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Melt inclusions from volcan de Colima, Mexico: complex examples of magmatic differentiation

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FLORIDA INTERNATIONAL UNIVERSITY

Miami, Florida

MELT INCLUSIONS FROM VOLCÁN DE COLIMA, MEXICO: COMPLEX
EXAMPLES OF MAGMATIC DIFFERENTIATION

A thesis submitted in partial fulfilment of the

requirements for the degree of

MASTER OF SCIENCE

in

GEOLOGY

by

Zachary D. Atlas

2001

To: Dean Arthur W. Herriott
College of Arts and Sciences

This thesis, written by Zachary D. Atlas, and entitled Melt Inclusions From Volcán de Colima, Mexico: Complex Examples of Magmatic Differentiation, having been approved in respect to style and intellectual content, is referred to you for judgment.

We have read this thesis and recommend that it be approved.

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DEDICATION

This thesis is dedicated to my loving wife Silvia. Her patience, support and steadfast realism have allowed me to keep my goals in sight and to strive for the highest possible achievement.

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ABSTRACT OF THE THESIS

MELT INCLUSIONS FROM VOLCÁN DE COLIMA, MEXICO: COMPLEX EXAMPLES OF MAGMATIC DIFFERENTIATION

by

Zachary D. Atlas

Florida International University, 2001

Miami, Florida

Professor Gautam Sen, Major Professor

Melt inclusions are minute magma bodies trapped within growing crystals. Their chemical compositions are useful in deciphering pre-eruptive conditions and magma evolution. The present study examined melt inclusions trapped in phenocrysts from the 3rd and 4th magmatic cycles (1869-1988) at *Volcán de Colima*, Mexico. Melt inclusions have highly evolved chemical compositions: 65-77% SiO₂, >12% Al₂O₃, 3-6% Na₂O and K₂O and less than 5.5% Fe and Mg. Major element compositions suggest that they are strongly differentiated magmas controlled by fractionation of plagioclase, opx, cpx and hornblende. Water concentrations were measured to be 2.7-3.5 wt. % in cpx hosted inclusions and 0.3-0.7 wt % in opx and plagioclase. Trace element compositions are anomalously low and inversely correlate with water. From this we deduce that Colima lavas and scorias simultaneously differentiate and degas. Moreover, hornblende rim growth rates constrain the ascent of the Colima magmas to ~100 days for passive eruptions and >4 days for plinian eruptions.

TABLE OF CONTENTS

CHAPTER	PAGE
1: INTRODUCTION.....	1
2: GEOLOGY AND HISTORY OF THE COLIMA VOLCANIC COMPLEX.....	3
2.1 Tectonic setting and development of the Colima Volcanic Complex.....	3
2.2 Magmatic products and their interpretation.....	7
3: MELT INCLUSION STUDIES.....	12
3.1 The significance of melt inclusions.....	12
3.2 Limitations of Melt Inclusion Studies.....	16
4: METHODS.....	19
4.1 Sample Preparation.....	19
4.2 Rehomogenization experiments.....	22
4.3 Microprobe analysis.....	25
4.4 Ion probe Analysis.....	29
4.5 Laser Ablation ICPMS.....	31
5: PETROGRAPHY.....	33
5.1 Whole Rock Samples.....	33
5.2 Melt Inclusion Petrography.....	37
6: RESULTS.....	44
6.1 Experimental Run Results.....	44
6.2 Major Elements.....	44
6.3 Trace and Rare Earth Elements.....	55
6.4 Water in Melt Inclusions.....	62
CHAPTER 7: DISCUSSION.....	64
CHAPTER 8: CONCLUSIONS.....	80
REFERENCES.....	81
APPENDICES.....	89

LIST OF TABLES

TABLE	PAGE
Table 4-1: Temperature, f_{O_2} and run duration for all controlled fugacity, 1 atm. experiments.	23
Table 6-1: Run results for DI-194 samples (1869 lava flow).	45
Table 6-2: Run results for controlled fugacity experiments.	45
Table 6-3: EPMA analysis of secondary crystals in plagioclase MI's.	46
Table 6-4: EPMA Analysis of melt inclusions in natural samples from <i>Fuego de Colima</i> (1869-1913)	48
Table 6-5: EPMA analysis of melt inclusions from controlled fugacity rehomogenization experiments on sample DI-194 (1869).	49
Table 6-6: EPMA analysis of 1 atmosphere re-homogenization experiments.	50
Table 6-7: Major and trace element analyses of melt inclusions.	56

LIST OF FIGURES

FIGURE	PAGE
Figure 2-1: Placement of Colima complex within the triple rift system and southern Colima graben.	4
Figure 2-2: Map view of the Colima rift valley.	5
Figure 2-3: Eruptive cycles of <i>Fuego de Colima</i> for the past 500 years.	9
Figure 2-4: SiO ₂ vs. K ₂ O diagram for <i>Nevado de Colima</i> and <i>Fuego de Colima</i> .	10
Figure 2-5: Plots of major element variation for the last two major magmatic cycles of <i>Volcán de Colima</i> .	11
Figure 3-1: Rhyolitic pumice, ignimbrite & tephra from the western United States and their respective melt inclusions	14
Figure 3-2: Andesite and Rhyolite Melt Inclusions.	15
Figure 3-3: Comparison of transmitted light image and back-scatter image.	18
Figure 4-1: Sample DI-194 showing plagioclase with clear core, zoned rim and ring of melt inclusions	20
Figure 4-2 Design of a platinum-wound one-atmosphere gas-mixing furnace.	21
Figure 4-3: Plot of log f_{O_2} versus temperature.	23
Figure 4-4: Probe diameter measurements.	27
Figure 4-5: Sodium loss calibration curves.	28
Figure 4-6: H ₂ O calibration graph for NIST 610, M3N and M6N standards.	30
Figure 5-1: Photomicrograph of samples DI-194, 1004-407 & 1004-414.	34
Figure 5-2: Photomicrograph of sample M82-11 and JT 91-68.	35
Figure 5-3: Photomicrograph of sample 1004-421.	36
Figure 5-4: Photomicrograph of plagioclase with a high concentration of melt inclusions in its core from sample Col-9D.	39

Figure 7-5: Plot of selected trace elements from <i>Colima</i> melt inclusions versus H ₂ O.	71
Figure 7-6: Plot of H ₂ O versus K ₂ O for <i>Colima</i> and <i>Popocatepetl</i> melt inclusions	72
Figure 7-7: Water saturation curves.	74
Figure 7-8: Plot of phase stability for <i>Colima</i> , <i>Mt. St. Helens</i> (MSH) and for basaltic andesite.	78
Figure 7-9: Ascent rate curves derived from hornblende rim growth.	79

CHAPTER 1: INTRODUCTION

Volcán de Colima (Mexico) is situated within a triple rift system adjacent to a subduction zone. This unusual tectonic setting coupled with a history of regularly occurring eruptions has made it an intriguing target of study for volcanologists. The *Colima-Cantaro* Complex consists of three volcanoes that have evolved southward along the rift valley through time. Historically, each of these volcanoes shows evidence for periods of effusive (strombolian type) activity punctuated by major eruptions (Luhr and Carmichael, 1990). Large dome collapse events have also destroyed each cone at least once during their development (Luhr, 1993). This pattern of activity makes *Colima* an important natural laboratory in need of further study. Furthermore, the proximity of these volcanoes to major population centers makes their study important for hazard mitigation.

Colima lavas form a tight group of medium K andesite (Luhr & Carmichael 1990, Luhr and Kyser 1989, Verma & Luhr 1992, Robin et al. 1993, Robin et al. 1990). Numerous chemical analyses on lavas and tephra have shown distinctive temporal trends for the past 200 years (Luhr & Carmichael 1990). To date, no study has fully addressed the melt inclusions found in the phenocrysts of *Colima* lavas. Melt inclusions have been used by a number of workers in other parts of the world to infer chemical conditions occurring in the magma chamber prior to eruption (Nielsen et al, 1985, Roedder, 1984 Sours-Page et al, 1998). Therefore, a study of *Colima's* melt inclusions was deemed important for deciphering the subsurface magmatic processes that cause its recurring major eruptions.

The present work focuses on melt inclusions entrapped in phenocrysts from block lavas erupted in 1869, 1880 and from one scoria sample erupted in 1913. This study presents the major element compositions of *Colima* melt inclusions. It also includes the first report of ion probe analysis of concentrations of water and trace elements in *Colima* melt inclusions. These data, in conjunction with analytical petrography of the hornblendes found in these lavas, are used to decipher crystallization versus degassing history of *Colima* lavas.

CHAPTER 2: THE COLIMA VOLCANIC COMPLEX

2.1 Tectonic setting and development of the Colima Volcanic Complex

Fuego de Colima is part of a system of volcanoes collectively known as the *Colima volcanic complex*. The complex includes *Volcán Cantaro*, *Nevado de Colima* and the presently active *Fuego de Colima* and is located within the southern arm of a triple rift system in west central Mexico (Fig 2-1). The triple rift includes the Colima graben, the Tepic-Zacoalco rift (including *Volcán Ceboruco*) and the Chapala Rift (including Lake Chapala and associated geyser fields). The *Colima* rift is approximately 50 km wide. It is bounded by a series of N-S trending normal faults in the north and NE-SW trending faults in the southern portion of the rift valley (Bandy et al., 1995). *Fuego de Colima* lies central in the rift valley located at 19° 30' 44" N, 103° 37' 02" W. The complex is approximately 150 km northeast of the Middle America Trench and is closer to the trench than any other volcanic center in the Trans-Mexican Volcanic Belt. *Fuego* and *Nevado de Colima* stand approximately 4000 m high making them the main topographic features in the surrounding valley. The remainder of the rift valley is spotted with small cinder cones typically less than 500 m high. Cinder cones have been intensely studied (Luhr & Kyser, 1989; Luhr & Carmichael, 1981, 1982) and many of these products are chemically unlike the magmatic products erupted from the main vents.

Volcanic activity in this area has shifted southward through time (Fig 2-2) from *Cantaro* in the north to *Nevado* (central) to *Fuego de Colima* at the southern end of the complex (Luhr & Carmichael, 1980 & 1990; Robin et al., 1982; Calanchi et al., 1995).

Figure 2-1. Placement of Colima complex within the triplereft system and southern Colima graben. Figure modified from Macias et al., 1993. The top part shows the location of Colima complex relative to the TMVB.

Figure 2-2: Map view of the Colima rift valley showing the location of Fuego de Colima, Nevado de Colima and V. Cantaro relative to the graben walls (fault scarps), cinder cones and nearby population centers. Red Dashed arrow show the relative direction of shift in the focus of activity for Colima over the past 1.7 Ma. (Figure modified from Luhr & Carmichael, 1980)

All volcanic activity in the area is of Quaternary age, with activity at *Volcán Cantaro* beginning approximately 1.7 Ma b.p. (Calanchi et al., 1995). *Fuego de Colima* began erupting contemporaneously with the most recent *Nevado* cone less than 80 thousand years ago (Robin et al., 1987). Unfortunately, all of the initial traces of activity at *Fuego de Colima* were buried or erased by the avalanche collapse of the prehistoric cone between ~9300 and ~4300 years ago (Robin et al., 1987; Luhr, 1993), and therefore, this age is still in question. Subsequent to the collapse, minor activity commenced on the southern flank of *Fuego de Colima* (*Los Hijos de Colima*), additionally illustrating the southern shift of volcanic activity (Luhr & Carmichael, 1990). During the last 100 years activity at *Colima* has been confined to the central vent of *Fuego de Colima*.

A number of hypotheses have been proposed to explain this interesting tectonic environment. One hypothesis proposes that part of the East Pacific Rise (EPR) has “jumped” eastward over the subduction zone and has induced rifting in western Mexico (Luhr et al, 1985). Further, this ridge shifting has been occurring for the past 10-12 Ma and is part of a major plate reorganization that initialed with the separation of Baja California in the Miocene. A recent paleomagnetic investigation offers evidence of continental rifting occurring in this area as far back as the Upper Cretaceous (Housen & Beck, 1999). The possibility exists that this area has long been a location for major continental plate reorganization. Other hypotheses suggest that tensional forces associated with subduction of the Rivera-Cocos plate boundary created the triple rift (Bandy et al., 1995). It has also been proposed that oblique subduction of the Cocos plate is causing shear forces that created and perpetuates the rift (DeMets & Stein, 1990). These issues have yet to be resolved.

2.2 Magmatic products and their interpretations

A considerable amount of work on lavas from the Colima complex has been performed (Luhr & Carmichael, 1980; Luhr & Carmichael, 1981; Luhr & Carmichael, 1982; Luhr & Carmichael, 1990; Luhr and Kyser, 1989; Verma & Luhr, 1992; Robin et al., 1990; Robin et al., 1992). A brief description of prehistoric and modern lava from the complex is given to provide a context in which we may interpret melt inclusion data. Lava from *Nevado de Colima* has varied from basic andesites (57.5 wt. % SiO₂, Robin et al., 1987) containing abundant phenocrysts of plagioclase, orthopyroxene and clinopyroxene to more silicic andesites and dacites (61-64.3 wt. % SiO₂) containing abundant amphibole (Robin et al., 1987; Calanchi et al., 1995). Like *Nevado*, Fuego's phenocrysts also vary over time. Some lava flows contain little or no amphibole, while others contain abundant amphibole (up to ~10% hornblende, Luhr & Carmichael, 1990). Still other flows contain only amphibole relics (Luhr and Carmichael, 1990). *Colima* andesites also contain varying amounts of orthopyroxene, clinopyroxene, plagioclase and titanomagnetite.

Since 1818 *Fuego de Colima's* eruptions have been well documented (e.g., Luhr & Carmichael, 1980 & 1990, Robin et al., 1982, Calanchi et al., 1995). Based on such documentation, Luhr and Carmichael (1990) state that major eruptions (VEI >3) occur every ~100 years with dome building events thereafter for 40-50 years. They define this as the eruptive cycle of *Fuego de Colima*. During the cycle the volcano is never dormant. Active degassing of the volcano occurs almost constantly. Measurements of SO₂ flux have been regularly monitored at *Colima* for the past few decades. Taran et al. (2000)

reports SO₂ flux to be approximately 100 tons/day during the period 1991-1994. In periods of activity SO₂ flux can be much higher and have been measured at >10,000 tons/day (Smithsonian GVP, volcanic activity report). For 40-50 years prior to a major eruption, lava flows and minor phreato-magmatic eruptions (VEI<2) occur (Luhr and Carmichael, 1990, Robin et al., 1987). Large eruptions (VEI = 4) close out the eruptive cycles and have occurred at *Fuego* in 1585, 1606, 1622, 1818, 1890 and 1913 (De la Cruz-Reyna, 1993). Figure 2-3 shows the eruptive cycle of *Fuego de Colima*.

Chemically *Fuego de Colima* lavas are mainly medium-K andesites (Fig. 2-4). As with eruptive style, major element composition can also be seen to cycle over the period of the last 200 years (Fig. 2-5) (Luhr & Carmichael, 1990; Robin et al., 1987). Each chemical cycle begins with the eruption of basaltic andesite magmas [these scorias have SiO₂ ≤ 59%, MgO ≈ 4.0%, FeO ≈ 6.5 % (Luhr & Carmichael, 1990)]. Succeeding eruptions become more evolved throughout this magmatic cycle (SiO₂ increases to 60-62%, MgO decreases to 2.5-3.5% and FeO decreases to 4.5-5.5%). We hereafter use the term “magmatic cycle” to refer to the chemical cycle. We also prefer magmatic cycle starting eruptions to refer to the violent eruptions of 1818 & 1913 eruptions because it is believed that these primitive magmas represent new magma batches fed into the chamber (Robin et al., 1991; Robin & Potrel, 1993). Hypotheses concerning the origin and evolution of these lavas include (1) fractional crystallization of plagioclase + hornblende + clinopyroxene + titanomagnetite (Luhr, 1992; Calanchi et al., 1995; and Macías et al., 1993); and (2) magma mixing coupled with fractional crystallization (Luhr & Carmichael, 1990; Verma & Luhr, 1993; Robin & Potrel, 1993; Robin et al., 1991)

Figure 2-3. Event cyclicity of Fuego de Colima for the past 500 years. Figure from Luhr and Carmichael 1990

Figure 2-4: SiO_2 vs K_2O diagram (After Gill 1981) showing lavas from Nevado de Colima and Fuego de Colima. Scorias and minettes from the Colima Cinder Cones. Colima MI's (this study) are also plotted for comparison (□ = 1 atm. runs, ♦ = CF runs, ● = inclusions in cpx, ● = in opx, ○ = in hornblende & ● = in plagioclase inclusions).

Figure 2-5: Plots of major element variation for the last two major magmatic cycles of *Volcan de Colima*. Also included is the first eruption of the second major magmatic cycle (1818). Data from Luhr and Carmichael (1990). Arrows point to lava flows analyzed for melt inclusions in this study. Dashed lined represent the boundary between each magmatic cycle and mark the first eruption of the new cycle.

CHAPTER 3: Melt Inclusion Studies

3.1 The significance of melt inclusions

Geochemical studies of igneous rocks provide us with an insight into the nature of volcanic processes. Although lavas record source characteristics, differentiation and contamination processes, important information about the magmatic history may be lost or overprinted by irreversible near surface processes, such as magma degassing. In some cases significant inferences about the pre-eruptive magmas may be obtained through the study of melt inclusions (MI's) (Nielsen et al, 1995; Lee & Stern, 1998; Gioncada et al., 1998). MI's are small pockets of silicate glass that are entrapped by rapidly growing crystals (Roedder, 1984). They are commonly found in olivine, plagioclase, quartz and diamond but can occur in almost any other crystal under the right conditions of formation (USGS melt inclusion page, from internet). Early studies reported MI's to be a sample of the initial melt from which the phenocryst grew (Sommer, 1977). More recent studies demonstrate that MI's are representative of the magma coexisting in equilibrium with the growing crystal at any given point in time (Kamenetsky et al., 1997; Vaggelli et al. 1993; Roedder, 1984, 1991; Vogel & Aines, 1996; Dunbar & Hervig, 1992).

Melt inclusions have been extensively studied in phenocrysts of mid ocean ridge basalts (MORB) (e.g., Sours-Page et al., 1999; Kamenetsky et al., 1997; Nielsen et al., 1995; Sinton et al., 1993), rhyolite (e.g., Vogel & Aines, 1996; Dunbar & Hervig, 1992), and mantle xenoliths (e.g., Schiano et al., 2000; Schiano & Bourdon, 1999; Varela et al., 1999). In these settings MI's are chemically similar to or more primitive than bulk magmas. They have been interpreted as being either primary melts (Nielsen et al., 1995;

Sinton et al., 1993) or residual melts whose composition has been controlled by crystal fractionation and magma mixing (Sours-Page et al., 1999; Vogel & Aines, 1996; Dunbar & Hervig, 1992; Webster & Duffield, 1991) (Figs. 3-1).

In contrast to the above, the present study is concerned with MI's found in phenocrysts of andesitic rocks. Unfortunately, only a few studies have been performed on andesite hosted MI's (Sisson & Layne, 1993; Finny, 2000; Wardell et al., 2001). These prior studies have shown that MI's in andesites are very different than MI's found in MORB, rhyolite and mantle xenoliths. The former have wide ranges of chemical composition and in many cases contain more than 75% SiO₂ even though their host magmas show no signs of this extreme composition (Figure 3-2). In general, andesitic MI's are highly evolved and seem to represent products of extreme magmatic differentiation.

Of particular importance to MI studies is their volatile content. If melt inclusions are representative of the melt from which the phenocryst grew (or representative of the evolving magma), then the volatile components contained in melt inclusions must represent magmatic volatile contents (Webster & Duffield, 1991). Because volatiles have drastic effects on silicate melts, knowledge of their pre-eruptive concentrations is crucial to our understanding of magmatic phase equilibria (Holloway & Blank, 1994). An important application of dissolved water content of melts is that the depth at which a MI is trapped may be deciphered from such data (Roedder 1984; Sommer, 1979, Dunbar & Hervig, 1992; Roggensack, 2001; Wardell et al., 2001). In this way, volatile contents may be helpful for determining the pressures at which differentiation processes occur.

Figure 3-1: Rhyolitic pumice, ignimbrite & tephra from the western United States and their respective melt inclusions plotted on a K₂O-SiO₂ variation diagram (Gill, 1981). Data represents samples from the Bandelier tuff, Taylor Creek and Southwest Nevada Volcanic Field (SWNVF). Colima lava (+) and MIs (□ 1 atm. runs, ◇ CF runs, ● non-homogenized samples are plotted for reference (see Fig. 2-3 to distinguish between cpx, opx, hbl., and plag. hosted inclusions in non-homogenized analyses)

Figure 3-2: Andesite and Rhyolite Melt Inclusions from several continental and ocean island arcs plotted with averaged MORB compositions from the Mid-Atlantic ridge (MAR), East Pacific Rise (EPR) and Indian Ocean ridge (IOR). Colima lavas and scorias + and Colima MI's [$\circ$ 1 atm. runs, $\bullet$ CF runs & $\circ$ non-homogenized samples are shown for reference (see Fig. 2-3 to distinguish between cpx, opx, hbl. & plag. hosted inclusions)]. (Colima data from Luhr & Carmichael, 1990 and this study)

3.2 Limitations of Melt Inclusion Studies

Although it is apparent that the study of MI's is highly desirable, certain obstacles remain. 1) Melt inclusions are difficult to analyze. Because MI's are very small (most MI's measure 100 μm to $<25 \mu\text{m}$), even with the smallest beam size, a "mixed signal" between MI and host phase is often possible. MI's are also difficult to detect in reflected light and often backscatter imaging (Fig 3-4).

2) MI's may be compromised by a through-going crack or they may contain secondary crystals. As with the above, this will cause significant variation in composition across the MI. Many authors re-dissolve (rehomogenize) secondary crystals in the MI's in a high temperature furnace or 1 atmosphere heating stage assembly (Nielsen et al., 1995, Roedder, 1990 and 1984, Sours-Page et al., 1999). However, at atmospheric pressures, the vapor typically exsolves from the melt in a MI creating a post-entrapment (or shrinkage) "bubble" (Roedder, 1984). Although vapor bubbles are indicative of rehomogenization (Roedder, 1984), this should not be considered confirmation of rehomogenization as vapor bubbles are common in many MI's (Nielsen et al., 1998; Dunbar & Hervig, 1992; Vogel & Aines, 1996; and others). Therefore, these "bubbles" also must be avoided during analysis for the same reasons stated for secondary crystals.

3) Re-heating is generally performed on mineral separates. During mineral separation fractures can develop. Upon re-heating melting can preferentially occur along these fractures, significantly altering their composition. Highly incompatible elements, volatile elements and vapor can leak out through these cracks (Nakamura & Shimakita, 1998). Furthermore, the furnace atmosphere (which is easily exchanged through the

fractures) can also significantly change iron content of the MI and host effecting equilibrium (Nielsen et al., 1998). This can lead to an inaccurate interpretation of MI's chemistry.

MI's visible in both visible light and Back Scatter Imaging

Figure 3-3: Comparison of transmitted light image and back scatter image (compositional mode) from EPMA. Sample 1004-414 from Smithsonian Institution.

CHAPTER 4: METHODS

4.1 Sample Preparation

The eight samples used in this study were obtained from the Smithsonian Institution and from Florida International University archives. These samples represent block lava erupted from the *Volcancito* (DI-194 & 1004-414) in 1869, the 1880 block lava eruption (M82-11, 1004-407 & JT 91-68) and one scoriaceous sample from the cycle starting 1913 eruption (1004-421). Each of the samples (with the exception of DI-194 and JT 91-68) has been well studied and their whole rock analyses appear in Luhr & Carmichael (1990). Thomas & Draper (1991 field study) collected sample DI-194 from the lava flow of the *Volcancito* near the base of the "Plyon" adjacent to the caldera wall on the NE flank of Fuego de *Colima*. Sample JT 91-68 was collected from the base of the 1880 flow nearest to the caldera wall (adjacent to the 1913 deposit). All samples were cut into thin sections and the abundance of melt inclusions were inspected visually (Fig. 4-1). Because melt inclusions are abundant in plagioclase only these grains were chosen for rehomogenization experiments. Whole rock samples of DI-194 were crushed by hand and plagioclase (between 40-70 mesh) was separated with a Frantz Magnetic Separator. Plagioclase containing devitrified inclusions tend to have higher magnetic susceptibility than expected for pure plagioclase. Therefore, the lowest two magnetic fractions were used. Plagioclase grains were then hand picked and placed in a platinum boat and set aside for experimentation. Atmospheric re-heating experiments were conducted in a platinum wound 1 atm. furnace (Fig. 4-2). Runs were conducted under atmospheric conditions and controlled oxygen fugacity (f_{O_2}) at various temperatures and

Figure 4-1: Sample DI-194 showing plagioclase with clear core, highly zoned rim and ring of melt inclusions (isotropic globules). Magnification is 100x in crossed polarized light

Figure 4-2 Design of a platinum-wound one-atmosphere gas-mixing furnace. (figure modeled after Gangopadhyay (1999) Masters Thesis). Figure not drawn to scale

for different run duration. Oxygen fugacity was controlled using a mixture of purified CO₂-H₂ gas.

4.2 Rehomogenization experiments

Six atmospheric runs (uncontrolled f_{O_2}) were performed at 1200° C, 1150° C, 1140° C, 1125° C, 1100° C, and 1000° C. Procedures used for these experiments were modified following methods used by Sinton et al, (1993) and Dunbar and Hervig (1992). Temperatures were adjusted after optical analysis of the prior run. Initial runs are approximately 4 hours duration (after Nielsen et al., 1995). However in a later study, Nielsen et al. (1998) recommended run duration less than 2 hours to preserve initial water contents. Therefore, all controlled fugacity runs were limited to 40 minutes to avoid volatile loss. Controlled oxygen fugacity experiments were performed at descending intervals of 10° C starting at 1200° C. Oxygen fugacity was set +1 Log unit above the NNO buffer (Fig. 4-3). Metered values of CO₂-H₂ gas and corresponding f_{O_2} values are given in Appendix A (Table A-1). The f_{O_2} buffer was chosen to closely match the conditions determined for the *Colima* magmas (Luhr & Carmichael, 1980). Table 4-1 lists the run temperatures, f_{O_2} conditions and run duration.

Starting temperatures were chosen from the wealth of available data on *Colima* lavas. Magmatic temperature of andesitic lava from *Volcán de Colima* have been reported to be between 880°C and 1200°C using pyroxene and Fe-Ti Oxide thermometry (Luhr & Carmichael, 1980 & Luhr, 1992). In general, Luhr & Carmichael (1980) report temperatures for *Colima* andesites to be between 940°C and 1052°C. Robin et al. (1990), concur with the above and report temperatures of crystallization of *Colima*'s acidic

Figure 4-3: Plot of $\log f_{O_2}$ versus temperature. Experimental runs and Colima lavas are shown relative to QFM and NNO buffers. Experimental runs plot approximately at NNO +1 (Dashed Line).

Run	T (°C)	f_{O_2}	Duration (min)
DI-194 CF 1	1140 ± 1	-8.41 ± 0.21	40
DI-194 CF 1b	1150 ± 1	-8.35 ± 0.23	31
DI-194 CF 1c	1150 ± 1	-8.35 ± 0.23	64
DI-194 CF 2	1130 ± 1	-8.75 ± 0.06	40
DI-194 CF 3	1120 ± 1	-8.81 ± 0.15	40
DI-194 CF 4	1110 ± 1	-9.89 ± 0.01	40
DI-194 CF 5	1100 ± 1	-9.34 ± 0.35	45
DI-194 CF 6	1090 ± 1	-9.11 ± 0.11	40
DI-194 CF 6b	1090 ± 1	-9.12 ± 0.04	62
DI-194 CF 7	1080 ± 1	-9.25 ± 0.18	45
DI-194 CF 7b	1080 ± 1	-9.24 ± 1.25	62

Table 4-1: Temperature, f_{O_2} and run duration for all controlled fugacity, 1 atm. experiments.

andesites, between 918°C and 982°C based on pyroxene thermometry and 1040°C - 1120°C using Fe-Ti Oxide thermometry. Accordingly, the run temperatures were chosen to be >900°C and < 1200°C. Rapid quenching of the runs (< 0.5 sec.) in distilled water ensured homogenization of the glass in the MI's. However, vapor bubbles were always observed to be present.

Calibration runs were conducted prior to each set of experiments. Temperature during calibration was measured using an internal Pt-Rh_{5%} thermocouple and external, Oxytrol Systems reference thermocouple. The calibration procedure determines the “hot spot” and the temperature where control and reference thermocouples are equal. This is accomplished by moving the reference thermocouple up, through the center of the furnace tube. This procedure determines temperature gradients within the furnace (Gangopadhyay, Masters Thesis, 2000). The external thermocouple was plugged into a voltmeter to simultaneously measure f_{O_2} . Final values for f_{O_2} were determined according to the Nernst equation (Appendix A). Finally, the liquidus temperature of pure diopside (1391.5°C) was used to calibrate exact temperature. Temperature variation in the furnace was found to be precise to $\pm 1^\circ\text{C}$. Appendix A contains data on the calibration runs and calculation of oxygen fugacity.

Experimental runs were inspected under a petrographic microscope and aliquots separated for mounting in epoxy. Plagioclase grains were randomly chosen for mounting to avoid sampling bias. Grains were then mounted on petrographic slides using petropoxy and polished to expose MI's. Procedures for exposing melt inclusions were difficult and sometimes resulted in breaching or destroying the MI itself. One method commonly used by other investigators is double polishing the phenocryst to expose the

MI on both sides. This method has been helpful in obtaining H₂O and CO₂ analysis using FTIR spectroscopy (Dixon & Clague 2001, Gioncada et al. 1998, Vogel & Aines 1996). Nielsen, et al (1998) cautioned that double polishing is an arduous, time consuming task that sometimes results in micro-fractures thereby breaching the MI. Therefore *Colima* samples were polished to expose MI's on one side only. All run products were analyzed for major elements, trace elements and water as outlined below.

4.3 Microprobe analysis

Melt inclusions were imaged and analyzed using a JEOL 8900R electron microprobe at Florida International University's Florida Center for Analytical Electron Microscopy. Analytical conditions on the JEOL 8900R were 15 KeV at 20 nA beam current, using a point mode beam. Count times for major elements are 10 seconds (Na, K, Ca, Mg, Fe, Ti, Mn & Al) and 20 seconds for silicon. Standard measurements were taken on the SPI-2 standard block. The standards used on SPI-2 were obsidian glass, diopside, albite, magnetite and rhodenite (see Appendix B for standard compositions). Standard compositions were gathered at least bi-weekly (due to current fluctuations, filament changes or consistently poor totals on host materials) and elements were re-peaked prior to daily analysis. Because volatility of Na in natural glasses tends to be quite high (Nielsen & Sigurdsson, 1981), this element was analyzed for 10 seconds during the first beam pass using a high-count crystal (PETH). Line scans were performed at ~5 μm intervals. Point analyses were performed after the line scans to confirm the chemistry of the MI's. Subsequent spots were located in compositional backscatter mode and are visually identical to the initial spot.

Sodium loss was measured for various beam diameters using obsidian glass standard from SPI-2 at 15 kV and 20 nA. Each beam diameter was measured by fluorescence on the benitolite and quartz standards. Beam diameter measured 5.8 μm for point mode and 22.8 μm for a beam diameter of 50. However, due to the distinct fluorescence of the different materials this measurement can be as much as 50% in error (Fig. 4-4). Therefore, the smallest beam sizes have a $\pm 3 \mu\text{m}$ error while the largest beam diameter is $\pm 7.5 \mu\text{m}$. These analyses reveal considerable Na loss (1.5 - 2.0 wt %) during the first 10 seconds of analysis using a point beam ($5 \pm 3 \mu\text{m}$) (Fig 4-4). Larger beam sizes (e.g., beam diameter = 20 & 50) tend to minimize Na loss in the obsidian glass but because of the large error associated with the measured beam size, these beam sizes are too large for analysis of MI's smaller than 50 μm in diameter.

Na loss was also measured as a function of time in obsidian glass (Fig. 4-5). After 10 seconds of analysis count rates drop to 1/2 the initial count rate observed at 2 seconds. However, because the beam needs 2 seconds in order to stabilize prior to the time when counting begins, this drop may reflect loss after four seconds. If beam stabilization is considered to have an effect on the sodium loss (as it should), then total beam time of 4 seconds corresponds to a loss of only 1/4 of the counts. It was found that Na loss in the obsidian glass was deficient by 0.9 wt % (if 4 second times are considered) and 1.8% (if 2 second times are considered). We expect these values to be intermediate because the beam strength fluctuates between 20 nA and zero almost randomly during stabilization. Thus, some sodium loss is expected, but not as much as if the beam is stable for the additional 2 seconds. Furthermore, the obsidian glass is closest standard to the MI's in composition and thus we assume a similar loss of Na_2O in the melt inclusions.

Figure 4-4: Probe diameter measured from the flourescenc of quartz and benitolite. Outer circles on benitolite represent area of decreasing brightness observed during measurement. Scale bar is shown for reference

Figure 4-5: Sodium loss in obsidian as a function of time. Analytical conditions are 15 kV and 20 nA using point beam mode. Open symbols represent sodium loss experienced when only counting times are taken into account. Closed symbols represent total beam time. Values used for sodium addition are intermediate between 2 and 4 second curves at 10 seconds (dashed line).

Therefore all sodium values reported reflect initial pass values using the point beam and were corrected for sodium loss using the Na₂O loss calibration curves. The extent of correction was the median value observed to be lost in obsidian after 10 seconds (e.g., 1.35%).

4.4 Ion Probe Analysis

Measurements of trace elements and H₂O were made on a Cameca IMS 3f Ion Probe at Arizona State University. Analytical conditions were outlined in detail by Hervig & Dunbar (1992); Dunbar & Hervig (1992) and Hervig et al. (1989). One inch, carbon coated, petrographic slides, analyzed earlier by EPMA, were placed in a high vacuum chamber where they are bombarded with negative ¹⁶O ions. Although gold coats are preferred for ion probe analysis carbon coats did not seem to hinder analysis. Carbon coats however do increase the sputtering time necessary for ion beam penetration and location of MI's. Primary accelerating voltage on the Cameca 3f was -12.55 kV, secondary accelerating voltage was 4500V and arc current was fixed at 80mA. A 75 V offset was used to limit molecular interference. Standard measurements for trace elements were taken on the NIST 610. Standard measurements for water were acquired from the M3N and M6N rhyolite standards at ASU. Calibration curves (Fig. 4-6) matched those calculated in Ihinger et al. (1994). ³⁰Si was used as an internal standard and was measured before and after element acquisition. Analytical uncertainty is <10% for trace elements and ± 0.5 wt % for H₂O (Hervig et al., 1989).

Figure 4-6: H₂O calibration graph for NIST 610, M3N and M6N standards

4.5 Laser Ablation ICPMS

An attempt was made to obtain trace element data using laser ablation, induced coupled plasma mass spectrometry (LA-ICP-MS). This technique is currently being developed for use with melt inclusions and has been successfully applied to analysis of natural glasses (Pearce et al., 1999; Schroeder, et al., 1998) and phenocryst analysis (Davidson et al., 2001; Christensen et al., 1995). Some advantages of using LA-ICP-MS are presented in the following section. First, sample preparation/analysis time for LA-ICP-MS is reduced from days or even weeks to hours as compared with standard microdrilling techniques followed by isotope dilution mass spectrometry. In a comparative study between TIMS and LA-ICP-MS, Davidson et al. (2001), reports that one day of analysis using LA-ICP-MS yielded the same number of results as years of analyses performed using the microdrilling-TIMS method. Another advantage of LA-ICP-MS is that it can be used to analyze very small areas ($\sim 5\mu\text{m}$ diameter spot). By varying the power of the laser, the depth of the pit can also be controlled to high precision (Schroeder, et al., 1998; Pearce, et al., 1999). This is very advantageous when studying melt inclusions because most melt inclusions measure 25-100 μm in length and 10-50 μm in width and depth.

One sample (a hornblende from 1004-421) was ablated with a Cetac LSX 200 Nd:YAG Laser operating in the UV spectra range (266nm) using methods outlined by (Davidson et al., 2001). Analysis of the hornblende was used as a test for accuracy of concentrations determined by LA-ICP-MS (compositions of hornblende has been well studied, Luhr & Carmichael, 1980) and to determine proper laser power needed to produce pit sizes no larger than chosen MI's. Power was varied from 100% at a 20 Hz

repetition rate to 25% at 10 Hz and two pits were created measuring $\sim 100\ \mu\text{m}$ and $30\ \mu\text{m}$ respectively. Ablated material was transferred to the HP 4500 Quadrupole Mass Spectrometer via a transfer tube in an Argon atmosphere. Calibration was performed using the NIST SRM 612 and 610 standards. Results for these two “test” analyses yielded grossly inaccurate results. Hornblende was found to contain 1150 ppm of La (compared to 2-3 ppm reported by Luhr & Carmichael, 1980) in spite of the fact that the signal (in counts per second) was identical to the blank (purified Ar gas). Similar results were found for all other elements analyzed. Therefore, this strategy was abandoned (for the time being) and no further discussion of LA-ICP-MS analysis is provided.

CHAPTER 5: PETROGRAPHY

5.1 Whole Rock Samples

Samples from the eruptive period 1869-1913 were inspected in thin section using transmitted and reflected light microscopy. As described earlier, all samples contain zoned plagioclase phenocrysts, with varying amounts of orthopyroxene (opx) clinopyroxene (cpx) and hornblende. Groundmass typically consist of >50% micro-phenocrysts of plagioclase. A small but significant amount of cpx, opx and iron-titanium oxide phase is also present throughout the groundmass (analyzed by Luhr & Carmichael, 1980 as Ti-magnetite). Appendix C describes these samples in detail and gives modal proportions phenocrysts and groundmass.

Petrographic analysis also reveals that hornblende is the most variable component. Modal percentages of amphibole tend to decrease in the years preceding the major eruption of 1913. Furthermore, dark oxidation rims around amphibole phenocrysts were observed to thicken towards the end of the magmatic cycle (e.g., in 1890 lava). Samples from the 1869 lava flow of the Volcancito (DI-194, 1004-407 & 1004-414) contain abundant hornblende with distinct oxidation rims (Fig. 5-1). These oxidation rims are a result of reaction breakdown of hornblende (Moore & Carmichael 1998; Devine & Rutherford, 1998; Rutherford & Hill, 1993). Oxidation rims were found to vary between 7 μm and 25 μm in thickness and average 18 μm in samples 1004-407 & DI194. It should also be noted that observed rim thickness is greater when the crystal is viewed parallel to the crystallographic c axis than if viewed perpendicular to the c axis. Rutherford & Hill (1993) recommend measuring the thickness of the rim perpendicular to

Figure 5-1: Photomicrograph of samples DI-194 (top), 1004-407 (center) & 1004-414 (bottom) shown in plane polarized light. All are oxyhornblende rich. Rims measure 7 to 25 μm and are larger parallel to the crystallographic c- axis. Magnification is 15x.

Figure 5-2: Photomicrograph of sample M82-11 Top and JT 91-68 Bottom, shown in plane polarized light. Samples range from hornblende poor to hornblende absent. Hornblende rims measure 8 to 10 μm in M82-11 and $\sim 200 \mu\text{m}$ in JT 91-68. Interior of red ring is $\sim 1\text{mm}$ in diameter. Magnification is 15x.

Figure 5-3: Photomicrograph of sample 1004-421 shown in plane and crossed polarized light. Large hornblende and clinopyroxene phenocrysts appear at left. Interior of red circle is ~1 mm in diameter. Magnification is 15x.

the c axis, although they observed no noticeable difference with respect to orientation in the hornblende rims measured from Mt. St. Helens (MSH) and Soufriere Hills. Rim thickness in sample 1004-414 is much larger and measures 80 μm to 100 μm when viewed either perpendicular or parallel to the c axis. Inner opacite rims on these phenocrysts are comparable in size ($\sim 20 \mu\text{m}$) to its smaller rimmed counterparts mentioned above. Sample M82-11 (from the 1880 block lava flow) contains a minor amount of hornblende. Hornblende rims in this sample measure 8-10 μm . Sample JT 91-68 (also reported to be from 1880 block lava, Thomas & Draper, field study, 1991) contains oxidized relics of amphibole distinguishable only by their original crystal form (Fig. 5-2). Oxidation rims in sample JT 91-68 were measured from the center of the relic to its edge and were found to be $200 \pm 25 \mu\text{m}$ (across basal sections). We must assume that because the relic hornblende retains its original form, the farthest extent of the rim is equal to the edge of the original hornblende phenocryst. Therefore, measurements in this sample may be regarded as the minimal size of the reaction rim. The 1913 scoria sample is rich in amphibole that does not have any oxidation rims (Fig. 5-3).

5.2 Melt Inclusion Petrography

As described in numerous papers, most plagioclase crystals contain reacted cores (Luhr and Carmichael, 1980, 1989, Luhr, 1992, Robin, et al., 1990). Melt inclusions commonly occur within these cores (Fig. 5-4). A much smaller population of MI's occurs within the rims. In general, core MI's are 50-150 μm in diameter while rim MI's are typically smaller (20-75 μm in diameter). The shapes of MI's are highly variable and

range from rectangular to round to polygonal (Fig. 5-5). Unhomogenized, plagioclase hosted MI's are often brown and contain post entrapment crystals (Fig. 5-6).

Experimentally rehomogenized MI's, hosted by plagioclase, have been optically analyzed prior to EPMA and ion probe analysis. After heating, MI's usually become vitreous and no significant change in shape is observed (Fig. 5-7). Most rehomogenized MI's contain at least one vapor bubble and in rare instances can contain 3 or more vapor bubbles (Fig. 5-8) (multiple vapor bubbles are generally limited to large MI's heated to lower temperatures). Secondary crystals can be seen in MI's rehomogenized under lower temperature (Fig. 5-9). Optically the secondary crystals are square, light green in color and show low birefringence (δ was difficult to estimate due to the small size). It should also be noted that some crystals contain both vitrified and devitrified inclusions (Fig. 5-10).

Hornblende and pyroxene also contain melt inclusions. These MI's are quite different than those found in plagioclase. MI's in hornblende and pyroxene are vitreous. Their shape is usually diamond to oval, and they typically contain one vapor bubble (Fig. 5-11). Inclusions in pyroxene and hornblende are rare and occur almost anywhere in the host crystal with no apparent spatial preference.

Figure 5-4: Photomicrograph of plagioclase with a high concentration of melt inclusions in its core from sample Col-9D. Shown in plane polarized light, magnification 100x.

Figure 5-5: Photomicrographs of plagioclase melt inclusions. Shapes vary from rectangular (left) to oval (right) to polygonal (center). Shown in plane polarized light, magnification is 400x.

Figure 5-6: Photomicrograph of brown de-vitrified inclusions in plagioclase from sample 1004-421. Photo was enhanced from 100x picture. Scale bar shown is for reference.

Figure 5-7: Melt inclusions after re-homogenization at 1100°C. Melt inclusions are rectangular to oval and show exsolved vapor bubbles. Picture is in plane polarized light, magnification 400x.

Figure 5-8: Melt inclusion after re-homogenization experiment showing multiple vapor bubbles. Picture is in plane polarized light, magnification 400x.

Figure 5-9: Melt inclusions after re-homogenization at 1090°C. Multiple secondary crystals are readily seen within the inclusions. Picture is in plane polarized light, magnification enhanced from 400x.

Figure 5-10: Plagioclase crystal after re-homogenization showing both vitreous and de-vitrified melt inclusions. Picture is in plane polarized light, magnification 400x.

Figure 5-11: Melt inclusions in hornblende and pyroxene from sample 1004-421. Shapes of MI's are rounded (right) to diamond like (left) and MI's contain a single vapor bubble. Magnification of both pictures is 400x.

CHAPTER 6: RESULTS

6.1 Experimental Run Results

Melt inclusions were reheated in the one atmosphere furnace homogenize in steps. Under atmospheric conditions, all melt inclusions became vitreous at 1200°C. At 1150°C and 1140°C more than 90% of MI's are homogeneous. At 1125°C, 75% of the melt inclusions in plagioclase cores and 100% of MI's in plagioclase rims are homogeneous (Table 6-1). Controlled fugacity experiments showed similar results. Table 6-2 summarizes these results and shows the amount of post-entrapment/post-experimental crystallization. Again, high temperature runs successfully rehomogenized all melt inclusions and no secondary crystals can be observed. Melt inclusions in runs conducted $\leq 1120^{\circ}\text{C}$ show secondary crystals. As seen in the previous chapter (Fig. 5-13), these crystals have a square shape and are light green in color. An analysis performed on one of the secondary crystals shows it to be clinopyroxene (Table 6-3). Runs conducted at 1110°C have between 1 to 3 secondary crystals per MI, while runs conducted at 1100°C and below have multiple secondary crystals.

6.2 Major Elements

Melt inclusions from *Fuego de Colima* are all highly silicic (SiO_2 : 60 – 75 wt. %). Unhomogenized MI's show the widest variation in all elements (Table 6-4). SiO_2 is 63-75%. Al_2O_3 is most variable in these samples. It ranges from 2.88-21.8 % by weight but most samples contain approximately 15 - 18 wt. % Al_2O_3 . MgO , FeO and TiO_2 range from trace levels, 0.25 wt. % and 0.017 wt. % to 5.57 wt. %, 5.39 wt. % and 1.08 wt. %, respectively.

Run #	Temp (oC)	Run Duration	Results
DI 194.1	1000	4 hrs. 16 min.	Homogenization of <10% rim MI's Core MI's de-vitrified
DI 194.2	1100	4 hrs. 8 min.	100% homogenization of rim MI's <50% homogenization core MI's
DI 194.5	1125	4 hrs. 1 min	100% homogenization of rim MI's 75% homogenization core MI's (dependant on crystal)*
DI 194.3	1200	4 hrs. 12 min.	100% of all MI's in cores and rims
DI 194.4	1150	4 hrs. 2 min.	100% of all MI's in cores and rims

Table 6-1 – Run results for DI-194 samples (1869 lava flow). * Certain crystals from this run were homogenized completely while others showed homogenization of rim MI's with de-vitrified core MI's similar to those seen in prior runs

Run #	Temp (oC)	-Log f_{O_2}	Duration (Min)	Results
DI-194-CF 1	1140	-8.41	40	Complete homogenization of MI's in all cores and rims
DI-194-CF 1b	1150	-8.35	31	Complete homogenization of MI's in all cores and rims
DI-194-CF 1c	1150	-8.35	64	Complete homogenization of MI's in all cores and rims
DI-194-CF 2	1130	-8.75	40	Homogenization in Cores and Rims Less than 5% of MI's contain a single green secondary crystal
DI-194-CF 3	1120	-8.81	40	Homogenization of rim MI's with >50% homogenization of core MI's. 2-3 green secondary crystals are common in larger MI's
DI-194-CF 4	1110	-9.89	40	Homogenization of 90% of rim MI's and <50 % of Core MI's. Large MI's contain multiple secondary crystals usually clustered near MI wall.
DI-194-CF 5	1000	-9.34	35	Homogenization of 50-75% of rim MI's. Core MI's are de-vitrified
DI-194-CF 6	1090	-9.11	40	Homogenization of <25% of Rim MI's. Core MI's are de-vitrified
DI-194-CF 6b	1090	-9.12	62	Homogenization of <25% of Rim MI's. Core MI's are de-vitrified
DI-194-CF 7	1080	-9.25	45	No homogenization of MI's was visible in plagioclase MI's
DI-194-CF 7b	1080	-9.24	62	No homogenization of MI's was visible in plagioclase MI's

Table 6-2 – Run results for controlled fugacity experiments. Runs conducted above 1130 were fully effective in homogenizing MI's and dissolving or hindering growth of secondary crystals

Green secondary crystal (Experimental run DI-194.4)	
Oxide	Wt %
SiO ₂	54.67
TiO ₂	0.486
Al ₂ O ₃	5.63
FeO*	11.11
MnO	0.342
MgO	9.59
CaO	14.09
Na ₂ O	2.59
K ₂ O	0.565
Total	99.073

Table 6-3: EPMA analysis of secondary crystals in plagioclase MI's.
Analysis on the green crystal is from DI-194 CF4.

respectively, and values are lowest in MI's hosted by the mafic minerals (e.g., cpx, opx and hornblende). K₂O, CaO and Na₂O (corrected values) also show wide variations in composition range from and 0.21 to 4.28 wt. %, 1.17 - 5.56 wt. % and 2.24 to 5.63 % wt. % respectively and there does not seem to be any systematic difference in concentration of these elements as a function of host mineral. Furthermore, none of the major elements analyzed in unhomogenized samples seem to correlate with SiO₂ (Fig. 6-1).

Experimental rehomogenization runs conducted at atmospheric conditions show a much narrower range of compositions (Table 6-5). SiO₂ ranges from 64.5 to 73.1 wt. %. Al₂O₃ is 15.4 to 18.6 wt. % and shows little variation with SiO₂. MgO (0.43 to 1.95 wt. %), CaO (2.14 to 4.40 wt. %) and FeO (1.03 to 4.26 wt. %) are most variable. These elements show an inverse relationship to SiO₂. TiO₂, and MnO show small variations in composition (0.358 – 0.642 wt. % TiO₂ and 0.025 - 0.123 wt. % MnO) and tend to be higher in MI's with low SiO₂. K₂O and corrected Na₂O concentrations range from 0.503 - 1.51 wt. % and 1.88 to 3.04 wt % respectively. They both show some increase with increasing SiO₂ but a large amount of scatter makes this relationship somewhat unclear (Fig. 6-2)

Melt Inclusions from controlled fugacity runs show a wider variation in major element composition than MI's from atmospheric experiments, but the variation was not as wide as unhomogenized runs (Table 6-6). SiO₂ ranges from 60.7 to 74.6 wt. %. FeO (1.88-5.68 wt. %), CaO (1.26-6.67 wt. %), Al₂O₃ (10.17-20.35 wt %), Na₂O (3.38 to 6.95 wt. %, after corrections) and for the most part MgO (0.121-2.17 wt %), all tend to have high values at low SiO₂ and decrease with increasing SiO₂ (Fig 6-3). K₂O show the opposite trend. It is highest (2.88 wt. %) with highest SiO₂ and lowest (1.36 wt %) at low

Table 6-4 EPMA analysis of melt inclusions in natural samples from Fuego de Colima (1869-1913)

Melt Inclusion Analysis - 1004-421 (1913)													
	SiO2	TiO2	Al2O3	FeO ^T	MnO	MgO	CaO	Na2O	Na2O†	K2O	Total	Total†	n*
1004-421PL-MI1	75.270	0.357	8.377	5.390	0.092	1.617	1.891	4.430	5.780	0.211	97.635	98.985	3
1004-421PL-MI2	63.470	1.080	13.390	8.530	0.160	2.590	2.190	2.720	4.070	1.630	95.760	97.110	1
1004-421PX-MI1	70.050	0.026	15.628	0.236	0.021	0.020	1.166	0.890	2.240	1.184	89.221	90.571	5
1004-421PX-MI2	70.635	0.017	15.080	0.260	0.014	0.000	1.198	1.073	2.423	1.370	89.646	90.996	2
1004-421HBL-MI	68.300	0.025	16.300	0.356	0.032	-	1.600	1.430	2.780	1.570	89.614	90.964	1
Melt Inclusion Analysis - M82-11 (1880)													
M82-11PL-MI-1	70.465	0.447	12.020	5.070	0.138	1.519	1.690	5.160	6.510	4.275	100.783	102.133	2
M82-11PX-MI-1	70.367	0.387	17.627	1.693	0.068	0.652	2.100	1.469	2.819	1.797	96.160	97.510	3
M82-11PX-MI-2	69.680	0.370	17.910	1.370	0.052	0.188	2.200	2.190	3.540	2.870	96.829	98.179	2
Melt Inclusion Analysis - 91-68 (1880?)													
91-68PX-MI-1	69.388	0.589	18.555	1.496	0.050	0.198	2.438	1.568	2.918	2.065	96.345	97.695	4
91-68HBL-MI-1	67.610	0.296	21.750	1.088	0.041	0.070	1.940	1.660	3.010	1.850	96.305	97.655	1
Melt Inclusion Analysis - 1004-414 (1869)													
1004-414PL-MI2	61.770	0.693	15.687	4.866	0.104	2.071	4.650	8.147	9.497	1.897	99.885	101.235	3
1004-414PX-MI1	69.190	0.270	17.780	0.834	0.033	0.340	1.880	1.800	3.150	1.680	93.808	95.158	1
1004-414HBL-MI	69.450	0.195	2.880	0.860	0.007	5.570	5.555	3.560	4.910	0.473	88.549	89.899	2

† These columns represent corrected values after 1.35 wt % Na₂O addition.

* n equals the number of analyses taken on individual melt inclusions (see appendix D for standard deviation of averages).

Table 6-5: EPMA analysis of melt inclusions from controlled fugacity re-homogenization experiments on sample DI-194 (1869).

Melt Inclusion analysis for controlled f_{O_2} run at 1150°C													
	SiO ₂	TiO ₂	Al ₂ O ₃	FeO ^T	MnO	MgO	CaO	Na ₂ O	Na ₂ O†	K ₂ O	Total	Total†	n*
DI194CF1b.2a	60.220	0.426	16.110	4.250	0.088	1.427	4.650	4.130	5.480	1.700	93.001	94.351	1
DI194CF1b.2b	65.700	0.441	16.310	3.970	0.064	1.213	3.460	3.610	4.960	1.900	96.668	98.018	1
DI194CF1b.2c	65.300	0.390	16.510	3.940	0.046	1.218	3.560	3.720	5.070	1.835	96.518	97.868	2
DI194CF1b.2d	62.120	0.607	16.780	5.055	0.096	1.551	5.180	1.459	2.845	1.238	94.086	95.436	2
DI194CF1b.2e	61.960	0.475	16.910	4.060	0.076	1.299	3.910	5.280	6.630	1.480	95.451	96.801	1
Melt Inclusion analysis for controlled f_{O_2} run at 1140°C													
DI194CF2.1a	68.120	0.512	15.960	2.290	0.029	0.277	2.250	4.570	5.920	2.620	96.629	97.979	1
DI194CF2.1c	72.680	0.441	12.060	1.880	0.045	0.121	1.264	3.720	4.620	2.330	94.541	95.891	1
Melt Inclusion analysis for controlled f_{O_2} run at 1130°C													
DI194CF3.1a	65.088	0.405	14.398	5.405	0.142	3.151	3.410	3.183	4.533	1.950	97.131	98.481	4
DI194CF3.2b	68.880	0.739	13.623	4.030	0.072	1.233	1.620	2.430	3.780	3.020	95.647	96.997	3
DI194CF3.2c	72.523	0.571	12.053	2.377	0.043	0.759	1.280	3.583	4.933	2.867	96.057	97.407	3
Melt Inclusion analysis for controlled f_{O_2} run at 1120°C													
DI194CF4.1b	62.370	0.677	14.860	5.720	0.112	1.980	4.710	5.030	6.830	1.830	97.289	98.639	1
DI194CF4.1c	64.365	0.447	14.990	4.460	0.099	1.790	3.720	4.700	6.050	1.765	96.335	97.685	2
DI194CF4.1d	60.775	0.618	15.795	5.425	0.097	2.165	4.960	5.595	6.945	1.425	96.855	98.205	2

† These columns represent values after addition of 1.35 wt % Na₂O

* n equals the number of analyses taken on individual melt inclusions (see Appendix D for individual analyses).

Table 6-6: EPMA analysis of 1 atmosphere[‡] re-homogenization experiments.

Spot/line	SiO ₂	TiO ₂	Al ₂ O ₃	FeO ^T	MnO	MgO	CaO	Na ₂ O	Na ₂ O [†]	K ₂ O	Total	Total [†]	n*
5	73.148	0.455	18.684	2.118	0.049	0.613	2.139	0.657	2.007	1.170	99.033	100.383	1
9	70.223	0.508	16.066	2.610	0.070	0.774	3.342	0.616	1.966	1.531	95.740	97.090	1
10	75.204	0.380	15.406	1.669	0.084	0.429	2.547	0.503	1.853	1.486	97.708	99.058	1
Ave for 184/219	67.172	0.642	17.166	4.262	0.123	1.947	4.400	0.605	1.955	1.293	97.610	98.960	2
Ave 185/216	64.465	0.620	17.053	4.056	0.094	1.946	4.371	1.429	2.779	1.232	95.266	96.616	2
Ave 191/222/224/229	72.725	0.395	16.254	2.934	0.097	0.743	3.040	1.273	2.623	1.372	98.833	100.183	4
Ave 202/239/240/241	71.434	0.525	18.273	2.761	0.063	0.773	3.130	1.446	2.796	1.499	99.904	101.254	4
Ave 226/227/242	68.160	0.462	16.863	3.059	0.066	0.886	3.915	1.922	3.272	1.510	96.843	98.193	3
280G	68.878	0.491	17.992	2.139	0.064	0.854	3.474	1.808	3.158	1.094	96.794	98.144	1
281G	69.208	0.463	18.027	2.377	0.056	0.794	3.333	1.379	2.729	1.174	96.811	98.161	1
282G	67.811	0.498	17.215	2.137	0.117	0.651	3.786	1.694	3.044	1.089	94.998	96.348	1
283G	70.006	0.346	16.845	2.427	0.083	0.780	3.483	1.427	2.777	1.169	96.566	97.916	1
284G	72.976	0.383	17.772	1.026	0.064	0.892	2.753	1.497	2.847	1.114	98.477	99.827	1
285G	70.642	0.500	17.785	2.004	0.039	0.820	3.085	1.535	2.885	1.098	97.508	98.858	1
286G	69.893	0.461	17.253	2.155	0.075	0.526	2.853	1.276	2.626	1.201	95.693	97.043	1
287G	68.302	0.488	17.394	2.554	0.025	0.624	3.356	1.794	3.144	1.181	95.718	99.086	1
288G	68.174	0.390	17.508	2.658	0.106	0.874	4.224	1.740	3.090	1.165	96.839	98.189	1

[†] These columns represent values after addition of 1.35 wt % Na₂O

* n equals the number of analyses taken on individual melt inclusions (see Appendix D for individual analyses).

[‡] All analyses from run DI-194-3 (Conditions were 1100°C and atmospheric *f*O₂).

Figure 6-1: Plot of major elements versus SiO_2 for un-homogenized melt inclusions. Symbols represent melt inclusions hosted by plagioclase $\square$, cpx $\diamond$, opx $\blacklozenge$ and hornblende $\blacksquare$. Na_2O represents corrected values (see table6-4).

Figure 6-1 (cont.): Plot of major elements versus SiO₂ for un-homogenized melt inclusions. Symbols represent melt inclusions hosted by plagioclase □, cpx ◇, opx ◆ and hornblende ■.

Figure 6-2: Major element plots for experimentally re-homogenized MI's. Heating runs conducted under atmospheric conditions. Na_2O values reflect 1.35% sodium addition (see Table 6-5).

Figure 6-3: Major element plots for experimentally re-homogenized MI's. Heating runs conducted under controlled f_{O_2} conditions listed in table 6-1 and Appendix A. Na_2O values reflect 1.35% sodium addition (see Table 6-6).

values of SiO_2 . Again, TiO_2 and MnO tend to show only small variations and are highest in low SiO_2 MI's.

Still noticeable is the deficiency in Na_2O from atmospheric runs when compared to controlled fugacity runs and natural analyses. We attribute this to further sodium loss occurring in furnace experiments. These runs were 4 hrs. duration (as opposed to 40 – 60 min. used for controlled f_{O_2} runs) and it seems that the loss of H_2O experienced by Nielsen et al, 1995 in long duration runs (>2 hrs.) applies well to other volatile species such as sodium. Unfortunately we have no way of quantifying Na loss in the 1 atmosphere furnace and can only expect that the concentration of Na_2O in these MI's would have been similar to the others if the length of the runs had been shorter.

6.3 Trace and Rare Earth Elements

Trace elements and rare earth elements (REE) were analyzed in four melt inclusions. Two are clinopyroxene hosted and come from the 1913 scoria sample (1004-421). One is hosted in orthopyroxene from M82-11 (1890 block lava flow) and one hosted by plagioclase from 1004-414 (1869 *Volcancito* flow). None of these MI's were reheated and data were obtained *in situ* from whole rock thin sections. Major element analysis of each of these MI's along with their trace element and REE concentrations appear in table 6-7.

Primitive mantle normalized trace elements trends are similar to the 1913 scoria (sample 1004-421 whole rock data analyzed by Luhr & Carmichael, 1990) (Fig. 6-4). This sample was chosen for comparison because of its primitive composition and because it is the host for two of the 4 MI's analyzed. Luhr and Carmichael (1990) suggest that the

Table 6-7: Major and trace element analyses of melt inclusions.

Sample	1004-414	M82-11	1004-421	1004-421
Host	Plagioclase	orthopyroxene	clinopyroxene	clinopyroxene
Year of Eruption	1869	1890	1913	1913
SiO ₂	60.400	69.680	70.050	70.635
TiO ₂	0.568	0.370	0.026	0.017
Al ₂ O ₃	17.550	17.910	15.628	15.080
FeO ^T	4.720	1.370	0.236	0.260
MnO	0.170	0.052	0.021	0.014
MgO	1.469	0.188	0.020	0.000
CaO	4.100	2.200	1.166	1.198
Na ₂ O [†]	11.590	3.540	2.240	2.423
K ₂ O	2.140	2.870	1.184	1.370
Total	101.398	96.829	89.221	89.646
H ₂ O	0.394	0.716	3.420	2.760
B	2.815	2.815	2.019	3.039
Be	34.463	34.463	46.486	65.327
Rb	44.463	44.463	30.706	64.029
Sr	156.191	156.191	165.106	232.695
Y	15.515	15.515	3.055	22.407
Zr	299.987	299.987	125.338	180.240
Nb	6.671	6.671	0.833	8.702
Ba	744.879	504.659	702.540	950.380
Ce	28.458	25.528	1.661	9.478
Nd	13.135	11.803	5.091	15.352
Sm	4.033	1.489	4.489	19.642
Eu	3.186	0.508	2.191	0.000
Tb	0.482	0.309	2.188	8.912
Ho	0.529	0.529	1.761	11.802
Tm	0.226	0.099	1.254	8.690
Cs	0.750	0.760	1.400	n / a
Th	3.279	1.528	2.725	n / a
U	0.783	0.710	0.549	n / a

Major elements analyzed by EPMA, Trace elements and water analyzed by Ion Probe.

n/a means data not available

† Na₂O represents corrected values after addition of 1.35 wt. %.

Figure 6-4: Plot of trace elements from Colima melt inclusions normalized to primitive mantle values of Sun & McDonough (1989). Thick gray line represents trace element data for Colima lavas from 1869-1913 (Data from Luhr & Carmichael, 1990)

more mafic scoriaceous andesites represent new batches of primitive magma from which the later lavas evolved. Unfortunately no trace element data exist for lava and scoria from the magmatic cycle starting eruption of 1818. Therefore, we must assume that trace element abundance in the 1913 scoria is similar to that of past magmatic cycle starting compositions (e.g., from the same source). Compared to the 1913 scoria, *Colima* MI's are slightly enriched in Rb, Ba and Th and somewhat depleted in Sr, Ce and Sm. It can be seen that these differences are most apparent in MI's hosted by clinopyroxene from the 1913 scoria sample itself. One of the two cpx hosted MI's (1004-421-PXMI 1) shows significant variations in almost all trace elements. It is highly depleted in Ce, Sm, Eu, Y, Nd and most likely Nb (no data exists for Nb in respective lava and scoria). Trace elements from sample M82-11-PXMI-1 match the bulk rock more closely except for slight depletions in Nd, Sm, Eu and Y. The MI hosted by plagioclase (sample 1004-414-PLMI, 1869) also provides a close match although Sr is somewhat depleted while Eu is fairly enriched. Sr depletion in all samples is probably due to plagioclase crystallization while the enrichment in Eu seen in 1004-414 may be due to hornblende dissolution.

Simple trace element fractionation models can be used to see if the differences in the trace element abundance are a direct result of fractionation. Calculations were performed using equations given in Rollinson (1995). Elemental abundance was calculated using both equilibrium (Equation 1) and fractional (Rayleigh) crystallization (equation 2).

$$C_l/C_o = 1 / [D + F(1-D)] \quad (1)$$

$$C_l/C_o = F^{(D-1)} \quad (2)$$

Where D is the bulk partition coefficient (values for K_d used to calculate D were obtained from Rollinson, 1995), F is the percent of melt remaining (or inversely the proportion of crystals) C_L is the concentration in the melt inclusion and C_o is the concentration of the element in the initial magma. Because many of the trace elements in low abundance fractionate to apatite, one model calculates the bulk partition coefficient using a small proportion of apatite (0.5% of total) while the other model is apatite free. The result of each of these models is shown in Figs. 6-5 a-d).

While both equilibrium and fractional models correctly predicted some trace elements these models cannot account for all the trace element concentrations measured in *Colima* MI's. For example Ba concentration is correctly predicted in models for all four MI's at $F \leq 0.5$ (50% crystallization or more). However, all models fail to predict the concentration of Ce and Nd in 1913 MI's. In general, 1913 MI's seem to be different for most elements than the predicted models and MI's from 1869 & 1890. Moreover, some elements are correctly predicted using a specific F value while the other elements require a different value of F to accurately account for their concentration. Finally, although matrix accounts for 54.6 to 74.4 % of the erupted lavas from the period 1869-1913 (Luhr & Carmichael, 1990) crystallization models require 75% crystallization to account for certain trace elements. Certainly, more data are needed to clarify these issues.

Figure 6-5 a & b: Observer Colima melt inclusion concentrations plotted against calculated trace element compositions for varying degrees of equilibrium crystallization. Initial compositions values used in the model are from whole rock data of Luhr and Carmichael (1990) for sample 1004-421 (1913).

Figure 6-5 c & d: Observer Colima melt inclusion concentrations plotted against calculated trace element compositions for varying degrees of fractional crystallization. Initial compositions values used in the model are from whole rock data of Luhr and Carmichael (1990) for sample 1004-421 (1913).

6.4 Water in Melt Inclusions

Water was analyzed in the four MI's discussed above. The pyroxene hosted MI's from the 1913 scoria (samples 1004-421-PXMI-1 & 1004-421-PXMI-2) contained 3.42 ± 0.5 wt. % and 2.76 ± 0.5 wt. % H_2O respectively. These water contents are nearly identical to those calculated by Luhr (1992) using albite-anorthite liquid equilibrium (2.5 - 4.8 wt. % H_2O with the higher values occurring in the 1913 scoria). Furthermore, the H_2O contents found in these MI's agree well with the experimentally derived phase equilibria for *Colima* Lavas at various water contents (Moore & Carmichael, 1998). They show the *Colima* assemblage to equilibrate at between 3 wt. % and 5 wt. % H_2O (Fig 6-6).

The melt inclusion from the 1890 lava flow (M82-11 PXMI 1) was nearly dry and measured only 0.716 ± 0.5 wt. %, while water in the MI hosted by plagioclase from 1869 (sample 1004-414-PLMI, 1869) near detection limits (0.394 ± 0.5). Based on the model by Moore et al. (1998) such water contents translate to nearly atmospheric pressure. On the other hand, based on data from Wardell et al. (2001), on MI's from New Zealand, we estimate pressures of 25-75 bars for our melt inclusions which contain similarly low water contents. Thus, we estimate that these block lavas had lost most of their original water content through degassing (this will be discussed further in the following chapter).

Figure 6-6: Stability fields for plagioclase, orthopyroxene, clinopyroxene, hornblende and magnetite as a function of pressure and temperature. Dashed numbered lines are water isopleths, orange lines are isopleths for measured water contents of colima MI's. Red line represents isothermal decompression for Colima MI's and red filled circles represent analyses of 1913 MI's. Also shown are average ranges of anorthite content for Colima and other Mexican plagioclases. Shaded areas represent fields of stability for phase present in Colima and other Mexican lava. Black area shows calculated temperatures of Colima lava. (Figure modified from Moore & Carmichael, 1998)

CHAPTER 7: DISCUSSION

In general most MI's contain higher SiO_2 and lower MgO and FeO than both bulk rock and groundmass. Many *Colima* MI's are related to bulk magmas and groundmass by fractionation of a combination of plagioclase, opx, cpx and hornblende. These relationships are shown in Fig 7-1a & b. In order to accommodate the large variation in MI compositions, trapped melts may undergo different degrees of reaction and loss of the solute component to the surrounding crystal, causing different diversity in the residual melt composition. This suggests that *Colima* MI's are trapped during different stages of crystal fractionation. Furthermore, if the trends are solely dependent on crystal fractionation they must show similar results on other element-element plots so long as 1) only one liquidus phase is present, and 2) that phase is the sole carrier of the MI. In many cases major element plots do indeed show similar results (Fig 7-1b). However, the plot SiO_2 vs. raw (uncorrected) Na_2O does not satisfy melt inclusions compositions no matter the proportion of crystal assemblage fractionated (Fig 7-2 bottom). As shown in chapter 5, the deficiency seen in Na_2O is mainly due to volatile element loss common to electron microprobe analysis (Nielsen & Sigurdsson, 1981). Addition of 1.35 wt % Na_2O to these analyses plus the addition of 0.3 wt % Na_2O to plagioclase (estimated from volatility of Na in albite) helps rectify most of these compositions so they plot near the plagioclase fractionation line (Fig. 7-2, top). While Na loss in the 1 atmosphere furnace may account for the location of 1 atmosphere experimental runs on the plot of SiO_2 vs. Na_2O (i.e., they plot below the plagioclase fractionation line) some of the natural samples are still outside of the allowable fractionation paths. Robin et al.

Figure 7-1: Plots of Al₂O₃ (a) and FeO (b) versus CaO. Fractionation of plagioclase and cpx dominate differentiation trends while hornblende and opx contributions are relatively small in both graphs. Lines drawn from mineral compositions represent liquid enrichment paths extending away from bulk rock. Filled circles represent un-homogenized MI's in plagioclase (full black), opx (lt. blue), cpx (red) and hornblende (yellow)

Figure 7-2: Plot of Na_2O vs. SiO_2 . Top graph is for corrected Na_2O values for all runs and plagioclase and show trends consistent with crystal fractionation from the mineral assemblage shown. Below is the same graph using raw Na_2O values. Filled circles represent unhomogenized MI's in plagioclase (full black), opx (lt. blue), cpx (red) and hornblende (yellow)

(1991) reports a single analysis of a plagioclase rim from 1913 containing a more sodic composition (6.78 % Na₂O at 58% SiO₂). Therefore, non-homogenized MI compositions may have been influenced by sodic plagioclase rim growth.

Other discrepancies involving the host phases are noticeable when we consider the trace element compositions of MI's. Certain trace elements cannot be related to host crystals nor can they be related to *Colima* lavas. In the prior chapter it was shown that the MI's from the 1913 eruption of *Fuego de Colima* do not match the host scoria as well as MI's in the 1869 and 1890 lava. Moreover, the addition of apatite is not sufficient to explain the deficiencies seen in the trace elements of the pyroxene hosted MI's from 1913. The REE abundance of the 1913 MI's is also completely different from that of the host rock (Fig. 7-3). These deficiencies may have been a result of hornblende dissolution and/or cpx fraction or may have been due to magma mixing as described by Luhr (1992) and Robin et al. (1993). Luhr and Carmichael (1990) suggested that the more mafic nature of the 1913 lava is due to an injection of basaltic or basaltic andesite magma. Furthermore, Luhr (1992) states that the initial source of the mafic magmas is the mantle wedge, which may explain the anomalous REE trends. To be sure, further investigation of the trace element and REE abundance in *Colima* MI's is needed.

Despite the above, trace elements offer us considerable clues to the processes occurring at *Colima*. H₂O and K₂O are generally considered to be incompatible with respect to *Colima* magmas below 1kb pressure (Moore & Carmichael, 1998). Therefore, they preferentially partition into the melt phase (Dixon & Clague, 2001). Accordingly, as H₂O increases, so should incompatible element concentration. However, this is not what is observed in the present study. Water correlates negatively with TiO₂ ($r^2 = 0.925$), FeO

($r^2 = 0.908$), CaO ($r^2 = 0.859$) and Na₂O ($r^2 = 0.751$) (Fig. 7-4). Water also correlates negatively with incompatible trace elements (Ce, U, and Zr). However, Cs and B correlate positively to water (Fig. 7-5). Thus, water concentrations decrease during differentiation of the andesitic melt. K₂O is also seen to inversely correlate to H₂O (Chapter 6, Fig. 6-6b). This trend is not unique to *Colima*. Finny (2000) showed that this same relationship exists at Popocatepetl Volcano in central Mexico. A comparison of *Colima* MI's and MI's from Popocatepetl is shown in Figure 7-6. In the Popocatepetl study, Finny (2000) concluded that differentiation continues to occur during ascent, causing volatile and other incompatible elements to partition to the melt. However, if the ascending magma degasses during rise H₂O will be preferentially lost from the differentiating melt. This would result in a negative correlation between H₂O and all other incompatible elements (inset in Fig. 7-4 and 7-5). Based on such behavior shown by *Colima* MI's, we could conclude that the trends seen in H₂O versus K₂O and incompatible trace elements are representative of magmatic degassing during differentiation.

Melt inclusions in clinopyroxene from 1913 contain a maximum of 3.4 ± 0.5 % H₂O and a minimum of 2.7 ± 0.5 % H₂O. While many studies relate water saturation to depth in basaltic and rhyolitic magmas (Burnham, 1979, Moore et al., 1998; Dixon 1995, Dunbar & Hervig 1992, Hervig et al, 1995), few studies report H₂O saturation in andesitic magmas. A recent model by Moore et al. (1995 & 1998) calculates the solubility of water in various magmas at temperatures and pressures more applicable to *Colima*. Their study included a dacite from Mt. St. Helens that closely matches the composition used to calculate the *Colima* phase equilibria mentioned above. Moore et al.

Figure 7-3: Rare earth element spidergram for Colima MI's and bulk lava of 1913 (thick dashed line). Eu anomaly in 1869 sample may be due to hornblende resorption. Trends for MI's from 1913 clearly differ from bulk rock, 1869 MI and 1890 MI.

Figure 7-4: Plot of major element compositions versus H_2O . Inset shows expected trend for compatible versus incompatible elements during fractionation

Figure 7-5: Plot of selected trace elements from Colima melt inclusions versus H₂O. Insets to right show expected trends for incompatible element abundance versus water for differentiation without degassing (top inset) and differentiation with degassing (bottom inset)

Figure 7-6: Plot of H_2O versus K_2O for Colima and Popocatepetl melt inclusions. Best fit line is for Colima MI's but also fits quite well the data of Finny (2000).

(1995 & 1998) show that andesitic melts with 3.4 wt % H₂O would be saturated with pure H₂O vapor at a pressure of 800 bars (Fig 7-7). Higher pressures would be required if the vapor contains significant amounts of CO₂ (Dixon & Stolper, 1995; Blank et al., 1993). This pressure is well in agreement with the stability field derived by Moore & Carmichael (1998). Therefore, assuming that P_{H₂O} is equal to the total pressure the pyroxene hosted MI's containing 3.4 wt. % H₂O must have been entrapped at ~2.4 km. Using the same reasoning we estimate ~500 ± 100 bar (1.5 ± 0.3 km) for the water poor MI. We have also shown in the previous chapter that the stability fields of the observed crystal assemblage found in the scoria that hosts these MI's (e.g., plagioclase, opx, cpx and hornblende) is not stable under equilibrium conditions below 600 bar and ~950°C (see figure 6-6)(Moore & Carmichael, 1998). Therefore if the water poor MI lies outside the stability field (i.e., has less than the maximum allowed by analytical error), then this implies one of three possibilities. 1) This melt inclusion has partially degassed after entrapment. This possibility is deemed unlikely because diffusion of H₂O through pyroxene is very slow (King et al 1999). 2) The field of hornblende stability must be extended to include this point or 3) the pyroxene started growing within the hornblende stability field and continued to grow during ascent where it entrapped the MI with 2.7 % H₂O. While extension of the hornblende stability field would be a very exciting hypothesis, degassing of magma upon ascent seems more plausible and agrees well with trace element and incompatible major element trends.

MI's trapped in plagioclase and orthopyroxene are relatively dry (0.3 ± 0.5% and 0.7 ± 0.5% H₂O respectively). Maximal pressures, based on water saturation, are far less than those observed in the water rich MI's. It is believed that these water poor MI's

Figure 7-7: Water saturation curves derived from Moore et al. (1995 & 1998) for dacite composition of Mt. St. Helens from Rutherford and Devine (1993). Top curve is for water undersaturated conditions while the lower (dashed) curve represents water saturated dacitic magma. Solid straight lines intersecting these curves are for Colima melt inclusions 1004-421 PXMI 1 containing 3.4 wt % H_2O and 1004-421 PXMI 2 containing 2.7 % H_2O .

likely represent degassed melts. Since these MI's occur in plagioclase and opx, this observation implies that these phases crystallized at shallow depths.

Seismic evidence suggests that a magma chamber may exist at 1-3 km depth (Núñez-Cornú et al 1994). Seismicity also occurs immediately below the surface. This information is compatible with our MI-based inference above. It seems that water rich magmas in such a magma chamber crystallized amphibole + cpx; and during ascent to the surface such magmas lose H₂O and crystallize plagioclase + opx. Amphibole also breaks down during ascent and degassing. However, the extent of the breakdown reaction depends on the ascent rate. In sum, relatively high H₂O in the amphibole & cpx hosted MI's versus low H₂O of the plagioclase and opx MI's indicates that amphibole and cpx crystallized early at a deeper level (~2.4 km), whereas plagioclase and opx crystallized much later under near surface pressures.

Ascent rate curves have been derived for andesitic and dacitic volcanoes of Mt St. Helens (MSH) and Soufriere Hills using experimentally grown hornblende oxidation rims under H₂O saturated conditions (Rutherford & Hill, 1998; Devine & Rutherford, 1998). Note that the stability field of hornblende in *Colima* magma is substantially different from the MSH case despite the fact that the starting compositions for the two magmas are similar (Fig 7-8). This difference is not understood; but it is apparent that the MSH amphibole field would be incompatible with the presence of amphibole in *Colima* lavas. That is, the MSH phase diagram is inapplicable to *Colima*. Unfortunately, to date, no study has measured hornblende rim growth rates for *Colima*, and therefore we must assume that the measured growth rates derived by Rutherford & Hill (1993) are applicable to *Colima*, which may or may not be correct in detail, but may still be useful.

As outlined in chapter 5, hornblende rims for samples from 1869 are ~15-20 μm wide and average 18 μm wide. Applying the MSH derived equations (Rutherford & Hill, 1993) we obtain magma ascent times of 9 - 15 days from 2.4 km depth (Fig 7-9) ($\sim 8.3 \pm 2.8$ m/hr). The 1869 sample, 1004-404 has comparatively thick rims (80 – 100 μm). For this sample we estimate ascent times of between 30 and 50 days (2.0 – 3.3 m/hr.). 1880 block lavas contain hornblende with thin rims (in M82-11 rims are 8-10 μm) and totally resorbed hornblendes with rims that measure 200 ± 25 μm from the center of the “crystal” (JT91-68). Ascent times have been estimated for sample M82-11 to be 5 days (ascending at ~ 20 m/hr.) while it is estimated to be a minimum of 65-120 days (0.83 – 1.5 m/hr.) for JT91-68. Rutherford and Hill (1993) report that the onset of hornblende rim growth begins after 4 days regardless of starting temperatures and pressures used. Therefore, the 1913 scoria sample containing rim free amphibole crystals must have ascended to the surface in less than 4 days at speeds ≥ 25 m/hr. This rapid ascent rate may also help explain why the modal percentage of phenocrysts in the 1913 scoria is low. Further, if the opx grows in the upper reaches of the magmatic column fast ascent helps explain why there is little opx found in the scoria.

In sum, MI's from *Colima* have highly complex histories. Their water contents dictate that some were entrapped at near surface conditions while others formed at depth. Their major and trace element compositions suggest that the magmas were simultaneously differentiating and degassing during ascent. Clinopyroxene and hornblende phenocrysts appear to have crystallized early from undegassed magma, at or below 2.4 km while orthopyroxene, and to some extent plagioclase, grew at near surface conditions. From the growth of hornblende rims we have estimated that *Colima's*

scoriaceous magmas must rise from 2.4 km depth to the surface at ≥ 25 m/hr. This ascent rate has a dramatic effect on both crystal content and the concentration of water in the MI. Magmas rising quickly (1913) correlate with MI's that contain more water and are associated with more violent eruptions. Those MI's with low water contents (found in eruptive products from 1869 and 1880) correspond to slow ascent rates (< 8 m/hr.), block lava flows with higher modal abundance of phenocrysts.

Figure 7-8: Plot of phase stability for Colima, Mt. St. Helens (MSH) and for basaltic andesite. Dashed lines represent water saturation isopleths. Arrows show isothermal decompression lines for both Colima and MSH. Also plotted are MI's analyzed for water (red circles and yellow squares) in relation to colima stability field shown in figure 6-6. Diagram modeled from Moore & Carmichael (1998) and Rutherford & Hill (1993).

Figure 7-9: Ascent rate curves derived by Rutherford & Hill (1993) for Mt. St. Helens and Devine and Rutherford (1998) for Soufriere Hills volcano plotted with measured hornblende reaction rims for Fuego de Colima (this study).

CHAPTER 8: CONCLUSIONS

[1] Melt inclusions from Volcan de *Colima* are highly differentiated and have complex histories. They show a wide range of compositions that depict an overall evolution of magma recorded by entrapment over time.

[2] Water contents found in melt inclusions are vastly different. Pyroxene and amphibole have the highest H₂O contents and reflect entrapment pressures of at least 800 bars (2.4 km). Melt inclusions in plagioclase and orthopyroxene are dry and must have been entrapped at near surface conditions.

[3] Incompatible major and trace elements and trace element correlate inversely with water indicating that magmas were simultaneously undergoing degassing and differentiation during ascent.

[4] Hornblende rims record different ascent rates for different eruptive events. Rim thickness for hornblende in 1869 lavas suggests ascent rates on the order of 8.3 ± 2.8 meters/hour. Ascent rates in 1880 lavas are much slower in one sample (0.83 – 3.3 m/hr.) relatively fast in another (20 m/hr.). Ascent rates for the cycle beginning eruption of 1913 are >25 m/hr. derived from the fact that hornblende in this sample has no oxidation rims. This correlates well to the VEI of each eruption with 1913 being the most violent (plinian) and 1880 being least (strombolian-effusive type) except for the rapid ascent of M82-11 which is seen to still be slower than 1913.

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APPENDIX A: Calibration and determination of f_{O_2} in 1 atm. furnace

Calibration Run 1 (23 Jun, 1999)

Depth (cm)	Control Temp. (°C)	Reference Temp (°C)
26.5	1000	998
25.5	1000	1005
24.5	1000	1012
23.5	1000	1011
22.5	1000	1013
21.5	1000	1007
20.5	1000	997

Diopside calibration at 21 cm depth and 1391°C (12 hrs. 35 min.)

Time	Temp. (°C)	Stage
19:55	989	Stage I Ramp (Start)
21:50	1395	Stage I Ramp
21:55	1400	Stage I Dwell (Start)
22:53	1398	Stage II Ramp
23:05	1392	Stage II Dwell (Start)
23:20	1392	Stage II Dwell
23:30	1391	Stage II Dwell
11:40	1391	Drop Quench

Result: Glass and crystals observed

Calibration Run 2 (15 Dec., 1999)

Depth (cm)	Control Temp. (°C)	Reference Temp (°C)
27.4	1150	1164
26.2	1150	1167
25.2	1150	1166
24.2	1150	1161
23.2	1150	1154
22.7	1150	1146

Diopside calibration at 22.9 cm depth and 1391°C (10 hrs. 37 min.)

Time	Temp. (°C)	Stage
11:23 (16 Dec. 99)	989	Stage I Ramp (Start)
13:23	1395	Stage I Dwell (Start)
13:35	1400	Stage I Dwell
14:20	1398	Dwell Check
14:23	1392	Stage II Ramp (Start)
14:43	1392	Stage II Ramp
14:53	1391	Stage II Dwell (Start)
16:05	1391	Stage II Dwell
10:00 (17 Dec. 99)	1391	Drop Quench

Result: Glass and crystals observed

Calibration Run 3 (7 Aug. 2000)

Depth (cm)	Control Temp. (°C)	Reference Temp (°C)
27	1000	1019
26	1000	1019
25	1000	1012
24	1000	1008
23.5	1000	999
23	1000	994
22	1000	980

Diopside calibration at 23.5 cm depth and 1391°C (4 hrs. 14 min.)

Time	Temp. (°C)	Stage
14:16 (8 Aug. 00)	968	Stage I Ramp (Start)
14:54	1221	Stage I Ramp
15:17	1388	Stage I Dwell (Start)
15:34	1399	Stage I Dwell
16:11	1399	Stage I Dwell
16:17	1398	Stage II Ramp (Start)
16:37	1386	Stage II Dwell (Start)
16:40	1391	Stage II Dwell
20:54	1391	Drop Quench

Result: Glass and crystals observed

Calculation of oxygen fugacity

Variation of Nernst Equation:

$$P_{O_2} = 0.209 \exp [-46.42 E \text{ (Volts)} / T \text{ (}^{\circ}\text{C)}]$$

Where: $P_{O_2} = -\text{Log } f_{O_2}$

Table A-1: f_{O_2} for corresponding E and T at NNO Buffer during experimental runs

Run	T (°C)	CO ₂	H ₂	E (Volts)	f_{O_2}
DI-194 CF 1	1140 ± 1	140 ± 9	3 ± 2	0.542 ± 0.015	-8.41 ± 0.21
DI-194 CF 1b	1150 ± 1	148 ± 0	30 ± 2	0.542 ± 0.016	-8.35 ± 0.23
DI-194 CF 1b	1150 ± 1	148 ± 0	30 ± 2	0.542 ± 0.000	-8.35 ± 0.00
DI 194 CF 2	1130 ± 1	148 ± 0	2 ± 1	0.562 ± 0.004	-8.75 ± 0.06
DI-194 CF 3	1120 ± 1	147 ± 1	3 ± 1	0.562 ± 0.011	-8.81 ± 0.15
DI-194 CF 4	1110 ± 1	147 ± 9	17 ± 8	0.632 ± 0.001	-9.89 ± 0.01
DI-194 CF 5	1100 ± 1	147 ± 1	8 ± 4	0.590 ± 0.024	-9.34 ± 0.35
DI-194 CF 6	1090 ± 1	148 ± 9	33 ± 3	0.570 ± 0.008	-9.11 ± 0.11
DI-194 CF 6b	1090 ± 1	150 ± 0	0.5 ± 0.5	0.571 ± 0.003	-9.12 ± 0.04
DI-194 CF 7	1080 ± 1	147 ± 0	1.5 ± 1.5	0.575 ± 0.008	-9.25 ± 0.18
DI-194 CF 7b	1080 ± 1	145 ± 2	0.5 ± 0.5	0.575 ± 0.084	-9.24 ± 1.25

APPENDIX B

Standard Compositions

Table B-1: Compositions of SPI standard used for EPMA calibration and analyses.

Standard compositions from SPI-2753						
	Obsidian	Albite	Diopside	Magnetite	Rhodenite	Rutile
SiO ₂	73.94	68.52	55.37		45.98	
TiO ₂	0.1		0.08			100
Al ₂ O ₃	13.11	19.54	0.09			
FeO	1.72		0.05	99.8*	1.55	
MnO	0.06		0.05		37.66	
MgO	0.07		18.62		0.87	
CaO	0.76	0.13	25.73		6.4	
Na ₂ O	4.06	11.59				
K ₂ O	5.04	0.22				

*Fe⁺² as FeO (31.03%), Fe⁺³ as Fe₂O₃ (68.76%)

APPENDIX C

Detailed petrography of lava samples

1869 “Volcancito” Samples

Samples 1004-414 and DI-194 are both from the 1869 flow of the Volcancito. They contain approximately 20% plagioclase phenocrysts 10% oxyhornblende, 3% opx and cpx phenocrysts. Generally plagioclase laths are zoned and glomerophorpyritic. Laths measure 1-3 mm. Anorthite content measured by the Michel-Lévy method ranges from An_{64} to An_{73} in 1004-414 (average An_{70}) and An_{45} to An_{75} in DI-194 (average An_{64}). Hornblende is typified by simple twinning, mild zoning and clean margins. Orthopyroxene and clinopyroxene are generally euhedral and phenocrysts measure 1-3 mm.

1004-407 is similar to the above. Plagioclase is present and accounts for 10-20% of the sample. Anorthite ranges from An_{38} to An_{71} (average $An_{63.5}$). Large hornblende crystals are present (5%) and display opacite rims. Minor amounts of pyroxenes can be seen as small (< 0.2 mm) euhedral grains and total 1-2%.

1890 Samples

M82-11 is very similar to the 1869 samples with the exception of hornblende. Hornblende is nearly absent in this sample. Relic opacite rims have degraded as well and measure less than 0.5 mm. In many places and oxide phase and CPX are all that remains

of the original hornblende. Plagioclase contents range from An₃₃ to An₇₂ (average An_{63.5}).

Sample JT91-68 was also collected from the 1890 flow. It contains ~20% plagioclase, 2-3% orthopyroxene and 1% clinopyroxene. Approximately 3 % of the sample is the oxidized relic of hornblendes. Most large phenocrysts of plagioclase have mottled cores surrounded by a zone of melt inclusions. Smaller plagioclase tends to be mildly zoned or not zoned and do not have mottled cores or abundant MI's. Anorthite content distribution in this sample is bi-modal. One group centers at An₃₉ (5 grains) while the other group is An₇₁ (4 grains). Pyroxenes are twinned and tend to cluster.

1913 Sample

Sample 1004-421 is from the 1913 pyroclast flow deposit. It also contains 20% plagioclase, with lesser amounts of hornblende and pyroxenes. Plagioclase ranges from An₃₃ – An₇₆ (average An₆₂). Hornblende shows opacite rim textures and is generally not twinned as seen before. Orthopyroxene is anhedral and generally measures less than 2 mm. Small CPX clusters are prevalent throughout the thin section and cpx is often twinned.

Table C-1 – Optic petrography of unhomogenized whole rock thin sections.

Sample #	Year of flow	Mode (norm. to 100%)	MI Information
DI – 194	1869	Pl _{20.0} , cpx _{1.5} , opx _{1.0} hbl _{1.0}	Abundant rectangular MI's in plag. cores. Rare MI's in Hbl and pyroxenes
JT91-68	1869	Pl _{20.0} cpx _{2.0} opx _{1.0} hbl _{0.1}	Abundant rectangular MI's in plag. cores and diamond shaped to oval MI's in pyroxenes
1004-404	1869	Pl _{18.6} cpx _{1.4} opx _{0.9} hbl _{7.4}	Abundant rectangular MI's in plag. cores
1004-407	1869	Pl _{20.2} cpx _{1.4} opx _{0.9} hbl _{4.0}	Rare oval MI's in pyroxenes Abundant MI's in plag. cores Rare diamond shaped to oval MI's in Hbl and pyroxenes
M82-11	1890	Pl _{23.6} cpx _{1.9} opx _{0.8} hbl _{3.2}	Abundant oval and rectangular MI's in Plag. rims and cores and throughout pyroxenes
1004-421	1913	Pl _{12.5} cpx _{2.6} opx _{1.6} hbl _{3.6}	Abundant MI's in plag. throughout Rare scattered diamond shaped to oval MI's in Hbl and pyroxenes
Col 9D	1979	Pl _{30.1} cpx _{1.8} opx _{2.9} hbl _{0.1}	Abundant MI's in plag. rims Rare scattered MI's in pyroxenes
Col 30	1982	Pl _{28.0} cpx _{5.2} opx _{2.4} hbl _{0.5}	Abundant MI's in plag. rims extremely abundant MI's throughout pyroxenes

Data for 1004 samples, Col samples and M82-11 is from point analysis done by Luhr and Carmichael (1990). Data for DI-194 and JT 91-68 is from estimated visual comparison to Philpotts (1989). Melt inclusion information from optical inspection.

APPENDIX D

Table D-1: Individual point analyses for experimentally homogenized MI's from sample DI-194 runs 3 (1200°C) and 4 (1150°C).

Spot/line	SiO ₂	TiO ₂	Al ₂ O ₃	FeO*	MnO	MgO	CaO	Na ₂ O	K ₂ O	Total
Spot 5	73.148	0.455	18.684	2.118	0.049	0.613	2.139	0.657	1.170	99.033
Spot 9	70.223	0.508	16.066	2.610	0.070	0.774	3.342	0.616	1.531	95.740
Spot 10	75.204	0.380	15.406	1.669	0.084	0.429	2.547	0.503	1.486	97.708
184G	66.760	0.687	17.596	4.415	0.156	2.110	4.409	0.815	1.355	98.303
185G	62.817	0.633	17.706	4.115	0.087	2.064	4.395	1.490	1.252	94.559
202G	72.003	0.402	19.657	2.814	0.040	0.737	3.087	1.996	1.440	102.176
216G	66.113	0.607	16.399	3.997	0.101	1.828	4.347	1.367	1.212	95.971
219G	67.584	0.597	16.735	4.108	0.090	1.783	4.390	0.395	1.231	96.913
222G	73.659	0.463	14.328	3.098	0.069	0.808	2.924	1.814	1.592	98.755
224G	66.971	0.415	18.002	2.927	0.096	0.841	2.768	0.641	1.502	94.163
226G	69.457	0.421	17.065	3.406	0.042	0.898	2.876	1.593	1.630	97.388
227G	71.653	0.433	14.802	3.179	0.120	0.901	3.098	1.770	1.631	97.587
229G	72.464	0.410	15.396	2.958	0.061	0.733	3.508	0.999	1.254	97.783
239G	72.370	0.776	17.996	2.969	0.090	0.904	2.826	1.528	1.629	101.088
240G	72.954	0.515	16.912	2.823	0.109	0.849	2.885	0.246	1.288	98.581
241G	68.408	0.406	18.528	2.439	0.011	0.601	3.720	2.012	1.639	97.764
242G	63.370	0.531	18.721	2.593	0.035	0.859	5.770	2.402	1.269	95.550
280G	68.878	0.491	17.992	2.139	0.064	0.854	3.474	1.808	1.094	96.794
281G	69.208	0.463	18.027	2.377	0.056	0.794	3.333	1.379	1.174	96.811
282G	67.811	0.498	17.215	2.137	0.117	0.651	3.786	1.694	1.089	94.998
283G	70.006	0.346	16.845	2.427	0.083	0.780	3.483	1.427	1.169	96.566
284G	72.976	0.383	17.772	1.026	0.064	0.892	2.753	1.497	1.114	98.477
285G	70.642	0.500	17.785	2.004	0.039	0.820	3.085	1.535	1.098	97.508
286G	69.893	0.461	17.253	2.155	0.075	0.526	2.853	1.276	1.201	95.693
287G	68.302	0.488	17.394	2.554	0.025	0.624	3.356	1.794	1.181	95.718
288G	68.174	0.390	17.508	2.658	0.106	0.874	4.224	1.740	1.165	96.839

Analysis performed on an ARL SEMQ microprobe at Florida International University. Analytical conditions were 15 kV and 10 nA using a point beam. Count times were 10 seconds for each element

Table D-2: Individual point analyses for controlled fugacity runs

	SiO ₂	TiO ₂	Al ₂ O ₃	FeO*	MnO	MgO	CaO	Na ₂ O	K ₂ O	total
DI194CF1b.2a	60.220	0.426	16.110	4.250	0.088	1.427	4.650	4.130	1.700	93.001
DI194CF1b.2b	65.700	0.441	16.310	3.970	0.064	1.213	3.460	3.610	1.900	96.668
DI194CF1b.2c	64.960	0.335	16.690	3.960	0.070	1.191	3.520	4.030	1.800	96.556
DI194CF1b.2c-pt2	65.640	0.445	16.330	3.920	0.021	1.244	3.600	3.410	1.870	96.479
DI194CF1b.2d	62.200	0.617	17.090	5.050	0.060	1.523	5.270	0.698	0.966	93.474
DI194CF1b.2d-pt2	62.040	0.597	16.470	5.060	0.132	1.578	5.090	2.220	1.510	94.697
DI194CF1b.2e	61.960	0.475	16.910	4.060	0.076	1.299	3.910	5.280	1.480	95.451
DI194CF2.1A	68.120	0.512	15.960	2.290	0.029	0.277	2.250	4.570	2.620	96.629
DI194CF2.1B	61.970	0.052	20.690	0.686	0.019	0.021	3.610	11.520	0.908	99.476
DI194CF2.1B-pt2	61.540	0.018	21.650	0.677	0.000	0.036	4.360	11.340	0.739	100.360
DI194CF2.1C	72.680	0.441	12.060	1.880	0.045	0.121	1.264	3.720	2.330	94.541
DI194CF3.1A	62.980	0.440	15.360	5.410	0.119	2.330	4.000	4.780	1.540	96.959
DI194CF3.1A-pt2	61.770	0.314	19.770	3.780	0.094	1.414	5.950	3.170	1.181	97.443
DI194CF3.2A-pt3	68.310	0.283	8.660	7.220	0.260	6.630	1.650	1.600	2.200	96.813
DI194CF3.2A-pt4	67.290	0.581	13.800	5.210	0.096	2.230	2.040	3.180	2.880	97.307
DI194CF3.2B	69.720	0.706	12.940	4.190	0.087	1.339	1.610	2.180	2.950	95.722
DI194CF3.2B-pt2	69.310	0.826	13.280	3.970	0.047	1.171	1.440	2.450	3.050	95.544
DI194CF3.2B-pt3	67.610	0.684	14.650	3.930	0.081	1.190	1.810	2.660	3.060	95.675
DI194CF3.2C	68.290	0.233	15.830	1.700	0.045	0.431	1.910	5.840	2.840	97.119
DI194CF3.2C-pt2	76.990	0.545	9.230	2.060	0.025	0.579	0.840	2.650	2.830	95.749
DI194CF3.2C-pt3	72.290	0.936	11.100	3.370	0.059	1.267	1.091	2.260	2.930	95.303

Table D-2:
Continued

	SiO2	TiO2	Al2O3	FeO*	MnO	MgO	CaO	Na2O	K2O	total
DI194CF4.1A	61.210	0.460	17.350	3.900	0.064	1.349	5.050	7.760	1.330	98.473
DI194CF4.1A-pt2	62.630	0.584	14.670	5.640	0.075	2.010	4.690	4.730	1.780	96.809
DI194CF4.1B	62.370	0.677	14.860	5.720	0.112	1.980	4.710	5.030	1.830	97.289
DI194CF4.1C	64.110	0.437	15.330	4.430	0.101	1.760	3.770	5.190	1.730	96.858
DI194CF4.1C-pt2	64.620	0.456	14.650	4.490	0.096	1.820	3.670	4.210	1.800	95.812
DI194CF4.1D	60.520	0.565	16.700	4.980	0.106	1.960	5.300	6.320	1.280	97.731
DI194CF4.1D-pt2	61.030	0.671	14.890	5.870	0.087	2.370	4.620	4.870	1.570	95.978
Ave (all MI's)	65.289	0.491	15.334	3.988	0.077	1.491	3.398	4.407	1.950	96.426
s.d. (all MI's)	4.247	0.203	2.923	1.559	0.049	1.209	1.515	2.503	0.704	1.571

Table D-3: Average microprobe analysis and standard deviations for multiple point analyses

1004-421						
Element	Ave Px MI1	s.d.	Ave PX MI 2	s.d.	Ave PI MI 1	s.d.
SiO2	70.050	2.185	70.635	0.460	75.27	12.057
TiO2	0.026	0.009	0.017	-	0.357	0.254
Al2O3	15.628	1.709	15.080	0.042	8.377	5.656
FeO	0.236	0.015	0.260	0.010	5.39	4.083
MnO	0.021	0.007	0.014	-	0.092	0.049
MgO	0.020	0.012	0.000	-	1.617	1.658
CaO	1.166	0.041	1.198	0.083	1.891	1.081
Na2O	0.890	0.204	1.073	0.037	4.43	2.69
K2O	1.184	0.237	1.370	0.000	0.211	0.159
Total	89.221		89.646		97.635	1.439

M82-11						
Element	Ave Plag MI1	s.d.	Ave Px MI1	s.d.	Ave Px MI2	s.d.
SiO2	70.465	0.148	70.367	0.420	69.680	0.184
TiO2	0.447	0.334	0.387	0.017	0.370	0.028
Al2O3	12.020	3.875	17.627	0.202	17.910	0.141
FeO	5.070	3.083	1.693	0.105	1.370	0.113
MnO	0.138	0.134	0.068	0.022	0.052	0.016
MgO	1.519	1.063	0.652	0.266	0.188	0.019
CaO	1.690	1.061	2.100	0.022	2.200	0.141
Na2O	5.160	2.022	1.469	0.469	2.190	0.141
K2O	4.275	0.148	1.797	0.074	2.870	0.071
Total	100.783	0.520	96.160	0.043	96.829	0.061

Table D-3 (Cont.)

1004-414				
Element	Ave Pl MI2	s.d.	Ave Hbl MI	s.d.
SiO2	61.770	1.468	69.450	2.051
TiO2	0.693	0.896	0.195	0.121
Al2O3	15.687	7.398	2.880	0.297
FeO	4.866	5.471	0.860	0.580
MnO	0.104	0.147	-	-
MgO	2.071	2.821	5.570	1.230
CaO	4.650	1.678	5.555	0.403
Na2O	8.147	4.272	3.560	1.358
K2O	1.897	0.663	0.473	0.027
Total	99.885	1.114	88.549	1.921

91-68		
Element	Ave Plag MI	s.d.
SiO2	69.3875	1.234
TiO2	0.5885	0.079
Al2O3	18.555	1.132
FeO	1.49575	0.413
MnO	0.05025	0.016
MgO	0.19775	0.199
CaO	2.4375	0.314
Na2O	1.5675	0.022
K2O	2.065	0.066
Total	96.34475	0.904

Table D-4: Average compositions used for plotting differentiation trends

	Average Groundmass	Average Bulk Rock (1869-1913)	Average Plagioclase*	Average Clinopyroxene	Average Orthopyroxene	Average Hornblende
SiO ₂	65.86	59.852	51.151	51.358	53.154	42.885
TiO ₂	0.746	0.665	0.02	0.559	0.1958	2.582
Al ₂ O ₃	15.98	17.735	29.35	2.673	1.4656	11.812
FeO*	4.247	5.47	0.637	8.455	18.04	11.915
MnO	0.177	0.101	0.012	0.238	0.5032	0.195
MgO	1.457	3.175	0.126	15.465	26.964	14.355
CaO	3.666	6.003	12.39	20.316	1.1726	11.242
Na ₂ O	5.299	4.611	5.771	0.449	0.0366	2.643
K ₂ O	2.204	1.34	0.159	-	0.0054	0.402
P ₂ O ₅	0.233	0.197	n/a	n/a	n/a	n/a
Total	99.869	99.149	99.616	99.513	101.5372	98.031

* Plagioclase analysis is a combination of numerous data points acquired from this study and Luhr & Carmichael, 1980. All other values are from Luhr and Carmichael 1980 & 1990.