a few to tens of kilometers of the surface, forming a magma chamber, and then propagating fractures or dikes during cooling that can lead to eruption. On icy worlds, however, there is typically a strong negative density contrast between the icy melts and the surrounding water-ice crust, meaning that Rayleigh-Taylor instabilities and buoyant plumes are unlikely to form. Although the eruption of basic magmas, often higher in density than at least much of the upper crust, is observed on Earth, the mechanism of magma ascent in this case is not diapiric, and therefore the presence of higher density melts at the surface is not necessarily problematic. However, it is unclear if these mechanisms are related to plate tectonism—which so far has been observed only on Earth—and so may not occur on other worlds. Ways have been suggested to overcome this issue, some very similar to volcanic processes on Earth. For Titan, this includes Mitri et al.'s (2008) model that proposes that large-scale tectonic stress (solid state convection, tides, nonsynchronous rotation, satellite volume changes, and/or topography) may promote eruption of cryovolcanic fluids. Also convection in the ice shell for icy satellites could have an important role for cryovolcanic activity of Titan and other icy satellites (Sotin et al., 2002; Tobie et al., 2006; Choukroun et al., 2007; Mitri et al., 2008; Mitri and Showman, 2008b), bringing heat, volatiles, and/or melts to the near-surface in a non-buoyant manner.

Despite these limitations, we might examine with more detail the difference between volcanic activity and geysers in the Earth. The term *geyser* refers to a very specific type of phenomenon on Earth, in which a special plumbing system is required. In order for a geyser to form (rather than a hot spring, mud pot, or fumarole) a geyser requires a narrow spot or constriction, usually close to the surface. Water above the constriction acts as a lid, helping to keep the boiling water underground and under pressure. When the geyser erupts, it blows off the lid. Since a geyser's eruption uses up all the heat and/or water available in the system, the geyser needs a period (interval) to recharge. Thus, geysers are highly cyclic, and unlike volcanic activity on Earth they do not involve the mobilization of mantle-derived liquids and gases in each eruption. Consequently, if we saw an extrusive process involving silicate rocks–melt on Earth classified as volcanism, and a similar process involving icy rocks–melt on an icy satellite, we would classify the latter as cryovolcanism. A silicate volcano on Earth is not a geyser, because water is not magma on Earth, but water-melts are magma on icy satellites where these materials are the main constituent of crust and "cryomantle."

To account for these subtleties, we subclassify volcanism as (1) primary volcanism, as the eruption of melts sourced directly from the mantle; (2) secondary volcanism, as the eruption of magmatic materials that have been melted or remelted in the crust; and (3) tertiary volcanism, as the eruption of volcanic volatiles (gas or liquid) alone, that have been brought to the surface by a rising body of magma (e.g., fumaroles). Based on this hierarchical classification scheme, we consider that the following processes are volcanic: (1) eruptions of magmatic materials on Earth, including basalts, rhyolites, carbonatites, etc.; (2) eruptions of similar materials on the other terrestrial-rocky worlds of the solar system; and (3) eruptions of water-ice mixtures on the icy moons, as the materials originate from the subsurface and play a similar role to magmas on the terrestrial planets. These definitions apply regardless of whether the melt was directly sourced from the mantle or from a subsurface liquid layer and/or partial melting in an ice shell, and regardless of how it was transported to the surface. In contrast, the following are not considered to be true volcanic processes: (1) eruptions of mud, sulfur, and water on Earth, unless these materials are juvenile, i.e., brought up from depth by the magma, in which case they could be classed as tertiary volcanism; (2) sulfur on Io, as sulfur is a volcanic volatile in this case, and the crust and mantle of Io are silicate; (3) hydrocarbons on Titan, such as methane, as this plays a similar role to water and CO_2 in the Earth system.

CONCLUSIONS

The nature of volcanism is highly variable over the solar system, and the traditional definition of a volcano as defined for Earth needs to be modified and expanded to include processes such as cryovolcanism. We suggest the following definition:

A volcano is an opening on a planet or moon's surface from which magma, as defined for that planetary body, and/or magmatic gas is erupted.

The composition of magma is variable, depending on the planetary body. For the terrestrial planets (Mercury, Venus, and Mars), as well as for the Moon and Io, the magma is silicate, likely basaltic in most instances. Io may also have eruptions of liquid sulfur and SO_2, but these are unlikely to be the primary magma. Sites where only juvenile volcanic gas but no magma is currently coming out, are considered volcanoes in this definition (tertiary volcanism, as defined above). This may be the case, for example, for Ra Patera on Io, from where a plume but not thermal emission was detected during the *Galileo* mission (Lopes et al., 2001) and for which there was no evidence of erupted solids or liquids. Titan and icy satellites have aqueous mixtures in their interior, and these are the magmas (cryomagmas) for those bodies.

We caution against the use of the term *geyser* for Enceladus plumes (Fig. 8), which we define as cryovolcanic. At present, we do not know enough about the near-surface environment at Enceladus to characterize the plumes as geysers, analogous to terrestrial geysers in its mechanism of eruption. Future research on this subject is required to clarify the real nature of this type of geological activity.

In any case, it is clear that understanding the eruption mechanisms on other planets is important for better constraining how eruptions behave on Earth under present and past conditions. It also allows us to study and model volcanic processes in a significantly broader context. Planetary exploration has significantly broadened our definition of what a volcano is, and future data may further challenge our definition.

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Sedimentary volcanoes: Overview and implications for the definition of a volcano on Earth

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ABSTRACT

In contrast to volcanologists, sedimentologists are not only interested in "classical" volcanoes, but also in a second type, viz. sedimentary volcanoes. This type of volcano is helpful for sedimentologists in understanding the processes that occur in the (commonly unconsolidated) subsoil, even after deep burial.

Sedimentary volcanoes can be grouped in three classes: mud volcanoes, sand volcanoes, and associated structures such as water-escape and gas-escape structures. Mud volcanoes on deep-sea floors can become partially abraded in the course of time by ocean currents, and then are recognizable on seismic profiles as so-called pockmarks. A brief review of the various types of sedimentary volcanoes is provided, and representative examples of an active mud volcano and of a dormant mud volcano are dealt with in some more detail.

Sedimentary volcanoes have several characteristics in common with "classical" volcanoes, including their shapes and the processes that contribute to their genesis; they therefore deserve the name *volcanoes,* which implies that the term *volcano* has to be redefined.

INTRODUCTION

Volcanology and sedimentology are commonly considered as entirely different disciplines within the earth sciences. A closer look into both disciplines shows, however, that there are numerous aspects (e.g., pyroclastics) in which both disciplines overlap, so that volcanologists and sedimentologists can profit from each others' knowledge. The present contribution focuses on features that are commonly called *sedimentary volcanoes*. Whether these structures—which are nowadays only a study object of sedimentologists—should also be studied by volcanologists is a matter of personal taste, but it is clear that these structures deserve to be studied in detail to appreciate them in their full complexity.

The precise definition of a volcano is not as easy as it would seem intuitively (see Borgia et al., this volume). Furthermore, increasing knowledge of structures on other planets (e.g., Burr et al., 2009) forces us to stretch those definitions (e.g., Lopes et al., this volume). To some extent, the same can be said about some sedimentary features on our planet that display a similar shape as volcanoes. In this chapter the three main types of sedimentary volcanoes and associated structures are dealt with: mud volcanoes, sand volcanoes, and escape structures. It is interesting, certainly for volcanologists, that sedimentary volcanoes have a genesis that in many respects resembles that of volcanoes that are due to outflow of magma (indicated in the following as "true" volcanoes).

MUD VOLCANOES

Mud volcanoes are the most common type of sedimentary volcanoes. Remarkably enough, however, they are much less commonly mentioned in the more general geological glossaries

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and encyclopedias than their sandy counterparts. Most mud volcanoes have sizes of the order of decimeters to a few meters, both in diameter and in height (Fig. 1), but examples of several hundreds of meters occur, and examples of kilometers in diameter exist on the deep-sea floor (among others, in the eastern Mediterranean), as indicated by reflection seismics or acoustics. Examples of mud volcanoes up to ~500 m high have been described, among others from the El Arraiche mud-volcano field in the Gulf of Cadiz (Foubert et al., 2008). The Conical Seamount in the forearc of the Marianas Arc, which has been found to be a mud volcano, is even larger, reaching 25 km in diameter and 2 km in height (White, 2005). Wherever deep-sea mud volcanoes are mentioned, they are commonly associated with seepings of gas (Fig. 2), not rarely along faults affecting the sea bottom (e.g., Westbrook, 2005; Bonini, 2009; Mazzini et al., 2009).

Figure 1. Concentration of mud volcanoes in the Crimea area. Note the resemblance in architecture to a "true" shield volcano. Photo courtesy of Adriano Mazzini.

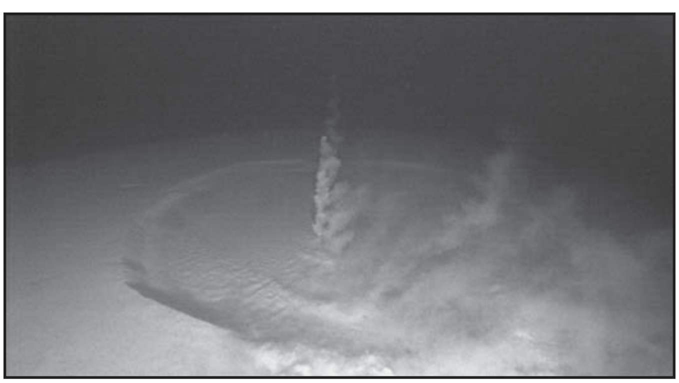

Figure 2. Methane release from an undersea mud volcano. Photo courtesy of the Institute for Exploration and the Rhode Island Institute for Archeological Oceanography (IFE/URIIAO).

Like "true" volcanoes, which build up because magma is pushed upward from a magma chamber with subsequent outflow of lava, most deep-sea mud volcanoes form under conditions when sedimentary particles are pushed upward together with gases and/or liquids (commonly water as the liquid, and air or hydrocarbons as gas) from a reservoir in the subsoil. This mixture follows an upward pathway (which may or may not represent a zone of relative weakness) and flows out at the sedimentary surface as a commonly low-viscosity material (Aslan et al., 2001; Mazzini et al., 2007) that spreads out like a shield volcano. There exist also examples, however, where thick piles of extruded material form, more or less as a stratovolcano, probably because a significant amount of already somewhat consolidated mud is included in the outflowing mass with a high particle–gas-and-liquid ratio.

Most small-scale mud volcanoes are formed when a rapidly deposited subaqueous succession with an alternation of thin sandy and fine-grained layers (that seal off the underlying water-saturated sediments) becomes affected by the weight of the sedimentary overburden. This overburden can increase the hydrostatic pressure of the pore water in the underlying material, forcing it to move into the direction of the lowest pressure (upward), commonly along zones of weakness defined by, for instance, the inclined lamination-stratification of sedimentary structures, but breaking through the overlying layers until it can flow out at the sedimentary surface. The upward moving water erodes the sediments along the "intrusive" pathway, thus becoming a mixture of water and mud. At the sedimentary surface this mixture flows out, commonly in the form of low-density mass flows, sometimes in the form of particle-laden currents. As soon as the ascending water-sediment mixture reaches the sedimentary surface, the current velocity diminishes, and particles start to settle, the largest ones settling first. This implies that much material is deposited directly around the opening in the sedimentary surface, with decreasing amounts of material being deposited per surface area farther away (from both loss of material "upstream" and an increasing surface area with increasing distance to the opening). Thus a body with the shape of a shield volcano is commonly formed, well comparable morphologically with "true" shield volcanoes.

Mud volcanoes occur also frequently where water-saturated fine-grained sediments undergo sudden changes in pressure, for instance, from an earthquake (Manga et al., 2009), so that the sediments fluidize. A third situation that favors the genesis of mud volcanoes is the presence of hydrothermal activity; upwelling hot waters (Fig. 3) may break through the sedimentary cover, resulting in "bubbling" mud masses that flow out unless kept in place by surrounding walls, but they may also result in mud geysers that could suddenly become active in an explosive way because of overpressure of superheated fluids (Gisler, 2009).

Many of the originally giant mud volcanoes on seafloors are present now only in the form of some remnants that are known as pockmarks (Fig. 4). As pockmarks occur as a rule at great depth, they are known almost exclusively from geophysical data obtained during cruises with research ships. The origin of pockmarks is still under debate, but most researchers are of

Figure 3. Gas (methane) bubbles rising in the crater lake of the Dashgil mud volcano in Azerbaijan. Photo courtesy of A. Kopf.

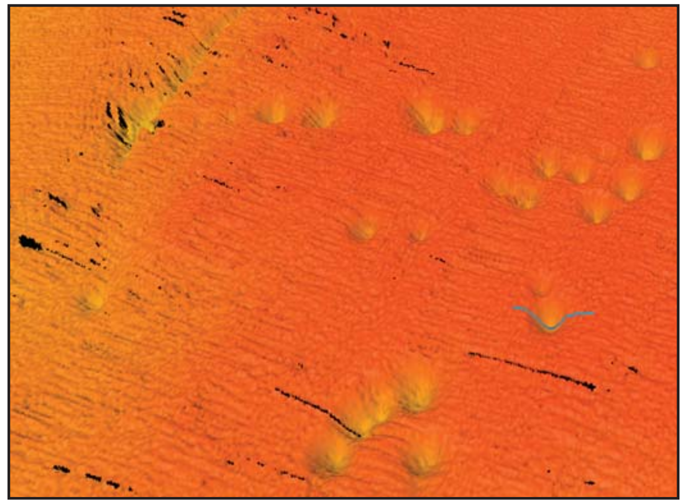

Figure 4. Pockmarks on a 500-m-deep seafloor, detected during a cruise by researchers from the University of New Hampshire (UNH) and the U.S. National Oceanic and Atmospheric Administration (NOAA).

the opinion that they are remnants of giant mud volcanoes (or rising mud-gas bubbles) that were formed on the seafloor as a result of the escape of huge amounts of gases from the subsoil (cf. Piper, 2005). These gases are probably hydrocarbons (mainly methane) formed from the dissociation of clathrates, owing to disturbance by processes that are not yet well understood. Nevertheless, the escaped hydrocarbons are considered a significant source of greenhouse gases in the atmosphere. For example, some mud volcanoes on the Mediterranean Ridge are known to release nowadays methane gas in amounts of the order of 30 million cubic meters per year (Hieke, 2004), and the Eocene Thermal Maximum is ascribed to such methane escapes by Svensen et al. (2004). Furthermore, the high temperatures of the Early Jurassic are ascribed to these gases by Kemp et al. (2005), and it has been observed that the submarine escape of methane may result in fire at the sea surface (Aquilina, 1886). This implies that this type of mud volcano has great ecological—and possibly climatic—significance.

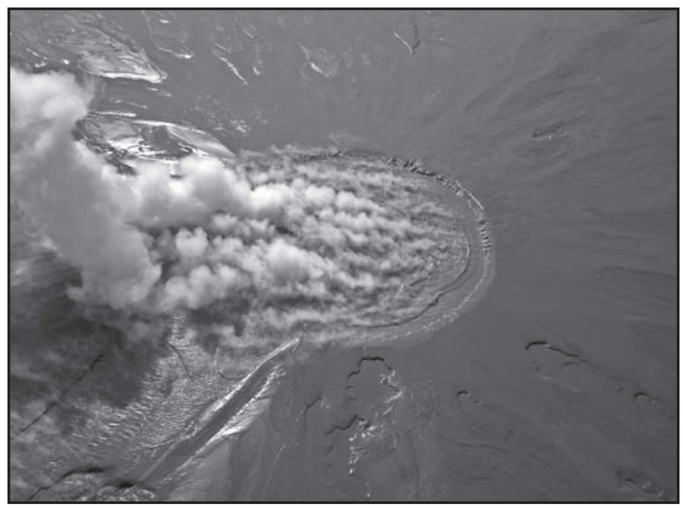

Figure 5. The Lusi mud volcano during a period of (relatively) low activity. From Mazzini et al. (2007); reproduced with permission.

An Active Mud Volcano: Lusi

Large mud volcanoes occur not only on the sea bottom but can also form on land. One example of a still active giant mud volcano has received much attention lately (see, among others, Mazzini et al., 2007; Davies et al., 2008); it was baptized *Lusi*. This mud volcano (Fig. 5), near the village of Sidoarjo in the northeastern part of the Indonesian island of Java, became active on 29 May 2006 and is still working in spite of several attempts to stop its activity. The amount of mud that has escaped so far is so large that the mud now covers a surface area of 7.5 km^2, which would have been considerably larger if the outflowing mud had not been stopped by purposely constructed dikes. The area covered by the mud has become uninhabitable (Fig. 6), and the ~30,000 people that lived there had to be evacuated. At present, the mud-covered area between the dikes is easily detectable on satellite images (Fig. 7A).

The eruption of Lusi (Fig. 7B) caused excitement among some mud volcanologists because it represents a unique scientific event: The geological conditions immediately prior to the eruption had been monitored in a gas-exploration well (BJP1), which was located 150–200 m away at the time of the eruption, and the birth and early stages of the evolution of such a large mud volcano had not been closely observed before (Davies et al., 2007). It has been debated whether the mud volcano originated as a result of a pressure front after an earthquake as advocated by,

Figure 6. Mud cover in the neigborhood of the Lusi mud volcano. Note the street light for scale. From Mazzini et al. (2007); reproduced with permission.

Figure 7. Details of Lusi. (A) False-color satellite image of the dam-contained area flooded by the mud erupted by Lusi. Color interpretation: red is vegetation; pale blue and green represent bare ground and/or fallow fields; black indicates water; the mud from Lusi is charcoal gray. NASA photo. (B) Beginning of Lusi (27 May 2006): Vapor and mud are erupted in the middle of a rice pond. From Mazzini et al. (2007); reproduced with permission. (C) Helicopter view, showing the position of Lusi near borehole BJP1. From Mazzini et al. (2007); reproduced with permission.

among others, Mazzini et al. (2007) or by the nearby drilling of the gas-exploration well (Davies et al., 2008) (Fig. 7C), but most recent data suggest—as will be detailed below—that the formation of this mud volcano was triggered by the drilling activity, which initiated the formation of a gas and liquid reservoir at a depth beyond 2833 m (Davies et al., 2008).

Lusi resembles a “true” volcano in many respects, including the outflow of hot material (though, obviously, not as hot as outflowing magma). Temperatures of up to 97 °C were measured at a few dozens of meters away from the vent, and the boiling character of the mud-water-gas mixture suggests that the temperature of the mixture, when reaching the surface, must be some 100 °C. Measurements in borehole BJP1 indicate, indeed, a temperature of 100 °C at a depth of 1700 m and a temperature of 138 °C at 2667 m. The high geothermal gradient of 42 °C per kilometer is probably related to the proximity of a volcanic arc.

The history of Lusi has been well documented. Seeping of water started, at several places, in the early morning of 29 May 2006; a true eruption followed, still during the morning, a few hours later. Steam rose up to some 50 m high (Fig. 7B), and a mixture of mud, ~60% boiling water, and gases (mainly water vapor, CO_2 and CH_4, but also some H_2S) was ejected several tens of meters high. Nearby craters developed during the next days, and the amount of outflowing mud increased from 5000 m^3 per day in the beginning to 120,000 m^3 per day in August. Peaks up to 170,000 m^3 per day followed in September, and a record of 180,000 m^3 per day was reached in December 2006. The intensity then slowed down somewhat but was still ~100,000 m^3 at the end of October 2008.

The total amount of mud erupted by Lusi amounted to >27 million m^3 by mid-2007. The accompanying amounts of water were also huge: until March 2007 it amounted to some 15 million m^3. Precise data of more recent dates are not available to the present author. The outflow of the mud-water-gas mixture has, in combination with the weight of the erupted mud, resulted in the geological exceptionally rapid subsidence of an area of 22 km^2 of 1–4 cm per day(!) until mid-2007 (Mazzini et al., 2007). The possible total subsidence that the area may undergo is now estimated at 44 m.

The development of Lusi is obviously related to the characteristics of the subsoil. The stratigraphy is well known from borehole BJP1. The successive units consist roughly of (from top to bottom): modern alluvial sediments (0–300 m), a Pleistocene alternation of sands and mud layers (300–1000 m), Pleistocene bluish gray clay (1000–2000 m), and Pleistocene volcaniclastic sand (2000–3000 m). Much less is known about the tectonic structures in the subsoil, but there are sufficient indications that faults occur and that fracture levels are present. These probably form the zones of weakness along which the water and mud in the subsoil have migrated.

The depth from which the mud comes has been established by comparison of the clay minerals in the Lusi mud masses with those from the various stratigraphic intervals encountered in the BJP1 well (from which 13 samples have been collected). In the borehole the clay minerals in the 1341–1432 m interval are almost pure smectite, indicating rapid volcaniclastic sedimentation under marine conditions. All other intervals contain mixtures of smectite, illite, and kaolinite, but in varying ratios. The Lusi mud has a clay composition that is similar to that in the 1615–1828 m interval in the borehole. Because the upwelling masses may have taken “extra” smectite from the marine volcaniclastics, it is possible that lower units also have contributed to the Lusi mud. Taking all clay-mineralogical data together, the mud must have come from layers between 1219 and 1828 m, likely from between 1615 and 1828 m, and probably from the lower part of this interval (Mazzini et al., 2007).

The clay mineralogy, in combination with the depth from where the mud is derived and with the high geothermal gradient, provides a clue for the source of the huge amounts of water that are produced by Lusi. A large part of this water must have originated from the dehydration of the clay minerals. The smectite in at least the 1109–1828 m interval, under the prevailing conditions, must have undergone diagenetic transformation into illite. As 1 m^3 of smectite can thus produce 0.35 m^3 of water, as some 65% of the smectite probably has been dehydrated, and as the smectite content is about one-third of the clay minerals, more than 10^9 m^3 of water must have been available. In addition, pore water that was already present in the undercompacted marine clays is a source of the water. This explains why such huge amounts of water could be produced, and why the production of water can still go on.

It is also interesting to note that the dehydration of the smectite has increased the volume of the total mass, and that the diagenetic process that took place at a depth of >1109 m thus must have contributed to an ever increasing pressure buildup. This explains the force of the eruption. In this context it should be mentioned that the ongoing outflow of the mud-water mixture threatened to cover such a vast area that walls were constructed to encompass the mud. The ongoing outflow went on longer than expected, however, and some of these walls failed. A subsequently constructed dam was capable of restricting the “flooded” area to 7.5 km^2 (see Fig. 7A). Attempts were made to stop the outflow by putting 398 clusters of chained concrete balls (with diameters of 20–40 cm) into the crater, but the flow rate was not significantly affected.

As mentioned above, there has been a hot discussion about what triggered the volcanic activity. Three possible origins have been mentioned: an earthquake, the nearby drilling activity, and a combination of both. The discussion was obviously of considerable economic importance, because a drilling-induced catastrophe would have great financial consequences for the drilling company.

The earthquake hypothesis has a logical background. An earthquake with a moment magnitude of 6.3, that had its epicenter at some 250 km from the Lusi site near Yogyakarta (where it caused almost 6000 casualties and left more than half a million people homeless), took place on 27 May 2006, only two days before Lusi became active. It is, in addition, well known that earthquakes can trigger mud volcanoes (Chigira and Tanaka, 1997; Mellors et al., 2007), and the magnitude of the Yogyakarta earthquake was above the empirically found threshold value for the triggering of mud volcanoes. Not only the magnitude is decisive, however, but also important is the distance between the earthquake's epicenter and the location of mud volcano. For the case of the Yogyakarta earthquake and the Lusi volcano, this distance was too great, considering the effect that was caused, even under optimal conditions. It has, in addition, been noted by Manga (2007) that two larger and closer earthquakes did not trigger a mud volcano in the neighborhood of the Lusi site. Estimations of the ground motion (Davies et al., 2008) also indicate that transient changes in pore pressure (~10 Pa) were far too small to induce the fluidization that must have taken place. An earthquake can therefore be ruled out as the cause (or a helpful condition) of the Lusi volcano.

This leaves the drilling activities (Fig. 7C) as the most likely cause, and there are good reasons to believe that these activities must be held responsible, indeed. On 28 May, one day before Lusi started, the drill bit and drill string were removed from the borehole. Their joint volume was replaced by drilling mud, but during this operation a significant influx of water and gas into the borehole must have occurred, as indicated by the fact that some 30%–50% of the drilling mud was pressed upward to the surface. The replacement in the borehole of the relatively heavy drilling mud by the lighter formation water-gas mixture must have further lowered the pressure in the well so that more overpressured formation water could flow in. Thus conditions were created that allowed Lusi to become an active mud volcano.

Lusi is still active. It is envisaged to keep the ever increasing volume of mud at its present site of 7.5 km^2 by constructing larger dams (Fig. 8), consisting of concrete cylinders of 120 m diameter with walls of 10 cm; this dam should be 50 m high and suffice to contain the mud-water mixture until the source of the water becomes exhausted and the eruption consequently comes to an end.

A Dormant Mud Volcano: Dashgil

Most of the >1500 known large-scale mud volcanoes are dormant. Some well investigated examples occur in Azerbaijan, in the Caspian Basin, which has the world's highest concentration of both continental and marine mud volcanoes. This must be attributed to a combination of factors: the quick Quaternary subsidence and infill (up to 2.4 km per million years), the diffuse methane generation in deeply buried clay units, and the compressional tectonics that led to anticlinal traps and the frequent seismicity that can trigger eruptions (Mellors et al., 2007).

The best investigated of these mud volcanoes is the Dashgil mud volcano (see, among others, Mazzini et al., 2009; Kopf et al., 2009), which is some 60 km southwest of Baku, on the crest of the Dashgil fold. It is accompanied by several more mud volcanoes (the most important being the Koturdag, Bahar, Satellite, and Delianiz mud volcanoes), but the Dashgil mud volcano (Fig. 9) is well accessible, and it shows the widest variety of seeps. It erupted for the last time in 1958 and should therefore be considered as dormant, even though some minor outflows still continue, with a short increase in the outflow in 2001 in the northwestern part of the crater (Aliyev et al., 2002).

The volcano has a low elevation with a morphology that resembles a pie, and its central crater is ~200 m in diameter (Hovland et al., 1997). The erupted mud covers an area of ~5.5 km^2. The crater area is partly covered by a large lake, from which methane bubbles are released. Outside the lake area it contains a large number of sites where seepage of mud and water occurs; this gives rise to peculiar forms, viz. gryphons, pools, and salsa lakes. Gryphons

Figure 8. High activity of Lusi during construction of the dam built to enclose the erupted mud. From Mazzini et al. (2007); reproduced with permission.

Figure 9. Center of the Dashgil mud volcano. Note the "ringwall" in the background. Photo courtesy of A. Kopf.

(Fig. 10) might be considered small-scale mud volcanoes: They are concentrated in the central part of the "mother" volcano, are commonly some 2–3 m high, and have a conical shape, more resembling a stratovolcano than a shield volcano. Mud, water, gas, and oil seep out and also erupt at irregular intervals. Thus they are ideal objects for studying the processes that formed the Dashgil mud volcano at a much larger scale. Pools are more or less circular seepage features, commonly at the feet of gryphons, where water and gas are released continuously with minor amounts of mud. These pools are up to ~2 m in size. So-called salsa lakes (the type of more or less saline lakes that are found within or close to large mud volcanoes because of erupted "fossil" seawater from deep-seated marine sediments) are true lakes with sizes up to 75 m and depths of 10 m. Gas and water, usually with only a minor amount of mud, erupt from the bottom under these lakes, usually in a vigorous way (Mazzini et al., 2009).

The release of hydrocarbons is large (Jakubov et al., 1971, estimated it at 15 million m^3 per year, but this may be an overestimation), like in many other (particularly deep-marine) mud volcanoes (the total amount of methane released by continental and shallow-marine mud volcanoes worldwide is estimated at 6–9 million tons per year: Etiope and Milkov, 2004; and the amount released by deep-marine mud volcanoes is probably higher). The dry gas from Dashgil consists of 94.9%–99.6% methane. The rest consists of ethane (<0.4%) and CO_2 (0.3%–5.1%).

The geochemistry of the gases suggests that they are derived from reservoirs that may lie several kilometers underneath the volcano, from where continuous leakage takes place. Molecular fractionation during the rising of the gas must be held responsible for the variations in gas wetness. Considering the uniform composition of the gas, it may well be that there is only one source of the methane. A different situation exists for the water: Its geochemistry is different for the gryphons, pools, and salsa lakes, and the gryphons erupt fluids that come from a deep-seated source, whereas the pools and salsa lakes show imprints of meteoric water; their higher salinity must be ascribed to in situ evaporation.

Field evidence, in combination with geochemical investigations, makes it likely that an intricate, interconnected plumbing system occurs in the shallow subsoil of the volcano. This plumbing system is continuously recharged from the deep-seated sedimentary reservoirs by a branched system of conduits (Mazzini et al., 2009).

CPTu (cone penetration testing with pore pressure measurement) tests, performed in 2007, indicate that the central part of the crater lake, which hosts the conduit for gas ascent, shows a low sediment shear strength of 5–20 kPa and excess pore-fluid pressures of 15–30 kPa at 1 m below the bottom (Kopf et al., 2009). The mud in the conduit appears to be much less stiff (150 kPa cone resistance as a measure for undrained shear strength) than elsewhere (300–700 kPa), which may be ascribed to the presence of deep-seated shales that partly liquefy and ascend (Fig. 11). The center of the conduit shows a low pore pressure, probably because of rapidly migrating gas, which increases to 30 kPa at the lake bottom and reaches hydrostatic values at the crater rim.

The fluid-filled crest of the Dashgil dome is underlain by an overpressured region. Kopf et al. (2009) therefore expect that updoming now takes place. In combination with the fact that explosive eruptions occurred in 1908 and 1928, they expect that a similar violent activity may occur in the near future.

SAND VOLCANOES

Sand volcanoes (Fig. 12) have an origin that is largely comparable with that of mud volcanoes, but there are some important

Figure 10. Gryphon in the center of the Dashgil mud volcano, showing characteristic mudflow deposits that are formed continuously. Photo courtesy of A. Kopf.

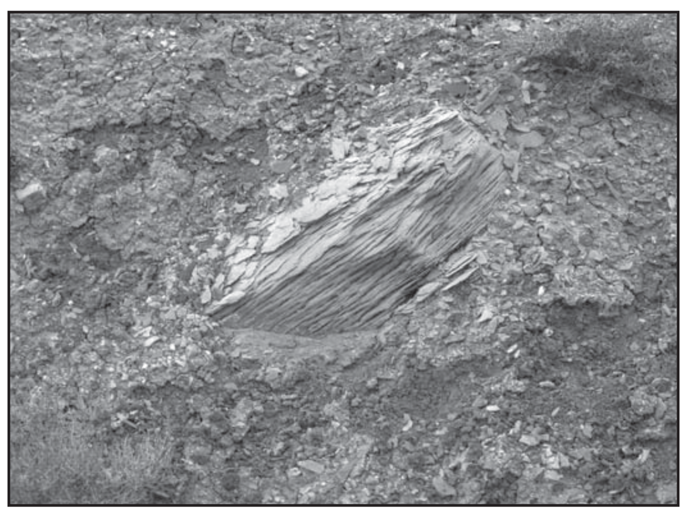

Figure 11. Slab of shale brought up from a deep-seated shale layer that was partly liquefied during upward transport in the main vent of the Dashgil volcano. Photo courtesy of A. Kopf.

Figure 12. Sand volcano in the Permo-Carboniferous Talchir Formation along the Nandir Jhor River near Bedasar.

differences from a sedimentological point of view. The granulometry of the sand volcano is, obviously, different from that of mud volcanoes, and consequently the shapes of the two types of sedimentary volcanoes differ. Whereas mud volcanoes consist of fine-grained material that can easily be transported by a particle-laden water flow over (relatively) large distances, sand volcanoes are composed of sand-sized particles. Such particles are not easily transported far away when the current velocity diminishes. Another difference with mud volcanoes is that sand volcanoes are generally much smaller (maximally a few meters in diameter and height) than mud volcanoes, but it must be noticed in this context that the precise granulometry of submarine sedimentary volcanoes is not well known. The differences in behavior between unconsolidated sand and clay make it, however, unlikely that long-lasting and huge eruptions of sand volcanoes can take place, even though sand-clay mixtures (with a surplus of water) might do so for some time. Consequently, sand volcanoes seem to be restricted to features with a "reservoir chamber" at maximally a few meters deep, which is a limiting factor for their size indeed.

Examples can, obviously, be studied best under subaerial conditions, and favorable places are where water-saturated sands occur. When the vertical pressure increases (for instance as a result of ongoing loading by sedimentation), sand volcanoes may form for the same reasons as mentioned above for mud volcanoes. In addition, sand volcanoes are known to form if water-saturated sands (preferably with some finer fraction included) are affected by a shock, for instance, an earthquake (Montenat et al., 2006). Analyses of lateral changes within one layer of the size of the sedimentary volcanoes (and of the other types of soft-sediment deformations) can result in indications of the direction of the epicenter of the responsible earthquake (Rodríguez-Lopez et al., 2007).

When sand-laden water leaves the vent, the sandy particles are almost immediately deposited, particularly under subaerial conditions, because the water disappears immediately—still on the volcano itself—into the sandy material. Consequently, a relatively steep volcano tends to build up, comparable with the shape of a stratovolcano (Figs. 13A, 13B). There are more resemblances between these two: The high-density flows that may leave the vent via the crater opening will form mass-flow lobes from which the water escapes almost immediately into the sand so that the lobes become water-free and do not move farther downward, thus giving an appearance of lobes that look much like lobes of cooling lava (Fig. 13C).

It thus turns out that sand volcanoes are the result of essentially the same processes that result in mud volcanoes, but that they are much less spectacular. They do not seem to occur more frequently either, nor are they found in rocks from a wider variety of environments. It is therefore not clear why sand volcanoes are more commonly mentioned in the literature than mud volcanoes. A possible explanation might be that sandy deposits receive in general much more attention from sedimentologists than muddy sediments, among other reasons because sedimentary features are much more easily observed in the field in sandy deposits than in muddy sediments.

ESCAPE STRUCTURES

Escape structures can be considered structures that are formed by essentially the same processes as those forming sedimentary volcanoes, but where insufficient solid particles were "erupted" together with the liquid or gas to build up a volcano-shaped body. The lack of solid particles in the outflowing material should not be ascribed to lack of unconsolidated material but rather to the fact that either more or less pure water escaped from the subsurface. This, in turn, must be due to the upward expulsion of a water stream that is insufficiently strong to erode the sediments surrounding the "vent," commonly because this material consists of too-coarse particles (Fig. 14A). A typical example at a very small scale are the escape structures formed in beach sand each time that backwash of a wave occurs: During the swash process the interstitial pressure in the top part of the beach sand is raised, and when the wave disappears the overpressured interstitial water and trapped air escape to the sedimentary surface, leaving millimeter-sized holes (Fig. 14B). Although the driving mechanism of the upward flow of liquid-gas in this case is the same as in the case of the mud and sand volcanoes presented above, the small scale of these phenomena and the absence of a construction around the vent eliminate any of the similarities that can be traced between sedimentary and "true" volcanoes. Consequently, these structures will not be further discussed here.

DISCUSSION AND CONCLUDING REMARKS

"True" volcanoes and sedimentary volcanoes have largely similar morphologies, they overlap each other in size (large sedimentary volcanoes reach sizes that compete with those of many "true" volcanoes), and they have a genesis that is comparable in

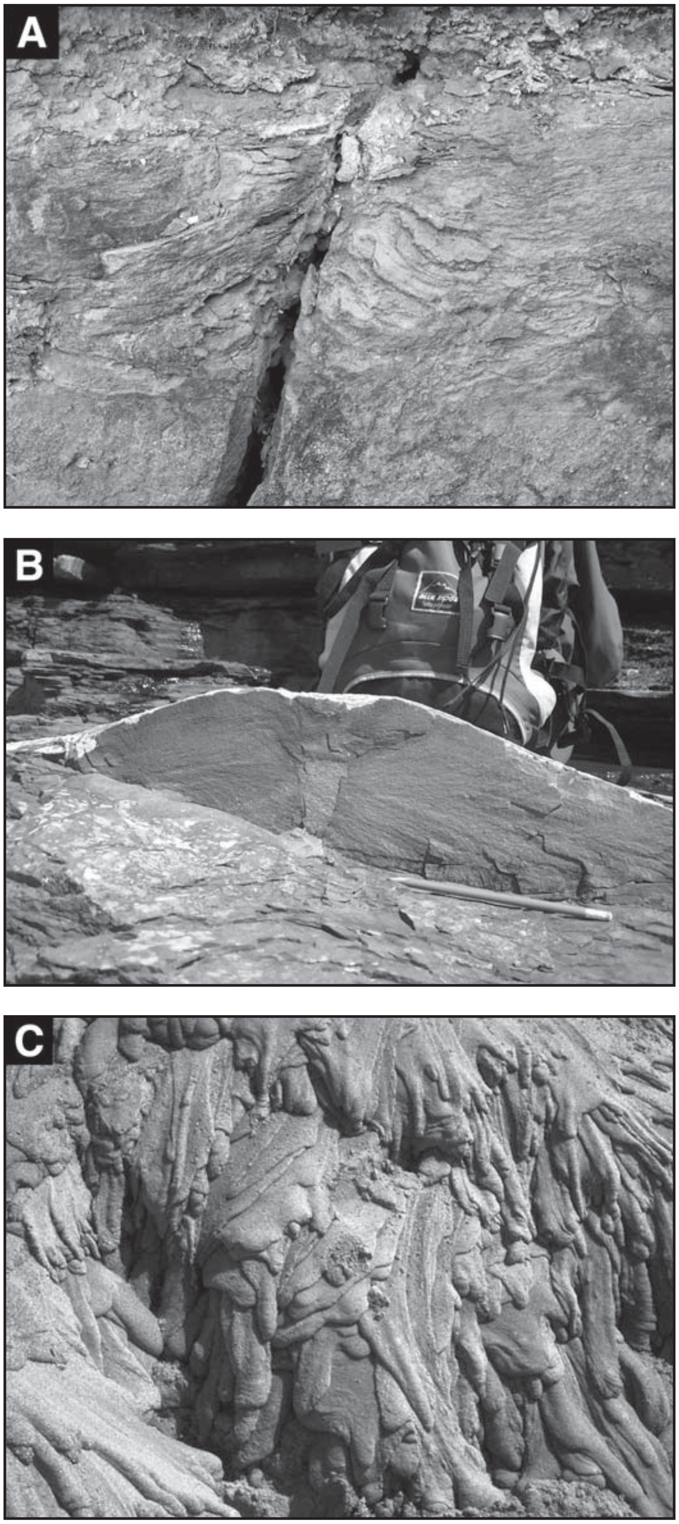

Figure 13. Characteristics of small-scale sand volcanoes with the shape of a stratovolcano. (A) Cross section showing curved interior of the flanks of the volcano. Flysch near Tylmanow (Poland). (B) Cross section through a sand volcano in County Clare, Ireland. Photo courtesy of Edward Lewis (University of Leicester). Note the well-developed vent. (C) Flow lobes on the flank of a recent sand volcano in a sand quarry at Uelsen (Germany), resembling flow lobes from a "true" stratovolcano.

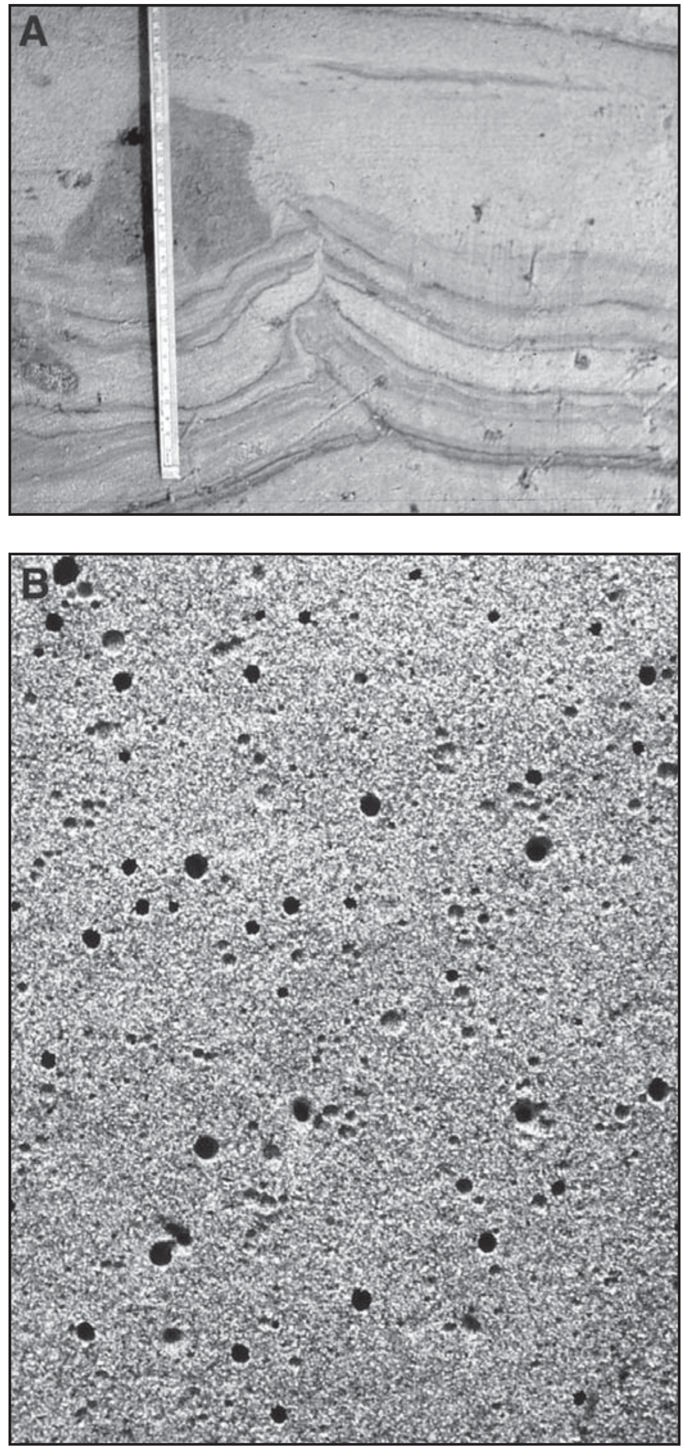

Figure 14. Water-escape structures. (A) In coarse Holocene sands in a reclaimed part (Noordoostpolder) of the former Zuiderzee (now Lake IJssel), central Netherlands. (B) Millimeter-sized vents in a sand beach between Llanes and Aviles (N. Spain), produced by the expulsion of air and water from the uppermost sand layer under the influence of pressure pulses resulting from successive swash waves.

many respects. For instance, comparison of the material flowing out from "true" and sedimentary volcanoes shows that the sedimentary outflows resemble basaltic lava in the case of mud volcanoes, and acid lava in the case of sand volcanoes; this results in sedimentary volcanoes that morphologically resemble shield volcanoes and stratovolcanoes, respectively.

It could also be established that sedimentary volcanoes have, like "true" volcanoes, some kind of "magma" chamber in the form of a gas- and/or liquid-bearing layer with increasing pressure (either continuously, for instance, from the weight of the ever thickening sedimentary overburden, or incidentally, for instance, from an earthquake-induced shock wave). If the pressure exceeds a threshold value, the pressurized gas and/or liquid (commonly pore water with dissolved air or hydrocarbons) breaks upward through the overlying sediment, often following an already existing zone of weakness (which can result from the configuration of sedimentary structures, earlier sedimentary deformations, etc., but also from fractures in a hard-rock unit), to finally flow out at the sedimentary surface, either subaerially or (more commonly) subaqueously.

In fact, the material that rises up through the connection between the source in the subsoil and the sedimentary surface behaves in several respects as magma on its journey from the magma chamber to the vent or crater opening: Magma becomes more fluid and develops gas bubbles while rising owing to decreasing pressure, and in sedimentary volcanoes with a deep-seated source the formation of bubbles also takes place. This must be deduced not only from the sizes of some submarine mud volcanoes at places where gas escape is a common feature but also from the enormous power of the rising mud that still prevents closure of the vent of the Lusi mud volcano.

Ideally, the many similarities between "true" and sedimentary volcanoes should promote a fruitful collaboration between volcanologists and sedimentologists that can be mutually beneficial. Volcanologists might, for instance, profit from the results of research into sedimentary volcanoes if they would pay more attention to the analyses of sedimentary volcanoes and of the processes that form them, as this would increase their insight into the processes that lead to the formation of "true" volcanoes. It is noteworthy to mention in this context that active sedimentary volcanoes can, as a rule, be studied much more easily than active "true" volcanoes, because the hazardous conditions that characterize most volcanic eruptions (high temperature, poisonous gases, falling tephra) do not commonly occur in the direct neighborhood of sedimentary volcanoes. These differences in conditions are important for research possibilities. It is, for instance, difficult to analyze with fairly good accuracy what processes facilitate, hamper, or block the ejection of magma. In contrast, sedimentary volcanoes, either active or in a still unconsolidated state, allow us to obtain a detailed picture of their 3-D buildup. It could thus be detected that irregularities in the vent's shape are fairly commonly present in sedimentary volcanoes; these may contribute to temporary or long-lasting blockage of the vent, resulting in the buildup of an ever growing pressure that may eventually lead to a sudden explosive release of the blocked material (see, among others, Taddeucci and Wohletz, 2001; Macías and Siebe, 2005); on the other hand, such blocking may also lead to the cessation of the volcanic activity. This is entirely comparable with models for plugs in volcanic vents, which are based on geophysical data and logical reasoning, but which have never been actually seen in action.

On the other hand, volcanic processes that can be actually observed have led already long ago to conclusions that can also be applied to sedimentary volcanoes and that therefore can be of interest to sedimentologists studying mud or sand volcanoes. It should be remembered in this context that "true" volcanoes were already studied in antiquity, but that sedimentology started not earlier than about half a century ago to be an earth-science discipline on its own. The study of features such as sedimentary volcanoes started even much later. It is therefore not surprising that, from the very beginning on, many more data have been collected from "true" volcanoes and that many of the hypotheses about the processes forming "true" volcanoes have long been taken for granted, even though no actual observations of the processes taking place inside an active volcano could be made. In this context the "discovery" of sedimentary volcanoes—which consist of unconsolidated sediments when being formed, and which therefore can be studied in three dimensions, regarding the processes both on their outer flanks and in their interior—must be considered most fortunate.

In addition, many sedimentary volcanoes (and gas- or water-escape structures) have such small sizes (order of meters or even decimeters or centimeters) that they can be studied easily not only in the field but also in laboratory experiments. This has greatly increased the insight into their genetic processes, and both sedimentologists and volcanologists can benefit from such experimental approaches. This proves that joint research efforts in both earth-science disciplines can contribute significantly to a better insight into the processes that shape Earth.

The similarity in both morphology and genesis implies that there is sufficient reason to state that sedimentary volcanoes indeed deserve the name *volcano,* with the consequence that the textbook definition of this term has to be modified to accommodate such structures. In any case the general modifications required for that purpose seem to be similar to the modifications required to include volcanic formations inferred to exist on other planets. As knowledge and understanding of the processes associated with all of these terrestrial and extraterrestrial formations increase, we are likely to reach a more satisfactory definition of the term.

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Monogenetic volcanic fields: Origin, sedimentary record, and relationship with polygenetic volcanism

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ABSTRACT

Monogenetic volcanism is commonly represented by evolution of clusters of individual volcanoes. Whereas the eruption duration of an individual volcano of a volcanic field is generally short, the life of the entire volcanic field is longer than that of a composite volcano (e.g., stratovolcano). The magmatic output of an individual center in a volcanic field is 1–3 orders of magnitude less than that of a composite volcano, although the total field may be of the same volume as a composite volcano in any composition. These features suggest that the magma source feeding both monogenetic volcanic fields and composite volcanoes are in the same range. Monogenetic volcanic fields therefore are an important and enigmatic manifestation of magmatism at the Earth's surface. The long eruption duration for an entire volcanic field makes this type of volcanism important for understanding sedimentary basin evolution. Accumulated eruptive products may not be significant from a single volcano, but the collective field may contribute significant sediment to a basin. The eruptive history of volcanic fields may span millions of years, during which dramatic climatic and paleoenvironmental changes can take place. Through systematic study of individual volcanoes in a field, detailed paleoenvironmental reconstructions can be made as well as paleogeographic evaluations and erosion-rate estimates. Monogenetic volcanoes are typically considered to erupt only once and to be short-lived; recent studies, however, demonstrate that the general architecture of a monogenetic volcano can be very complex and exhibit longer eruption durations than expected. In this way, monogenetic volcanic fields should be viewed as a complex, long-lasting volcanism that in many respects carries the basic characteristics similar to those known from composite volcanoes.

INTRODUCTION

Volcanic fields are important manifestations of volcanism in almost every tectonic setting, although they tend to be most commonly formed in continental regions. Also, volcanic fields can be formed by products of essentially every composition, although most commonly they are basaltic (Valentine and Gregg, 2008). Volcanic fields are composed of individual, short-lived volcanoes that are commonly, but not exclusively, small in eruptive volume. Monogenetic volcanic fields commonly consist of large numbers of volcanic clusters and/or alignments that include hundreds of structures (Condit and Connor, 1996; Connor, 1987, 1990; Connor et al., 1992, 2000; Connor and Conway, 2000; Conway et al., 1998; Valentine et al., 2006). The eruptive history of individual volcanoes within a volcanic field might be short, but the total duration of volcanism in a single volcanic field might exceed the total life of a composite volcano, expanding over millions of years. The total eruptive products of an individual volcano of a

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volcanic field (0.001–0.1 km^3 of dense rock equivalent, DRE) is generally 1–2 orders of magnitude less than in an average sized composite volcano, but the total eruptive volume of a whole volcanic field is in general comparable with the total volume of eruptive products of a composite volcano. Growing evidence from field studies demonstrates that the eruptive history of a volcanic field can be as complex as that of an average composite volcano. Moreover, similar complexity has been recognized in many individual volcanoes of some volcanic fields. These recent findings highlight the need to view volcanic fields in the same manner as we view composite volcanoes.

Monogenetic volcanic fields commonly involve explosive eruptions (phreatomagmatism) driven by interaction of magma with shallow or deep groundwater and/or surface water (White, 1991a). Thus, landforms typically found in volcanic fields include small-volume volcanic eruptions form tephra cones, rings, or mounds consisting of bedded pyroclastic deposits of fallout, pyroclastic density currents, and/or downslope remobilization of tephra (Connor and Conway, 2000; Vespermann and Schmincke, 2000). In many cases, seasonal climatic changes can 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). For these reasons a great variety of volcanic landforms develop especially in lowlands where the hydrogeology of the country rocks may be very complex (White, 1991a, 1991b). The resulting volcanic landforms in such settings are strongly dependent on the nature of the pre-eruptive surface (e.g., depositional environment), lithology, and mechanical properties of volcanic conduit wall rocks, vent geometry, and type and availability of external water (Lorenz, 1987; Valentine and Groves, 1996).

Most volcanological studies of 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. In some cases the exposed inner architecture of monogenetic volcanoes reveals volcanic lithofacies that bear important information on the eruptive mechanism of the volcano (Keating et al., 2008; Lorenz and Kurszlaukis, 2007; Németh and White, 2003). Owing to the long periods involved in the history of a single volcanic field, some individual volcanoes may erode significantly while others are being formed. Thus a volcanic field may be a group of variably eroded volcanic landforms preserved on a gradually eroding landscape (Konecny and Lexa, 2000; Konecny et al., 1999; Lorenz and Haneke, 2004; Németh and White, 2003; White, 1991b). Over longer time periods, a relatively uniform landscape could be dissected, lowered, and commonly inverted, preserving clusters of formerly low-lying zones of the syn-eruptive landscapes in elevated positions (Németh and Martin, 1999). From this point of view, monogenetic volcanic fields are very useful for characterizing the course of erosion of the surrounding syn-eruptive landscape. Erosional remnants of clustered volcanoes of the same age may help to define a geomorphic horizon (Németh and Martin, 1999). With substantial age clusters of volcano remnants, and precise age data for such clusters, a refined erosional history of large areas (hundreds of square kilometers) can be reconstructed.

This paper reviews the physical characteristics, sedimentary processes, and landform evolution of monogenetic volcanic fields. Here the monogenetic volcanic fields are treated in a “holistic” view, where deep, intermediate, shallow, and surface processes are considered as parameters that control the manifestation of volcanism. An introduction to the general structural setting of the fields, their typical shape, vent density, eruption recurrence rates, and relationship between volcanic processes and resulting volcanic landforms is given with special reference to unusual processes and volcanic landforms associated with volcanic fields. Throughout the paper it is argued that formation of volcanic fields is controlled by a combination of deep and shallow processes and physical environments. The eruptive mechanism and the resulting volcanic landforms as part of a volcanic field are dominantly controlled by the magma generation, structural scenario, and the magma’s physical parameters (chemistry, magmatic gas content) commonly referred to as “dry” processes (Valentine and Gregg, 2008). Dry processes form various scoria cones and lava flows. In spite of the predominance of volcanic landforms resulting from dry processes, almost every volcanic field on Earth contains volcanic landforms controlled by the shallow architecture of the substrate and its physical parameters, commonly involving external water. Such processes are referred to here as “wet” volcanic processes. Because of the growing number of information about the role of dry processes in the formation of individual volcanoes of a volcanic field (Valentine and Gregg, 2008), relatively limited attention was given before to the wet processes responsible for volcanism even when the latter commonly represents the most violent and therefore the most dangerous type of volcanism in a volcanic field. Such wet processes are commonly controlled by the physical conditions of the shallow substrate, and therefore there is a major need to understand the state of the sediment substrate (lithification, saturation strength, etc.) beneath a volcanic field in relation to the type of volcanism.

This review starts from the current understanding of the origin of monogenetic volcanism, followed by some aspects of the spatial distribution patterns of monogenetic volcanic fields. The basic volcano-sedimentological characteristics of monogenetic volcanism are given in relation to the surface and shallow subsurface hydrogeological conditions. Finally a summary is presented on the current knowledge of erosional processes in volcanic fields and the use of such data in paleoenvironmental reconstructions.

VOLCANIC FIELD CHARACTERISTICS

Fundamental physical characteristics of volcanic fields that are the focus of current research include (1) the distribution of vents and volcanic complexes (Connor, 1987); (2) the timing and recurrence rates of the volcanic eruptions in a given volcanic field (Condit and Connor, 1996; Conway et al., 1998; Tanaka et al., 1986); (3) the number, type, and eruption history of individual vents (Connor, 1990; Siebe et al., 2005; Valentine and Perry, 2007; Valentine et al., 2006); 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, 2000; Stamatakos et al., 1997).

Volcanic fields may develop as loose clusters or alignments of small-volume volcanoes such as the Hopi Butte in Arizona (White, 1991b), or rarely around a central volcano, such as the volcanic field around Lamongan, in Java (Carn, 2000). Continental volcanic fields are commonly characterized by low magma supply rates, over relatively long periods of time (millions of years) (Connor et al., 2000; Takada, 1994). They typically consist of broadly scattered volcanic vents, such as the Springerville Volcanic Field (Condit et al., 1989; Connor et al., 1992). Strongly aligned volcanic vents, however, are also possible, as for example in the volcanic field near the Yucca Mountains in Nevada (Connor and Hill, 1995; Connor et al., 2000). Vent clustering and individual vent alignment within a field have been documented from the Eifel in Germany (Büchel, 1993; Schmincke et al., 1983), the Western Pannonian Basin in Hungary (Martin and Németh, 2004), and the Auckland Volcanic Field in New Zealand (Cassidy and Locke, 2004; Edbrooke et al., 2003; Lindsay and Leonard, 2007; Magill et al., 2005). In systems with strong fissure behavior, lava tends to be erupted initially over the length of the fissure before being focused at one or two points as larger fire-fountains, or in extreme cases as explosive Strombolian eruptions, as for example at the Parícutin eruption (Luhr and Simkin, 1993). Development of single explosive centers from an initially long fissure is accompanied by a decrease of partial melting and may be related to the length scales, or "magmatic footprints," of mantle source zones for individual volcanoes (Valentine and Perry, 2006). In any case, volcano distribution within a field is generally inferred to be a result of structural control on magma rise.

Volcanic fields generally display changes in the predominant style of eruptive activity throughout their entire life span. For instance, at the Pliocene Crater Flat, Nevada, there was a gradual transition from large-volume, lava-producing effusive and Hawaiian-style lava spatter eruptions along fissures toward eruptions producing single cones via Strombolian (poorly fragmented, discrete eruptive burst) or violent Strombolian (sustained, well-fragmented eruption columns) eruptions (Valentine and Perry, 2006). Similar trends are also documented at other continental volcanic fields such as the Bakony-Balaton Highland Volcanic Fields in Hungary, where large lava fields and shields tend to be older and capped by young, small-volume scoria cones (Martin and Németh, 2004; Wijbrans et al., 2007). This appears to show that in the waning phase of a volcanic field there is a reduction in the magmatic flux and a decreasing effusion rate and eruptive volume (Valentine and Perry, 2006).

Classification of volcanic processes can also be made based on the detailed study of the time-averaged volumetric volcanic output rates in relationship with compositional variations and tectonic settings (White et al., 2006). A worldwide compilation of repose periods between successive eruptions and intrusive to extrusive ratios of volcanoes and volcanic fields over long periods (>10^4 years) showed that average long-term volcanic output rates tend to decrease from basaltic to more evolved compositional ranges, and in this regard to show differences between oceanic and continental settings. In this case, volcanic output rates and intrusive to extrusive eruptive products are controlled by local crustal thickness, tectonic settings (magnitude and orientation of principal stresses), magma composition, and melt generation rates in the source region. Continental volcanic fields in this view differ significantly from large volume, commonly silicic volcanic systems, as the former occupy large areas (hundreds of square kilometers), are usually erupted over long periods (millions of years), and are emplaced over thick (>30 km) crust (White et al., 2006).

Continental volcanic fields are commonly grouped according to their total eruptive products over time. In this classification scheme, two end members can be identified: (1) low eruptive volume flux, and (2) high flux fields. Typical low flux fields are the Auckland Volcanic Field in New Zealand (Briggs et al., 1994; Magill et al., 2005) and the Southwestern Nevada Volcanic Field (Valentine and Perry, 2007). A high-volume flux field is exemplified by the Eastern Snake River Plain Field (Kuntz et al., 1986). These end-member volcanic fields have fundamentally different time-volume relationships of their eruptions as well as different relationships between shallow dikes, sills, vents, and preexisting structures. These fields also appear to be time-predictable, in that a linear relationship exists between the timing of an eruptive episode and the cumulative volume of earlier episodes (Fig. 1). This implies that the repose time after a volcanic event is proportional to the event's volume, and therefore the timing of future episodes can be predicted on the basis of the preceding volumes (Valentine and Perry, 2007).

Although such tendencies are correct in general, the time-volume relationship is not a precise indicator of eruption onset, as other major volcanism-triggering processes, such as major tectonic or magmatic events, may overshadow the general trend. In some fields, tectonic control on magma flux seems to be very low. Magma generation therefore seems to be controlled by regional tectonic strain as a result of slow magmatic pressure buildup in localized zones. In each melt release, magma forms dikes that propagate toward the surface, releasing local stress. As large volumes of melt leave the system, longer periods are needed to "reload" the stress before a new melt-release event can take place. Crustal strain can also be released by faulting and/or magmatism (e.g., dike emplacement). In low-volume flux fields, faulting is the dominant strain relief; hence these structures capture rising melts, producing aligned dikes and vents. Furthermore, the similarities of the overall melt volume calculated for long-lived single polygenetic volcanoes and entire monogenetic volcanic fields (Connor and Conway, 2000) can be explained if total volumes of available melt are controlled by large scale mantle movements and anomalies in the same volume range regardless of the tectonic situation.

The volcanoes typical of volcanic fields are generally small in volume, but on closer inspection they often show signs of

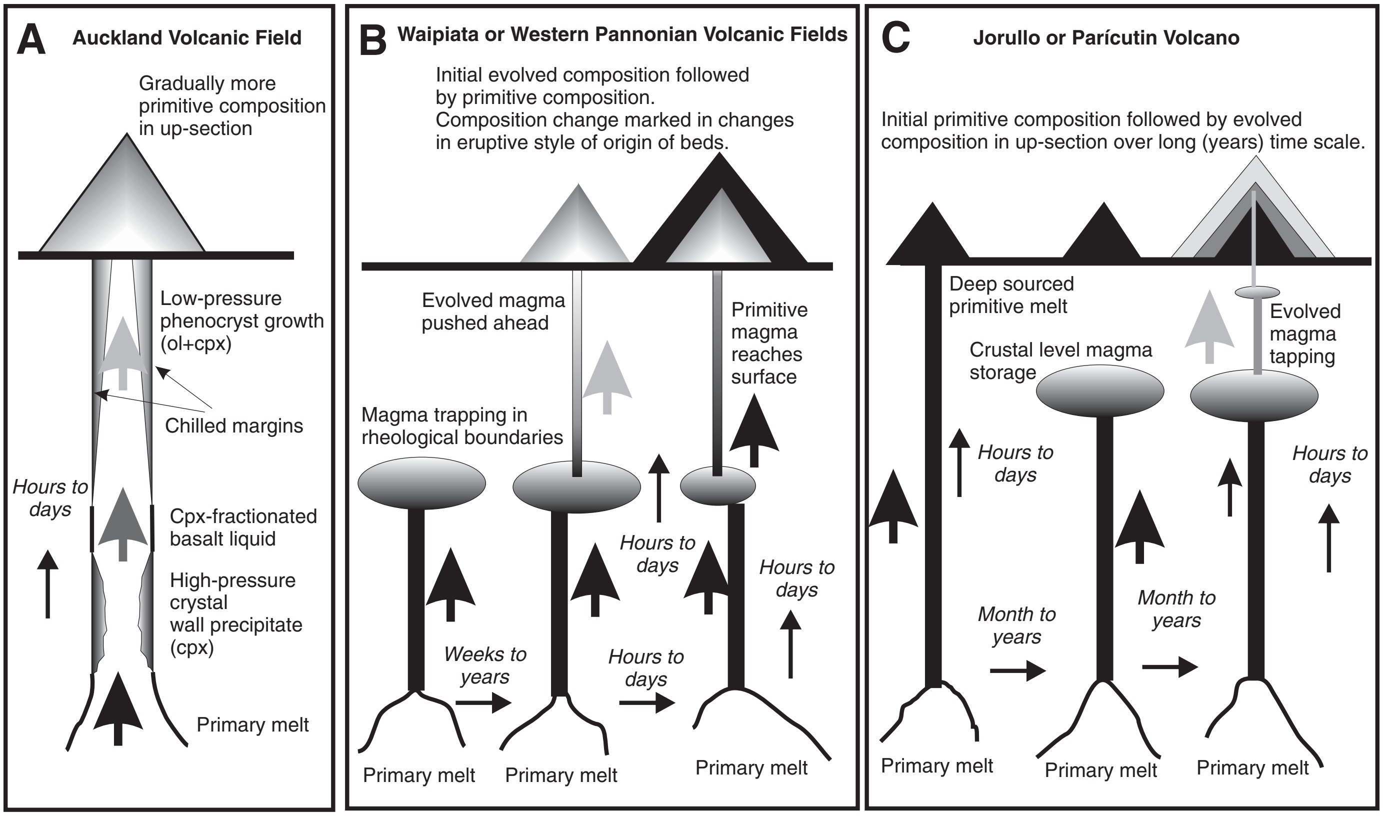

Figure 1. Three theoretical models for the plumbing system of monogenetic continental volcanoes based on chemical signatures of erupted materials. (A) Model based on Crater Hill in the Auckland Volcanic Field (Smith et al., 2008) operates on the identified chemical evolution from evolved to more primitive compositions in a single uninterrupted sequence of a phreatomagmatic to magmatic explosive volcano proximal succession. (B) Model based on analysis of basal phreatomagmatic volcanic glass shards and late magmatic explosive juvenile pyroclasts, demonstrating trends from evolved to more primitive compositions in the sequence of proximal complex phreatomagmatic volcanoes from southern New Zealand and across the western Pannonian Basin, Hungary (Németh et al., 2003). (C) Model based on the two largest and longest lived young scoria cones (Parícutin and Jorullo in Mexico), demonstrating initial early primitive eruptive products gradually overlain by more evolved younger eruptive products over 10 yr of activity (Johnson et al., 2008). In each diagram evolved magmas are represented by light-gray patterns; ol—olivine; cpx—clinopyroxene.

multiple eruption phases that may have lasted over several years. Therefore their architecture can be complex regardless of their small size, such as has been documented in the Rothenberg scoria cone in Germany (Houghton and Schmincke, 1989) or the Sinker Butte (Fig. 2) in Idaho (Brand and White, 2007). It is also notable that volcanic fields in continental settings are often associated with shield volcanoes and lava flow fields, products that commonly account for half the total eruptive volume in these cases (Greeley, 1982; Hasenaka, 1994; Németh, 2004; Walker, 1993).

In magmatically controlled volcanic fields, each eruption occurs when pressure builds up in the magma storage reservoirs, and therefore the magma rise will purely depend on melt accumulation, fractionation, and concentration of volatiles (Valentine and Perry, 2007). As a result, the volume of each eruptive episode will be dependent on the magma flux and the repose time, as the previous eruption possibly generated a volume-predictable behavior for some fields. At magmatically controlled fields the melt production rate is purely related to the thermal structure of the mantle with less dependence on the tectonics (Valentine and Perry, 2007).

Only very small volumes (0.01 km^3 or less) of primary magma are needed to generate monogenetic volcanoes. Such small volumes of magma may "freeze" in the feeding system in only a few days to weeks. Despite this, some scoria cones show a gradual transition toward composite volcanoes (Fig. 3) and are difficult to classify in terms of monogenetic and polygenetic systems (McKnight and Williams, 1997). On the basis of a handful of historic eruptions, these volcanoes are large in volume and have produced tephra commonly with a great diversity in composition and fragmentation.

In general there are three major elements to be considered in the ascent and emplacement of magma either on Earth or other planets: (1) magma generation and buoyancy, (2) rheological boundaries in the lithosphere, and (3) density boundaries in the lithosphere (Walker, 1989). Each of these factors strongly depends on the physical properties and structure of the

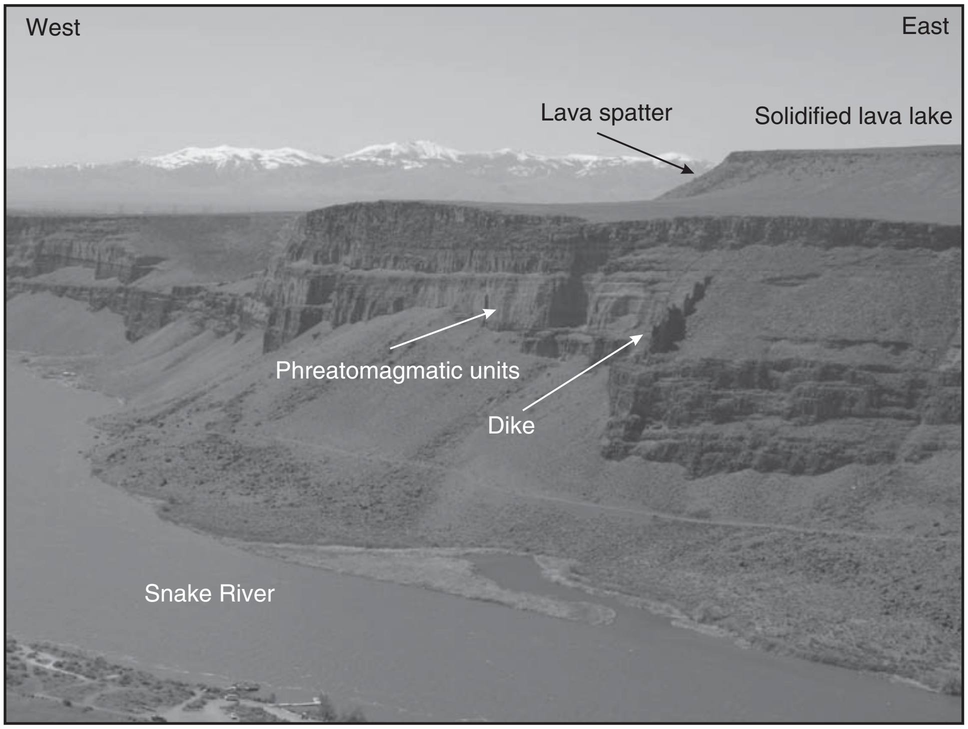

Figure 2. Sinker Butte (Idaho) is a large phreatomagmatic volcano representing a long-lasting eruptive history from initial phreatomagmatic to effusive eruptions. The lava spatter dominated the top section, and its lava prevented most of the volcanic complex from erosion.

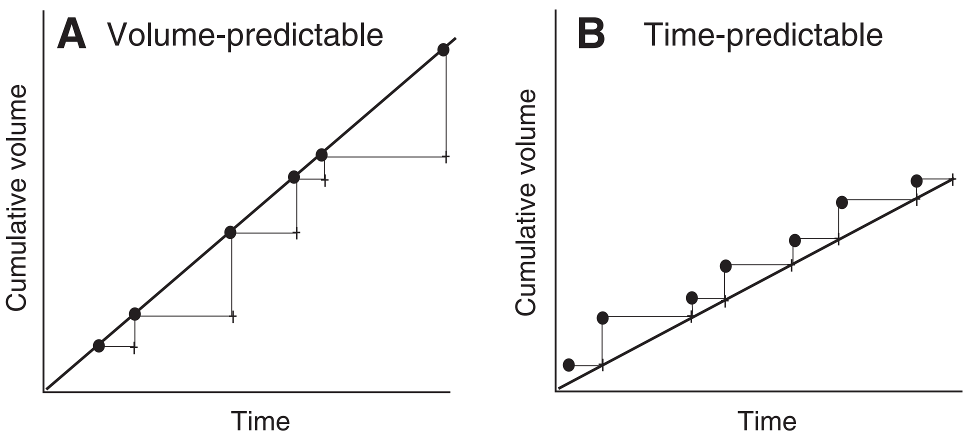

Figure 3. Diagrammatic representation of volume- and time-predictable volcanic field behavior based on models of Valentine and Perry (2007). Filled circles represent the ages of episodes and the cumulative volume of those episodes and all preceding episodes. Crosses represent the ages of episodes and the cumulative volume of the preceding episodes only.

lithosphere encountered by the magma. 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 (Parsons et al., 1992; Parsons and Thompson, 1991; Valentine and Krogh, 2006). Local fissure lengths are inferred to reflect the aerial extent of the source melt anomaly, and their measurement can bear key information about the evolutionary stage of a volcanic field (Valentine and Perry, 2006). Nevertheless, the dike-vent orientation may not relate in a systematic way to the least principal stress (Gaffney et al., 2007; Valentine and Perry, 2007), because faults are commonly long-lived features that may reflect a previous stress regime of the area. On the other hand, in fields characterized by high magma fluxes (commonly referred to as magmatically controlled fields), frequent dike injection can produce comparable or larger magnitude strain release than faulting, and therefore faults may be rare in such areas. The resulting volcano-distribution pattern may be random (but might be clustered), and the dike orientation in a field will show perpendicular orientation with the prevailing least principal stress because the dikes are not captured by any existing background fault network.

Monogenetic versus Polygenetic Volcanism

A simple tectonic-based classification will place most volcanoes on Earth into four settings: (1) rift volcanoes (Walker, 1999); (2) subduction volcanoes, including front- and backarc volcanoes; (3) hot-spot volcanoes (Jellinek and Manga, 2004); and (4) intraplate volcanoes (Johnson, 1989). These end members can explain many features associated with magmatism in relation to chemistry and magma supply rates. Complexities arise, however, because volcanic fields can be found in each of the four mentioned settings, and individual volcanoes within a field are commonly considered to be monogenetic. The search for understanding the relationship between tectonic settings and type of volcanism (polygenetic versus monogenetic), therefore, is a long-standing goal in volcanology.

The formation of monogenetic versus polygenetic volcanoes has been explained on the basis of direct influence of the tectonic regime of the volcanism, based on the liquid-filled crack interaction theory (Takada, 1994). Owing to the generally small volumes of magma involved during scoria cone forming and other continental monogenetic volcanic eruptions (days to 15 yr) it is generally assumed that these magmas undergo limited evolution through fractional crystallization and assimilation. However, weak compositional variations in eruptive products of monogenetic volcanism are increasingly commonly recognized phenomena (Németh et al., 2003; Smith et al., 2008). Small-volume eruptions in alkali basalt fields such as the Auckland Volcanic Field, New Zealand, are inferred to be compositionally zoned as a result of changing degrees of crystallization along the walls of dikes that carry primitive magmas from deep source regions (Smith et al., 2008; Brenna et al., 2010). The small-volume magma involvement in most of the monogenetic volcanoes of a continental volcanic field is the likely reason that the effect of crystallization at different levels in the conduit can be inferred (Smith et al., 2008). For rare, longer lived continental monogenetic volcanoes such as Jorullo or Parícutin in Mexico, more complex plumbing systems are expected (Johnson et al., 2008). Disparities between the compositions of the melt inclusions and the whole-rock lava samples at Jorullo gave evidence for a two-stage crystallization process of the rising magma (Johnson et al., 2008). Both Jorullo and Parícutin show evidence that magmas evolved during the course of the several year-long eruptions during fractional crystallization and assimilation; the exact level of such processes, however, is under debate. Residence time calculations from olivine crystals at Jorullo indicated an up to 1300-day period before they reached the surface (Johnson et al., 2008). Eroded scoria cones and other monogenetic volcanoes in many places are associated with shallow-level complex networks of dikes and sills where magmas could evolve (Németh and Martin, 2007; Valentine and Krogh, 2006); however, the development and evolution of such plumbing systems during the course of an eruption are not known, or are hard to test. At Jorullo, during the 15 yr eruptive history, early eruptive products were more primitive basalt that changed to more evolved basaltic andesite over time (Luhr and Carmichael, 1985). The early Jorullo melts are inferred to have risen from depth through a complex network of dikes and sills where olivine crystallized ~16 km below the surface (Johnson et al., 2008). The later eruption phase records olivine crystallization at a very shallow level (80–700 m) and the addition of plagioclase ± clinopyroxene as crystallizing phases, all indicating some sort of accumulation zones of melt beneath the volcano (Johnson et al., 2008). Such shallow storage zones of melts are predicted to be associated with long-lived monogenetic volcanoes. The above-mentioned two cases can be viewed as two end members of small-volume continental volcanoes, either as small-volume short-lived (Auckland Volcanic Field) or as large(r)-volume, long-lived (Jorullo and Parícutin) volcanoes. In Auckland (small-volume, short-lived eruption) the fractionation is inferred to be taking place in the melt column tapping primitive melt from the source region, whereas at Jorullo or Parícutin (large[r]-volume, long-lived eruption) had a significantly longer time frame (longer than the time required to solidify a dike from a deep source to the surface), and therefore the eruption products represent eruption episodes, each fed by individual magma batches and gradually forming shallow magma storage places. A somewhat opposite trend to Jorullo and Parícutin, but similar to the Auckland chemical evolutionary trend, was documented in continental volcanic fields of the Waipiata Volcanic Field in southern New Zealand and in the central Pannonian Basin, Hungary (Németh et al., 2003). In these volcanic fields the early eruptive products commonly associated with phreatomagmatism demonstrate more evolved character than those formed in the later stage but within a very limited time frame (e.g., hours to days). This evolutionary trend has been explained by short-term trapping processes potentially at the crust-mantle boundary, where melt evolution took place. Further melt propagation was triggered by renewed primitive magma input, which pushed the small volume of evolved melt ahead to become involved in phreatomagmatism in the initial stage of the eruption. However, no further study has been conducted to compare in detail the compositional evolution of the uppermost, mostly magmatically fragmented successions where trends similar to the Auckland Field may be expected. It seems that magmatic plumbing systems of continental volcanic fields are complex, and the Auckland, Jorullo, and Waipiata-Pannonian Basin examples demonstrate only a small fraction of possible variations in relation to regional tectonic settings (Fig. 4).

The volcanic-system model of Cañón-Tapia and Walker (2004) gives a significantly different perspective of the formation of monogenetic versus polygenetic volcanoes. This model of volcanism incorporates the views of Walker (1993), Gudmundsson (1995), and the results of melt generation models of Marsh (2000). In this combined approach, the birth of a volcanic system is defined by the time when the hydraulic fracturing allows a significant amount of magma to be transported out of the source region. A basic hypothesis of this model is that large zones completely filled with melt are unlikely to exist in nature, and considers it more realistic to assume that melt exists in mineral channel networks that might extend laterally for large distances. The tapping process therefore is largely influenced by the geometry of this channel network. Cañón-Tapia and Walker (2004) envisioned three basic scenarios of tapping. In the first of these scenarios it is assumed that there is enough interconnected melt in the source

Figure 4. Half section of a transitional volcanic complex from the Western Snake River Plain Volcanic Field. Along the Snake River, half sections of mafic volcanoes expose complex evolution from phreatomagmatic to magmatic explosive eruptions over time. The durations of the eruption of these volcanoes are largely unknown; however, there are no significant time break indicators, such as erosional surfaces, that could suggest that these volcanoes were long-lived. The preserved eruption volume, however, indicates that they may have been active longer than a moderate-sized scoria cone.

region, and consequently that the pressure at the entrance of the conduit can be maintained high enough to pump magma at constant or increasing levels. If the minimum component of stress is horizontal, then the tapped magma will be able to rise to the surface. Tapping will continue until magma pressure at the entrance of the conduit drops because the melt is exhausted locally around its entrance, therefore ending eruptive activity at the surface. The availability of magma laterally in this case would provide slight shifts of locations where hydraulic fracturing can take place in a short time scale. Each shift would produce a slightly different single volcanic edifice at the surface to define monogenetic volcanoes. If shifts of the magma entrance to a vertical conduit take place rapidly, newly segregated rising melt could use the same, still “warm” path to the surface, resulting in complex volcanoes that are predominantly amalgamated monogenetic edifices, a commonly observed volcanic architecture associated with long-lived volcanic fields. Over longer periods of time, if the mantle anomaly changes its level slightly (e.g., rising), a gradual compositional evolution can be observed over the life of the volcanic field. This gradual shift in composition of the erupted products has also been documented geologically, as discussed above.

A second tapping scenario proposed by Cañón-Tapia and Walker (2004) also assumes similar availability of interconnected melt channels from a large-volume region of partial melt. The difference from the first scenario is that the minimum principal stress changes to a subvertical position along the way to the surface. Owing to this change in orientation of the minimum stress, magmas will never reach the surface. Instead, the tapped magma would be diverted to lateral zones such as sills. Changes in principal stress orientations (e.g., switching between horizontal to vertical), triggered by regional tectonic events, may result in rhythmic evolution of a monogenetic field and large-scale sill-forming periods. The third tapping scenario assumes that the interconnected magma volume in the source region is not large enough to sustain the pressure at the entrance of the conduit, regardless of the orientation of the actual stress field. In this scenario, large volumes of magma could remain trapped between the source and surface, and these shallow zones of magma storage may evolve for long periods. According to Cañón-Tapia and Walker (2004), the presence of such shallow magma storage regions controls the movement of other magmas to the surface, and a shallow central conduit system will develop upon continued feeding of new magmas, eventually producing a polygenetic volcano with evolved magmas.

One important aspect in the model of Cañón-Tapia and Walker (2004) is that the melting rate seems assumed to be fixed and relatively low regardless of tectonic scenarios. However, this model of volcanism is not explicit concerning important aspects that include compositional variations of the erupted products. If the melt is considered to be produced in a mantle of homogeneous composition, the model of volcanism proposed by Cañón-Tapia and Walker (2004) would be in conflict with observations because there is increasing evidence suggesting that the mantle sources of volcanism are heterogeneous. Consequently, the surface manifestation of magmatism can be very diverse simply on the basis of this heterogeneous source (Anderson, 2006; Bergmanis et al., 2007; Davaille, 1999; Ito and Mahoney, 2005). Nevertheless, if the heterogeneous character of the mantle source is incorporated into the model of volcanism proposed by Cañón-Tapia and Walker (2004), the coexistence of diverse types of magma within a single volcanic field can be explained relatively easily, as the model would allow each of these different heterogeneities to be tapped somewhat independently of the others. The assumptions made concerning melting rates, however, might be more difficult to incorporate in this model and might prove to be critical and probably at the center of the argument concerning how monogenetic and polygenetic volcanoes can form. Unlike the assumed uniform rate of melting, melting rates can be demonstrated to be very different in different locations. These variations may or may not be related to a geotectonic setting, but clearly they are related to the structural set of a region and primarily depend on heat flux, fluid supply, mantle convection, and mantle fertility (Ballmer et al., 2007; Cannat et al., 2008; Cohen and Onions, 1993; Eggins, 1992; Huang and Lundstrom, 2007; Montesi and Behn, 2007; Regelous et al., 2008; Smith and Lewis, 1999; Storey et al., 2007; Stracke et al., 2006; Turner et al., 2006). Future research objectives therefore could be to find the relative role of each of the above parameters in the formation of poly- versus monogenetic volcanism.

CHARACTERISTICS OF INDIVIDUAL VOLCANOES IN VOLCANIC FIELDS

Dry Eruptive Processes in Monogenetic Volcanic Fields

Dry eruptive processes are considered to be those in which the volcanic eruptions are primarily controlled by the gas content and physical parameters (e.g., viscosity, temperature, magma flux) of the rising magma. The external water of the shallow subsurface environment has no influence on this type of volcanism. Dry eruptive processes range from lava effusion to low and then to high explosivity processes, all of which are reflected in very different volcanic landforms at the surface. The volcanic landforms formed by dry eruptive processes preserved at the surface have distinctive geometry, architecture, and deposits (in the case of explosive eruptions) as well as transitional features that can be observed and measured.

The individual lava flows associated with continental volcanic fields tend to be ~1–10 km long (Connor and Conway, 2000; Kilburn, 2000; Valentine and Perry, 2006). Exceptionally long lava flows, however, are also known such as in many lava fields in the Cenozoic volcanic fields in Patagonia, Argentina (Gorring et al., 1997). The total thickness of lava could reach several tens of meters and could blanket significant parts of a volcanic field, as in the Western and Eastern Snake River Plains in Idaho (Greeley, 1982; Hughes et al., 2002). Strombolian scoria cones and Hawaiian spatter cones are commonly sources of lava flows. Basaltic volcanic fields are 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, sky rise, whaleback humps, lava tubes, and pressure ridges (Kilburn, 2000). Among these features, tumuli are whaleback-shaped uplifts and 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), and in eastern Australia (Ollier, 1964; Wilmoth and Walker, 1993). 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 with their distance from their source (Rossi and Gudmundsson, 1996). Large tumuli are comparable in size to small lava-spatter cones, and therefore their recognition is important in establishing the eruption history of volcanic fields. It has been demonstrated that recognition of tumuli features and their characteristic surface textures may help quantify eruption duration (Mattsson and Hoskuldsson, 2005).

Flow localization of rising melt captured in fractures tends to form a cylindrical upper conduit (100 m) that commonly feeds scoria and spatter cones (Gaffney et al., 2007; Keating et al., 2008; Valentine and Krogh, 2006). Spatter cones consist of near-vent, strongly baked, red, slightly bedded stacks of ejecta, dominated by spindle or highly vesiculated fluidal bombs. These deposits usually reflect piling and deformation of volcaniclastics in near-vent positions. Spatter cones and scoria cones can also build up steep piles that may be over-steepened and collapse gravitationally. Parts of cones can also be rafted away by moving lava flows initiated from the base of the cone. Among many examples, Los Morados in southern Mendoza, Argentina, shows one of the best exposed and preserved half-collapsed scoria cone (Fig. 5).

Magmatic explosive and/or degassing processes as the result of fragmentation of the uprising mafic magma lead to the formation of scoria cones, often with welded core zones (Fig. 6) (Vespermann and Schmincke, 2000). The textural characteristics of the pyroclasts, such as high vesicularity, fluidal shape, and dark, often red color indicate a magmatic degassing and fragmentation history within Strombolian-style explosive eruptions (Jaupart and Vergniolle, 1988; Sumner, 1998; Valentine and Keating, 2007; Valentine et al., 2005, 2006; Vergniolle and Brandeis, 1996; Vergniolle et al., 1996; 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 (Sumner et al., 2005; Thordarson and Self, 1993; Vespermann and Schmincke, 2000; Wolff and Sumner, 2000). The common intercalation of scoria beds with welded fall-out deposits and/or clastogenic lava flows indicates sudden and repeated alternations in eruption style between Strombolian and Hawaiian volcanoes, and vice versa (Parfitt and Wilson, 1995; Parfitt et al., 1995; Valentine et al., 2006; Wilson et al., 1995). In addition, pyroclastic breccias, lapilli tuff, and tuff interbeds in scoria cones are common signs of a phreatomagmatic influence on the eruptions (Doubik and Hill, 1999; Houghton and Hackett, 1984; Houghton et al., 1984, 1999; Martin and Németh, 2006; Risso et al., 2008).

In spite of the numerous scoria cones associated with volcanic fields and central volcanoes (e.g., along rift zones), only

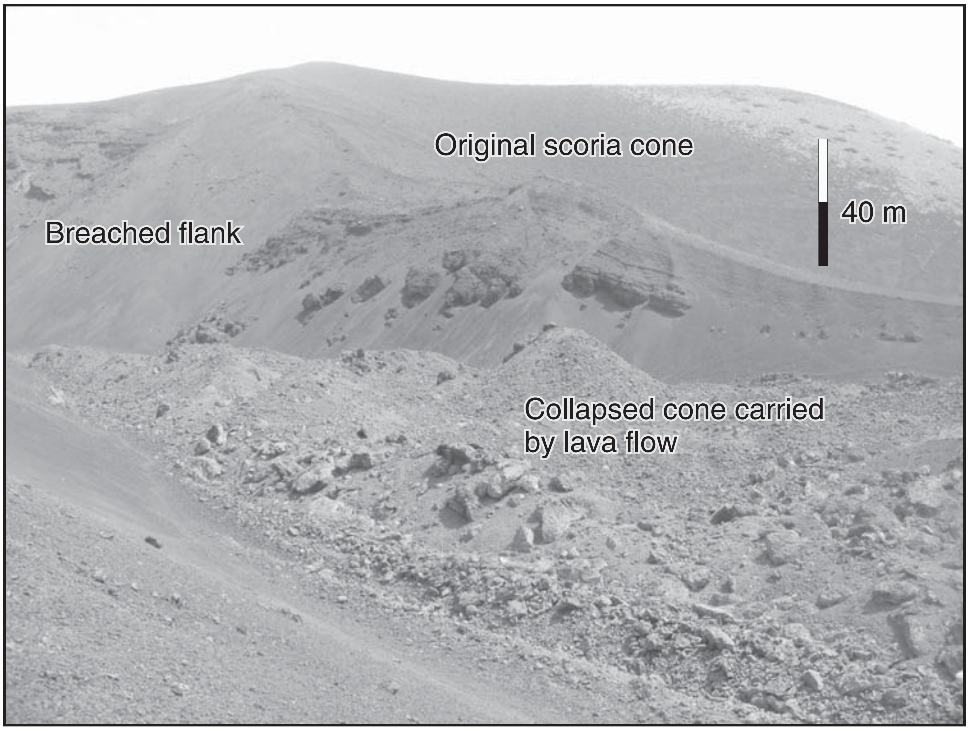

Figure 5. Los Morados (Argentina) scoria cone has an open side. This crater wall breaching is inferred to be a result of the combination of gradual over-steepening of the lava spatter dominated near the vent cone flank and initiation of a flank lava flow. The moving lava flow gradually carried away the collapsed lava-spatter-dominated fragments from the cone flank.

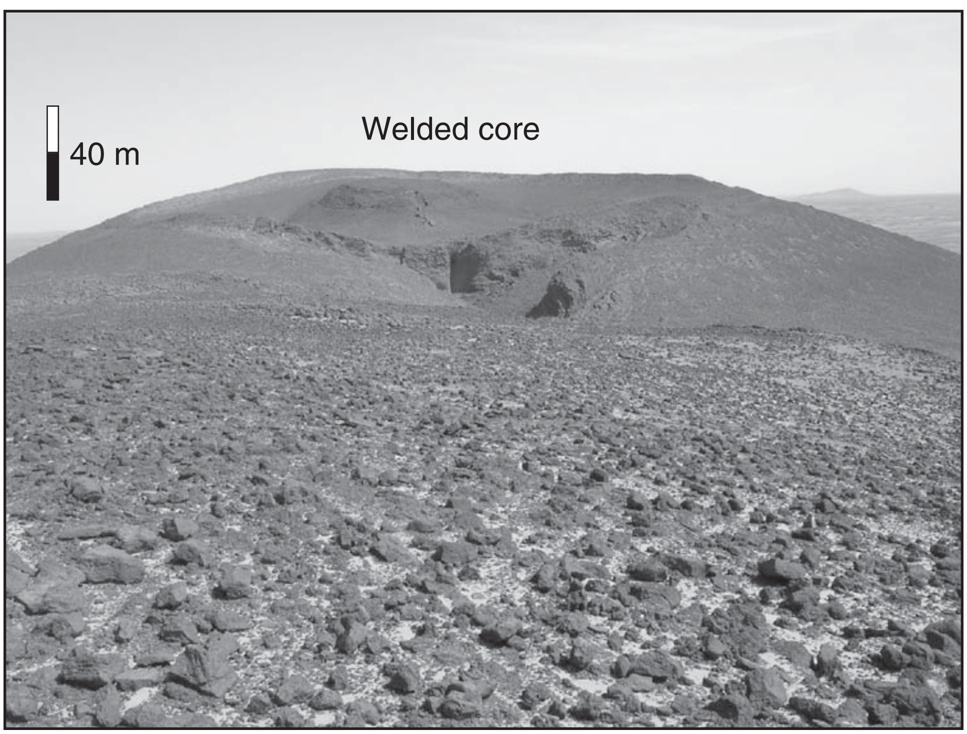

Figure 6. Welded core zone of a scoria cone from the Al Haruj, Libya, shows a lava-like texture.

a few detailed studies have been carried out on their architecture (Chouet et al., 1974; Head and Wilson, 1989; Keating et al., 2008; McGetchin and Settle, 1975; McGetchin et al., 1972, 1974; Riedel et al., 2003; Valentine and Keating, 2007; Valentine et al., 2006, 2007). The inner parts of scoria cones usually consist of welded agglutinate, which is more resistant to erosion and hence can be preserved for long periods, leaving a steep sided feature that mimics that (in shape, not size) of young cones (Fig. 7). Such eroded remnants are common in arid climates such as central Libya in the Haruj Volcanic Field (Martin and Németh, 2006; Németh, 2004). Detailed analyses of deposits preserved on scoria cones are used to examine the role of shallow-seated magmatic systems in controlling explosive eruptions of such volcanoes (Houghton et al., 1999; Keating et al., 2008; Valentine and Keating, 2007). Variations in degassing patterns, magma-ascent rates, and degrees of interaction with external water are thought to be responsible for sudden changes in eruption sequences from deposits representative for both wet and dry eruption conditions (Houghton et al., 1999; Keating et al., 2008). In general, scoria cone–forming eruptions are linked to Strombolian-type activity driven by magmatic fragmentation that occurs in the near-surface region of the open volcanic conduit (Blackburn and Sparks, 1976; Houghton et al., 1999; Keating et al., 2008). Among scoria cones, a great variety has been observed and described, reflecting gradual transitions between Hawaiian lava fountaining to moderate Strombolian-type eruptions. It has been suggested that variations in the magma-ascent rate is the most important factor that causes 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). Further decreases in gas content cause a transition to more passive effusion of vesicular lava (Parfitt and Wilson, 1995). Some authors suggest that a change from the Hawaiian to the Strombolian style requires a significant reduction in magma ascent speed (Parfitt and Wilson, 1995). “Violent” Strombolian eruptions are explosive eruptions of mafic magma characterized by eruption column heights of <10 km, voluminous ash production, and simultaneous lava effusion (Houghton et al., 2004). The mechanism of generation, fragmentation, transportation, and deposition of ash in these eruptions has been investigated in detail at Volcán Parícutin, Mexico (1943–1953). Parícutin eruptive products include thick tephra deposits of alternating ash and lapilli beds distributed several kilometers from the cone-building pyroclastic units (Luhr and Simkin, 1993; Newton et al., 2005). Comparative studies show 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 evidence from the western Transmexican Volcanic Belt indicates that violent Strombolian-style mafic eruptions among scoria cones may be more common than heretofore suspected (Martin and Németh, 2006).

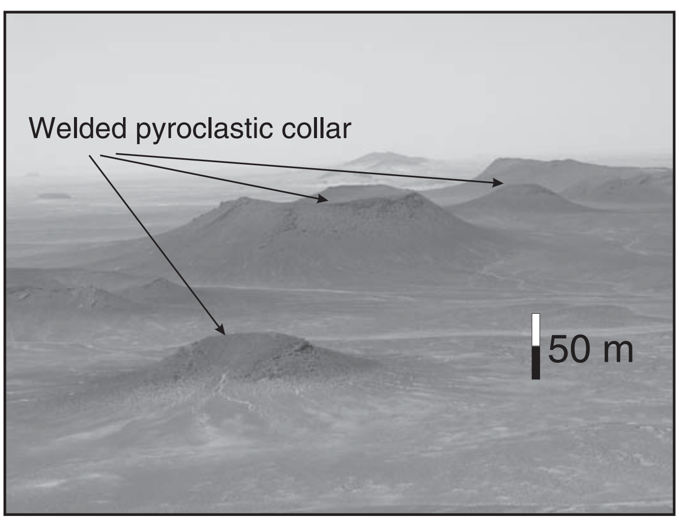

Figure 7. Welded top of scoria cones (Al Haruj, Libya) are more resistant against erosion, and they function as a cap over the loose ash- and lapilli-dominated volcanic structure. The erosion-modified morphology from a first site indicates a young morphological feature for this type of cone in spite of the large volume that has already been removed from the original cone.

Wet Eruptive Processes in Monogenetic Fields

Phreatomagmatic volcanoes in a volcanic field are commonly associated with lowlands or valleys (Lorenz, 1973, 1986; 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 fluviolacustrine basin such as the western Pannonian Basin Miocene-Pliocene fields (Martin and Németh, 2004) and the Snake River Plain (Godchaux et al., 1992; Wood and Clemens, 2004) volcanic fields are of great volcanological and paleogeomorphological interest. The relatively long volcanic history of such volcanic fields and the adjacent lake-fluviatile environment make them ideal for studying sublacustrine, perilacustrine, and postlacustrine volcanism, such as at the Western Snake River Plain in Idaho (Godchaux and Bonnichsen, 2002; Godchaux et al., 1992). Such fields provide an excellent opportunity for developing our knowledge of eruptive mechanisms that result from a wide spectrum of possible magma-water interactions with a special relationship to the paleoenvironment, the paleohydrology, and the physical characteristics of pre-volcanic units.

Maars are small-volume volcanoes and the second-most common volcanic landforms on Earth (Lorenz, 1985, 1986). Maar volcanoes have characteristically wide and deep craters, commonly referred to 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 (Buttner et al., 2002; Lorenz, 1986; Zimanowski et al., 1991, 1997). The sudden formation of steam during magma and water interaction produces explosions that disrupt the country rocks (Zimanowski et al., 1991, 1995, 1997, 1998) and create a mass deficit in and around the explosion center, which leads to the formation of a collapse crater. The explosion locus gradually migrates downward owing to the gradual exhaustion of water sources to fuel explosions (Lorenz, 1986). The various individual explosion sites or chambers jointly form the root zone of the volcano (Lorenz, 2000, 2002; Lorenz and Kurszlaukis, 2007). Via these processes a diatreme forms, and, in principle, it represents a collapse feature like a sinkhole (Lorenz and Kurszlaukis, 2007). Downward explosive penetration of the root zone on its own feeder dike and consequent collapse phases of the diatreme leads to a growing diatreme and a growing maar crater above (Lorenz, 1986; Lorenz and Kurszlaukis, 2007). The eruptive environmental conditions (including magma and host-rock characteristics) may cause variability in size and shape of maars and the characteristics of their deposits. Maar volcanism can also involve eruptive stages of magmatic explosivity, especially in fields with limited water availability such as the Quaternary Llancanelo Volcanic Field in Argentina (Risso et al., 2008) (Fig. 8). Magmatic phases at the end of a maar eruption may even generate a lava lake. Because the formation of maars is driven by the interaction of groundwater and uprising melt, after the collapse of the crater groundwater 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, as has been documented from the Ukinrek maar (Pirrung et al., 2008) that formed in 1977 (Self et al., 1980) or from the Oligocene Messel maar, Germany (Pirrung et al., 2001; Schulz et al., 2005).

A diatreme is the substructure of a maar crater and its tephra ring (Lorenz, 1986; Lorenz and Kurszlaukis, 2007; White, 1991b). In maar-diatreme volcanoes a large amount of fragmented country rocks and commonly juvenile lapilli and bombs are ejected (Lorenz and Kurszlaukis, 2007; Lorenz et al., 2002). In Argentina, for instance, in the center of the Patagonian mafic Cenozoic plateau lava fields, newly discovered diatremes stand ~100 m above the surrounding plain (Fig. 9), exposing their feeding dike systems (Martin and Németh, 2007; Martin et al., 2005; Németh et al., 2007b). Diatremes themselves are cone-shaped volcanic structures cut into pre-eruptive rocks (Fig. 10). They are as deep as 2.5 km and up to 1–2 km in upper diameter. 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. Uprising magma from the underlying feeder dike 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 and Kurszlaukis, 2007; Lorenz et al., 2002; Németh et al., 2007b).

In a "soft substrate" environment maar volcanoes are broad (Fig. 11) and are underlain by "champagne glass"–shaped diatremes (Auer et al., 2007; Lorenz, 2003; Sohn and Park, 2005). In contrast, the crater wall of maar volcanoes that erupted through a "hard rock" environment are steep, filled with volcaniclastic delta deposits, and underlain by deep diatremes (Auer et al., 2007; Lorenz, 2003; Sohn and Park, 2005).

The West Eifel Volcanic Field, Germany, represents a classic example of how preexisting country rock structures have

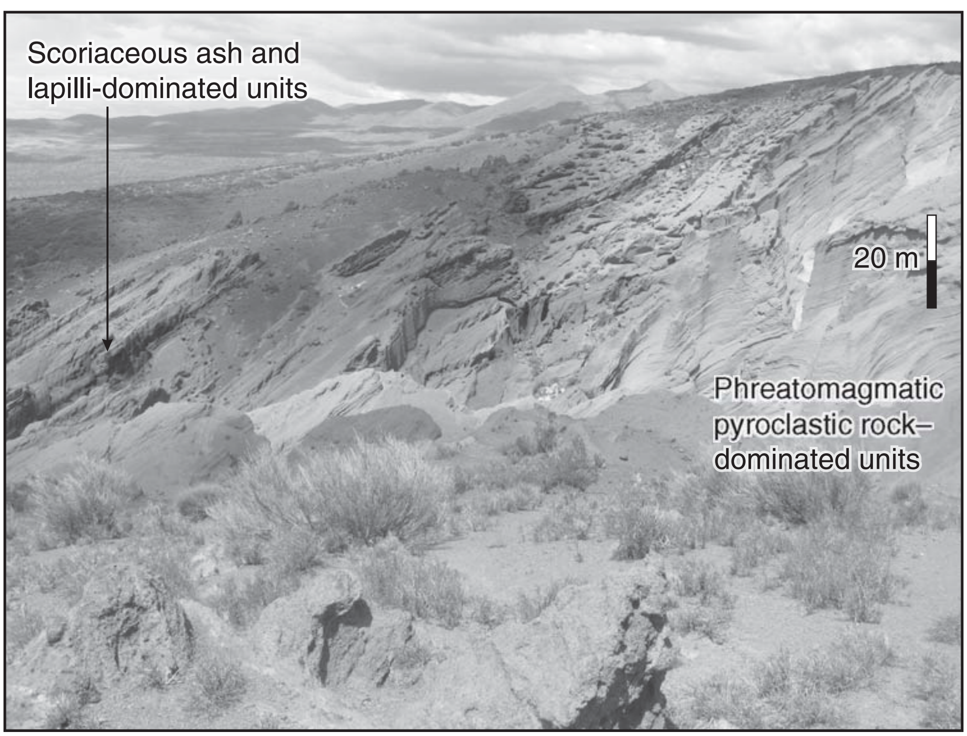

Figure 8. Malacara Volcano is a complex transitional, "wet-dry" mafic explosive volcano, part of the Llancanelo Volcanic Field, Argentina. The volcano is characterized by a complex basal succession of pyroclastic units indicating magma-water interaction. Such beds are intercalated with magmatic gas-driven explosion-generated "dry" pyroclastic beds. The topmost succession of the volcano is fine lapilli, coarse ash succession, indicating sustained eruptions of lava fountaining and Strombolian style eruptions.

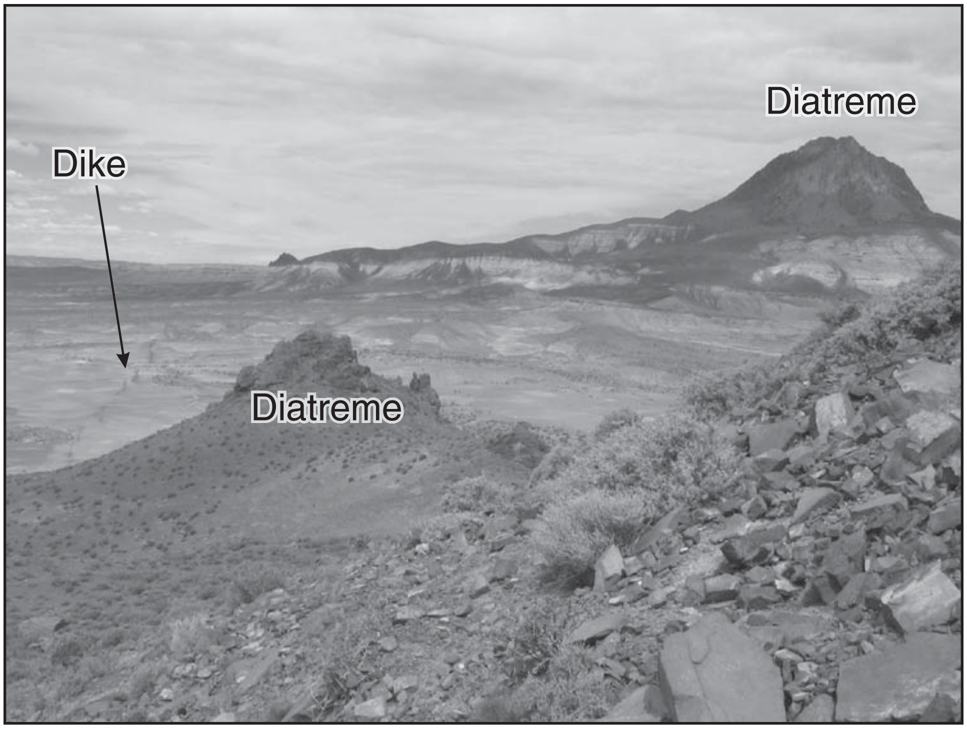

Figure 9. Diatreme field in Chubut, Argentina, represents deeply eroded maar-diatreme volcanoes erupted in a region with good water availability. The recognition of such diatremes can help to conceptualize the syn-eruptive paleoenvironment of a field.

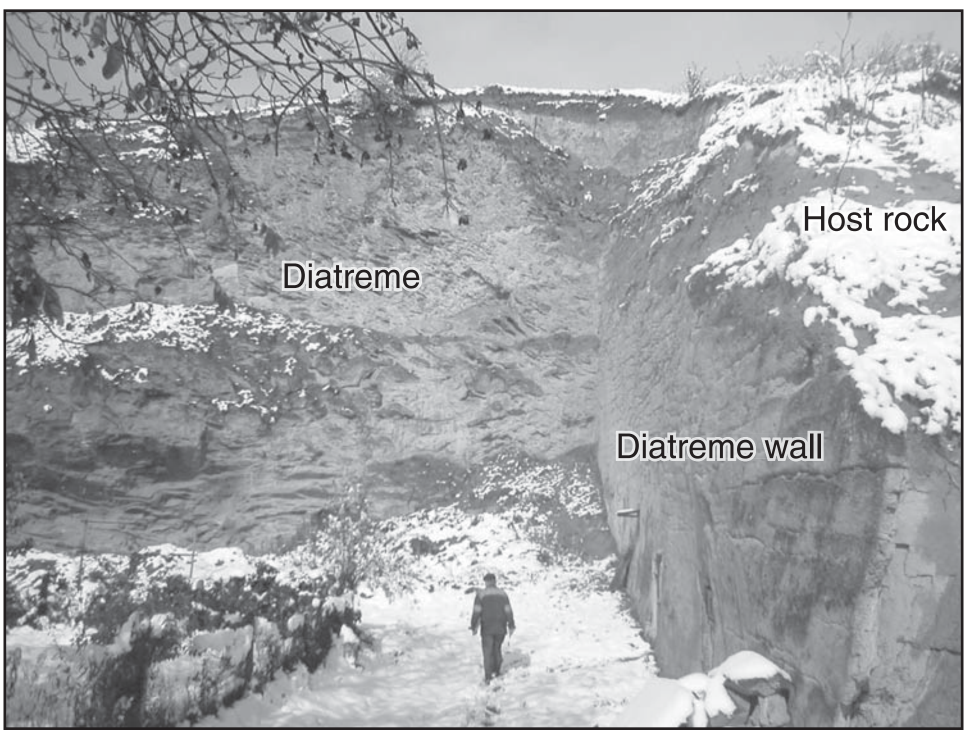

Figure 10. A sharp contact between diatreme fill (right) and host siliciclastic sediments (left) from the Southern Slovak Basaltic Volcanic Field (Slovak Republic).

Figure 11. Flat and broad maar craters of the Pali Aike Volcanic Field, Argentina, are typical and expected features of maar-diatreme volcanism through "soft substrate" such as the sand and silt succession of the pre-volcanic-rock units of the Pali Aike Volcanic Field.

influenced and controlled the position and emplacement behavior of rising magma in the uppermost crust (Büchel, 1993; Lorenz, 1984; Lorenz and Büchel, 1980; Schmincke, 1977). Here the maar-diatreme volcanoes were formed at the intersections of mostly basaltic dikes, especially with local water-bearing faults or joints, below fault-controlled valley floors (Lorenz, 1984). Where meteoric water was not available, scoria cones were formed (Lorenz, 1984).

Peperite results from the interaction between magma and wet sediment and exhibits a range of complex textures (Skilling et al., 2002; White and Houghton, 2006; White et al., 2000). 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 traveled through wet sediments (Skilling et al., 2002). Peperite and their associated pyroclastic environment are common in small-volume phreatomagmatic vent-filling deposits and/or along the contacts between sediment and mafic intrusions (Fig. 12) and lavas, as in more complex stratovolcanoes and/or caldera volcanic systems (Martin and Németh, 2007; Németh and Martin, 2007). In the western Pannonian Basin, most of the lava flows erupted inside 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 water saturated. The identification of peperite in such settings therefore does not guarantee the reconstruction of the eruptive environment or the establishment of the syn-volcanic landscape position onto which the volcano erupted.

Transitions between Dry and Wet Eruptions within Monogenetic Volcanism

Maar-diatreme volcanoes are created by phreatomagmatic eruptions arising from a contact of ascending magma with groundwater in available aquifers. If this contact is eliminated, magma continues its ascent toward the surface, giving rise to other volcanic forms and products. Usually this occurs in the advanced or closing stages in the evolution of the volcano. However, there are examples when the maar-forming eruptions were preceded by effusive and/or Strombolian activity (Gutmann, 2002). Phreatomagmatism is encouraged if the magma flux is low relative to the rate of water supply and if the top of the magma column has subsided below the water table. A general evolutionary sequence from hydromagmatic eruptions during formation of the maars, through Strombolian eruptions, and finally to lavas reflects a declining influence of magma-groundwater interactions with time, documented in many volcanic fields such as Hopi Buttes, Arizona (White, 1991b). Transient hydromagmatic events occurred relatively late in the 1975 eruption sequence of the Tolbachik Volcano in Kamchatka, Russia (Doubik and Hill, 1999). The general time sequence of eruption events (from wet to dry explosive processes) and the distribution of the volcanic vent type (dry and wet) with regard to the geomorphology of the pre-volcanic landscape is well established,

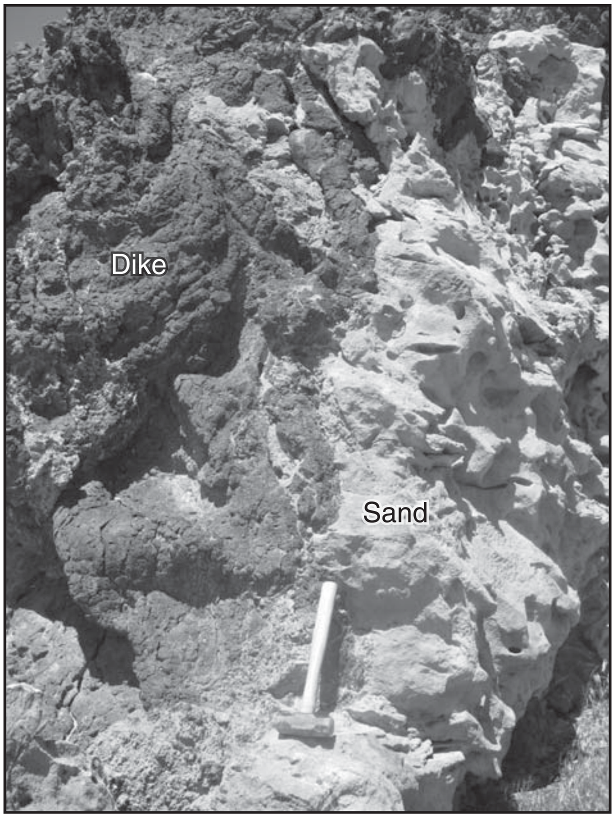

Figure 12. Globular peperite along a feeder dike in contact with the host siliciclastic sediments of a small phreatomagmatic volcano (71 Gulch volcano) in Idaho.

having phreatomagmatic volcanoes in low-lying areas and scoria cones in highlands (White, 1991a). Intermittent phreatomagmatic activity from sudden groundwater access to the volcanic conduit in the waning stage of an eruption has also been described in large ocean islands such as at Kilauea in Hawaii (Dzurisin et al., 1995). The most widespread of these deposits at Kilauea (Uwekahuna Ash Member) is a basaltic surge and fall deposit a few meters thick (Dzurisin et al., 1995).

Simultaneous magmatic and phreatomagmatic explosive events in the same volcano have been recognized from tuff ring sequences. For instance, the Crater Hill tuff ring in New Zealand contains unit(s) that can only be interpreted as the products of the mixing of ejecta from simultaneous wet and dry explosions at different parts of a multiple vent system (Houghton et al., 1999). Other types of mixed deposits from the Eifel are interpreted to represent mixing from two point sources (e.g., vents) of quite different but stable character (Houghton and Schmincke, 1986, 1989).

GEOMORPHIC ELEMENTS AND THE SEDIMENTARY RECORD OF VOLCANIC FIELDS

Monogenetic volcanoes can form in any type of geological environment; their volcanic landforms, however, strongly depend on water availability and the substrate geology. In fully subaqueous environments, either in sublacustrine or submarine settings, lens-shaped pyroclastic mounds form (White, 1996). These mounds consist of flat-lying pyroclastic density-current deposits (White, 2000). In a shallow subaqueous environment after the construction of a pyroclastic mound, eruption clouds and directed pyroclast jets can breach the water surface and form capping steep-flanked tephra cones (Belousov and Belousova, 2001; White, 2001), such as the Pahvant Butte in Utah (White, 1996, 2001). In fully subaerial settings, when magma interacts 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 tephras (Vespermann and Schmincke, 2000). When magma interacts with groundwater, maar volcano forms are surrounded by flat-lying tephra beds (Vespermann and Schmincke, 2000). Maar-diatreme volcanoes are commonly the only sites where already eroded pre-volcanic sediments may have been preserved; thus their study may provide information for understanding the stratigraphy. Phreatomagmatic explosive phases occur in many small-volume volcanoes, especially those in valley settings with good water availability. Subsequent magmatic explosive eruptions commonly occur during the waning phase of the eruption history (Aranda-Gomez et al., 1992; Martin and Németh, 2004; Woerner et al., 1988). This phase commonly produces large scoria cones, accompanied by lava effusion inside the crater such as at the La Brena maar in Durango, Mexico (Aranda-Gomez et al., 1992) (Fig. 13). Phreatomagmatic volcanic fields commonly occur within fluviolacustrine basins; therefore the recognition of a great diversity of phreatomagmatic volcanoes in ancient settings almost certainly represents basinlike and/or valley settings. Such an interpretation may alter the extrapolation of the estimated landscape erosional data (Németh et al., 2007a).

The sedimentary record associated with monogenetic volcanism can be divided into two major groups of volcanic sedimentary facies: (1) cone- or edifice-building units, and (2) distal, interedifice successions. The first volcanic group characteristically represents the deposits generated by primary explosive-eruption–related processes. These predominantly pyroclastic units that form the volcanic edifices are interbedded, with beds representing reworking in the course of the eruption or sudden changes of eruption styles (Fig. 14). The second group of volcanic successions is characterized predominantly by distal tephra units (e.g., tephra fall blankets) and fluvial and eolian sediment (White, 1991a). The facies architecture interedifice areas are poorly studied. The general perception, however, is that the preservation potential of volcanic material in the sedimentary record in a volcanic-pitted basin is low (Ufnar et al., 1995), and therefore sites such as maar lakes are especially important for preservation of distal tephra (Guo et al., 2005; Muhs et al., 2003; Sabel et al., 2005; Wang et al., 2007).

The tephra ring successions of a phreatomagmatic volcano are commonly interpreted to be deposited predominantly from pyroclastic density currents, such as base surges, as well as

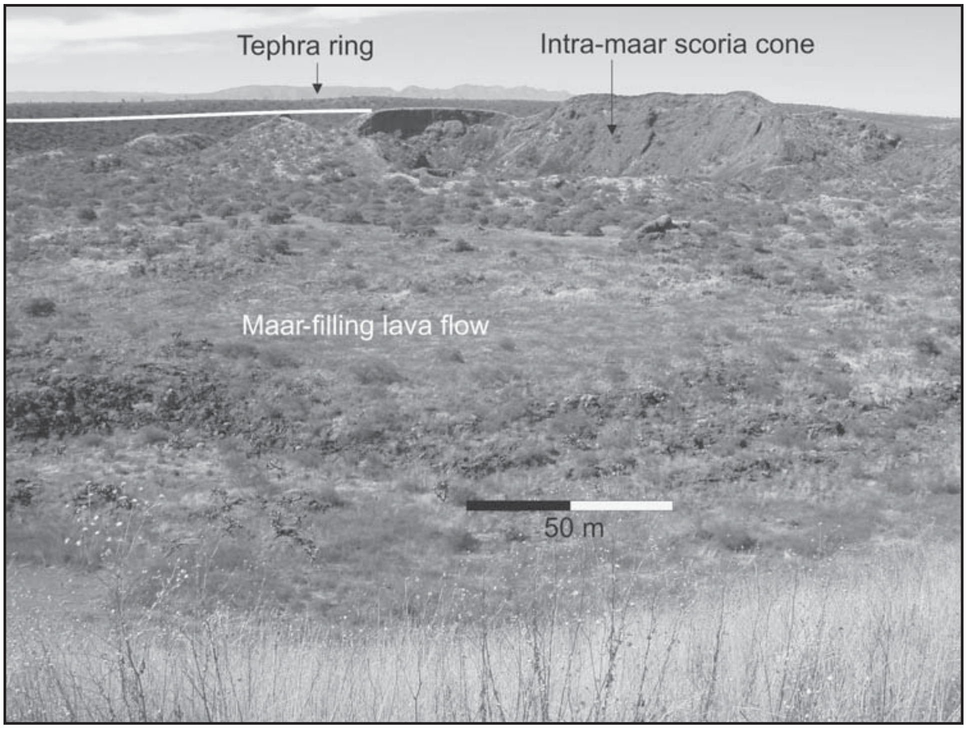

Figure 13. La Brena maar in Durango, Mexico, filled with a lava field and a lava-spatter-dominated scoria cone.

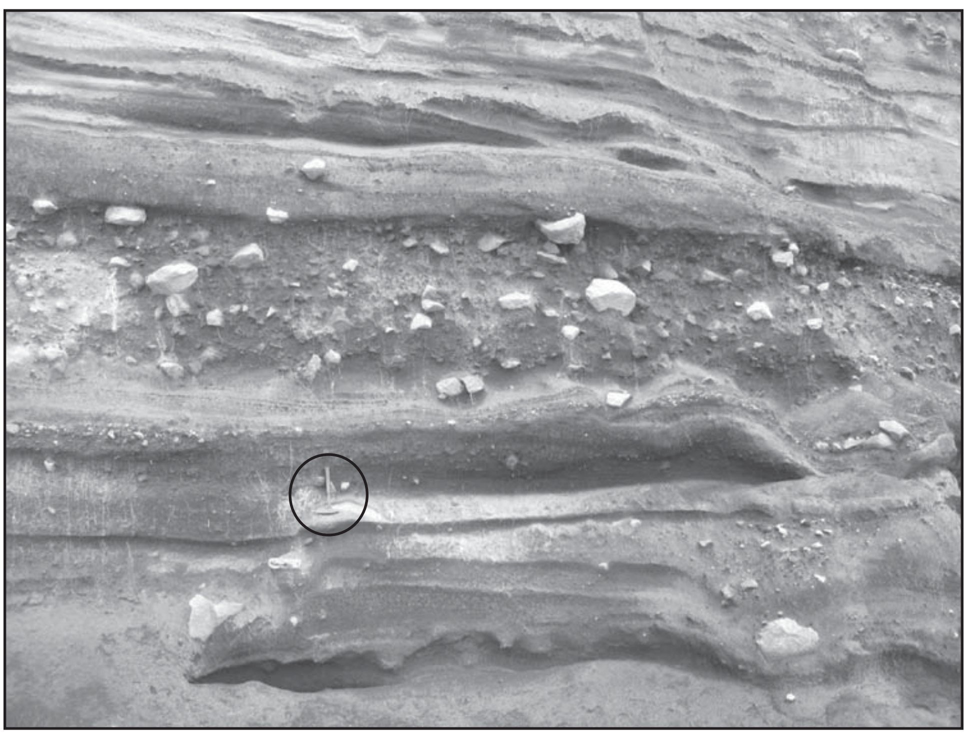

Figure 14. Coarse-grained tuff-breccia zone in a phreatomagmatic primary pyroclastic succession of a maar formed in 1913 in West-Ambrym, Vanuatu. The bed is inferred to represent an event zone in the course of the eruption associated with some syn-eruptive remobilization of debris on the flank of the growing tephra ring. Hammer circled for scale (~30 cm).

input of fall material into the passing base surge currents (Dellino, 2000; Dellino et al., 1990, 2004a, 2004b; Dellino and La Volpe, 2000). Interbedded scoriaceous fall deposits are common in tephra ring successions (Fig. 15) and are interpreted to result from an intermittent change in eruptive style, where water availability to fuel magma-water interaction was suppressed (Houghton et al., 1999). In distal sections, base-surge-dominated beds are gradually replaced by reworked decimeter-thick beds of fine ash (Fig. 16) deposited from syn-volcanic debris flows and/or hyperconcentrated mass flows (Lajoie et al., 1992; Sohn, 1996; Vazquez and Ort, 2006). Facies variation in tuff ring sequences has been used to define characteristic changes in energy regimes. From the Crater Elegante in Mexico, proximal sand-wave bed facies are replaced by massive facies in the medial regions, and by planar bed facies in the most distal regions (Wohletz and Sheridan, 1979). A facies variation of a single surge unit from the Hopi Butte showed a different trend, having disorganized beds in the rim section, stratified beds and sand-wave beds in the medial sections, and plane-parallel beds in distal regions (Vazquez and Ort, 2006).

When the phreatomagmatic explosions of a maar-diatreme volcano finally come to an end, the crater typically fills with water. The resulting maar lakes are deep in relation to their diameter and are closed depressions cut off from the surrounding landscape by a crater rim. This special architecture affects the lake and its sediments, as they drain material from an extremely small catchment area (Kotthoff and Schmid, 2005; Mingram et al., 2004a, 2004b; Sabel et al., 2005; Scharf et al., 2001; Wilde and Frankenhauser, 1998; Zolitschka et al., 2000). Allochthonous clastic material reaches the crater mainly as turbidity 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 (Belis et al., 1999; Brukner-Wein et al., 2000; Mingram, 1998; Schabetsberger et al., 2004). Maar lakes often develop a meromictic division of their water column, providing in this way an exceptional condition for the preservation of sedimentary structures as well as fossils. Laminites within maar sediments commonly are strongly characteristic of the paleoenvironment and as such are highly powerful reconstructive tools (Crausbay et al., 2006; Garcin et al., 2006; Vazquez et al., 2004).

Large tuff rings and maars can host a significant thickness of lacustrine sediments accumulated over long periods (thousands of years) in their craters. Such lacustrine sediments are initially saturated, and over time they compact and subside (Suhr et al., 2006), resulting in convolute and distorted bedding, water escape structures, or unusual blocks atop one another (Németh et al., 2008). After erosion and exhumation the position of a lacustrine succession may be used to estimate degrees of erosion, once compaction of the crater-filling succession is accounted for.

If rejuvenation of volcanic activity occurs within a water-filled maar, eruptions may take place purely in subaqueous to emergent conditions (Németh et al., 2007a), forming complex sedimentary facies similar to those forming in normal subaqueous sedimentary basins. To distinguish between sedimentary

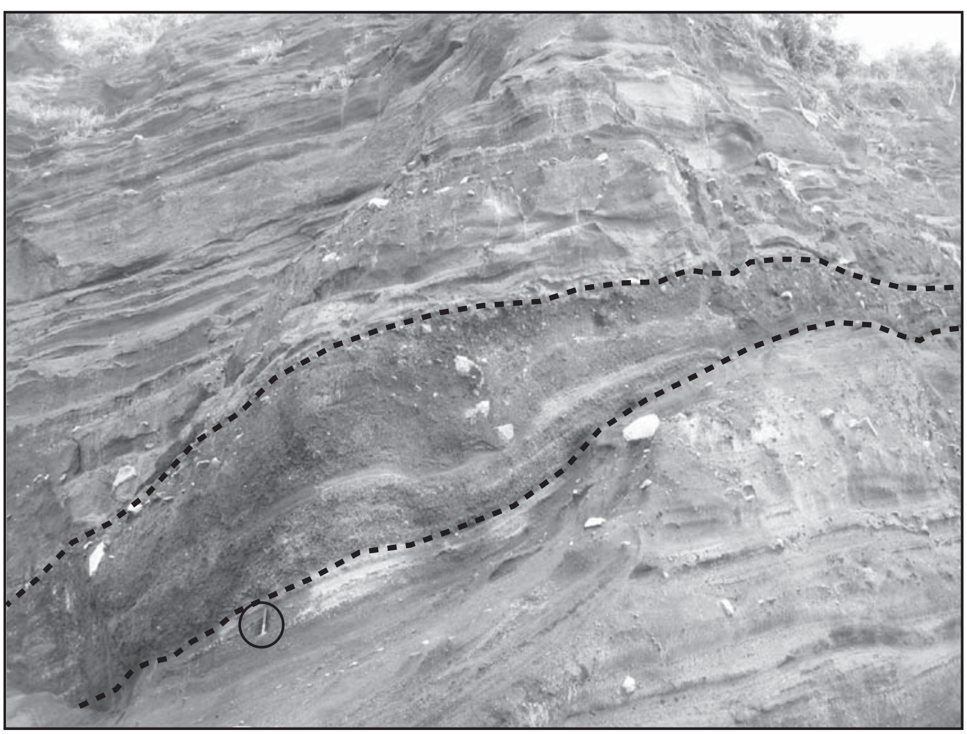

Figure 15. Interbedded scoria beds in a phreatomagmatic succession (between dotted lines) are common in phreatomagmatic volcanoes such as the tuff rings of the 1913 eruption of West-Ambrym, Vanuatu. Such beds can represent sudden changes in the eruption style (phreatomagmatic to magmatic), or simultaneous eruptions from nearby vent(s) that are commonly part of the same growing “monogenetic” volcano. Hammer circled for scale (~30 cm).

Figure 16. Tabular beds of debris flow and hyperconcentrated mudflows in the distal wedge of the 1913 phreatomagmatic volcanoes of West-Ambrym, Vanuatu.

environments related to basinal settings from those created by volcanism is important in reconstructing the syn-volcanic landscape evolution of the region.

INTEGRATED SOURCE-TO-SURFACE MODEL

Volcanic fields are generally very complex volcanic systems that can be active over millions of years. Distribution patterns of individual vents (e.g., scattered, clustering, or aligned), the chemical changes of eruptive products over time, the associated intrusive processes of the volcanism (e.g., dike swarms, fissure eruptions) and their relationships with central vents are features that bear information about the magmatic feeding systems of the volcanic field. The existence of volcanic fields in association with or without large-volume composite volcanoes, their potential development in several tectonic settings, and their common total magmatic output similar to those calculated for composite (e.g., stratovolcanoes) volcanoes are evidence to accept that volcanic fields result from an interconnected channel network that pumps magmas to the surface over long periods without significant pause en route to the surface. The potential to repeatedly inject new magmas more or less in the same pathway to the surface is not impossible, but it requires substantial melt availability, and special local tectonic conditions. With increasing field data, some edifices of a volcanic field that were viewed as monogenetic can now be seen as complex volcanoes of small size. Because the total eruptive duration of a volcanic field is long (millions of years), it is inferred that tectonic stress regime changes can take place and lead to even the pausing of melts in the crust to initiate the "birth" of composite volcanoes (with evolved chemical signatures). In other cases, "failed" composite volcanic complexes can form, made up of closely spaced vents (Fig. 17). Volcanoes such as Los Loros in Argentina have a complex eruptive history but a small volume (e.g., evolved chemical composition and development of a sustained Plinian eruption column to feed pumiceous pyroclastic flows).

Volcanic fields are in many places heavily populated such as the Auckland Volcanic Field in New Zealand, which is also the largest city of that nation. Volcanic hazard studies of a volcanic field commonly target an understanding of the recurrence rate, the eruption frequency, and the style of possible future eruptions (Connor and Hill, 1995; Connor et al., 2000; Cronin et al., 2001; Cronin and Neall, 2001; Edbrooke et al., 2003; Ho, 1992; Ho and Smith, 1998; Hurst and Smith, 2004; Magill and Blong, 2005a, 2005b; Magill et al., 2006). In this perspective the volcanic system of a volcanic field involves the magmatic source, the magmatic plumbing, the tectonic stress field, and the eruptive surface environment. All of these aspects need to be studied and understood in order to develop realistic volcanic hazard assessments of a volcanic field.

CONCLUSIONS AND OUTLOOK

Volcanic field studies should be concentrated on the overall understanding of the magmatic plumbing system, tectonic stress field variations over time, and sedimentary basin evolution.

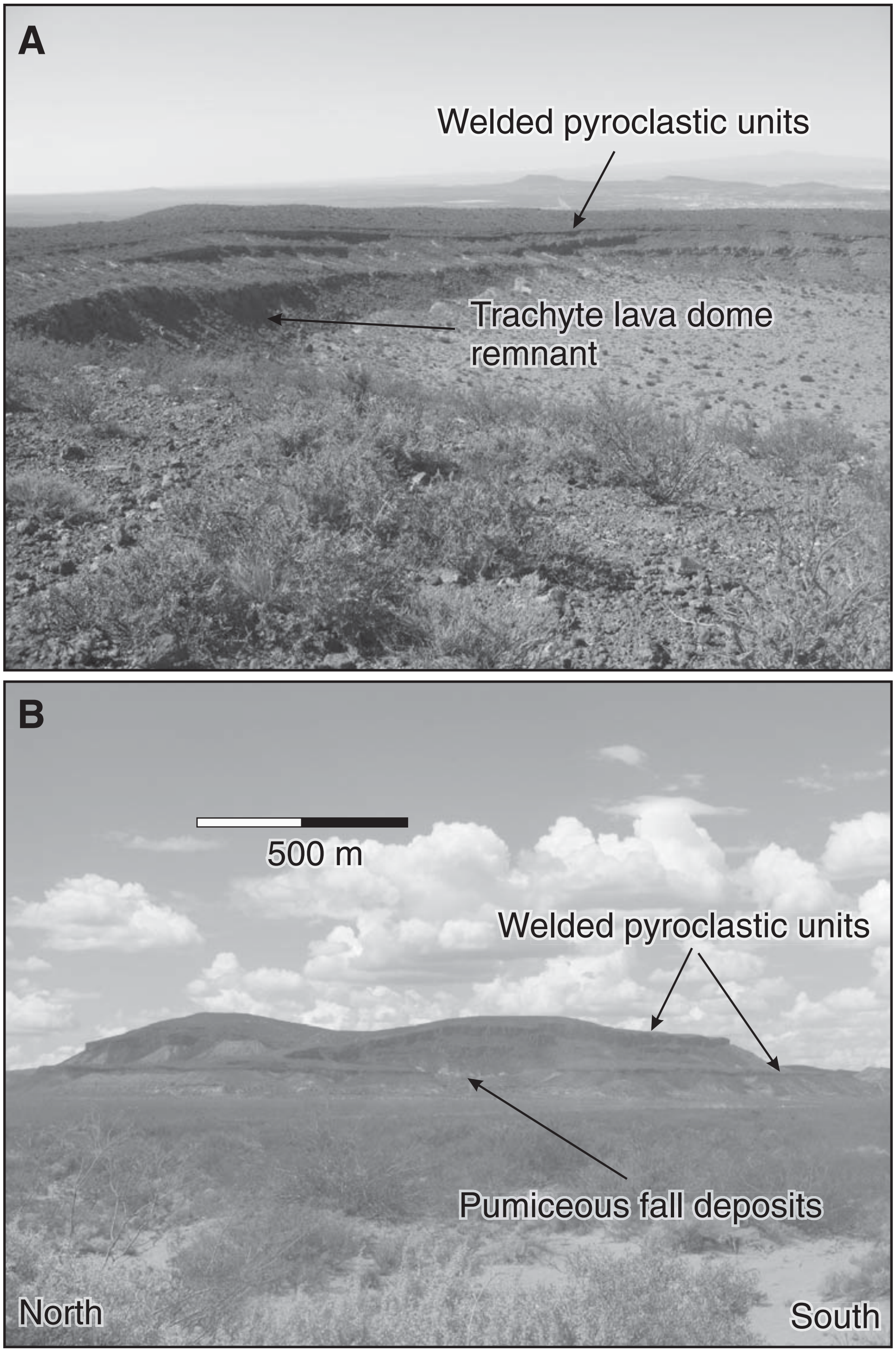

Figure 17. Los Loros (Argentina) is a small-volume volcano with a wide, well-preserved crater (A) in part of a volcanic field. Its composition is trachytic, and its eruptive history documents Plinian phases of accumulated pumice fall deposits (B), erosional phases leading to volcanic gravel-filled channels, and highly explosive phases leading to small-volume welded pyroclastic flow deposits (A and B). This complex nature of the otherwise 2-km-wide volcano (~1 km diameter of its crater) highlights the difficulty in classifying volcanoes as monogenetic based purely on their small volume, morphology, or composition.

This holistic view is necessary for volcanic hazard assessment studies of volcanic fields. As discussed in this paper, the question of origin of monogenetic versus polygenetic volcanism is still unsolved. Mosaic-like information exists about key potential controlling parameters of volcanism, including the heterogeneous versus homogeneous nature of the melt source, the role of the melting rate in the formation of volcanoes, and the role of melt migration through complex channel networks. The exact calculation of total magma flux in volcanic fields also calls for a better understanding of the magma migration to the surface. More detailed studies of volume- versus magmatic-controlled models for volcanism in volcanic fields need to be undertaken with new approaches that employ statistical analyses of the time and space distribution of volcanic events in a volcanic field.

From a volcanic-hazard-management point of view, special care is necessary for understanding those volcanoes erupted through wet processes. Such phreatomagmatic volcanoes are commonly significant in their volume, although the magma needed for their formation in most cases is very small. Phreatomagmatic volcanoes may contribute only a quarter of the total number in a volcanic field, but they are commonly the most hazardous. The number of such volcanoes in a volcanic field is larger in low-lying areas that are commonly the most populated. The role of the physical conditions of the sediment substrate cut through by volcanoes includes their water-saturation level, lithification, and strength, as well as the variation of these parameters beneath a volcanic field. All of these factors need more systematic study to understand the role of shallow-level conditions in the formation of specific volcano types in a volcanic field.

Applying these new ideas in volcanic field studies would put this type of volcanism in a "source-to-volcano" model and would make volcanic field studies as important as other studies centered on polygenetic volcanoes.

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From a definition of volcano *to conceptual volcanology*

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ABSTRACT

Volcanology textbooks either use the classical "opening-in-the-ground" definition of a volcano or simply avoid a definition. A possible systemic redefinition of the term *volcano* is considered in this paper. Starting from the classical Aristotelian requirements of a definition, it is shown that only a definition that is part of a hierarchically organized system of definitions can be accepted. Thus, conceptual constructs should reflect the same type of makeup as nature's processes, which are hierarchically organized. Such a line of reasoning implies that a volcano should be defined by making an explicit mention of the hierarchy of systems to which it belongs. Therefore, volcano can be defined as either a subsystem (i.e., the eruptive subsystem) of the broader igneous system or as a particular type of igneous system (i.e., one reaching the surface of Earth). A volcano, viewed as a volcanic system, is composed of a magma-generation subsystem, a magma transport subsystem, magma storage subsystem(s), and an eruptive subsystem. The accurate definition and identification of each subsystem should allow distinction between individual volcanoes in both space and time. Minimal conventional requirements need to be agreed upon by volcanologists to identify and recognize a particular volcano from other volcanoes (including those partially occupying the same space but separated in time, or those partially overlapping in both space and time). An accurate definition of a volcano using the systemic approach involves definitions of other basic terms and concepts of volcanology in a similar way, eventually resulting in a hierarchical system of definitions that would lend volcanology a solid, consistent, and coherent conceptual core, increasing its scientific maturity. Conceptual volcanology can be envisaged as addressing the issue of accurate definition of basic terms and concepts, besides nomenclature and systematics, aiming at reaching the conceptualization level of more basic sciences.

INTRODUCTION

Definitions are boring but necessary.
—G.A. Macdonalds (*Volcanoes*, 1971)

The theoretical edifices of scientific disciplines are built up around a generally small number of basic concepts. These fundamental entities constitute what one may call the conceptual "hard core" of that discipline. For instance, basic concepts in physics include precise definitions of concepts like *matter, energy, wave,* and *field.* Other concepts, such as *atom, electron, boson, meson,* etc., are shared by physics and other natural disciplines, like chemistry or particle physics. Biology also has a set of basic concepts—*cell, organ, organism, species, population,* and *evolution,* just to name a few—and in this sense Earth sciences are not different from other scientific fields. What is common to all of these disciplines is that the basic concepts (1) tend to be accurately defined, (2) are specific to that discipline, and (3) constitute a

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coherent assemblage of concepts that facilitates the progress of that particular discipline.

But what about volcanology? It is obvious that volcanology as a scientific discipline should have its own conceptual core, which should include basic concepts and terms such as *volcanism, volcano,* and *(volcanic) eruption.* Close inspection of these terms, however, reveals that these concepts are not well defined and that it is difficult to consider them assembled within a coherent and logical system.

In this paper I attempt to demonstrate that current formulations of the most basic concept of this discipline (i.e., a volcano) do not fulfill the requirements of a sound definition, and are internally inconsistent. To avoid these problems I suggest a possible way to redefine the term *volcano* on systemic grounds. This approach to some extent leads to the need for the creation of a new branch of volcanology, called here *conceptual volcanology.* This subdiscipline of volcanology is required if we want this discipline to obtain a truly universal language. Evidently, this does not imply that advances in volcanology cannot be made in the absence of such a subdiscipline, but rather that those advances can be reached much more efficiently. To some extent, this would be analogous to the effect that "strong inference" has proved to have in other fields of science (Platt, 1964). Adoption of this strategy is therefore not mandatory, but it might be decisive to catapult volcanology beyond its current status.

ABOUT "DEFINITION"

Before attempting any definition of the term *volcano,* it is convenient to examine the requirements of a sound, accurate definition. To be generally accepted by the scientific community, scientific definitions must obey some logical requirements, most of which were stated long ago by philosophers. In particular, Aristotle offered a system of definitions that has to be tightly related to systematics and classification. Biology has used very successfully such an approach for a long time. Starting from Linné, a hierarchical system of definitions became implicitly embedded in the biological sciences: any species belongs to a *genus,* which belongs to a *family,* which belongs to an *order,* etc. In geology, the Strunz classification of minerals (mineral species, order, subclass, class) provides another example of a hierarchically organized system of definitions. Unfortunately, the construction of such a hierarchical system has not been adopted by other geological disciplines, as for example, volcanology. Consequently, from this point of view it is fair to say that concepts such as *volcano* remain ill defined until the present. The last statement does not imply the complete absence of a "definition" of such term, because there are various forms in which a volcano has been "defined," as discussed below. Nonetheless, all of those "definitions" have a certain embedded ambiguity. Such characteristics could be avoided if the Aristotelian approach is adopted.

The Aristotelian concept of definition is unfolded and discussed in detail in the Appendix. A reader familiar with the concepts of Aristotelian philosophy might skip that section; those willing to expand their philosophical knowledge of such fundamental aspects of any natural science might find it interesting reading. In any case, to follow the arguments discussed in the main sections of this paper it suffices to remember that according to Aristotle the most basic requirements of a definition include: (1) the denomination of a *genus proximus* (the closest class) to which the object to be defined belongs, and (2) a list of *differentia specifica* (i.e., specific differences) by which that object differs from other objects belonging to the same class (Ross, 1927).

CURRENT DEFINITIONS OF *VOLCANO*

There is no doubt that *volcano* is the very pivotal concept of volcanology around which all theoretical and practical approaches of this discipline are built up. Despite this, the concept of *volcano* is poorly defined, if it is defined at all. Notably, important works on the field have preferred to avoid any definition of the term *volcano* (e.g., the *Encyclopedia of Volcanoes,* Sigurdsson et al., eds., 2000).

Certainly there is a classical and currently accepted definition of a volcano found in many volcanology textbooks. This definition represents the "mainstream" idea of what a volcano is, and it will be referred to in this paper as the "current definition." In general, volcanology scholars and researchers are—implicitly or explicitly—in agreement with the current definition. In a companion article Borgia et al. (2010, this volume) review the relevant literature, including Internet sources. The picture that emerges from that review is disheartening. Almost all of the available definitions of a volcano are in fact variations of the current definition, according to which a volcano is "a place/site, vent, or opening, on Earth's surface through/at which magma reaches the surface." A typical example of this classical volcano definition is offered by Francis (1993, p. 2): "A 'volcano' … is simply a site at which material reaches the surface of the planet from the interior." This statement is frequently complemented with another sentence: "Volcanoes are also a commonly conical mound or mountain built up by the eruption products around that vent." Sometimes a volcano is defined by making reference to only this latter meaning (e.g., Fisher et al., 1997, p. 43), therefore considering volcanoes basically as landforms.

Criticism of the Current Definition of *Volcano*

A number of shortcomings of the current definition of *volcano* can be listed. In an early attempt to challenge this generally accepted concept of a volcano, it was stated that the current definition is inconsistent with the requirements of a scientific definition, furthermore being inaccurate and confusing (Szakács, 1993). In this section I elaborate more on this topic.

First, it is noted that the current definition has an inherent duality: a volcano is a "vent" *and* simultaneously it is a "landform." Such duality cannot be sustained on logical grounds, because an object cannot be two mutually exclusive things at one time. Specifically, this dual definition claims that *volcano* is simultaneously something characterized by an absence of matter (a hole: opening or vent) and something characterized by a prominent presence of matter (a mound or mountain). Furthermore, the part of the definition denoting a volcano as a positive landform excludes caldera volcanoes (which are essentially negative landforms) as part of the *genus proximus,* and the definition as such, is therefore a false statement.

Second, the current definition reduces what we call a volcano to its surface expression, ignoring all its components located underground. However, it is now obvious that volcanic activity implies not only what is happening at the level of Earth's surface, but it includes processes taking place below the surface, down to the magma chamber and involving the whole magma plumbing system. Therefore, the subsurface parts of the eruptive system, its roots, cannot be ignored when defining a volcano. This would be equivalent to omitting the roots from the definition of a *tree.*

Third, by reductio ad absurdum, it is easy to demonstrate that on the basis of the current volcano definition one should accept that in many cases "a volcano is many volcanoes." For instance, in a recent article reporting on the March 2006 eruption of Raoul volcano it is stated that "at least 21 vents were involved in the eruption with most ejecta coming from five of these" (Christenson et al., 2007). That means, if we accept the current definition of *volcano* (volcano as a vent) that on 17 March 2006 at least 21 distinct volcanoes erupted on Raoul Island. However, volcanologists consider the Raoul caldera as part of a single, unique volcano (e.g., Simkin and Siebert, 1994). Similarly, multivent volcanoes, such as Fuji, Etna, Kilauea, etc., should be considered assemblages of various volcanoes instead of being single volcanoes, as they are currently recorded in the relevant volcanological literature (Simkin and Siebert, 1994).

Fourth, the confusing character of the current definition is strongly apparent when considering two closely spaced cinder cones with their related lava fields in a field of monogenetic volcanoes such as, for instance, in the Michoacán-Guanajuato area in western Mexico, as compared with two cinder cones with their related lava fields of similar size and similar spacing on the flanks of either a large composite volcano (such as Fuji), or a shield volcano (such as Kilauea) (Fig. 1). In the former case all agree that there are two distinct volcanoes (e.g., Jorullo and Parícutin) and are listed as such in volcano records. In contrast, in the latter case those cinder cones are thought to be parts—lateral or satellite vents—of a large volcano rather than being considered individual volcanoes. Consequently, two cones with similar characteristics as the cones in the Michoacán-Guanajuato region are not listed as separate entries in volcano records. A tentative solution to avoid the contradictions illustrated by these examples is to consider volcanic landforms as composed of small edifices overlying large ones (i.e., a fractal approach). Although this fractal perspective may be interesting to explore in more detail, and might contribute to eliminate the problem described above of one volcano being many volcanoes at a time, it fails to explain why the same volcanic feature (i.e., a cinder cone) is treated as one individual volcano in some cases, and just as part of an individual volcano, even if fractal in nature, in other cases.

All these examples lead us to the conclusion that "opening" or "vent" cannot be accepted as *genus proximus* for "volcano" because, according to the Aristotelian approach, objects belonging to the same class have to be closely related to each other, sharing most of their essential features ("attributes"). By following blindly such logic we would be led to think that if a volcano

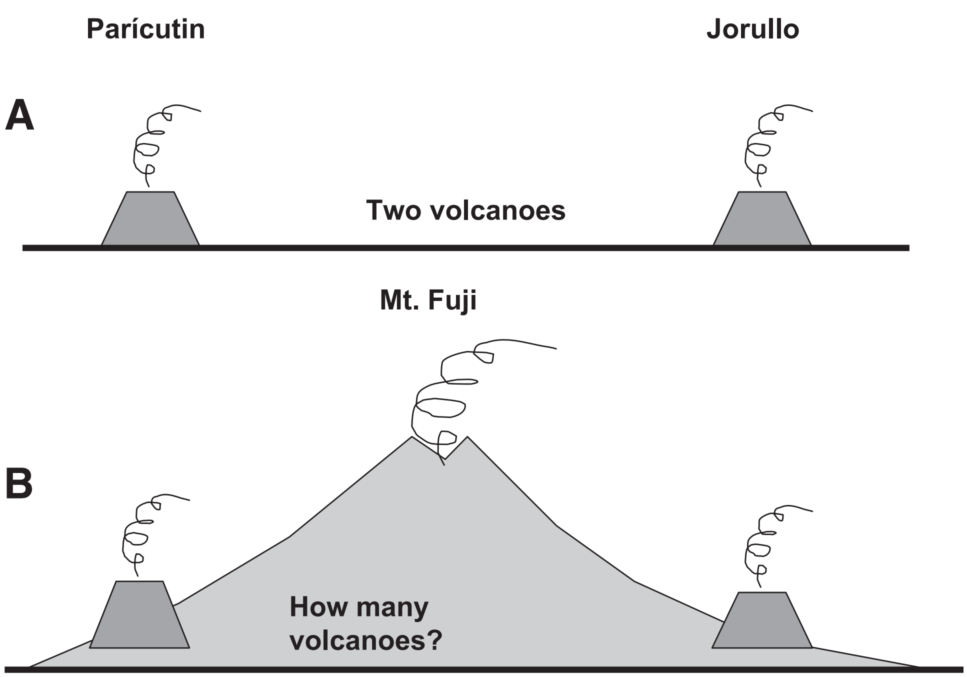

Figure 1. Illustration of the type of conceptual problems posed by the current definition of *volcano.* (A) Two cinder cones in a field of monogenetic volcanoes are unanimously viewed as two distinct volcanoes, having their own names and being recorded as such in volcano catalogues. (B) Two cinder cones of the same size and at the same distance from each other are viewed as satellite vents if they are located on the flanks of a large polygenetic volcano. In this case they are generally unnamed and are not recorded as individual volcanoes in volcano catalogues.

is a "hole in the ground" through which some material passes from Earth's interior to the surface, one should accept any kind of holes in the ground, such as golf holes or worm holes, as being included in the same class of things as a volcano. In addition, the current definition cannot be included in a hierarchically organized system. From the point of view of logic, this problem has more important consequences than the previous one because it prevents the proper advance of the discipline. This issue is therefore discussed more extensively below.

Alternative Definitions of *Volcano*

A few volcanologists and textbook authors, aware of the shortcomings of the current definition of *volcano,* have attempted to provide alternative definitions. However, those attempts have been sporadic and are recorded invariably in "out-of-mainstream" publications or textbooks. Commonly, these alternative definitions differ from the current definition, because instead of focusing on the surface expression of the object (i.e., the vent or topographic aspect of a volcano) they aim at capturing the essential features of volcanoes. Most of such sporadic attempts have portrayed volcanoes basically as thermal anomalies relative to their surroundings, and therefore being visualized as kinds of hotspots at Earth's surface. For instance, in his volcanology textbook, *Volcanoes Today and in the Geological Past,* Rădulescu (1976) states that "being scientifically more precise but much less suggestive [than the current definition] the definition of volcano could be: the place on the surface of the [Earth's] crust, where permanently or rhythmically the temperature is much more elevated than in neighboring points…. The definition is satisfactory excluding geothermal areas in which a) the temperature is not much more elevated as compared to neighboring areas and b) much higher temperatures are located exclusively at depth" (p. 18). Although this "thermal" definition of *volcano* avoids some of the logical flaws of the current definition, it is still unsatisfactory from the point of view of the Aristotelian approach.

An entirely novel approach to defining *volcano* was undertaken in a companion article of this volume (Borgia et al., 2010). According to these authors, there is no terrestrial physical limit to what one may call a volcano. As a consequence, volcano is defined by Borgia et al. as *a particular type of terrestrial environment.* In extremis, adopting such a definition would imply that the Earth itself can be considered as a giant global volcano. Although interesting, such a definition leads to some conflicts with the restrictions imposed by the Aristotelian concept of a "definition" mainly because a possible systemic approach is precluded. The limitations of this definition reside in the need for any system to have physical limits. If volcanoes are thought of as having no limits at the scale of the whole Earth, as implied in Borgia et al.'s definition, then volcanoes cannot be defined as part of a system. A further problem arising from the adoption of an integrationist approach in which the object of study fades within a general indistinct background loosely referred to as "environment" is that such an absence of physical limits can be deceptive and even frustrating. For example, arguing that the dispersion and long-range presence of volcanic products (i.e., fine ash) in the atmosphere means that a volcano is physically extended indefinitely in the atmosphere is equivalent to saying that a human body has no limits because its output "products," such as expired volatiles, are dispersed in the atmosphere around it. Thus, one could "define" humans analogically as "human environments," which is a poor definition of the nature of any human being. More specifically, "environment" cannot be accepted, from the standpoint of the philosophical requirements of a scientific definition, as the closest higher-ranked category (or *genus proximus*) of things to which "volcano" belongs. For these reasons, a limitless volcano concept as such proposed by Borgia et al. (2010, this volume) does not comply with the requirements of an Aristotelian definition.

No Definition as an Alternative Approach

Often it is found that in volcanology textbooks or in science divulgation books there is no definition of a volcano at all. The authors of such works apparently rely on the assumption that the reader already knows what a volcano is, therefore obviating the need to define it. Alternatively, these authors may have been well aware of the difficulties in accepting the current definition, having chosen simply to avoid any definition to circumvent those problems.

In any case, the "no definition" solution is unacceptable, because a pivotal term at the very core of a scientific discipline cannot be in use without a sound definition. All mature sciences have their basic terms well defined, and volcanology should not be different. For this reason, instead of avoiding the problems raised by the current and the available alternative definitions of *volcano* I address this issue by adopting a systemic approach in the following section.

A POSSIBLE SYSTEMIC APPROACH TO THE DEFINITION OF *VOLCANO*

Premises for a Systemic Definition

The Aristotelian requirements of a definition call for an assemblage of hierarchically organized system in which each entity is defined in relation to a hierarchically higher-ranked entity and, in turn, it is a *genus proximus* for hierarchically lower-ranked entities. In other words, in a systemic approach any object is a component (subsystem) of a larger system and, in turn, is a system itself composed of other component parts (subsystems).

Actually, conceiving all objects and processes involving magma as parts of a hierarchically organized Earth system is not an outstandingly exotic approach. For instance, the term *igneous system* is frequently used in petrology textbooks, and the usage of volcanic system is not unprecedented. In particular, G.P.L. Walker (1993, 2000) was a pioneer of the concept of the volcanic system in volcanology. His volcanic system model, and that of Gudmundsson (1995), were included in a more general version inserted in the plate tectonic paradigm (Cañón-Tapia and Walker, 2004). These approaches, however, did not pass beyond the physical-theoretical application of the volcanic system concept toward a more abstract conceptualization level such as that leading to a systemic definition of the term *volcano* itself. This is better appreciated by examining the following quote taken from the work by Walker (2000): "When describing volcanoes it is of great merit to recognize the existence of a larger entity, namely, the volcanic (or magmatic) system. A system may embrace the melting anomaly at the magma source, the conduits through which magma rises toward the surface, the magma chambers, intrusions, and geothermal zones, and the volcano itself" (p. 284). Although not explicitly stated as a definition, from this quotation it is obvious that in Walker's approach *volcano* is part of a volcanic (or magmatic) system, and consequently can be considered a subsystem of that larger system. However, it remains elusive what a volcano is, other than one of the subsystems of a larger entity (volcanic-magmatic system). Nevertheless, this is an excellent starting point if one aims to obtain a new *volcano* definition on systemic grounds, because the great merit of Walker, and of his followers, is that he recognized and made evident the hierarchical systemic organization of the magmatic phenomenology to which all volcanic activity and volcanoes can be related.

Two Possible Systemic Definitions

Let us consider a magmatic (or igneous) system reaching Earth's surface and extending from the deep magma source region to the surface vent(s) where magma erupts. Following Walker's scheme, one may denominate its different components as *subsystems,* namely, (1) the magma generation subsystem, including the asthenospheric or lithospheric location(s) of, and process related to, the magma source; (2) the magma transport subsystem, including all magma paths and conduits and processes, from source to surface; (3) magma storage subsystem(s), encompassing all magma chambers and related processes at various levels, including those found within volcanic edifices; and (4) the eruptive subsystem, involving all features and processes related to the surface apparition of magma, from the shallowest feeding magma chamber to vent(s) and crater(s), including what is called the volcanic edifice (Fig. 2A). With this sketch in front, it is easy to attempt a definition of *volcano* that satisfies the Aristotelian approach, because one may acknowledge the case of an igneous system which does not reach the surface and define it as an "intrusive system," whereas, as Walker and his followers did previously, the case of the surface-reaching igneous system becomes a "volcanic system." From this point forward, there are two possible options for defining the term *volcano.*

One possibility is just to identify *volcano* with *volcanic system.* In this view, *volcano* can be defined as a particular type of igneous system, and a formal definition would be: "A volcano is an igneous system that reaches Earth's surface." In this case, it is implicitly assumed that all the subsystems listed above are parts—subsystems—of a volcano. As a consequence, a volcano extends from the deepest roots of the igneous system—magma source region—up to the surface, including all magma paths, conduits, and chambers, and the edifice built up by eruption products and the topographic feature(s) (either prominent or subdued, positive or negative) resulting from volcanic activity (Fig. 2B, option 1).

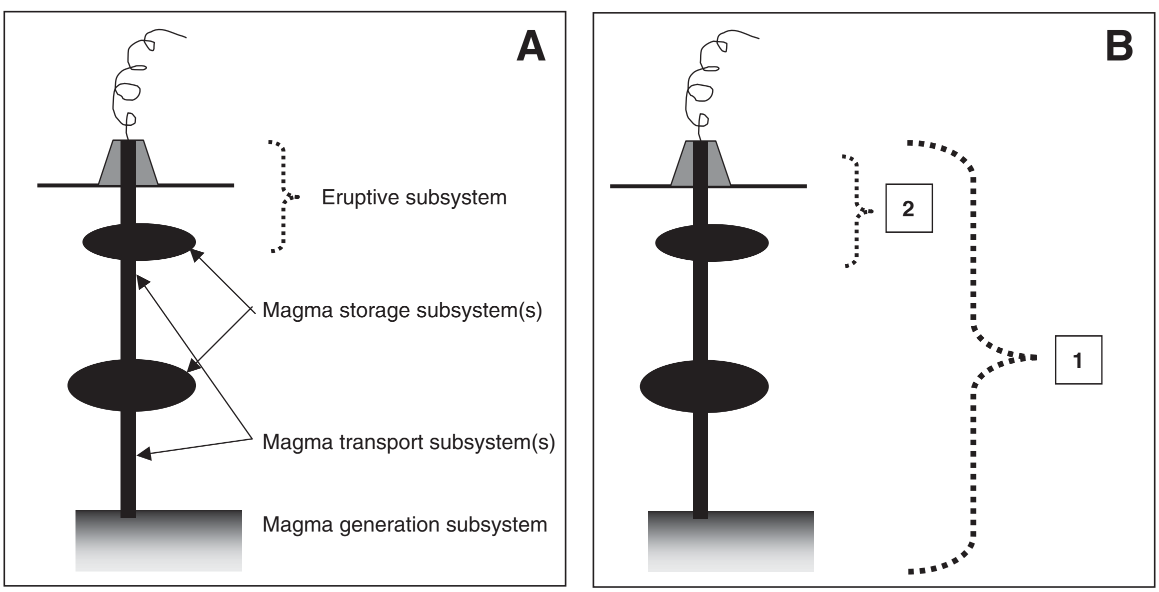

Figure 2. Volcanoes viewed in terms of an igneous (magmatic) system. (A) Subsystems of a surface-reaching magmatic system. (B) Two possible ways to define *volcano* in a systemic approach: (1) *volcano* as a particular type of magmatic system (one that reaches Earth's surface); (2) *volcano* as the eruptive subsystem of a surface-reaching magmatic system.

The second option is to define *volcano* only as the uppermost part of the *volcanic system.* In this case, the formal definition would be: "A volcano is the eruptive subsystem of the volcanic system extending from, and including, the shallowest magma reservoir that feeds eruptive activity, all magma plumbing pathways and conduits tapping that reservoir, and the volcanic edifice at the surface, and also including possible intrusions within such an edifice" (Fig. 2B, option 2). In the particular case where magma chambers are lacking between the magma source and the surface, this definition of *volcano* coincides with *volcanic system* and becomes identical to option 1. Nevertheless, it is noted that these two options are not completely identical.

Note that both of the above definitions of *volcano* satisfy the philosophical requirements of "definition," including the Aristotelian "principle of parsimony" (i.e., to use only the necessary essential words in the definition). In addition, both of these definitions are perfectly compatible with a hierarchical system of definitions; i.e., they are defined in relation to higher-ranked entities, which in turn are (already) defined in a similar way. Also, the definition at hand allows for further definitions of lower-ranked entities, serving as reference terms for them (e.g., *volcanic edifice, stratovolcano, shield volcano, active volcano,* etc.). Such a system of definitions closely reflects the hierarchical organization of natural entities that definitions are meant to designate. Therefore, the two definitions advanced in this section are acceptable and equally valid on philosophical and conceptual grounds.

DISCUSSION

The "volcano-as-volcanic-system" definition is conceptually more rewarding, as it accounts for all interconnected parts of the system and does not introduce artificial boundaries in the volcanic system, discriminating between a surface "eruptive subsystem" named *volcano* in contrast to the deeper part of the volcanic system that lacks a particular name. However, the "volcano-as-eruptive-subsystem" definition seems to be more practical because it may help in understanding and conceptualizing complicated patterns of surface volcanic features, such as compound and multivent volcanic edifices. Furthermore, it allows the introduction of the concept of "unit volcano" as a theoretical simplification in order to reconstruct complex volcanic constructs by a combination of a number of elementary features. The unit volcano could be envisaged as a hypothetical simple volcano (an "elementary volcano"), composed of a unique magma transport path connected at its lower end to a unique magma reservoir or directly to a unique magma source, and at its upper end to a surface vent or a number of vents opened in a unique crater. It also includes all eruption products issued through that/those vent(s) and related volcanic landform(s).

Fields of monogenetic volcanoes are particularly sensible with respect to the issue of *volcano* definition and conceptualization. Individual monogenetic volcanoes in volcanic fields can be readily viewed as "unit volcanoes" irrespective of their possible common magma source. However, according to the systemic approach to the volcano definition in version 2, fields of monogenetic volcanoes can be viewed either as an assemblage of volcanic systems, every monogenetic volcano representing a distinct volcanic system (Fig. 3A), or as a huge unique volcanic system with numerous individual volcanoes (i.e., eruptive subsystems) (Fig. 3B). Actually the two above alternative interpretations of fields of monogenetic volcanoes in the systemic approach can be thought of as end members of a spectrum of possible combinations of "unit volcanoes" and volcanic systems. Complex fields including polygenetic volcanoes among prevailing monogenetic volcanoes, such as the San Francisco Volcanic Field (Arizona, USA) can also be understood, represented, and modeled in terms of a combination of "volcanoes" and "volcanic systems" deriving from the systemic conceptual approach to volcano definition.

Within a sound conceptual framework based on the systemic definition of *volcano,* individual volcanoes should be identified and distinguished from each other in both space and time. That task is sometimes extremely difficult in the case of complex and compound volcanic constructs in which different generations of edifices coexist in the same limited space: i.e., the case of "overlapping volcanoes" (Fig. 4). In the case of Fuji, Japan, for instance, volcano mapping and chronological studies revealed the existence of an old stratovolcanic edifice (Old Fuji), almost completely covered and engulfed by a younger cone (Young Fuji) (e.g., Endo et al., 2003). Moreover, the cone itself and its feet are densely dotted by a large number of cinder cones (i.e., monogenetic "volcanoes"). Conceptually, this situation can be interpreted either as (1) a single volcano with a major discontinuity in its evolution, including numerous "parasitic" vents; or (2) two distinct (in time) but overlapping (in space) stratovolcanoes, each one (or only Young Fuji) with its own related assemblage of "parasitic" cinder cones; or (3) three different volcanoes, two overlapping stratovolcanoes and one field of monogenetic volcanoes; or (4) a large number of volcanoes, two of which represent overlapping stratovolcanoes and the rest a number of individual monogenetic volcanoes occupying collectively roughly the same surface area as the stratovolcanoes. Despite this complexity, Fuji as a whole looks like a simple, beautifully symmetric stratovolcano, and nobody doubts its uniqueness as a volcano. Other compound volcanoes such as Teide (Canary Islands, Spain), Tongariro (New Zealand), and Lascar (Chile) (Davidson and De Silva, 2000), just to name a few, display much more complex topographical features, reflecting an even more complex internal makeup and evolution. In such complex cases the usage of the conceptual approach, including that of the "unit volcano," deriving from the systemic definition, is the only way to unravel volcanic history and structure. Figure 4B displays one possible combination of individual "volcanoes" on the conceptual model of the Fuji case.

Of course, actual information coming from detailed mapping of volcanic areas, combined with geochronological investigation and dedicated geophysical surveys, are needed to select the most suitable conceptual model as best approaching natural reality. In such endeavors a conceptual framework built up around the systemic definition of *volcano* will help greatly when evaluating and

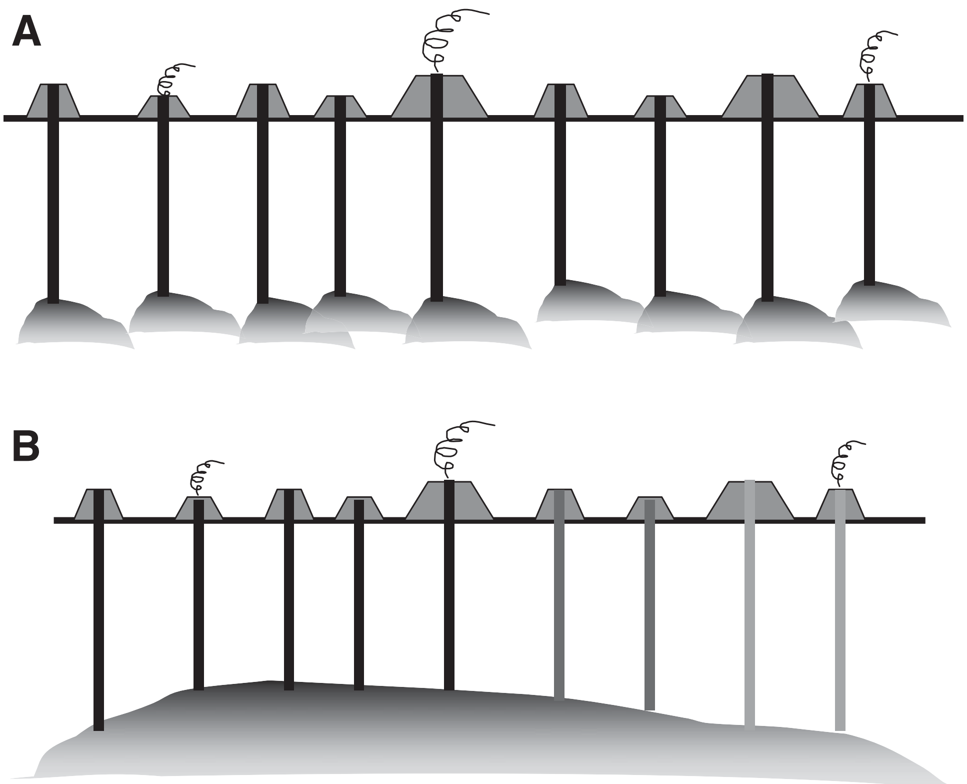

Figure 3. Two end members of a spectrum of possible interpretations of a field of monogenetic volcanoes in the systemic approach: (A) each monogenetic volcano is the surface expression of a simple individual "magmatic system"; (B) all monogenetic volcanoes of the field are parts of a huge, unique and complex magmatic system. Many intermediate combinations of simple and compound systems can be envisaged to explain the "systemic structure" of the field of monogenetic volcanoes.

interpreting field and laboratory data collected either in modern or ancient volcanic areas.

Toward a Conceptual Volcanology

In a mature science or scientific discipline, basic terms and concepts should be defined according to Aristotle's principle of definition. Furthermore, all definitions in that discipline should be organized hierarchically in a coherent system of definitions inseparable from a hierarchically organized systemic approach. If we examine volcanology from this point of view, it is obvious that this discipline does not satisfy such a criterion of scientific maturity. Therefore, as surprising as it might seem, modern volcanology is still not a mature science, and it will not reach such a status unless it develops its own robust and coherent system of definitions and a related hierarchically organized classification system. A different measure of the maturity of a science is the manner in which abstract concepts or scenarios are dealt with, the way in which existing concepts are managed to accommodate new discoveries, and the form in which new names are devised and introduced within the discipline. For instance, introduction of new concepts and terms is a common development as new insight in volcano phenomenology is obtained. Concepts such as "volcano instability," "sector collapse," and "debris avalanche" were unknown 40 years ago. Others such as "volcano spreading" is even a younger concept. They all designate newly discovered phenomena in volcano evolution. More general and abstract concepts are also introduced as tools of understanding complex phenomena. "Pyroclastic density current" is just one example. "Capable volcano" (McBirney et al., 2003) is another. Concepts of even higher abstraction levels could be used in volcanology. The systemic approach proposed in this paper for volcano definition, for example, calls for the introduction of new higher-generalization concepts, such as "unit volcano," inspired from the "unit basin" concept used in hydrogeology as an effective tool for understanding complex flow patterns of subsurface water in real sedimentary basins. The "unit volcano" as a theoretical abstraction designates the smallest and simplest subsystem of the surface-reaching magmatic system, which still can be viewed as a "volcano" (i.e., composed of one single vent, one single conduit rooted in a single magma chamber—or magma generation area—and with a simple surface topographic expression or edifice). It is an "elementary volcano." Complex volcanic systems can be viewed, understood, and modeled as combinations of "unit volcanoes." Theoretical advance in volcanology by introducing such new concepts could be substantial.

In addition, it is noted that besides classical terms, more or less well defined and accepted by the volcanological community, the impetuously developing specialized subfields of the discipline introduce a number of new terms, frequently used in an ad hoc manner. Usage of multiple terminologies for the same phenomenon (e.g., edifice collapse, edifice failure, flank failure, sector collapse) is not an isolated case nowadays. Even the meaning of some classical terms (e.g., *ignimbrite, caldera*) under-

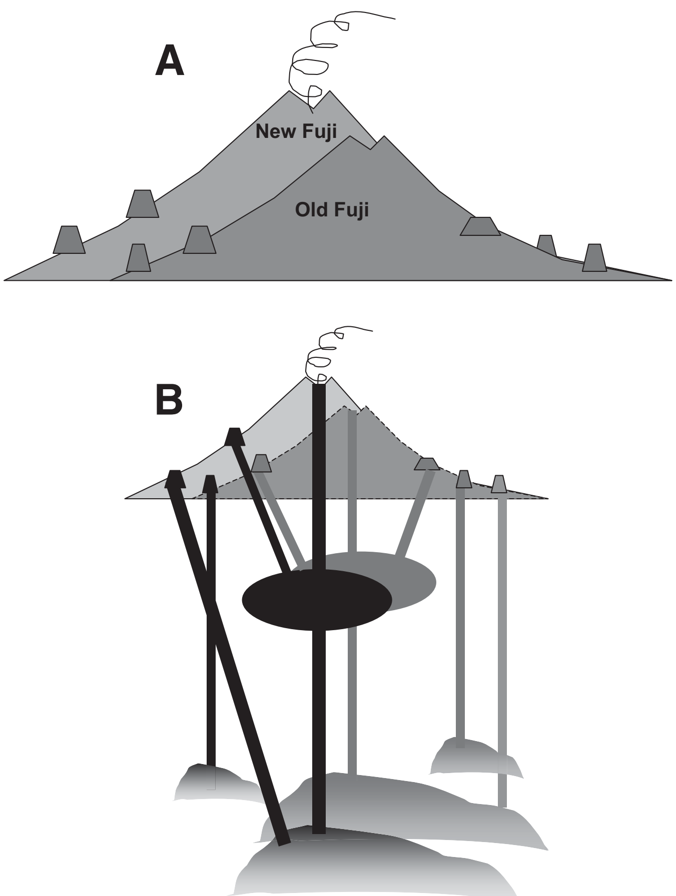

Figure 4. Illustration—using Mount Fuji, Japan, as a virtual example—on how the systemic approach of volcano definition may help in unraveling and understanding complex volcanic structures in which individual volcanoes overlap in space: (A) a generic pattern of organization of the complex (in spite of apparently simple conical geometry) volcanic structure of Mount Fuji, consisting of two spatially overlapping composite volcanoes distinct in time (an Old Fuji and a Young Fuji), and a number of small-sized satellite vents represented by cinder cones; (B) one possible explanatory configuration in a systemic approach: the complex structure results from a specific time-space pattern of a combination of volcanic systems.

went changes of meaning with time, most commonly acquiring genetic connotations after an initial purely descriptive meaning. The semantic content of terms evolves with time, as new details, and more profound understanding of the phenomena they designate is obtained. In other cases, the meaning of classical terms is artificially and inadequately expanded to accommodate new phenomena (e.g., *sector collapse caldera*) instead of introducing new names (as, for example, proposed by Szakács and Ort, 1998, for edifice failure or flank failure depressions). Unfortunately, in volcanology one may find many controversial or inadequately used terms. That reflects, for certain, a kind of related conceptual obscurity. A recent debate, for instance, on the meaning and usage of *erosional caldera* (Karátson et al., 1999; Karátson and Thouret, 2001, Szakács and Ort, 2001) is symptomatic in this respect.

For this reason I propose the creation of a branch of volcanology named *conceptual volcanology* that would have as a basic purpose the clarification of such nomenclature-related issues.

Personally, I would envisage *volcanology* as consisting of three main branches or sections: conceptual volcanology, theoretical volcanology, and applied volcanology. Applied volcanology, in turn, might be composed of (1) a "geological volcanology" dealing with all subjects related to the geological record of volcanism and volcanoes, including related phenomena such as hydrothermalism and ore genesis; and (2) a "societal volcanology" dealing with volcanic hazard, risk, monitoring, and other (adverse or benign) societal effects of volcanic phenomena, including climatic effects, historical and cultural side effects, etc. Such a classification of volcanology itself is a conceptual volcanology topic.

"Conceptual volcanology" is clearly distinct from "theoretical volcanology," which aims at understanding the physical processes of volcanic phenomenology, at expressing them in mathematical formulations, and, finally, at modeling them in a credible way. As such, theoretical volcanology is more technical. The recent book of Parfitt and Wilson (2008), *Fundamentals of Physical Volcanology,* is an excellent example of a theoretical approach in volcanology. Such syntheses are extremely necessary and welcomed. However, they cannot replace conceptual approaches. For instance, despite the many advances in theoretical volcanology, it is not an easy task to answer challenging questions, such as, "Why are geysers considered volcanoes on other planets or moons but not on Earth?" Actually, the answer to this particular question asked of the author by Alexander Belousov (January 2008, written commun.) basically depends on a long sequence of definitions that have little to do with the scope of theoretical or applied volcanologies as delimited above. First, it depends on the definition of *volcano* itself. If "volcano" is defined either as a particular type of magmatic system or as the eruptive subsystem of a magmatic system, the answer to the geyser question resides in the definition of "magmatic system." This system relies on the previous existence of a definition for *magma,* which in turn could be defined as molten rock generated below Earth's surface. Therefore, the answer to the geyser question relies on the definition of *rock* as a natural assemblage of minerals, and on the previous definition of *mineral,* and so on. Therefore, a geyser—whose definition is also needed—can only be considered a volcano if (1) water that erupts through its activity is accepted as magma, which could be the case if water is considered to be a melt of ice; and (2) ice is considered a rock, or more precisely a monomineralic rock composed of the mineral ice (this apparently uncommon definition of a rock type by using the same name as the mineral is not as uncommon as it might seem, because geologists are used to the dual nature of dolomite both to describe a mineral and a rock type, dolomite). In any case, the relevant question is then whether ice can be accepted as a rock. If it is considered that a rock should be stable at geological time scales, then we eliminate seasonal ice of temperate climate regions as a rock, but the term still will include long-lived polar ice and glacier ice in Arctic and high mountain environments. If *magma* is defined as a rock-melt only if it is generated by natural processes under Earth's surface, then only natural subsurface ice-melts can be viewed as magma. Such a reasoning leads, however, to the paradoxical result that geysers could be considered as volcanoes in some places on Earth (those areas of perennial ice) but not in others, and only under particular circumstances (when ice-derived "magma" is produced beneath Earth's solid surface). Alternatively, if a more restrictive "hard" definition of rock is accepted, according to which rocks are mineral assemblages of natural chemical substances that are present only, or prevailingly, as solid phases at Earth-surface and near-surface conditions (this is why native mercury—a liquid—is not a mineral on Earth), then ice will not qualify as rock. Consequently, terrestrial ice melts cannot be considered to be magmas, and consequently geysers are not volcanoes on Earth. However, on other celestial bodies where temperature-pressure (T-P) conditions allow for the presence of water only, or dominantly, as a solid phase, ice becomes a rock beyond any doubt, and the subsurface occurrence of the ice-rock may allow for the formation of water-magma able to erupt on the planet's or the Moon's surface. That is exactly what is already named *cryovolcanism* in planetary sciences (Francis, 1993; Lopez et al., this volume). In this case any water spring or geyser coming out at the surface from below is a volcano on those celestial bodies. On the other hand, if the definition of *geyser* includes some specific requirements (e.g., the presence of a shallow subsurface void in which liquid water is stored and heated by a local heat source, allowing intermittent explosive evacuation of that system), a geyser will not qualify as a volcano even in extraterrestrial environments, except possibly for very particular cases.

The above example suggestively shows how crucial definitions are in understanding and recording natural processes on Earth and other planets, and how deeply they influence the final outcome of a scientific reasoning process. This also is applicable to any conceptual approach in science. The discussion of the above example strongly suggests that devising isolated and unrelated definitions is not an acceptable solution when dealing with basic terms in volcanology at a conceptual level. In other words, any proper definition must be integrated in a self-consistent system of definitions. That system, built up around such pivotal concepts as "volcano," "volcanism," and "eruption" would comprise not only strictly volcanological terms but will be integrated in the conceptual framework of the broader Earth sciences. That is exactly what I mean when proposing "conceptual volcanology" as a distinct subdiscipline of volcanology. As such, conceptual volcanology is not a simple collection of acceptable definitions; it is much more than a glossary of well-defined terms. Those terms and their definitions make sense only if they are organized hierarchically as a coherent system of definitions and terms. Definitions organized in a system and nomenclature call for systematics and classifications; hence conceptual volcanology cannot avoid them. As a consequence, the three main subdivisions of a conceptual volcanology could be (1) definitions, (2) nomenclature, and (3) systematics.

Evidently, conceptual volcanology is not a job in which only one or a few persons are involved. Issues, such those mentioned above, should be analyzed and debated by (at least part of) the volcanological community before arriving at generally acceptable conclusions eventually with normative value. A specialized IAVCEI (International Association of Volcanology and Chemistry of the Earth's Interior) Commission, for example, could be an effective and visible forum for such debates. In my opinion, conceptual volcanology, once introduced and accepted as a subdiscipline of volcanology, would raise the intellectual level of scientific debate in volcanology and also provide technical approaches with a more intellectual flavor. This in turn would contribute to making volcanology more visible and attractive for decision makers and the cultivated general public alike. That could be a rewarding intellectual challenge not only to volcano theorists but to other scientific thinkers as well.

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APPENDIX

The Classical Philosophical Concept of *Definition*

The criteria of a correct, philosophically acceptable definition were established millennia ago by the classical Greek philosophy, and in particular by Aristotle (384 B.C.–322 B.C.). Although it is clear that the original meaning of the Aristotelian terminology has changed throughout the years, if attention is focused on the most basic aspects of his discourse it is also true that many aspects of the Aristotelian approach remain valid at present. In particular, several aspects related to the process of achieving a correct definition, and those related to classification systems, are relevant in the context of present-day scientific disciplines. This is well exemplified by the continued use of the classification system of living organisms in biology (and other branches of the natural sciences) discussed in more detail below. Consequently, despite the fact that the Aristotelian discourse was produced many centuries ago, it still can provide important insights that can help us to create more mature scientific disciplines. In particular, it can help us to better appreciate the problems faced by modern volcanology concerning the definition of one of its most basic concepts (i.e., volcano).

In the views of Aristotle, the philosophical requirements of a definition include (1) the denomination of a *genus proximus* (the closest class) to which the object to be defined belongs, and (2) a list of *differentia specifica* (i.e., specific differences) by which that object differs from other objects belonging to the same class (Noica, 1994). One important—and yet to be fully explored—implicit consequence of Aristotle's approach to the issue of "definition" is that any definition has only sense if it is part of a hierarchically organized system. When one defines a term designating a thing and names the *genus proximus*—the higher class of things to which it belongs—this implicitly recognizes that the reference term has been defined properly in advance. In other words, the predefined reference term has its own *genus proximus,* which, in turn, has been predefined, and so on. Similarly, after we define a term in the proper way, that term could serve later as *genus proximus* for the next generation of terms. As a consequence, definitions cannot be obtained independently from each other, but rather must be organized in a hierarchical structure that can be called a system of definitions. This structure is indispensable for defining any object unequivocally.

One of the later (A.D. sixth century) Neoplatonic commentators of Aristotle known as David states that the definition is "a concise enouncement *[sic]* which clarifies the nature of an object" (p. 14) and "delimitates the thing subjected to knowledge and separates it from those unknown" (Liiceanu, 1977, p. 19). Explaining Aristotle's approach, David further discusses the details of a definition by emphasizing the differences between a "definition" and other related concepts. For example, he states the equivalence between a name and the corresponding definition by saying that "what a definition is doing using several words, the name does using a single word" (p. 14). Nevertheless, he also stresses the major difference between "name" and "definition" by saying: "the name is a simple, synthetic and concise definition, while the definition is an unfolded name" (p. 14). And "definition consists of essential words and clarifies the substance and nature of the object … while description is conceived from accident and indicates the accessories of the thing which are exterior to its very nature" (p. 15). He further points to the difference between what is essential and accidental in definitions: "essential is what, being present [in the definition], preserves the integrity of the concept … accidental is what, being present [in the description] neither preserves nor annihilates its integrity" (p. 15). Aristotle himself warns against the confusion between *attribute* (i.e., essential property) of a thing which must be considered as the "specific difference" in a definition and *accident* (i.e., a casual property) which must not (Ross, 1927). David gives subtle explanations for discriminating between "(true) definition" and "descriptive definition" by saying, "a definition differs from a descriptive definition; while a definition is simple and is composed merely of essential words … a descriptive definition is a mixture … being conceived of both essential and non-essential words and … it is composed of a definition and a description" (all quotations from Liiceanu, 1977, p. 16). By making all of these precisions, David also warns against the abuse of descriptive words to be introduced in a definition, encouraging us to minimize the number of words of a definition by using only those essential to highlight the attributes of the object. To some extent, this is only possible if the *genus proximus* is properly chosen and well defined. Consequently, the adequate selection of a *genus proximus* is a necessary condition to satisfy David's claim about the desirable "(true) definition" as different from the undesirable "descriptive definition."

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Origin of Large Igneous Provinces: The importance of a definition

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ABSTRACT

In the original definition of a Large Igneous Province (LIP) much emphasis was put on the "extraordinary" character of these provinces. Such emphasis might have contributed to bias the form in which these provinces are commonly visualized, and consequently has contributed to the selective acceptance of genetic models. To avoid such bias, in this chapter the various available definitions are examined, taking into consideration the rules of logic that help us to avoid fallacies. From these definitions, the most critical parameters are identified, and an alternative model of formation of LIPs is advanced. The model developed here envisages LIPs as an extreme of a continuum in volcano-magmatic processes that are produced by essentially the same underlying processes. The differences between LIPs and non-LIPs are conceived as the result of different conditions present in a particular region, but that nonetheless have nothing extraordinary. Although the model developed here is one of several possible alternatives, by having identified the most common fallacies surrounding a LIP, even from its very definition, it might be possible to assess those alternative models in a more equilibrated form in future works.

INTRODUCTION

Large Igneous Provinces, or LIPs, were defined more than a decade ago as "voluminous emplacements of predominantly mafic *extrusive and intrusive* rock whose origins lie in processes *other than 'normal'* seafloor spreading" (Coffin and Eldholm, 1992, p. 17). This definition has marked much of the research done on the subject ever since, as LIPs have been almost unanimously considered to represent periods of *anomalously high magma production rates*. Thus, the essential interpretation of LIPs as events of an extraordinary character likely "to record periods when the outward transfer of material and energy from the Earth's interior operated in a *significantly different mode* than at present" (Mahoney and Coffin, 1997b, p. ix) has been a common feature of the vast majority of papers published on this subject over the past 15 yr. Actually, the emphasis put on the presumed anomalous magma production rate required to explain both the extrusive and intrusive components of LIPs is shared by essentially all genetic models, whether they are associated with the arrival of a mantle plume to the surface of the Earth (Eldhom and Coffin, 2000; Hooper, 2000; Richards et al., 1989; White and McKenzie, 1995), or to different processes of global scale (e.g., Abbott and Isley, 2002; Coltice et al., 2007; Hales et al., 2005; Jones et al., 2002; King and Anderson, 1995; Mutter et al., 1988; Sheth, 1999a; van Wijk et al., 2001).

To better appreciate the influence of the above definition of LIPs in the research made on the subject, I marked with italics key words that somehow predispose us to look for "extraordinary" explanations for these natural phenomena in exactly the same form that saying "do not think of a white elephant" almost invariably brings to the mind of the person hearing that message the image of such an animal, even if briefly. To some extent the problem surrounding any definition of a LIP was identified by Menard (1969) when he said that "the central problem is satisfactorily defining normal." Actually, this is a central problem that we need to face every time that we attempt to organize natural

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phenomena (not only LIPs) in any sort of classification scheme, therefore making it necessary to examine very briefly what the purpose of any classification is.

In the words of Best (1982, p. 20), "Classification is a human endeavor that attempts to recognize . . . common or contrasting features of related things . . . [although it should not be] an end in itself, but a means of seeing more clearly, simply and unambiguously the interrelations of the different properties of different rocks." According to him (p. 20), the objective of classification is therefore subdividing the continuous spectrum of a given property "in some meaningful way" that could help us to make a genetic interpretation of such diversity. As Best further recognized, usually there are thousands of possible forms to conceive a classification scheme, yet not all of the possible classifications are equally helpful for the identification of factors that are significant in the understanding of the genesis of the object of study. A corollary of such diversity is that some classification schemes might predispose us to make selective judgments based in the apparent order artificially introduced by the classification scheme itself. Consequently, the classification scheme exerts an often overlooked influence in the creation of genetic models, and it is important to keep in mind such influence if an unbiased interpretation of observations is really to be made.

The predisposition for a selective interpretation of observations introduced by a particular definition scheme not only represents a typical example of circular reasoning, but actually it could mark the beginning of what Dickinson (2003) has described as the *modern mythic style of thinking in geosciences*. Using the vocabulary of logic (Copi and Cohen, 1994), the characteristic aspect of mythical thinking is the selective assignment of truth values to some of the premises used in the interpretation of observations, sometimes in a very subtle form, but nevertheless favoring an a priori accepted conclusion. Consequently, to avoid mythical thinking it is extremely important to have definitions leading to classification schemes that are as unbiased as possible, yet at the same time allow us to recognize meaningful aspects that can be interpreted genetically. In this chapter I examine several aspects of the current, and two other more recent, definitions of LIPs, aiming to identify the elements with the largest potential to yield significant clues that could help us to better understand the genesis of LIPs and their relation to other manifestations of volcano-magmatic activity in our planet. In the second part of the chapter, I develop an alternative model for the genesis of LIPs that takes into consideration some of the physical constraints identified as more significant in the first part. As it turns out, the model proposed here is a special case of the general model of volcanism proposed by Cañón-Tapia and Walker (2004), and consequently it also contributes to a better understanding of the relation that exists between LIPs and other manifestations of volcano-magmatic activity in our planet. At the onset, it should be clear that the proposed model does not pretend to be a final and definitive answer to every possible question concerning the formation of LIPs, and it has much room for improvement concerning its predictive capabilities in particular from a compositional point of view. Nevertheless, I consider that this model provides an example of the form in which mythical thinking can be avoided in the study of this type of province. Other consequences of having good definitions in science, and in particular in the context of volcanic activity, are examined in the chapter by Szakács (this volume).

TWO RECENT DEFINITIONS OF A LIP

Very recently Sheth (2007) suggested the need to reexamine the currently accepted definition of a LIP quoted at the beginning of this chapter. The motivation for Sheth's suggestion seemed to be unrelated to the above considerations concerning the key role played by a definition and its associated classification scheme, but nevertheless reflecting some concern for such issues. In any case, Sheth's (2007) proposed definition of a LIP comprises the area covered by the igneous rocks present in a given province, so that any place where more than a threshold area is covered by igneous rock (he proposed this threshold to be >50,000 km^2) should be called automatically a LIP. Evidently, by adopting this definition, the bulk of the present-day ocean floor becomes the largest LIP that has ever existed in the geologic history of our planet, which is diametrically opposed to the explicit exclusion of ocean spreading from the group of LIPs made by the definition of Coffin and Eldhom (1992). Other departure from the original definition of a LIP found in the work by Sheth (2007) is that he devised a hierarchical system in which LIPs are subdivided in different categories depending on (1) whether the rocks of the province are extrusive or intrusive and (2) on the predominant composition of the rocks found in that province. Thus, at the first hierarchical level LVPs would stand for "Large Volcanic Provinces" and LPPs for "Large Plutonic Provinces" independently of rock composition. At the second hierarchical level, terms such as LRPs standing for "Large Rhyolitic Provinces," LGPs for "Large Granitic Provinces," or LBPs for "Large Basaltic Provinces" would be required. A most appealing aspect of such a classification scheme is that it contains more subdivisions than a scheme based in a "LIP" versus a "non-LIP" scheme inherent in the definition of Coffin and Eldhom (1992). The increased number of groups with contrasting differences in Sheth's definition might in principle facilitate the task of identifying the significant aspects of the formation of each type of province more easily than it could be possible if only two large groups are defined. In turn, such distinction might prove to be an advantage if the mechanisms controlling the genesis of each type of province are really different among the various groups of the classification scheme. Thus, by allowing ourselves to work with different subtypes of provinces we are more likely to identify processes that might not apply to every subtype, therefore increasing our understanding of these natural phenomena more rapidly than we would have done had we insisted in keeping together all of the provinces in one single class.

The revised definition of a LIP proposed by Sheth (2007), however, is not the only definition that has been advanced recently. Very soon after Sheth's work was published, Bryan and

Ernst (2008) proposed an alternative revised definition of a LIP. The classification scheme proposed by these authors also recognizes that the original definition of the term *LIP* made by Coffin and Eldhom (1992) might have become inadequate to convey the most recent discoveries. Consequently, Bryan and Ernst considered that a new set of criteria should be adopted before assigning to a particular province the status of a LIP. In particular, Bryan and Ernst (2008) suggested that age, crustal and tectonic settings, the predominant intrusive or extrusive character of the rocks, their composition, area, volume, and rapidity of emplacement should all be considered an integral part of the definition of a LIP. Thus, by combining all of these factors Bryan and Ernst (2008) proposed the following definition: "LIPs are magmatic provinces with areal extents >0.1 Mkm^2, igneous volumes >0.1 Mkm^3, and maximum life spans of ~50 Myrs that have *intraplate tectonic settings or geochemical affinities,* and are characterized by igneous pulse(s) of short duration (~1–5 Myrs), during which a large proportion (>75%) of the total igneous volume has been emplaced." Although this definition is much more complex than the original definition issued by Coffin and Eldhom (1994), the hierarchical scheme of classification associated with the new definition only has one hierarchical type allowing distinction of oceanic and continental LIPs, because the rest of the criteria incorporated in the definition are used simultaneously to give place to the various categories listed in the second hierarchical level. In practice, this means that although criteria such as age or intrusive versus extrusive character might be used to attach some labels, all of them have the same hierarchical weight and consequently do not favor the identification of independent genetic processes.

At first sight, the Bryan and Ernst definition seems to be an improvement relative to the definition made by Coffin and Eldhom (1992). On closer inspection, however, it is seen that these two definitions have the same weakness, indicated by the italics that I inserted in them both. Indeed, including a tectonic setting and presumably an associated geochemical affinity in the form done by Bryan and Ernst (2008) actually favors the selective interpretation of observations from within the classification scheme. Note that such a bias is not found in the scheme proposed by Sheth (2007), in particular regarding geochemical composition, because the latter scheme is descriptive and allows for the inception of different LIP subtypes should the need arise (i.e., the lack of enough data to form a hierarchical subtype at present does not exclude its probable eventual creation if future observations reveal that a large enough group with such a distinctive characteristic does indeed exist in nature). In contrast, in the definition proposed by Bryan and Ernst (2008), geochemistry is used as an exclusion criterion leading to a "LIP" versus a "non-LIP" classification scheme that might not account for a diversity of independent LIP subtypes. Consequently, this aspect of the definition proposed by Bryan and Ernst (2008) seems to be opening the doors for the occurrence of mythical thinking in the subject, in exactly the same form that the previous definition by Coffin and Eldhom (1992) did when referring to LIPs as "due to processes other than normal."

Having said this, it is necessary to recognize that the strong emphasis made on the size of the area covered by the rocks in the scheme proposed by Sheth (2007) might hamper the identification of common or contrasting features that might turn to have a genetic significance. The main question is not whether the threshold value should be 50,000 km^2 or 100,000 km^2, inasmuch as the fact that the definition advanced by Sheth (2007) is independent of time. As pointed out by Bryan and Ernst (2008), given sufficient time basically all processes responsible for the generation of magma will produce igneous rocks of LIP-scale dimensions. Consequently, the introduction of time as a parameter in the classification scheme seems to be an important feature that might contain clues concerning the genesis of this type of provinces. Unfortunately, such a parameter is excluded from Sheth's definition.

In summary, while it is apparent that the original definition of a LIP proposed by Coffin and Eldhom (1992) has been superseded by the research made in the past 15 yr, it would seem that we still lack a satisfactory form for classifying this type of natural phenomena. From my point of view, such a lack of clarity in a classification scheme has contributed at least in part to favor mythical thinking in the study of LIPs for all of these years. Whereas the more recent definitions of a LIP take steps in order to avoid such biases, there are still voids in the current definitions that need to be addressed before actually being able to have a truly unbiased interpretation of observations. Some of these issues are examined in the following sections.

OBSERVATIONS AND INFERENCES ON LIPs AND NON-LIPs

An underlying issue in the debate of the origin of LIPs is their probable relation with an extraordinary behavior of Earth's interior during their formation. Establishing what is "normal" and what is not, however, is a rather difficult task for several reasons. For instance, almost every geoscientist would agree when saying that at present there is no evidence suggesting that a LIP is being formed anywhere in the world. Based on such an observation, we might conclude that the present situation represents the "normal" case. Nevertheless, it is equally valid to assume that the scenario leading to the formation of LIPs is the normal situation, and that we are nowadays passing trough a time of "abnormal" activity. In this sense it can be argued that LIPs have been fairly common throughout Earth's history when regarded as a group and not on a one-to-one basis (see references in Ernst et al., 2005; Macdougall, 1988a; Mahoney and Coffin, 1997a), therefore further justifying the idea that the anomalous behavior actually is represented by the present-day scenario. In fact, many other arguments can be used to support either the normality or abnormality of the processes that generate LIPs, all of which depend on the frame of reference that is being used. Consequently, it is suggested that a first step for avoiding mythical thinking in the study of LIPs is that instead of referring to "normal" and "extraordinary" or "abnormal" events when describing these provinces it is wiser to restrict our judgment to distinguish two types of volcano-magmatic activity without making reference to their status of "normality."

To avoid mythical thinking, the following step is to establish as objectively as possible the characteristics of each of the two identified types of volcano-magmatic activity. Although apparently simple, it is in the comparison between the two types of activity that the risks of recreating the processes of mythical thinking become very large. This is the case because a large number of sometimes unidentified assumptions might influence the form in which some evidence is presented and compared with the other extreme of the spectrum. For instance, one of the presumed distinctive characteristics of LIPs, until now, has been their high rate of magma production. Magma production rates, however, cannot be directly measured either in LIPs or in present-day volcanic provinces, as these processes take place in a part of the Earth that remains inaccessible for direct observation despite recent advances in technology. Consequently, any judgment concerning LIPs or any other volcanic province around the world that is based on magma production rates necessarily contains an underlying set of previous assumptions that are necessary to infer such magma production rate in the first place.

Understanding the role played by such underlying assumptions is critical to avoid a logical error during the process of inference that might result in the construction of a formal fallacy. As some readers might not be very familiar with the formal nomenclature of logic, it might be convenient to open a parenthesis in the presentation that is devoted to examining in more detail the various forms in which a fallacy can be committed. Readers familiar with such rules of the process of reasoning might skip this parenthesis.

Anatomy of a Fallacy

There are many ways in which formal errors in logical reasoning can take place. Some of these errors are somewhat difficult to identify, as the argument (or syllogism, in the nomenclature of logic) may seem to be correct at first sight, and these are generally referred to as fallacies (Copi and Cohen, 1994). One such error is to construct a categorical syllogism that gives the appearance of containing three terms (two premises and a conclusion) when it actually contains more. This error commonly takes place when one of the premises actually contains a second premise that is presented in a cryptic form, being embedded in the premise that is easily identified. Alternatively, this error can be made when a given premise is considered to have a fixed truth value when in fact its truth value depends on the truth value of another, non-explicitly mentioned premise. The exact name of the fallacy that is committed in this form depends on the definition of *syllogism* that is used. Nevertheless, these general groups of fallacies can be detected if proper attention is given to some simple rules.

In particular, it is noted that the use of an ambiguous statement as a premise in the construction of another syllogism may result in an error for three reasons (Copi and Cohen, 1994). First, an error in the syllogism is produced because the truth value of the ambiguous premise depends on a different premise. Consequently, failure of detecting such a cryptic premise leads us to commit the "fallacy of quaternio terminorum," or the fallacy of four terms (note that the name remains regardless of the real number of hidden premises). Second, if the truth value of one of the cryptic premises turns out to be false, then the conclusion of the second syllogism necessarily must be false. Failure in acknowledging this possibility will lead us to commit the fallacy of "drawing an affirmative conclusion from a negative premise." Third, failure to recognize the existence of the hidden premise might contribute to committing the fallacy of "equivocation" when one of the terms is used in different senses in each of the two premises explicitly stated in the syllogism.

To illustrate the three types of fallacies in a context relevant to the present chapter it is convenient to consider the form in which seismic imaging is sometimes used to make inferences concerning the characteristics of the Earth's interior, and how these inferences are sometimes used in connection with the origin of LIPs. Although some workers might consider seismic imaging of the Earth's interior an unbiased and very objective source of information, it turns out that there are several assumptions made in the interpretation of the actual data (for a recent and extensive discussion of such assumptions see, e.g., Thybo, 2006). Discrepancies concerning some of those assumptions can actually lead to discrepancies concerning the interpretation of the actual data in significant forms. Furthermore, regardless of the final interpretation concerning the probable occurrence of melt at depth that is reached when conducting a seismic survey of a region, it is clear that measured seismic data only contain information concerning the physical state of the rocks through which seismic energy actually traveled. If that physical state changes in time, then the conclusion reached by the seismic survey would be invalid. Realistically we do not expect that the physical state of large portions of Earth's interior will change in lapses of minutes or even of days, but if lapses of thousands or even millions of years are involved, however, then the occurrence of such a change becomes a real possibility. Therefore, it should be clear that seismic information only provides some constraints concerning the probable physical state of the Earth's interior at times not much different from that of measurement, and even in this case it is possible to reach two contrastingly different conclusions based on the same type of data, as illustrated by comparing the conclusions reached by Thybo (2006) with those reached by Priestley and McKenzie (2006) concerning the presence of melt within the mantle.

In the present context, the relevant fact is not to discuss which of those conclusions is "true" but to focus on the fact that completely different conclusions (either a "true" or a "false" conclusion in logical parlance) can be reached by using the same type of observations. The relevant part is therefore that the difference between the two possible truth values of the conclusion depends on the assumptions that are used for the interpretation of the observations. Consequently, although seismic interpretation might be a reasonable source of information, there is an ambiguity in its truth value, as this value depends on the truth value of the premises used in the interpretation of the signals. Consequently, failure in acknowledging the relevance of the assumptions made when interpreting seismic information facilitates the completion of the first type of fallacy (fallacy of four terms) as those premises become hidden and constitute a source of ambiguity that is not acceptable in a valid syllogism.

The second type of fallacy (drawing an affirmative conclusion from a negative premise) might be extremely difficult to identify because commonly we overlook the truth value of a premise that is not explicitly stated in the syllogism, and, even worse, the truth value of the hidden premise might become accepted "de facto" more as the result of habit than as the result of a real exercise of logical inference. Actually, this type of error is very common in mythical thinking, as it promotes the selective acceptance of some facts, rejecting any questioning about them, and many examples of this type of fallacy can be found in the literature dealing with the existence of mantle plumes in Earth, some of which were examined by Dickinson (2003).

A practical example of the third type of fallacy (of equivocation) in a context relevant for the discussion about the origin of LIPs can be found when the term *melt* is under scrutiny. For instance, some workers have argued that the amount of melt produced in a given setting can be inferred from the composition of the erupted products, and that such volumes of melt can be corroborated by using seismic signals (e.g., Korenaga et al., 2002; White et al., 1992, 2001). Despite their apparent appeal, these works have the problem of combining two disparate sources of information (seismic and geochemical), each of which has a different set of premises that may or may not be true, therefore resulting in two conclusions (one seismic and the other geochemical in nature) that are used as premises for a new syllogism despite the existing ambiguity in terms of each of their truth values. Actually, the source of the problem (at a logical level) is that the term *melt* in each of the original approaches has a slightly different sense. In the seismic study *melt* actually denotes *crustal thickness,* which in turn has been assumed to be the result of the collection of a liquid phase that (1) was extruded in its entirety from the region of origin but (2) remained trapped at depth to form the observed crustal thickness. In contrast, in the geochemical approach, *melt* denotes a "cumulative volume of liquid" that was formed within the region of origin before an eruption and that was expelled all the way to the surface. Therefore, when comparing seismic and geochemical evidence we are comparing the inferred thickness of a solid layer that we think was produced as the result of a complex process of melt extraction out of the region of genesis, but that nonetheless was not sufficient to move such liquid all the way to the surface to be later eroded, with an inferred volume of liquid that was expelled all the way to the surface. Evidently, many more factors were involved in the creation of the seismic "melt" than in the creation of the geochemical "melt," and any numerical agreement related to the volumes of "melt" produced in the two cases might be a coincidence rather than being a direct measurement of a given melt volume. Failure in recognizing this possibility, inherent to the slight change of meaning of the term *melt* in both methods, leads to the "fallacy of equivocation."

Before returning to the main subject of this chapter, it is important to note that identification of fallacies must not be confused with undue criticism to any of the methods used to make inferences concerning the internal state of Earth. For instance, although the conclusion reached in the sense that the agreement of both "thicknesses of melt" in the third example given above (i.e., one seismically and the other geochemically determined) gives place to a fallacy, such fallacy of equivocation does not allow us to make any judgment of truth concerning the validity of each of the methods if considered independently of each other. This is the case because the fallacy is actually formed when both types of information are forced to be part of the same syllogism, and not because any of the parts is necessarily false. In other words, it might be that the seismic method yields a true crustal thickness, whereas the geochemical method yields a true fractional distribution of melt as a function of depth, even if the former is not necessarily related to the volume of melt produced in a single region of partial melt (RPM) at a given short time interval, and the latter does not necessarily correspond with the estimated crustal thickness measured by seismic methods. Consequently, it should be clear that the use of both methods of obtaining information about some of the characteristics of Earth's interior will still be valid (at a logical level) as long as the conclusions reached by each method are not invoked as "corroborations" of the truth value of the conclusions reached by the other method.

As a summary of this parenthesis it can be said that if the set of assumptions made by any method of observation is not the same for the two types of volcanic activity being compared, then the comparison might be biased, and it might result in a fallacious conclusion. Consequently, a critical step that needs to be taken to avoid mythical thinking in relation to the origin of LIPs is to be certain that we are comparing the same type of evidence gathered through equivalent means and with the same set of underlying assumptions for both LIPs and non-LIP provinces. In the following sections I examine with some detail some of the commonly used sources of information, and the form in which these sources of information can allow us to compare LIPs and non-LIP volcanic provinces in a relatively unbiased form, starting by punctuating the meaning behind some key terms.

"CFBs" instead of "LIPs" and "Modern" instead of "Non-LIP" Volcanic Provinces?

One of the most pressing restrictions faced when attempting to characterize LIPs is that because of their various levels of exposure and ease of access not all of these provinces have been studied with the same detail. Consequently, it might be convenient to restrict the universe of studied provinces to those that can provide the most complete record of evidence obtained independently of any genetic interpretation (note that *universe* is used throughout this chapter in the mathematical-logic sense, particularly in set theory, where this term denotes the set that contains as elements all the entities described by the class). In particular, it is noted that among the best documented LIPs there is a group of provinces that have erupted most of their products over continental crust (continental flood basalts, or CFBs), and for which a flow-by-flow stratigraphy and a set of relatively extensive radiometric ages is available in many cases. As in all LIPs, CFBs are characterized by having large volumes of lava (>10^6 km^3) usually erupted in very short times (<1 m.y.) (Hooper, 2000). Consequently, we might consider that CFBs are good representatives of LIPs even when LIPs and CFBs are not completely equivalent terms (the latter being a subgroup of the former, regardless of which definition of a LIP is adopted).

On the other hand, characterizing the "non-LIP" provinces is not as simple as it might seem at first sight, especially if this type of province is defined by different criteria that do not become the complement of the "LIP" definition. Good candidates that can be considered to *represent* this type of volcano-magmatic activity are all of the volcanic provinces actively forming at present, because most geoscientists would easily accept that no LIP is being formed at present. It is emphasized that we are concerned with finding a group of provinces that can be considered as representative of the "non-LIP" type of activity, rather than defining the whole universe of provinces belonging to this group, and in this sense the "present-day" subgroup seems a good candidate. It should be noted, however, that not all of the present-day provinces have been studied with the same detail, that the amount and quality of available information might vary from province to province, and that "present-day" might be a far too restrictive time frame. For instance, although some direct observations allowing us to determine magma extrusion rates very precisely have been made in a few "present-day" provinces, these are restricted to a limited number of volcanoes. Furthermore, the period of observation available for some volcanoes in some cases might be extremely short to be considered as representative of the average behavior of that particular volcano. Consequently, it would be unwise to restrict this type of volcanic activity only to those examples where direct observation of an eruption has taken place. Thus, the second type of volcano-magmatic activity (the one that will be compared with CFBs) might be reasonably formed by "modern" volcanic provinces, understanding by this term all examples of relatively recent volcanic activity, regardless of whether they can be considered part of presently active provinces or not, provided that they cannot be considered akin to any LIP in an obvious form.

An advantage of comparing CFBs and modern volcanic provinces (as just defined) instead of the larger universe of LIPs and non-LIPs is that we allow ourselves to compare information gathered through the same methods in both instances. In other words, instead of being forced to compare information gathered through, say, seismic and geochemical methods, therefore risking the danger of committing a fallacy of equivocation (see above), we can compare both provinces by using information that is gathered by exactly the same methods. In this form we also diminish the possible bias that could be caused by committing the fallacy of four terms, because even when undoubtedly there might be hidden premises not explicitly stated in our analysis, those hidden premises will be shared by both groups entering the comparison. Consequently, no selective bias will enter our comparison, therefore diminishing the possibility of recreating the patterns of mythical thinking.

The only fallacy that cannot be completely eliminated from any analysis of a natural system is the fallacy of drawing an affirmative conclusion from a negative premise. As pointed out by Oreskes (1999, p. 3), "the history of science demonstrates that the scientific truths of yesterday are often viewed as misconceptions, and, conversely, that ideas rejected in the past may now be considered true." Actually, this condition is inherent to all natural systems, because a definitive proof is only possible in closed systems. In all open systems positive proof is not possible, and we are limited to eliminate alternative hypotheses only by disproving them as new evidence becomes available (Oreskes et al., 1994). For this reason, extrapolation is a procedure that always has the possibility of leading to a fallacious syllogism. Evidently not all of the extrapolations are fallacies, but it is extremely important to be aware at all times of the original range for which factual observations were used as well as of the assumptions behind the extrapolation. In this sense, while there are many advantages in reducing the universe of LIPs to the representative group of CFBs as discussed above, there is also an increased danger of drawing a fallacy when conclusions reached by studying CFBs are extrapolated to other LIPs. Evidently, the larger the differences between a given LIP and the bulk of CFBs, the larger is such a danger. For instance, although there is no doubt that giant dike swarms (GDSs) or Archean greenstone belts (AGBs) might have all the requirements to be considered LIPs, there are significant differences between these types of provinces and the group represented by CFBs. In contrast, the differences between many oceanic plateaus (OPs) and ocean basin flood basalts (OBFBs) on the one hand and CFBs on the other are probably not that significant. Therefore, the danger of a fallacy will be larger when interpreting GDSs and AGBs than when interpreting OPs or OBFBs. In consequence, it would seem that from the point of view of logical reasoning it is convenient to have criteria that could be used to distinguish between these types of LIPs in the classification scheme. Similar arguments can be used to sustain a case of fallacious reasoning when extrapolating conclusions reached by using the subgroup of modern volcanic provinces to other, more ancient "non-LIP" provinces. Nevertheless, if we compare the benefits of reaching some insights into the processes by comparing representative subgroups of each type of volcanic activity, rather than comparing the whole universe of provinces forming each group, the danger of extrapolation becomes justified.

In summary, it should be clear that by adjusting our universe of observations to a range that allows us to have equivalent sources of information across a given threshold, the danger of committing a formal fallacy is reduced. This also should help us to eliminate the more unlikely hypotheses much more easily, because the number of parameters entering our analysis is also likely to be reduced. The problem is then to decide which of the proposed criteria entering the definition of a LIP lead to more significant thresholds that can be used to make a significant distinction between LIP and non-LIP types of volcano-magmatic activity in genetic terms.

TOWARD AN UNBIASED YET USEFUL CLASSIFICATION SCHEME OF LIPs

Reducing the universe of examples that need to be considered when trying to establish the main differences between LIPs and non-LIPs, as done in the previous section, is an important step in avoiding biasing the analysis, and at the same time ensuring that we have a large enough number of examples that can be considered as representatives of each group of contrasting volcano-magmatic activity. To really avoid the most frequent fallacies, however, it is necessary to examine with some care the actual information that will serve as the basis for the intended comparison to be sure that we are actually comparing equivalent types of information (or in the language of logic, to be sure that we are dealing with the same number of premises in each case). One of the parameters suggested both by Sheth (2007) and by Bryan and Ernst (2008) is to consider whether the province is characterized by intrusive or extrusive rocks. Nevertheless, although Sheth (2007) uses this parameter as a foundation for establishing his hierarchical classification, Bryan and Ernst (2008) only use it to distinguish different types of provinces that are at the same hierarchical level. It is considered here that a hierarchical classification that explicitly distinguishes between predominantly extrusive and predominantly intrusive LIPs provides a natural breakpoint in the continuum of volcano-magmatic activity, and consequently this criterion seems to be well justified as an integral part of the classification scheme of LIPs.

Among the other parameters that have been proposed to enter the definition of a LIP there are four that can be directly measured without the need to make reference to any genetic model, and also requiring a lower amount of previous assumptions. These parameters are (1) the age of the rocks, (2) a time lapse of activity, (3) the area covered by the rocks, and (4) the volume of igneous rocks. A fifth, relatively unbiased parameter that can be obtained from those direct estimates is an average time of activity. Although undoubtedly there are uncertainties concerning the estimation of all of these parameters in every type of volcanic province (i.e., LIPs and non-LIPs alike), when all things are considered these uncertainties do not exert any influence in the validity of the inferences made from the perspective of logical reasoning. Nevertheless, their potential value to yield a significant classification scheme is not the same in all cases. These parameters will be examined in more detail next.

Age of the Province

Both the age of the province and the time lapse of activity commonly are determined from radiometric measurements made on the rocks of that province. Whereas definitively the uncertainties are not the same for each radiometric method available, from the point of view of logical reasoning all of the different methods are completely equivalent because the type of assumptions made by all of them are essentially the same (e.g., the system remained closed to mass transfer of a given isotope after formation of the rock, the decay rates are well known, the contents of the isotopes of interest can be determined accurately, etc.). Consequently, despite the fact that the numerical uncertainties can change depending on the method used to obtain the date (and this in turn might depend on the age of the rocks themselves), all radiometric ages can be considered as completely equivalent sources of information at a logical level. Therefore, in these two cases we can discuss the relevance of the parameters measured without concern about the specific uncertainties associated with the method of measurement.

Although at present some questions need to be addressed in which the age of the various LIPs might be important; e.g., links with exogenic or terrestrial forcing processes or mass extinction events, periodicity, and clustering of LIP emplacement (see references in Ernst et al., 2005) there does not seem to be a significative difference in the genesis of older versus younger LIPs that can be identified from other data. For instance, a plot of the age against the area of LIPs does not reveal any particular trend (Fig. 1). Although this plot does not include a few LIPs with a documented extension larger than 10^5 km^2, the exclusion of these data does not alter the absence of a pattern. Also, the fact that the age used to plot the data was always the oldest age in the database should not modify this conclusion, because some of the reported

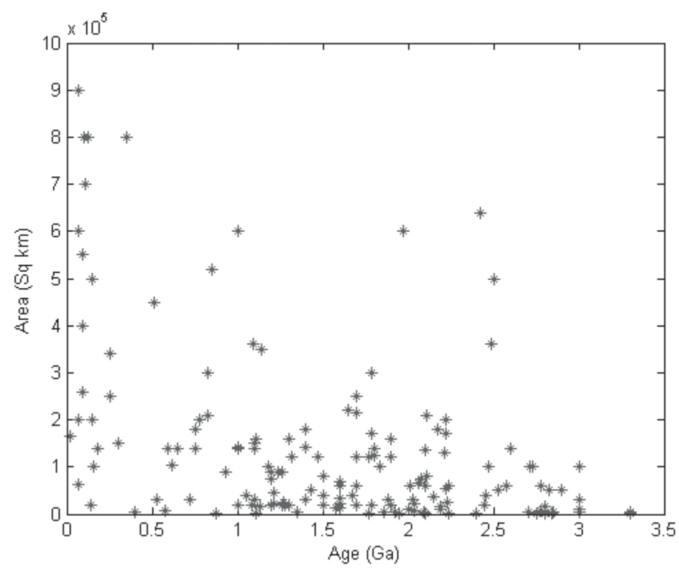

Figure 1. Plot showing the age versus the area of LIPs for which determinations of both variables have been reported in the database of the Large Igneous Provinces Commission (accessed May 2010). Exclusion of five LIPs for which the area exceeds the limits on the vertical axes of the figure does not change the absence of any trend, and actually makes it more difficult to observe the scatter of the data shown in this figure.

age ranges are very small relative to the total age scale of the figure. Thus, at least at present, it would seem that using the average age of the province as a parameter in the classification scheme does not provide any advantage from a genetic point of view.

Duration of Activity

Perhaps the biggest difference in the definition of a LIP proposed by Sheth (2007), relative to those of Coffin and Eldhom (1992) and Bryan and Ernst (2008), is that Sheth's definition does not consider the duration of activity as an important parameter. Because we are interested in a classification scheme that can be used to obtain an insight into the genesis of these provinces, and considering that given enough time any province might reach dimensions that would classify it as "large," as pointed out by Bryan and Ernst (2008), it would seem that contrary to Sheth's suggestion, the duration of activity is a parameter that should enter the definition of a LIP at essentially any hierarchical level. Failure to include this parameter might result in ambiguities even in the most basic definition of a LIP. To illustrate this point, consider that although the Hawaiian-Emperor Seamount chain can be included as an example of a LIP if only size is considered, this requires the inclusion of products of the Meiji Seamount that were erupted >80 m.y. ago, and probably extending to more than 100 Ma (Scholl and Rea, 2002). Such a time frame exceeds the limits introduced in the definition of a LIP made by Bryan and Ernst (2008), and therefore the whole chain of seamounts should not be considered a LIP even when it is probable that such chain has been created by essentially the same process. Evidently, we could impose the constraints on the time of activity as set by Bryan and Ernst (2008) and split the seamount chain into LIP and non-LIP parts, but this would seem a rather arbitrary breakpoint that has nothing to do with the genesis of the province. Furthermore, if the chain of seamounts is split in time intervals, each of 5 m.y. (note that this continues to be an arbitrary time lapse without any other apparent justification behind it), we would conclude that the most recent LIP in the chain is the most productive, which in turn leads to another series of problems concerning the presumed plume head and tail production, as already discussed by Foulger (2007). Perhaps a more natural breakpoint in this case would be to consider the time scale associated with the formation of one individual island, as this period is more likely to represent a pulse of volcanic magmatic activity. Evidently, if only the more recent volcanoes of this chain (say, ca. 1 Ma) are considered as part of the province, instead of considering the whole volume of the chain of seamounts, we would need to consider the volume of one of the islands. Note that the 1 Ma breakpoint selected here is slightly larger than the interval of formation of the Big Island of Hawaii, because the oldest volcano of this island (Kohala) seems to be younger than 0.78 Ma (Sherrod et al., 2007), but such a difference does not alter the following conclusion because the volume of products is much smaller than the normally accepted threshold values in either definition of a LIP. A representative volume of the Big Island can be obtained by considering a circular cylinder having the same basal surface than the present-day island of Hawaii, ~10,500 km^2, and a height of 9 km that would account for the submerged part of the island that nonetheless is over the seafloor. Both in terms of the surface area and the associated volume, it is clear that the island of Hawaii should not be considered as a LIP because both area and volume are about one order of magnitude smaller than the accepted thresholds; this remains to be the case even if the much lower threshold set by the LIP definition of Sheth (2007) is used. Therefore, as illustrated by this example, the time used as reference is extremely important to define whether a given province can reach the required size (whether areal extent or estimated volume) to be considered a LIP. On the other hand, this example also illustrates that selecting an arbitrary threshold for time in the definition of a LIP can lead to an unjustified segmentation of a province that could mask relevant clues concerning the genesis of these provinces.

Another set of problems is derived from the fact that we would need to decide which time scale is really significant in a LIP classification scheme if we want to assign the status of a hierarchical marker to this parameter. For instance, we might decide to include in our estimations the oldest and youngest available radiometric ages for one particular province, because there is no doubt that radiometric determinations are unbiased in terms of any genetic interpretation of the provinces. Adoption of the radiometric record in this form, however, would imply neglecting information coming from detailed stratigraphic or morphologic studies that suggest that extremely long periods of quiescence might have occurred in at least some of the CFB provinces (Jerram and Widdowson, 2005). Therefore, by adopting the radiometric age range without considering the local stratigraphy, we risk estimating an extremely large period of activity that might not convey an accurate representation of real processes occurring in the province at any time. At the other extreme of the temporal scale, we could consider the possibility of estimating the time of activity as reflected by one eruptive event, and with enough information in assessing typical times of quiescence. This approach faces a practical problem, however, that precludes the real application of this criterion in a deterministic form. The problem is that we do not have the resolution power to date with radiometric means and with the required accuracy the products of a single eruptive event. This lack of resolution is associated with the probable fact that most eruptive events take from less than a day to less than ten years to be completed (Simkin and Siebert, 2002). Thus, even when the direct observation of active volcanoes tells us there is the occasional eruption that might last over 20 yr, it is really risky to extrapolate such long durations as typical values for eruptions taking place in LIPs. Consequently, it is clear that obtaining a really accurate estimate of the real time of duration of any single eruption observed in the geologic record is almost impossible at present, and therefore it would not be recommended to use this time frame as the basis of a hierarchical classification of LIPs.

In summary, the exclusion of a time frame in the definition of a LIP seems to introduce an unwanted source of uncertainty that hampers the identification of significant trends in a genetic interpretation of LIPs, but uncertainties in the determination of relevant time frames preclude the use of this parameter as a marker that can be used to define a hierarchical category in the classification scheme. Consequently, whereas it is recommended that the duration of activity be included in the definition of a LIP, this should be done with extreme care, keeping in mind that at least two time scales might become important in further analysis. One of these time scales is related to the whole duration of activity in the province; the other is related to the probable duration of individual events. The first of these relevant time scales formed part of the original definition of a LIP and therefore has been incorporated by most studies made on the subject in the past 15 yr. In contrast, the second of these time scales has been excluded from the analysis of the evidence in most cases. The relevance of such a time scale will be further discussed later in this chapter.

Area and Volume of Erupted Products

Sheth (2007) considered that the area covered by igneous rocks in a given province should be used instead of using their volume as a parameter required in the definition of a LIP. The justification for this preference was that estimation of volumes are more uncertain and are affected more than areas by erosion in the older provinces. Although undoubtedly it is certain that the volumes of igneous rocks in LIPs must include uncertainties derived from their old age and tectonic influences, it is also true that measurements of erupted volumes in modern volcanic provinces are not devoid of difficulties. Actually, from the point of view of logical reasoning, the fact that such uncertainties in the estimation of volumes does increase with the age of the province is irrelevant, because such uncertainty will influence the position of the threshold value, but it does not affect the fact that this parameter can be used to define such thresholds. For instance, some of the uncertainties found when estimating the volumes of erupted products in modern volcanic provinces include the burying that results from still more recent products as well as some important effects of erosion. In addition, the lack of continuity in the exposed products of either CFBs or modern volcanic provinces makes both types of provinces equally susceptible to spurious correlations based on geochemical arguments (if the regional source of magma is more or less homogeneous) or paleomagnetic arguments (if two different eruptions took place either in a relatively short time period or at two very different times that nonetheless had similar orientations of the paleomagnetic field). Consequently, the possibility of committing an error of judgment that could affect the true value of a particular premise (or in this case the number associated with the volume of erupted products in each province) is essentially the same in both CFBs and modern volcanic provinces, and therefore it does not result in a selective bias. Relatively similar arguments can be used to extrapolate this result to the realm of other LIPs, although in this case the uncertainties in the isotopic ages can also become important sources of spurious correlations.

On the other hand, it would seem that volume can bear a more direct relationship than area from the point of view of genetic processes. The area covered by some erupted products might depend strongly on factors such as previous topography or vent distribution, but it also might depend on the viscosity of the magma, the mechanism of growth of the products (endogenous versus exogenous), and rate of eruption. Undoubtedly, topography and vent distribution are also likely to influence the estimation of volumes of igneous rocks, but all the other parameters will have less influence in those determinations. Besides, the volume of igneous rocks can be related (directly or indirectly) to the amount of magma produced in a given part of the mantle, which is likely to bear some information concerning the genesis of those provinces. Consequently it is considered here that volume is a better parameter to be used for the characterization of the size of any province than area. Other reasons to prefer volume over areal coverage were examined elsewhere (Bryan and Ernst, 2008).

Although these considerations indicate that volume is indeed a required parameter that needs to be included in the definition of a LIP, it remains to be determined whether this parameter can be used to establish a finer classification of the products (i.e., as a hierarchical indicator). At this time it seems that such finer division might not be possible because the number of provinces for which volume estimates are documented is relatively small. In effect, most of the provinces included in the database of the Large Igneous Provinces Commission (accessed May 2010) at the time of this writing lacked this estimate of their size. Consequently, identification of significant breakpoints in volume estimates for the whole universe of LIPs would seem premature. In addition, it should be noted that some of the volume estimates are somewhat speculative in terms of the relevant ages involved, especially in those cases when the volume estimates have been based in indirect observation through geophysical means. As an example, it is interesting to note the case of the Okavango dike swarm in Botswana. Jourdan et al. (2004) documented the presence of two populations of dikes of contrastingly different radiometric ages in this dike swarm, suggesting that the younger, Early Jurassic dikes were emplaced along a reactivated zone of lithospheric weakness marked by the older, Proterozoic dikes. From this evidence it is clear that the combined volume of all of the dikes in the swarm is an overestimation of the real volume of magma involved in the formation of a single LIP. Furthermore, as pointed out by Jourdan et al. (2004), the presence of two contrasting ages in dike swarms might not be unique to Okavango, making unclear what proportion of any swarm has been emplaced as the result of a unique event of LIP dimensions, or rather it contains dikes emplaced during two events separated in time for more than 50 m.y. (which is the time frame specified in the definition of a LIP by Bryan and Ernst, 2008). Consequently, any fine subdivision of LIPs based in volume as a criterion would be somewhat misleading at this time, and therefore the volumes of whole provinces should enter the definition of a LIP only as a rough indicator that would allow us to divide the continuum of volcano-magmatic activity in two categories: “Large” and “no-Large” provinces. Considering the uncertainties associated with the quantitative determination of this parameter, it seems that discussions concerning the exact location of the breakpoint dividing both categories are somewhat useless at this time.

It was noted above that there are two broad time scales that should be kept in mind. The situation with the volumes of igneous rocks is not different. In the preceding lines I have examined the relevance of volumes of LIPs regarded within the entire province. However, a different form of comparing volumes of igneous rock is to consider the products of single eruptive events. In this comparison it might be useful to use data from the representative provinces of each end member as defined above (i.e., CFBs and modern volcanic provinces) rather than attempting a comparison drawing examples from the whole universe of LIPs and non-LIPs. In the case of a typical modern volcanic province the volume of extruded magma during one single extrusive event can be considered to be ~<0.5 km^3, having as a typical upper limit that portion extruded during an eruption with a volcanic explosivity index (VEI) of ~5. Nevertheless, it is noted that although very seldom, the volumes extruded in this type of provinces might achieve much larger values, as exemplified by the 15 km^3 associated with the Laki eruption of 1783 (Pyle, 2000). For CFBs, the volumes involved in a single extrusive event are also variable, as revealed by the data compiled by Tolan et al. (1989) for the Columbia River Basalt Province. These data, shown in Figure 2, indicate that a large proportion of single eruptive events involved relatively small volumes of magma, having a mean value of ~60 km^3. More voluminous events (~750 km^3) have also been documented, although these are a very reduced proportion in terms of their frequency of occurrence. Finally, the data in this figure show that a still lower proportion of events might have record values of volumes exceeding 2000 km^3. Thus, even when only in a rather approximate form, a difference of two to three orders of magnitude might seem to distinguish both provinces at the scale of the volume of individual eruptions. As will be argued below, this form of comparing sizes of different types of volcano-magmatic activity contains important clues from a genetic point of view that are not easy to identify when comparing total volumes of erupted products in the various provinces, as has been commonly done in the analysis of LIPs until now.

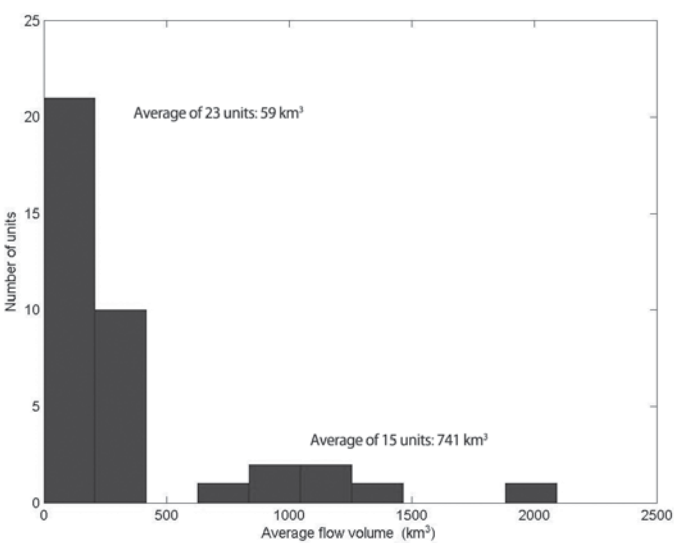

Figure 2. Histogram showing the volumes of individual lava flows in the Columbia River Basalt Province (data from Tolan et al., 1989).

Eruption Rates, Magma Production Rates, and Magma Volumes

Another parameter that has been suggested to be critical for the correct definition of a LIP is the rapidity of emplacement of at least a significant part of the province. Although the potential for committing a fallacy by including the intrusive part in the comparison of CFBs and modern volcanic provinces was pointed out earlier, it is necessary to examine in more detail the form in which eruption rates, magma production rates, and magma volumes relate to each other in order to better appreciate the extent to which these parameters might influence the definition of a LIP, promoting a bias in the analysis of observations. First it is noted that unlike the volumes of erupted products and the associated extrusion rates, the volumes of magma and magma production rates in volcanic provinces cannot be measured directly. The fact that we deal with factors requiring additional assumptions for their estimation increases the risk of committing a fallacy because the number of premises is increased as well as the chain of associated syllogisms. Furthermore, although it might be tempting to consider the calculated average extrusion rate as directly representative of the magma production rate of a volcanic province, it is important to be aware that such an association is incorrect and might favor the unnoticed introduction of a strong bias in the form in which we envisage both types of activity. This is better appreciated if we consider the diagram of Figure 3. In this figure the volume of melt stored in a region of partial melting (RPM) is plotted as a function of time for several situations. A constant rate of magma production without any event of extrusion results in the discontinuous line drawn from t0 to t6, whose slope is the specified rate of magma production in this case. If periods of melt tapping are allowed to exist while magma production continues to take place at the same rate, the amount of melt available at any time in the RPM will no longer be indicated by the discontinuous line, but rather it will be indicated by the solid line with different slopes. The changes in the slope of this line found in the intervals t1–t2, t3–t4, and t5–t6 are associated with tapping events E1 to E3, respectively, displayed at the lower part of the diagram. E1 is a tapping event that has a rate equal to the magma production rate. Consequently, during the time interval t1–t2 the volume of melt produced in the RPM is effectively canceled out by the volume of melt tapped out, therefore resulting in the lack of melt accumulation in the RPM during this time. Thus a tapping event like this one might yield the impression of a period of zero magma production. As the tapping event comes to an end at t2, the curve of melt volume resumes its previous trend, yet the volume of magma

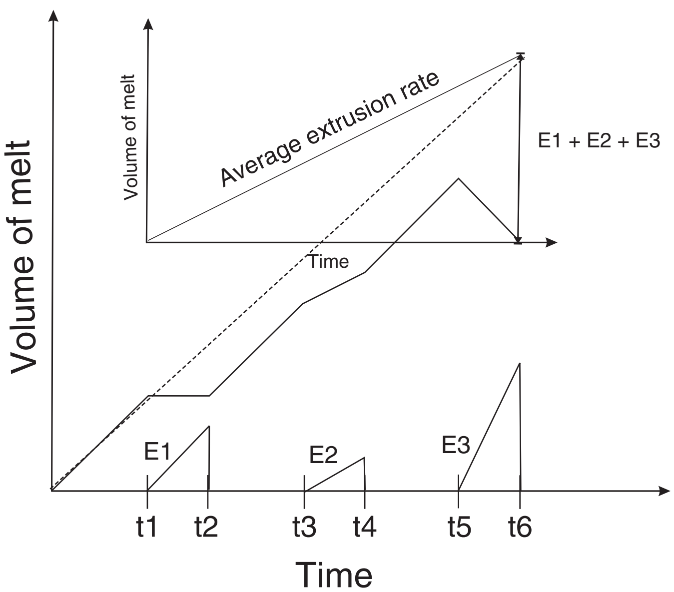

Figure 3. Plot showing the general relationship between magma production rate, volume of magma stored in a given region of partial melt (RPM), and magma tapping events. See text for details.

in the region of melting is smaller than it would have been if no tapping event had occurred. A new tapping event initiating at t3 again influences the slope of the curve representing the melt volume accumulated in the RPM. For illustrative purposes, in this case the tapping rate is set to be smaller than the magma production rate. The disparity between magma production and magma tapping rates causes the magma volume in the region of partial melt to keep increasing between times t3 and t4, despite the occurrence of the tapping event. Again, it is remarked that in this case the real magma production rate was actually the same as it was before the occurrence of E2, yet the apparent magma production rate obtained from the slope of the curve marking the melt accumulation is smaller. Once again, as the tapping event comes to an end, the slope of the curve representing the amount of melt in the zone of storage resumes its original value. The tapping event starting at t5 has a tapping rate faster than the magma production rate. During the interval t5–t6 the melt content in the region of partial melt will decrease, even if during this time the rate of melt production remains unaltered. Thus, even when physically there is some melt being produced during this time, the magma production rate would seem to be negative because the total melt volume stored within this region does not increase but effectively decreases with time. Evidently a tapping event like this one cannot continue indefinitely, because the melt in the RPM will eventually be exhausted. In consequence, this type of tapping event is limited by the amount of magma previously accumulated in the RPM rather than by the magma production rate itself.

Although highly schematic, the various situations depicted in Figure 3 illustrate several important aspects concerning the relation between magma production rate, magma tapping or extrusion rate, and magma volumes. In particular, this diagram shows the dangers of estimating a magma production rate with data that strictly belong to magma extrusion rates, even if all of the magma extruded from the region of melting was accurately measured at the surface. First, the diagram shows that a variety of extrusion rates can be supported by a single value of the magma production rate. Second, extrusion rates faster than production rates are sustainable only if there is a time gap between the moment of initiation of magma production and the initiation of the extrusive event; the larger this gap, the longer the interval with a high extrusion rate can be sustained before exhausting the melt available for extrusion. An alternative form to visualizing this is that a situation in which magma is extracted from the RPM as soon as it is formed is not capable of sustaining a large extrusion rate. Third, although the cumulative extruded volumes can be used to calculate something that could be interpreted as an average magma production rate (light axes at the top of the figure), the slope of this curve does not necessarily represent the actual magma production rate, because in fact we ignore the exact moment when magma started to be produced (i.e., t0 is not equal to t1, the time of the initiation of the first eruptive event). Fourth, the total volume existing at a given time at depth cannot be determined from information concerning the volume of erupted products. Fifth, even very slow production rates may sustain large rates of extrusion, given the adequate time for magma accumulation. Sixth, if we disregard the presence of the extrusive events and somehow are able to examine the solid line of the diagram, we would conclude that the magma production rate actually changed at t1, t2, t3, etc., because these points mark a shift in the slope of the melt volume content as a function of time, when in fact the figure was constructed by having a fixed value of magma production rate throughout all of the interval. Thus, this figure shows that even if we know accurately a magma extrusion rate we can say very little concerning the actual magma production rate at a given time. As a general corollary, it should be clear that much care should be exerted when interpreting the actual observations in attempting to characterize any volcanic province. For this reason, it is concluded here that magma production rates are not a good source of unbiased information, and that therefore such a parameter should be avoided when attempting to compare the volcanic activities forming CFBs with those of modern volcanic provinces, or to assess any genetic models proposed for LIPs as a whole.

Figure 3 makes reference to an extrusion rate, which to some extent can be considered equivalent to the emplacement rate of an igneous province. In the strictest sense an "extrusion rate" as used in this figure, however, only concerns processes taking place in the region of genesis of the igneous products, but once the magma is extruded from such a region it may be emplaced as an igneous intrusion, or alternatively it may be emplaced as a volcanic rock. The ability to identify the conditions favoring either of these alternatives remains a challenging subject of modern volcanology and is beyond the scope of this chapter. Actually, taking into consideration the uncertainties concerning the real age of the events leading to the emplacement of many intrusive rocks, as exemplified by the Okavango dike swarm mentioned above, the results are extremely uncertain when trying to establish significant breakpoints in the estimation of rates of emplacement of any province formed predominantly by intrusive components unless they have been exposed by erosion and have been extensively dated (which is seldom the case). Although estimates of the age of eruptive products are also relatively insufficient for many volcanic provinces, some of the correlations made for these types of products are relatively more certain than for intrusive rocks. Nevertheless, we might assess the value of adopting rates of eruption as a parameter, entering the formal definition of a LIP and its probable use as a marker in a better classification scheme. Again, this can be better achieved by restricting ourselves to examining information gathered from CFBs and modern volcanic provinces rather than by using the whole universe of LIPs and non-LIPs.

Although some variations can be found when comparing results obtained from two different CFBs, the accuracy obtained in radiometric ages of many of these provinces leads us to consider that extrusion rates during the peak of activity might have exceeded 100 km^3/yr in some extreme cases (Coffin and Eldhom, 1994). The same information, however, indicates that extrusion rates of CFBs are more commonly in the range of half a cubic kilometer to a few tens of cubic kilometers per year (Coffin and Eldhom, 1994), i.e., at least one order of magnitude smaller than the "extreme" cases. On the other hand, the rate of extrusion of modern volcanic provinces typically quoted even for the most productive volcanoes such as Kilauea and Mauna Loa are <0.05 km^3/yr, (Lipman, 1995). However, as pointed out by Harris et al. (2007), the exact definition of an *effusive rate* has been a source of confusion until recently, and much variability can be found in the reported values from the use of various measurement methods (note that the problems described by Harris et al., 2007, are an example of fallacies of equivocation, because the same term is used in slightly different forms in each case). When proper attention is given to the various sources of potential errors (note that the list of factors suggested by these authors to eliminate the fallacies are but three different forms of saying that all terms used in the construction of the syllogism should have an uniform meaning), the data shown by these authors indicate that extrusion rates in modern volcanic provinces are in the range of 0.01–1 km^3/yr, which might be more than one order of magnitude larger than the limit associated with Kilauea. In any case, the relevant point in the present context is to show that if one decides to compare CFBs and modern-day provinces by using the high end of the reported range of extrusion rates for CFBs (>100 km^3/yr) and the lower end of the range of modern volcanic provinces (<0.05 km^3/yr), the difference between both types of volcanic activity is strikingly large. If attention is focused on the other extremes of both ranges (0.5 and 1 km^3/yr for CFBs and modern provinces, respectively), however, it turns out that the difference between both types of activity is not that large. Furthermore, it could be concluded in the latter case that extrusion rates in modern volcanic provinces are actually higher than in CFBs. Consequently, acceptance of the first comparison and denial of the second constitute a clear example of selective focusing on information that yields a result expected because of a previously accepted notion of an extraordinary character of CFBs.

Interestingly, it has been documented that the extrusion rate for the Laki eruption of 1783–1785 might have reached a peak value of 8.7×10^3 m^3/s (Thordarson and Self, 1993). This value is equivalent to >274 km^3/yr, which is comparable to even the highest extrusion rates calculated for the more productive LIPs. From a morphological point of view, it has been documented that although a large variety of volcanic styles and architectures are found in CFBs, pahoehoe flows are not uncommon in these provinces (Jerram and Widdowson, 2005). This morphological type of lava seems to be favored by eruption rates <100 m^3/s (Griffiths and Fink, 1992), which are equivalent to just 3.2 km^3/yr, corresponding to the higher end of the range commonly accepted to be typical of modern volcanic provinces. In addition, it has been documented that the rate of arrival of magma to the crust in CFBs may be similar to that documented to occur in mid-ocean spreading ridges when large enough time intervals are considered (Macdougall, 1988b; Thompson, 1977). Nevertheless, these results tend to be dismissed without further examination, despite the fact that all of them would seem to suggest that the case for extraordinarily large extrusion rates in CFBs might not be as clearly established as commonly assumed. This attitude toward such lines of evidence contradicting the expected result is typical of mythical thinking and provides further support to the conclusion that a selective comparison between CFBs and modern volcanic provinces has been made in many instances.

In any case, the analysis made until now in this section indicates that "rapidity of emplacement" is a parameter that can be understood in many different forms. Consequently, this parameter should be used with extreme caution when attempting to distinguish between LIPs and non-LIPs. Most importantly, "rapidity of emplacement" must not be confused with "fast magma production rate," because the processes described by both terms are entirely different. Furthermore, although available evidence might seem to support the notion that LIPs are characterized by having erupted relatively large volumes of magma in relatively small periods of time, the rate of eruption of individual events in these provinces might have been not much different than the rate of eruption of individual events in other provinces. Therefore, the term *rapidity of emplacement* should be used to describe the formation of the whole province without extrapolating this term to the description of individual events. Actually, recognizing the possibility of having relatively uniform rates of eruption across the universe of igneous provinces constitutes a clue that could be significant when attempting to interpret the available evidence in genetically oriented models, as will be shown in the last part of the paper.

Tectonic and Crustal Settings and Geochemical Composition of Rocks

Among the list of parameters that were considered critical for the definition of LIPs by Bryan and Ernst (2008), we find tectonic and crustal setting as well as the composition of rocks. According to these authors, in order to be considered a LIP an igneous province should have been formed remotely from contemporaneous plate boundaries, in stable crustal regions with long histories of no prior deformation or contractional deformation, or undergoing extension, and with an "intraplate" geochemical signature. As these authors noted, some LIPs seem to have been emplaced near the edges of Archean cratons (Anderson, 1994). Although these settings might be defined as "intraplate" because of the distance to contemporaneous plate margins, nonetheless such regions represent areas that were undergoing extension at the time of LIP emplacement. Determining whether such extension was the cause of the LIP or vice versa has been one of the most debated subjects of the LIP literature over the years (Sheth, 1999a, 1999b). Consequently, adopting an evidently ambiguous indicator as a crucial part of the definition undoubtedly favors the selective interpretation of observations. Furthermore, as also noted by Bryan and Ernst (2008), "an intraplate tectonic setting is particularly problematic for the Cenozoic LIPs of North America." Nevertheless they dismissed such peculiarity by invoking the distinctive characteristics of these provinces in terms of their extent (including both area and volume) and rapidity of eruption, among other things. In other words, the really critical observations in this case would not be the tectonic setting but actually the other parameters. Consequently, arguing that the intraplate character of a province is critical for its classification as a LIP is a contradictory statement.

Concerning the composition of the rocks that form a province, it is noted that although there is the common perception that LIPs are remarkably homogeneous, a detailed examination reveals significant compositional variation taking place both spatially and temporally (e.g., Jerram and Widdowson, 2005). Furthermore, associating a "distinctive intraplate geochemical signature" to any rock type might involve a significant amount of circular reasoning (Anderson, 2000), which is another type of fallacious argument. Consequently, if the definition of a LIP is to be useful for providing an unbiased grouping of common features that aim to identify real patterns or trends of a given property, it would seem better to avoid including the geochemistry of the rocks in a province as a critical argument.

Finally, concerning crustal setting, Bryan and Ernst (2008) noted that the difference between silicic and other continental mafic-dominated LIPs is thought to be the crustal setting. Although such a distinction might justify the broad and genetically unbiased scheme dividing provinces based on the dominant composition proposed by Sheth (2007), it is uncertain whether such a subdivision of provinces based on this parameter can provide clues concerning their genesis without entering discussions concerning whether there is a "distinctive geochemical signature" for each setting. For this reason, it is considered here that although inclusion of this parameter in a classification scheme of LIPs in the form made by Sheth (2007) is acceptable, assigning a higher hierarchical value to such a parameter might increase the chances of falling into the realm of mythical thinking. Consequently, it would seem better to exclude this parameter from the list of critical ones, at least until more information becomes available.

How Many Parameters Are Needed to Create a Significant Classification of LIPs?

The analysis made throughout this section indicates that distinction between the intrusive and the extrusive components of a LIP is an important criterion that can help us to avoid unwanted fallacies in the interpretation of other characteristics of these provinces. Consequently, the first hierarchical level of the classification scheme proposed by Sheth (2007) seems to be well justified. Two other parameters that need to be included in the definition of LIPs are size (whether area or volume) and time. These two parameters need to be considered simultaneously in order to make a useful discrimination between LIP and non-LIP types of volcano-magmatic activity, but the available evidence does not seem to grant their use as hierarchical markers that can be used to construct a classification scheme with finer subdivisions at this time. Furthermore, it was noted that "rapidity of emplacement" is a somewhat ambiguous term that can be interpreted in various alternative forms, each leading to different interpretations. The more useful of such interpretations is when this parameter is used to describe the time of emplacement of one extrusive province. Thus size, time of activity, and the resulting rate of emplacement might be useful at present only to broadly distinguish LIPs from non-LIPs. Nevertheless, noting that emplacement rates of individual eruptive events can be similar across many types of igneous provinces might provide important clues in a genetic context. These clues will be explored with more detail in the following sections. The other criteria analyzed (age, crustal and tectonic settings, and the dominant composition of the rocks in the province) do not seem to play a crucial role in such LIP versus non-LIP distinction. Actually, these parameters might favor the occurrence of fallacious arguments when analyzing the observations. Consequently, they are not considered critical elements of the definition of a LIP in this work.

AN ALTERNATIVE MECHANISM OF FORMATION OF CFBs: THE VOLCANIC SYSTEMS APPROACH

As mentioned in the introduction, the definition of a LIP advanced by Coffin and Eldhom (1992) was highly influential in determining the orientation of much of the research done on the subject for slightly more than 15 yr. Perhaps the aspect in which this influence has been stronger has been in the association of LIPs with events of extraordinary character, implying with this that such provinces should be created by mechanisms entirely different than those operating in the formation of other provinces without the LIP characteristics. In particular, essentially all of the current explanations for the genesis of these provinces have always invoked the need for the *rapid production of magma* to explain the formation of LIPs. In the analysis made in the previous section, however, it has been shown that this perception is incorrect to the extent that little can be said about magma production rates by observing magma extrusion rates (Fig. 3). Furthermore, it was shown that determining the volume of an intrusive suite and assuming that all of that volume was emplaced almost coevally also could be unjustified in some cases, and that such presumption could lead to a fallacious estimate of the relevant magma extrusion rate. In addition, the previous section also presented some evidence suggesting that the rates of extrusion of individual eruptive events could have been similar both in LIP and non-LIP provinces, further supporting a perception of a LIP as the result of essentially the same underlying principles that control the formation of other volcano-magmatic provinces with no-LIP characteristics, rather than being the product of extraordinary processes.

Conceiving LIPs as the result of similar mechanisms than those underlying the formation of non-LIPs is not the same as saying that there are no differences between those two types of volcano-magmatic provinces. Actually, much of the analysis of the previous section was oriented toward identifying some parameters that could help us to identify breakpoints in the continuum of volcano-magmatic provinces that could be used as guides when attempting to unravel the details leading to the genesis of all of those provinces. This approach seems to have been successful, because it allowed us to identify an important clue that seems to have been overlooked until now. This particular clue is the observation that eruption rates of individual events might have been similar across all the universe of volcano-magmatic provinces, despite the undeniable differences in the volumes extruded when time intervals between 1 and 5 Ma are considered. The model developed in this section was constructed with this important clue in mind.

Based on the relationship that should exist between magma production rate, magma volume stored within the region of melt genesis (hereafter referred to as the region of partial melt, or RPM), and the rate of extrusion schematically illustrated in Figure 3 and described with more detail above, it becomes evident that one possibility for explaining the formation of LIPs while having an extrusion rate not that different from the extrusion rate of non-LIPs is to have a large amount of magma stored at depth before the onset of an eruptive event. Thus a first task in developing the genetic model of this section is to establish the volumes of magma that can be available for tapping even in present-day conditions, and to establish whether such volumes of magma have the potential to feed one event of LIP characteristics. If this task is successful, then we can consider that probably the difference between the LIP and the non-LIP event would be that the mechanism of extrusion in the former allowed tapping a larger volume of magma, whereas in the latter the tapped volume is smaller. The second task therefore would be to explore mechanisms that could explain such differences.

How Much Magma Can Be Stored under the Surface?

Based on seismic observation beneath zones of active volcanism, for example, underneath Italy or Japan, zones of low velocity likely to contain some melt can be defined as having lateral extensions of 200 km or even larger, and thicknesses in the order of 50–100 km (Nakajima et al., 2005; Panza et al., 2003). If we consider that melt proportions can be somewhere between 2% and 15% (e.g., Sato and Ryan, 1994) it can be concluded that typical volumes of present-day RPMs are of the order of 10^6 km^3. A similar volume would be found if, instead of considering the regional zones of seismic attenuation mentioned above, we focus on the zones of seismic attenuation beneath Hawaiian and Icelandic volcanoes, because the latter seem to have a diameter of ~130–175 km and equivalent thicknesses (e.g., Ryan, 1990; Watson and McKenzie, 1991). Even still larger dimensions of low-velocity zones could be obtained if attention was focused under other tectonically active regions (e.g., Grand, 1994). Although constraining the dimensions of an RPM and the amount of melt contained in it can be done within a certain degree of confidence from observation of seismic-wave propagation times for present-day volcanic provinces, this source of information should not be used to attempt a direct comparison with the volume of melt produced under any CFB during its formation because such comparison leads to the fallacy of four terms. Nevertheless, the danger of committing such a fallacy can be eliminated if magma volumes are estimated by using exactly the same method and identical premises in both CFBs and modern volcanic provinces alike. Consequently, at least from the point of view of logical reasoning, such comparison would be more acceptable than comparing magma volumes estimated by using different premises, even if the magma volumes estimated by our single method are not quite accurate.

As it turns out, it has been suggested that measurement of the content of rare earth elements (REE) in the volcanic products can be inverted to constrain the distribution of melt as a function of depth (White and McKenzie, 1995). Although the melt distributions obtained with this method are undoubtedly influenced by the choice of the mantle source and the form in which the melting process and transport mechanisms are envisaged, the key aspect of this method is that it can be used by assuming exactly the same premises to constrain melt distribution beneath any volcanic province, modern or past, and therefore the results of such inferences obtained from CFBs and modern volcanic provinces can be compared directly with each other without resulting in a fallacy associated with the existence of different premises. In other words, regardless of the limitations inherent in this particular method, the melt distributions obtained by using REE of the erupted products has the advantage of providing some information that can be used to make a quantitative comparison between modern and CFB provinces, thus avoiding any biases that could favor a particular genetic interpretation, even if only subtly.

Figure 4 shows a series of melt distributions as a function of depth obtained from various modern provinces that include mid-ocean-ridge (MOR) and intraplate settings (White and McKenzie, 1995). Although according to this method the magmas produced in MORs tend to be formed at shallower depths than the intraplate magmas (Fig. 4A), the same data also reveal that the

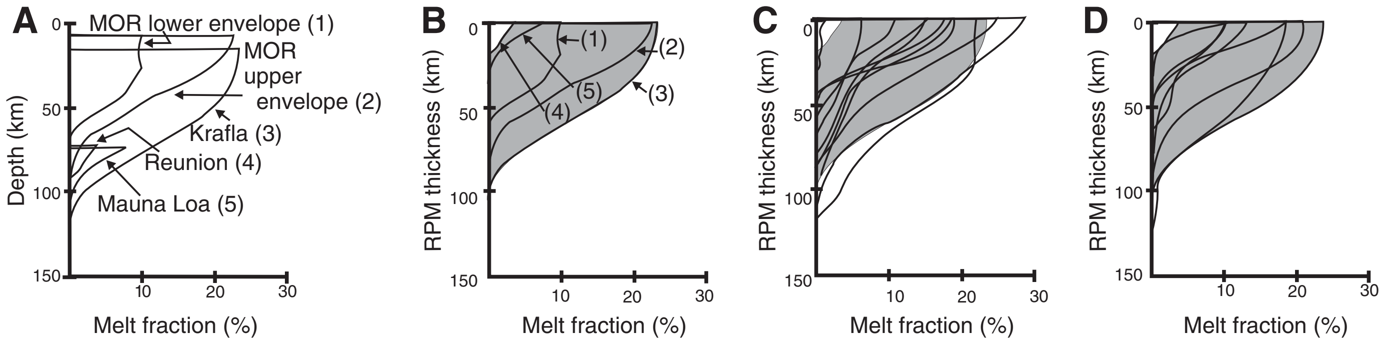

Figure 4. Melt distributions with depth inferred from rare earth element (REE) inversions made by White and McKenzie (1995). (A) Curves obtained from basalts from mid-ocean ridges (MORs) and intraplate volcanoes. For mid-ocean-ridge basalts (MORBs) only the upper and lower limits of the melt distributions are shown for clarity. The same curves of diagram A are shown in B, having their upper limit displaced, therefore emphasizing the dimensions of the various regions rather than the depths at which they are generated. The shadowed area is an envelope that is used for comparative purposes in C and D. (C) Comparison of the melt distributions of diagram B with those inferred from Siberian, Keweenawan, North Atlantic, Columbia River, Deccan, Etendeka, Paraná-Karoo, and Coppermine River basalts (representative examples from White and McKenzie, 1995). (D) Comparison of the melt distributions of diagram B with those inferred from various basalts and dikes ranging in age from 0 to 2700 m.y. RPM—region of partial melt.

inferred melt distributions define a layer of some 100 km regardless of tectonic setting (Fig. 4B). Such melt distribution curves imply that the average melt fractions (estimated as the areas to the left of the inferred melt distribution curves normalized by the thickness of the corresponding RPMs) are also relatively constant (~5%–15%) irrespective of the tectonic scenario.

In Figure 4C the depth-shifted melt distributions of several major CFBs, also obtained by using the REE method by White and McKenzie (1995), are compared with the depth-shifted distributions obtained from modern volcanic provinces. In this figure it can be observed that from the nearly 30 CFB-related melt distributions examined by White and McKenzie (1995) (only a representative selection was included in the figure for clarity), neither the associated thickness of the RPM nor the average melt content are extraordinarily larger than the upper limits of the present-day volcanically active provinces. Furthermore, it is found that even one of the CFBs, the Keweenawan Province, yields a melt distribution that is smaller than the lower limit of the modern volcanic provinces represented by the Reunion Volcano in these diagrams. Actually, the similarities between the melt distributions inferred to have existed under major CFB provinces can also be extended to include other minor basalt provinces, including lavas and dikes ranging in age from 0 to 2700 Ma, as shown in Figure 4D. In summary, the melt distributions inferred from REE of erupted products of both CFBs and modern volcanic provinces seem to have been produced in regions of partial melt with (1) a typical thickness of ~100 km, and (2) average melt contents between 5% and 15%. Such similarities further reinforce the possibility of having similar processes controlling volcano-magmatic activity in both types of provinces, rather than having extraordinary processes behind the formation of LIPs.

Although the REE method suggests that the percentage of melt might change drastically with depth, a conservative estimate of the typical melt content valid for all types of volcanic provinces can be considered to be 5% melt constituted uniformly in a layer 100 km thick. It is remarked that this approximation concerning the average melt content in a column of mantle rock is made only for the purposes of estimating representative volumes and does not provide an accurate representation of the rare element content of any particular eruption. Similarly, it is remarked that the diagrams of Figure 4 suggest that melt contents might exceed 20% locally (such large proportions of melt are restricted to layers ~40 km thick). Thus, a conservative estimate of the dimensions of a present-day RPM, and of its melt content, are used in the following calculations and are depicted in Figure 5.

A simple calculation based on the constraints shown in Figure 5 indicates that a regional zone of partial melting found beneath present-day active volcanic provinces conservatively might contain ~10^5 km^3 of melt. Using a different set of arguments, the presence of large volumes of melt trapped beneath the surface at a global scale was also noted by Schmeling (2000). Therefore, the argument of such melt storage is not entirely novel in the literature, although its significance in the genesis of LIPs has been overlooked until now. In particular, it has not been noted previously that the volume of melt trapped under certain regions of the Earth might be much larger than the volume of melt extruded in any single eruptive event either in a modern volcanic province or a CFB province. Actually, a typical eruption in a modern volcanic province extrudes a volume of magma of ~0.5 km^3, whereas that of a single extrusion event in the history of a CFB province might reach 600 km^3. It is observed that a typical present-day RPM conservatively can sustain >300 eruptive events of CFB proportions, and an incredibly larger number of eruptive events of modern volcanic province proportions. Remarkably, the number of eruptions that could be fed from the already existing magma in a present-day RPM would account for nearly 20% of the total volume of extrusive products characteristic of CFB provinces without requiring the production of a

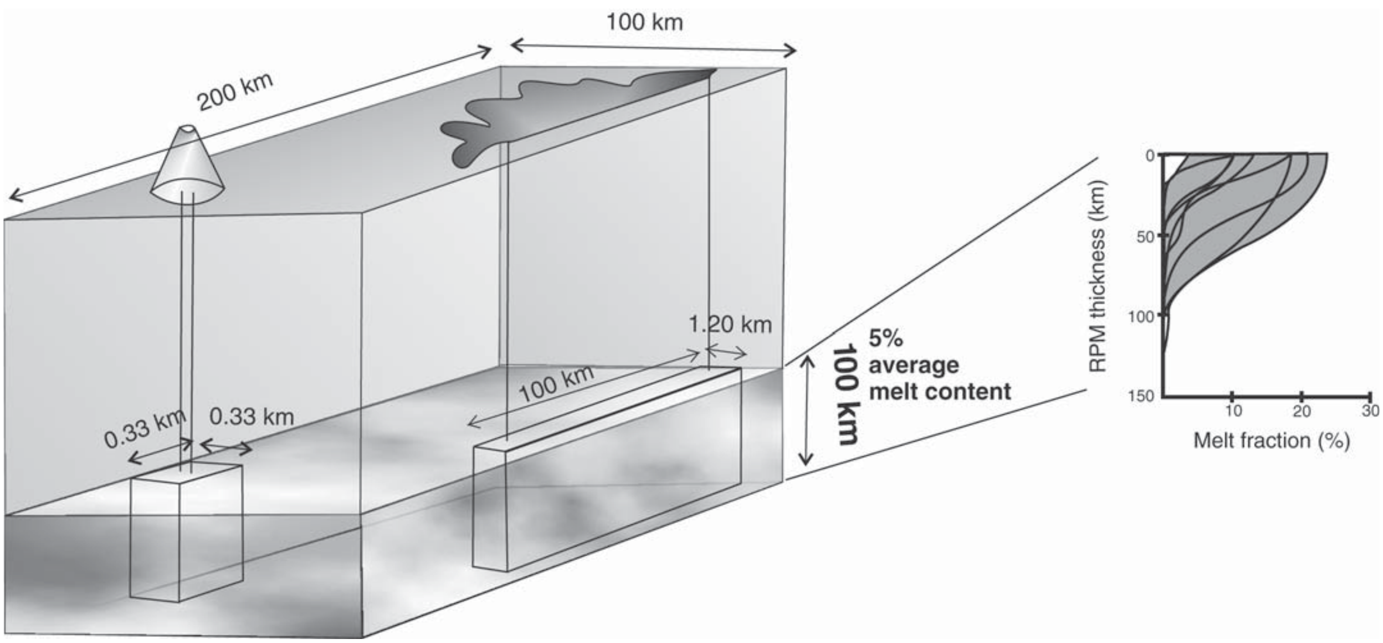

Figure 5. Schematic relation of the volumes of mantle rock required to produce the erupted volumes in one eruption of a normal volcano (cone) and a CFB-forming province (elongated feature on the surface). The average melt content of the melt zone is consistent with the constraints imposed by melt distributions inferred from rare earth element (REE) compositions of erupted products shown in Figure 4 and reproduced at the right of the diagram; the dimensions of the regional low velocity zone are shown in the upper part of the box. RPM—region of partial melt.

single additional drop of melt. Although this volume of melt is not enough to explain the formation of any CFB province, the percentage of volume obtained in this form is not at all negligible. Furthermore, if the calculations were made by using the limit of 15% melt (well within the limits allowed by observations), the volume of melt potentially existing at present beneath many RPMs would account for 60% of the volume of a typical LIP. Thus, it is clear that present-day volumes of magma trapped at depth have the potential to feed one event of LIP characteristics.

Mechanisms of Tapping Magma from the Deepest RPMs

If a typical present-day RPM has the potential to feed an eruptive event as volumetrically impressive as any event found in any CFB province, we need to explain (1) why is all of that melt not being extruded at present, and (2) what parameters could be responsible for the occurrence of such an efficient extraction process from time to time, so that a CFB province is formed. A clue to answer both of these questions can be found by focusing on key aspects that control the processes of melt extraction from its source and transport it to the surface as portrayed in the general model of volcanism advanced by Cañón-Tapia and Walker (2004). These authors argued that the conditions of melt interconnectivity within a region of partial melting (rather than the total amount of melt) and the overall state of stress of the rock overlying it are the two most important factors that control the style of volcanic activity at the surface. According to their model, a large degree of melt interconnectivity and a decrease in the magnitude of the compressive stress on the overlying rock promote a more efficient form of magma extraction than could take place if interconnectivity is small, or compressive stress is large. Therefore, their model predicts that an eruptive event capable of tapping large volumes of magma, like those forming CFB provinces, can take place only if an adequate combination of interconnectivity and stress is reached.

Establishing quantitative limits for the first of these two parameters (melt interconnectivity) depends on the knowledge of the dihedral angles relevant to all of the mineral phases that are likely to be found in a given parcel of mantle rock. As a first approximation, dihedral angles <60° yield a fully interconnected network of fluid-filled channels even for very small amounts of melt present in the system (Laporte and Provost, 2000b; von Bargen and Waff, 1986). Such an approximation, however, is valid only for monomineralic rocks with single valued solid-solid and solid-liquid interfacial energies. In natural systems some complications arise because these rocks are polymineralic, and in addition the solid-melt interfacial energies depend on the orientation of the interface relative to the crystalline lattice (Jurewicz and Jurewicz, 1986; Laporte and Provost, 2000a). Consequently, real magmatic systems might display a connectivity threshold even if the dihedral angles are in the range of 30°–60°, and perhaps more importantly, might display a strong 3-D dependence that will influence the interconnectivity of melt at a local scale (Holness, 2005). In particular, the degree of interfacial anisotropy is more pronounced for olivine, amphibole, and clinopyroxene than for quartz and plagioclase (Holness, 2005; Laporte and Provost, 2000b). Consequently, a marked influence of interconnectivity might be expected for mantle rocks, especially in those in which a preferred mineral orientation is suspected. Unfortunately, some discrepancies are found in the published dihedral angles of various mineral species (Holness, 2005), and therefore development of a truly quantitative model of a volcanic system that incorporates this parameter might be premature. Nevertheless, based on the current knowledge concerning the characteristics of the dihedral angle of the most relevant mineral species, some key premises seem to be justified. For instance, interconnectivity of the melt is unlikely to be uniform across the whole extension of an RPM. Variations are expected at various scales, depending on the distribution and orientation of the various mineral phases present in the original rock. Such variations in the interconnectivity of melt within an RPM are likely to influence the percolation process, and might prove to be an important factor that controls the total amount of melt that is capable of leaving an RPM upon onset of a tapping event. In particular, those regions of the mantle that had undergone the continuous influence of tectonic forces, and consequently that are likely to have developed a well-defined mineral fabric, are more likely to have an anisotropic interconnectivity than regions of the mantle exempt from those tectonic influences. For this reason, it is convenient to examine the consequences of such anisotropy.

Melt interconnectivity in a vertical direction contributes to the development of the conditions necessary for the initiation of a magma tapping event that allows the rapid raising of magma through a hydraulic fracture. The difference in density between the solid and liquid phases within the RPM produces a vertical force that promotes the ascent of the less dense phase (commonly the liquid) and the descent of the denser phase (usually the solid). Part of this process is better described as a two-phase flow taking place in a porous medium, where the viscosity of the liquid and the permeability (which in a sense is a measure of the interconnectivity of melt in the vertical direction) are important parameters (e.g., Fowler, 1990; McKenzie, 1985; Scott and Stevenson, 1986; Stevenson and Scott, 1991). Superimposed on the process of porous flow, the difference in density between solid and liquid promotes a different process that also contributes to the vertical migration of the liquid phase, and actually provides the only mechanism through which the melted rock is capable of reaching the surface without cooling back down to the solid state while traveling across a region of the mantle where melt is not thermodynamically favored. In simple terms, in addition to defining the vertical gradient of pressure controlling the porous flow (and compaction of the solid matrix), the lower density of the liquid phase within the RPM also is responsible for the onset of a vertical component of stress that is exerted upon the solid rock that forms the upper boundary of that region. The magnitude of that vertical stress is proportional to the difference in density between solid and liquid phases multiplied by the vertical dimensions of a column of liquid formed below the boundary of the RPM (Cañón-Tapia, 2009). In turn, such a vertical component of stress induces a horizontal component of stress within the solid rock directly on top of the liquid column, the magnitude of which is controlled by the elastic properties of the solid (in particular, its Poisson's ratio) and by the loading conditions. Owing to the large confining pressures commonly found on top of the deepest RPMs, the solid rock undergoing the vertical component of stress induced by the column of liquid underneath it would be unable to deform, and therefore the induced horizontal stress would be of a tensional nature with a much smaller magnitude than the vertical component inducing it (Turcotte and Schubert, 1982). Importantly, the total horizontal stress underwent by the solid rock at the boundary of the RPM is not necessarily tensional, because the induced tensional component is counterbalanced by the horizontal compression exerted by the adjacent rock. Nevertheless, the presence of a vertical column of liquid within the RPM induces a stress differential in the solid rock at its upper boundary, leading to a situation of metastable equilibrium (Cañón-Tapia, 2007). Such a metastable situation will be sustained as long as the vertical extension of the column of liquid does not reach a threshold value, determined by the combined effects of the confining pressure and the tensile strength of the overlying solid rock. When such a threshold value is reached, the solid boundary of the RPM starts to fracture, opening a space that will be filled in by the liquid in the column beneath it. This moment represents the initiation of a magma tapping event through the formation of a hydraulic fracture that eventually might lead to the emplacement of a dike or another tabular intrusion, depending on the orientation of the plane of the fracture relative to the vertical.

Although establishing the conditions of melt interconnectivity and internal stress required to initiate a magma tapping event is an important step toward a better understanding of the reasons for having volcanism akin to either CFB or modern volcanic provinces, we still need to examine the processes taking place after the initiation of such a tapping event in order to identify the most important parameters controlling the different styles of volcanic activity typical of both types of provinces. Following the general framework outlined by Cañón-Tapia and Walker (2004) we can identify three different processes starting to take place simultaneously after the onset of the fracturing event that initiates the tapping of magma. One of these processes concerns the movement of magma in the vicinity of the vertical column of magma that induced the first fracturing, and the other two concern the propagation of the fracture front away from the site of nucleation. The first of these processes is determined by the interconnectivity of melt in a horizontal direction, as such interconnectivity determines the effective permeability of the RPM, and in combination with the relevant viscosity and the resulting stress gradient determine how much melt can enter the forming fracture, and consequently the duration and size of the tapping event. The second of these processes is important, because the propagation of the fracture away from the RPM determines the fate of the moving magma either as an igneous intrusion stalled at a shallower level, or alternatively as an eruptive event that leads the magma from the RPM all the way to the surface of the planet. The third of these processes (the propagation of the fracture front away from the site of nucleation but remaining along the boundary of the RPM) exerts a strong influence in controlling the pressure conditions within the RPM, therefore determining the total volume of magma that can be extruded from the RPM in a given tapping event, and the dimensions of the igneous intrusion, which indirectly also determine whether a magma tapping event ends up as an eruptive event or not (Cañón-Tapia and Merle, 2006).

To illustrate the form in which all of the various parameters examined so far might lead to the formation of a CFB or to a volcanic province more akin to modern examples, it is convenient to present two examples that include some numerical calculations. The first step is to constrain the minimum amounts of magma required to initiate a magma tapping event. It can be shown that in general critical heights <100 km will suffice to initiate an event of hydraulic fracturing in most conditions of geological relevance (Cañón-Tapia, 2008, 2009). Actually, as shown in Figure 6, a column of liquid of 50 km should be able to induce an excess vertical component of stress exceeding 50 MPa almost independently of the depth at which the base of such a column of liquid is found. Owing to the fact that typical mantle material has a Poisson's ratio of ~0.25, according to linear elastic theory, such a vertical stress should induce a tensile horizontal stress >10 MPa in solid rock at the upper boundary of an RPM. Importantly, 10 MPa marks the typical tensile strength determined for most rock types

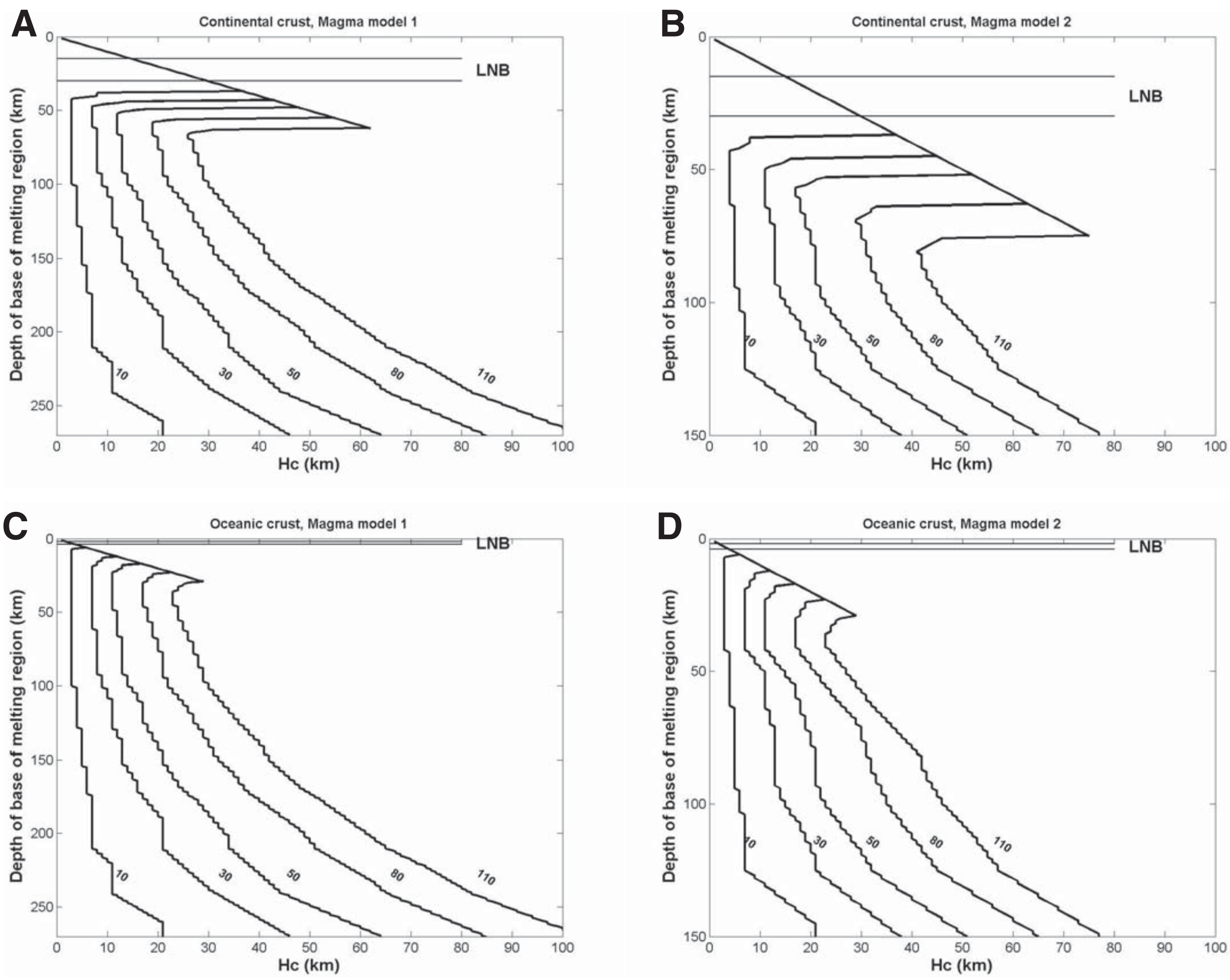

Figure 6. Diagrams showing the size of the critical heights of a column of magma required to exert a vertical stress of the amount shown numerically by each curve, as a function of the depth of the base of such a column. The two magma models correspond to densities of melt calculated from the two examples of Stolper et al. (1981), each applied considering a column of solid rock having either oceanic or continental crust above the melt (from Cañón-Tapia, 2009). LNB—level of neutral buoyancy; Hc—critical height.

(Jaeger and Cook, 1976; Turcotte and Schubert, 1982). Consequently, a minimum requirement for initiating a magma tapping event is to have uninterrupted continuity of the liquid phase in the vertical direction for at least 50 km below the upper boundary of any RPM. Importantly, this constraint in the vertical dimensions of the column of magma is independent of the total amount of magma present in the RPM. For example, 500 m^3 of magma would suffice to initiate a magma tapping event if all of this volume were contained in a narrow conduit having a horizontal surface area of 0.01 m^2 and a vertical extent of 50 km, but in contrast, even 10 km^3 of magma would not suffice to initiate a magma tapping event if this volume of magma were contained in a large pool with 1 km^2 of horizontal area and a depth of only 10 km. The difference between these two examples is that for the former the column of magma creates a sufficient vertical stress to induce a horizontal tensional stress slightly larger than tensile strengths of most rock, whereas a larger pool of magma only results in a vertical excess stress of <10 MPa in most areas of geological interest, which would induce a tensional horizontal stress <3MPa in most cases, being less than the typical tensile strength of most rocks. Consequently, whereas the minimum volume of magma required to initiate a magma tapping event can be constrained using hydrostatic principles, there are no constraints that can be imposed on the maximum volumes involved in any tapping event other than perhaps specifying a maximum melt fraction for a given volume of mantle rock. Nevertheless, the exact distribution of this liquid within the RPM, controlled to some extent by the degree of horizontal interconnectivity, plays an important role in controlling the outcome of any tapping event, as discussed next.

A detailed description of possible outcomes of magma tapping events is outside the scope of the present chapter, but a general examination of two extreme situations should serve to illustrate the two extreme behaviors likely to result from an eruptive event akin to either a CFB or a modern volcanic province. For the two situations to be considered here, I assume that an RPM has lateral dimensions of 100 km × 100 km, a thickness of 50 km, and contains an average 1% melt (Fig. 5). As discussed above, these dimensions are well within the range deduced from both seismic signals and the REE content of the erupted products. From the total 500 km^3 of melt contained in the RPMs of each of the two examples, I focus attention on only 120 km^3 of liquid for simplicity. The same calculations can be done by using any other prescribed volume considered representative of the volume of individual eruptive events in CFBs without altering the result.

In one case, I consider an RPM having 120 km^3 of melt concentrated in two zones, each having 20 km × 1.2 km in the horizontal direction and 50 km in the vertical direction. The other example RPM has the same 120 km^3 of magma dispersed in 445 parcels of rock, each having horizontal dimensions of 0.33 × 0.33 km and with a vertical extension of 50 km. In both RPMs the remaining 380 km^3 of magma is considered to be uniformly distributed in the remaining 47,600 km^3 of mantle rock that defines the RPM. In other words, although the average melt content of each of the example RPMs is equal to 1%, the melt is concentrated in regions having 5% melt concentration and in regions having 0.8% melt, and those regions of higher melt concentration are distributed differently in each case.

Considering that the liquid is assumed to be distributed in volumes of mantle having the same vertical extension in both scenarios, the probability of having an uninterrupted column of magma capable of initiating a tapping event will be slightly higher in the volumes of mantle with higher melt concentrations, and therefore I focus on these zones. Where the zones of melt concentration are small, the volume of magma likely to be tapped by any propagating fracture will be restricted to the melt contained by the region directly surrounding the first fracture. This volume of melt is on the order of 0.27 km^3, corresponding to the volume of magma erupted in a volcanic eruption of a modern volcanic province. Actually, even if the fracture extended beyond the horizontal limits of one region of high melt concentration, the lateral extent of this fracture would add only very small volumes of melt to the tapping event because it would be piercing the surface of a volume of mantle with very small amounts of magma. Because the mechanism driving the propagation of the fracture front both in the horizontal and vertical dimensions depends on the volume of melt that is entering the newly formed crack (Cañón-Tapia and Merle, 2006), it is unlikely that one single fracture could propagate long enough along the surface of the RPM to intersect a second region of high melt concentration. Consequently, any tapping event taking place in this RPM is likely to involve very small volumes of magma at a time.

For the RPM having only two zones of relatively high melt concentration, any fracture that starts tapping magma is likely to involve up to 60 km^3 of liquid in one single event. The tapping of such a volume of magma could take place through a conduit of relatively limited horizontal extension, provided that the horizontal interconnectivity is sustained at all times within the zone of melt concentration, or alternatively it might involve a fracture with a horizontal extension of a few kilometers in one direction if the fracture allowing the mobilization of magma grows along the top of the RPM parallel to the longest horizontal extension of the zone of melt concentration. In either case, a volume of liquid corresponding to one volcanic eruption of a CFB province will have been tapped in one single event with relative ease.

Despite the fact that both examples shown in Figure 5 are highly artificial, they serve to illustrate the type of processes that might take place during a real tapping event as well as the roles of the various parameters involved. At a minimum, the examples provided here illustrate the fact that, at least in theory, two contrastingly different eruptive behaviors can originate from RPMs that contain identical melt fractions and total amounts of magma. Variations of the internal distribution of such melt, rather than of actual melt content, might suffice to explain the different eruptive behaviors typical of CFBs and modern volcanic provinces. In addition, these examples suggest that such variations in the internal concentration of melt within a given RPM can also explain why a present-day RPM might not be feeding a CFB, giving some clues concerning the processes that could lead to the achievement of the conditions required for the creation of such a province from time to time.

Geological Context

According to the previous findings, in order to explain with more detail we need to focus attention in the processes that could lead (1) to the uneven accumulation of melt within an RPM, and (2) to favor the rupture of the upper boundary of the RPM for long distances. One possible alternative to satisfy both conditions certainly is related to the action of mantle plumes. These explanations, however, are not unique. A very reasonable scenario that could lead to the same result can be derived from the observed coincidence between CFBs and zones of lithospheric discontinuity, such as craton boundaries or ancient shear zones (Anderson, 1994). King and Anderson (1995) showed that a marked difference in lithospheric thickness can induce a small-scale mantle convection capable of producing an RPM with zones of melt concentration exceeding 2%. According to their results, the more marked the difference in lithospheric thickness, the more extensive the zones of melt accumulation, and consequently the larger the volume of melt that could be stored within the RPM. Furthermore, their model results show that the regional zone of partial melt would not have a homogeneous distribution of melt, but rather that melt will tend to be concentrated in two or three broad zones (depending on the actual difference in lithospheric thickness), separated by zones with a lower concentration of melt. Although those results evidently represent a highly idealized situation, the general feature of an extensive regional zone of partial melt with zones of melt concentration separated by zones of lower melt concentration is precisely one of the conditions required by the model developed in the previous section.

Actually, the same mantle geometry explored by King and Anderson (1995) has characteristics that favor the rupture of the RPM for long distances. Analogue models of overpressured magma chambers with different geometries (Cañón-Tapia and Merle, 2006) have shown very clearly that rupture events are more likely to nucleate in those places on the surface of the chamber with a larger angularity and that such angularities guide the propagation of the fracture along the surface of the chamber once it had started to form. For this reason it seems reasonable to consider that the angularities associated with the step geometry resulting from a marked difference in lithospheric thickness could serve as nucleation sites for the occurrence of fractures that allowed the rapid tapping of the magma formed as the result of the small convection cells. Finally, it is also noted that the vertical dimensions of the zones of higher melt concentration in the numerical results of King and Anderson (1995), are well within the limits of the 100 km required to initiate a tapping event based solely on the difference in density of the magma with its surroundings, as shown above. Consequently, unlike the conclusions reached by King and Anderson (1995), it is considered here that a mechanism for the melt to reach the surface does indeed exist, even without having an external source of lithospheric extension. In any case, if such an extension occurred as the result of other larger scale processes, the initiation of the fracturing event will be facilitated, and actually could be achieved by columns of magma of <50 km height.

In summary, the model presented in this section envisages CFBs and modern volcanic provinces as the result of the same basic processes. The most striking difference between both provinces—the large difference in the typical volumes of the erupted products—can be explained as the result of some special conditions, which nonetheless have nothing extraordinary or anomalous. For this reason, it is concluded here that CFB provinces, and by extension, LIPs, are not necessarily the manifestations of a significantly different mode of operation of the Earth but rather one extreme of a spectrum of possible outcomes that can take place during the combination of the various parameters involved in controlling volcanic activity. Relatively small variations in some of these parameters are likely to have been responsible for the wide diversity of features displayed by LIPs and modern volcanic provinces alike, including those related to chemical composition.

DISCUSSION

Until now the debate concerning the origin of LIPs has centered on the mechanisms required to produce large amounts of magma in a relatively short time interval. For the various reasons discussed in the first sections of this chapter, this feature of LIPs is prone to lead to fallacious judgments, and consequently it promotes mythical thinking. Perhaps the most notable exceptions to this trend have been the works by Silver et al. (2006), who envisaged CFBs as "drainage events" rather than "melting events," and that by Jerram and Widdowson (2005), who focused more on the internal facies architecture and structure of CFBs, pointing out that these provinces are not as uniform as commonly portrayed.

In a way, this work has the same general conceptual framework that is found behind those two works because more attention is given to the extrusive expression of those provinces, and such expression is explained more in terms of a mechanism that allows a more efficient drainage of magma, avoiding any dependence on magma production rates. Nevertheless, important differences exist between the model proposed by Silver et al. (2006) and the model advanced here, which deserve further examination. First, it is noted that the model developed here is a particular example of the model of volcanism developed by Cañón-Tapia and Walker (2004), which is of a more general nature because it concerns all expressions of volcanism and is not constrained exclusively to explain the occurrence of cratonic flood basalts. Second, the model developed here does not make a distinction between the stage of formation and maintenance of a reservoir of magma, and the stage of drainage. Unlike the model developed by Silver et al. (2006), in the model developed here the region of partial melt is considered to be the natural expression of the tectonic evolution of the Earth because it simply marks the region where the pressure-temperature conditions of any parcel of mantle are such that they can sustain the melting of some of its constituent minerals. These conditions seem to be globally met below an average depth of 100 km under continents, coinciding in extent with the globally detected seismic discontinuity (Thybo and Perchuc, 1997), and therefore are not exclusive of the cratonic environments. Actually, the presence of those regions is responsible for the occurrence of volcanic activity around the world, whether related or not to a LIP. Third, Silver et al. (2006) divide the drainage stage in two separate and sequential substages. In the first of these substages, porous-flow migration of magma is envisaged to collect at depth, and in the second substage the collected magma ascends, forming dikes in the process. Such a distinction between the collection and drainage substages is not present in the model presented here because the initiation of a dike does not rely on the previous collection of magma in a pool on top of the RPM. Actually, in my model the porous flow of magma within an RPM can be envisaged as being responsible for the uneven lateral transport of the melt within the zone of partial melting, therefore defining the dimensions of the zone of influence of a tapping event, which explains the observed differences in the volumes of erupted products in the various provinces around the world. Consequently, unlike the model of Silver et al. (2006), the model developed above explicitly accounts for the episodic character of the many events that form a CFB province. As pointed out by Jerram and Widdowson (2005), such an episodic character is commonly neglected in calculating average rates of formation of CFB provinces, but it is marked by the fact that not all of the magma was extruded during one single eruptive event. Also, the model developed here can explain in simple terms the occurrence of an extended period (i.e., beyond the 1 m.y. limits) of volcanic activity at or very close to the areas of formation of CFBs, the occurrence of which is sometimes neglected in the interpretation of these provinces, as pointed out by Sheth (1999a).

Incidentally, the model developed here not only accommodates the occurrence of an extended period of volcanism beyond the peak of activity in forming a CFB, but it also serves to explain the occurrence of portrayed volcanism in other types of tectonic settings. For example, the same basic principles have been used to explain the occurrence of volcanism a long time after subduction had ended along the Peninsula of Baja California (Negrete-Aranda and Cañón-Tapia, 2008), and this might serve to explain the intermittent occurrence of volcanism in places like the Tibetan Plateau. In all of these instances the protracted period of volcanic activity represents the "normal" magmatic activity of the region being disrupted by a transient effect acting as stress concentrator. For the CFB, events such as a global plate reorganization might provide the required external disruption, but a more localized disruption might suffice to trigger a bout of volcanic activity. Such a renewed (or enhanced) bout of volcanism comes to an end once the effects of such a transient event on the Earth's surface wanes. In this context, it is also noted that the observations suggesting that deformation style, and hence stress distribution, varies progressively along the rift axis during rift propagation (van Wijk and Blackman, 2005). This should explain why CFBs, or LIPs, do not form an uninterrupted chain along places of continental breakout. Either the places where no volcanic activity of this type occurred were characterized by a smoother difference in lithospheric thickness across the rift, or the local stress distribution was not enough to allow the tapping of magma all the way to the surface. In particular, the latter alternative is an aspect of the model developed here that might need some clarification.

In the model developed above, emphasis was placed on the conditions favoring the initiation of a diking event, allowing the rapid tapping of magma out of its region of origin or storage. Although some mention was made about the role played by the propagating front of such a fracture away from the RPM, so far no constraints concerning the final end of the tapped magma had been imposed, and somehow it could have been that the model made the implicit assumption that the tapped magma was directly erupted at the surface. Actually, the conditions examined in the previous section only concern the initiation of a fracturing event for the deepest RPM that was justified from all available evidence (seismic, geochemical, petrological, etc.). Nevertheless, those conditions do not suffice to justify an assertion in the sense that the magma tapped from those depths is erupted at the surface as a result of a single tapping event. Actually, using the same hydrostatic arguments that were used to constrain the critical heights of magma required to initiate a fracturing event shown in Figure 6, Cañón-Tapia (2009) also assessed the probability of a single tapping event originating at the deepest RPM to reach the surface. In particular it was noted that the vertical end of a propagating fracture needs to pierce rock of different mechanical properties upon its ascent. The mechanical state of those rocks depends on the lithology, the tectonic setting, and even their prior history, and consequently, a wide range of situations is likely to have taken place in nature. Without attempting to be exhaustive, however, some constraints can be derived from general situations such as those shown in Figure 7. As shown in the Figure 7 diagrams, the vast majority of propagating fluid-filled fractures will be unable to pierce the upper layers of rock under most conditions. This situation can lead to two possible alternatives. One alternative is that the magma coming from below simply stalls at the rheological boundary, forming a sill or other large pluton, and so the magma never reaches the surface. The other alternative is that the stalled magma eventually finds its way to the surface owing to a change in the mechanical condition of the rock above it. The first alternative evidently would explain the occurrence of very large intrusive complexes, such as the Bushveld (Cawthorne and Walraven, 1998) or the Skaergaard intrusion (Tegner et al., 1998). The second alternative would explain some of the geochemical and petrological signatures of most LIPs that suggest some residence time of magmas at relatively shallow levels, and depending on the heterogeneity of the source region it could also explain some of the geochemical trends observed in some LIPs (e.g., Smith, 1992). Which of these alternatives takes place in every case, might depend on a combination of several factors, among which might be the mechanical weakening of the overlying crust owing to the repeated input of magma from below (Annen and

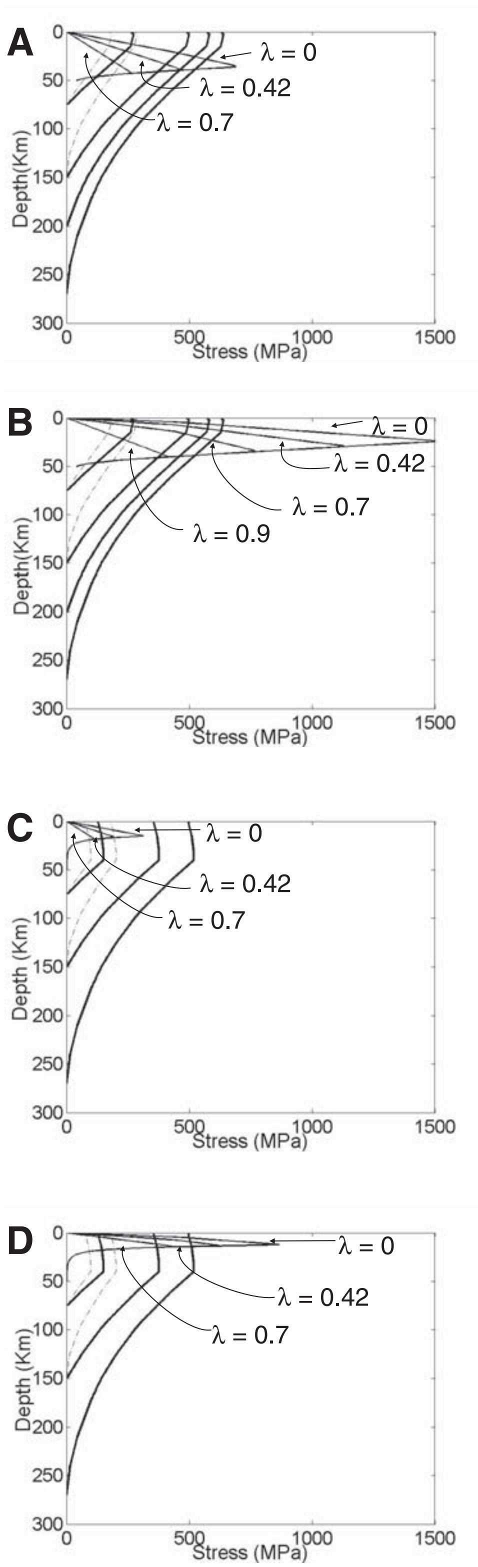

Figure 7. Diagrams showing the relationship between the hydraulic head of an ascending column of magma and the strength of the solid rock as a function of depth (from Cañón-Tapia, 2009). Only the magma columns to the right of the rock-strength curves are able to rise all the way to the surface, with the rest remaining trapped at depth by a rheological boundary.

Sparks, 2002), or a reduction of the horizontal stresses from external, larger scale processes. Consequently, only a detailed analysis of the evidence on a case by case basis could provide the required elements for answering such a question. Such a level of interpretation is beyond the scope of the present work, although the lack of specific examples does not invalidate the general constraints outlined here.

Finally, it is noted that much of the discussion so far has assumed that magma production is completely halted within the RPM of reference, and we had focused only on the processes of tapping such magma. If we remove such constraint, by acknowledging that mantle rocks are likely to undergo some movement when time scales of the order of 1Ma are considered, and that such movement might contribute to the creation of an additional amount of melt in some cases, it turns out that the volume of magma that can be extruded from a region of partial melt analogous to those active today in a 1 m.y. interval can be increased substantially. For instance, consider that the eruption rate for Hawaiian volcanoes has been calculated to be somewhere between 0.03 and 0.1 km^3/yr (Lipman, 1995; Vogt, 1979). Following Figure 5, this magma can be assumed to represent the volume of magma produced by the prism of mantle rock directly beneath each of the active volcanoes in Hawaii. Taking into consideration the surface area of these volcanoes (from 6 to 10 × 10^3 km^2 including the submarine portion), it can be concluded that the region of partial melt depicted in Figure 5 could produce 2 to 4 times as much magma as underneath a single Hawaiian volcano in the same period of time (i.e., from 0.06 to 0.4 km^3/yr). Therefore, in 1 m.y. the amount of magma would extend from 6 × 10^4 to 4 × 10^5 km^3, which would add an additional 40% of the volume of a CFB.

In summary, by comparing the melt distributions as a function of depth inferred from the composition of erupted products, and combining this information with the dimensions of present-day regions of partial melt under zones of active volcanism, it is found that at present an RPM has the potential to produce a similar volume of extruded products in a CFB province in a time interval of 1 m.y. Consequently, this model shows that the difference between CFBs and modern volcanic provinces might not be the rapid production of melt under the surface, but probably it might be related to a more efficient form for extracting that melt from its region of origin. The basics of such a mechanism were outlined in the previous section.

CONCLUDING REMARKS

Undoubtedly, more detailed studies need to be undertaken to provide a more robust model than the one developed here to fully explain the origin of LIPs. In particular, geochemical and petrological aspects of the erupted products need to be examined with more detail in the light of this model. Nevertheless, the approach followed here avoids as much as possible many of the fallacies that have been commonly made when addressing the origin of LIPs, highlighting the fact that there is no need to invoke an extraordinary or abnormal mechanism to explain the occurrence of such volcanic provinces.

In the strictest sense, the model developed here does not invalidate models explaining the origin of LIPs in terms of the occurrence of mantle plumes, but it is also true that such models are not so robust as to exclude any other non-plume explanation without further inquiry. Consequently, it is only fair to say that the mantle-plume origin of LIPs remains a good working hypothesis rather than being a well-established fact beyond any possible objection. Thus, plume and non-plume hypotheses deserve equal treatment, which is only possible if mythical thinking is avoided.

As pointed out by Dickinson (2003), avoiding mythical thinking is important to foster a better understanding of a particular phenomenon for several reasons. First, it helps us to acknowledge uncertainty forthrightly, which is a necessary step in the search for real solutions. Second, by allowing ourselves to entertain alternative explanations, even if some of these might be based on relatively inconclusive evidence at the beginning, we might be able to identify clues that could have been ignored otherwise. Third, by entertaining alternative explanations we do not imply that the dominant explanation is straightforwardly incorrect. Fourth, the strengths of the dominant approach might be better appreciated once the real weaknesses of the alternative models have been objectively established. Consequently, to avoid mythical thinking it is necessary to examine alternative hypotheses, each of which should allow us to delineate a series of critical observations that can be used in future studies to test its validity. However, it is also equally important to consider that some of the critical observations required to test an alternative model might not be available at a given time, not because it is technologically impossible to make them but because the dominant model did not require them. Under such circumstances, it would be unfair to reject the alternative hypothesis only because it has not been explored as deeply as the dominant model, which would therefore reinforce the selective bias in favor of the predominant model typical of mythical thinking. Consequently, to avoid as much as possible such biasing we need to be careful not to ask for more conclusive evidence from an alternative hypothesis than we asked from the dominant model when it was in its initial stages, and furthermore we should accept the challenge of making the necessary observations that can help us to decide in the future whether this alternative approach is reasonably valid or not. The model developed here is a step in that direction, and hopefully the discussion made here concerning the form in which premises might influence our thinking concerning a given subject will be helpful in focusing on the role played by various premises behind the various alternative models proposed mainly in the past 10 yr in explaining the origin of LIPs by a mechanism other than mantle plumes.

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Do pyroclastics form part of a volcano?: A sedimentologist's view

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ABSTRACT

Pyroclastic rocks form part of the products of most volcanoes. It is not self-evident, however, that they consequently form part of their "mother" volcano, not even if the material comes to rest on the volcano's flanks in the form of volcanic ash, ignimbrites or—in a reworked state—as lahars. The question of whether such rocks should be considered part of the volcano is approached here from a sedimentological point of view. For this purpose the sedimentological characteristics of pyroclastic deposits, as well as their significance for sedimentological research, are briefly mentioned. Some situations are sketched that show how non-volcanogenic processes can affect the relationship between pyroclastic deposits and their "mother" volcano. Considering all arguments, it must be deduced that pyroclastic deposits do not form part of a volcano unless covered by effusive rocks (lava) on the slope of the volcano itself.

INTRODUCTION

The term *pyroclasts* denotes particles produced during volcanic eruptions, and the genesis of pyroclastic rocks is therefore by definition volcanic. These rocks were, as mentioned by Fisher and Schminke (1984), for a long time studied almost exclusively by igneous petrographers, but in the course of time they have became increasingly discussed in sedimentological literature as well. Together with clastic, chemical, organic, and organogenic rocks, sedimentologists nowadays consider pyroclastic rocks as one of the five categories of sediments (Middleton, 2003). Pyroclastic rocks thus have become an intermediate type of rock that attracts the attention of volcanologists and sedimentologists alike. Whereas volcanologists can still claim justifiably that pyroclastic rocks are volcanic, sedimentologists are also correct in considering these rocks as sedimentary because pyroclastic deposits consist of material that has been transported through the air (just like, for instance, the particles building loesses) or that has moved downslope from the volcano through mass flowage (just like, for instance, debris flows that come down a mountain slope). Furthermore, such materials may become "contaminated" during transport with other (non-volcanogenic) particles (cf. Buesch, 1992), and consequently the material forming a pyroclastic deposit might not be derived from a volcanic eruption in its entirety.

One might argue that transport of volcanic particles through the air is not sufficient to consider the deposits formed after settling from the air as sediments (or, after lithification, as sedimentary rocks), but this is merely a matter of semantics. How many sedimentary characteristics and what percentage of non-volcanogenic particles should such deposits have to be called a sediment? In fact, loose material on a volcano's slope commonly is transported downslope over a more or less significant distance, sometimes only by rolling over a few centimeters, sometimes by gravity-induced sliding over a few meters, sometimes as bed load in small streams that can run for kilometers and that can produce current ripples in the top part of reworked ash (Raj, 2008). Furthermore, sedimentologists recognize the importance of the various transport processes and diagenetic changes that particles may undergo, and therefore distinguish between volcanic ash (if fine grained) and pyroclastic fall deposits (without a grain-size indication) if the material is deposited directly by settling from the air after a volcanic eruption (the consolidated or cemented equivalent is called *tuff* in this case). If this material has been transported grain by grain down a slope (for instance in streams

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that may develop after rainfall), the resulting sediments are called *tuffite*. Importantly for sedimentologists, tuffites can contain non-volcanogenic particles (in various amounts) that have been eroded from the subsoil. If either pyroclastic fall deposits (or broken-up tuff particles) or tuffites move downslope in the form of mass transport (for instance because the water-saturated top layer becomes unstable at the slope of a volcano), the deposit that originates after the mass flowage stops is called a *lahar* or *volcanic mudflow deposit*. Consequently, from a sedimentological point of view, there is sufficient reason to consider such rocks as sediments, and actually sedimentologists consider pyroclastic fall deposits, tuffs, tuffites, and lahars, as well as ignimbrites, to be pyroclastic sediments (Brown and Calder, 2005; Neuendorf et al., 2005).

It is interesting in this context that the present author is not aware of any sedimentological study in which lava flows are considered as sediments, yet, lava flows are emplaced by gravity as a kind of mass flow, which is a typically sedimentary process (like transport in the form of mass flows). In addition, on their way downslope, either subaerially or subaqueously, lava flows may erode the subsoil, which results in (not rarely sedimentary) masses of material that become embedded within the lava flow (see, among others, Flower and Strong, 1969; Houghton and Hackett, 1984; Kano, 1989; Nasir, 1992; Clocchiatti et al., 2004; Sottili et al., 2009). This implies that lava flows do not always consist exclusively of volcanic material. Particularly when an acid lava flow is cooling and moves downslope as forceful viscous lobes, it may induce considerable erosion. The embedded sedimentary masses may not become heated enough to melt, so that a lava unit can originate in a deposit in which sedimentary masses are well recognizable. Even if the amount of embedded sedimentary material is significant, however, the lava is considered by sedimentologists, as a rule, as a volcanic rock. To some extent, such a different appreciation reflects the nature of the material that is transported downslope. Whereas pyroclastic material is transported as a mixture of gas and solid particles, lava flows form a coherent mixture in which the liquid being expelled out of the volcano defines a continuum. Consequently, from the sedimentologist's point of view the particulate nature of pyroclastic rocks determines its affinities with sedimentological processes rather than the transport processes by themselves.

Although seemingly trivial at first sight, the volcanic-versus-sedimentary affinity of pyroclastic rocks results in an interesting problem regarding the definition of volcanoes, because stratovolcanoes are formed by superimposed pyroclastic deposits and lava flows, sometimes portrayed as cakes with various alternating layers. The exact proportion of pyroclastic deposits and lava flows on a given stratovolcano is rarely well documented, but most volcanologists would concede that the amount of pyroclastic material is near to half the volume of the volcano (and sometimes even more) in many cases. Consequently, if attention is addressed to the depositional mechanism of the particulate products that build the edifice, stratovolcanoes should be considered sedimentary formations in their own right, and these formations should be included in the category of sedimentary volcanoes (cf. Van Loon, this volume). In any case the intermediate nature of pyroclastic products also represents a problem in the sense of deciding which tools are more appropriate for their study. Should this type of rock be approached by using the volcanologist's or the sedimentologist's approach? To what extent do we need to ask a sedimentologist studying pyroclasts to become a volcanologist or vice versa? What is the loss for a sedimentologist (volcanologist) if he/she decides to ignore the tools of the other field of study? Commonly these questions are not addressed in the published literature, and therefore there seems to be a gap (or even worse, a bias) in the curricula of many students potentially interested in studying pyroclastic rocks.

An additional problematic aspect of pyroclastic particles is that they can travel too far away from the volcano that ejected them to the surface of the planet. The long distances traveled by some of these particles, in whatever form, may blur the boundaries of a volcano, as in general it would be difficult to have a definitive answer to the question of whether a given deposit is part of the volcanic edifice or not. Although this problem might seem trivial at first sight, it can have important implications for the definition of a volcano, and consequently a closer examination of this problem might help to reach a more definitive answer to the general question of what a volcano is.

In this contribution, I address the two latter problems. First, I examine the importance of some pyroclastic products from a sedimentologist's point of view, outlining the types of studies that volcanologists commonly do. The idea of this section is not to give an extensive account of pyroclastic rock facts for volcanologists (the book by Fisher and Schminke, 1984, does an excellent job of this) or for sedimentologists, but rather to pinpoint the advantages of combining strategies in the study of such a hybrid type of rocks. In the second part of the paper I incorporate the sedimentologist's point of view of pyroclasts in a discussion concerning the relationship of pyroclastic rocks with the boundaries of a volcano.

THE SEDIMENTOLOGIST'S APPROACH TO STUDYING PYROCLASTICS

Pyroclastic Fall Deposits

From the various types of pyroclastic deposits, pyroclastic fall deposits (for the sake of brevity called *ash* in the following, regardless of the grain size) carry the most important sedimentological information. Ash is important for sedimentological research for several reasons. The most important in the context of this chapter are (1) it commonly covers extensive areas, (2) it is a most valuable marker horizon, and (3) volcanic minerals often provide detailed source-area information.

Volcanic ash can, in the case of a large eruption, easily travel hundreds of kilometers (Woods et al., 1995; Turner, 2001) while gradually settling from the air, thus producing a layer or at least a horizon that is traceable over an area of the order of some hundreds of thousands of square kilometers (as in the case of the famous Thera and Krakatau eruptions). The areas covered by the products of a single eruption may include both marine and continental settings, thus offering an excellent and reliable method for correlating continental and marine deposits (Siani et al., 2004). Furthermore, extensive continental areas may include a wide variety of sedimentary environments, and these may form part of geologically entirely different entities (e.g., a number of unconnected basins) divided from each other by mountain ranges, inland seas, great lakes, etc.

The relevance of volcanic ash as a correlation tool is even more evident in ancient deposits (weathered ash even received a special name: *tonstein,* a German word meaning *claystone*), where it may represent key correlation units, particularly in continental settings. Even in a small depositional basin where rapid facies shifts occur that prevent easy correlations, ash (tonstein) can be the only reliable tool for correlation (Eden et al., 1963); highly detailed correlations are possible if several tonstein layers are present (e.g., Zhou et al., 2000). Evidently, the more that detailed knowledge of a given area is available, the easier it is to obtain a clear picture of the general environmental conditions of the pertinent volcanic activity.

Nevertheless, it is not always necessary to have a true ash layer or an ash-rich horizon to extract useful sedimentological information using volcanic knowledge. If an area is eroded, and the eroded particles are transported to come to rest at a depositional site, the mineral content of the deposits provides information about the source area(s). Particularly heavy minerals are commonly used for this purpose (Mange and Wright, 2007), and because some heavy minerals from volcanic ash show typical characteristics, they commonly allow the source area of a sediment to be identified, which implies that the (rough) transport direction can be reconstructed, which is of great importance for analysis of the paleogeographical development (Ross, 1999). A well-known example deals with the Quaternary of The Netherlands. The Dutch Quaternary stratigraphy and sedimentology have, for many decades, been investigated in extreme detail, probably in more detail than in any other country (De Jong, 1967; Wong et al., 2007). Many sedimentary units, however, are difficult to distinguish from one another (for instance, because they form part of fluvial successions deposited by different rivers such as the Meuse and the Rhine). In addition, these Quaternary (Pleistocene) sediments—deposited mainly under glacial and periglacial conditions—cannot be dated themselves (for instance, on the basis of microfossils) with sufficient accuracy. For this reason, heavy minerals are very important for determination of the stratigraphic position. In particular, the volcanic mineral augite is an excellent example. For instance, the sudden occurrence of this mineral in one of the Dutch Pleistocene fluvial units (Urk Formation) could be traced back to result from volcanism in the Eifel Mountains in Germany. Not only could the unit be classified as Rhine deposits (which was possible already on the basis of both the entire heavy-mineral suite and the composition of the gravel), but it was also possible to date the unit with fairly good accuracy, as the volcanism in the source area could be dated precisely (500–420 ka: Lippolt et al., 1986).

Another characteristic of pyroclastic fall deposits is that they show, as a rule, a wide range of grain sizes. Consequently, they are commonly poorly sorted in the direct vicinity of the volcano (Fig. 1). Transport through the air increases sorting, so that it is possible to reconstruct in which direction the volcano that produced an ash layer should be located (Sukumaran et al., 1999). Even if the wind has a strong directional component during a volcanic eruption—and the stronger the wind, the more eccentric the ellipsoid of the depositional surface becomes—some ash layers or horizons are commonly deposited on all sides of the volcano. A well-known example of a strongly ellipsoid depositional area is the ash deposited after the giant eruption of Thera (Santorini, Greece) ca. 1613 B.C. (Balter, 2006). The ash can, owing to the then prevailing northerly winds, be traced even farther south, into Egypt. The well recognizable ash leaves no doubt that this ash was derived from the Thera eruption (Eastwood et al., 1999), so it

Figure 1. Imprint made by a volcanic bomb. Particularly near a volcano, ejected particles of different sizes settle because no sorting mechanism has significant influence. Golan Heights, Israel.

is possible to map the deposits in detail. If the Thera eruption had occurred in the remote geological past, and if the volcano itself had been eroded away, its exact position could still, in spite of the ellipsoid shape of the ash deposits, have been reconstructed on the basis of grain-size analysis: Thus the volcano must have been positioned at the point at which the ash became finer grained in all directions.

As in non-volcanogenic clastic sediments, many types of sedimentary structures that indicate some kind of transport can be found in volcanic ash (this is why such deposits are regarded by sedimentologists as sedimentary rather than as volcanic). The most common, obviously, is the irregular bedding (Fig. 2) that originates because deposition from the air is not continuous (Aalto and Miller, 1999) but irregular (commonly not because of pulses in the expulsion of particles from the volcano but to variations in both wind power and wind direction). Under relatively quiet conditions, however, the bedding can be quite regular (Fig. 3). From a sedimentological point of view, more interesting are the various types of structures—relatively rarely studied by volcanologists—that also can be found in clastic sediments and that provide information about current directions (for instance, ripple marks that indicate the paleoslope direction, and sole marks (Aalto and Miller, 1999) such as grooves, formed by objects such as large blocks that slid down the slope. Many of these structures indicate that the particles found in ash layers, after having settled from the air, have been reworked by currents running down from the volcano's slope. In ancient deposits this can help to reconstruct the paleoslope, and thus it becomes possible to infer the direction at which the vent must have been positioned, even if the remnants of the volcano or the lava flows are not exposed or have disappeared by erosion. Consequently, volcanologists benefit by studying pyroclastic deposits using sedimentology-related tools, because such tools allow them to reconstruct the details of an eruption or to assess the general characteristics of a volcanic province (including vent distribution and/or paleoslopes).

Figure 3. Regularly bedded volcanic ash, indicating quiet depositional conditions. Golan Heights, Israel.

For sedimentologists, however, some structures in ash also give information that is commonly less relevant for volcanologists. Examples are the imprints that can be left in ash by organisms, such as dinosaur tracks in ancient rocks and bear tracks in recent ash (Fig. 4). Even more spectacular than the famous (but few) Laetoli footsteps are the thousands of footsteps made by prehistoric man in Mexican ash of at least 40,000 years old (Fig. 5) (González et al., 2006) and the oldest dated human footsteps (345,000 years old) that constitute the Devil's Trail in lithified ash of the Roccamonfino volcano (Campania, southern Italy) (Fig. 6) (Mietto et al., 2003; Scaillet et al., 2008). In any case, imprints left in fresh ash by animals and plants, in combination with body fossils that were preserved because of quick burial—sometimes followed by deposition of a mudflow, sealing off the air and therefore resulting in anoxic conditions—help sedimentologists to reconstruct the fauna and flora of the area at

Figure 2. Strongly irregular bedding in volcanic ash because of changing wind directions and variations in wind intensity. Golan Heights, Israel.

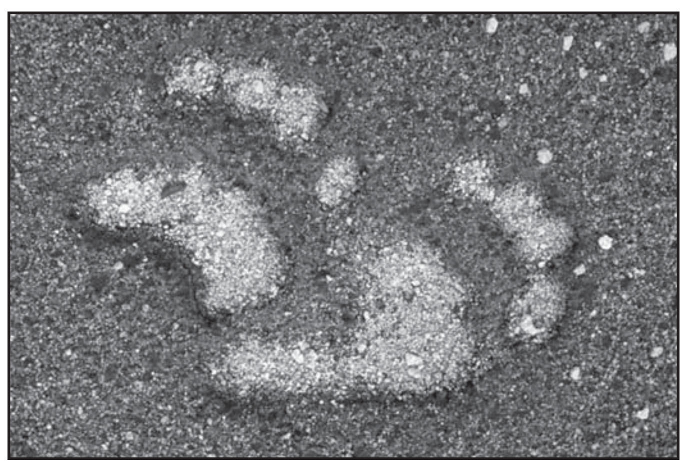

Figure 4. Bear tracks in recent ash (U.S. Geological Survey photograph taken in October 1980 by Lyn Topinka).

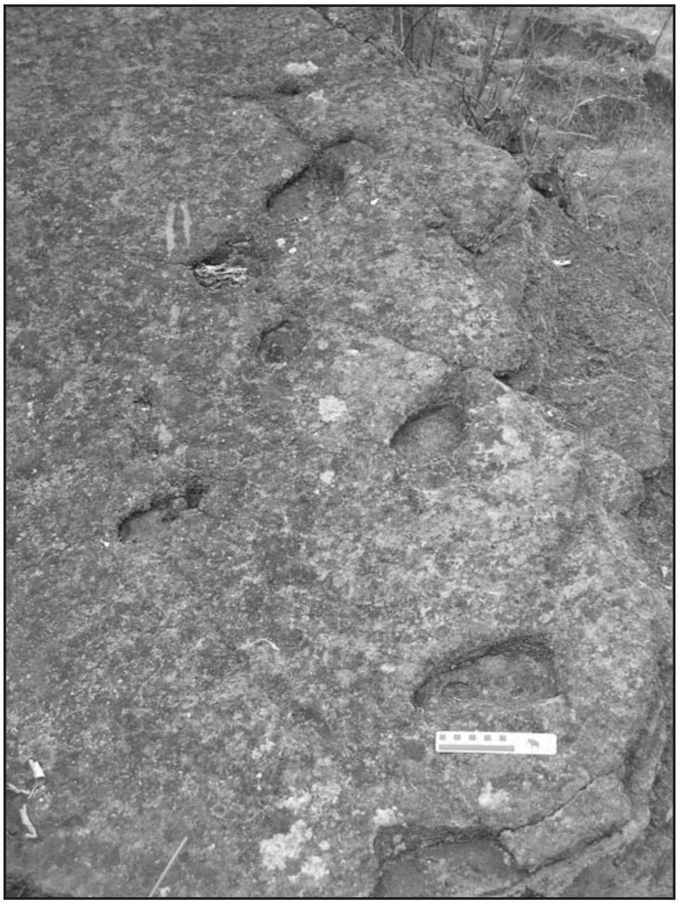

Figure 5. Lithified volcanic ash in de Barranca La Mina del Frances with various imprints of prehistoric man (from González et al., 2006; reproduced with permission).

Figure 6. The oldest dated human footprints, in ash 345,000 years old of the Italian Roccamonfino volcano (from Scaillet et al., 2008; reproduced with permission).

the time of the ash fall; these data provide a proxy for environmental and paleoclimatic reconstructions.

Other Types of Pyroclastic Sediments

Lahars (volcanic mudflows) and ignimbrites (deposits that result from so-called *nuées ardentes*; ignimbrites are also called *pumiceous pyroclastic flow deposits* after Sparks et al., 1973, and Cas and Wright, 1988) are the two other types of pyroclastic rocks that are of interest for sedimentologists, particularly in the context of hazard prevention. In addition to volcanic ash (in the sense used above), these two types of pyroclastic rocks also form a common study object for volcanologists and sedimentologists alike, as they offer more examples of how an interdisciplinary approach may be fruitful. Both types therefore will be briefly examined in this section.

Lahars

Lahars are subaerial forms of mass transport (Fig. 7), commonly high-density flows that, in a nonvolcanic context,

Figure 7. Lahar (dark, on the snow) on Mount Saint Helens, formed after an explosive eruption on 19 March 1982. It flowed into the North Fork of the Toutle River valley. Part entered Spirit Lake (lower left), but most material went down the Toutle River, eventually reaching the Cowlitz River, 80 km downstream (U.S. Geological Survey photograph taken 21 March 1982 by Tom Casadevall).

would be called *mudflows*. They tend to grow quickly during their downslope transport, as they easily erode water-saturated freshly fallen ash as well as other sediments (Thouret et al., 1998); in the latter case, a sediment is formed that consists partly of volcanogenic, partly of non-volcanogenic material (and even entire trees can be included: Fig. 8), thus showing that this type of pyroclastic rock is also intermediate between effusive (igneous) and clastic (sedimentary) rocks. Lahars flow down through valleys, reaching often densely populated areas at the foot of the volcano. This happened, for instance, with the Roman village of Herculaneum in A.D. 79. Large blocks up to 10 m in diameter form part of the deposit that came to rest when the flow lost its velocity, possibly due to the many obstacles formed by the Herculaneum buildings. The Roman village of Pompeii, which was long considered to have become buried exclusively by volcanic ash, was found by sedimentological analysis by Luongo et al. (2003) to have been destroyed partly also by lahars, a more sedimentary type of threat.

Mass flows and the behavior of muddy sediments have been the subject of intensive research since the middle of the last century (Boswell, 1948; Kuenen and Migliorini, 1950), and it has been well known for decades (Crowell, 1957; Dott, 1963; Van Loon, 1970) that mudflows can suddenly "freeze." Although still many questions have to be answered, it turns out that changes in both water content and slope play a dominant role. This insight, however, has not helped sedimentologists thus far to develop methods that can be used to freeze lahars when they threaten a populated area; one of the main problems is that they tend to have such a high velocity that they reach populated areas before they have been noticed (e.g., lahars of the 1977 and 2001 eruptions of Popocatépetl in Mexico reached 50 km/h: Muñoz-Salinas et al., 2007). For the time being, the "solution" seems therefore, as in the case of lava flows, to build constructions that force (or at least help) the flow to take another route (Lavigne, 1999).

Figure 8. Tree trunk embedded in a lahar, showing the erosive power of mudflows running down a volcano's slope and characterizing the poor granulometric sorting within such a deposit.

Another interesting aspect of lahars is their ambiguous association with volcanic eruptions. For instance, one of the largest catastrophes associated with volcanic activity in human history took place during the 1985 eruptive activity of the Nevado del Ruiz volcano in Colombia (Barberi et al., 1990; Pierson et al., 1990; Voight, 1990), although it was not due to the material directly ejected by the volcano at that particular time. Actually, the intensity of the eruption was moderately small if measured only in terms of the amount of ejected material, but these products triggered the formation of lahars that destroyed the town of Armero, killing ~23,000 people. The Nevado del Ruiz lahars (which had sedimentologically highly interesting characteristics: Calvache, 1990; Thouret et al., 1990; Kohlbeck et al., 1994) were clearly associated with the ongoing eruption, even though the material that formed the lahar was for the most part not ejected by that eruption itself. In contrast, a change in the hydrology of Mount Pinatubo in the Philippines took place from the deposition of volcanic deposits even prior to its 1991–1992 eruption. This change promoted the formation of lahars that involved the recently deposited pyroclastic sediments. Nevertheless, the formation of lahars on this volcano continued to take place many years after the eruption had ended (Chorowicz et al., 1997). Consequently, even though the lahars may have involved the pyroclastic deposits formed during one particular eruption, the lahars themselves cannot be considered, in the Pinatubo case, to have been the direct result of such an eruption. For this reason, it is clear that lahars are a good example of an intermediate type of rock that defies the common division established between volcanic and sedimentary fields.

Finally, for the purpose of considering whether a lahar forms part of its "mother" volcano, in the next section, it should be realized that these mass flows can travel not only with a high velocity but also quite far. The 13 November 1985 Nevado del Ruiz lahar traveled some 50 km down the Lagunilla River valley before devastating Armero. Mount Rainier (USA) produced some 55 lahars during the last 10,000 years, and many of them traveled some 40 km, whereas Mount Saint Helens produced lahars that traveled even more than 120 km (Burns, 2005).

Ignimbrites

Even more devastating (though fortunately much scarcer) than lahars are the nuées ardentes (Fig. 9), nowadays more commonly indicated by the fairly ambiguous term *pyroclastic flows* or *pyroclastic density currents* (Branney and Kokelaar, 2002). These mixtures of fine particles and volcanic gases can run down with velocities of >150 km/h. Numerous ancient deposits of this type are known, which are called *ignimbrites* (Fig. 10). Nuées ardentes can erode the slope from which they run down, so that clastic material may be found incorporated in an ignimbrite (Buesch, 1992), thus supporting the view that this type of pyroclastic rock, too, is intermediate between igneous and sedimentary.

Figure 9. Nuée ardente from the 7 August 1980 eruption of Mount Saint Helens, stretching from the crater to the valley floor (U.S. Geological Survey photograph taken 7 August 1980 by Peter W. Lipman).

Figure 10. Polished surface of an ignimbrite from the Oslo Graben (Norway).

Ignimbrites are studied by sedimentologists (e.g., Giannetti and De Casa, 2000; Mandeville et al., 1996) particularly because they occur commonly interbedded between lahars and volcanic ash. These studies indicate that ignimbrites can show vertical and lateral changes that allow a type of stratigraphy to be established (Choux et al., 2004; Carrasco-Núñez and Branney, 2005; LaBerge et al., 2006; Caffe et al., 2008). The grain-size distribution within ignimbrites can also help to reconstruct the transport mechanism (Sparks, 1976).

Like lahars, nuées ardentes can travel great distances. Ignimbrites are known to cover vast areas, up to some 45,000 km^2 (Brown and Calder, 2005), which implies travel distances of the order of at least 50 km. They can consequently reach areas that are situated far away from the volcano itself.

Tsunamites

For the sake of completeness, it is noteworthy in this context that sedimentologists also study another type of sediment that is (sometimes) related to volcanism (Begét et al., 2008). This concerns tsunamites (also called "tsunamiites": Shiki et al., 2008a), which are coastal sediments formed by the reworking of coastal material resulting from the action of tsunamis. Such deposits were formed, among others, following the 1883 eruption of Krakatau (Pelinovski et al., 2006) and the eruption of Thera (Santorini), which was likely responsible for the termination of the Minoan civilization on Crete (Taddeucci and Wohletz, 2001; Friedrich et al., 2006). An eruption on the Kurile Islands in 1737 generated a tsunami that swept up a fjord to a height of 65 m (McCall, 2005). Tsunamites, which are characterized by a vertically saw-toothed granulometry (Shiki et al., 2008b) from the "shuttle movement" of repeated flooding and backwash events (Shiki and Cita, 2008), are not pyroclastic in nature, however. Consequently, these types of deposits are not discussed further in this paper.

In summary, all the examples included in this section show that volcanic ash is of interest for volcanologists and sedimentologists alike, but possibly most important is the conclusion that a spiraling effect, leading to ever more insight, is created if volcanologists and sedimentologists benefit from each others' expertise. Consequently, through its intermediate nature, volcanic ash promotes the fruitful cooperation between researchers in both earth-science disciplines.

ARE PYROCLASTIC DEPOSITS PART OF THEIR "MOTHER" VOLCANO?

It can be deduced from the above that sedimentologists have a great interest in pyroclastic deposits and their genesis. Ash, lahars, and ignimbrites are studied in detail for their sedimentological characteristics. One might therefore expect that sedimentologists are also interested in the study of volcanoes. It appears, however, that sedimentologists rarely deal with volcanoes themselves, apart from the pyroclastic deposits that may be found on them. This raises the question of whether such pyroclastics do form part of the volcano, or rather have they become an independent object unrelated to volcanic activity at all. Although this may seem a semantic problem at first sight, it is also of practical importance, as ambiguous terminology may hamper efficient communication.

In fact, one should distinguish between two questions that may be asked: (1) are lava flows and pyroclastic deposits in the foreland of a volcano part of that volcano, and (2) is all material that is deposited on a volcano also part of that volcano?

Rocks in the Foreland

The relationship between lava flows and volcanoes is generally straightforward. In considering the solidified lava that has come to rest on the flank of the "mother" volcano, there can be hardly any discussion about the question of whether it is part of that volcano. Actually, considering any lava flow as part of its "mother" volcano is in fact in accordance with the commonly used (formal or informal) definitions of a volcano. Nevertheless, it is noted that some lava flows may travel over huge distances. For instance, even though there is still a debate concerning the identity of the fissures from where the lavas now forming the Siberian Traps (Reichow et al., 2009) and the Deccan Plateau (Shrivastava and Ahmad, 2005) were issued—and there may have been numerous places relatively far away from one another—these examples show that lava flows can travel hundreds of kilometers from their vent. In these cases, the limits of the volcano producing the lavas are not straightforward to obtain. Leaving aside these extreme cases, similar problems can be found concerning lavas that do not come to rest on the flanks of the volcano. For instance, what about lavas that have come to rest on a sedimentary cover in the foreland of the volcano? One might argue that the lava flow is part of the volcano in these cases because the lava flow is still physically connected with the volcano. But what if such a lava flow becomes disconnected from the parent volcano, for instance as a result of tectonics or due to erosion? It seems to me that one should not consider such a flow as a distal parts of the "mother" volcano. It should be realized in this context that the disconnection from the "mother" volcano can, certainly in cases of strong and/or repeated tectonic activity, result in configurations that make it practically impossible (certainly in the field) to find out with which volcano a specific outcrop of lava may have been connected in the geological past. Nevertheless, it seems against logic to state that a remote lava flow is part of its "mother" volcano until the moment that it becomes isolated from the volcano by tectonics or erosion. Thus, tentatively it can be considered that if the boundaries of a volcano are marked by a break in the slope of its flanks, then any lava flow that extends beyond this break starts to be detached from the "mother" volcano. As the volcano grows (by repeated eruptions and by extending the area over which a relatively steep slope is found) this lava flow might become part of the volcano again. Evidently, if the lava flow traveled far away from the main volcano-edifice, then the lava ceases to be an integral part of the mother volcano. In the cases where a break of slope is absent (e.g., on the fissure eruptions mentioned above), whether the lava flow is still part of a volcano remains ill defined.

The same reasoning holds, obviously, for pyroclastic deposits, perhaps even more, because ash layers tend to pass gradually, with increasing distance from the volcano, into ash horizons (which are themselves constituted of non-volcanogenic material in which ash fragments are embedded), whereas lahars and ignimbrites may even result from processes that leave no deposits on the volcano's flanks itself, but only in the foreland where the slope is so slight that the material running down comes to rest.

The above considerations are supported by the various definitions and descriptions of a volcano (see, e.g., McCall, 2005; Allaby, 2008), although some older definitions leave room for discussion (Visser, 1980). Whatever definition of a volcano is used, however, as long as such a definition emphasizes the relevance of volcanoes as a geomorphological feature, the conclusion seems unavoidable that pyroclastic rocks (just like lava) should not be considered part of their "mother" volcano if they have come to rest elsewhere than on the flanks of the volcano itself. Other definitions of volcano advanced in this volume (e.g., Borgia et al., this volume; Szakács, this volume) cannot be used to address this issue. In any case, it is noted that the relationship of pyroclastic rocks with the flanks of a volcano expressed above does not imply, however, that all pyroclastic rocks deposited on the flanks of the "mother" volcano should be considered by definition as part of that volcano; the restrictive conditions will be elaborated in the following subsection.

Pyroclastic Deposits on the Flanks of a Volcano

A volcano is, according to Neuendorf et al. (2005), "A vent in the surface of the Earth through which magma and associated gases and ash erupt; also, the form or structure, usually conical, that is produced by the ejected material...." In the present context, where the vent is no point of discussion, the second descriptive definition is best applicable. This seems to leave little doubt: In this definition, pyroclastic deposits on a volcano's flank are part of the volcano.

In the above definition (and in many other definitions of a volcano) the morphology plays a role. A volcano may, for instance, undergo erosion and still be called a volcano, but the erosion should not remove all lava flows and pyroclastic deposits so that only a neck remains; a neck is not considered a volcano but rather a volcano remnant. This, however, raises the question whether volcano-shaped morphological features that owe their genesis predominantly to volcanic activity should be considered volcanoes in their entirety. What if non-volcanogenic components come to rest on a volcano and thus form part of the morphological feature? Such a situation can occur when volcanoes are situated in an arid environment (which can result in windblown dust on the volcano; see, among others, Von Suchodoletz et al., 2009, who mention the occurrence of Saharan dust on the volcanoes of the island of Lanzarote) or if a reef develops where a volcano rises above sea level (and where such a situation may eventually lead to the formation of an atoll: McManus, 2009).

One might distinguish between two types of non-volcanogenic components: rock and other material. Non-volcanogenic rock material that can be found on the flank of a volcano will consist of particles that have settled from the air. Particularly volcanoes in the vicinity of deserts or periglacial areas may receive (sometimes significant) amounts of (predominantly silt-sized) particles, most commonly consisting of quartz. Should such eolian material on a volcano be considered part of that volcano? The answer to this question depends on the approach: For geomorphologists the answer may be positive; for volcanologists the answer may be negative.

The question becomes still more complex if the material that settles on top of a volcano does not consist of rock material but of snow. Snow (if present for a long time it may partly be transformed into ice) is found on top of many large volcanoes (Fig. 11), and it does not essentially change the morphology of the volcano. Yet, few earth scientists will consider snow or ice as part of a volcano. This may seem logical at first sight, but it is not. Water (and consequently also snow and ice) is a compound that should be considered (according to most definitions) as a kind of sediment (albeit, admittedly, an exceptional type of sediment). It would not be very scientific to consider one type of sediment, if deposited on a volcano, as part of that volcano, if other sediments in the same position would not be. It seems nevertheless most practical to exclude snow (and ice) from the components that build a volcano, particularly since parts of the snow and/or ice may disappear temporally from melting to re-appear after the next snowfall. Considering the snow as part of a volcano would therefore imply a continuous change in the volcano (in the form of alternating growth and decreasing size), even if the volcano is dormant or extinct.

Figure 11. Perfectly shaped Fujiyama (Mount Fuji) in Japan, one of the many volcanoes showing a snow-covered top (photo courtesy of Hiroki Suzuki).

Excluding snow and ice as part of a volcano would, however, also pose problems. It is well known that snow-covered volcanoes may eject ash that settles after having cooled off high in the atmosphere on the snow cover that has not melted away completely (which may be the case if the ejection of the particles occurs at only one side of a large volcano). The activity of a snow-covered volcano can thus lead to a succession of ash and snow layers, which successions are well known (Fig. 12), among others from the Copahue volcano (Argentina) that erupted in July 2000 and has been seismically active ever since (Ibáñez et al., 2008). If the snow is not considered part of the volcano, the latter would in this case consist of, for instance, an igneous body covered by snow (which forms no part of the volcano), followed by an ash layer (that might be considered as being part of the volcano, followed by a snow layer (not a part of the volcano), etc. In fact, if the ash layers would be considered part of the volcano, and the snow layers would not, the volcano would consist of an igneous body with several layers of ash that "hang in the air" and have no contact with the igneous body or with the other ash layers. This approach seems absurd from a sedimentological, stratigraphical, geomorphological, and volcanological view. The conclusion must therefore be that either both snow and ash (or other pyroclastic deposits) on the flank of a volcano should be considered part of the volcano, or none of them should. It seems, for all the above reasons, the most logical—from both a theoretical and a practical point of view—to consider neither snow (nor ice) nor pyroclastic deposits as part of the volcano on which they rest.

Implications for the Definition of a Volcano

It was concluded in the above subsection that pyroclastic deposits should not be considered part of their "mother" volcano, irrespective of how they have come to rest in the volcano's foreland or on the volcano's flanks. This apparently logical conclusion is, however, not tenable either, as it would have illogical implications for stratovolcanoes that are built by both lavas and pyroclastic deposits. If the latter are not considered part of the volcano, lava streams that cover them—and the lava on the volcano should be considered by definition as part of the volcano—would in this case cover material (i.e., a pyroclastic rock) that is not part of the same volcano. The consequence would be that the morphological feature that is called a volcano would, from a volcanological and sedimentological point of view, only partially be a volcano with something else inside. But what should that "else" be called, or how should it be classified? Even if one would opt for such an approach, it would have no practical meaning, as one cannot be sure (unless very detailed and expensive action is taken, for instance, in the form of a dense network of borings) what kind of material is present underneath the surficial lava.

Consequently, the conclusions reached in the previous section must be adapted to include "buried" pyroclastic deposits. On the basis of all the above considerations, one might define a volcano as "a body of rock with a more or less conical shape that is built of volcanic material, excluding pyroclastic rocks and

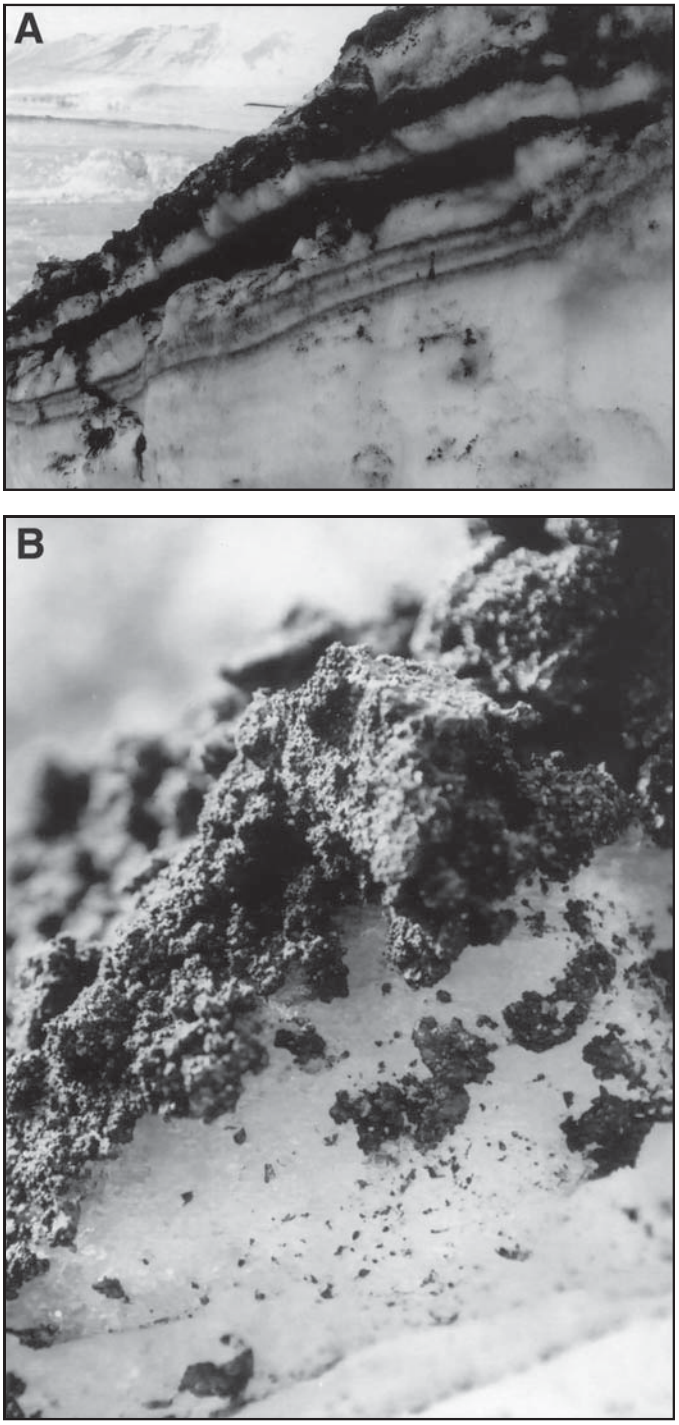

Figure 12. A rare sedimentary succession: alternating snow and ash layers, resulting from the July 2000 Copahue (Argentina) eruption. Photos courtesy of Elizabeth Rovere. (A) Succession six days after the eruption. (B) Detail of the irregular contact between ash and snow, owing to the impact velocity of the particles on the fresh snow.

non-volcanogenic materials that cover the most surficial lava." The question of whether pyroclastic deposits are part of a volcano should therefore be answered negatively, unless they have been deposited on the volcano itself and have later become covered by a lava flow. Such an answer might satisfy all earth scientists, and it might encourage such scientists, like volcanologists and sedimentologists, to cooperate in the ever ongoing effort to unravel the history of the Earth by applying the expertise gained in the different earth-science disciplines.

CONCLUSIONS

Pyroclastic rocks can be considered either sedimentary or volcanic, depending on the criteria used. None of these views is intrinsically better than the other, and the interest in these intermediate rock types resides in the different aspects of the rock that are important for the purpose of the study. Nevertheless, an often overlooked fact is that a richer appreciation of these rocks can be gained by adopting the approach of the "other" discipline from time to time. The hybrid nature of pyroclastic rocks gives place to interesting problems that might influence the very definition of what a volcano is. Keys for the solution of these problems, and many others, are the study of pyroclastics, which can provide most valuable information for both volcanologists and sedimentologists alike. This proves that joint research efforts in both earth-science disciplines can contribute significantly to a better insight into the processes that shape Earth.

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Is Tharsis Rise, Mars, a spreading volcano?

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ABSTRACT

A theoretical similarity derived for spreading volcanoes remains valid for over three orders of magnitude in size, from small slumping volcanoes to large ocean plates. This similarity means that Mount Etna, Sicily, can be used as a terrestrial analogue of Tharsis Rise on Mars. A quantitative comparison of surface structures suggests that Tharsis has been spreading outward as a result of movement of plates at a planetary scale, producing rifting through the summit of the Rise, folding around its periphery, and with radial tear-fault systems connecting the summit rifts to the basal folds. The tear-fault systems form the fossae, of which Valles Marineris is the largest, and decouple the various plates of the Rise, so Tharsis appears to be analogous to a terrestrial mid-ocean ridge, but without any corresponding subduction zone. Instead, like volcanoes, the Tharsis plates are obducted over the Martian crust. The maximum viscosity at the bottom of the Martian lithosphere, derived from our spreading analysis, is in the order of 10^{21} Pa s, so if the mantle of Mars is hot and ductile, we suggest that the spreading of Tharsis could still be occurring.

INTRODUCTION

Volcanoes of a large size range occur on Earth and the other terrestrial planets and satellites. Their geological evolution is controlled not only by the erupting magma but also by the interaction between the volcanic edifice and its basement (Van Wyk de Vries and Borgia, 1996; Borgia et al., 2000a). Small volcanoes such as cinder cones tend to have radial symmetry and usually do not spread because they are too small. Larger volcanoes, like Concepciòn, Nicaragua, or Vesuvius, Italy (Borgia and Van Wyk de Vries, 2003; Borgia et al., 2005), will tend to sink and spread when they are built on a weak basement. These volcanoes become radially faulted, and the basement is pushed and thrusted outward, becoming itself faulted and folded.

For even larger volcanoes, like Etna or Mauna Loa, spreading also may be induced by the intrusive complexes that commonly accumulate inside them. These complexes, being hot and ductile, tend to flow laterally, tearing apart the volcanic summit to generate rift zones, and pushing aside their spreading flanks (Borgia, 1994). The inception of these larger-scale edifice dynamics results in an adjustment of the overall symmetry of the volcano, causing it to change from radial to bilateral.

Therefore, from a dynamic viewpoint, it is reasonable to propose that a continuity in gravity-driven processes exists over many orders of magnitude in space and time, ranging from small, slumping cones to large, spreading ocean ridges. This is the approach taken by Borgia and Treves (1992) in attempting to interpret some ophiolitic complexes found in accretionary prisms as remnants of obducted ocean shield volcanoes. This analogy, we believe, may apply also to planetary volcanism, and it is assumed

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in the definition of *volcano* given by Borgia et al. in this volume (Chapter 1). In particular, we think that Tharsis Rise on Mars offers us the opportunity to test these ideas. Former work on Tharsis concentrated on explaining its topographic and gravitational anomalies by modeling the stress distribution generated by volcanic lithospheric loading (Banerdt et al., 1982), buoyant uplift of the lithosphere (Banerdt et al., 1992), isostatic density compensation (Sleep and Phillips, 1979), or mantle convection (Banerdt et al., 1992; Schubert et al., 1990). These authors then compared the calculated stress distribution to the observed tectonic deformation. Each of these models explains some of the tectonic features of Tharsis, but none has achieved general consensus.

Here we take a different approach. First, we describe the theoretical foundations of the similarity that characterizes spreading volcanoes, and we derive a fundamental dimensionless number for the spreading process. We show that this number maintains consistency from small spreading volcanic cones to ocean plates at a planetary scale. Using this similarity relation, we compare Etna volcano to Tharsis Rise on Mars, showing that, as on Earth, the derived similarity holds also on Mars from large spreading volcanoes like Olympus Mons to the whole Tharsis province. We conclude that the definition of a volcano by Borgia et al. (this volume) could be extended also to include spreading crustal structures at a planetary scale.

A SIMPLE RELATION FOR VOLCANIC SPREADING

The topographic load of a volcano (subscript v) may induce flow in a layer of ductile basement (subscript d) under a brittle upper layer (subscript b). Following Carena et al. (2000), we may describe this flow by means of the two-dimensional, rectangular-coordinated, lubrication approximation of the Navier-Stokes equations as (see Fig. 1):

$$\left\{\begin{array}{lll} \dfrac{\partial v_x}{\partial x}+\dfrac{\partial v_z}{\partial z}=0 & \text{- conservation of mass} & \text{(1a)} \\ -\dfrac{\partial p}{\partial x}+\mu_d\dfrac{\partial^2 v_x}{\partial z^2}=0 & \text{- conservation of x-momentum} & \text{(1b)} \\ -\dfrac{\partial p}{\partial z}-\rho_d g=0 & \text{- conservation of z-momentum} & \text{(1c)} \end{array}\right\},$$

where x is the horizontal coordinate, and z is the vertical (positive upward) coordinate; v_x and v_z are the velocities in the x and z directions, respectively; p is the hydrostatic pressure within the layer of ductile basement, and g is the acceleration of gravity. ρ_d and μ_d are, respectively, the density and the Newtonian viscosity of the ductile layer. In this equation we have neglected, to a first order approximation, the flexural stiffness of the crust and basement. We also consider the crust, below the basement, to be rigid relative to the ductile basement. The boundary conditions for equation 1a are:

$$v_z\big|_{z=0}=0, \qquad \text{(2a)}$$

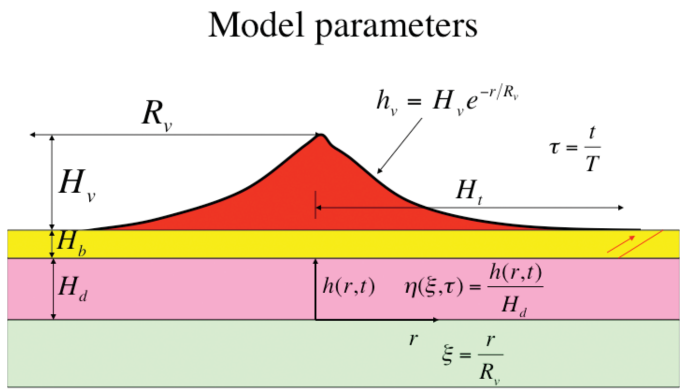

Figure 1. Diagram of a spreading volcano, with variables used in the text.

$$v_z\Big|_{z=h_d}=\int_{v_z|_{z=0}}^{v_z|_{z=h_d}}\partial v_z=\frac{\partial h_d}{\partial t}, \qquad \text{(2b)}$$

where $h_d = h_d(t,x)$ is the thickness of the ductile layer, a function of time (t) and x. These boundary conditions state that the ductile layer adheres at all times to the lower boundary (the crust) and that the vertical velocity at the top of the ductile layer describes the time and space evolution of the interface between ductile and brittle layers (free-surface kinematic boundary condition).

The boundary conditions for equation 1b are:

$$v_x\big|_{z=0}=0, \qquad \text{(3a)}$$

$$v_x\big|_{z=h_d}=0 \ \text{(3b1)} \quad \text{or} \quad \frac{\partial v_x}{\partial z}\Big|_{z=h_d}=0. \ \text{(3b2)}$$

The first of these two boundary conditions implies that the ductile layer is at all times attached to (i.e., does not slip over) the crust. The second of these conditions can be set so that the brittle layer may restrain the flow of the ductile layer below (equation 3b1), thus balancing the applied viscous stresses (τ_{xz}):

$$\tau_{xz}\big|_{z=h_d}=-\mu\frac{\partial v_x}{\partial z}\Big|_{z=h_d} \qquad \text{(3b1a)}$$

by elastic compression within it. Alternatively, the brittle upper layer may be passively carried on top of the ductile layer, resulting in an inviscid type of boundary condition (equation 3b2) and in the upper layer extension:

$$\tau_{xz}\big|_{z=h_d}=0. \qquad \text{(3b2a)}$$

It is important to point out that the interface between the lower and upper layers can take any condition between the two extreme sets by equation 3b. The boundary condition for equation 1c is:

$$p\big|_{z=h_d}=\rho_b g H_b+\rho_v g h_v+P_0, \qquad \text{(4)}$$

where ρ_b and H_b are, respectively, the density and the thickness of the upper layer; ρ_v and $h_v = h_v(x)$ are respectively the density and the topographic relief of the volcano above the surrounding areas, and P_0 is the atmospheric pressure.

Integration of equation 1 leads to:

$$\begin{cases} \dfrac{\partial h_d}{\partial t} = -\dfrac{\partial}{\partial x}\displaystyle\int_{z=0}^{z=h_d} v_x \partial z & \text{- conservation of mass} \quad (5a) \\ v_x = \dfrac{\partial p/\partial x}{2\mu_d}\left(z^2 - shz\right) & \text{- conservation of x-momentum} \quad (5b) \\ p = \rho_d g(h_d - z) + \rho_b g H_b + \rho_v g h_v + P_0 & \\ & \text{- conservation of z-momentum} \quad (5c) \end{cases},$$

where s = 1 if the first boundary condition (equation 3b1) is applied, or s = 2 if the second boundary condition (equation 3b2) is used. Substituting equations 5b and 5c into equation 5a leads to:

$$\frac{\partial \eta_d}{\partial \tau} = \frac{\partial}{\partial \xi}\left\{\eta_d^3 \frac{\partial}{\partial \xi}\left[\eta_d + \left(\frac{\rho_v H_v}{\rho_d H_d}\right)\eta_v\right]\right\}. \tag{6}$$

In equation 6 we used the following dimensionless scaled variables:

$$\xi = \frac{x}{R_v},\ \eta_d = \frac{h_d}{H_d},\ \eta_v = \frac{h_v}{H_v},\ \tau = \frac{t}{T}, \tag{7}$$

and we have set

$$T = \frac{3c\mu_d R_v^2}{\rho_d g H_d^3}. \tag{8}$$

R_v and H_v are the maximum radius and original topographic relief above the surrounding areas of the volcano, respectively. H_d is the original thickness of the ductile layer.

The value of c depends on the boundary condition between the brittle upper layer and the lower ductile layer; that is, c = 4 if the upper layer restrains the flow (using the first boundary condition, equation 3b1), and c = 1 if the upper layer does not restrain lateral horizontal flow (using the second boundary condition, equation 3b2). The boundary conditions necessary to solve equation 6 are defined at the origin, where there is no horizontal flow, that is

$$\frac{\partial \eta_d}{\partial \xi}\Big|_{\xi=0} = 0, \tag{9a}$$

and at large distances from the volcano, where η remains constant over time, that is

$$\eta_d\Big|_{\xi=\infty} = 1. \tag{9b}$$

The initial condition is assumed to be a constant thickness of the ductile layer at the beginning of the deformation, that is

$$\eta_d\Big|_{\tau=0} = 1. \tag{9c}$$

A solution to equation 6 was applied to the gravitational spreading process at Vesuvius by Borgia et al. (2005).

Following equation 4, the outward flow of the ductile layer seen at the beginning of the deformation will drag the overlying brittle layer, creating a compressive stress (σ_{xx}) within it that accumulates, increasing away from the origin, and must be balanced in the far-field by fixed boundaries (Borgia et al., 2000a; Kilty, 2000). Assuming that this stress is a reasonable approximation of the maximum principal stress σ_1, we may write

$$\sigma_1 \approx \sigma_{xx} = -\frac{1}{H_b}\int_{\vartheta=0}^{\vartheta=x} \tau_{xz}\Big|_{z=h_d} \partial\vartheta, \tag{10}$$

where ϑ is a dummy variable. If these stresses are high enough, the brittle layer may rupture at a distance L from the origin, according to the Coulomb fracture criteria with no cohesion, according to the relation

$$\frac{\sigma_1}{\sigma_3}\Big|_{x=L} = N_\phi = \frac{1+\sin\phi_b}{1-\sin\phi_b}, \tag{11}$$

where ϕ_b is the angle of internal friction of the brittle layer; σ_3, the minimum principal stress, can be approximated by the lithostatic pressure in the brittle layer (σ_{zz}) as:

$$\sigma_3 \approx \sigma_{zz} = \rho_b g H_b + \rho_v g h_v + P_0. \tag{12}$$

In the approximations made to write equation 1, the shear stress at the base of the brittle layer is given by equation 3b1a. Thus, substitution of equations 5b, with s = 1, and 5c for the case of $z = h_d$ into this equation, leads to

$$\tau_{xz}\Big|_{z=h_d} = \frac{\rho_v g}{2} h_d \frac{\partial h_v}{\partial x}. \tag{13}$$

Therefore, at the beginning of the deformation when $h_d = H_d$ is independent of x, equation 10 solves to

$$\sigma_1 \approx \frac{\rho_v g}{2H_b} L\left[h_v\right]_{\vartheta=0}^{\vartheta=x} = \frac{\rho_v g H_v}{2}\left(\frac{H_d}{H_b}\right)\left(1-\eta_v\right). \tag{14}$$

It is interesting to note that at $x = \infty$ the maximum horizontal stress is comparable to the lithostatic pressure at the base of the volcano. Substituting equations 10, 13, and 14, for $x = L$, into equation 11, neglecting P_0, which is small compared with the other terms, and solving for L, we finally obtain

$$L = R_v\left[\ln\frac{1+2N_\phi \dfrac{H_b}{H_d}}{1-2N_\phi\left(\dfrac{\rho_b}{\rho_v}\right)\dfrac{H_b^2}{H_d H_v}}\right]^{\frac{1}{2m}} \approx R_v. \tag{15}$$

Within the limits of the assumptions made above, the upper brittle layer may fail close to the peripheral base of the original volcano (that is, at a distance from the axis comparable to the radius of the volcano).

We also observe that in equation 6, for the term containing the volcanic topography to have a significant influence in the flow of the ductile layer, such as in the case of small spreading volcanoes, it must be:

$$\left(\frac{\rho_v H_v}{\rho_d H_d}\right) = O(1), \tag{16}$$

that is

$$H_v = \left(\frac{\rho_d}{\rho_v}\right) H_d \Rightarrow H_v \approx H_d = H. \tag{17}$$

Therefore, solving equation 8 for H_d, after substitution of equation 15 and of equation 17 where relevant, leads to:

$$H = \sqrt[3]{\frac{3\mu_d}{\rho_d g T}} L^{\frac{2}{3}}. \tag{18}$$

Figure 2 is a log-log plot of equation 18 for terrestrial and Martian volcanoes. The data for the terrestrial volcanoes and ocean plates, for which the set is best constrained (Table 1), fall on a straight line with an angular coefficient given by:

$$\sqrt[3]{\frac{3\mu}{\rho g T}} \approx 4, \tag{19}$$

and a correlation coefficient $R^2 = 0.996$. The reason for the straight line is that the angular coefficient (equation 19) is a cube root, so it is fairly insensitive to variations in the parameters contained in the root. Furthermore, among these parameters, only the viscosity of the ductile layer and the spreading time may have significantly different values, but because a higher viscosity will typically correspond with a longer characteristic spreading time, and since these two parameters are the numerator and denominator of the same fraction, their variations will tend to cancel each other out.

TABLE 1. GEOMETRIC PARAMETERS OF VOLCANOES USED IN FIGURE 2 AND THEIR ERRORS

Name of volcano	L (m)	±σ(L) (m)	H (m)	±σ(H) (m)
Earth				
Maderas*	7.00E+03	1.00E+03	1.39E+03	1.00E+02
Concepciòn E*	7.00E+03	5.00E+02	1.61E+03	1.00E+02
Concepciòn W*	8.00E+03	5.00E+02	1.61E+03	1.00E+02
Poàs N†	1.50E+04	1.00E+03	2.30E+03	2.00E+02
Poàs S†	1.20E+04	1.00E+03	1.80E+03	2.00E+02
Etna S§	2.80E+04	2.00E+03	3.30E+03	3.00E+02
Etna E§	3.30E+04	3.00E+03	4.80E+03	5.00E+02
Kilauea#	6.50E+04	5.00E+03	6.50E+03	1.00E+03
Mauna Loa#	9.70E+04	1.00E+04	9.50E+03	1.00E+03
Atlantic**	3.30E+06	1.00E+05	1.00E+05	2.00E+04
Pacific E**	1.60E+06	1.00E+05	5.00E+04	1.00E+04
Pacific W**	8.30E+06	3.00E+05	1.50E+05	2.50E+04
Mars				
Tharsis Tholus max††	7.50E+04	1.00E+04	6.00E+03	1.00E+03
Tharsis Tholus min††	6.00E+04	1.00E+04	6.00E+03	1.00E+03
Olympus Mons NW†	4.00E+05	5.00E+04	2.30E+04	2.00E+03
Olympus Mons SE†	3.00E+05	5.00E+04	1.80E+04	2.00E+03
Tharsis Rise§§	3.90E+06	5.00E+05	1.00E+05	2.00E+04

*Van Wyk de Vries and Borgia (1996).
†Borgia et al. (1990).
§Borgia et al. (1992).
#Borgia and Treves (1992).
**Chapman and Pollack (1977).
††Borgia et al. (2000a).
§§Esposito et al. (1992); Scott and Tanaka (1986).

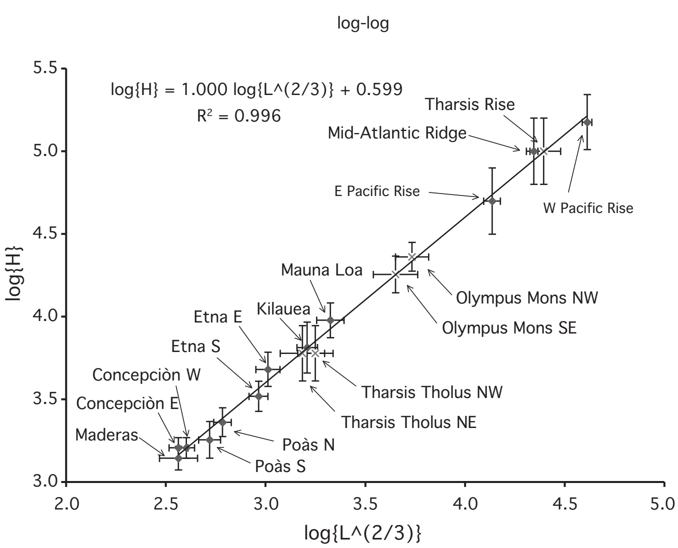

Figure 2. Similarity relation for spreading volcanoes. Filled rhombs are terrestrial spreading volcanoes, and ×'s are Martian ones. Data sources are given in Table 1. See text for details.

The validity of the scaling relation given by equation 18 indicates that large spreading volcanoes can indeed be considered good analogues of mid-ocean-ridge dynamics. Of course, there are specific differences that arise because the ridges are larger in absolute size. But there are many more differences in magnitude and topology between cinder cones and large ocean shield volcanoes, such as the Hawaiian volcanoes (Delaney et al., 1998), than between ocean shields and mid-ocean ridges.

On Mars, volcanic spreading has been suggested to occur in the relatively small volcano Tharsis Tholus (Borgia et al., 2000a) and the large Olympus Mons (Francis and Wadge, 1983; Borgia et al., 1990; McGovern et al., 2007), both of which are on Tharsis Rise. These volcanoes also plot close to the terrestrial trend (Fig. 2).

The relation expressed by equation 18, which is derived from first principles (i.e., conservation of mass and momentum), supports the Borgia et al. (this volume) proposition for a holistic definition of *volcano*. In this definition the emphasis is set on process and on the whole volcanic environment rather than on any specific part of it.

APPLICATION TO THARSIS RISE

As equation 3 appears to apply to the whole realm of spreading, we suggest that Etna Volcano and its well-studied structures may be an acceptable scaled-down terrestrial analogue for Tharsis Rise on Mars. A similar assumption was made by Borgia et al. (2000b) and Webb and Head (2002).

In observing Figure 2, one may question why we compare Tharsis Rise to Etna Volcano and not directly to terrestrial ocean ridges, which plot much closer to Tharsis. The reason concerns the topology of their spreading dynamics. As we will show in the following sections, the periphery of Tharsis is obducted over the Martian crust. This makes Tharsis more similar to Etna, whose periphery is thrusted over the surrounding terrains, than to mid-ocean ridges, at the periphery of which the ocean crust is generally subducted at convergent plate margins. Another reason is that although Mount Etna can unquestionably be called a volcano, considering a mid-ocean ridge a volcano might be more debatable. Thus, it may be easier to accept that Tharsis Rise can effectively be considered a volcano, once we can show that a scaling analogy exists between the geometric features of Mount Etna and Tharsis Rise.

Tharsis Rise

Tharsis Rise is 7000 by 7000 km in maximum dimensions (Scott and Tanaka, 1986; Fig. 3). It covers the border between the northern lowlands and the southern highlands from about +60°

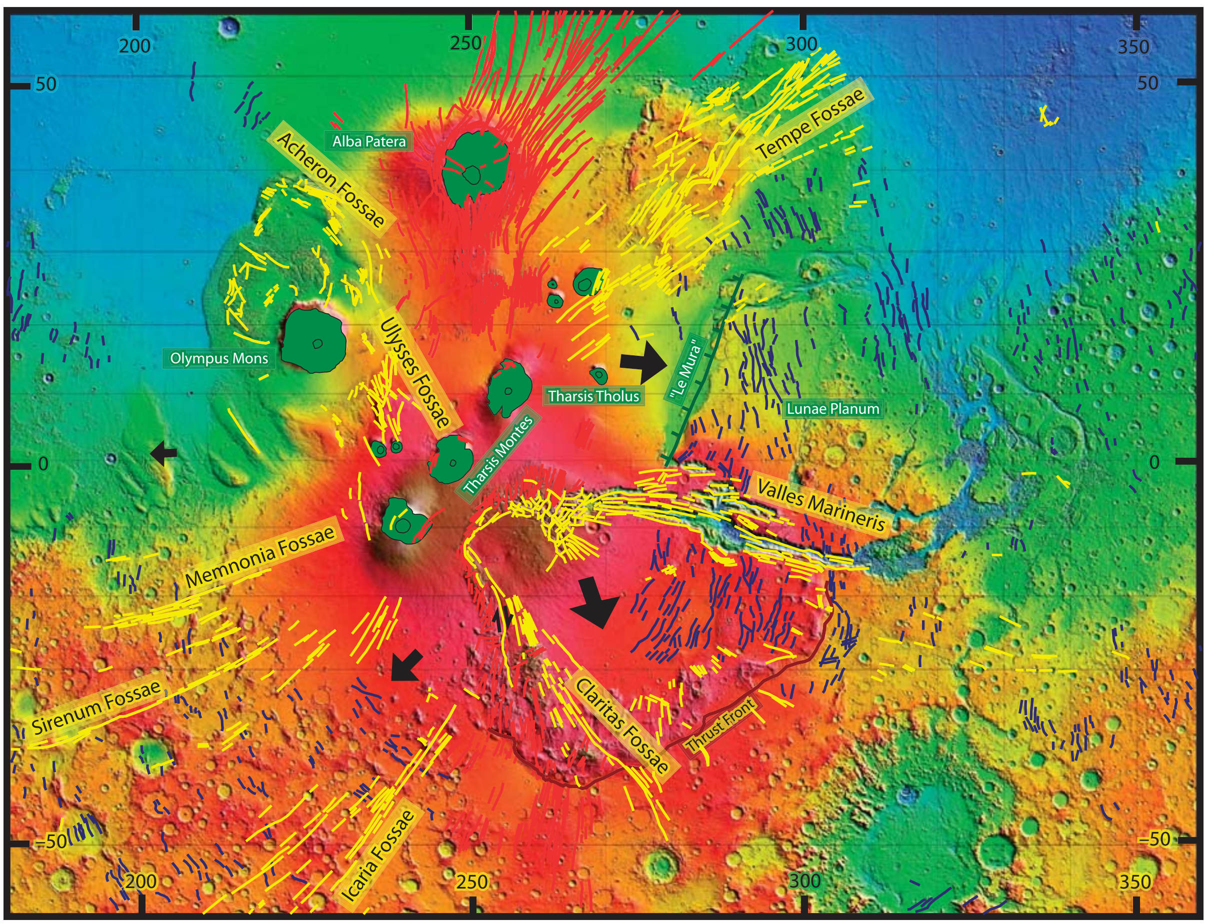

Figure 3. Topography of Tharsis Rise (MOLA, 1999) and geology and structure of Tharsis (after Scott and Tanaka, 1986). Note the N-S major rift system that cuts Tharsis Rise (in red), covered by the deposits of the Tharsis Montes. The various fossae (in yellow; Tempe trending NE, Valles Marineris, the largest, trending ESE, Claritas trending SSE, Icaria trending SSW, Sirenum and Memnonia trending WSW, and Ulysses and Acheron trending NW) are radial to the Rise. The peripheral tangential fold belt (in blue) is particularly evident to the east of the Rise. The Syria "plate" appears to slide SSE, forming a major thrust front (brown). It also forms a left-lateral transtensional crustal fault system that is the origin of Valles Marineris, and the right-lateral transpressional crustal system that forms the Claritas Fossae. A normal fault (dark green) seems to uplift the Lunae Planum from the Tharsis slope and is similar to "Le Mura" faults of Amiata Volcano, Italy, where spreading occurs on a particularly thick underlying ductile layer (Delcroix et al., 2006).

to −60° latitude and from 40° to 160° west longitude (just less than one-third of the total surface of Mars). The central part of the Rise (excluding volcano summits) stands >10,000 m above the northern lowlands and >4000 m above the southern highlands (MOLA, 1999). Surmounting the Rise are the NE-trending Tharsis Montes volcanoes, some of which reach above 22 km in elevation. A major system of N-S–trending grabens that cut through the middle of the Rise, covered in the center by the Tharsis Montes deposits, measures 500–1000 km wide and is 7000 km long. These grabens are highly disrupted with a ridge-groove topography (Banerdt et al., 1992). Other systems of grabens, 50–100 km wide and 1500–2500 km long, tend to be radial to Tharsis Rise, forming many of the fossae (Banerdt et al., 1992). Some of them, such as Tempe Fossae, Valles Marineris, and Claritas Fossae, resemble complex terrestrial rifts and seem to cut through the entire Martian lithosphere (Banerdt et al., 1992). The periphery of the Rise, particularly in the NE-SE and the SW, is occupied by tangential systems of folds and wrinkle ridges that appear to be related to thrust faulting of the crust (Banerdt et al., 1992). These are interrupted by the radial fossae. Commonly these folds and ridges are asymmetric, suggesting tectonic transport away from the Rise. In some cases, particularly on the SE periphery of the Rise, the ridges achieve substantial topographic relief (>3000 m above the surrounding areas); there may have been similar high ridges to the NE and NW, now hidden beneath younger deposits. The tectonic history of Tharsis suggests that the N-S rifting, the radial faulting, and the tangential folding are all associated with the uplift and evolution of the Rise (Banerdt et al., 1992).

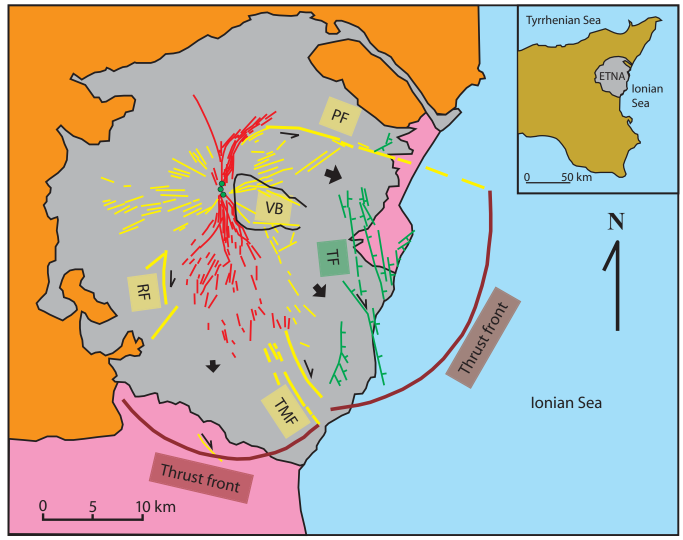

Figure 4. Map of major geological units and structural features of Mount Etna (after Borgia et al., 1992; Merle and Borgia, 1996). The edifice of Etna (gray) is spreading on top of the ductile sub-Etnean clays (pink), driven by its weight and the creeping intrusive complex. Note the basal anticlines (solid brown lines) that run along the base of Etna from SSW to NNE. The rifts (red lines) and the radial fissures and faults (yellow lines) are the loci of most eruptions. The Valle del Bove (VB) results from massive erosion of a summit leaf graben, Pernicana (PF), Trecastagni-Mascalucia (TMF), and Ragalna (RF) are transtensional fault systems. The Timpe normal faults (TF, green lines) appear related to the Malta Escarpment.

Etna Volcano

Mount Etna is a complex volcano in eastern Sicily (Italy), ~40 km across and ~3300 m above sea level. Three main structural features characterize the edifice and the geodynamics of Etna (Fig. 4): (1) the volcanic rift system that crosses the summit of Etna from SSE to NNE (Garduño et al., 1997), which is covered in the highest central section by summit volcanic cones (Chester et al., 1985); (2) the transtensional fault systems of Pernicana (Azzaro, 1997), Trecastagni-Mascalucia (Lo Giudice and Rasà, 1992), and Ragalna (Rust and Neri, 1996) that connect the summit rift to (3) the peripheral compressional belt surrounding the base of the volcano from ENE to SSW.

These physical boundaries identify three main sectors on the volcano (Borgia et al., 1992). Two sectors spread laterally on a basal décollement in approximately eastward and southward directions with rates in the order of 1–2 $*10^{-2}$ m a^{-1}. The third sector tends to be stable, being buttressed by the Maghrebian-Apennine Chain. The spreading process appears to be controlled by the load of the volcano on the clayey substratum and the creeping extension of the intrusive complexes (Borgia et al., 1992), which lie at ~5 km depth (Hirn et al., 1997). The summit rift extension, resulting from spreading, is compensated by shortening across the peripheral compressional belt that forms a festoon of anticlines. The tangential extension owing to the radial outward motion is mostly compensated within the transtensional fault systems.

Comparing Etna with Tharsis Rise

In spite of the difference in dimensions (Etna is ~200 times smaller than Tharsis) the geometry and topology of the structures that characterize the surface of Mount Etna closely resemble those found on Tharsis Rise: the SSE-N-NNE summit rift system of Etna is analogous to the major N-NNE rift system of Tharsis, the transtensional fault systems on Etna are analogous to the radial fossae, and the peripheral compressional belt at the eastern and southern base of Etna is analogous to the belt of folds and ridges that surrounds the base of Tharsis.

In order to compare Etna with Tharsis, we have developed a set of dimensionless parameters concerning the relations between geometric variables of characteristic volcanic features found in both areas (Table 2). We have not developed similar dimensionless parameters for the dynamic evolution of these features, because rates of deformation are unknown for Mars. From our comparison of these geometric dimensionless parameters, we suggest that more than a simple morphological analogy exists. In fact, all corresponding dimensionless ratios are essentially equal, apart from the total width of the peripheral fold belt for Tharsis, which is relatively wider than for Etna. This difference is smaller if the wavelength

TABLE 2. GEOMETRIC SIMILARITY BETWEEN THARSIS RISE (DATA FROM SCOTT AND TANAKA, 1986) AND MOUNT ETNA (BORGIA ET AL., 1992)

Structural element	Tharsis Rise	Mount Etna	Ratio (Tharsis/Etna)
Summit rift			
Length (km)	7000	20	350
Width (km)	750	2	375
Length/width	9.3	10	
Radial fault systems			
Length (km)	2000	30	67
Width (km)	75	1	75
Length/width	26	30	
Peripheral folds			
Length (km)	2000	35	57
Max. wave length (km)	150	3	50
Width (km)	1500#	5	300
Length/max. wave length	13.3	11.7	
Length/width	1.3	7	

#Including all trains of peripheral folds.

of the folds is considered instead of the total width of the fold belt. This fold belt on Tharsis is made of trains of folds, similar to terrestrial duplex structures, instead of being composed of only one major fold, as on Etna. This difference could perhaps be a result of the brittle crustal layer of Tharsis, which may be relatively thinner than for Etna (Van Wyk de Vries and Borgia, 1996).

Other differences in the scaling parameters (Table 2) are the N-S rift of Tharsis that is ~360 times larger than the Etna rift, whereas the lateral faults and peripheral folds are only 60 times larger. This fact suggests that the Tharsis rift has a relatively deeper source than Etna; i.e., the Tharsis rift could be controlled by processes occurring in the martian mantle instead of by the spreading of intrusive complexes, as on Etna.

Following the analogy, the fossae may behave as transtensional fault systems that decouple the various sectors of the Rise with different spreading velocities and accommodate the tangential extension by radial outward motion of the sectors. In places there is evidence of sub-radial transpressional fault systems, such as the Gordii Dorsum (Forsythe and Zimbelman, 1988) and the area south of the eastern end of Valles Marineris (Shultz, 1989), both with left-lateral motion, or the distal end of Claritas Fossae (Fig. 3). All these strike-slip motions are consistent with tectonic stress and deformation that is directed radially away from Tharsis. They are also consistent with the model proposed here. In addition, compression across the basal tangential fold-and-ridge systems compensates the summit extension of the major N-S rift and the radial spreading of the flanks of the Rise. Finally, as the Maghrebian-Apennine Chain, which deepens eastward and southward, buttresses the NW and W flanks of Etna, perhaps the top of Mars' mantle that slopes southward (Zuber et al., 2000) could buttress Tharsis Rise, inhibiting spreading toward the N and NW.

ANALYSIS AND CONCLUSIONS

The base of the Martian lithosphere, which corresponds to the brittle-ductile transition, is estimated to lie at ~80–100 km depth below Tharsis (Esposito et al., 1992; Zuber et al., 2000). From the scaling similarity shown in Figure 2, the spreading of Tharsis requires that this full lithospheric thickness be involved. Thus the spreading sectors of Tharsis appear to be like terrestrial lithospheric plates that spread outward from the Rise away from the major N-S rift system. One of these, the SE plate, seems to be an immense crustal plate that slumps southeastward, extending Noctis Labyrinthus and Valles Marineris, and compressing the distal part of Claritas Fossae (Fig. 3), where it is obducted over the Martian crust, as opposed to subducted under it, as occurs in the case of terrestrial plates.

We may also derive from equation 19, using average parameters for terrestrial gravity and crustal density, the characteristic time of deformation for terrestrial mid-ocean-spreading ridges as follows:

$$T \approx 1.5 \times 10^{-6} \mu . \qquad (20)$$

Assuming a viscosity of the asthenosphere in the order of 10^{19}–10^{20} Pa s (Turcotte and Schubert, 1982), the characteristic time of spreading for terrestrial ocean plates is in the order of 1–10 Ma. That is, a value 10–100 times smaller than the maximum age of plates.

Reversing the argument and using average parameters for the Martian gravity and crustal density (Esposito et al., 1992; Zuber et al., 2000), equation 19 leads to

$$\mu \approx 2 \times 10^{5} T . \qquad (21)$$

Tectonic activity on Tharsis is recorded from the Early Noachian, dated at ca. $4*10^3$ Ma (Tanaka et al., 1992). This value corresponds to 10–100 times the maximum time scale for spreading on Tharsis. It follows that the viscosity at the bottom of the Martian lithosphere below Tharsis could well be between 10^{20} and 10^{21} Pa s, a value similar to that suggested by Turcotte and Schubert (1982). This value is only slightly larger than the viscosity of the terrestrial asthenosphere, but lower than the value of 10^{22} Pa s discussed, but rejected, by Zuber et al. (2000).

The low viscosity value derived above for the Martian lithosphere suggests that the tectonic processes on Tharsis could perhaps still be active. Therefore, we conclude that, just as mid-ocean ridges may be pictured as extremely "large and flat" volcanoes, Tharsis Rise perhaps could be an actively spreading volcano itself.

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Some challenging new perspectives of volcanology

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ABSTRACT

As a result of the rapid advances in technology experienced during the last half of the twentieth century, methods used to study volcanoes have diversified enormously. Such diversification has created the illusion of a rapid advance in volcanology. Indeed, it is undeniable that the types of information and the amount of data available for studying diverse aspects of volcanoes have increased in the last decades, and are very likely to continue their growth in the near future. All of this information, however, does not seem to have solved many of the fundamental questions related to volcanic processes, not only in our planet, but in other parts of our solar system as well. Actually, the rapid development of new technologies and methods of measurement may have had a negative impact in the development of volcanology as a whole, because the super-specialization favored by the development of new technologies and methods has also limited communication among volcanologists with varying orientations. As a consequence, some of the fundamental problems in volcanology have been obscured by the quest for ever increasing accuracy of specific variables at the expense of deeper understanding of the general context. In this chapter we address this situation by identifying some of the challenging topics that, in our opinion, need to be explored in more detail by the community of volcanologists as a whole. Aspects such as the long-term evolution of composite volcanoes, volcano failure, and the internal structure of volcanoes are included in this list. An effort is made to formulate specific questions that can be solved in each of these sub-branches of volcanology, not by claiming the need for more specialized data, but rather by keeping an eye on the general context in which these problems occur. Thus, the topics examined in this work provide specific examples of how we envisage a solution for the biggest challenge of volcanology: fostering the communication among workers with different specializations with the goal of solving questions that are of a general nature.

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INTRODUCTION

Volcanology is a rapidly growing science. Its advance has been substantial particularly during the last few decades. By adopting the techniques and procedures of analogue and numerical modeling, volcanologists have gained powerful tools that allow them new and unprecedented insight into volcanic phenomena. Further, volcanology has also become an experimental science and has expanded into the extraterrestrial environment, where it now contributes substantially to the field of planetary sciences. Ever more sensitive volcano-monitoring instruments and techniques have been developed, and the amount of data related to individual volcanoes has grown at a rapid pace. As a consequence, volcanology at the turn of the twenty-first century has diversified enormously, becoming much more interdisciplinary. Numerous "new branches" have developed, permitting interdisciplinary research to be promoted in concert with fields such as social sciences, medicine, or history. Such interactions were unthinkable only a few decades ago. Undoubtedly, such a level of activity reveals a field of science that is healthy and very much alive.

Growing knowledge, in general, results in an increasing number of questions. These questions may promote still new approaches, whose results may provide even newer avenues of research. Amidst this rising spiral of knowledge, however, we find that theoretical and conceptual issues in volcanology may be lagging behind. Although some attempts have been made recently, new syntheses are urgently needed. The *Encyclopedia of Volcanoes* (Sigurdsson et al., 2000), for instance, was intended to be an up-to-date collection of volcanological knowledge at the turn of the millennium. Nevertheless, because of the multitude of authors involved in its writing, the *Encyclopedia* does not offer a uniform theoretical oversight of the discipline. Two other examples are the books *Volcanic Processes* (Dobran, 2001) and *Fundamentals of Physical Volcanology* (Parfitt and Wilson, 2008). Both of these books, however, despite inclusion of a summary of the basic physics of volcanic phenomena, fail to provide a synthesis of the conceptual developments that have occurred in the past few decades.

Somewhat paradoxically, the rapid advance of technology applied to volcanic studies has focused on the development of new methodologies but has lost sight of the really fundamental questions within the field, thus postponing an assessment of the relevance of more precise measurements relative to the understanding of fundamental problems. Such introspection, if done self-critically, reveals that despite all of the new developments, volcanologists as a group may have not advanced much in understanding the very fundamental questions in our field of study. Furthermore, we may discover that sometimes we are moving in the wrong direction by inadvertently increasing confusion within the field as increasing specialization forces creation of new subdisciplines that eventually tend to disrupt communication between scientists.

The dangers of the rapid super-specialization occasioned by recent developments in volcanology are captured by the words of John Platt (1964), "Beware of the man of one method or one instrument, either experimental or theoretical. He tends to become method-oriented rather than problem-oriented. The method-oriented man is shackled" (p. 351). The shackling imposed by each method results in a community that, as a whole, starts to lose sight of the most important questions of the field, and although there are no guarantees that a problem-oriented scientist can find the correct answers every time, "at least [he is] reaching freely toward what is most important" (Platt, 1964, p. 351). Thus, it is no exaggeration to state that the rapid growth of volcanology might result in important conceptual losses if the community of super-specialized volcanologists does not make an effort to put aside its preferred methods and learn new ones with a fresh perspective.

Particularly, super-specialization might cause us to overlook the fundamental role played by qualitative constraints imposed by causality in a general context, uncritically favoring the constraints imposed by any quantitative method. Actually, many qualitative issues are sometimes dismissed as being unscientific, or at best as being part of the "philosophy of science." This label commonly reflects the belief that philosophical matters are not important at all, because "philosophy" contains no mathematics and consequently is not a true science. Whereas it is clear that an important component of modern science is its quantitative side, it is wrong to conclude that any quantitative approach is a correct one, for the result might be irrelevant for the particular phenomenon of interest. In the words of Albert Einstein, "So far as the laws of mathematics refer to reality, they are not certain, and so far as they are certain, they do not refer to reality" (Einstein, 1921, p. 385–386, our translation). Finding the correct equilibrium between quantitative and qualitative approaches is not a simple task. Clearly such equilibrium cannot be reached by systematically denying the importance of any qualitative approach that departs from the currently accepted paradigm, even if such paradigm does not provide a good explanation for some of the quantitative observations.

As readers may have noted, a philosophical theme underlies all of the previous chapters of this book. As a comparison of several of these chapters will reveal, striving to define what a volcano is not a simple task. Actually, when considering such a fundamental question, it is almost impossible to avoid some "philosophical thinking." Whether we as editors have succeeded in conveying the importance of such philosophical matters to the community of volcanologists is uncertain, but we are certain that this book attempts to make this point evident to our community. In this final chapter of the book we briefly address a number of additional subjects that might be considered to be somewhat "philosophical" in nature. Some of these subjects emerge from the development of already "accepted" ideas (e.g., volcanic facies and volcano instability). Others have already been enunciated elsewhere, although they seem to have remained largely ignored (e.g., long-term evolution of volcanoes viewed as self-organized critical phenomenon). We do not claim that the ideas developed in this chapter are definitive solutions to pending problems in volcanology. On the contrary, we intend to present them a challenge by suggesting possible new approaches to fundamental questions in the general field of volcanology. Actually, some of these ideas have received so little attention that they are more easily discussed over a few pints of beer than presented in a traditional scientific form. Nevertheless, all of these ideas are challenging and thought-provoking. If reading these ideas becomes a stimulating intellectual task, we will have accomplished our goals.

LONG-TERM EVOLUTION OF COMPOSITE VOLCANOES

Volcanic hazard studies commonly address the problem of reconstructing volcanic evolution and history of the most hazardous volcanoes on Earth: composite volcanoes. A major problem faced when attempting to reconstruct the volcanic history of composite volcanoes, however, is their long-lived nature; they have active lifetimes of 10^5 to 10^6 yr. Such a long history can be divided into a recent segment, documented by both historical and geological records, and an ancient segment, documented only through the geological record. The record in both recent and ancient segments is usually incomplete, but such incompleteness is time-dependent: the older the record, the more incomplete it is. Consequently, the history of volcanic events has an uneven reliability throughout the entire evolution of a particular volcano.

Some volcanic hazard assessments may not include the long-term evolution of a volcano, focusing rather on its most recent history (e.g., Houghton et al., 1987). Although this approach might facilitate coping with the most immediate dangers, a deeper understanding of the hazards, including the long-term history posed by a particular volcano, is highly desirable in most cases because such a record might reveal the occurrence of infrequent events that otherwise would be overlooked. In general, the more complete and better understood the volcanic history of a particular volcano is, the more reliable forecasts on its future behavior can be elaborated. Nevertheless, knowledge of the history of any volcano does not constitute proof that the volcano will behave cyclically (i.e., the fact that a given event took place in the past does not guarantee that a similar event will take place in the future) nor does it provide clues warning of the possible occurrence of first-time events. Consequently, any forecast of eruptive behavior will always have an inherent uncertainty, although the form in which such uncertainty is dealt with differs among the volcanological community. In this section we suggest a possible way to optimize information according its relevance to (1) the long-term evolution of the volcano, and (2) volcanic hazard assessment.

Conceptualizing Volcano Evolution

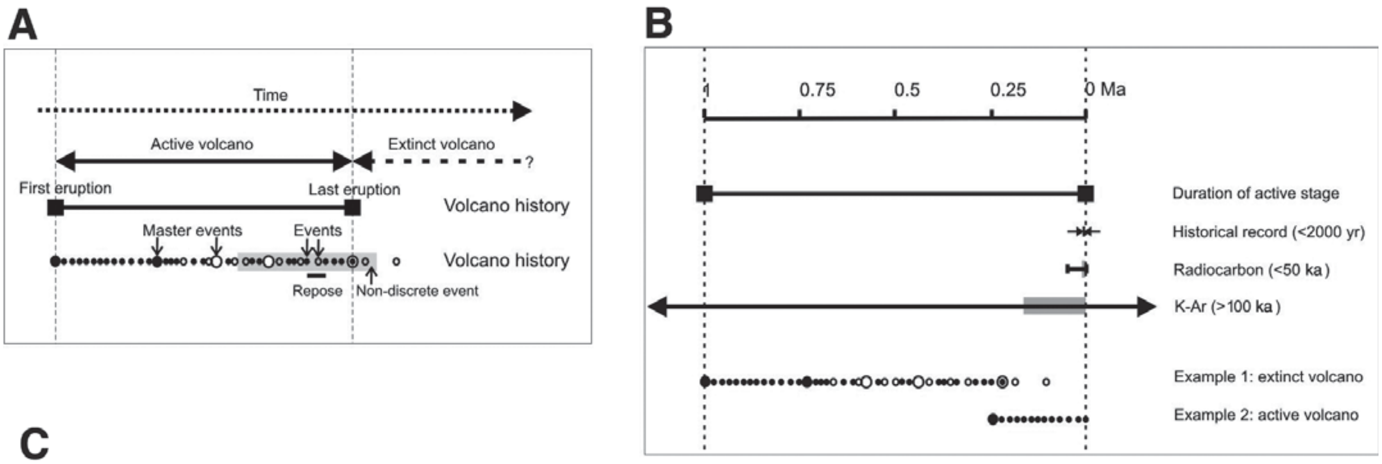
Let us first define the term *volcano evolution.* The simplest definition would be: “The history of events that have occurred in the past and which could happen in the future of a volcano.” Thus, we can look at volcanic evolution as a sequence of two types of events: “eruptive” and “non-eruptive.” Commonly the first and the last eruptive events are considered to mark the “birth” and “death” of a volcano, and the time bounded by these two events is referred to as the active stage. After the last eruption has occurred in a volcano it is said that it has entered an “extinct” stage. However, the volcano does not disappear after its last eruption, and actually it can be the source of further events, many of which can present large threats to nearby populations. Thus, from the point of view of prevention of natural disasters, events taking place within the extinct stage need to be included in the volcano’s history.

In practice, following an eruption, it is difficult to know whether a volcano has entered its extinct stage or not, because there is always the possibility of a new eruption taking place in the distant future. Nevertheless, in the theoretical case examined here, it can be assumed that we can be certain that it has been determined that the last eruption of that particular volcano has occurred, and that therefore the “extinct” stage is well defined.

Figure 1A shows a conceptual model of a possible volcano evolution as a succession of discrete events along the time axis. The eruptive events are marked by solid circles, and the non-eruptive events are marked by open circles. All of these events are separated by intervals in which nothing occurs in the volcano (note that these are not “repose” or “inactive” intervals in the sense commonly given to these last two terms, which are almost exclusively associated with the absence or presence of an eruption, respectively). Importantly, this model assumes that the beginning and end of any event can be determined with confidence, which may not always be the case. For example, eruptions generally can be considered to be discrete events, although accurately defining the beginning and end of an eruption might not be easy. The case of Stromboli is illustrative in this context, because it has been in a state of persistent activity for almost 2000 yr (Rosi et al., 2003). The main activity of this volcano consists of small-intensity explosions occurring at intervals of tens of minutes during which jets of incandescent material are ejected into the air (these explosions define the typical Strombolian eruption). Superimposed on this “normal” activity are more intense events of effusive and explosive nature, with a frequency of ~10 yr. Consequently, it is not clear whether the eruptive activity of this volcano consists of discrete events (separated by intervals of either 10 min or 10 yr), or whether it is better described as a continuous eruption.

The difficulty in identifying the beginning and end of a volcanic event is not only associated with eruptive events but also includes the non-eruptive category. For instance, erosion and processes related to edifice-basement interaction, such as volcano spreading, in general can be considered as non-discrete events. Nevertheless, periods of anomalously high-intensity erosion, and probably events such as edifice collapses, might be better considered as discrete erosional events.

For these reasons, *eruptive events* and *discrete events* are generally not equivalent terms, nor are the terms *erosional events* and *continuous events* equivalent. Thus, it might be useful to distinguish between discrete and continuous events as two categories that are not necessarily identified with eruptive and non-eruptive events. In any case, despite the difficulties of unambiguously defining the meaning of *event* in Figure 1A, the fact remains that a volcanic history can be envisaged as a succession of events in the form there illustrated.



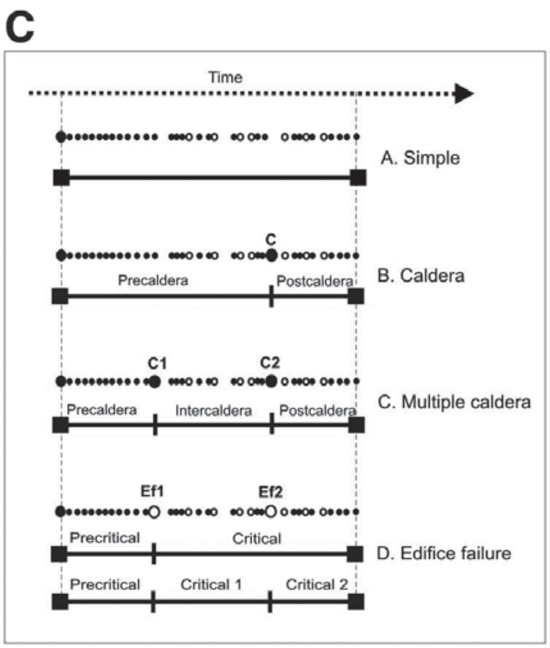


Figure 1. Conceptualization of volcano evolution. (A) Volcano evolution includes both active volcano stage (developed between the first eruption, i.e., volcano "birth," and last eruption, i.e., volcano extinction) and extinct volcano stage (following last eruption). Volcano evolution can be conceptualized as a succession of events, both non-discrete events (e.g., volcano-basement interaction, light-gray shade) and discrete events (circles) of different kinds: eruptive events (filled circles) and non-eruptive events (e.g., flank or edifice failure, open circles). Small circles symbolize normal events, and larger circles symbolize master events. Note that non-eruptive and non-discrete events may continue after the active volcano stage ends. Continuous erosion is not considered here as an "event." (B) Comparison between a time scale of typical composite active volcano evolution and time scales of various dating methods. Note that short-range methods (historic record and radiocarbon) are able to cover but a fraction of volcano history, whereas long-range methods (e.g., K-Ar and Ar-Ar) have error bars (light gray) larger than event densities. Error bars are shown only as an order of magnitude. A typical duration of composite volcano evolution of ca. 1 Ma is considered. (C) Possible ways of dividing active volcano history in stages of evolution. A—simple, one-stage (or non-critical) evolution with no master event; B, C—volcano evolution stages defined by caldera formation master events (C, C1, C2); D—volcano evolution stages defined in two possible versions by edifice failure master events (Ef1, Ef2). See text for additional explanation.

Yet another distinction between events relates to the size and importance of each event. Some events are remarkable, as, for example, a caldera forming event or a volcanic-edifice failure. Such events are noteworthy, not only because of their size but also as a result of their large-scale consequences for the future of a volcano; following such large-size events, eruptive style and behavior usually change. By their profound influence on the entire volcanic system and its environment, these types of events constitute milestones in volcano evolution. As such, they deserve to be clearly distinguished from "common" events. For this reason, the term *master event* is proposed to name them appropriately. In general, the frequency of master events is much lower than that of common events at a particular volcano.

Notably, "non-events" can also be viewed as master events. Long-term periods of quiescence may result in such extensive continuous erosion that significant changes occur to both the volcanic edifice and the subsurface parts of a volcanic system. In these cases, the long quiescence that commonly would be considered a non-event exerts such a strong influence on further volcano behavior that it becomes a master event. Consequently, not all of the master events are discrete, nor are all of the continuous events common, in the sense given above.

In summary, volcano evolution can be viewed as a sequence of eruptive and non-eruptive events, some of which will be discrete events and some of which are non-discrete; all of these are punctuated by some master events amidst a background of common events. This description constitutes one example of the form in which volcano evolution can be conceptualized. Some of the advantages and uses of adopting this conceptualization are described in the next sections.

Reconstructing Long-Term Volcano Evolution

Ideally, one should identify and date all events constituting volcano evolution in order to reconstruct the whole volcanic history. What can be done in practice depends on the available historical and geological record and on the accuracy of the dating techniques used. Because not all events, especially those of small scale (e.g., small-scale ash eruptions or phreatic eruptions) leave behind a recognizable geological record, any ambitious plan to fully reconstruct volcano evolution is unrealistic. The situation is even worse if we consider the problems associated with dating techniques. For example, one problem is that there is not any single technique able to cover the entire time range of active volcanoes. The historic record is uneven and goes back in time to at most two millennia. Most historical records are much shorter than this, and vary in their completeness depending on where the volcano is located. Similar limitations apply to the various radiometric dating methods, because no single method is able to cover the long-term history of a volcano (Fig. 1B). Another problem is related to dating accuracy versus eruptive frequency. For radiometric dating techniques used for old volcanic rocks (in the range of 0.1 to slightly over1 Ma) analytical errors are simply larger than the time gaps between eruptions. Therefore, a number of events, even large ones, cannot be resolved in time because of poor dating resolution. Consequently, these small events are lost from the record. The net result of these shortcomings is that, in general, even in the best cases the oldest segment of volcano evolution is more poorly constrained than its younger segment. Also, it might turn out that the details of various segments of volcanic history are known with different degrees of confidence, thus resulting in a most inhomogeneous volcanic record.

Because of these limitations, we are commonly limited to reconstruction of the "master-event history" of a volcano. Although certainly incomplete (relative to the whole evolution of a volcano as illustrated in Fig. 1A), this master-event history should be sufficient to allow segmentation of volcanic evolution into a number of well-defined evolutionary stages. In principle, each of these stages will be characterized by a predominant behavior that is typical of that stage. Thus, even if the complete record of the history of events within each stage is incomplete, the geological record may provide a reliable characterization of the style and intensity of volcanic activity between consecutive evolutionary stages (Fig. 1C).

For the younger stages we might obtain additional details by examining separately the "event-content" of periods bounded by intra-stage large (or second order) events using the appropriate dating techniques (e.g., radiocarbon for the last ~50 ka). For the youngest segment of active volcano history, all third order events might be identified and dated, allowing an even more detailed history to be reconstructed. Consequently, by explicitly acknowledging the presence of master and continuous events in the history of a volcano, it is possible to construct a hierarchical history that can be used to assess the level of accuracy that can be reached, depending on which part of the history of the volcano is being examined. Such a hierarchical history allows study of the long-term evolution of a volcano using nontraditional volcanological approaches, as described below.

Long-Term Evolution of Composite Volcanic Edifices Viewed as Self-Organized Critical (SOC) Phenomenon

SOC Theory

The physical theory of self-organized criticality (SOC) (Bak et al., 1987) successfully addresses complexity in nature by explaining the behavior of dynamic systems in terms of their spontaneous development toward a critical state as an attractor. In this case the critical state is considered to be an attractor because it is reached regardless of the initial conditions of the system, provided that a long enough time is considered. This critical state is reached because of simple local interactions that may follow very different paths but that nonetheless eventually lead to the same end result. SOC is typically observed in slowly driven nonequilibrium systems with a high level of nonlinearity. Systems showing SOC tend to develop similar patterns on many scales, from the very small to the very large; they emerge from repetitive, nonlinear interactions, which lead to a progressive buildup of large-scale correlations and ultimately to crises (Sornette, 2006).

Using the concepts of SOC, a number of natural and societal phenomena as diverse as solar flares (Podlazov and Osokin, 2002), earthquakes (Bak et al., 2002), landscape formation and landslides (Turcotte and Malamud, 2004), forest fires (Krenn and Hergarten, 2009), epidemics (Rhodes et al., 1997), biological evolution (Bak and Sneppen, 1993), and fluctuations in economic systems (Sornette, 2006) have been modeled so far.

The most common experimental demonstration of SOC in physics is related to the "sandpile model," which is based on a cellular automaton (Bak et al., 1987). In this model the sandpile grows as new grains are added to it until it reaches a critical slope beyond which any further sand grain added may, or may not, trigger "slides" or "avalanches" whose magnitude is limited only by the size of the pile itself. By reaching its critical slope—the slope being the "control parameter" of the sandpile—the system reaches its critical state as well. In the critical state the system responds to any perturbation in a way unpredictable in detail, but it keeps its critical parameter constant (i.e., at the value of the angle of repose corresponding to average sand-grain size) during further growth history. As a result, the property of critical behavior of the sandpile emerges spontaneously, that is, it is "self-organized," and the critical point in the evolution of this system is viewed as its attractor.

Applicability of SOC to Composite Volcanic Edifices

Because of their gradual buildup by continuous addition of new material, and the presence of mechanisms that progressively initiate and amplify instabilities, edifices of composite volcanoes are particularly appealing for consideration in terms of SOC theory (Szakács and Şuţeanu, 1995). In a systemic approach (e.g., Szakács, this volume), volcanic edifices are viewed as subsystems of the larger volcanic system, which can be envisaged as a system with a slow evolution rate possessing a high level of nonlinearity that continuously pushes it beyond the equilibrium stage. Thus, volcanic systems are dynamic open systems displaying a certain structural complexity that emerges during volcanic evolution. Increasing instability accompanies edifice growth (McGuire, 1996; Szakács and Seghedi, 2000), resulting frequently in slope failure events, which are rather the rule than the exception at large mature composite volcanoes such as those in the Central Andes (Francis, 1994). It is thus apparent that these volcanic edifices spontaneously evolve toward a state of instability, leading to failure or toward a critical state. Once criticality is reached, the edifice remains unstable during further volcano evolution, as revealed by repeated flank or edifice failure events.

Beyond these general analogies, volcanic edifices differ notably from sandpiles in a number of ways. First, they are made of more or less coherent volcanic products, unlike sandpiles that are built of loose granular material. Second, unlike sandpiles, no unique control parameter accountable for self-organization of the critical state can be identified at growing volcanic edifices. For instance, concave-upward steady-state slope profiles characterize most composite volcanoes (Davidson and Da Silva, 2000) irrespective of whether they have undergone edifice failure events or not. Rather, volcano instability is a multifactorial feature involving a number of intrinsic and extrinsic instability factors (e.g., Szakács and Seghedi, 2000). Therefore, control parameters related to each instability factor should be considered both separately and collectively. Gravitational edifice loading, hence instability, for example, can be quantified by the level of the gravity center of the edifice. Instability from asymmetrical mass distribution within the edifice can be parameterized by eccentricity of the gravity center. Edifice weakening by fracture networks is dependent on fracture density, whereas loss of material coherence within the edifice, generated by hydrothermal alteration, can be expressed as volume fraction of the cone affected as well as the position and geometry of rock alteration areas.

In any case, each instability parameter, taken alone, has its own critical threshold value leading to a particular critical state. The net result of the combined action of all involved instability factors will lead to the actual critical state of the particular volcanic edifice. Thus, the attractor of self-organized criticality at composite volcanoes is a "complex attractor" whose structure is extremely difficult to determine. Once acknowledged, this complexity, however, does not preclude application of the SOC theory to volcanic edifice evolution. Much theoretical and experimental work is needed to assess the relative contribution, and combined action, of particular instability factors in the emergence of critical state at composite volcanoes. Nevertheless, the perspectives of adopting such an approach to study volcanoes are extremely appealing, as will be discussed next.

Perspectives of Application of SOC Theory to Understanding Long-Term Evolution of Composite Volcanoes

Viewed in the light of the SOC theory, the first edifice failure or sector collapse event at composite volcanoes is of particular relevance, because it marks the entrance of the system into the critical state as a result of the combined effects of instability factors. This master event separates the pre-failure "pre-critical" stage and the "post-critical" (or simply "critical") stages of volcanic evolution. All further sector collapses, whatever their magnitude, are symptoms of the critical stage already reached. Other master events such as caldera formation, however, could mark attainment of the critical state of a composite volcano when they initially occur in the volcano history. In this case, different control parameters have to be considered Following the above reasoning, long-term volcanic evolution can be divided into two first-order stages, according to its criticality status. This leads to a classification into either a pre-critical or a critical stage. Furthermore, this approach allows for a new set of systematics for composite volcanoes according to their SOC status. Volcanoes that have undergone at least one edifice failure or sector collapse (or caldera-forming) event qualify as "critical volcanoes"; others do not. The latter category can be further subdivided into a "precritical" type, to which all active volcanoes with no evidence for sector collapse or caldera formation belong, or into a "non-critical" type, including those extinct volcanoes with no failure or caldera formation during their evolution. Conceptually, the precritical and non-critical types do not apply to active volcanoes with no prior failure events, because they might eventually become critical during their future eruptive history.

Discriminating between precritical and non-critical active volcanoes will be a great challenge for future research, because such discrimination will provide a first-order selection criterion for prioritizing hazard assessment studies. Once the critical state of a volcano has been identified, further studies should be aimed at understanding the nature of criticality and the identification of the critical parameter(s). Such identification, in turn, would influence the planning and realization of detailed hazard and risk evaluation procedures and, eventually, would guide the choice of the most adequate monitoring techniques.

Because instability-driven large volcano failure events are infrequent, they apparently are less hazardous as compared with smaller but more frequent eruptive events. Large-scale events, both eruptive and non-eruptive, however, have a destructive potential that is orders of magnitude higher than that of smaller events, and thus should not be ignored. In the light of the SOC theory, critical volcanoes can produce such large catastrophic events at any time owing to the nonlinear nature of their intrinsic unstable dynamic state. Any small-scale "crisis" may trigger a disproportionately large avalanche-type failure event. Therefore, identification and priority surveillance of critical volcanoes in populated regions is crucial for risk mitigation.

VOLCANO INSTABILITY

Recognition and study of edifice instability and related processes prompted one of the major developments in volcanology toward the end of the twentieth century. The 18 May 1980 eruption of Mount St. Helens, Western United States, began with a giant multiple landslide and quickly transformed into a debris avalanche that removed the whole upper part of the volcanic edifice and left behind a large horseshoe-shaped depression (Lipman and Mullineaux, 1981). Research that initially focused on the deposit left behind by the debris avalanche event at Mount St. Helens led to the discovery of similar deposits worldwide at numerous extinct volcanoes. The study of such deposits has resulted in the understanding of the trigger, transport, and deposition mechanisms of volcanic debris-avalanche deposits (DADs), and the findings have been published in a plethora of papers. Consequently, a number of instability factors have been identified and evaluated (e.g., McGuire, 1996) and tentatively classified as intrinsic and extrinsic factors (Szakács and Seghedi, 2000).

Thus, by the end of the twentieth century, volcanologists recognized that edifice failure and debris avalanche processes are not exceptional, isolated phenomena but rather are common features of large, long-lived mature volcanoes, including both composite and shield types. In this section we highlight the fact that volcano instability is amenable to study by nontraditional approaches, such as the SOC theory discussed in the previous section.

We present a discussion of volcanic instability independently of volcano collapse, because volcano instability may initiate other processes as well. For example, volcano-basement interaction may result in a complex suite of processes, undergone simultaneously by both edifice and basement, that have been described as "volcano spreading" by Borgia (1994). Thus, volcano spreading is a different response to edifice instability, showing that flank or edifice failure is not the only possible outcome when a volcanic edifice continues to grow. The relationships between flank or edifice failure and volcano spreading—both relieving edifice instability—are as yet not fully understood. For instance, at Socompa volcano (Central Andes) it is apparent that volcano spreading triggered a huge edifice collapse event involving basement rocks (Wadge et al., 1995). In contrast, at other volcanoes such as those in the East Carpathians, Romania, it seems that those processes might be mutually exclusive. In this section we examine with some detail aspects of volcano stability other than those related to the SOC approach.

Instability Factors and Stability Factors

It is now generally accepted that edifice instability is the expected outcome of the evolutionary history of large long-lived volcanoes, which naturally evolve toward an unstable stage. A number of factors lead to instability. Some of them are inherently related to volcanic activity and edifice growth (intrinsic instability factors). Others have an external origin (extrinsic instability factors). According to Szakács and Seghedi (2000), all of these factors can be summarized as follows. (The main critical parameter related to the given instability factor is indicated in each case below by the italics preceded by the initials "C.p," standing for "Critical parameter."):

A. Intrinsic instability factors

AI. Gravitational instability (gravitational overloading)

1. Near-vent accumulation of effusive products *C.p. weight center*
2. Shallow intrusions within the edifice *C.p. weight center*

AII. Structural instability (structural "weakening")

3. Fissuration, jointing, faulting of edifice rocks *C.p.: joint frequency*
4. Eruption-related seismicity
5. Dyke-sill-intrusion *C.p. dyke/sill frequency*
6. Hydrothermal activity and alteration *C.p. pore-P, % clay minerals*

AIII. Geometrical instability

7. Departure from symmetry

B. Extrinsic instability factors

8. Basement topography, structure and lithology
9. Active basement tectonics
10. Non-volcanic seismicity
11. Hydrogeological and climatic conditions
12. Butressing/debutressing, etc.

All of these factors act, separately or in particular combinations, toward growing edifice instability during volcano evolution. However, the stability status of a particular volcano cannot be evaluated and modeled by considering instability factors alone. It was pointed out that ~75% of composite volcanoes in the Central Andes, having relative heights of at least 2500 m, underwent at least one failure event during their past volcanic history (Francis, 1994). Twenty-five percent of those volcanoes, however, survived (at least up to now) without failure events. More relevantly, there are extinct volcanic areas in which composite volcanoes with no record of failure prevail over those with such a record (e.g., in the Carpathian Neogene volcanic arc, Central-Eastern Europe). Moreover, one may find examples of stable and unstable volcanoes close to each other in the same volcanic region, where a large volcano (e.g., Mayon, Philippines) has not undergone edifice failure as yet, whereas a much smaller volcano nearby (e.g., Iriga) has. The above examples may suggest processes or features that counteract instability. One may name these *stability factors*.

Symmetry, both topographic and gravitational (axisymmetrical distribution of masses within a volcanic edifice) might be viewed as such a stability factor, whereas asymmetry is a factor of instability. Buttressing could be another stability factor, at least for sectors of an edifice. Fully buttressed edifices may exist within clusters of closely spaced volcanoes. Further, it seems that there is a link between magma type and edifice stability, irrespective of volcano size: volcanoes built of products of more fluid mafic magmas develop more stable edifices than those made of rocks from more viscous felsic magmas, as observed in Kamchatka in Russia (Belousov et al., 1999). If so, mafic magma acts as a stability factor, whereas felsic magma is an instability factor in the development of composite volcanoes.

In light of the above discussion, it is extremely tempting to consider that the actual stability status of a particular volcano—stable or unstable—depends on the balance between the instability factors and stability factors acting on it. Unfortunately, at present we know much more about processes favoring instability than about those favoring stability. In other words, somewhat paradoxically, it might turn out that volcanologists may learn a lot about edifice instability by investigating stable volcanoes (i.e., large composite volcanoes that have not undergone failure events). For this reason, we consider that comparative study of stable and unstable volcanoes of both similar and different magma composition in the same volcanic area appears to be a promising avenue of future volcano research. Geophysical methods, gravity surveys in particular, appear as adequate tools in investigating symmetry-asymmetry of mass distribution, hence gravitational edifice stability-instability, at mature long-lived volcanoes. Such investigations, not yet undertaken to the best of our knowledge, might reveal to what extent symmetry-asymmetry of internal mass distribution is reflected in volcano topography; eventually, this should allow a quantification of edifice stability status, at least from the gravitational point of view. Remote sensing studies may also provide clues concerning the stability state of an edifice by allowing a timely detection of changes in topographic features by a diversity of methodologies (e.g., Garvin, 1996; Massonnet and Sigmundsson, 2000; Zebker et al., 2000).

Solving Instability: Re-Equilibration Processes

Several processes can be listed as accounting for re-equilibration of unstable volcanoes: erosion, caldera formation, edifice-flank failure, and volcano-basement interaction. Erosion actually acts continuously against instability by reducing mass-loading from the volcanic edifice. Its re-equilibration effectiveness is higher at low-eruption-frequency volcanoes and in high-erosion-rate climatic conditions. Its contribution is, however, low as compared with other re-equilibration processes, but it increases significantly after the volcano becomes extinct.

Caldera formation, when it occurs, removes edifice instability extremely efficiently by dramatically reducing volcano height and gravitational loading. Caldera collapse, however, requires special conditions such as a large, shallow magma chamber and a large-volume eruption. Consequently, its re-equilibration capability is effective only at caldera-forming composite volcanoes.

Flank or edifice failure seems to be the most common means through which unstable volcanoes re-equilibrate their edifices. It may occur through a large spectrum of scales, from minor rockfall events on volcano slopes, to collapse of larger parts of a flank sector, to collapse and removal of the entire summit part of the edifice. Theoretically, the magnitude of such events is only limited, according to the SOC theory, by the size of the volcano itself (Bak et al., 1987). The edifice is strongly coupled to its substrate, however, basement may also be involved in the failure event. This may occur when the detachment surface excavates below the topographic base of the edifice within basement rocks, as occurred at Socompa volcano (Wadge et al., 1995; van Wyk de Vries et al., 2001). Fortunately, from the point of view of hazard assessment, such events are extremely infrequent at the scale of human history. Removal of instability by flank or edifice failure is commonly only partial, so that pre-failure instability conditions are reestablished, as witnessed by numerous edifices recording multiple failure events.

Incomplete removal of instability contributes to the maintenance of an unstable (or critical) state in the long term, so that further repeated failure events can be expected. On the other hand, large collapse events themselves contribute to further instability of the remnant part of the edifice, adding more instability factors, such as, for example, unbutressing. Composite volcanoes with evidence of multiple (up to six or seven) edifice flank or failure events are not uncommon (e.g., at Colima volcano, Mexico, Komorowski et al., 1994, 1997; and Shiveluch volcano, Kamchatka, Belousov et al., 1999).

Volcano-basement interaction and related processes can also be viewed as acting toward edifice re-equilibration. Volcano spreading in particular lowers both edifice height and position of the gravity center by subsidence of the central-summit part of the volcano, whereas lateral expansion increases its basal diameter. Hence the edifice height-diameter ratio significantly decreases, providing more stability. The main difference, as compared with flank or edifice failure, is that it is a long-term and gradual process, whereas failure is an instant phenomenon. This circumstance may have important consequences for the extent and manner in which these two types of re-equilibration processes will influence the further evolution of a volcano.

Consequences of Re-Equilibration by Flank or Edifice Failure

To what extent and in what manner flank or edifice failure will influence the volcanic system itself, as well as its immediate environment, depends on the size of the event. The size can be quantified by the volume, or mass, removed from the edifice. In turn, removed mass is a measure of the extent of unloading and re-equilibration. Large-scale events (involving volumes on the order of km^3) should have profound and long-lasting effects on the volcanic system. In particular, the first such event, as a master event, may mark a turning point in volcano evolution from a pre-critical stage to a critical stage of evolution. In contrast, small-scale events (e.g., <0.01 km^3) will have limited effects on the volcanic system. From this point on, we will focus only on the consequences of large-scale events.

The key factors that determine the consequences of a flank or edifice failure event are gravitational unloading, depressurization, and the instantaneous occurrence of these processes. These factors act on various parts, or subsystems, of the volcanic system. They also act on other neighboring volcano-related systems: the magma plumbing subsystem, including the shallow magma chamber; the eruptive subsystem and eruptive behavior; the volcano hydrothermal system; volcano topography and erosion; sedimentation around the volcano; and the basement. The full characterization of each of these interactions has not been investigated but constitutes a fruitful avenue for research in volcanology.

Effects on Shallow Magma Chamber and on the Magma Plumbing System

Sudden gravitational unloading and depressurization of the volcanic edifice from large-scale flank or edifice failure events are expected to strongly influence magma-chamber processes by causing a pressure drop in the chamber and, by favoring crystallization, allowing concurrent volatile loss from the whole shallow plumbing system. Rapid decompression-driven crystallization has been reported at Mount St. Helens volcano through melt inclusion studies (Blundy and Cashman, 2005) following eruptive events. If decompression by normal eruptive events involves such discernible effects in the shallow magma chamber, as recorded in volcanic rocks, one may expect that decompression events triggered by edifice failure would be orders of magnitude stronger. Thus, these events should have significant aftermaths in the shallow magma plumbing system, including the complete "freezing" of the post-eruptive, volatile-depleted magma chamber, and later, as a consequence, the reorganization of shallow magma pathways and storage beneath the volcano. Whether such changes in the shallow parts of a volcanic system might propagate downward, influencing the deeper parts of the volcanic system, is uncertain and remains to be quantified. Nevertheless, it seems that a large edifice failure event, particularly the initial one, may result in the reorganization of the subsurface magma transport and storage subsystems. This, in turn, may influence magma diversification processes and consequently compositional aspects of magma ascending within the eruptive subsystem and erupting at the surface during subsequent volcano evolution. Comparative petrochemical (including stable isotope geochemistry) investigation of pre-first-failure and post-first-failure volcanic products could be an effective new approach in the understanding of petrogenetic processes induced by large-scale flank or edifice failure events at composite volcanoes. To our knowledge, no such study has been undertaken. A possible outcome of such studies could be a unique geochemical "failure signature" of a large failure event (Fig. 1A), which might be identified in the products resulting from the first failure. Old extinct volcanoes that underwent a single large edifice failure during their history might be better targets for investigation in this respect than active volcanoes—such as Mount St. Helens—for which the post-failure record includes fewer eruptive events than during its pre-failure evolution. Collecting the required evidence to assess this possible scenario would therefore be an interesting avenue of research.

Effects on Eruptive Behavior

Large-scale failure events will strongly influence the eruptive behavior of volcanoes in both the short term and the long term. Their short-term effects are obvious when failure occurs at or near the beginning of an eruption, as was dramatically observed at Mount St. Helens on 18 May 1980, when the sudden unloading and depressurization of the edifice set the initial conditions for, and governed the further development (in terms of mechanisms, their succession, directivity, devastation, etc.) of, the eruption (i.e., a lateral blast followed by the Plinian phase). Edifice failure had an instantaneous effect here by driving the eruption along a certain path. Flank or edifice collapse may occur, however, at any stage of an eruption, or it could happen independently of any eruption (e.g., Unzen volcano in 1792; Hoshizumi et al., 1999), in which case its short-term influence on eruptive behavior would be much less important.

Reorganization of the shallow magma plumbing system, and possibly even deeper effects on the magma-generation system, following an instability-related master event occurring at some point of volcano evolution, will likely influence subsequent volcano behavior for the long term. One may expect that the first large-scale failure event in particular will be very effective in changing long-term eruptive behavior because of its profound impact on the shallow magma storage and magma feeding subsystems, as well as on the reorganization of magma transport paths. For example, post-failure re-localization of the subvolcanic magma chamber(s) owing to the radical changes of the lithostatic pressure conditions in the shallow basement environment of the volcanic system is an expectable outcome that will govern the post-event eruptive style of the volcano. As a major event, the first failure may separate volcano evolution in distinct—pre-failure and post-failure—stages that could be named pre-critical and critical, respectively, according to the SOC theory (see above). Comparing pre- and post-first-failure eruptive styles by investigating eruption products in the geological record at long-lived extinct volcanoes could be an efficient tool in understanding the long-term influence of flank or edifice failure events on volcano behavior.

Effects on Topography, Erosion, and Sedimentation

Both cone topography and ring-plain topography around the cone will be strongly modified as a consequence of large-scale flank or edifice failure (i.e., lateral collapse) events, because new topographic features, both negative and positive landforms, will be created. A large open depression, referred to as "debris avalanche depression," "amphitheater," or "horseshoe-caldera," is left behind by an edifice-failure event on the upper summit part of the volcano, which deeply excavates the edifice (Ui et al., 2000). Flank collapse results in smaller, yet significant, slide-shaped and shallower depressions (e.g., Sciarra del Fuoco at Stromboli, Italy; Marsella et al., 2009). Such prominent, newly created, asymmetric negative topographic features focus and channelize further volcanic activity and the distribution of resulting volcanic products. As a consequence, both topographic and mass-distribution symmetry of the edifice will be strongly altered. As a result, asymmetry might be enhanced, and hence further instability can be induced. Alternatively, asymmetry might be reduced, and therefore instability reduced, in the middle term.

The appearance of a large depression in the central part of the edifice will dramatically increase the recharge and influx of meteoric water in the central-axial part of the edifice. Occasionally, lakes form at the bottom of such depressions. As a consequence, conditions for subsurface magma-water interaction develop, and phreatic-phreatomagmatic eruption frequency could increase.

Lowering the edifice height of composite volcanoes by lateral collapse and removal of the steepest upper flanks results in lower erosion rates on the outer cone slopes. In contrast, inward-oriented erosion, transport, and sediment accumulation are enhanced because of the steep scarps bounding the debris avalanche amphitheater. As a consequence, erosional mass transfer shifts from dominantly outward to dominantly inward, at least until further eruptive activity eventually restores the pre-failure axisymmetrical, cone-shaped topography. Accumulation of erosional debris and sediments at the depression floor, which may, at least partially, coincide with post-failure vent locations, could additionally influence eruption style and mechanisms. For example, increased loading might require a corresponding increase in volatile and magmatic pressure, i.e., inducing a shift from open-vent eruption mechanisms (e.g., Strombolian) to closed-vent mechanisms (e.g., Vulcanian) (Parfitt and Wilson, 2008).

Emplacement of large-volume DADs results, by a number of processes, in dramatic topographic changes around the conical edifice itself. First, the sector facing the amphitheater where the DAD is deposited will acquire a higher elevation, corresponding to DAD thickness, with a unique, uneven hummocky surface. Second, the whole hydrographic system of that sector will be entirely reorganized; local small-sized lacustrine basins that form become local centers of deposition. Erosional-depositional processes intensify as a consequence of DAD emplacement in the sector facing the amphitheater, in contrast to other sectors where erosion-sedimentation will be attenuated.

As for the overall volcanic facies distribution, which displays a general axisymmetrical pattern at pre-critical composite volcanoes, an edifice failure event introduces an important element of asymmetry by the sectorial emplacement of DAD and its further influences on the development of the erosional-depositional system around the volcano.

Possible Effects on Basement

Loading of the basement by the weight of large-volume volcanic edifices is the primary cause of volcano-basement interactions, as pointed out by recent studies (Borgia, 1994; Merle and Borgia, 1996). Because edifice growth and loading is gradual, so also are the consequent volcano-basement interaction processes, which produce significant structural consequences in the edifice itself and its basement, both centrally and peripherally. They can be recognized mostly as peculiar structural features in both volcano and basement. Volcano spreading, involving subsidence of the central part, is the major mechanism of volcano-basement interaction in which ductile rocks are present in the shallow basement (van Wyk de Vries and Matela, 1998). Otherwise, flexure of a brittle shallow layer is the most common basement response to gravitational volcano loading (van Wyk de Vries and Matela, 1998).

In contrast, flank or edifice failure events result in sudden unloading of both the volcanic edifice and, implicitly, its basement. Large-scale master events of instant unloading should produce effects of large magnitude, not only within the volcanic edifice (as presented in previous sections) but also within the basement. Surprisingly, no investigation has been conducted so far to understand basement response to sudden gravitational unloading by edifice failure. One may expect that, depending on its structure and composition, basement deformation in response to unloading could be either ductile or brittle. More importantly, basement deformation feeds back to the edifice itself, inducing further deformation within it.

In the long term, repeated basement loading-unloading cycles might have important effects on volcano structure, stability, and behavior by volcano-induced basement deformation and its feedback to volcano deformation. Particular structural features should arise in both the basement and the edifice as a consequence. Analogue and/or numerical modeling of such processes, likely to occur in nature, may shed new light on volcano-basement interaction processes and on particular structural features they generate.

Effects on Volcano-Hydrothermal System and Ore Genesis

Evolved composite volcanoes commonly host hydrothermal systems, frequently with related ore deposits, within their edifices and shallow basement. The nature and extent of the influence that instability-related processes exert on these hydrothermal systems and ore genesis are as yet poorly understood, but intrinsically complex. These influences are probably as important as the changes induced by caldera formation on the volcano-related ore-forming hydrothermal systems.

One may assume that edifice or flank failure of a mature volcano might result in dramatic changes of its associated hydrothermal system, in both the short term and the long term. Short-term effects are basically related to sudden depressurization of the system (López and Williams, 1993). The key factors that determine the response of the hydrothermal system to such a destructive event are its depth, maturity, and the amount of depressurization. Deep excavation of the volcano, and related dramatic depressurization, can lead to the explosive disruption of the pressurized edifice-hosted hydrothermal system. An example is the lateral blast phase of the 18 May 1980 eruption of Mount St. Helens, USA, interpreted as a "hydrothermal eruption" by Lipman and Mullineaux (1981). Further, the extended evisceration, dispersion of the volatiles into the atmosphere, and incorporation of solid-phase components in the resulting DAD may develop into a DAD-hosted ore deposit when mature. Otherwise, DADs might include hydrothermally altered rocks, which could mislead exploration geologists looking for mineralization at deeper levels at the site of the DAD emplacement.

The outcome on the deeper, basement-hosted hydrothermal system depends on its maturity. The evolution of an immature system could be aborted as a consequence of premature, depressurization-driven boiling, preventing formation of ore-grade mineralization. Mature systems, however, will benefit from a pressure drop and induced boiling by massive deposition of pressure-sensitive ore minerals and formation of high-grade ore. This is true for the Ladolam high-grade gold deposit on Lihir Island in Papua-New Guinea, as pointed out by Simmons and Brown (2006). This deposit is hosted in an amphitheater-shaped depression left behind by lateral collapse of the volcano. Additional favorable conditions for ore deposition resulted from the explosive depressurization of the magmatic-hydrothermal system, which "produced a diatreme breccia complex and highly permeable rocks, which now host the ore" (Simmons and Brown, 2006), formed from a "gold juice" (Heinrich, 2006) already present in the system.

Questions to be asked in connection with the above examples are (1) what level or threshold of depressurization (and volume of material removed from the edifice) is needed to trigger massive ore mineral deposition, (2) what is the optimal composition and what are the elemental concentrations of the hydrothermal fluids subjected to decompression to generate ore-grade mineralization, (3) what is the optimal depth of the hydrothermal system on which a failure event will exert a favorable influence in terms of ore formation? Numerical and analogue modeling may offer clues by establishing a set of key parametric values optimal for ore formation triggered by volcano edifice failure.

One of the major long-range effects of edifice failure is related to the increase of meteoric water amounts contributed to the hydrothermal system because of the formation of a large depression and reorganization of the surface hydrologic regime. This, in turn, induces a change in the hydrological balance of the subsurface hydrothermal system: An increasing recharge area within the debris-avalanche depression favors higher inflow fluxes. A shift from a high-temperature vapor-dominated regime to a low-temperature dilute hydrothermal regime is the expected outcome of such a change in the volcano-related hydrothermal evolution. Those processes will imprint the resulting hydrothermal products (mineral assemblages) and their thermal, chemical, and isotopic features, including those of the ore minerals. A thorough investigation of hydrothermally altered areas and related ore deposits at collapsed volcanoes might result in the discovery of an "edifice-failure signature" in the hydrothermal evolution and its products.

The influence on the hydrothermal system of a gradual increase in edifice stability by volcano-basement interaction, such as volcano spreading and related phenomena, has not yet been studied. Since volcano-basement interaction involves gradual processes, their effects on the volcano-hydrothermal system are much more subtle than those related to edifice or flank failure events. Deformation induced in both edifice and basement would result in changes in fluid pathways according to the shift of local stress regimes between compressional and tensional. These, in turn, will depend on a number of factors, e.g., the presence, depth, and thickness of a plastically deformable layer in the basement. Depending on the interplay of these factors, extensional structures may form both centrally and peripherally with respect to the volcanic edifice, within the edifice itself, and/or in its basement (van Wyk de Vries and Matela, 1998). The distribution of such structures could determine the location of ore mineral deposition. In principle, extensional settings are favorable for vein-type mineralization, and compressional settings for porphyry-type mineralization. As a consequence, the type and location of hydrothermal ore formation may be strongly influenced by local tectonic features generated through volcano-basement interaction processes. Peripheral volcano re-equilibration–related compression may generate reverse faulting, up-doming and enhanced salt diapirism, such as those pointed out in the eastern Transylvanian Basin, Romania, by Szakács and Krézsek (2006), creating unusual structural traps for hydrocarbon accumulation.

In summary, an understanding of instability-related processes could greatly improve mineral exploration methodologies at both extinct and old composite volcanoes. Interdisciplinary approaches at the fringes of volcanology, structural geology, tectonics, and ore genesis seem to be a promising avenue for future research.

INTERNAL STRUCTURE OF A VOLCANO

Although current definitions of a volcano commonly focus on the external structures produced during eruptive events (Borgia et al., this volume), it is becoming more widely accepted that an important aspect of volcanic activity is controlled by processes taking place under the planetary surface. Whether this hidden part is considered a subsystem or an integral part of the volcanic system itself (Szakács, this volume) is not important. What is clear is that the understanding of volcanic phenomena must necessarily address processes that take place in the magma chambers and intrusions that emanate from them. Detailed characterization of the relationship between all of these different components of the volcanic system is an important goal of modern volcanology. In the following section we examine some issues related to these topics.

What Are the Physical Characteristics of Magma Chambers That Feed Volcanic Eruptions?

Volcanic activity requires the presence of melted rock beneath the surface of the planet. For this reason, many geologists studying volcanic activity in the nineteenth century envisioned Earth as a planet with a solid crust and a largely molten interior (Sigurdsson, 1999). By the 1920s, various sources of geophysical information, predominantly seismic, had shown that the only part of the Earth that could be considered to be completely molten was restricted to be >2900 km beneath the surface, and thus a most unlikely source of the liquid rock that feeds volcanic activity. At approximately the same time it was more or less well established that decompression could produce melting of the solid rock, but, according to Sigurdsson (1999), geologists of the time were unable to find an acceptable decompression scenario to produce melting at the scale required to feed the observed volcanic activity around the world. By the 1970s this situation had changed, as plate tectonics provided a plausible mechanism for the decompression of the Earth's interior along divergent plate margins, and by the 1980s it was established that water plays an important role in modifying the melting conditions of the mantle at convergent margins (Tatsumi and Eggins, 1995). As >94% of the active volcanoes on Earth are close to either a convergent or a divergent plate margin, the paradox of having a solid Earth that nonetheless is capable of feeding volcanic eruptions is therefore considered to be solved, because the mechanisms for locally melting the rock associated with tectonic activity explain the presence of almost all of the observed volcanic activity on Earth (Bardintzeff, 1998; Bogatikov et al., 2000; Wilson, 1989).

Although decompression and water addition to the solid mantle along divergent and convergent margins, respectively, provide plausible mechanisms for the formation of a liquid phase under the planetary surface, the amount of melt formed and its spatial distribution continue to be a matter of debate. Seismic information along these margins commonly reveals the presence of a low-velocity zone that could be interpreted to be a region containing some amounts of melt (e.g., Hasegawa et al., 1994; Mégnin and Romanowicz, 2000; Zhao, 2001; Miller et al., 2006). Temperature and grain-size differences and the presence of water in the mantle, however, might explain the observed low-velocity zones without invoking the presence of melts (e.g., Karato and Jung, 1998; Faul and Jackson, 2005). Consequently, the amount of melt beneath the surface of the planet remains uncertain in many cases.

Similar uncertainties are found, on a smaller scale, in determining the characteristics of magma chambers capable of feeding volcanic activity. Although conceptual models used to visualize magma chambers have evolved dramatically through time, the use of one model or another is not as easy to pinpoint in the literature as would be desired.

Most conceptual models prior to the 1990s depicted magma chambers as large, predominantly liquid reservoirs of various sizes and shapes (Turner and Campbell, 1986; Marsh, 1989; Bloomer and Meyer, 1992). However, many lines of evidence have suggested that this conceptual model may not represent common types of magma chambers found in nature (Iyer, 1984, 1992). On the contrary, the magma chamber model suggested by most lines of evidence is more likely to be a relatively widespread region not filled with liquid magma but rather containing a mixture of melt and crystals in which the solid fraction occupies >70% of the volume (Rosendahl, 1976; Sinton and Detrick, 1992). This model, although derived mainly from observations made along oceanic ridges, seems to be equally valid for volcanoes in other settings, involving chambers of very different sizes (e.g., Sharp, 1982; Marsh, 2000; Soosalu and Einarsson, 2004; Nunziata et al., 2006).

Visualization of magma chambers as regions where a mixture of melt and crystals coexist, rather than as cavities completely filled with liquid, may lead to different interpretations of the geological record. For example, the compositional or mineralogical sequences displayed by some plutonic or volcanic rocks may be interpreted as the result of in situ processes or the result of a series of injection events, depending on whether the classical model or the more recent model of a magma chamber is adopted (Marsh, 2000). Additionally, the selection of a magma chamber model can influence visualization of processes leading to the presence of volcanic activity on the surface of a planet. Nevertheless, the consequences of a shift in the conceptual model of a magma chamber remain to be fully explored.

Formation of Tabular Conduits

One example of the way in which the conceptual model of a magma chamber can influence the manner in which other volcanic processes are visualized involves the formation of conduits that allow the transportation of the molten material in the chamber away from it (Fig. 2). In general, these conduits will have one dimension much smaller than the other two, and therefore can be called *tabular conduits*. Depending on the orientation of

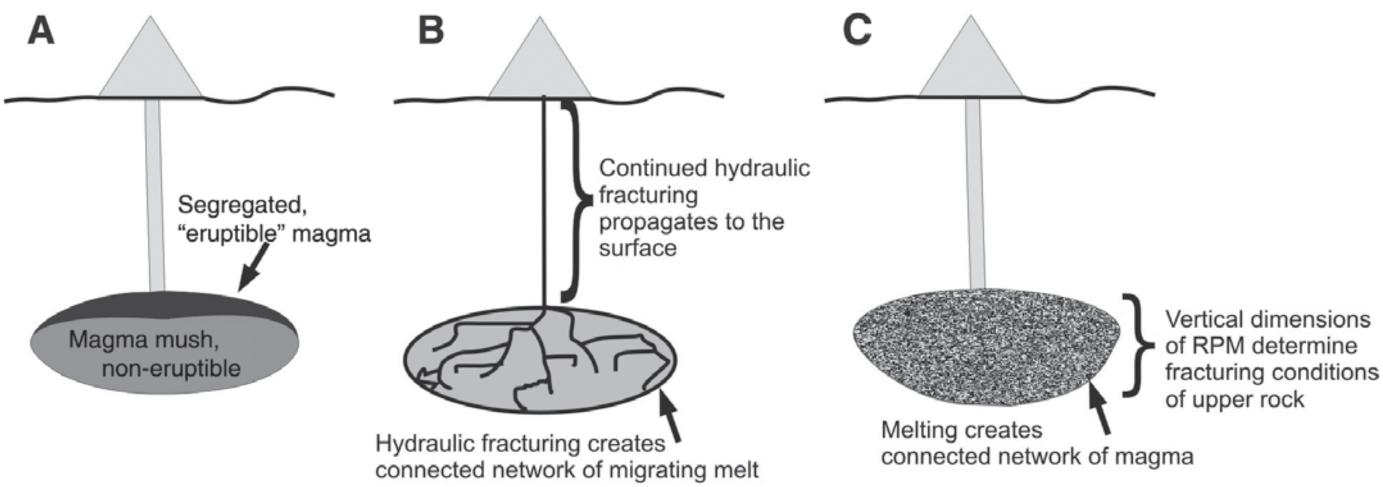


Figure 2. Schematic representation of three alternative models used to visualize magma chambers and their relationship with volcanic activity. (A) The classical model, in which chambers are considered a fluid-filled cavity, at least in its upper part. (B) The arborescent model, in which an interconnected melt fraction starts to fracture the solid fraction inside the chamber to originate the conduits that eventually tap magma out of the chamber. (C) The hydrostatic model, in which the excess pressure required to initiate the fracture of the overlying solid rock is achieved by the difference of densities between solid and liquid inside the chamber. See text for details.

the conduit relative to the stratification of the rock outside the magma chamber, these conduits can be called dikes, sheets, or even sills. If one of these conduits might temporally connect the magma chamber and the surface of the planet, the tabular conduit is called a feeder dike. Regardless of the name given, the important fact in the present context is that the mechanisms of formation of such conduits are likely to be the same independently of whether they serve to feed a volcanic eruption or only provide a pathway to transport the liquid phase within the chamber to a different location in the lithosphere. For this reason, we avoid calling them dikes, and adopt the more general term *tabular conduits*.

As illustrated in Figure 2, there are at least three alternative mechanisms of nucleation of tabular conduits that involve the two alternative models of a magma chamber. In the first case it is considered that tabular conduits can be produced only if the magma chamber either is fully filled with a liquid phase or it has developed a zone in which the liquid phase accumulated, becoming a pool of "eruptible magma." In the two other cases shown in Figure 2, it is considered that tabular conduits can be formed anywhere at the walls of the chamber, even in the absence of a pool of eruptible magma, or alternatively that tabular conduits can be nucleated anywhere within the chamber (i.e., within the magmatic mush). Although the second and third possibilities do not rely on the accumulation of a pool of eruptible magma, each of those alternatives implies important differences in the mechanisms that need to be invoked, and on the conditions that need to be satisfied to initiate the nucleation of a tabular conduit.

The concept of a pool of liquid accumulated at the top of a magma chamber, envisaged as a parcel of eruptible magma was examined in some detail previously (Marsh, 1996). This idea finds support when the long-term result of porous flow within a region of partial melt is taken into consideration, as it can be shown that the long-term evolution of a region of partial melt bounded by an impermeable zone, where melt is thermally unstable, leads to the formation of regions with liquid volumes of >60% above the magma mush regions (e.g., Rabinowicz et al., 2001). Furthermore, although perhaps not the most common type of magma chamber documented to exist beneath active volcanoes, it is possible to find some chambers that display this upper accumulation of magma underneath some regions of active volcanism (e.g., Sinton and Detrick, 1992).

This conceptual model, however, cannot be used easily to assess the conditions required for the nucleation of a tabular intrusion in general terms. In this sense, it is noted that the accumulation of liquid by itself does not increase the internal pressure within the confines of the chamber, as such accumulation only consists in a redistribution of already existing solid and liquid phases. Consequently, a re-accommodation of material inside the chamber does not provide by itself the physical mechanism required to achieve an "overpressure" capable of fracturing the rock that forms the roof of the chamber.

Gas exsolution as the pool of liquid forms is therefore required to justify such overpressures. For this reason, even when this conceptual model seems to consider the most common physical characteristics of magma chambers, in essence it is a mere transposition of the old concept of a magma chamber visualized as a cavity completely filled with liquid. The only difference between both models is that in the more recent model the cavity has been reduced to a region of more limited dimensions. Thus, no new insights concerning the mechanism of formation of tabular intrusions can be obtained by using this conceptual model. In addition, eruptive activity might take place in some volcanoes that lack clear evidence of a pool of eruptible magma above a mushy chamber (e.g., Soosalu and Einarsson, 2004). This also casts some doubts on the universal validity of this conceptual model as envisioned by Bachmann and Bergantz (2008).

A model associated with a second hypothesis concerning the origin of tabular conduits is illustrated in Figure 2B. This model describes the process of dike nucleation as the evolution of an interconnected percolative regime into an interconnected fracture regime. The physical processes associated with the initiation of such arborescent tendencies have been examined in some detail by various authors (Daines and Kohlstedt, 1994; Nicolas, 1986; Nicolas and Jackson, 1982; Sleep, 1988), but a model that describes both the percolative transport of magma and the subsequent evolution of the rapidly ascending dike in a coupled form is still lacking (Fowler, 1990a, 1990b). In any case, according to this conceptual model, tabular conduits can form directly from the magma mushes as the result of hydrofracturing. This presents certain problems, because from the energetic point of view, a partially molten material is already fractured, and consequently the fracturing required to originate a tabular intrusion is only a modification of the existing melt geometry. Nevertheless, as discussed by Fowler (1990b), a more precise understanding of fracturing (perhaps involving criteria that are more complicated than the Coulomb or Griffith models) might pinpoint more accurately the location of the initiation of a tabular conduit within the chamber. In any case, this level of detail is still a matter for future research.

The third model, illustrated in Figure 2C, is a viable alternative that avoids the problems faced by the evolutionary percolative model, yet at the same time retaining the basic aspects that envisage magma chambers as regions consisting of a mush rather than a pool of magma. This hypothesis considers that hydraulic fracturing that leads to the formation of a tabular conduit takes place only within the completely solid rock that lies outside the magma chamber. One obvious advantage of shifting the site of nucleation of the tabular intrusion to the roof (or walls) of the chamber is that in this case the Coulomb or Griffith criteria, commonly used to describe the conditions necessary for the brittle fracturing of a solid, are valid without any further modification. Although at first sight this approach might seem to be equivalent to the model of Figure 2B, by avoiding the definition of what a fracture is, this approach constitutes an independent model and should not be confused with the model proposed by Nicolas (1990).

Another advantage of the approach illustrated in Figure 2C is that it allows us to constrain some of the physical characteristics associated with the magma chamber required for the initiation of a tabular intrusion if it is assumed that the overpressure required to initiate the fracturing event is due to the buoyancy of a vertical column of magma. The quantitative constraints imposed by this model on the vertical extent of regions of partial melt (or magma chambers), in turn, can be useful to determine whether a particular chamber has the potential to initiate an extrusive event or not, as delineated by Cañón-Tapia (2008, 2009). Consequently, this conceptual model leads to the formulation of critical observations that can be used to validate its use. Evidently, the full consequences of this model are still unexplored, requiring more work to achieve this aim. Nevertheless, given its relative simplicity and flexibility, this model has the potential to provide a much better answer concerning the origin of tabular conduits than the other two alternatives examined above.

Where Are Magma Chambers Formed?

Asking where magma chambers are located is, to some extent, similar to asking where magma originates. Consequently, a thorough answer to this question would require a review of many geodynamic models. In the present paper we concentrate on two key pieces of information that can be constrained empirically: (1) the depth at which magma chambers can exist, and (2) the state of stress likely to surround these chambers.

Constraints on the Depth of Magma Chambers

The question concerning the depth of a magma chamber is deceptively simple, yet it is extremely important. Part of the apparent simplicity comes from ample evidence suggesting that most active volcanoes erupt products that have spent some time in chambers at depths not exceeding 20 km, and in many instances shallower than 10 km (e.g., Baker, 1987; Clarke et al., 2007; Peccerillo et al., 2007; Rutherford and Gardner, 2000; Saito et al., 2005). The occurrence of very shallow (<5 km) magma reservoirs is so common that in some cases any reservoir between 10 and 20 km is referred to as a deep magma chamber (e.g., Nunziata et al., 2006; Smith et al., 2005). Thus, one could be tempted to restrict the relevant depths for the presence of magma chambers to anywhere between 2 and 30 km depth. This depth range, however, is in striking contrast with the evidence, such as the presence of xenoliths or the equilibrium conditions of various mineral phases (e.g., McGetchin and Ullrich, 1973; Meyer and Svisero, 1987; O'Reilly and Griffin, 2006), that magma originates at much greater depths. Indeed, petrological evidence suggests that in many cases magma could have risen to the surface from chambers as deep as 200 km, and possibly in some cases reaching 300 km depth (Evans, 1997; Zhang and Liou, 2003). Thus, although it is true that the occurrence of eruptions from such deep-seated sources are not common on Earth, one is forced to extend the depth range for the presence of magma chambers to at least 250 km below the surface. Moreover, it is likely that beneath many volcanoes, complex networks of reservoirs lie at various depths. The evidence for this complexity, however, is overlooked except in a few cases (e.g., Galipp et al., 2006). Nevertheless, it must be remarked that the meaning of *deep reservoir* depends on the context in which it is used.

State of Stress Surrounding Magma Chambers

Similar to concepts about the depth of magma chambers, ideas about the state of stress surrounding magma chambers are also prone to oversimplification and broad generalization. From a geodynamic and geochemical point of view there is a tendency to consider that magma generation takes place in three well-defined settings: convergent boundaries (arcs), divergent boundaries (rifts), and above plumes (hotspots). This generalization can be extended to include the characteristic state of stress surrounding a magma chamber, and therefore it would be common to say that chambers associated with arc volcanoes are more likely to undergo a compressional stress regime, whereas those associated with volcanic activity in rifting zones are more likely to undergo an extensional regime. These generalizations, however, could be very misleading, as the stress structure of each of these tectonic scenarios can be far more complex than the simple associations would suggest.

Considering that the range of depths of interest in the present context extends to 250–300 km below the surface, not only the state of stress in the lithosphere, but also the short-time state of stress of parts of the asthenosphere, is germane to models explaining the occurrence of volcanic activity. This may seem surprising (or even erroneous) at first sight, because geostatic pressure is generally used to describe the large values of compressive normal stresses and relatively high temperatures likely to prevail at great depths.

This common view, however, overlooks another condition required for the stress field to be considered geostatic: Stresses must be maintained for long periods of time so that shear stresses are reduced to such low values that can be neglected. This condition is likely to be satisfied only when time intervals on the order of millions of years are considered. Nevertheless, when the relevant scale of time is much reduced, even rocks undergoing large confining pressures and high temperatures react as solids, as shown by the behavior of seismic waves. Consequently, when the time scale relevant for the formation of tabular conduits is considered (from minutes to a few days), even the asthenosphere must be considered to behave like a solid (Middleton and Wilcock, 1996). This is equivalent to saying that shear stresses may not be null around magma chambers in general, and (perhaps more importantly) that all three principal stresses may not be equal.

Unfortunately, understanding and characterizing the state of stress in the Earth's outer shells has proved to be a rather difficult task that has required contributions from a variety of sources. Ignoring the problem of the temporal scale mentioned above, balancing the required global and regional information and/or models used to describe lithospheric stresses is extremely complicated, and only limited success was achieved until recently (e.g., Lithgow-Bertelloni and Guynn, 2004). Thus, it would be presumptuous to say that the real state of stress surrounding magma chambers can be described with great accuracy. The consequences of considering a situation in which the three principal stresses around a magma chamber are not equal to each other, however, deserves to be explored in more detail to better constrain the probability of the occurrence of a volcanic eruption in a general context.

INFORMATION TRANSFER THROUGH VOLCANOES: NEW APPROACHES?

The final section of this paper is highly speculative, for no formal work has ever been done to support the subject here discussed. We decided to include it in the present chapter because the societal benefits of earthquake prediction justify the exploration of diverse avenues of research, even those that may seem highly speculative. It should be understood that we do not advocate approaches other than rigorous treatment of information, and therefore we do not promote the a priori acceptance of the relationship proposed here.

Volcanoes are extremely effective transmitters of matter, energy, and information from the deep Earth to its surface. They can be viewed, in a symbolic way, as Earth's "chakras" (i.e., points where energy is concentrated). In a more physical sense, ascending matter and energy fluxes are recorded at active and dormant volcanoes alike. Heat transfer toward the Earth's surface, and related transfer of volatile matter, remain long-lasting processes after a volcano becomes extinct. While carrying magma and energy upward, volcanoes implicitly deliver information from depth to the surface in a number of ways. Much of what is currently known about Earth's mantle and crust has been derived from information embedded in volcanic products and processes. Volcanic conduits can be viewed as rodlike or sheetlike vertical features with relatively homogeneous composition and structure that cut across geological structures of greater complexity and compositional heterogeneity (Fig. 3). In addition, they are more long-lived stable structures at the geological time scale than surface features. Furthermore, they physically connect the Earth's surface to deeper regions of Earth, across the crust or lithosphere. As a consequence, conduits of volcanoes can be viewed as channels of information transfer from depth to the surface. While active, and for some time afterward, information is carried by magma and related fluids as well as by thermal energy fluxes. Later on, these channels may serve as solid media for transmission of other types of information. Information-carrying signals, such as earthquake precursor signals originating deep below the Earth's surface, may be transmitted with much less loss of information (and implicitly much higher signal-to-noise ratio) through homogeneous, vertically extended structures than through the horizontally segmented heterogeneous lithosphere or crust. Volcanic conduits can thus be viewed as upside-down "antennas," which can be used as privileged pathways of possible earthquake precursor signals (Fig. 3).

Conduits of monogenetic volcanoes are promising transmitters of deep Earth information that could be received and decoded at surface monitoring stations, because these volcanoes have a more homogeneous conduit (even if already solidified) than those of polygenetic volcanoes, as the conduit is formed as a single event. Among monogenetic volcanoes, those with dominantly effusive activity are the best candidates for privileged earthquake-monitoring sites. Further selection criteria may include the presence of mantle xenoliths in surface volcanic products, indicating a direct link between the deep lithospheric mantle and surface through the conduit. Sensors emplaced on selected extinct monogenetic volcanoes might increase the effectiveness of earthquake prediction strategies. The accuracy of the widely debated VAN earthquake prediction method, based on recognition and interpretation of electromagnetic precursors (e.g., Varotsos et al., 1986), for example, could be tested and improved substantially in this way.

An important prerequisite for using volcanic conduits as privileged sites of seismic monitoring activities is prior detailed geological and geophysical investigation. The extension, morphology, composition, homogeneity, and other depth-dependent features of conduits should be thoroughly understood before their use in seismic monitoring. Thus, there are ample opportunities for new and pioneering studies of volcanic conduits.

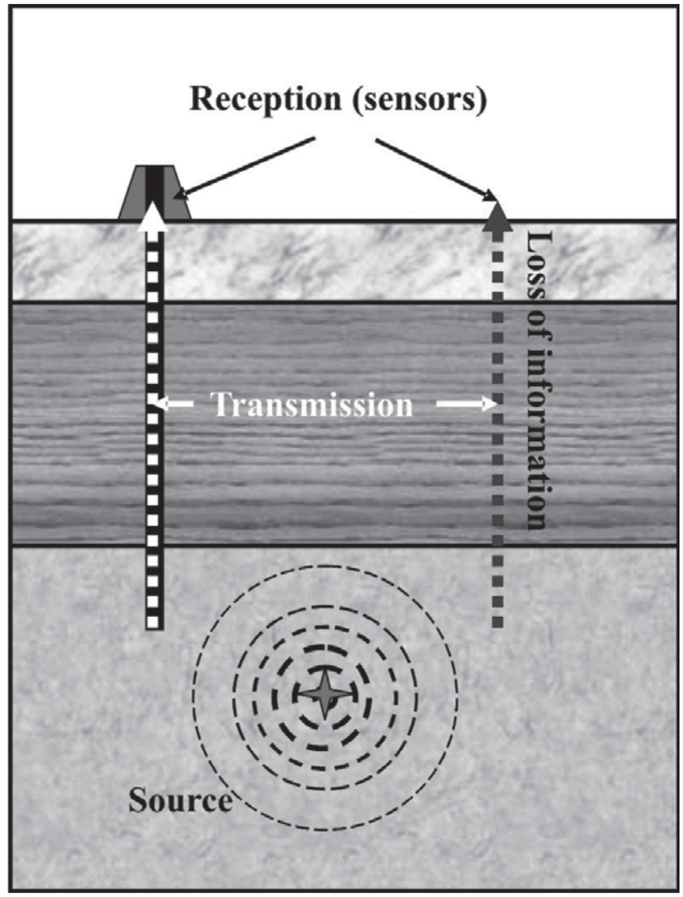


Figure 3. Conceptual model showing transmission of information (e.g., a seismic precursory signal) through the heterogeneous lithosphere and through a volcanic conduit.

CONCLUDING REMARKS

Throughout this chapter we have presented examples of some of the challenges faced by volcanology in the near future. The list of topics examined in this chapter is by no means exhaustive, and may seem unrepresentative of the interests of much of the volcanological community. It is unavoidable that the list of topics presented above reflects the bias of the authors. Nevertheless, in each of the cases examined in this chapter we have striven to provide as broad a view as possible, balancing the advantages of adopting new technologies with the need of continuing the search for answers to questions that have a general scope. In any case, we hope that this chapter stimulates some readers to formulate their own list of pending questions that need to be addressed by volcanology in the future. If that list has nothing in common with ours, but expresses the need to design research from a general, rather than from a highly specialized perspective, we will have succeeded.

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