THE INFLUENCE OF VOLCANIC SYSTEMS ON THE MORPHOLOGICAL EVOLUTION OF LAVA FLOW FIELDS

A thesis submitted to the University of London for the degree of Doctor of Philosophy

by

Jenkin Wyn Hughes

University of London Observatory Planetary Image Centre
(Department of Physics and Astronomy, University College London)
33/35 Daws Lane
Mill Hill
LONDON, NW7 4SD
I fy mham, ac er cof am fy nhad
(To my mother, and to the memory of my father)

Dyfal donc a dyrr y garreg
(Constant blows break the stone)
ABSTRACT

This thesis presents the results of a detailed study of the volcanic system of Mount Etna, Sicily, in the 1600 to 1689 and post-1750 periods of magmatic output. The conditions which prevailed within the volcanic system are ascertained, and their influence on the intra-volcano movement of magma constrained. The examination is extended to the superficial emplacement and planimetric evolution of the lava flow fields.

In the post-1750 period, there was a marked sectorial control on eruptive styles at the surface, with large volume ( > 55 x 10^6 m^3), long duration (32 days) and flow field (Type B) producing eruptions being generally restricted to the eastern sector. Western sector eruptions generally produced flow unit (Type A) lava flows and were of smaller volume and shorter duration. This dichotomy is interpreted to reflect interaction between the high level plumbing system and an asymmetric high level gravitational stress field. No sectorial dichotomy existed in the 17th century period, an observation which is attributed to the superposition of a more prominent basement tectonic stress field on the gravitational stress field. These changes in internal conditions are interpreted to have coincided with a shift in the regional tectonic stress regime below Etna.

Though the predominant stress fields have varied, it is argued that the factors controlling the emplacement of the lava flows were common to both systems. A morphometric study of post-1600 lava flows establishes that the planimetric evolution of lava flows occurs in response to a temporally decaying effusion rate, and that the variety of final flow field morphologies, from aa channel-fed flow units to the pahoehoe tumulus flow fields, mostly reflect stages of systematic planimetric development along a single evolutionary trend. The change in effusion rate is itself, considered to be related to a temporal change in the magnitude and location of the principal eruptive mechanism within the volcanic system.
Table of Contents

Abstract .............................................3
Table of Contents ....................................4
List of Tables .......................................8
List of Figures ......................................9
Acknowledgements ..................................12
Chapter 1. Introduction .............................13
Chapter 2. The Volcanology of Mount Etna Volcano ...17
  2.1 Introduction ..................................17
  2.2 Geological Setting .............................18
    2.2.1 Regional and Volcano Tectonics ............18
    2.2.2 Basement Geology .........................22
  2.3 Volcanic History .............................23
  2.4 The Volcanic Edifice ..........................25
    2.4.1 Volcano Morphology .......................25
    2.4.2 Eruptive Styles ............................28
  2.5 The Plumbing System of Mount Etna ............30
    2.5.1 Post-1971 ..................................30
    2.5.2 Pre-1971 .................................35
  2.6 Summary ......................................38

PART I POST-1600 LAVA FLOWS: A MORPHOMETRIC STUDY ...40
Chapter 3. The Morphological Evolution of Etnan Lava Flow Fields ...41
  3.1 Introduction ..................................41
  3.2 Data Collection ................................42
    3.2.1 Altitude ..................................43
    3.2.2 Lateral Distance (Feeder Dyke Length) ...43
    3.2.3 Flow Length ................................45
    3.2.4 Surface Area ...............................45
    3.2.5 Erupted Volume ............................46
    3.2.6 Eruption Duration .........................48
3.2.7 Eruption Rate ..... 49
3.3 Method of Analysis ..... 50
3.4 Lava Flow Morphologies of the Post-1750 Period. ..... 58
   3.4.1 Rheology ..... 61
   3.4.2 Volume ..... 66
   3.4.3 Eruption Rate and Duration ..... 71
   3.4.4 Discussion ..... 79
3.5 Lava Flow Morphologies of the Early 17th Century ..... 80
   3.5.1 Rheology ..... 81
   3.5.2 Volume ..... 82
   3.5.3 Eruption Rate and Duration ..... 85
   3.5.4 Discussion ..... 87
3.6 Summary ..... 88

PART II THE POST-1600 VOLCANIC SYSTEM OF MOUNT ETNA ..... 90
Chapter 4. The Post-1750 Volcanic System of Mount Etna: Eruptive Activity and Lava Flow Emplacement on the Western Sector ..... 91
   4.1 Introduction ..... 91
   4.2 Method of analysis ..... 91
   4.3 Spatial distribution ..... 93
   4.4 Erupted Volume of Lava ..... 99
      4.4.1 Below 2,000 m asl ..... 99
      4.4.2 Above 2,000 m asl ..... 102
   4.5 Eruption Duration ..... 103
      4.5.1 Below 2,000 m asl ..... 103
      4.5.2 Above 2,000 m asl ..... 108
   4.6 Eruption Rate ..... 108
      4.6.1 'Peak' Effusion Rates ..... 111
      4.6.2 Temporal Variation of Real-Time Effusion Rates ..... 116
   4.7 Lava Flows: Emplacement and Planimetric Evolution ..... 120
      4.7.1 Below 2,000 m asl ..... 122
      4.7.2 Above 2,000 m asl ..... 123
   4.8 Discussion ..... 124
Chapter 5. Post-1750 Eruptive Activity and Lava Flow Emplacement on the Eastern Sector

5.1 Introduction
5.2 Method of Analysis
5.3 Erupted Volume of Lava
  5.3.1 Below 2,000 m asl
  5.3.2 Above 2,000 m asl
5.4 Eruption Duration
  5.4.1 Below 2,000 m asl
  5.4.2 Above 2,000 m asl
5.5 Eruption Rate
  5.5.1 Below 2,000 m asl
  5.5.2 Eruptive Mechanism
  5.5.3 Above 2,000 m asl
  5.5.4 'Valle del Bove' Group of Eruptions
5.6 Lava Flows: Emplacement and Planimetric Evolution
  5.6.1 Lengthening
  5.6.2 Widening and Thickening
5.7 Discussion
5.8 Summary: The Post-1750 Volcanic System

Chapter 6. The 1600-1689 Volcanic System of Mount Etna: Eruptive Activity and Lava Flow Emplacement

6.1 Introduction
6.2 Method of analysis
6.3 Spatial Distribution
6.4 Erupted Volume
  6.4.1 Below 2,000 m asl
  6.4.2 Above 2,000 m asl
  6.4.3 High Level Volumetric Capacity
  6.4.4 Volumetric Capacity of 17th Century Volcanic System
6.5 Eruption Duration
  6.5.1 Below 2,000 m asl
  6.5.2 Above 2,000 m asl
6.6 Eruption Rate
   6.6.1 Eruption Durations < 32 days ..206
   6.6.2 Eruption Durations > 32 days ..208
   6.6.3 Eruptive Mechanism ..213
6.7 Lava Flows: Emplacement and Planimetric Evolution ..218
   6.7.1 Lengthening ..218
   6.7.2 Widening and Thickening ..219
6.8 Discussion ..222
6.9 Summary: The 17th Century Volcanic System ..224
Final Comments ..226
Chapter 7: Conclusions and Suggestions for Further Work ..228
  7.1 General Conclusions ..228
    7.1.1 Lava Flow Emplacement and Planimetric Evolution ..228
    7.1.2 The Post-1600 Volcanic System ..230
  7.2 Suggestions for Further Work ..234
    7.2.1 Eruption Surveillance ..234
    7.2.2 Tectonism and Stress Fields ..236
    7.2.3 Volcanic Hazard ..236
Appendix 1: Extruded Volume Calculations ..238
References ..241
Supplementary Papers Inside Back Cover

LIST OF TABLES

2.1: Stratigraphy of Etnean Basement Rocks ...
2.2: Mount Etna: Volcanic Succession ...
3.1: Mount Etna: 17th Century Eruption Data ...
3.2: Mount Etna: Post-1750 Eruption Data ...
3.3: Mount Etna: Eruption Information from Lava Flow Morphology ...
5.1: Mount Etna: a table showing the relationship between Post-1750 eruptive styles above 2,000 m asl on the tensile eastern sector and variable magma pressure decay rates ...
6.1: Mount Etna: a table showing the relationship between 17th century eruptive styles at various elevational intervals and variable magma pressure decay rates ...

LIST OF PLATES

Plate 1 Aerial Photograph of the 1947 Lava Flow Field ...
Plate 2 Aerial Photograph of the 1865 Flow Field ...
LIST OF FIGURES

Chapter 2

Figure 2.1. Geological and tectonic map of Southern Italy  ...18
Figure 2.2. Rose diagram of structural lineaments on Mount Etna  ...20
Figure 2.3. Geological map of Mount Etna  ...26
Figure 2.4. Mount Etna: structural lineaments and vent distribution  ...27
Figure 2.5. Size frequency distribution of plagioclase phenocrysts  ...35
Figure 2.6. Volumetric output of Mount Etna since 1535 AD  ...37

Chapter 3

Figure 3.1. Summit of Etna  ...44
Figure 3.2. Schematic diagram showing variation of effusion rate with time  ...49
Figure 3.3. Sketches of post-1750 flow morphologies  ...59
Figure 3.4. Apparent viscosity - shear rate diagram showing stability fields of pahoehoe and aa surface textures  ...64
Figure 3.5. Etna (Post-1750): logarithm of flow length versus the logarithm of erupted volume  ...68
Figure 3.6. Etna (Post-1750): logarithm of flow length versus the logarithm of eruption rate  ...72
Figure 3.7. Etna (Post-1750): logarithm of flow length versus the logarithm of theoretically calculated eruption rate with contoured time intervals  ...75
Figure 3.8. Etna (17th Century): logarithm of flow length versus the logarithm of erupted volume  ...84
Figure 3.9. Etna (17th Century): logarithm of flow length versus the logarithm of eruption rate  ...86

Chapter 4

Figure 4.1. Flank profiles of Etna  ...92
Figure 4.2. Spatial distribution (Post-1750): lava flow morphology (A): <2,000 m asl, (B): >2,000 m asl  ...94
Figure 4.3. Spatial distribution (Post-1750): erupted volume (A): <2,000 m asl, (B): >2,000 m asl  ...95
Figure 4.4. Spatial distribution (post-1750): eruption duration (A): <2,000 m asl, (B): >2,000 m asl  ...96
Figure 4.5. Spatial distribution (post-1750): eruption rate
(A): <2,000 m asl, (B): >2,000 m asl

Figure 4.6. Western sector: vent elevation versus volume erupted
Figure 4.7. Western sector: volume erupted versus distance of vent from the central conduit
Figure 4.8. Western sector: vent elevation versus duration
Figure 4.9. Western sector: duration versus distance of vent from the central conduit
Figure 4.10. Stress distribution on a volcanic edifice
Figure 4.11. Western sector: vent elevation versus eruption rate from the central conduit
Figure 4.12. Western Sector: eruption rate versus distance of vent from the central conduit
Figure 4.13. General: vent elevation versus theoretical eruption rate
Figure 4.14. General: theoretical eruption rate versus distance of vent from the central conduit
Figure 4.15. Vent elevation versus flow length (all data)
Figure 4.16. Western sector: vent elevation versus flow length
Figure 4.17. Western sector: flow length versus distance of vent from the central conduit

Chapter 5

Figure 5.1. Eastern sector: vent elevation versus volume erupted
Figure 5.2. Eastern sector: volume erupted versus distance of vent from the central conduit (A): <2,000 m asl, (B): >2,000 m asl.
Figure 5.3. Distribution of vents at the head of the Valle del Bove
Figure 5.4. Eastern sector: vent elevation versus eruption duration
Figure 5.5. Eastern sector: eruption duration versus distance of vent from the central conduit (A): <2,000 m asl, (B): >2,000 m asl.
Figure 5.6. Eastern sector: vent elevation versus eruption rate
Figure 5.7. Eastern sector: eruption rate versus distance of vent from the central conduit (A): <2,000 m asl, (B): >2,000 m asl.
Figure 5.8. Schematic diagram of a volcanic system showing relationship between magma input and output during a volcanic eruption
Figure 5.9. Eastern sector: vent elevation versus flow length
Figure 5.10. Eastern sector: flow length versus distance of vent from the central conduit (A): <2,000 m asl, (B): >2,000 m asl
Figure 5.11. Schematic diagram showing stages in the planimetric evolution of the 1983 flow field
Figure 5.12. Schematic diagram showing stages in the planimetric evolution of 1865 flow field.
Figure 5.13. Schematic diagram showing stages in the planimetric evolution of the 1950-51 flow field
Chapter 6

Figure 6.1. Spatial distribution (17th century): lava flow morphology
(A): <2,000 m asl, (B): >2,000 m asl

Figure 6.2. Spatial distribution (17th century): erupted volume
(A): <2,000 m asl, (B): >2,000 m asl

Figure 6.3. Spatial distribution (17th century): eruption duration
(A): <2,000 m asl, (B): >2,000 m asl

Figure 6.4. Spatial distribution (17th century): eruption rate
(A): <2,000 m asl, (B): >2,000 m asl

Figure 6.5. Plot of 17th century erupted volume data versus
(A): vent elevation, (B): lateral distance

Figure 6.6. Plot of 17th century lava flow/flow fields length data versus
(A): vent elevation, (B): lateral distance

Figure 6.7. Plot of 17th century eruption duration data versus
(A): vent elevation, (B): lateral distance

Figure 6.8. Plot of 17th century eruption rate data versus
(A): vent elevation, (B): lateral distance
ACKNOWLEDGEMENTS

This thesis represents part of an ongoing quest, which I have pursued from an early age, to become a volcanologist. I have been fortunate along the way to have studied and worked alongside experts in the field of geology and volcanology, each of which has greatly influenced my way of thought, and career development. I shall not attempt to acknowledge everyone for fear of omitting individuals, but I am in debt to you all.

As for this thesis, I would like to express my sincerest thanks to the following individuals for their contribution towards its completion.

John Guest (UCL), for giving me the opportunity to conduct this research and for sharing his wealth of volcanological experience with me; Angus Duncan (Luton College) for useful discussion on new and existing ideas, Chris Kilburn (JPL) for showing unfaltering encouragement and lively discussion; Dave Rooks (UCL) for professional photographic assistance and his patience when confronted by my tedious tasks; Soren-Aksel Sorensen (UCL) who in 10 minutes dispelled a months worth of doubt, and Gil Thornhill (Open University) for reading my thesis and for always having a sympathetic ear.

As for the people in the front line who have incurred my frequently changing moods, and whose friendship and understanding have secured my sanity on several occasions, I would like to thank everyone at the University of London Observatory Annexe and at the Observatory itself. However, there are individuals who have been particularly kind and supportive to me during my studies. To Maureen Evans, Valerie Peerless, and Peter Thomas, I would like to say a special thank you- diolch yn fawr.

These acknowledgements would not be complete without mention of my family and relations in Wales, and my friends from the year of 86 at Lancaster, who have stood by me through 'thick and thin', and who have catered for my every needs. My choice of career has been entirely of my own making and I am eternally grateful to my parents for having giving me this 'free hand' with my life.

Finally, this thesis could not have been completed without the financial support of The Natural Environment Research Council, and a 6-month bursary from The Department of Physics and Astronomy, University College London.
CHAPTER 1

Introduction

"A short time ago a severe shock of earthquake of an undulating kind was felt. The eruption of Prince of Naples Mount is increasing in severity. The lava is running rapidly in a double stream towards Nicolosi, to the great danger of the town. Panic amongst inhabitants". Nature, 27th May 1886

This extract briefly describes the fear and uncertainty which accompanied the 1886 eruption of Mount Etna volcano, Sicily, as the lava flow encroached on the town of Nicolosi on the southern flank. Fortunately, the lava flows did not advance beyond the outskirts of the town. Remove the date from the above report and it could equally well have been written a few years ago to accompany the Etnean eruptions of 1981 and 1983. Over the last two decades, Mount Etna has been in a state of remarkable activity which has frequently seen land and property threatened by lava flows. Techniques for diverting active lava flows are being perfected but, to date, these have generally proved ineffective, relying on the slow movement of the lava flow. In the case of the 1981 eruption, all the existing lava flow diversion techniques would have failed because of the rapid speed at which the lava flow was emplaced. With all the developments in the field of volcanology, the possibility of accurately predicting the extent of lava flows on the flanks of this volcano, or any other, remains beyond our grasp.

Periods of eruptive surveillance and morphometric studies have constrained the eruption parameters and external processes that influence the emplacement of individual flows, and the planimetric evolution of flow fields. General empirical relationships have been established from these studies which relate eruptive parameters (eruption rate, volume, duration) with flow dimensions (length, width) e.g. Malin
Walker (1973), Pinkerton (1987), Kilburn and Lopes (1988). In the case of Etna, these empirical relationships have been incorporated into an analysis of hazards posed by lava flows (Chester et al. 1985, Chapter 8). To supplement these studies, theoretical rheological models of lava flow emplacement have also been developed, but they await further testing and refinement (Hulme, 1974, Pieri and Baloga, 1986, Pinkerton and Wilson, 1988).

Further improvements to our knowledge of lava flows must be directed towards understanding the manner in which magma is delivered to the eruption site, for ultimately it is processes which are occurring within the volcanic system (plumbing and stress fields) which determine the overall size of an eruption (volume, duration and real-time effusion rate). It follows that temporal variations in the magnitude of eruptive parameters that control the emplacement of lava flows and their morphological evolution into flow fields must, by continuity, be related to changes in the rate of magma transport through the volcanic system. During the last few decades, the coordinated integration of a variety of scientific techniques, including those of geophysics (Einarsson, 1978; Ryan et al., 1981; Klein et al., 1987; Ryan, 1988), petrology (Scott, 1983; Wright and Helz, 1987), surveillance of eruptive activity, theoretical modelling (Shaw, 1980; Wilson and Head, 1981; Whitehead, 1986) and studies of the historical development of volcanoes, have allowed considerable advances to be made in understanding the internal plumbing of individual volcanoes. Research into the stress fields of volcanoes is at an earlier stage (Bousquet et al. 1988), being hindered by logistical difficulties on the ground and the dynamic nature of the problem. Orbital satellites, such as the planned Earth Observing System (EOS) may, by their greater scope, and potential for real-time observation, prove useful in the detection of changes in a volcano’s stress fields (Mouginis-Mark et al. 1989). However, this will depend on the accuracy and precision of the measuring instruments, which must be capable of detecting movements on the scale of a few millimetres. Despite all the available information, these two aspects of volcanology have generally developed apart, with little integration.

In this thesis, I present a study of the post-1600 volcanic system of Mount Etna, Sicily in which I examine the spatial distribution and temporal variation of eruptive activity on the volcano. A model of the volcanic system is developed, and
of the way in which it releases magma in eruption. The model is constrained as much as possible by information derived from individual eruptions and from modern eruption monitoring techniques (geophysics, petrology, geodetic, eruption surveillance and fluid dynamics). The role of the volcanic system in influencing the planimetric evolution of the lava flow/flow fields is then examined by integrating the eruptive model with the conclusions of a morphometric study of the lava flows.

A general review of the volcanology of this volcano is presented in chapter 2, which covers such details as its geological and tectonic setting, the evolution of the edifice, and the status of our present knowledge of the volcanic system. The present study of Etna is restricted to the post-1600 period, for which the greatest information is available. Since 1600, the volcano has been characterised by two time spans with different rates of output from flank eruptions: a period of high magmatic output between 1600 and 1689, and a more moderate output between 1750 and the present. During each of these periods, the magmatic output was broadly constant and the continuity of eruptive style suggests that the configuration of the magma transportation system was roughly stable. If the difference in output was the result of changes in the rate of magma generation in the mantle or enhanced magma ascent rates, then this suggests that the Etnean volcanic system, at least at depth, differed between the output periods. Differences in the high level plumbing configuration between the early 17th century and subsequent output periods have also been argued on the basis of lava petrography (Guest and Duncan, 1981; Duncan and Guest, 1982). This study therefore examines each output period separately.

In Part I of the thesis, a morphometric study of post-1600 Etnean lava flow fields is undertaken, in which the data of the early 17th century and post-1750 magmatic output periods are considered separately. This is reported in chapter 3. In addition to determining the relative influence of a variety of eruption parameters (lava volume, duration, and eruption rate) on lava flow emplacement and the planimetric evolution of flow fields, the conclusions of previous morphometric studies are also examined. The 'output' factor is discussed in order to establish whether this had any significant influence on the planimetric evolution of the lava flow fields in each output period.
In Part II, models of the post-1600 volcanic system are constructed from eruption information, lava petrography and, for the more recent period, geophysical information. The post-1750 (chapters 4 and 5) and the early 17th century (chapter 6) periods are considered separately in order to find out whether the volcanic system underwent significant change between the two periods of differing magmatic output. Amongst the components of the volcanic system to be discussed are eruptive mechanisms, potential storage areas, the magma transportation system, and regional and gravitational stress fields. The conclusions of the morphometric studies conducted in Part I, are then integrated into the model. Part II ends with a comparison of the post-1750 and early 17th century volcanic systems.

The principal conclusions of the whole thesis are summarised in chapter 7, and suggestions are made as to the possible direction of future work.
CHAPTER 2

The Volcanology of Mount Etna Volcano

2.1 Introduction

This chapter reviews our current understanding of the Mount Etna volcanic system. Starting with an overview of the volcano’s tectonic setting, this is followed by a brief discussion of its historical development, the processes which have controlled its construction, and the variety of high level plumbing configurations which may have existed during its lifetime.

The study of any volcanic system is strongly dependent on the availability of a comprehensive body of data, covering a broad spectrum of individual disciplines within the field of volcanology, to form a solid platform upon which volcanological ideas can be developed. The basaltic volcano of Mount Etna, Sicily was chosen for this study specifically for this reason. This volcano is virtually unrivalled on Earth as a place to study volcanological processes, having a record of historical eruptions which covers over 2,500 years (Chester et al. 1985). The application of modern volcano monitoring techniques to Mount Etna (Murray and Guest, 1982; Sanderson, 1982; Consentino et al. 1982) is at an early stage, compared with the more comprehensively monitored Hawaiian volcanoes. Nevertheless, Etna has been subjected to considerable scrutiny in other respects, namely petrology (Cristofolini, 1973; Rittmann, 1973; Duncan and Guest, 1982; Scott 1983; Armienti et al. 1984), magma output (Wadge et al. 1975; Wadge and Guest, 1981; Mulargia et al. 1987) and morphological examinations of its lava flow fields (Walker, 1974; Wadge, 1978; Lopes and Guest, 1982; Romano and Sturiale, 1982; Lopes, 1985; Romano and Vaccaro, 1986; Guest et al., 1987; Kilburn and Lopes, 1988). Based on the conclusions of these studies, it has been possible to place some constraints on the operative eruptive mechanisms (Wadge, 1977; Guest and Duncan, 1981; Armienti et al. 1984).
2.2 Geological Setting

2.2.1 Regional and Volcano Tectonics

The majority of geological and tectonic features of the Mediterranean region are the remnants of the Cenozoic continental collision between the Eurasian and African continental land masses. The northward subduction of the African plate beneath Eurasia resulted in the closure of the Mesozoic Tethys ocean, causing intense deformation and uplift to the north, culminating with the Alpine orogeny.

![Generalised geological and tectonic map of Sicily showing the distribution of volcanic centres (modified after Grindley, 1973).](image)

**Figure 2.1** Generalised geological and tectonic map of Sicily showing the distribution of volcanic centres (modified after Grindley, 1973).

The current geological framework of southern Italy is shown in Figure 2.1. The collision zone is considered to pass through Sicily (McKenzie, 1970) but there is little evidence for active subduction beneath the island. Here, the collision front has been masked by a series of southwardly emplaced thrust nappes which now form part of the 'outer' Calabrian-Sicilian arc. To the east, a seismically active Benioff zone was identified by Caputo et al. (1970) plunging steeply to the north west from beneath the Calabrian arc, in the direction of the volcanically active Eolian Island 'inner' arc.
The calc-alkaline volcanism of the Eolian Islands is probably associated with the consumption of the African lithosphere beneath the volcanic arc (Barberi et al. 1974) but the shoshonitic nature of the present volcanic products may indicate that the area is in a senile stage of development (Keller, 1980). This point is reinforced by the absence of oceanic type crust on the floor of the Ionian Sea to the south of the inferred collision front (Barberi et al. 1974). To the north of the Eolian Islands, the continental margin gives way to the 3-4 km deep Tyrrhenian Basin, which is floored by oceanic type crust. In a general plate tectonic context, this basin has been interpreted by Barberi et al. (1974) as representing a back arc basin that formed in conjunction with the volcanic island arc. However, as pointed out by Di Girolamo (1978), this association is most unlikely because the volcanism of the Eolian Islands is less than 1 million years old (m.y. old) whereas spreading within the Tyrrhenian Basin commenced some 5 m.y previously.

Whilst several plate tectonic models have been proposed for the region (McKenzie, 1970; Dewey et al. 1973; Barberi et al. 1974; Biju-Duval et al. 1977), close examination of the temporal and spatial sequence of geologic events reveals gross inconsistencies within each model. These arise partly due to a lack of detailed investigation (Chester et al. 1985), but generally reflect the complexity of a region where post-orogenic, micro-plate movements have had a major influence on the present tectonism (Alvarez, 1972). A detailed discussion of these models is presented in chapter 3 of Chester et al. (1985).

The location of Mount Etna within this tectonic framework (Figure 2.1) is anomalous since it lies in front of the Eolian volcanic arc, in what is normally a non-volcanic (compressional) region (Grindley, 1973). However, eastern Sicily marks a zone of local distension in which tholeiitic to alkali magmas have been erupted. These magmas are more akin to 'hot spot' mantle diapiric volcanism rather than the subduction related Eolian arc volcanism. Though the main tectonic environment in the north of Sicily is one of compression, with the dominant stress, $\sigma_1$, directed NE-SW (Lo Giudice et al. 1982), this is gradually being replaced along the eastern seaboard of Sicily by a northwardly propagating extensional regime (Bousquet et al. 1988). The extensional regime is caused by the gradual anticlockwise rotation of the Calabro-Sicilian arc about a fulcrum located within the Straits of Messina (Ghisetti
and Vezzani, 1979; Frazzetta and Villari, 1981). This has been accompanied by a contemporaneous and continuous northward migration of volcanic activity along the eastern seaboard of Sicily; commencing with Upper Miocene volcanism in the Iblean Plateau region of southern Sicily, with the present activity of the Etnean region representing the most northerly manifestation.

Figure 2.2 Rose diagram, by Lo Giudice et al. (1982) illustrating the alignment of (A) tectonic lineaments in the Etna region, in terms of a regional tectonic stress regime with the maximum compressive axis (σ1) aligned NNE-SSW and (B) linear volcanic elements on the whole of the volcano

Mount Etna lies at the intersection of several deep-seated tectonic lineaments, the summit area marking the epicentre of a tensional/dilational stress regime. Rittmann (1973) noted three principal azimuthal directions for these lineaments:

(a) the major NE-SW trending Messina fault system,
(b) the E-W trending Monte Kumeta-Alcantara fault system and
(c) the ESE-WNW trending Tindari-Leojanni fault system.

With tectonic activity continuing in the Calabro-Sicilian arc region, differential shearing movements along these conjugate lineaments (Frazzetta and Villari, 1981) generates tension at the locus of their intersection, a situation which is exploited by ascending magma (Cristofolini et al, 1977). Following the principles proposed by Nakamura (1977), Lo Giudice et al. (1982) and Lo Giudice and Rasa’ (1986) argue that the azimuthal orientation of volcanic lineaments (dykes and eruptive fissures) on the volcano correlates closely with the orientation of regional tectonic lineaments in the surrounding region and can be accommodated within a regional stress regime dominated by compression (Figure 2.2). The deficiencies of this argument are two-fold.

1. Within the main mass of the volcanic pile, McGuire and Pullen (1990) demonstrated that the effect of regional tectonic stresses becomes secondary above the basement, to the effect of gravitationally induced stresses of the volcanic pile. Above the basement, volcanic lineaments show broad radiality about the summit. Regional stresses only have a control on the azimuthal orientation of volcanic lineaments at low levels on the flanks where volcanic products only form a thin veneer on the basement rocks.

2. In situ neotectonic investigations by Bousquet et al. (1988) reveal that there is a spatial variation in the distribution of principal stress fields on the volcano, with a compressive stress regime on the northern flanks of the volcano giving way to an extensional regime on the volcano’s other flanks. This extensional regime is accentuated along the eastern seaboard of the volcano by a northwardly propagating and actively collapsing continental margin (Lentini, 1982; Lo Guidice and Rasa’, 1986) which is occurring along the Iblean-Maltese fault system. On the volcano, this feature manifests itself as a set of shallow tectonic lineaments with normal fault characteristics, trending NNW-SSW and ENE-WSW (Cristofolini, 1979), and downthrowing towards the sea. The limited vertical extent of these faults was inferred by Lombardo and Patané (1982) from changes in the attitude of seismic isoanomals with depth. These changed from NW-SE trends at shallow levels (less than 4 km) to the regional tectonic trend of NNE-SSW at deeper levels. They interpreted the shallow foci seismic events to result from sliding along rotational slip planes. In the
vicinity of Etna, the gravitational instability of this eastern seaboard is accentuated by volcanic loading (Guest et al. 1984b), and general intrusive/extrusive volcanic activity which is exploiting these high level weaknesses. More recently, activity has been located at higher levels on the volcano, on an orthogonal set of fissures, trending SSE and ENE on the western wall of the Valle del Bove.

In summary, the deep-seated tectonism which is influencing the migration of magma upwards from the mantle may bear close azimuthal orientation with the lineaments identified by Rittmann (1973). However, at high levels on the volcano, the stress fields are more complex and cannot be readily resolved into principal directions. In addition to the gravitational stress field of the volcanic pile itself, there are probably several subsidiary and localised tectonic stress fields superimposed upon the regional trend.

2.2.2 Basement Geology

The volcano is built on continental crust between 35 and 40 km thick (Lentini, 1982). These consist of a series of allochthonous Alpine thrust nappes that were emplaced southwards during the Eocene and Quaternary over the carbonate sequence of the African, or Iblean foreland. The separation surface between these sequences is estimated to be between 9 and 10 km below the present summit (Lentini, 1982). To the north of the volcano, the exposed geological sequence reveals overthrusting from the core of the nappe so that the stratigraphically highest units, that expose the more internal zones of the orogenic belt, are also the oldest (Lentini, 1982). The sequence of units, starting with the highest structural unit in the north, is shown in Table 2.1.

The Iblean foreland in southern Sicily consists of a thick sequence of carbonate deposits of Pliocene to Oligocene age, with volcanic intercalations towards the top. Away from the collision front, the Iblean foreland forms a rigid slab which has been largely unaffected by the Alpine orogeny. However, in the proximity of the collision front, there is evidence that the carbonate succession is significantly reduced, with maximum thinning occurring beneath Etna (Lentini, 1982).
### Table 2.1 Stratigraphy of Etnean basement rocks (after Lentini, 1982)

<table>
<thead>
<tr>
<th>Stratigraphy</th>
<th>Unit</th>
<th>Interpretation</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>External Zone</td>
<td>Basal Complex</td>
<td>Numidian Flysch-miogeosynclinal deposition</td>
<td>Triassic-Miocene</td>
</tr>
<tr>
<td></td>
<td>Panormide complex</td>
<td>Carbonates</td>
<td>Upper Cretaceous - Oligocene</td>
</tr>
<tr>
<td></td>
<td>Scilicide Complex</td>
<td>Allochthonous Flysch deposits</td>
<td>Upper Jurassic-Eocene</td>
</tr>
<tr>
<td>Internal Zone</td>
<td>Calabride Complex</td>
<td>Crystalline sequence - granites, gneisses</td>
<td>Variscan Age</td>
</tr>
</tbody>
</table>

### 2.3 Volcanic History

The earliest volcanic materials in the Etna region are the Basal Tholeiitic volcanics of Pleistocene age which were erupted about 300,000 years before the present (B.P) from a series of fissures along the southern margin of the present volcano (Rittmann, 1973) - Figure 2.3. To the west, subaerial lava eruptions filled the Simeto Valley with a thick lava sequence of pahoehoe morphology (Duncan, 1978). Further east, where the eruptive fissures cut the palaeocoastline, the magma was intruded into shallow submarine clays, and in the vicinity of Aci Trezza, pillow lavas and hyaloclastites formed where submarine extrusions occurred (Cristofolini, 1973). Stratigraphically, these are the lowest units on the volcano (Romano, 1980). A limited volume of alkali olivine basalt lavas were also erupted during these early stages. These lavas, which are exposed in the SW at Paterno and in the NE near Piedimonte have been dated as 210,000 years B.P. - Table 2.2.

About 150,000 years ago, following a considerable period of apparent quiescence, the composition of the erupted materials changed from the early tholeiitic phase to the Alkaline Series. There followed eruptions from a series of overlapping polygenetic centres (Cristofolini et al., 1982) producing lavas and tephra of a range of compositions, the majority being hawaiites, mugearites and benmoreites (Klerkx, 1970; Romano and Guest, 1979; McGuire, 1982) - Table 2.2.
<table>
<thead>
<tr>
<th>ALKALIC SERIES</th>
<th>Unit</th>
<th>Centres</th>
<th>Composition</th>
<th>Event</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Recent Mongibello</td>
<td>Present Centre</td>
<td>Hawaiites</td>
<td>Pit Collapses</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Piano</td>
<td></td>
<td>Caldera collapse</td>
<td>1669</td>
</tr>
<tr>
<td></td>
<td>Ancient Mongibello</td>
<td>Leone</td>
<td>Hawaiites, Basic Mugearites, Mugearites, Benmoreites.</td>
<td>Caldera collapse</td>
<td>5,000 years B.P.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Ellittico</td>
<td></td>
<td>Caldera collapse</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Belvedere</td>
<td></td>
<td>Caldera collapse</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Vavalaci</td>
<td></td>
<td>Caldera collapse</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Trifoglietto</td>
<td>Trifoglietto II</td>
<td>Basic Mugearites, Mugearites</td>
<td>Phreatomagmatic activity</td>
<td>26,000 years B.P.</td>
</tr>
<tr>
<td></td>
<td>Pre-Trifoglietto</td>
<td>Calanna (Tardaria, Trifoglietto I ?)</td>
<td>Mainly Hawaiites</td>
<td>Major caldera collapse</td>
<td>106,000 years B.P.</td>
</tr>
<tr>
<td></td>
<td>Paterno Cone</td>
<td></td>
<td>Alkali Olivine basalt</td>
<td>Eroded cone - Paterno</td>
<td>210,000 years B.P.</td>
</tr>
<tr>
<td>BASAL THOLEIITIC VOLCANICS</td>
<td></td>
<td></td>
<td>Subaerial and submarine tholeiites</td>
<td>Acicastello/Acitrezza</td>
<td>300,000 years B.P.</td>
</tr>
</tbody>
</table>

Table 2.2 The Mount Etna volcanic succession as summarised from Table 3.2 in Chester et al. 1985
The present Mongibello construct probably started to take shape some 20,000 years ago to the northwest of, and superimposed on, an older centre - Trifoglietto II (Table 2.2). Several centres subsequently developed erupting a broad range of alkalic basalts. Autoclastic domes of evolved composition on the lower flanks have been associated with these centres. Activity at each centre was largely terminated by major caldera collapse and this was followed by lava eruptions which infilled the calderas. The rims of these ancient calderas can be identified as annular breaks of slope on the volcano profile. It is possible that at least one of these caldera forming episodes was associated with the emplacement of the Biancavilla ignimbrite on the SW flank (Duncan, 1976; Chester et al. 1985).

Some 5,000 years ago, as the volcano approached its present form, a large portion of the eastern sector collapsed through a series of gravity sliding episodes. This resulted in the horseshoe-shaped hollow known as the Valle del Bove, some 1 km deep and 5 km across, that opens to the sea (Guest et al. 1984a). The mere existence of the Valle del Bove provides much of the exposure for the ancient alkalic centres and their structural relations (Chester et al. 1985). Major fault scarps sub-parallel to the coast with downthrows to the east are interpreted as strong evidence of the gravitational instability of the unsupported eastern flank (Guest et al., 1984a).

Since the formation of the Valle del Bove, the style of volcanism has been dominated by effusive eruptions of basaltic lava with subordinate strombolian activity. The summit cone has been in a state of continuous activity, certainly through much of recorded history back to nearly 700 BC. In consequence, the summit region is subjected to rapid changes in morphology as a result of strombolian cone-building phases and pit collapse (Guest, 1973; Chester et al. 1985 chapter 4).

2.4 The Volcanic Edifice
2.4.1 Volcano Morphology

The morphology of the present edifice reflects the superposition of a variety of volcanic products on major tectonic basement structures and the remnants of ancient volcanic centres. The present Etnean edifice has a basal diameter of about 40 km and rises to over 3,300 m asl. The active Mongibello edifice is displaced to the northwest of the ancient centres, giving shorter slopes on the northern and northwest
Figure 2.3 Geologic map of Mount Etna (after Hughes et al. 1990). UCL Negative Number 89/9/110.
Figure 2.4 Map of Mount Etna showing the number-density distribution of vents (modified after Guest and Murray, 1979) and tectonic lineaments (modified from Romano et al., 1979). UCL Negative Number 89/9/13.
flanks which abut against the Peloritani mountains. The eastern and southeast flanks of the volcano extend down to the coast. Generally, the slopes of the volcano take on a concave profile, with gentle slopes on the lower flanks of about 5° increasing to about 10° at 1,800 m asl. (Guest, 1980). Above 1,800 m asl., slopes steepen and may exceed 20° on the eastern and western flanks as the 10 km diameter cone of recent Mongibello activity is encountered. At these intermediate levels, there is a strong departure from the conical symmetry to the northeast and south. The topography to the northeast is dominated by the northeast rift zone. This extends some 10 km from the summit as a prominent ridge built up by marginal faulting and the accumulation of lavas and pyroclastic material (Figure 2.4). To the south, the profile of the volcano is affected by the prominence of the older Vavalaci centre and the Valle del Bove. Though a southern rift does exist, it is more subdued than its northeast counterpart, its presence being indicated by a region of high cone density on the southern flank (Figure 2.4). At 3,000 m asl., the remains of the young Piano caldera is visible as a break in slope around the summit cone. Arguably active up till the great eruption of 1669 (Chester et al. 1985), this feature is now filled with young lavas and has been surmounted by the present summit cone. Though the volcano is dominated by the steep-sided Mongibello cone, more than 80% of the surface area occupied by the volcano has slopes less than 10° (Walker, 1977). For this reason, the present edifice takes on the appearance of a strato volcano (Mongibello) built upon a 'primitive' shield volcano (Tanguy, 1978). However, a comparable gross morphology could be achieved by spatially varying the eruptive style on the edifice and increasing the frequency of summit eruptions so that steep slopes develop around the summit (Guest and Murray, 1979; Lopes and Guest, 1982).

2.4.2 Eruptive Styles

**Summit Activity**

Activity at the summit is dominated by mild strombolian activity and almost continuous magmatic outgassing through the summit craters. This is known as 'persistent' activity (Rittmann, 1962). Effusive eruptions occurring at the summit are termed terminal eruptions and last for periods of time ranging from days to years. The eruptive fissure, which may be exploited several times, is supplied by magma
ascending through the central conduit system of the volcano. Though relatively
violent explosive activity often occurs over the vent, lava is generally extruded at low
overall rates (< 1 m$^3$s$^{-1}$) (Guest, 1982). For a sustained eruption, this approximates to
the mean output of the volcanic system (Chester et al. 1985). Occasional high
effusion rate eruptions have occurred from the NE crater, but these tend to last only
a few hours. A series of these eruptions occurred between the 16th July, 1977 and
the 28th of March 1978 (Chester et al. 1985). The majority of lava flows erupted
from the summit achieve lengths of 3 to 4 km on the steep sides of the summit cone,
but a few longer flows, such as the 7 km March 1978 lava flow on the NW flank
(S.E.A.N. 1989), have been produced during the high effusion rate summit eruptions.

Flank Activity

Flank eruptions (Rittmann, 1962), defined as those eruptions occurring away
from the summit cone, occur less frequently, roughly averaging 1 every 10 years
(Guest, 1980), though in the 1980’s there were five. Lava extrusion occurs initially
at high effusion rates (in excess of 10 m$^3$s$^{-1}$), the eruptive vent being supplied by
lateral dykes that tap the central conduit magma column at various depths below the
summit (Wadge, 1977; Guest and Duncan, 1981; Murray and Pullen, 1984). Normally, flank eruptions are preceded by a lull in persistent activity (Sharp et al.
1981). Explosive activity is normally established at the eruptive vent once the
eruption has commenced, though it is not uncommon for degassing to take place at
an intermediate location along the eruptive fissure e.g. 1928 eruption; Imbó (1928) and
1983 eruption; Frazzetta and Romano (1984). As the energy of the eruption dwindles,
persistent activity becomes re-established at the summit craters.

Above 1,800 m asl., the eruptive fissures of flank eruptions display a broad
radiality about the summit cone (McGuire and Pullen, 1989), but at lower elevations,
they are generally restricted to well-defined zones. To the northeast and south, flank
eruptions exploit the tectonic basement weaknesses of the northeast and southern rift
zones (Figure 2.4). Re-opening of the rift zones is normally followed by eruptions
from the distal portions of the rift. Subsequent eruptive episodes along the same
fissure system tend to emanate at higher elevations eg. 1883-1886-1892 eruptions
(Ricco, 1910), as the fractures are gradually infilled with 'frozen' dykes.
Another style of eruption, which shares some of the characteristics of both flank and terminal activity, is sub-terminal activity (Rittmann, 1962; Armienti et al., 1989). This style of activity is generally characterised by the quiet effusion of lava on the upper flanks bordering the central craters. Degassing mainly occurs through the summit craters, though explosive activity at the vent is not uncommon e.g. April, 1978 eruption near the SE crater (S.E.A.N., 1989). It is difficult to place an exact constraint on this style of activity because of its transitional characteristics between the eruptive style end members. It is therefore not surprising that eruptions occurring outside the spatial limits suggested by Rittmann (1962) have been ascribed to this type of activity, e.g. 1985 (A) eruption on the southern flank (Romano and Vaccaro, 1986).

Eruptions on the flanks with a magma supply independent of the volcano’s central conduit system were termed ‘eccentric’ by Rittmann (1962). At least two of the more recent eruptions, those of 1974 on the western flank, have been classified accordingly (Tanguy and Kieffer, 1974; Duncan and Guest, 1982). Tanguy (1974) distinguished these particular eruptions from the majority of flank eruptions by their high explosive activity; unusual macroseismic activity; the sluggishness of the erupted lavas, and the lack of any noticeable correlation between summit and flank activity. Further evidence for the ‘eccentric’ nature of these eruptions are revealed in the petrography. The lack of plagioclase phenocrysts in the 1974 lava is considered to be due to the magma rising directly from an environment at depth where the $P_{\text{H_2O}}$ was sufficiently high to suppress plagioclase crystallisation to near solidus temperatures (Tanguy and Kieffer, 1976; Duncan and Guest, 1982). The 1974 feeder system was therefore not linked to the central conduit feeder network at shallow depth, where the storage or slow ascent of magma allows plagioclase crystallisation to occur.

2.5 The Plumbing System of Mount Etna
2.5.1 Post-1971

The high level plumbing system of Etna was first explored in 1971 (Wadge et al., 1975) with the initiation of a geodetic survey around the summit region. Radial tilts recorded in this early period were interpreted by Wadge (1976) as reflecting strain induced by the presence of a cylindrical column of magma within the central conduit system of the volcano. Wadge (1977) envisaged the central conduit system of the
volcano above 1800 m asl as an important storage area with a volumetric capacity of 42 x 10^6 m³.

The areal extent of the geodetic survey was increased in 1975 and a permanent network of benchmarks established (Murray and Guest, 1982). The observed pattern of deformation over a five year period did not conform to that already established at the Hawaiian volcano of Kilauea (Fiske and Konoshita, 1969), where ground deformation is interpreted as reflecting discharge from, and recharge into, a high level storage area (Mogi, 1958). However, the pattern of deformation on Etna did correspond to the filling and storage of magma in high level dykes, some of which subsequently propagated to the surface and fed effusive eruptions e.g. the 1978 and 1979 eruptions (Murray and Guest, 1982).

More recently, geophysical studies have supplemented the available geodetic data (Consentino et al., 1982). The intrusion and movement of magma or volcanic fluids typically generate detectable earthquakes and/or vibrations (Banks et al., 1989). Not only does this provide precursory warnings of possible eruptive activity, it also provides valuable insight into the high level storage and migration of magma. On the Hawaiian volcanoes, a broad aseismic region exists immediately beneath the summit calderas. This has been interpreted as a magma-filled storage area (Ryan, 1988). No comparable region exists for Mount Etna where earthquakes take on a random distribution on the volcano (Gresta and Patanè, 1987). Eruptive and seismic phenomena occur mainly along the principal structural trends, but often the directions of the eruptive fractures and earthquake concentrations during the same eruption do not coincide (Gresta and Patanè, 1987). The stress induced into the volcanic pile by the injection of a new dyke is therefore relieved by the release of seismic energy along other tectonic lineaments on the volcano. During most eruptions, seismic epicentres become particularly concentrated along eastern flank lineaments (Lombardo and Patanè, 1982; Consentino and Lombardo, 1984; Cristofolini et al., 1985; Gresta et al., 1990). The inability of this flank to accumulate great stresses adds weight to arguments that this region of the volcano has been weakened by gravitational movements.

Seismic events associated purely with the injection of magma which is propagating feeder dykes tend to follow a characteristic pattern. The majority of flank
eruptions are preceded by a seismic crisis as the feeder dyke propagates to the surface (Sharp et al. 1981, Glot et al. 1984), the pattern of seismic release being concentrated around the propagating dyke. Volcanic tremor (low level seismic signals caused by the rapid motion of a vapour-gas-magma mixture at high levels in the volcanic system (Seidl et al. 1981 and Schick et al. 1982)) increases from low levels before the eruption to a maximum soon after magma flow has been established (Tanguy and Patanè, 1984). As the energy of the eruption diminishes, this gradually falls back to background levels. Though the above pattern is generally repeated before the majority of flank eruptions, there are exceptions. The 1985 eruption on the southern flank was not preceded by a seismic crisis. It is conceivable that the feeder dyke entered an open fissure (Gresta and Patanè, 1987), possibly the remnant of the 1983 fissure.

Seismic evidence therefore suggests that the feeder dykes of flank eruptions are emplaced during or just before the eruption commences. Certainly, this is suggested by those eruptions where the feeder dyke underwent incremental extension over several days, such as 1928 (Wadge, 1980) and 1971 (Rittmann et al. 1971). However, gravimetric (Sanderson, 1982) and geodetic surveys (Murray and Guest, 1982; Murray and Pullen, 1984) provide evidence to the contrary. Variations in the local gravity field and geodetic movements were detected in the vicinity of the 1981 and 1983 eruption sites up to a few years before the actual eruption occurred. In both cases, the observed phenomena was interpreted in terms of the passive movement of magma into a fissure, opened in response to a dilational stress regime. If this is the case, the seismic crisis associated with these eruptions might relate to the release of energy around a magma pressure pulse which propagates rapidly along the feeder dyke towards the eventual eruption site. Improved monitoring and more research is required to resolve this problem.

Apart from a comparable increase in volcanic tremor, summit eruptions, in contrast to flank eruptions, are characterised by a lack of any significant seismic phenomena prior to the event. It is most probable that the magma feeding these type of eruptions ascends rapidly from depth, exploiting established pathways within the central conduit system (Sharp et al. 1981). Rapid magma ascent rates have also been advocated by Sanderson (1982) to explain the absence of detectable gravity changes prior to the 1980 summit eruptions. Sanderson argued that any change of mass
associated with the ascending magmas probably occurred over short time scales which could not be detected by the infrequently repeated gravity resurveys.

Inferences about the deeper levels of the Etnean volcanic system were first made by Machado (1965). The localised attenuation of P-wave phases from the 1908 Messina earthquake as they passed below Etna, led Machado to infer the presence of a deep reservoir. A systematic seismological search for deeper storage areas was made in 1977 using an approach which utilised the travel times of both natural and artificial seismic waves as they passed below the volcano. A theoretical model fit to the data suggested a large storage area with an ellipsoidal shape, at a depth of 20 km below the volcano. Optimum fit for the model constrained the dimensions of the storage area to 22km x 31km x 4km, with elongation in a NNE-SSW direction in close correspondence with the predominant regional tectonic trend (Sharp et al., 1980). This body was interpreted to contain about 14% of melt (1600 km$^3$) trapped in a complex network of fissures.

Recent petrological work has contributed further to the understanding of Etnean magmatic processes. Historical lavas erupted since the 14th century have broadly uniform hawaiitic compositions, indicating limited differentiation. However, the strongly porphyritic nature of the lavas, with a minimum phenocryst content of 30%, indicates that the magma undergoes significant crystallisation before eruption. The predominance of plagioclase as the main phenocryst population suggests that crystallisation occurs at moderate to low pressures (Guest and Duncan, 1981). Experimental work on hawaiitic melts indicates that plagioclase crystallisation is suppressed to near solidus temperature at pressures of about 5 kbars (Green et al., 1967). For a melt with a density of 2,700 kgm$^{-3}$, this pressure is equivalent to a depth of 22 km. Therefore, lavas in this period undergo substantial crystallisation at high levels in the volcanic system, but are erupted before any significant differentiation can occur. This is in agreement with the views of Rittmann (1973).

Examination of the petrographic nature, trace element and isotope chemistry of recent lavas reveals more subtle variations. Detailed analyses of the textures and compositions of plagioclase phenocrysts show that in some lavas more than one plagioclase population is present. This has been interpreted by Duncan and Preston (1985) as the result of magma mixing. Chemical variations observed during the 1981
eruptive episode (Scott, 1983) support the magma mixing hypothesis, the detected variations being related to the mixing of fresh magma introduced into the volcano in 1980 with residual magma from the 1979 eruption. Furthermore, Gyopari (1988), on the basis of incompatible trace element ratios, identified mixing between several end member compositions within a single eruptive episode. In certain examples, one end member can be related to an earlier eruptive episode. In the 1971 eruption, fresh magma appears to have mixed with residual magmas that had incompatible trace element ratio signatures of lavas erupted in 1923 and 1928. If the proposal of Gyopari is correct then magma batches may remain isolated within the volcanic system for at least 50 years, and thus be available for subsequent mixing in later intrusive and eruptive episodes. The limiting factor would be the residence times of magma before cooling and solidifying. Theoretical calculations by Wilson and Head (1988) show that dykes up to 5 m in thickness, emplaced into a volcanic pile, would solidify within 2.5 years. The magma batches on Etna would therefore either need to be thick and stored at depth to reduce the effects of cooling, or stored in zones where high magma transport rates keep the pile 'hot'.

The analysis of lava samples collected regularly over some of the more recent, long duration eruptions (>50 days) reveals that the erupted lavas become chemically more primitive as the eruption proceeds. This phenomenon, which was detected during the 1983 and 1985A eruptions (Armienti et al. 1984 and 1987), may represent the influx of fresh magma into high levels.

Geodetic, geophysical and petrological evidence therefore suggests that no large capacity high level magma storage area has existed on Mount Etna over the last 20 years. Any storage which does occur is limited to within the central conduit system and a network of high level dykes (Rittmann, 1973; Wadge, 1977; Murray and Guest, 1982; Guest and Duncan, 1981; Duncan and Guest, 1982). However, these observations may not be representative of the entire historical period.
2.5.2 Pre-1971

The petrological examination of Etnan lavas provided much of the basis for discussion of Etnas’ plumbing system prior to the advent of geodetic and geophysical surveys. Cristofolini (1973), in order to account for the evolved suites of prehistoric lavas present in the volcano’s stratigraphy, suggested that the magmas underwent significant high level storage before eruption. In addition, he argued that the volumes

![Diagram showing size frequency distribution of plagioclase phenocrysts > 0.4 mm (longest axis) for: (A) a 17th century lava flow, 1669, showing ‘cicirara’ texture and (B) a post 17th century lava, 1843. For both (A) and (B), the area of measurement was 250 mm². (Diagram from Hughes et al. 1990)
of magma required for the process were greater than could be stored in a simple dyke system. As described in section 2.3, the remnants of several prehistoric summit calderas testify that high level storage areas have been present at intervals during the past (Guest, 1980).

The time periods over which these changes in plumbing occur may be quite short. Changes in volumetric output and lava petrography which have occurred in the last 400 years of Etna’s history have been attributed to differing high level plumbing conditions (Wadge et al. 1975; Guest and Duncan, 1981; Duncan and Guest, 1982). Lavas erupted in the first eighty nine years of the 17th century are characterised by a distinctive porphyritic texture in which large plagioclase phenocrysts are enclosed within a fine grained groundmass. Locally, this petrographic texture is termed 'cicirara' (Rittmann, 1973). The lavas erupted after 1689 show quite a different size frequency distribution pattern in the plagioclase phenocrysts (see Figure 2.5). The petrographic texture of these lavas is generally characterised by smaller, but more abundant plagioclase phenocrysts, though, as described in section 2.4.2, the 1974 lavas are an exception.

The contrasting plagioclase crystallinity of the 17th century lavas and those erupted after 1700 clearly indicates a difference in the nucleation rate and crystal growth histories of the magmas. The 17th century lavas crystallised under conditions which favoured a low nucleation rate but enhanced crystal growth. Guest and Duncan (1981) and Duncan and Guest (1982) suggest that such conditions could be met if the magmas, prior to being erupted, were stored within a substantial high level reservoir, where cooling was inhibited. Lavas erupted since the 1689 eruption have abundant small plagioclase phenocrysts, indicating their crystallisation occurred under conditions which favoured a higher nucleation rate. For this petrographic texture to develop, the magma probably experienced more substantial undercooling prior to eruption. A scenario in which magma ascends rapidly through the volcanic system, with limited storage in small dyke-like bodies provides a feasible environment in which this texture could develop. This is consistent with the conclusions of the geodetic and geophysical studies of the last 20 years and may indicate that the volcanic system, as far as magma crystallisation is concerned, has behaved fairly uniformly since 1689.
Figure 2.6 Volumetric output of Mount Etna since 1535 AD (modified after Wadge et al., 1975 using an updated data set (section 3.2)). The calculated output, with the exception of the period from 1971 to 1989, is for volumetric contributions from flank activity only. The mean output for the first half of the 17th century period (a) is 1.2 m$^3$s$^{-1}$ compared with that from the mid 18th century onwards (b) of 0.2 m$^3$s$^{-1}$.

These periods of contrasting petrography coincide with periods of variable magma output (Wadge et al., 1975). Since 1535, Etna has shown four periods of differing output - Figure 2.6. The first, from 1535 to 1600 was a period of low output and was followed by a period of higher output (1.2 m$^3$s$^{-1}$) which lasted from 1600 to 1689. Following the small eruption of 1689 on the eastern flank, a low output period occurred during which there was no major flank eruption. The resumption of flank activity soon after 1750 heralded the start of the present period of moderate output (0.2 m$^3$s$^{-1}$) which is continuing to the present day.
Interpretation of the output diagram (Figure 2.6) is not as straightforward as it appears because the data used to construct each output trend are different. Within the high output period of the early 17th century, the volcano was in flank eruption for over 20% of the time. There is little evidence for any other style of activity. In contrast, within the present post-1750 output period, eruptions are known to have occurred both at the summit and on the flanks. However, the measured volumetric output for this particular period was primarily constructed from flank eruptions data only (Figure 2.6), volumetric contribution from summit activity being unavailable prior to the 1950’s. It is difficult to quantify the contribution from ‘persistent’ activity prior to the 1950’s but, if the ratio of volumetric contributions from persistent and flank eruptions measured over the last 10 years is considered representative of the whole output period, then an estimate of the total output can be achieved by adding the flank and persistent volume contributions together. Chester et al. (1985) (chapter 4) demonstrate that the affect of this correction was only to slightly increase the post-1750 total volumetric output. There was still a discernible difference between the outputs of the two periods.

The low output from flank eruptions in the period between 1689 and 1750 may have been because activity was concentrated at the summit. Immediately following the eruption of 1689, the volcano may have been undergoing a period of reconstruction at the summit following the major collapse which occurred in 1669 (Wadge et al. 1975). At the other end of the scale, the apparent increase in output since 1971 may indicate a change in the rate of magma resupply from depth, but it could also be a consequence of including contributions from terminal and sub-terminal activity in the cumulative volume totals.

2.6 Summary

In summary, generalised models of the plumbing system of Mount Etna reveal it to be complex and variable with time. Petrographic and volumetric output arguments suggest that major changes have occurred within the volcanic system over the last 400 years (Wadge, 1977; Guest and Duncan, 1981; Duncan and Guest, 1982). For the period from 1600 to 1689, the volcanic system was characterised by a high magma production rate, and long magma residence times at high levels. The variation
in the observed output with time for the post-1600 period could relate to changes in mantle conditions (Wadge et al., 1975), but it is also possible that conditions within the high level plumbing system may have varied, with extra magma reservoirs, in addition to the central conduit system, being available to supply the higher output, 17th century period (Wadge, 1977; Guest and Duncan, 1981; Duncan and Guest, 1982). In the following period of moderate output, which is still continuing, evidence suggests that magma rises rapidly to the surface with little storage at high levels.
PART I: Post-1600 Etna Lava Flows: A Morphometric Study.
CHAPTER 3

A Morphometric Study of Etnan Lava flow fields

3.1 Introduction

This chapter reports a morphometric study of the post-1600 lava flows of Mount Etna, Sicily, which identifies those eruption parameters that have controlled their emplacement, and planimetric evolution into flow fields.

As a means for understanding the emplacement histories of lava flows, morphometric studies have proved a valuable tool in developing general empirical relationships between eruption parameters and flow dimensions (Walker, 1973 and 1974; Malin, 1980; Lopes and Guest, 1982; Romano and Sturiale, 1982; Lopes, 1985; Pinkerton, 1987; Kilburn and Lopes, 1988). Up till now, the lava flows of only three basaltic volcanoes, those of Mount Etna, Sicily, and Mauna Loa and Kilauea on Hawaii have provided most of the data for these studies. There is clearly a need for similar studies to be carried out on the lava flow fields of other volcanoes covering a greater compositional range e.g. the carbonatite lavas of Ol Doinyo Lengai, Tanzania (Dawson et al. 1990, Keller and Krafft, 1990).

A large data set was a pre-requisite for the statistical approach adopted by previous Etnan and Hawaiian morphometric studies. However, in choosing such large data sets relating to all the available data, these general studies may have overlooked potential modifying factors which might have influenced the relationships established. For example, conditions within the Etnan volcanic system, as evidenced by periods of contrasting output and petrography, have shown some variation since 1600 and it is conceivable that eruptive conditions may have been different in each output period. Similarly, the data used in the Hawaiian study were an amalgamation of information from two individual volcanic systems (namely Kilauea and Mauna Loa), each of which has been characterised by a distinctive output (which has shown
independent temporal variation (King, 1989)) and petrology (Wright, 1971; Tilling et al. 1987).

A comparative examination of the existing studies also reveals some basic differences in their approach to data presentation. The Hawaiian study of Malin (1980) was restricted to flow units only whilst the Etna study of Lopes and Guest (1982) and Lopes (1985) did not discriminate between flow units and flow fields. As they stand, the conclusions of these studies are not compatible.

In this study, the Etna lava flow data are streamlined so that they are consistent with the data used in the morphometric studies of other volcanoes. In addition, in contrast to previous Etnean morphometric studies, the lava flows of each output period are considered separately.

3.2 Data Collection

Though Mount Etna has a documented historic record extending back over 2,000 years (Chester et al. 1985), the post-1600 period was chosen for this study as it represents the most complete part of the historical record. There is a decrease in the quality and detail of eruption reports before this time, and many of the lava flows have been covered by younger flows. Even within the post-1600 period, the reporting of individual eruptions is highly variable. The best documented eruption reports of the pre-1750 period are those where the lavas encroached on centres of population. Eruptions high on the flanks away from the centres of population went largely unnoticed. For this reason, the dating of a few eruptions, such as the eastern eruption of 1651, have been questioned (Tanguy, 1979; Romano and Sturiale, 1982). Unless stated otherwise, the author follows an eruption chronology based on the geological map of Romano et al. (1979).

A concerted effort was made to obtain as much information as possible from original eruption reports. When this proved difficult, data compiled during previous morphometric studies was called upon (Romano and Sturiale, 1982 and Lopes, 1985). Duration data were supplemented, where necessary, from the eruption chronicles of Tanguy (1981). Length and altitude measurements were taken primarily from the 1:50,000 geological map of Romano et al. (1979), examined in conjunction with historical records, and a set of aerial photographs. The early geological map of Von
Walterhausen (1848) proved useful when examining flows erupted during the early 18th century.

The eruption information used in all chapters is presented in this section. For each of the eruptions considered, the following parameters were measured where possible: vent elevation, elevation of flow front, horizontal distance of eruptive vent from central conduit, flow length, lava flow surface area, erupted volume, eruption duration and eruption rate. The criteria followed when making each measurement are outlined in the following subsections.

3.2.1 Altitude (m. asl.)

The altitude of the eruptive vent was taken as that point where the principal flow emerged from the ground. Since this flow did not always emerge from the lowest point of the eruptive fissure (as revealed by aerial photographs), there is some discrepancy between the quoted altitude and that shown on the geological map of Romano et al. (1979). Where the eruptive vent was masked by cinder cones, the flow source was taken to be that point where the principal flow emerged from beneath the cone.

The elevation of the flow front was taken as that point on the flow/flow field most distant from the vent. Where the flow front was buried and could not be seen, as in the case of the 1643 lava flow, the flow front elevation of the exposed flow length is presented. The corresponding error assumes that the flow front of the unexposed portion of the flow coincides with the flow front elevation of the overlying flow.

For the majority of lava flows, elevation errors are considered no greater than the topographic contour interval of the geological map (±25m), but this may increase for some summit flows where the exact locations of the vents are not accurately known.

3.2.2 Horizontal Distance between Eruptive Vent and Central Conduit, Lv, (km)

Though it is not possible to directly measure the length of feeder dykes, the radial to sub-radial orientation of fissure trends to the summit, and the fact that most eruptive fractures propagate to the eruption site from the locality of the summit area
suggests that there is a connection between the central conduit system of the volcano and the eruption site in the form of a dyke. Making this assumption, the length of the feeder dyke can be approximated to the horizontal distance between the central conduit system and the eruption site.

Figure 3.1 Sketch of the summit area of Mount Etna. The star indicates the assumed axis of the central conduit system from which all horizontal distance measurements were taken.

For the purpose of this study, the axis of the central conduit system in both the post-1750 and 17th century magmatic output periods is assumed to lie in the middle of the present central craters, as located on the geological map of Romano et al. (1979) - see Figure 3.1. Dyke length was measured from this fixed point to the eruption site along a line generally perpendicular to the contours. Though eruptive fractures have been known to leave the summit area from the locality of the northeast (1911 fissure, Platania, 1912) and southeast (1983, Murray and Pullen, 1984; and 1989
fissures, Frazzetta and Lanzafame, 1990) craters instead of the central craters, the fact that persistent activity was occurring contemporaneously within the central craters during some of these eruptions, points to an interconnection. For these particular eruptions, measurements were first taken between the inferred axis and the summit crater from which the dyke propagated, and subsequently, from the summit crater to the eruption site.

In some cases, as in 1983, dyke length was obtained by directly following the line of dry surface fractures (Bousquet et al. 1984). For those eruptions where lava effusion occurred simultaneously from spatially independent locations, only the distance to the lowermost active vent was considered eg. 1978, 1979 (Guest et al. 1980a). There is therefore a tendency to underestimate the length of active dyke involved in these eruptions. Measurements were made using a cartographic rule to a accuracy of ±50m.

3.2.3 Flow Length, L, (km)

Flow length was taken to represent that distance between where the flow emerged from the vent and the most distal point on the flow front. Measurements were made perpendicular to the contours. Where the vent areas and flow fronts of flow fields had been covered, and for those flows erupted subsequent to the publication of the geological map of Romano et al. (1979), individual eruption reports were consulted. Once again, measurement was made using the cartographic rule.

3.2.4 Surface Area, A, (km²)

Several different methods have been used in the literature for measuring surface area. The standard method has been the grid/point technique which involves counting mm squares, but in a more recent study, Lopes (1985) used a planimeter. The precision of the two methods is considered comparable but the accuracy is largely dependent on the degree of flow exposure. For this study, the surface area measurements of Lopes (1985) were used. These were checked and supplemented where necessary by measurements made using the grid/point technique. Lopes (1985) considers the accuracy of the planimeter technique to be in the order of 5% for those flows with good exposure, but may decrease to about 25% for some poorly exposed
flows. The map of Von Walterhausen (1848) proved useful for delineating the outline of buried early 18th century flows. However, tracing the outlines of a few buried 17th century flows, such as 1651-53 (9), involved some interpolation.

3.2.5 Erupted Volume, V, (x $10^6 \text{ m}^3$)

The method most commonly used to calculate erupted volume has been to multiply the surface area of a flow field by its average thickness. The thicknesses of simple flows on steep slopes are easily measured, but it is more difficult to estimate a representative average thickness for an entire flow field because of the numerous superimposed flow units, and the uneven distribution of mass (Wadge, 1977). Errors using this technique may exceed 50% in the worst instances. More recently, a cartographic approach has enabled a more accurate form of thickness measurement. In this method, the thickness of the flow field is obtained by subtracting the pre­eruption topography (obtained where possible from cartographic maps) from the post­eruption topography (as obtained by resurveying). This has been done for the 1950-51 flow field (Tanguy, 1983) and the 1981 flow field (V.E.S.T., 1981). This technique, which is argued to produce volumetric data with an accuracy of ±10%, is dependent on the availability of accurate and regularly updated cartographic maps.

During the 1983 eruption, Frazzetta and Romano (1984) estimated the daily volumetric output at the vent by multiplying the surface velocity of the lava by the cross-sectional dimensions of the lava channel. Though few measurements were made in any one day, the resulting values were extrapolated over 24 hours. The volume of the final flow field was extracted from a volumetric output-eruption duration diagram (area under the curve). They combined this technique with the former to reduce possible errors to ±20%.

On Etna, the cartographic approach can only be applied to the more recent period of the historical period, for which suitable maps are available. Cartographic maps of Etna do not exist prior to the mid-19th century. Whilst this approach does provide a means for checking the accuracy of recent volumetric estimates, the above limitation restricts its use in this study of the post-1600 period. The surface area - flow thickness technique is the only approach available which can be used to estimate the volumes of eruptions back to 1600.
It was felt that no improvement in accuracy would be achieved by repeating the surface area - flow thickness technique for volume calculation, without the implementation of an independent programme of extensive flow thickness measurement on the volcano. For this study, the existing volumetric data set was used by the author to calculate, where possible, a weighted mean volume value for each eruption. This involved the collation of available volumetric values for individual eruptions, allocating each individual volume value an error (according to the relative accuracy of the calculation technique), and then calculating a weighted mean.

The volumetric data had to undergo considerable pre-sorting in order to achieve a consistent data set. A few of the literature sources presented volume values which reflected the total volume of material extruded (both explosive (pyroclastic) and effusive (lava flows), Wadge, 1977; Romano and Sturiale, 1982), whilst others provided only the volume of the effusive material (Cumin, 1954). In order to maintain a consistent data set, only contributions from the latter were included in the volumetric data set. In the majority of Etnean eruptions, the volumetric contribution from explosive material is seldom greater than the volume error on the effused material, and can therefore be ignored. None of the volumetric values were converted to their dense rock equivalent. Precision is limited to \(1 \times 10^6 \text{ m}^3\) for the larger volume flows (>5 \(\times 10^6 \text{ m}^3\)) and 0.5 \(\times 10^6 \text{ m}^3\) for flows smaller than this value.

The weighted mean equation is given by:

\[
\mu = \frac{\sum (x_i / \sigma_i^2)}{\sum (1/\sigma_i^2)}
\]  

(3.1)

where \(x_i\) is the measured parameter (in this case volume) and \(\sigma_i\) is the error on the measured parameter. For each weighted mean, the uncertainty on the mean can be calculated using the equation

\[
\sigma_\mu = \left( \frac{1}{\sum (1/\sigma_i^2)} \right)^{0.5}
\]

(3.2)
As with all calculations of this nature, the significance of the result improves with a large data set. In this study, a maximum of only five data points were available for consideration in the mean. Care was taken not to include volumetric data which had propagated through several data sets. For those studies which had used the surface area - flow thickness method for calculating the erupted lava volume, whilst little could be done to check the flow thickness data, the accuracy of the surface area measurements used in the calculations could be compared with those of the author and Lopes (1985). Where in the consideration of the author, the error on the volume estimate was greater than 50%, the data point was rejected. On a cautionary note, many of the volume estimates for the pre-20th century flows were made after the flow fields had been significantly covered. Though there is an apparent consistency in the range of estimates provided in the literature for a few of these flows, little can be done to check their reliability.

The accuracy of 17th century volumetric data are difficult to determine due to a number of factors. Whilst some measurement of surface area has been possible, a realistic estimate of flow thickness is very difficult to determine because the pre-eruption topography is unknown. In addition, many flows are only partially exposed and have been repeatedly covered. This may explain why there are fewer volumetric estimates for these flows, and for those estimates which are available, the considerable range in the quoted values. Errors are generally considered not to exceed 30% though on some of these values, such as 1634-38 (6), they could be as much as 100%.

The data used in the post-1600 weighted volume calculations are presented in Appendix 1.

3.2.6 Eruption Duration, \( T \), (Days)

As with most other eruption data, the quality of the duration data decreases with age. Duration is normally quoted to the nearest day, though over the last decade of improved eruption surveillance, precision and accuracy has improved to a few hours, or even the nearest 30 minutes as in the case of the 1981 eruption (Guest et al. 1987). Eruption durations, with the exception of persistent effusions, are well documented for the majority of post-1750 eruptions, but the data become less reliable for the 17th century period, with possible inaccuracies of several months. However,
on a comparative basis, the longer duration of the majority of the 17th century flows, up to an order of magnitude greater than those of eruptions in the post-1750 period, largely counteracts the problem of inaccuracy.

3.2.7 Eruption Rate, $Q_a$ (m$^3$s$^{-1}$)

The real-time rate at which a volume of lava is erupted from a volcano is known as the effusion rate. In the field, this is typically determined by measuring the surface velocity of the lava flow near the vent and multiplying it by the cross-sectional area of the channel. However, even amongst recent eruptions, logistical problems generally make it a difficult parameter to measure. A more commonly calculated parameter is the average effusion rate or eruption rate (Wadge, 1981). Though this parameter has the same units as the actual effusion rate, it differs from it in that it is

![Diagram of effusion rate and eruption rate](image)

Figure 3.2 Variation of the actual effusion rate with time for a hypothetical basaltic eruption (Curve A). The curve consists of a waxing phase, in which the effusion rate rises rapidly to a peak, and a waning phase, in which the effusion rate decreases progressively with time. The eruption rate (B), calculated by dividing the total volume of material by the duration of the eruption, is also shown (modified from Wadge, 1981).
the product of the total volume erupted divided by the eruption duration. It is favoured to the actual effusion rate in that it can readily be calculated at the end of an eruption. It is the *eruption rate* which is presented here.

Care must be taken not to associate these two effusion rate parameters too closely because there are clearly important differences between them (Figure 3.2). The actual effusion rate is known to vary significantly during the course of an eruption (Cumin, 1954; Walker, 1973; Wadge, 1981); a point which will be discussed further in a later section. For the eruption rate, all errors propagate from the volume and duration data, with greatest errors, possibly in the region of 50%, occurring at low volumes and short durations.

*The data for the 17th century period are tabulated in Table 3.1, whilst those for the post-1750 period are tabulated in Table 3.2. All eruptions have been given an identification number in these tables (column 1). Where there is reference to a particular eruption in the thesis, this number appears in brackets after the eruption date e.g. 1832 (30).*

### 3.3 Method of Analysis

For this study, the eruption data were firstly divided into one of two 'output' bins relating to the high output period from 1600 to 1689, and the present period of moderate output which commenced in 1750. Subsequently, the data within each 'output' bin were separated into 'morphological' bins relating to lava flows composed essentially of a single flow unit, and flow fields which consisted of several flow units.

The analysis then followed the same format used in previous morphometric studies. The data were plotted on simple two-dimensional diagrams in order to test relationships found by other authors (Walker, 1973 and 1974; Wadge, 1978; Malin, 1980; Lopes and Guest, 1982; Lopes, 1985; Pinkerton, 1987; Kilburn and Lopes, 1988) and to look for new relationships. Where necessary, fully-weighted linear regression lines were computed between variables, and correlation coefficients calculated in order to test existing relationships. The formulae for these calculations were kindly made available to the author in the 'Monticor' statistical package developed by Dr. P.C.T. Rees. Relationships were considered to be significant when
the calculated probability of a random correlation was less than 1% (P < 0.01) (Langley 1968).
<table>
<thead>
<tr>
<th>Id. No.</th>
<th>Year</th>
<th>Vent Alti. (m asl.)</th>
<th>Front Alti. (m. asl)</th>
<th>$L_1$ (km)</th>
<th>$L$ (km)</th>
<th>$A$ ($\text{km}^2$)</th>
<th>$V$ ($\times 10^4 \text{m}^3$)</th>
<th>$T$ (Days)</th>
<th>$Q_a$ ($\text{m}^3\text{s}^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1607</td>
<td>2075±25</td>
<td>800±25</td>
<td>3.7±0.05</td>
<td>10.7±0.05</td>
<td>6.25±0.62</td>
<td>57±10</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>1610</td>
<td>2500±25</td>
<td>1250±25</td>
<td>2.5±0.05</td>
<td>6.8±0.05</td>
<td>5.56±0.56</td>
<td>27±5</td>
<td>86_{1}^{±2}</td>
<td>3.6±0.7</td>
</tr>
<tr>
<td>3</td>
<td>1614-24</td>
<td>2550±100</td>
<td>960±25</td>
<td>3.3±0.15</td>
<td>8.5±0.15</td>
<td>7.61</td>
<td>-</td>
<td>3650_{1}^{±182}</td>
<td>3.8±0.8</td>
</tr>
<tr>
<td>4</td>
<td>1634-38</td>
<td>1950±25</td>
<td>810±25</td>
<td>5.6±0.05</td>
<td>7.8±0.05</td>
<td>19.99_{2}^{±2.0}</td>
<td>1200±260</td>
<td>1225_{1}^{±182}</td>
<td>1.9±0.6</td>
</tr>
<tr>
<td>7</td>
<td>1643$^7$</td>
<td>1250±25</td>
<td>1,000±400</td>
<td>10.7±0.05</td>
<td>1.0±4.7</td>
<td>1±0.3</td>
<td>~1_{±1}^{±2}</td>
<td>11.6±6.8</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>1646-47</td>
<td>2000±25</td>
<td>575±25</td>
<td>7.2±0.05</td>
<td>8.6±0.05</td>
<td>10.02_{1}^{±1.0}</td>
<td>170±30</td>
<td>58_{1}^{±1}</td>
<td>33.9±6.0</td>
</tr>
<tr>
<td>9</td>
<td>1651-53</td>
<td>2100±25</td>
<td>650±25</td>
<td>2.8±0.05</td>
<td>11.4±0.05</td>
<td>24.97_{3}^{±3.75}</td>
<td>440±80</td>
<td>1095_{1}^{±182}</td>
<td>4.7±1.2</td>
</tr>
<tr>
<td>10</td>
<td></td>
<td>825±25</td>
<td>0</td>
<td>15.0±0.05</td>
<td>8.9±0.05</td>
<td>39.74_{2}^{±4.0}</td>
<td>850±130</td>
<td>126_{1}^{±18}</td>
<td>78.0±12.0</td>
</tr>
<tr>
<td>11</td>
<td></td>
<td>275±25</td>
<td>255±25</td>
<td>10.0±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>12</td>
<td></td>
<td>1400±100</td>
<td>220±25</td>
<td>6.25±0.5</td>
<td>8.1±0.4</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 3.1 Eruption data for effusive activity which has occurred on Etna between 1600 and 1689.
<table>
<thead>
<tr>
<th>Id. No.</th>
<th>Year</th>
<th>Vent Alti. (m asl)</th>
<th>Front Alti. (m asl)</th>
<th>$L_1$ (km)</th>
<th>$L$ (km)</th>
<th>$A$ (km$^2$)</th>
<th>$V$ ($\times 10^4$ m$^3$)</th>
<th>$T$ (Days)</th>
<th>$Q_0$ (m$^3$/s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>14</td>
<td>1763(W)</td>
<td>1725±25</td>
<td>1600±25</td>
<td>5.15±0.05</td>
<td>0.85±0.05</td>
<td>0.21</td>
<td>20±4</td>
<td>32±2</td>
<td>7.2±2.6</td>
</tr>
<tr>
<td>15</td>
<td></td>
<td>1610±25</td>
<td>1575±25</td>
<td>6.35±0.05</td>
<td>3.70±0.05</td>
<td>2.13±0.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>16</td>
<td></td>
<td>1650±25</td>
<td>1763(S)</td>
<td>2450±25</td>
<td>4.70±0.05</td>
<td>3.50±0.05</td>
<td>2.00±0.3</td>
<td>66±15</td>
<td>9.1±2.1</td>
</tr>
<tr>
<td>18</td>
<td></td>
<td>1764-65</td>
<td>2600±25</td>
<td>2.90±0.20</td>
<td>5.10±0.20</td>
<td>-</td>
<td>60±30</td>
<td>550±30</td>
<td>1.3±0.6</td>
</tr>
<tr>
<td>19</td>
<td></td>
<td>1766</td>
<td>1975±25</td>
<td>950±25</td>
<td>5.90±0.05</td>
<td>6.50±0.05</td>
<td>9.42±0.9</td>
<td>90±16</td>
<td>193±2</td>
</tr>
<tr>
<td>20</td>
<td></td>
<td>1780</td>
<td>2150±25</td>
<td>1750±25</td>
<td>6.45±0.05</td>
<td>1.60±0.05</td>
<td>0.38±0.04</td>
<td>0.5±0.2</td>
<td>&lt;1</td>
</tr>
<tr>
<td>21</td>
<td></td>
<td>1792-93</td>
<td>1900±25</td>
<td>1000±25</td>
<td>4.05±0.05</td>
<td>5.95±0.05</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>22</td>
<td></td>
<td>1809</td>
<td>3000±25</td>
<td>1700±25</td>
<td>1.10±0.05</td>
<td>7.50±0.05</td>
<td>3.83</td>
<td>5.67±1.5</td>
<td>28±4</td>
</tr>
<tr>
<td>23</td>
<td></td>
<td>1811**</td>
<td>1950±25</td>
<td>1100±25</td>
<td>4.00±0.05</td>
<td>5.50±0.05</td>
<td>4.74±0.70</td>
<td>43±11</td>
<td>180±2</td>
</tr>
<tr>
<td>27</td>
<td></td>
<td>1819**</td>
<td>2400±25</td>
<td>975±25</td>
<td>3.20±0.05</td>
<td>6.50±0.05</td>
<td>4.91±0.74</td>
<td>42±6</td>
<td>66±2</td>
</tr>
<tr>
<td>28</td>
<td></td>
<td>1832</td>
<td>1700±25</td>
<td>1275±25</td>
<td>4.45±0.05</td>
<td>5.00±0.05</td>
<td>1.35</td>
<td>6</td>
<td>43±5</td>
</tr>
<tr>
<td>29</td>
<td></td>
<td>1838**</td>
<td>3150±25</td>
<td>2000±25</td>
<td>0.45±0.05</td>
<td>3.10±0.05</td>
<td></td>
<td>9-18</td>
<td>210±30</td>
</tr>
<tr>
<td>30</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>31</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Id. No.</td>
<td>Year</td>
<td>Vent Alti. (m asl)</td>
<td>Front Alti. (m asl)</td>
<td>$L_1$ (km)</td>
<td>$L$ (km)</td>
<td>$A$ (km$^2$)</td>
<td>$V$ ($\times 10^4$ m$^3$)</td>
<td>$T$ (Days)</td>
<td>$Q_0$ (m$^3$s$^{-1}$)</td>
</tr>
<tr>
<td>-------</td>
<td>------</td>
<td>-------------------</td>
<td>-------------------</td>
<td>----------</td>
<td>--------</td>
<td>---------</td>
<td>-----------------</td>
<td>--------</td>
<td>-----------------</td>
</tr>
<tr>
<td>32</td>
<td>1843</td>
<td>1950±25</td>
<td>525±25</td>
<td>2.90±0.05</td>
<td>13.90±0.05</td>
<td>6.38±0.64</td>
<td>54±7</td>
<td>9±1</td>
<td>69.4±12</td>
</tr>
<tr>
<td>33</td>
<td>1852-53</td>
<td>1725±25</td>
<td>525±25</td>
<td>4.30±0.05</td>
<td>8.50±0.05</td>
<td>10.00±1.5</td>
<td>125±21</td>
<td>280±2</td>
<td>5.2±0.9</td>
</tr>
<tr>
<td>34</td>
<td>1865</td>
<td>1700±25</td>
<td>750±25</td>
<td>6.40±0.05</td>
<td>8.20±0.05</td>
<td>7.22±0.72</td>
<td>96±9</td>
<td>151±2</td>
<td>7.4±0.7</td>
</tr>
<tr>
<td>35</td>
<td>1874</td>
<td>2150±25</td>
<td>1900±25</td>
<td>6.10±0.05</td>
<td>1.10±0.05</td>
<td>-</td>
<td>1±0.2</td>
<td>7h±1h</td>
<td>39.7±9.8</td>
</tr>
<tr>
<td>36</td>
<td>1879S</td>
<td>2425±25</td>
<td>1950±25</td>
<td>2.75±0.05</td>
<td>1.75±0.05</td>
<td>0.06±0.006</td>
<td>0.5±1h</td>
<td>12h±3h</td>
<td>11.6±5.5</td>
</tr>
<tr>
<td>37</td>
<td>1879NE</td>
<td>1850±25</td>
<td>550±25</td>
<td>7.60±0.05</td>
<td>9.35±0.05</td>
<td>2.41±0.24</td>
<td>30±4</td>
<td>10±4</td>
<td>34.7±4.6</td>
</tr>
<tr>
<td>38</td>
<td>1883</td>
<td>1050±25</td>
<td>-</td>
<td>11.50±0.05</td>
<td>0.25±0.05</td>
<td>-</td>
<td>0.05±4</td>
<td>2±4</td>
<td>-</td>
</tr>
<tr>
<td>39</td>
<td>1886</td>
<td>1400±25</td>
<td>750±25</td>
<td>8.50±0.05</td>
<td>6.50±0.05</td>
<td>6.0±0.9</td>
<td>57±7</td>
<td>19±1</td>
<td>34.7±4.6</td>
</tr>
<tr>
<td>40</td>
<td>1892</td>
<td>1800±25</td>
<td>950±25</td>
<td>6.50±0.05</td>
<td>6.90±0.05</td>
<td>10.10±1.01</td>
<td>100±16</td>
<td>172±2</td>
<td>6.7±1.1</td>
</tr>
<tr>
<td>41</td>
<td>1908</td>
<td>2200±25</td>
<td>1575±25</td>
<td>3.30±0.05</td>
<td>2.70±0.05</td>
<td>0.52±0.05</td>
<td>1.2±0.3</td>
<td>18h±4h</td>
<td>18.5±9</td>
</tr>
<tr>
<td>42</td>
<td>1910</td>
<td>2300±25</td>
<td>1950±25</td>
<td>4.25±0.05</td>
<td>1.30±0.05</td>
<td>0.07±2</td>
<td>0.3±3</td>
<td>12h</td>
<td>6.9</td>
</tr>
<tr>
<td>43</td>
<td>1911</td>
<td>2000±25</td>
<td>700±25</td>
<td>5.90±0.05</td>
<td>10.50±0.05</td>
<td>3.91±0.4</td>
<td>40±6</td>
<td>26±1</td>
<td>17.8±2.7</td>
</tr>
<tr>
<td>44</td>
<td>1911</td>
<td>1625±25</td>
<td>550±25</td>
<td>8.90±0.05</td>
<td>7.50±0.05</td>
<td>5.05±0.5</td>
<td>53±7</td>
<td>11±4</td>
<td>55.8±7.8</td>
</tr>
<tr>
<td>45</td>
<td>1923</td>
<td>2000±25</td>
<td>1375±25</td>
<td>6.90±0.05</td>
<td>3.00±0.05</td>
<td>0.87±3</td>
<td>6.01±0.64</td>
<td>62±9</td>
<td>20.0±22.4</td>
</tr>
<tr>
<td>46</td>
<td>1923</td>
<td>1800±25</td>
<td>625±25</td>
<td>7.90±0.05</td>
<td>7.90±0.05</td>
<td>5.14±2</td>
<td>3.84±0.6</td>
<td>59</td>
<td>20.9±3.3</td>
</tr>
<tr>
<td>47</td>
<td>1928</td>
<td>1700±25</td>
<td>1000±25</td>
<td>5.95±0.05</td>
<td>3.45±0.05</td>
<td>1.11±2</td>
<td>4.38±1.08</td>
<td>40±7</td>
<td>23.1±25.7</td>
</tr>
<tr>
<td>48</td>
<td>1928</td>
<td>1200±25</td>
<td>25±25</td>
<td>10.15±0.05</td>
<td>9.00±0.05</td>
<td>3.27±2</td>
<td>40±7</td>
<td>40±7</td>
<td>26.2±4.7</td>
</tr>
<tr>
<td>49</td>
<td>1942</td>
<td>2600±25</td>
<td>1725±25</td>
<td>2.35±0.05</td>
<td>2.60±0.05</td>
<td>0.80±0.04</td>
<td>2±4</td>
<td>13h±4h</td>
<td>42.7±16.9</td>
</tr>
<tr>
<td>Id. No.</td>
<td>Year</td>
<td>Vent Alti. (m asl)</td>
<td>Front Alti. (m asl)</td>
<td>( L_h ) (km)</td>
<td>( L ) (km)</td>
<td>( A ) (km(^2))</td>
<td>( V ) (x10(^4) m(^3))</td>
<td>( T ) (Days)</td>
<td>( Q_x ) (m(^3)/s)</td>
</tr>
<tr>
<td>--------</td>
<td>------</td>
<td>------------------</td>
<td>------------------</td>
<td>-------------</td>
<td>-------------</td>
<td>-------------</td>
<td>----------------</td>
<td>-------------</td>
<td>-------------</td>
</tr>
<tr>
<td>50</td>
<td>1947</td>
<td>2275±25</td>
<td>1825±25</td>
<td>5.20±0.05</td>
<td>2.20±0.05</td>
<td>0.30(^2)</td>
<td>1.64±0.2</td>
<td>0.6</td>
<td>10±2</td>
</tr>
<tr>
<td>51</td>
<td></td>
<td>2225±25</td>
<td>900±25</td>
<td>6.15±0.05</td>
<td>6.40±0.05</td>
<td>1.34(^2)</td>
<td>10</td>
<td>7</td>
<td></td>
</tr>
<tr>
<td>52</td>
<td>1949</td>
<td>3000±25</td>
<td>2275±25</td>
<td>0.60±0.05</td>
<td>3.10±0.05</td>
<td>0.41(^2)</td>
<td>1.72±0.2</td>
<td>1.5</td>
<td>8±1</td>
</tr>
<tr>
<td>53</td>
<td></td>
<td>2150±25</td>
<td>1450±25</td>
<td>3.60±0.05</td>
<td>4.45±0.05</td>
<td>1.31(^2)</td>
<td>6.5</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>54</td>
<td>1950-51</td>
<td>2250±25</td>
<td>775±25</td>
<td>2.90±0.05</td>
<td>8.50±0.05</td>
<td>8.17±0.8</td>
<td>121±14</td>
<td>372±1</td>
<td>3.8±0.3</td>
</tr>
<tr>
<td>55</td>
<td></td>
<td>1964</td>
<td>3200±25</td>
<td>0.05±0.05</td>
<td>3.10±0.05</td>
<td>0.71</td>
<td>0.85±0.05</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>56</td>
<td></td>
<td>3200±25</td>
<td>1925±25</td>
<td>0.05±0.05</td>
<td>3.10±0.05</td>
<td>0.71</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>57</td>
<td></td>
<td>3000±25</td>
<td>1925±25</td>
<td>0.05±0.05</td>
<td>3.10±0.05</td>
<td>0.71</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>58</td>
<td></td>
<td>1964</td>
<td>3200±25</td>
<td>0.05±0.05</td>
<td>3.10±0.05</td>
<td>0.71</td>
<td>0.85±0.05</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>59</td>
<td></td>
<td>3000±25</td>
<td>1675±25</td>
<td>0.80±0.05</td>
<td>3.60±0.05</td>
<td>0.67</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>60</td>
<td></td>
<td>2950±25</td>
<td>1675±25</td>
<td>0.80±0.05</td>
<td>3.60±0.05</td>
<td>0.67</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>61</td>
<td></td>
<td>3000±25</td>
<td>2350±25</td>
<td>1.25±0.05</td>
<td>3.30±0.05</td>
<td>0.67</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>62</td>
<td></td>
<td>3000±25</td>
<td>2350±25</td>
<td>1.25±0.05</td>
<td>3.30±0.05</td>
<td>0.67</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>63</td>
<td>1971</td>
<td>2675±25</td>
<td>1950±25</td>
<td>1.85±0.05</td>
<td>1.80±0.05</td>
<td>0.28(^2)</td>
<td>6.19±0.58</td>
<td>0.5</td>
<td>42±9</td>
</tr>
<tr>
<td>64</td>
<td></td>
<td>2575±25</td>
<td>1950±25</td>
<td>1.85±0.05</td>
<td>1.80±0.05</td>
<td>0.28(^2)</td>
<td>-</td>
<td>9.5</td>
<td>-</td>
</tr>
<tr>
<td>65</td>
<td></td>
<td>2550±25</td>
<td>1900±25</td>
<td>2.70±0.05</td>
<td>1.80±0.05</td>
<td>0.28(^2)</td>
<td>-</td>
<td>9.5</td>
<td>-</td>
</tr>
<tr>
<td>66</td>
<td></td>
<td>2300±25</td>
<td>2000±25</td>
<td>3.55±0.05</td>
<td>2.00±0.05</td>
<td>0.28(^2)</td>
<td>-</td>
<td>10</td>
<td>-</td>
</tr>
<tr>
<td>67</td>
<td></td>
<td>1800±25</td>
<td>600±25</td>
<td>5.75±0.05</td>
<td>7.30±0.05</td>
<td>2.40(^2)</td>
<td>-</td>
<td>32</td>
<td>-</td>
</tr>
<tr>
<td>Id. No.</td>
<td>Year</td>
<td>Vent Alti. (m asl)</td>
<td>Front Alti. (m asl)</td>
<td>$L_t$ (km)</td>
<td>$L$ (km)</td>
<td>$A$ (km$^2$)</td>
<td>$V$ (x10$^4$ m$^3$)</td>
<td>$T$ (Days)</td>
<td>$Q_*$ (m$^3$s$^{-1}$)</td>
</tr>
<tr>
<td>---------</td>
<td>---------</td>
<td>--------------------</td>
<td>--------------------</td>
<td>------------</td>
<td>----------</td>
<td>-------------</td>
<td>---------------------</td>
<td>------------</td>
<td>---------------------</td>
</tr>
<tr>
<td>68</td>
<td>1974(A)</td>
<td>1675±25</td>
<td>1450±25</td>
<td>1.75±0.05</td>
<td>0.15</td>
<td>0.75</td>
<td>5</td>
<td></td>
<td>2.3</td>
</tr>
<tr>
<td>69</td>
<td>1974(B)</td>
<td>1675±25</td>
<td>1475±25</td>
<td>5.85±0.05</td>
<td>1.00±0.05</td>
<td>0.30±0.03</td>
<td>2.4±0.5</td>
<td>4</td>
<td>17±1.4</td>
</tr>
<tr>
<td>60</td>
<td>1976/77</td>
<td>1400±25</td>
<td>0.55±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>71</td>
<td>1975(A)</td>
<td>1675±25</td>
<td>1400±25</td>
<td>6.25±0.05</td>
<td>1.20±0.05</td>
<td>0.20±0.02</td>
<td>2.1±0.5</td>
<td>18±1.4</td>
<td>1.4±0.3</td>
</tr>
<tr>
<td>72</td>
<td>1975(B)</td>
<td>2625±25</td>
<td>-</td>
<td>2.95±0.05</td>
<td>-</td>
<td>-</td>
<td>13±4</td>
<td>200±2</td>
<td>0.8±0.2</td>
</tr>
<tr>
<td>73</td>
<td>1976/77</td>
<td>2975±25</td>
<td>-</td>
<td>1.65±0.05</td>
<td>-</td>
<td>-</td>
<td>12±4</td>
<td>191±2</td>
<td>0.7±0.2</td>
</tr>
<tr>
<td>74</td>
<td>1978(A)</td>
<td>2900±25</td>
<td>1625±25</td>
<td>1.10±0.05</td>
<td>3.70±0.05</td>
<td>1.56</td>
<td>23</td>
<td>35</td>
<td>7.6</td>
</tr>
<tr>
<td>75</td>
<td>1978(B)</td>
<td>2750±25</td>
<td>2150±25</td>
<td>1.90±0.05</td>
<td>1.10±0.05</td>
<td>0.69±0.16</td>
<td>27±1</td>
<td>37±1</td>
<td>-</td>
</tr>
<tr>
<td>76</td>
<td>1978(C)</td>
<td>2600±25</td>
<td>2400±25</td>
<td>2.45±0.05</td>
<td>0.35±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>77</td>
<td>1978(A)</td>
<td>3000±25</td>
<td>2050±25</td>
<td>0.65±0.05</td>
<td>1.35±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>78</td>
<td>1978(B)</td>
<td>2725±25</td>
<td>1650±25</td>
<td>1.55±0.05</td>
<td>3.30±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>3±1</td>
</tr>
<tr>
<td>79</td>
<td>1978(B)</td>
<td>2450±25</td>
<td>-</td>
<td>2.30±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>3±1</td>
</tr>
<tr>
<td>80</td>
<td>1978(B)</td>
<td>2450±25</td>
<td>-</td>
<td>3.30±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>4±1</td>
</tr>
<tr>
<td>81</td>
<td>1978(C)</td>
<td>2600±25</td>
<td>1500±25</td>
<td>2.30±0.05</td>
<td>3.50±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>4</td>
</tr>
<tr>
<td>82</td>
<td>1978(C)</td>
<td>1700±25</td>
<td>-</td>
<td>3.85±0.05</td>
<td>4.00±0.05</td>
<td>-</td>
<td>-</td>
<td>3.5</td>
<td>5±1</td>
</tr>
<tr>
<td>83</td>
<td>1978(C)</td>
<td>1650±25</td>
<td>-</td>
<td>4.45±0.05</td>
<td>-</td>
<td>-</td>
<td>2.5</td>
<td></td>
<td>11.6±5.2</td>
</tr>
<tr>
<td>Id. No.</td>
<td>Year</td>
<td>Vent Alti. (m asl)</td>
<td>Front Alti. (m asl)</td>
<td>$L_i$ (km)</td>
<td>$L$ (km)</td>
<td>$A$ (km$^2$)</td>
<td>$V$ ($\times 10^4$ m$^3$)</td>
<td>$T$ (Days)</td>
<td>$Q_*$ (m$^3$s$^{-1}$)</td>
</tr>
<tr>
<td>--------</td>
<td>------</td>
<td>-------------------</td>
<td>--------------------</td>
<td>----------</td>
<td>--------</td>
<td>-----------</td>
<td>----------------------</td>
<td>---------</td>
<td>---------------------</td>
</tr>
<tr>
<td>86</td>
<td>1979</td>
<td>2800±25</td>
<td>1800±25</td>
<td>1.25±0.05</td>
<td>3.00±0.05</td>
<td>-</td>
<td>0.9</td>
<td>1</td>
<td>10.4</td>
</tr>
<tr>
<td>87</td>
<td>1979</td>
<td>1700±25</td>
<td>875±25</td>
<td>5.35±0.05</td>
<td>6.00±0.05</td>
<td>1.41</td>
<td>6.6</td>
<td>4</td>
<td>19.1</td>
</tr>
<tr>
<td>88</td>
<td>1979</td>
<td>2500±25</td>
<td>1850±25</td>
<td>3.30±0.05</td>
<td>2.50±0.05</td>
<td>-</td>
<td>-</td>
<td>1</td>
<td>-</td>
</tr>
<tr>
<td>89</td>
<td>1979</td>
<td>2150±25</td>
<td>1725±25</td>
<td>5.20±0.05</td>
<td>1.25±0.05</td>
<td>0.03</td>
<td>0.05</td>
<td>2</td>
<td>0.3</td>
</tr>
<tr>
<td>90</td>
<td>1981</td>
<td>1400±25</td>
<td>600±25</td>
<td>8.90±0.05</td>
<td>8.25±0.05</td>
<td>4.00$^3$</td>
<td>18.5</td>
<td>20±2</td>
<td>40h±4h</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>4.59±0.46</td>
<td></td>
<td></td>
<td>6$^*$_12h</td>
</tr>
<tr>
<td>91</td>
<td></td>
<td>1150±25</td>
<td>900±25</td>
<td>10.60±0.05</td>
<td>1.46±0.05</td>
<td>0.59</td>
<td>1.5</td>
<td>6</td>
<td>126.6±13.8</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>38.6±5.0</td>
</tr>
<tr>
<td>92</td>
<td>1983</td>
<td>2275±25</td>
<td>1075±25</td>
<td>4.60±0.05</td>
<td>7.50±0.05</td>
<td>5.18±0.52</td>
<td>91±14</td>
<td>131$_{\pm 14}$</td>
<td>8.0±1.2</td>
</tr>
<tr>
<td>93</td>
<td>1984</td>
<td>3000±25</td>
<td>1950±25</td>
<td>0.75±0.05</td>
<td>3.10±0.05</td>
<td>1.44±0.14</td>
<td>10±3$^*$</td>
<td>172±1</td>
<td>0.7±0.2</td>
</tr>
<tr>
<td>94</td>
<td>1985(A)</td>
<td>2500±25</td>
<td>1825±25</td>
<td>3.90±0.05</td>
<td>3.05±0.05</td>
<td>4.30±0.43</td>
<td>30±10$^*$</td>
<td>123±1</td>
<td>2.8±0.9</td>
</tr>
<tr>
<td>95</td>
<td>1985(B)</td>
<td>2750±25</td>
<td>1700±25</td>
<td>1.90±0.05</td>
<td>3.50±0.05</td>
<td>0.70±0.07</td>
<td>0.7$^*$</td>
<td>6$^*$_12</td>
<td>1.9±0.6</td>
</tr>
<tr>
<td>96</td>
<td>1986-87</td>
<td>3050±25</td>
<td>2350±25</td>
<td>0.70±0.05</td>
<td>1.85±0.05</td>
<td>0.14</td>
<td>0.7</td>
<td>1</td>
<td>120±1</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>5.10±0.51</td>
<td>93</td>
<td>4.9±1.0</td>
</tr>
<tr>
<td>97</td>
<td>1986-87</td>
<td>2500±25</td>
<td>1350±25</td>
<td>2.25±0.05</td>
<td>5.10±0.05</td>
<td>-</td>
<td>-</td>
<td>89</td>
<td>-</td>
</tr>
<tr>
<td>98</td>
<td>1986-87</td>
<td>2300±25</td>
<td>1325±25</td>
<td>3.10±0.05</td>
<td>4.45±0.05</td>
<td>-</td>
<td>-</td>
<td>26</td>
<td>-</td>
</tr>
<tr>
<td>99</td>
<td>1986-87</td>
<td>2200±25</td>
<td>1325±25</td>
<td>3.30±0.05</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>100</td>
<td>1989</td>
<td>2600±25</td>
<td>1100±25</td>
<td>2.25±0.05</td>
<td>6.50±0.05</td>
<td>-</td>
<td>15±5$^*$</td>
<td>11$^*_4$</td>
<td>16±5.4</td>
</tr>
</tbody>
</table>
3.4 Lava Flow Morphologies of the Post-1750 Period

Virtually all the lavas erupted in the post-1750 period have broadly the same petrographic character and chemical composition, and yet the resultant flow fields have a variety of final planimetric morphologies. The majority of flow fields erupted in this period have aa surface textures, though patches of pahoehoe are found on many, especially in the proximity of the vent areas. Only a few flow fields occur where pahoehoe surface textures predominate. The range of observed morphologies are listed below.

(i) Essentially simple flows (as defined by Walker, 1971) that are long by comparison with their width e.g. 1947 (51) lava flow on NE flank (Figure 3.3 (A); Plate 1).

(ii) Compound (as defined by Walker, 1971) flows that are narrow by comparison with their length, and have a well-defined central channel with numerous overflows e.g. 1780 (22) lava flow on SW flank.

(iii) a. Compound flow fields that are wide by comparison with their length. These flow fields are composed of several flow units, each having well-developed channels and overflows. Individual flows emanate from the vicinity of the vent area e.g. 1865 (34) flow field, NE flank (Figure 3.3 (B); Plate 2), or downslope of the main vent where the main flow has been checked by a break in slope.

(iii) b. Pahoehoe textured flow fields that are broad by comparison with their length, and are compound in character. Tumuli occur over the flow field where there has been a localised concentration of effusion from ephemeral boccas e.g. 1792-93 (24) lava flow, SE flank.

(iv) Short thick flows with well-developed ogives eg. 1974 (68) (Figure 3.3 (C)).

(v) A short flow field consisting of a narrow ridge, crested by a channel/tube system.

The extreme thickness of this flow type is interpreted to result from the repeated extrusions of short flow units which piled up around the vent area (Chester et al. 1985) e.g. 1763 (18) flow field, S flank (Figure 3.3 (D)).

Of the 47 discernible flow fields erupted between 1750 and 1990, 41 are of the aa-textured, type (i), (ii) or (iii) variety. Extensive pahoehoe flows of type (iiib) were only produced during the flank effusions of 1764-65 (19), 1792-93 (24) and 1975 (72), whilst only the 1763 (18) eruption on the southern flank produced the type (v) variety.
Examples of type (iv) morphologies are only found within the 1763 (18) and 1974 (68) flow fields.

For the purposes of this study, the flows are grouped into two types: Type A, that are narrow compared with their length (to include types (i) and (ii)) and Type B, that are wide by comparison with their length (to include types (iii), (iv) and (v)).

Figure 3.3 Example sketches of four of the morphological flow types observed in post-1750 flank eruptions; (A) long, narrow flow of 1947; (B) complex, broad flow field of 1865; (C) one of the thick 1974 flows with surface ogives; and (D) the short thick flow of 1763 with a channel-tube system along the flow crest. The arrow indicates the direction away from the vent (from Hughes et al. 1990).

These are broadly comparable to the two Etnean flow field populations which Kilburn and Lopes (1988) isolated during a statistical analysis of inter-relationships between eruption parameters and flow dimensions.
Plate 1 Aerial photograph of the 1947, Type A eruption on the northern flank and the compound pahoehoe tumuli flow field of 1614-24.

Plate 2 Aerial photograph of the 1865, Type B flow field on the eastern flank. The scale is the same as for Plate 1.
In the following sections, those factors which influence the morphological evolution of lava flow fields, as identified by Walker (1973), Malin (1980) and Lopes and Guest (1982), will be re-examined in the context of the present study. They are: lava rheology, erupted volume, effusion rate, eruption duration and the effects of topography.

3.4.1 Rheology

The rheological characteristics of magmas are influenced by many factors, amongst which are temperature, crystallinity, vesicularity and liquid polymerisation. For a magma at a temperature above its liquidus (temperature at which crystals first appear in the melt), its rheological properties may approach that of a newtonian fluid (Shaw, 1969), so that when it is acted upon by an applied stress, it deforms instantaneously. At temperatures below the liquidus, the number of phases within a magma increases with the appearance of crystals and vesicles, and the rheological properties of the magma change towards those of a non-newtonian fluid (Shaw, 1969). Movement is initially impaired, for in addition to its viscosity (resistance to flow), a finite strength or yield strength must be overcome before it will flow.

The majority of Etna lavas are erupted at temperatures between 1,080 and 1,125°C (Archambault and Tanguy, 1976), and have undergone significant pre-eruption crystallisation (30-40%, Guest and Duncan, 1981). It is therefore most probable that Etna lavas probably have non-newtonian rheologies upon eruption (Kilburn, 1984). This conclusion has been reinforced by laboratory (Gauthier, 1973) and field based rheological experiments (Pinkerton and Sparks, 1978) on Etna magmas at temperatures between 1,080 and 1,200°C. Interpretation of the results in terms of a non-newtonian, Bingham fluid, produced yield strengths ranging between 4.5 and 370 Pa, and Bingham viscosity values between 90 and 9400 Pa s. Highest viscosities and yield strengths were associated with those magmas at the lower end of the temperature interval.

Once on the surface, sub-aerial cooling (radiative and conductive) and volatile loss further reduce the temperature of the lava, promote crystal growth, and increase the yield strength and viscosity of the lava as it moves away from the eruptive vent. Over an uniform ground slope, this progressive change in rheology manifests itself to
an observer as a gradual increase in flow thickness between the eruptive vent and flow terminus (Pieri and Baloga, 1986; Guest et al. 1987; Dragoni, 1989). In order to sustain movement, the basal shear stresses at the flow front (as controlled by the rate of flow thickening) must exceed the yield strength.

The only way in which lava rheology can be directly responsible for the diversity of flow field morphologies is if the rheology of the lavas being erupted at the vent showed substantial variation during, and between individual eruptive episodes. Whilst some subtle variations in the rheological properties of Etnean magmas may occur between eruptive episodes (due to slight variations in the temperature, chemical composition and volumetric phenocryst content of the magmas), dramatic rheological changes during individual eruptive episodes can be induced by volatile loss. The effect of volatile loss is to undercool a body of magma relative to its liquidus. This leads to a sudden enhancement of the nucleation rate which triggers an all-pervasive growth of quench crystals in the magma. This process leads to a rapid increase in viscosity and the development of a high yield strength in the magma (Sparks and Pinkerton, 1978; Lipman et al. 1985). Sparks and Pinkerton (1978) highlighted three processes by which degassing leads to the undercooling of a lava. Firstly, decompression of the gas-phase cools the magma system down adiabatically. Secondly, the solution of gases has the effect of increasing the liquidus of the melt towards that of an anhydrous magma. The amount of undercooling is therefore exaggerated. Thirdly, fire-fountaining provides some opportunity for cooling the magma by radiative heat loss with the atmosphere.

Whilst explosive activity has been a major component of all post-1750 eruptive activity, degassing does not appear, on the whole, to have effected major rheological variations during, and between individual eruptive episodes. The fact that Etnean lavas, once on the surface, retain a minimum thickness on a given incline led Robson (1967) to infer that either upon eruption or soon after, Etnean lavas behave as non-newtonian liquids and acquire a yield strength of the order of $10^3$ Pa (Kilburn 1984). This implies that most Etnean magmas are brought to the surface within a restricted range of rheological states (Kilburn, 1984; Chester et al. 1985), a point which is reinforced by the broadly uniform chemical composition and phenocryst volumetric content of the lavas upon eruption.
There are only a few examples within the post-1750 output period, where strong degassing has led to clear rheological variations in the lavas erupted within a single eruptive episode. In these instances, it is possible to identify individual flow units within a flow field whose morphology and surface features are more typical of viscous lavas. These flow units, which are generally short, thick and have ogives on the flow surface, can be found within flow fields extruded during some of the most explosive eruptions of this period. Examples occur within the 1763 (15) (West flank, upper eruption site), the 1763 (18) (South flank) and the 1974 (69) (West flank, first phase) flow fields (Figure 3.3 (C)).

During the first phase of the 1974 eruption, the extrusion of the viscous flow unit (unit 5 of Guest et al., 1974) followed the most explosive period of the eruption. Subsequent flow units, erupted during periods of reduced explosive activity, have comparable morphological characteristics to the majority of Etnean flow units. The magma supplying the 1974 eruptions (as discussed in chapter 2), is considered to have risen rapidly along an ascent path which was largely independent of the central conduit system (Tanguy and Kieffer, 1976). The high explosivity of this particular eruption was probably a consequence of the fast ascent. This resulted in the rapid decompression of the magma which in turn effected the quick release of a large quantity of volatiles from the magma.

Surface Texture

If, as inferred above, the rheology of post-1750 Etnean lavas have been broadly uniform, then some other explanation is necessary to account for the existence of both aa and pahoehoe-textured flow fields in this output period.

The formation of aa and pahoehoe texture is determined by a complex interrelationship between lava rheology and applied shear stress (Williams and McBirney, 1979; Peterson and Tilling, 1980 and Kilburn, 1981; Rowland and Walker, 1987). This relationship is illustrated graphically in Figure 3.4. Though pahoehoe and aa flow regimes occupy different parts of the diagram, under certain critical rheological and/or stress conditions, a flow of pahoehoe will transform to aa. Field observations have determined that the critical relation between viscosity and shear rate is inverse so that at low yield strengths and viscosities, a high rate of shear is required to begin
the transition to aa. Conversely, if the yield strengths and viscosities are high, a much lower rate of shear will induce the transition. These critical conditions are delineated by the transition threshold zone (TTZ) in Figure 3.4.

A large number of Hawaiian lava flows are initially erupted as pahoehoe, but as they move away from the vent, they change to aa. At the vent, the lavas are characterised by high temperatures (1,140°C) (that approach the liquidus temperature (1,200°C) (Neal et al. 1988)) and fluid rheologies (which approach the newtonian state (Moore, 1987; Wolfe et al. 1988)), but as they move away, the rheology of the lavas tends towards that of non-newtonian fluids as they cool and crystallise. The existence of pahoehoe textures in the vicinity of the vent area, even at high effusion rates reflects, in this instance, the predominance of lava rheology as the factor which controls the transition process. Since the rate of flow movement generally decreases with increasing distance from the vent, the only way in which the critical threshold
can be exceeded and the transition from pahoehoe to aa effected, is if the yield strength and viscosity of the lava is increasing at a faster rate than the rate at which the flow is slowing down. In the case of long duration, tube-fed eruptions (Greeley, 1987), pahoehoe textures can exist from vent to flow front because the low effusion rates involved never exceed the transition threshold. In addition, lava tubes are very efficient insulators and can transport lava virtually isothermally for long distances. In this way, the rheology of the lava being added to the flow front may closely resemble that of the fluid lava being issued at the eruptive vent. Flow fields characterised entirely by aa surface textures on Hawaii are generally restricted to explosive eruptions (Lipman et al. 1985, Rowland and Walker, 1990). For these eruptions, the rheology of the pre-eruptive magma has been modified through volatile loss. Whilst effusion rates at the vent may be comparable to those which pertained during the formation of pahoehoe channel-fed flows, the modified rheology of these lavas was sufficient to induce an earlier transition through the critical threshold.

Etnean lavas may, as mentioned previously, have higher yield strengths upon eruption than Hawaiian lavas. It is not surprising therefore, that aa textures predominate at high effusion rates. Pahoehoe textures only become established on these aa flow fields during the later stages of the eruption, once the effusion rate has decreased below the critical deformation threshold. During the 1975 NE rift eruption, Pinkerton and Sparks (1976) determined this threshold to be equivalent to an effusion rate of $2 \times 10^3$ m$^3$s$^{-1}$; pahoehoe textures only forming below this value. The 1792-93 (24) and 1764-65 (19) pahoehoe-textured flow fields are characterised by toey lobes, ropey structures and tumuli (Guest, 1982). Whilst the 1975 (72) flow field was not supplied for a period long enough for tumuli to develop, it possesses all the characteristics of the other post-1750 pahoehoe textured flow fields. No pahoehoe-textured, channel-fed flows are discernible on these flow fields; most of the pahoehoe structures are small scale and are associated with small lava flows which issued from ephemeral boccas. It is possible that the pahoehoe flow fields were supplied by more fluid magmas. However, the scaling of individual flow units and surface features show broad similarities with some of the 1975 flow units, and are therefore more likely to be the products of sustained low effusion rates. For the lava flow fields of
Mount Etna, lava rheology appears to be a secondary factor to flow kinetics in controlling the pahoehoe-aa transition.

In general, whilst many features of lava flow fields, from the smallest structures (Peterson and Tilling, 1980; Kilburn, 1981; Rowlands and Walker, 1987, 1990) to the overall gross appearance (Hulme, 1974; Pinkerton and Sparks, 1976; Chester et al. 1985), have been explained in terms of variations in the non-newtonian rheology of the flowing lava, the principal effect of rheology in the emplacement of a lava flow is merely to control the thickness and width of a flow. This affects the length in that more of the erupted volume goes into the thickening (Walker 1973) and widening of the flow.

3.4.2 Volume

A relationship between flow volume and the morphometric development of lava flow fields should be expected if only for the fact that a lava flow cannot lengthen without the addition of extra volume. For Hawaiian lava flows, Malin (1980) found a correlation between the logarithm of flow unit volume and the logarithm of the length, and established the relationship

\[ L = c(V)^m \]  

where \( L \) is the length (km) of a lava flow unit, \( V \) is the volume (x10⁶ m³) of that unit and \( m \) and \( c \) are constants. The value of the exponent \( m \), provides information about how the volume is distributed along the flow. Malin (1980) gave three examples.

1. For flows where the depth, width and length vary equally and proportionally with volume, the value of \( m \) approaches 0.33.
2. For flows where the cross-sectional area varies proportionally with volume (and length), \( m \) approaches 0.5.
3. For flows whose cross-sectional area is fixed (channel or tube situation), \( m \) approaches 1.
For Hawaiian lavas, the exponent obtained by Malin (1980) approached 0.55 ± 0.03. This implied that the second case applied to Hawaiian lavas and that the erupted volume contributed as much to the cross-sectional area of the lavas as to their lengthening. An exponent of 0.37 ± 0.03 was obtained by Lopes (1985) from a general study of all Etnean lava flow fields suggesting that in the case of Etnean lavas, the depth and width of flows vary equally and proportionally to volume.

**Type A**

The relationship established for Etnean lava flows is re-examined in Figure 3.5, but this time the data have been separated into the equivalent of flow units (Type A) and flow fields (Type B). In order to maintain consistency with the Malin study, only Type A lava flow data (essentially flow units) are included in the calculation to establish equation 3.3. A fully weighted regression line fitted to the Type A data in Figure 3.5 returns an exponent of 0.43 ± 0.03; a value intermediate to cases 1 and 2 above. However, even with this analysis, the value of the exponent must be viewed with some caution because:

1. Many of the larger volume Type A flows such as 1843 (32) (Gemmellaro, 1843) and 1928 (48) (Imbò, 1928), achieved their maximum eruption length before the end of the eruption, and yet lava continued to be supplied from the vent throughout the eruption, albeit at a reduced rate. The final volume of these larger flows is not, therefore, the volume that contributed to lengthening alone, but to a lesser degree, widening and thickening as well. These Type A group of flows cannot therefore be strictly classified as flow units.

2. The lengthening phase of a few of the Type A lava flows was impeded by topographic obstacles. The 1886 (39) lava flow encountered several cinder cones during its emplacement and underwent considerable branching (Silvestri 1886) whilst the 1832 (30) lava flow on the western flank encountered a small, topographically high, inlier which forms part of the surrounding Peloritani mountains.

3. Some of the scatter observed in Figure 3.5 may be due to variable underlying ground slope. This can be separated into two components:

   a. Over short distances, a flow advancing down a gentle, but constant slope would be shorter and wider than one which had descended a constant, but steeper gradient,
Figure 3.5 Logarithmic diagram of length of flow (km) as a function of erupted volume (x $10^6$ m$^3$) for post-1750 Etna lava flows. For the Type A (flow unit) data, the relationship between volume and flow length (i) is of the form $L = 1.87 \pm 0.07 V^{0.43\pm0.09}$. The regression coefficient of this line is $r = 0.799$ for 27 data, and the probability of a chance relationship is < 0.5%. Numbers associated with the data points are their identification numbers, as presented in Table 3.2. Horizontal lines are error bars.
for more of the erupted volume would be distributed to the flow's cross-sectional area than to its length (Wadge, 1978). This effect would be most pronounced for flows occurring near the summit region where ground slopes are steeper than the average. However, further quantification of this effect is not possible with the existing data set due to the limited precision and accuracy of the volumetric data at low values. Better volumetric data are required in order to improve the calibration of this diagram.

b. Over longer distances (> 7 km), underlying ground slope does not remain constant but progressively decrease towards the foot of the volcano. For long lava flows which start from vents located at mid-elevations on the volcano, it is noticeable that when they encounter the shallow slopes around the foot of the volcano, their rate of lengthening decreases, with more of the available volume being taken up in widening around the flow front. This effect is most pronounced along the short northwest and northeast margins of the volcano where the slopes rapidly decrease in response to a shallowing basement. The 1911 (44) and 1923 (46) lava flows provide good examples (Platania, 1912 and Ponté, 1923).

Cumulatively, the points listed above affect the distribution of volume within a lava flow. In terms of equation 3.3, they would have the effect of artificially decreasing the value of the calculated exponent. Taking this into consideration, it may therefore be the case that Etnean lavas are similar to Hawaiian lavas in that the erupted volume contributes as much to the cross-sectional area of the lavas as to their lengthening.

**Type B**

Type B lava flows can be distinguished from Type A flows in Figure 3.5 by their greater volumes, though the final lengths attained by each flow type are similar. Historical records and recent observations of lava flow development (Guest et al., 1987) do not reveal any major differences in the way in which these lava flow fields initially evolve. For the majority, the first flow emplaced during the initial stages (first week), is normally the longest. However, in contrast to Type A eruptions, there is enough lava volume available to supply the Type B eruptions beyond this initial stage. Though some more limited lengthening may occur, the additional volume contributes to the widening and thickening of the flow field (Lopes and Guest, 1982;
Lopes, 1985; Kilburn and Lopes, 1989) through a process of new flow unit generation. Whilst these new flow units seldom exceed the length of the primary flow unit, it is conceivable that the distribution of volume within each new flow unit is described by equation 3.3.

Not all the Type B lava flow fields have evolved along this systematic pattern. The 1950-51 (54) (Cumin, 1954), 1983 (92) (Frazzetta and Romano, 1984) and 1986-87 (99) (Caltabiano et al. 1987) flow fields underwent extension after the emplacement of the primary flow unit. The lava flows of the above eruptions did not flow over an uniform ground slope but had to traverse a marked break of slope. The extensional stage of the first flow unit was checked by the break of slope and resulted in a period of widening. The longest flow unit emerged as a lateral breakout from the first unit at the break of slope. All subsequent morphological changes for these flow fields occurred at the break of slope and not at the vent. In these three instances, the differing growth pattern can largely be attributed to this topographical factor. The erupted volume therefore contributed to lengthening with some widening and thickening, followed by a repeat cycle. For the 1764-65 (19) and possibly the 1792-93 (24) pahoehoe flow fields, the stratigraphic relationship of individual tumuli within these flow fields suggests that they underwent progressive lengthening throughout the eruption, during which the systematic growth cycle was repeated several times.

The 1763 (18) Type B flow field on the southern flank (indicated on Figure 3.4) is often quoted as anomalous on volume - flow length diagrams because it is very short for its volume. Though the morphology of this flow field differs from that of the majority (section 3.4), there are indications that even this flow reached its maximum length during the early stages of the eruption (Sturiale, 1970). However, instead of contributing to the progressive lengthening of the flow field, the additional volume piled up about the vent area forming a thick lava mound. Chester et al. (1985) suggest that this eruption may have progressed as a series of intermittent short effusions, with each effusive phase giving rise to a short flow unit of limited volume.
3.4.3 Eruption Rate and Duration

The importance of effusion rate to the emplacement of lava flows was first realised by Walker (1973) from a logarithmic relationship he established between eruption rate and flow length. In his study, Walker found that the lengths of lava flows emplaced during eruptions which had lasted between 30 hours and 9 months, occupied a restricted range on an eruption rate-flow length diagram. This was irrespective of the composition of the magmas supplying them. His conclusions have been questioned in some subsequent morphometric studies which have restricted their data set to the lava flows of individual volcanoes (such as Hawaii, Malin, 1980). However, most of the discrepancies between the various studies arise because different data selection criteria were used. It is noticeable that when an uniform selection criterion is applied, such as that followed by Walker (1974), the end results are similar. This has been demonstrated for Etna by Lopes and Guest (1982) and Lopes (1985), and for Hawaii data by Pinkerton (1987) (using the data of Malin, 1980).

Type A

The eruption rate - flow length relationship determined for the lava flows of Mount Etna by Lopes and Guest (1982) is re-examined in Figure 3.6. Resolving the data into the equivalent of flow units (Type A) and flow fields (Type B) has the effect of reducing the scatter within the limiting envelopes. This was predicted by Walker (1973). As observed with the volumetric data, there is a conspicuous separation between flow types; for comparable lengths of flow, Type A flows nearly always have higher eruption rates than Type B flows. The solid boundary in Figure 3.6 separates the two populations. The remaining spread observed within each group is partly due to inaccuracies in eruption rate values (considered to be no greater than 30%), but mainly reflects the fact that the eruption rates have been calculated over differing durations. Within the Type A group of flows, the eruption rates have been calculated over durations ranging from 40 hours to 26 days. The influences of ground slope are largely masked by this effect and by the inaccuracies in the effusion rate data.

The influence of duration on the eruption rate - flow length relationship can be further examined by modifying the empirical relationship established between
Figure 3.6 Logarithmic eruption rate - flow/flow field length diagram for the post-1750 lavas of Mount Etna. The dashed lines represent the confining limits of Walker (1973). The dotted line is an arbitrary boundary which separates Type A and Type B flow field data. Numbers accompanying each data point corresponds with its identification number in Table 3.2.
erupted volume and flow length (equation 3.3). Since the eruption rate is derived by dividing the total erupted volume by the eruption duration, erupted volume can be re-expressed in equation 3.3 in terms of these two parameters. This yields:

$$L = c \left( \frac{Q_0}{10^6} \right)^m t$$

(3.4)

where $c$ and $m$ are the constants (derived for equation 3.3), $Q_0$ is the eruption rate ($\text{m}^3\text{s}^{-1}$), and $t$ is the eruption duration (s). (The $10^6$ is a scaling factor required to balance the units of volume). This equation has several potentially useful applications in that:

1. It should be possible to contour any eruption rate - flow length plots according to specific time intervals.
2. If the distance between an active flow front and the eruptive vent is known at specific time increments during the emplacement of a flow unit, then it should be possible to monitor changes in the eruption rate (to a first approximation) during the emplacement of the flow unit (Pinkerton, 1987). This type of approach is not restricted to Type A flows alone and can also be used as a means of quantifying the eruption rate of primary flow units in Type B flow fields, provided that the relevant information is available within eruption reports. Unfortunately, this is not always the case.

A detailed, but by no means comprehensive examination of historical records yielded the data listed in Table 3.3. The eruption rate - flow length data are displayed in Figure 3.7 and the diagram has been contoured in terms of increasing time. It is clear from Figure 3.7 that at a specific duration, the eruption rate can take on a range of values. However, when examined in conjunction with flow length, a more significant conclusion becomes apparent. For those lava flows in Figure 3.7, where it is possible to follow the paths of changing eruption rates across time contours, the longest flow at any pre-determined time is always associated with the highest eruption rate. The final length of a lava flow unit must be related in some way to the effusion rate. In addition, where it has been possible to track the emplacement progress of an individual flow unit, such as 1910 (43), it is clear that the eruption rate does not vary
Table 3.3 Information collected from historical eruption reports, identifying position of flow fronts at various time intervals after the start of the eruption. Eruption rates, $Q_*$, were calculated using Equation 3.4.
<table>
<thead>
<tr>
<th>Year</th>
<th>Time</th>
<th>Flow Front Position</th>
<th>Flow Length</th>
<th>$Q_r$ (m$^3$/s)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>1843</td>
<td>39h</td>
<td></td>
<td>9.3 km</td>
<td>297</td>
<td>Gemmellaro (1843)</td>
</tr>
<tr>
<td></td>
<td>96h</td>
<td></td>
<td>12.5 km</td>
<td>240</td>
<td>Tanguy (1981)</td>
</tr>
<tr>
<td>1852</td>
<td>7d</td>
<td>Ballo</td>
<td>8.5 km</td>
<td>56</td>
<td>Gemmellaro (1854)</td>
</tr>
<tr>
<td>1865</td>
<td>7d</td>
<td>Sciara di Scoriavacca</td>
<td>8.2 km</td>
<td>51</td>
<td>Fouqué (1865)</td>
</tr>
<tr>
<td>1879</td>
<td>45h</td>
<td>Linguaglossa - Randazzo Road</td>
<td>6.8 km</td>
<td>124</td>
<td>De Saussure (1879)</td>
</tr>
<tr>
<td>1886</td>
<td>3d</td>
<td></td>
<td>4.0 km</td>
<td>23</td>
<td>Silvestri (1886)</td>
</tr>
<tr>
<td>1892</td>
<td>57h</td>
<td>Rinazzi Vineyard</td>
<td>4.6 km</td>
<td>39</td>
<td>Riccò and Arcidiacono (1904)</td>
</tr>
<tr>
<td></td>
<td>57h</td>
<td>Mte. Albero</td>
<td>4.1 km</td>
<td>30</td>
<td>Mercalli (1893)</td>
</tr>
<tr>
<td>1910</td>
<td>7½h</td>
<td>Mte Leo</td>
<td>5.4 km</td>
<td>436</td>
<td>Riccò (1912)</td>
</tr>
<tr>
<td></td>
<td>14½d</td>
<td>Cisterna della Regina</td>
<td>10.5 km</td>
<td>44</td>
<td>Oddoné (1911)</td>
</tr>
<tr>
<td>1911</td>
<td>11½h</td>
<td>Linguaglossa/ Randazzo Road</td>
<td>5.6 km</td>
<td>310</td>
<td>Riccò (1912), Platania (1912)</td>
</tr>
<tr>
<td>1923</td>
<td>7h</td>
<td>800 m asl.</td>
<td>5.8 km</td>
<td>552</td>
<td>Ponte (1923)</td>
</tr>
<tr>
<td>1928</td>
<td>41h</td>
<td>Circum Etna Rail Bridge</td>
<td>7.5 km</td>
<td>171</td>
<td>Imbò (1928)</td>
</tr>
<tr>
<td></td>
<td>5d</td>
<td>Messina to Catania Rail</td>
<td>9.0 km</td>
<td>89</td>
<td></td>
</tr>
<tr>
<td>1947</td>
<td>24h</td>
<td>1,250 m. asl.</td>
<td>5.5 km</td>
<td>142</td>
<td>Cucuzza Silvestri (1949)</td>
</tr>
<tr>
<td></td>
<td>65h</td>
<td>1,068 m. asl.</td>
<td>5.7 km</td>
<td>57</td>
<td></td>
</tr>
<tr>
<td>1950-51</td>
<td>7h</td>
<td>1,250 m. asl.</td>
<td>4.9 km</td>
<td>373</td>
<td>Cumin (1954)</td>
</tr>
<tr>
<td></td>
<td>81h</td>
<td>1,100 m. asl.</td>
<td>6.4 km</td>
<td>60</td>
<td></td>
</tr>
<tr>
<td>1981</td>
<td>3½h</td>
<td></td>
<td>5.3 km</td>
<td>964</td>
<td>Guest et al. (1987)</td>
</tr>
<tr>
<td></td>
<td>21h</td>
<td>Alcantara River Bed</td>
<td>7.9 km</td>
<td>377</td>
<td></td>
</tr>
<tr>
<td>1983</td>
<td>10d</td>
<td>1,420 m. asl.</td>
<td>4.8 km</td>
<td>10</td>
<td>Frazzetta and Romano (1984)</td>
</tr>
<tr>
<td>1985i</td>
<td>60h</td>
<td>2,080 m. asl.</td>
<td>2.1 km</td>
<td>6</td>
<td>Romano and Vaccaro (1986)</td>
</tr>
<tr>
<td>1985ii</td>
<td>18h</td>
<td>1,700 m. asl.</td>
<td>3.9 km</td>
<td>85</td>
<td></td>
</tr>
<tr>
<td>1986-87</td>
<td>8h</td>
<td>1,750 m. asl.</td>
<td>1.8 km</td>
<td>32</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.3
Figure 3.7 Logarithmic theoretical eruption rate (m³/s⁻¹) - flow length (km) diagram for the post-1750 output period. The theoretical eruption rate was calculated using equation 3.4. The diagram is contoured according to increasing time intervals (at order of magnitude intervals), and the dashed confining limits of Walker (1973) are included. Data points are joined where it was possible to calculate the eruption rate at intervals during the emplacement of individual flow units. Each data point is identified by its identification number (Table 3.2).
linearly with time (note that successive time contours represent an order of magnitude increase each time). Since changes in the eruption rate with time reflect sympathetic changes in the real-time effusion rate, an inverse relationship suggests that the real-time effusion rate must decrease with time. This statement is important to understanding the emplacement of lava flow units.

In general, the first few hours of an eruption are characterised by a high effusion rate during which a new flow unit undergoes rapid extension. The rate of lengthening decreases after the first few days in response to a decreasing underlying effusion rate, but also because a cooling induced crust forms at the flow front which inhibits movement. A point must be reached where lengthening cannot be sustained because the real-time effusion rate at the vent decreases to the extent that it cannot supply new lava to the flow front at a rate which is fast enough to overcome the increasing yield strength of a thickening crust. In this way, lava backpressure will develop within the main channel, possibly extending back to the vent area. This is either relieved by breaching about the flow front or breaching of a weaker levee further upstream, thereby generating a new flow. For the majority of Type A flows, this stage is never reached because the eruption terminates. The final length of a Type A lava flow is therefore directly related to effusion rate; the higher the effusion rate the greater the distance travelled by the lava flow before the crust becomes thick enough to resist forward movement.

Not all lava flows are sustained for durations long enough for cooling to be important. As first noticed by Lopes and Guest (1982), a few eruption rate data points lie below the lower limit of Walker (1973) - see Figure 3.6. Similar patterns were observed for Hawaiian lava flows by Malin (1980) and Pinkerton (1987). By virtue of the original selection criteria, the lengths of these flows cannot be cooling-limited for they were emplaced in less than 30 hours. They must therefore be volume-limited, an insufficient volume of magma being available to feed the lava flow to its cooling limited length. An example of a volume-limited flow is provided in the northern flank eruption of 1981 (90) (Guest et al. 1987). This lava flow underwent a period of rapid extension at high effusion rate during the first few hours of the eruption, but this phase suddenly ceased after a period of only 40 hours when the lava supply ended. The path followed by the changing eruption rate of this flow is shown in Figure 3.7.
On the basis of a heat conduction model, Guest et al. (1987) and Pinkerton (1987) argue that had the supply of lava continued to this particular flow, it could have continued extending beyond its 8.25 km length.

In summary, Type A flows are associated with high effusion rates and short eruption durations.

**Type B**

An implication of the above flow development model is that all lava flows will have a Type A profile during the early stages of an eruption. This point is largely borne out by the Type B flows, where all available lines of information converge on the fact that these flows initially follow the same systematic pattern of growth. Historical reports of Type B eruptions indicate that for the majority, the initial flow of each eruption reached its maximum length within the first few days e.g. 1852 (Gemmellaro, 1854); 1865 (Fouqué, 1865); 1892 (Riccò and Arcidiacono, 1904). However, unlike Type A flows, Type B flows continue to be fed after the initial phase. Once the principal flow has reached what is assumed to be its cooling limited length (though other 'chance' factors could effect the generation of new flows before this point is reached (Kilburn and Lopes, 1988)), there is a sufficient supply of magma available to feed new flows which might sprout from the principal channel in the vent area. The effusion rate at which these new flows are emplaced must be less than that of the primary flow because

1. The new flows never exceed the length or volume of the first (Figure 3.3) and
2. This is implied from the inverse relationship between eruption rate and eruption duration (Figure 3.7).

Examination of Type B flow fields on aerial photographs confirms that the primary flow has the largest dimensions of all the flow units in the flow field and is similar in size to that of individual Type A flow units.

Successive flow units generated during the course of the eruption evolve to their maximum thermal length and contribute to widening the flow field (Lopes and Guest, 1982; Kilburn and Lopes, 1988). Planimetric evolution continues (unless topographically restricted) until the lava forms a field through which every available path which can take lava away from the vent, equals the maximum thermal length.
Once the flow field has reached its maximum width, new flows become superimposed on the flow field, adding to its overall thickness. Large thicknesses of lava accumulate around the vent area and complex tube systems develop within the pile. These supply lava to ephemeral boccas (Guest et al. 1980b, Chester et al. 1985). During these latter stages, the development of the flow field is restricted to the proximity of the vent area, which takes on the morphology of a pahoehoe-textured tumulus e.g. 1852-53; 1865. There is therefore an important duration consideration in the formation of Type B flow fields. Clearly, not all Type B flow fields are supplied for long enough for them to evolve to this final stage: the range of observed aa-textured flow fields must therefore reflect the various stages along this evolutionary trend.

On Etna, lava tubes develop in regions of the flow field where lava movement is sustained for several weeks. They are generally formed by roofing over of the principal supply channel above the level of branching a few weeks after the eruption commenced e.g. 1950-51 (Cumin, 1954); 1983 (Frazzetta and Romano, 1984). The transport of lava within a lava tube is more efficient than in an open channel because the lava is constrained by its cross-sectional area and is subject to reduced rates of cooling (Swanson, 1973). In this way, lava flows supplied by tubes should be able to attain greater lengths (Malin, 1980). However, a decreasing underlying effusion rate generally counteracts this effect for Etnan lava flows, thereby restricting the length achieved.

As mentioned in the volumetric section of 3.4.2, not all Type B flow fields initially follow this systematic pattern of growth during the early stages of their development. The first lava flows to be emplaced during the 1950-51 (54), 1983 (92) and 1986-87 (99) eruptions were exceeded in length by a later flow. However, once the lengthening stage was complete, these flow fields evolved in a similar way to all the other flow fields, though the morphological changes were displaced from the vent to the break of slope. A histogram by Frazzetta and Romano, (1984) showing the variation of eruption rate with time for the 1983 eruption reveals that the emplacement of the longer second flow unit was accompanied by an increase in the eruption rate. No such 'double peak' appears in the eruption rate histogram of Cumin (1954) for the comparable 1950-51 eruption. In these three instances, the differing growth pattern
is probably a result of this topographic factor rather than any specific changes in the effusion rate with time.

A further enigma is encountered when deciphering the real-time effusion rate histories of the two pahoehoe flow fields of this period. No major channel fed flows are discernible on the 1792-93 (24) and 1764-65 (19) pahoehoe flow fields which could be related to the initial high effusion rate phase observed for the aa flow fields. The dimensions of individual flow units and the pahoehoe surface textures are consistent with low effusion rates. It is possible that these eruptions were characterised by low effusion rates throughout, the flow fields undergoing progressive lengthening between periods of tumuli construction.

3.4.4. Discussion

Eruptions that produce both Type A and Type B flow fields normally start with relatively high effusion rates (> 10 m$^3$s$^{-1}$) and an initial flow usually travels several kilometres in the first day or few days. The distance travelled by an individual flow is dependent on the rate of effusion at the vent; the higher the rate of effusion, the greater the length achieved. Lengthening ceases when either the volume of magma available for eruption is exhausted, for example the 1981 flow (Guest et al. 1987) or the rate of lava supply to the flow front is insufficient for the basal shear stresses to overcome the increasing yield strength of a thickening crust, for example the 1983 flow field (Guest et al. 1987). Although some branching may occur as a result of overflows and topographic effects, the main bulk of the flow field is fed by a single open channel. Such a situation normally exists for the first few days of the eruption. Termination of the eruption at this stage results in a Type A morphology flow field.

If the eruption continues, substantial branching may occur from the main channel by overflows, and breaching of the levees. This is normally the result of back pressure within the main channel which develops as a result of channel blockage downstream or topographic effects (Guest et al. 1987). Diversion of lava into the new branches terminates the supply of lava to the main channel which may then drain. Successive branching may occur to form a compound flow field of Type B. With continued effusion, the main channel above the level of branching usually roofs over to form a tube and, as effusion rate decays, a complex tube system may develop
giving rise to numerous short flows superimposed over the original flow field. Once
lengthening has ceased, the additional lava volume contributes to the thickening and
widening of the flow field around a fan of distributary tubes.

The difference in morphology between Type A and Type B flow fields of the
post-1750 period is thus the result of duration, changing effusion rate with time and
available volume.

3.5 Lava Flow Morphologies of the Early 17th Century

Lavas erupted between 1600 and 1689 have broadly the same chemical
composition as those lavas of the following period, but they differ petrographically.
Morphologically, the flow fields display a greater variety of final shapes, and
pahoehoe surface textures predominate. Amongst the morphologies observed are:
(i) Compound (as defined by Walker, 1971) aa-textured flows that are narrow by
comparison with their length and have a well-defined central channel with numerous
overflows e.g. 1607 (1) lava flow SW flank.
(ii) Compound, aa-textured flow fields that are wide by comparison with their length.
These flow fields are composed of several flow units, each having well developed
channels and overflows. Individual flows emanate from the vicinity of the vent area
e.g. 1646-47 (8), NE rift and 1669 (10), S Rift.
(iii) Pahoehoe-textured flow fields that are broad by comparison with their length and
are compound in character. Tumuli occur over the flow field where there has been
a localised concentration of effusion from ephemeral boccas e.g. 1651-53 (9) W flank.
(iv) Pahoehoe-textured compound flow fields which are characterised by mega-tumuli,
perched lava lakes and lava terraces, as described by Guest et al, 1984b e.g. 1614-24
(3), (Plate 1).

In order to maintain consistency with the post-1750 morphological analysis, the
flow fields are once again grouped according to their morphology into Types A and
B. Only (i) qualifies for the Type A group of narrow flow fields while the remainder
((ii), (iii) and (iv)) are collated to the Type B group of broader flow fields.
Palaeomagnetic evidence reveals that the flow field attributed to the eastern flank
eruption of 1651 may be considerably older than this date (Tanguy, 1980). This flow
field is therefore omitted from this study.
The aim of the following section is to establish whether differences between the final morphologies of 17th century and post-1750 flow fields are a product of differing eruptive styles, or that they merely reflect varying stages of evolution along the same systematic pattern of growth. An independent study of the 17th century flow fields, along the lines of that conducted for the post-1750 period in section 3.4 is not worthwhile due to the small size of the data set. Instead, the 17th century data are superimposed on the post-1750 diagrams and a comparative approach is adopted to see whether the relationships found between eruption parameters and flow dimensions in the latter period are representative of the former period. The analysis follows the same procedure implemented in section 3.4.

3.5.1 Rheology

Some indirect measure of 17th century lava rheology can be obtained by examining certain morphological aspects of the flow fields. One such feature which is suitable for this purpose is surface texture. The predominance of pahoehoe textures during this output period might suggest more fluid lava rheologies. If the lavas were more fluid then, as discussed in section 3.4.1, the pahoehoe flow regime should be able to avoid the transition to aa until higher rates of deformation, and pahoehoe channel morphologies might be supported. However, the dimensions and morphology of individual pahoehoe flow units within flow fields such as 1614-24 (3), 1634-38 (6) and 1651-53 (9), are consistent with them having been fed at low effusion rates (Guest et al. 1984a), probably comparative to those of the post-1750 period.

For those eruptions of the 17th century where high effusion rates are known to have occurred, aa flow textures predominate (e.g. 1646-47 (8) and 1669 (10)). These particular eruptions were also highly explosive, and it is possible that the rheology of the lava may have differed from that which supplied the pahoehoe flows. Nevertheless, some explosive activity is reported from the summit craters during the low effusion rate emplacement of the 1614-24 pahoehoe flow field of the northern flank (Romano and Sturiale, 1982) and yet, any rheological changes which might have occurred as a result of this activity were not sufficient to have any drastic affect on the pahoehoe-aa transition threshold for this eruption. Therefore, for the 17th century
period flow kinetics had a greater effect on the transition threshold between pahoehoe and aa than any rheological changes induced by degassing.

An additional source of information about the rheology of these lavas can be obtained from their cooling histories. Petrographically, the porphyritic 'cicirara' textures of these lavas developed under conditions where crystal nucleation was inhibited but where the growth of plagioclase crystals was favoured. The lower degree of undercooling suggested by the abundance of large crystals and the paucity of smaller ones may indicate that the 17th century lavas had higher temperatures upon eruption compared to the post-1750 lavas. It is therefore possible that the lavas were also more fluid. Even so, from a rheological viewpoint, the fact that well-developed crystals were already present in the lavas upon eruption indicate that the pre-eruptive magmas were at temperatures below the liquidus. This increases the likelihood that the lavas were behaving as non-newtonian fluids at the vent.

Overall, whilst it is possible that lava rheology may have varied between the two output periods, its role in determining the final morphology of the 17th century flow fields is likely to have been secondary.

3.5.2 Volume

*Type A*

Most of the eruptions in this output period produced lava flows that eventually developed Type B final morphologies. For these particular eruptions, most of the evidence for the initial stages of their morphological development has been lost. To accentuate the problem, of the three eruptions which did produce lava flows of Type A morphology, the two lava flows of 1643 (7) and 1689 (13) have, for the greater part, been buried beneath younger lavas. There is therefore considerable uncertainty in the final lengths and volumes of these flows. For the solitary Type A flow remaining, that of 1607 (1), it plots in Figure 3.8 close to the volume-length correlation (equation 3.3) of the post-1750 period.

A single reliable data point does not provide sufficient evidence to support a conclusion which might suggest that the distribution of volume within a 17th century, Type A flow is the same as within a post-1750 flow. However, considering the general consensus reached by independent morphometric studies about the question
of volume distribution within basaltic lava flows, notably Hawaii (Malin, 1980) and Etna (Lopes, 1985), the eruptive style, and physical and chemical properties of the 17th century lavas would have to be significantly different from the post-1750 lavas to greatly affect this relationship. Based on the available information, there is little evidence to suggest that this was the case.

**Type B**

With the exception of the small volume 1610 (2) eruption, all other 17th century Type B flow fields exceed the volume of the largest post-1750 Type B flow field. The 17th century period saw the eruption of the longest observable sub-aerial flow since systematic records began - the 17 km lava flow of 1669 (10). By the end of the 126 day eruption, some 850 x 10^6 m^3 of lava had been erupted (Figure 3.8) into a vast Type B flow field; a volume which easily exceeds that of any Type B flow field in the post-1750 period. Emplacement of the longest channel-fed flow took in excess of 43 days (Chester et al. 1985, table 3.3). The long duration may have been because the erupting vent was initially supplying three lava flows simultaneously. Assuming the distribution of volume within this initial flow to vary according to equation 3.3, some 170 x 10^6 m^3 of lava was erupted into this flow. In reality, the emplacement of the longest flow was probably more complex than assumed by equation 3.3; the morphology of the flow on the outskirts of Catania indicates that the flow may have consisted of several channels. The above value is therefore probably a conservative minimum volume for this particular flow. Other aa-textured, Type B flow fields of this period, such as 1646-47 (8) also have discernible flow units within the flow field which can be traced from the vent area to the flow field front. This suggests similar early growth patterns. A volume of approximately 35 x 10^6 m^3 is suggested from equation 3.3 for the longest flow in this flow field.

As observed with the pahoehoe flow fields of 1764-65 (19) and 1792-93 (24) in the post-1750 period, it is difficult to ascertain the growth pattern of the 17th century pahoehoe flow fields. During the initial stages of the 1651-53 (9) eruption on the western flank, Tanguy (1981) and Chester et al. (1985) report the rapid emplacement of a long flow unit which reached the outskirts of Bronte. Evidence for this flow can only be seen in the vicinity of Bronte where the flow has a distinctive
Figure 3.8 Logarithmic diagram of volume erupted during 17th century eruptions against the length achieved by the flow/flow field. The filled square is the data point for the most voluminous eruption of the post-1750 period, that of 1950-51 (54). The relationship established between the volume erupted and the flow unit lengths of post-1750 Type A eruptions (equation 3.3) is also included (i). Each datum point can be identified in Table 3.1 by its identification number.
aa-textured, channel morphology; closer to the vent it is buried beneath superficial tumuli cover. For this particular pahoehoe flow field, it appears that it may initially have developed along the lines of the aa Type B flow fields of the post-1750 period, and that the tumulus stage of development occurred later.

Elsewhere, Guest et al. (1984b) document that the flow which determines the length of the 1614-24 (10) flow field is an aa channel-fed flow but they consider this unit to have originated as a breakout from the front of a tumuli. For the 1614-24 flow field, the complex stratigraphy of the flow field is best explained by a complex growth pattern dominated by repeated cycles of flow extension, widening and thickening. During each cycle, the pattern of flow field growth favoured the construction of a tumulus. Successive tumuli were constructed downslope following breakouts either from the flow front, or from behind the tumuli (Guest et al. 1984b), but the formation of a new tumulus did not necessarily terminate the growth of the preceding one. Though Guest et al. are only describing the 1614-24 flow field, the stratigraphic relationships of tumuli observed within other pahoehoe flow fields of this period, such as 1634-38 (6) and 1651-53 (9), are comparable. This suggests that at some stage during these eruptions, the lava flow fields underwent similar patterns of growth. In these instances, the distribution of volume was not principally directed into a single flow unit.

3.5.3 Eruption Rate and Duration

Type A

An examination of the relationship between eruption rate, duration and flow length for the 17th century lava flows is somewhat restricted because of the lack of duration data (Figure 3.9). This is particularly a problem for the Type A lava flows for which there is virtually no reliable duration information. However, where there are volumetric data available, it should be possible to place a lower estimate on the eruption rate of a Type A flow by constraining the duration to the maximum observed for post-1750 Type A eruptions (32 days). For the 1607 (1) eruption, a minimum eruption rate of 25.8 m³s⁻¹ is suggested.
Figure 3.9 Logarithmic diagram of eruption rate versus the length of 17th century lava flow/flow fields. The dashed lines are the confining limits of Walker (1973) and the solid diagonal line is the post-1750 boundary which separates Type A and Type B flow fields. For the two Type A data of unknown duration (1607 (1) and 1689 (13)), their eruption rates have been calculated for an eruption duration of 32 days. This duration value represents the maximum associated with a post-1750 eruption which produced a Type A morphology flow field at an equivalent vent elevation. Error bars are included for all length measurements (though these are generally smaller than the size of the data point symbol) and for most of the eruption rate data (apart from (1) and (13)). The 1643 (7) lava flow has a large uncertainty associated with its length. The length value used is that for the exposed portion of the lava flow. The large positive vertical error bar assumes that the flow front of the unexposed portion of the flow coincides with the flow front of the overlying flow unit.
Figure 3.9

Flow Length (km) vs. Eruption Rate (m³s⁻¹)
Type B

From the available eruption rate data for the Type B flow fields of the 17th century period, it is clear that over comparable time periods, the eruption rates are much greater than their post-1750 equivalents. For example, the 126 day 1669 (10) eruption has an eruption rate of 78.0 m$^3$s$^{-1}$ compared to the 8.0 m$^3$s$^{-1}$ rate of the 131 day 1983 (88) eruption of the post-1750 period. This eruption rate value is also greater than that for most of the shorter duration (<26 day) Type A eruptions of the following period. This can be seen from Figure 3.9 where the 1669 (10) data point plots within the field of the post-1750, Type A flows. The same is true for the other Type B, aa flow field of this period, that of 1646-47 (8).

The eruption rates of the Type B pahoehoe flow fields clearly overlap with the Type B region of the post-1750 eruptions in Figure 3.9. However, this is once again considered fortuitous, for the eruption rates of these particular eruptions were calculated over durations up to an order of magnitude greater than those of the following output period. For example, the longest duration eruption of the post-1750 period, the 1½ year 1764-65 (19) pahoehoe eruption has an eruption rate of 1.3 m$^3$s$^{-1}$ compared to the 3.8 m$^3$s$^{-1}$ rate for the 10 year 1614-24 (3) eruption.

If the available 17th century, Type B data are examined as a whole, the higher eruption rates over comparable time periods imply that, for the greater part of these eruptions, the real-time effusion rates must have been higher than those of the post-1750 period. In order to account for the presence of two Type B 17th century data in the post-1750, Type A field (to the right of solid boundary) in Figure 3.9, these particular eruptions must have characterised by slower rates of effusion rate decay compared to the post-1750 and 17th century Type A eruptions.

3.5.4 Discussion

The absence of helpful eruption reports, and the predominance of Type B flow fields, makes deciphering the eruptive conditions which prevailed during the early stages of 17th century eruptions difficult. Of the information which is available, there are indications that the early stages of most of these eruptions were dominated by high effusion rates during which a Type A lava flow unit was emplaced e.g. 1646-47 (8), 1651-53 (9) and 1669 (10). Though higher overall real-time effusion rates have been
advocated for the more voluminous Type B eruption, the fact that the lengths of the majority of these Type B flow fields and Type A flows are no longer than those for the post-1750 period, suggests comparable initial emplacement conditions. The 1669 lava flow is clearly an exception.

The planimetric development of the flow fields after this initial stage probably followed the same pattern of growth as the post-1750 flow fields, with additional lava contributing to the thickening and widening of the flow field instead of lengthening, as the underlying effusion rate decreased. Whilst this stage may represent the evolutionary limit of post-1750 Type B flow fields, the majority of 17th century flow fields progressed further in their planimetric evolution with the construction of pahoehoe textured tumuli, superimposed on the original aa textured flow field. The development of these tumuli probably commenced in the proximity of the vent area or in that region of the flow field where the flow first encountered a pronounced break of slope. Development of the tumuli was probably at low effusion rates, (as evidenced by the pahoehoe texture and small size of individual flow units), with the progressive construction of a mound of lava. Based on the stratigraphy of the tumuli within the flow fields, the construction of individual tumuli was largely terminated by a breakout from their base, and the lava within the tumulus drained. The cycle was then repeated with the construction of a parasitic tumulus, generally downslope of the original. The numerous tumuli on the 17th century lava flow fields suggests that this sequence of events recurred several times.

The morphology of the 17th century flow fields is once again considered to be the result of duration, changing effusion rates with time and the available volume. During much of the later stages of the long duration, pahoehoe eruptions, the effusion rate at the surface must have approximated to the rate of magma input from depth, and may therefore have been fairly constant.

3.6 Summary

The range of lava flow morphologies which have formed on Etna since 1600 probably reflect flow fields at various stages of evolution along a systematic pattern of growth, Type A representing the juvenile stage of development, and the pahoehoe Type B flow fields of the 17th century representing the latter stages. Overall, the
variation of effusion rates with time probably followed the same pattern, with high initial effusion rates rapidly giving way to a protracted period of low effusion rates. Though eruption rates are greater for the 17th century, Type B eruptions, the initial effusion rates may have been comparable between the two output periods. This hypothesis is based on the comparable dimensions of the initial flow units. The greater magnitude of 17th century eruption rates was probably determined by higher effusion rates within the protracted waning phase which followed the initial high effusion rate phase. This was caused by a higher rate of magmatic input from depth during the course of the eruption. The effect of higher effusion rates was probably to enhance the progression of flow development rather than to dramatically change the natural planimetric evolution of the flow field.

Though the planimetric evolution of the 17th century flow fields may not have been greatly affected by the higher effusion rates, the very fact the magmatic output and effusion rates were higher within this period must indicate that conditions within the volcano have changed between the two periods under discussion. This is reinforced by the contrasting petrography of the lavas erupted, and by the differences in the magnitude of individual eruptions, in terms of volume and duration. Clearly, in any study of Mount Etna, whether it be a morphometric study of the lavas (as described above), or a general examination of the volcanic system, these two historical periods of contrasting magmatic output, from 1600 to 1689 and post-1750, must be considered separately.
PART II: The Post-1600 Volcanic System of Mount Etna.
CHAPTER 4

The Post-1750 Volcanic System of Mount Etna: Eruptive Activity and Lava Flow Emplacement on the Western Sector

4.1 Introduction

The next two chapters present a detailed examination of the Mount Etna volcanic system during the post-1750 period of magmatic output. Information derived from individual eruptions, such as their volume, duration, eruption rate and spatial distribution on the volcano, are used alongside petrological, geodetic and geophysical observations, to develop a model of the volcanic system. In light of previous eruptive models which have been proposed for Mount Etna, the location of potential magma storage regions within volcano, and eruptive mechanisms are discussed.

The influence of the volcanic system on the morphological evolution of the lava flows is then examined by integrating the conclusions of the post-1750 morphometric study (discussed in chapter 3, section 3.4) into the model.

4.2 Method of Analysis

In this analysis, all the post-1750 eruptions are examined in situ on the flank of the volcano. As a starting point, and as a means of obtaining a general overview, post-1750 flank eruptions (expressed in terms of flow morphology, erupted volume, eruption duration and eruption rate) are mapped out on an areal plan of the volcano in order to test for the possible presence of systematic distribution patterns.

In subsequent sections, individual eruption parameters are examined in greater detail to see whether there are any vertical (elevation) or horizontal (distance from central conduit) restrictions to their magnitude. For a symmetrical volcano, all points at a given elevation will occur at the same radial distance from the centre of the volcano. It therefore follows that information about either dimension can be obtained
through the examination of only one dimension. It is necessary to examine Etna in both dimensions because the edifice displays a strong departure from symmetry (Figure 4.1). For a restricted altitudinal range, eruptions can occur at a range of distances from the central conduit. Conversely, for a restricted lateral range, eruptions can occur over a range of elevations. This adds to the complexity of the analysis.

![Figure 4.1](image)

**Figure 4.1** Four profiles of Mount Etna volcano showing the north, south, east and west flanks (After Wadge, 1977)

A model of the volcanic system is gradually developed from the conclusions of each section. The morphometric study is blended into the model by examining, in the first instance, how the lengths of individual lava flows vary with vent elevation and the lateral distance between the vent and the central conduit. Changes in the planimetric evolution of individual flow fields are then examined to see whether they coincide with variations in the way in which magma is transported through the volcanic system.
4.3 Spatial Distribution

The spatial distribution of the two flow field populations (Type A and B) erupted on the flanks of Etna since 1750 is shown in Figure 4.2A and B. Below 2,000 m asl, the majority of Type B flow fields occur on the east and southeastern flanks of the volcano, within and surrounding the Valle del Bove (Figure 4.2A). Only two eruptive episodes have produced Type B flows on the western flanks, those of 1763 (17) and 1974 (72). All other flows erupted on the north, west and northeast flanks, and at low elevation on the eastern and southeastern flanks have been of Type A.

Above 2,000 m asl, the sectorial distribution of the flow fields is still apparent with Type B flow fields being restricted to the east of the summit. Whilst the Type B flow fields of 1764-65 (19), 1975/76 (74) and 1976/77 (75) were emplaced to the northwest, the vent areas of these eruptions occur within the confines of the northeast rift zone. Type A flows occur mainly on the western flanks, but they are also found on the upper eastern flanks, in association with an orthogonal set of eruptive fissures at the head of the Valle del Bove (omitted from Figure 4.2B for clarity).

The spatial distribution of the two flow field types suggests that lava is erupted in a different way on the eastern sector (those flanks on the seaward side of the rift systems) as contrasted with the western sector (those flanks on the landward side of the rifts). All Type A flows on the western sector (and Type B), and the majority of those Type A low on the eastern sector have volumes less than $60 \times 10^6$ m$^3$ (Figure 4.3A and B). The 1923 (46) eruption of the northeast rift is an exception. The majority of Type B flow fields occurring below 2,000 m asl on the eastern sector have volumes greater than $60 \times 10^6$ m$^3$ (Figure 4.3A) though above 2,000 m asl, there is a general decrease in the volume of lava erupted into these flow fields. This is particularly true above 2,500 m asl where the volumes of Type B flow fields do not exceed $60 \times 10^6$ m$^3$ (Figure 4.3B).

A similar pattern emerges on the duration and eruption rate distribution diagrams. As first noted by Wadge (1981), most of the eruptions on the eastern flank, within and surrounding the Valle del Bove, tend to be of long duration (i.e. more than 50 days). This is illustrated in Figures 4.4A and B. These eruptions are also associated with low eruption rates ($< 10$ m$^3$s$^{-1}$) (Figures 4.5A and B). All eruptions
Figure 4.2 Spatial distribution of the two morphological flow types that characterise post-1750 flank eruptions. Diagram (A) shows the distribution below 2,000 m asl, and (B), above. The narrow directional lines represent Type A flows whilst the broader directional lines represent Type B flow fields. In order to avoid cluttering in that region encompassed by the Valle del Bove, the 1978 (78, 82, 84) and 1979 (87) flow fields have been omitted from this, and the remaining spatial distribution diagrams. The keys refer to both diagrams. Each flow in this diagram, and in the following three diagrams, is accompanied by an eruption identification number which permits each eruption to be traced to its eruption data in Table 3.2.
Figure 4.3 Spatial distribution of post-1750 flank eruptions, expressed in terms of the volume erupted. (A) and (B) are the distributions below and above 2,000 m asl respectively. The data are divided into two samples about a limiting volume of $60 \times 10^6$ m$^3$, a value deliberately chosen to exceed the largest volume associated with an unrestricted Type A flow (as determined in Section 3.4.2). Bold directional lines represent eruptions which exceeded the limiting volume whilst the narrow directional lines represent eruptions with volumes less than $60 \times 10^6$ m$^3$. 
Figure 4.4 Spatial distribution of post-1750 flank eruptions expressed in terms of eruption duration. (A) and (B) are the distributions below and above 2,000 m asl respectively. The data are divided into two samples about a bounding duration of 50 days, a value deliberately chosen to exceed the longest duration observed for a Type A flow. Bold directional lines represent eruptions which exceeded 50 days duration whilst the narrower directional lines represent eruptions which did not last this long.
Figure 4.5 Spatial distribution of the post-1750 flank eruptions expressed in terms of eruption rate. (A) and (B) are the distributions below and above 2,000 m asl respectively. The data are separated into two samples about limiting values of 10 m³/s. Bold directional lines represent eruptions with eruption rates less than the limiting value whilst the narrower directional lines represent eruptions which exceed this value.
on the western sector, and those which produced Type A flows of the lower eastern sector, lasted less than 50 days (Figure 4.4A and B) and had eruption rates in excess of $10 \text{ m}^3\text{s}^{-1}$ (Figures 4.5A and B). The western sector eruptions of 1763 (17), 1974 (72) and 1947 (51) are exceptions in that their eruption rates are less than $10 \text{ m}^3\text{s}^{-1}$.

Based on the above observations, it can be deduced that sectorial processes are operating on the volcano in this period of magmatic output which are allowing long duration, large volume but low eruption rate effusions on the eastern sector but are preventing similar eruptions from occurring on the western sector.

Within the western sector group, the 'eccentric' eruptions of 1974 (72), which were supplied by conduits detached from the central conduit at high levels (Tanguy, 1974), are easily distinguished from the flank eruptions that were fed directly from the central conduit. Other anomalous eruptions have occurred on the western flank in close proximity to the 1974 eruption sites. Though the lavas of the 1763 (17) eruptions have petrographies common to the majority of Etnan lavas, the erupted products have morphologies which are similar to those of the 1974 eruptions i.e. thick, 'viscous' flow units of limited dimension, large ratio of pyroclastic to effusive products. It is this latter factor which distinguishes the 1763 eruptions from the other western sector flank eruptions. Since the eruptive style at the surface is primarily controlled by the ascent of magma through the plumbing system, similarities between the 1763 and 1974 eruptive products may therefore reflect comparable ascent conditions. One implication of this is that the magma which supplied the 1763 eruption sites also ascended along paths which were largely independent of the central conduit system. For this reason, the western eruptions of 1763 and 1974 are not considered any further in the context of this study.

*The remainder of this chapter will concentrate on examining eruptive behaviour on the western sector. Eastern sector eruptive activity is examined in chapter 5.*
4.4 Erupted Volume of Lava

In situations where the central conduit of the volcano is fully charged with magma before an eruption (as inferred by magma standing at a high level in one of the summit craters), a dyke tapping into the conduit at the lowest elevation should have access to the greatest proportion of stored magma. Within this output period, the most voluminous eruptions on the western sector have occurred at mid-elevations, circa 2,000 m asl (Figure 4.6). Above and below this elevation there is an overall pattern of decreasing erupted volume, though eruptions at the same elevation can have a range of values (in excess of errors on individual data points). A limit, A, is added to Figure 4.6 which constrains the maximum volume erupted at each elevation below 2,000 m asl. A similar limit for those eruptions occurring above 2,000 m asl is not viable due to the small number of available data. It is clear from this diagram that there is some process operating below 2,000 m asl on the western sector which is acting to retain a greater proportion of the available volume of magma within the edifice.

4.4.1 Below 2,000 m asl

The distance between the eruptive vent and the central conduit, or expressed another way, the inferred dyke length, appears to have a stronger control on the maximum volume of lava extruded by western sector eruptions occurring below 2,000 m asl (Figure 4.7). For eruptions with vents at the same elevation, such as 1843 (32) and 1910 (43), or at differing elevations, such as 1832 (30) and 1879 (37), the largest volume extruded at a given lateral distance is constrained by a single, well-defined relationship represented by the negative regression line, I. As the length of the dyke connecting the eruptive vent to the central conduit increases so the magnitude of the lava volume extruded decreases. The gradient of the negative regression line, I, is therefore a function of the material retained within the volcano after an eruption. Provided that the rate of recharge from depth exceeds the rate at which magma is solidifying within the volcano, the retained magma is then available for mixing with subsequent magma injections. Evidence for magma mixing on Etna is supported by the work of Scott (1983) who, in a detailed chemical study of the western sector
Figure 4.6  Elevation of post-1750, western sector eruptive vents (m asl) plotted against the volume of lava erupted (x $10^6$ m$^3$). Limit (A) constrains the maximum volume erupted by vents at elevations below 2,000 m asl. There are insufficient data to constrain a similar limit for those eruptions above 2,000 m asl. The values plotted here are the total volume erupted along the fissure versus the altitude of the lowermost active vent. Horizontal lines are error bars. The eruption data in all the following diagrams can be traced to Table 3.2 by an eruption identification number which accompanies each datum.

Figure 4.7  Horizontal distance of post-1750 western sector eruptive vents from the central conduit (inferred feeder dyke length, km) plotted against the total volume of lava erupted (x $10^6$ m$^3$). The dashed regression line I, constrains the maximum volume extruded by eruptions occurring below 2,000 m asl with vents at various distances from the central conduit. This is of the form $V = -3.8L + 59$. ($r = -0.979$; $n = 6$ and $P = < 0.6\%$). (NOTE: The axes of this diagram, and all subsequent diagrams where one of the axes parameters is the horizontal distance between the eruptive vent and the central conduit, are plotted in this way to aid comparison with the altitude diagrams. The summit of the volcano is at the origin).

Since I is a straight line in Figure 4.7, the non-constant slope of the volcano dictates that this limit cannot be represented by a straight line on an elevation diagram (A). Insufficient data prevents a more accurate representation of the limit on this diagram: this is a problem which arises in most of the following lateral distance - vent elevation diagrams.
eruption of 1981, attributed the observed chemical variation to mixing of residual magma from the 1979 eruption with fresh magma which rose from depth in 1980.

It follows from the above relationship that the largest fraction of the available magma volume which can be extruded on the flanks of the western sector is represented by that point on the linear regression line in Figure 4.7 at a distance of 2.9 km from the central conduit. This is the shortest possible lateral distance between the central conduit and the surface of the volcano at elevations less than, and equal to 2,000 m asl. A maximum magma volume of some \(48 (\pm 7) \times 10^6\) m\(^3\) is suggested. It is conceivable that most of this volume could be stored at a high level in the volcanic system, possibly within the volcanic edifice itself. Wadge (1977) estimated that when full, the Chasm (one of the summit craters), could hold some \(42 \times 10^6\) m\(^3\) of magma above 1,800 m asl, and proposed that the magma feeding flank eruptions drained from this stored volume. The estimated magma volume is close to the largest lava volume erupted during a single western sector eruption (1843 (32)). However, Tanguy (1979) highlights several problems with Wadge’s theory. Firstly, on the basis of information provided by Kasser et al. (1978), he questioned the volumetric capacity of the Chasm, arguing that Wadge had grossly overestimated the internal dimensions of the crater. On top of this, Tanguy noted that the assumption that the Chasm is nearly brimful before eruptions is not always fulfilled. In 1971, the Chasm was empty to a depth of 1 km prior to the eruption (Murray and Guest 1982). Indeed, it is frequently the case that summit activity prior to flank eruptions is concentrated away from the inactive Chasm, at the northeast or southeast craters. Of greater importance is the fact that within the post-1750 period, the configuration of the summit craters has changed dramatically over relatively short periods of time. For example, the Chasm has only been in existence since the early 1950’s (Guest, 1973). The high level storage of magma within the volcanic system cannot therefore be attributed to a single crater but is probably accommodated within the whole central conduit and, as suggested by the geophysical evidence, a few high level dykes. The lateral extent of the central conduit is not known but it probably exists beneath the present summit cone and may possibly extend to beneath the Piano Caldera and a short way into the rifts.
On the basis of the above arguments, whilst gravitational drainage may be an important mechanism in the initial release of magma during eruption, it is highly improbable that the 54 (±7) x 10^6 m^3 of lava extruded during the 1843 (32) eruption at 1,950 m asl drained from above this elevation. The greater proportion of the erupted volume probably ascended with the assistance of buoyancy forces, from below the eruption site.

4.4.2 Above 2,000 m asl

The data for western sector eruptions occurring above 2,000 m asl depart from the trend established for eruptions below 2,000 m asl (Figure 4.7). At equivalent distances from the central conduit, smaller volumes of lava were extruded. Though these eruptions occurred from vents at differing elevation, there is no well-defined pattern of variation between the two parameters in Figure 4.7. For instance, the 1949 (53) eruption occurred at an elevation (2,150 m asl) which is lower than that of the 1947 (51) eruption, and from a vent which is closer to the central conduit. Whilst it might be expected that these conditions are more favourable for the extrusion of a large volume of lava, the eventual eruption only managed to extrude some 8 x 10^6 m^3 of lava, a value which is less than that extruded by the 1947 eruption. Similar anomalous, small volume eruptions occurred in 1874 (35) and 1942 (49). Reports of the 1942 (Cumin, 1943) and 1949 (Cumin, 1950) eruptions do not suggest that the pre-eruption status of the volcano was any different for these particular eruptions. The 1949 eruption commenced with a small terminal eruption within the central craters and was followed shortly after the end of flank activity, by the resumption of terminal activity at the northeast crater (Cumin, 1950). It can only be concluded that there is some process operating above 2,000 m asl which is preventing the extrusion of a larger fraction of the available magma volume.
4.5 Eruption Duration

Western sector eruption durations have been highly variable in the post-1750 period, ranging from the 13 hour eruption of 1942 (49) to the 26 day eruption of 1910 (43). The longest duration eruption occurred at 2,000 m asl (Figure 4.8). Above and below this elevation, there is an overall decrease in the duration of individual eruptions, though for each elevation, eruptions can have a range of durations. A limit (A) is applied to the western sector data occurring below 2,000 m asl in Figure 4.8, which constrains the longest eruption durations yet seen at each elevation. A similar limit for those eruptions occurring above 2,000 m asl is not viable due to the small number of data. Though this pattern appears to be similar to that established for erupted volume (Figure 4.6), the eruption determining the maximum is not the same one in both diagrams.

4.5.1 Below 2,000 m asl

The relationship between eruption duration and \( L_i \) (Figure 4.9) is not as straightforward as that established between erupted volume and this parameter. Though there is a general pattern of increasing duration with decreasing lateral distance and increasing elevation (as outlined by limit line I), most of the western sector eruptions which occurred below 2,000 m asl did not last as long as the duration maximum equivalent to their lateral distance. It also does not follow that the duration of western sector eruptions occurring at the same elevation decrease as \( L_i \) increases. For example, considering the proximity of the 1843 (32) eruptive vent to the central conduit (2.9 km), and the volume of lava extruded by the eruption, it might be expected that this eruption would have been one of the longest duration eruptions of the period. Instead, it lasted only 9 days. The 1910 (43) eruption which occurred at the same elevation (2,000 m asl) but at a greater distance from the central conduit (5.9 km), lasted 26 days. It can be inferred from the distribution of the data in Figures 4.8 and 4.9 that the factors influencing the duration of an eruption are not simply a function of parameters associated with vent elevation and dyke length alone.
Figure 4.8 Elevation of post-1750 western sector eruptive vents (m asl) plotted against eruption duration (days). Limit (A) constrains the longest eruption durations associated with vents occurring below 2,000 m asl. There are insufficient data to constrain a similar limit for eruptions which occurred from vents above 2,000 m asl. The duration value used here is the total duration for which the eruptive fissure was active and not the cumulative duration based on the duration of activity at individual eruptive vents. These duration values are plotted versus the lowermost active vent.

Figure 4.9 Horizontal distance of post-1750 western sector eruptive vents from the central conduit (km) plotted against eruption duration (days). The dashed limit line I, constrains the longest eruption duration at a given lateral distance for those western sector flank eruptions which occurred at, and below 2,000 m asl.
Figure 4.8

Figure 4.9
Influencing Factors

Figure 4.10 Stress distribution within, and below a hypothetical volcanic pile. A column of magma fills the central conduit of a volcano. At a depth Z below the top, the hydrostatic pressure $P_{hz}$ acts against the pressure of the enclosing rocks $P_{rz}$. Below the volcanic pile, the stress distribution approaches lithostatic pressure, and the horizontal and vertical stresses are approximately equal. Above the volcano-basement interface, the stresses are less than lithostatic; the horizontal stresses ($\sigma_2, \sigma_3$) are less than the vertical stress $\sigma_z$ such that $\sigma_z (= \sigma_1) > \sigma_2 > \sigma_3$ (modified from Wadge 1977).

During the initial stages of an eruption, a dyke will only propagate when conditions are satisfied such that:

$$P_m > \tau + \sigma_3$$

(4.1)

where the magma pressure $P_m$, exceeds the combined sum of the tensile strength of the enclosing rocks, $\tau$, and the minimum compressive stress $\sigma_3$, acting to close the fissure (McGuire and Pullen, 1989). Once fracturing has occurred, $\tau$ becomes zero. If dykes are intruded perpendicular to $\sigma_3$, then a lithostatic stress state provides an
upper bound on the horizontal stress (Rubin, 1990). The lithostatic stress state is represented by equation 4.2. At a depth \( z \) below the top of the volcano

\[
\sigma_z = \int_0^z \rho_g dz
\]  

(4.2)

where \( \rho \) is the density of the country rock (inclusive of volcanic pile). This scenario, which assumes an isotropic stress distribution (the three principal stresses are equal in magnitude), probably only applies to conditions at depth within the volcanic system - well below the volcanic pile (Figure 4.10). Though conditions may arise at shallow depth where the horizontal compressive stress at shallow depths in a volcanic pile can exceed \( \sigma_1 \) (Dieterich 1988), it generally follows that the morphology of the volcano dictates that \( \sigma_2 \) and \( \sigma_3 \) are generally smaller than the vertical principal stress - see Figure 4.10. If the volcano is assumed to be an elastic solid constrained by its basement not to expand laterally, and \( \sigma_2 \) equals \( \sigma_3 \), then according to Wadge (1977), an approximate value for \( P_{rz} \), or volcanic load is given by:

\[
\sigma_3 - \sigma_2 = \left[ \frac{\nu}{1 - \nu} \right] \sigma_z
\]  

(4.3)

where \( \nu \) is Poisson’s ratio (the ratio of lateral contraction to longitudinal extension). For rocks with a high tensile strength, a common value for \( \nu \) is 0.25. (Griffiths and King, 1983).

**Interpretation of Duration Data**

An implication of the assumption that the volcanic pile behaves elastically is that as the excess magma pressure within a dyke decreases towards the value of the least compressive stress, the volcanic load will close the eruptive fissure. Taking a literal interpretation, it can be inferred from the fact that the duration of western sector eruptions occurring below 1,900 m asl (Figure 4.8) increase with decreasing dyke
lengths (Figure 4.9) that magmatic pressures can exceed the compressive forces acting to close the fissure for a greater length of time. These favourable conditions probably arise because:

(i) a shorter dyke length reduces the distance between the eruptive vent and the pressure source within the central conduit and,

(ii) from a trigonometric viewpoint, a decrease in the lateral distance between the eruptive vent and the central conduit must also entail a decrease in the volcanic load acting on the fissure.

On the basis of the above inference, the brevity of the 1843 (32) eruption, whose vent is close to the central conduit, could not have been influenced by a greater compressive force acting on the eruptive fissure. The predominant control on the eruption duration in this instance had to be the magmatic driving pressure. A smaller initial magmatic pressure can be discounted by virtue of the large volume of lava erupted in the eruption. Therefore, the shorter duration must indicate that the rate at which the excess pressure was released with time must have been greater during this eruption, so that its magnitude quickly reduced to values below the volcanic load as the magma supply decreased. The extended duration of the 1910 (43) eruption relative to that of eruptions below 1,900 m asl occurring at the same distance from the central conduit, must indicate that the compressive force effecting the closure of the fissure was smaller. The location of the 1910 eruptive fissure on the western edge of the southern rift zone (McGuire and Pullen, 1989) may be important. The two rift zones: northeast and southeast, are the surface manifestation of the interface between the buttressed western sector and the gravitationally unstable eastern sector.

Whilst there is potential variability in the length of time taken to erupt a given volume of lava by those eruptions occurring below 2,000 m asl on the western sector, the eruptive forces are only strong enough to erupt that volume of magma stored in the upper levels of the volcanic system before the fissure closes. A maximum volume of 48 (± 7) x 10^6 m^3 was inferred in Section 4.4.1. The variability in the duration hinders any attempt at precisely quantifying the length of time required to completely erupt the available volume of magma. At best, it is only possible to bracket a duration range which encompasses the extremes observed for the more voluminous eruptions (9 to 26 days).
4.5.2 Above 2,000 m asl

The majority of western sector eruptions with vents above 2,000 m asl did not last as long as those eruptions below 2,000 m asl occurring at the same lateral distance from the central conduit (Figure 4.9). Only the 1947 (51) eruption, at 14 days, lasted longer than 2 days. The general decrease in the magnitude of eruption durations above 2,000 m asl indicates that the magmatic pressures exceeded the compressive force acting to close the fissure for only a short period of time. Since an increase in elevation results in a progressively smaller load acting on the eruptive fissure, the curtailed durations demonstrate that not only does \( P_m \) decrease with increasing elevation, its magnitude must also decrease at an equal, or faster rate with elevation compared to the volcanic load. In order to explain the longer duration of the 1947 eruption, the magnitude of the compressive force acting to close the 1947 (51) fissure must have been smaller than that which was acting on the fissures of other eruptions above 2,000 m asl. As inferred for the 1910 eruption (see page 107), the path followed by this eruptive fissure along the northwest margin of the northeast rift zone may have influenced the magnitude of \( \sigma_3 \).

4.6 Eruption Rate

The variation of eruption rate with vent elevation for post-1750 western sector eruptions is shown in Figure 4.11. The distribution pattern of the data in this diagram is, on the whole, the inverse of that established between vent elevation and the volume (Figure 4.6) and duration (Figure 4.8) eruption components which constitute the eruption rate. In apparent contradiction to the conclusions of Walker (1974) and Lopes and Guest (1982), it does not follow that eruption rates increase with lower vent elevation: indeed it could be argued that eruption rates decrease in magnitude below 2,000 m asl (Figure 4.11). To highlight this point, the equivalent of Lopes and Guest (1982) dotted limit line has been included on Figure 4.11. (It must be emphasised that Lopes and Guest determined their limit line using data covering the entire historical period and not post-1750 alone). There is, however, a tendency towards lower eruption rate values at elevations approaching 2,000 m asl in association with the longer duration, larger volume eruptions. The 1843 (32) datum is an exception.
Figure 4.11  Elevation of post-1750, western sector eruptive vents (m asl) plotted against eruption rate (m³s⁻¹). The eruption rate value predominantly used here is the total volume erupted along the fissure divided by the duration of the effusion. This value is plotted against the lowermost active vent. The bracketed number which follows the eruption identification number in this, and the following diagrams, is the duration over which the eruption rate value was calculated. This is expressed in days (d) and hours (h). The dotted limit line is this study’s equivalent of Lopes and Guest’s 1982 limit which constrains the highest eruption rate at a given elevation.

Figure 4.12  Horizontal distance of post-1750, western sector eruptive vents from the central conduit (inferred dyke length, km) plotted against eruption rate (m³s⁻¹). The limit has been transposed from Figure 4.12 and connects the same data points. The eruption rate is plotted against the active vent which was most distant from the central conduit.
Figure 4.11

Figure 4.12
For the short duration eruptions above 2,000 m asl, their eruption rates exceed those of some of the longer duration eruptions below 2,000 m asl. However, over comparable periods of time, there is a decrease in the eruption rates with increasing elevation towards the summit i.e. a smaller volume is erupted in a given time interval. As first noted by Walker (1974), this suggests that over the initial stages of an eruption, the magnitude of real-time effusion rates increase as the elevation of the vent gets lower on the flanks of the volcano.

It would be unwise to over-emphasise the significance of this limit, for the eruption datum it connects were not calculated over periods of equal duration. There is over 8½ days difference between the final durations of the 1843 (32) and 1942 (49) eruptions. Considering the rapidity at which the real-time effusion rate is known to vary over the initial few hours of an eruption (as noted in chapter 3, section 3.4.3), this small difference in duration could be significant to the magnitude of the eruption rate.

The relationship between eruption rate and the lateral distance between the eruptive vent and the central conduit is shown in Figure 4.12. Included on this diagram is the transposed eruption rate maximum of Figure 4.11. In this instance, this limit constrains the largest eruption rates observed at a given lateral distance from the central conduit. Below 1,900 m asl, there appears to be a pattern of decreasing overall eruption rate as the lateral distance between the central conduit decreases. However, this pattern does not hold for the 1843 (32) and 1910 (43) eruptions, which occurred from vents between 1,900 and 2,000 m asl. For instance, considering the proximity of the 1843 (32) eruptive vent to the central conduit, this eruption extruded a large volume of lava in a disproportionately short duration, hence its relatively large eruption rate. A higher real-time effusion rate must have prevailed during this eruption. Over similar time periods, eruptions above 2,000 m asl with vents at the same distance from the central conduit, have smaller eruption rates relative to eruptions below 2,000 m asl (Figure 4.12).

The absence of any well-defined systematic pattern in the variation of the overall eruption rate with vent elevation and inferred dyke length clearly indicates that those factors which influence the magnitude of the real-time effusion rate and its variation with time are not simply a function of these parameters.
4.6.1 'Peak' Effusion Rates

In addition to logistical difficulties, the technology is not yet available to allow the direct and accurate measurement of real-time effusion rate in the field. However, an estimate of the initial effusion rate can be obtained by examining eruption rates which have been calculated over short time intervals (few hours). In effect, the eruption rate approaches the real-time effusion rate. An immediate drawback of this method is that eruptions seldom last less than 24 hours duration, hence there is little suitable direct eruption information available. The indirect method of eruption rate evaluation presented here utilises the theory and eruption rate data obtained through equation 3.4 in section 3.4.3. For individual eruptions, estimates of the eruption rate at differing time intervals were calculated from information relating to the position of the propagating flow front. The data, which are listed in Table 3.2, are presented in Figure 4.13. No distinction is made between western and eastern sector data in this diagram.

An elevational constraint on the magnitude of the initial effusion rate is suggested by this diagram. In general, for those eruptions where lava fed a single flow unit, the eruption rate over the first few hours shows an increase with lower vent elevation (Figure 4.13). The dotted line connects eruption rate datum calculated for a duration of 7 hours. The largest theoretical eruption rate of nearly 964 m$^3$s$^{-1}$, calculated for the first 3¾ hours of the 1981 eruption, corresponds quite well with the peak effusion rate of 700 m$^3$s$^{-1}$ which was determined independently for this eruption by V.E.S.T. (1981). Another notable feature of this diagram when examined alongside Figure 4.11, is that there is a strong variation in the rate at which the eruption rate changes with time (as demonstrated by the interconnected datum). At low elevation, the eruption rate decreases by over an order of magnitude in a very short period of time. This contrasts quite sharply with those eruptions at a higher elevation, where the difference between the initial and final eruption rate is generally much smaller. This serves to indicate that for these high elevation eruptions there must be a smaller difference in the temporal variation of the real-time effusion rate compared to those eruptions at a lower elevation.

Information relating to the influence of dyke length on the initial effusion rate can be obtained by re-plotting the eruption rate data used in Figure 4.13 in the
Figure 4.13 Elevation of post-1750 eruptive vent (m asl) plotted versus theoretical eruption rate (m$^3$s$^{-1}$) calculated using Equation 3.2. The dotted line connects eruption rate datum which were calculated for a duration of 7 hours. Data are connected by a horizontal line where several eruption rate values (calculated over differing durations) were available for a single eruptive episode. No distinction is made between western and eastern sector data.

Figure 4.14 Horizontal distance of vent from the central conduit (inferred dyke length, km) plotted against theoretical eruption rate (m$^3$s$^{-1}$) calculated using Equation 3.2. The nomenclature is the same as that in Figure 4.12. The position of the ditted limit (which has been transposed from Figure 4.14) is uncertain. Since the limit is considered to be a straight line in Figure 4.14, the non-constant slope of the volcanic edifice dictates that in this dimension, the limit cannot be linear. However, lack of data prevents a more accurate representation of this limit.
Figure 4.13

Theoretical Eruption Rate, $Q_m$ (m$^3$/s)

Figure 4.14

Theoretical Eruption Rate, $Q_m$ (m$^3$/s)
horizontal dimension. This is presented in Figure 4.14. It is clear from this diagram that over the initial stages of an eruption there is a general increase in the magnitude of the initial eruption rate with increasing dyke length. However, eruptions which occur at the same elevation, but at differing lateral distances from the central conduit (brought about by the asymmetry of the volcano) have, over similar time intervals, comparable eruption rates. Based on the available data, this demonstrates that vent elevation, and not lateral distance of the vent from the central conduit is the important factor which influences the magnitude of the effusion rate over the first few hours of an eruption. It also explains why some eruptions, such as 1981 (90/3h), which lasted less than 7 hours, can be constrained by limit line in Figure 4.14. Any increase in frictional resistance to flow which might be expected for a longer dyke is not observed during these early stages. This indicates that flow rates were sufficiently high to minimise any frictional effects. The magnitude of initial magmatic driving pressures must also have greatly exceeded the magnitude of the volcanic load acting to close the fissure.

**Eruptive Mechanisms**

One implication of the observation that initial eruption rate increases with decreasing vent elevation is that the force which effects the initial effusion rate may also be a function of altitude (Walker, 1974). One such force is hydrostatic pressure. For a column of magma standing in the central conduit of a volcano, the hydrostatic pressure at a given point within the column is determined by the weight of the overlying magma acting down on that point. The magnitude of this driving force increases for eruptions whose dykes plumb into the central conduit at the lowest level below the top of the column. For a free-standing, gas-free magma column of density $\rho_m$, the pressure $P_{hz}$ at any point within it is given by:

$$P_{hz} = \int_0^z \rho_m g dz$$

(4.4)

where $z$ is the height of the column above that point (Figure 4.10).
Hydrostatic pressure has been proposed as the predominant eruptive force on Etna by several authors (Walker, 1974; Wadge, 1977 and Lopes and Guest, 1982). Where observations have been possible, these confirm that magma is generally at high levels within the central conduit system before eruption, and that the magma level decreases rapidly upon eruption e.g. 1979 eruption (Guest et al., 1980a). A similar eruptive mechanism was proposed by Perret (1924) for the initial stages of the 1906 Vesuvian eruption.

Another altitude dependent pressure source of equal, or possibly greater short-term importance, is that due to volatile exsolution. It has long been realised that volatile exsolution is an important means for increasing the rate of magma emission (see Williams and McBirney, 1979 for a summary). Following a breach in the walls of the central conduit, a degassing column of magma initially in quasi-equilibrium would undergo rapid decompression as the magma level fell. This would have the immediate effect of pushing the nucleation level (point at which volatiles exsolve from the melt) to a deeper level in the volcanic system. As a result, the sudden exsolution of a large quantity of volatiles would accentuate the initial pressure by adding to the magma buoyancy force. The initial contribution of this pressure source to an eruption is dependent on the amount of depressurisation. This increases for dykes which tap into the central conduit at deep levels i.e. largest fall of the magma column. The involvement of this type of force in an eruption would manifest itself as an increase in the explosivity, either at the vent or through the summit craters, soon after the eruption commenced. For Etna, this type of eruptive phenomenon has been recorded during several eruptions, notably 1886 (Daubrée, 1886), 1911 (Platania, 1912), 1950-51 (Cumin, 1954), 1981 (Villari, 1983) and 1983 (Romano and Vaccaro, 1986). This eruptive mechanism, which compliments the hydrostatic force, was first proposed by Perret (1924) to explain the explosive phase of the Vesuvian eruption of 1906. More recently, it has been forwarded by Scandone and Malone (1985) to explain the 1980 eruptive activity of Mount St Helens volcano, Washington State. The equations which describe the role of exsolving volatiles in the ascent of basaltic magma are summarised by Wilson and Head (1981) and will not be elaborated on any further here.
A third force, whose magnitude is not constrained by vent elevation, is that due to overpressure within a dilated magma storage area. Following the failure of the walls enclosing the storage area, pressure is gradually reduced through the relaxation of the elastic strain stored by the enclosing rocks. Whilst this mechanism may be important in volcanic systems such as Hawaii, where a high level storage area is present (Machado, 1974; Wadge, 1981; Dvorak and Okamura, 1987), its importance as a prevailing force within the post-1750 volcanic system of Mount Etna must be limited. Though the central conduit system may undergo some dilation when it is filled with magma (Wadge, 1976), there is no evidence, as discussed in chapter 2, for a high level storage area in this period of magmatic output.

Conduit Parameters

In all the above instances, there is a basic assumption that magma pressure is the sole factor influencing the magma flow rate. However, there are other contributory factors, such as magma rheology and conduit parameters, which are potentially of equal, or greater importance. Taking the simplest case, the inter-relationship between these controlling factors can be approximately modelled in terms of the laminar flow of a viscous Newtonian fluid between parallel plates - as described by the Hagen-Poiseuille equation (Massey, 1983). For initial eruption conditions, this equation is of the form:

\[ Q_0 = \frac{\Delta P_0 h w^3}{12 \eta L} \]  

(4.5)

where \( Q_0 \) is the peak volume flow rate of a magma of viscosity \( \eta \), through a dyke of length \( L \), width \( w \) and height \( h \). \( \Delta P_0/L \) is the peak pressure gradient along the dyke between the central conduit and the eruptive vent. Of all the other parameters in the equation other than the driving pressure \( \Delta P \), the volume flow rate is most sensitive to the width of the conduit. This is because of the power nature of the proportional relationship. A 50-percent decrease in conduit width would decrease the flow rate by a factor of 8.
Where a fracture has a tectonic origin, the width is determined by the amount of stress to be relieved (Wilson and Head, 1981; Rubin, 1990), though it may be the situation that the pressure of magma within the conduit increases the width beyond that determined purely by stress release. The width of a magma-propagated, hydraulic fracture is determined by the amount of excess magmatic pressure relative to both the compressive force acting to close the fissure (Wadge, 1977). In a situation where magma is standing at a high level in the central conduit of a conical volcano, the width of any hydraulically induced fracture should be fairly uniform at all elevations provided that the volcanic pile consisted of rocks of uniform strength. With decreasing elevation below the summit, any mechanical advantage arising from an increase in magmatic pressure would be countered by an increase in the weight of the volcanic load acting to close the fissure.

It is highly probable that both hydraulic fracturing (Wadge, 1977) and tectonic stress release has been an important means for dyke injections on Etna (the latter, especially on the rift zones), but this is difficult to substantiate. It is presently impossible to directly measure the widths of subterranean Etnean dykes of the post-1750 period. However, some measure of representative widths can be obtained from prehistoric dykes exposed within the Valle del Bove (assuming that they were emplaced under similar conditions). Wadge (1977) quotes a dyke width range of 0.2 to 2 m, with a mean value of 1 m, though McGuire (1983) notes the presence of an exceptionally wide dyke (12 m) in a pyroclastic formation in the Trifoglietto II volcanic sequence.

In order to reduce the number of variables, the initial dyke width in the following discussion, unless stated otherwise, is assumed to be the same for all eruptions. It is further assumed that magmatic pressure is the sole factor influencing the real-time effusion rate.

4.6.2 Temporal Variation of Real-time Effusion Rates

Now that some information is available about the peak real-time effusion rates (Figure 4.13) and final eruption rates (Figure 4.11 and 4.12), it should be possible, with the aid of observations made in other sections, to gain some measure of the way in which the volcanic system has influenced the variation of real-time effusion rate
with time. The manner in which an eruption progresses following this initial high effusion rate phase is strongly dependent on the way in which the magnitude of the driving pressure varies with time relative to the influence of the stress fields, conduit parameters and magma rheology. To demonstrate this inter-relationship, when the magma pressure within the dyke begins to fall below the least compressive stress acting on the fissure (Equation 4.1), the width of the dyke reduces until the flow of magma can no longer be sustained. Further retardation of magma flow is induced by viscous frictional effects which increase as the magma flow rate decreases.

Whilst the magnitude of the real-time effusion rate may be closely related to the vent elevation over the first few hours of an eruption, it is the proximity of the vent to the pressure source in the central conduit, which primarily controls the period for which a western sector effusion can be sustained. Below 1,900 m asl, though the magmatic pressure at low elevation is initially high and effects high initial effusion rates, it rapidly decreases to levels below the volcanic load hence, the eruptive fissure closes. Therefore, at low elevations and for long dykes, there is a tendency for short duration and relatively high eruption rate eruptions such as 1981 (90) (Figures 4.12 and Figures 4.9).

At progressively higher elevations, whilst initial magmatic pressures are lower, longer eruption durations are possible. The increase in the magnitude of eruption duration towards 1,900 m asl (Figure 4.9) with progressively shorter dykes must indicate that the magmatic forces can exceed the compressive forces acting to close the fissure for longer. This allows for the eruption of greater volumes of lava. For example, the 1879 (37) eruption occurred from a vent at 1,850 m asl, at a distance of 7.6 km from the central conduit system and erupted some $30 \times 10^6$ m$^3$ of lava in 10 days. Whilst the initial effusion rate of this particular eruption was probably smaller than that of the 1981 (90), the dyke remained open for a longer duration (Figure 4.9) thus allowing the extrusion of a greater volume of lava (Figure 4.7). The final eruption rate was therefore smaller (Figure 4.11).

The two eruptions which occurred above 1,900 m asl, those of 1843 (32) and 1910 (46), depart from the above patterns (Figure 4.9 and Figure 4.12). The relatively short duration in which the 1843 eruption extruded a relatively large volume of lava indicates that high effusion rates persisted throughout this eruption. Whilst the initial
driving pressures may have been smaller than for 1981 (86) (as inferred from Figure 4.12), the rate at which the excess magmatic pressure was released must have been higher for this eruption. As a consequence, the material stored at high levels in the volcanic system was evacuated a lot faster during this eruption compared to the eruptions at lower elevation, ultimately leading to an earlier end to the eruption. This resulted in a high overall eruption rate.

A simple way of sustaining a protracted period of high effusion rate without altering the magnitude of the initial driving pressure would be either to increase the initial width of the dyke (this need not be a large change as argued in section 4.6.1) or, if the dyke width closes with time, to maintain an initial dyke width comparative to that of the other eruptions, but reduce the rate at which the dyke closed. Two factors probably contributed to the increased dyke width:

1. The proximity of the vent to the pressure source within the central conduit i.e. a short feeder dyke.
2. A relatively smaller volcanic load acting on the eruptive fissure. This resulted from a change in the morphology of the volcano on the western flank circa 1,900 m asl, from the 'shield' structure of the lower slopes, to the steeper slopes of the Mongibello cone. This would have invoked smaller horizontal stresses (σ₂, σ₃) acting against the column. The 1843 (32) vent occurred on the lower slopes of the Mongibello cone, immediately above the break of slope.

The 1910 (43) eruption occurred at a comparable elevation to that of 1843 (32) but in this instance, the vent was at a greater distance from the central conduit. Whilst a short duration might be expected, the 1910 (43) lasted 26 days. This is the longest duration seen for an eruption on the western sector within this period of output. Whilst the 1910 (43) eruption may have commenced with a rate approximately comparable to 1843 (32), and eventually erupted a volume of lava greater than most eruptions at a lower elevation, it took over twice the duration to accomplish this feat. The real-time effusion rate during this eruption, whilst smaller than for most eruptions at lower elevation, might be more typical of 'normal' eruptive conditions at this elevation, but this is disguised by the fact that the magnitude of σ₃ acting on the fissure was smaller. This allowed the eruption to continue to a lower effusion rate, hence the longer duration.
Above 2,000 m asl, it is significant that the magnitude of magmatic pressures at the start of an eruption decrease towards the summit. Initial real-time effusion rates for these eruptions though high were, over a similar time period, probably smaller than those for eruptions at lower elevation. The fact that these eruptions are predominantly characterised by high eruption rates (Figure 4.12) and low volumes (Figure 4.7) is a reflection of the short time period for which the magmatic pressure exceeded the volcanic load. An example of how the effusion rate would vary over an extended period of time (i.e. situation where \( \sigma_3 \) is smaller) is presented by the 1947 (51) datum. Though the duration of this eruption is similar to that of eruptions occurring at the same lateral distance from the central conduit below 2,000 m asl, the eruption rate is much smaller. This is indicative of lower real-time effusion rates during the course of the eruption.
4.7 Lava Flows: Emplacement and Planimetric Evolution

As part of a research paper looking into the hazard posed by lava flows, Walker (1974) determined that, on the whole, the lengths of Etnean lava flows tend to increase as the elevation of the vent decreases. A similar conclusion was reached by Lopes and Guest (1982) in a subsequent study of Etnean lava flows which utilised a more comprehensive data set. This is illustrated in Figure 4.15, the limit A, defining the longest lava flow extruded from a vent at a given elevation.

Figure 4.15 Elevation of eruptive vents (m asl) plotted against flow length (km) based on Etnean eruptions covering the whole historical record. The solid line (i) constrains the longest flow extruded from vents at given elevations (after Lopes and Guest, 1982).

Figure 4.15 is redrawn in Figure 4.16, using data restricted to the post-1750 western sector only. Perhaps the most important inference which can be drawn from Figure 4.16 is that the observed variation of flow length with vent elevation differs from that seen by Walker (1974) and Lopes and Guest (1982) (Figure 4.15). Above 2,000 m asl, lava flow lengths are constrained by the limit (A) of Lopes and Guest.
Figure 4.16  Elevation of post-1750, western sector eruptive vents (m asl) plotted against flow length (km). Only the longest flow extruded during a single eruptive episode is considered. The solid limit (A) is that of Lopes and Guest (1982), whilst the maximum lengths of lava flows extruded below 2,000 m asl are constrained by a new (dashed line) limit (B).

Figure 4.17  A plot of distance between the vent and the central conduit versus the final length achieved by post-1750 lava flows of the western sector. The dashed limit (I) has been transposed from Figure 4.16. The Solid limit line (II) constrains the longest flow lengths attained by western sector lava flows which were erupted from vents below 2,000 m asl but at differing lateral distances from the central conduit.

Since (II) is a straight line in Figure 4.7, the non-constant slope of the volcano dictates that this limit cannot be represented by a straight line on an elevation diagram (B). Insufficient data prevents a more accurate representation of the limit on this diagram: this is a problem which arises in most of the following lateral distance - vent elevation diagrams.
(1982) but, below 2,000 m asl, whilst lava flows at a given elevation can have a range of lengths, there is an overall decrease in the maximum flow length observed at a given elevation (constrained by the new limit (B) in Figure 4.16). This new observation has implications for the morphological construction of the volcanic edifice. If the lengths of individual flow units did increase with decreasing vent elevation, there would be a tendency for the edifice to develop shallow, 'shield-like' slopes and the extent of the volcano would be immense. This is not the case for Mount Etna in the post-1750 period where the construction of the volcano has mainly been limited to above 500 m asl.

4.7.1 Below 2,000 m asl

On the basis of the relationship established between the volume of Type A lava flows and their final lengths in section 3.4.2, it is not surprising that the lengths of western sector lava flows, in keeping with that seen for volume (Figures 4.6 and 4.7), show a stronger relationship with the inferred dyke length (Figure 4.17) compared to vent elevation (Figure 4.16). Below 2,000 m asl, the longest flow at any elevation is associated with the vent which is nearest the central conduit. As the length of the feeder dyke connecting the eruptive vent to the central conduit increases, so the longest final flow length decreases. According to the gradient of limit line (II), this occurs at a rate of some 955 m per km of dyke. In the worst possible case, where a western sector flank eruption extruded the maximum eruptible fraction of $48 \times 10^6$ m$^3$, the resulting lava flow would have a length between 13 and 14 km.

Not all the data which define the gradient of I on the western sector volume - inferred dyke length diagram (Figure 4.7) attained the maximum flow length equivalent to their lateral distance from the central conduit in Figure 4.17. For instance, considering its volume, the 1832 (30) lava flow in Figure 4.17 is shorter than might be expected. The main flow, which was erupted onto the shorter western slopes, encountered a topographic obstacle on the outskirts of Brönte which delineates the volcanic limit on this flank of the volcano. A supply of lava was maintained to the flow front even after it had encountered this natural limit, but the new lava was accommodated by widening at the flow front, rather than by further lengthening. If the emplacement of the 1832 lava flow had not been impeded, then from the
relationship in Figure 4.17, it had the potential of reaching a distance of some 12 km from the eruptive vent. From a hazard point of view, this would have taken the lava flow through the town of Bronte on the western slopes of the volcano.

4.7.2 Above 2,000 m asl

Above 2,000 m asl, the absence of any systematic pattern in the variation of volume with inferred dyke length (Figure 4.7) manifests itself in the final lengths of the lava flows (Figures 4.17). At best, it is only possible to constrain the longest flow seen at each elevation between 2,000 m asl and the summit. This is defined by the limit I, of Lopes and Guest (1982) which in Figure 4.17, has been transposed to the horizontal dimension. Since A in Figure 4.16 is a straight line, the non-constant slope of the volcano dictates that this limit cannot be represented by a straight line on a lateral distance diagram (Figure 4.17). Insufficient data prevents a more accurate representation of the limit on these diagrams.

The fact that only Type A flows were produced during western sector eruptions is an indication that the rate of effusion at the vent during the course of the eruption was sufficient to supply a propagating flow front throughout the eruption. It does not follow from the fact that the longest flows are associated with the largest volumes that these flows were emplaced and supplied at the highest effusion rates. To the contrary, in the post-1750 period, the dependence of effusion rates on the elevation of the vent is reflected in the general emplacement rate of the lava flows, with the highest rates of emplacement and the most hazardous lava flows emerging from those vents at lowest elevation. For example, the principal lava flow of the 1981 (90) eruption, which emerged at 1,400 m asl, travelled 8.25 km in less than 2 days. Though the 1910 (43) lava flow, which emerged from a vent at 2,000 m asl, was more voluminous and some 2.5 km longer than the former lava flow, it did not reach its maximum length until 14 days after the start of the eruption.
4.8 Discussion

The volumes and durations of eruptions on the western sector of Etna in the post-1750 period of magmatic output are considered to have been determined by the duration for which the magnitude of a temporally decaying magmatic driving pressure ($P_m$) can exceed the compressive force ($\sigma_y$) acting to close the eruptive fissure.

In the present period, the most voluminous eruptions have occurred from vents close to the central conduit and within the 1,700 to 2,000 m asl elevation interval. According to the hydraulic fracturing model of Wadge (1977), these favourable eruption conditions at mid-elevations, which allow for the most efficient exploitation of the stored volume, occur because the largest excess of magma pressure within the central conduit relative to the volcanic load at the start of an eruption corresponds with this elevational interval. It must follow that after the initial breach in the central conduit, the decaying pressure force should, under normal circumstances, be able to exceed the compressive forces acting to close the fissure for longest at this elevation. Short duration eruptions occurring in this elevation range, such as the 1843 (32) eruption, do not reflect smaller initial magma pressures but rather, a quicker release of the excess pressure. Even the largest of eruptions on the western sector terminate once that volume stored at a high level has been erupted. This occurs because the pressure exerted by any ascending magma is insufficient to overcome the compressive force acting to close the fissure.

At progressively lower elevations, any mechanical advantage arising from an increase in magmatic driving pressures is largely counteracted by an increase in the volcanic load acting on the fissure source. The shallower slopes of the lower volcano also result in an increase in the length of dyke required to connect the eruptive vent to the central conduit. This has the effect of increasing the mechanical resistance to flow as pressures decrease during eruptions. Therefore, at progressively lower elevations, though initial real-time effusion rates increase, the duration of eruptions decrease. Whilst the magnitude of the volcanic load decreases at increasing elevations above 2,000 m asl, this occurs at a rate which is slower than that at which the magnitude of magmatic pressures decrease. The magnitude of individual eruptions are therefore small because the difference between the excess pressure and the compressive force acting to close the fissure are also small.
CHAPTER 5

Post-1750 Eruptive Activity and Lava Flow Emplacement on the Eastern Sector

5.1 Introduction

This chapter, which deals with eruptive activity and lava flow emplacement on the eastern sector, represents the second part of the examination into the post-1750 volcanic system of Mount Etna, Sicily. It has already been established in section 4.3 that eastern sector eruptions have the potential of being more voluminous, of longer duration and of lower eruption rate compared to western sector eruptions. In addition, there is a tendency for lava flows to develop Type B flow field morphologies. An attempt is made to identify those conditions within the volcanic system which allow for the strong dichotomous sectorial control of eruptive styles in the post-1750 period of magmatic output. This entails a discussion of potential magma storage areas, the magma transportation system, and eruptive mechanisms.

5.2 Method of Analysis

With the exception of the spatial distribution component (which is not repeated here), the analysis of the post-1750 eastern sector data followed the same approach as that applied to the western sector data. This is described in chapter 4, section 4.2. In order to simplify the presentation of the information contained within the diagrams of this chapter, the data are grouped according to the predominant lava flow morphology (Type A flow units and Type B flow fields) which were produced during the eruptions. For comparative purposes, the western sector limits are included on all the relevant diagrams.
5.3 Erupted Volume of Lava

Whilst eastern sector eruptions which have occurred at the same elevation as western sector eruptions are, on the whole, more voluminous (Figure 5.1), the same distribution pattern emerges, with the most voluminous eruptions occurring towards mid-elevations on the volcano, peaking somewhere between 1,700 and 2,300 m asl. Eruptions at the same elevation may have a range of values. Limits are added to this diagram (B and B') which constrain the maximum volume erupted on the eastern sector in the post-1750 period.

![Diagram showing the distribution of erupted volume with vent elevation for eastern sector eruptions of the post-1750 period. Limits (B and B') constrain the maximum volume erupted whilst limit A, is that for the western sector (Figure 4.6). For eruptions where lava effusion occurred from more than one point along the fissure, the cumulative volume is presented. This value is plotted against the eruptive vent which was active at the lowest elevation. Horizontal lines are error bars. Data points on this and all subsequent diagrams which are accompanied by a black dot are part of the 'Valle del Bove' group of eruptions.]

5.3.1 Below 2,000 m asl

For those eruptions which occurred below 2,000 m asl on the eastern sector, the relationship between the volume of lava extruded and the lateral distance between
Figure 5.2

A. The horizontal distance of an eruptive vent from the central conduit (km) plotted against the extruded volume of lava (x 10⁴ m³) for post-1750 eastern sector eruptions occurring below 2,000 m asl. The cumulative erupted volume for each eruption is presented against the most distant eruptive vent. The relationship described by the regression line II, and is of the form V (x 10⁴) = -14.5Lₖ(km) + 183. For the 10 data, r = -0.885 and Π < 0.01. (Omitted from the calculation were the 'Valle del Bove' group of eruptions (black dot), and the 1811 (27) eruption). Regression line I, is that calculated in chapter 4, section 4.4 for western sector eruptions.

B. This diagram is the equivalent of A but restricted to post-1750 eastern sector eruptions occurring above 2,000 m asl. The regression lines are those presented in A.
the eruptive vent and the central conduit is shown in Figure 5.2A. At any given
elevation (Figure 5.1), the volume of lava extruded at a given lateral distance can take
on a range of values (Figure 5.3). However, for the majority of eruptions, irrespective
whether they occur at the same elevation, such as 1892 (40) and 1923 (46), or at
differing elevation, such as 1852-53 (33) and 1892 (40), there is a maximum volume
which can be extruded at any given value of $L_i$. This single, well-defined relationship
is represented by the negative regression line (II) in Figure 5.2A. The form of this
relationship is such that the most voluminous eruptions are associated with the vents
nearest the central conduit: at increasing values of $L_i$ there is a decrease in the
magnitude of the largest volume which can be extruded. Lateral distance, or dyke
length, therefore has a greater influence on the maximum volume which can be
extruded at a given elevation.

Only the 1811 (27) eruption, and a series of eruptions (1971 (67), 1978c (85),
and 1979 (87)) which occurred on an orthogonal set of northeast and southeast
trending fissures at the head of the Valle del Bove (Figure 5.3) failed to extrude the
maximum volume of lava equivalent to their lateral distance from the central conduit
(Figure 5.2A). Clearly, vent elevation and dyke length did not exert as much control
on the quantity of lava extruded during these eruptions. The volcanic system may
have only been partially-charged within the volcanic edifice before these eruptions.
Taking a specific example, the chasm was empty down to a depth of over 1 km prior
to the 1971 (67) eruption.

Comparing sectors, in the proximity of the central conduit, the largest volume
which can be extruded below 2,000 m asl on the eastern sector clearly exceeds that
erupted from a dyke of comparable length on the western sector (Figure 5.2A).
However, at increasing dyke lengths, because the rate at which magma is retained
within the volcanic system per unit length of dyke is much higher on the eastern
sector (steeper gradient of II), at 11 km distance from the central conduit, there is
little difference between the volume erupted in both sectors. The intersection of the
steeper eastern sector regression line, II, with the lateral distance axis occurs at 12.9
km from the central conduit system. This lateral distance corresponds well with the
horizontal distance measured for the 1883 (38) eruption on the southern flank. During
this eruption a small volume of magma, about 50,000 m$^3$, was effused at a distance
of 11.5 km. Though this value of $L_4$ is some 1.4 km shorter than that outlined above, there is evidence that the actual fissure was almost 2 km longer (Silvestri, 1883). In terms of the volcanic system, this point may be the limit beyond which no magma can be extruded from the system.

The largest fraction of the available magma supply which can be erupted by eastern sector flank eruptions is $140 \times 10^6$ m$^3$ (see footnote 1). This is some $92 \times 10^6$ m$^3$ more than could be erupted by western sector flank eruptions. In addition to the central conduit storage area, some tens of intra-volcano dykes would be required to store such a large volume of magma at high levels within the edifice. Recent periods of geophysical and geodetic surveillance have found no evidence for a large storage area at a high level on Etna (Murray and Guest 1982). It seems reasonable therefore, that most of the excess volume ascends into the volcanic edifice after the magma stored at high level has been erupted. Intensive sampling during recent eruptions (Armienti et al. 1984; 1989) has revealed compositional variations during the course of individual high volume eruptions. In general, the first products to be erupted have slightly evolved compositions and give way during the course of the eruption to more primitive compositions. This is probably indicative of new magma rising into the volcano during the course of these high volume eruptions. If this interpretation of the petrological information is correct, then it follows that the steeper gradient of the eastern sector regression line cannot be a simple function of the material retained at high levels within the volcanic system, but must relate to the volume retained within the volcanic system as a whole (to include the deeper levels).

---

1The morphology of the volcano at 2,000 m asl is such that the minimum distance between a potential eruption site on the eastern sector of the volcano and the axis of the central conduit is 2.9 km. It follows that the largest fraction of the available magma supply which can be extruded by eastern sector flank eruptions is that point on the regression line II at a lateral distance of 2.9 km from the central conduit. This value is $140 \times 10^6$ m$^3$. This value differs from that published by Hughes et al. (1990). The published volume of $170 \times 10^6$ m$^3$ was deduced by incorrectly extrapolating the regression line back to the volume axis. The basic calculation was also incorrect because it included two data points from eruptions which had occurred above 2,000 m asl, those of 1950-51 (54) and 1983 (92).
5.3.2 Above 2,000 m asl

As seen for the equivalent eruptions on the western sector, those eruptions occurring above 2,000 m asl on the eastern sector did not extrude as much lava as those eruptions below 2,000 m asl at equivalent lateral distances (Figure 5.2B). It is difficult to find any systematic pattern within the distribution of the whole data in Figure 5.2B. However, looking at the data over smaller elevation intervals (Figure 5.1), the following points become apparent:

1. At a given elevation above 2,000 m asl, eruptions occurring from vents at differing values of $L_a$ can extrude a range of lava volumes but, for that elevation, there does appear to be a maximum volume which can be extruded at a given lateral distance. This maximum may decrease with increasing lateral distance. For example, at 2,250 m asl, the maximum volume which can be extruded at a given lateral distance is defined by the 1950-51 (54) and 1983 (92) eruptions.

2. The same pattern may hold at progressively higher elevations (approaching the summit), but it is more difficult to determine the attitude of the upper limit (maxima) because of insufficient data. Nevertheless, at equivalent distances from the central conduit (Figure 5.2B) there is a general decrease in the magnitude of the volume maxima at progressively higher elevations above 2,000 m asl. Until this diagram can be better calibrated with more eruption data, further inferences must be considered tentative. However, it is tempting to speculate that the interrelationship between the maximum volume and inferred dyke length at elevational intervals between 2,000 m asl and the summit might take the form of a series of linear relationships parallel to the dashed line (II), and spaced at intervals between (II) and the origin (the summit), each line representing the maximum volume-inferred dyke length relationship for a given vent elevation.

The fact that there is a progressive decrease in the maximum volume which can be erupted at a given lateral distance with increasing elevation must indicate that between 2,000 m asl and the summit (the origin), there is a decrease in the volume fraction which can be extruded by flank eruptions, from approaching $140 \times 10^6$ m$^3$ at 2,000 m asl, towards smaller fractions at the summit. A few of these eruptions, such as the 1764-65 (19) eruption, extruded a volume of lava in excess of the $48 (\pm 7) \times 10^6$ m$^3$ maximum which could be erupted on the western sector (Figure 5.2B). A
supplementary magma supply, in addition to that which could be stored at a high level, was accessed by these eruptions. Only those eruptions near the summit could have been supplied by magma previously stored at high levels (those data points which lie to the left of the western sector limit, (I), in Figure 5.2B).

Figure 5.3 Spatial distribution of vents towards the head of the Valle del Bove. These eruptions, restricted to an orthogonal set of fissures trending northeast and southeast, form a separate structural grouping distinct from western and eastern flank eruptions (see text for discussion).

Whilst vent elevation and dyke length might exert an underlying control on the magnitude of the volume maximum, these factors alone cannot explain why, for a given vent elevation, most eruptions extrude a volume of lava which is smaller than the maximum at a given lateral distance from the central conduit. These variable volume eruptions have been particularly prevalent around the summit, and along the western edge of the Valle del Bove, in association with a localised orthogonal set of
fissures trending northeast and southeast (Figure 5.3). The southeast set of fractures have been intermittently active since at least 1792, but the northeast set only became established during the 1971 eruption. With these eruptions, effusion may take place from several widely spaced localities, from dykes trending in both directions. This was observed during the 1978b (81), 1978c (85) and 1979 (87) eruptions (Guest et al. 1980a). This particular collection of eruptions will herein be known as the 'Valle del Bove' group.

5.4 Eruption Duration

The duration of eastern sector eruptions generally exceed those of western sector eruptions at the same elevation (Figure 5.4) though, as seen with the volumetric diagram, some of the eruptions which occurred from the orthogonal set of fissures at the head of the Valle del Bove are an exception. However, in contrast to the pattern seen for the western sector, the magnitude of eastern sector eruption durations do not peak at 2,000 m asl but continue to increase above this elevation. The longest duration eruption of the post-1750 period was the 18 month eruption of 1764-65 (19) which occurred from a vent at approximately 2,600 m asl.

A striking feature of Figure 5.4 is that below 2,000 m asl, there is a marked variation in the way in which eruption duration varies with increasing vent elevation and flow field morphology. Eruptions which produced Type A lava flow morphologies have durations which progressively increase with elevation, from the 2 day eruption of 1883 (38) at 1,040 m asl to the 32 day eruption of 1923 (46) at 1,800 m asl. At elevations below 2,000 m asl the longest durations which are associated with these particular eruptions are constrained by limit line B. Above 1,700 m asl, and coinciding with a change in the predominant flow morphology from Type A to Type B, individual eruption durations increase dramatically over a small elevational interval, from 32 days at 1,800 m asl (1923 (45)) to the 380 day eruption of 1792-93 (24) at 1,900 m asl. No Type B flow fields presently exist below 1,700 m asl. This increase in duration also coincides with an increase in the erupted volume beyond the

---

2The vent area of the 1764-65 (19) eruption is located on the NE rift, but its exact position is not known because it has been buried beneath younger lavas. The vent elevation of 2,600 m asl quoted here may be too low.
Figure 5.4 Elevation of post-1750 eruptive vents (m asl) on the eastern sector versus eruption duration (days). The duration value used here is the eruption duration and not the cumulative duration based on the period of activity at each separate eruptive vent. Where effusive activity occurred from several locations, and at differing elevation on the same fissure system, the duration value is presented against the lowermost active vent. The maximum duration of eruptions which occurred below 2,000 m asl and produced Type A flow morphologies, are constrained by limit line B, whilst the maximum duration of those eruptions which produced Type B flow field morphologies, and whose extruded volumes exceed the volume maximum (48 x 10⁶ m³) of western sector flank eruptions (this includes 3 eruption data which produced Type A flow morphologies) are constrained by limit line C. Data accompanied by an asterisk denote eruptions which produced flow fields predominated by pahoehoe surface textures. All other eruptions produced flow fields of aa surface texture. The duration of the 1764-65 (19) datum is expressed in months (18m).

maximum which could be erupted by a western sector flank eruption (Figure 5.2A). Though eruptions at the same elevation can have a range of durations - the 32 day 1923 (46) eruption at 1,800 m asl is half an order of magnitude smaller than the 172 day 1892 (40) eruption at the same elevation - limit line C constrains the maximum duration of eruptions which extruded more than 48 (± 7) x 10⁶ m³ of lava. Vent
Figure 5.5
A  Horizontal distance of the eruptive vent from the central conduit (km) versus eruption duration (days) for post-1750, eastern sector eruptions occurring below 2,000 m asl. Duration values are presented against the most distant eruptive vent. In this diagram, the variation of eruption duration with lateral distance is not uniform and a marked inflexion occurs in the relationship at a distance of between 7.5 and 9.0 km from the central conduit. The maximum duration of those eruptions which produced aa-textured, Type A morphology flow fields; and whose data define the maximum volume limit (II) in Figure 5.2A, are constrained by limit line (II). These data only occur at lateral distances greater than 7.5 km from the central conduit. For those eruptions which produced aa-textured, Type B morphology flow fields, whose data define the maximum volume limit (II) in Figure 5.2A, and whose volume exceed the volume maximum ($48 \times 10^6$ m$^3$) of western sector flank eruptions (this includes 3 eruption data which produced Type A flow morphologies), these data are constrained by the limit (III). This relationship is of the form $t$ (days) $= -64.6L$ (km) + 569. For the 7 data, the regression coefficient $r = -0.989$ and $P < 0.5\%$.

B  This diagram is the equivalent of A, but restricted to those post-1750, eastern sector eruptions occurring above 2,000 m asl. The nomenclature used in this diagram is the same as for Figure 5.4, and the regression lines are the same as those presented in A above.
Figure 5.5

**Figure A**

- **Eruption Duration (Days)**
- **Distance of Vent from Central Conduit (km)**

**Key**
- < 2,000 m. asl
- Eastern Sector
- ■ Type A
- ■ Type B

**Figure B**

- **Eruption Duration (Days)**
- **Distance of Vent from Central Conduit (km)**

**Key**
- > 2,000 m. asl
- Eastern Sector
- ■ Type A
- ■ Type B
elevation therefore appears to have had little control on those factors which influence the duration of individual eruptions.

5.4.1 Below 2,000 m asl

The relationship between the duration of eastern sector flank eruptions occurring below 2,000 m asl and the lateral distance of their eruptive vents from the central conduit \(L_j\) is shown in Figure 5.5A. In general, the longest eruption duration associated with a given lateral distance increases as \(L_j\) decreases. However, this distribution is not uniform. For those eruptions with vents within 7.5 km of the central conduit there is a marked, order of magnitude increase in the maximum eruption duration over a small lateral distance range. This contrasts sharply with those eruptions which occurred at lateral distances in excess of 7.5 km whose durations do not exceed 32 days. With the exception of the 1978c (85) and 1979 (87) eruptions of the 'Valle del Bove' group, the remainder of these eastern sector eruptions lasted longer than western sector eruptions supplied by feeder dykes of similar length.

Further information can be drawn from Figure 5.5A when it is examined alongside the equivalent volume diagram (Figure 5.2A), and flow field morphology is taken into consideration. For those eruptions which define the volume maximum in Figure 5.2A, and produced Type A morphology flow fields (data restricted to \(L_j > 7.5\) km), it is conspicuous that there is an underlying increase in the maximum duration of individual eruptions as the lateral distance decreases (limit line II). This coincides with an increase in the volume of lava erupted (Figure 5.2A). When the volcano is interpreted in terms of an elastic solid, an implication of this observation is that as the eruptive vent gets nearer the central conduit, magmatic pressures within these eastern sector dykes can exceed the compressive stresses acting to close the eruptive fracture for a longer period. However, the variation of eruption duration is not uniform, and for a given lateral distance, eruptions can have a range of durations. Instances occur where an eruption with a vent near the central conduit has a shorter duration than a more distant eruption. For example, the 11 day, 1911 (44) eruption occurred at a lateral distance of 8.9 km from the central conduit compared to the 18 day, 1928 (48) eruption at lateral distance of 10.15 km. On the basis of proportionality, the differences in the duration of these respective eruptions is not
reflected by a commensurate increase and decrease in the volume of lava they erupted; the 1911 (44) eruption extruded some $55 \times 10^6$ m$^3$ compared to the $38 \times 10^6$ m$^3$ of the 1928 (48) eruption. As seen for the western sector eruptions, variation in the extrusion efficiency of a given volume of lava can only be achieved by changing the temporal rate at which the magmatic pressure decreases. This can be effected in a variety of ways, including changing the width of the feeder dyke. This point will be discussed further in section 5.5.

On the basis of their short duration (< 32 days) and small volume (< $60 \times 10^6$ m$^3$), it must follow that eastern sector eruptions with vents at lateral distances in excess of 7.5 km from the central conduit, were primarily controlled by high level eruptive forces. However, that these particular eastern sector eruptions are more voluminous (Figure 5.2A) and lasted longer (Figure 5.5A) than western sector at equivalent distances from the central conduit, implies that differences must exist in the way in which each sector responds to eruptive events. The larger erupted volumes are a consequence of the longer eruption durations. Longer eruption durations indicate that the feeder dykes remained open for a longer time period on the eastern sector compared to the western sector. In order to achieve this, either the magmatic pressure $P_m$, within the feeder dykes of the eastern sector eruptions is greater than that of the western sector eruptions, or the least compressive stress $\sigma_3$, acting across the feeder dykes is smaller (equation 4.1).

It is difficult to envisage a means by which the 'bulk' of the eastern sector could accommodate higher magmatic pressures. The higher degree of shallow seismicity (<5 km) on the eastern sector compared to the western sector indicates that the former has a reduced inherent strength and cannot accumulate a great amount of strain (Gresta et al. 1990). Supplementary evidence for the lower strength and instability of this unsupported sector of the volcano is provided by major fault scarps along the coastline which downthrow to the sea, and by the Valle del Bove depression, which has been interpreted to have formed during a series of gravitational sliding episodes (Guest et al. 1984a). Under these conditions, the eastern sector of the volcano would rupture at lower magmatic pressures.

It is more fruitful to explain the longer eruption durations as resulting from a smaller least compressional stress $\sigma_3$, acting to close the feeder dyke. The distribution
of volcanic load on the eastern sector is not uniform due to the presence of the Valle del Bove depression. Therefore, at all elevations and lateral distances on the eastern sector, the value of $\sigma_3$ acting on feeder dykes which propagate beneath, or in the proximity of the Valle del Bove, would be smaller than the equivalent value of $\sigma_3$ acting on western sector feeder dykes. Under these conditions, eruptions can be sustained for a longer period at a lower magmatic driving pressure, thus allowing more voluminous eruptions. Even so, for eruptive vents at lateral distances beyond 7.5 km, the short duration of the ensuing eruptions (<32 days) suggest that the feeder dykes close once that volume stored at high levels has been erupted.

The critical lateral distance from the central conduit within which the maximum volume of lava extruded by eastern sector eruptions exceeds the $48 \times 10^6$ m$^3$ maximum of western sector flank eruptions, is equal to 9.0 km (Figure 5.2A). The duration of the 1928 (48) and 1886 (39) eastern sector eruptions which straddle this important volumetric limit for western sector flank eruptions are 18 and 19 days respectively. This duration range is consistent with the 26 day limiting eruption duration established from the western sector data (Figure 4.9). Within 9.0 km of the central conduit, whereas the maximum volume which can be extruded at a given lateral distance increases uniformly as $L_j$ decreases (Figure 5.2A) there is, in contrast, a sudden, order of magnitude increase in the maximum eruption duration (Figure 5.5A). This relative difference is accompanied by a change in the predominant flow field morphology that characterise the eruptions which define the maximum. Type A flow morphologies predominate in that interval between 7.5 and 9.0 km whereas within 7.5 km, Type B flow field morphologies are dominant.

On first examination, there appears to be little correlation between the way in which these two eruption parameters vary against the inferred length of dyke at values of $L_j$ less than 9.0 km. However, when the data are grouped according to the predominant surface texture that characterise their lava flows/flow fields, a more apparent relationship is observed. Concentrating on those eastern sector eruptions which: 1. had vents within the above lateral distance range; 2. produced aa-textured lava flow fields, and 3. whose data define the maximum volume limit of II in Figure 5.2A, it is conspicuous that there is a well-defined maximum to the value of the eruption duration at a given value of $L_j$. The form of this relationship is inverse: there
is a decrease in the magnitude of the maximum duration at a rate of some 93 minutes per metre of dyke (gradient of \( \text{III} \) in Figure 5.5A). This coincides with a progressive, uniform decrease in the volume of lava they erupted (Figure 5.2A). As seen with the equivalent volume diagram (Figure 5.2A), the 'Valle del Bove' group of eruptions (1971 (67), 1978c (84) and 1979 (87)) and the 1811 (27) eruption did not persist to the maximum duration equivalent to their lateral distance (Figure 5.5A).

Several interesting inferences can be drawn from a comparison of \( \text{II} \) on Figure 5.2A and \( \text{III} \) on Figure 5.5A. Firstly, the shortest time period in which the largest fraction of the available magma volume (140 x 10^6 m^3) can be extruded by an aa producing flank eruption on the eastern sector is 381 days (value of \( \text{III} \) at 2.9 km). A second inference which can be drawn from the sudden increase in eruption durations relative to the volume extruded is that in the interval between 18 and 32 days, there is a temporal decrease in the efficiency at which the volcano erupts lava. This must reflect a progressive decrease in the magnitude of the magmatic forces driving the eruption. In addition, it follows from the volumetric arguments presented in section 5.3 that any material erupted after 32 days could not have been stored at high levels, and must have risen into the volcano during the course of the eruption. This interpretation of the data gains further credence when it is considered that the volume of lava extruded by those eruptions occurring in the transition region between 7.5 and 9.0 km (55 - 65 x 10^6 m^3) is similar to the quantity of lava which was extruded by the 1983 eruption (Frazzetta and Romano, 1984) before more primitive magmas were detected at the eruptive vent (Armienti et al. 1984). Whatever the cause of the temporal decrease in the magnitude of the magmatic pressures, after 32 days, the principal eruptive forces could not have been located within the volcanic pile.

The fact that eastern sector eruptions occurring below 2,000 m asl, with feeder dykes less than 7.5 km in length, are sustained at lower magmatic pressures indicates that the eruptive fissures remain open after the initial stage. This was first noted by Wadge (1981). The least compressive stress, \( \sigma_3 \), cannot therefore be an important factor influencing the duration of these eruptions. Relatively dilate eruptive fissures are easily envisaged in this volcano-tectonic environment, considering the gravitational instability of the eastern flanks. It follows from equation 4.1 that the predominant factor that determines the duration of an eruption must be the magma pressure, the
eruption terminating when the flow rate through the dyke (as controlled by $P_m$) is insufficient to overcome the rate at which the magma is cooling within it. The decrease in eastern sector eruption durations at increasing lateral distances along III in Figure 5.5A indicates that as magma pressures decay, the ability of the eruptive forces in sustaining eruptions through open fissures decreases.

Of all the eruptions whose data define II in Figure 5.2A, only the 1792-93 (24) eruption produced a pahoehoe-textured flow field. On Figure 5.5A, the duration of this particular eruption is a factor of 2 greater than the other eastern sector eruptions with vents at the same distance from the central conduit, and yet, there was no proportional increase in the volume erupted (Figure 5.2A). In effect, this eruption only extruded half the volume of the other eruptions over the same time period. Eruptive conditions must have differed for the 1792-93 (24) flank eruption. It can be inferred that the magmatic pressure driving this eruption was smaller, and underwent a more uniform temporal variation. The latter point was probably a significant factor in dictating the duration. The fact that this eruption lasted so long clearly indicates that the magma flow rate, though small, was still sufficient to exceed the rate of magma cooling within the dyke for a protracted period. Though an isolated example, this eruption demonstrates that under certain conditions, protracted eruptions can be sustained on the flanks.

5.4.2 Above 2,000 m asl

Most of the effusive eruptions which occurred above 2,000 m asl on the eastern sector do not exceed the duration of those eruptions below 2,000 m asl with vents at the same distance from the central conduit (Figure 5.5B - limit III). Of all the flank eruption data presented, only the 18 month eruption of 1764-65 (19) exceeds this limit. Though not shown of Figure 5.5B, a few terminal eruptions have lasted durations in excess of limit (C). For example, the northeast crater was in a state of continuous effusive activity between 1966 and 1971 (Guest, 1982).

Closer examination of the data in Figure 5.5B does not reveal any simple relationship between lateral distance and eruption duration. The following points of detail emerge from a comparison of Figure 5.4 and 5.5B.
1. At a pre-determined elevation, eruptions occurring from vents at differing lateral distances from the central conduit can have a range of durations which appear randomly related. This is well-demonstrated by those eastern sector eruptions occurring between 2,200 and 2,300 m asl (1908 (41), 1950-51 (54), 1983 (92) and 1986-87 (99)).

2. Eruptions with vents at the same elevation and at the same lateral distance from the central conduit can have differing durations e.g. 1975-76 (74) and 1976-77 (75); 1908 (41) and 1986-87 (99). Terminal eruptions can behave accordingly. For instance, between 1966 and 1971, the northeast crater was in a state of almost continuous effusive activity, whilst throughout 1977, the same vent was the focus of intermittent short duration (< 5 days) eruptions (Guest, 1982).

3. At a pre-determined elevation, there does not appear to be an upper limit to the duration of eruptions at a given lateral distance from the central conduit, and there is no conspicuous decrease in the maximum duration with increasing elevation. For example, the 18 month 1764-65 (19) eruption occurred from a vent at approximately 2,600 m asl: its duration exceeds that of all other eruptions occurring below 2,000 m asl. However, it is probably significant that this eruption, like that of 1792-93 (24), produced a lava flow field predominated by a pahoehoe surface texture.

Initial observations therefore indicate that in contrast to that seen below 2,000 m asl, vent elevation and dyke length appear to have little control on the duration of individual eruptions. For a given elevation, if the rate at which the pressure decreased with time was uniform for all eruptions, then, on the basis of previous arguments, it might be expected that there would be a well-defined relationship between eruption duration and lateral distance for that elevation. The very fact that the duration of eruptions at a given lateral distance from the central conduit appear random indicates that either the temporal pressure decay rate was variable, or the temporal pressure decay rate was constant for each eruption but some unidentified 'chance' factor randomly curtailed eruptions, or both. This is discussed further in section 5.5.

Since an increase in elevation entails a decrease in the magnitude of the magmatic driving pressure and volcanic load (section 4.5), the presence of long
duration eruptions above 2,000 m asl on the eastern sector signifies that the stress regime which exists below 2,000 m asl and within a lateral distance of 7.5 km of the central conduit, must prevail all the way up to the summit. This is in contrast to that observed on the western sector where the volcanic load influenced the duration of eruptions. In this stress environment, the duration of the majority of these eastern sector eruptions must be dictated by magmatic pressure.

The fact that short duration eruptions can occur from vents above 2,000 m asl on the eastern sector next to those of long duration eruptions does not immediately imply that the high level stress regime is prone to sudden changes. On the contrary, it is sometimes noticeable that the vent from which a short duration eruption has occurred is frequently the site of polygenetic effusive activity. This not only indicates that the feeder dyke remained open between successive eruptive episodes, but the rate of magma recharge must have been sufficiently high to counteract any cooling which had occurred within the dyke. This is well demonstrated by the 1978 'Valle del Bove' group of eruptions. Following the end of the first 1978 eruption episode on June 5th, the feeder dyke was still reported to be glowing and emitting jets of gas at high pressure in late July (SEAN Bulletin, 1978). Phase 2 of the 1978 eruptive sequence, which lasted some 5 days, commenced on August 24th from the same eruptive fissure. This was followed by a further short duration eruptive phase on the 18th of November. The magma which supplied the second and third 1978 eruptive episodes was the same magma, though more evolved in nature, which fed the first eruptive episode (Armienti et al. 1984). Taking the longest repose period between successive eruptions (phase 2 - phase 3, 92 days), and assuming that the country rock at the end of the second eruptive phase was only some 100k less than the temperature of the magma, the theoretical calculations of Wilson and Head, 1988 (Figure 2) demonstrate that a dyke less than 1 m in width would have solidified in the repose period.

In most of the above instances, the early termination of the eruptions was probably due to a magma driving pressure which was too small to sustain the eruption, rather than the volcanic load acting to close the fissure. This therefore explains how it was possible to sustain a short duration eruption in a gravitationally unstable stress regime.
5.5 Eruption Rate

The variation of eruption rate with vent elevation for post-1750 eastern sector eruptions is shown in Figure 5.6. At a given elevation, eruptions are characterised by a range of eruption rates, but for that elevation, there is no simple systematic pattern of decreasing eruption rate with increasing eruption duration. For instance, the 1892-93 (40) and 1971 (67) eruptions which have vents at 1,800 m asl, have comparable eruption rates at $\sim 6.4 \text{ m}^3\text{s}^{-1}$ and yet the 1892-93 (40) eruption lasted 106 days longer than the 66 day 1971 eruption. For the above reason, it is not surprising that over periods of comparable duration, it is difficult to find a simple pattern of decreasing eruption rate with increasing elevation. For example, the 1886 (38) and 1928 (48)

![Figure 5.6](image)

**Figure 5.6** Elevation of post-1750 eastern sector eruptive vents (m asl) plotted against eruption rate ($\text{m}^3\text{s}^{-1}$). The eruption rate value used here is the total volume erupted along the fissure divided by the duration of the effusion. This value is plotted against the lowermost active vent. The dotted line represents the highest eruption rate at a given elevation for eruptions in the 12 hour to 9 day duration range (Figure 4.11). Datum A relates to a 7 hour terminal eruption which occurred on 18th December 1977 (Wadge and Guest, 1981), and the 'Valle del Bove' group of data are identified by a black dots.
Figure 5.7
A. Lateral distance of post-1750 eastern sector eruptive vents occurring below 2,000 m asl from the central conduit (km) plotted against eruption rate (m$^3$s$^{-1}$). The eruption rate for each eruption is plotted against the lateral distance of the most distant eruptive vent. The limit has been transposed from Figure 5.6.

B. This diagram is the equivalent of A above but restricted to those post-1750 eruptions occurring above 2,000 m asl. The nomenclature is the same as for Figure 5.6.
Figure 5.7
data were both calculated over a duration value of 19 days, but the eruption rate of the 1886 (39) eruption which occurred from a vent at 1,400 m asl, exceeds that of the 1928 (48) eruption, which occurred from a vent at 1,200 m asl, by some 10.3 m$^3$/s$^{-1}$. This is counter to what was largely observed on the western sector and emphasises the complexity of eruptive activity on the eastern sector. The above observations demonstrate that the factors influencing the temporal variation of 'real-time' effusion rates are not solely a function of vent elevation. However, this does not (as demonstrated in chapter 4, section 4.6.1), necessarily negate a relationship between the initial effusion rate and vent elevation. Most of the eastern sector eruption rate data do not exceed the western sector eruption rate maxima, but this is largely because the majority of the eruptions lasted longer than the 12 hour - 9 day time period for which the eruption rate maximum limit line applies. Only the A datum, which relates to a 7 hour northeast crater eruption, exceeds this limit.

Whilst the overall variation pattern of eruption rate and vent elevation shown in Figure 5.6 appears to be similar to that seen for the western sector (Figure 4.11), there are important differences. Firstly, above 1,700 m asl on the eastern sector, the magnitude of the smallest eruption rate at a given elevation is much lower than that seen on the western sector ($< 10$ m$^3$/s$^{-1}$). These data are associated with those eruptions which lasted in excess of 32 days. Secondly, though the eruption rates of some eastern and western sector eruptions with vents at the same elevation may be similar, this is largely fortuitous, for the eruption rates were not calculated over periods of equal duration. For instance, the 1923 (46) eastern sector and 1832 (30) western sector eruptions had vents in the 1,700 to 1,800 m asl elevation range, and comparable eruption rates (22.6 and 23.5 m$^3$/s$^{-1}$ respectively), but the duration of the 1923 (46) eruption is 13 days greater than the 19 day, 1832 eruption. The volume of erupted lava also differs, the 1923 (46) eruption extruding some $22 \times 10^6$ m$^3$ more lava than $43 \times 10^6$ m$^3$ extruded by the 1832 (30) eruption.

5.5.1 Below 2,000 m asl

On first inspection, it is difficult to find a coherent relationship between the eruption rates of eruptions occurring below 2,000 m asl and their lateral distance from the central conduit (Figure 5.7A). However, when this diagram is examined alongside
the corresponding volume (Figure 5.2A) and duration (Figure 5.5A) diagrams, the following groups emerge:

1. At lateral distances beyond 7.5 km of the central conduit, whilst eruptions have volumes less than $60 \times 10^6$ m$^3$ and durations less than 50 days, there is considerable variability in the magnitude of eruption rates at a given lateral distance.

2. Within 7.5 km of the central conduit, individual eruptions have volumes in excess of $60 \times 10^6$ m$^3$ and durations in excess of 50 days, and there is a more regular relationship between eruption rate and lateral distance. In general, $Q_\sigma$ decreases as the lateral distance decreases.

3. Superimposed on the trend of (2), are the data of the 'Valle del Bove Group' of eruptions. The volume and duration of these eruptions (which are identified by a black dot), do not generally exceed $60 \times 10^6$ m$^3$ and 50 days respectively. The 66 day 1971 (67) eruption is the only exception. Since the stress field environment in which these eruptions occurred can sustain protracted eruptions (section 5.4.1), not only are the durations of these eruptions generally very short in contrast to (2), their eruption rates are also very small for their duration. This group of eruptions will be discussed separately in section 5.5.4.

**Lateral distances beyond 7.5 km**

Similarities between the limiting maximum volume and duration of these particular eastern sector eruptions and those of western sector eruptions (Figure 4.12), suggests that the magnitudes of their eruption rates were primarily influenced by high level processes. In view of the association established between vent elevation and 'peak' effusion rates (chapter 4, section 4.6.1), and the linear increase observed between the volume of lava erupted by the majority of eastern sector eruptions occurring below 2,000 m asl and lateral distance (Figure 5.2A), if conditions were uniform during all these eruptions, then it might be expected that those eruptions at lowest elevation would be associated with the shortest durations and highest eruption rates (see western sector, (section 4.6.2)). To account for the large spread between the data occurring beyond 7.5 km from the central conduit (Figure 5.7A), it must be concluded that conditions within the volcano differed during individual eruptions. For example, the 1911 (44) eruption has a higher eruption rate than the 1886 (39) eruption
even though it occurred from a vent at a higher elevation (1,625 m asl compared to 1,400 m asl) and at a greater lateral distance from the central conduit (8,900 m compared to 8,500 m). The rate at which the excess magma pressure was relieved during the 1911 (44) eruption must have been greater than during the 1886 (39) eruption. This increased eruptive efficiency is best explained in terms of a wider eruptive fissure. Since both these eruptions occurred on the rift zones (NE and S respectively) which delineate the extensional interface between the western and eastern sectors, variable rates of gravity sliding and/or differential block movements within the eastern flank (Hughes et al. 1990) provide the necessary environment in which variable initial dyke widths can be accommodated.

The 1809 (26) and 1928 (48) eruptions, which occurred from vents at the lowest elevation (Figure 5.6) and greatest lateral distance from the central conduit (Figure 5.7A) on the eastern sector represent, with their relatively long durations and low eruption rates, the most significant departure from the inferred pattern. Examining each eruption individually, the 1809 (26) eruption rate is misleading for it represents the cumulative eruption rate based on effusive activity which occurred at two separate eruption sites. The main activity during this eruption occurred close to the site of the present northeast crater whilst the effusive activity which occurred at 1,350 m asl on the northeast rift was late stage. It is unfortunate that there is no independent volume data for both eruption sites, only a cumulative volume for the whole eruption is available.

As for the 1928 (48) eruption which occurred off the rift zone on the eastern flank, it is difficult to explain why, considering its elevation, the eruption rate was not greater. The low real-time effusion rate which is suggested for this eruption implies that magma pressure was released very slowly. Though it could be argued that a relatively narrow dyke may have been responsible, this situation would have required a larger than normal volcanic load acting on the fissure. The above situation is unlikely for firstly, the eruptive fissure was emplaced into a region of the volcano already weakened by tectonic activity and secondly, the protracted duration of this eruption compared to a western sector eruption occurring at an equivalent distance from the central conduit suggests a smaller volcanic load acting on the fissure. However, it may be significant that the top of the 1928 (48) feeder dyke first
intersected with the surface at an elevation of 1,700 m asl (Imbó, 1928). Initial effusive and explosive activity took place at this elevation before the feeder dyke propagated downslope to the site of the lower eruptive vent. It is therefore conceivable that the 1928 (48) feeder dyke, instead of propagating horizontally from the central conduit to the level of the lower eruption site, left the central conduit at a higher elevation (at level of upper eruption site) before descending at a shallow level downslope (Wadge, 1980) to the site of the lower eruption site. In this way, the real-time effusion rate would have the characteristics of an eruption which occurred at a higher elevation.

*Lateral distances within 7.5 km*

Within 7.5 km of the central conduit, eruptions are sustained for longer durations and to lower effusion rates. In contrast to that observed beyond 7.5 km, and for those eruptions which produced aa surface textures and whose data define the maximum volume limit of II in Figure 5.2A, there is a general decrease in the magnitude of the eruption rate with shorter lateral distances. This coincides with a linear increase in both the volume erupted (Figure 5.2A) and eruption duration (Figure 5.5A). Therefore, as the inferred dyke length decreases, so eruptions can be sustained at lower effusion rates. The 1811 (27) and 1792-93 (24) data are exceptions. In view of its location on the volcano, not only is the duration of the 1811 (27) eruption very short (section 5.4.1), the magnitude of the eruption rate is also very low for its duration. The magnitude of the 1792-93 (24) eruption rate is half that of other eruptions at the same lateral distance from the central conduit. This implies that the eruption was characterised by lower real-time effusion rates. The way in which this eruption fits into post-1750 eruptive activity is discussed in section 5.5.2.

The change, circa 7.5 km lateral distance from the central conduit, from an apparently random relationship between eruption rate and lateral distance to a strong linear relationship, coincides with an increase in the magnitude of the eruption duration beyond the duration for which an eruption could be sustained by high level activity alone. This provides an important temporal constraint on the data. It was argued above that most of the scatter observed amongst those eruptions which occurred beyond 7.5 km and lasted less than 32 days (Figure 5.5A), was largely a
product of variable dyke widths which effected variable eruptive efficiencies. Amongst those eruptions which occurred within 7.5 km of the central conduit and exceeded 32 days duration, the stronger association between inferred dyke length and eruption rate suggests that dyke width was a subsidiary factor to dyke length in controlling the magnitude of these eruption rates. There is therefore a temporal change in the predominant dimension of the dyke which influences the magnitude and temporal variation of the real-time effusion rate. For this to occur, the geometric shape of the dyke must evolve.

Evidence for conduit evolution has been found in the geometric shapes of relic feeder conduits (Delaney and Pollard, 1981) and phenocryst alignment within 'frozen' basaltic dykes in Hawaii and northwest of Scotland (Williams, 1989). Conduit evolution occurs in response to a natural improvement in hydraulic efficiency. For a given pressure gradient, flow rates are higher for circular conduits than they are for dykes of equal cross-section (Delaney and Pollard, 1981). Therefore, as the magma flow rate decreases, only those thicker sections of the dyke, where heat loss is insufficient to solidify the magma will remain open, and flow becomes concentrated in one or more intra-dyke conduits. Since the width in this instance is not imposed on the dyke by external influences, but evolves naturally, its dimension must be determined by the magma flow rate through the dyke. In effect, the width becomes self-regulatory, increasing and decreasing in response to changes in the magma flow rate (Bruce and Huppert, 1990). However, at low and decreasing values of magma flow rate, (i) heat transfer between the dyke and the country rock becomes important (Delaney and Pollard, 1981) to the extent that magma solidifies onto the wall of the dyke, thereby reducing the width of the dyke and, (ii) viscous effects also become prominent within the molten dyke and retard the magma flow rate. The effect of these two processes is to increase the probability that the eruption will terminate. Though there has been no published study of Etnean dykes, a similar process is envisaged to occur during effusive Etnean eruptions (Wadge, 1981). The localisation of effusion to a few vents soon after the initial fissure effusion is a common characteristic of many eruptions. This has been documented for the 1892 (40) (Riccò and Arcidiacono, 1904) and 1911 (44) (Platania, 1912) eruptions.
Once magma flow rate and dyke length become mutually dependent, the main factor dictating the eventual size of an eruption must be the pressure gradient between the pressure source within the central conduit and the eruptive vent on the flank. This is directly related to the length of the dyke. The longer the dyke, the greater the effects of frictional and viscous resistance to flow as the driving pressure decreases, hence this increases the probability of a blockage developing in the dyke.

5.5.2 Eruptive Mechanism

The characteristics of eastern sector eruptions which occur within 7.5 km of the central conduit differ from those which occurred from vents outside this lateral distance range. The fact that eruptions can exceed 32 days duration, and are sustained at lower effusion rate suggests that the dyke remains open after the initial phase: however, more important is the observation that the volume erupted exceeds that which can be stored at high levels in the volcanic system. This high level capacity is inferred to be within the range 48 - 60 x 10⁶ m³. Most of the erupted material must have risen into the volcano during the course of the eruption.

The combined hydrostatic head and volatile exsolution mechanism which is responsible for the initial effusion rate is only effective over relatively short time scales and cannot therefore be responsible for the observed phenomena. Another eruptive force must preside over Etna eruptions after 32 days. Whatever its nature, the fact that effusion rates decrease exponentially with time (as shown for the 1950-51 eruption by Cumin (1954) and the 1983 eruption by Frazzetta and Romano (1984)) must reflect a comparative decrease in the magnitude of the driving force. What is more, this force must be located below the volcanic pile. An exponential decrease of the effusion rate has been observed in a number of basaltic terrestrial eruptions (Machado, 1974; Scandone, 1979; Wadge, 1981; Dvorak and Okamura, 1987) but this has generally been attributed to the gradual release of elastically stored energy from a contracting high level reservoir. The only large scale storage area which has been inferred for Etna is located at a depth of 20 km (Sharp et al. 1981).
Figure 5.8 Schematic diagram showing relationship between magma output at the surface and magma input at depth for eruptions occurring within 7.5 km of the central conduit on the unbuttressed eastern sector. For terminal eruptions \( Q_{\text{in}} = Q_{\text{out}} \): the column remains at a constant height in the central conduit and the eruption lasts as long as magma is entering the system at depth. For flank eruptions \( Q_{\text{in}} \ll Q_{\text{out}} \): depressurisation of the magma column results in an initial output which far exceeds the magnitude of the magma input at depth. There is a time delay before the source responds to the passing of the depressurisation front downwards through the system. This is accentuated by a narrow transportation conduit. \( v_s \) and \( v_e \) are the velocities of the magma at the source and the eruptive vent respectively, and \( \pi r^2 \) is the cross-sectional area of the transportation conduit.

Assuming that the height to which gas-free magma ascends within a volcanic system is determined by a balance between the lithostatic pressure forcing magma up the column, and hydrostatic magma pressures acting down through the column on the source, if the system behaved ideally and had no viscosity, then any event which reduced the magnitude of the hydrostatic pressure acting down on the source (such as a dyke draining the column at a level below its top) would invoke an immediate counter-response at source of equivalent magnitude which would force more magma into the system. In this way, changes in the effusion rate at the surface would, by continuity, directly reflect pressure changes at source. It is only under these ideal circumstances that the deep level storage area could be responsible for initial effusive
activity at the surface. However, the assumption of a viscous-free system detracts from reality. The principal affect of viscosity is to dramatically increase the response time of the deeper levels of the system to any changes which have occurred in the upper levels. Therefore, in a real system, the rate of magma output during a flank eruption would initially greatly exceed the rate of magma input (Figure 5.8). In this situation, the deeper levels of the system cannot be responsible for phenomena observed at the surface; this negates the influence of the deeper level storage area during the initial stages of Etna flank eruptions. The implications of the above inferences are two-fold; firstly, if output exceeds input then eruptions must be limited to a finite duration and secondly, the force effecting the exponential decay rate of the magma flow rate has to be located within the central conduit system.

In view of the above constraints, the eruptive mechanism advocated here is one where eruptive activity at the surface is related to a magma column which is undergoing decompression from the top downwards. This eruptive mechanism is a modified version of mechanisms previously forwarded to explain eruptive phenomena at Vesuvius (Perret, 1924) and Mount St. Helens (Scandone and Malone, 1985). Following a breach in the walls of the central conduit at a level below the top of a magma column, the gravitational drainage of the magma column overlying the point of failure effectively decompresses that portion of the column below the point of failure. An immediate response, as mentioned in chapter 4, section 4.6, would be for the nucleation level at which the gas exsolves from the magma to descend to a deeper level in the volcanic system. The depth to which it descends will depend on the initial amount of depressurisation, for the nucleation level of a volatile is a function of the confining pressure. The sudden appearance of a large quantity of volatiles in the magma would add to the buoyancy force, causing the magma to rise upwards. In doing so, this depressurises that parcel of magma at a deeper level causing it to ascend, and so forth. Thus, a chain of events is initiated which slowly affects progressively deeper levels of the system. In essence, this can be visualised as a decompression front moving down through the magma transport system towards the source. Because of the viscosity, the passing of the decompression front downwards through the system would not induce a uniform response at all levels. In general, the most immediate response would occur near the top of the column whilst a much
slower response would be felt at deeper levels as the magma overcomes its viscosity. Therefore, by the time a parcel of magma initially at depth in the volcanic system reached the nucleation level, its velocity is much lower than that of a parcel initially located at that level. Cumulatively, this manifests itself as an exponential decay in the effusion rate at the eruptive vent. The time period of the decay is therefore a function of the viscosity of the system.

Even under these circumstances, the decompression front must eventually reach the source region and invoke a response which would push magma into the system. Once this had occurred, the eruption at the surface would 'see' an infinite source of magma. That this is not the situation for post-1750 Etna flank eruptions is demonstrated by the fact that they are only capable of erupting $140 \times 10^6$ m$^3$ of the available magma volume. This indicates that there is a parameter which is compounding the viscosity factor and limiting the rate of magma ascent through the central conduit. This parameter must be the diameter of the central magma transport system connecting the volcanic pile with the source. For a narrow conduit, in addition to viscous effects, the magma flow rate is also subjected to frictional wall effects which increase the resistance to flow (Massey, 1983). Frictional wall effects are dissipated in a wider conduit, thereby allowing the freer movement of magma through the conduit. Circumstantial evidence therefore suggests that for the post-1750 period, the diameter of the central magma transport system has been relatively narrow. A narrow conduit and the viscosity of the system does not stop flow completely, it merely slows it down. Therefore, even after the largest flank eruptions, magma reappears at high levels in the central conduit a short while after the end of the eruption. Following the large volume 1852-53 (40) eruption, strombolian activity was recurring from the summit crater after a repose period of only 4 months (Tanguy, 1981).

Since the width of the conduit system at depth in the volcanic system is determined by the regional tectonic stress regime, a narrow conduit width is consistent with the present compressive regional tectonic stress regime which exists beneath the volcano (Scarpa et al., 1983). A compressive stress regime with a positive gradient (compression increases with depth) superimposed on the hydrostatic field also has the affect of artificially increasing the density contrast between the magma and the
country rock (Takada, 1989) hence, magma can ascend to the surface with little trapping in storage areas on the way. Takada cites an example where if a stress with a gradient of 2 MPa per km is superimposed on a hydrostatic system where the magma has a density of 2,800 kg/m³ and the lower crust has a density of 3,000 kg/m³, the net effect is to increase the apparent density of the lower crust to 3,200 kg/m³. This increases the buoyancy force.

In view of all the constraining conditions, the eruption which would exploit the volume of magma stored in the volcanic system most efficiently (largest volume, shortest time) would be the one occurring at the lowest elevation on the unbuttressed eastern sector but within a 7.5 km radius of the central conduit (so as to effect the largest amount of depressurisation and disruption to the system), and connected to the central conduit by the shortest dyke, (so as to withstand cooling for the longest time period and survive to lowest effusion rates). The latter parameter for Etna is dictated by the morphology of the volcano: it is not surprising that the most efficient exploitation of the post-1750 volcanic system to date was during the 1852-53 (33) eruption at 1,700 m asl. Like the 1843 (32) western sector eruption, this eastern sector eruption occurred near the break of slope where the cones of recent Mongibello activity gives way to the gentler sloping lower flanks.

5.5.3 Above 2,000 m asl

Assuming that similar eruptive conditions prevailed above 2,000 m asl, and that eruptive fissures did not close as the magmatic pressure decreased, then it might be expected that for a given elevation, uniform relationships might exist between eruption rate and lateral distance. The fact that the eruption rate at a given lateral distance from the central conduit appears variable (Figure 5.6 and 5.7B) demonstrates that this is not the case and indicates that either the pressure decay rate with time is random or, the temporal pressure decay rate is constant, but there is some 'chance' factor which is randomly curtailing eruptions, or both. On the assumption that the effusion rate is controlled solely by the magmatic driving pressure, evidence for both the above conditions can be extracted from a comparison of the volume and duration eruption data.
All the above points are aptly demonstrated by eruptions which have occurred between 2,500 m asl and 2,600 m asl. The 1985A (94) eruption extruded $30 \times 10^6$ m$^3$ of lava in 123 days whereas the 1975 (72) eruption took 68 days longer to extrude only $12 \times 10^6$ m$^3$. Furthermore, the 1989 (100) eruption extruded 15-20 $\times 10^6$ m$^3$ of lava in only 11 days. Clearly, there is a marked difference between the way in which the pressure, and hence the real-time effusion rate, decayed with time during these eruptions even though they occurred at approximately the same elevation. As for the other consideration of curtailed durations, the 1764-65 (19) eruption extruded some $60 \times 10^6$ m$^3$ over a period of 18 months. Whilst the volume and duration of this eruption exceeds that of the 1975 (72) by some $48 \times 10^6$ m$^3$ and $11\frac{1}{2}$ months, over comparable durations the real time effusion rate, whilst small, must have been similar. This is reflected in their comparable eruption rates of 1.3 m$^3$s$^{-1}$ and 0.8 m$^3$s$^{-1}$. Therefore, here we have two eruptions where the real-time effusion rate, hence real-time pressure decay rate, was probably equal, and yet one eruption lasted much longer than the other.

All the above observations demonstrate that eruptive conditions in the post-1750 period have not been uniform. By comparing the volume, duration and eruption rate data of eruptive episodes which have occurred at a given elevation, it has been possible to compile a table (Table 5.1) which describes the optimum characteristics of eruptions at three elevations above 2,000 m asl, under three different magma pressure decay rate conditions. Specific examples are given where appropriate but, when this is not possible, 'best' examples are provided of eruptions whose characteristics closely match those of the specified characteristics. The magma pressure decay rate condition which is considered to characterise the predominant style of effusive activity at a given elevation, is indicated by the stippling. (This will herein be known as the 'normal' style of activity at a given elevation).

Expanding on the model to include activity above 2,000 m asl, because there is a progressive decrease in the vertical distance through which a magma column can fall after a breach in the walls of the central conduit, it follows that the amount of initial depressurisation felt by the magmatic column at a level below the breach in the central conduit must also decrease. The initial magmatic pressures, and hence, initial real-time effusion rates therefore decrease with increasing elevation. This is in
Table 5.1 This table describes the optimum characteristics of eruptions at three different elevations above 2,000 m asl on the eastern sector, under high, moderate and low magma pressure decay rate conditions. The inferred predominant eruptive conditions at a given elevation are denoted by the stippled boxes. For each box, examples, and typical eruption characteristics (in terms of eruption duration, D, erupted volume, V and eruption rate, Q,) are provided.
<table>
<thead>
<tr>
<th>Vent Elevation</th>
<th>High dP/dt</th>
<th>Moderate dP/dt</th>
<th>Low dP/dt</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summit ~ 3,000 (m asl)</td>
<td>(A) D = (&lt; 2 days) V = (&lt; 1.5 x 10⁶ m³) Q₂ = (&gt; 50 m³/s⁻¹) <strong>Examples</strong> 1977 (7:11:1977) from NE crater.</td>
<td>(B) D = (&lt; 40 days) V = (&lt; 20 x 10⁶ m³) Q₂ = (&gt; 10 m³/s⁻¹) <strong>Examples</strong> 1971 (60); 1978(A) (76) both from near SE crater.</td>
<td>(C) D = (∞) V = (∞) Q₂ = (&lt; 1 m³/s⁻¹) <strong>Best Examples</strong> 1966 to 1971 from NE crater; 1984 (93) from SE crater.</td>
</tr>
<tr>
<td>~ 2,500 (m asl)</td>
<td>(D) D = (&lt; 84 days) V = (~ 60 x 10⁶ m³) Q₂ = (~ 10 m³/s⁻¹) <strong>Examples</strong> 1763 (18) S flank</td>
<td>(E) D = (&lt; 150 days) V = (~ 35 x 10⁶ m³) Q₂ = (~ 4 m³/s⁻¹) <strong>Examples</strong> 1985A (94) S flank</td>
<td>(F) D = (∞) V = (∞) Q₂ = (&lt; 2 m³/s⁻¹) <strong>Best Examples</strong> 1764/65 (19); 1975/76 (74).</td>
</tr>
<tr>
<td>~ 2,000 (m asl)</td>
<td>(G) D = (&lt; 150 days?) V = (&lt; 180 x 10⁶ m³?) Q₂ = (~ 10 m³/s⁻¹?) <strong>Best Examples</strong> None Observed in post-1750 period</td>
<td>(H) D = (&lt; 380 days) V = (&lt; 130 x 10⁶ m³) Q₂ = (~ 5 m³/s⁻¹) <strong>Best Examples</strong> 1950-51 (54)</td>
<td>(I) D = (∞) V = (∞) Q₂ = (~ 3 m³/s⁻¹) <strong>Best Examples</strong> 1792-93 (23).</td>
</tr>
</tbody>
</table>

Table 5.1
agreement with inferences made in chapter 4, section 4.6. However, even though initial effusion rates decrease with increasing elevation, they still exceed the rate of magma flow into the volcano. Since the volume extruded during a flank eruption is a function of the amount of initial depressurisation, the reduced amount of initial disruption to the system would dramatically increase the response time of the magma transport system at depth to an event which has occurred in the upper levels. Hydrostatically, this effect is compounded by the fact that the volcanic system at depth has to force magma to higher elevations on the volcano whilst acting against a greater downward hydrostatic magma pressure. The consequence of this to an eruption would be scale down the time period which would pass before solidifying magma would block off the fissure. Therefore, at progressively higher elevations above 2,000 m asl, there is a reduction in the finite duration of eruptions. On lateral distance diagrams, at a given distance from the central conduit, there would be a progressive decrease in the volume erupted and eruption duration with increasing elevation, but more importantly, because it takes longer to erupt a given volume of lava, the eruption rate also decreases. Under these conditions, it follows from the association established below 2,000 m asl that the most voluminous, longest duration and hence lowest eruption rate flows are associated with the shortest dyke lengths, that a similar pattern of variation should persist at a given elevation above 2,000 m asl.

At the summit, the predominant style of eruption is that where lava is extruded at very slow rates from one of the summit craters. Since activity at the summit effectively occurs from the top of the magma column, there can be no high level depressurisation component to the overall magma driving pressures. All the driving pressures must act upwards through the central conduit. Only two obvious driving forces are apparent: firstly the exsolution of volatiles at high levels within the central conduit which accentuate the buoyancy force and secondly, the lithostatic pressure sustaining the magma column at the top of the volcano. Assuming the summit of the volcano is at its equilibrium height, so that the hydrostatic pressure of the whole magma column acting down on the source region is approximately equal to the lithostatic pressure forcing magma up the column, then the buoyancy pressure forces must be quite small. In consequence, the rate of magma output at the surface must equate to the rate of magma input into the volcano (Walker, 1967 and Guest, 1973).
Under these conditions, once an eruption has commenced, it should, in theory, be capable of persisting for a protracted duration because the eruption would 'see' a continuous supply of magma (Figure 5.8). Long duration eruptions seldom occur at the summit: the 1966-1971 episode of persistent activity is amongst the longest of the post-1750 period. Other summit eruptions which erupted lava at approximately the same rate have relatively curtailed durations - the 172 day 1984 (93) eruption for example. This variability indicates that 'chance' factors operate on the volcano which randomly terminate eruptions at the summit. Amongst such factors would be failure of the walls of the central conduit and dyke emplacement at a level below the summit.

_Low \( \frac{dP}{dt} \)_

'Normal' eruptive conditions have not always persisted at a given elevation and instances occur of eruptions during which the temporal rate of pressure release was greater and/or less than normal. For example, several flank eruptions have been characterised by effusion rates at the vent whose magnitudes have been comparable to terminal effusion rates. For example, the 1764-65 (19) and 1975-76 (74) eruptions. In Table 5.1, these eruptions have been grouped into the low pressure decay rate category (Boxes (C), (F) and (I)).

In order to explain the long duration of the 1764-65 (19) eruption relative to 'normal' flank eruptions occurring at the same elevation and lateral distance from the central conduit (Figures 5.4 and 5.5B), the rate of effusion at the eruptive vent must have been steady for most of the 18 month eruption. These conditions, in which magma input equals magma output, provide the only means by which this eruption could have withstood the cooling effects that close 'normal' eruptive fissures. Considering the elevation of the eruptive vent, to account for this unusual activity on the flank, either the volcano was only partially charged before the eruption, with magma ascending from directly beneath the eruption site (no initial depressurisation mechanism), or the eruptive vent was fed by a dyke descending from the summit area. In the absence of suitable eruption reports, the conditions which prevailed within the volcanic system during this particular eruption cannot be ascertained. However, circumstantial evidence to support the concept of descending feeder dykes is forthcoming from a study of an eruptive sequence which occurred at the summit of
the volcano between 1975 and 1977. In this period, effusive eruptive activity on the volcano alternated intermittently between the northeast crater and eruptive vents located on the upper northern flanks, (Guest, 1982, Chester et al. 1985). Initial flank activity occurred from a vent at 2,600 m asl on the northeast rift (see Pinkerton and Sparks, 1976) but from the 29th of November 1975, this was restricted to a vent located at about 2,950 m asl near Punta Lucia. Though the duration of individual flank eruptive events were highly variable, being controlled by the period of inactivity at the northeast crater, it is significant that the rate of effusion at all eruption sites (northeast crater and flank) were very similar. This is reflected in the magnitude of the eruption rates (see the eruptive sequence from 72 through to 76 in Table 3.2). A logical interpretation of the above eruptive sequence is that in that interval when the northeast crater was blocked, excess magma pressure was released by diverting magma along shallow descending dykes to eruption sites on the flanks. The release of excess magma pressure reverted to the summit following the re-opening of the northeast crater, and flank activity terminated.

In relation to all other post-1750 activity, one of the most anomalous flank eruptions of this period was that of 1792-93 (24) on the southeast flank. Compared with other eruptions which have occurred at the same elevation and lateral distance from the central conduit, this eruption lasted twice as long but its eruption rate was a factor of two smaller. In view of these points, and drawing an analogy with 1764-65 (19), the effusion rate at the vent must have been low, but steady, throughout its 380 day duration. Eruptive conditions in which the effusion rate at the vent is equal to the rate of magma input into the volcano are once again suggested, but it is more difficult to envisage those attributes of the volcanic system which gave rise to the conditions necessary for this unusual flank eruption. Not only is it more difficult to visualise a dyke descending from the summit to the eruptive vent at 1,900 m asl, the concept of a partially-charged volcano can also not be accommodated. Tanguy (1981) documents that prior to this eruption, explosive activity was occurring from the summit which was giving rise to ash clouds. Therefore, the magma column must have been at a fairly high level before the eruption.

For all the eruptions which fall into this low pressure decay rate category, since magma input is considered to equal magma output, the eruption at the vent should in
theory, have been capable of persisting indefinitely for it would 'see' a continuous
supply of magma. The fact that eruptions terminate, even under these apparently ideal
conditions, is a clear indication of the dynamic nature of a volcanic system which is
undergoing perpetual readjustments in response to changing regional and volcano-
tectonic stress fields.

*High dP/dt*

At the other extreme of eruptive activity which has occurred above 2,000 m
asl are those eruptions whose eruption rates, hence real-time effusion rates, have been
higher than would be anticipated at a given elevation (Table 5.1, boxes (A), (D) and
(G)). The other characteristics of these particular eruptions are that they have a higher
than 'normal' explosivity and are of shorter duration than other eruptions at the same
elevation. The fact that they were able to erupt a greater volume of lava than
'normal' in a given time interval indicates that the initial magma pressure excess must
have been greater for these eruptions. In the context of the eruptive model, this would
have the effect of accentuating the initial difference between magma output and input
at a given elevation thereby resulting in initial real-time effusion rates more akin to
those observed at a lower elevation. However, any increase in eruptive efficiency
arising from an effective increase in the magnitude of the initial amount of
depressurisation has no effect on the ambient rate of input for that elevation. The
effect of this is to decrease the magnitude of the finite duration of these eruptions with
increasing elevation.

Differences between the two end categories of temporal pressure decay rates
are most clear-cut amongst summit eruptions. Under 'normal' eruption conditions,
lavas are extruded at very low effusion rates from one of the summit craters and can
persist for several years but, at the other extreme of post-1750 terminal activity are
the infrequently recurring high effusion rate, short duration eruptions. The initial
excess magma pressures during these eruptions must have been higher than for
'normal' persistent activity so that when it was eventually released, the upper levels
of the magma column was extruded at such a high rate that the rate of magma input
could not keep up. This resulted in small volume, short duration eruptions. A series
of these short duration, small volume (1 x 10^6 m^3) eruptions occurred between July
16th, 1976 and 28th March, 1978 from the northeast crater, each effusive episode being separated by a short repose period during which the column was re-charged (Wadge and Guest, 1981). Conditions necessary to accentuate the initial excess of magma pressure probably resulted from a blockage within the main surface magma transport conduit. Eruptions occurred when the build-up of pressure was sufficient to overcome the resistance of the obstacle. It is interesting to note that this series of eruptions immediately followed that period of summit activity during which activity on the volcano altered intermittently between the northeast crater and the northern flank. Whereas a build-up of pressure due to a blockage within the northeast crater was relieved by lava effusion on the flank in the preceding period, the absence of a similar pressure release mechanism in the following period indicates that the feeder dykes supplying the flank vents were sealed. Either the feeder dyke had solidified or they had been pinched shut during a period of high level readjustment to the volcano-tectonic stress field.

Pre-eruptive conditions must have been different for the flank eruptions which fall into this category for a mechanism involving the development of large magma overpressures within the volcano is more difficult to reconcile. In situations where magma is standing at a high level in the central conduit, an increase in magma pressure within the volcano would be compensated for by an adjustment to the level of magma in the central conduit. In effect, the magma column behaves as a manometer which regulates the amount of pressure stored in the volcano. In the event of too much pressure being present, this is relieved by a terminal eruption. A different mechanism is therefore required for generating the large magma overpressures observed during this category of flank eruption. One viable mechanism is magma over-depressurisation. To effect this mechanism, a confined gas-rich dyke would have to undergo almost 'catastrophic' depressurisation following its sudden exposure to the atmosphere during a tectonic event. The excess amount of initial overpressure would therefore be effected by the release of a larger than normal quantity of volatiles and the resulting eruption would be characterised by a more explosive eruptive style. These characteristics typify the 1763 (18) southern flank eruption but, whereas volatile exsolution increases the effusion rate at the vent under normal conditions, this effect was masked during this eruption because the sudden loss
of a considerable quantity of volatiles changed the rheology of the erupting lava (see chapter 3, section 3.4.1). Fortunately this form of eruption has not recurred frequently in the post-1750 historical period.

5.5.4 'Valle del Bove' Group of Eruptions

Of all the regions on the volcanic edifice, eruptive styles have been most variable amongst the 'Valle del Bove' group of eruptions which occurred along the orthogonal set of fissures at the western edge of the Valle del Bove (Figure 5.3). In relation to the other post-1750 activity which has occurred on the eastern sector at the same elevation and lateral distance from the central conduit, the volume of lava they erupted and the periods of their activity are generally much smaller. For example, the vent of the southern flank eruption of 1985A (94) and that of the 1978b (82) eruption, which are located on the SE branch of the orthogonal fissure system, occur at approximately the same elevation (2,500 and 2,450 m asl respectively) and at approximately the same distance from the inferred axis of the central conduit, and yet their final durations, at 123 days and 4 days, and their extruded volumes, at $30 \times 10^6$ m$^3$ and $3 \times 10^6$ m$^3$, differ by an order of magnitude. Another characteristic of these particular eruptions is that activity at any one time during an eruptive episode is not necessarily restricted to a single eruptive vent. Sometimes, activity occurred simultaneously from several localities along a single eruptive fissure, such as in 1986-87 (99), or from widely spaced localities along feeder dykes emplaced along both the SE and NE set of orthogonal fissures, such as occurred in 1979 (83). For this reason, it has been difficult to categorise some of these eruptions.

Information accumulated from observations of the temporal sequence and character of eruptive behaviour along these orthogonal fissures, the petrology of the erupted products (Armienti et al., 1984) and ground movements surrounding the eruptive fractures (Mackey and Scott, 1980), all suggest that these two fracture systems delineate the edge of an unstable block at the head of the Valle del Bove which is undergoing differential movement. Magma enters the fracture systems near their point of intersection close to the southeast crater and descends to the site of the eruptive vents. However, whilst there is a common magma source, there is a marked
difference between the eruptive style observed on each fissure set. This is considered to reflect differing resident stress conditions on each fissure system.

The southeast fracture system marks a pivotal shearing plane about which the unstable block lying to the north of the fracture system is undergoing eastward forward tilting (Mackey and Scott, 1980). Lubrication of this shear plane by magma is probably accentuating the movement of the block. Since 1971, it is noticeable that there is a general decrease in the duration of eruptive episodes which have occurred on this fissure system as the eruptive vent gets further away from the southeast crater. The longest eruptive episode, at 37 days, was in 1978 (76).

In contrast, the northeast trending fissure system outlines the collapsing backwall of the unstable block. Accordingly, some of the eruptive fissures which have occurred along this fissure system have, compared to the southeast trending fissure set, lasted for considerably longer periods. The 1986-87 (99) eruption lasted in excess of 100 days. This is not true for all eruptions; the 1989 (100) eruption whose vent was in close proximity to that of the above eruption, lasted only a few days. These differing styles of eruptive activity may be a consequence of variable displacement rates which would influence such factors as the degree of depressurisation and the dimensions of the eruptive fissure.

The concentration of much of the recent activity along these orthogonal fissures highlights the instability of this sector of the volcano, and the importance of the fissures as a pressure control valve for the volcano. If the post-1971 sequence of eruptive activity at the summit is examined (Guest, 1982, Figure 5.3; Chester et al. 1985; Romano and Vaccaro, 1986; Caltabiano et al. 1987; Global Volcanism, 1989 and Bertagnini et al. 1990), it is noticeable that activity along these fissure systems generally occurs when there is a peak in activity at the southeast crater, but inactivity at the other summit craters (chasm, bocca nuova and northeast crater). Greatest anticorrelation appears to be with activity at the northeast crater. When pressure release at the summit takes place exclusively through the southeast crater, the magnitude of the magmatic forces are sufficient to dislodge the unstable block further eastward. This phenomena warrants further study.
5.6 Lava Flows: Emplacement and Planimetric Evolution

5.6.1 Lengthening

The final lengths attained by lava flows/flow fields erupted on the eastern sector (Figure 5.9) show a variation with vent elevation which is comparable to that seen for the western sector (Figure 4.16). The final lengths of these flows show a progressive increase between the summit and 2,000 m asl, though on no occasion is the limit, (A), of Lopes and Guest (1982) exceeded. Below 2,000 m asl, there is a pattern of decreasing flow length with lower elevation, the majority of flows being shorter than the western sector limit, (B). The 1928 (48) lava flow is an exception.

Below 2,000 m asl

The relationship established between the lengths of western sector flows occurring below 2,000 m asl and the lateral distance of the eruptive vent from the central conduit, is not seen for the equivalent eastern sector flows/flow fields (Figure 5.10A). Most eastern sector lava flows, especially Type B flow fields and those flows erupted from an orthogonal set of fissures at the head of the Valle del Bove e.g. 1971 (67), 1978 (80) and 1979 (83) are shorter than the western sector limit, (II). Only the 1928 (48) lava flow exceeds the western limit.

Whilst a deficiency in volume of available magma may explain the short final lengths of a few of the Valle del Bove flows (see section 5.5.2), the same cannot be true for the remaining eastern sector flows because the volume erupted into these cooling limited flows easily exceeds that erupted into western sector lava flows at the same elevation (Figure 4.6 and 5.1), and at the same distance from the central conduit (Figure 4.7A and 5.2A). If all the erupted lava contributed to the propagation of a cooling-limited lava flow then, considering the association established between volume and flow length in chapter 3, section 3.4.2, eastern sector lava flows should be longer than western sector lava flows. This should at least be true for the eastern sector Type A flows. Only the 1928 (48) lava flow, which was erupted from a vent low on the eastern flank reflects this empirically derived relationship.

Three of the remaining Type A flows, those of 1809 (26), 1911 (44) and 1923 (46) were erupted from the distal end of the northeast rift and flowed down the northeast slopes of the volcano. Evidence that the propagation of the 1911 (44) and
Figure 5.9 Elevation of post-1750 eastern sector eruptive vents (m asl) plotted against flow/flow field length (km). The western sector limits of Figure 4.16 are included: the solid limit (A) is that of Lopes and Guest (1982), whilst the maximum lengths of western sector lava flows extruded below 2,000 m asl are constrained by the dashed limit (B). Most flow units/flow fields have aa surface textures; those predominated by pahoehoe surface textures are indicated with an asterisk. A black dot accompanying the eruption identification number indicates a lava flow erupted from the orthogonal fissures at the head of the Valle del Bove. The letter A identifies the longest flow unit emplaced during one of several high effusion rate eruptive episodes which occurred from the northeast crater in 1977. Data joined by a dashed line highlight lava flows which underwent extension following the emplacement of the initial flow unit. For these particular eruptions, the open square signifies the length of the initial flow unit.

1923 (46) lava flows was somehow impeded is forthcoming from their broad, fan-like fronts. Some measure of how much the emplacement of the 1923 (34) lava flow was restricted can be gained from the following extract by Ponte (1923). He states, "In ten hours, the lava had travelled about 7 km, falling in that distance 1,200 metres, but as soon as it had reached the plain, as has happened in other eruptions of Etna, it slackened the speed of its advance, and swelled like the carapace of a tortoise. Thus
Figure 5.10  Horizontal distance of eruptive vent from the central conduit (km) versus flow/flow field length (km) for post-1750, eastern sector eruptions occurring (A) below and (B) above 2,000 m asl. For an explanation of the annotations, see the caption of Figure 5.9. The western sector limits of Figure 4.17 are included. Re-iterating an earlier point, the fact that (A) in Figure 5.9 is a straight line, the non-constant slope of the volcano dictates that (I) on this diagram cannot be a straight line. Insufficient data prevents a more accurate representation of this limit on these diagrams.
it happened that the front of the lava, which on the evening of June 17 was about 1 km from the Circum-Etna railway, reduced its speed, and did not invade the station of Castiglione until the night of the June 19." When the whole of Ponte's eruption report is examined, it emerges that a supply of lava was sustained to the flow front even after it had reached its greatest length, but the shallowing slopes at the foot of the northeast flank checked the velocity of the advancing lava flow so that lengthening was impeded and widening effected. This emplacement scenario can equally be applied to the 1911 (44) lava flow (Platania, 1912). Had these particular Type A lavas flowed down slopes comparable to those which the 1928 lava flow traversed, it is likely that their final lengths would have exceeded those of the western sector lavas erupted from vents at the same elevation, and at the same distance from the central conduit.

In addition to the possible influence of ground slope on the emplacement of individual flow units, eruption reports reveal that during the early stages of some eastern sector effusions, several independent eruptive vents were active, each vent giving rise to a flow unit which retained its individuality throughout emplacement. The redistribution of lava which would normally feed a single flow into several flow units effectively decreases the maximum potential length which the flow unit could have achieved (Kilburn and Lopes, 1988). Though lava effusion becomes localised at a single vent with time, this tends to occur after the initial peak in effusive activity. This sequence of events is reported for the 1865 (Fouqué, 1865) and 1892 (Riccò and Arcidiacono, 1904) eruptions. The complex geometry of eastern sector vent areas during the initial stages of some eruptions is a result of en-echelon offsets to the eruptive fissure. This feature is particularly prevalent on the flanks to the east of the rift zones (Wadge, 1977). To the northeast of the Valle del Bove, fractures are characterised by sinistral movements (offsets to the left) whilst those to the south of this feature show dextral movement (offsets to the right). This type of fracturing increases the probability of several vent areas developing during the early stages of an eruption, each with the potential of giving rise to a lava flow.

Lava flows on the southern flank also have to contend with adverse topography resulting from a high density of cinder cones (Guest and Murray, 1979) - Figure 2.4. These act to decrease the overall length of the initial flow units by promoting
bifurcation and branching (Kilburn and Lopes, 1988). The 1886 (39) lava flow is an example. However, within this period, the resurfacing rate on this southern flank has been high and the topographic obstacles are becoming increasingly subdued compared to their early prominence at the beginning of this period of output (see geological map of Von Walterhausen, 1848). This is increasing the possibility that lava flows emplaced on this flank during future eruptions may achieve lengths which are greater than those currently attainable, and have the potential of invading agricultural and populated regions on the lower flanks.

**Above 2,000 m asl**

Above 2,000 m asl, there is a progressive decrease in the maximum length attained by eastern sector lava flows, though at a given elevation, individual lava flows have a range of final lengths (Figure 5.9). This scatter persists in the horizontal dimension (Figure 5.10B): lava flows erupted from vents at the same elevation, but at differing lateral distances from the central conduit, can have a range of lengths. Nevertheless, at a given elevation, there does appear to be an upper limit to the longest flow which can be erupted at a given lateral distance, there being a gradual decrease in the length of the longest flow with increasing lateral distance. In addition, with increasing elevation, there is a general decrease in the size of the longest flow extruded at a given lateral distance from the central conduit, though some lava flows erupted from the summit are an exception. Whilst flows erupted from the summit have lengths which are normally in the 3-4 km range, a few flows of greater length (> 7 km) have occurred e.g. 1809 (25) and 1977 (A) (Figure 5.9 and 5.10B). These flow units were emplaced during the high effusion rate, short duration eruptions.

With the occurrence of more eruptions in this elevation range, it may be possible to contour Figure 5.10B with bounding limits which outline, for a given elevation, the longest flow erupted at a given lateral distance from the central conduit. There are presently insufficient data to accurately complete this task. Even with further eruptions, there is no guarantee that the emplacement of individual lava flows would be unrestricted. On the whole, most of the lava flows associated with the 'Valle del Bove' group of eruptions are shorter than might be expected for their elevation, an observation which is related more to eruptive conditions within the
volcano, rather than to some external factor such as topography. The same is probably true for the pahoehoe-textured lava flows of 1764-65 (19) and 1792-93 (24). No prominent flow units are visible on these lava flow fields which might give an indication of a high effusion rate initial phase. Instead, stratigraphic relationships between individual flow structures within the flow fields are consistent with them having undergone progressive lengthening during their emplacement. Effectively, the recorded length is that of the flow field rather than the initial flow unit.

Topography has affected the emplacement of several lava flows. For instance, as a result of the shallow underlying ground slope, lava issued from the 1985A (94) vent area did not go directly into forming a flow unit, but accumulated around the eruption site. When the lava flow did begin to lengthen, it did so along a wide front. Numerous flow units developed where the lava encountered a sudden steepening of the underlying topography (Romano and Vaccaro, 1986). In this example, a considerable fraction of the erupted volume contributed to flow widening instead of lengthening. Even when lengthening did take place, the fact that numerous flow units were active simultaneously resulted in shorter than expected final flow unit lengths. More clear-cut examples of topography affecting the emplacement of flow units are provided by those eastern sector flow fields erupted from vents on the flanks of the steep-sided Mongibello cone. At the start of these particular eruptions, of which the 1950-51 (54), 1983 (92) and 1986-87 (97) are examples, the lengthening stage of the first aa-textured, channel-fed flow was interrupted by a period of widening as the lava flows descended from the steep slopes of the Mongibello constructs onto the shallow gradients of the lower flanks. Following the emplacement of the first flow unit, the channel transporting lava from the vent area to the break of slope slowly developed into a tube. Subsequent flow units adding to the flow field were not generated from the vent area but emerged as lateral breakouts or overflows from the exit of the channel/tube system at the break of slope.

In all the above instances, the first new flow unit to emerge at the break of slope following the emplacement of the principal flow unit, superseded the length of the initial flow unit and determined the length of the overall flow field. The greater length of this second major flow unit was probably aided by the fact that it did not have to form its own channel in the proximal region of the vent, but exploited the
channel system established during the initial stages of the eruption down to the break in slope. If the first new flow unit had emerged from the vent area, and not at the break of slope, then it is unlikely that it would have exceeded the length of the initial

Figure 5.11 A schematic diagram from Guest et al. (1987) showing the planimetric evolution of the 1983 flow field. Solid lines indicate the main channels that were active up to the date indicated, and after the date presented by the previous diagram. Dashed lines represent channels no longer active. The bar chart indicates the time after the start of the eruption (in days).

unit. Had the lengthening stage of the first flow unit not been hindered, it is possible that it would have achieved a comparable length to that of the final flow field. This sequence of events is clearly demonstrated for the 1983 (86) eruption by Frazzetta and Romano (1984) and Guest et al. (1987) - see Figure 5.11, and is documented for the 1950-51 (54) and 1986-87 (96) flow fields by Cumin (1954) and Caltabiano et al. (1987) respectively. The length of the first flow unit and that of the final flow field have been included on Figure 5.9 and Figure 5.10B, and are joined by a dashed line.
5.6.2 Widening and Thickening

Below 2,000 m asl

Eruptions on the eastern sector which were fed by dykes less than 7.5 km in length, continue to be supplied after the initial phase of lengthening but, as discussed in chapter 3, section 3.4.2, the additional lava volume contributes to the development of a flow field (Type B) through a process of new flow unit generation. In nearly all examples, the first aa, channel-fed flow unit has the largest dimensions with subsequent units decreasing in size and becoming concentrated at the vent area.

The timing of the initial morphological change probably reflects the attainment of an equilibrium between a declining effusion rate at the vent and the rate of cooling at the flow front (Kilburn and Lopes, 1988), though other critical factors may influence this critical period, as discussed in section 5.6.2. Since the initial effusion rate shows a dependence on the elevation of the eruptive vent (section 4.4.3), greatest disequilibrium between those forces favouring flow propagation and those impeding flow propagation should manifest itself in the flow fields produced from the vents at the lowest elevation. For example, during the 1852-53 (33) eruption at 1,700 m asl, Gemmellaro (1854) reports that the longest flow unit of the eruption, which was erupted towards Zafferana Etnea, was emplaced in the 7 days following the start of the eruption. Another long flow unit was generated from the vent after 20 days and flowed towards Milo for 5 days. However, after 46 days, effusive activity was restricted to the vicinity of the vent area.

A similar sequence of events can be inferred for the 1865 eruption from the eruption reports of Fouqué (1865). Though Fouqué’s eruption chronicles are patchy for the latter stages of the eruption, based on observations made 65 days after the eruption commenced, he reports that the direction of activity had changed from an easterly direction towards the northeast. On the basis of these observations, and with the aid of aerial photographs of the final flow field, it has been possible to distinguish between these two phases of the eruption (Figure 5.12 and Plate 1B). The first phase of this eruption was dominated by the emplacement of flow units of variable dimension, with strong aa, channel-fed characteristics. However, the second phase consisted of numerous small flow units with aa/pahoehoe textures which were extruded from ephemeral boccas located throughout the flow field.
Figure 5.12 Reconstruction of the phases in the development of the 1865 lava flow field based on information derived from Fouqué (1865) and aerial photographs. After 65 days, the planimetric development of the flow field changed, from the strong channel morphology of phase 1 characterised by numerous large Type A flows, to the more subdued ephemeral bocca morphology of phase 2.

Figure 5.13 A schematic diagram showing the various stages in the development of the 1950-51 flow field (modified after Cumin, 1954). The numbers represent the sequence of events: 1. is the first flow to be extruded on the 25th November 1950. It was emplaced within 2 hours. 2. following propagation of the fissure system downslope, the principal flow field was formed. 3. the third stage was erupted from the same vent as 2, and consisted of flows emerging from ephemeral boccas which overrode flows of phase 2. This third stage, which equates to phase 2 of the 1983 flow field above, started after 62 days.
Though several flow units were simultaneously active during the early stages of the 1892 (40) eruption (Ricco and Arcidiacono, 1904), this does not detract from the fact that the underlying style of effusive activity was similar to that of other eruptions which produced flow fields of Type B flow morphology. The dimensions of the early aa, channel-fed flow units are similar to other eastern and western sector flow units. With increasing time, new flow units decreased in size and, after 45 days, accumulated around the vent area. This led to the development of localised, pahoehoe-textured tumuli structures in its vicinity (Guest et al. 1980b).

**Above 2,000 m asl**

Above 2,000 m asl, it might be expected that the change between the two phases of flow field development would occur earlier because the contribution of the initial real-time effusion rate to the eruption, and the duration for which it predominated over the rate of cooling at the flow front, would be smaller. A smaller initial real-time effusion rate (which would reflect a smaller initial driving force) would manifest itself by way of a shorter and hence, less voluminous, first flow unit. For most of the eruptions which occurred immediately above 2,000 m asl, such as 1950-51 (54), 1983 (92) and 1986-87 (97), quantification of this theory is difficult because, as discussed in section 5.5.1, the emplacement of the initial flow unit was influenced by topography. Nevertheless, it is still possible to identify two distinct phases in the morphological evolution of these flow fields, the only difference being that the generation of new flow units did not occur from the vent area. In general, they emerged as lateral breakouts or overflows from the exit of the tube system at the break of slope. In effect, the vents of these eruptions became displaced downslope as the eruption progressed.

During the 1983 eruption, the first new flow to emerge from the bocca of the tube system at the break of slope occurred 32 days after the start of the eruption, and the extrusion of some $55 \times 10^6$ m$^3$ of lava (Frazzetta and Romano, 1984). The onset of this period of extension, which occurred after 32 days, accompanied the arrival of a new pulse of magma at the vent which was petrographically distinct from the first output (i.e. more primitive chemistry (Armienti et al. 1984)). The timing of the
petrographical change is consistent with that time period established in section 5.3.1 which must pass before the new influx of magma would occur from depth.

Between the 32nd and 62nd day of the 1983 (92) eruption, effusive activity on the flow field gradually diminished, with the generation of sequentially shorter flow units (Figure 5.11). After 62 days, all activity was restricted at the break of slope to the vicinity of the tube exit, leading to the development of a lava mound. Where occlusions developed within the lava mound which prevented the effective distribution of the lava entering it, this led to the development of backpressure within the main feeder tube. Eventually, the hydrostatic pressures were sufficient to physically lift and fracture the lava mound, thereby releasing lava from within via a series of ephemeral boccas. The importance of ephemeral boccas in the distribution of lava within a mature flow field has been previously established during the 1971 eruption (Guest et al. 1980b). By the end of the 1983 eruption, that area of the flow field around the break of slope was slowly developing into a pahoehoe-textured tumulus (Figure 5.11). Essentially, the post-62 day activity is Phase 2.

It can be no coincidence that the 1950-51 (54) lava flow field underwent similar morphological changes at approximately the same time. It is fortunate that Cumin (1954) included a schematic plan drawing of the various phases in the evolution of the 1950-51 flow field. This diagram is redrawn in Figure 5.13. For this eruption, Cumin identified three stages in the morphological evolution of the flow field. Essentially, the first two stages relate to phase 1, and the third stage is phase 2 of the 1983 flow field. The third stage of the flow fields’ development commenced after 62 days, and resulted in the development of large, pahoehoe-textured tumuli at the break of slope. This observation is confirmed on aerial photographs of the flow field. Though the 1950-51 eruption lasted nearly three times the duration of the 131 day, 1983 eruption, the extra volume extruded in the excess period did not lengthen the flow field but contributed to the construction of the tumuli around the break of slope. This observation confirms the conclusions of chapter 3, section 3.4 and of Kilburn and Lopes (1988), that lava flows appear to follow the same evolutionary growth pattern. Over a comparative time period, the 1950-51 and 1983 flow fields probably looked very similar.
At the summit of the volcano, where the contribution of the high level forces is probably quite small compared to lower elevations, the overall eruptive forces must be predominated by the deeper level forces from an early stage. This is reflected in the evolution of the flow fields. For eruptions at the summit which are characterised by continuous lava effusion, there is a tendency for the resulting flow fields to develop a phase 2 morphology from an early stage. There is little to distinguish phase 1 from phase 2. For example, Romano and Vaccaro (1986) report that the initial stage of the 1984 (89), southeast crater eruption, was characterised by the emplacement of a short flow unit which rapidly reached a length of about 2 km. Within 10 days of the start, widening had occurred as the result of the formation of numerous superimposed and parallel flow units. By the beginning of June, some 30 days after the eruption commenced, effusive activity on the flow field was occurring from ephemeral boccas distributed throughout the flow field.

Other eruptions at elevations between 2,000 m asl and the summit, such as 1975 (Pinkerton and Sparks, 1976) at 2,625 m asl, and 1985 (Romano and Vaccaro, 1986) at 2,500 m asl, produced flow fields which evolved in much the same way. Once again, it is notable that phase 1 in the development of these flow fields was curtailed compared with phase 1 of those eruptions at lower elevation.

Some insight into what the appearance of a flow field would look like following phase 2 is provided by the 18 month eruption of 1764-65 (19) on the northeast rift. In contrast to eruptions at lower elevation, which used a considerable amount of their available volume in extruding phase 1, and exhausted their supply after a short period in the extrusion of phase 2, the location of the 1764-65 (19) flow field at a high elevation on the volcano meant that phase 1 was negligible. Therefore, most of the available volume contributed to phase 2, which predominated from an early stage. In contrast to the other flow fields, the 1764-65 (19) flow field consists of numerous pahoehoe-textured tumuli. Successive tumuli appear from aerial photographs, to have developed progressively downslope, away from the vent area, so that the most distant tumulus is the youngest on the flow field. An evolutionary growth pattern similar to that described by Guest et al. (1984a) for the 1614-24 flow field of the 17th Century, may be applicable to this flow field. The continued effusion of lava after the formation of the first tumulus probably resulted in the development
of a breach at the base of the tumulus, on its downslope side. The release of lava from within the tumulus, combined with the introduction of fresh lava into the flow field at the vent, then resulted in the formation of new flow units. These piled up around the new vent because of their low emplacement rates. This set up the necessary conditions for the formation of a new tumulus. This cycle of events was probably repeated several times, gradually extending the flow field downslope. Of those flow fields emplaced on the flanks of the volcano which have prominent phase 1 and phase 2 components, the author considers that the 1950-51 flow field progressed furthest along the evolutionary trend.

The only other Type B flow field of the post-1750 period which possesses a comparable pahoehoe, tumuli morphology was emplaced during the 1792-93 (24) eruption. However, this flow field did not result from a summit eruption, but was extruded from a vent at 1,900 m asl on the outer southern wall of the Valle del Bove. The location of this flow field, in the midst of prominent two phase flow fields is therefore anomalous. There is little evidence from the morphology and stratigraphy of flow units within the flow field to suggest that the early stages of planimetric development was dominated by a period of major flow extension, unless this had been buried beneath the tumulus cover. This is unlikely because at 380 days, the duration of the eruption is comparable to a few of the two phase flows but they did not last long enough for pahoehoe-textured tumuli to form a superficial cover over the original aa-textured, channel-fed flow units.

In summary, for the majority of effusive Etnean eruptions in the post-1750 period of magmatic output, the planimetric evolution of the lava flows from Type A to Type B morphologies occurs primarily in response to a progressive diminution in the magnitude of the real-time effusion rates, though subsidiary external factors, such as variable underlying ground slope may influence the timing of the initial morphological change. The temporal decrease in real-time effusion rates reflects the volcanic systems attempt at re-establishing an equilibrium between the rate of output at the surface and the low, but constant rate of input.
5.7 Discussion

The volume extruded during eruptions on the western sector of Etna in the post-1750 period of magmatic output are determined by the duration for which the magnitude of a temporally decaying magmatic driving pressure ($P_m$) can exceed either the compressive force ($\sigma_c$) acting to close the fissure and/or, the rate at which magma is cooling within the feeder dyke connecting the eruptive vent to the central conduit. The latter parameter is influenced by the rate of magma ascent through the magma transportation system; conditions which are controlled by the width of the conduit and the viscosity of the magma.

The most efficient exploitation of the volume available in the whole volcanic system (not high level alone) was effected by the 1852-53 (33) eruption at 1,700 m asl. This eruption, which occurred within 7.5 km of the central conduit on the unbuttressed eastern sector, did not necessarily induce the greatest amount of depressurisation on the system, but the morphology of the volcano in the vicinity of the eruptive vent was such that only a short dyke was required to connect it to the central conduit. The proximity of the vent to the pressure source resulted in a protracted eruption because effusion rates were able to withstand viscous and cooling effects for longer. At lower elevation and at lateral distances up to 7.5 km from the central conduit, the sizes of eruptions decrease as viscous and cooling effects become important at progressively earlier time intervals during individual eruptions. Beyond 7.5 km on the eastern sector, the volume of eruptions is inferred to be limited to that magma volume stored at high level because eruptive fissures close before new magma can be introduced from depth. Accordingly, the size of eruptions in terms of their volume and duration generally decrease with increasing lateral distances.

Eruptions occurring above 2,000 m asl have vents within 7.5 km of the central conduit and, for the majority, it is inferred that their fissure remain open to low magmatic pressures. Nevertheless, the eruptive efficiency of individual eruptions decreases towards the summit as a result of two factors: firstly, there is a smaller initial depressurisation component and secondly, in the ensuing eruption magma has to be raised to a higher elevation. This decrease in eruptive efficiency results in an earlier curtailment of eruptions because cooling and viscous effects are experienced much earlier by the eruptions. Least efficient eruptions generally occur at the summit.
but, because these eruptions are sustained by forces at the base of the magma column and not by forces located at the top, the rate of effusion at the vent during these eruptions is approximately constant and equates to the rate of magma input, these eruptions have the potential of lasting for protracted durations. However, whilst the above conditions probably reflect normal conditions, circumstances such as blockages in the conduit and sudden tectonic events, sometimes dictate that conditions vary at a given elevation, and eruptive styles may differ.

5.8 Summary: The Post-1750 Volcanic System

Since A.D. 1750, magma has been entering the volcanic system of Mount Etna at a mean rate of 0.2 m$^3$/s$^1$. This uniform rate of magma supply, combined with the similar chemistry and petrography of most of the lavas suggests that the magma transportation system conveying magma between the source and the volcanic pile has been uniform.

In contrast, the marked sectorial control on activity at the surface demonstrates that the uniform basement conditions are not transposed to the volcanic pile. It is considered that this sectorial constraint is a consequence of an east-west orientated asymmetric gravitational stress field which is operating at high levels on the volcano, above the basement. On the buttressed western sector, the distribution of volcanic load is fairly uniform. As a result, the duration of eruptions occurring on this sector of the volcano are limited to that time period for which magma pressure can exceed the volcanic load acting to close the eruptive fissure. To date, eruptions on this flank have only lasted long enough to erupt that volume of magma stored at high levels in the volcano. No eruption has exceeded 26 days duration, or erupted a volume in excess of $53 \times 10^6$ m$^3$. The high rates of effusion which characterise these eruptions give rise to flow fields of Type A (flow unit) morphology.

Conditions differ on the eastern sector. Not only are the flanks on this sector gravitationally unstable, the distribution of volcanic load is not uniform due to the presence of the Valle del Bove depression, which nestles deep in the upper eastern flank. Eruptions on this sector of the volcano can therefore survive to lower magma pressure values compared with western sector eruptions. Within 9.0 km of the central conduit, eruptive fissures remain open after the first phase of activity, and eruptions
are sustained by magma rising up through the volcanic system. In consequence, eruptions on this flank of the volcano have the potential of lasting for longer than 32 days, and of extruding in excess of $60 \times 10^6$ m$^3$ of lava. These eruptions end when the rate of magma ascent through the volcano is insufficient to overcome the rate at which magma is cooling within the feeder dyke which connects the eruptive vent to the central conduit though near the summit, sudden tectonic events and blockages within the 'open' surface conduits may terminate the supply of magma to an eruption site unexpectedly. Overall the final morphology of the resulting lava flows differ from those observed on the western sector for under the conditions of decreasing real-time effusion rate, the initial lava flow undergoes planimetric evolution. This results in the formation of a Type B lava flow field.
CHAPTER 6

The 1600-1689 Volcanic System of Mount Etna: Eruptive Activity and Lava Flow Emplacement

6.1 Introduction

The first eighty nine years of the 17th century were characterised by a period of high output (Figure 2.6) during which the volcano was in a state of flank eruption for over 20% of the time. Compared to the post-1750 period of magmatic output, this represents a seven-fold increase in the percentage time spent in flank eruption. This period of output was preceded, during the second half of the 16th century, and succeeded, for the sixty one years after the small 1689 eruption, by periods of low magmatic output.

Several aspects of the eruptive activity which occurred between 1600 and 1689 differ from that observed in the post-1750 period of magmatic output. Individual eruptive events were capable of lasting for longer durations (up to ten years) and could erupt larger volumes of lava (up to 1.5 km³). In addition, the petrography of the lavas, as discussed in chapter 2, section 2.5.2, indicate that the magmas of the two periods were not subjected to the same crystallisation histories. Elsewhere, this has been taken to reflect distinctively different regimes of magma ascent and storage prior to eruption (Guest and Duncan, 1981; Duncan and Guest, 1982). Collectively, the above observations converge on the notion that conditions within the 17th century volcanic system were different from those which pertained in the post-1750 volcanic system.

In this chapter, these internal conditions are constrained. A model of the volcanic system is developed which describes the way in which magma was transported through the volcano and released in eruption, and how this influenced the planimetric development of 17th century lava flow fields. In the final comments
section at the end of this chapter, a comparative summary of the 17th century and post-1750 volcanic systems is presented which constrains those attributes of the Etnian volcanic system which may have undergone change between these two contrasting periods of magmatic output.

6.2 Method of Analysis

To enable direct comparison with the conclusions of the post-1750 period, the 17th century eruption data were subjected to the same analytical procedures. These methods are described in chapter 4, section 4.2 and are not repeated here. To aid comparison, post-1750 western and eastern sector maxima are included on all the relevant diagrams. It is assumed that the summit of the volcano in this period was at the post-1750 elevation, and that the axis of the central conduit was located about its post-1750 position (see Figure 3.1).

6.3 Spatial Distribution

The spatial distribution of the two flow field populations erupted on the flanks of Etna between 1600 and 1689 is shown in Figure 6.1 A and B. Below 2,000 m asl, lava flows are generally restricted to the northeast and southern rift zones (Figure 6.1 A). Only the eastern flank, Type A flow field of 1689 was erupted off the rift zones. Compared to the post-1750 period, the progression from Type A to Type B flow field morphologies with higher vent elevation is not as apparent in this output period. Type B flow fields occur at all elevations. However, there is a progressive change in the predominant surface texture of the flow fields, with aa giving way to pahoehoe with increasing elevation. Lava flows erupted at the distal ends of the rifts, such as the 1643 (7) and 1646-47 (8) flow fields on the NE rift and the 1669 (10) flow field on the southern rift, have aa surface textures. For most of these eruptions, lava extrusion was accompanied by strong degassing at the vent. In contrast, the 1634-38 (6) pahoehoe flow field was erupted fairly passively from a vent at an elevation approaching 2,000 m asl on the southern flank (Figure 6.1 A).

All flow fields occurring above 2,000 m asl are concentrated on the north, west and southwest flanks of the edifice (Figure 6.1 B). No eruptions are documented to have occurred on the upper eastern flank within and surrounding the Valle del Bove.
Figure 6.1 Spatial distribution of the two morphological flow types that characterise eruptions which occurred between 1600 and 1689. Diagram (A) shows the distribution below 2,000 m asl and (B), above. The narrow directional lines represent Type A flows whilst the broader directional lines represent Type B flow fields. An asterisk next to Type B flow fields indicates that they are predominated by pahoehoe surface textures in contrast to the other flow fields which are predominated by aa surface textures. Each flow in this diagram, and in the three diagrams that follow, is accompanied by an eruption identification number which permits each eruption to be traced to its eruption data in Table 3.1.
Figure 6.2 Spatial distribution of eruptions which occurred between 1600 and 1689 expressed in terms of the volume erupted. (A) and (B) are the distributions below and above 2,000 m asl respectively. The data are divided into two samples about a limiting volume of $60 \times 10^6$ m$^3$. Bold directional lines represent eruptions which exceeded the limiting volume whilst the narrower directional lines represent eruptions with volumes less than $60 \times 10^6$ m$^3$. A question mark next to an individual flow indicates uncertainty in the volume erupted.
Figure 6.3 Spatial distribution of eruptions which occurred between 1600 and 1689 expressed in terms of eruption duration. (A) and (B) are the distributions below and above 2,000 m asl respectively. The data are divided into two samples about a limiting duration of 50 days. Bold directional lines represent eruptions which exceeded 50 days duration whilst the narrower directional lines represent eruptions which did not last this length of time. A question mark next to an individual flow indicates uncertainty in the actual duration.
Figure 6.4 Spatial distribution of eruptions which occurred between 1600 and 1689 expressed in terms of eruption rate. (A) and (B) are the distributions below and above 2,000 m asl respectively. The data are divided into two samples about a limiting eruption rate of 10 m$^3$s$^{-1}$. Bold directional lines represent eruptions with eruption rates less than the limiting value whilst the narrower directional lines represent eruptions which exceeded this value. A question mark next to an individual flow indicates uncertainty in the eruption rate.
On the whole, pahoehoe-textured, Type B flow fields predominate, even on the western flanks. With the exception of the aa-textured lava flow of 1607 (1), Type A flows are conspicuous by their absence.

The general associations established for the post-1750 period between various eruption parameters and final flow morphology, are largely maintained for the 17th century period. Below 2,000 m asl, only the two eruptions which produced Type A flow fields did not erupt more than $60 \times 10^6$ m$^3$ of lava (Figure 6.2A) and exceed 50 days duration (Figure 6.3A). The remaining eruptions, which all produced Type B flow fields, erupted volumes in excess of $60 \times 10^6$ m$^3$, and lasted for more than 50 days, even those which occurred at low elevations. Above 2,000 m asl, eruptions occurring on the western and northern flanks, such as 1651-53 (9), were capable of extruding in excess of $60 \times 10^6$ m$^3$ of lava (Figure 6.2B) and could last for more than 50 days (Figure 6.3B). The 1607 (1) eruption, which produced the only Type A flow of this elevation range, erupted a smaller volume and lasted less than 50 days. Of the eruptions which produced Type B morphology flow fields and lasted in excess of 50 days, only the 1610 (2) eruption, extruded less than $60 \times 10^6$ m$^3$ of lava.

Below 2,000 m asl, the association of low eruption rates ($< 10$ m$^3$ s$^{-1}$) with the production of Type B flow fields breaks down for those 17th century eruptions which occurred from vents situated towards the distal ends of the two rift zones (Figure 6.4A). Irrespective of their volume, duration or final flow field morphology, all the eruptions have eruption rates in excess of $10$ m$^3$ s$^{-1}$. The post-1750 association is only restored for eruptions which occurred at a higher elevation. For example, the 1634-38 (6) eruption on the southern flank, which produced a Type B flow field is characterised by a low eruption rate ($< 10$ m$^3$ s$^{-1}$). Above 2,000 m asl, only the 1607 (1) eruption is inferred to have had an eruption rate in excess of the above value (Figure 6.4B). The remaining eruptions, even those which occurred on the western flank, had eruption rates less than $10$ m$^3$ s$^{-1}$.

A fundamental conclusion which can be drawn from the above observations is that the processes controlling eruption have differed between the 17th century and post-1750 eruptive periods. This inference is based on the following points:

1. Eruptions occurring below 2,000 m asl are generally constrained to the rift zones.
2. The volcano was capable of sustaining long duration, large volume and high eruption rate effusions at low elevations.

3. There is a distinct absence of any sectorial constraint on the final size of individual eruptions. Long duration, large volume and low eruption rate effusions were occurring on all sectors.

6.4 Erupted Volume

The elevational distribution of the volume of lava extruded by 17th century eruptions is illustrated in Figure 6.5A. Most of the eruptions which occurred on the western sector extruded volumes of lava whose magnitude are constrained by the post-1750 maximum (A) for that sector. However, the 1651-53 (9) eruption, which extruded some $440 \times 10^6$ m$^3$ of lava onto the western flank, represents a significant departure.

With the exception of the 1689 (13) eruption, all other 17th century eastern sector eruptions extruded quantities of lava which exceeded the post-1750 maximum at a given elevation (B and B'). However, in contrast to the post-1750 distribution, there is little systematic variation in the distribution of the erupted volume with increasing elevation; voluminous eruptions occurred across the entire elevation range. More marked is the fact that the quantity of lava extruded by these flank eruptions easily exceeds the largest volume eruptible by a post-1750 flank eruption ($140 \times 10^6$ m$^3$). Amongst these eruptions, the 1669 (10) is further anomalous for it occurred at an elevation (825 m asl) below which it was impossible to sustain an eruption in the latter output period.

6.4.1 Below 2,000 m asl

Whereas the largest volume erupted at any elevation below 2,000 m asl on the post-1750 eastern sector was constrained by the length of the inferred feeder dyke (Figure 6.5B, limit II), the same is not true for 17th century eastern sector eruptions. There is a considerable variation in the volume erupted at a given lateral distance. The majority of these eastern sector eruptions extruded a larger volume of lava than the post-1750 maximum at a given lateral distance. Only the 1689 (13) eruption is
Figure 6.5

A. A plot showing the relationship between the vent elevation of 17th century eruptions which occurred between 1600 and 1689, and the volume of lava they extruded. Open circles and filled squares relate to western and eastern sector eruptions respectively. Included on the diagram are error bars and the post-1750 maximum limits for the western sector (A) and eastern sector (B and B'). All the data in the following diagrams are accompanied by an identification number which allows each datum to be traced to its eruption data in Table 3.1.

B. The volume erupted by 17th century eruptions plotted against the lateral distance of their eruptive vents from the axis of the central conduit. The annotations are the same as for A. The post-1750 limits, which constrain the volume erupted below 2,000 m asl on the western (I) and eastern (II) sectors, are included on this diagram.
inferred to have extruded a volume of lava which, for its lateral distance, is smaller than the post-1750 western sector maximum (Figure 6.5B, limit I).

As mentioned above, the volume extruded by many of these flank eruptions easily exceeds the post-1750 maximum of $140 \times 10^6$ m$^3$. This in itself indicates that 17th century flank eruptions either had access to a larger volume of magma in the volcanic system, or, if the volume of magma was no greater than that which was available to supply post-1750 eruption, they could extrude a greater fraction of this volume. Moreover, it is particularly significant that some of these large volume eruptions occurred on areas of the volcano where, in the post-1750 period, it was impossible to extrude volumes of lava other than that stored at high levels in the volcanic system. The 1669 (10) eruption, which occurred at a lateral distance beyond the lateral limit of post-1750 volcanic activity, provides the best example of this point. For the deeper levels of the 17th century volcanic system to be capable of supporting surface activity to greater lateral distances from the axis of the central conduit, the configuration of the basement stress field must have been different to that which prevailed in the post-1750 volcanic system.

For those western sector eruptions which occurred below 2,000 m asl, a relationship between the volume of lava extruded and lateral distance, similar to that observed for the post-1750 period, cannot be ascertained for the 17th century period because of the lack of data. Nevertheless, the solitary western sector eruption which occurred below 2,000 m asl, that of 1643 (7), extruded a volume of lava which, even after allowances for the largest possible error on the datum, is constrained by the post-1750 western sector limit.

In summary, the constraining influence of dyke length on the volume extruded on the lower flanks of Etna (below 2,000 m asl) is not as apparent in this output period compared to that seen in the post-1750 period. This is particularly the case for the eastern sector.

6.4.2 Above 2,000 m asl

The volume erupted by post-1750 flank eruptions decreased as the elevation of the vent increased above 2,000 m asl so that at a given lateral distance from the central conduit, the volume erupted never exceeded that extruded by eruptions
occurring below 2,000 m asl. Those 17th century eruptions which occurred above 2,000 m asl on the western sector generally did not exceed the post-1750 western sector maximum (I). However, it should be noted that the volume extruded by these eruptions exceeded the volume of the largest post-1750 western sector eruption occurring at the same elevation above 2,000 m asl, and at the same lateral distance. Further elaboration on potential relationships is not possible because of insufficient data. The 1651-53 (9) western sector data is the only datum which departs from the above pattern. Whilst it could be argued that most 17th century western sector eruptions were supplied by magma stored at a high level in the volcanic system (because they are constrained by the post-1750 western sector maximum), it is most unlikely that all the volume erupted by the 1651-53 (9) eruption was stored at a high level in the volcanic system. Most of the erupted volume must have ascended into the volcanic edifice from depth during the course of the eruption. The uniqueness of this eruption indicates that in the 17th century period, there must have been some spatial variation in the mechanical response of the western sector of the volcano to an eruptive episode.

The 1614-24 (3) eruption was the only 17th century eruption to occur above 2,000 m asl on the eastern sector, but even this eruption shows a marked departure from that observed in the post-1750 period. This eruption, which occurred on the northeast rift, is the most voluminous of the 17th century period, and like most of those eruptions which occurred below 2,000 m asl on the eastern sector, was probably supplied for most of its duration by magma which ascended from below the volcanic pile. Clearly, dyke length and eruption elevation were not important considerations in influencing the volume of this, and the 1651-53 (9), 17th century eruptions.

6.4.3 High Level Volumetric Capacity

In the post-1750 period, some insight into the volumetric capacity of the upper levels of the volcano was obtained from an examination of the western sector volumetric data. The approach assumed that the volume of lava which supplied these particular eruptions resided mainly in the upper levels of the volcano prior to eruption, and that eruptions always terminated before additional material was introduced into the upper levels of the volcanic edifice from elsewhere. The short duration of these
eruptions, which did not exceed 26 days, and the available petrological information (Scott, 1983; Armienti et al. 1984) appears to support this interpretation. In the post-1750 period, the maximum volume erupted on the surface below 2,000 m asl was constrained by the length of the feeder dyke, the size of the erupted volume decreasing as the length of the dyke connecting the central conduit to the eruptive vent on the surface increased (Figure 6.5, limit I). As a consequence, western sector flank eruptions in the post-1750 period could only extrude a small fraction of the available magma supply, equivalent to some 48 (± 7) x 10^6 m^3 (i.e. that volume value of the regression line (I) at a lateral distance of 2.9 km from the central conduit in Figure 6.5B). This was considered to be the volumetric capacity of the post-1750 volcanic system at high levels. For most eastern sector eruptions, new material was introduced into the edifice from below during the course of the eruptions.

An estimate of the high level volumetric capacity for the 17th century volcanic system cannot be determined by the same approach because of the paucity data. However, it may be possible to gain an indirect insight into the high level status of the 17th century volcanic system from an examination of cooling-limited lava flow/flow field lengths. This point follows on from observations of post-1750 effusive activity where it was inferred that in an eruption, that volume of magma which was stored in the upper levels of the volcanic system prior to an eruption went into the propagation of the first flow unit, the distribution of volume with length varying according to equation 3.3. Flow length can therefore be used as a surrogate measure of volume. Since the length of a lava flow can be measured separately from the volume, this approach has the added advantage in that it releases for inclusion in the analysis, those data whose final volume was not restricted to that stored at a high level alone. However, this approach does assume that the following criteria are satisfied:
1. eruptions at a given elevation experienced the same initial eruption conditions
2. the initial flow unit was not impeded, and
3. that the length of the lava flow/flow field was determined by the first flow unit to be emplaced.

The potential usefulness/drawbacks of this technique, as applied to 17th century eruptions, are discussed in the following pages.
Figure 6.6

A. A plot of vent elevation (m. asl) versus flow/flow field length (km) for the 17th century lava flows of Mount Etna. Open circles refer to Type A morphology lava flows whilst filled squares denote Type B lava flow fields. In order to avoid cluttering of the data, only a single length measurement is presented for each eruption: that for the flow field or the length of the longest flow unit. Included on the diagram are the flow length limits of Figure 4.16 (A), and the post-1750 western sector limit (B) for flows which occurred below 2,000 m asl. Pahoehoe textured flow fields are indicated by an asterisk.

B. The lengths of 17th century lava flows/flow fields (km) plotted against the lateral distance of their outflow vents from the axis of the central conduit (km). The post-1750 western sector limits (I and II) limits of Figure 4.17 are included. See the caption of (A) for an explanation of the annotations.
Figure 6.6
Flow Lengths of 17th Century Lava

For 17th century lava flows, the relationship between eruption elevation and the maximum lengths attained by the lava flows is shown in Figure 6.6A. In this dimension, it is clear that at equivalent elevations, the vast majority of lava flows produced during 17th century eruptions are no longer than those of the post-1750 period. At a given elevation, lava flows are characterised by a range of final flow lengths, but even the longest cooling-limited flow is constrained by the bounding limits of the latter period. The 1669 (10) lava flow field, at 17 km in length, is a strong departure from the above observation, but this lava flow was issued from a vent outside the elevation range of post-1750 eruptions.

For the western sector flows of the post-1750 period, the unrestricted lengths attained by those lava flows extruded below 2,000 m asl showed a stronger association with dyke length than with vent elevation, there being a general shortening of the final lengths attained by the lava flows with increasing dyke lengths. It was argued that this observation was primarily not an artefact of topographic obstacles hindering the lengthening of the lava flow (though a few flows did show some evidence that they had been hindered), but mirrored an underlying decrease in the volume of lava which could be extruded. The volume decrease was a result of increased magma retention in the volcanic system. Certain eastern sector lavas had the potential of achieving greater lengths than western sector lava flows at their respective lateral distances from the central conduit, but the vast majority were contained by the western sector limits.

With the exception of the 1669 (10) lava flow, only the 1643 (7) and 1689 (13) lava flows of the 17th century managed to attain lengths anywhere near comparable to those reached by the post-1750 western sector lavas extruded below 2,000 m asl (Figure 6.6B). However, these particular data points are associated with the greatest uncertainties in terms of their measurable lengths. The 1643 (7) lava flow is buried at its distal end beneath the younger 1646-47 (8) lava flow field. The length given in Table 3.1 is that for the exposed flow length: the error bar assumes that the flow extends from the vent to the flow front of the overlying 1646-47 (8) flow field. For the 1689 (13) lava flow on the eastern flank, the opposite applies. The flow front is visible, but the vent area is not. The length given in Figure 3.1 assumes a vent at
1,400 m asl, which is the vent elevation given in most literature sources (Chester et al. 1985).

The only pahoehoe flow field occurring below 2,000 m asl, that of 1634-38 (6), and the aa textured flow field of 1646-47 (8), are shorter than the post-1750 flow length limit (II). For some post-1750 eastern sector eruptions whose lava flows did not reach this limit, external factors on the surface of the volcano rather than internal constraints on magma flow were forwarded as the cause. Amongst such factors were adverse topography, and multiple vents at the eruption site. Similar factors may have influenced the emplacement of 17th century flow fields. Though the vent area for the 1646-47 (8) eruption has been concealed by a cinder cone that was built-up during the eruption, there is little evidence to suggest that multiple vents were present during the early stages of the eruption. However, the location of the eruption site at the distal end of the NE rift ridge where the contours suddenly open out, probably acted as a focal point for the emergence of divergent lava flows. It may therefore be the case that several flow units were active at the same time during the early stages of the eruption. This would have reduced the maximum potential length which the lava flow would otherwise have attained if only a solitary flow unit had been fed. The resulting effect was similar to that of having multiple vents.

Deciphering the events which occurred during the emplacement of the 1634-38 (6) flow field is even more difficult to comprehend because a large portion of the flow field has been covered by younger lavas. This adds to the fact that in the exposed portions, evidence of the initial stages has been shrouded beneath a cover of tumuli. On what exposure there is available, there is little evidence of any prominent flow units which might suggest that the eruption was characterised by an initial high effusion rate phase. Nevertheless, it might be important that the flow field developed along two separate branches, one which propagated directly south from the vent, and another which developed in a southeasterly direction. If the two branches were active simultaneously during the initial phase of the eruption, then this may have acted to reduce the final length of the flow field.

For eruptions which occurred at differing elevations above 2,000 m asl, but at the same lateral distance from the central conduit, there is a general pattern of decreasing maximum flow lengths with progressively higher elevations towards the
summit (Figure 6.6B). This pattern of variation is similar to that observed above 2,000 m asl in the post-1750 period. When the data are examined in greater detail, minor differences can be resolved between the final lengths attained by 17th and post-1750 eruptions which occurred from vents at the same lateral distance from the central conduit, and at the same elevation. For example, the 1614-24 (3) lava flow which occurred from a vent at 2,600 m asl, though constrained by the post-1750 limit at its particular elevation, exceeds the length of the post-1750 flow of 1989 (100) at that elevation. However, this difference may be an artefact of the data for the measured length parameters are different. Whilst the 6.5 km length of the 1989 (100) eruption is that of the initial flow unit, the measured length of the 1614-24 (3) lava flow is inferred to be that of the flow field and not the initial flow unit. There is evidence in the morphological stratigraphy of the flow field that it underwent progressive extension during the course of the eruption. The measured length therefore overestimates the length of the first flow unit.

**Discussion**

Apart from the 1669 (10) lava flow (which occurred below the elevation range of post-1750 eruptions), over the same elevational interval, there is little to distinguish 17th century eruptions from their post-1750 counterparts in terms of the lengths attained by their lava flows. Taking these observations on their own, and bearing in mind the association established between flow length and volume, there is little to indicate that conditions at a high level in the 17th century volcanic system were any different from those which prevailed in the post-1750 period. Even after making allowances for the fact that not all the 17th century lava flows conform to the list of assumptions presented on page 190, for those which approximately do (1607 (1), 1614-24 (3) and 1651-53 (9), the distribution of the data in Figures 6.5 and 6.6A suggest that during the 17th century period, there was a progressive increase in the volume taken up by the first flow unit at elevations down to 2,000 m asl, and that below this elevation, this volume decreased. This is similar to what was observed in the post-1750 period.

At face value, it could be argued that a representative value for the volumetric capacity of the 17th century volcanic system is some 48-60 x 10^6 m^3. However, this
value is based on the volume of the longest post-1750 flow unit to emerge from a vent at or below 2,000 m asl. Eruptions which occurred below 2,000 m asl probably had access to a greater volume of lava, but adverse stress field conditions within the volcanic pile prevented them from exploiting this volume to the full. Such conditions may have prevailed during the 1643 (7) and 1689 (13) eruptions of the 17th century, thereby concealing the presence of the greater volume.

Evidence for the existence of a larger volume of magma is provided by the 1669 (10) lava flow. It is clear from the large volume of lava extruded by the 1669 (10) eruption at a low elevation, and at a large lateral distance from the central conduit, that this eruption was not subjected to the same process of magma retention inside the volcanic system. Differing stress field conditions must have prevailed during this eruption. On the basis of its exposed length alone, a minimum lava volume of some 170 x 10^6 m³ is suggested for the principal flow unit, a volume fraction which easily exceeds the size of the most voluminous post-1750 flow unit.

In view of the close similarities seen between the lengths attained by lava flows of the two periods, the length and inferred volume of the 1669 flow unit raises some important questions about the high level capacity of the 17th century volcanic system, and the suitability of using the volume of individual flow units as tools for the indirect estimation of the high level volumetric capacity. If a post-1750 eruption had occurred under the same circumstances, the volume taken up by the first flow unit alone would have effectively exhausted the post-1750 volcanic system of its eruptible magma fraction. If this situation did occur, on the basis of the geophysical information presented in earlier chapters, it is unlikely that the entire volume which went into the flow unit could have been resident in the upper levels of the system prior to the eruption. The argument that the volume of the initial flow unit can be used as a measure of the volumetric capacity of the upper levels of the volcano therefore breaks down. It may therefore be purely fortuitous that the volume of the longest flow unit in the post-1750 period was comparable to the high level capacity.

In summary, as a surrogate measure of erupted volume, and a tool in the evaluation of the high level volumetric capacity of the volcanic system, the similar lengths of 17th century and post-1750 lava flows suggest (even after making allowances for possible modifying factors) that either the capacity of the high level
plumbing system in the two output periods was not too dissimilar, or the capacities were different but the eruptive mechanisms were similar, or both.

6.4.4 Volumetric Capacity of 17th Century Volcanic System

In the post-1750 period of magmatic output, only eastern sector eruptions with vents within 7.5 km of the central conduit could access that volume of lava stored in the deeper levels of the volcanic system. Even so, the volume of lava extruded at the surface was constrained by the length of dyke connecting the eruption site to the central conduit; the size of the erupted volume decreasing as the length of the dyke increased (Figure 6.5, limit II). In view of these constraints, the largest fraction of the available magma volume which could be extruded by post-1750 eruptions was some $140 \times 10^6$ m$^3$, a value determined by that point on the regression line II at a distance of 4.5 km from the central conduit.

Determining a volumetric capacity for the 17th century volcanic system is more difficult because of the absence of any clear-cut relationship between dyke length and the volume of lava extruded. Nevertheless, it is plainly clear that these eruptions could erupt a greater fraction of the available magma volume, easily in excess of $140 \times 10^6$ m$^3$. Though the 1614-24 eruption extruded the greatest volume of lava in the 17th century period, effusing some $1.3$ km$^3$ of lava, this value cannot be taken as the real-time volumetric capacity of the volcanic system, for this volume was extruded over an interval of 10 years. During most of this eruption, the volcanic system must have been behaving as an open system, with magma entering the system at a rate equivalent to the rate at which it left. Therefore, a considerable fraction of the volume extruded during this eruption may not have been present in the volcanic system when the eruption commenced. To establish a reliable estimate for the real-time volumetric capacity, it is necessary to only consider the volumetric data of those eruptions where the rate of magma output during the eruption clearly exceeded the rate of magma input at depth (i.e. where the volcano was effectively behaving as a closed system). Accordingly, the largest eruption of the 17th century period which satisfies this condition is the 126 day, 1669 (10) eruption. Without making allowances for the volume of magma remaining in the system after the termination of the eruption, the 1669 (10) eruption extruded some $850 \times 10^6$ m$^3$ of the available magma.
volume. Considering the low elevation of the eruptive vent on the flank, the rapidity at which the lava was extruded, and the long repose period which followed this eruption, this brief period of volcanic activity may have effectively emptied the 17th century volcanic system. The extruded lava volume is therefore probably representative of the real-time volumetric capacity of the 17th century volcanic system.

6.5 Eruption Duration

In the post-1750 period of magmatic output (chapters 4 and 5), it was argued that the principal factors influencing the duration of an eruption varied according to the relative difference between the magnitude of the magmatic pressure in the feeder dyke and the volcanic load. On the buttressed western sector (uniform distribution of volcanic load), an eruption could only persist for that period during which the magmatic pressure in the feeder dyke exceeded the volcanic load acting to close the fissure. Typically, eruptions occurring in such a stress field did not exceed 32 days duration. In contrast, on the unbuttressed eastern sector, the reduced amount of volcanic load meant that eruptions could continue for a longer period, and survive to lower magmatic driving pressures. Within 7.5 km of the central conduit, feeder dykes did not close and an eruption could continue for as long as the rate of magma movement through the eruptive fissure exceeded the rate of magma cooling and solidification. Under these less restrictive conditions of magma flow, eruptions could last for longer than 32 days.

It is immediately clear from the 17th century distribution of eruption duration against eruption elevation (Figure 6.7A) that the configuration of the stress regime in this period must have differed from that which prevailed in the post-1750 period. On the western sector, the 1651-53 (9) eruption which occurred above 2,000 m asl, lasted 3 years. This is a marked departure from that observed in the post-1750 period. However, the vertical extent of the differing stress environment on the western sector may have been restricted, for the duration of the solitary eruption to occur below 2,000 m asl, that of 1643 (7), is constrained by the post-1750 western sector maximum for its elevation.

On the eastern sector, the vertical extent of the differing stress regime was much greater in the 17th century period, with long duration eruptions in excess of 32
Figure 6.7

A. The elevation of 17th century volcanic eruptions on Mount Etna plotted against their duration. Open circles refer to eruptions which occurred on the western sector whilst filled squares relate eastern sector eruptions. The post-1750 eastern sector duration limits (B and C) of Figure 5.4 are included on this diagram but, for the purposes of clarity, the post-1750 western sector duration limit is omitted. In the absence of a known duration for a particular eruptive episode, a duration estimate was determined on the basis of flow field morphology. For the two data of this period of unknown duration, those of 1607 (1) and 1689 (13) (highlighted by a 'plus' sign), a duration estimate of 32 days is assumed on the basis of their exposed Type A morphology. (This follows from the fact that no post-1750 Type A flow unit which has occurred at the same elevation as these eruptions has lasted more than 32 days).

B. The duration of 17th century volcanic eruptions plotted against the lateral distance of their eruptive vents from the axis of the central conduit. Included on this diagram are the post-1750 eastern sector, maximum duration limits of Figure 5.5A. See the figure caption of 6.7A for an explanation of the annotations.
Figure 6.7

Assumed Summit

Vent Elevation (m asl)

Eruption Duration (Days)

Key

<2,000 m asl
>2,000 m asl

WEST

EAST

Distance of Vent from Central Conduit (km)

Eruption Duration (Days)

Key

<2,000 m asl
>2,000 m asl

WEST

EAST
days occurring down to low elevations on the flanks of the volcano. The vent of the 126 day eruption of 1669 (10) was at an elevation at 825 m asl on the southern rift (Figure 6.7A). Even at those elevations above 1,700 m asl on the eastern sector and within 7.5 km of the central conduit, individual 17th century eruptions are generally of greater magnitude than comparable post-1750 eruptions (Figure 6.7A). This difference is most pronounced at 2,600 m asl, where the 10 year duration of the 1614-24 (3) eruption belittles the 18 month, 1764-65 (19) eruption. Nevertheless, as seen for the post-1750 period, eruptions at the same elevation could have a range of durations, some of which - such as the 58 day eruption of 1646-47 (8) - did not exceed the post-1750 duration maximum (C) at their respective elevation.

6.5.1 Below 2,000 m asl

Eastern Sector

Compared to the post-1750 situation, the ability of the 17th century volcanic system to sustain a 126 day eruption at a lateral distance of 15 km from the central conduit (Figure 6.7B) could only have been possible if the relative magnitude of the basement regional stress regime to the gravitational stress regime was different. In the post-1750 period, the volcano’s gravitational stress field and basement tectonics controlled eruptive style which occurred on that segment of the eastern sector outside a 7.5 km radius of the central conduit to the extent that it was impossible to support eruptive phenomena on the flanks outside a lateral distance of 13 km (Figure 5.2A) from the central conduit. There is little to indicate that the morphology of the volcano’s flanks in the 17th century period was any different from the present morphology, but it is possible, but by no means certain, that the summit was dominated by a caldera and not a cone (Guest, 1973). The gravitational stress field induced by the bulk of the volcanic edifice could not, therefore, have been that much different from the present regime. To account for the 17th century activity on the lower flanks, it must therefore be concluded that the high level gravitational stress field was being superseded by a more dilational/tensile stress field in the volcano’s basement. In order to account for the fact that open fissures could persist in the Etnean basement at greater lateral distances from the central conduit, the value of the least compressive stress ($\sigma_3$) must have been smaller in the 17th century basement.
For those post-1750 eastern sector eruptions which occurred below 2,000 m asl and lasted in excess of 32 days, the duration range seen at a given elevation was mainly due to a stronger association between the duration eruption parameter and lateral distance (regression line III in Figure 6.7B). However, this relationship only manifested itself because:

1. the eruption data which delineated the relationship occupied a restricted elevational range on the flanks of the volcano (less than 200 m).
2. the data was restricted to those eruptions which produced predominantly aa textured flow fields.
3. the temporal rate of magma pressure decay was considered to have been uniform for all these eruptions.

With the inclusion of data from elevations above 2,000 m asl which had been comparably affected by uniform magma pressure decay rates, this relationship (III) did not hold. It generally followed that the duration of an eruption at a given lateral distance from the eruptive vent decreased with increasing elevation. Individual relationships between duration and lateral distance may have existed for each elevation interval (though not for those data associated with eruptions which produced pahoehoe textured lavas).

When it is considered that the effects of the basement stress regime was more extensive (in both the vertical and lateral dimensions) on the eastern sector in the post-1750 period, and that the few eruptive events which occurred below 2,000 m asl were distributed likewise, it is not surprising that there is no coherent relationship between eruption duration and lateral distance from the central conduit. The potential of looking for relationships between duration and lateral distance at a given elevation is limited because too few elevations were the sites of multiple eruptions. However, for the one elevation which did support multiple eruptive events in the 17th century (2,000 m asl), the relationship between duration and lateral distance is not at all straightforward. When compared with post-1750 observations which occurred at the same elevation, the 1646-47 (8) eruption lasted less than the duration maximum for its lateral distance, whilst the 1634-38 (6) pahoehoe producing eruption easily exceed the same duration maximum at its respective distance from the central conduit. Assuming that these two 17th century eastern sector eruptions occurred in a similar
stress environment, and terminated with the solidification of magma in the feeder dyke, the variable magnitude of individual eruption durations could only reflect differing eruptive conditions. Evidence for this hypothesis is forthcoming when the duration data in Figure 6.7B are examined alongside the equivalent volume data in Figure 6.5B. At a given elevation or lateral distance, there is no coherent relationship between the magnitude of these two eruption parameters for individual 17th century eruptions. Though extruding similar volumes of lava (200 x 10^6 m^3 against 172 x 10^6 m^3), the extrusion efficiency of the 1646-47 (6) eruption was much greater, with effusion being complete in only 58 days. In comparison, the 1634-38 (8) eruption took approximately 4 years to extrude the same volume of lava.

Despite the fact that eruptive efficiencies may have varied, the capacity of the 17th century volcanic system to sustain a 4 year eruption on its flanks, compared to a duration maximum of little over 1 year for post-1750 flank activity at the same elevation, hints at differing magma ascent conditions within the volcanic system. This hypothesis is particularly relevant if it is assumed that the initial excess pressure, and its temporal rate of decay was similar to that which characterised post-1750 flank activity at the same elevation. Clearly, the factor which was effecting flow discontinuity amongst flank eruptions in the post-1750 volcanic system was not as prominent in the 17th century period. This is discussed in greater detail in the following eruption rate section.

When examined in combination, the volume and duration characteristics of the 1689 (13) eruption differ markedly from other 17th century eastern sector eruptions which occurred below 2,000 m asl. Though the duration of this particular eruption is not well-constrained, the aa texture, and channel morphology of the exposed lava flow, suggests emplacement over a short time interval, probably in the 10-32 day duration range. A 32 day limit is still very short compared to the other eastern sector eruptions. In addition, the volume of lava extruded during this eruption was also very small in comparison (Figure 6.5B), and may not have exceeded that volume of magma stored in the upper levels of the volcanic system. Unless the volcano was only partially charged, it is therefore difficult to reconcile the 1689 (13) eruption as having occurred in a similar stress environment. In relation to the 17th century eruptive succession, this eruption occurred at the tail end of the high output period (Figure 2.6),
and effected the final 'flushing out' of the 17th century volcanic system prior to the onset of the subsequent period of modified output (Hughes et al. 1990). It may be that conditions within the volcanic system had changed after the final high output eruption of the period, that of 1669 (10), and that the 1689 (13) eruption occurred in that period when the Etnean basement was changing towards its post-1750 configuration; the lateral extent and predominance of basement tectonics was being reduced; and an asymmetric gravitational stress regime was being established as the predominant stress field at low elevations on the eastern sector. Nevertheless, even this explanation is not wholly satisfactory for this eruptive episode still lies just within that region of the post-1750 eastern sector where long duration eruptions could be sustained.

**Western Sector**

The 1643 (7) eruption which occurred from a vent at a lateral distance of 10.7 km from the central conduit did not exceed the post-1750 western sector duration maximum for its lateral distance. In light of the conditions which are inferred to have influenced post-1750 western sector activity, it may have been the case that the lower western sector was relatively unaffected by the different basement stress regime, and that post-1750 conditions applied.

**6.5.2 Above 2,000 m asl**

**Eastern Sector**

Evidence that the influence of basement tectonics may have extended all the way to the top of the volcano on the eastern sector is provided in the 10 year eruption of 1614-24 (3). For its lateral distance from the central conduit, this eruption exceeded the duration of all other eastern sector eruptions which had occurred at lower elevations. To have sustained an eruption of this duration near the top of the edifice, not only must the real-time effusion rate have been equal to the ambient input flow rate for most of the eruption, the stress fields of the volcanic system and magma supply must have been very stable.
Western Sector

Whereas the solitary western sector eruption which occurred below 2,000 m asl possessed many characteristics akin to post-1750 eruptions at the same lateral distance from the central conduit, those eruptions which occurred above 2,000 m asl bear fewer similarities. All these 17th century eruptions exceeded the duration of the longest post-1750 western sector eruption at the same lateral distance (Figure 6.7B): the 1651-53 (9) eruption even exceeded the duration of the longest post-1750 eastern sector eruption for its lateral distance. The ability of these western sector eruptions to have durations which exceeded 32 days¹ demonstrates that the high level gravitational stress field must have been subordinate to some other stress field in the 17th century period. In view of what was occurring on the eastern sector, it may also have been that the 17th century western sector was also being affected by the dilational/tensile basement tectonics which were being transmitted upwards into the upper volcanic pile. The presence of a reduced regional least compressive stress (σ3) acting on the upper western sector is the only way to satisfactorily explain the long duration eruptions. However, the vertical and lateral extent of basement control on high level eruptive activity appears to have been much smaller on the western sector compared to the eastern sector, being constrained elevations above 2,000 m asl and to a radial distance of less than 5 km from the central conduit. The different response of the two sectors to the upward transmission of the basement tectonics was probably influenced by the superimposed high level gravitational stress field. The gravitational instability of the eastern sector probably accentuated the influence of basement tectonics on surface activity, whereas the more stable, buttressed western sector had a more stabilizing control on the extent to which basement tectonics controlled surface activity.

¹ The duration value presented for the 1607 (1) eruption is assumed to be 30 days. This estimate was determined from the lava flow morphology working off the fact that no Type A lava flow unit at the same elevation has taken greater than 32 days to be emplaced in the post-1750 period. However, Tanguy (1981) questions the existence of the 1607 eruption and argues that the lava flow attributed to this year was extruded during a second 1610 eruption which occurred some 3 months after the first. Activity during this second phase lasted 104 days.
Collectively, the available data suggest that in the 17th century period, most of the eruptive activity at elevations above 2,000 m asl was occurring in a stress environment which allowed for the presence of open fissures at low magmatic driving pressures. Even so, it is clear from the range of eruption durations at a given elevation that conditions must have varied between individual eruptive events. There is little to suggest that the differences in duration were a result of varying extrusion efficiencies. For instance, whilst the 1614-24 (3) eruption at 2,600 m asl extruded a proportionally greater volume of lava than the 1610 (2) eruption at the same elevation (Figure 6.6B), and similarly lasted a proportionally greater duration (Figure 6.7B), in real-time, the quantity of extruded lava was probably similar. It must be concluded therefore that the eruptions were being randomly curtailed. This will be discussed further in the next section.

6.6 Eruption Rate

The relationship between eruption elevation and 17th century eruption rates is shown in Figure 6.8A. None of the eruptions had eruption rates which exceeded the post-1750 maximum eruption rate limit at their respective elevations, but when it is considered that most of the eruptions lasted for periods which exceeded the 12 hour to 9 day duration range for which the eruption rate maximum limit was calculated, this is not surprising. The only eruption to last less than 24 hours in this period was that of 1643 (7). However, its eruption rate is much smaller than the post-1750 maximum at its elevation.

At a given elevation, 17th century eruptions are characterised by a range of eruption rates, which decrease as the eruption duration (number in brackets) increase. However, when comparing eruption rates over a range of elevations, there is no simple relationship between these two parameters and the duration for which the eruption rate was calculated. For example, the eruption rate of the 58 day 1646-47 (8) eruption at 2,000 m asl exceeds that of the 30 day 1607 (1) eruption at 2,100 m asl even though it might be expected that the longer duration eruption would have the smaller eruption rate. Even with the uncertainty (underestimation) which is associated with the duration of the 1607 (1) eruption, a longer duration would decrease the eruption rate and accentuate the difference between the two values. As a further example, the
Figure 6.8

A. The elevation of 17th century volcanic eruptions on Mount Etna (m asl) plotted against their eruption rates (m$^3$s$^{-1}$). The eruption rate value used here is the total volume erupted along the whole fissure divided by the duration of the effusion. Those eruption rate data whose duration value was estimated from the exposed lava flow morphology, are highlighted by a 'plus' sign. The number included in brackets after the identification number refers to the eruption duration. This is expressed in days (d) or years (y). Open circles and filled squares refer to eruptions which occurred on the western and eastern sectors respectively. The post-1750 dotted line of Figure 4.11 is included on this diagram.

B. The eruption rate (m$^3$s$^{-1}$) of 17th century volcanic eruptions on Mount Etna plotted against the lateral distance of their eruptive vents from the axis of the central conduit. The dotted limit has been transposed from Figure 6.8A. See the caption of A for an explanation of the annotations.
**Figure 6.8**

**Diagram A**
- **Assumed Summit** vs. **Vent Elevation (m, asl)**
- **Key**:
  - <2,000 m asl
  - >2,000 m asl
- **Legend**:
  - WEST: ○
  - EAST: ■

**Diagram B**
- **Eruption Rate (m^3 s^-1)** vs. **Distance of Vent from Central Conduit (km)**
- **Key**:
  - <2,000 m asl
  - >2,000 m asl
- **Legend**:
  - WEST: ○
  - EAST: ■
eruption rates of the 1610 (2) and 1614-24 (3) eruptions, which occurred from vents at 2,500 and 2,600 m asl respectively, are very similar even though the latter eruption lasted for 10 years compared to the 86 days of the 1610 (2) eruption. Collectively, the above information demonstrates that the temporal variation of the real-time effusion rate during the 17th century was not simply a function of eruption elevation alone. However, as stated in chapter 5, section 5.5.1, whilst the eruption rate can be affected by a whole host of differing factors, this does not necessarily negate a relationship between the initial real-time effusion rate and vent elevation.

Final 17th century eruption rates are comparable to those of the post-1750 period at equivalent elevations (Figure 4.11 and 5.6), but this is largely fortuitous. When individual eruption rates are inspected alongside their component volume and duration eruption parameters, some major differences become apparent. They are:

1. At elevations below 2,000 m asl, individual eruption rates in the post-1750 period decreased very rapidly over the first few days of the eruption. High individual eruption rates, such as the 127 m$^3$s$^{-1}$ of the 1981 (86) eruption, reflected durations less than 2 days. For those eruptions which progressed beyond a duration of 50 days, individual eruption rates decreased below 10 m$^3$s$^{-1}$. The 80 m$^3$s$^{-1}$ eruption rate of the 126 day, 1669 (10) eruption represents a marked departure from the post-1750 pattern.

2. In that category of 17th century eruption characterised by eruption rates less than 10 m$^3$s$^{-1}$, the final eruption rates exceeds that of post-1750 eruptions occurring at the same elevation. This is despite the fact that the eruptions of the former period were of considerably longer duration. This important observation indicates that the ambient rate of magma input during the waning phase of the 17th century eruptions must have been greater than that of the post-1750 period.

6.6.1 Eruption Durations < 32 days

One of the contributory factors which indirectly influenced the final eruption rate of eruptions in the post-1750 period was the ability to maintain an open fissure. On the buttressed western sector and at lateral distances from the central conduit beyond 7.5 km on the unbuttressed eastern sector, it generally followed that eruptive fissures closed as the magma driving pressure decreased to the levels of the least compressive stress. As a result, individual post-1750 eruption rates did not decrease
below 10 m³s⁻¹. Though the initial real-time effusion rate may have been the same for all eruptions (its magnitude increasing as the eruption elevation decreased on the flanks of the volcano), for the eruptions which occurred on these areas of the volcano, the natural decrease of the real-time effusion rate was subjected to some enforced 'dampening' as the magmatic pressure decreased, and the feeder dyke narrowed. This manifested itself most clearly in the lateral dimension (Figure 4.12 and 5.7A), there being a general increase in the eruption rate with increasing lateral distance in response to a progressively earlier curtailment of the eruptions. Overall, those eruptions which occurred from vents nearest the central conduit persisted for the longest durations and had the smallest eruption rates. The distribution of the data was not entirely uniform with lateral distance and a certain amount of scatter was introduced as a result of variable eruption efficiencies. This was inferred to have been primarily induced by variable dyke widths.

Of the two 17th century eruptions which are considered to have been subjected to similar stress conditions, the 1643 (7) eruption had a comparable duration to the post-1750, 1981 (90) eruption and occurred at the same lateral distance from the central conduit, but the eruption rate, at 11.5 m³s⁻¹, is very small in comparison (Figure 6.8B). Considering the fact that the lower end of the eruptive fissure cannot be seen, the eruptive fracture which supplied this eruption may have extended itself to that critical length where the least compressive stress (σ₃) was equal in magnitude to the value of the excess magma pressure at that elevation. The resulting eruption only extruded a small quantity of lava in the short eruptive period. An analogy can be drawn with the 1883 (38) eruption which occurred on the southern flank in the post-1750 period.

As for the other 1689 (13) eruption, considering the rapidity at which the eruption rate is known to vary over the initial few days of an eruption, definite statements about this datum cannot be made with any confidence because of the uncertainty associated with the duration. Nevertheless, a general statement about the real-time effusion rate of this eruption is possible from an examination of the flow length data in Figure 6.6A and B. Assuming that the length of the 1689 (13) lava flow presented in these diagrams represents the cooling-limited flow length, the
comparable final length of this flow unit to corresponding post-1750 lava flows suggests that a similar eruption mechanisms may have operated during this eruption.

6.6.2 Eruption Durations > 32 Days

For those 17th century eruptions which lasted greater than 32 days, the distribution of their eruption rate with lateral distance is shown in Figure 6.8B. The data are not separated into those which occurred above and below 2,000 m asl. Though the number of data are small, it is still possible to make some interesting inferences from the available information. In the post-1750 period, only eruptions which occurred within 7.5 km of the central conduit on the eastern sector could survive to low magmatic driving pressures. For the eruption rate data which occurred in this lateral distance range, it generally followed that below 2,000 m asl, the distribution of the data associated with aa producing eruptions was more uniform than that seen for eruptions which occurred outside a lateral distance of 7.5 km, there being a gradual decrease in the eruption rate with decreasing lateral distance in response to increasing eruption durations (Figure 5.7A). It was argued that the more even distribution of the data was a direct result of a decrease in the importance of dyke width with time, combined with an uniform decay of magma flow rate (Figure 5.7A). The inclusion of data from those eruptions which occurred above 2,000 m asl introduced complications because of variable rates of magma flow rate decay (Figure 5.7B), but there may have been individual eruption rate relationships with lateral distance for each elevation.

When the eruption rate data of Figures 6.8A and B are examined alongside their component erupted volume (Figure 6.5B and eruption duration (Figure 6.7B) data, the following points of detail emerge:

1. there is a general increase in eruptive efficiency with decreasing elevation below the summit. The largest quantity of lava extruded in a given period of time was effected during the 1669 (10) eruption. In contrast to the eruptions which occurred at a low elevation on the post-1750 western sector, the temporal decay of the real-time effusion rates during this eruption were not subjected to 'dampening' but decayed naturally. In terms of the eruptive model, the location of the eruptive vent at a low elevation in a more dilational/tensile basement stress field favoured a high initial
hydrostatic driving pressure. This effected a high initial effusion rate whose magnitude decayed slowly during the course of the eruption. However, the eruption terminated before the real-time effusion rate could decrease to the level of the ambient rate of magma input for its elevation. This explains its large eruption rate. At the other extreme of 17th century activity, the 1614-24 (3) eruption which occurred from a vent at 2,600 m asl, extruded a slightly greater volume of lava compared to the 1669 (10) eruption, but the latter was 29 times more efficient at extruding the volume.

2. With so few elevations being the site of repeated eruptions in the 17th century period, there is little prospect of looking for individual relationships between eruption rate and lateral distance for a range of elevations. However, for individual elevations, it is clear that the distribution of eruption rate with lateral distance is not uniform. For example, the 1634-38 (4) and 1646-47 (8) eruptions occurred at approximately 2,000 m asl, but there is little coherent variation in their eruption rates with lateral distance. In the post-1750 period, for eruptions which were characterised by an uniform magma flow rate decay, scatter between individual eruption rate data at a given elevation was a reflection of a stronger relationship between eruption rate and lateral distance. The absence of such a relationship between these two 17th century eruption data must infer that the two eruptions were not subjected to the same rates of magma flow rate decay. As stated in section 6.5.1, this interpretation is reinforced when the erupted volume (Figure 6.5B) and eruption duration (Figure 6.7B) components of the eruption rate are examined alongside each other. If the estimates of the extruded volumes are correct, the 1646-47 (8) eruption was 25 times as efficient as the 1634-38 (6) eruption at extruding the same volume of lava. Variable rates of magma flow rate decay are also suggested for the 1607 (1) and 1651-53 (9) eruptions which occurred at 2,100 m asl, but the uncertainty associated with the duration of the former eruption prevents any further exploration of this relationship.

In an exercise similar to that performed on the data of those post-1750 eruptions occurring above 2,000 m asl, the volume, duration, eruption rate data and eruptive styles of individual 17th century eruptive episodes were examined and a table compiled (Table 6.1) which described the optimum characteristics of eruptions at three elevational intervals under three different magma pressure decay rates. In Table 6.1,
the predominant eruptive conditions at a given elevation are highlighted by the stippled boxes.

Dealing with each category of magma pressure decay rate separately, the eruptions which characterise the **High** $dP/dt$ category bear many resemblances to the equivalent post-1750 category in that the eruptions were restricted to a finite duration and associated with abnormal explosive activity. This gave rise to a large cinder cone over the vent area. In addition, the lavas produced were strongly aa in texture and were predominated by large channel-fed flow units. The principle difference between the eruptions of the two periods is that the 17th century eruptions are generally larger. For the equivalent post-1750 eruptions occurred on the flanks, it was suggested that the high excess developed as a result of the sudden over-depressurisation of a gas-rich dyke following a tectonic event. A similar mechanism probably gave rise to the 17th century eruptions of 1646-47 (8) and 1669 (10). Both eruptions, which are located on the rift zones, were probably initiated by an event which effected the seaward displacement of the eastern flank. The mass which was moved during these events must have been tremendous: the eruptive fissure of the 1669 (10) eruption is 15 km in length.

All the remaining 17th century eruptions are considered to represent examples of the **Moderate** $dP/dt$ category. However, even this category is markedly different from that which characterised post-1750 eruptions. In the latter period, eruptions in this category were restricted to a finite duration. In contrast, 17th century eruptions in this category which initially proceeded at effusion rates which exceeded the ambient rate of magma input, were capable of surviving for protracted durations. Once steady state conditions of magma flow had been attained, the eruption could be terminated at any time by a change in the supply conditions at source, or a change in the high level stress field. This random process is reflected in the varied duration of the eruptions which fall into this category. The differences between the two periods are discussed in section 6.6.3.

3. Despite there being differences in the rate of magma flow rate decay, for those eruptions which survived to the stage where the rate of magma output at the surface was equal to the ambient rate of magma input from depth, it is clear that after a certain time interval, the period of an eruption has little influence on the final eruption
Table 6.1

This table describes the optimum characteristics of 17th century eruptions which occurred at three different elevational intervals in a tensile/gravitationally unstable stress environment, under high, moderate and low magma pressure decay rates conditions. Inferred normal conditions at a given elevation are denoted by the stippled boxes. For each box, examples and typical eruption characteristics (in terms of eruption duration, $D$, erupted volume, $V$ and eruption rate, $Q_a$) are provided.
<table>
<thead>
<tr>
<th>Vent Elevation</th>
<th>High dP/dt</th>
<th>Moderate dP/dt</th>
<th>Low dP/dt</th>
</tr>
</thead>
<tbody>
<tr>
<td>High Elevations</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2,000-3,000 m asl</td>
<td>(A) ( D = (\infty \text{ days}) ) ( V = (\infty \text{ m}^3) ) ( Q_a = (&lt; 4 \text{ m}^3\text{s}^{-1}) )</td>
<td>(B) ( D = (\infty \text{ days}) ) ( V = (\infty \text{ m}^3) ) ( Q_a = (&lt; 2 \text{ m}^3\text{s}^{-1}) )</td>
<td>(C) ( D = (\infty \text{ days}) ) ( V = (\infty \text{ m}^3) ) ( Q_a = (&lt; 2 \text{ m}^3\text{s}^{-1}) )</td>
</tr>
<tr>
<td></td>
<td>Examples: None Observed</td>
<td>Best Examples: 1614-24 (3) on NE rift</td>
<td>Best Examples: None Observed</td>
</tr>
<tr>
<td>Mid-Elevations</td>
<td>(D) ( D = (&lt; 20 \text{ days}) ) ( V = (\sim 10^6 \text{ m}^3) ) ( Q_a = (&gt; 50 \text{ m}^3\text{s}^{-1}) )</td>
<td>(E) ( D = (\infty \text{ days}) ) ( V = (\infty \text{ m}^3) ) ( Q_a = (4 \text{ m}^3\text{s}^{-1}) )</td>
<td>(F) ( D = (\infty) ) ( V = (\infty) ) ( Q_a = (&lt; 2 \text{ m}^3\text{s}^{-1}) )</td>
</tr>
<tr>
<td>1,000-2,000 m asl</td>
<td>Examples: 1646-47 (8) on NE rift</td>
<td>Best Examples: 1651-53 (9) on W flank; 1634-38 (6) on SE flank?</td>
<td>Best Examples: 1634-38 (6) ?</td>
</tr>
<tr>
<td>Lower Elevations</td>
<td>(G) ( D = (&lt; 20 \text{ days}) ) ( V = (\sim 1 \text{ km}^3) ) ( Q_a = (70 \text{ m}^3\text{s}^{-1}) )</td>
<td>(H) ( D = (\infty \text{ days}) ) ( V = (\infty \text{ m}^3) ) ( Q_a = (5 \text{ m}^3\text{s}^{-1}) )</td>
<td>(I) ( D = (\infty) ) ( V = (\infty) ) ( Q_a = (&lt; 3 \text{ m}^3\text{s}^{-1}) )</td>
</tr>
<tr>
<td>&lt;1,000 m asl</td>
<td>Examples: 1669 (10) on S rift</td>
<td>Best Examples: None Observed</td>
<td>None Observed</td>
</tr>
</tbody>
</table>

Table 6.1
rate. This is demonstrated by the long duration eruptions (> 1 year) of 1634-38 (6), 1651-53 (9) and 1614-24 (3) which occurred above 2,000 m asl. Considering that these eruptions occurred over an elevation range of 600 m, a lateral distance range of over 5.6 km, and lasted durations which differed by some 7 years, their final eruption rates are very similar (Figure 6.8B). The significance of this is that given a long enough duration, eruptions at all elevations and at varying lateral distances from the central conduit approach a minimum eruption rate which approximately equates to the ambient rate of magma input into the volcanic system.

The length of time taken for an eruption to attain this status is clearly a function of the difference between the initial real-time effusion rate at the vent and the ambient rate of magma input. This is a function of eruption elevation (chapter 5, section 5.5). For eruptions near the summit, the difference between these two real-time effusion rate parameters is much smaller than for those eruptions which occurred at the foot of the volcano, hence the period of time which must pass before these two parameters are equable is much reduced. For the 1610 (2) eruption at 2,500 m asl, the comparable eruption rate of this 86 day eruption to that of the 10 year 1614-24 (3) eruption suggests that at this elevation interval, it took less than 86 days for the two parameters to become approximately equal. In contrast, by the end of the 126 day eruption of 1669 (10) at 850 m asl, the large eruption rate suggests that the real-time effusion rate either had not reach the levels of the ambient rate of magma input or if it had, the eruption could not have been sustained at that rate of magma flow for any great period of time otherwise the final eruption rate would have been much smaller. However, on a cautionary note, this particular eruption should not be directly compared with those examples provided for higher eruption elevations because differing magma flow rate conditions prevailed during these eruptions (Table 6.1). For instance, whereas the examples for higher eruption elevations were associated with the moderate magma pressure decay rate category in Table 6.1 (boxes B and E), the 1669 (10) is considered to be an example of a category of eruption which occurred at a low elevation, but was characterised by a high magma pressure decay rate (G). In terms of the affect this would have on the time interval required to attain the status of steady state ambient flow rates, this time period would be much greater for the
1669 (10) eruption because of the accentuated differences between the initial real-time effusion rate and the ambient rate of magma input.

6.6.3 Eruptive Mechanism

Despite there being major differences between the final volumes and durations of 17th century and post-1750 effusive eruptions, several aspects of their eruptive styles are similar. Firstly, a temporal decrease in the eruption rate suggests an underlying decrease in the real-time effusion rate for 17th century eruptions. Secondly, in view of the association established between flow length and effusion rate, if the lengths of the flow fields are assumed to reflect the length of the initial flow unit to be extruded, the comparable final lengths attained by 17th century and post-1750 lava flows at the same elevations suggest similar real-time effusion rates over the early stages of the eruption. It is only in the latter stages of eruptions that differences emerge between the two periods, with higher ambient flow rates for the 17th century period compared to the post-1750 period. This supports the idea that the eruptive mechanisms which operated during the initial stages of these eruptions were similar.

The concept of similar eruptive mechanisms is not wholly unreasonable in view of the finite number of physical variables within a volcanic system which can actively contribute to the eruptive process. However, it has been postulated on the basis of lava petrography, that the configuration of the high level plumbing system differed between the two periods, with a high level storage area being present in the 17th century period (Guest and Duncan, 1981; Duncan and Guest, 1982; Hughes et al. 1990). If such a situation existed in the 17th century period, then it is possible that the initial eruptive mechanism may have involved the deflation of this storage area, and not the depressurisation of a magma column within the central conduit. It is only when 17th century eruptive activity is compared with that of volcanic systems such as Mauna Loa and Kilauea on Hawaii, for which high level storage areas are known to exist, that supportive evidence for a high level storage area in the 17th century volcanic system breaks down.

For the Hawaiian volcanoes, eruptions are generally preceded by the inflation of a high level storage area (3-4 km depth) in response to the influx of new magma.
from depth. At the onset of an eruption, the dilated storage area is punctured, and the contracting storage area forces magma into the rift zones where, if conditions are favourable, it escapes onto the surface of the volcano. Typical durations for eruptions resulting from this high level phenomena lie in the range \( \frac{3}{4} - 38 \) days for Kilauea (Dvorak and Okamura, 1987), whilst the voluminous 1950 and 1984 eruptions of Mauna Loa lasted 12 and 20 days respectively (Malin, 1980; Lockwood et al, 1985 and King, 1989). These eruption duration values are no greater than those associated with high level phenomena on Etna in the post-1750 period. No high level storage area was present in that period. Long duration eruptions do occur on the Hawaiian volcanoes, for example the 450 day, 1859 eruption of Mauna Loa and the 1969-74 Mauna Ulu and the currently active 1983 Pu'u O'o eruptions of Kilauea. However, like the long duration eruptions of Mount Etna, there is evidence that the Hawaiian volcanoes behave as open systems with new magmatic material (of contrasting composition to that initially erupted) ascending into the volcano during the course of the eruption (Wright, 1971 and Tilling et al. 1987).

For the Etnean eruptions of the post-1750 period, even if a high level storage area had been present, it is unlikely that it would have any markedly different effect on the duration of eruptions which resulted from high level eruptive activity, or on the change-over duration range at which deep level phenomena began to take over from high level phenomena as the predominant control on surface activity. It is not unreasonable to assume therefore, that the 32 day change-over limit of the post-1750 period is also applicable to the 17th century period. Those long duration eruptions in excess of 32 days must therefore have been supplied, at least in part, by magma introduced into the upper levels of the volcanic system during the course of the eruption. It is highly improbable that the 4 year, 1634-38 (6) and the 126 day, 1669 (10) eruptions were sustained by eruptive mechanisms located at a high level in the volcanic system. As discussed in section 6.4, the presence of a high level storage area would also manifest itself through other eruption parameters. For example, the volume extruded during the initial high effusion rate phase would be greater if the volumetric capacity of the system at high levels was higher. This in turn would reveal itself as longer initial lava flow units. These characteristics of 17th century activity are not to dissimilar to those observed in the post-1750 period. Based on these
collective observations, the existence of a substantial high level storage area in this output period is therefore debatable.

In the absence of a high level storage area, the 'cicirara' petrographic texture of the 17th century lavas could not have formed in the upper levels of the volcanic system. The uniformity of this texture in the 1669 lavas must indicate that it was already in existence in the magma as it was ascending from depth. This constrains its location of formation in the volcanic system to that region below the volcanic pile, but above 22 km depth (below this depth, plagioclase crystallisation is suppressed by the pressure). The requirement of conditions which favoured low nucleation rates but enhanced crystal growth for the development of the 'cicirara' texture limit the number of potential locations which could harbour its development. It may have been that plagioclase crystallisation was occurring in the deep level storage area of Sharp et al. 1981.

An eruptive mechanism for the 17th century period, similar to the combined hydrostatic/volatile exsolution and magma depressurisation mechanism proposed for the post-1750 activity, is not negated by the eruption data, but some aspect of the volcanic system must have been different to allow for the longer duration of these eruptions, and the higher ambient input flow rates. Reiterating previous observations based on the final cooling-limited lengths of the flow fields, initial eruptive conditions do not appear to have been that different, the planimetric dimensions of those lava flow units erupted on areas of the volcano which supported long duration eruptions generally increasing with decreasing eruption elevation. This point, which is indicative of increasing initial real-time effusion rates with decreasing eruption elevation, can be explained in terms of the proposed mechanism, the higher effusion rates reflecting a greater initial pressure contribution from the elevationally dependent hydrostatic/volatile exsolution pressure components. The differences must therefore be associated with the waning phase of the eruptions.

Further constraint on the cause is forthcoming when the data associated with eruptions which occurred from vents above 2,000 m asl in Table 6.1 are compared with the equivalent data in Table 5.1. In essence, it is as if the characteristics of every box in Table 5.1 of the post-1750 period has been displaced once to the left for Table
6.1 of the 17th century period. The characteristics which are common to both Tables are:

1. those eruptions which commenced with a high excess magma pressure were restricted to a finite duration and
2. those eruptions which commenced with a low excess magma pressure could survive for protracted durations.

The critical grouping in Table 6.1 which shows up the differences most clearly is the one for eruptions which commenced with a moderate excess magma pressure (box B). Whereas post-1750 eruptions within this excess magma pressure range were restricted to a finite duration (Table 5.1, boxes B, E and H), 17th century eruptions could last for a protracted duration and had the potential of lasting indefinitely.

The inability of post-1750 flank eruptions to survive for protracted durations was attributed to a narrow magma ascent conduit, and the viscosity of the magmas. In circumstances where the rate of magma output at the surface exceeded the ambient rate of magma input, the collective effect of these two factors was to limit the rise of magma to a finite ascent rate (as determined by the cross-sectional area of the conduit), and to slow down the response time of the volcanic system's deeper levels to and eruptive event which had occurred at the surface. This is inferred to have led to a discontinuity in the flow of magma through the volcanic system which allowed the magma in the feeder dyke connecting the central conduit to the eruptive vent to solidify before magma continuity could be re-established.

In the 17th century period, it is clear that under the same conditions of pressure decay, eruptions where the initial rate of output at the surface exceeded the ambient rate of magma input from depth could survive to the stage where these two parameters were equal without a discontinuity developing in the volcanic system. In terms of the eruptive mechanism, once steady state output had been achieved, the eruption at the surface would 'see' an infinite supply of magma, and have the potential of surviving indefinitely. Only changes in basement conditions, or variation in the rate of magma supply at source could terminate these eruptions. In the 17th century, magma must have been able to ascend more freely through the volcanic system.
Under steady state conditions, the rate at which magma ascends between the source region and the volcanic pile can be described by the Hagen-Poiseuille equation for laminar flow through a circular conduit (Massey, 1987). This is given by:

\[ Q = \frac{\pi r^4 \frac{dP}{L}}{8\eta} \]  

where \( Q \) is the steady-state volume flow rate, \( r \) is the radius of the circular conduit, \( L \) is its length, \( \eta \) is the magma viscosity, and \( \Delta P/L \) is the pressure gradient along the conduit connecting the source to the volcanic pile. Several of these parameters were probably the same for the two volcanic systems. For instance, if both volcanic systems had a common source, it is unlikely that the length of the conduit varied dramatically for eruptions at the same elevation. Similarly, it is difficult to reconcile a higher pressure gradient operating in the 17th century volcanic system, especially when it is considered that the lithostatic pressure acting down on the source and pushing magma into the system, could not have been that different. If it is assumed that the rheology of the magmas were similar throughout the post-1600 period, an increase in the radius of the ascending conduit provides the only plausible means of easing the flow of magma. The effect of a wider conduit would be three-fold:

1. there would be a dissipation of frictional wall-effects,
2. the finite flow rate capacity of the ascent conduit would also be increased because of the larger cross-sectional area.
3. there would be an increase in the ambient rate of magma input. Circumstantial evidence for a higher ambient flow rate in the 17th century is forthcoming from the eruption rates of the long duration (> 1 year) eruptions.

Of the three effects outlined above, the most important with regards to 17th century activity must be point (2). By increasing the finite flow rate capacity of the ascent conduit, this offsets the threshold initial flow rate above which a discontinuity is incurred towards higher values. Therefore, in the mid-elevation vertical range considered here, whereas a post-1750 eruption whose initial real-time effusion rates exceeded the ambient flow rate is restricted to a finite duration (1852-53 (33) - 280
days), a 17th century flank eruptions occurring under a comparable pressure decay rate condition could survive for protracted durations without experiencing a discontinuity (1651-53 (9) - 3 years).

Even the larger finite flow ascent rate capacity of the 17th century magma ascent conduit was not great enough to maintain flow continuity in the volcanic system during the 1646-47 (6) eruption which occurred in this elevation range. The higher than normal initial pressures which characterised this eruption resulted in effusion rates at the surface which exceeded the finite flow rate capacity of the ascent conduit, and resulted in a flow discontinuity after only 58 days (Table 6.1, box D). In that time period, it had extruded a quantity of lava comparable in magnitude to that erupted by some of the protracted, multiple-year eruptions.

The interpretation of the differing eruption data in terms of a wider magma ascent conduit is reinforced when it is considered alongside the prevailing basement stress regime. Compared to the compressive stress regime which persisted in the post-1750 period, a wider ascent conduit for the 17th century period could best be accommodated in a relatively more dilational/tensile basement regional stress environment. Such changes in the basement stress field have been inferred to have occurred throughout Etna's history (Lo Guidice et al. 1982).

6.7 Lava Flows: Emplacement and Planimetric Evolution
6.7.1 Lengthening

A detailed examination of the how the final lengths of lava flows vary with eruption elevation (Figure 6.6A) and lateral distance (Figure 6.6B) has already been performed in the volumetric section of 6.4.3 and is not repeated here. Elaborating on the how these observations integrate with the erupted volume and real-time effusion rate arguments, for those lava flows which persisted to flow field status, it is apparent, after due consideration is given for those lava flows whose emplacement may have been hindered, that the unhindered 'cooling limited' lengths of the lava flows increased with decreasing elevation (Figure 6.6A). As seen for the lava flows of the post-1750 period, this increase in length correlates well with an inferred increase in the initial real-time effusion rate with decreasing elevation below the summit. Therefore, at elevations below the summit, the increase in the initial real-time effusion
rate meant that lava flows could achieve a greater length in that time interval during which the real-time effusion rate exceeded the restraining forces of crustal thickening at the flow front.

As for the lava flows which did not progress beyond flow unit, Type A status, the observations of shortening flow lengths with decreasing elevation and increasing lateral distances between the vent and the central conduit, may indicate that in addition to the restraining influence of a thickening crust at the flow front, these eruptions may also have been influenced by a 'minimum compressive stress' factor which was dampening the natural temporal decay rate of the real-time effusion rate by gradually reducing the width of the fissure.

In summary, the lengths attained by 17th century lava flows were influenced by two important constraining factors: the first being the nature of the predominant stress field during the eruption, and the second being the elevation of the eruptive vent on the volcano which influenced the real-time effusion rate. From a hazard viewpoint, a combination of low vent elevation and a stress environment which allowed for open fissures, gave rise to the conditions which produced the longest flow unit of the 17th century period - that of 1669 (10).

6.7.2 Widening and Thickening

Only the development of a few of the 17th century lava flows ended after the emplacement of the first flow unit (e.g. 1607 (1)): for the rest, planimetric evolution continued with phases of widening and thickening, culminating with the growth of a tumulus. For some of the Type B flow fields produced, the dense pahoehoe tumulus cover suggests repeated cyclical phases of planimetric growth along the lines of the above pattern.

Though only a few of the flow fields retain evidence for the initial stages of their planimetric growth, if, as inferred in previous sections, high level conditions in the 17th century volcanic system were similar to those present in the post-1750 period, it is therefore also likely that the lava flows underwent a similar planimetric growth pattern. Though the rate of magma input from depth was greater during the 17th century period, this would not manifest itself as a major controlling influence on lava flow emplacement at the surface during the initial stages of an eruption, for its
magnitude would be minuscule compared to that of the initial real-time effusion rate at the vent. Its effect, if any, would only be felt during the later stages of flow field development once the real-time effusion rates had declined to its magnitude. Therefore, in view of these constraining conditions, and drawing comparisons with the post-1750 period, at mid-elevations on the volcano, the initial lengthening and widening phases of the 17th century flow fields probably occurred in the first 62 days of the eruption, with that period after 62 days, during which the real-time effusion rate gradually decreased towards the levels of the magma input from depth, being occupied by tumulus development around the vent area. At elevations towards the top of the volcano, the tumulus building phase may have started before 62 days because of a smaller lengthening and widening phase, whilst near the base of the edifice, the greater magnitude of the initial two stages may have deferred tumulus construction to after 62 days. This duration range for the planimetric development pattern of the first growth cycle is a reflection of an elevational constraint on the initial real-time effusion rate factor which has the greatest controlling influence on initial lava flow emplacement.

After the first 62 days, it is likely that the real-time effusion rate at the vent was approaching the rate of magma input from depth. Therefore, for all eruptions which occurred above mid-elevations on the volcano’s flanks, it is likely that subsequent planimetric growth was occurring under conditions of steady magma flow. Systematic planimetric growth patterns should therefore be expected.

The stratigraphic relationship of flow structures within these flow fields suggest that planimetric growth following the initial cycle took the form of repeated tumulus construction. The construction of each tumulus was probably similar to that of the first, the growth cycle having a lengthening, widening and thickening phase. However, in contrast to the first cycle, the first two phases of the second and subsequent cycles were much reduced in size. In a detailed examination of the 1614-24 (3) lava flow field, Guest et al. (1984a) concluded that successive tumuli developed as parasitic growths from the front or side of the preceding tumulus. In this way, the tumulus cover was able to advance downslope over the products of the first growth cycle and, in the case of the 1614-24 (3) flow field, extended the length of the flow field beyond that determined by the first flow unit. A similar growth sequence is
suggested for the 1634-38 (6) and 1651-53 (9) lava flow fields but, in the case of the latter, because the first lengthening phase was greater than that of the 1614-24 (3) flow field, the presence of the first flow unit, which determines the overall length of the flow field, can still be seen in the distal portion of the flow field.

Not all 17th century flow fields evolved beyond the first growth cycle. Though the quantity of lava erupted into the 1669 (10) flow field was amongst the largest of 17th century period, the initial lengthening and widening stages was so large that it took up all the available magma. The magma supply was exhausted before the planimetric development of the flow field could evolve into the tumulus construction phase of the first cycle. Compared to the pahoehoe-textured, tumulus-dominated flow field of 1634-38 (6), the morphology of the 1646-47 (6) flow field, which was also erupted from a vent at 2,000 m asl, is quite marked, the enhanced magnitude of its first lengthening and widening phase reflecting its higher real-time effusion rates. In consequence, though the final volume of the two flow fields were similar, when the supply of magma to the 1646-47 (8) eruptions was discontinued, the real-time effusion rates had not decreased down to the levels of magma input rates required to initiate the tumulus construction stage.

In summary, it is concluded the variety of final morphologies displayed by 17th century flow fields are a reflection of various stages of planimetric evolution along a single evolutionary trend. Planimetric evolution occurred in response to a decreasing underlying real-time effusion rate. Post-1750 lava flow fields developed along a similar trend, but 17th century flow fields could progress further along the evolutionary growth pattern. This pattern takes the form of repeated growth cycles, each cycle being dominated by a lengthening, widening and thickening phase, and ending with the construction of a tumulus. Principal differences between flow fields reflect differences in the lengthening and widening phases of the first cycle. For eruptions whose planimetric evolution progresses beyond the first stage, it is noticeable that all cycles subsequent to the first, whilst smaller in magnitude, were of comparable size to each other.
6.8 Discussion

The final volume and duration of 17th century eruptions on Etna are affected by a variety of factors whose relative influence vary according to the stress field regime which prevailed at the time of the eruption. Eruptions which occurred at low elevations on the western sector (and possibly also the eastern sector) terminated when the magmatic pressure in the feeder dyke decreased below the volcanic/lithostatic load. This effected the closure of the eruptive fissure. Elsewhere on the volcano, where the controlling stress field allowed for open fissures, eruptions ended only when a magma flow discontinuity developed in the volcanic system. The conditions necessary to bring about such a discontinuity probably varied according to the relative difference between the real-time rate of magma output at the surface, and the real-time rate of steady state magma input at depth.

1. In circumstances where these two flow rate parameters are roughly equal, or differ by a small amount, volcano parameters such as dyke length, width and height, and the diameter of the ascending conduit, have little effect on the eventual size of the eruption at the surface. The steady conditions of magma output which generally characterise these types of eruptions, are more an influence of source conditions at depth in the volcanic system. Eruptions terminate in response to a change in source conditions, or a change in the high level stress field conditions.

2. In circumstances where the rate of magma output at the surface greatly exceeds the rate of magma input, a discontinuity arises either because the finite rate of magma ascent, as constrained by the maximum diameter of the ascent conduit connecting the source region to the volcanic pile, is persistently less than the rate of magma output at the surface, or the magma supply becomes exhausted.

The most efficient exploitation of the volume available in the whole volcanic system was effected by the eastern sector eruption of 1669 (10), at 825 m asl. The occurrence of this eruption, which coincided with a major tectonic displacement in the Etnean basement, decompressed the volcanic system to such an extent that in the ensuing 126 days, the 17th century volcanic system was effectively emptied of its available magma.

At progressively higher elevations on the volcano, the efficiency of eruptions in extruding the available volume of lava decreased (i.e. a smaller volume of lava
erupted in a given time interval). Drawing analogies with the post-1750 period, this was probably due to:

1. a reduction in the magnitude of the initial depressurisation component,
2. in the ensuing eruption, the volcanic system had to expend a greater amount of energy in raising magma to a higher elevation.

Nevertheless, in contrast to that seen in the post-1750 period, it is noticeable that 17th century eruptions with a high initial effusion rate component could persist for durations greater than one year. Post-1750 eruptions with comparable initial eruption characteristics ended up with curtailed durations because a narrow ascent conduit restricted the rate of magma flow upwards from depth. This resulted in an eventual discontinuity in the flow of magma to the eruption site. In the 17th century period, even though the initial rate of magma output probably exceeded the rate of magma input, the larger dimensions of the magma ascent conduit allowed for enhanced finite magma ascent rates. Therefore, certain 17th century flank eruptions which occurred at elevations above 2,000 m asl were able to continue after the initial high effusion rate phase, effusive activity gradually settling down to steady flow rate conditions as the effusion rate at the vent approached magma input rates. Once these conditions had been attained, protracted eruptions were possible as reflected by the 3 to 10 year duration of some of the eruptions.

However, in departure from normal conditions, even the larger finite flow ascent rate capacity of the 17th century magma ascent conduit was not great enough to maintain flow continuity in the volcanic system during the 1646-47 (6) eruption at 2,000 m asl. A sudden seaward movement of the eastern flank on the seaward side of the northeast rift resulted in the over-depressurisation of a gas-rich dyke resident in the rift. The high effusion rates which ensued as a result of the release of a large excess of magma pressure, probably exceeded the finite flow rate capabilities of the magma ascent conduit and a flow discontinuity developed in the system. Though the eruption was over in 58 days, in that time period, it had extruded a quantity of lava comparable in magnitude to that erupted by some of the protracted, multiple-year eruptions.

The most inefficient eruption of the entire period was that of 1614-24 (3). With a vent at 2,600 m asl, this eruption was elevationally one of the highest of the
period. Steady flow rate conditions prevailed during most of this eruption, the low vent effusion rate reflecting the rate of magma input into the volcanic system. However, because of the protracted 10 year duration of this eruption, the final quantity of extruded lava (some 1.3 km$^3$) exceeded that erupted during the most efficient eruption of the period, that of 1669 (10).

6.9 Summary: The 17th Century Volcanic System

In the 17th century period, magma was leaving the Etnean volcanic system at a mean rate of some 1.2 m$^3$s$^{-1}$. This is assumed to equate to the rate of magma input into the volcano. Throughout this period, the steady rate of output was mirrored by uniformity in the chemistry and petrography of the lavas: an observation which suggests that the transportation system conveying magma between the source and the volcanic pile was roughly static. However, the higher rate of output relative to that observed for the post-1750 period implies that magma could ascend more freely through the volcanic system. Assuming that the magmas of the two output periods had similar rheological properties, a wider ascent conduit could account for the increase in output of the 17th century period. To effect a wider conduit, the stress field in the Etnean basement must have differed, with a tendency towards a more dilational/tensile regime.

Above the basement, the distribution of eruptive activity at the surface was more uniform in the 17th century period, with long duration (> 32 days) and large volume (> 60 x 10$^6$ m$^3$) eruptions being sustained on both the buttressed western sector, and the unbuttressed eastern sector. A notion of higher magmatic pressures generated within a capacious high level storage area, which were sufficient to override the asymmetric high level stress field, cannot be used to explain this phenomenon for there is little evidence to suggest that the high level volumetric capacity of the two volcanic systems were dissimilar. A more fruitful explanation advocates that a more dilational/tensile basement stress regime was being transmitted through to the upper levels of the volcanic pile and was over-riding the asymmetrical high level gravitational stress field. This is the only situation which can readily explain the more extensive presence of eruptions on the lower eastern flanks. Further circumstantial evidence alights from the fact that many of the 17th century eruptions occurred on
lineaments which parallel known major basement tectonics. On the western sector, it is argued that the lateral predominance of the basement stress field may have been limited, for eruptions occurring on the lower western flanks were comparable to those of the post-1750 period. It is important to state that the high level gravitational stress field was not passive during this period. To generate the decompressive block movements required to explain the violent eruptions which occurred at the distal ends of the two rift zones, tectonic basement movements must have been accentuated by gravitational sliding of the eastern flank.

The predominance of a dilational/tensional basement stress field during this period meant that eruptions only terminated when the rate of magma ascent through the volcanic system was insufficient to overcome the rate at which magma was cooling within the feeder dyke connecting the eruptive vent to the central conduit. In consequence individual eruptions were able to survive the initial high effusion rate stage of the eruption and had the potential of continuing to lower real-time effusion rates which reflected the ambient magma input rate from depth. As seen for comparable post-1750 eruptions, the resulting lava flows underwent planimetric evolution from essentially flow unit (Type A) to Type B flow field morphologies. For the protracted 17th century eruptions of several years duration, planimetric evolution continued to a stage of pahoehoe tumulus development. Lava flows which did not evolve beyond the first Type A flow unit were only produced by those eruptions which occurred in the gravitational stress field, the eruption terminating before evolution could occur.
Final Comments

The large eruption of 1669 with associated major summit collapse, heralded a change in the configuration of the regional stress regime in the Etnean basement. This had a direct influence on those factors influencing the ascent and storage of magma in the volcanic system.

In the 17th century period which ended with this eruption, circumstantial evidence suggests that a dilational/tensile regional stress regime presided in the Etnean basement. In this regime, the rate of magmatic output was relatively high at 1.2 m$^3$s$^{-1}$, indicating that magma could move freely through the volcanic system. Near the surface, this particular regime was transmitted into the volcanic pile, and predominated over the gravitational stress field induced by the bulk of the volcanic pile itself. Basement controls on volcanic activity at the surface were most pronounced amongst those eruptions whose fissures were aligned with the major regional tectonic structures. In the absence of an externally imposed stress field, the configuration of the internal plumbing probably reflected hydrostatic conditions (Takada, 1989), with storage areas developing in those regions of the crust where the density contrast between the melt and the country host rock approached zero. Limited storage is suggested from the petrography of the erupted lavas (Guest and Duncan, 1981, Duncan and Guest, 1982).

Following the 1669 (10) eruption, activity on the flanks of the volcano quietened down with only one eruption occurring in the next 80 years. In this period, flank activity may have been subordinate to terminal activity as the volcano was undergoing reconstruction at the summit. When activity resumed on the flanks circa 1750, the rate of magma output was greatly reduced compared to the 17th century period and the erupted magmas showed little sign that they had been subjected to high level storage. These characteristics of the dynamic volcanic system are consistent with the presence of a more compressive regional stress regime in the Etnean basement, superimposed on the hydrostatic field (Takada, 1989). A compressive stress regime would act to reduce the width of the magma transportation conduit but would also effect an apparent increase in density contrast between the melt and the country rock so that magma could ascend to high levels with little trapping on the way. Near the surface, the affect of the compressive stress regime was to restrict the extent to which
basement tectonics could be transposed to the upper levels of the volcanic pile. In consequence, this resulted in the emergence of an asymmetric high level gravitational stress field (which had been secondary to the basement regional stress field in the 17th century period) as the predominant stress field which controlled eruptive styles at the surface in the post-1750 period of magmatic output.
CHAPTER 7

Conclusions and Suggestions for Further Work

7.1 General Conclusions

The aim of this work was to examine the inter-relationship between the way in which magma moves through a volcanic system and the emplacement of lava flows on the surface of the volcano. To this end, two periods in the post-1600 eruptive succession of Mount Etna volcano, Sicily, were chosen to study this problem. The mean magmatic output from flank activity was the prime distinction between the two periods, the output from 1600 to 1689 exceeding that of the post-1750 period. Using a variety of scientific means, the conditions which prevailed within the volcanic system during each period were constrained, and relationships sought between the timing of morphological changes in the development of a lava flow field and changes, as affected by the volcanic system, in the magnitude of time-dependent eruption parameters.

The results of the study, whilst revealing similarities in the way in which the post-1600 lava flows have planimetrically evolved, confirm that conditions in the volcanic system were different during each output period. These conclusions are presented in more detail in the following sub-sections.

7.1.1 Lava Flow Emplacement and Planimetric Evolution

1. The principal eruption parameter which exerts greatest influence on the emplacement and planimetric evolution of Etnean lava flows is the real-time effusion rate and its variation with time. A lava flow will only lengthen provided the basal shear stresses at the flow front exceed the yield strength of the cooling lava. It therefore follows that the higher the rate of effusion at the vent the farther the flow can lengthen before its motion is arrested. Assuming uniform cooling conditions,
there must therefore be a critical threshold effusion rate below which the rate of lava supply at the eruptive vent cannot feed the flow front at a fast enough rate to overcome the influences of the thickening crust. Once this stage is reached, any additional lava finds it easier to generate a new flow unit rather than to continue feeding the stagnating principal lava flow. The new flow generally develops as a lateral breakout in the vicinity of the vent area. The recurrence of this sequence of events leads to the planimetric development of the lava flow field.

2. The variety of lava flow field morphologies which have formed on the flanks of Etna since 1600 are interpreted to reflect various stages of planimetric development along a single evolutionary trend. The evolutionary sequence is predominated by cyclical phases of lengthening, widening and thickening, more or less in that order, the number of cycles being determined by the ability of the volcanic system to sustain the eruption. Each cycle ends with the development of a tumulus structure. The first tumulus develops in the vicinity of the vent area or, as in the case of the 1950-51 (54) flow field, at a break of slope below the vent area. The second tumulus generally forms from the first, on its downslope side. This precedent is followed by subsequent cycles, with new tumuli emerging from the front of the one that formed previously, though it may be that several tumuli were undergoing construction at the same time. For all the flow fields which evolve beyond the first cycle, it is noticeable that the second cycle is much smaller than the first. However, all cycles after the second, irrespective of the elevation of the eruptive vent, are of comparable magnitude. It is therefore likely that by this stage in the evolutionary cycle, the real-time effusion rate is steady, and uniform for all eruptions. In essence, where lava flow fields have a restricted planimetric evolution, the only difference between them is the size of the lengthening and widening phases of the first cycle. This difference arises because the magnitude of the initial real-time effusion rate is variable, being dependent on the elevation of the eruptive vent. The lower the eruptive vent, the greater the difference between the initial real-time effusion rate and the ambient rate of magma input into the system, hence, the greater the magnitude of the lengthening and subsequent widening phase of the first cycle.
For the post-1750 period of magmatic output, the degree of progress along the evolutionary sequence is influenced by the sector on which the eruption occurred. On the western sector, individual lava flows do not evolve beyond the lengthening phase of the first cycle. Only eastern sector lava flows have the means of progressing further but even here, flank eruptions rarely develop beyond the first cycle because most of the lava volume which these eruptions can access is taken up in the first lengthening and widening phase. It is mainly only near the summit, where the size of the first lengthening and widening phase is small, that eruptions can progress beyond the first cycle. The morphology of the resulting flow fields take on a pahoehoe textured, tumulus appearance.

A higher rate of magma input from depth allowed some of the 17th century flank eruptions, such as 1651-53 (9), to extrude lava flows which incurred several cycles of planimetric growth. Within these flow fields, evidence of the first cycle (whose lengthening phase was the largest), is largely buried beneath an expansive superficial tumulus cover which extends down to a low elevation on the flank. In the case of the 1614-24 (3) lava flow field, it may be that several cycles of tumulus construction extended the flow field beyond that which was achieved by the first lengthening phase.

7.1.2 The Post-1600 Volcanic System

1. Above the basement, eruptive activity in the post-1750 period has been characterised by a marked sectorial dichotomy of eruptive styles. Eruptions to the west of the rift zones have been limited to durations less than 32 days and volumes less than $60 \times 10^6$ m$^3$. In contrast, to the east of the rift zones, eruptions which occurred within 7.5 km of the central conduit, were capable of lasting durations and extruding lava volumes in excess of the western sector maximum. This sectorial dichotomy is inferred to be due to an asymmetric gravitational stress field which is present within the volcanic edifice at elevations above the basement. The asymmetry is induced by a non-uniform spatial distribution of volcanic load, and the gravitational instability of the eastern flank. On the buttressed western flank, eruptions end when the magma pressure within the volcano’s plumbing system falls below the level of the volcanic load. Conditions differ within 7.5 km of the central conduit on the
unsupported eastern flank because eruptive fissures on this region of the volcano can remain open at low magmatic pressures.

2. Between 1600 and 1689, the spatial distribution of eruptive activity on the flanks of Etna was more uniform, with long duration (> 32 days) and large volume (> 60 x 10^6 m^3) eruptions occurring on the western sector, and at low elevations on the eastern sector of the volcano. In addition, 17th century eruptions were, on the whole, of greater volume and longer duration than post-1750 eruptions at the same elevations and lateral distances. Activity in this 17th century period is inferred to have been dictated by a dilational/tensile basement stress field which was being transmitted into the volcanic pile, and was superseding the asymmetric high level gravitational stress field of the post-1750 period. On the buttressed western sector, the high level gravitational stress field was being over-ridden to a lateral distance of about 5 km from the central conduit but it may have predominated outside this lateral distance range. On the eastern sector, gravitational sliding was accentuating the effects of the basement stress field.

3. The ability of the volcano to sustain an eruption on its flanks is dependent on the inter-relationship between the magma driving pressure (P_m) and the minimum compressive stress (σ_3) acting across the fissure. On the post-1750 western sector, it was inferred that the short duration of the eruptions was a consequence of the fissure closing as the magma pressure in the dyke decreased below σ_3. Optimum eruptive conditions (largest erupted volume and longest eruption duration) on the post-1750 western sector of Etna occurred at an elevation of about 1,900 m asl. An elevational change in the magnitude of the volcanic load is considered to have been an important factor in giving rise to these conditions, for the optimum elevation coincides with a change in volcano morphology, from the steep-sided Mongibello construct above 2,000 m asl, to the gentler sloping lower flanks.

4. In situations where eruptive fissures remained open even at low magmatic driving pressures (as a result of a smaller σ_3) the persistence of the eruptions is inferred to be ultimately linked to a continuity of magma flow to the surface. When the rate of
magma flow through the eruptive fissure connecting the eruption site to the central conduit cannot overcome the rate at which the magma is cooling, the fissure becomes sealed and the eruption ends.

The major factor which limits flow continuity is considered to be the finite flow rate capacity of the ascent conduit which transports magma into the volcanic pile. The finite flow rate capacity is controlled by the cross-sectional area of the ascent conduit. Under conditions where:

a. the rate of magma output at the surface at any one time does not exceed the finite flow rate capacity of the ascent conduit, eruptions can, in theory, last indefinitely.

b. the rate of magma output initially exceeds the finite flow rate capacity of the ascent conduit, the ensuing eruptions are constrained to a finite duration.

In the post-1750 period, condition (a) was only satisfied for those eastern sector eruptions whose feeder dykes tapped into the top of the magma column in the central conduit. Though some eruptions were sustained for over a year, protracted eruptions did not materialise even under these ideal eruptive conditions because of the inherent weakness of the volcanic construct. Eruptions which tapped into the central conduit at lower elevations were constrained to a finite duration, the magnitude of which increased with decreasing elevation. This coincided with an increase in the eruptive efficiency of individual eruptions i.e. largest volume erupted in shortest time. Therefore, under the conditions of (b), the 'discontinuity factor' was deferred longest for those eruptions which caused greatest initial disruption to the equilibrium of the volcanic system, and whose vents were the shortest distance from the central conduit.

Similar initial conditions are inferred to have prevailed in the 17th century volcanic system, with eruptions low on the edifice being restricted to a finite duration. However, above 2,000 m asl on the flanks, eruptions which are considered to have commenced with a high effusion rate phase (of similar magnitude to post-1750 eruptions at the same elevation) could last for protracted durations of several years. Compared to the post-1750 period, the fact that a discontinuity did not arise during these 17th century eruptions is taken to indicate that the finite flow rate capacity of the magma ascent conduit was greater in this period (larger cross-sectional areas). Discontinuity conditions could therefore be deferred to higher magma flow rates, such
as attainable on the lower flanks. The notion of a magma ascent conduit with a larger cross-sectional area would also explain the higher ambient real-time effusion rates which characterised the waning stages of 17th century eruptions. A dilational/tensile basement stress regime is also consistent with these inferences.

5. The distinctive porphyritic textures of the 17th century 'cicirara' lavas have been interpreted as having formed within a high level storage area. The pervasive distribution of this texture throughout lavas which could only have been erupted when the volcano was behaving as an open system does not support this notion. This in itself negates the need for a high level storage area in this period of output. The eruption data itself does not provide any useful information about the high level internal plumbing of the volcano. On the premise that the lengthening stage of a lava flow's emplacement only takes up that volume stored in the upper levels of the volcanic system, the comparable final lengths of 17th century and post-1750 lava flows/flow fields at the same elevation may be taken to indicate that the high level volumetric capacity of the volcanic system was similar in both output periods. However, an alternative interpretation is that the comparable flow field lengths is a reflection of similar eruptive mechanisms. However, even this observation indirectly suggests similar high level plumbing systems between the two periods for differing eruptive mechanisms might otherwise be expected.

6. As previously observed by Walker (1974) and Lopes and Guest (1982), this research supports the notion that the magnitude of the initial real-time effusion rate increases as the eruption elevation decreases below the summit. An eruptive mechanism is proposed which attributes the magnitude of the initial driving force to a combination of hydrostatic and volatile exsolution pressures. In the absence of a contracting high level storage area, the exponential decay of real-time effusion rates are attributed to the decompression of a viscous column of magma located in the central conduit system of the volcano.

7. All the differences which exist between the 17th century and post-1750 volcanic systems are considered to be genetically linked to differences in the presiding
basement stress fields. The various characteristics of the volcanic system in these two output periods can be induced by changing this one variable. The change in the configuration of the presiding basement stress field probably accompanied a change in the regional tectonic stress field of eastern Sicily following the great eruption of 1669, towards one of greater compression.

7.2 Suggestions for further work

Though the following suggestions are forwarded with the eruption surveillance and eruption forecasting on Mount Etna volcano specifically in mind, they are equally applicable to other terrestrial volcanic systems, such as Mauna Loa and Kilauea on Hawaii and Piton del la Fournaise on Réunion Island, which possess similar eruptive styles.

7.2.1 Eruption Surveillance

Though some improvements to the methods followed in this thesis should be possible, and that some new developments could be spawned from this work, any real progress is dependent on a refinement of detail. This could only be brought about by improvements in eruption surveillance techniques, in the reporting of volcanic phenomena, and in the measurement accuracy of eruption parameters.

Despite the logistical difficulties associated with the surveillance of eruptive activity, existing reports available for the Etnan eruptions of this period contain a wealth of information about eruptive phenomena and the evolution of lava flow fields. A considerable amount of information remains to be extracted but, it is somewhat unfortunate that the information relevant to the exploitation of the present method is either poorly recorded, or not recorded at all. The availability over the last decade of helicopters for aerial surveillance and intercom communications on the ground has resulted in a considerable improvement in the monitoring of lava flow field evolution, as shown during the 1983 eruption of Mount Etna (Frazzetta and Romano, 1984) and the 1984 eruption of Mauna Loa, Hawaii (Lockwood et al. 1985). Nevertheless, many of the prospective gains from an improvement in technique are lost because the approach to eruption surveillance has, to date, lacked coordination.
In preparation for future eruptions, it is imperative that a standard approach to eruption surveillance is implemented. Amongst those measurements and information to be recorded should be:

*1 surveillance of the eruptive vent noting, in real-time, the effusion rate, degree of explosivity (and at summit), morphological changes to vent area and number of active lava flows being fed.

*2 simultaneous with *1, note the elevation of the active flow front(s) and its distance from the vent, velocity of advance, morphology of flow, flow width and thickness, and timing of planimetric changes to the flow field.

*3 frequent and systematic collection of lava flow samples (quenched) from the proximity of the vent area (explosivity allowing).

*4 Collection of real-time geophysical information for integration with *1 and *2.

The introduction of a standard approach to eruption surveillance would make large differences to the quality and quantity of information available for analysis, and a more detailed picture of the Etnean volcanic system could be obtained. For example, the frequent and systematic sampling of lavas during an eruption would allow the 'real-time' variation of petrological parameters (magma composition, degree of evolution, crystallinity and trace element content) to be observed. If it were possible to develop a technique to measure the 'real-time' variation of effusion rate at the vent the means would therefore be available, through the integration of eruption data with simple fluid dynamical models, to constrain the placement of an erupted sample of lava to a pre-eruption location within the volcanic system. This would help to identify positions within the volcanic system where magma mixing and differentiation of the volcanic products may be occurring. Presently, this is assumed to occur in the upper levels of the volcanic system. Clearly, this method need to be tested on a future eruption.

As an extension of the above method and the techniques initiated by Scott (1983) and Gyopari (1988), for a flow unit whose emplacement sequence is known within an eruptive episode, and whose flow front is well-constrained in space and time, a systematic sampling run along the length of the flow unit, from its flow front to the vent, should reveal samples which originate from progressively deeper regions of the volcanic system. Provided the information contained in eruption reports is of
sufficient detail to constrain the temporal sequence of individual flow unit
emplacement within a eruptive episode, this method would not only provide
information about the pre-eruption volcanic system but a comparison of information
from several eruptive episodes would permit some insight into any temporal changes
which might be occurring within the volcanic system between eruptions. In practice,
this method has significant advantages over the former in that the raw data is already
available.

7.2.2 Tectonism and Stress Fields

The importance of gravity sliding and the buttressing effect of adjacent
topographic highs has been long recognised in Hawaii (Fiske and Jackson, 1972;
Wadge, 1981). Etna has a wider geographical range of potential eruption sites than
Kilauea presumably as a result of complex basement tectonics interacting with the
local gravity fields. In addition, Etna’s location in a complexly active tectonic region
probably favours temporal changes in the high level plumbing. Accordingly, an
understanding of the relationships between eruptive activity, volcano-tectonism and
regional tectonism could well provide additional important clues to understanding the
changing tectonic regime of this region, and its interaction with the volcanic system.
Such information could be forthcoming from real-time seismological studies. The
present status of seismological research on Etna is reviewed by Gresta and Patane

7.2.3 Volcanic Hazard

This study also has implications for volcanic hazard analysis. For a volcano
like Etna with a long recorded history of flank eruptions, or for volcanoes where there
is a good chronology based on radio-carbon dating, statistical analysis of past events
provides a useful technique for developing models that can be used in volcano
forecasting. However, such studies could be misleading if they are averaged over two
or more periods of differing conditions. Assuming the internal conditions of Etna are
the same today as they have been since the mid-1700’s, forecasts for the immediate
future should therefore be based only on the study of this eruptive period. There is
therefore a need for a re-appraisal of the existing Etnean hazard maps of Guest and
Murray (1979) and Chester et al. (1985), to accommodate for these differing conditions.

As an important aside, it should be noted that between 1983 and 1990 there has been a sharp increase in flank eruption productivity. Only time will tell whether this is a small scale aberration or an indication of a distinct change in the effective supply. Monitoring of magma output, combined with a consideration of the composition and phenocryst content of lavas is of importance in recognising such changes. A return for example, to the production of 'cicirara' lavas could herald the onset of large volume eruptions, possibly low on the flanks, such as that of 1669.

Future work on Etna should involve the continued study of both the short term and long term changes in the styles of activity and their implications for the internal state. Such studies will enhance our ability to refine the Etna model.
APPENDIX 1 : VOLUMETRIC LAVA FLOW DATA FOR MOUNT ETNA; CALCULATIONS

This appendix presents the data used in the lava volume calculations. The final volume parameter used in both Part 1 and Part 2, is, where possible, a weighted mean of the available volume information. Details of the procedures followed in preparing the data for the various calculations, and the equations themselves are presented in chapter 3, section 3.2.5, and are not repeated here. For each eruption, each datum used in the volume calculation can be followed to its literary source by the superscript prefix. The reference referred to by each prefix is as follows:

*Scuito-Patti, 1867
b Silvestri, 1867
c Silvestri, 1879
d Oddone, 1910
e Cumin, 1943
f Cumin, 1950
g Di Rie, 1961
h Imbò, 1965
i Huntingdon, 1972 (unpublished data)
j Romano and Sturiale, 1975
kBottari et al., 1976
k Wadge, 1977
l Romano and Guest, 1979
m Tanguy, 1979
n Wadge, 1979
o Romano and Sturiale, 1982
p Wadge, 1981
q Wadge and Guest, 1981
r Romano and Sturiale, 1982
t Tanguy, 1983
u Frazzetta and Romano, 1984
v Patenté et al., 1984
w Tanguy and Patenté, 1984
x Lopes, 1985
y Romano and Vaccaro, 1986
z Caltabiano et al., 1987
** Bertagnini et al., 1990
APPENDIX 1: VOLUMETRIC LAVA FLOW DATA FOR MOUNT ETNA; CALCULATIONS

<table>
<thead>
<tr>
<th>Part 1: Mount Etna Lava Flows of the 17th Century</th>
<th>Year</th>
<th>Volume (x10^6 m^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1607</td>
<td>50±15  66±20  58±17</td>
</tr>
<tr>
<td></td>
<td>1610</td>
<td>31±9   23±7   29±9</td>
</tr>
<tr>
<td></td>
<td>1614-24</td>
<td>1500±450  1050±315</td>
</tr>
<tr>
<td></td>
<td>1634-38</td>
<td>200±100</td>
</tr>
<tr>
<td></td>
<td>1646-47</td>
<td>150±45  173±52  180±54  184±55</td>
</tr>
<tr>
<td></td>
<td>1651-53</td>
<td>480±144  426±128  414±124</td>
</tr>
<tr>
<td></td>
<td>1669</td>
<td>795±239  937±281  760±228  980±294</td>
</tr>
<tr>
<td></td>
<td>1689</td>
<td>15±5</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Part 2: Mount Etna Lava Flows from 1750 to 1990</th>
<th>Year</th>
<th>Volume (x10^6 m^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1763W</td>
<td>17±5  36±18</td>
</tr>
<tr>
<td></td>
<td>1763S</td>
<td>70±21  100±50  51±25</td>
</tr>
<tr>
<td></td>
<td>1766</td>
<td>122±37  75±23  94±28</td>
</tr>
<tr>
<td></td>
<td>1780</td>
<td>28±8  66±33  26±8  41±12</td>
</tr>
<tr>
<td></td>
<td>1792</td>
<td>80±24  97±29</td>
</tr>
<tr>
<td></td>
<td>1809</td>
<td>20±10  36±11  28±5</td>
</tr>
<tr>
<td></td>
<td>1811</td>
<td>40±20  48±24  50±25  38±19</td>
</tr>
<tr>
<td></td>
<td>1819</td>
<td>48±14  35±11  100±50  41±12  40±12</td>
</tr>
<tr>
<td></td>
<td>1832</td>
<td>54±11  40±8  39±8</td>
</tr>
<tr>
<td></td>
<td>1843</td>
<td>52±10  56±11</td>
</tr>
<tr>
<td></td>
<td>1852-53</td>
<td>120±36  125±25  130±30</td>
</tr>
<tr>
<td></td>
<td>1865</td>
<td>120±36  92±18  87±17  96±19  108±22</td>
</tr>
<tr>
<td></td>
<td>1879</td>
<td>42±13  31±16  25±7  26±7  38±11</td>
</tr>
<tr>
<td></td>
<td>1886</td>
<td>62±12  66±13  48±10  56±11</td>
</tr>
<tr>
<td></td>
<td>1892</td>
<td>63±31  111±22  120±35</td>
</tr>
<tr>
<td></td>
<td>1908</td>
<td>1.2±0.5  3±1.5  1±0.5</td>
</tr>
<tr>
<td></td>
<td>1910</td>
<td>44±13  40±12  60±18  31±10</td>
</tr>
<tr>
<td></td>
<td>1911</td>
<td>42±13  88±26  43±13  63±19  71±21</td>
</tr>
<tr>
<td></td>
<td>1923</td>
<td>53±15  97±30  49±15  78±23  90±30</td>
</tr>
<tr>
<td></td>
<td>1928</td>
<td>35±10  40±12  53±16</td>
</tr>
<tr>
<td>Year</td>
<td>Value A</td>
<td>Value B</td>
</tr>
<tr>
<td>---------</td>
<td>---------</td>
<td>---------</td>
</tr>
<tr>
<td>1947</td>
<td>11°±3</td>
<td>7°±3.5</td>
</tr>
<tr>
<td>1949</td>
<td>11°±3</td>
<td>8°±2</td>
</tr>
<tr>
<td>1950-51</td>
<td>116°±12</td>
<td>157°±47</td>
</tr>
<tr>
<td>1971</td>
<td>75°±35</td>
<td>50°±15</td>
</tr>
<tr>
<td>1974A</td>
<td>2.4±0.5</td>
<td></td>
</tr>
<tr>
<td>1974B</td>
<td>2.1±0.5</td>
<td></td>
</tr>
<tr>
<td>1975A</td>
<td>13°±4</td>
<td></td>
</tr>
<tr>
<td>1975B</td>
<td>5°±2</td>
<td></td>
</tr>
<tr>
<td>1975/76</td>
<td>12°±4</td>
<td></td>
</tr>
<tr>
<td>1976/77</td>
<td>6°±2</td>
<td></td>
</tr>
<tr>
<td>1978A</td>
<td>27°±8</td>
<td></td>
</tr>
<tr>
<td>1978B</td>
<td>3°±1</td>
<td></td>
</tr>
<tr>
<td>1978C</td>
<td>5°±2</td>
<td></td>
</tr>
<tr>
<td>1979</td>
<td>12°±4</td>
<td></td>
</tr>
<tr>
<td>1981</td>
<td>20±2°</td>
<td>18±3°</td>
</tr>
<tr>
<td>1983</td>
<td>100±20°</td>
<td>80±20°</td>
</tr>
<tr>
<td>1984</td>
<td>10±3°</td>
<td></td>
</tr>
<tr>
<td>1985A</td>
<td>30±10°</td>
<td></td>
</tr>
<tr>
<td>1985B</td>
<td>0.7°</td>
<td></td>
</tr>
<tr>
<td>1986/87</td>
<td>51±10°</td>
<td></td>
</tr>
<tr>
<td>1989</td>
<td>15±5°</td>
<td></td>
</tr>
</tbody>
</table>
REFERENCES


Changing styles of effusive eruption on Mount Etna since AD 1600

J.W. Hughes, J.E. Guest
Department of Physics and Astronomy,
University College London,
Gower Street,
London WC1E 6BT, UK

and

A.M. Duncan
School of Earth Sciences,
Luton College of Higher Education,
Park Square,
Luton LU1 3JU, UK

Abstract

The large eruption of 1669 with associated summit crater collapse, brought to an end a period in Etna’s history, starting in 1610, that was characterized by large volume eruptions and high output. A period of low output followed until the mid-eighteenth century, when it became moderately high and constant. These changes were broadly paralleled by changes in the petrography of the lava.

Investigations of the plumbing of Etna should properly account for these phases of differing eruption rate. Examination of the styles of eruption during the period 1750 to present shows that two types of lava flow-fields were produced: those that are relatively narrow, and those that are relatively broad. The distribution of these two types shows sectorial constraint, the latter type being absent on the western and northern flanks where eruptions are of shorter duration and less voluminous than those to the east.

Geological evidence suggests that the unsupported eastern seaward flank is less stable than the western and northern flanks that abut the mountains. Thus lower magmatic pressures are needed to sustain effusion on the eastern flank causing longer duration eruptions.

No sectorial distribution of flow types is found for eruptions during the seventeenth century and the much higher output during this period may suggest that the magmatic pressure generated within the volcano was high enough to maintain open fissures for long periods even on the buttressed western and northern flanks.

Contents

Introduction ................................................................. 386
General Geology of Etna .................................................. 386
Plumbing Models for Etna ................................................. 387
Post-1750 Activity ....................................................... 392
  Flow-field morphology ............................................... 392
  High-level magma movement and storage ......................... 396
Activity Between 1600 and 1700 ........................................ 400
Discussion .................................................................. 403
Acknowledgements ....................................................... 403
References ................................................................. 404
Introduction

During the last few decades, considerable advances have been made in understanding the plumbing of individual volcanoes. These advances have depended on the integration of geophysics (Einarsson, 1978; Ryan, Koyanagi and Fiske, 1981; Klein et al., 1987; Ryan, 1988), petrology (Scott, 1983; Wright and Helz, 1987), and the surveillance of eruptive activity. In addition, theoretical models (Shaw, 1980; Wilson and Head, 1981; Whitehead, 1986) of many aspects of volcanic processes have been developed and must await further testing and refinement.

Mount Etna, in eastern Sicily, has been the subject of an increasing research effort by Italian and other European scientists since the end of the Second World War. These studies have provided a considerable fund of information about its geology and eruptive activity. For a comprehensive review see Chester et al. (1985). Most of the techniques available for the study of volcanoes have been applied to Etna, although the geophysical studies have been less comprehensive than, for example, Kilauea volcano. Nevertheless, studies so far have led to a generalized model of the plumbing of Etna, showing it to be complex and variable with time (Guest and Duncan, 1981; Whitehead, 1986) of many aspects of volcanic processes have been developed and must await further testing and refinement.

Mount Etna is unique in that unlike other active basaltic volcanoes, it has a historic record of documented eruptions going back over 2000 years. In this chapter we discuss the changes in the style of activity since the beginning of the seventeenth century, the spatial distribution of different types of activity and the implications for near-surface plumbing of the volcano. The period 1600 to present is specifically chosen as this represents the most complete part of the record. Before this many of the products are covered by subsequent lavas.

General Geology of Etna

Etna has a maximum basal diameter of about 40 km and rises over 3300 m above sea level (a.s.l.). It rests on a southeasterly sloping basement and the outer flanks on the northern and western sides are against pre-existing mountains (Figure 1). The eastern and southeastern flanks extend to the coast. Etna is situated at the inter-section of three major regional fault trends: NW–SE, NE–SW and E–W (Grindley, 1973; Lo Giudice et al., 1982). The basement stress-field and associated fractures are thought to control the ascent of magma into the volcanic edifice and influence the distribution of flank eruption sites (Rittmann, 1973).

The earliest materials erupted from Etna were the Basal Tholeiitic Volcanics of Pleistocene age (Figure 2), which are considered to have formed several small overlapping basaltic shields; pillow lavas and hyaloclastites were produced in the east where eruptions took place under the sea (Cristofolini, 1973). A change from the eruption of tholeiites to lavas of the Alkalic Series occurred about 150 000 years BP (Condomines et al., 1982). Subsequently, a number of centres produced lavas and tephra of hawaiite, mugearite and benmoreite composition (Klerkx, 1970; Romano and Guest, 1979; McGuire, 1982).

The present volcanic edifice probably started to form 20 000 years BP. The volcano was truncated several times by major caldera collapse events (Guest, 1980) (Figure 2). A major sector collapse event occurred 5000 years BP on the eastern flank, producing an amphitheatre-shaped hollow, known as the Valle del Bove. It is some 1 km deep and 5 km across, and opens to the sea (Guest, Chester and Duncan, 1984). Major fault scarps subparallel to the coast are interpreted as strong evidence of the gravitational instability of the unsupported eastern flank (Guest, Chester and Duncan). The limited vertical extent of these faults was inferred by Lombardo and Patanè (1982) from the attitude of seismic isoanomalies on the eastern flank, which change from NW–SE trends at shallow levels (< 4 km) to the regional tectonic trend of NNE–SSW at deeper levels. They interpreted the shallow foci seismic events as being due to sliding on rotational slip planes.

Since the formation of the Valle del Bove, the style of volcanism has been dominated by effusive eruptions of basaltic lava with subordinate strombolian activity (Plates I and II). The summit cone has been in a state of near continuous historical activity, since at least 700 BC. In consequence, the summit region has rapidly evolved morphologically as a result of strombolian cone-building phases and pit collapse (Guest, 1982; Chester et al., 1985, chapter 4). Principally, because summit eruptions are not preceded by seismic crises, it has been suggested that summit vents are
northernly open to the central conduit of the volcano. The semi-continuous activity at the summit is known as ‘persistent activity’ (Rittmann, 1962).

Flank eruptions, in contrast, occur every few years. They are generally restricted to specific areas. One is a rift that extends from the summit to the northeast and another a rift extending in a southerly direction (Figure 3). Other areas of high vent concentration lie on the western flank and at low elevations on the southeast flank. In addition, a well-defined orthogonal pattern of fissures trend northeast and northwest at the head of the Valle del Bove and produce eruptions within the Valle and on its outer flanks (Cucuzza Silvestri, 1957; Guest and Murray, 1979).

It is clear from the above brief description of Etna that it has had a complex history and that the inferred plumbing is not therefore simple. This might be expected, since Etna sits in a region of tectonic complexity near the boundary between the African and Eurasian plates (Barberi et al., 1973; Lentini, 1982; Chester et al., 1985, chapter 3; Cristofolini et al., 1985).

**Plumbing Models for Etna**

Until the mid-1970s discussions of Etna’s plumbing were based on available petrological evidence. The Royal Society meeting about Etna in London during 1972 (Guest and Skelhorn, 1973), focused on the question of the existence of a high-level magma chamber. Rittmann (1973) argued that magma entered the high-level plumbing directly from depth with no major storage within the volcano. Cristofolini (1973), however, suggested that substantial high-level storage was required to account for the more evolved lavas and that the volumes required were greater than could be reasonably stored in a simple dyke system. Guest and Duncan (1981) consider that during the last 200 years there has been no significant high-level
Figure 2. Geologic map of Etna (modified from Romano et al., 1979)
Figure 3. Map of Mount Etna showing the number–density distribution of vents (modified after Guest and Murray, 1979) and tectonic lineaments (modified from Romano et al., 1979)
storage of magma but that in prehistoric times high-level storage of magma occurred associated with the generation of the more evolved magmas. It is likely that these high-level magma reservoirs were related to the caldera collapse events described above (Duncan and Guest, 1982).

Recent petrologic work has contributed further to an understanding of Etnean magmatic processes. Though lavas erupted since the fourteenth century have all been hawaiites of broadly uniform composition, analyses of lavas erupted during the last 20 years, suggest that larger eruptions tend to be more primitive. These may represent the influx of fresh magma into high levels (Armienti et al., 1984, 1987). The chemical differences are, however, very slight and there is still a pressing need for a careful assessment of analytical precision to ensure the validity of such temporal variations.

The variation in the petrographic nature and trace element chemistry of historic lavas is, in contrast, more apparent. Lavas erupted in the latter part of the sixteenth century and throughout the seventeenth century, are petrographically distinct from those erupted after 1700. These lavas are characterized by relatively large and abundant plagioclase phenocrysts, giving them a distinctive porphyritic texture. Locally this is termed ‘cicirara’ (Rittmann, 1973). The lavas erupted after 1700 show quite a different pattern of size-frequency distribution in the plagioclase phenocrysts. These differences are discussed below. (The lavas of the 1974 flank eruption are an exception in that they are virtually aphyric with no plagioclase phenocrysts (Tanguy and Kieffer, 1976).)

Detailed analyses of the textures and compositions of plagioclase phenocrysts have revealed that many of the lavas have more than one population of plagioclase. This has been interpreted (Duncan and Preston, 1985) as the result of magma mixing. That magma mixing is an important process in the high-level plumbing of Etna is also supported by the work of Scott (1983), who attributed the observed chemical variation in the 1981 lavas to the mixing of residual magma from the 1979 eruption with fresh magma that ascended in 1980.

An increase in K and Rb in lavas erupted since 1970 has also been noted (Tanguy and Clocchiatti, 1984; Cristofolini et al., 1984). Clocchiatti, Joron and Treuil (1988) argue that this enrichment is due to the selective contamination between magma and sedimentary country rocks. The contamination has to be selective since Sr-isotope ratios and other element compositions are unaffected. It is difficult to envisage quite how such a process of selective contamination would operate, and other workers (Cristofolini et al., 1987; Gyopari, 1988) consider that the enrichment results from mantle processes.

Based on a study of horizontal ground deformation in the summit region, Wadge (1976, 1977) suggested that the central conduit below the summit craters was an important storage area and that smaller volume (less than 50 x 10⁶ m³) lavas were fed by the drainage of the central conduit through radially oriented fissures. The volume erupted would then be limited by the amount of material available. This, in turn, would be a partial function of the altitude of the vent and the height of the magma column. According to Wadge, larger volume eruptions require a storage chamber near the base of the volcano.

In 1975 a programme of precise levelling was set up on the volcano to test for the presence of a high-level storage reservoir (Murray and Guest, 1982). The pattern of deformation over a 5-year period did not correspond to the well-established pattern at Kilauea (Fiske and Kinoshita, 1969), where it is interpreted as reflecting discharge from, and recharge into, a high-level magma storage area. However, the deformation on Etna did correspond to the filling and storage of magma in high-level dykes.

Seismic monitoring since the early 1970s provides further insights into the plumbing system (Cosentino et al., 1982). Flank eruptions are generally preceded by several days of increased seismicity, until magma breaks through at the surface (Cosentino et al., 1982). Summit eruptions, however, are generally preceded by harmonic tremor only. These patterns support the view that the summit craters are normally open to the central conduit, whereas flank-eruption magmas generally enter newly formed fissures.

The geophysical evidence therefore suggests that during the period of observation there has been no substantial magma reservoir within the volcanic pile. Although high-level storage has occurred in dykes, these are not concentrated in one specific area, and the position of active dykes varies temporally. In addition, it appears from geochemical evidence, that there can be interconnections between such storage dykes. This is
Plate I. A series of photographs illustrating the styles of activity observed at Mount Etna. (i) Top. Main flow channel of the 1983 eruption near the vent where it is beginning to form a tube. (ii) Bottom. The 1983 flow front invading woodland at a distance of about 3.5 km from the vent. (Photographs by (i) A.M.D. and (ii) J.E.G.)
Plate II. (i) Top. An oblique aerial view of the Etna summit taken in 1983 showing fumarolic activity at the Northeast Crater in the foreground and degassing from the Bocca Nova pit located centre right. The 500-m wide Chasm pit (centre) and the Southeast Crater (upper left) are relatively inactive. (ii) Bottom. Strombolian activity at the Northeast Crater in 1969 building a cone.

(Photographs by J.E.G.)
Plate I. A series of photographs illustrating the styles of activity observed at Mount Etna. (i) Top. Main flow channel of the 1983 eruption near the vent where it is beginning to form a tube. (ii) Bottom. The 1983 flow front invading woodland at a distance of about 3.5 km from the vent. (Photographs by (i) A.M.D. and (ii) J.E.G.)
ETNA'S STYLES OF EFFUSIVE ERUPTIONS

apparently the case for the 1981 eruption (Scott, 1983). However, the summit craters and their high-level feeding systems are thought to operate independently. For example, between 1966 and 1971, the Northeast Crater was erupting lava, whereas less than 300 m away, the main summit pit was drained of lava to some 100 m below its rim (Guest, 1973).

In 1908, the Messina earthquake produced compressional waves that reached Mount Etna. The local attenuation of P-wave velocities from the earthquake led Machado (1965) to infer the presence of a deep magma reservoir. A systematic search for a deeper magma storage area was made in 1977 using the travel times of natural and artificially induced seismic waves below the volcano. Analysis of the results are consistent with a large ellipsoidal storage area at 20 km depth with dimensions of $22 \, \text{km} \times 31 \, \text{km} \times 4 \, \text{km}$ (Sharp, Davis and Gray, 1980). The body may be interpreted to contain roughly 14% of melt trapped in an inferred network of fissures. This corresponds to about 1600 km$^3$ of magma. It appears that this reservoir region feeds the volcano, and produces a steady state output rate that during the 1970s was $0.7 \, \text{m}^3\text{s}^{-1}$ (Wadge and Guest, 1981).

Because geophysical observations have been available for Etna for a limited time, the interpretations may not of course correspond to the long-term plumbing. Changes in output, for example, have occurred before. For instance, between 1600 and 1669 the eruption rate was high, and resulted

![Figure 4. Volumetric output of Mount Etna since AD 1600 from flank activity only (modified after Wadge, Walker and Guest (1975) using an updated data set). Volumetric data for eruptions were collected from Cumin (1954), Wadge (1977, 1979, 1981), Tanguy (1979), Wadge and Guest (1981), Romano and Sturiale (1982), Frazzetta and Romano (1984), Tanguy and Patanè (1984), Lopes (1985), and Romano and Vaccaro (1986); each data point represented the weighted mean. The mean eruption rate for the seventeenth century period (A) is 1.19 m$^3\text{s}^{-1}$ compared with that for the eighteenth century onwards (B) of 0.18 m$^3\text{s}^{-1}$. The apparent increase in output from 1983 may indicate a change in the rate of magma resupply from depth.](image)
in several eruptions of long duration and large volume. Since about 1750 to the present, however, the output has been at a moderate and steady rate (Figure 4). Changes in output were accompanied by changes in the petrography of plagioclase phenocrysts discussed above.

Since 1600, therefore, the volcano has been characterized by two time spans with different rates of output from flank eruptions: 1600–1669 and 1750 to the present. During each of these periods, the eruption rate was broadly constant and the continuity of eruptive style suggests that the volcanic internal structure was roughly static. It is the purpose of this chapter to consider the styles of activity during these two periods in terms of the volcanic mechanisms operating. In particular, we assess factors controlling inferred plumbing regimes before and after the 1669 eruption.

Post-1750 Activity

Flow-field morphology

Most lavas erupted from the flanks during this period are aa. These lavas may be subdivided into four types. (i) Essentially simple flows (as defined by Walker, 1971) that are long relative to their width; (ii) complex (Walker, 1971) flows that are relatively narrow and have a well-defined central channel with numerous overflows; (iii) complex flow-fields that are relatively wide and consist of a number of branches having well-developed channels and overflows; and (iv) short, thick flows with well-developed ogives. Flow-fields that have mainly pahoehoe surface textures are also relatively broad, are complex in morphology and were fed by lava tubes giving rise to numerous ephemeral boccas. A further type of flow consists of a narrow ridge built up from overflows from a channel and tube system along the crest.

Of the 36 flank eruptions between 1750 and 1987, only one, that of 1974 (Figure 5(c)), produced a flow of type (iv). Extensive pahoehoe flows were only produced during the 1792 eruption. The long narrow ridge type flow (Figure 5(d)) was only produced in 1763. With these few exceptions, all the flow-fields are of types (i), (ii) or (iii). For the purposes of this study, we group the flows into two types (Figures 5 and 6): type A that are narrow compared with their length (to include type (i) and (iii)) and type B that are wide by comparison with their length (to include type (iii) and pahoehoe flows). The flows developed during the 1974 eruption and the unique 1763 south flank flow-field are excluded. The flow-field types A and B appear to correspond to the two populations described by Kilburn and Lopes (1988) identified from a statistical analysis of all historical flow-fields on Etna. The relatively long and narrow flow-fields (type A) are generally related to eruptions that had higher effusion rates, and shorter durations than those that produced the wide flows (type B).

Although many features of lava flows can be attributed to variation in the rheology of the flowing lava, the similar chemical composition and petrographic nature of the lavas suggest that they were within a restricted rheological range (Kilburn, 1984; Chester et al., 1985). The different morphologies observed are thus unlikely to be the result of their rheological properties.

Figure 7 shows the distribution of these two populations of flow-fields, erupted on the flanks of Etna since 1750. The type B flows only occur on the east and southeastern sector of the volcano within and surrounding the Valle del Bove, whereas on the north and western flanks only type A flows occur.

The spatial distribution of the two types of flow suggests that lava is erupted in a different way in the Valle del Bove sector as contrasted with the northern and western flanks. As first noted by Wadge (1981), most of the eruptions in the Valle del Bove tend to be of long duration (i.e., more than 50 days). This is illustrated in Figure 8, which shows that no eruptions on the western and northern flanks had durations of more than 50 days. The same pattern is also reflected in the volumes of the flow-fields (Figure 9). Flows on the northern and western flanks do not exceed $60 \times 10^6$ m$^3$ of lava, whereas most of those in the Valle del Bove are larger.

Both type A and type B flow-fields are normally produced by eruptions that start with a relatively high effusion rate (typically $>10$ m$^3$ s$^{-1}$); the initial flow usually travelling several kilometres in the first day or so. During that time the effusion rate decays and this is reflected in a decreasing rate of flow-front advance. Although some branching may occur as the result of overflow and topographic effects, the bulk of the flow-field is fed by a single open channel. The resultant flow-field is of type A. If the eruption continues, back
Figure 5. Example sketches of the four morphological flow types observed in post-1750 flank eruptions: (a) long, narrow flow of 1947; (b) complex broad flow-field of 1865; (c) one of the thick 1974 flows with surface ogives; (d) the short thick flow of 1763 with a channel-tube system along the flow crest. Arrows indicate the direction away from vent.

pressure or topographic effects (Guest et al., 1987) may cause substantial branching and the development of new flows. Diversion of lava into the new branches then terminates supply to the main channel, which subsequently drains. Successive branching may occur, forming a complex flow-field of type B. With continued effusion, the main channel above the level of branching usually roofs over forming a lava tube and, as the effusion rate decays, a complex tube system may ultimately develop giving rise, in turn, to numerous short flows superimposed on the original flow-field. Once lengthening has ceased, the additional lava volume contributes to the thickening and widening of the flow-field around a fan of distributary tubes. The difference in morphology between type A and type B flow-fields is thus the result of duration, changing effusion rates and available volumes.

On the eastern flank, therefore, although long duration eruptions produced relatively wide flow-fields starting with a high effusion rate, this rapidly decays over a matter of days to a steady and low effusion rate representative of most of the eruption. For eruptions on the western and northern flanks, however, eruptions stop before a sustained period of low effusion rate occurred.

A fundamental conclusion of the above obser-
Figure 6. Aerial photographs of (a) the 1947 type A flow on the northern flank and (b) the 1865, type B flow on the eastern flank.

The difference in eruption style is interpreted as reflecting interactions between the volcanic system and the regional and localized gravitational stress fields. The uniform distribution of lithostatic load on the buttressed western and northern flanks effects the closure of eruptive fissures on these flanks once the magma pressure within the feeder dyke has fallen below a critical threshold.

As already noted, geological evidence from the unsupported eastern and southeastern flank indicates that considerable seaward gravitational sliding has been occurring for more than 15,000 years. Guest, Chester and Duncan (1984) identified several factors that have had some influence on the rate of sliding, including excess loading by lavas on the flanks and repeated high rates of magma injection into the rift zones. In addition, the presence of the deep topographic depression of the Valle del Bove has modified the local gravi-
tional stress field, and there is evidence that this has at least had some influence on the attitude of eruptive fissures in its vicinity (McGuire, 1983; McGuire and Pullen, 1989). Therefore, eruptive fissures in the eastern sector, in the proximity of the central conduit system and along the rift zones, have the potential of remaining relatively dilated at lower magmatic pressures, allowing longer eruption durations. Since the eruptive fissure does not close after the initial phase of high effusion rate, as happens with the western and northern sector eruptions, the eruptions can continue, albeit at a reduced rate, by magma ascending from beneath the volcano. The effusion of lava ceases when the magma pressure is insufficient to maintain eruption. The magnitude of eastern sector eruptions is therefore inferred to be strongly dependent on the volume of the ascending magma batch, and the efficiency by which it migrates through the volcanic system to the eruption site.

It should be noted that a few of the eruptions that have occurred in the eastern flank were of short duration (Figure 8), especially those in the rift zones. This is to be expected since the rate of gravity sliding of the eastern flank may vary and there may be differential block movements within.
High-level magma movement and storage

Evidence suggests that most flank eruptions are fed through dykes in contact with the central conduit at a relatively high level. Apparent exceptions occur, such as the eruption of 1974 (Guest et al., 1974). The lack of plagioclase phenocrysts in the 1974 lava is thought to be due to magma rising directly from depth where the $P(H_2O)$ was high enough to suppress plagioclase crystallization to near solidus temperatures (Tanguy and Kieffer, 1976; Duncan and Guest, 1982). The 1974 feeder system, therefore, is not considered to be linked to the central conduit at shallow depth, where storage or slow ascent of magma at a low pressure and degassing through the summit craters would promote plagioclase crystallization. For eruptions since 1750, the observation that summit activity ceases at the onset of flank eruptions, supports the view that flank dykes are fed from the central conduit. Making this assumption, the minimum horizontal length of the dyke may be taken to approximate to that of the horizontal distance of the lowermost flank vent from the central crater. In some cases, such as the 1983 eruption, the dyke length may be obtained directly by following the path of surface fractures (Murray and Pullen, 1984).

The inferred dyke lengths feeding each of the eruptions since 1750 are plotted in Figure 10.
against the total volume of material erupted from the fissures. For those eruptions where lava effusion occurred simultaneously from separate locations, the maximum length to the lowermost active vent was considered only. There is a tendency, therefore, to underestimate the total length of dyke involved. Most of the data plotted in Figure 10 cluster on one of two separate trends. The steeper trend represents eruptions in the eastern sector and the shallower trend is for those of the western and northern sector. Only eruptions that have occurred at the head of the Valle del Bove (as discussed below), and the western eruptions of 1947 and 1949 depart from the observed trends. For both sectors, Figure 10 indicates that the closer the vent is to the central conduit, the greater the volume of lava erupted. It follows from this general relationship that the maximum volume of magma available for eruption in each of the two sectors is represented by the point where the best-fit lines meet the ordinate (Figure 10). For eruptions close to the central conduit, most lava must come directly from the central conduit. The further the vent is from this conduit, the greater the length of dyke. In addition, dykes that have their 'head' above the vent altitude will, of course, drain gravitationally.

The linear relation between the volume erupted and the length of the feeding fissure, supports the view that there were well-defined constraints on

Figure 8. Spatial distribution of post-1750 flank eruptions in terms of eruption duration
the magnitude of an eruption at a given radial distance. If this is correct, the maximum volume of magma available for an eruption during this period was about $170 \times 10^6 \text{ m}^3$. The lower ‘maximum volume’ of $60 \times 10^6 \text{ m}^3$ for western and northern flank eruptions suggests that after the largest possible eruption there may still be up to some $110 \times 10^6 \text{ m}^3$ of magma available for subsequent activity. The value of $170 \times 10^6 \text{ m}^3$ for eastern sector eruptions may represent the eruptible fraction of the largest arriving batch. The negative slope of each of the two curves may therefore be a function of material that remains in the system after eruption. Thus the longer the undrained flank-feeding dyke, the greater the ratio of material remaining in the system to that erupted.

For western and northern sector eruptions, it is conceivable that the entire volume of material erupted had prior storage in the edifice. Wadge (1977) estimates that the central conduit magma volume is about $42 \times 10^6 \text{ m}^3$. This is close to the volume required. In 1971, however, the Chasm within the central crater was empty to the 1 km depth level prior to eruption (Murray and Guest, 1982) indicating that the central conduit was only partially charged within the edifice. Only those 1971 vents that were at a lower elevation than the magma level within the central conduit could have conceivably been fed by a falling magma column.

Figure 9. Spatial distribution of post-1750 flank eruptions in terms of erupted volume.
A substantial volume could be held in dykes as suggested by the geophysical evidence and this combined with that in the central conduit could thus explain the observed volume. For eastern flank eruptions, the situation is different as some tens of major dykes would be required to hold the necessary magma. It appears reasonable then, that for eruption in this section, up to $110 \times 10^6 \text{m}^3$ of new magma is supplied once the stored magma has been drained.

Residual magma left in dykes after eruption may be subsequently available for future mixing. Evidence for the process of magma mixing, from the textures and compositional ranges of plagioclase phenocrysts, has already been discussed. The limiting factor would be the residence time of the magma before cooling and solidifying. A recent study of the trace element content of lavas erupted between 1910 and 1983 provides further evidence of mixing (Gyopari, 1988). On the basis of incompatible trace element ratios, Gyopari (1988) has identified mixing between several end-member compositions within single eruptive episodes. In certain examples, one end member can be related to an earlier eruption. In the 1971 eruption, for instance, fresh magma appears to have mixed with magma that had incompatible trace element ratios that were characteristic of magma erupted in 1923 and 1928. If the model of Gyopari is correct, then batches of magma may remain isolated within dyke systems for at least 50 years, and thus be available for subsequent mixing in later intrusive and eruptive events. Theoretical calculations (Wilson and Head, 1988) show, however, that dykes up to 5 m in thickness emplaced into a volcanic pile are likely to solidify within 2.5 years. These batches of mixing magma, therefore, would need to be relatively thick (perhaps pod-like) and stored deeply. This is somewhat similar to the situation in the East Rift Zone of Kilauea.

![Graph](image-url)

Figure 10. The total volume of lava erupted during an individual eruption as a function of the horizontal distance of the most distant eruption vent from the central conduit. The best-fit line A only applies to those western sector flank eruptions that occurred at or below 2000 m a.s.l. Wadge (1977) considers this elevation as that where high-level eruptive forces favour optimum disruption and exploitation of the central conduit storage system. Above 2000 m a.s.l. western eruptions become increasingly inefficient at releasing the whole volume of magma available for eruption, for example, 1947 and 1949. For the eastern sector, an optimum exploitation level of 2300 m a.s.l. is suggested for these flank eruptions from the data used to fit B. The Valle del Bove group eruptions include only those that were erupted from an orthogonal set of fissures (see Figure 11 and text).
Figure 11. Distribution of eruptive vents in the Valle del Bove. These eruptions, restricted to an orthogonal set of fissures trending northeast and northwest, form a separate structural grouping distinct from the eastern and western flank eruptions (see text for discussion). The 1985b 'summit' eruption was only included to highlight the fissure trends where there is evidence that fresh magma mixes with isolated batches of residual magma (Wright and Helz, 1987).

It was noted earlier that a small number of eruptions do not fit the pattern described above (Figure 10). Each of these eruptions occurred near the head of the Valle del Bove (Figure 11) in association with two orthogonal fissure and dyke systems trending northeast and northwest (McGuire, 1983). With these eruptions, effusion takes place successively from several widely spaced localities from dykes that trend in both directions. Thus the injection of dykes to feed these effusions appears to be complex. Taking the distance from the central conduit to the eruption site does not necessarily represent the total path length traversed. Thus, lack of conformity of these eruptions to the trends shown in Figure 10, and the fact that less volume is erupted than would be predicted for the eastern sector, suggests that the complex fissure pattern retains more magma than for other eruptions.

### Activity Between 1600 and 1700

As shown in Figure 4, the first 70 years of the seventeenth century was characterized by a high output. This period was preceded by a period of low output during the second half of the sixteenth
century. It, in turn, was succeeded, after the 1669 eruption, by another period of low output, lasting into the mid-eighteenth century. During the seventeenth century, in addition to the high output, the volcano was in a state of flank eruption for over 20% of the time between 1600 and 1669. This represented a sevenfold increase in the percentage of time spent in flank eruption from 1750 to present.

The difference in the size frequency of plagioclase phenocrysts between seventeenth century lavas and those erupted after 1700 is illustrated in Figure 12. The lavas of the 1689 eruption show 'cicirara' texture and petrographically relate to lavas produced during the period of high eruptivity that ended in 1669. The contrasting plagioclase crystallinity between the seventeenth century lavas and those erupted after 1700 clearly indicate a difference in the nucleation and crystal growth history of the magma. In silicate melts, the rates of crystal growth and nucleation are virtually zero at the liquidus and, with progressive undercooling ($\Delta T$), increase to a maximum and then decrease (Kirkpatrick, 1977). The peak of the growth rate curve occurs at a smaller $\Delta T$ than that of the nucleation rate curve. The seventeenth century magmas may, therefore, have risen to a relatively shallow depth where the pressure was sufficiently low for plagioclase to be a near liquidus phase. The large plagioclase phenocrysts, however, indicate a lower $\Delta T$, thus inhibiting nucleation but enhancing crystal growth. Such a growth history could be explained by a high magma input rate and storage in a substantial high-level reservoir, both of which would inhibit cooling. After 1700, there was a lower rate of replenishment and limited storage of magma in the central conduit system and the associated dykes. This could allow for greater cooling during high-level ascent and lead to a larger $\Delta T$ that would promote higher nucleation rates. This, in turn, should result in numerous small plagioclase phenocrysts.

Three spectacular eruptions account for most of the output during the seventeenth century. The first of these was the 10-year eruption that started in 1614 (Guest, Wood and Greeley, 1984). This originated on the northeast rift and produced some 1-2 km$^3$ of lava covering some 21 km$^2$ of the northern flank. Most of the flow-field is surfaced by pahoehoe fed from extensive tube systems. The average eruption rate was relatively low, about 6 m$^3$s$^{-1}$, but effusion rates at individual bocca operating contemporaneously must have been much lower. The flow-field is particularly outstanding because of the presence of mega-tumuli and perched lava lakes, as described by Guest, Wood and Greeley, (1984). A similar flow-field was produced during the 1651–1653 eruption on the west flank of the mountain with similar effusion rates.

The third major eruption destroyed part of Catania in 1669. It erupted from relatively low on the southern flank (about 700 m a.s.l.), and produced an extensive type B flow-field, with a volume of about 1 km$^3$. This eruption, like that of 1646 on the northern flank, was accompanied by violent magma degassing and the production of a large cinder cone.

Although these three eruptions dominate the record, nearly all the eruptions during this period were of long duration, comparable to those that
occurred in the eastern sector from 1750 onwards. Most occurred on the rifts but three were western flank eruptions. In contrast to the post-1750 lavas, the seventeenth century lavas do not, in terms of flow morphology, duration, or volume (Figure 13), show strong sectorial control in their distribution.

The large phenocryst lavas, the high output, the occurrence of long duration eruptions separated by short repose periods, the styles of eruption and the distribution of flow-field types, all suggest that the volcano's stress field during the seventeenth century may have been markedly different from that of the mid-eighteenth century onwards.

Compared with the period 1750 to the present, the seventeenth century plumbing of Etna appears to require a higher magma production rate, longer high-level residence times and stress conditions within the volcanic pile permitting long duration eruptions from the rift systems and the western flank. Clearly, the interplay in the gravitational and tectonic stress field within the volcano and

Figure 13. Spatial distribution of seventeenth century flank eruptions around the volcano in terms of the two morphological flow types. Similar distributions are revealed when eruption duration and erupted volume are considered; type A eruptions are characterized by durations <50 days and volumes <60 x 10^6 m^3. All type B eruptions exceed these volumes and durations.
the magmatic pressure of stored material was different in the two periods.

Although the long duration eruptions on the rifts may be related to a seaward gravitational sliding of the eastern flank and the resulting relatively open conduits within the rift systems, the 1651–1653 west flank eruption requires that the magmatic pressure remained high enough to maintain a long duration eruption on the buttressed western flank. The existence of such a system is consistent with high rates of eruption during this period and the petrographic features of the lava.

Hence, from 1600 to 1669, magmatic pressures within the volcano were high enough to dominate the distribution and style of eruptions occurring at that time. The rapid emptying of the volcanic system during the 1669 eruption relatively low on the flanks appears to have brought this situation to an end, either as the result of rapid draining, or a change in tectonically induced strain (or both). Nearly 100 years later, the volcano resumed its steady state eruptive pattern at a much lower rate.

Discussion

It is clear from the geological studies of Mount Etna and the historical record that different styles of activity, thought to represent different states of the plumbing, have occurred at various times.

Several conclusions may be made. Although it is generally assumed that a major control on flow-field morphology is the rheology (and thus the composition) of the erupted material, this work indicates that lavas of similar composition and rheology can produce different flow-field forms dependent on such factors as the eruption rate and the duration of the eruption. These factors are not independent and in turn are controlled by the combined characteristics of the ascent and storage pathways (the volcano's plumbing). In consequence, consideration of flow-field morphology can be a relevant tool in making broad inferences about the internal state of a volcano (see also Wadge, 1977).

The importance of gravity sliding and the buttressing effect of adjacent topographic highs has been long recognized in Hawaii (Fiske and Jackson, 1972). Etna has a wider geographical range of potential eruption sites than Kilauea, presumably as a result of complex basement tectonics interacting with the local gravity stress fields. Etna's location in a complexly active tectonic region probably favours temporal changes in the high-level plumbing. Accordingly, an understanding of the relations between eruptive activity and regional tectonism could well provide additional important clues to understanding the changing tectonic regime of this area.

The study also has implications for volcanic hazard analysis. For a volcano such as Etna, with a long recorded history of flank eruptions, or for volcanoes where there is a good chronology based on radiocarbon dating, statistical analysis of past events provides a useful technique for developing models that can be used in volcano forecasting. Such studies could be misleading, however, if they are averaged over two or more periods of different conditions. Assuming the internal conditions of Etna are the same today as they have been since the mid-1700s, forecasts for the immediate future should therefore be based only on the study of this eruptive period. It should be noted, however, that between 1983 and 1987 there was a sharp increase in flank eruption productivity. Only time will tell whether this is a small-scale aberration or an indication of a distinct change in the effective supply. Monitoring of the eruption rates, combined with a consideration of the composition and phenocryst content of lavas, is of importance in recognizing such changes. A return for example, to the production of 'cicirara' lavas could herald the onset of large volume eruptions, possibly low on the flanks, such as that of 1669.

Future work on Etna should involve the continued study of both the short-term and long-term changes in the styles of activity and their implications for the internal state. Such studies will enhance our ability to refine the Etna model.

Acknowledgements

The authors thank C.R.J. Kilburn for lively discussion on this topic, as well as M.P. Ryan, R.I. Tilling and anonymous reviewers for helpful comments on the manuscript. We are grateful to Valerie Peerless for assistance with typing and D. Rooks for photographic work. The work of J.W.H. was supported by a UK Natural Environment Research Council Postgraduate Studentship.
References


