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Field and Remote Sensing Analysis of the 2015 Pyroclastic Density Currents at Colima (Mexico) and Calbuco (Chile) Volcanoes: Implications for Hazard Assessment and Crisis Management

Elodie Macorps

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Field and Remote Sensing Analysis of the 2015 Pyroclastic Density Currents at Colima (Mexico) and Calbuco (Chile) Volcanoes: Implications for Hazard Assessment and Crisis Management

by

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A dissertation submitted in partial fulfillment of the requirements for the degree of
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DEDICATION

I would like to dedicate this dissertation to my husband, my parents and my brother, for their immense support along the way, and without whom I would never have made it to the end.

I would also like to dedicate my work to Sylviane and Gérard Fesseau, my science teachers from middle and high school, and now friends. They introduced me to all the wonders of science as a kid, they encouraged me, and shared their passion for exploring the World and for humanitarianism. They inspired my thirst for wanting to make a difference in the World and working on Disasters.

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ABSTRACT

Although one of the most spectacular phenomena of active volcanoes, Pyroclastic density currents, or PDCs, are considered the most dangerous volcanic hazards. PDCs are avalanches of hot volcanic gases, ash, and larger volcanic fragments that travel at incredible speed down the flank of a volcano. High dynamic pressures, high temperatures, and high velocities are the primary dangers associated with PDCs and lead to near-complete destruction and death.

I use a multi-disciplinary approach to study the deposits left behind by PDCs, in order to understand their dynamics, their interactions with the receiving landscape, and their final distribution, starting on the ground and going up to space. The study cases of this research are two PDC events that occurred in 2015 at Volcán de Colima in Mexico, and at Volcán Calbuco in Chile. The main goals of this dissertation are (1) to add to the collection of field deposit studies of PDCs, (2) to quantify the control of the topography on the path and distribution of PDC deposits, (3) to demonstrate the advantages of remote sensing techniques for both long-term hazard assessment and short-term crisis management of PDCs, as well as lahars which are frequent secondary hazards.

The first part of this work focuses on field-based observations and a detailed analysis of these PDC deposits. Although the PDC-forming eruptions at Colima and Calbuco were different (i.e., dome-collapse vs. column-collapse), the common conclusion drawn from both sets of PDCs is that they were primarily controlled by topography. Mainly, the morphology of the valleys in which the PDC deposited, which includes the cross-sectional area, the volumetric capacity, the
sinuosity, and the slope of the channels, had significant impacts on the final footprint of the PDC deposits.

These field-based results are used as validation for the second part of this research, which focuses on spaceborne remote sensing data. A combination of optical, synthetic aperture radar, or SAR, and thermal satellite sensors is successfully used to track PDC and lahar deposits from space, thus opening an avenue for application to post-eruption crisis management efforts. Optical data at medium and very high spatial resolution (30 to 1 meter) are also used to build digital elevation models or DEMs. The DEMs are used (1) to measure channel morphology parameters, (2) to retrieve volumes of PDC deposits and other erupted products such as lava dome and lava flows at Colima, and (3) in a time-series to track topographic changes caused by eruption at these two volcanoes.

The techniques developed and presented in this dissertation show reproducibility across two different types of PDC events, and therefore suggest applicability to other, both past and future PDC-forming eruptions. For instance, tracking topographic changes at active volcanoes could help update hazard maps, considering the effects of the topography on the final impacts of PDCs and lahars that were demonstrated here. Using remote sensing data to map PDC and lahar deposits from previous eruptions offers an alternative to costly and time-consuming field campaigns, and it allows for retrieval of critical parameters (e.g., erupted volumes, deposit footprint) that can be used as model confirmation and as input boundary conditions for testing computational models of PDCs. Moreover, with the growing availability of near-real-time satellites, there is potential for using the methods introduced in this dissertation to support emergency officials during and immediately after an eruption, by providing accurate extents of PDC and lahar deposits.
CHAPTER ONE:
INTRODUCTION

Rationale and Motivation

Pyroclastic density currents (PDCs) are among the most spectacular but also highly destructive and dangerous phenomena at active volcanoes. PDCs are hot avalanches of volcanic particles, known as pyroclastic fragments, and gas that can travel great distances at high speed across the landscape under the effect of gravity. They are the most complex types of natural multiphase gravity currents, where the mixing of the solid phase (particles) and fluid phase (gas) defines a natural continuum from concentrated (solid-dominated) to dilute (fluid-dominated) PDCs (Sulpizio et al., 2014; Dufek, 2016). The consensus for a conceptual model of PDCs is a dense (i.e., concentrated), coarse-grained, ground-hugging basal undercurrent overridden by a dilute, turbulent ash-cloud surge (Sulpizio et al., 2014 and references therein).

The danger of these flows comes from their ability to cause near complete devastation of large areas. The high dynamic pressures can destroy infrastructures, the high temperatures are lethal and can burn everything on their path, and the high velocities make it impossible to outrun the flows (Baxter, 1990; Jenkins et al., 2013; Brown et al., 2017). Nearly 100 million people live at risk from PDCs (Small and Naumann, 2001; Chester et al., 2000). Roughly 90% of the fatalities caused by PDCs occur within a 20-km radius from the initiation point (Brown et al., 2017) and over 29 million people worldwide live within 10 km of active volcanoes (Brown et al., 2015). Most people killed by volcanoes are victims of PDCs and lahars (i.e., sediment-charged flows), totaling about 120,000 deaths over the last 500 years (Brown et al., 2017). PDCs account for one third of
volcanic fatalities recorded in history (Auker et al., 2013). The highly destructive and deadly nature of PDCs warrant the need to better forecast their impacts on surrounding populations, livestock, and infrastructures for hazard assessment and crisis mitigation purposes.

PDC hazards and risks mitigation are based on forecasting of possible flow paths or inundation areas, which is most often accomplished via computational modeling of PDCs (Esposti Ongaro et al., 2020). PDC hazard models are utilized to generate forward predictions of inundation areas which, in conjunction with field observations and the mapping of the distribution of past deposits, are applied for the mitigation of hazards and potential impacts from future PDCs. Well-constrained data sets from past PDC events are used for model confirmation (Esposti Ongaro et al., 2020) and are critical for defining the input and boundary conditions used in the models (Lube et al., 2020).

Field observations and field mapping of PDC deposits provide information regarding the dynamics involved during transport and deposition of the flows, as well as the interactions with the local topography. The stratigraphy and architecture of a PDC deposit record the processes that occurred within the flow prior to freezing, and therefore the interpretations can be used to infer physical models and compare with computer model outputs. Moreover, careful mapping of inundated areas is used with well-defined metrics to compare with the results from inverse modelling of that particular event for model confirmation (e.g., Charbonnier et al. 2018). Therefore, a first motivation for this study is driven by the needs for well-constrained and comprehensive datasets from past PDCs using the remnant deposits. This is accomplished in this dissertation by presenting the results from field studies conducted on two sets of PDC-forming eruptions that occurred in 2015: (1) the April 22-23, 2015 explosive eruption and column-collapse
at Calbuco Volcano in Chile and (2) the July 10-11, 2015 dome-collapse eruption at Volcán de Colima in Mexico.

Although PDC deposits provide invaluable information, field campaigns can be costly and time-consuming, and the loose and fragile deposits may be rapidly eroded and remobilized by water. Accessing pristine deposits shortly after an eruption may be extremely dangerous due to ongoing activity, the remaining high temperatures and gases trapped in deposits, and/or the potential for hazardous secondary lahars that form from the mixing of remobilized deposits and water flows. Moreover, direct access to active volcanoes can often be restricted due to the remote location, presence of strong relief, dense vegetation or even due to the presence of local populations that control the areas. In these context, remote sensing tools provide multiple advantages to map PDC deposits when access to the field is impractical or when deposits are no longer present. In cases where field observations are possible, such as those considered in this work, they can provide ground-truthing for interpretations derived from remote sensing observations.

Remote sensing data can also be used to derive models of the topography, commonly known as digital elevation models (DEMs). They provide information regarding the pre-eruptive topography and allow for measurements of volume changes either by deposition or morphological changes experienced during an eruptive event. Therefore, a second motivation for this dissertation is to demonstrate the advantages in applying remote sensing techniques to extract complete datasets from past PDCs including deposit distributions and volumes, as well as tracking syn- and post-eruption topographic changes. This is done by compiling results from multiple satellite sensors (radar, thermal and optical remote sensing tools) to look at the distribution of PDC deposits, and by looking at time-series of DEMs at coarse- (30 m) and high-resolution (5 m) to
investigate the pre-eruptive topography, the changes caused by eruptive events, and to calculate volumes of volcanic deposits.

The topography of volcanoes where PDCs occur is generally characterized by high and complex relief involving numerous valleys and channels that carve the landscape in various directions from the vent. Small-volume (< 1 km$^3$) concentrated PDCs preferentially follow these topographic drainage pathways, but they can also *escape* channel confinement, overspill and spread laterally, and deposit on top of valley interfluves, or re-channelize into adjacent valleys. These flows are referred to as *overbank* flows and initiate directly from the parent basal undercurrent. When overbank flows comprise only the dilute upper part of a PDC, they have been referred to in the literature as ash-cloud surges, or surge-derived pyroclastic flows (e.g., Druitt et al., 2002; Ogburn et al., 2014). Detachment of the surge component (i.e., ash-cloud surge) leads to independent lateral motion from the parent flow, with the ability to traverse topographic barriers, and sometimes to re-channelize and behave as concentrated flows in lateral channels.

The hazardous overbank flows and detached surges have been associated with numerous fatalities over the years. On June 3rd, 1991, 43 people died at Unzen (Japan), including 3 volcanologists, while on the valley interfluves because of the detachment of the ash-cloud surge from the block-and-ash flows (i.e., dome-collapse PDCs; Fisher, 1995). Similarly, at Soufriere Hills Volcano (Montserrat) in 1997, 19 lives were claimed by overbank flows that overspilled the valley confines on June 25, inundating a large area on the lower flanks of the volcano (Loughlin et al., 2002a; Ogburn, 2015).

At Merapi (Indonesia), the decoupling of the ash-cloud surge during the 22 November 1994 block-and-ash flows caused the deaths of 64 people and seriously injured/burned dozens more people (Abdurachman et al., 2000; Bourdier and Abdurachman, 2001). The renewed activity in
2006 caused overbank flows that led to additional fatalities and destruction of a village located ~5 km from the summit along the main valley of the southern flank (Charbonnier and Gertisser, 2008; Gertisser et al., 2012). Again in 2010, the dome collapse and large overbank flow volume resulted in over 200 deaths, and more than 2200 damaged buildings in the villages up to 15 km from the summit (Jenkins et al., 2013; Charbonnier et al., 2013). Most recently, in 2018, the eruption of Volcán de Fuego (Guatemala) resulted in more than 100 deaths and over 1000 injuries caused by PDCs and overbank flows when it devastated villages and a golf course resort on the SW flank (Charbonnier et al., 2019).

The large number of fatalities resulting from overbank flows and detached ash-cloud surges justifies the need for improving the understanding of these processes. Previous studies of the mechanisms involved with the generation of overbank flows and surge detachment have primarily highlighted the effects of localized topography. Most particularly, the morphology of the valleys in which the parent flows travel has been shown to control both the runouts and potential for overbank flows. Records in the FlowDat database (Ogburn, 2012) indicate a link between channelization and increased flow mobility, which coincide with findings from numerically modeled PDCs (Stinton, 2007). Reduced channel cross-sectional area from progressive infilling by prior deposits, channel constriction (either natural or by anthropogenic structures), sinuosity and slope are parameters that have been identified on multiple occasions to be related to the generation of overbank flows and/or surge detachment at Soufriere Hills Volcano (Loughlin et al., 2002b; Ogburn et al., 2014), at Merapi (Bourdier and Abdurachman, 2001; Charbonnier and Gertisser, 2008, 2011; Thouret et al., 2010; Lube et al., 2011; Charbonnier et al., 2013; Komorowski et al., 2013; Jenkins et al., 2013), and at Unzen (Yamamoto et al., 1993; Fujii and Nakada, 1999). The majority of these studies, however, are generally all limited to one type of
PDC initiation mechanism (i.e., dome collapse) and analyze single events. Therefore, a third motivation for this research is to add onto the catalog of records for overbank flow generation as a result of topographic controls, and further highlight the extent to which the relationships between topography and deposit extent can be generalized. This is done by comparing two types of PDC-forming eruptions to derive relationships between PDC deposit distribution and topographic parameters by means of a Geographic Information System (GIS).

While the capabilities of computational modeling have substantially improved in the last few decades – for example the successful simulation of the emplacement of both the concentrated and dilute phases of a PDC in the two-layer version of VolcFlow (Kelfoun, 2017; Gueugneau et al., 2019, 2020) – the limitations and complexity from using computer models for hazard assessment of PDCs come from the unpredictability of eruption scale, duration, and source characteristics (Lube et al. 2020). It is crucial for these models to account for the diversity of generation mechanisms of PDCs and eruption intensity in order to encompass the range of possible outcomes.

Computational models are also time-consuming and require high computing power that is not always available, especially in low-income countries. PDC-forming eruptions often occur without much warning, and there is rarely enough time to run complex modeling programs. The lack of time during a crisis, absence of accurate inputs and the difficulties associated with these models justify the need for additional methods to help emergency officials and risk management groups determine the main areas at risk prior to an eruption. Hence, a fourth motivation for the work presented in this dissertation is the intent to find new ways to support hazard mapping and crisis management for PDC-related hazards. This is attempted here using knowledge of
topographic controls on the distribution of PDCs and utilizing high-resolution DEMs to determine potential areas at risk of overbank flows and/or surge detachment processes.

*While research focused on hazard and risk mitigation is important ahead of a crisis, my final motivation for this study is oriented towards post-PDC events.* I hope to demonstrate the usefulness of remote sensing tools to help recovery efforts determine the extent of affected areas, and where to look for survivors. Moreover, as mentioned earlier, subsequent remobilization of loose deposits and generation of lahars are common phenomenon that occur post-PDC emplacement. Lahar hazards are the result of fresh and loose eruptive deposits, PDCs or tephra, being remobilized on steep slopes experiencing heavy rain or if these deposits emplaced into river streams, or from summit glacier melts during the eruptive event. The steepness of slope and recent vegetation loss as well as loose burnt/damaged vegetation amongst the deposits, all exacerbate the lahar hazards.

Remote sensing data can assist in determining areas where subsequent lahars are most likely to occur, thus helping prioritizing evacuation efforts. This is particularly useful if the population is engulfed in ash clouds that limits visibility on the ground. Furthermore, the combination of sensors is critical when one fails to provide sufficient information. For instance, in tropical regions where cloud cover at the volcano summit is common, or if an ash-cloud is masking the surface, optical images will have poor observational capabilities. This shortcoming can be compensated with the use of radar data with day/night and all-weather acquisitions (Henderson and Lewis, 1998). This is also true if an eruption occurs at night, in which case thermal imagery can help trace the extent of recent and hazardous hot materials on the ground.

The combination of optical and radar imagery during a volcanic crisis proved very useful at Merapi during the 2010 eruption for crisis management (Ratdomopurbo et al., 2013). In 2006,
commercial optical images were capable of capturing the early growth and subsequent collapse of a newly erupted lava dome, but in 2010, the rainy season resulted in a persistent cloud cover that limited the use of optical imagery. Immediate actions were taken by the Volcano Disaster Assistance Program (VDAP) to acquire Synthetic Aperture Radar (SAR) imagery (Pallister et al., 2013). They tracked the progress of the precursory activity by conducting qualitative visual change detection in the summit area using intensity images provided by RADARSAT-2 and TSX. They used high-resolution GeoEye-1 and Worldview-2 infrared images, when available without cloud cover, as well as thermal images from the ASTER instrument to increase the observations count with time. The combinations of these data permitted estimates of lava dome discharge rates and the extents of PDCs after the paroxysmal phase of the eruption, which helped the Indonesian Center for Volcanology and Geologic Hazard Mitigation (CVGHM) in forecasting the magnitude of the eruption and the necessary extent of evacuations and recovery efforts (Pallister et al., 2013).

Similarly, SAR data from Sentinel-1A/B and Thermal data from ASTER were used at Fuego Volcano (Guatemala) in June 2018 to rapidly determine the extent of areas inundated by PDCs and help recovery efforts on the ground. The data were subsequently used to define the targets of aerial surveys for very high-resolution imagery (https://maps.disasters.nasa.gov).

Scope of Thesis

The objectives of this dissertation are to (1) add to the collection of field deposit studies of PDCs, (2) further the understanding and demonstrate the relationships between topographic controls and the distribution of PDCs, (3) demonstrate the use of topography analysis for rapid assessment of hazard areas, (4) present the advantages of remote sensing tools for deposit mapping and generating digital elevation models, and (5) to show the applicability of DEM time-series,
both at coarse and fine-resolution, for tracking topographic changes through time and for determining volumes of volcanic deposits.

Two research directives are being investigated with different timelines: (1) the application of datasets for long-term hazard assessment work, and (2) the use of remote sensing data for immediate to short-term application in risk mitigation, crisis management and recovery efforts. As such, this work explores the combination of field-based and remote sensing observations to provide a comprehensive study of two PDC-forming eruptions from 2015, and to investigate the dynamics and hazards of PDCs. To address these objectives, the dissertation is structured as follows:

Chapter 2 provides background information for this dissertation. First, an overview of PDCs is introduced, along with the varying generation mechanisms and common conceptual models presented in the literature. Second, the geological settings and a summary of the eruptive history of Volcán de Colima and Calbuco Volcano are discussed, along with a description of the eruptive events that led to the generation of PDCs being studied in this work. The chronology of the major events at both Colima and Calbuco is reconstructed based on local monitoring reports and online reports from the Global Volcanism Program (Smithsonian Institution).

Chapter 3 presents the results from field investigations of the block-and-ash flow (i.e. PDC, see Chapter 2) deposits from the dome-collapse eruption that occurred in July 10-11, 2015 at Volcán de Colima in Mexico. Deposit architecture, stratigraphic sequence and sedimentology results are used to infer dynamics of transport and deposition and to derive a conceptual model for depositional mechanisms of these block-and-ash flows. Qualitative relationships between channel capacity and the generation of overbank flows are highlighted, along with evidence for dynamic changes in response to channel enlargement. This chapter is a reprint of the published work

Chapter 4 catalogues the results from field investigations of the 2015 column-collapse PDCs at Calbuco Volcano in Chile. Due to limited field access, the observations were confined to two valleys of interest on the NE and SW flanks. In a similar manner to the field work conducted at Volcán de Colima (see Chapter 3), the stratigraphy and sedimentology of the deposits are analyzed, along with the deposit architecture and distribution with respect to the pre-eruptive topography. The ensemble is used to infer flow dynamics of these PDCs, which are then used in Chapter 5 to better interpret the relationships with the topographic controls. Although field observations are limited to two valleys only, they give insights onto the dynamics of the eruption and allow to derive some general relationships with topographic parameters, similar to the work presented in Chapter 3.

Chapter 5 presents the remote sensing analysis of the PDCs described in Chapters 3 and 4. In this chapter, multiple sensor types are combined to provide a comprehensive overview of the PDCs and demonstrate the usefulness of a combined study for crisis management purposes. High-resolution remote sensing data are used to derive pre-eruption DEMs of each volcano, which are then used for the analysis of valley morphology. I created a GIS-based algorithm to be used in the open source QGIS software, for extracting channel capacity, cross-sectional area, sinuosity and slope of the valleys of interest. The results are compared with the distribution of PDCs mapped from both field and remote sensing data, paying particular attention to overbank and surge deposits, in order to derive relationships between topographic parameters and the generation of these hazardous flows.
Chapter 6 investigates the topographic changes through time, by means of DEM time-series at both coarse and high-resolution. The goal is to demonstrate the importance of monitoring such changes at active volcanoes in order to provide up-to-date hazard assessments. Coarse-resolution DEM datasets have frequent and repeated observations that allows for data rich time-series to analyze trends of elevation changes through time. High-resolution datasets, from commercial imagery, on the other hand are more costly to acquire, but complement the analysis by providing snapshot of elevation changes caused by particular events in time. I show the potential for open source data to support emergency management work for rapid topography analysis prior to or during a volcanic crisis.

Finally, the implications of this work and findings on hazard assessment and crisis mitigation are discussed in Chapter 7, along with the main conclusions of each of the previous chapters and perspectives for future research directives.

References


CHAPTER TWO: 
BACKGROUND

Pyroclastic Density Currents

Among the different types of surface expression of volcanic phenomena, Pyroclastic density currents (PDCs) are one of the most devastating and least predictable (Blong, 1986). They are ground-hugging highly mobile mixtures of gas and volcanic fragments that can travel over long distances, up to tens of kilometers (Dade and Huppert, 1998) and with speed greater than 100 km/h. PDCs are gravity-driven ‘pseudo-fluid’ that are generally the product of violent eruptions resulting from magma fragmentation and explosion. The deposits left behind can be of variable volumes depending on the eruption and the generation mechanism, from a few million cubic meters to more than 1,000 km$^3$ (Brown and Andrews, 2015). The structure and morphology of these deposits are recordings of the PDC dynamics during emplacement (Sparks et al., 1976; Fisher, 1966).

Historical evidence and past studies have shown that PDCs can cause near complete destruction of widespread areas, often covering more than 1 km$^2$ up to 20,000 km$^3$ (Brown and Andrews, 2015) and represent serious threats to surrounding populations and infrastructures as they are often deadly and can results in short term economical and agricultural problems (Tilling and Lipman, 1993; Tanguy et al., 1998). Some major eruptions are prime examples: Montagne Pelée in 1902 (Lacroix, 1904), Mount Unzen in 1991 (Yamamoto et al., 1993), Soufriere Hills Volcano in Montserrat in 1997 (Calder et al., 1999; Voight et al., 2002; Sparks et al., 2002), Merapi in 1994, 2006 and 2010 (Bourdier and Abdurachman, 2001; Charbonnier and Gertisser, 2008;
Komorowski et al., 2013). PDCs pose great threats because their intensity and frequency can be highly variable. Similarly to volcanic eruptions, high intensity PDCs are less frequent but highly destructive, while PDCs of lower intensities are more frequent, and still maintain a destructive power. Jenkins et al. (2013) uses the example of the 2010 eruption at Merapi to show the highly destructive power of even small-volume PDCs. Due to the seriousness of PDC hazards, their assessment and mitigation has been a main focus of the international volcanology community during the last 40 years. A few studies have focused their effort on regrouping information available in the literature in an attempt to characterize and combine observations (Fisher and Schminckke, 1984; Cas and Wright, 1987; Druitt, 1998; Branney and Kokelaar, 2002; Sulpizio et al., 2014; Dufek et al., 2015; Dufek, 2016). In the following sections we describe the initiation mechanisms for PDCs and then focus on the structure and nomenclature of PDCs and their deposits. We only discuss small-volume end-member PDCs with typical volumes < 1 km$^3$ as opposed to large volume ignimbrites. Small-volume PDCs are relatively short-lived and occur in discrete pulses.

**Initiation Mechanisms of PDCs**

The four main initiation mechanisms identified and illustrated in Figure 2.1 are (1) the collapse of a sustained pyroclastic column, (2) the collapse of a short-lived fountain, (3) lava dome collapse and (4) the lateral or radial expansion of over-pressurized jets from flank collapse inducing sudden and violent decompression of a lava dome or pressurized chamber. This dissertation focuses on two types of generation mechanisms with the collapse of a sustained column at Calbuco volcano in Chile and the dome collapse eruption at Volcán de Colima in Mexico.
Figure 2.1. Simplified sketches of the four initiation mechanisms of PDCs. See text for explanations.
**Column collapse**

In the case of high-intensity eruption, the high degree of magma fragmentation leads to the explosion and injection of pyroclastic materials in the atmosphere forming a plume or column. The plume stability is relative to the mass flux of injected material, the speed of the jet, the conduit opening diameter, the temperature and air entrainment (Sparks et al., 1978). When the jet speed and mass flux decreases, or when the conduit diameter increases (by way of erosion of the inner wall), the plume does not incorporate enough material to sustain its vertical stability. The destabilization of the column triggers multiple PDCs (Figure 2.1a; Bursik and Woods, 1996; Branney and Kokelaar, 2002) and the complete collapse under its own weight is responsible for the largest PDC volumes (e.g., ignimbrites Carey et al., 1988). Rapid sedimentation of dilute PDCs can also lead to secondary concentrated PDCs (Calder et al., 1999; Druitt et al., 2002).

**Fountain collapse**

Also called ‘boiling over’, pyroclastic fountains are similar to plume with the difference that the air entrainment is not being sufficient to sustain a stable jet. Thus, the fountain does not rise and immediately collapses generating PDCs (Figure 2.1b). Pyroclastic fountains are often associated with short-lived explosions such as Vulcanian eruptions (e.g., Montserrat, Druitt et al., 2002). The PDCs resulting from lower intensity eruption and short-lived fountains are more localized spatially and called pumice-and-ash flow due to the rich pumice composition of the flows.

**Dome collapse**

Volcanic lava domes are solidified or semi-solidified extrusion of highly viscous lava. Lava dome eruptions can last many years or even decades, in which phases of long-term effusive eruptions (dome-building activity and lava flows) are interrupted by dome collapses and/or
explosions of ash and gas (Figure 2.1c). Lava domes, when exposed on the crater floor, are susceptible to gravitational instabilities, which can then lead to partial or complete collapse of their structure. Gravitational instabilities can be triggered by pure dome growth, by over-pressurization of the magmatic gases trapped inside the dome, or a combination thereof.

Based on the notion of balance between pressurization and mechanical resistance of the dome, a classification of the types of dome-collapse eruptions has been proposed by Sato et al. (1992): (i) Merapi-type with almost purely gravitational collapse, either by exogenic growth or by the pressure caused by endogenic dome growth (Ui et al., 1999), (ii) Pelean-type with moderate intensity explosion-driven collapse, and (iii) Soufriere-type with complete dome explosion.

The dome permeability exerts an important role over the explosivity and the types of associated PDCs (Boudon et al., 2015): a lava dome with impermeable outer layer (shell) will contribute to over-pressurization of the gases trapped within the dome, which will tend to result in explosion-driven collapse with more dilute PDCs (Pelean-type) because of a more intense fragmentation of the dome. Conversely, a more permeable dome structure with fractures will allow a continuous degassing, thus leading to pure gravitational collapse due to overgrowth (Merapi-type). As it collapses, the dome fragments rapidly, forming locks to ash size particles, coming together to form PDCs, generally referred to as ‘block-and-ash flows’, with relatively small volumes and localized to one flank of the volcano (Cole et al., 2002; Komorrowski et al., 2013; Ogburn et al., 2014).

**Radial/lateral expansion of over-pressurized jets**

This type of PDC generation mechanism is less common and has only been observed three times, during the famous Mount St. Helens eruption on May 18, 1980 (Kieffer, 1981; Lipman and Mullineaux, 1981), at Soufriere Hills Volcano on December 26, 1997 (Sparks et al., 2002; Voight
et al., 2002; Woods et al., 2002) and for the Bezymiyan eruption in 1956 (Gorshkov, 1959; Belousov et al., 2007). The growth of a lava dome, whether external or internal (cryptodome), may cause gravitational instabilities for the whole volcanic edifice causing a large part of the volcano flank to collapse eroding part of the dome with it (Figure 2.1d). The sudden release of the mechanical constraints over the pressurized internal portion of the dome creates a violent explosive decompression and fragmentation of the materials (Woods et al., 2002; Belousov et al., 2007). This explosion leads to a high intensity and high energy blast, typically composed of highly fragmented pyroclastic material, and may transform into a sustained Plinian column if juvenile magma is present within the conduit (Lipman and Mullineaux, 1981; Belousov et al., 2007). The flank collapse also transforms into a debris avalanche, which is composed of a wider range of particle sizes with much larger fragments (e.g., blocks with tens of meters in diameter) (Sparks et al., 2002; Voight et al., 2002).

**Nomenclature of PDCs and Deposits**

Fisher (1966) first described dense PDCs in the literature, referring to them as pyroclastic flows. Shortly after, Sparks et al. (1973) identified the dilute part and referred to them as pyroclastic surges. The differences in their deposits were attributed to the difference in transport and deposition mechanisms, and the two parts were further described as being the two extremes of one density continuum (Fisher, 1979; Carey, 1991; Valentine, 1987; Druitt, 1992; Branney and Kokelaar, 1992). Only then were pyroclastic flows and pyroclastic surges described as belonging to one entity called pyroclastic density current (PDC), in which each part follows a different flow dynamic (Druitt, 1998; Branney and Kokelaar, 2002).
Figure 2.2. Schematic illustration of the anatomy of a generalized PDC. a) Generalized morphology of a moving PDC in side view with the ground-hugging dense-basal avalanche and its overriding ash-cloud surge. Air entrainment of the ash-cloud surge result in thermal expansion, which transforms into a rising buoyant plume. b) Close-up on the internal structure of a moving PDC depicting the particle concentration gradient resulting in a bed-load and a suspended-load region separated by a transition zone. The different regimes for flow dynamics are indicated for the different regions of the flow. In the gas dominated system the particles are primarily influenced by particle-gas drag. In the bed-load region, momentum is redistributed by particle-particle collisions and as concentration increases, by frictional interactions. c) Cross-section of a channelized PDC showing the particle concentration gradient from the concentrated dense-basal avalanche to the dilute ash-cloud surge. Particles settling from the bed-load region result in massive deposits in the channel bed, while the dilute current results in stratified deposits. Sketches modified after Dufek (2016) and Gueugneau (2019).
The general model of the anatomy of a PDC is (i) a dense-basal part (i.e., the bed-load region), with high particle concentration that moves in contact with the ground because of its higher density with respect to the surrounding atmosphere, and (ii) a dilute part (i.e., the suspended-load region), where particles are transported within a gas matrix by turbulence (Figure 2.2). The suspended-load region likely grows with time as particles settles into the bed-load region and gas from the atmosphere is entrained into the current due to its high shear conditions (Dufek, 2016). The dilute part of the PDC may transform into a buoyant plume that rises upward as volcanic gases and the heated entrained atmospheric air expand and if enough particles have sedimented.

Studies of flow dynamics of PDCs show that the particle concentration of each part is the driving parameter that control their dynamics because it defines the physics involved in the particle transport (Burgisser and Bergantz, 2002; Sulpizio et al., 2014; Dufek et al., 2015; Dufek, 2016). The dynamic of the concentrated flow, also called dense-basal avalanche, is driven by a collisional and frictional regime in which the momentum transfer of kinetic energy occurred by interaction between particles (Figure 2.2b), and result in thick and massive deposits.

In the dilute flow, which we refer to as ash-cloud surge in this dissertation, the turbulent gas-particle drag regime controls the dynamic and results in thinner and stratified deposits (Figure 2.2c). In a very turbulent and energetic dilute PDC, the flow thickness of the bed-load region may remain negligible, and the particles will transition rapidly from suspension to deposition (Branney and Kokelaar, 2002).

The critical particle concentration at which one regime transitions to another is not well constrained. It was proposed at 30 wt.% by Dade and Huppert (1996), challenged by Weit et al. (2018) who propose a much lower percentage, being function of the Reynolds number. Studies by
Breard et al. (2016) and Breard and Lube (2017) experimentally showed the existence of a transitional zone half-way between a collisional regime and a turbulent regime.

Physical processes occurring during the late stages of particle transport, just before the solid load comes to rest, are recorded within the deposit architecture and sedimentological features. Focus of field studies of PDCs is therefore often centered on the deposits left by the concentrated, lower part of the current (bed-load region) that is never accessible for direct observations due to ubiquitous cover by large volumes of overriding ash clouds (e.g., Cole et al., 2005).

Concentrated PDC deposits typically consist of massive poorly sorted mixtures of large blocks set within a lapilli/ash matrix (e.g., Cas and Wright, 1987; Druitt, 1998). Sedimentary structures, granulometry and clast componentry are used to determine lithofacies of the PDC deposits. These lithofacies can then be used to infer depositional regimes occurring in the flow-boundary zone (Branney and Kokelaar, 2002; Sulpizio and Dellino, 2008; Sulpizio et al., 2014).

Moreover, PDCs preferentially channelized into the pathways carved on the flank of a volcano and are subject to the influence of the topography (e.g., Calder et al., 1999; Stinton, 2007; Andrews and Manga, 2011; Esposti Ongaro et al., 2012; Doronzo and Dellino, 2014; Ogburn et al., 2014; Brand et al., 2016). The extent of the influence of the topography primarily depends on the volume and energy of the PDCs relative to the complexity of the topography. Changes in the deposit lithofacies are commonly observed from topographical control (e.g., Schwarzkopf et al., 2005; Belousov et al., 2007; Lube et al., 2007, 2011; Sulpizio et al., 2008, 2014; Cronin et al., 2013; Komorowski et al., 2013; Doronzo and Dellino, 2014; Capra et al., 2016; Macorps et al., 2018; Marti et al, 2019; Ogburn and Calder, 2017).

Previous studies of deposits at various active volcanoes (e.g., Charbonnier and Gertisser, 2008, 2011; Gertisser et al. 2012; Schwarzkopf et al. 2005; Komorowski et al. 2013; Cole et al.,
1998; Calder et al., 1999; Saucedo et al., 2004; Sulpizio et al., 2010; Sarocchi et al., 2011; Miyabuchi, 1999) used detailed field observations for stratigraphic reconstructions and sedimentological analysis of the deposits as ways to better understand the dynamics and physical processes involved within PDCs. Crucial findings regarding PDC dynamics highlight the non-uniformity of the current in time and space, and show that unsteadiness of flow conditions arise from complex interactions with topography and obstacles (e.g., Schwarzkopf et al., 2005; Sulpizio and Dellino, 2008; Stinton et al., 2008; Charbonnier and Gertisser, 2011, 2013; Komorowski et al., 2013; Solikhin et al., 2015; Ogburn, 2015).

More specifically, the unpredictability of PDC dynamics involves the generation of overbank flows, which have been recently observed and described at Merapi for the 2006 and 2010 eruptions (Lube et al., 2011; Charbonnier et al., 2011, 2013; Komorowski et al., 2013) and at Volcán de Colima (Macorps et al., 2018). Overbank flows designate valley-derived, unconfined flows generated when the primary valley-confined PDCs escape channel confines and spread laterally away from the channel, sometimes over spilling onto valley interfluves and partially filling adjacent channels.

This complex behavior of PDCs can pose great threats to the surrounding population and infrastructures due to lateral and unconfined spreading of the main valley-confined flows, for which the flow path can then become highly difficult to predict. In addition, generation of overbank flows can impact the mobility of the main PDC body, hence increasing its runout and hazard potential (Gertisser et al., 2012). These studies converge towards evidence of a control of the topography on the flow dynamics and physical processes of deposition of PDCs and demonstrate the strength of field investigations to better understand these effects.
Volcán de Colima – Mexico

Geological Context and Tectonic Settings

Volcán Colima (19°31 N; 103°37 W; 3860 m above sea level, hereafter a.s.l.) is located at the western end of the Trans-Mexican Volcanic Belt (hereafter TMVB), a 1200 km-long E-W active continental volcanic arc extending across Mexico between the Gulf of Mexico and the Pacific Ocean (Figure 2.3).

![Figure 2.3. Location of the Trans-Mexican Volcanic Belt (TMVB) in Mexico. Volcán de Colima is located in the Western TMVB (white star).](image)

The TMVB is one of the two major volcanic arcs, along with the NNW-trending Sierra Madre Occidental (SMO) silicic province, to have formed during the Cenozoic as a consequence of the subduction of the eastern Pacific plates (i.e., Rivera and Cocos plates) beneath the North America plate at the Mid-American Trench (Atwater, 1989; Ferrari et al., 2000).
A counterclockwise rotation of the volcanic arc resulted in a progressive transition of the volcanic activity from the SMO to the TMVB, which activity began in the middle to late Miocene (~ 11.6 Ma) (Ferrari et al., 1999). The quaternary volcanism in Mexico has been primarily confined to the TMVB with large andesitic composite volcanoes, intervening scoria cones and rare rhyolitic complexes. Volcán Colima is Mexico’s most historically active andesitic composite volcano.

**Figure 2.4.** Tectonic settings of the Western TMVB. Schematic representation with the triple junction of the Tepic-Zacoalco Graben, the Chapala Graben and the Colima Graben. The latter is divided into a northern and a southern sector by the Tamazula Fault zone. The Colima Volcanic Complex (CVC) sits in the southern sector of the Colima Graben, and is composed, from north to south of Volcán Cantaro, Nevado de Colima, and Volcán de Colima (indicated by a red triangle).

The western TMVB consists of three major linear regional structures (Figure 2.4): the Colima graben, the Chapala graben and the Tepic-Zacoalco Graben, which together form a triple junction system (Luhr et al., 1985). The Colima graben extends SSW for about 90 km starting ~50 km south of Guadalajara. The Chapala graben extends eastward for 75 km and the Tepic-Zacoalco
Graben is a NW-trending 180 km-long chain of small composite volcanoes (Luhr and Carmichael, 1981). The Colima graben is thus bordered by N-S normal faults running from the junction with the Chapala and Zacoalco grabens to the north and by the Pacific Ocean coast to the south. The superposition of two tectonic regimes in this area, the subduction of the Rivera plate and the rifting process along the Colima graben, has generated composite volcanism (alkaline and calc-alkaline) during the last 4.6 Ma (Allan, 1986).

Figure 2.5. The Colima Volcanic Complex. The CVC is composed from North to South of Volcán El Cantaro, Nevado de Colima and Volcán de Colima. The City of Colima is indicated with a large red circle and is located ~ 30 km SSW of Volcán de Colima. The villages in which ash layers were recorded during the July 2015 paroxysmal event, are indicated with small red circles.
The Colima Volcanic Complex (CVC) sits in the southern sector of the Colima graben (Figure 2.4) which cuts the limestone platform of the Colima basin (Cortes et al., 2010). The CVC and the NE-SW Tamazula fault are the two geological features that separate the southern from the northern sector of the Colima graben (Garduño et al., 1998). The CVC is built upon the Lower Cretaceous Tecalitlán, Encino, Tepames and Coquimatlán Formations, as well as volcanic and plutonic Tertiary rocks (Cortes et al., 2005). The CVC is composed, from N to S of Volcán El Cantaro, Nevado de Colima and Volcán de Colima (Figure 2.5).

During the Quaternary, in the CVC, the andesitic volcanic centers migrated southward with time. It began with the emplacement of Volcán El Cantaro 1.7 Ma (Allan et al., 1991), followed by Nevada de Colima 0.53 Ma (Robin et al., 1987). During the late Pleistocene and coeval to Nevado de Colima activity, Paleofuego volcanic complex began its formation 5 km to the south (Robin et al., 1987). Robin et al. (1987) suggests that the summit of Paleofuego may have been higher than the current Volcán Colima and possibly taller than Nevado de Colima.

Paleofuego was destroyed during a Mount St Helens-style flank collapse, leaving behind a caldera structure of 5 km in diameter with an opening axis towards the south (Luhr and Prestegaard, 1988; Cortes et al., 2010). This collapse destroyed the previous volcanic structure and generated a large debris avalanche that traveled towards the south for over 65 km. The age of the collapse event is argued to have occurred at 9370 +/- 400 yr B.P. (Robin et al., 1987), at 4300 yr B.P. (Luhr and Prestegaard, 1988) or at 3600 +/- 120 yr B.P. (Komorowski et al., 1997).

Volcán Colima began erupting inside the caldera left by the gravitational collapse of Paleofuego (Cortes et al., 2010), of which the northern wall remains visible to date. Basic alkali magmatic centers migrated in time from W to E towards Volcán El Cantaro, with activity recorded during the late stages of Nevado de Colima’s life and through the early stage of Volcán Colima.
The most recent basic alkali magma scoria cone is Volcán Apaxtepec, which is the only one east of the andesitic Colima-Cantaro axis. Luhr and Carmichael (1981) describe the southern end of the Colima graben as the volcanic analog to classical, post-plutonic, hypabyssal lamprophyre localities.

The present edifice of Volcán de Colima (Figure 2.6) is a stratovolcano consisting of pyroclastic flow deposits, tephra fall deposits and lava flows that have reached up to 5 km from the summit. It occupies an area of 20 km² with a total volume estimated around 10 km³ (Luhr and Carmichael, 1990a). On June 13, 1869, the ascent of a lava spine found an opening on the NE flank of the main cone and led to the formation of the “parasitic” cone Volcancito (Luhr and Carmichael, 1990a). Blocky lava flows from Volcancito flooded the NE portion of the caldera floor.

Figure 2.6. Ground photograph of Volcán de Colima. The photograph was taken from the SSW during field work in February 2016, to show the relative proximity of the volcano to farmlands.
Historical Eruptions

Volcán Colima, also referred to as “Fuego de Colima” or “Volcán de Fuego” in the literature, is one of the most historically active volcanoes in North America. The oldest known reference to its eruption dates from 1560, which is two decades after the Spanish Conquest of Mexico (Medina-Martinez, 1983). This would lead to a total record of activity of ~ 460 years. Medina-Martinez (1983) used compiled information from the 16th and 17th centuries from the reports of the “Archivo de Indias” and from a catalogue published by Arreola (1915). Continuous activity since the late 18th century led several naturalists to write reports, some of which were published in the proceedings of the Sociedad Cientifica Antonio Alzate in the late 1800s (Medina-Martinez, 1983).

The frequent and often spectacular activity of Volcán Colima during the 19th century also led to the installation of two observatories that carried out observations continuously for 12 years (Medina-Martinez, 1983). Additionally, daily reports were kept between 1893 and 1905 (Diaz, 1906) and Waitz (1936) wrote a complete report of the Plinian eruption of 1913. The earliest reliable available report of an eruption dates back to 1576 (Luhr and Carmichael, 1980), although Medina-Martinez (1983) argued that the 1560-1750 period has scarce information on the details of activity with only the dates of the most relevant signs of activity (e.g., explosive eruption with noticeable flows). While data should be taken and used with caution, it is worth noting that eruptions were reported as consisting of “abundant ash and hot material fall out”, reaching the city of Colima 30 km away (Arreola, 1915) and “causing darkness and need for candlelight during the day” (from Medina-Martinez, 1983). This suggests explosive activity of relatively high intensity with a Volcanic Explosion Index (VEI) ≥ 3, and is in agreement with historical eruption record interpretations from Luhr and Carmichael (1990b), De la Cruz-Reyna (1993) and Luhr (2002).
Historical compilation records count 57 total eruptions, six of which (1585, 1606, 1622, 1818, 1890 and 1913) were classified as VEI 4 eruptions (De la Cruz-Reyna, 1993). The VEI is a scale of increasing explosivity with index values from 0 to 8, each representing about an order of magnitude increase (Newhall and Self, 1982).

Since the first recorded eruption, Volcán Colima has displayed a pattern of eruptive cyclicity characterized by effusive activity that builds lava flows and a lava dome inside the open crater, and explosive activity that partially and/or completely destructs the summit lava dome and generates tephra and pyroclastic density currents. The effusive phase is often referred to as a “constructive phase” while the explosive phase is considered “destructive phase” and the paroxysmal (short cataclysmic event) phase of a cycle. The effusive phase may be continuous for months to decades prior to the explosive phase. Such behavior is commonly observed at andesitic edifices. Robin et al. (1991) suggests that the explosive phase is the result of a magma mixing process while the effusive phase corresponds to the magma differentiation phase. Robin et al. (1991) argues that the cycles begin with the explosive phase while Luhr and Carmichael (1980) suggest that it ends with explosive paroxysmal event.

**Last Eruptive Activity Prior to the Events of July 2015**

**2004 – 2005 activity**

The 2004-2005 activity began at the end of September 2004 with the formation of a new lava dome in the crater. The dome overflowed the northern and northwestern crater rim, forming two blocky lava flows on the N and WNW flanks that reached 2.4 km in November 2004 (GVP, 2005a). A partial dome collapse on October 6, 2004, generated BAFs in La Lumbre ravine that reached 6 km, which was, at the time, the longest runout recorded at Volcán de Colima since the
Plinian eruption of 1913 (Sulpizio et al., 2010). The termination of lava effusion at the end of November 2004 was followed by intermittent explosive activity dominated by small gas-and-ash Vulcanian eruptions (GVP, 2005a). Between late December 2004 and February 2005, there were at least 15 explosions with eruptive columns up to 3 km h.a.c. (height above crater) recorded, suggesting a gradual increase in the pressurization of the system (Zobin et al., 2006).

Between February and September 2005, the activity showed several major Vulcanian explosions with eruptive column up to 5 km h.a.c., accompanied by an undetermined number of PDCs with maximum runouts within 3 km from the summit (Sulpizio et al., 2010; GVP bulletin, 2005b). The recurring explosions during that time completely removed the 2004 lava dome and left behind a 30-m-deep crater (GVP, 2005b). On June 5 and 9, 2005, there were two remarkably long PDCs emplaced along the Montegrande and La Arena ravines with runouts of 5.1 and 5.4 km respectively (Gavilanes-Ruiz et al., 2009; Sulpizio et al., 2010). The eruptive activity between October 2005 and December 2006 remained of small steam-and-ash Vulcanian-type, with decreasing intensity.

2007 – 2011 activity

Beginning in January 2007, there was an increase in the number of small explosions but a decrease in both its energy and ash fraction. This was characteristic of new dome growth in the crater, which was first observed in February 2007 (GVP, 2007). The dome growth was continuous at a slow rate between February and September 2007. The dome growth increased exponentially between October 2007 and March 2008, then increased again between August and November 2008. By March 2009, the dome was occupying about 80% of the crater (GVP, 2009). There were multiple daily explosions of gas that did not disrupt the dome.
In the fall 2009, the new dome growth reached the western rim of the crater, and in February 2010, a new lobe formed out of the western dome margin promoting an increasing number of rockfall events (GVP, 2010). This illustrated a change in dome growth mechanism from endogenic to exogenic. A steady growth of the dome continued through March 2011 accompanied by numerous rockfalls and small gas-and-ash explosions, with the western lobe reaching ~ 55 m in length and the main dome overspilling on the southern flank (GVP, 2011). The same type of activity, i.e., dome growth, rockfalls and small explosions, continued with a decline in growth rate and explosions, until a final explosion on June 21st, 2011 that removed part of the dome. This final explosion was followed by a period of repose with dropped seismicity and all visible signs of dome growth ceased (GVP, 2013a).

Figure 2.7. Aerial photo of the 2013 lava dome at Volcán de Colima. This annotated photograph shows the filled crater and the lava flow on the western slope. It was taken on April 19, 2013, during a flight of Civil Protection of Jalisco State. The black arrow is pointing south. Figure modified from the Bulletin 38:12 from the Global Volcanism Program (GVP, 2013b).
2011 – 2013 activity

After a 1.5 year-long period of relative calm, the eruption resumed with small Vulcanian explosions on January 6th, 2013. The explosions generated a depression inside the dome-crater and large ejecta near the crater rim (GVP, 2013a). Two more explosive events occurred on the 13th and 29th of January, excavating a 250,000 m$^3$ crater (GVP, 2013b). A new rapid dome growth was recorded between February and the end of March 2013, accompanied by an increase in the frequency and energy of small explosions. The dome filled the crater by April 2013 (Figure 2.7) after which a small lava flow traveled down the W flank together with frequent rockfalls due to the flank steepness (GVP, 2013b).

Dome growth ceased during the period of April through November 2013, although daily small explosions and rockfalls remained constant. The rainy season was characterized by 14 recorded block-rich lahars that descended the flanks of the volcano including a 6 km-long lahar that descended into the Montegrande ravine on June 11th, 2013, and a 6 hour-long lahar recorded on September 16th, 2013 during the tropical cyclone Manuel on the Pacific coast (GVP, 2013b).

2014 activity

Between January and May 2014, there were 10 reports of intermittent ash emission issued by the Washington VAAC (Volcanic Ash Advisory Center). The frequency of rockfalls, which had not stopped since April 2013, remained constant until early March 2014. Similarly, the frequency of explosions remained stable until March 2014 after which both rockfall and explosion events tapered off. In May 2014, the effusive activity resumed and formed a lava flow on the W flank of the volcano associated with the extrusion of a new lava dome. The lava dome overspilled from the crater and led to the emplacement of a 2.4 km-long lava flow on the SW flank in late September 2014 (Davila et al., 2019). This effusive activity was accompanied by an abrupt
increase of rockfalls in July 2014, reaching a peak of 218 daily events on September 16\textsuperscript{th} (GVP, 2015). The rainy season (June – September) was characterized by multiple lahars, including ten major lahars recorded in the Montegrande ravine and three in the La Lumbre ravine. On the 21st of November 2014, the explosive activity resumed with an ash-plume that rose ~ 7 km (h.a.c.) and collapsed into a 3.1 km-long PDC that descended into the San Antonio ravine and the dilute cloud travelled 2.9 km along the Montegrande ravine (GVP, 2015a). Another larger explosion was recorded on November 30\textsuperscript{th}, 2014 and two smaller eruption followed on December 4\textsuperscript{th} and 24\textsuperscript{th} with associated ash-plumes up to 5 km (h.a.c.). The November – December explosions destroyed the dome formed in early 2013 (GVP, 2015a).

**Chronology of the Paroxysmal Events of July 2015**

**Prior to the paroxysmal event**

The explosive activity of 2015 was extensive overall, resulting in 751 advisories by the Washington VAAC (GVP, 2016). It began on January 3\textsuperscript{rd}, 2015 with an explosion that generated an ash-plume up to ~ 3 km (h.a.c.), dispersing ash up to 175 km away, towards the NE (GVP, 2015b). Another large explosion occurred on January 21\textsuperscript{st}, with a plume up to 4 km (h.a.c.) and ash dispersal up to 74 km in the NE direction (GVP report, 2015b). This series of explosions in January 2015 caused the partial destruction of the summit lava dome and left a 140 m-wide crater. Active lava flows on the W and WNW flanks were reported by the Unidad Estatal de Protección Civil de Colima (UEPCC) on January 24 (GVP, 2015c).

Between February and April 2015, several ash plumes were reported daily with altitudes between 5.5 and 7.3 km (a.s.l.), and persistent ashfall within a 30 km-radius, although generally trending towards the NE (GVP, 2016). The highest plumes of 8.5. km (a.s.l.) were recorded in late
April (GVP, 2015d), after which the frequency and intensity of ash plumes decreased through early July, with ash emissions recorded every few days as opposed to multiple times a day (GVP, 2016). This change in activity was concurrent with a new dome extrusion phase, which was reported in May 2015 with the detection of a thermal anomaly inside the crater (GVP, 2015e). The lava dome was overflowing the crater on the S rim by the beginning of June 2015. In early July, there were two main lava flows advancing rapidly (few hundreds of m/day) on the northern and southern flank (Davila et al., 2019). Between the 7th and the 10th of July, the frequency of ash plumes, rock falls and small BAFs increased (GVP, 2015f). BAFs travelled at most distances of 2.5 km on the N, W and S flanks, and were observed with the ejection of incandescent material from the crater. By the 10th of July, the lava flow on the southern flank was ~ 700 m long.

**Paroxysmal event**

The paroxysmal phase of the eruption began on the 10th of July at 12:00h (24h, local time) with a large explosion that produced an ash plume that rose up to an altitude of 7.6 km (a.s.l) and drifted W for 150 km (GVP, 2015f). The activity further increased at 20:17h with the dramatic 52 min-long sustained collapse of the summit lava dome, as recorded by the seismic signals from a broadband station from the Seismological Network at Colima University (RESCO) and the Centro Nacional de Prevención de Desastres (CENAPRED) located 6.5 km from the summit in the Montegrande ravine (Reyes-Davila et al., 2016; Capra et al., 2016). The sustained-collapse generated a first series of BAF pulses that emplaced along the Montegrande ravine and stopped before the large volcaniclastic plain area located 9 km from the summit (Macorps et al., 2018). Coevally, a 4 km-high (h.a.c) ash plume associated with the overriding ash-cloud of the dense BAFs (Figure 2.8a-b) rose up and drifted to the SW, as first recorded by the video monitoring equipment from RESCO and CENAPRED in La Lumbre ravine (Capra et al., 2016 and personal
communications with L. Capra). Ash fall was then recorded in towns up to 18 km SW from the volcano, including La Yerbabuena where a 5 cm-thick ash layer blanketed the town, La Becerrera, San Antonio, Carrizalillo, El Naranjal and Suchitlán (Figure 2.5). The UEPCC ordered the evacuation of residents in La Yerbabuena and assisted in the voluntary evacuation in nearby villages, totaling 70 people (GVP, 2015f; Capra et al., 2016). No precursory seismic activity was recorded prior to the first dome-collapse.

At 13:30h on the 11th of July, a second 1h47 min-long sustained collapse of the lava dome and large portion of the southern crater rim triggered the emplacement of a second series of BAF pulses (Figure 2.8c) into the Montegrande ravine up to 9.3 km, which were also partially rechanneled into the adjacent San Antonio ravine with a maximum runout distance of 6.1 km (Macorps et al., 2018). Figure 2.9 shows photographs of the aftermath in the volcaniclastic plain at the bottom of the Montegrande ravine. Ash fall was reported in towns up to 30 km to the SW, including Comala, Villa de Alvarez and Colima (GVP, 2015f). Davila et al. (2019) suggest that the second event was caused by the collapse of the new fast-growing dome, which can be agreed upon with the sustained lava flow that began to flow immediately after on the southern flank, reaching a maximum distance of 2.5 km by the 6th of August (Capra et al., 2016). The V-shaped scarped left behind in the summit area suggests that the partial collapse of the southern flank also contributed to the generation of the second BAF event (Macorps et al., 2018).

**Post paroxysmal event**

Over the following days, ash emissions, incandescent rockfalls from the advancing lava flows and small BAFs continued at a moderate level and evacuations continued (GVP, 2016). Four days later, on July 15th, Washington VAAC reported that ash emissions had ceased, however, lava flow rockfalls continued. Ash plumes resumed on July 31st with daily to hourly occurrences, and
continued throughout the year (GVP, 2016). By October 2015, the summit crater was 200 m-wide and 50 m-deep according to a summit overflight carried out by the UEPCC. The rainy season was marked by Hurricane Patricia in October 2015, which contributed to the generation of multiple stream-flows and lahars, cutting into the existing deposits and reworking them to longer runout distances.

Figure 2.8. Photographs of the PDCs from July 10, 2015 at Volcán de Colima. a) Photograph of the moving ash-cloud surge during descent of the block-and-ash flows at Volcán de Colima taken on July 10, 2015 by a worker of the Civil Protection. b) Photograph of burning tree trunks transported to the bottom of the Montegrande ravine by the block-and-ash flows. c) Photograph of the moving ash-cloud surge, near the electric tower at the bottom of the Montegrande ravine, taken on July 10, 2015 by a worker of the Civil Protection. d) Snapshot of the moving block-and-ash flow recorded on July 11, 2015 at 13h33 by a webcam posted on top of a seismic station in the Montegrande ravine, two seconds before its complete destruction.
Figure 2.9. Aftermath of the July 10-11, 2015 dome-collapse at Volcán de Colima. a) Photograph of block-and-ash flows in the alluvial plain at the end of the Montegrande ravine, taken on July 13, 2015 by a worker of the Civil Protection from the top of the electric tower shown in Figure 2.7a. b) Photograph of skeleton remains of farm animals trapped during the descent of the block-and-ash flows (credit – E Macorps). c) Photograph of a snapped and charred tree caused by the block-and-ash flows in the distal part of the Montegrande ravine (credit – E. Macorps).

During the following year of 2016, the activity was characterized by small explosions and the emplacement of PDCs with maximum runout of 2.5 km. For the period of January – April 2016, there were hundreds of ash emissions recorded (Figure 2.10), and a new, slow-growing lava dome was observed on February 17th (GVP, 2017). The activity slowed down between May and September, although there were still multiple explosions per week. On September 30th 2016, the lava dome overflowed the crater rim, creating a new lava on the SW flank and reached a distance of 2.2 km by the end of October 2016 (GVP, 2017). A second lava flow began on the S flank in mid-November 2016 and reached 1.7 km in length on December 5th (Figure 2.11; GVP, 2017).
Frequent explosive activity was common between the end of 2016 and early 2017, often generating ash plume up to 4 km (h.a.c.) and ash fall in nearby towns. The strong explosion of February 3rd 2017 generated a 6 km (h.a.c.) ash plume and small pyroclastic flows on the E flank (GVP, 2017). Between February and March 2017, the intensity and frequency of explosions decreased significantly. After mid-March 2017, the seismicity decreased steadily and explosive activity ceased completely, leading to a calm period of two years until renewed activity in April 2019 (GVP, 2019).

**Figure 2.10.** Field photographs of the activity at Volcán de Colima. The photos were taken during the 2016 field campaign and show the frequent but small explosions of ash and gas.

**Figure 2.11.** Photographs of the southern lava flow of Volcán de Colima a) from February 2016 showing the opened crater rim onto the Montegrande ravine and the extent of the lava flow that began after the July 2015 dome collapse (credit – E. Macorps). b) From January 2017 showing the filling of the channel and closure of the crater rim by the progressing lava flow (courtesy of L. Capra).
Calbuco Volcano – Chile

Geological Context

Calbuco Volcano (1974 m a.s.l.) is a composite stratovolcano located in the Southern Andes of Chile (41°20’S, 72°37’W; Figure 2.12) and is among the most active and most hazardous volcanoes in Chile (Petit-Breuilh, 1999; Clavero et al., 2008; Castruccio and Clavero, 2015). The Southern Volcanic Zone of the Andes (SVZ; Figure 2.12) is the result of the subduction of the oceanic Nazca Plate under the continental South American Plate (Figure 2.12; López-Escobar et al., 1992, 1995). The volcanic arc extends from latitude 33° to 46° and is bounded by the intersection of the Juan Fernandez Ridge with the continental margin at ca. 32°S and the Chile Rise triple junction at ca. 47°S (Figure 2.12; López-Escobar et al., 1992).

Calbuco Volcano is located in the southern SVZ (SSVZ; 37° - 46°S), which is characterized by a 30 km-thick continental crust and dominant basaltic and basaltic-andesitic volcanic products (Stern, 2004 and references therein), although Calbuco volcano is almost exclusively of andesitic composition, similar to the geochemical signatures of the northern SVZ (Hickey-Vargas et al., 1995). The basement underneath Calbuco volcano comprises meta-sedimentary rocks (Upper Paleozoic; Parada et al., 1987), plutonic rocks (tonalities, diorites and granites; Neogene, Parada et al., 1987), and Early Pleistocene volcanic and volcanoclastic sequences (Lahsen et al., 1985).

The edifice morphology is a truncated cone shape covering an area of 150 km² (Lahsen et al., 1985) that has been built over the last ~ 300 ky (Selles and Moreno, 2011). The cone mainly consists of blocky and “aa” type andesitic lava flows interbedded with block-and-ash flow deposits from dome-forming eruption, as well as “hot” and “cold” lahars (López-Escobar et al., 1992; Castruccio et al., 2010; Selles and Moreno, 2011; Castruccio and Clavero, 2015; Castruccio et al., 2016). Hot lahars are particularly characteristic of Calbuco volcano, as opposed to cold lahars that
occur at the vast majority of volcanoes in the Southern Andes (Moreno et al., 2006). The lower northern flank of the volcano is characterized with a distinctive hummocky terrain that is the result of two major sector collapses that occurred in post-glacial times (Clavero et al., 2008).

Figure 2.12. Location of Calbuco Volcano in the Andes South Volcanic Zone (SVZ) and regional tectonic settings.

The morphologic evolution of Calbuco, recording its activity from Mid-Pleistocene to present (Lahsen et al., 1985; Hickey-Vargas et al., 1995; Selles and Moreno, 2011), has been reconstructed by López-Escobar et al. (1992) and Selles and Moreno (2011) who distinguish four stages, named Calbuco 1, 2, 3 and 4 (Figure 2.13).
Calbuco 1 (Mid – Upper Pleistocene; ca 340-110 ka) mainly consists of porphyritic basaltic-andesitic lava flows from effusive activity, interbedded with volcaniclastic deposits (“volcanic breccias” in Romero et al., 2016) (Lahsen et al., 1985; Selles and Moreno, 2011).

Calbuco 2 (Upper Pleistocene, ca. 110 – 14.5 ka) constitutes the main cone and is made of glacially-eroded andesitic to dacitic (56 – 61 wt.% SiO$_2$) lava flows interbedded with thick pyroclastic rocks, flow and fall deposits from dome-forming eruptions with associated PDCs (block-and-ash flows), and volcanic avalanche deposits from recurrent lahars (Selles and Moreno, 2011). One of the two major sector collapses, which was dated at ~ 14 ka, produced a debris avalanche of 2-3 km$^3$ covering an area of ca. 55 km$^2$ in the NNW direction, reaching a maximum thickness of 200 m, forming an elliptical caldera open toward the NNE and contributing to shape the distinctive hummocky terrains present on the lower northern flank of the volcano (Figure 2.12; Lahsen et al., 1985; Selles and Moreno, 2011).

During the Calbuco 3 stage (Upper Pleistocene – Holocene, ca. < 14 ka – 1893 BC) the activity was focused inside the caldera of Calbuco 2. The stratigraphic sequence of Calbuco 3 comprises andesitic to basaltic andesite (54 – 60 wt.% SiO$_2$) and dacitic (64.5 wt.% SiO$_2$) lava flows interbedded with breccias and tuffs, a basaltic-andesitic ignimbrite emplaced about 6.5 ka, and volcanic breccias, PDC and lahar deposits, dominantly distributed towards the S and N sectors of the volcanic edifice (Selles and Moreno, 2011). The products from Calbuco 3 stratigraphic sequence, mainly the coarse-grained tephra fallout deposits, suggest sub-Plinian to Plinian eruption types (Watt et al., 2011).

Finally, Calbuco 4 sequence (Historical eruptions < 1893 BC – present) records the youngest history from the volcano including a series of basaltic andesite to andesitic (54.8 – 59.3% SiO$_2$) blocky lava flows, a series of lava dome eruption with associated block-and-ash flows and
tephra falls, historic and recent lahar deposits as well as sedimentary deposits due to glacier environment (e.g., moraines, fluvial, lacustrine and alluvial deposits) (Figure 2.13; López-Escobar et al., 1995; Selles and Moreno, 2011). The glaciers on the edifice have a small volume and a seasonal snow cap above 1000 m a.s.l. (Lahtsen et al., 1985).

Figure 2.13. Geological map of Calbuco Volcano. The map shows the four stages of Calbuco described in the text, and the underlying units. *(Modified from López-Escobar et al., 1995)*
**Historical Eruptions**

The historical activity at Calbuco volcano was summarized by Petit-Breuilh (1999). It started in the mid-XIX century and included Strombolian to sub-Plinian eruptions during three major eruptive cycles (1893-1895; 1929 and 1961) with the emission of lava flows, tephra fallout and block-and-ash flows (Petit-Breuilh and Moreno, 1997; Moreno, 1999). It also included Vulcanian eruptions (1906-1907, 1917, 1932, 1945 and 1972).

According to Petit-Breuilh (1999), explosive eruptions occurred in 1792, 1845, 1893, 1894-95, 1906-07, 1917, 1927, 1929, 1932, 1945, 1961, and 1972. Three of these eruptions have been classified with a VEI > 3. The historical eruption of Calbuco with the highest intensity was the Plinian eruption of 1893, which produced tephra fallout, PDCs and hot lahars, and was the focus of some historical volcanology studies (e.g., Pohlmann, 1893; Fisher, 1893; Martin, 1895, Lahsen et al., 1985).

Prior to April 2015, the last major eruption occurred in February 1961, about 9 months after the 1960 magnitude 9.5 Valdivia earthquake, and lasted for about 3 months (Klohn, 1963; Gonzales-Ferran, 1995; Petit-Breuilh, 1999). The eruption developed a volcanic plume of ~ 12 km high above the crater and dispersed ~ 0.07 km$^3$ Dense Rock Equivalent (~ 0.2 km$^3$ bulk) volume of tephra, which affected the NE sector of the volcano including cities in Argentina (Gonzales-Ferran, 1995; Petit-Breuilh, 1999; Daga et al., 2014, Romero et al., 2013).

After the last eruption in 1972, the only activity reports were those of a weak fumarole emission in May 1995 and a strong fumarole emission in August 1996 (GVP, 1996). No activity had been reported since, until the sudden renewed seismic activity and the eruption of April 2015 (GVP, 2015g).
The April 2015 Eruption

Between December 2014 and March 2015, there was an increase in seismic activity recorded on the few stations around Calbuco (SERNAGEOMIN report, 2015a, https://www.sernageomin.cl/). Three hours before the eruption, seismic activity and underground noises were felt and reported by inhabitants near the volcano (~ 5 – 10 km from the summit). Meanwhile, the stations recorded a significant increase above background levels of the shallow seismicity below the volcano with more than 200 volcano-tectonic and long-period earthquakes (SERNAGEOMIN report, 2015b; Castruccio et al., 2016).

Then, for the first time in 43 years, a new eruptive cycle began on April 22, 2015, at 18:05h local time with a 90-min sub-Plinian eruptive phase (Figure 2.14a-b). Within a few minutes it produced a sustained ash-rich column up to 15 km above the crater, dispersing tephra primarily towards the ENE direction (SERNAGEOMIN report, 2015c). Minor column collapses on all flanks of the volcano and primary lahars on the southern flank were occasionally reported (Mella et al., 2015).

After a 5h30min-break, a second phase of greater intensity began on April 23, 2015, at 01:00h local time, producing a sustained ash-rich column that reached a maximum height of 28 km above sea level according to weather radar (Vidal et al., 2015; Bertin et al., 2015; Van Eaton et al., 2016; SERNAGEOMIN report, 2015d). Although the second phase occurred in the middle of night, the impressive lightning and incandescence observed (Figure 2.14c) suggested an increased occurrence of PDCs within two and half hours, i.e., ~ 3:30h local time (Van Eaton et al., 2016; Castruccio et al., 2016). Witness accounts report incandescent ejecta observed up to 5-km away from the summit (SERNAGEOMIN report, 2015d). This second phase lasted for about 6 hours, with a similar tephra dispersion to the NE.
Figure 2.14. Photos of the April 22, 2015, eruption at Calbuco Volcano. a) Photograph taken from Puerto Montt (~30 km SSW). Credits: photo by Keraunos ob – posted online on the Earth of Fire blog by Bernard Duyck – obtained from the Global Volcanism Program website. b) and c) Photographs of the eruption from Puerto Varas (~ 25 km E). Photos posted online by the BBC (https://www.bbc.com/news/world-latin-america-32425370)
In the next few days, the activity decreased in intensity, with sporadic weak plumes between 400 m- and 2 km-high (above crater). On April 30, 2015, a third eruptive phase of much lower intensity and magnitude than the first two, started at 23:30h local time and lasted 2 hours. The eruption column reached about 5 km above crater with tephra dispersion to the SE (Segura et al., 2015; SERNAGEOMIN report, 2015e). The low juvenile content of the tephra released from this third explosion indicates a steam-driven eruption with little involvement of fresh magma (Van Eaton et al., 2016).

**Figure 2.15.** Infrastructure damages from tephra and lahars at Calbuco Volcano during the April 2015 eruption. Photographs were taken during the field campaign in November 2016. a) and b) are photographs of a fish farm to the south of Calbuco that was destroyed by lahars. c) and d) are photographs of a house covered by tephra fall that led to the partial collapse of the structure to the NE of Calbuco.
During the first two eruptive phases, PDCs reached 6.8 km from the vent on the NE flank in the Rio Blanco Este and on the SW flank in the Rio Blanco Sur. Subsequent lahars reached Chapo Lake on the southern flank, destroying numerous houses and fish farming buildings (Figure 2.15a-b). The erosional impacts by lahars consisted in bedrock erosion, incision of alluvial channel, erosion of surficial deposits due to the loose nature of fresh PDC deposits and felling of large areas of forest (Russell et al., 2015).

Large amounts of tephra fall deposits resulted in collapsed buildings and blocked roads up to 25 km NE of the volcano with 15 to 40 cm-thick accumulated tephra (Figure 2.15c-d; Castruccio et al., 2016). Over 6500 people were evacuated within 20 km of the volcano while the ash dispersed towards the NE across Chile, Argentina and Uruguay, causing damages to buildings, rendering roads and bridges impassable, affecting the agriculture and resulting in a severely disturbed air-traffic (Van Eaton et al., 2016). The Oficina Nacional de Emergencia del Ministerio del Interior (National Office of Emergency of the Interior Ministry, ONEMI) also called for people to stay within 200-m of the main drainage networks (valleys, rivers) due to lahars (GVP Bulletin, 2015g).

In the weeks following the paroxysmal eruption, the eruptive and seismic activity decreased progressively, and the alert level from SERNAGEOMIN was lowered back to orange on May 19, to yellow on May 28 and back to green on August 2015 (Castruccio et al., 2016; GVP Bulletin, 2015g).

Regarding the eruption dynamics, Castruccio et al. (2016) suggested that the triggering mechanism was either volatile oversaturation due to crystallization or volatile addition a small intrusion into the base of the magma chamber, without significant magma mixing or with an intruding magma compositionally similar to that of the residing magma. In either case, the triggering mechanism generated convection and sufficient over-pressurization of the system,
which favored fracturing of the surrounding rock and opening of a conduit to the surface, to cause the onset of the first eruptive phase. The start of the eruption resulted in decompression of the magma chamber, thus promoting intense vesiculation of the remaining magma and an increase in eruption rate towards the end of the eruption.

This is consistent with the timing of the deposits proposed by Van Eaton et al. (2016) using satellite observations of the eruption. The low content of lithic clasts together with the morphology and vesicularity of the majority of juvenile clasts indicate that the first and second eruptive phases were essentially magmatic with little to no involvement of external water (Castruccio et al., 2016). During the third eruption phase, however, the white eruption plume and the low juvenile content of the deposits suggest steam-driven explosion with little involvement of magma (Van Eaton et al., 2016). The variation in vesicularity between the types of juvenile clasts (i.e., brown scoria and dense dark-grey juvenile) but with consistent phenocrysts assemblages can be attributed to variable ascent rates, decompression and degassing of the magma chamber during an eruptive phase (Castruccio et al., 2016).

References


CHAPTER THREE:

STRATIGRAPHY, SEDIMENTOLOGY AND INFERRED FLOW DYNAMICS
FROM THE JULY 2015 BLOCK-AND-ASH FLOW DEPOSITS
AT VOLCÁN DE COLIMA, MEXICO

Abstract

The July 2015 block-and-ash flow (BAF) events represent the first documented series of large-volume and long-runout BAFs generated from sustained dome collapses at Volcán de Colima. This eruption is particularly exceptional at this volcano due to (1) the large volume of BAF material emplaced (0.0077 ± 0.001 km3), (2) the long runout reached by the associated BAFs (max. ~10 km), and (3) the short period (~18 hours) over which two main long-sustained dome collapse events occurred (on 10 and 11 July, respectively). Stratigraphy and sedimentology of the 2015 BAF deposits exposed in the southern flank of the volcano based on lithofacies description, grain size measurements and clast componentry allowed the recognition of three main deposit facies (i.e., valley-confined, overbank and ash-cloud surge deposits). Correlations and lithofacies variations inside three main flow units from both the valley-confined and overbank deposits left from the emplacement of the second series of BAFs on 11 July provide detailed information about: (1) the distribution, volumes and sedimentological characteristics of the different units; (2) flow parameters (i.e., velocity and dynamic pressure) and mobility metrics as inferred from associated deposits; and (3) changes in the dynamics of the different flows and their material during

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emplacement. These data were coupled with geomorphic analyses to assess the role of the
topography in controlling the behaviour and impacts of the successive BAF pulses on the volcano
flanks. Finally, these findings are used to propose a conceptual model for transport and deposition
mechanisms of the July 2015 BAFs at Volcán de Colima. In this model, deposition occurs by rapid
stepwise aggradation of successive BAF pulses. Flow confinement in a narrow and sinuous
channel enhances the mobility and runout of individual channelized BAF pulses. When these
conditions occur, the progressive valley infilling from successive sustained dome-collapse events
promote the overspill and lateral spreading of the upper and marginal regions of the main flow
body, generating highly mobile overbank flows that travel outside of the main valley. Volume-
and distance-dependent critical channel capacities for the generation of overbank flows are used
to better estimate the inundation area of these hazardous unconfined pyroclastic flows. These
results highlight the importance of including and correctly assessing the hazards posed by large
volume and long runout BAFs associated with frequent, small VEI, sustained dome-collapse
eruptions.

Introduction

Pyroclastic density currents (PDCs) are among the most devastating and least predictable
volcanic phenomena on Earth. They result from explosive eruptions by the collapse of either
sustained pyroclastic columns, short-lived fountains or lava domes, or by the lateral/radial
expansion of over-pressurized jets. PDCs are mixtures of volcanic fragments, ash and gas that are
initially denser than the surrounding atmosphere and travel across the ground under the effect of
gravity (Carey, 1991; Druitt, 1998; Freundt and Bursik, 1998; Sulpizio et al., 2014; Dufek, 2016).
Historical evidence and past studies have shown that PDCs can cause near complete destruction
of widespread areas and represent serious threats to surrounding populations and infrastructure (Tilling and Lipman, 1993; Tanguy et al., 1998). Due to these risks, assessment and mitigation of PDC hazards has been a main focus of the international volcanology community during the last 30 years. Block-and-Ash flows (BAFs) are small volume (< 0.5 km$^3$) end-member PDCs of high particle concentration that are frequently generated during gravitational collapse or explosive disruption of growing lava dome complexes (Cas and Wright, 1987; Druitt, 1998). They are characterized by complex, gravity-controlled, multi-phase flow dynamics where momentum transfer is controlled predominantly by particle collisions and frictional interactions (Dartevelle, 2004; Bursik et al., 2005; Dufek and Bergantz, 2007a, 2007b; Dufek, 2016).

In recent years, progress has been made towards a better understanding of the dynamics of transport and deposition of BAFs through laboratory and large scale experiments (e.g. Felix and Thomas, 2004; Cagnoli and Romano, 2010; Roche, 2012; Breard et al., 2016), remote sensing and numerical modelling (e.g. Denlinger and Iverson, 2001; Iverson and Denlinger, 2001; Pitman et al., 2003; Patra et al., 2005; Saucedo et al., 2005; Lube et al., 2007; Charbonnier and Gertisser, 2009; Kelfoun et al., 2009). However, direct field investigations remain one of the most powerful tools for obtaining crucial information regarding the depositional behaviour of BAFs, against which the applicability of observations from other sources (e.g. numerical simulations) can be assessed. Physical processes occurring during the late stages of particle transportation, just before the solid load comes to rest, are recorded within the deposit architecture and sedimentological features. Due to the fact that the lower part of the current (bed load region) is never accessible for direct observations due to ubiquitous cover by large volumes of overriding ash clouds (e.g. Cole et al., 2005), field studies of BAFs is therefore often limited to the associated deposits left after flow emplacement.
BAF deposits typically consist of massive poorly sorted mixtures of large blocks set within a lapilli/ash matrix (e.g., Cas and Wright, 1987; Druitt, 1998). Sedimentary structures, granulometry and clast componentry are used to determine lithofacies of the PDC deposits. These lithofacies can then be used to infer depositional regimes occurring in the flow-boundary zone (Branney and Kokelaar, 2002; Sulpizio and Dellino, 2008; Sulpizio et al., 2014). Previous studies of BAFs at Merapi, Indonesia (Charbonnier and Gertisser, 2008, 2011, 2012; Gertisser et al. 2012; Schwarzkopf et al. 2005; Lube et al. 2011; Komorowski et al. 2013), Soufriere Hills, Montserrat (Cole et al., 1998, 2002; Calder et al., 1999), Volcán de Colima, Mexico (Rodriguez-Elizarraras et al., 1991; Saucedo et al., 2002, 2004, 2005; Sulpizio et al., 2010; Sarocchi et al., 2011) and Unzen, Japan (Miyabuchi, 1999) used detailed field observations for stratigraphic reconstructions and sedimentological analysis of the deposits as ways to better understand the dynamics and physical processes involved within BAFs. Schwarzkopf et al. (2005) presented a conceptual model for an unsteady transport mechanism with kinetic granular waves, and deposition of flow waves through freezing from the base upward and from the rear moving forward due to the loss of kinetic energy and dispersive pressure. Sulpizio et al. (2007) generalized the early concept of pulse aggradation introduced by Schwarzkopf et al. (2005), and presented the stepwise aggradation model for multiple BAF pulses. This model is supported by Charbonnier and Gertisser (2011), who additionally considered simple freezing from the base to the surface for a single pulse. Particle shape and textural analyses performed by Sarrochi et al. (2011) led to distinction between frictional and collisional regimes during transport. Furthermore, grain size distribution with variations in fines was used to confirm the stepwise aggradation model of different BAF pulses during deposition.
Crucial findings regarding BAF dynamics highlight the non-uniformity of the current in both time and space, and show that the unsteadiness of flow conditions arises from complex interactions with the topography and obstacles (Schwarzkopf et al., 2005; Sulpizio and Dellino, 2008; Charbonnier and Gertisser, 2011, 2013; Komorowski et al., 2013; Solikhin et al., 2015). More specifically, the unpredictability of BAF dynamics involves the generation of overbank flows, which have been recently observed and described at Merapi for the 2006 and 2010 eruptions (Lube et al., 2011; Charbonnier et al., 2011, 2013; Komorowski et al., 2013; Cronin et al., 2013). Overbank flows refers to unconfined flows generated when the primary valley-confined BAFs escape their channel confines and spread laterally, sometimes over-spilling onto valley interfluves and partially filling adjacent channels. This complex behaviour of BAFs can pose a great threat to the surrounding population and infrastructure.

The July 2015 eruption at Volcán de Colima offers a new well-constrained case study for investigating BAFs associated with the generation of overbank flows, and allowed direct access to pristine flow architecture shortly following flow emplacement. Moreover, a particularity of this eruption was the emplacement of a large volume of pyroclastic material \(0.0077 \pm 0.001 \, \text{km}^3\), with a series of long runout flows over a short period of time (i.e. 18 hours), generated from long-sustained dome collapses. These long-lasting dome collapses are similar to those observed on June 14, 2006, at Merapi Volcano (Charbonnier and Gertisser, 2008, 2011). Emplacement of such large BAF volumes with the generation of multiple overbank flows over such a short period of time stresses the complexity of establishing appropriate hazard mitigation plans prior to the initiation of this type of event. This paper presents the results obtained from field studies of the July 2015 BAF deposits at Volcán de Colima, which aimed to investigate the internal processes associated with the rapid emplacement of such large-volume and long-runout BAFs. It first focuses on the
stratigraphy and sedimentology of the 2015 BAF deposits based on grain size measurements and clast componentry. Correlations between stratigraphic units and lithofacies variations provide detailed information about: (1) the distribution, volumes and sedimentological characteristics of the different units; (2) flow parameters (i.e., velocity and dynamic pressure) and mobility metrics as inferred from associated deposits; and (3) changes in the dynamics of the different flows during emplacement.

**Geological Background**

Volcán de Colima (19°31 N; 103°37 W; 3860 m a.s.l.) is located in the western portion of the Trans-Mexican Volcanic Belt (hereafter TMVB), toward the southern edge of the Colima Volcanic Complex (Figure 3.1). The TMVB is a 1200 km-long E-W active continental volcanic arc extending across Mexico between the Gulf of Mexico and the Pacific Ocean. Volcanism in the TMVB began in the middle to late Miocene and is the consequence of the subduction of the Rivera-Cocos plates beneath the North American plate at the Mid-American Trench (Atwater, 1989; Ponce et al., 1992; Ferrari et al., 1999, 2000).

Volcán de Colima is currently the most active volcano in the TMVB with at least 57 eruptions since 1519, six of which (1585, 1606, 1622, 1818, 1890 and 1913) were classified as VEI 4 eruptions (Luhr and Carmichael, 1990; De la Cruz-Reyna, 1993; Saucedo et al., 2005). Its past activity includes explosive-type activity up to sub-Plinian or Plinian eruption-style with the generation of PDCs from column collapse, as well as repeated major gravitational collapses of the edifice that produced large debris avalanches (Luhr and Prestegaard, 1988; Stoopes and Sheridan, 1992; Komorowski et al., 1997; Capra and Macias, 2002; Cortes et al., 2010; Saucedo et al., 2005, 2010; Roverato et al., 2011; Gavilanes-Ruiz et al., 2009).
Figure 3.1. Map of the three deposit facies identified during the 2015 block-and-ash flows at Volcán de Colima together with the locations of stratigraphic sections (labelled SA, IN and Mont) discussed in the text. The locations identified with the letters A to E refers to the overspill points identified on the topographic profile on Figure 3.3. In the blue boxes, $v$ and $P_D$ represent the calculated values of velocity and dynamic pressure respectively. The insert map locates Volcán de Colima in the Trans-Mexican Volcanic Belt (TMVB).
Figure 3.2. Pre- and post-eruption (i.e. April 2013 and January 2016 respectively) tri-stereo Pleiades satellite images showing two different zones along the Montegrande ravine affected by both the valley-confined and overbank deposits in July 2015. 

a) Valley-confined and interfluve overbank deposits at the overspill point B between the San Antonio and Montegrande ravine (see text for explanations).

b) Valley-confined deposits and wedge-shaped, coarse- and fine-grained interfluve overbank deposits further downstream.
In the last 25 years, activity at Volcán de Colima has been dominated by recurrent dome growth and often with subsequent gravitational collapse events generating short runout BAFs (1991-92, 1998-1999 and 2004, Saucedo et al., 2005) as well as small-volume PDCs generated from column-collapse events following Vulcanian explosions of VEI 2-3 (Macias et al., 2006; Varley et al., 2010).

**The July 2015 Eruption**

A preliminary report of the events of July 2015 at Volcán de Colima was published by Capra et al. (2016). On 10 July, 2015, at 12:00 (local time), an ash plume rose up to 7.7 km a.s.l. and drifted due to strong westward winds. At 20:17 that evening, according to seismic signals recorded from a broadband station from the Seismological Network at Colima University (RESCO) located in the Montegrande ravine (Reyes-Dávila et al., 2016), the paroxysmal activity began with a 52-minute-long sustained collapse of the summit lava dome and southern part of the crater rim. This sustained dome-collapse event generated the first series of BAF pulses that were channelized into the Montegrande ravine on the southern flank of the volcano, and stopped before the large volcaniclastic plain area located 9 km from the summit (Figure 3.1), as witnessed by photographs (by José Velazquez) taken at the end of the first day of the paroxysmal phase. Meanwhile, a 4 km ash-plume rose up and dispersed ash over the villages to the SW of the volcano. On 11 July at 13:50, a second 1h 47-minute-long sustained collapse of the dome and large portion of the southern crater rim resulted in the emplacement of a second series of BAFs along the Montegrande ravine up to the end of the volcaniclastic plain reaching a maximum runout of 9.3 km (calculated using a straight-line distance). This second series of BAF pulses was partially rechannelled into the adjacent San Antonio ravine at an overspill point that separates these two
ravines located 4.1 km from the summit and reach a total runout of 6.1 km (Figures 3.2a and 3.3). Overbank flow deposits from the July 2015 eruption at Volcán de Colima were also observed lying directly on top of the outer banks of the Montegrande ravine (Figures 3.2b and 3.3), and at greater lateral distances as rechanneled overbank flow deposits in narrow channels adjacent to the main ravine. After the emplacement of the second series of BAF pulses, a lava flow descended on the southern flank up to 3 km from the summit. During the next few months following the July 2015 events, the annual rainy season with many large rain storms including Hurricane Patricia (October 2015) were responsible for the generation of multiple streamflows and lahars that cut narrow gullies into the BAF deposits as they ran along the Montegrande and San Antonio ravines, remobilizing some of deposits and partially exposing the stratigraphic sequence of the July 2015 BAFs. Further investigations revealed that only deposits from the second series of BAF pulses were exposed by streamflow and lahar erosion in early 2016.

![Figure 3.3. Topographic profile along the Montegrande ravine showing the distribution of overbank flow deposits along the channel, and overspill locations A to E located on Figure 3.1.](image-url)
The July 2015 BAF events represent the first documented series of large-volume and long-runout BAFs generated from sustained dome collapses at Volcán de Colima. The volume and mobility of the July 2015 BAFs are comparable to that of the 1913 PDCs generated from column collapse during a Plinian eruption, and are much greater than past BAFs recorded at Volcán de Colima during the last century (Figure 3.4). These events emphasize the complex behaviour of such large–volume BAFs when flowing over a complex topography and within a narrow channel, as evidenced by multiple processes of flow decoupling and over spilling leading to surge detachment and the emplacement of voluminous, highly-mobile overbank flows.

**Figure 3.4.** Volumes and runout of historical pyroclastic density currents at Volcán de Colima. The runout and volume data for block-and-ash flows generated from dome- and fountain-collapse are compiled from Saucedo et al. (2005). Runout and volume data for the 1913 column-collapse PDC phase I, II and III are obtained from Saucedo et al. (2010). Data from the July 2015 block-and-ash flows (red diamonds) are included.
Methods

Field Observations and Terminologies

A field campaign in February 2016 allowed for an extensive investigation of the July 2015 BAF deposits along the deep gullies partly eroded by water runoff, some of which transformed into lahars. A total of 16 stratigraphic sections were studied in the Montegrande and the San Antonio ravines (Figure 3.1), and 25 samples from different flow units were collected for granulometry and component analysis. The term ‘flow unit’ refers to a single BAF deposited as one lobe. One flow unit refers to a deposit layer bound by erosive or gradational contact (e.g. ash layer, trains of clasts and/or breakout in grain size, sudden change in clast composition). These observations, coupled with high-resolution satellite images captured shortly following the eruption, provide valuable information regarding the nature and timing of these BAF deposits.

The July 2015 deposits are thus classified into three main facies of deposition: (1) valley-confined deposits in the Montegrande ravine, (2) overbank flow deposits, classified into four sub-facies categories (i) coarse-grained interfluve overbank deposits, (ii) fine-grained interfluve overbank deposits, (iii) wedge-shaped overbank deposits, (iv) rechannelled overbank deposits, and (3) ash-cloud surge deposits along the valley margins and in proximity to overbank deposits (Figures 3.1 and 3.2). Stratigraphic units from the valley-confined deposits are named “VC” and the numbers are correlated with the order of deposition. Because the BAF deposits were only partially exposed at the time of investigation, the lowermost visible flow unit was assigned the first number: “VC1”. The flow units of overbank deposits are named “OB” with a similar numbering system. The distribution of the collected samples along the flow paths within the different deposit facies was determined by accessibility and availability of erosive channels. Steep slopes and the current state of volcanic activity prevented an approach to the immediate proximal
area of the BAF deposits (i.e. within the first 2.5 km from the summit). Based on these criteria for sampling sites, 16 samples were collected from nine stratigraphic sections located in the major gullies of the Montegrande ravine, all corresponding to valley-confined deposits. One section (three samples) located in the vicinity of overspill point B between the San Antonio and Montegrande ravines exposes the interfluve overbank deposit facies. Three sections (three samples) in the San Antonio ravine and two sections (two samples) in an adjacent channel to the Montegrande ravine expose the sub-facies of rechannelled overbank deposits.

Following the lithofacies scheme introduced by Sulpizio et al. (2007) based on sedimentary structures, componentry, grain size distribution, coarse-tail grading and sorting of the flow units, six main lithofacies are distinguished and describe the architecture of the deposits: mBLAms (massive, Block, Lapilli and Ash with moderate clast segregation), mBLAws (massive, Block, Lapilli and Ash with weak clast segregation), mLA (massive, Lapilli and Ash), mA (massive Ash), sLA (stratified Lapilli and Ash) and sA (stratified Ash) (Table 3.1).

Grain Size Analysis

Grain size analysis of the July 2015 BAF deposits was carried out using a multiple-step method. Field sampling consisted in defining a ~ 0.5 x 1 m rectangular area within a specific unit where the outcrop was described as representative of the entire unit at a given section site. Large blocks within the sampling area were collected into the sample box when protruding more than 50%. These were all measured and weighed in the field with a field scale. Then, manual dry sieving was performed on clast sizes between -6 phi (64 mm) and -2 phi (4 mm) in full-size phi intervals and each fraction weighed in the field. The remaining < -2 phi (i.e. the matrix) fraction was weighted but only a quarter of the total mass was collected and carried back to the laboratory for
wet-sieving analysis at 1 phi intervals, up to 5 phi (32 μm). Based on the work of Schumacher et al. (1990), the cone and quarter method was used in the field to ensure that each sample fraction collected was representative of the whole sample matrix. The riffle splitting method (using closed bins) was used in the laboratory in order to obtain similar distributions of the matrix fraction during splitting. Fractions smaller than 5 phi were not integrated into the grain size distribution to facilitate comparison between samples. However, estimates based on the loss of fines during wet-sieving showed that the fraction smaller than 5 phi never exceeded 7 wt. % of the total sample mass. The final mass percentages of each sample were plotted using the software GRADISTAT© (Blott and Pye, 2001) to calculate grain size parameters. The DECOLOG software (www.DECOLOG.org; Borselli and Sarocchi, 2009) was also used to apply deconvolution to the grain size distributions in order to facilitate the comparison between the different peak components (i.e. sub-populations of particle sizes) of the polymodal distributions. The deconvolution process consists in dividing a polymodal distribution into a mixture of lognormal distributions based on three peak components that are the most statistically representative of the different particle size sub-populations. The results of grain size analyses for the 25 samples collected is presented in Table 3.2.

Componentry

A component analysis of each sample matrix (grain size < -2 phi) was performed under a binocular microscope by point counting of particles for fractions between -2 and 3 phi. For each size fraction, a total count of about 300 grains were handpicked and assigned to one of three component categories: (i) angular dark to light grey-coloured, dense to slightly vesicular unaltered volcanic fragments, interpreted as freshly extruded magma from a fresh dome growth or from a superficial magma batch in the conduit (Luhr, 2002; Atlas et al., 2016); (ii) lithic fragments, which
are angular, dense to vesicular volcanic clasts identified as older dome or lava fragments from the 1913 or earlier eruptions by their hydrothermally altered and/or oxidized surface/interior, and (iii) free crystals derived from larger volcanic fragments of type (i) or (ii). For each sample, the proportion of each component is displayed in Table 3.2.

**Volume Calculations and Geomorphic Analysis**

The extents of the deposit facies were mapped with the ArcGIS™ software (ESRI, Redlands, CA, USA) based upon field investigations and using two stereo pairs of 50 cm high resolution PLEIADES-1A satellite images from April 2013 and January 2016 respectively (source: Airbus Defence and Space). Volume calculations and geomorphic analysis were performed using two 1-m spatial resolution digital elevation models (DEMs) obtained by photogrammetry of the two stereo pairs.

The total volume of deposits was obtained from DEM difference between the pre- and post-eruption DEMs using the ArcGIS™ software. Areas of negative volume difference, due to the presence of extensive tree-covered terraces on the pre-eruptive DEM that were stripped by the 2015 BAFs, were corrected by (1) creating polygons around each tree-covered terrace and (2) assigning a constant deposit thickness derived from field measurements to calculate individual deposit volumes that were then used to adjust the total deposit volume (Table 3.3).

Analysis of channel geometry (slope gradient and channel cross-sectional area) of the Montegrande and San Antonio ravines was performed using the ArcGIS™ software ‘spatial analyst’ tool on the 1-m spatial resolution DEM from April 2013. About 120 cross-sectional profiles were constructed along the affected area on the southern flank of the volcano to capture variations in channel geometry with distance from the summit.
Table 3.1. Classification of flow unit lithofacies observed inside the July 2015 Block-and-Ash flow deposits at Volcán de Colima. This classification is based on sedimentological features and grain size characteristics of individual flow units. Each lithofacies is interpreted in terms of associated flow transport and deposition mechanisms.

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Description</th>
<th>Interpretation</th>
<th>Unit and Section</th>
</tr>
</thead>
<tbody>
<tr>
<td>Massive Blocks, Lapilli and Ash with moderate clast segregation (mBLAms)</td>
<td>Thickness &gt; 2 m. Very poorly sorted with blocks set in a lapilli-ash matrix. Maximum range of grain size distribution. Moderate coarse-tail inverse grading. Clast trains and long axes of elongated clasts oriented parallel to the flow direction</td>
<td>Deposits from a BAF pulse in which granular temperature due to grain interaction and fragmentation processes is moderate, although sufficient enough to produce moderate clast segregation and orientation (kinematic squeezing and kinetic sieving of large particles)</td>
<td>Unit VC1 and VC2 at sections Mont1 and Mont2; Unit VC1 at section Mont 3</td>
</tr>
<tr>
<td>Massive Block, Lapilli and Ash or Massive Lapilli and Ash with weak clast segregation (mBLAws)</td>
<td>Thickness &gt; 1 m. Very poorly sorted with blocks set in a lapilli-ash matrix. Massive with weak or absent coarse-tail grading. Rare or absent clast trains</td>
<td>Deposits from a BAF pulse in which granular temperature and grain dispersive pressure dropped resulting in weak or absent clast segregation. Sudden drop of energy observed with very coarse deposition facies</td>
<td>Unit Mont2; Unit VC2 and VC3 at sections Mont3, Mont5, Mont6, Mont7 and Mont8; Unit VC1, VC2 and VC3 at section Mont4</td>
</tr>
<tr>
<td>Massive Lapilli and Ash or Massive Ash (mA, mLA)</td>
<td>Thickness &lt; 20 cm. Moderately to well-sorted lapilli-ash or ash layer. No stratification and absence of coarse-tail grading</td>
<td>Vertical settling of the overriding ash-cloud surge accompanying the basal BAF body. Can also result from lofting and entrapment of fine particles from passage at break in slopes and sharp bends of the channel</td>
<td>Unit VC2 at section Mont9. Unit OB1 at section IN1 and at surge units. Unit OB3 at section SA3</td>
</tr>
<tr>
<td>Stratified to Cross-stratified Lapilli and Ash or Stratified Ash (sLA, sA, csLA)</td>
<td>Thickness &lt; 30 cm. Stratified and cross-stratified lapilli and ash layers of millimeter size (&gt; 15 layers)</td>
<td>Deposition from the lateral motion of the overriding ash-cloud surge and/or fine-grained overbank flows</td>
<td>Section IN2 Section Mont2</td>
</tr>
</tbody>
</table>
Table 3.2. Grain size parameters and component data of the studied samples. $Md_\phi$ = Median size particle (phi); $\sigma_\phi$ = Sorting. The first, second and third components correspond to the three main peaks in grain sizes, obtained from deconvolution of the polymodal grain size distributions. See text for details about matrix components.

<table>
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<tr>
<th>Distance from summit</th>
<th>Section</th>
<th>Unit</th>
<th>Blocks (wt. %)</th>
<th>Lapilli (wt. %)</th>
<th>Ash (wt. %)</th>
<th>$Md_\phi$</th>
<th>$\sigma_\phi$</th>
<th>1st mode</th>
<th>2nd mode</th>
<th>3rd mode</th>
<th>Unaltered fragments</th>
<th>Accidental Lithics</th>
<th>Loose crystals</th>
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<td>VC2</td>
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<td>32.7</td>
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<td>73.1</td>
<td>18.0</td>
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<td>20.1</td>
<td>32.2</td>
<td>47.7</td>
<td>-1.3</td>
<td>3.8</td>
<td>-5.4</td>
<td>0.7</td>
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<td>-2.9</td>
<td>3.6</td>
<td>-6.5</td>
<td>-4.9</td>
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<td>-2.7</td>
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<td>4.0 km</td>
<td>IN1</td>
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<td></td>
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<td>OB2</td>
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<td>2.9</td>
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<td>SA1</td>
<td>OB3</td>
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<td>47.2</td>
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<td>-0.8</td>
<td>3.0</td>
<td>-4.0</td>
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<td>0.0</td>
<td>71.5</td>
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<td>5.9 km</td>
<td>SA2</td>
<td>OB3</td>
<td>4.3</td>
<td>29.0</td>
<td>66.6</td>
<td>0.1</td>
<td>3.0</td>
<td>0.7</td>
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<td>3.7</td>
<td>63.8</td>
<td>25.8</td>
<td>10.4</td>
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<td>OB2</td>
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<td>43.3</td>
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<td>67.5</td>
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<td>Mont-E2</td>
<td>OB2</td>
<td>13.5</td>
<td>28.8</td>
<td>52.7</td>
<td>-0.3</td>
<td>3.3</td>
<td>-6.4</td>
<td>1.3</td>
<td>-2.9</td>
<td>65.9</td>
<td>21.7</td>
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</table>
Results

Valley-Confined Deposits

Valley-confined or channelled deposits of the 11 July, 2015, BAFs were observed within the narrow erosional channels in the Montegrande ravine (Figure 3.1). A total of nine stratigraphic sections were described in the valley-confined deposit facies: section Mont1 in the proximal area (up to 4 km from the summit), sections Mont2 and Mont3 in the medial area (between 4 and 6.5 km), and sections Mont4 to Mont9 in the distal area (up to 9.2 km from the summit) (Figure 3.1, Table 3.2).

Sections Mont1, Mont3, Mont4, Mont5, Mont7 and Mont8 (Figure 3.5) expose the flow units in sections parallel to the flow direction. In contrast, sections Mont2, Mont6 and Mont9 (Figure 3.5) exhibit the flow units perpendicularly to the flow direction. Furthermore, section Mont2 is located in a super-elevation deposit area, against the valley-wall on the outside of a sharp bend, while section Mont9 is located in the volcaniclastic plain area at the bottom of the Montegrande ravine where the channel significantly widens and becomes shallower and less confined (Figure 3.1). Three flow units were identified in the valley-confined deposit facies for the events of 11 July (i.e., VC1, VC2 and VC3).

Table 3.3. Area, volume, geometric (A/V^{2/3}) and plan-shape (A/L^2) parameters of the July 2015 Block-and-Ash flow deposits. VC = valley-confined deposits; OB = overbank deposits.

<table>
<thead>
<tr>
<th>Deposit facies</th>
<th>Area (km²)</th>
<th>%</th>
<th>Volume (km^3)</th>
<th>%</th>
<th>A/V^{2/3}</th>
<th>A/L^2</th>
</tr>
</thead>
<tbody>
<tr>
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<td>0.0058</td>
<td>75.3</td>
<td>29.0</td>
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<td>Overbank</td>
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<td>0.0019</td>
<td>24.2</td>
<td>38.3</td>
<td>0.007</td>
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<td>Surge</td>
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<td>37.3</td>
<td>0.00004</td>
<td>0.55</td>
<td>742.6</td>
<td>0.011</td>
</tr>
<tr>
<td>VC + OB</td>
<td>1.52</td>
<td>62.7</td>
<td>0.0077</td>
<td>99.4</td>
<td>39.0</td>
<td>0.018</td>
</tr>
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<td>Total</td>
<td>2.42</td>
<td>100.0</td>
<td>0.0077</td>
<td>100.0</td>
<td>62.0</td>
<td>0.028</td>
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</tbody>
</table>
Figure 3.5. Stratigraphic sections Mont1 to Mont8 from the proximal to distal areas showing the lithofacies variations of the valley-confined BAF units with distance.

The lowermost exposed flow unit, unit VC1, is a massive, light-grey unit with angular to sub-angular blocks embedded in a lapilli-ash matrix, and observed in sections Mont1, Mont2, Mont3, Mont4 and Mont6 (Table 3.2, Figures 3.5 and 3.6). Its thickness varies from 2.5 m in the proximal area to ~ 2 m in the distal area, although its base was not exposed except in section Mont2. In the proximal area, unit VC1 exhibits a subtle inverse coarse-tail grading with well-developed clast imbrication with long axes of tabular and elongated clasts oriented parallel to the flow direction (lithofacies mBLAms). In the medial area, the unit shows a weak inverse coarse-tail grading (lithofacies mBLAws) in section Mont2 that disappears in section Mont3.

Grain size analyses of unit VC1 shows at first a shift and progressive overlapping of the first (main) and second peak components of the deconvoluted grain size distributions towards the coarser grain sizes from section Mont1 to Mont2, followed by a shift of the main peak component of the distribution to the finer grain sizes (1.5 phi) up to section Mont3 (Table 3.2 and Figure 3.7).

Component analyses show a decrease in lithic content coupled with an increase in unaltered volcanic fragment content (Table 3.2 and Figure 3.6). In section Mont4 located 6.9 km from the
summit, however, unit VC1 becomes very poorly sorted, coarser (with a narrow second peak component focused at -7.7 phi), richer in lithic content (Table 3.2 and Figures 3.6 and 3.7), and do not exhibit any coarse-tail grading (lithofacies mBLAws, Figure 3.5).

The middle flow unit, unit VC2, is observed and interpreted as the flow unit that reached the maximum flow runout (i.e. up to the end of the volcaniclastic plain area, Figure 3.1). Unit VC2 is a massive, very poorly sorted ($\sigma$ ranges from 2.8 to 4.1), medium-grey unit with angular to sub-angular blocks set in a lapilli-ash matrix. Its thickness gradually decreases from 2.1 m in the proximal section (Mont1) to 1.5 m in the medial area (Mont2 and Mont3), to 0.9 m in the distal area (Mont9), although it completely filled the 6 m-deep and narrow pre-existing erosional channels that cut within the distal volcaniclastic plain.

In the proximal section (Mont1), unit VC2 shows a normal coarse-tail grading (lithofacies mBLAms) and the presence of a train of coarse-clasts, roughly at mid-thickness of the unit, where elongated large clasts are oriented approximately parallel to the direction of the flow. The presence of degassing pipes was also noticed at the contact with the basal unit.

In the medial and distal areas, from sections Mont4 and Mont7, unit VC2 exhibits subtle inverse coarse-tail grading (lithofacies mBLAws). In the medial area, unit VC2 shows evidence of clast entrainment at the bottom with vertical trains of larger clasts protruding upward from the underlying unit VC1 (Figure 3.5). In the distal sections (Mont5 and Mont7), clast imbrication is found in the basal part of the flow unit, and in section Mont8, the upper part of unit VC2 contains metre-sized blocks which exhibit abundant friction marks.

Grain size analyses and componentry of unit VC2 show at first a shift of the peak components towards finer grain sizes from proximal to medial areas together with a decrease in unaltered volcanic fragments coupled with an increase in lithic content (Table 3.2 and Figures 3.6
and 3.7). Then, in the distal area, the grain size distribution of unit VC2 fluctuates to become first coarser at section Mont4 (first and second peak component skewed towards coarser particle sizes), finer further downstream and then coarser again in section Mont7, after which the peak components shift towards finer grain sizes again with distance (Table 3.2 and Figure 3.7). A similar trend is obtained from the componentry, with decreasing content in unaltered fragments coupled with an increase in lithic fragments from section Mont4 to section Mont6 and then again from section Mont7 to section Mont9 (Table 3.2 and Figure 3.6). The loose crystal content also increases in the distal sections.

The uppermost unit, unit VC3, is massive, dark-grey, very poorly sorted (σØ ranges from 3.2 to 3.9) unit with sub-angular blocks set in a similar lapilli-ash matrix to that of the previously described flow units. Unit VC3 was completely reworked by lahars and streamflows in the proximal area, limiting our observations to the medial and distal sections. The front of unit VC3 was observed downstream of section Mont7. Its thickness is constant at 0.7 m in the medial sections, thins out to 0.5 m in section Mont7 (not sampled), before pinching out. Sampling of this unit was only possible at sections Mont2, Mont3 and Mont4.

Grain size analyses show significant shift of the main peak component from fine to coarse grain sizes with distance, as well as a decrease in loose crystal content (Table 3.2 and Figures 3.6 and 3.7). Visual observations of the distal part of unit VC3 after section Mont4 indicate an increase in fines and decrease in block content with distance.

In the medial sections, abundant charcoal fragments and recycled pumices from past eruption products were found inside this unit. In section Mont2, a 3-cm-thick surge layer of fine to coarse-ash (lithofacies mA) is interbedded between unit VC3 and VC2.
Figure 3.6. Summary of the grain size and componentry in the valley-confined BAF units. The results are presented for all three units (vertical stratigraphy from VC1 at the bottom to VC3 at the top) and at different stratigraphic sections with increasing distance from summit.
Figure 3.7. Summary of the grain size distribution data for the three valley-confined BAF units and their variation with distance. The histogram of grain size distribution are overlaid with the results from deconvolution of the polymodal distributions. The bold line, the thin line and the dashed line represent the 1st, 2nd and 3rd principal peak components respectively.
Overbank Deposits

A total of six stratigraphic sections were located and described in the overbank deposit facies: section IN1, located 4 km from the summit, exposes coarse-grained and poorly sorted interfluve overbank flow units at the overspill point between the Montegrande and the San Antonio ravines (location B on Figures 3.1 and 3.2a). Rechannelled overbank flow units are exposed in the San Antonio ravine at sections SA1 and SA2, which are respectively located at 4.9 and 5.9 km from the summit, and in a narrow erosional channel located on the eastern side of the Montegrande ravine at sections Mont-E1 and Mont-E2, about 8 km from the vent (Figures 3.1 and 3.8). The fine-grained overbank deposit facies was observed in the San Antonio ravine in section SA3 and on the interfluve of the Montegrande ravine in section IN2 (Figures 3.1 and 3.8). Finally, the wedge-shaped overbank deposit facies was found and described at section WS1, located 5.8 km from the vent (Figures 3.1, 3.2b, 3.3 and 3.8). Similarly to the valley-confined deposits, three flow units were identified in the overbank deposit facies (OB1, OB2 and OB3).

The San Antonio ravine is 200 to 250-m-deep and the deeper eroded channel with steep walls complicated field investigations in this area. Three units were identified along the narrow and deep erosional channels, but only the flow units OB2 and OB3 were accessible for sampling. The sections inside the small side branch to the east of the Montegrande ravine (i.e. sections Mont-E1 and Mont-E2, Figure 3.1) only expose one overbank unit, which was interpreted as unit OB2.

The lowermost overbank flow unit, OB1, is a massive, very poorly-sorted, light-grey unit with a thickness ranging from 15 cm when exposed as the interfluve sub-facies, to 2.1 m in the rechannelled deposit sub-facies, and up to 3.3 m in the wedge-shaped deposit sub-facies. Inverse coarse-tail grading (lithofacies mBLAms) is found in the wedge-shaped and rechannelled facies. In section IN1 (interfluve sub-facies), unit OB1 is capped with a 6-cm-thick fine-grained unit
(Figure 3.8b) interpreted as a surge layer, and its grain size distribution ranges from ash to coarse-lapilli (lithofacies mLA). Componentry reveals that unit OB1 contains higher proportions of unaltered volcanic fragments in contrast to the overlaying units OB2 and OB3.

The middle overbank flow unit, OB2, is a massive, very poorly-sorted, medium-grey unit with a thickness ranging from 20 cm when exposed as the interfluve sub-facies, to 1.1 - 2.5 m in the rechannelled deposit sub-facies, and up to 4.7 m in the wedge-shaped deposit sub-facies. Similarly to unit OB1, unit OB2 displays inverse coarse-tail grading (mBLAms), well-pronounced inside all three overbank deposit sub-facies, and is capped with a 5-cm-thick surge layer in section IN1 (Figure 3.8b). Moreover, unit OB2 in section WS1 contains coarse-clast trains well visible in the upper half of the unit (Figure 3.8a). This unit also contains the highest proportions in accidental lithics and greater proportions in loose crystal content than flow units OB1 and OB3 (Table 3.2). Grain size analyses of the rechannelled deposit samples show a shift of the main peak component of the distribution towards finer sizes with distance from the summit (Table 3.2).

The uppermost overbank flow unit, unit OB3, is a massive, very poorly-sorted, darker-grey unit with a thickness ranging from 17 cm when exposed as the interfluve deposit sub-facies, to a minimum thickness of 1 m in the rechannelled deposit sub-facies and up to 2 m in the wedge-shaped deposit sub-facies. Unit OB3 is also characterized by an inverse coarse-tail grading (mBLAms) which is absent in the interfluve deposit sub-facies (lithofacies mBLws and mLA instead).

At several locations along the Montegrande ravine, fine-grained overbank deposits are observed on top of the outer valley banks and characterized by a 20 to 30-cm-thick unit of stratified coarse-ash to lapilli matrix (lithofacies sLA), sometimes containing small blocks, covered with a 3.5 to 9-cm-thick layer of very fine stratified ash (lithofacies sA). Section IN2 exposes these fine-
grained overbank deposits perpendicularly to the flow direction and shows cross-stratifications of lapilli and coarse-ash layers (lithofacies csLA) with imbricated angular blocks that display evidence of rotation following the flow direction (Figure 3.8d).

![Diagram of stratigraphic sections](image)

**Figure 3.8.** Stratigraphic sections of the overbank deposit sub-facies: (a) wedge-shaped deposits, (b) coarse-grained interfluve deposits, (c) rechannelled deposits and (d) fine-grained deposits. Yellow arrows indicate flow direction.

### Ash-Cloud Surge Deposits

Surge deposits were observed in the forested areas on the edge of the ravine and often accompanying fine-grained overbank deposits onto channel interfluvess. They are characterized by massive to stratified ash layers (lithofacies mA and sA) (Table 3.1) of thicknesses of a few centimetres. In the distal area, singed trees up to 30 m-high suggest a vertical motion of the ash-cloud surge due to the elutriation of fines at breaks in slopes. Stratified lithofacies of the deposits
indicate deposition from the lateral motion of the ash-cloud, while the massive lithofacies is commonly associated with gentle settling of the overriding ash cloud on top of the BAF system. The outlines of the surge deposits reported in Figure 3.1 show that the lateral spreading of the ash-cloud surge was fairly limited along the entire affected area as it did not extend more than 150 m away from the valley-confined and overbank deposits.

**Interpretations and Discussion**

**Correlations between Units and Granulometry**

Although field observations are limited to the partial exposure of the BAF deposit sequence, correlations between the exposed units, which are interpreted as deposits from the latest BAF pulses that were generated towards the end of the paroxysmal phase of the eruption on 11 July, are presented in Figure 3.9. The preliminary report by Capra et al. (2016) infers that a total of six pulses descended along the main channel of the Montegrande ravine and, based on real-time data acquisition from a camcorder and a geophone and witness reports, the first series of BAF pulses on 10 July were all confined inside the Montegrande ravine and stopped just before the entrance to the volcaniclastic plain area, around 8.3 km from the summit (Figure 3.1).

Three flow units of valley-confined deposits can be identified in the Montegrande ravine and correlated to the emplacement of the second series of BAF pulses on 11 July. Similarly, three flow units of overbank deposits were also observed in the three sub-facies recognized from field investigations, which are also inferred to have been generated from the second series of BAF pulses. With a quite consistent flow unit thickness and lithofacies, unit VC2 represents the BAF pulse that was transported the farthest, reaching the maximum runout of 9.3 km (Figure 3.9). Additionally, unit VC2 exhibits the highest content in accidental lithics compared to unit VC1 and
VC3, which could correspond to strong erosion of the collapsing portion of the southern crater rim during the emplacement of this pulse. Similarly, the overbank flow unit OB2 also exhibits a higher content in accidental lithics compared to units OB1 and OB3. These evidence lead to the fact that the three units of valley-confined deposits found during field investigations are correlated to the three units of overbank deposits described above, and that all these units were deposited during the same event (i.e. the second sustained dome collapse phase on 11 July).

The first sustained collapse on 10 July generated a first series of BAF pulses that were solely confined into the Montegrande ravine and resulted in the progressive infilling of the main channel. Therefore, only the second series of BAF pulses on 11 July were able to overspill the valley margins and generate overbank flows. Grain size analyses of the valley-confined and overbank deposit samples, displayed in the Walker diagram (Figure 3.10a), show that units of overbank deposits are better sorted and finer-grained than their respective stratigraphically correlated valley-confined units (i.e., OB1 and VC1). Although there were only two flow units of rechannelled overbank deposits found inside the narrow erosional channel to the east of the Montegrande ravine (sections Mont-E1 and Mont-E2), these flow units are interpreted as OB2 and OB3 because of their similar componentry as those found in sections IN1, SA1 and SA2 (Table 3.2). The large infilling of the channel by the BAF pulse that deposited unit VC2 may have also triggered the formation of overbank flows in the distal area and enabled the late stage emplacement of the uppermost overbank flow unit (OB3) in the adjacent channel afterwards.

Two sections in the distal area (i.e. sections Mont4 and Mont7, Figures 3.1 and 3.5) show units with coarse-rich grain size distributions and high block content (Table 3.2 and Figure 3.7). These two sections, located 6.9 km and 7.8 km from the summit, respectively, are interpreted as locations of topography-induced sedimentation of coarse-rich material from the passage of
successive BAF pulses. The rough, uneven coarse-rich surface of these BAF deposits is clearly visible on high-resolution satellite images taken shortly after the eruption. Interestingly, an increase in median particle size is observed in all three valley-confined flow units up to the first location at 6.9 km from the summit, then in flow units VC2 and VC3 up to the second location at 7.8 km from the summit (Table 3.2, Figures 3.1, 3.7 and 3.10b). A sudden decrease in median grain size inside the two upper valley-confined units is also observed downslope of the first location in section Mont5 and downslope of the second location in section Mont8. Based on these grain size variations, we can interpret the first location of topography-induced sedimentation of coarse-rich material (i.e. section Mont4) as the lobate front of flow unit VC1 (see Figure 3.9 for stratigraphic correlation).

Facies Variations and Interpretations of Flow Dynamics

In the proximal area, units of valley-confined deposits show lithofacies of moderate clast segregation (mBLAms) suggesting internal flow dynamics governed by high grain-dispersive pressure that induced segregation processes such as kinetic sieving (Middleton, 1970; Savage & Lun, 1988) and kinematic squeezing (Le Roux, 2003) of the largest clasts. In the medial and distal areas, clast segregation becomes more subtle and discrete (lithofacies mBLAws), which implies a progressive decrease in grain-dispersive pressure due to fewer grain interactions and therefore a decrease in kinetic energy produced inside the flow during transport. The overall decrease in accidental lithic content in flow unit VC1 can be interpreted as a decrease in flow competence during transport (Charbonnier and Gertisser, 2011), although a sudden increase in accidental lithics was observed at section Mont4 (Figure 3.6), where sudden changes in channel morphology induced rapid sedimentation of coarse-rich material.
Figure 3.9. Stratigraphic correlations in the valley-confined BAF deposits. See text for details.
Sedimentary features like vertical trains of larger clasts protruding upward, as observed between unit VC2 and underlying unit VC1 in the medial sections, provide evidence for clast entrainment. The rough surface of the underlying VC units promoted the entrainment of large particles into the overlying unit (Roche et al., 2015; Fauria et al., 2016; Pollock et al., 2016). In the distal area, the increased deposit thickness of BAF unit VC2 in the volcaniclastic plain, filling up to 6-m deep gullies, suggests a sudden increase in the sedimentation rate in this area due to the loss of channel confinement and flow momentum.

Deconvolution of the grain size distributions from all three valley-confined units shows a general increase in the dispersion of fine grain size populations with distance (while the total weight percent of ash do not show any trend, see Table 3.2), together with lower proportions of coarse grain size populations (Figure 3.7), which could be due to several factors depending on the location along the Montegrande ravine: (1) the progressive break up of larger clasts into smaller fragments with transport; (2) coarse-clast segregation related to overbank processes thus leaving locally finer-grained valley-confined deposits (see following conceptual model); (3) earlier deposition of the coarse-rich frontal lobe of a single pulse, leaving a finer-grained main body and tail to be transported further downstream. Additionally, the distal VC units contain higher proportions of free crystals (Figure 3.6), also suggesting processes of particle interactions and increasing fragmentation processes with distance. Similar conclusions can be drawn for the overbank flow units (Table 2), which further reinforces the idea that overbank deposits are directly derived from their parent valley-confined flow units (i.e., Charbonnier and Gertisser, 2011; Gertisser et al., 2012).

Changes in flow dynamics with distance from the summit interpreted from variations in deposit lithofacies are further supported by the changes in flow velocities with distance as well.
Flow velocity in the proximal area can be calculated from the equation $v = \sqrt{2gh}$ with $h$ the flow runup at the overspill point between the San Antonio and Montegrande ravines (Figure 3.1) and $g$ the gravity constant (Sheridan et al., 2005). The runup of the BAF pulse correlated to unit VC2 is estimated at ~7 m, which gives a minimum flow velocity of ~11.7 m/s. In the medial and distal areas, the sinuosity of the Montegrande ravine led to the presence of pristine super-elevation deposits from unit VC2 for which the equation $v = \sqrt{(ghr)/b}$, where $h$ is the difference between the channelized deposits and the super-elevation deposit height, $r$ is the radius of curvature and $b$ is the channel width, can be used to estimate minimum flow velocities within the channel (Zanchetta et al., 2004; Sheridan et al., 2005). Using this method along a ~500 m highly sinuous distal segment of the valley, the minimum flow velocities calculated for the BAF pulse correlated to unit VC2 decrease from 5.3 to 4.7 m/s (Figure 3.1).

This decrease in flow velocity with distance is consistent with the decrease in grain dispersive pressure and flow kinetic energy inferred from longitudinal lithofacies variations of unit VC2. Velocity measurements can also be used to estimate flow dynamic pressures, which is a measure of the lateral stress exerted by the current (Valentine, 1998a, b; Doronzo, 2013) using the following equation: $P_{dyn} = 0.5\rho_f v^2$, where $\rho_f$ is the flow density (estimated at 1580 kg/m$^3$ following the method proposed by Zanchetta et al., 2004) and $v$ is the velocity (Jenkins et al., 2013; Sulpizio et al., 2014). Using the velocity measurements in the proximal, medial and distal areas, the dynamic pressures of the BAF pulse correlated with unit VC2 vary from 108 kPa in the proximal area, to 23.0 kPa in the medial area and 17.4 kPa in the distal area.

The higher content in free crystals together with the finer median particle size measured inside samples from overbank deposits are interpreted here as evidence of lofting and entrapment of fines when the valley-confined flows encountered areas of channel enlargement, sharp bends
and/or break in slopes. In this model, the upper region of the valley-confined flow became progressively enriched in fines and free crystals downstream due to strong particle density segregation processes occurring when flowing over these areas of topography-induced sedimentation. This can be proposed as one of the mechanisms responsible for the generation of overbank flows, more specifically for explaining the presence of the fine-grained overbank deposit facies at these locations. Componentry reveals that interfluve and rechannelled overbank deposits have an overall higher accidental lithic content than valley-confined deposits (Table 2), which could be due to the erosive behaviour of overbank flows spilling and spreading over the valley interfluves and into adjacent channels. The ten-fold increase in flow unit thickness observed between the interfluve overbank deposits (i.e. section IN1) and the rechannelled overbank deposits located further downstream inside the San Antonio ravine brings further evidence for the transitional behaviour of overbank flows (from erosion- to deposition-dominated) as they spilled over the valley margins and became rechannelled into adjacent tributaries. In contrast, the wedge-shaped overbank deposits are characterized by greater flow unit thicknesses, a wider range of grain sizes and often a non-erosive basal contact with the presence of undisturbed underlying soil layers, which suggest that they were generated from the passive overspill and rapid en masse deposition of the main valley-confined flows onto the adjacent valley banks. In contrast to the high-energy overbank flows generated from an overspill jump over a valley wall, passive overspill behaviour occurs when the vertical distance between the top of the main valley-confined flow body and the valley banks is small, allowing the marginal flow body to gently overspill alongside the channel without requiring strong inherited momentum.

In the proximal and medial areas, the lithofacies variations described above for unit VC2 imply that flow dynamics inside the main BAF body during transport were mainly driven by
particle-particle interactions which generated a sufficient granular temperature to induce kinetic sieving and kinematic squeezing of the coarse clasts. In the more distal areas, momentum transfer inside the main BAF body is maintained through an increase in particle collisions, probably caused by the presence of a zone of channel constriction and flow confinement (from 6.9 to 7.8 km, see Figure 3.1) due to the presence of large lahar terraces on one or both sides of the distal part of the valley (Figure 3.10b). Eventually, frictional forces overcome the driving forces due to increase in grain interlocking at the base of the current, resulting in a flow transition from erosional to depositional regime. Furthermore, the presence of zones of topography-induced sedimentation of coarse-rich material observed in the distal sections of the valley-confined BAF deposits suggests a sudden drop of energy of the basal flow region due to increasing resistance forces such as basal shear stress, and therefore a freezing of the basal underflow. The lack of ash-layers in between individual VC units can be explained by the mechanical erosion exerted by the shear of the basal portion of subsequent pulses as they travelled over previously deposited units.

**Effects of the Geomorphology**

Previous field-based studies of concentrated PDCs including BAFs (i.e. Sulpizio and Dellino, 2008; Charbonnier and Gertisser, 2011; Komorowski et al., 2013; Sulpizio et al., 2014) have shown that topography can strongly influence their dynamics, such as changes of fluid turbulence within the current, variations in particle concentration in the flow-boundary zone, or the presence of topographic barriers to partially block the current causing stripping and decoupling of the upper flow region (Sulpizio et al., 2014 and references therein). Although some significant advances have been made through both large- and small-scale experiments, interactions between the topography and the dynamics of BAFs remains complex and not fully understood.
Figure 3.10. Walker diagram and changes in grain size with channel morphology. a) Walker diagram (sorting versus median grain size (Md)) of the three valley-confined and associated overbank units. b) Median grain size of the valley-confined unit VC2 (black) and pre-2015 channel width (blue) with distance from the vent (C = channel constriction; W = channel widening). The red arrows indicate the location of the two coarse-rich lobate fronts found inside the valley-confined deposits. Two photographs taken in October 2011 depict the difference in channel width (credits: L. Capra)
In the line of research of similar studies at Merapi (Lube et al., 2011; Charbonnier and Gertisser, 2011, 2013; Komorowski et al., 2013; Cronin et al., 2013; Solikhin et al., 2015), Unzen (Miyabuchi, 1999) and Montserrat (Calder et al., 1999), the July 2015 BAF deposits at Volcán de Colima present evidence of these complex interactions with the rugged pre-eruptive terrain.

Although the source mechanism and eruption characteristics of the 2006 BAFs at Merapi (Charbonnier and Gertisser, 2008, 2011) and 2015 BAFs at Volcán de Colima share multiple similarities (i.e. sustained collapse of a lava dome, multiple BAF pulses with the largest volume ones emplaced over a single flank of the volcanoes, and generation of overbank flows), the resulting deposit architecture and inferred flow dynamics are quite different. For instance, stronger clast segregation was observed in flow units of the 2006 deposits at Merapi with, however, less variability in lithofacies with distance and therefore less variability in their transport and deposition mechanisms.

These differences can be in part attributed to differences in the geomorphology of the valleys (i.e., slope, sinuosity, channel geometry) in which these BAFs were deposited (i.e., the Gendol valley at Merapi versus the Montegrande ravine at Volcán de Colima). These differences are: (1) maximum channel width, which is wider at Merapi by about 100 m; (2) sinuosity, which is notably more pronounced at Volcán de Colima than at Merapi; and (3) breaks in slope, which are more abundant at Merapi with steeper initial slopes than at Volcán de Colima. Furthermore, both the Merapi 2006 and the Volcán de Colima 2015 BAF events resulted in the generation of overbank flows escaping the confines of the Gendol valley and Montegrande ravine, respectively. However, the percentages of volume and area covered from the overbank deposits at Merapi (i.e. 4.4% and 11.7% respectively, Charbonnier and Gertisser, 2011) were significantly lower than the percentages calculated at Volcán de Colima (Table 3.3).
Figure 3.11. Pre-2015 channel capacity (black line) and width of the July 2015 overbank deposits (shaded blue) with distance from the vent. The dashed line indicates the critical channel capacity to overcome in order to start generating overbank processes.

A comparison between the channel capacity (i.e. cross-sectional area of the channel) and the width of overbank deposits from 120 transects drawn across the Montegrande ravine (Figure 3.11) reveals a positive correlation between the two parameters: as the channel capacity decreases, the width of overbank deposits increases. Previous work at Merapi identified sinuosity, valley infilling and channel confinement to be direct factors on the generation of voluminous overbank flows (Charbonnier and Gertisser, 2011; Lube et al., 2011; Cronin et al., 2013).

In the case of Volcán de Colima, complications arise when trying to estimate the effects of sinuosity (i.e., changes in channel direction) on the generation of overbank flows. For instance, changes in the channel direction by 120 degree in the proximal area of the volcano on slopes between 10 and 15° promoted the overspill of a small portion of overbank flows (Figures 3.1 and 3.3, location A), whereas the presence of 80 to 60 degree-sharp turns in the distal area only produced small overbank flows over the valley walls (locations C and D Figures 3.1, 3.3, and
Areas affected by overbank flows in the medial and distal areas of the volcanoes also suggest a complex effect from changes in channel direction combined with the reduction of channel capacity due to the presence of erosion terraces within the channel, the asymmetry of the channel geometry and local breaks in slope. While this may also have been the case at Merapi, the higher DEM resolution used for Volcán de Colima allowed to better identify the local channel geometry and the presence of these finer topographic features.

The presence of overbank deposits dominantly on one side of the valley suggests the potential for flow acceleration followed by lateral overspill when pairs of bends are present in the channel. However, the lack of accurate flow metrics such as velocity measurements to estimate the Froude number and the change in flow regime does not allow to calculate optimum change in channel direction for the generation of overbank flows. The correlations found in Figure 3.11 highlight the presence of a critical value of channel capacity defined at ~ 400 m² for which overbank flows start being generated. Similar correlations were found for the Gendol valley before and after the 2006 and 2010 eruptions at Merapi (Lube et al., 2011; Solikhin et al., 2015) with respective critical channel capacity values at 1100 m² and 3300 m².

The total volume of BAFs that were emplaced on the southern flank of Merapi during the 2006 eruption was significantly lower than the one obtained for the 2015 eruption at Colima (Table 3.3), which first suggests a volume- and/or mass-flux-dependence of the critical channel capacity for overbank generation. It should be noted that, due to the lack of syn-eruptive data regarding individual volumes and mass fluxes of the successive BAF pulses at Colima, a cumulative critical channel capacity with integrated values over the entire sequence of the July 2015 BAFs was calculated instead. Finally, during the 2006 Merapi eruption, overbank deposits were only observed in the proximal and medial areas on the southern flank. In the more distal areas, lower
flow velocities and a much larger channel capacity inhibited overbank processes, which implies that the critical channel capacity is a dynamic value that also varies with distance and local mass flow rate. This idea is supported by the presence of overbank flow units in the distal areas of the Montegrande ravine at Volcán de Colima, which correspond to the emplacement of the BAF pulse with the largest volume and mass flow rate (flow unit VC2). The significantly greater proportions of overbank flows generated in the distal areas during the VEI 4 Merapi eruption in 2010 corroborate these hypotheses as well (Charbonnier et al., 2013).

**Conceptual Model of BAF Transport and Deposition**

Increase in channel confinement (i.e., decrease in channel capacity) with distance induces flow convergence, which favours the processes of particle collisions over frictional interactions while inhibiting lateral spreading and thinning. Thus, the flow maintains high grain dispersive pressure and granular temperature with transport mainly driven by momentum transfer between particles through frequent collisions. Conversely, the decrease in channel confinement and the lateral spreading of the flow is accommodated by an increase in distance between particles within the flow, which results in a decreasing number of collisions between particles per unit of time (Dufek, 2016). Decreasing frequency of collisions leads to decreasing momentum transfer between clasts, hence the progressive loss of flow kinetic energy, which promotes rapid deposition and freezing of the main channelized flow body. The loss of flow kinetic energy and flow deceleration is further enhanced when channel enlargement is coupled with a break in slope, which induces a sudden increase in grain interlocking and normal bed frictional forces, thus promoting deposition.

This model is consistent with the following interpretations, as deduced from field investigations carried out along zones of topography-induced sedimentation of coarse-rich
material observed at two locations inside the distal July 2015 Colima BAF deposits (see section 5.1). A sudden drop of flow kinetic energy after channel enlargement results in rapid deposition and freezing of the basal portion of the current, which contains abundant large solid particles due to processes of flow density stratification through grain-dispersive pressure and fragmentation processes that occur during transport of the bulk material in the proximal and medial areas.

The correlations between valley-confined flow units described above suggest that this process of freezing of the basal portion of the current can occur more than once within the same unit (e.g. unit VC2). In such a model, a coarse-rich pulse front is formed by a frozen portion of the basal, highly concentrated flow region while the upper region of the flow, which is finer-grained with lower particle concentration, is transported further by decoupling and ramping over the frozen base (Figure 3.12). Downstream from the first coarse-rich lobe, channel constriction and confinement promotes further flow density stratification through grain dispersive pressure. Then, a similar dynamic process of rapid deposition and freezing of the basal underflow, as well as ramping of the upper flow region can repeat at the next channel enlargement area downstream.

This transport and deposition model for the July 2015 BAFs at Volcán de Colima could explain the fluctuations of grain size distributions observed inside flow unit VC2 between sections Mont4 and Mont8 (Figures 3.6 and 3.7, and Table 3.2) and is also consistent with the presence of the fine-grained lithofacies mLA for flow unit VC2 found in the distal portion of the volcaniclastic plain (section Mont9, Figures 3.7 and 3.9). In this model, deposition occurs stepwise (Figure 3.12), with aggradation of coarse-rich BAF pulse fronts, which is similar to the emplacement model initially proposed for all types of PDCs by Sulpizio et al. (2007) and revised by Charbonnier and Gertisser (2011) for long-runout BAFs at Merapi.
Lithofacies correlations between the valley-confined and wedge-shaped overbank flow units show that the arrival of each BAF pulse at a given site in the main valley controls the generation of overbank flow pulses on top of the valley margins. Reduction in channel capacity and progressive valley infilling from earlier BAF pulses facilitate the passive overspill of the main flow body over the valley margins, which rapidly experiences freezing and en masse deposition to form the wedge-shaped overbank deposit facies.

Similar mechanisms can be inferred for the emplacement of coarse-grained interfluve overbank deposits. The finer-grained characteristics of the overbank flow units in contrast to their respective valley-confined units suggest that they were generated from the overspilling of the upper and marginal regions of the valley-confined BAF pulses. However, due to conservation of kinetic energy inside the valley-confined BAF pulses in the proximal and medial areas, and due to strong processes of density particle segregation like kinematic squeezing occurring inside the flow when passing over steep slopes, the resulting overbank flows still contain a significant proportion of large clasts (mBLA lithofacies) in these areas.

In distal areas, progressive widening of the main channel resulted in the loss of kinetic energy of the BAF pulses, lower grain-dispersive pressures and weaker clast segregation, hence promoting the development of finer-grained upper and marginal flow regions. High channel sinuosity facilitates the overspill of these flow portions onto the valley interfluves as the flow accelerates on the outside of the bend due to centrifugal forces and encounters the valley walls, resulting in the fine-grained interfluve overbank deposit facies. Similar processes are inferred to explain the presence of fine-grained overbank deposits in the distal margins of coarse-grained interfluve and rechannelled overbank deposits.
Implication for Hazard Assessment

The mobility parameter ($\Delta H/L$) describes the ability of a gravity-driven mass flow to travel downslope (Iverson 1997). Calculation of the mobility parameter for the longest 2015 BAF pulse at Volcán de Colima (BAF unit VC2) gives a value of 0.25, while similar calculations made for the three overbank flows (generated at overspill locations A, B and E, Figure 3.1) give lower values between 0.13 and 0.23. The lower values obtained highlight a greater mobility of this type of flows compared to their valley-confined counterparts. Similar observations were made for the overbank flows generated on 14 June 2006 at Merapi (Charbonnier and Gertisser, 2011) and for the surge derived pyroclastic flows on the 25 June and 26 December 1997 at Soufriere Hills (Calder et al., 1999), and it was concluded that several factors like (1) the finer-grained character of these overbank flows, (2) the inherited momentum from their parent flows during overspill and/or decoupling processes and (3) their rapid rechannelization into narrow, adjacent tributaries, could have all contributed to their enhanced mobility.

Furthermore, comparison with mobility parameters obtained from other BAFs at Merapi, Unzen, Soufriere Hills and past eruptions at Volcán de Colima (Charbonnier and Gertisser, 2011) shows that the values obtained here are within the range of those from small-volume concentrated PDCs and dome-collapse BAFs at Soufriere Hills and Unzen volcanoes, but is lower (i.e., showing greater mobility) than past BAFs at Volcán de Colima and Merapi. In terms of other mobility metrics like the total affected area, the aspect ratio $A/V^{2/3}$ and plan-shape parameter $A/L^2$, the valley-confined and overbank deposits of the July 2015 BAFs at Colima fit within the trend of increasing mobility values with increasing volume (Table 3.3). Total values of these mobility metrics (i.e. Area, $A/V^{2/3}$, $A/L^2$) plot at the minimum end of the trend for cold debris avalanches and to the maximum end of the trend for dome-collapse concentrated PDCs (Charbonnier and
Moreover, the aspect ratios $A/V^{2/3}$ obtained for the three overbank flows at Volcán de Colima are similar to those obtained for the 14 June 2006 overbank flows at Merapi while their plan-shape ratios $A/L^2$ show higher values than their valley-confined counterparts, thus confirming the highly mobile character of such overbank flows.

![Figure 3.12](image)

**Figure 3.12.** Conceptual model for transport and deposition mechanisms of the July 2015 block-and-ash flow pulses at Volcán de Colima. **a)** Three-dimensional sketch illustrating the emplacement of two block-and-ash flow pulses inside a synthetic channel characterized by progressive widening and a break-in-slope. Magnification circles highlight the schematic three-dimensional particle organization inferred inside each pulse front at two different locations in the channel. **b)** Inferred transport mechanisms in the XZ plane. **c)** Inferred deposition mechanisms in the XZ plane.
With a long runout (9.3 km) and large volume (0.0077 km$^3$) emplaced over a relatively short time-period (~ 18 hours), the July 2015 BAF events at Volcán de Colima highlight the challenges posed by concentrated PDCs from sustained dome-collapse events in terms of hazard assessment and risk mitigation. First, the short amount of time (few hours) needed for such large flow volumes to be generated from sustained dome collapses constitutes a new and unrecognized threat at Colima that need to be taken into account by emergency planners and stakeholders. Furthermore, this eruption resulted in the largest volume and runout for BAFs encountered during historical times, within the range of those generated from column collapse during the 1913 Plinian eruption (Figure 3.4). In terms of frequency of activity, dome collapse eruptions and BAFs are more likely to occur than Plinian eruptions, and therefore, any robust short-term hazard assessment should have an emphasis on the threat posed by frequent, small-volume and short-runout BAF events. However, the July 2015 eruption at Volcán de Colima shows that large volume and long runout hazardous PDCs also need to be included in the list of threats associated with more frequent, small VEI, Vulcanian column-collapse and dome-collapse eruptions.

The implications of July 2015 Colima BAF events for hazard assessment also involve the occurrence of highly mobile overbank flows, which, as previously mentioned, have been examined from a few BAF eruptions before at other volcanoes such as Merapi and Soufriere Hills. This study has emphasized the direct relationship between changes in channel geometry (and more specifically changes in channel capacity and sinuosity) and the generation of overbank flows. Generation of high energy BAFs in a narrow and sinuous channel increases the potential of overspilling and lateral spreading of the main flow body when sudden changes in channel morphology occur. The overspill of the upper and marginal parts of the valley-confined flows also occur more frequently in areas of reduced channel capacity and high sinuosity. Natural topography
obstacles, such as lahar erosion terraces and irregular slopes within the main valley, and anthropogenic obstacles such as sabo-dam structures, increase the risk of overbank flows by reducing the valley-retaining capacity. Lateral spreading of these concentrated and highly mobile overbank flows results in flow paths that are difficult to predict and needs to be accounted for in current hazard maps.

Conclusion

Stratigraphy, sedimentology, grain size and component studies of the July 2015 BAF deposits at Volcán de Colima provide detailed information about:

(1) the distribution, volumes and sedimentological characteristics of the different units;

(2) flow parameters (i.e., velocity and dynamic pressure) and mobility metrics as inferred from associated deposits;

(3) changes in the dynamics of the different flows and their material during emplacement. These data were coupled with geomorphic analyses to assess the role of the topography in controlling the behaviour and impacts of the successive BAF pulses on the volcano flanks.

Lithofacies variations of the valley-confined BAF deposits were used to interpret the changes in flow dynamics with distance, and to suggest a conceptual model for transport and deposition mechanisms of such long-runout BAFs at Volcán de Colima. Channel confinement and sinuosity maintain high kinetic energy and momentum inside the channelized flows through high-grain dispersive pressures and clast segregation processes.

Conversely, a sudden decrease in channel confinement favoured lateral spreading and thinning (due to mass conservation) of the flow, in which the decrease in particle collisions together with the increase in grain interlocking in the coarse-rich basal region of the flow resulted
in the loss of kinetic energy and flow momentum. Sudden increase in frictional forces at the base led to rapid freezing of the coarse-rich basal flow region while the upper part of the flow ramped over the frozen base and moved further downstream.

In this model, deposition occurs by rapid stepwise aggradation of successive BAF pulses. Flow confinement in a narrow and sinuous channel enhances the mobility and runout of individual channelized BAF pulses. When these conditions occur, the progressive valley infilling from successive sustained dome-collapse events promote the overspill and lateral spreading of the upper and marginal regions of the main flow body, generating highly mobile overbank flows that travel outside of the main valley. Volume-, mass-flux- and distance-dependent critical channel capacities for the generation of overbank flows can be used to better estimate the inundation area of these hazardous unconfined pyroclastic flows.

Hazard assessment at Volcán de Colima and other active volcanoes presenting similar type of activity should involve extensive studies of the topographic and geomorphic characteristics of the volcano flanks, which could be pursued using newly available high resolution remote sensing tools (e.g., TanDEM-X, high-resolution optical imagery).

Using these new remote sensing tools, time-series analyses of the major and rapid topographic changes that often occur at active volcanoes (i.e., small sector collapse of the summit area, new dome growth and/or crater, formation of lahar erosional channels and depositional terraces in the surrounding valleys) can be undertaken.

Future work will aim to develop an extensive database of high-resolution DEMs at active explosive volcanoes that will account for significant topographic changes and will focus on evaluating the role of the topography on the transport and deposition dynamics of future BAFs.
References


CHAPTER FOUR:
FIELD INVESTIGATIONS OF THE PYROCLASTIC DENSITY CURRENT DEPOSITS FROM THE APRIL 2015 ERUPTION AT CALBUCO VOLCANO, CHILE

Introduction

This chapter presents results and interpretations of the pyroclastic density currents (PDCs) at Calbuco Volcano (hereafter referred to as Calbuco) and is one of the only studies to catalogue detailed observations of the PDC deposits in the Rio Frio – Rio Blanco Este confluence system and in the Rio Blanco Sur (Figure 4.1). A few other studies looked at the PDC deposits shortly after the eruption, but their respective observations and interpretations were restricted to a couple of outcrops, with a larger focus on tephra fallout (e.g., Clavero et al., 2015; Bertin et al., 2015; Castruccio et al., 2016). Castruccio et al. (2016) looked at two PDC outcrops, but interpretations were focused on the eruption dynamics and the magmatic composition using geochemical signatures of the deposits, rather than the dynamics of these PDCs.

In a similar manner as presented in Chapter 3 at Volcán de Colima (simplified hereafter as Colima), the sedimentary structures and sedimentology analysis are used to derive lithofacies of the PDC deposits, which can in turn be used to infer flow dynamics (Branney and Kokelaar, 2002; Sulpizio and Dellino, 2008; Sulpizio et al., 2014). However, some important differences between Colima and Calbuco are worth noticing:

1) The PDCs at Calbuco descended into many river catchments that radiate around the volcano, whereas the PDCs at Colima were limited to a single flank of the volcano (see Chapter 3). Satellite observations showed that the majority of PDCs and tephra fall primarily deposited
towards the NE direction (Figure 4.1). This coincides with the NE directionality of the volcanic plume during the first and second eruption phases on April 22 and 23 (see Chapter 2). A more complete description of the remote sensing observations and deposit distribution, along with a thorough description of the topography of Calbuco is presented in Chapter 5.

(2) The pre-eruptive morphology of the valleys at Calbuco is more complex than the Montegrande ravine at Colima (refer to Chapter 3 and Chapter 5 for further details). At Calbuco, the Rio Frio – Rio Blanco Este and the Rio Blanco Sur are wide valley structures (up to 240 m in the Rio Blanco Sur and 300 m in the Rio Frio), with low slopes (i.e., < 10° for the studied portion), and enclosed with high and steep walls of ~ 100 m. The valley floor is itself incised with narrow (up to 20 m-wide) and sinuous channels carved by the active rivers and streams, with walls up to 25 m-high and highly vegetated (pre-eruption). For this reason, the distribution of deposit facies of PDCs is discussed with respect to the pre-eruptive depositional environments/structures rather than in terms of valley-confined and overbank deposits as it was done for Colima.

Methodology

Field Observations and High-Resolution Remote Sensing Mapping

A field campaign was carried out in November 2016, allowing for extensive field observations of the April 2015 PDC deposits along the erosional channels formed from syn- and post-eruption lahars in the valleys of the Rio Blanco Este (NE flank) and the Rio Blanco Sur (S flank). The eruption of Calbuco was associated with hot lahars from the melting of glacier ice and snow at the summit of the volcano (see Chapter 2). Post-eruption lahars were generated from the continuous carving of the deposits by the existing rivers and streams in most valleys.
Figure 4.1. Distribution of the April 2015 eruption deposits at Calbuco draped over a 50-cm resolution Pleiades-1A image. Only concentrated PDCs are mapped with valley-confined deposits outlined in blue, and overbank deposits in pink.
The erosion with incised gullies and lahars deposits downstream was particularly strong in the Rio Blanco Este with significant erosion in the distal reaches, and part of the distal deposits in the Rio Blanco Sur were completely removed by lahars. In both valleys, the majority of the erosion channels cut the primary deposits longitudinally, in the flow direction, and only a few transversal sections were formed. The remobilization of deposits upstream results in the accumulation of rounded coarse materials deposited downstream, within a wet matrix with associated wavy stratifications and cross-bedding. These features are used to differentiate the reworked deposits and lahar deposits from the fresh PDC deposits.

The pre-eruptive topography was analyzed using a DEM at 4-m resolution generated from very high-resolution (VHR) DigitalGlobe WorldView-1 data acquired on May 22, 2014. Both the VHR orthorectified image and the DEM were used to interpret the pre-eruptive landscape. As support for field observations, VHR optical images acquired from the Pleiades-1A satellite on April 30, 2015 and on December 1, 2015 were used to map the extent of the PDC deposits and individual Lobes in the Rio Blanco Sur and Rio Frio – Rio Blanco Este, using the ArcGIS™ software (ESRI, Redlands, CA, USA). These high-resolution optical images were also used to digitally map deposits in all the other valleys (Figure 4.1). The comparison between the images acquired 7 days after the second eruptive phase (i.e., 04/30/2015) and 8 months after the eruption ended (i.e., 12/01/2015) shows the strong deposit erosion caused by lahars. The Pleiades stereo pair from 12-01-2015 was also used to generate a 1-m DEM post-eruption, which was subsequently used with the 2014 pre-eruption DEM to derive estimates of deposit volumes. However, this topic is the focus of Chapter 6 and therefore, I only present volume estimates for the respective segments of the Rio Frio - Rio Blanco Este confluence system and the Rio Blanco Sur where field work was performed (Table 4.1). I also calculated the mobility metrics such as the
total area of the mapped deposits in both valleys, and mobility metrics such as the geometric parameter \( A/V^{2/3} \) and the plan-shape parameter \( A/L^2 \) (Table 4.1).

**Table 4.1.** Mobility parameters for the studied PDCs at Calbuco. Area, volume, geometric \( (A/V^{2/3}) \) and plan-shape \( (A/L^2) \) parameters of the distal PDCs of the Rio Frio – Rio Blanco Este confluence system and the Rio Blanco Sur. The maximum runout distance of the PDC deposits is measured as a straight line from the crater rim. The volume estimates include the presence of tephra but was also calculated after erosion of the deposits had begun.

<table>
<thead>
<tr>
<th>Segments</th>
<th>Maximum Runout (km)</th>
<th>Area ( (x10^6 \text{ m}^2) )</th>
<th>Volume ( (x10^6 \text{ m}^3) )</th>
<th>( A/V^{2/3} )</th>
<th>( A/L^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rio Frio – Rio Blanco Este</td>
<td>6.3</td>
<td>0.6</td>
<td>5.2 ± 1.3</td>
<td>21.9</td>
<td>0.066</td>
</tr>
<tr>
<td>Rio Blanco Sur</td>
<td>4.2</td>
<td>0.2</td>
<td>2.3 ± 0.4</td>
<td>10.8</td>
<td>0.042</td>
</tr>
</tbody>
</table>

Following the naming convention used in Chapter 3 at Colima, the term ‘flow unit’ is used to refer to a single PDC deposited as one Lobe. One flow unit refers to a deposit layer bound by erosive or gradational contact (e.g., lapilli and/or ash layer, trains of clasts and/or breakout in grain size, sudden change in clast composition). In the Rio Blanco Este, a total of 24 stratigraphic sections were studied for stratigraphic and sedimentological characteristics of the different PDC flow units along a 1.6 km-long segment from 3.9 to 5.5 km distance from the summit (Figure 4.2). In the Rio Blanco Sur, 13 stratigraphic sections over 1.5 km in the most distal portion of deposits, were studied to map all the individual Lobes on high-resolution images (Figure 4.3).

**Grain Size and Componentry Analyses**

The methodology used for grain size and componentry analyses is similar to that employed for samples at Colima and presented in Chapter 3. A total of 11 samples were collected in the Rio Frio – Rio Blanco Este. For a given unit, where the outcrop was considered to be representative of the entire unit, a \( \sim 0.5 \times 1 \text{ m} \) rectangular area was defined to be sampled. Then, a \( 1 \text{ m}^3 \) box was used in-situ to perform manual dry sieving and subsequently to weight fractions between -8 and -
4 phi (i.e., 256 to 8 mm size clasts) in full-size phi intervals. The remaining fraction < -4 phi, considered as the matrix, was weighed on the field, then split using the cone-and-quarter method in order to carry a reasonable weight and representative matrix fraction back to the laboratory. The matrix fractions were then sieved using wet sieving equipment and each fraction weighed at 1 phi intervals, up to +5 phi (32 mm). Fractions smaller than +5 phi were not integrated into the grain size distribution to facilitate the comparison between samples. The estimated loss of fines during wet-sieving never exceeded 8% of the total sample mass. In the Rio Blanco Sur, 12 samples were collected, but only the matrix fraction was brought back to the lab for grain size analyses, without in-situ sieving. Therefore, the block-size fraction (i.e., larger than -4 φ) was neither weighted in the field nor included in the grain size analysis back in the lab. For a given sample, however, if the matrix fraction carried back to the laboratory included clasts larger than lapilli-size clasts (i.e., larger than -4 φ or 8 mm) they were weighted and included in the total grain size distribution.

For all samples, the final mass percentages were plotted into the GRADISTAT© software (Blott and Pye, 2001) to compute grain size parameters. The DECOLOG software (www.DECOLOG.org; Borselli and Sarocchi, 2009) was also used to apply deconvolution of the polymodal grain size distribution, which consists of dividing each distribution into a mixture of lognormal distributions based on three peak components. These peak components are statistically representative of the different particle size sub-populations and facilitate the comparison between samples. The grain size analyses of the samples from both valleys are summarized in Table 4.2.

A component analysis of each sample matrix was performed on clasts sizes between -2 and +3 phi under optical binocular microscope. For each fraction, a minimum of 500 gains were handpicked and assigned to one of three component categories: (1) juvenile, (2) lithics (altered clasts) and (3) free crystals. The juvenile category was divided into three sub-categories (1a) brown
scoria, (1b) dark-grey dense scoria, and (1c) dense whitish clasts, according to our field observations and following the results from petrological analyses by Castruccio et al. (2016). Both the brown scoria and the dark-grey dense scoria are basaltic andesite (54.7 – 55.4 wt. % SiO₂). The brown scoria presents a higher bulk vesicularity (30 – 60 vol. %) and lower density (1.3 – 1.4 g/cm³) than the dark-grey dense scoria (15 – 20 vol. % and 1.6 – 1.8 g/cm³ for vesicularity and density respectively). The whitish dense clasts, however, are almost completely crystalline without vesicles, and are andesitic (58.10 wt. % SiO₂). The proportion of each categories and sub-categories were then recalculated based on the weight percentage of each size fraction. The matrix portion was sieved in the laboratory using wet sieving technique, and componentry analyses to identify the percentage of the particle types was performed under a binocular microscope. The proportion of each component for each sample is displayed in Table 4.2.

Results

*Rio Frio – Rio Blanco Este*

The PDCs that descended onto the NE flank, reached the longest runouts in the Rio Frio – Rio Blanco Este confluence system, with a straight-line distance at 6.3 km between the crater rim and the terminal front Lobe (Figure 4.2). The erosion channel that cut through the deposits was caused by the syn-eruption lahars from glacier melts at the summit and post-eruption lahars from the underlying streams and rivers within the Rio Blanco Este and Rio Blanco Frio. The stronger erosion and removal of deposits at the confluence of the two valleys is a result of higher volume flux of water. The sustained eruption plume during the eruption also resulted in multiple tephra layers from fallouts emplaced during both the first and second eruption phase (Castruccio et al., 2016).
Table 4.2. Grain size parameters and componentry data of the studied samples at Calbuco.

$Md_0$ = Median size particle ($\phi$); $\sigma_0$ = Sorting. The 1st, 2nd and 3rd modes correspond to the three main peaks in grain sizes, obtained from deconvolution of the polymodal grain size distributions. Samples from the Rio Blanco Sur only consisted of the matrix portion and blocks were not included unless $< -5 \phi$. Matrix components are 1a = brown scoria; 1b = dark-grey dense scoria; 1c = whitish dense clasts; 2 = lithics; 3 = free crystals. See Table 4.3 for description of individual units.

<table>
<thead>
<tr>
<th>Distance from crater</th>
<th>Section</th>
<th>Unit</th>
<th>Sample Grain Size</th>
<th>Matrix Component (wt. %)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Blocks (wt.%)</td>
<td>Lapilli (wt.%)</td>
</tr>
<tr>
<td>3.9 km</td>
<td>RBE-2</td>
<td>Lobate</td>
<td>26.1</td>
<td>32.7</td>
</tr>
<tr>
<td>3.9 km</td>
<td>RBE-2</td>
<td>Lobate</td>
<td>8.9</td>
<td>46.5</td>
</tr>
<tr>
<td>4.0 km</td>
<td>RBE-3</td>
<td>Lobate</td>
<td>9.4</td>
<td>40.7</td>
</tr>
<tr>
<td>4.0 km</td>
<td>RBE-3</td>
<td>PDC-III</td>
<td>19.9</td>
<td>43.3</td>
</tr>
<tr>
<td>4.9 km</td>
<td>RBE-10</td>
<td>Lobate</td>
<td>42.5</td>
<td>38.1</td>
</tr>
<tr>
<td>5.0 km</td>
<td>RBE-11</td>
<td>PDC-III</td>
<td>18.5</td>
<td>33.9</td>
</tr>
<tr>
<td>5.3 km</td>
<td>RBE-13 PDC-III</td>
<td></td>
<td>12.3</td>
<td>49.4</td>
</tr>
<tr>
<td>5.3 km</td>
<td>RBE-14 PDC-III</td>
<td></td>
<td>9.0</td>
<td>52.2</td>
</tr>
<tr>
<td>5.5 km</td>
<td>RBE-15 PDC-II</td>
<td></td>
<td>24.2</td>
<td>35.3</td>
</tr>
<tr>
<td>5.5 km</td>
<td>RBE-15 PDC-I</td>
<td></td>
<td>1.4</td>
<td>39.9</td>
</tr>
<tr>
<td>5.5 km</td>
<td>RBE-16 PDC-II</td>
<td></td>
<td>16.4</td>
<td>29.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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Table 4.3. Description of the stratigraphic units from concentrated PDCs in the Rio Frio – Rio Blanco Este system. The most representative lithofacies characteristics of each flow unit are summarized along with results from grain size analyses.

<table>
<thead>
<tr>
<th>Flow Unit</th>
<th>Descriptions</th>
</tr>
</thead>
<tbody>
<tr>
<td>PDC-I</td>
<td>Massive brown layer ~1.5 to 2 m thick on top of a tephra layer from fallout deposits at the beginning of the eruption, which blankets the pre-eruptive soil (i.e., debris avalanche). The unit is depleted of blocks (1.4 wt. %) with a lapilli-ash matrix that appears consolidated. The matrix is composed of &gt; 50 wt.% of brown scoria, a few dark-grey scoriaceous clasts and dense whitish clasts. Although poorly-sorted ($\sigma \bar{d} = 2.8$), it is better sorted than any of the other PDC units sampled here. There are subtle lapilli trains, degassing pipes (gas segregation) cut by the overlying deposits, and charred wood fragments at the base that suggest high temperatures of emplacement. This unit only outcrops in the most distal part of the valley, at the confluence between the Rio Frio and Rio Blanco Este.</td>
</tr>
<tr>
<td>PDC-II</td>
<td>Massive dark-grey unit with a sharp basal contact with unit PDC-I underneath, and ~ 1.8 m – up to 4.5 m thick. The matrix is rich in dark-grey cauliflower bombs and dense dark-grey scoriaceous fragments (&gt;85 wt. %), but contains very little brown scoria (&lt; 1 wt. %). It presents large vertical segregation pipes with lapilli and blocks that cut through the whole unit. The unit is topped by a lapilli-ash tephra layer.</td>
</tr>
<tr>
<td>PDC-III</td>
<td>Massive brown unit with large whitish dense blocks (up to 1 m in diameter) and cauliflower bombs (up to 80 cm in diameter). It has a highly variable thickness, sometimes with sub-layers caused by coarse-tail grading or coarse-clast trains. The sub-layers have undulated and blurry contacts that make them difficult to track throughout the whole outcrop. The unit can present coarse-rich meter- to decameter-size lenses, and degassing pipes with vertical clast segregation that cut the whole layer. I make the interpretation that this unit is the main unit that outcrops throughout most of the sections. The coarse-rich front of this unit pinches out in the distal part of the Rio Frio – Rio Blanco Este.</td>
</tr>
<tr>
<td>Lobate PDC</td>
<td>This unit is mainly observed at the surface of other deposits with a lobate surface morphology and coarse-rich bulbous frontal snout and lateral levees. The snout and lateral levees are dominantly composed of brown scoria and scoriaceous bombs up to 60 cm diameter, and sometimes supporting whitish dense clasts up to meter size diameter. The unit is dominantly massive and poorly-sorted, with occasional coarse lenses of cauliflower bombs up to 50-cm diameter. It contains charred wood fragments from the burnt vegetation. The deposits consist of multiple overlapping Lobes, although we refer to it as a single unit, with varying thickness (between 1.2 and 9.9 m thickness measured). Individual lobate units may be separated with a thin lapilli layer. The most surficial lobate unit is capped by 5 – 9 cm-thick tephra layer from the end of the eruption.</td>
</tr>
</tbody>
</table>
Figure 4.2. Field locations in the Rio Frio and Rio Blanco Este valleys. a) NE flank of Calbuco on 50-cm Pleiades-1A image from 04-30-2015. Strong degassing emitting from the crater appears as a white plume. The bounding box indicates the area enlarged in the next image. b) GPS locations of the stratigraphic sections in the Rio Frio – Rio Blanco Este valleys referred to in the text. The terminal Lobe of the PDC deposits in the Rio Blanco Este is indicated with an arrow.
Although strong erosion from the lahars occurred shortly after the eruption ended, the considerable volume of total PDC and tephra fall deposits allowed for stratigraphic reconstruction of the flow units. The results presented hereafter focus first on the concentrated PDCs, then looking at surge deposits, and finally, additional observed surface features.

**Concentrated PDCs**

A total of 16 stratigraphic sections were described for the concentrated PDC flow units from the most proximal (i.e., 3.9 km) to most distal (i.e., 5.5 km) locations, and referred to as RBE-1 to RBE-16 (Figure 4.2). Based on stratigraphic reconstruction and sedimentology analysis, there were 4 PDC units recognized along the 1.8 km-long segment (i.e., PDC-I, PDC-II, PDC-III, and PDC-IV). Their respective lithofacies and main characteristics are summarized in Table 4.3. Hereafter, each section is described with interpreted units and sampling results when carried out.

**RBE-1**: The section, shown in Figure 4.3, cuts the deposits in the direction almost parallel to the flow direction. The stratigraphic sequence is ~ 4.1 m-thick but is not fully exposed at the base, and the flow units are deposited over what used to be a forested embankment next to the river stream path. Comparing the VHR pre- and post-eruption images with field observations, I concluded that the flow units are overlapping Lobes from the Lobate PDC, for which a few Lobe fronts can be discerned on the VHR image from April 30, 2015 (i.e., before erosion began). The two bottom units are inversely graded with scoriaceous bombs towards their respective upper contact. The top-most unit presents a decameter-size coarse lens in a finer brown matrix. Its surface is capped with the tephra layer from fallout. Superelevation deposits are present along the vertical valley wall with felled trees (Figure 4.3e). Superelevations occur as a result of centrifugal forces causing the flows to rise up against the outside bend of the valley. Finally, there is evidence of vertical segregation pipes from degassing that cut through most of the flow units.
Figure 4.3. PDC deposits at section RBE-1. **a)** Photograph of the most proximal location accessed on foot during field work. The yellow arrows indicate flow direction, including where the flow likely jumped a cliff. The red X is a point of reference to correlate the other images. **b)** 50-cm Pleiades image from April 30, 2015 showing the extent of the stratigraphic sections observed. The yellow arrows indicate flow direction similar to image a). **c-f)** The 3 panels are photographs of the stratigraphy of the deposits indicated with the red box on image b. The different flow units are separated with a dashed line and the corresponding flow units are linked from one panel to the next when possible. **g)** Corresponding units (Lobes) in map view on the VHR image from April 30, 2015. **h)** VHR Worldview-1 image from May 2014 showing the pre-eruptive valley floor and the path of the incised channels.

*Figure 4.3 continues on the following page.*
**RBE-2**: This section exposes the deposits at an angle of about $45^\circ$ with flow direction (Figure 4.4a). This section comprises three flow units that were interpreted as three Lobes from the Lobate PDC unit (Figure 4.4b). Lobe 1 is $\sim 1.7$ m-thick, although not fully exposed, and Lobe 2 is $3.0$ m-thick. The contacts between Lobes 1 and 2, and between Lobes 2 and 3 are lapilli layers of $\sim 5$-cm thick. All three Lobes present similar lithology, although the surface of Lobe 3 was reworked by lahars, for that reason, only Lobes 1 and 2 were sampled (Figure 4.4c-d; see Table 4.2 for results). Lobes 1 and 2 are both massive, very poorly sorted ($\sigma_\phi$ at 3.3 and 3.6 respectively) with a brownish grey matrix, rich in brown scoria, with a few dark-grey dense scoria and whitish dense clasts. Results from componentry analyses show similar composition with brown scoria accounting for more than 50 wt. % of both samples and very low lithic content (Table 4.2). Lobes 1, 2 and 3 present meter-size lenses of cauliflower bombs, and Lobe 2 contains a $1.4$ m-diameter dense whitish block toward the top of the unit. During sieving of the samples from Lobes 1 and 2, the presence of charred wood fragments was noticed. Results from grain size analyses show that Lobe 2 presented a higher block content than Lobe 1 (Table 4.2).
**Figure 4.4.** PDC deposits, grain size and componentry results at section RBE-2. a) Overview of the outcrop (red bounding box) with the lobate surface morphology. The yellow arrow indicates flow direction. b) The outcrop shows the sequence of three flow units, separated by thin lapilli-ash layers (dashed lines). The blue lines circle the coarse-rich lenses. Note the meter-size whitish dense block embedded in the middle unit. The granulometry and componentry results from sampling of the lowermost and middle flow units are presented in panels c) and d). The grey-scale pie charts present the granulometry results with B, L, A for Block, Lapilli and Ash content (wt.%) respectively. The colored pie charts present the componentry results with the following legend: brown = brown scoria; dark grey = dark-grey dense scoria; white = whitish dense clasts; blue = free crystals; red = lithics. This legend is applicable to all granulometry and componentry results in other figures.

**RBE-3:** This section presents overlapping Lobes from the Lobate PDC unit, which was sampled on a small breach within the deposits that was perpendicular to the flow direction (Figure 4.5). The lobate unit is massive, very poorly sorted ($\sigma_\phi = 3.3$) and inversely graded with a few blocks (30-15 cm in diameter). It also contains scoriaceous cauliflower bombs similar to those observed in the Lobe front at the surface (Figure 4.5c).
Figure 4.5. PDC deposits, grain size and componentry results at section RBE-3. a) Overview of the outcrop along the northernmost valley wall. The bounding box indicates the location of the other panels. The yellow arrow indicates flow direction. b) Photograph of a coarse-rich Lobe front at the surface of the outcrop. The thickness was estimated at 4.2 m. c) Photograph of the sampled unit interpreted as lobate PDC unit. The yellow measuring stick is 1-m long. d) Outcrop with two flow units: the topmost unit corresponds to the lobate PDC unit sampled and shown in image c. The lowermost unit is sampled here and interpreted as unit PDC-III. e-f) Results from grain size and component analyses of sampled units in c and d respectively. See figure 4.4 for legend of the granulometry and componentry pie charts.

RBE-3 (Cont’d): Grain size and componentry results are very similar to Lobes 1 and 2 in section RBE-2, although it contains a greater amount of accidental lithics (i.e., 4.6 wt. %; Table 4.2). At the sampling location the unit is about 2.0 m-thick, although not fully exposed. Other measurements of the Lobes’ thicknesses, using the distinctive surface feature, give values of 4.2, 8.0 and 9.9 m (Figure 4.5b).
There are also superelevation deposits against the lava cliff at ~4.8 m above the deposit surface, similar to those observed at section RBE-1. About 100 m downstream, along the same outcrop, the lobate unit at the top is fully exposed with thickness of 1.3 m and overlays a dark-brown unit ~1.7 m-thick but not fully exposed (Figure 4.5d). This unit is very poorly sorted ($\sigma_0 = 3.7$), massive with weak normal grading and segregation pipes that cut through the unit but not into the Lobate unit.

**Figure 4.6.** Photographs of the outcrops from sections RBE-4 to RBE-9. The order of the panels from a) to g) follows the flow direction from right to left, indicated with yellow arrows. When possible, the boundaries of the flow units are marked with dashed lines and named according to visual interpretations. In image a) the blue lines indicate degassing pipes. For image c), a zoom over the upper part of the section is shown next to it. The units are thin and in contact with the pre-eruptive soil, shown with a vegetation-rich layer at the bottom. This raised area was confirmed on a pre-eruption VHR image and DEM (see Figure 4.7).
**RBE-3 (Cont’d):** The unit is coarser with dark-grey and whitish dense blocks. Sampling results show a median grain size at -3.1φ confirming the coarser nature of the unit. Componentry results show dominance of brown scoria (> 50 wt. %), with a greater amount of dark-grey dense scoria than the Lobate unit (25 wt. %). Based on sedimentology results, this unit was interpreted as PDC-III.

![Figure 4.7](image)

**Figure 4.7.** PDC deposits at section RBE-6 and valley topographic profiles for sections RBE-4 to RBE-9. **a)** Overview of the outcrop at section RBE-6, indicated with a downward arrow and displayed in Figure 4.6c). The flow direction is indicated by the yellow arrow. **b)** Close-up of the center of the channel near section RBE-6. The red line indicates the pre-eruptive ground, observed with a thick lapilli-ash tephra fallout layer and vegetation-rich level. The units observed below the pre-eruptive ground level correspond to where the channel once was and banked against the wall. **c)** Pre-eruption VHR image showing the path of the channel and the locations of sections RBE-4 to RBE-9. The dark blue lines are channel cross-sections displayed in panel **d)**. The profiles were obtained with a 1-m pre-eruption DEM.
Figure 4.8. PDC deposits, grain size and componentry results at section RBE-10. a) Photograph of the coarse-rich Lobe fronts, looking upstream. The yellow arrows indicate flow direction with the divergence onto the valley interfluve. b) Sampling of the lobate PDC unit, through the tephra fall. c) Close-up of the exposed unit with the thick tephra cover at the top. d) Map view of the overbank lobate PDCs visible on VHR image. e) Results from sampling of the lobate PDC unit. Granulometry legend: B, L, A for Block, Lapilli and Ash content (wt.%); Componentry legend: brown = brown scoria; dark grey = dark-grey dense scoria; white = whitish dense clasts; blue = free crystals; red = lithics.

**RBE-4 to RBE-9:** These sections were used to reconstruct the stratigraphy of PDC deposits, but access to the deposits was impossible due to the height of the outcrops and the dangerously loose nature of the deposits along the stream. Therefore, the interpretations for each unit rely solely on visual observations (Figure 4.6). The lowermost PDC unit, PDC-I, is only visible when the erosion by the river was maximal (Figure 4.6.a, b, d). The contact with unit PDC-II is sometimes difficult to perceive. Unit PDC-II was capped with a tephra layer, which was used
as a marker. At section RBE-5 (Figure 4.6.b), the middle unit interpreted as unit PDC-II, presents a train of coarser particles with whitish dense blocks. At the top, the overlapping of two Lobes from the Lobate PDC is shown by the contact against a large block that was carried by the previous Lobe. At section RBE-6 (Figure 4.6.c), the deposits are much thinner and only expose units PDC-III and Lobate PDC, on top of the valley interfluve. Topographic profile of the valley pre-eruption indicates higher walls covered by vegetation after a strong break in slope (Figure 4.7). Inside the riverbed, it seems that units PDC-I and PDC-II were deposited and subsequently eroded. This suggests infilling of the channel prior to deposition on top of the valley interfluves.

Figure 4.9. Photos of the lobate surface of PDCs and tree damages. a) Photograph of a coarse-rich Lobe front at section Oth-4. The yellow arrow indicates flow direction. b) Superelevation deposit from the lobate PDC unit near section Oth-4. c) Close-up on the blown down and bent trees along the superelevation deposit. The felled trees are a great proxy for the direction of the flow during transport.
**RBE-10:** This location is used to look at the grain-size and componentry of the surface of a Lobe front from the Lobate PDC (Figure 4.8). Sampling was difficult because of the ~ 40 cm-thick tephra layer that covers the unit, and we could only expose 30 cm of the Lobate PDC (Figure 4.8b-c). Measurement of the Lobe thickness, however, gives 2.3 m near the frontal snout. The matrix consists of 60 wt. % of brown scoria, similar to the other sampled Lobes at RBE-2 and RBE-3. It is very poorly sorted and much coarser than the other samples with a median particle size at -4.7 φ and a proportion of blocks greater than the proportion of lapilli (Table 4.2). This is consistent with the coarse-rich frontal snout composed of brown block-size scoria.

**Table 4.4.** Thickness measurements of surface Lobe fronts at sections in the Rio Frio – Rio Blanco Este (refer to Figure 4.2 for locations). The values are minimum thicknesses because the base of the Lobes could not be seen from the surface.

<table>
<thead>
<tr>
<th>Section</th>
<th>Oth-2</th>
<th>Oth-3</th>
<th>Oth-4</th>
<th>Oth-5</th>
<th>Oth-6</th>
<th>Oth-7</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lobe thickness (m)</td>
<td>5.0</td>
<td>2.1</td>
<td>2.1</td>
<td>1.2</td>
<td>1.3</td>
<td>2.3</td>
</tr>
</tbody>
</table>

Measurements of Lobe thicknesses were carried out at various locations along the Rio Frio – Rio Blanco Este at sections Oth-2 to Oth-7 (Table 4.4). These same locations were used to look at the damage to the vegetation. There is a superelevation unit at section Oth-4 that appears to be similar in composition to the Lobate PDC unit (Figure 4.9). The trees are blowdown and bent in the direction of the flow, and there is an upward decrease in tree damage where trees are left standing but wilted, dead and unburnt.

**RBE-11:** This section cuts the deposits perpendicular to the flow direction and is located at the point of channel constriction of the Rio Frio, prior to the confluence with the Rio Blanco Este (Figure 4.10). The outcrop shows a top greyish-brown unit 1.4 m-thick, interpreted as a Lobate PDC unit, and a bottom brown unit with a minimum thickness of 1.3 m, although not fully exposed, separated by an 8 cm-thick lapilli layer (Figure 4.10a).
Figure 4.10. PDC deposits, grain size and componentry results at section RBE-11. a) Stratigraphic sequence at section RBE-11, showing two flow units separated with a lapilli-ash layer. The yellow stick is 1 m tall. b) Granulometry and Componentry results from sampling of the lower unit. See Figure 4.4 for legend of the granulometry and componentry pie charts. c) Another outcrop a few meters downstream showing the same two units separated with the lapilli-ash layer. The top unit is recognized as the lobate PDC unit, while the bottom unit is interpreted as PDC-III unit. d) Post-eruption VHR image showing the location of section RBE-11 and the other outcrop in image c. Note the strong degassing of the deposits that was occurring at the time of image acquisition (i.e., April 30, 2015). e) Pre-eruption VHR image of the same area in image d. The locations of the two outcrops are shown with yellow dots. The three lines indicate the cross-section profiles shown in image f). The profiles of the channel highlight the narrow morphology and the constriction prior to the confluence with the Rio Blanco Este channel to the east.
Figure 4.11. Comparison between pre-eruption (a) and post-eruption (b) very high-resolution images at the confluence of the Rio Frio and Rio Blanco Este. The location of sections RBE-12 to RBE-16 are indicated. The pre-eruptive river is shown with a blue dashed line. The post-eruption degassing of the deposits and the hot lahars are indicated as well.
**RBE-11 (Cont’d):** The bottom unit is massive, very poorly sorted ($\sigma_\phi = 3.7$) with a majority of brown scoria (51.7%, Table 4.2) and up to 26 wt. % of dark-grey dense scoria. Although the median grain size distribution is lower than that of PDC-III at section RBE-3 (i.e., -1.3 $\phi$), the relatively similar componentry and facies led to interpretations that this unit is the upper part of the PDC-III unit within the confined path. Strong confinement of the PDC may have resulted in vertical expansion, thus allowing for the upper dilute part of the flow to deposit over the valley wall. The thin lapilli-layer at the top before the Lobate PDC unit suggests time-lapse with accumulation of tephra before deposition of the Lobe.

**RBE-12:** This section is located on the outer bend of the confluence between the Rio Frio and Rio Blanco Este, which consists of significant channel enlargement (Figures 4.11 and 4.12). The deposits are emplaced over a forested embankment (Figure 4.11a) and are not fully exposed at the base, covered with lahar deposits that eroded and reworked the lowermost PDC units. The stratigraphy shows the distinctive dark-grey unit PDC-II at the base, with a thin lapilli layer at the contact with the brown unit PDC-III and finally the Lobate PDC unit (Figure 4.12c). Unit PDC-III shows three inversely graded sub-units (PDC-IIIa to c), with a higher content in whitish dense blocks and lapilli (Figure 4.12d). The block fabric orientation is oriented in a direction parallel to the flow and to the section layering. Sub-units IIIa and IIIc present larger blocks than sub-unit IIIb. There is evidence of segregation pipes that cut through all the sub-layers. The total unit thickness is 5.3 m with the sub-unit thicknesses at 2.3, 1.2 and 1.8 m respectively from bottom to top at this section. The whole stratigraphic sequence thickness varies between 7.4 and 9.4 m.

The upper unit (Lobate PDC) seems to have a consistent thickness of ~ 0.8 to 1 m. With distance, moving downstream, there is a longitudinal transition from three to two inversely graded sub-units and eventually a single unit. The longitudinal change in lithofacies of PDC-III is also
seen on the inner bend of the valley across the river, although the outcrops could not be accessed on foot (Figure 4.13). Seemingly, the unit transitions from inversely graded to normally graded with deposition of large blocks, after which a coarse-free PDC-III unit travelled further downstream (Figure 4.13b).

**RBE-13:** as previously stated, this section shows a highly disorganized unit PDC-III with large angular whitish dense blocks within the whole unit and large degassing pipes with clast segregation (Figure 4.14). The unit thickness varies between 4.5 and 8.4 m, it is very poorly sorted ($\sigma_\phi = 3.5$) with a median particle size at -2.6 phi. The Lobate PDC unit is present at the top, with a thickness of ~ 1 m. The matrix composition seems to be equal part brown scoria and dark-grey dense scoria. Unit PDC-III seems to be pinching out between dark-grey unit PDC-II and the Lobate PDC unit above (Figure 4.14a).

![Figure 4.12. PDC deposits at section RBE-12. a) Location of section RBE-12 on post-eruption VHR image. b) Exposed PDC deposit units in the confluence area. The yellow arrows indicate flow direction. c) Close-up of the outcrop at section RBE-12, with the interpreted stratigraphy. d) Close-up of the deposits focused on unit PDC-III showing the sub-units with strong clast segregation and degassing pipes. The yellow stick is 1 m tall.](image-url)
Figure 4.13. PDC deposits along the inner bend across section RBE-12. a) Photograph of the deposits on the inner bend of the confluence area, across the erosion channel from section RBE-12. The deposits are seen in contact with the substrate on top of an embankment. The yellow arrow indicates flow direction. The deposits show evidence of coarse-tail inverse grading and a progressive decrease in thickness. b) Continuation of the outcrop with the deposits thinning out and transitioning from coarse-tail normal grading to fine-grained. The base of the deposit shows a vegetation rich level.
Figure 4.14. PDC deposits, gain size and componentry results at section RBE-13. **a)** Photograph of the PDC deposits around section RBE-13 along the outer bend of the confluence area. The different interpreted units are outlined when possible. **b)** Close up of the outcrop at section RBE-13. **c)** Zooming-in on the sampled section. **d)** Close up of the large degassing pipes with clast segregation. In all three photographs of the outcrop, the yellow stick is 1 m tall. **e)** Granulometry and componentry results from sampling unit PDC-III at section RBE-13. See Figure 4.4 for legend of the granulometry and componentry pie charts.
**Figure 4.15.** PDC deposits, gain size and componentry results at section RBE-14. a) Photograph from the top of the partially collapsed deposits at section RBE-14. The yellow arrow indicates flow direction. b) Close-up of the sampled deposit unit. The yellow stick is 1 m tall. Note the tephra layer that caps the PDC deposits. c) Grain size and componentry results from sampling. Granulometry legend: B, L, A for Block, Lapilli and Ash content (wt.%); Componentry legend: brown = brown scoria; dark grey = dark-grey dense scoria; white = whitish dense clasts; blue = free crystals; red = lithics.

**RBE-14:** this section exposes the deposits perpendicularly to the flow direction (Figure 4.15). The top unit corresponds to the Lobate PDC unit, overlaying a fine-grained PDC-III unit. It is in contact with the dark-grey PDC-II unit capped with the tephra layer marker. Although the block content is lower, the lithology (grain size and component results, Figure 4.15c) is consistent with that of the other samples from PDC-III (Table 4.2).

**RBE-15:** this section is the most distal valley-confined outcrop that was studied during the field campaign (Figure 4.16). The pre-eruptive soil, an old debris avalanche, can be seen below the layer of tephra fallout from the beginning of the eruption. The tephra layer is ~ 20-25 cm-thick, well sorted, composed of sub-angular lapilli and ash fragments, and seemingly consistent...
throughout the whole outcrop. This lapilli-ash layer exhibits 3 sub-units, finer (7 cm-thick), coarser (10 cm-thick) then finer again (4-5 cm-thick). Above this tephra fall layer, is unit PDC-I which is a massive, with a brown-colored matrix, 1.5 to 2 m-thick, poorly sorted ($\sigma_\phi = 2.8$) with a low block content ($\sim 1$ wt.%) and a fine median particle size at 0 phi. There are subtle lapilli trains within the unit. The second PDC unit corresponds to the dark-grey PDC-II unit capped with the tephra layer of $\sim 15$ cm. Unit PDC-II is $\sim 4.5$ km-thick and seems consistent throughout the whole outcrop.

![Figure 4.16. Most distal outcrops of the PDC deposits observed during field work and PDC deposits at section RBE-15. a) – c) Three panels showing the longitudinally exposed units around section RBE-15. The individual flow units are traced and named according to interpretations. D.A = Debris Avalanche. d) Outcrop at section RBE-15 where the bottom unit and second (dark-grey) unit (i.e., PDC-I and PDC-II) were sampled. Granulometry and componentry results are shown in e) and f) for PDC-I and PDC-II respectively. Granulometry legend: B, L, A for Block, Lapilli and Ash content (wt.%); Componentry legend: brown = brown scoria; dark grey = dark-grey dense scoria; white = whitish dense clasts; blue = free crystals; red = lithics.](image-url)
**RBE-15 (Cont’d):** It is very poorly sorted ($\sigma_\phi = 3.9$) with a block content greater than 25 wt. % and a median particle size at $-2.3 \, \phi$. The componentry analysis indicates that 86 wt.% of the juvenile clasts are dense dark grey scoria, which gives the distinctive dark grey, almost black coloration to the matrix of this PDC unit. The remaining juveniles are whitish dense clasts, and there are dark-grey cauliflower bombs. There is no apparent grading although the larger bombs and blocks seem to be mainly distributed towards the lower and upper edges of the unit. The third unit corresponds to PDC-III, similar in appearance to the sampled unit at section RBE-14. It is massive, brown, ~1.5 m-thick with a lapilli matrix. Finally, there is the Lobate PDC unit, ~ 80 cm-thick covered with the final tephra layer.

**Figure 4.17.** Overview of the deposits along the inner bend of the confluence area across from section RBE-15. **a)** Flow direction and jump over the embankment interpreted from deposits and blown down trees. **b)** Dark-grey unit PDC-II observed inside the channel. The unit above is largely reworked by lahars (hot and cold). The raised channel bank displays blown down trees and wilted trees from the Lobate PDC unit. **c)** Massive unit PDC-III is only partially reworked at the surface, and displays a longitudinal transition from coarse-tail inverse grading to coarse-tail normal grading. The Lobate PDC unit is observed on top of the channel bank, with blown down trees and wilted trees. The damage to trees decreases upward.
**RBE-15 (Cont’d):** Across the erosion channel from section RBE-15, along the inner bend of the confluence area, photographs of the deposits show the presence of reworked units that appear to be PDC-II in contact with PDC-III (Figure 4.17). The Lobate PDCs are also observed along the valley wall with superelevation deposits. Unit PDC-II is not fully exposed, with a minimum thickness of ~1.5 – 2 m, and is located at the point that used to be in the river bed pathway on the pre-eruptive terrain. Unit PDC-III is largely reworked towards the center of the erosion channel, where it is interpreted as overlaying unit PDC-II, but appears less affected towards the raised area along the inner bend. The contact with the substrate is sometimes observed with a discontinuous non-carbonized vegetation-rich horizon.

**Figure 4.18.** Overbank deposits, grain size and componentry at section RBE-16. a) Photograph overlooking the deposits on the outer bend of the confluence area, with the main flow direction and that of the overbank flow. b) Close-up on the location of section RBE-16 behind trees. c) Outcrop sampled at section RBE-16, where the overbank unit is in contact with the pre-eruptive surface. d) Sampling results with grain size and componentry analyses. Granulometry legend: B, L, A for Block, Lapilli and Ash content (wt.%); Componentry legend: brown = brown scoria; dark grey = dark-grey dense scoria; white = whitish dense clasts; blue = free crystals; red = lithics.
**RBE-15 (Cont’d):** The upper few decimeters of the deposits are reworked by lahars, but the remaining bulk of the deposit unit PDC-III displays sedimentary features similar to those observed along the outer bend at section RBE-12: there is a transition from inverse coarse-tail grading and coarse clast trains to normal coarse-tail grading. The deposits of unit PDC-III also display degassing pipes that cut through the reworked part of the deposits. This suggests that degassing was occurring during erosion by the hot lahars that descended with the PDCs. Observations on VHR images confirm the presence of hot lahars at that location that were rapidly reworking the deposits a few days after the eruption. Finally, the lobate PDC unit covers the channel embankment and overlays unit PDC-III. Damage to the vegetation includes bent and felled trees in the direction of the flow, as well as singed and standing trees where the superelevation deposits occurred. The VHR image from April 30, 2015 shows lobate morphologies of the surface of the deposits up to the terminal front of the PDC deposits in the Rio Blanco Este.

**RBE-16:** this section is located on top of the valley interfluve, about 60 m laterally away from section RBE-15 (Figure 5.18). There is only one unit, massive, ~ 1.6 m-thick in contact with the pre-eruptive soil and covered with the tephra fall deposits from the end of the eruption. At this outcrop, the unit is exposed perpendicularly to the flow direction and contains cauliflower bombs similar to those observed in unit PDC-II. The block and lapilli contents are lower than the valley-confined unit PDC-II by 7.7 and 5.7 wt. % respectively and the ash content is greater by 13.4 wt. %. The median grain size is -0.5 φ (Table 4.2). Additionally, componentry analysis of the matrix reveals that this overbank unit is similar to that of the valley-confined unit with 68 wt. % of the juvenile material being dense dark grey clasts. The content of whitish dense clasts is slightly higher than the valley-confined unit and there is a ~6 wt. % increase in the lithic content. These results support the idea that this overbank unit observed on the valley bank was generated from the
overspill of valley-confined unit PDC-II. During the overbanking process, a smaller proportion of blocks are transported over the valley edge, which explains the lower block content. Additionally, the increased proportion of lithic particles can be explained by the erosive behavior of the overbank flow during overspill, particularly while travelling over a forested surface.

**Surge deposits**

Although not the primary focus of this field campaign, there are three sections (i.e., ASC-1 to 3) where surge deposits and their impact to vegetation were described.

![Surge deposits images](image-url)

**Figure 4.19.** Photographs of superelevation and surge deposits at sections ASC-1 to ASC-3. **a)** Photograph of the superelevation deposits at section ASC-1. **b)** Photograph of the superelevation deposits at section ASC-2 with the decreasing tree damages upward. **c)** Bottom of the superelevation deposits at section ASC-2 with the bent trees in the direction of the moving flow. **d)** Close-up of the fine-grained lapilli-ash deposits at section ASC-2. **e)** Looking downstream of ASC-2. **f)** Surge deposits over the promontory at section ASC-3. The trees are bent in the direction of the flow. The yellow arrow in each panel represents the flow direction.
ASC-1: this location corresponds to a superelevation of surge deposits up to 24.1 m against the valley wall on the inner bend (Figure 4.19a). There is an upward decrease in the damage to the vegetation, with blowdown and burnt trees at the base of the superelevation, and above stripped but still standing towards the top. The felled trees are oriented with the direction of the valley bend.

ASC-2: this section is ~ 300 m downstream from section ASC-1 and also presents superelevation of surge deposits up to 14.1 m high (Figure 4.19b). The blowdown trees at the base of the superelevation are oriented in the direction of the bend (Figure 4.19c). Upward, the trees are left standing although burnt and stripped from branches and foliage (Figure 4.19e). At the base of the superelevation, the surge layer is about 9 cm-thick, between two tephra layers (Figure 4.19d). The surge deposit contains dead and dried vegetation fragments and is composed of a lapilli-ash matrix.

ASC-3: This section is located downstream of section RBE-15 in the distal part of the valley, on top of the riverbank. The trees are bent in the direction of the forward motion of the ash-cloud surge (Figure 4.19f).

Other deposit features

Lahar deposits were described at section Oth-1 with two types of lahar deposits recognized: hot and cold lahars (Figure 4.20). Hot lahar deposits are thin deposits with lobate morphology, rich in brown scoria and accidental lithics and appear moderately sorted (Figure 4.20a). Hot lahars were seen on VHR imagery immediately after the end of the second eruptive phase on April 23 (Google Earth imagery). The cold lahar deposits are very coarse-grained, light grey with a thin lobate morphology that cuts through PDC deposits and the hot lahar deposits (Figure 4.20b-c). The unit is depleted of fines and does not present surface evidence of degassing pipes or tephra blanket layer, suggesting deposition after the eruption. This was confirmed with VHR Pleiades-1A
imagery from December 1, 2015, showing the deposit erosion and a contrast between the light-grey-colored lahars and the dark-colored PDC and tephra deposits (Figure 4.20d).

Other surface features of the PDC deposits include the undulated (concave and convex) surface of the Lobate PDCs (Figure 4.21a) and the surface expression of degassing pipes with hydrothermal alteration of the deposits (sulfur crust) in a circular pattern (Figure 4.21b). In the distal part of the deposits, the strong degassing of the PDC deposits accentuated the concave surface and sinking of the deposit surface, particularly along the erosion path. Degassing was visible on the VHR image from April 30, 2015, in the distal part of the deposits, along the portion of deposits that was eroded the fastest afterwards.

**Figure 4.20.** Hot and Cold lahar deposits and lobate surface morphology. **a)** Photograph of the surface of cold and hot lahar deposits with finger-like morphology at section Oth-1. The deposits of cold lahars are deposited over the hot lahar deposits. The flow direction is indicated with a yellow arrow **b)** Enlargement over the cold lahar deposits cutting through PDC deposits. **c)** Photograph of the surface of cold and hot lahar deposits cutting through the lobate PDC deposits at the surface of section RBE-2. **d)** VHR Pleiades-1A image from December 2015 showing the contrast between PDC deposits in brown and cold-lahar deposits in lighter color during the erosive process.
Figure 4.21. Photographs of the surface of deposits. a) Undulated and convex surface of the lobate PDC unit in the Rio Frio. b) Surface expression of degassing pipes with hydrothermal alteration of the deposits in a circular pattern.

Rio Blanco Sur

The PDC deposits in the Rio Blanco Sur experienced less erosion than those of the Rio Frio – Rio Blanco Este. Erosion of the deposits occurred mainly along the valley walls on both sides of the valley. Access to the deposits was restricted to a 1.5 km segment in the most distal portion due to the presence of a vertical cliff at 2.7 km from the crater rim.

In contrast to the PDC deposits of the Rio Frio – Rio Blanco Este, the deposits of the Rio Blanco Sur were not blanketed by a tephra fall layer. The deposits present a distinct lobate morphology with finger-like Lobes exhibiting lateral levees coarser and higher than a central channel, terminating with a coarse and bulbous frontal snout. The Lobes are overlapping in a retrograde fashion (i.e., later Lobes stopping in the more proximal distance), and the multiple Lobes are observable both on VHR images and in the field (Figure 4.22a).

Individual Lobes were mapped on a VHR Pleiades-1 image from April 30, 2015 (i.e., a week after the second paroxysmal eruptive phase), using validation from the field observations to interpret the stratigraphic relationships between the Lobes both in map and cross-section views (Figure 4.22b). A comparison of the VHR images from immediately after the eruption (04-30-
2015) and from 9 months later (12-01-15) allows for the identification of the portions of the deposits that were completely remobilized by lahars. The erosion observed during field work is comparable to the erosion observed on the VHR image from December 2015, and can be considered a match for correlation with field notes. Figure 4.23 gives an overlook of the deposits and relative erosion.

Field observations show a strong inverse grading in each Lobe, which allows for identifying individual flow units in cross-section. For a given Lobe, its levees and frontal snout are much coarser than the remaining part of the flow unit. Moreover, observations suggest a longitudinal progression in the clast assemblages of the Lobes with time: in the most distal part of the deposits (i.e., earlier deposits) the composition of the Lobes is relatively constant with dominance of brown scoria and scoriaceous bombs, as well as a few small dark-grey dense scoria. Moving upstream in the Lobes (i.e., later deposits), the composition shows an increase in dark-grey dense scoria and an increase in both size and proportion of whitish dense clasts. The most proximal cross-section of the Lobes shows large whitish dense blocks up to 60-cm diameter. Other meter-size boulders were present in the proximal Lobes, as well as silicic blocks from the lava flow that forms the vertical cliff and pre-eruptive substrate.

Hereafter, Figures 4.24 to 4.40 are used to show cross-sectional views of the Lobes mapped in Figure 4.22b, with the corresponding section indicated. The legend of each figure will serve as a summary of observations and interpretations. PDC deposits are fragile, easily erodible and invaluable information may become lost over time. These incredible lobate features are infrequently recorded with such detail. Therefore, my intent for this section is to constitute an archive for all of the data collected during the field campaign in November 2016, by means of annotated and well-located photographs.
Figure 4.22. The Rio Blanco Sur and mapped PDC deposit Lobes.  
a) SW flank of Calbuco on 50 cm Pleiades-1A image acquired on April 30, 2015. Strong degassing emitting from the crater appears as a white plume. The bounding box indicates the area enlarged in the next image. 
b) Mapped PDC Lobes in the Rio Blanco Sur and the locations of the relevant stratigraphic sections used for deposit interpretation and/or sampling.
Figure 4.23. Photo overview of the deposits in the Rio Blanco Sur and distal sections. a) Overview of the Rio Blanco Sur from the west side of the valley interfluve, looking north toward the deposits and the volcano. b) Close-up of the lobate deposits and location of the distal RBS sections, numbered according to the nomenclature in Figure 4.22.
Figure 4.24. Cross-section of the Lobes at section RBS-1. a) Massive units with strong coarse-tail inverse grading corresponding to Lobes 1 to 3. The maximum unit thicknesses measured in the field are indicated. Note the large whitish dense block (> 60 cm in diameter) supported in Lobe 3. b) Exposed Lobes downstream of image a, along the same outcrop displaying the intricate overlapping of the Lobes. We note that the Lobe composition is strongly dominated by brown scoria, with scoriaceous cauliflower bombs at the Lobe margins.
Figure 4.25. Cross-sections of Lobes 2 and 3 at section RBS-2, exposed longitudinally with flow direction. Note the presence of decimeter-size whitish dense blocks in Lobe 3. Lobe 2 is only partially exposed. Lobe 3 shows strong inverse grading.
**Figure 4.26.** Cross-section of Lobes 1 to 3 exposed longitudinally, a few meters downslope of section RBS-2.

**Figure 4.27.** Cross section of Lobes 1 to 4 exposed at a 45° angle and then longitudinally with flow direction at section RBS-3. The inverse coarse-tail grading of each Lobe is well-displayed. Note the coarse-rich levee of Lobe 3 against the inclined rough substrate caused by the old andesitic lava flow.
Figure 4.28. Lobes 2 to 4 at section RBS-4. a) Overview of Lobes 2 to 4 at section RBS-4, which are exposed longitudinally and perpendicularly to the flow direction. Note the coarse-rich, raised levees at the surface of Lobe 4. The bounding boxes indicate the close-ups in images b, c and d. The deposits are rich in brown scoria and show overlapping finger-like Lobes, with coarse-rich Lobe perimeters, particularly at the surface. The yellow stick is 1-m tall for scale.
Figure 4.29. Deposits of Lobes 1 to 4 at section RBS-5. The Lobes are exposed longitudinally and perpendicularly to flow direction, which goes from panel 3 to 1. Field hammer for scale (circled), and the blue arrows are representing flow direction. This section is located on the easter side of the valley, at the same distance from the summit as section RBS-4.

Figure 4.30. Cross-section of Lobes 4 and 3 at section RBS-7. The flow direction is indicated with the blue arrows and goes from the left to the right panel (note the summit of Calbuco to the left corner of the first panel). The yellow stick is 1-m tall for scale.
Figure 4.31. Stratigraphic section and surface view of the Lobes at section RBS-6. Panels a) and b) are an overview of Lobes 3 to 5 longitudinally at section RBS-6. In panel a) the surface of Lobe 5 can be seen above the surface of Lobe 4 in the distance, across the Lobes from where I was standing (circled). The lobate, finger-like morphology of the deposits is clearly visible. Panels c), d) and e) show the surface of the Lobes while walking from W to E across the deposits. The flow direction is indicated with white arrows, and the Lobes are indicated when correlation with the map was possible. The yellow circled X is a point of reference on images a) and c) for 3D positioning. The yellow stick is 1-m tall for scale.
Figure 4.32. Stratigraphic section and surface view of the Lobes at section RBS-8. Panels a) and b) show the cross-sections of Lobes 3 and 4 exposed parallel to flow direction at section RBS-8. The yellow stick is 1-m tall for scale. Panels c), d) and e) are photographs of the deposit surface, taken from above section RBS-8, looking upslope (c), across (d) and downslope (e), respectively. When possible, the surface extent of the individual Lobes was drawn on the photographs, with colors matching those of the mapped lobes in Figure 4.22.
Figure 4.33. Outcrops of Lobes 1, 2 and 4 at section RBS-9, parallel to flow direction. The double red lines represent the matching point to follow the first panel into the second one. Field hammer for scale in the second panel, and me sitting in the third panel (blue circles).

Figure 4.34. Outcrop with Lobes 9 and 10 exposed parallel to flow direction at section RBS-10. Note the finer-grained appearance of the flow units in contrast to the distal Lobes. These deposits are still displaying inverse grading. The outcrop is located against the pre-eruptive substrate, into the water stream pathway.
Figure 4.35. Dilute deposits at section RBS-11. a) Map of the Lobes on VHR Pleiades-1A image from April 30, 2015, at section RBS-11. The yellow dots indicate the location of the pictures for panels c to e. b) Pre-eruption VHR Worldview-1 panchromatic image showing the riverbed path (blue dotted line) and the direction that the detached dilute flows at section RBS-11 may have taken due to the sinuous channel (thick blue arrows). My interpretations are that the dilute upper part of the flows detached and travelled in a straight direction while the dense-basal avalanche remained confined to the riverbed path. c) Evidence of tree damage on the edges of the dilute deposits, which are exposed perpendicularly to flow direction (blue arrow). The singed and leafless vegetation suggests it was impacted by dilute flows rather than by the concentrated ones. d) Outcrop of dilute flows along the valley wall. Myself next to the outcrop for scale (circled). e) Close-up of the outcrop of dilute deposits exposed parallel to flow direction (indicated with a blue arrow), and progressively thinning from a thickness of 2.8 m to < 1 m. f) Outcrop of Lobe 6, exposed parallel to flow direction, and located across the erosion channel from the dilute deposits (i.e., shown on panel e). The yellow stick is 1-m tall for scale.
Figure 4.36. Surface overview of the lobes from section RBS-11. a) Map of the Lobes on VHR Pleiades-1A image from April 30, 2015, at location RBS-11. The arrows indicate the viewing directions of the photographs in panels b), c), d) and e). Panels b) and c) show a vertical cliff along the western side of the Rio Blanco Sur, where post-eruption lahars descended and severely damaged the vegetation. The flow direction of the dilute deposits is indicated with blue arrows. The vegetation is bent and broken in the direction of the dilute deposits, further supporting the idea that the dilute part of the flows followed a straight trajectory, while the dense basal avalanche remained confined to the channel path. Panels d) and e) show the surface of Lobes 16, 17 and 18, which are the most proximal lobes that were mapped, with respect to the summit (see Figure 4.22). The 20 m-high vertical cliff can be spotted in the background on panel e), annotated and indicated with a black arrow.
Figure 4.37. Overview of Lobes 18, 16, 14, 13 and 12 from section RBS-12 to section RBS-13. Close-up of the deposits exposed parallel to flow direction. The deposits appear enriched in whitish dense clasts in comparison to earlier Lobes in the distal sections. The deposit surface also shows enriched content in whitish dense clasts. Myself, sitting for scale (circled).
Figure 4.38. Cross-section and surface view of the Lobes at section RBS-13. a) Photograph overview of the Lobes’ surface at section RBS-12, about 200-m downstream from the point of valley constriction with the vertical cliff. b) Close-up photograph of an outcrop along the erosion channel, where the deposits of Lobes 13, 14 and 15 (from bottom to top), are exposed parallel to the flow direction. c) Close-up photograph of the vertical cliff and the waterfalls at the point of constriction of the valley. The position from where this photograph was taken is the furthest and highest point of the Rio Blanco Sur that I was able to access during field work.
Figure 4.39. Singed vegetation from the heat damage of the ash-cloud surge. The singed area, which is delineated with the dotted line, is limited to the outer bend, and near the vertical cliff and valley bottleneck geometry. The dilute deposits were emplaced over the eastern side of the valley (refer to map in Figure 4.22).

Figure 4.40. Surface of Lobes 16 and 17 at section RBS-14. Note the distinct finger-like convex shape and surface enriched in coarse clasts, including an increased number of dense clasts.
Discussion

Timing of Deposits and the Eruption Dynamics

As presented in Table 4.3, four PDC units with varying lithological facies were differentiated and described in the Rio Frio – Rio Blanco Este. However, it is difficult to quantify the true number of PDCs that descended the valley because of the sub-units in PDC-III and the numerous overlapping Lobes of the Lobate PDC. Moreover, the deposits described in this chapter are restricted to the medial and distal parts of the Rio Frio – Rio Blanco Este, and an unknown number of PDCs may have deposited in the more proximal channels.

Castruccio et al. (2016) and Romero et al. (2016) counted 8 PDC units, but only analyzed two stratigraphic sections at the confluence between the Rio Frio and the Rio Blanco Este, which had already been eroded by the time field work was conducted for this study (Nov. 2016). In the Rio Blanco Sur, a total of 18 flow units were mapped, but the more proximal deposits may have been composed of other Lobes that could neither be accessed nor mapped.

Prior studies of the tephra layers from the 2015 eruption of Calbuco suggest that there were four tephra layers (layers 0 – 3), which were related to the two paroxysmal phases of the eruption (e.g., Bertin et al., 2015; Van Eaton et al., 2016; Castruccio et al., 2016). Layer 0 was associated with the first eruption phase, while layers 1 to 3 were correlated with the second phase. The presence of tephra layers 0 and 1 at the base of the PDCs stratigraphic sequence, in contact with the pre-eruptive soil in the Rio Frio – Rio Blanco Este, suggests that these PDCs were generated during the second eruption phase.

This is consistent with findings by Van Eaton et al. (2016) who described the second phase as pulsatory with greater intensity and noted a stopping of the umbrella cloud expansion at ~ 3:00 am (local time), which could suggest decreasing eruption rate with a column more susceptible to
partial collapses. Van Eaton et al. (2016) also described a dramatic rise in the rates of proximal volcanic lightning stroke after ~ 6:30 am (local time), which could have been triggered by the rise of low-level ash-clouds from PDCs.

It is more difficult to determine the time of occurrence of the PDCs in the Rio Blanco Sur, because there were no tephra layers to use as a marker. However, the deposits showed a gradual change in clast assemblages, with an increase in dark-grey dense scoria and whitish dense clast content in later PDCs, similar to the change observed in tephra layers.

According to Castruccio et al. (2016), tephra layers 0 and 1 were dominated by brown scoria, and layers 2 and 3 presented an increase in dark-grey dense scoria as well as whitish dense clasts. Similarly, in the PDCs of the Rio Frio – Rio Blanco Este, there was a transition between unit PDC-I and PDC-II with a brown scoria-dominated unit PDC-I and a dark-grey dense scoria-dominated unit PDC-II. Tephra layer 2 was located above unit PDC-II, with an increased content in dark-grey dense scoria relative to layers 0 and 1.

Furthermore, unit PDC-III presented large whitish dense blocks, and both units PDC-III and Lobate PDCs presented an increased content in whitish dense clasts. Because the composition of the earlier Lobes in the Rio Blanco Sur present similar clast assemblages as unit PDC-I in the Rio Blanco Sur, i.e., dominance of brown scoria, it can be proposed that the Lobes of the Rio Blanco Sur also correspond to the second phase of the eruption.

It can also be argued, however, that the coarser grain-size of the scoriaceous Lobes in the Rio Blanco Sur and the greater content in medium to large scoriaceous bombs differ from unit PDC-I (i.e., finer-grained, block content ~1.4 wt. %), and therefore might have begun depositing during the first eruption phase.
Deposit Facies and Interpretations of Flow Dynamics

The strong channelization of the PDCs in these two valleys, and more generally in all impacted valleys (as seen on VHR imagery), indicates a predominant transport as unexpanded high concentration granular flows. The extent of surge deposits detected from VHR images suggest that ash-cloud surge expansion was restricted to the proximal areas with steep slopes, where it was able to overtop topographic barriers (e.g., 200 m-high ridges). In the medial and distal areas, ash-cloud surges were detected with vegetation damage (i.e., burnt versus unaffected vegetation) and seemed confined to the lateral extent of the concentrated PDCs as fringes. This is consistent with findings of surge behavior at Soufriere Hills at Montserrat by Ogburn et al. (2014).

There are two main deposit lithofacies being recognized in the studied valleys of Calbuco. First there are the massive deposit units comprising the bulk of the channel-confined deposits in the incised channels that carve the large valley floor of the Rio Frio and the confluence area of the Rio Frio and Rio Blanco Este. Second there are the scoria rich lobate deposits with levee – channel morphologies and bulbous coarse-rich Lobe fronts, overlaying the massive deposits in the Rio Frio – Rio Blanco Este and comprising the entirety of the deposits in the Rio Blanco Sur.

Channel-confined massive deposits

The massive units present moderate to strong clast segregation with coarse-tail inverse grading and clasts trains. Clast segregation suggests high granular temperature from intense particle collisions, which generated high grain-dispersive pressure that subsequently induced segregation processes of particles such as kinetic sieving (Middleton, 1970; Savage and Lun, 1988) and kinematic squeezing (Le Roux, 2003). Kinetic sieving results in the migration of small clasts towards the bottom part of the flow unit, while kinematic squeezing favors the migration of large clasts towards the top part of the flow unit.
The strong clast segregation facies is primarily observed when the deposits are confined to the river beds in the Rio Frio. The hypothesis is that channel confinement induces flow convergence, which in turn increases particle-particle collisions while increasing flow thickness. This is in agreement with the deposition of the upper part of the confined flow units over the valley interfluves. Numerical simulations by Stinton (2007) showed that a simulated PDC travelling through channel constriction presents an increased thickness by a factor equal to that of the change in channel width. This is consistent with the unsteadiness of individual pulses with deposit thickness changing with distance.

The coarse-tail inverse grading within each sub-unit, the coarse-clast trains and the development of a lower coarse-free part at the base of each sub-unit indicate that the effects of density clast segregation processes were still high when the flow entered the wide confluence area. Stinton (2007) also showed the energization of flows caused by constriction, because of an increase in grain concentrations that results in an increase in the frequency of grain collisions and grain dispersive pressure (granular temperature). This leads to higher flow kinetic energy able to maintain momentum and therefore resulting in longer runouts. The channelization of the massive units in the small channels of the Rio Frio, and the constriction prior to the arrival at the confluence with the Rio Blanco Este may have enhanced flow thickness and kinetic energy, which maintained high granular temperature sufficient for maintaining strong clast segregation.

The presence of sub-units also suggests progressive aggradation of granular-flow pulses generated during transport, similar to the model described by Sulpizio and Dellino (2008). Clast trains may disappear locally and be combined with lenses of coarse-clasts at the base of the lowermost sub-unit, which indicate unsteady flow conditions. The granular-flow pulses that create sub-units may be due to pulses at the source, or caused by flow unsteadiness during transport with
the development of kinematic waves (Schwarzkopf et al., 2005). The increased particle-particle collisions caused by channel confinement is also observed with the increasing content in free crystals with distance, which suggest increased fragmentation inside the flow pulses.

The presence of a massive, thin, fine-grained and well-sorted lapilli-ash layer between stacked flow units can indicate a time-lag between deposition of individual units, or entrapment and settling of the ash-cloud by the overlaying unit, similar to the model proposed by Sulpizio and Dellino (2008). The presence of these deposit facies with sedimentary structures such as clast trains and coarse-tail grading are suggestive of a depositional mechanism by *en masse* freezing of individual sub-units. Each pulse stops *en masse* when the driving forces fall below the resistance forces due to friction at the base and grain interlocking.

In the distal part of the massive deposit units, however, evidence of progressive aggradation is suggested by the longitudinal transition from multiple sub-units to a single unit with inverse coarse-tail grading, which transitions to normal coarse-tail grading, and finally, a coarse-free unit that thins out progressively downslope. This is characteristic of PDCs waning in intensity (e.g., Druitt, 1998; Torres-Orozco et al., 2018).

The depositional mechanism is similar to the conceptual model described in Chapter 3 at Colima where the dense undercurrent drops momentum, depositing the large clasts, and the upper part (finer-grained) ramps over and travels forward with progressive deposition that results in deposits thinning out.

The presence of inverse coarse-tail grading and transition to normal grading in the most distal deposits at the wide confluence area was observed for deposits that remained somewhat channeled, thus supporting the theory of maintained high momentum caused by channelization of the flow. There is also evidence of fluidization and high pore-fluid pressure shown by gas escape
structures, both vertical degassing pipes cutting through the deposits and circular hydrothermally altered patches affecting the surface of the deposits.

Fluidization in a granular flow describes the reduction of inter-particle frictions caused by high interstitial gas pore pressure, also referred to as high pore-fluid pressure. Fluidization, or high pore-fluid pressure, has been proposed as one of the mechanisms to explain long runout distance of dense granular flows as well as coarse clast support (e.g., Sparks 1978; Wilson, 1984; Branney and Kokelaar, 2002; Roche, 2012).

The combination of high pore-fluid pressure as well as high granular temperature may have contributed to the strong clast segregation facies observed in massive deposits. Branney and Kokelaar (2002) and Komorowski et al. (2013) note that the combination of high deposition rates of individual pulses and high pore-fluid pressure can trap gases in the deposits, which can subsequently escape as degassing pipes within seconds after deposition.

Soft sediment deformations in dry pyroclastic flows have been associated with deflation processes caused by diffusion of pore-fluid pressure (Valentine et al., 2020). In their study, Valentine et al. (2020) estimated diffusion time scales for individual flow units to range from a few seconds to ~ 10 min, considering bed thickness up to 2 m. Although the calculations were performed for fine-grained ignimbrites, using experimentally-derived pressure diffusion coefficients determined for ignimbrite materials with similar fine ash contents (Druitt et al., 2007; Roche et al., 2016), the order of magnitude of the time scales can be applied to the coarse-grained and poorly sorted deposits at Calbuco.

This means that deposit deflation and compaction may have occurred within hours to days after deposition, which is consistent with the evidence of surface degassing observed on VHR imagery a few days after the eruption. Moreover, surface morphology of the deposits emplaced
over channel interfluves, which consist of inclined slopes in a direction perpendicular to flow
direction, show irregular collapse scars. These can be interpreted as unstable deposits that
compacted in an irregular manner shortly after deposition, and is largely caused by the highly
rugged pre-eruptive shape of the valley floor.

**Lobate deposits and levee-channel morphology**

The characteristic levee-channel morphologies of the lobate deposit facies, which can also
be referred to as scoria flows, have been described in other PDCs, and reported as typical of
friction-dominated dense granular flows (Rowley et al., 1981; Lube et al., 2007a; Jessop et al.,
2012). The formation of levee-channel morphologies has been obtained experimentally and
numerically with dry granular flow dynamics, where particle segregation causes the development
of lateral quasi-static zones enclosing a finer-grained central channel where the frictional forces
are reduced, allowing increased runouts (e.g., Felix and Thomas, 2004; Mangeney et al., 2007;
Kokelaar et al., 2013).

Felix and Thomas (2004) also showed that levee-channel morphology can occur for self-
channelized unconfined dry granular flows. Coarse-rich lateral borders of the flow induce finger-
producing granular-front instabilities in polydisperse dry or fluidized granular flows (Pouliquen
and Vallance, 1999). In these studies, however, the levees can only be achieved on slopes around
the repose angle of the particles (27° – 29°), which is significantly higher than natural slopes over
which these deposits are observed. For example, at Lascar Volcano in Chile, the 1993 pumice
flows displayed levee-channel morphologies in the distal deposits emplaced over slopes <5°
(Sparks et al., 1997; Calder et al., 2000; Jessop et al., 2012). At Ngauruhoe in New-Zealand, the
1975 PDC deposits presented levee-channel morphologies in the distal facies for gentle to
moderate slopes < 20° (Lube et al., 2007b).
Other lobate morphologies of PDC deposits have generally been reported in depositional environments that consist of broad, high and steep-walled valley with low slopes (i.e., unconfined environment), for example at Merapi (e.g., Schwarzkopf et al., 2005; Charbonnier and Gertisser, 2008); at Colima (e.g., Rodriguez-Elizarraras et al., 1991; Saucedo et al., 2002), at Unzen (e.g., Miyabuchi, 1999; Ui et al. 1999), and at Soufriere Hills (e.g., Cole et al., 2002).

Using the average levee thickness from field measurements of pumice flows at Lascar volcano, Jessop et al. (2012) calculated the theoretical minimum slope angle for motion to be possible (i.e., friction angle of pumice flows). They found that for slopes below 6°, the flows would be unable to travel far as the frictional forces would be too great. The friction coefficient for natural flows is much smaller than that of dry granular material, which may reflect fluidization of the granular mass (e.g., Sparks, 1978; Wilson, 1980; Roche et al., 2004).

This theory was confirmed in a recent study by Gueugneau et al. (2017), in which experimental and numerical simulations showed that fluidized granular flows, i.e., with initial pore-fluid pressure, are capable of reproducing the levee-channel morphologies for gentle slopes between 12° and 6°. The model used a Coulomb rheology with pore-pressure as opposed to dry Coulomb rheology. Their numerical simulations, however, resulted in a nearly-drained center channel. This can be explained by the fact that deposit aggradation (i.e., overlapping of Lobes) and particle segregation are not taken into account in the simulations, but these phenomena may change material permeability, hence the diffusion timescales of pore-fluid pressure (e.g., Roche, 2012). The combination of dense granular flow behavior with pore-fluid pressure would explain the greater flow mobility (i.e., long runouts) and deposits with levee-channel morphologies over low slopes <5° (i.e., below the friction angle).
Now applying these concepts at Calbuco, a difference in flow regime can be determined between the lobate deposits of the Rio Frio – Rio Blanco Este and those of the Rio Blanco Sur. The lobate deposit facies in the Rio Frio – Rio Blanco Este showed evidence of gas escape structures and longer runouts for individual flow units than the scoria flows in the Rio Blanco Sur, which do not show gas escape structures. The longest Lobes deposited up to the terminal front of the PDC deposits in the Rio Blanco Este at 6.3 km, while the maximum runout for Lobe 1 in the Rio Blanco Sur stopped at 4.1 km.

These observations suggest that the lobate deposit facies in the Rio Frio – Rio Blanco Este resulted from partially fluidized granular flow dynamics, while the scoria flows in the Rio Blanco Sur resulted from friction-dominated dense granular flow dynamics. It can be argued, however, that the larger supply of materials during column collapse towards the NE sector of the volcano, likely contributed to the greater mobility and longer runouts of the lobate PDCs. In fact, the presence of superelevation deposits from the lobate PDCs observed up to the most distal part of the deposits in the Rio Blanco Este can be interpreted as evidence for greater flow energy as a result of increased particle concentrations from the greater supply at the source.

Looking at the slope angles of the underlying surface showed that in both valleys, deposition of the lobate deposit facies occurred on slopes between 8°-5°. These slopes were measured on the pre-eruptive topography, but it is also very likely that the progressive deposition of PDC Lobes decreased the slope angles locally, increasing the frictional stress for subsequent Lobes. This theory explains well the retrogressive deposition of the Lobes in the Rio Blanco Sur, where the accumulation of Lobes near section RBS-14 decreased the local slope by 3°. In the Rio Frio, the presence of coarse-rich lenses observed in the lobate deposits may represent buried Lobe fronts, which also highlights the overlapping of multiple Lobes both in progressive and
retrogressive aggradation. Moreover, in the Rio Frio – Rio Blanco Este, the infilling of the channels contributed to increased surface smoothness and changed the depositional environment into a wide (almost unconfined), low slope terrain.

In a similar way, the Lobes in the Rio Blanco Sur were deposited in the area of valley enlargement, after a zone of constriction and a break in slope from 18° to 8°, hence in a similar depositional environment as previously described with a wide, relatively unconfined, and low slope surface. Furthermore, the break in slope and the channel constriction also present a vertical cliff ~ 20 – 30 m high (Figures 4.36 and 4.38). The transport over the cliff may have resulted in a hydraulic jump that promoted the separation between the dense and dilute regimes, with the lofting of fine particles and the sudden decrease in velocity of the dense bed-load (Burgisser and Bergantz, 2002). The hydraulic jump may have also contributed to the release of gas-pore pressure, thus reducing the gas fraction, increasing flow density, and therefore transitioning towards the dry granular flow end-member regime.

This interpretation is supported by the evidence of ash-cloud surge damages to vegetation on the valley wall near the vertical cliff, as well as the deposition of dilute PDC deposits over the promontory to the east of the valley (Figure 4.39). The loss of flow confinement after the passage through channel constriction may have led to radial dispersion which may have promoted degassing and the release of pore-fluid pressure as well.

The presence of thick dilute deposits along the western valley wall (section RBS-11, Figure 4.35) also suggests separation of the dilute phase, which I interpret as resulting from the bend in the valley. Earlier Lobes were likely less affected as they followed the carved path on the rugged slopes caused by the old andesitic lava flows and water streams, hence resulting in narrower and longer Lobes. It is worth noting, however, that the strong vertical clast segregation observed in
individual Lobes, with scoriaceous bombs in the upper margins of the flow units, supports the idea of partial fluidization of the granular mass for clast support. Furthermore, there were several occurrences of meter-sized dense blocks supported within a single flow unit, which implies high pore-pressure to generate sufficient energy for supporting large clasts. Dense clast support was also observed in the lobate PDCs of the Rio Frio – Rio Blanco Este, which is in agreement with dense granular flows with high pore-fluid pressure as previously suggested.

The lobate deposit facies seems to occur with minimal disturbance to underlying units, either with the massive PDC units or with other Lobes. This suggests limited eroding capabilities, which prevented successive flow units from entraining materials and therefore from increasing their kinetic energy by bulking. The low lithic content of Lobe 1 in the Rio Blanco Sur, which supposedly travelled over the fresh pre-eruptive substrate, further supports the weak capability of the flows to abrade and entrain local materials.

**Interpretations for transition of flow dynamics**

Observations of two main deposit facies (i.e., massive and lobate) are interpreted in terms of transitional flow dynamics as a response to changes in the geomorphology of the depositional environments, as well as changes at the source. The inferred sequence of flow dynamics of PDCs in the Rio Frio – Rio Blanco Este and the Rio Blanco Sur is summarized as follows:

1. Partial column collapses generated the majority of PDCs into the valleys of the NE flank of the volcano. The high concentration PDCs that reached the Rio Frio and descended towards the confluence with the Rio Blanco Este were primarily transported as valley-confined flows, following the carved pathways within larger valley structures. High particle concentration in the dense underflow combined with channel confinement led to a high collision rate between particles and increased particle fragmentation, which developed high granular temperature and
upward gas flow released that increased pore-fluid pressure (i.e., fluidized bed). The combination of high pore-fluid pressure and high granular temperature contributed to strong clast segregation processes and the development of kinematic waves. The transport dynamics of the PDCs into the Rio Frio and Rio Blanco Este is therefore described as unsteady pore-pressure modified granular flows, where the high kinetic collisional regime and the reduced frictional interactions from high pore-fluid pressure generated highly energetic and unsteady PDCs. The energization of the flows under channel constriction promotes long runouts and a temporarily sustained high dispersive pressure even as the channel width increases. The deposition in the confined channels occurs as rapid aggradation of sub-units, similar to “en masse” freezing, maintaining the inverse grading and clast trains in the final deposit structures. The unsteadiness of the flows and generation of kinematic waves result in variable deposit thicknesses longitudinally, and passive overbanking with thin deposits over the interfluves. When the width of the channel increases in the confluence area, the deposition regime rapidly transitions to progressive aggradation as the flows experience a drop in momentum, with a progressive thinning and decreased grain size of the flow units.

(2) The infilling of the channels by the massive deposits transforms the depositional environment into a wide valley, with low to absent lateral confinement and smooth gentle slope angles. This leads to radially spreading and thinning flows that lose pore-fluid pressure and progressively decelerate. The flow dynamics transitions toward the dry granular flow end-member, with a transport regime dominated by grain interlocking and frictional forces that overcome the driving forces, inducing a drop of kinetic energy and dispersive pressure. Deposition occurs with steep and coarse-rich lobate fronts enriched in low density clasts (scoriaceous cauliflower bombs), and the formation of levee-channel morphology, characteristic of dry granular flows.
Particle segregation in the levees engenders self-channelization by reducing permeability of the flow borders, which maintains a partially fluidized central channel that can travel over slope angles well below the angle of repose for dry pyroclastic materials (~ 20°). Transition to friction-dominated granular flow behavior also occurs after a strong break in slope, which creates a hydraulic jump, enhancing the escape of pore-fluid pressure and lofting of fine particles upward.

(3) High kinetic energy of lobate PDCs was maintained for longer runout distances in the Rio Frio and Rio Blanco Este because of the higher mass flux at the source during the final column collapses that allowed for sustained high pore-fluid pressure. Under the effects of gravity and centrifugal forces, the high energy lobate PDCs were able to “slosh” against the valley walls, creating superelevation deposits.

**Quantitative Estimates of Velocity, Dynamic Pressure and Temperature**

The presence of superelevation deposits can be used to estimate minimum flow velocity when centrifugal effects cause the mass to rise up and bank against the outer bend using the following relationship (Zanchetta et al., 2004; Sheridan et al., 2005):

\[
v = \sqrt{ghr/b}
\]

(1)

where \(h\) is the height difference between the channelized deposits and the top of the superelevation deposit, \(b\) is the channel width and \(r\) is the radius of the curvature. Using this method, minimum flow velocities obtained for the lobate PDC units in the Rio Frio – Rio Blanco Este decrease from 10.5 to 5.2 m/s with distance.
Another method that also uses the balance of centripetal and gravitational forces, consists in using the morphology of the levee-channel deposits that bank against the valley wall, as proposed by Evans et al. (2001):

$$v = \sqrt{(gh_{diff}r)/W} \quad (2)$$

where \(h_{diff}\) is the height difference between the deposit levees (considering the superelevation as one of the levees) and \(W\) is the horizontal distance between the levees. With equation 2, a minimum flow velocity of 6.0 m/s was obtained near section Oth-4 in the Rio Frio – Rio Blanco Este.

Jessop et al. (2012) used morphological measurements of pumice Lobes at Lascar to derive velocity estimates, using the following equation:

$$v \approx \beta \frac{g\gamma^3h_{ave}^3}{h_c} \quad (3)$$

\(\beta\) is a dimensionless constant that depends on the material grain size, with a typical value \(\beta = 0.65\) for sand (Forterre and Pouliquen, 2003), which Jessop et al. (2012) considered appropriate for pumice flows. \(\gamma\) is also a dimensionless constant of 1.25 obtained experimentally for polydisperse flows. \(h_{ave}\) is the average levee thickness calculated as \(h_{ave} = \frac{1}{2}(h_l + h_r)\) with the subscript \(l\) and \(r\) referring to the left and right levee respectively. \(h_c\) is the thickness of the central channel. Equation 3 was applied on Lobe 5 in the Rio Blanco Sur (see Figure 4.X) because of the well-formed levee-channel morphology that could be measured on high-resolution DEM and in the field. Using a range of \(h_{ave}\) and \(h_c\) values from multiple measurements along Lobe 5, we obtained a range of velocities between 3.0 and 7.0 m/s (decreasing when approaching the snout).

The velocity estimates in both the Rio Blanco Este and Rio Blanco Sur are in good agreement with velocity estimates for pumice flows by Jessop et al. (2012). These velocities also
roughly correspond with observations of pumice flows from other eruptions around ~ 10 m/s (Hobblit, 1986; Cole et al., 2002), which were estimated in proximal areas.

Dynamic pressure is a measure of the lateral stress exerted by the current (Valentine, 1998; Doronzo, 2013), and is often used in reference to the potential impacts and damages that PDCs can have on vegetation and infrastructures (e.g., Valentine, 1998; Baxter et al., 1998; Clarke and Voight, 2000; Cole et al., 2002; Dellino et al., 2010; Jenkins et al., 2013). Flow dynamic pressure can be calculated from velocity estimate using the following equation:

\[ P_{\text{dyn}} = 0.5 \rho_f v^2 \]  

with \( \rho_f \) as the flow density, estimated at ~ 1490 kg/m\(^3\) following the method by Zanchetta et al. (2004). This yields a range of dynamic pressure between 74.5 kPa and 18.6 kPa. Dynamic pressures < 3 kPa are sufficient for blown down trees (Cole et al., 2002; Clark and Voight, 2000). Dynamic pressures in excess of 15 kPa and 5 kPa are required for transporting blocks larger than 1 m and 30 cm respectively. This is consistent with the large blocks transported into flow units in the distal area.

Temperature measurements were not performed directly in the field, but the damage to the vegetation is typically used as a proxy for estimates. Charred or carbonized wood was rarely found in the PDC deposits, suggesting that the flows were not hot enough to turn healthy wood into charcoal, which typically occurs above 250 – 300°C. Blown down trees along the PDC deposits were dead but unburnt, while the standing trees displayed a decrease in heat damage from the base upwards: the transition was observed from bent bare tree trunks, to standing but de-limbed tree trunks, to less wilted but unburnt dead vegetation, to healthy trees. Similar low temperatures of PDCs (i.e., < 300°C) have been recorded for other eruptions, such as Merapi (e.g., Jenkins et al., 2013) and Tungurahua (Hall et al., 2015).
Conclusion

This chapter provided a comprehensive analysis of the PDC deposits from the 22-23 April 2015 eruption at Calbuco. Although the field observations were restricted to the deposits from two valleys, in the medial to distal areas, valuable information was extracted from deposit facies. The variation of deposit facies, derived from assemblages of sedimentary structures and sedimentological characteristics, were interpreted in terms of transport and deposition mechanisms. These mechanisms were correlated to the depositional environments, leading to inferred relationships between the morphologies of the valleys and the flow dynamics.

It was concluded that the earlier PDCs remained confined to the main channels, with pore-pressure modified granular flow dynamics, where momentum transfer occurs via particle collisions, which also leads to high pore-fluid pressure that reduces frictional forces. Rapid aggradation of stacked sub-units and massive flow units, depending on flow unsteadiness, contribute to infilling of the topography. The reduced surface roughness and slope angles of the valley floor, combined with the loss of lateral confinement leads to a transition to friction-dominated dense granular flows of later PDCs. This change in flow dynamics is accentuated by strong break in slope (e.g., hydraulic jumps). These PDCs deposit in a characteristic levee-channel morphology with steep bulbous Lobe front. The mass flux and particle concentration sets the initial degree of fluidization, which in turn controls the runout distance of the dense granular flows.

In the Rio Blanco Sur, the distinct levee-channel morphology of the deposits makes it possible to investigate and corroborate existing scaling laws between morphometric characteristics (shape of the front, thickness of the levees and central channel, etc.) and the slope of the topography provided for similar pumice flows (e.g., Jessop et al., 2012). These results may help constrain future numerical modelling and experiments of granular flows. For instance, the deposit
morphology and sedimentary structures were used to validate the likely presence of high pore-fluid pressure in dense granular flows as modelled by Gueugneau et al. (2017). This rheology was also used to explain the long deposit runouts for low slope angles in the Rio Frio – Rio Blanco Este.

The description of these PDC deposits with extensive field observations provide a unique and invaluable record, first because almost all of the studies of this eruption focused on the tephra fall deposits, and second because PDC deposits are incredibly fragile and these spectacular features may have disappeared already. My intent is for all of these data and observations to be archived and hopefully used as part of the larger database of worldwide records of PDC deposits.

References


CHAPTER FIVE:

REMOTE SENSING ANALYSIS OF THE 2015 PYROCLASTIC DENSITY CURRENTS (PDCs) AT COLIMA AND CALBUCO VOLCANOES

Introduction

In Chapters 3 and 4, I presented the results from field investigations of pyroclastic density current (PDC) deposits at Volcán de Colima in Mexico (Figure 5.1) and Volcán Calbuco in Chile (Figure 5.2). Moving forward, I will refer to these volcanoes more simply as Colima and Calbuco, respectively. My observations showed that the flow dynamics of PDCs during their transport and deposition are intricately linked to the shape of the receiving landscape, most specifically the morphology of the valleys. The interpretations made from Chapters 3 and 4, however, were limited to the few areas that could be accessed on the ground during field work. In this chapter, I take a step back and use remotely-sensed data from satellite imagery, namely optical, synthetic aperture radar (SAR) and thermal data, to map the overall extent of the 2015 PDC deposits at both volcanoes. Then, I compare the distribution of these deposits with the pre-eruptive topography, in order to derive relationships applicable across multiple events rather than being volcano- or event-based.

The benefits of using remote sensing tools are highlighted at active volcanoes where ground-based monitoring and field work are dangerous, and sometimes impossible, due to the remote characteristics of the location, natural hazards (eruptive activity, dense vegetation, and rugged relief) or socio-political reasons (Head et al., 2012). Optical, SAR and thermal satellite imagery can be applied for mapping volcanic deposits and structural features, which are critical to
produce hazard maps and improve forecasting of the potential impacts from future events (Hooper et al., 2012). Datasets of flow paths, volumes and hazard footprints from past PDC-forming eruptions provide critical information to better understand the complex dynamics involved and calibrate hazard assessment models (Lube et al., 2020 and references therein). However, PDC deposits are fragile and may be eroded or remobilized rapidly after emplacement, so in some cases, applying remote sensing techniques may be the only viable option for retrieval of these parameters.

High-resolution optical data have been used to map pyroclastic deposits and volcanic structures from the 2002-2003 eruption of the Semeru volcano, Indonesia (Solikhin et al., 2012) and from the 2006 dome-collapse of Merapi volcano (Thouret et al., 2010). Temporal changes in intensity and coherence signals from radar data have been used to map pyroclastic flow deposits at Soufriere Hills Volcano in Montserrat (Wadge et al., 2002, 2011), to track PDCs and lahars at Unzen volcano (Terunuma et al., 2005), to track pyroclastic flow deposits syn- and post-eruption at Merapi Volcano, Indonesia (Solikhin et al., 2015a) and to map lahar deposits at Redoubt Volcano, Alaska (McAlpin et al., 2013). Krippner et al. (2018) used a combination of optical and thermal remote sensing data from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) to analyze the lava dome-collapse and associated PDCs at Shiveluch volcano in Kamchatka. Pallister et al. (2019) applied a combination of optical, radar and thermal imagery to map the PDC deposits at Sinabung volcano (Indonesia) and used it to retrieve deposit distribution from successive lava flow front and margin collapse, volumes and eruption rates. The importance of combining multiple remote sensing methods for classification of land cover and deposit mapping can be particularly valuable during or immediately following an eruption because rain and ash cloud cover can mask terrains in optical images and reduce thermal anomalies of the terrains.
Figure 5.1. Topographic map of Calbuco with annotated valleys of interest. These valleys are the ones inside which PDC deposits were recognized and will be referenced throughout the text. The inset map shows the location of Calbuco Volcano in Chile, in South America.
Figure 5.2. Topographic map of Colima with the Montegrande and San Antonio ravines indicated. The inset map shows the location of Colima in Mexico, in Central America.
Finally, recent studies have highlighted the topographic controls on the dynamics of small-volume PDCs (i.e., < 1 km$^3$), particularly for the generation of hazardous overbank flows and ash-cloud surge detachment (e.g., Loughlin et al., 2002; Charbonnier and Gertisser, 2008, 2011; Komorowski et al., 2010, 2013; Lube et al., 2011; Charbonnier et al., 2013; Ogburn et al., 2014; Macorps et al., 2018; Marti et al., 2019). It is crucial to further our understanding of the mechanisms involved in the distribution of PDC deposits to better predict the extent of inundated areas by future PDC events. One way to look at the effects of the topography is to compare sets of topographic parameters, resolved on digital elevation models (DEMs), with the corresponding extent of PDC deposits mapped using remote sensing data.

Using the 2015 PDCs from Colima and Calbuco as case studies, I demonstrate how the combination of multiple sensors, namely optical, SAR and thermal, can be used to map PDC deposits and investigate the relationships between the distribution of deposits and the topography. The details of the PDC-forming events for Colima and Calbuco were compiled in Chapter 2, and summarized hereafter.

At Calbuco, the sub-Plinian eruption began abruptly on the evening of April 22, 2015, with a pulsatory ash-rich column, sustained for 90-min, which rose up to 15 km (height above crater) in the atmosphere and formed an umbrella cloud expanding and dispersing tephra to the NE. After a 5h30min-break, in the night of April 23, 2015, the second and more violent phase of the eruption began and generated a 6 hour-long sustained eruption plume that reached a maximum height of 28 km, with an umbrella cloud dispersing tephra again to the NE. The partial and eventually complete collapses of the two successive eruption plumes resulted in multiple PDC pulses that descended into the main drainage pathways around the volcano. The volume of PDCs was calculated from the elevation differences between pre- and post-eruption VHR DEMs (see Chapter 6), which
returned a total volume of $60.7 \pm 5.4 \times 10^6$ m$^3$. Concurrently, the melting of the summit glacier contributed to the generation of syn- and post-eruption lahars, also referred to as hot and cold lahars (see Chapters 2 and 4), that were bulking from remobilization of both PDC and tephra deposits. Such secondary lahars can be extremely devastating because they occur after the eruption has ended and can run twice the length of the flows they are eroding.

At Colima, the eruption of July 10-11, 2015, was of lower intensity than Calbuco, with a dome-collapse event that sent a series of PDC pulses on a single flank (i.e., southern flank) of the volcano into just two main ravines for a total deposit volume of $6.2 \pm 1.1 \times 10^6$ m$^3$. Although the total volume of PDCs at Colima was 10 times lower than that of Calbuco, it was the longest recorded runout distance at Colima for a dome-collapse event. This raised the necessity of updating hazard maps accordingly. The lesson from the Calbuco eruption also raises the same awareness for the need to update hazard maps: there were few to no warning signs of the high intensity eruption, and many infrastructures were completely destroyed by the PDCs, thick accumulation of tephra fall and lahars.

Datasets and Methodology

In the following section, I briefly describe the optical, SAR, and thermal remote sensing datasets and data processing methods used to map the extent of PDC deposits at Colima and Calbuco. When referring to the acquisition dates of remote sensing images, the term “eruption” (as in pre- and post-eruption) is used in reference to the respective events that generated the studied PDCs at each volcano. More details regarding the data processing are provided as supplementary materials in Appendix I.
Optical Imagery

The optical imagery dataset consisted of two sets of resolution: (1) medium-resolution (30 m) Landsat-8 imagery, and (2) very-high-resolution (VHR, 1-5 m) commercial imagery from Pleiades-1A, SPOT-6 and Worldview-1 satellites.

The Landsat-8 Operational Land Imager (OLI) Band 4 (red channel, spectral range 0.64 – 0.67 µm) and Band 5 (near-infrared NIR channel, spectral range 0.85 – 0.88 µm) were used to compute Normalized Difference Vegetation Index (NDVI) values for pre- and post-eruption scenes at Colima and Calbuco using Equation 1:

$$NDVI = \frac{\rho_{NIR} - \rho_{R}}{\rho_{NIR} + \rho_{R}}$$

where $\rho_{NIR}$ and $\rho_{R}$ are the surface reflectance values for NIR band 5 and red band 4, respectively.

In order to perform radiometric calibration and atmospheric correction, and obtain surface reflectance, each Landsat-8 scene was processed with the Atmospheric and Radiometric Correction of Satellite Imagery (ARCSI) Software (Bunting and Clewley, 2018). The ARCSI processing steps are detailed in Appendix I. Radiometric calibration and atmospheric correction are required when working with multi-dates images, to remove differences in land surface and atmospheric conditions, as well as solar position and sensor geometry during the different image acquisitions (Lu et al., 2004; Pu et al., 2017).

The NDVI is comparable to a greenness index of the land surface and provides information regarding land cover/land use (LULC) types (e.g., forest, farmland, bare ground). When applied in a time-series, the variation in NDVI values allows detection of LULC changes through time, including the deposition of volcanic deposits and vegetation changes. However, seasonal changes of vegetation (i.e., plant phenological stages) can cause changes in NDVI values that are independent of LULC changes. Therefore, it is important to further normalize the NDVI images
in order to remove non-LULC changes for a reliable time-series analysis. I used a method described by Pu et al. (2017), which is based on a linear regression of NDVI values for polygons of no-LULC changes. The difference with the work of Pu et al. (2017) is that radiometric and atmospheric normalization of the OLI bands was performed before calculating the NDVI images, which were in turn submitted to the normalization procedure with the linear regression method.

The concept behind this normalization method is that the NDVI values of unchanged LULC areas for a given NDVI image acquired at one time, are a linear function of the values of the same areas in an NDVI image from another time. The slope and intercept coefficients of the linear regression are then used to calculate a predicted NDVI image based on an earlier one, had there not been any surface changes. Finally, the NDVI image differencing and ratio are done between the predicted NDVI and its corresponding actual NDVI image. This allows to reduce noise in NDVI changes caused by plant phenological changes and in this particular case, to focus on LULC changes caused by an eruption and deposition of volcanic products. Details of the normalization procedure with linear regression are provided in Appendix I, as well as the list of NDVI image pairs analyzed (Table I.1). The workflow for processing the Landsat-8 OLI bands and producing time-series of normalized NDVI images is presented in Figure 5.3.

The VHR datasets for Colima and Calbuco are summarized in Appendix I (Table I.2). These VHR images were used in Chapters 3 and 4 to match deposit features observed in the field, thus allowing further comparison and validation of the features observed via other remote sensing sensors, which are being presented throughout this chapter. These same datasets were also used to generate VHR DEMs for pre-eruption topography analysis (see Channel Morphology Analysis section), and for retrieval of deposit volumes (see Chapter 6).
Dual-Polarization Synthetic Aperture Radar (SAR) Data

ALOS-2 PALSAR dataset

I used SAR data that were acquired by the Phased Array L-band Synthetic Aperture Radar (PALSAR) on board the Japanese Advanced Land Observing Satellite 2 (ALOS-2). The dataset consisted of images acquired in both ascending and descending orbits for Colima, and only in ascending orbit for Calbuco (see Appendix I, Table I.3). The data were acquired using Fine Beam Dual-Polarization, in right-looking geometry and with an incidence angle of about 40.55 degrees and 31.55 degrees for Colima and Calbuco data respectively. The dual-polarization included ‘horizontally polarized wave transmission – horizontally polarized wave reception’ (HH) and ‘horizontally polarized wave transmission – vertically polarized wave reception’ (HV) modes. The HH images were used to generate temporal coherence images, and both HH and HV images were used to look at changes in SAR intensity.

Data Processing

The processing of ALOS-2 data was performed with the GAMMA software package (Werner and Strozzi, 2000) and is summarized in Figure 5.3. The Level 1.1 data were used, which are Single Look Azimuth compressed (SLC) data that preserve the intensity and phase information. Absolute radiometric calibration of the SLC images was achieved using the external calibration coefficient of -115 dB (Werner et al., 2000). Pairs of SLC images with the same orbits (ascending or descending) were used to generate interferograms (Appendix I, Table I.3). For each pair, the SLC images were co-registered, and the intensity images were multi-looked using four looks in azimuth and two looks in range, for a pixel spacing of 13.2 m in slant range geometry. The coherence images were generated as part of the standard co-polarization interferometric processing with the same window size as multi-looking. The intensity and coherence images were geocoded
using the recently released NASADEM (2020) at 30-m resolution. Finally, the intensity images were converted from linear values \(pwr\) to decibels \(dB\) using Equation 2, with \(\pm 0.5\) dB radiometric accuracy (Motohka et al., 2018):

\[
dB = 10 \times \log_{10}(pwr)
\]

(2)

**Thermal Imagery**

**Landsat-8 Thermal Infrared Sensor dataset**

The Thermal Infrared Sensor (TIRS) onboard the Landsat-8 satellite was used to retrieve Land Surface Temperatures (LST). This sensor comprises two TIRS bands, namely Band 10 (spectral range 10.6 – 11.19 \(\mu m\)) and Band 11 (spectral range 11.5 – 12.5 \(\mu m\)) at 100-m resolution. Note, however, that only TIRS band 10 was used for LST retrieval, following the January 6, 2014, recommendation from the USGS (Montanaro et al., 2014) because of larger calibration uncertainties for band 11.

**Data processing**

For each Landsat-8 scene, each TIRS band 10 was preprocessed to convert digital number values to at-sensor spectral radiance values (see Appendix I, Table I.4 conversion coefficients). Then, the thermal at-sensor radiance was used in the Radiative Transfer Equation (RTE) method, as described in Yu et al. (2014), with inversion and simplification of Planck’s Law such that:

\[
T_s = \frac{K_2}{ln \left[ \frac{K_1}{L_\lambda - L^{\downarrow}_\lambda - \tau (1 - \varepsilon_\lambda)L^{\uparrow}_\lambda} + 1 \right]}
\]

(3)

where \(T_s\) is the LST in Kelvin (K), \(L_\lambda\) is the at-sensor radiance (W.m\(^{-2}\).sr\(^{-1}\).\(\mu m\)\(^{-1}\)), \(\varepsilon_\lambda\) is the Land Surface Emissivity (LSE), \(\tau\) is the atmospheric transmittance, \(L^{\downarrow}_\lambda\) and \(L^{\uparrow}_\lambda\) are the downwelling and
upwelling atmospheric radiances respectively (W.m⁻².sr⁻¹.µm⁻¹), and \(K_1\) and \(K_2\) are the calibration constants for Landsat-8 TIRS band 10 provided in Appendix I (Table I.4).

**Figure 5.3.** Workflow process for generating NDVI Ratio images and for Land Surface Temperature (LST) retrieval using Landsat-8 data. This is a detailed workflow of the pre-processing of Landsat-8 bands 4, 5 and 10, through the ARCSI software (steps in grey circles). Further processing of these bands was used to create NDVI images, which were then normalized and analyzed with temporal image ratios for NDVI changes through time. The NDVI images were also used in the calculations of proportion of vegetation \(P_V\) and land surface emissivity (LSE) to retrieve the LST images as final products. See text for details.

The atmospheric parameters \(\tau, L_{\Delta}^\uparrow\) and \(L_{\Delta}^\downarrow\) were obtained using the Atmospheric Correction Parameter Calculator developed by Julia Barsi at the NASA Goddard Space and Flight Center (https://atmcorr.gsfc.nasa.gov/; Barsi et al., 2005, 2003). The values of \(\tau, L_{\Delta}^\uparrow\) and \(L_{\Delta}^\downarrow\) for each Landsat-8 scene are summarized in Appendix I (Tables I.5 and I.6).
The LSE (ε, in Equation 4) was calculated using the method from Sobrino et al. (2008), which is based on the NDVI (Equation 1) and on the proportion of vegetation, \( P_V \) (Carlson and Ripley, 1997) calculated with Equation 4:

\[
P_V = \left\{ \frac{\text{NDVI} - \text{NDVI}_{\text{min}}}{\text{NDVI}_{\text{max}} - \text{NDVI}_{\text{min}}} \right\}^2
\]

in which \( \text{NDVI}_{\text{min}} \) and \( \text{NDVI}_{\text{max}} \) represent minimum and maximum NDVI, obtained from the histogram of the NDVI image (refer to Optical Imagery section for generation of Landsat-8 NDVI images). The complete workflow for processing Landsat-8 images for LST retrieval is summarized in Figure 5.3.

**Channel Morphology Analysis**

Channel morphology analysis was performed on the pre-eruption topography at each volcano using their respective VHR DEM. For Colima, the pre-eruption Pleiades-1A stereo pair (Appendix I, Table I.1) was utilized for the generation of a DEM at 1-m resolution using the OrthoEngine module in PCI Geomatica® software, with semi-global matching (SGM) algorithm (Hirschmuller, 2005). For Calbuco, the pre-eruption Worldview-1 stereo pair at Calbuco (Appendix I, Table I.1) was used to generate a 4-m resolution DEM using the stereogrammetry routine from the NASA Ames Stereo Pipeline (ASP) v. 2.5.1 (Shean et al., 2016).

The measurements for channel morphology were performed using a GIS-based Python algorithm I developed for the open-source QGIS software (the link to my Github repository can be found in Appendix II). The inputs required to run this algorithm are (1) the pre-eruption DEM in raster format, and (2) individual vector shapefiles of the right bank(s) and left bank(s) of the channel(s) of interest. For a given channel, the algorithm generates perpendicular transects at a constant interval along the channel. In this analysis, I tested intervals at 50-m and 100-m to reduce
noise in contiguous measurements. Next, vertices are created at each intersecting points of the transects with the left and right channel banks. These vertices are then used to create individual polygons that dissect the channel at those same regular intervals. Finally, the algorithm creates the channel centerline for the entirety of the channel length, which is dissected in 50 m and 100 m-long segments.

The elevation values are extracted from the DEM along each transect to retrieve the cross-sectional area of the channel (in m²) every 50 and 100 m, and along the channel centerline to compute the longitudinal slope of the channel floor every 50 and 100 m as well. The 50 and 100-m segments of the centerline are also used to compute channel sinuosity, which I chose to quantify in terms of changes in azimuth direction of each segment. Lastly, the polygon vertices are used to interpolate a surface at the top of the valley, allowing the computation of the volumetric channel capacity (in m³) for each polygon. The results are stored into comma separated value (CSV) files to track changes of each channel morphology parameter (i.e., cross-sectional area, volumetric capacity, slope and sinuosity) with distance along the channel.

The results are displayed in QGIS, using threshold values for percentage change (% change) in volumetric channel capacity and channel sinuosity to modify the color of the polygons and centerline segments, respectively, with classic color scheme (i.e., green: 0 – 50 % change, orange: 50 – 100 % change, red: > 100 % change). The choice for the % change threshold was based on prior findings described in Chapters 3 and 4, where the greater potential for hazardous overbank flows was associated with a decreased channel capacity and a change in channel direction > 100 %. The classic color scheme follows the conventional color code for an increasing range of hazard levels.
Results

Note that for each result section, I performed analysis for the entirety of the datasets described in the *Datasets and Methodology* section but decided on sharing images that yielded the most explicit results for the purpose of clarity and in the interest of brevity.

**Optical Images**

First, I examine the changes between pre- and post-eruption NDVI images at Colima (Figure 5.4a and 5.4b). Low NDVI typically corresponds to bare soil, as opposed to high NDVI for vegetated surfaces. The Montegrande ravine presents a decreased NDVI caused by the deposition of PDCs. Near the summit, however, the absence of vegetation before the eruption complicates the differentiation between new and old volcanic deposits. The absolute NDVI ratio of the pre- and post-eruption NDVI images, hereafter referred to as $|Q_{NDVI}|$, in Figure 5.4c is used to amplify the surface changes between the two image acquisitions. The PDC deposits inside the Montegrande ravine have low $|Q_{NDVI}| \leq 0.2$, while the unaffected surfaces have $|Q_{NDVI}|$ values close to 1. Therefore, the $|Q_{NDVI}|$ image has been reclassified using that same threshold value of 0.2 to better outline the extent of the PDC deposits (Figure 5.4d).

Similar results are obtained for Calbuco, where $|Q_{NDVI}|$ between the pre- and post-eruption Landsat images (Figure 5.5) emphasizes the area of deposition for the PDC and tephra fall deposits. The NE directionality of the eruption is highlighted with the decreased NDVI in that direction (Figure 5.5b-c). The increased NDVI around the summit crater is attributed to the increase in snow cover between the two image acquisitions. The decreased NDVI in the major valleys on the S, W, N and NE flanks is also associated with deposition of new materials, although it does not allow differentiation between deposits from PDCs, tephra fall and lahars.
Figure 5.4. Post-eruption and predicted pre-eruption Landsat-8 NDVI images of the southern flank of Colima. **a)** Actual post-eruption NDVI image from January 10, 2016. **b)** Predicted NDVI image from March 12, 2015. High NDVI values represent vegetated surfaces and contrast with bare ground and/or volcanic deposits represented with low NDVI values. **c)** Absolute NDVI ratio image obtained by dividing the actual post-eruption NDVI image by the predicted NDVI image. **d)** Reclassified NDVI ratio image with threshold values that emphasize the deposition of new volcanic materials.
Figure 5.5. Post-eruption and predicted pre-eruption Landsat-8 NDVI images of Calbuco. **a)** Post-eruption NDVI image from October 11, 2015. **b)** Predicted NDVI image from February 13, 2015. **c)** Absolute NDVI ratio image obtained by dividing the actual post-eruption NDVI image by the predicted NDVI image. **d)** Reclassified NDVI ratio image with threshold values that emphasize the deposition of new volcanic materials and lahars.

The post-eruption NDVI images have also been used to look at post-eruption changes in the vegetation patterns. Ash-cloud surge deposits are thinner and contain smaller particle sizes, meaning they can be eroded and washed away faster than the thick and coarse concentrated PDC
deposits. Moreover, the vegetation affected by ash-cloud surges is often characterized by burnt foliage, and sometimes leafless tree limbs that remain standing, thus contrasting with zones of felled trees from concentrated PDCs (see Chapter 4). The burnt vegetation will rapidly regenerate, and the erosion of thin surge deposits will allow for faster vegetation regrowth. Therefore, post-eruption increase in NDVI can be used as a proxy for differentiating ash-cloud surge deposits from concentrated PDC deposits.

**Figure 5.6.** Post-eruption temporal changes of NDVI at Colima. The NDVI images are zoomed-in over the most distal part of the Montegrande ravine. a) Post-eruption NDVI image from July 18, 2015, (i.e., a week after the eruption). b) Post-eruption NDVI image from January 10, 2016. c) Absolute NDVI ratio image obtained from the later NDVI image divided by the earlier NDVI image. The bright pixels indicate increasing NDVI values whereas dark pixels indicate a decrease over time. Red polygons indicate areas where the increased NDVI values represent vegetation regrowth where ash-cloud surge deposits were previously emplaced along the concentrated PDCs (with low NDVI values).

At Colima, the post-eruption increase in $|Q_{NDVI}|$ along the edges of the Montegrande ravine highlights vegetation regrowth, thus suggesting areas previously covered by ash-cloud surge deposits (Figure 5.6). The changes in post-eruption vegetation pattern of areas subjected to ash-cloud surges compared to those covered by thick concentrated PDC deposits are further
highlighted on VHR images in the proximal area of the Montegrande ravine (Figure 5.7). The
singed vegetation outlined on the VHR image acquired one week after the dome-collapse eruption
appears healthy in the VHR image acquired 6 months later, and with visible deposit erosion.

![Figure 5.7](image)

**Figure 5.7.** Post-eruption temporal changes of vegetation pattern at Colima. VHR optical imagery
from SPOT-6 at 2-m resolution acquired on 2015-07-25 (left) and Pleiades-1A at 0.5 m resolution
acquired on 2016-01-10 (right) in the Montegrande ravine at Volcán de Colima, at the junction
with the San Antonio ravine. The images show the extent of the ash-cloud surge with singed
vegetation in brown on the left and the change back to healthy vegetation on the right.

At Calbuco, the post-eruption increase in $|Q_{NDVI}|$ on the NE flank and in the valleys on
Figure 5.8 contrasts with the decrease in $|Q_{NDVI}|$ on Figure 5.5, indicating deposit erosion and/or
vegetation growth. However, it is difficult to determine if these areas correspond to ash-cloud
surge or tephra fall deposits, which blanketed most of the N-NE sector. Furthermore, as previously
stated, lahars and concentrated PDCs could not be differentiated using $|Q_{NDVI}|$, and the post-
eruption increase in $|Q_{NDVI}|$ may indicate erosion of lahar deposits rather than vegetation regrowth through ash-cloud surge deposits. During field investigations of the PDC deposits in the Rio Frio – Rio Blanco Este valley complex and in the Rio Blanco Sur, I observed distinct lobate and levee/channel surface morphologies of concentrated PDC deposits (see Chapter 4). Therefore, the presence of these surface features observed on VHR optical images can be used to identify concentrated PDC deposits where they could not be accessed in the field (Figure 5.9).

**Figure 5.8.** Post-eruption temporal changes of NDVI at Calbuco. a) Post-eruption NDVI image from January 31, 2016. b) Post-eruption NDVI image from March 19, 2016. c) Absolute NDVI ratio image obtained from the later NDVI image divided by the earlier NDVI image. The bright pixels indicate increasing NDVI values whereas dark pixels indicate a decrease over time. Red polygons indicate areas where the increased NDVI values represent vegetation regrowth where ash-cloud surge and/or tephra fall deposits were previously emplaced, as well as deposit erosion in the valleys.

It is particularly useful for differentiating PDC deposits from lahar deposits on the S flank, where the pre- to post-eruption $|Q_{NDVI}|$ is < 0.2 (i.e., threshold for newly deposited materials) as shown on Figure 5.5. Furthermore, field observations in the Rio Blanco Sur showed the maximum PDC deposit runouts at ~ 4 km from the summit (straight line method) while the lahars reached the Chapo Lake 11 km from the summit (see Chapters 2 and 4).
Figure 5.9. Lobate surface morphology of PDCs at Calbuco from VHR image. a) Rio Sur – N branch, b) Rio Sur – S branch, c) Rio Blanco Sur, d) channel within the 1961 lava flow, e) and f) Rio Frio – Rio Blanco Este complex. The presence of these deposit morphologies was used to infer and map the presence of concentrated PDC deposits as opposed to ash-cloud surge deposits or lahar deposits.

Therefore, the post-eruption increase in $|Q_{NDVI}|$ along the Rio Blanco Sur starting at ~ 4 km from the crater (Figure 5.8c) indicates vegetation regrowth through areas previously impacted by lahars only. Hayes et al. (2019) described the presence of lahars in the Rio Colorado, Rio Este and Rio Amarillo valleys, on the S flank, east of the Rio Blanco Sur, that also reached Chapo Lake, consistent with pre- to post-eruption decrease in $|Q_{NDVI}|$ on Figure 5.5. The post-eruption increase in $|Q_{NDVI}|$ (Figure 5.8) combined with the absence of evidence for concentrated PDC deposits on VHR images (Figure 5.9), suggest that these valleys almost strictly experienced lahars. One can
argue, however, that ash-cloud surges and/or tephra fall could have been deposited in the most proximal areas of these valleys and later eroded by lahars.

**Coherence Images**

Surface changes occurring between two SAR acquisitions induce a temporal decorrelation, such that areas subjected to these changes appear as dark (low coherence) in the coherence image. Decorrelation may also be associated with changes in the vegetation, either by disappearance during the deposition of PDCs (both dilute and concentrated) or following re-growth over areas affected by dilute PDCs or tephra fall deposits.

At Calbuco, the pre-eruption SAR image pair in Figure 5.10a shows relatively high coherence on the volcano’s flanks indicative of stable surfaces, and low coherence at the summit, which may be the result of changes in snow cover. In Figure 5.10b, strong decorrelation at the summit and on the NE flank occurs between the pre- and post-eruption SAR coherence images. The low coherence signal coincides with the general direction of the ash dispersion and the main volcanic flows towards the NE, as observed with VHR optical images and with the NDVI changes in the previous section. The valleys on the SW and W flanks also display decorrelation on Figure 5.10b, which is in agreement with the presence of PDCs interpreted from NDVI changes and confirmed on VHR optical images (Figures 5.5 and 5.9c). Figures 5.10c and 5.10d display the coherence signal for post-eruption SAR image pairs, and show decorrelation focused around the summit crater and along the Rio Frio – Rio Blanco Este and Rio Tepu valleys on the NE flank.
Figure 5.10. Interferometric coherence images at Calbuco. Light = high coherence; dark = low coherence. a) Pre-eruption coherence image between February 4 and March 4, 2015; b) syn-eruption coherence image between March 4 and April 29, 2015; c) post-eruption coherence image between April 29 and July 8, 2015; d) post-eruption coherence between July 8 and November 25, 2015. For each coherence image, the pair name is written as 'primary image'–'secondary image', both in YYYYMMDD format. The coherence images were obtained for direct polarization HH only.
Figure 5.11. Interferometric coherence images at Colima prior to the dome-collapse a) between September 11, 2014, and January 29, 2015; b) between February 25 and June 8, 2015. For each coherence image, the pair name is written as ‘primary image’–‘secondary image’, both in YYYYMMDD format. The coherence images were obtained for direct polarization HH only. The yellow lines delineate the decorrelated areas where volcanic products were deposited.

Figure 5.12. Interferometric coherence images at Colima post dome-collapse a) between April 11 and June 6, 2016; b) between January 28, 2016, and January 26, 2017; c) between April 10 and June 4, 2017. For each coherence image, the pair name is written as ‘primary image’–‘secondary image’, both in YYYYMMDD format. The coherence images were obtained for direct polarization HH only. The yellow lines delineate the extent of the lava flows, while the green line delineate the decorrelation induced by PDC deposits during the January 2017 eruption.
The loss of coherence at the summit is attributed to the return of the summit snow and glaciers during winter months (July, south hemisphere), similar to the low coherence values around the crater on Figure 5.10a. The low coherence values in the Rio Frio – Rio Blanco Este and Rio Tepu valleys are attributed to the continuing lahars that ran down and eroded the PDC deposits in the valley over the course of a year. The field campaign carried out in December 2016 confirmed the presence of lahar deposits and the strong erosion that occurred in these valleys (see Chapter 4). In contrast, the high coherence values in the Rio Blanco Sur are attributed to the limited erosion sustained by the lobate scoriaceous PDC deposits, also confirmed by field work (see Chapter 4).

At Colima, the low coherence values on the volcano’s flanks are caused by the significant changes in vegetation because of the longer temporal baseline for the image pairs (Figures 5.11 and 5.12). This interpretation is supported by the areas of high coherence values on the upper slopes of the volcano where vegetation does not grow. In Figure 5.11a, the decorrelation at the summit near the dome/crater can be attributed to volcanic activity, i.e., rockfalls, lava flows and small PDCs that occurred during this time period. In Figure 5.11b, the loss of coherence at the summit coincides with the lava flows that began from the lava dome overspilling the crater confinement. The lava flow descended onto the S flank during the month of June, reaching ~700 m by early July (GVP report, 2016; Chapter 2). Figure 5.12 shows post-eruption coherence images, with the July 2015 PDC deposits showing high coherence that suggests relatively stable deposit surface. In Figure 5.12a, the decorrelation near the summit is caused by the growing lava flow that was progressing onto the southern flank. The further progression of the lava flow onto the southern flank is outlined on Figure 5.12b, with a wider area of low coherence. In Figure 5.12c, the area near the junction between the Montegrande and San Antonio ravines shows lower coherence values in comparison to the high coherence inside the Montegrande ravine. This area corresponds
to the one where the optical images previously indicated the presence of ash-cloud surge deposits (Figure 5.7), thus allowing me to conclude that the lower coherence values are caused by vegetation regrowth and/or deposit erosion, contrasting with the stable surface of the valley-confined PDC deposits.

Intensity Images

The intensity of the return signal in radar imagery is influenced by the local terrain, including local slope, surface roughness and terrain composition. A smoother surface will produce a weaker backscatter intensity (given the same local slope), thus resulting in a weaker signal (darker pixels), and vice versa. The water content and surface temperature will also result in different backscatter intensity, particularly for crop fields and growing vegetation. While some ground features appear similar in HH and HV images, rough or inhomogeneous surfaces can appear brighter on HV than on HH because of the stronger depolarization that occurs over vegetation.

Volcán de Colima

Figure 5.13 shows georeferenced SAR intensity images of Colima acquired from ascending passes before and after the dome-collapse eruption with HV polarization. The areas covered by the 2015 PDC deposits are characterized by low intensity (darker pixels) along the Montegrande ravine, and contrast with high intensity (brighter pixels) from the vegetation on the volcano’s flank (Figure 5.13b-c). The infilling of the two ravines by PDCs contributes to a smoother surface, thus reducing the intensity signal (comparison between Figure 5.13a and 5.13b-c). Figures 5.13d through 5.13f are close-ups of the summit area delineated with a blue rectangle in Figure 5.13a. The lava flow that began in 2014 from the overflowing lava dome onto the S flank can be tracked.
through the series of intensity images (yellow dashed line, Figures 5.13d-f). The lava flow causes an increased surface roughness detected through higher backscatter intensity. The post-dome-collapse depression inside the summit crater is also visible on Figures 5.13e-f.

**Figure 5.13.** Intensity images at Colima, pre- and post-eruption. Top row: Georeferenced ALOS-2 intensity images from ascending passes in HV polarization acquired at 1 year intervals a) before the July 2015 PDCs on June 8, 2015, b) 1 year later on June 6, 2016 and c) 2 years later on June 5, 2017. Bottom row: Georeferenced intensity images zoomed-in on the south flank of the summit area (blue rectangle in image a) to show the progression of the lava flow (yellow dashed lines). d) HH intensity image from June 8, 2015; e) HV intensity image from January 28, 2017; f) HV intensity image from June 5, 2017.
Figure 5.14 shows georeferenced false-color composite (Red, Blue, Green band combinations) SAR intensity images for Colima. The band combination is obtained from pairs of intensity images to better illustrate the surface changes with time, where the red channel is the earlier intensity image, the blue channel is the later intensity image, and the green channel is the ratio of later and earlier intensity images. This band combination provides a legend, in which the red colored pixels indicate signal loss from increased surface smoothness, whereas the blue color indicates signal gain from increased surface roughness, and yellow is for stable surfaces.

In Figure 5.14a, the flanks of Colima pre-eruption are characterized by yellow-colored pixels, thus showing relatively stable surfaces. The summit area in Figure 5.14a shows an increase in surface roughness (light blue color) caused by the sparse deposition of volcanic materials from rockfalls, ash fall, small PDCs and lava flows during the eruptive activity of September to December 2014 (see Chapter 2 for details of these events).

Figures 5.14b and 5.14c contrast with Figure 5.14a and show an increase in surface smoothness in the Montegrande and San Antonio ravine on the south flank (red pixels), caused by the infilling of these valleys by the 2015 PDC deposits. Brighter red pixels, i.e., milder intensity signal loss, are also observed on the interfluvies of the Montegrande ravine, specifically towards the more proximal area and at the junction with the San Antonio ravine. These areas correspond to ash-cloud surge and overbank flow deposits according to field observations presented in Chapter 3, and the VHR optical imagery in the previous section. The intensity change can be related to the partial vegetation stripping and to the thinner, yet ash-rich deposits to explain a smaller increase in surface smoothness in comparison to thick PDC deposits in the valleys.

Near the summit to the south of the crater, the area of increased roughness (blue) indicates the presence of the post-dome collapse lava flow (Figure 5.14b-c). The wider area covered by the
lava flow in Figure 5.14c is related to the longer timespan (i.e., nearly 2 years) between the pair of intensity images used to generate the RGB image.

Figure 5.15 shows the false-color composite SAR intensity images using HH polarization for two image pairs acquired post-eruption. The areas of increased surface roughness (light blue colored pixels) in Figures 5.15a and on the close-up in Figure 5.15c, are interpreted as recolonization by the vegetation, after erosion of deposits or simply by regrowth of foliage from stripped and singed trees. These observations support the interpretations for ash-cloud surges in these areas, as described with VHR optical images and NDVI changes in the previous section. In Figures 5.15b and 5.15d, the relatively stable surfaces (yellow pixels) for those same areas suggest the lack of change in the vegetation pattern after regrowth.

![Figure 5.14. Georeferenced ALOS-2 false-color composite intensity ratio images of the south flank of Colima. The false-color composite (R: Earlier intensity image, G: Later intensity image, B: Result of the later intensity image divided by the earlier intensity image) images were obtained using a) a pair intensity images from before the July 2015 PDCs on September 11, 2014 and January 29, 2015; b) a pair of intensity images from before (on January 29, 2015) and after the July 2015 PDCs on January 28, 2016; and c) a pair of intensity images from before (on May 25, 2015) and after the July 2015 PDCs on April 10, 2017. All image pairs are in HV polarization. In images b) and c) the July 2015 PDCs appear in red while the lava flow appears in light blue.](image)
Figure 5.15. Georeferenced ALOS-2 false-color composite intensity ratio images post-eruption from vegetation change at Colima. The false-color composite images were generated with the same RGB combination as mentioned in the caption of Figure 5.14 and were obtained using pairs of intensity images acquired after the July 2015 PDCs in HH polarization. The changes in intensity signal are used to demonstrate the changes in surface vegetation a) for a 2 month-period between April 10, 2016, and June 6, 2016; b) for a 10 month-period between June 6, 2016, and April 10, 2017. Images c) and d) are enlargements of the areas delineated by the black rectangles on images a) and b) respectively.
Figure 5.16. Georeferenced ALOS-2 intensity changes of the south flank of Colima. The changes in intensity values are expressed in decibels (dB). The intensity difference maps were obtained using the same pair of intensity images from before (May 25, 2015) and after (April 10, 2017) the July 2015 PDCs comparing a) the intensity changes in HH polarization with b) the intensity changes in HV polarization. Brown pixels indicate no-data values.

Figure 5.16 shows the changes in the intensity values (hereafter referred to as $\Delta dB$) at Colima using the difference between a SAR intensity image from 2017 (post-eruption) and an intensity image from 2015 (pre-eruption), both with HH polarization (left) and HV polarization (right). Similar to the results shown in Figure 5.14, the $\Delta dB < 0$ (dark pixels) along the Montegrande and San Antonio ravines indicates increased smoothness caused by the PDC deposits. The $\Delta dB > 0$ (brighter pixels) incising through the PDC deposits is due to the remobilization of deposits by post-eruption lahars and runoff. The reworked PDCs and their
erosion generate increased surface roughness, which is better observed with HH polarization (Figure 5.16a), while the HV polarization better captures the changes caused by PDC deposits (Figure 5.16b). The greater magnitude of ΔdB near the summit is the result of increased surface roughness from the emplacement of the post dome-collapse lava flow, which is observed with both HH and HV polarizations.

In order to discuss the changes in a quantitative manner, I look at the average ΔdB for each feature discussed in the previous paragraphs, in both HH and HV polarizations (Table 5.1). HV polarization is more effective in displaying intensity evolution caused by the presence of volcanic deposits, while HH polarization is best suited for changes in the vegetation pattern and erosion. The ΔdB for volcanic deposits, either PDC or lava flow, is about two times higher in HV than in HH polarization.

**Table 5.1.** Average change in SAR intensity for individual features at Colima, in co- and cross-polarization. For a given feature, the column ‘image pair acquisition’ indicate if the intensity change occurred between a pre- and a post-eruption image, or only between post-eruption images. The average intensity changes are measured in decibels (dB)

<table>
<thead>
<tr>
<th>Features</th>
<th>Image pair acquisition</th>
<th>Polarization</th>
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<tbody>
<tr>
<td></td>
<td></td>
<td>HH</td>
</tr>
<tr>
<td>PDC deposits</td>
<td>Pre- and post-eruption</td>
<td>↓ -4.2 ± 3.6 dB</td>
</tr>
<tr>
<td>Ash-cloud surge deposits</td>
<td>Pre- and post-eruption</td>
<td>↓ -1.2 ± 3.1 dB</td>
</tr>
<tr>
<td>Lava flows</td>
<td>Pre- and post-eruption</td>
<td>↑ +3.3 ± 4.6 dB</td>
</tr>
<tr>
<td>Remobilized PDCs</td>
<td>Pre- and post-eruption</td>
<td>↑ +2.5 ± 5.1 dB</td>
</tr>
<tr>
<td>Vegetation regrowth</td>
<td>Post-eruption only</td>
<td>↑ +3.5 ± 2.6 dB</td>
</tr>
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**Calbuco Volcano**

Figure 5.17 shows georeferenced SAR intensity images for Calbuco with HV polarization acquired before the sub-Plinian eruption (Figure 5.17a), and during the third and smaller eruption...
Phase, 6 days after the paroxysmal eruption phases (Figure 5.17b). The geometric effects from the satellite track and incidence angle lead to one illuminated side and one shadow side for steep slopes. Similar to the interpretations made from the intensity images at Colima, the areas affected by PDCs, tephra fall and lahars (purple dashed lines) are characterized by low intensity values (dark pixels) in the syn-eruption intensity image, contrasting with the surrounding unaffected areas (bright pixels).

**Figure 5.17.** Georeferenced intensity images from ascending passes at Calbuco in HV polarization acquired a) pre-eruption and b) during the third eruption phase, but post-paroxysmal eruption phases. The yellow dashed line highlights the low intensity values of the paddy fields, the blue dashed line emphasized the changes in the summit area with the formation of a crater, and the pink dashed lines roughly contour the low intensity values from PDC and tephra fall deposits that contrast with the pre-eruption intensity image.
The summit depression area (blue dashed line) shows the formation of a crater at its center during the eruption in Figure 5.17b, which contrasts with the relatively smooth pre-eruption dome in Figure 5.17a. The paddy fields to the north of the volcano (yellow bounding-box) are also characterized by low intensity signals both before and after the eruption, which can be attributed to higher moisture content, as well as smoother surfaces than the surrounding vegetation on the volcano’s flanks. Similarly, the low intensity near the summit area in the pre-eruption intensity image (Figure 5.17a) contrasts with the vegetated slopes and may be caused by smoother surfaces and high moisture content from ice and snow.

**Figure 5.18.** Georeferenced intensity changes with time at Calbuco. The changes in intensity values are expressed in decibels (dB). a) Intensity difference calculated between a pre-eruption (2015-03-04) and a syn-eruption (2015-04-29) intensity image with HV polarization. b) Intensity difference calculated between a syn-eruption (2015-04-29) and a post-eruption (2015-11-25) intensity image with HH polarization. Brown pixels indicate no-data values.
Figure 5.18a shows the $\Delta dB$ obtained by differencing the same pre- and syn-eruption intensity images as in Figure 5.17. The infilling of the valleys by PDCs, and the blanketed topography by thin ash-cloud surge deposits and tephra fall deposits are characterized by $\Delta dB < 0$ that is consistent with increased surface smoothness. I note the greater signal loss for valley-confined PDCs contrasting with the open slopes covered with surge and/or tephra fall deposits. Although syn-eruption lahars were recorded in the N, NE and SW valleys (see Chapters 2 and 4), the SAR intensity images do not allow for differentiation between PDC and lahar deposits. The $\Delta dB > 0$ in the crater/summit area can be explained by increased roughness from crater excavation and by the removal of snow/ice cover. On the SW side, however, the unusually high $\Delta dB$ (i.e., up to $+20$ dB) cannot be explained solely by changes in snow cover. The comparison with VHR optical images shows the accumulation of PDC and tephra fall deposit in the summit depression, and changes in elevation values obtained from VHR DEMs (see Chapter 6), show increased elevations averaging $+30.3 \pm 10.3$ m. This suggests that the increased intensity in that area is a result of geometric changes in the direction of the radar antenna receiving a stronger return signal. Figure 5.18b shows the post-eruption $\Delta dB$ for a period of ~7 months.

In the valleys, the $\Delta dB > 0$ contrasts with the negative values observed in the same areas in Figure 5.17a and is therefore interpreted as increased surface roughness caused by lahar erosion and remobilization of the PDC deposits. The areas interpreted as surge and tephra fall deposits from Figure 5.17a also display $\Delta dB > 0$ but of lower magnitude than the valleys, which can be interpreted as increased surface roughness from recolonization by the vegetation, similar to what was observed at Colima. The $\Delta dB < 0$ near the summit may be the result of changes in snow cover, where snow and ice accumulated in the months following the eruption. For each of the features discussed above, the average $\Delta dB$ in both HH and HV polarizations are summarized in Table 5.2.
Table 5.2. Average change in SAR intensity for individual features at Calbuco, in co- and cross-polarization. For a given feature, the column ‘image pair acquisition’ indicate if the intensity change occurred between a pre- and a post-eruption image, or only between post-eruption images. The intensity changes are measured in decibels (dB).

<table>
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<td></td>
<td>HH</td>
</tr>
<tr>
<td>PDC deposits</td>
<td>Pre- and post-eruption</td>
<td>↓-8.0 ± 3.3 dB</td>
</tr>
<tr>
<td>Ash-cloud surge and/or tephra fall deposits</td>
<td>Pre- and post-eruption</td>
<td>↓-3.1 ± 3.3 dB</td>
</tr>
<tr>
<td>Summit changes</td>
<td>Pre- and post-eruption</td>
<td>↑+6.8 ± 6.4 dB</td>
</tr>
<tr>
<td>Snow cover change</td>
<td>Post-eruption only</td>
<td>↓-4.3 ± 4.5 dB</td>
</tr>
<tr>
<td>Lahar erosion</td>
<td>Post-eruption only</td>
<td>↑+7.7 ± 4.0 dB</td>
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Figures 5.19 to 5.21 show georeferenced false-color composite images of intensity changes using the same band combination as described for SAR intensity imagery at Colima in the previous section. In the pre-eruption RGB intensity image (Figure 5.19a), most of the area surrounding the volcano is characterized by relatively stable surfaces displayed in yellow-colored pixels. The paddy fields to the north and the summit area appear in dark blue, interpreted as changes in soil moisture and changes in snow cover respectively. On the NE and SE flank, the 1961 lava flow appears in light red color (pink dashed lines in Figure 5.19a), which suggests a smoother surface and/or changes in vegetation pattern on its surface.

In Figure 5.19b, the valleys infilled by PDC and lahar deposits, and the tephra fall and surge deposits on the proximal part of the NE flank appear in dark red (pink dashed lines) from increased surface smoothness. The crater area appears in bright blue, most likely from geometric changes caused by the eruption as discussed previously with Figure 5.18a.

In Figure 5.20a, the zoom over the Rio Blanco Sur shows the scoriaceous lobate PDC deposits in dark red pixels. The partial erosion of the PDC deposits in the Rio Blanco Sur appear in blue (Figure 5.20a) and can be observed on VHR optical image (Figure 5.20b) acquired 5 days
after the SAR intensity image used in Figure 5.20a. Figure 5.21a shows the intensity changes on the NE flank of Calbuco between pre- and post-eruption images, on which the lahars and remobilization of deposits in the Rio Frio and Rio Blanco Este can be detected. The erosion and remobilization of deposits increase surface roughness, which appear in bright blue pixels cutting through the red pixels of the PDC deposits. This is validated on the VHR optical images (Figures 5.21b and 5.21c) where the erosion channels from remobilization of PDC deposits by the lahars appear in dark grey.

**Figure 5.19.** Georeferenced ALOS-2 false-color composite intensity ratio images of Calbuco. The false-color composite (R: Earlier intensity image, G: Later intensity image, Blue: Ratio of the later divided by the earlier intensity image) images were obtained using a) a pair of intensity images from before the April 2015 PDCs; and b) a pair of intensity images from before the April 2015 PDCs and during the third eruption phase, but after the occurrence of PDCs on April 29, 2015. In image a) the pink dashed lines contour the 1961 lava flow. In image b) the PDCs and lahars appear in dark red and are highlighted by the dashed lines.
Figure 5.20. Comparing observations of PDC deposits from SAR intensity and VHR optical images in the Rio Blanco Sur at Calbuco. 

a) Georeferenced false-color composite intensity image of the Rio Blanco Sur obtained with images from before and after the April 2015 PDCs in HV polarization. The PDC deposits appear in red while the eroded surfaces appear in blue. 

b) High-resolution optical image of the PDC deposits in the Rio Blanco Sur from Pleaides-1A image at 50-cm resolution acquired on December 1, 2015. The PDC deposits appear in brown while the eroded surface appears in grey.
Figure 5.21. Comparing observations of erosion channels from SAR intensity and VHR optical images on the NE flank of Calbuco. a) Georeferenced false-color composite intensity image obtained with images from before and after the April 2015 PDCs in HH polarization. The PDC deposits appear in red while the eroded surfaces appear in blue. b) and c) High-resolution optical image from Pleiades-1A image at 50-cm resolution acquired on December 1, 2015, from the areas in image a). The erosion channels from remobilization and erosion of PDC deposits appear in dark grey and are pointed with arrows on both the optical and intensity image.

**Thermal Images**

**Volcán de Colima**

At Colima, the pre-eruption LST image on Figure 5.22a shows a positive thermal anomaly of ~ +90°C at the summit that corresponds to new extrusion of lava when the new dome was forming. The thermal anomaly is comparable to values estimated by Thiele et al. (2017) using an
aerial thermal camera during dome growth. The vegetated flanks of the volcano have an average LST of 17°C while the bare ground and farm fields at the bottom of the volcano are around 30°C on average. Vegetated surfaces tend to be cooler than bare ground. The contrasting average temperatures between the S and N flank are due to the difference in sun exposure time and shadows created by the topography. The N flank receives less sun radiation on average and the steep landscape accentuates the shadowing effect, thus decreasing the average LST by about 10°C.

The post-eruption LST image on Figure 5.22b shows high LST values of ~40°C on the southern flank inside the Montegrande ravine while the surrounding landscape displays average temperatures of 17°C for the forested areas and 30°C for the farmlands at the foot of the volcano. LST values at the summit reach up to 75°C inside the crater.

**Figure 5.22.** Pre- and post-eruption Land Surface Temperatures (LST) on the south flank of Colima. LST images were generated from Landsat-8 TIRS images, which were acquired a) pre-eruption on March 12, 2015, and b) post-eruption on January 10, 2016. c) Relative LST change between pre- and post-eruption LST images, where red indicates ≥100% increase in LST values and blue indicates ≥ -100% decrease in LST values.
Figure 5.23. Relative and absolute post-eruption LST changes at Colima. **a)** Relative changes in LST between two post-eruption Landsat TIRS images acquired on January 10 and November 25, 2016. In the color bar, red indicates ≥100% increase in LST values and blue indicates ≥-100% decrease in LST values, while yellow indicates constant LST. **b)** Absolute changes in LST values between the same two post-eruption Landsat TIRS images from January 10 and November 25, 2016.

Using the ratio of the pre- and post-eruption LST images, we can outline the changes resulting from the eruption itself (Figure 5.22c). The strongest increase in LST corresponds to the PDC deposits of the Montegrande ravine and to the lava flow that was descending the southern flank during the next year. The decrease in LST at the summit in the crater area corresponds to the destruction of the dome, although lava continued to be generated.

The relative changes between two post-eruption LST images show that increased LST are focused on the descending lava flow of the southern flank (Figure 5.23b). The absolute LST
changes between these same post-eruption on Figure 5.23b, may be used to determine which areas of the PDCs cooled down more rapidly than others.

The lava flow front displays a stronger decrease in LST of ~ -15°C, contrasting with an increase of +30°C in the active portion of the lava flow. The valley-confined PDC deposits show a small decrease in LST of ~ -3°C whereas the ash-cloud surge areas exhibit a stronger decrease of -5°C. This difference is attributed to the faster cooling of thin ash-cloud surge deposits while heat remains trapped inside the thick valley-confined PDC deposits. This suggests that post-eruption LST changes can be used as proxy to coarsely differentiate as-cloud surge from concentrated depositional areas. This is further supported by a study by Pensa et al. (2018) who found that channel confinement of the 2015 PDCs inhibited heat dispersion, thus resulting in higher emplacement temperatures between 345 – 385°C, while lower temperatures (170 – 220°C) were recorded for unconfined deposits.

**Calbuco Volcano**

The pre-eruption background LST on Figure 5.24a are between 13°C on vegetation and 20°C on bare soil on average. The western side of the summit displays lower LST with an average of +2°C which may be due to the snow and glacier cover. On the post-eruption LST image, I detect a positive thermal anomaly up to +45°C within the crater and positive anomalies of lower magnitude inside the major valleys surrounding the volcano.

The positive anomalies in the valleys are a result of the PDC and tephra fall deposits emplaced during the eruption, but the contrast with background LST is reduced by the volcanic plume that dispersed over wide areas during the eruption. The strongest positive thermal anomalies occur mainly in the NE direction, which is consistent with the directionality of the eruption as well as the observations from optical and SAR imagery described in previous sections.
I also note the temperature gradient from the summit towards the more distal slopes, which may be the results of exponential thinning of tephra fall deposit thickness with distance from the vent, as well as additional ash-cloud surge deposits in the proximal area, contrasting with only tephra fall deposits in the more distal areas.

**Figure 5.24.** Pre- and post-eruption LST at Calbuco. Absolute LST at Calbuco Volcano before the eruption on April 11, 2015 (**left**) and after the eruption on April 27, 2015 (**right**). Elevated LST relative to the background LST values are highlighted with the dashed line and correspond to the newly deposited PDCs and tephra fall. The PDCs in the valleys are annotated as 1. Rio Blanco Sur, 2. Rio Sur – S branch, 3. Rio Sur – N branch, 4. Rio Blanco Norte, 5. Rio Tepu, 6. Rio Frio – Rio Blanco Este. The highest LST values were recorded inside the active crater (black rectangle), while the lowest LST values correspond to the presence of the volcanic plume, mostly composed of water vapor (**see Chapter 2**).
Discussion

Remote Sensing Mapping of PDC Deposits

The results from deposit mapping using optical, SAR and thermal remote sensing data provide a comprehensive overview of the distribution of PDC deposits for both Colima and Calbuco volcanoes. Data fusion is of value particularly in steep valleys where shadowing and coarser resolution often mask the deposit footprints.

It is important to be able to distinguish dense concentrated PDCs from ash-cloud surges, because the two features cause different kinds of damages to infrastructures and vegetation (e.g., Cole et al., 2002; Jenkins et al., 2013). Thus, for the purpose of mapping hazards, it is important to evaluate the extent of each flow type. I find that SAR intensity images and LST images are good tools to differentiate between areas of ash-cloud surge deposits and dense PDC deposits, mainly due to their differences in deposit thickness and impact to vegetation. At Colima, the post-eruption changes in the vegetation pattern, specifically vegetation regrowth, is a good tool to delineate the areas covered with ash-cloud surge deposits. These patterns are detectable using trends of increasing SAR intensity values and changes in vegetation health on VHR optical images and Landsat NDVI images.

The advancement of slow-moving lava flows is also important for hazard purposes as they often generate small but frequent rockfalls and sometimes short-runout block-and-ash flows. This phenomenon was observed at Colima during the months following the progression of the S lava flow (GVP, 2016). My results show that SAR coherence images are well-suited to track syn-eruption motion of lava flows, but their applicability becomes limited when applied to fast event such as PDCs and lahars.
Figure 5.25. Signature fields of the average L-band SAR intensity changes in co- and cross-polarization for individual volcanic features at Colima and Calbuco. The average SAR intensity changes for each volcanic feature are measured in decibels (dB) and are used with their respective uncertainties to determine the signature fields. These fields are compared with the intensity change measured at Merapi by Solikhin et al. (2015a). Note that the average intensity change for vegetation regrowth was measured on a pair of post-eruption images.

In this study, I related reduced surface roughness in the valley to the deposition of dense PDCs. I also related increased surface roughness to reworked and/or eroded PDC deposits in those same valleys. These findings are in agreement with those made by Solikhin et al. (2015a) at Merapi Volcano. Furthermore, the average values of ΔdB that I obtained for concentrated PDCs at Colima and Calbuco are comparable to the ΔdB measured by Solikhin et al. (2015a) at Merapi and fall within the same signature fields (Figure 5.25).
For ash-cloud surges and tephra fall deposits, however, Solikhin et al. (2015a) recorded $\Delta dB > 0$, which they interpreted as increased roughness caused by the removal of vegetation. This is the opposite of my findings at Colima and Calbuco, which showed that the deposition of ash-cloud surges and tephra fall reduced surface roughness initially caused by the presence of vegetation.

For reworked PDC deposits, Solikhin et al. (2015a) found that the higher water content resulted in $\Delta dB > 0$ in HH polarization but $\Delta dB < 0$ in HV polarization. This is consistent with my findings for the reworked PDCs at Colima, and the $\Delta dB$ values obtained by Solikhin et al. (2015a) fall within the range of the measured uncertainties (Figure 5.25).

At Calbuco, however, the lahars resulted in $\Delta dB$ values closer to those measured for summit deformation (Figure 5.25). This may suggest that the changes in backscatter intensity observed cutting through the PDC deposits reflect the intense erosion, similar to what is observed at the summit, as opposed to watery reworked PDCs or lahar deposits. Furthermore, the PDCs at Colima and Merapi are both block-and-ash flows that resulted from a dome-collapse eruption, which may also explain the similarity in $\Delta dB$ results.

In conclusion, the changes in backscatter intensity measured with L-band images in direct (HH) and cross (HV) polarization appear consistent for a variety of PDC deposits, both fresh and reworked. Therefore, the signature fields derived from these changes (Figure 5.25) may be used for the detection of PDCs at other volcanoes. Further testing should be done for the other volcanic features such as lava flows, tephra fall and ash-cloud surge deposits. Overall, these results are highly encouraging and imply that the use of L-band band imagery is a reliable method for medium-resolution mapping of PDC deposits.
Figure 5.26. Probability maps of PDC deposit extent at Colima (top row) and Calbuco (bottom row), from NDVI changes (a and c) and SAR intensity changes in HV polarization (b and d).

Going one step further, the $\Delta dB$ obtained with ALOS-2 data (Tables 5.1 and 5.2; Figure 5.25) for detecting PDC deposits, combined with $|Q_{NDVI}|$ from Landsat-8 data, are used to derive probability maps of deposit extents (Figure 5.26). This is done with linear rescaling of the $|Q_{NDVI}|$ map and the SAR $\Delta dB$ map, to probability values in the range $[0, 1]$ using the threshold values estimated from mapping results such that:

- $P(X)_{SAR} = 0 < f(x) < 1$, with $f(x) = \frac{(x-\Delta dB_{min})}{(\Delta dB_{max}-\Delta dB_{min})}$
- $P(Y)_{R_{NDVI}} = 0 < f(y) < 1$, with $f(y) = \frac{(x-|Q_{NDVI_{min}}|)}{|Q_{NDVI_{max}}-|Q_{NDVI_{min}}|}$
$\Delta dB_{\text{min}}$ and $\Delta dB_{\text{max}}$ are the lower and upper thresholds of intensity change for probabilities of PDC deposits between 0 and 1. For the lower threshold value for $P(X) = 0$, I use the average $\Delta dB$ measured for ash-cloud surge deposits in HV polarization (Figure 5.25), which is -3 dB at Colima and -4 dB at Calbuco. For the upper threshold value for $P(X) = 1$, I use the average $\Delta dB$ measured for PDC deposits in HV polarization, which is -7 dB at Colima and -9 dB at Calbuco.

$|Q_{\text{NDVI, min}}|$ and $|Q_{\text{NDVI, max}}|$ are the lower and upper thresholds of $|Q_{\text{NDVI}}|$ for PDC deposits observed at both Colima and Calbuco. The lower threshold value of $|Q_{\text{NDVI}}|$ was set at 0.2 and the upper threshold was set at 0.8.

The NDVI probability map obtained with $|Q_{\text{NDVI}}|$ at Colima shows false positives in the proximal area where rockfalls and lava flows were present. The SAR probability map obtained with $\Delta dB$, however, shows low probability values where lava flows were emplaced, due to the differences in $\Delta dB$ values between PDCs and lava flows (Figure 5.25). Therefore, I apply a method developed by my collaborators to combine optical and SAR data into a weighted joint probability map (Jo and Osmanoglu, 2020). The authors successfully applied the algorithm to map water flood extents, and they successfully demonstrated the advantages of joint probability maps to improve the mapping accuracy, particularly in urban areas.

Here, I apply the weighted joint probability algorithm to combine the $|Q_{\text{NDVI}}|$ and $\Delta dB$ probability datasets in order to refine the mapping of PDC deposit extent (Figure 5.27) using Equation (6):

$$P(X|Y) = \frac{P(X) \cdot P(Y)}{P(X) \cdot P(Y) + [1 - P(X)] \cdot [1 - P(Y)]} \quad (6)$$
Figure 5.27. Joint probability maps of PDC deposit extents at a) Colima and b) Calbuco. Histograms of the independent (optical and SAR) and joint probability maps for the areas of interests 1-3 are displayed below the respective map.
The comparison of the histograms of the optical and SAR probability maps (Figure 5.27), with the joint probability histogram, are used here to check the performance of the technique. Background values in the $|Q_{NDVI}|$ probability map are significantly skewed towards 0, which is in agreement with the absence of any deposits. The histogram of $\Delta dB$ probability map, however, is spread over the whole range $[0, 1]$, although with higher frequencies for lower probability values. This suggests higher background noise in the probability map from SAR data. Combining the results into the joint probability map is valuable to reduce the background noise and emphasize the extent of PDCs. Moreover, I find that it reduces the occurrence of false positives in the proximal area where lava flows and rockfalls showed as high probability pixels on the optical dataset, but with low probability values on SAR dataset. The success of this technique may be employed for future PDC eruptions to rapidly produce damage extent maps and outline the recently deposited PDCs.

**Deposit Distribution and Dynamics with Valley Morphology**

Using the results from remote sensing data observations presented in this chapter, and the field descriptions of deposits presented in Chapters 3 and 4, together with the analysis of the pre-eruptive topography, I classify the mapped PDC deposits into three categories: (I) valley-confined PDCs; (II) overbank PDC deposits and (III) ash-cloud surge deposits.

A common characteristic of PDCs generated from column-collapse and dome-collapse eruptions is their preferential deposition into the main drainage pathways that carve the landscape, thus forming the deposits recognized as valley-confined PDCs (category I). First, I note that the directionality of the valley-confined PDCs is affected by a combination of factors, including the shape of the vent area and the eruption type.
For instance, the southward overspill of the lava dome at Colima, and the collapse of the southern crater rim, a portion of the southern flank and of the lava flow front, all contributed to the uni-directionality of the PDCs into the Montegrande ravine (Figure 5.28).

**Figure 5.28.** 3D view of the south flank of Colima from VHR post-eruption SPOT-6 image acquired on July 25, 2015, and draped over 5-m DEM. The PDCs appear in brownish grey inside the Montegrande and San Antonio ravines. The summit is masked by clouds, and the lava flow front that was descending between 2015 and 2017 is outlined in yellow. The white arrows show two overspill locations with rechanneled overbank flows.

At Calbuco, PDCs are distributed radially in the valleys surrounding the volcano from the vent, but the longest deposit runouts and volumes are recorded in the valleys in the N-NE direction (i.e., Rio Tepu, Rio Blanco Este and Rio Frio). This coincides with the directionality of the eruptive plume during both paroxysmal eruptive phases on April 22 and 23, and is also with the direction of the open summit depression in which the crater was excavated.
Overbank flow deposits (Category II) are more complex to map and require both prior knowledge of the valley morphology and validation from field observations to differentiate them from ash-cloud surge deposits (Category III). Overbank flows are concentrated flows and are the result of valley-confined PDCs that overspill from the channel confines to deposit onto valley interfluves, and can also spread laterally away from the valley and become rechanneled into adjacent channels. The deposit facies (i.e., sedimentology and granulometry) of overbank flows closely resemble that of the valley-confined dense-basal avalanche deposits from which they originated and separated, and contrast with stratified and thin fine-grained facies of ash-cloud surge deposits.

Ash-cloud surges, in contrast, can either settle over the dense-basal avalanche inside the valley, or detach from the dense-basal avalanche, spilling out of the channel and travel laterally, depositing on the valley interfluves and often surmounting steep slopes during surge expansion. Detached ash-cloud surges can even become rechanneled into adjacent channels, then known as surge-derived pyroclastic flows (e.g., Druitt et al., 2002; Ogburn et al., 2014). During settling of the ash-cloud surge deposits appear as fringes along the main valley. This is observed on VHR optical images with areas of singed vegetation along the valley pathway, and with thin, massive and laminar (i.e., non-stratified) ash deposits in the field.

At Calbuco, areas stripped of vegetation are used as markers for surge extent as opposed to tephra deposits on VHR optical imagery. Detached surges are observed at both Calbuco and Colima volcanoes with the presence of blown down trees at the surface of stratified to cross-stratified ash-rich deposits that suggest lateral motion (i.e., contrasting with vertical settling of ash-cloud surges) (see Chapter 4).
At Colima, deposits of detached surges are described as *dilute overbank flows* (see Chapter 3). Both overbank flows and detached surges are hazardous and highly destructive, particularly when they become unconfined with unpredictable trajectory, but the damages to vegetation and infrastructures often differ and can help determine the types of deposits as well. Most specifically, the damages to trees in the vicinity of the valley often suggest changes in dynamic pressures from $< 3$ kPa for standing trees, to $>7$ kPa for blown down trees (e.g., Cole et al., 2002).

**Figure 5.29.** Comparison of the valley morphologies at Colima and Calbuco. Schematic valley profiles in **a)** the Rio Frio at Calbuco Volcano, and **b)** the Montegrande ravine at Volcán de Colima. The vertical scale difference shows the large valley structure with an incised channel path at Calbuco, which contrasts with the shallow and narrow channel at Colima.

The proportion of overbank flows relative to the volume of valley-confined deposits at Calbuco is much lower than that of Colima. Overbank deposits account for 6% of the area covered by concentrated PDCs at Calbuco, whereas they account for 38% at Colima. This is largely caused by the differences in valley morphologies (Figure 5.29): At Colima, the morphology of the
Montegrande ravine is a narrow-incised channel with high sinuosity and slope breaks. The complexity of the channel is caused by the lahar erosion that carved into an existing path, creating terraces. At Calbuco, the valleys were carved by rivers from glacier runoffs, resulting in incised pathways within larger valley structures with high walls, and generally greater channel capacities. Moreover, the pre-eruptive topography of the proximal area on the Calbuco E-NE flanks shows a complex system of valleys and channels, ridges and old lava flows that added complexity to the pathways of PDCs (Figure 5.30).

**Figure 5.30.** 3D view of the NE flank of Calbuco from VHR optical image showing the PDC deposits and flow paths. The VHR Pleiades-1A image was acquired syn-eruption (i.e., during the third eruption phase on April 27, 2015) and is draped over 5-m DEM. The black arrows highlight the complexity of the paths taken by the PDCs as they descended onto the NE flank. Note also the presence of the 200-m-high ridge to the SE, which explains the lack of PDC deposits on the SE flank.
Figure 5.31. Surface morphology of PDC deposits with valley morphology at Calbuco. VHR Pleiades-1A images from April 27, 2015, at Calbuco volcano showing the presence of lobate PDC deposit morphologies in a) the Rio Blanco Sur, and b) in the south branch of the Rio Sur, with respect to valley morphology changes. Specifically, channel enlargement is highlighted in both valleys, as well a strong break in slope in the Rio Blanco Sur and the presence of a vertical cliff (image a). The ash-cloud surge expansion zone is shaded in yellow and is associated with the steep slopes of the proximal area in the Rio Blanco Sur. The steeper slopes in the proximal area of the Rio Blanco Sur where PDCs have high potential energy may have promoted transport over deposition, but also promoted ash-cloud surge expansion. Ash-cloud surge lateral motion is also observed in correlation to the vertical cliff. In the Rio Sur – S branch (image b) the orange dashed line indicates the ash-cloud surges with deposit fringes along the channel path.
Figure 5.32. Channel morphology analysis and relationship to PDC deposits at Calbuco in the proximal portion of the Rio Blanco Este. a) GIS outputs from the channel morphology algorithm displayed over VHR pre-eruption Worldview-1 panchromatic image from May 22, 2014. The color of each polygon section of the channel reflects the percentage change in channel capacity with distance (green: increased channel capacity or reduction by <50%, orange: reduction by 50-100%, red: reduction >100%). Results show the strong reduction in channel capacity with the combination of a 40° turn in channel direction that led to the overspill over the valley wall, spreading onto the valley interfluve and rechanneled towards the Rio Frio (red arrows). b) VHR post-eruption Pleiades-1A image from April 27, 2015 showing the PDC deposits within the proximal portion of the Rio Blanco Este and onto the valley interfluves. Note the lobate morphology of the PDC deposits over the interfluve area between the Rio Blanco Este and the Rio Frio, as well as the rechanneled deposits towards the Rio Frio. The yellow square indicates the area displayed in images c) and d). c) Google Earth image from 2014 showing the pre-eruptive landscape. d) Details of the PDC deposits in the proximal Rio Blanco Este area showing the lobate morphology of the valley-confined deposits after the enlargement of the channel, as well as the presence of a vertical cliff that promoted overbank flows onto the terrace and possibly enhanced ash-cloud surge expansion to the SE. The orange dashed arrow indicates the direction of ash-cloud surges that detached from the lobate PDCs.
Figure 5.33. Channel morphology analysis and relationship to PDC deposits at Calbuco in the Rio Tepu and Rio Frio – Rio Blanco Este 

a) GIS outputs from the channel morphology algorithm displayed over VHR pre-eruption Worldview-1 panchromatic image from May 22, 2014. The summit is in the bottom left corner. Reduction of channel capacity in the Rio Frio is correlated with the distribution of overbank deposits, outlined in pink over the VHR post-eruption Pleiades-1A image in figure b). Valley-confined deposits are outlined in blue and the direction of unconfined flows is shown by the red arrow. c) PDC deposits in the Rio Tepu with the yellow bounding box of enlarged area in image d). The proximal area shows evidence of ash-cloud surge expansion that blanketed the upper slopes along the valley. d) Zoom in the Rio Tepu that shows two vertical cliffs associated with evidence of ash-cloud surge deposits in the upper part of the valley walls, suggesting hydraulic jumps where air entrainment resulted in upward dispersion of the dilute part of the PDC.

Changes in flow dynamics of the dense-basal avalanche in response to changes in the valley morphology are observed at both Colima and Calbuco. At Colima, sudden enlargement of the channel and break in slope at Colima resulted in en masse deposition of the basal undercurrent (see Chapter 3).
At Calbuco, channel enlargement and break in slope (> 20°) are observed in correlation to lobate morphologies of PDC deposits observed on VHR imagery in the Rio Blanco Sur, Rio Sur (S-branch) and proximal portion of the Rio Blanco Este (Figures 5.31 and 5.32) and are interpreted as changes in flow dynamics from high pore-fluid pressure-driven flows to dense granular flow behavior (see Chapter 4). This can be attributed to a decrease in momentum from the rapid loss of channel confinement, and from decreasing pore-pressure during the elutriation of fines and gas escape caused by the break in slopes. The relation between unconfined depositional environment and lobate morphologies of PDCs is also observed for the overbank deposits on the interfluve of the proximal portion of the Rio Blanco Este (Figure 5.32) and in the Rio Frio (Figure 5.33).

![Figure 5.34](image)

**Figure 5.34.** Channel morphology analysis and relationships to PDC deposits at Colima in the proximal part of the Montegrande ravine. **a)** VHR pre-eruption Pleiades-1A image showing the pre-eruptive morphology of the Montegrande ravine. **b)** VHR post-eruption SPOT-6 image showing the PDC deposits including the valley-confined, overbank and ash-cloud surge deposits. The directionality of the valley-confined PDCs is shown by the black arrows, while the blue arrows indicate overbank flow. Note the correlation between overspill points and the strong sinuosity of the channel that resulted in overbank flows and detached surges. Ash-cloud surge expansion is also observed in association with the steep slopes of the proximal area. **c)** GIS outputs from the channel morphology algorithm in the proximal channels (yellow area in image b) leading to the Montegrande ravine displayed over the VHR pre-eruption DEM. Results show the reduction in channel capacity in conjunction with the location of overspill points and where sinuosity is the strongest.
The Effects of Channel Morphology Parameters

Channel capacity and cross-sectional area

The correlation between reduced cross-sectional area of the Montegrande ravine and the generation of overbank flows at Colima has been previously introduced in Chapter 3. Results from channel morphology analysis are consistent with these earlier findings (Figures 5.34 and 5.35) and show a strong correlation between the reductions of channel cross-sectional area and volumetric channel capacity, and the increased overbank deposit widths (Figure 5.36).

Figure 5.35. Channel morphology analysis and relationships to PDC deposits at Colima in the medial part of the Montegrande ravine. a) VHR pre-eruption Pleiades-1A image showing the pre-eruptive morphology of the Montegrande ravine. b) VHR post-eruption SPOT-6 image showing the PDC deposits including the valley-confined, overbank and ash-cloud surge deposits. The overspill points are indicated with blue arrows, and the differentiation between passive and active overbank deposits are highlighted. The decreasing effect of the sinuosity with distance is observed with the transition from detached surges to ash-cloud surge deposit fringes with superelevation of dense-basal avalanche. c) GIS outputs from the channel morphology algorithm in the proximal channels leading to the Montegrande ravine displayed over the VHR pre-eruption DEM. Results show the reduction in channel capacity in conjunction with passive overbank flows (see text for explanations) while sudden turn in channel direction corresponds to active overbank flows.
Figure 5.36. Longitudinal changes in channel capacity and overbank deposit width in the Montegrande ravine at Colima. **a)** Changes in the cross-sectional area with distance from the summit compared with the width of overbank deposits. Measurements of the cross-sectional area of the valley were performed at 50-m (full line) and 100-m (dashed line) intervals. The widths of overbank deposits are measured as the width of deposits beyond the valley edges and along each cross-section profile. **b)** Changes in volumetric channel capacity with distance from the summit compared with the width of overbank deposits. The volumetric channel capacity is measured as the volume of space below the top of the valley walls, for each polygon, at 50-m intervals.

There is a distance delay between the point of channel capacity reduction and the maximum increase in overbank width, which can be attributed to the downslope motion of the flow as it escapes the channel confines. The presence of overbank flows is observed when the channel cross-sectional area decreased below 7000 m² (Figure 5.36a) and when the volumetric capacity decreases below 15,000 m³ (Figure 5.36b). At Calbuco, the partitioning of the total PDC volumes into multiple valleys result in a smaller proportion of overbank flows. In the proximal area of the Rio Blanco Este at 2.6 km from the vent, the generation of overbank deposits that dispersed onto the
valley interfluves and rechanneled towards the Rio Frio (Figure 5.32) correlates with the reduction of cross-sectional area below 40,000 m² (Figure 5.37a). In Figure 5.37b, the presence of overbank is associated with a strong decrease in volumetric channel capacity below 20,000 m³. In the Rio Frio, the overbank deposits are limited to a 2 km-long segment.

**Figure 5.37.** Longitudinal changes in channel capacity and overbank deposit width in the proximal part of the Rio Blanco Este at Calbuco. **a)** Changes in cross-sectional area with distance from the summit compared with the width of overbank deposits. Measurements of the cross-sectional area of the valley were performed at 50-m (full line) and 100-m (dashed line) intervals. The widths of overbank deposits are measured along the profile from the cross-sectional area, beginning at the valley edge. **b)** Changes in volumetric channel capacity with distance from the summit compared with the width of overbank deposits. The volumetric channel capacity is measured as the volume of space for each polygon, at 50-m (full line) and 100-m (dashed line) intervals, below the valley edges.
In Figure 5.38a, the localized reduction in channel cross-sectional area coincides with increased overbank deposit widths, although the significant decrease in cross-sectional area after 5.2 km did not generate overbank flows. This lack of overbank flows is associated with the lower valley edges combined with increased channel width, in comparison to other sites with narrower incised channel paths where overbank deposits are observed. In Figure 5.38b, overbank deposits occur where the volumetric channel capacity decreases below 20,000 m³, which is similar to the value obtained for the proximal Rio Blanco Este (Figure 5.37b).

Figure 5.38. Longitudinal changes in channel capacity and overbank deposit width in the Rio Frio and distal part of the Rio Blanco Este at Calbuco. a) Changes in cross-sectional area with distance from the summit compared with the width of overbank deposits. Measurements of the cross-sectional area of the valley were performed at 50-m (full line) and 100-m (dashed line) intervals. The widths of overbank deposits were measured along the profile from the cross-sectional area, beginning at the valley edge. b) Changes in volumetric channel capacity with distance from the summit compared with the width of overbank deposits. The volumetric channel capacity is measured as the volume of space for each polygon, at 50-m (full line) and 100-m (dashed line) intervals, below the valley edges.
The inverse relationship between channel capacity and the generation of overbank flows makes intuitive sense: when the volume of a flow unit exceeds the cross-sectional area of the channel, there is overspilling of the dense basal avalanche onto the interflues. Therefore, the effect of the channel capacity is also dependent on the flow volume and mass flux, as a small-volume flow or with small mass-flux are most likely to remain confined to the channel bed.

**Sinuosity**

While channel capacity is strongly correlated to overbank deposits at both Colima and Calbuco, a combined effect from channel sinuosity is observed as well.

![Figure 5.39](image)

**Figure 5.39.** Longitudinal changes in channel sinuosity and slope with changes in overbank deposit width in the Montegrande ravine at Colima. **a)** Changes in channel sinuosity with distance from the summit compared with the width of overbank deposits. **b)** Changes in channel slope with distance from the summit compared with the width of overbank deposits. The sinuosity and slope were measured along the centerline of the channel for segment length of 50-m (full line) and 100-m (dashed line).
At Colima, the largest overbank deposit widths correspond to the rechanneled overbank flows, with overspill points located where changes in channel directions exceed 20° (Figures 5.34, 5.35 and 5.39a). The effect of sinuosity is limited, however, to the proximal and medial portions of the Montegrande ravine, suggesting a need for high flow momentum to enable the overspilling of the dense-basal avalanche.

**Figure 5.40.** Longitudinal changes in channel sinuosity and slope with changes in the width of overbank deposits in the proximal part of the Rio Blanco Este at Calbuco. a) Changes in channel sinuosity with distance from the summit, compared with the width of overbank deposits. b) Changes in channel slope with distance compared with the width of overbank deposits. The sinuosity and slope are measured along the centerline of the channel for segment length of 50-m (full line) and 100-m (dashed line).
At Calbuco in the proximal Rio Blanco Este and in the Rio Frio, the overbank deposits that are described as corresponding with reduced channel capacity in the previous section, are also associated with changes in channel direction >20° (Figures 5.32, 5.33, 5.40a and 5.41a).

Surge detachments are also observed in correlation to increased sinuosity in the proximal to medial areas in the Montegrande ravine at Colima (Figures 5.34 and 5.35). In distal areas, however, high sinuosity results in deposition in fringes on the interfluvies along the Montegrande ravine, associated with superelevation deposit structures (Figures 5.35 and Chapter 3). Similarly, at Calbuco, surge detachments are observed in association to strong sinuosity (> 20°), even in distal areas of the Rio Blanco Este due to the large flow volumes that descended the valley (Figure 5.33). In the Rio Sur (N-branch), detached surges are observed after a change in channel direction by 15° at 2.8 km from the summit (Figure 5.42).

**Slope**

Figures 5.39b, 5.40b and 5.41b show the changes in local slope of the channels at Colima and Calbuco with distance and the width of overbank deposits. Results suggest that break in slope has little correlation to the overbank deposit width, although localized break-in-slope >10° may increase the potential for overbank flow generation. However, observations from VHR optical images show that slope and slope changes can be correlated with ash-cloud surge expansion and surge detachment.

At both Colima and Calbuco, surge expansion is maximal in the proximal areas where slopes are highest (Figures 5.30–5.32 and 5.34). At Calbuco, I posit that during the later stages of the column collapse, the PDCs that travelled towards the ENE are those that resulted in significant surge expansion, which in turn travelled and deposited over the ~200-m high ridge to the SE of the Rio Blanco Este proximal depression (Figure 5.30). This is consistent with prior
findings at Soufriere Hills Volcano by Cole et al. (2002) and Ogburn (2015) who proposed that steep slopes enhanced air entrainment and thus increased thermal expansion of the ash-cloud surge, enabling it to surmount topographic barriers. At Colima, the surge detachment footprint is of lower intensity when slopes decreased below 15 degrees. Moreover, in the proximal and medial areas, surge expansion footprint and detached surges are associated with break in slope >10° (Figure 5.39b).

Figure 5.41. Longitudinal changes in channel sinuosity and slope with changes in the width of overbank deposits in the Rio Frio and distal part of the Rio Blanco Este at Calbuco. a) Changes in channel sinuosity with distance from the summit, compared with the width of overbank deposits. b) Changes in channel slope with distance compared with the width of overbank deposits. The sinuosity and slope are measured along the centerline of the channel for segment length of 50-m (full line) and 100-m (dashed line).
At Calbuco, the presence of vertical cliffs (>60° slope break) in the Rio Tepu, Rio Blanco Este (proximal) and Rio Blanco Sur is also associated with ash-cloud surge expansion (Figures 5.31 – 5.33). These are suggestive of hydraulic jumps, which have been proposed as major driving factor in the development and detachment of surges during previous eruptions at Colima (Saucedo et al., 2004, 2005), and at Merapi Volcano (Charbonnier and Gertisser, 2008, 2011; Bourdier and Abdurachman, 2011; Lube et al., 2011; Jenkins et al., 2013).

Figure 5.42. Evidence of detached surges associated with a change in channel direction in the Rio Sur – N branch at Calbuco shown on VHR Pleiades-1A images. a) The white lines highlight the change in channel direction and the arrows indicate the direction of the detached surges. The yellow bounding box is the area enlarged in image b) where the detached surges are indicated with partially stripped vegetation in the downslope direction suggesting horizontal motion of the ash-cloud surges.

Summary and Implications for Hazard Assessments

The parameters of channel morphology (i.e., cross-sectional area, channel capacity, slope and sinuosity) display correlation with the distribution of PDC deposits, namely the generation of overbank flows and surge detachment. The reduction of channel capacity, either longitudinally caused by the channel narrowing, or vertically by progressive infilling from successive PDC pulses or lahar erosion terraces, is associated with the presence of overbank flow deposits. Without the presence of slope breaks and sudden turns in the channel direction, the overbank deposits remain
close to the main channels, depositing as fringes. Conversely, when sinuosity and/or breaks in
slope occur, the overbank flows spread laterally and frequently become rechanneled into adjacent
channels. Changes in channel direction greater than 20° and break in slope by 10° are correlated
with greater generation of overbank flows. The proportion of overbank is dependent on the
combination of the intensity of channel constriction, sinuosity and local slope, but also depends on
the proximity to the source of PDCs. For instance, high-energy and large mass flux on the NE
flank at Calbuco resulted in most channels being filled in the proximal area, blanketing the
surrounding topographic obstacles.

Based on the observed controls by the valley morphologies, I classify overbank flows into
two types: (1) passive and (2) active overbank flows. Detached surges are considered as part of
type II flows while ash-cloud surge expansion without significant motion independent of the basal
undercurrent corresponds to type I flows in terms of behavior response to topographic parameters.

- Type I passive overbank flows occur from limited accommodating volume inside the channel
  which results in the inundation of valley banks. The volume and density of the deposits from
  passive overbank is dependent on the proportion of the moving flow that can be entrained
  onto the sides of the valley.

- Type II active overbank flows occur as momentum is generated for part of the flow to
  overcome topographic barriers and expand away from the main channel. Active overbank
  flow deposits are often accompanied by surge detachment when decoupling of the dilute part
  travels farther than its dense parent flow. Surge detachment may also occur on its own if the
  channel capacity is high enough to channelize the dense-basal avalanche and if momentum is
  initially reduced (e.g., in distal areas) when the flow encounters a topographic obstacle.
  Therefore, active overbank flows are more likely to occur in proximal to medial areas.
The results presented here are consistent with prior studies for the control of the topography, although these studies dominantly focused on dome-collapse PDCs. At Merapi during the June 2006 eruption, Charbonnier and Gertisser (2008) and Lube et al. (2011) found that channel constriction, either naturally occurring or from the presence of lahar prevention structures (sabo dams), channel depth reduction from progressive infilling by the successive pulses of block-and-ash flows, and the presence of sharp channel bends, were all factors that enabled the generation of overbank flows.

Lube et al. (2011) found reverse stratigraphic sequence (i.e., deposition of the upper part of the flow before the basal part of the flow) in overbank deposits, which suggested a time-lag of overbank generation after deposition of the ash-cloud surge. They proposed that the ash-cloud surge largely exceeded the depth of the valley and deposited onto the valley interfluves immediately upon arrival at the jump locations, which were recognized as location of channel constriction and change in direction, while the dense-basal avalanche remained channelized until the flow depth of later pulses could exceed the channel capacity, thus generating passive overbank flows.

The inverse relationship between channel capacity and the generation of overbank flows was also observed during the November 2010 eruption of Merapi (Charbonnier et al., 2013). The remaining deposits from June 2006 as well as the progressive infilling during the 2010 eruption contributed to the reduction of channel capacity, which subsequently led to large and deadly overbank flows (Charbonnier et al., 2013).

Superelevation deposit features were also observed by Lube et al. (2011) during the June 2006 eruption of Merapi, and by Charbonnier et al. (2013) and Solikhin et al. (2015b) during the November 2010 eruption, and were attributed to flow acceleration in sharp channel bends, thus
enhancing the potential for overspill. The concept of surge detachment in response to topographic obstacles has also been discussed in other studies of PDC deposits, and found to be well correlated to slope, reduced cross-sectional areas of the valleys, and sinuosity (Loughlin et al., 2002; Ogburn, 2015).

My research shows the reproducibility of analyses across different types of PDCs and emphasizes that a priori knowledge of the topography can be employed to gauge zones potentially at risk from overbank and surge inundation. For hazard mapping purposes, the algorithm utilized here to derive changes in channel morphology with distance can be used to identify channel constrictions for a range of PDC volumes in order to predict where overbank flows and detached ash-cloud surges are likely to overspill the valleys. Similarly, sinuous areas of the valleys and break-in-slope can be used to determine areas that will likely experience these hazardous flows. This information will be useful to derive “hot spots” for hazardous flows and therefore update hazard maps accordingly.

This application in hazard assessment and crisis mitigation, however, will require updated DEMs to account for the frequent topographic changes that occur at active volcanoes. For example, the main valleys at Colima were carved by many lahars in the year prior to the paroxysmal eruption in July 2015, modifying the channel morphology significantly with erosion terraces and depositing block rich deposits (obstacles). Moreover, in 2017, renewed activity generated PDCs that descended onto the SE flank in La Arenal ravine because of the crater rim closure to the south caused by the lava flow (see Chapter 2). These are examples of topographic changes that need to be taken into account when updating hazard maps. In Chapter 6, I discuss a method for tracking topographic changes using time-series of medium-resolution to VHR DEMs.
References


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CHAPTER SIX:
TRACKING TOPOGRAPHIC CHANGES AT ACTIVE VOLCANOES WITH A
SPACEBORNE TIME-SERIES OF DEMS: CASE STUDIES OF THE 2015 ERUPTIONS
OF VOLCÁN DE COLIMA (MEXICO) AND CALBUCO VOLCANO (CHILE)

Introduction

In Chapter 5, I identified the important role played by the topography on the distribution
of PDCs and related hazards. In this chapter, I explore ways to track topographic changes through
time, and to quantify volume changes caused by eruptive products.

The evolution of the landscape in an active volcanic system is rapid, complex and often
dramatic (Cas and Wright, 1987; Siebert, 1996; Thouret, 1999). Volcanic phenomena are both
constructive and destructive; the highest erosion rates on Earth have been recorded at active
volcanoes (Major et al., 2000; Lavigne, 2004). Recurring volcanic events such as lava dome
growth, lava flows, and PDCs contribute to changes in elevations.

The volume of PDCs generated during a single event is variable and is dependent upon the
initiation mechanism and intensity of the eruption. Small-volume PDCs (i.e., < 1 km³) are
topographically controlled (see Chapter 5) and more frequent than large volume PDCs, which are
not discussed here.

As I demonstrated in Chapter 5, the topographic changes that occur during and between an
eruption can have important hazard implications. For instance, valley infilling reduces the ability
to contain subsequent PDCs, thus favoring the generation of overbank flows (e.g., Lube et al.,
2011; Charbonnier et al., 2013; Ogburn, 2015; Macorps et al., 2018). Breaks in slope and sinuosity
of the valleys also affect the dynamics of subsequent PDCs, increasing the likelihood for ash-cloud surge detachment and overbank flows.

The presence of snow or ice at the summit of volcanoes in cold or mountainous regions, and the heavy rainfalls for volcanoes in tropical regions are events that increase the risk of lahars or sediment-charged flash floods when remobilizing the unconsolidated materials from PDCs and tephra fall. The erosion from such mass flows largely contribute to changes in the morphology of the valleys. They incise new paths, often accentuating the sinuosity of the channels, and can form erosion terraces that reduce the volumetric channel capacity. These lahars terraces may also facilitate the generation of overbank flows by acting as a ramp. Extreme events may also cause landslides, which contribute to major changes of the volcano’s morphology. It is therefore important to understand the topographic changes that occur at a volcano during and between eruptive events. Investigating and quantifying topographic changes at the scale of a volcano and for eruption timescale of a few days to a few years can be achieved with digital elevation model (DEM) time-series.

I analyzed two types of DEM time-series derived from stereogrammetry applied to spaceborne along-track image pairs to examine topographic changes between eruptive events. The first was a medium-resolution (30 m) time-series with frequent observations from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) satellite, and the second was a time-series consisting of a single pre- and post-eruption acquisition derived from high-resolution spaceborne imagery (HRSI). Each time-series method is applied to two PDC-forming eruptions that occurred in 2015 at Volcán de Colima in Mexico and Calbuco Volcano in Chile to examine the magnitude and variability of topographic changes at these study sites. (Figure 6.1).
Figure 6.1. Location of the two study sites: Volcán de Colima in Mexico, and Calbuco Volcano in Chile. The insets show each volcano in detail with HRSI data, revealing the main valleys inside of which the deposition of PDCs occurred. The PDC deposits are outlined for both valley-confined flows (blue) and overbank flows (pink), and the outline of the respective craters are displayed at the summit.

Study Sites

The two volcanoes of interest are shown in Figure 6.1. The first study site is Volcán de Colima (19°31 N; 103°37 W), one of the most active volcanoes in Mexico. The activity at Volcán de Colima is characterized by the alternance between effusive activity with lava dome growth and lava flows, and explosive activity with volcanic plumes of ash and gas. The choice for this study
site was motivated by the dome-collapse eruption on July 10-11, 2015, which generated PDCs, referred to as block-and-ash flows, into the main valley of the southern flank (Figure 6.1). The block-and-ash flows reached a maximum distance of 10.5 km, which was the longest runout distance observed at Volcán de Colima for a dome-collapse eruption. The second study site is Calbuco Volcano (41°20’S, 72°37’W) in the Southern Andes of Chile. In contrast to Volcán de Colima, Calbuco Volcano was in a period of quiescence since the last reported eruption in 1972, until renewed activity with an explosive eruption on April 22-23, 2015. The explosions created a volcanic plume that eventually collapsed and generated PDCs that travelled into the main valleys around the volcano. For each study site, I investigate the magnitude of elevation changes for different types of volcanic features, namely the summit and crater area, lava flows, and valleys with PDCs and erosion.

**Data and Methods**

**Medium-Resolution DEM Time-Series: ASTER DEMs**

**DEM generation**

The processing of a series of ASTER scenes is based upon the work of Shean et al. (2020) who used a time-series of ASTER DEMs to study the change in glacier mass balance in the High Mountain-Asia region. I use a similar workflow with modifications adapted specifically to our case study. The overall workflow process is summarized in Figure 6.2. The VNIR subsystem of the ASTER instrument on the TERRA satellite carries a nadir-looking and backward-looking sensor. The simultaneous acquisition of bands 3N (nadir) and 3B (backward), in the spectral range 0.78 – 0.86 µm, produces along-track stereoscopic images that allow for the generation of DEM at 15-m resolution (Toutin, 2002).
I selected ASTER Level 1A (L1A) scenes from the ‘ASTER L1A Reconstructed Unprocessed Instrument Data V003’ EarthData archives for our two study areas Volcán de Colima and Calbuco Volcano. I queried scenes with day-time acquisition and a maximum cloud cover of 50%, and for the time periods of August 2004 – December 2019 and January 2009 – December 2019 for Colima and Calbuco respectively. The queries resulted in 308 and 87 L1A ASTER scenes for Volcán de Colima and Calbuco Volcano respectively.

Both datasets were downloaded and pre-processed with Ames Stereo Pipeline v 2.6.1 (ASP) software (Beyer et al., 2018a). The pre-processing tool ‘aster2asp’ applies radiometric

**Figure 6.2.** Workflow for ASTER DEM time-series. This workflow summarizes the processing steps used to retrieve the trend maps of elevation changes from the time-series.
corrections contained in the metadata (rational polynomial coefficient (RPC) model) of each scene and prepares the stereo pairs to be processed with ‘stereo’. Then I use the ASP semi-global matching (SGM) algorithm for stereo correlation with default parameters (7 x 7 pixel window), and I use the publicly available TanDEM-X (TDX) 1-arcsec (90-m) global DEM as a reference DEM (Rizzoli et al., 2017) for initial orthorectification of the input stereopairs (to improve the speed and accuracy of stereogrammetric correlation in highly variable terrain). While the TDX DEM is relatively coarse, its horizontal and vertical (~ 3.5m absolute and < 2 m relative) accuracy likely improved correlation of steep terrain.

The SGM algorithm improves disparity matches in areas of low or repetitive texture (Beyer et al., 2018b). Using the SGM algorithm also allows to discern features of finer resolution by using smaller matching kernels (NASA Ames Research Center, 2019). I used ASP’s default SGM disparity map filters (3 x 3 pixel median filter and a 3 x 3 pixel “texture-aware smoothing filter” with scaling factor 0.13) to remove residual artifacts from the generated point clouds. Isolated clusters of < 50 pixels surrounded by ‘nodata’ values were removed from the disparity map. I use the RPC model contained within each scene to perform stereo triangulation and then filter out any triangulation errors exceeding 3 times the 75th-percentile. The point clouds are converted to output gridded DEMs posted at 30-m with elevations relative to the WGS84 ellipsoid. Finally, I perform 2-pixel erosion of outer DEM boundaries to remove any residual edge artifacts. Additionally, I use a threshold of ± 100 m elevation offset relative to the reference TDX DEM to filter-out clouds and erroneous elevation that occur in cloud shadows. Correlations between stereopairs failed when too many points were filtered out (either due to clouds, shadows or artifacts) or when the disparity maps failed to find enough correlations. The stereo processing resulted in 306 (Colima) and 85 (Calbuco) DEMs with cloud-removed elevations.
DEM coregistration

I coregistered the set of processed ASTER DEMs to the reference TDX DEM. Co-registration of all individual DEMs is required to remove horizontal offsets of points and elevation (vertical) biases relative to the filtered reference. Prior to co-registration, the TDX DEM is filtered using the auxiliary products bundled with the DEM tile to mask the artifacts/errors that persist over mountainous terrain, water bodies and areas with limited overlapping InSAR-derived DEM strips. The masking functionality is documented by Shean (2019; https://github.com/dshean/tandemx). It intends to preserve pixels with limited errors (ideally <1 m horizontal and vertical, especially for planar surfaces) and thus provides an excellent reference DEM to be used for robust co-registration of other DEM products.

The coregistration process consists in identifying static control surfaces that I assume did not change between the DEM timestamp and the reference DEM timestamp, and aligning each DEM to the reference TDX DEM. I selected polygons over areas with slopes < 5 degrees, and with low vegetation (i.e., grassland or farmland) or bare soil for increased accuracy. Then both the input DEM and the reference DEM were clipped to a common intersection and resampled to 30-m resolution using bicubic interpolation. The outlier elevation change values over the static surfaces were identified and removed. Then I used the Nuth and Kääb (2011) method with robust filtering and outlier removal. If the resulting translation exceeded 200 m or if more than 30 iterations were required for alignment, the DEM was excluded from the final dataset. For all other DEMs, a translation was applied to the DEM and the median vertical bias was removed. I then quality-filtered the full set of co-registered (aligned) DEMs. Using statistics on error metrics from all individual DEMs before and after co-registration, I identified outliers and removed DEMs with anomalously high residual bias (Figure 6.3).
Figure 6.3. Temporal coverage and error metrics for ASTER DEMs at Volcan de Colima and Calbuco Volcano, before (a, c) and after (b, d) co-registration. Each point shows the median elevation offset between a single DEM and the reference TanDEM-X over static surfaces. The error bars indicate the 16-84% spread of the elevation offset values. The orange points/bars ("inliers") represent the final set of DEMs used for trend-fitting.
Figure 6.4. ASTER DEM time-series at Volcán de Colima between 2004 and 2019. a) Trends of elevation change in meters/year. b) Uncertainty elevation map calculated on the DEM stack using Normalized Median Absolute Deviation (NMAD) for each pixel. c) DEM count per pixel used to compute each trend value.

Figure 6.5. ASTER DEM time-series at Calbuco Volcano between 2010 and 2019. a) Trends of elevation change in meters/year. b) Uncertainty elevation map calculated on the DEM stack using Normalized Median Absolute Deviation (NMAD) for each pixel. c) DEM count per pixel used to compute the trend value.

Trends of elevation change

I generated two time-served stacks of co-registered and filtered ASTER DEMs for Volcán de Colima and Calbuco Volcano at 30-m resolution. Using these stacks, I computed the per-pixel linear elevation change through time (dh/dt in meters per year, m/yr). For each pixel within the volcano extent, the changes were computed when a minimum of 5 measurements with a time span
of 3 years between the first and last measurement was available. I used the Thiel-Sen estimator to evaluate trends in elevation changes, which is robust for noisy data. Finally, the elevation change trend maps were filtered for residual outliers and to smooth the output trend maps using a 3x3-pixel median filter and a 3x3-pixel Gaussian filter. The final products consisted of the elevation change trend map at each volcano (Figures 6.4a and 6.5a), the elevation uncertainty maps (Figures 6.4b and 6.5b), and the total DEM counts (Figures 6.4c and 6.5c).

The final time-series stacks were composed of 190 DEMs between 2004 and 2019 at Volcán de Colima and 62 DEMs between 2010 and 2019 at Calbuco Volcano. The lower total DEM count at Calbuco Volcano is due to the frequent cloud cover for more than 50% of the images. I only selected the time-period between 2010 and 2019 because the eruption of April 2015 was the only activity reported at the volcano since 1972.

**Fusion of HRSI Optical Data for DEM Generation**

**Volcán de Colima**

The HRSI data included one pre-eruption stereopair from Pleiades-1A satellite at 0.5-m resolution acquired on April 8, 2013, one post-eruption stereopair from SPOT-6 satellite at 2-m resolution acquired on July 25, 2015 (i.e., 15 days after the PDC-forming eruption), and a post-eruption trio from Pleiades-1A at 0.5-m resolution from January 10, 2016. For each dataset, the corresponding DEMs were generated using stereogrammetry processing with PCI Geomatics 2018 OrthoEngine on the panchromatic bands only for highest resolution. I used the existing RPC associated with each scene's metadata file in the Rational Function math model. The data were projected in UTM 13N WGS84 projection.
Figure 6.6. Fusion of very-high resolution DEMs at Volcán de Colima. The left and middle panels are the hillshades of the Pleiades-1A and SPOT-6 DEMs respectively, and utilized for DEM fusion to obtain a complete high-resolution DEM of the south flank of Volcán de Colima, shown on the third panel (to the right).

I used the automatic tie point collector for co-registration of the left and right image, and I manually removed the points with high RMSE. I also manually added ground control points (GCPs) to improve registration. Selecting the proper left and right images, I generated the Epipolar Image and extracted DEM automatically using SGM extraction method with a sampling interval of two times the image resolution (e.g., 1-m DEM resolution for 0.5 m image resolution). I applied a low smoothing filter and used epipolar tracking. Finally, I generated a geocoded DEM. The left and right images were then orthorectified using this geocoded DEM.

The post-eruption DEMs were post-processed to merge both SPOT and Pleiades DEM in order to obtain full coverage of the southern flank of the volcano. The presence of clouds on the
SPOT DEM resulted in high elevation errors towards the summit. Using the “score channel” produced in the Geomatica Software during the DEM extraction process, which represents the correlation score for each pixel, I determined the area to be masked out on the SPOT DEM.

Both DEMs were co-registered to a filtered reference TDX DEM similar to the ASTER DEMs processing. In contrast to ASTER processing, I used a mask for non-static surfaces, which I estimated using the eruptive history for the period between 2016 and the reference TDX DEM timestamp (~ 2011 – 2014).

Then I merged the post-eruption Pleiades and SPOT DEM by converting them into point clouds, re-aligning the clipped SPOT DEM to the Pleiades DEM and using weighted average values for the overlapping area. I used weights of 0.2 and 0.8 for SPOT and Pleiades elevation values respectively as per the relatively low correlation values obtained for the SPOT DEM during epipolar image tracking. The final product is a near continuous 5-m DEM that covers the whole southern flank of the volcano (Figure 6.6).

**Calbuco Volcano**

The pre-eruption DEM was generated with a HRSI stereopair from Worldview-1 acquired on May 22, 2014. The same processing steps described for ASTER stereopairs were applied on the panchromatic bands, and the resulting DEM was generated at 4-m posting. I used a stereopair from Pleiades-1A acquired on December 1, 2015, to generate a post-eruption DEM at 1-m resolution.

The same processing steps in PCI Geomatics software described for the Pleiades imagery of Volcán de Colima were applied here. I note that because the pair of images used to generate the post-eruption DEM were acquired in December 2015 when lahar erosion of the deposits was
strong, the modelled surface includes the erosional features carved into the deposits and volume estimates from DEM differencing will be a minimum estimate.

**Topographic change retrieval**

I derived topographic changes from the HRSI DEM time-series stack generated at both Calbuco and Colima volcanoes. In a similar manner as described for ASTER DEM time-series, I first coregistered the HRSI DEMs for each site using polygons of static control areas (i.e., areas where no changes occurred between the two acquisitions of optical images) and resampled to a common resolution of 5 m. Then elevation values of the pre-eruption DEMs were subtracted from the post-eruption ones, such that positive values indicate increases in elevation whereas decreases in elevation are represented by negative values. The alignment (or co-registration) process allows to reduce the biases in elevation differences caused by either sensor geometry or processing uncertainties. The HRSI elevation difference maps for Volcán de Colima and Calbuco Volcano are shown in Figure 6.7. I describe topographic changes for volcanic features mapped using a combination of field observations and remote sensing data from prior and ongoing studies of these two eruptions (Macorps et al., 2018; Chapter 5). Mainly, the topographic changes are described for the summit area, which includes the crater/lava dome and lava flows (only for Volcán de Colima), and the main valleys.

For estimates of lava dome volumes at Volcán de Colima, I use simple geometric volume of an oblate ellipsoid:

\[ V = \frac{4}{3} \pi b^2 c \]

where \( b \) is the is the equatorial radius (half-width) and \( c \) is the polar radius (crater depth). The results are divided by 2 to get a dome with a flat underside as commonly observed for lava domes.
Figure 6.7. HRSI elevation difference map at a) Volcán de Colima and b) Calbuco Volcano calculated between before and after their respective PDC-forming eruption. I note the greater range of elevation differences at Calbuco Volcano from the greater eruption intensity.

**Uncertainty Estimates**

I estimated the elevation change error for each study site. To do this, I assumed that static/stable surfaces should have an elevation difference of 0, and then determined the random and systematic error component to obtain the total error. Even after removing the bias, there were remaining elevation differences in the stable surfaces that need to be accounted for. For each polygon representing a static surface, the mean and standard deviation of elevation differences was computed. The random error is estimated from the standard deviation of elevation differences for each static polygon, while the systematic error (or bias) is given by the mean of elevation
differences over the same static surfaces. Then the total error can be computed as the root-mean-square error (RMSE) as:

$$\sigma_{dh} = \sqrt{\sigma_{random}^2 + \sigma_{systematic}^2}$$

For the map of elevation difference at Volcán de Colima obtained from the pre-eruption Pleiades and the post-eruption SPOT-Pleiades fusion DEM difference, after bias removal, the remaining total measured error was calculated at ±1.5 m. For Calbuco volcano, the total measured error was estimated at ± 1.6 m. For the ASTER DEM time-series, the uncertainty at Calbuco Volcano is ±0.3 m/yr, and ±0.1 m/yr at Volcán de Colima.

Results and Interpretations

Volcán de Colima

Summit area – lava flows

The larger morphological changes of the volcano on the ASTER DEM time-series were observed in the summit area on the W and S flanks. The elevation trend map shows an increase of up to +2 m/yr on the SW flank, +1.5 m/yr on the W flank and +1 m/yr on the S flank. The trends of elevation increase with time on the W and S flanks are indicative of the main areas of emplacement of lava flows for the period of 2004 to 2019. Elevation trend of +0.5 m/yr on the N flank indicates that lava flows descended on the N flank as well although for a shorter time. The decreasing elevation trend of -0.3 m/yr towards the south can be attributed to the rim collapse that occurred during the July 2015 eruption (Capra et al., 2016; Macorps et al., 2018, Davila et al., 2019).

I calculated the statistics of elevation trends for the summit area, using a circle of 2 km radius from the summit, to estimate volume change. I found a mean value of +0.4 m/yr with a
standard deviation of ±0.2 m/yr, for an area of 12.5 \(10^6\) m². This yields a volume change of 5.0 ± 2.5 \(10^6\) m³/yr. The order of magnitude for volume change is comparable to volumes calculated for the single eruption of July 2015 (i.e., 7.7 \(10^6\) m³ estimated from field and digital mapping, Macorps et al., 2018).

![2013 - 2016 Elevation Difference (m)](image)

**Figure 6.8.** Pre- to post-eruption HRSI elevation difference map focused on the summit area of Volcán de Colima. The footprint of the three lava flows mentioned in the text are shown with the dotted lines.

The HRSI elevation difference map in the summit area also shows increased elevation values on the S and W flanks (Figure 6.8), in agreement with the positive elevation trends observed
with ASTER DEM time-series. These positive changes reflect the presence of lava flows that began shortly after the beginning of a new dome growth in early 2013. By April 2013, which is the acquisition date of the pre-eruption Pleiades image, the dome had filled the crater and a dome-coulee was recorded on the W flank. In May 2014, the progression of the W lava flow continued and a 2.4 km-long lava flow descended on the SW flank in late September 2014 due to an overflowing dome. The 2013 dome was destroyed during a series of explosions in November-December 2014, which also generated a PDC that descended 3.1 km in the San Antonio ravine. W and WNW lava flows were reported in early 2015. The newly built lava dome was overflowing the crater on the south rim a month before the paroxysmal eruption. The lava flow was reportedly advancing fast during early 2015 (Davila et al., 2019).

The elevation difference map reflects the total changes that resulted from these lava flows. Looking at each flow separately, I compute the statistics of elevation change. The SW lava flow presents the largest increase of elevation with a maximum change of $65.9 \pm 1.5$ m towards the lava flow front and an average surface increase of $23.4 \pm 12.1$ m. The W lava flow is characterized by a maximum elevation increase of $56.4 \pm 1.5$ m with an average increase of $22.6 \pm 14.3$ m comparable to the SW lava flow. The S lava flow shows a lower increase with a maximum elevation change of $43.3 \pm 1.5$ m and an average increase of $12.3 \pm 7.7$ m.

I note that the area immediately below the crater rim to the south is characterized by negative elevation differences, similar to the negative trend of elevation change observed on ASTER DEM time-series. This area corresponds to the portion of the rim and flank that collapsed during the July 2015 eruption. The negative elevation differences where the beginning of the lava flow would be expected suggest that part of the lava flow that descended the south flank during May–June 2015 was part of the collapse events that generated the block-and-ash flows.
The increase of elevation values on the W flank between the W and the SW lava flows, and on the N flank, both with an average of $8.8 \pm 5$ m are most likely caused by the frequent rockfalls and short-distance PDCs that occurred in late 2014, destroying the 2013 dome. I use the elevation difference and the area to compute volumes for the three lava flows (Table 6.1). I found a volume of $13.4 \pm 0.8 \times 10^6$ m$^3$ for the SW lava flow, $6.1 \pm 0.4 \times 10^6$ m$^3$ for the W lava flow, and $5.8 \pm 0.7 \times 10^6$ m$^3$ for the S lava flow.

**Table 6.1.** Comparison of estimated volumes for the three lava flows with results from Davila et al. (2019)

<table>
<thead>
<tr>
<th>Lava Flows</th>
<th>This study</th>
<th>Davila et al. (2019)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Area ($10^5$ m$^2$)</td>
<td>Volume ($10^6$ m$^3$)</td>
</tr>
<tr>
<td>W Lava Flow</td>
<td>2.7</td>
<td>6.1 ± 0.4</td>
</tr>
<tr>
<td>SW Lava Flow</td>
<td>5.7</td>
<td>13.4 ± 0.8</td>
</tr>
<tr>
<td>S Lava Flow</td>
<td>4.7</td>
<td>5.8 ± 0.7</td>
</tr>
</tbody>
</table>

The volumes calculated for the W and SW lava flows have comparable values to those estimated by Davila et al. (2019), and the comparison of these results is presented in Table 6.1. The higher volumes obtained with our elevation difference map may be due to DEMs of higher resolution (i.e., 5-m resampled) compared to 10 to 20-m resolution used by Davila et al. (2019). The volume estimate for the S lava flow is lower by a factor of 3 while the area is smaller by a factor of 2, which I attribute to two factors, (1) the collapse of a portion of the 2015 lava flow during the July events removing a portion of the flow that Davila et al. (2019) could resolve with an intermediate date in the time-series, and (2) the fact that the flow continued spreading after January 2016. Davila et al. (2019) used a DEM from April 2017 where a larger covered area was detected. This is in agreement with both the thermal and radar data from 2016 and 2017 on which I can detect the S lava flow with a larger footprint (*see Chapter 5*).
Lava dome

In the ASTER DEM time-series, the increase of 1 m/yr immediately on the dome and towards the W can be associated with the general trend of dome growth and the dome coulee that began in April 2013. The general increase in elevation observed in the dome area suggests that for the period between 2004 and 2019, the volcano experienced more episodes of dome growth rather than dome collapse or destruction during eruptive events.

In Figure 6.9a, I show the elevation differences between the 2013 HRSI DEM and another pre-eruption DEM generated from airborne LiDAR data acquired in 2005 (courtesy of L. Capra). The 2005 LiDAR DEM is a surface model (i.e., includes elevation of the canopy) at 5-m resolution. The elevation difference between the 2005 and 2013 DEMs (Figure 6.9a) shows an increase in elevation up to 23 m inside the crater, and up to 25 m on the W flank. These positive changes in elevations correspond to a lava dome formed in 2013 and a dome coulee caused by the overflowing dome after filling the crater (Figure 6.9c).

A decrease in elevation is observed along the east side of the crater, likely caused by the enlargement of the crater area during the period of 2005 to 2013. The 2013 lava dome structure observed on Figure 6.9c was later destroyed during the November-December 2014 explosions, hence shows a different lava dome than the 2015 dome that collapsed during the July 2015 events. Therefore, the elevation difference cannot be used directly to derive the volume of the lava dome that collapsed to generate the block-and-ash flows.

I can, however, estimate the volume that used to fill the wide empty crater shown by the elevation changes between 2013 and 2016 (Figure 6.9b). The negative elevation difference to the south reflects the collapse of the crater rim, and the resulting shape at the summit is an opened crater to the south (Figure 6.9d).
Figure 6.9. HRSI map of elevation differences between 2005 and 2013 at Volcán de Colima a) showing the 2013 lava dome and dome coulee to the west with high positive elevation changes and b) showing the excavated crater and crater rim after the July 2015 dome-collapse. c) 3D view of the 2005 - 2013 elevation changes overlaying the 2013 topography showing the lava dome and coulee. d) 3D view of the 2013 - 2016 crater and flank excavation towards the south overlaying the 2016 topography.

The crater shape left after the collapse of the 2015 dome has dimensions of 200 x 250 m with a depth of 60 m from the crater floor to the rims. Photographs acquired on February 4, 2015, shows that material filled the crater up to the rims, and reports from the observatory and emergency officials a few days prior to the collapse noted a rapid growth with a dome overtopping the crater rim and descending onto the southern flank (GVP, 2015). The elevation difference between the summit of the 2013 lava dome and the post-collapse 2015 crater is -100 ± 1.5 m. This would have
resulted in a dome height of 40 m above the crater rim, providing an upper boundary for volume estimate.

The comparison between a single ASTER DEM from the time-series acquired on April 2015 and the post-eruption DEM from January 2016 suggests a maximum elevation difference of -90 m for a dome dimensions of 200 x 150 m, which is closer to the measurements obtained from aerial photogrammetry by Thiele et al. (2017). I use these dimensions as the lower volume boundaries in the volume formula of oblate ellipsoid as described in the methodology section, and I obtain a minimum volume of $1.8 \times 10^6$ m$^3$ and a maximum volume of $2.9 \times 10^6$ m$^3$. The estimates are higher than the volume of $1.2 \times 10^6$ m$^3$ obtained by Thiele et al. (2017) using photogrammetry on aerial DSLR photographs.

I also used the elevation difference map between the ASTER DEM and the post-eruption high-resolution DEM to extract a volume estimate and I obtained $1.9 \times 10^6$ m$^3$, which is close to the lower bound volume estimated using the oblate ellipsoid approximation. Similarly, calculating the volume of excavated material using the 2013 – 2016 elevation difference map yields $2.4 \times 10^6$ m$^3$, which is also in the range of values obtained with geometric approximation. I only used a polygon covering the crater floor and did not account for the elevation change of the crater rims.

Using the elevation difference map between the 2005 LiDAR DEM and the 2013 DEM (Figure 6.9a) I can approximate the volume of the 2013 dome as of April 2013 with the associated dome coulee. I measure a volume of $0.4 \times 10^6$ m$^3$, which is 1.25 times lower than the volume estimates made by Thiele et al. (2017) with thermal and DSLR optical aerial data acquired in June 2013 who obtained a volume of $0.5 \times 10^6$ m$^3$. The difference between the two estimates may be the results of the employed method, the two-month difference between image acquisitions, or a combination of these two factors.
Montegrande and San Antonio ravines

The elevation trend map from ASTER time-series shows a general decrease in elevation of -0.3 m/yr on average along the Montegrande and San Antonio ravines. This can be attributed to different factors: first erosive behavior by multiple lahar events recorded over the years contributed to decreasing surface elevation despite the channel infilling during the 2015 block-and-ash flow events. Additionally, the removal of vegetation following PDC events may also have contributed to decreasing elevations. There is a portion of the Montegrande ravine, in the distal area towards the foot of the volcano that displays elevation trends of up to +1.7 m/yr. This area was greatly filled by the 2015 block-and-ash flows, and I therefore interpret the detected trend as resulting from the deposits.

On HRSI DEM time-series, the changes inside and along the Montegrande and San Antonio ravines are more complex. I observe both increasing and decreasing elevation values that are the results of the block-and-ash flows descent on July 10-11, 2015. DEMs generated from optical imagery are by nature surface elevations rather than terrain models, which means that the elevation values recorded include any objects that lie on a given surface, including vegetation and man-made structures. The areas of negative elevation changes around the channels are attributed to vegetation removal. This was confirmed by field work where deposits were found in those areas with knocked down trees either within or along the deposits and from HRSI data (Figures 6.10b and 6.10c). The increase in elevations inside the channels however are caused by the PDC deposits, although it is important to note that erosion by streams and subsequently lahars occurred prior to the image acquisition used for the post-eruption DEMs for the proximal area. The increase of elevation values inside the Montegrande ravine suggest deposit thickness of tens of meters in most areas.
Figure 6.10. Pre- to post-eruption HRSI elevation difference map at Volcán de Colima focused on the Montegrande ravine. a) 3D view of the HRSI elevation differences between 2013 and 2016 at with two polygons indicating the areas of interest for comparison between pre- and post-eruption changes. b) and c) show pre- and post-eruption HRSI of the intersection between the Montegrande and San Antonio ravines to highlight the destruction of canopy from the deposition of PDCs. d) and e) show pre- and post-eruption HRSI in the distal part of the Montegrande ravine where high values of elevation changes were observed on both ASTER and HRSI DEM time-series from the significant infilling of the valley.

The largest elevation increase is observed in the most distal part of the Montegrande ravine just before the channel opening onto the alluvial fan. The elevation difference is between 20 and 25-m (Figures 6.7a and 6.10a). This change coincides with the observation made on the ASTER DEM time-series where a trend of +1.7 m/yr was detected. Considering a time-series over 15 years, the trend would result in a total elevation difference of +25.5 m. Although the trend observed with
ASTER is non-linear and occurred during a single event, the ASTER time-series was able to detect areas of significant topographic changes. Using visual analysis of the pre- and post-eruption images that were used to generate the respective DEMs, I confirm the significant infilling of the channel in the distal portion of the Montegrande ravine covering the lower banks and portion of the trees (Figures 6.10d and 6.10e). Measurements of tree heights on the field yielded an average height of 15 m. Given that most of the trees were covered and the fact that they were standing on a small bank tends to validate the estimates for deposit thickness of 25 m in the center of the channel. Despite field observations of PDC deposits in the San Antonio ravine, the elevation difference map did not capture these surface changes (Figure 6.7a).

**Calbuco Volcano**

**Summit area**

I analyze the trends of elevation change and I interpret volume changes given the 10-year period covered by the time-series. The area at the summit of Calbuco volcano shows a mean decreasing trend of -7.8 m/yr, which would yield a total volume change of \(-39.1 \pm 1.5 \times 10^6\) m\(^3\). The directionality of the area towards the NE is in agreement with the morphological changes observed after the April 2015 eruption. The eruption resulted in an amphitheater-like depression towards the NE and a new crater opening in its center (Figure 6.7b). The southern side of the summit is characterized by a decreasing trend of \(-2.9 \pm 1.7\) m/yr on average. This depression was most likely caused by the removal of the summit glacier during the eruption as well. The volume change for ten years would be \(38.5 \pm 4.0 \times 10^6\) m\(^3\). The remaining portion of the summit displays positive elevation trends with an average of \(+1.4 \pm 1.8\) m/yr, resulting in a volume change of \(+12.3 \pm 2.6 \times 10^6\) m\(^3\).
Using HRSI, the changes at the summit also show the new crater created by the April 2015 eruption within an amphitheater-shaped depression opened to the ENE (Figure 6.11). The maximum decrease in elevation values was -105.2 m. With the area of negative elevation difference, I estimated a volume difference of \( 29.6 \pm 0.8 \times 10^6 \) m\(^3\). The style of the eruption at Calbuco was very different from that of Volcán de Colima and the volume change at the summit does not reflect the erupted materials, but merely helps to describe the topographic changes experienced during the eruption. The estimated volume is lower than that obtained from ASTER DEM time-series by \( 10 \times 10^6 \) m\(^3\) which could be due to averaging the trend of elevation changes and the fact that the trends are linearized rather than detected for a particular event.

**Figure 6.11.** Pre- and post-eruption elevation difference map focused on the NE flank of Calbuco Volcano. a) 3D view of the HRSI elevation differences between 2014 and 2015, overlaying the 2015 topography. b) Photograph of the summit area taken during field work in December 2016. The numbered arrows indicate the topographic features referenced on the elevation change map in a). c) Photograph of the exposed PDC deposits in the distal part of the Rio Blanco Frio – Este taken during field work in December 2016. Field measurements indicated a total thickness close to 20 m. The numbered arrows refer to the same summit features at images a) and b).
Valleys

Both the ASTER DEM time-series and the HRSI elevation difference map show positive elevation changes towards the NE that is in agreement with the directionality of the eruption (Figures 6.5a, 6.7b and 6.11a). The increase of elevations with time indicates areas of deposition, and I compare the polygons of PDC deposits mapped with a combination of remote sensing and field methods to evaluate the changes.

The Rio Blanco Este presents a large elevation trend in the proximal area between 1 and 2.5 km from the crater with a maximum trend at +12 m/yr and an average increase of +6.6 ± 3.1 m/yr. This is also observed on the HRSI elevation maps with the largest elevation increase observed between 1.1 and 2.5 km from the crater with a maximum elevation difference of +130 m. The comparison between pre- and post-eruption HRSI optical data confirms the infilling of the channel (see Chapter 5).

In the more distal part of the Rio Blanco Frio – Este, the elevation trend is +1.1 ± 0.9 m/yr on average on the ASTER time-series, while the HRSI elevation difference gives an average of +20 ± 5 m, which is in agreement with field observations. During a field campaign carried out in December 2016, the carving of the PDCs by lahars in the Rio Frio – Rio Blanco Este exposed more than 20-m thick deposits (Figure 6.11c).

The results from field and digital mapping of the PDC deposits in the Rio Frio – Rio Blanco Este terminate 6.7 km from the crater (straight-line method). The positive trend of elevation changes in the Rio Frio – Rio Blanco Este on both the ASTER DEM time-series and HRSI elevation difference map is observed beyond the extent of the polygons for PDC deposits, with an average elevation difference of +15 ± 5 m. The increase in elevation beyond the end of the PDC deposits reflects the thickness of the lahars that descended during and immediately after the
eruption, carving into the fresh and unconsolidated PDC deposits and remobilizing the materials further downslope. Deposit mapping from radar and optical images tracks the lahars up to 16.7 km from the crater where they reached the riverbed.

Figure 6.12. Pre- and post-eruption elevation difference map focused on the Rio Blanco Sur at Calbuco Volcano. a) 3D view of the HRSI from 2015-04-30 (i.e., 7 days after the deposition of PDCs) overlaying the 2016 topography. b) Zoom over the lobate PDC deposits in map view, corresponding to the polygon in a). c) 3D view of the HRSI elevation difference map focused on the Rio Blanco Sur with the same polygon area of the lobate PDC deposits as reference. The HRSI elevation difference map shows increased elevation values where the deposits are located. d) Photograph from within the Rio Blanco Sur showing the lobate PDC deposits and the summit area. The arrows on images a, c and d indicate the channel opening towards the Rio Blanco Sur from the summit.
The Rio Tepu also exhibits positive elevation trends in the distal part of the valley whereas the proximal part is very noisy. The positive trend is estimated at $3.6 \pm 3.1$ m/yr on average, but I note that the positive trend is offset from the deposit polygons and may be caused by the steep valley walls rather than topographic changes from deposition. The HRSI elevation difference map shows positive values suggesting deposition up to 4.3 km from the crater, although deposit mapping yielded a runout of 5.9 km. The negative values observed in the distal portion of the Rio Tepu is attributed to lahar erosion of the deposits. The elevation difference are up to +43 m in the proximal area of the channel and between 15 and 20 m in the narrow part of the channel.

The Rio Blanco Norte and the N- and S-branches of the Rio Sur did not show clear trends of elevation changes on the ASTER DEM time-series. The proximal area of the Rio Blanco Norte shows elevation trend with an average of $+0.2 \pm 0.3$ m/yr. HRSI elevation differences suggest deposit thickness between 10 and 15 m. In the Rio Sur, the lack of proper signal for elevation trends related to deposits can be interpreted from the narrow valleys with steep and high walls that resulted in shadows and larger uncertainties at coarse resolution. HRSI data show elevation differences at $+8.1 \pm 2.8$ m in the N-branch, and $+5.8 \pm 3.9$ m in the S-branch. The S-branch shows maximum thickness of 19 m in the proximal area and 12 m in the distal area.

In the Rio Blanco Sur, the average elevation trend for the lobate deposits on ASTER DEM time-series is $+0.0 \pm 0.8$ m/yr. The HRSI data gives a maximum thickness of 12-m that is consistent with field measurements (Figure 6.12). The proximal area of the valley exhibits larger trend values of $+1.9 \pm 1.8$ m/yr and the HRSI elevation difference map gives a maximum values of $+30$ m. This area was not accessible on the field for measurements to compare with these values. Polygons of concentrated PDC deposits determined from deposit mapping (Figure 6.1) were used to estimate
volumes of deposits, from both ASTER DEM time-series and HRSI elevation differences, which are summarized in Table 6.2.

### Table 6.2. Volumes of PDC deposits in the valleys at Calbuco Volcano from high-resolution topographic change maps and from ASTER DEM time-series

<table>
<thead>
<tr>
<th>Valleys</th>
<th>Volume (x 10^6 m^3)</th>
<th>Uncertainty (x 10^6 m^3)</th>
<th>Volume from ASTER DEMs (x 10^6 m^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rio Tepu</td>
<td>5.6 ± 0.9</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>N-2</td>
<td>4.4 ± 0.8</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Rio Sur – N branch</td>
<td>1.0 ± 0.2</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>W-2</td>
<td>0.7 ± 0.3</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Rio Sur – S branch</td>
<td>0.4 ± 0.1</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Rio Blanco Sur – Lobate deposits</td>
<td>1.0 ± 0.3</td>
<td>1.5 ± 0.5</td>
<td></td>
</tr>
<tr>
<td>Rio Blanco Sur – Proximal deposits</td>
<td>2.1 ± 0.4</td>
<td>5.8 ± 0.9</td>
<td></td>
</tr>
<tr>
<td>Rio Blanco Frio – Este</td>
<td>9.2 ± 1.0</td>
<td>10.9 ± 0.2</td>
<td></td>
</tr>
<tr>
<td>Rio Blanco Frio – Proximal deposits</td>
<td>36.3 ± 1.6</td>
<td>45.2 ± 2.0</td>
<td></td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>60.7</strong> ± <strong>5.6</strong></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The remaining channels to the NE in the proximal area and the area around the crater also show increased elevation (Figure 6.5a and 6.7b), although our analysis from deposit mapping did not permit us to identify the nature of deposits. I could not differentiate dilute PDCs from tephra fallout and from concentrated PDCs.

I estimated the overall volume from the proximal area using positive elevation differences and obtained 39.5 ± 4.0 x 10^6 m^3. In addition to the volume estimated from concentrated deposits, this brings a total deposit volume of 100.2 ± 9.6 x 10^6 m^3. I note, however, the presence of a glacier at the summit during acquisition of the Pleiades stereo pair used to generate the DEM, which potentially resulted in elevation increases of greater amplitude than that caused by eruptive products.

The volume for the distal portion of the Rio Frio – Rio Blanco Este is very close to the estimate from the ASTER DEM time-series (Table 6.2). The large volume increase in the proximal
part of the Rio Blanco Este is also comparable to the large positive trends of elevation change observed with the ATER DEMs. For instance, a maximum of 12 m/yr for ten years would result in an elevation difference of 120 m, which is in the range of values detected with DEM difference.

The volume obtained from high-resolution DEM is lower than the one estimated from ASTER DEM time-series. The averaging trend values to compute a volume change may explain the higher volume. In the Rio Blanco Sur, the volume of lobate PDC deposits is within uncertainty of the volume obtained from ASTER DEMs (Table 6.2). The proximal part of the Rio Blanco Sur, however, returned a volume that is half the one obtained from ASTER DEMs. The steep slopes and high relief on the proximal areas below the summit are complex to be resolved on optical data and may explain the difference.

**Discussion**

**Retrieving Volume of Erupted Products**

The limitation for computing linear trends of elevation changes is that eruptive activity is rarely, to not at all, linear. Using the ASTER time-series however, helps determine the morphological changes that have occurred at volcano-scale, and provides information on the main flanks that have been affected by eruptive events over the year. This information is critical for refining hazard maps with respect to the surrounding villages.

At Volcán de Colima, the HRSI elevation difference map is used to estimate the total deposited volume of PDCs, including the valley-confined and overbank flow deposits when the elevation difference was positive. Most of the overbank deposits appear as areas of negative elevation difference as a result of vegetation removal and therefore could not be included in the calculations. The proximal area of deposition is covered with the S lava flow that advanced after
the emplacement of the 2015 block-and-ash flows, and therefore, no volume estimates of the deposits were obtained for the first 2 km from the summit. Because the slopes in the proximal area are $> 35^\circ$, however, I can assume that there was little to no deposition and conclude that the volumes would be within the uncertainty.

The total volume obtained from the elevation difference map is $4.3 \pm 1.1 \times 10^6$ m$^3$. This volume is lower than previous estimates from Macorps et al. (2018), which is partly due to the omission of overbank flow deposits along the Montegrande ravine and in the San Antonio ravine in the calculations. It is, however, similar to the volume estimated by Davila et al. (2019). The volume of overbank deposits previously estimated from field investigations (Macorps et al., 2018), was $1.9 \pm 1 \times 10^6$ m$^3$. Combining these volumes, I obtain a total volume of $6.2 \pm 1.1 \times 10^6$ m$^3$.

The volume of excavated material from the crater obtained with HRSI time-series includes the removal of the south rim and is still 2.5 times lower than the volumes of PDC deposits. The elevation difference map reveals a depression on the south rim where the lava flow that had begun prior to the July 2015 events would have been emplaced. This suggests that part of the flank and the lava flow collapsed during the eruption and contributed to the added volume.

Computing the total volume of material removed from both the crater and the upper part of the south flank yields $3.6 \pm 0.2 \times 10^6$ m$^3$. The estimate is still lower than the total volume of PDC deposits, which can be attributed to the absence of the lava flow thicknesses into the calculations, since the pre-eruption DEM was generated prior to the descent of the flow. These findings is in agreement with the results from a study by Pensa et al. (2018) who determined the emplacement temperatures of the block-and-ash flows using the charcoal reflectance and partial thermal remnant magnetization analyzes. Their study differentiated two types of lithic clasts (i.e., non-juvenile materials) based on paleomagnetic behavior that reflected cold and hot clasts from
two origins. Their conclusion was that it was from two different portions of the dome, i.e., the hot inner dome versus the cold outer shell, but I suggest that it reflected the difference between rim / flank materials (cold) and lava dome / lava flow materials (hot) instead. The results from these data are important because they revise previous interpretations for the origin of the July 2015 block-and-ash flows that only accounted for lava dome and crater rim collapse, and omitting the lava flow collapse.

At Calbuco Volcano, the total volume of concentrated PDCs 0.0607 ± 0.0056 km$^3$ obtained from deposit mapping and the elevation difference map is likely underestimated because of the lahar erosion. Van Eaton et al. (2016) suggested a PDC volume of 0.1 km$^3$, which, by adding the volume from the summit is comparable to our results. The volume from the Rio Frio – Rio Blanco Este was estimated at 0.038 km$^3$ by Castruccio et al. (2016) who considered an average deposit thickness of 15 m for an area of 0.65 km$^2$. This estimate is comparable to our estimate, although greater, of 0.045 ± 0.0026 km$^3$ for an area of 1.7 km$^2$ considering the proximal area of deposition and a side channel towards the E.

Although no pre-eruption deformation was measured by interferogram up to one and a half day before the eruption, a ~ 12 cm co-eruptive deflation was measured by the Sentinel-1 interferogram on 26 April, corresponding to a volume change of 0.054 km$^3$ about 5 km SW of the volcano and at a depth of 9.3 km (Delgado et al., 2017). Our volume estimate focuses on PDC deposits, and the uncertainties include the possibility of having included tephra fallout, or having missed dilute deposits in the calculations. I compare our results with the estimates for total erupted volume from other studies in Table 6.3. Overall, the estimated volume of PDC deposits represents ~ 20% of the total erupted volume during the Calbuco eruption.
The relative agreement between field measurements of deposit thickness and the elevation difference from DEM difference suggests that our method for computing topographic changes may be quite robust, although significant noise is detected for steep valley walls and in the narrow channels, similarly to the San Antonio ravine at Volcán de Colima.

**Table 6.3. Summary of total volume erupted from the Calbuco eruption in different studies.**

<table>
<thead>
<tr>
<th>Volume ($\text{km}^3$)</th>
<th>Applied method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.38</td>
<td>Weibull law on tephra thinning</td>
<td>Castruccio et al. (2016)</td>
</tr>
<tr>
<td>0.28</td>
<td>Exponential law on tephra thinning</td>
<td></td>
</tr>
<tr>
<td>0.29</td>
<td>Power law on tephra thinning</td>
<td></td>
</tr>
<tr>
<td>0.28</td>
<td>Umbrella Cloud Expansion</td>
<td>Van Eaton et al. (2016)</td>
</tr>
<tr>
<td>0.58 ± 0.28*</td>
<td>Exponential tephra thinning</td>
<td></td>
</tr>
<tr>
<td>0.27</td>
<td>Tephra thinning (Weibull, exponential, and power law)</td>
<td>Romero et al. (2016)</td>
</tr>
</tbody>
</table>

* The 50% uncertainty is associated with the local effects of deposit compaction, wind remobilization and plausible range of assumptions associated with the calculations.

**Uncertainties in Topographic Changes**

The elevation trends in the summit area needs to be considered carefully as the uncertainty map shows that greater elevation errors were observed (Figures 6.4b and 6.5b). This is due to the lower DEM count in the summit (Figures 6.4c and 6.5c), which is a result of the cloud removal process during stereogrammetry. Cloud cover at the summit is a common issue for optical datasets at Volcán de Colima and Calbuco Volcano. Moreover, the continuous eruptive activity at Volcán de Colima with small explosions that produced ash plumes also contributed to errors during the stereogrammetry processing. The steep relief of the valleys was also affected by higher elevation uncertainties as a result of shadows that limits the performance of point matching for two stereo images. This effect was also observed on the rugged relief to the E of Calbuco Volcano that resulted in a noisy signal on the ASTER DEM time-series (Figure 6.5).
At Volcán de Colima, the generally low elevation uncertainties for the Montegrande ravine on the S flank is encouraging for looking at topographic changes. At Calbuco Volcano, the elevation trend map and the HRSI elevation difference map show that surface changes are generally focused on the NE side of the volcano, while the surroundings are close to 0, which is in agreement with the directionality of the eruption on April 2015.

For DEMs generated from optical data, the heights of vegetation are accounted for in the surface elevation values. DEM time-series are therefore affected by changes in the vegetation pattern, which can add noise when looking at fine-scale topographic changes such as erosion inside the valleys or deposition of PDCs.

**Implication of Topographic Changes for Hazard Assessments**

Topographic changes of a volcano influence the directionality of PDCs. The dimensions of a lava dome and the areas of lava dome growth can influence the direction of collapse (Calder et al., 2002; Watts et al., 2002; Stinton et al., 2014) and crater walls can be eroded, breached or collapsed (Stinton, 2007). This was observed at Volcán de Colima where the 2015 lava dome was growing towards the southern flank, eventually resulting in an overflowing dome with a lava flow, both of which collapsed during the July 2015 PDC-forming eruption.

After the July 2015 events, a new lava dome began forming in February 2016 and was overflowing onto the south flank by October 2016 (GVP, 2017a). The south descending lava flow filled the proximal depression of the south flank and created a new rim to the summit area. This new topographic barrier resulted in a change in PDC directionality during an eruption in January-February 2017 that sent PDCs into La Lumbre ravine on the East flank (GVP, 2017b).
Figure 6.13. Evolution of the Montegrande ravine at Volcán de Colima over one year. **a)** Photograph of the Montegrande ravine taken in January 2016 during our field campaign showing the 2015 PDC deposits that filled the valley. **b)** Photograph at the same location in the Montegrande ravine taken in November 2017 (courtesy of L. Capra) showing the intense erosion of the deposits after one year.

Significant erosion of PDCs in the months to years after emplacement creates new channels and paths that can modify the impacts of subsequent PDCs. Photographs of the Montegrande ravine taken during the Fall of 2017 show the large erosion that carved into the deposits, exposing deposit thickness greater than 10 m in most areas (Figure 6.13). The intense erosion also suggests a large degree of topographic changes to account for in future DEMs for numerical models and for hazard assessments because of the newly created pathways and new channel morphologies that can impact the generation of overbank flows.

Furthermore, computational models of PDCs require an initial volume of material as input parameter. Estimates of volume from past events can serve as a calibration parameter to test these models (Lube et al., 2020). Moreover, the mobility (runout distances, inundation areas, dropped-
height over length) of a flow is dependent on its volume. Empirical models that utilize the relationships between volume and mobility may be used to update hazard maps, but accurate estimates are required for calibration of these empirical laws.

Something to keep in mind, however, is that PDCs are pulsatory, and individual pulses may behave differently depending on volume and topographic conditions from prior pulses filling the channels. The difficulty is posed by the timescales of the transport and deposition of individual pulses, which are instantaneous (minutes to hours) and cannot be resolved on DEM time-series (few days at a minimum). It is therefore unlikely to be able to retrieve volumes of individual PDC pulses, with our current methods.

My work shows, however, the potential for retrieving mass flux of slow-moving lava flows and volumes of lava domes, which may be used as source input in geophysical mass flow models. For shorter occurring events such as PDCs and lahars, the sum of yearly trends of elevation changes obtained with ASTER DEM time-series can provide a good estimate of the total surface changes experienced from volcanic activity.

Similar work by Sean et al. (2020) was successful in retrieving elevation changes of glaciers in the High-Mountain Asia region (Shean et al., 2020). At Calbuco Volcano, I find that the sum of elevation changes per year obtained with ASTER time-series provides a good estimate of the total deposit volume in the valleys, despite the large changes at the summit. The negative trends of elevation changes interpreted as glacier removal at the summit and erosion in the valleys also provide information for the rate of change that can be expected post-eruption. Changes within the valleys are likely to impact future PDCs and gauging these changes prior to an event could significantly improve hazard assessments for the inundation area and runouts.
Conclusions

This study applied two types of DEM time-series to investigate the topographic changes associated with the PDC-forming eruptions of Volcán de Colima (July 2015 dome-collapse) and Calbuco Volcano (April 2015 sub-Plinian eruption). The high frequency of observations and free access to medium-resolution ASTER data allow for building time-series to observe trends of elevation changes through time. HRSI are often more difficult to access because of data cost and availability (i.e., commercial data), but DEM differencing between a pre- and post-eruption product can help identify important topographic changes at fine resolution.

Although topographic changes caused by eruptive activity are not linear, the trends of elevation changes from medium-resolution DEM time-series does allow tracking of the major changes occurring through time, thus showing promising results for applications related to monitoring topographic changes at active volcanoes.

The limitations to using optical data are the frequent cloud covers at the summit, which is why data products like ASTER which have frequent repeat-pass over the same areas can provide sufficient datasets to mitigate the lack of signal during cloudy days. Similarly, the dense vegetation that often covers the flanks of volcanoes in tropical regions can pose problems with surface models as they mask underlying topographic features and can add to the estimates of volume change on elevation difference maps. Using data acquired during different seasons can help determine the changes resulting from vegetation as opposed to volcanic activity. Moreover, destruction of vegetation from PDCs can provide important information regarding the dynamics of these flows. Differences between charred/delimbed trees vs. tree felling indicates different dynamic pressures of the flows which can in turn be used in numerical models for benchmarking (e.g., Esposti-Ongaro et al., 2020).
References


CHAPTER SEVEN:
CONCLUDING REMARKS AND PERSPECTIVES

It is important to distinguish between the terms hazard and disaster: the former refers to potentially damaging phenomena such as floods, earthquakes, and volcanic eruptions that only become classified as a disaster when it occurs in a populated area and brings mortality, damage, loss and destruction (Van Westen, 2000). Effective strategies for disaster management are therefore required to prevent a hazardous event from becoming a disaster and are based upon an integrated four-phase planning system that includes reduction, readiness, response and recovery (Figure 7.1).

**Figure 7.1.** The four phases of the disaster management planning system.
This four-phase planning system can essentially be broken down into two timelines: one that takes place before the occurrence of an event with long- to short-term hazard assessment and preparedness work, and one that takes place after the event with short-term crisis management and long-term recovery efforts (Joyce et al., 2009).

Hereafter I discuss the implications of the work presented in this dissertation for hazard assessment, preparedness and response to disasters from pyroclastic density currents (PDCs) and lahars. The fourth phase of the cycle, recovery, typically focuses on restoration of essential utilities and building reconstruction for short-term efforts, and on assessing the socio-economic damages for long-term recovery efforts with the affected populations. This last phase is not discussed here, as it concerns social sciences more than natural sciences.

**Hazard Assessments and Preparedness**

The pre-event disaster management strategies are predominantly based on the zonation of hazards indicating the potential areas at risk in the future. In Chapters 3 and 4, I provided comprehensive records of two PDC-forming eruptions to be employed into a database of evidence of past disasters, which can in turn be used for hazard assessments of future PDCs.

These records may be used for two types of future work: (1) updating the hazard maps at our study sites (i.e., Colima and Calbuco volcanoes) given prior knowledge of PDC deposit extents from these 2015 events; (2) using these data as confirmation datasets to test computational models that may be used for future eruptions, either at Colima and Calbuco, or anywhere else.

At Volcán de Colima, the July 2015 PDCs had the longest recorded runouts of the historical PDCs from dome collapse, only ever topped by those generated from the Plinian eruption of 1913 (Saucedo et al., 2005). The high frequency of dome-forming and dome-collapse eruptions at
Volcán de Colima requires regular updates of the hazard maps, which will now be extended to further distances than previous maps.

At Calbuco Volcano, the damages extended beyond the footprint of PDCs and were caused by the secondary lahars that remobilized the fresh PDC deposits of the April 2015 eruption. Hazard maps will be updated to account for the long runouts these hazardous flows may reach in the future, considering the proximity to farmlands, habitations and infrastructures.

The field evidence used to interpret dynamics of transport and deposition at Colima and Calbuco may be used in future work to further the understanding of flow dynamics. Changes in deposit facies of PDCs with respect to changes in the local morphology of the channel were used to propose conceptual models for the changing dynamics. This information can be used to inform the behavior of future flows. For instance, strong channelization of the flows results in longer runouts, while shorter flows tend to occur on unconfined slopes due to transition to a granular flow regime where frictional forces dominate.

Moreover, in Chapter 5 I presented evidence for similar flow behavior in response to topographic parameters, namely the slope, sinuosity, and capacity of the valleys in which PDCs deposited. I identified different types of hazardous flows derived from their depositional environment, namely outside of the valley-confines, as active versus passive overbank flows. Passive overspilling of the concentrated PDCs was observed as the result of reduced channel capacity either from pre-existing geomorphology or by progressive infilling of the channel by successive PDC pulses. Active overbank flows and detached ash-cloud surges on the other hand, were found to be the result of local changes in slopes and sinuosity, or from interaction with topographic obstacles. Reduced valley cross-sectional area also contributed to enhanced active overbanking processes, but was not the sole driver.
These results are important for forward modeling and hazard assessments, and I demonstrated the usefulness of simple and reproducible GIS-based measurements of valley morphologies as means to extract ‘hot spot’ areas where hazardous PDCs are likely to occur. The commonalities that were found between the two different types of PDCs (i.e., dome-collapse vs. column-collapse) are highly encouraging for the reproducibility of these measurements. Moreover, it suggests the potential for applying hazard assessment models based on topography analysis, which would then be applicable to any PDCs rather than being solely volcano- or scenario-based.

In Chapter 6, I emphasized the need for tracking topographic changes at active volcanoes and updating digital elevation models (DEMs) in order to increase the accuracy of hazard zones for the generation of overbank flows and detached surges specifically. Furthermore, topographic changes of the volcanic edifice, particularly at the summit, may impact the direction of PDCs and need to be accounted for in the generation of hazard maps. For instance, at Volcán de Colima, the changes in the crater area during 2016 resulted in the closure of the crater rim to the south, with a new opening to the east where subsequent PDCs descended during the next eruption of February 2017.

During hazard assessment and preparedness efforts, the method proposed in Chapter 6 for tracking topographic changes may be employed to better organize targeted aerial topographic surveys over areas that require new, accurate and high-resolution DEMs. In Chapter 6, I also demonstrated the potential for using medium- and high-resolution DEMs to retrieve volume of erupted materials from past eruptions. These volumes can in turn be used to better calibrate the input and source conditions of numerical models, for example the volume of a lava dome for the modeling of PDCs.
**Disaster Response and Crisis Management**

The cost of a disaster, both in terms of economic loss and fatalities, is dependent upon the rapidity and efficacy of the event response. A properly orchestrated disaster response requires an accurate and comprehensive overview of the affected area in the critical hours following an event, and rapid damage assessment can significantly reduce public health consequences (Garshnek et al., 1998). In the case of PDCs and lahars disasters, remoteness of the terrains combined with potentially incapacitated lifelines (e.g., disturbed transportation network caused by collapsed bridges) hinder ground-based surveys for timely assessment of damage extents.

During disaster response and crisis management, the most critical information is the extent of the inundated areas by the flows (considering PDCs and lahars), as well as the damage to life and infrastructure (including access to the area) and the total number of people potentially affected. Post-disaster mapping of damage extents, whether by PDCs or lahars, can also support relief agencies to determine the setup of relief camps and temporary shelters, while addressing the immediate danger posed by secondary lahars permits the prioritization of evacuation efforts.

It is also important to assess the damages to local communities regarding their food supply and livelihoods. These flows are damaging beyond the loss of life, and their effects may be longer term, particularly in low-income countries where food insecurities are persistent and can be exacerbated after a disaster. The flow extent mapping can provide information regarding the farms and infrastructures directly affected by these events and help agencies such as World Bank determine the post-disaster economic impacts, and insurance groups to provide monetary support to the most impacted. Finally, whether during or after a disaster, a direct communication line between scientists and stakeholders requires simple and easy-to-use products in order to respond as rapidly as possible to the event.
In Chapter 5, I presented the advantages of using a combination of optical and Synthetic Aperture Radar (SAR) data for providing comprehensive overview of flow extents, which can be used to target and organize the response efforts. I demonstrated the capabilities of multi-sensor analysis to alleviate the limitations from each sensor type and I showed that combined products may be used to derive probability damage extent maps. Optical data require cloud-free skies for maximizing the observations of the surface, which is often restricted during an event due to heavy ash clouds or rain. SAR data compensate for these limitations with all-weather and all-day imaging capabilities. Finally, for night-time events, thermal images can provide rapid assessment of the location of hot and hazardous flows to outline the dangerous zones and where search and rescue efforts will need to be directed.

My results showed similar values of pre- and post-eruption NDVI (normalized difference vegetation index) changes caused by the presence of erupted materials at both Colima and Calbuco. Results from SAR intensity images also showed similar changes caused by the deposition of PDC deposits and allowed for differentiation between lava flows and PDC deposits. The joint probability maps of PDC deposit extents allowed for the refinement of deposit mapping and is an encouraging method to be employed for future PDC-forming eruptions. With this methodology, coarse- to medium-resolution imagery can be processed relatively rapidly, both to provide overviews of the damaged areas and to organize more targeted efforts. For instance, requisitioning unmanned airborne vehicles (UAV) overflights to acquire very high-resolution (VHR) data and determine the damage extents more accurately. This can provide critical data-driven support to emergency responders for timely, improved, and informed responses.

Examples of initiatives that deliver products to the major stakeholders are The International Charter on Space and Major Disasters, signed by the French, Canadian and European space
agencies, which provides rapid image delivery in emergency situations; The US Federal Emergency Management Agency (FEMA), which has set up partnerships with the remote sensing community to actively provide image-derived map products; and The NASA Disasters Program, although not in an operational capacity, which also contributes to providing remote sensing imagery and maps to support various stakeholders during a disaster.

**Future work**

**Remote sensing and volcanic hazards**

According to the 2020 report on the human cost of disasters by the United Nations office for Disasters and Risk Reduction (UNDRR, 2020), volcanic activity accounts for 1% of total disaster events between 2000 – 2019 with 102 disaster events recorded. However, in 2018 it resulted in more deaths than in previous 18 years combined. For example, the Fuego eruption (Guatemala) in June 2018 killed 485 people (from PDCs and post-eruption lahars) and affected 1.7 million people. A report by ACAPS (www.acaps.org) summarized the damages to crops and livestock in villages around Fuego (Chimaltenango, Sacatepéquez and Escuintla) by PDCs and lahars, while ash and lava flows reportedly limited the supply of foodstuff to the evacuated communities. It also destroyed and damaged schools, energy network, a bridge and roads, thus limiting access to the devastated areas. The rescue operations were slowed down by the presence of volcanic flows and lahars. These reports highlight once more the need for further research on both hazard assessment of volcanic flows and for improving syn- and post-disaster work, specifically for utilizing remote sensing capabilities and reducing delivery times to decision-makers and emergency responders.
Because of the continuous monitoring of the globe with an ever-increasing number of sensors at multiple resolutions and using different capabilities, future work should further develop the joint analysis of multi-sensor data. The potential for mapping volcanic flows using SAR coherence and intensity combined with optical data was highlighted in this dissertation and should be further explored using data from other satellites with more frequent passes and higher spatial resolution (e.g., Copernicus Sentinel-1A/B and Sentinel-2 satellites operated by the European Space Agency). Future efforts should also focus on increasing the level of automation for efficient processing of large volumes of data, taking into account the sense of urgency that comes in dealing with natural hazards.

Ongoing efforts by the Advanced Rapid Imaging and Analysis (ARIA) team – a collaboration between NASA Jet Propulsion Laboratory and Caltech that exploits SAR and optical remote sensing data for hazard response (http://aria.jpl.nasa.gov) – are encouraging for developing new tools related to damage mapping of volcanic flows and others (e.g., floods, earthquake damages, wild fires). For example, the ARIA team produces post-disaster damage proxy maps depicting areas most likely damaged using changes in SAR coherence data from Copernicus Sentinel-1 satellites.

Another example of ongoing research applied to near real-time processing of SAR data is the SARVIEWS project (Meyer et al., 2019). SARVIEWS processing system is a cloud-based hazard monitoring system with fully automated processing algorithms for Sentinel-1 SAR data to generate hazard information products. For volcanic hazards and other types (e.g., floods, wildfires, earthquakes), future research should build from these ongoing projects to automatically process multi-sensor data in near real-time conditions in order to provide high quality damage extent maps.
Additional research avenues should consider Machine Learning (ML) or Artificial Intelligence (AI) for improved automation of data processing and analysis, and for dissemination of the results. An example of AI applied to volcanic flow mapping was done by Kadavi et al. (2017) where the artificial neural network approach was used on Landsat-7 imagery for land classification and detection of PDC deposits at Mount Sinabung and Merapi volcano (Indonesia). A similar approach could be considered for a combination of SAR and optical imagery.

For longer-term applications in hazard assessments, future research should continue to build onto the combined use of multi-sensor data for monitoring active volcanoes. A summary of the relevant sensor for volcanic activity monitoring, including degassing, thermal and deformation signatures, is provided by Reath et al. (2019). Integrating data from these sensors with the processing and analysis of topographic changes time-series using the methodology developed in this dissertation would add extensive capabilities to an operational monitoring system of volcanic activity.

Furthermore, accurate and up-to-date DEMs of active volcanoes are often lacking, particularly in developing regions. DEMs are often the core datasets for any disaster-related analysis, and need to be representative of the contemporaneous landscape. During hazard assessment and preparedness, or during disaster response, medium-resolution (≤ 30 m) DEMs are often sufficient (NASA Disasters Program, personal communications). With this dissertation, we have shown that such DEMs may be rapidly created, with robust statistical analysis for accuracy assessment. Moreover, these DEMs can constitute the basis for targeted surveys at high-resolution if needed, or for acquiring high-resolution commercial data on small areas, thus reducing the cost of data.
Broader research applications

In a broader context, this dissertation has developed reproducible workflows for processing and analyzing both commercial and open-source satellite data, which can be scalable and applied to other geoscience questions that relate to surface and topographic changes (e.g., deforestation, glacier melt, earthquake damage, flooding). Future research should take advantage of the stereo capabilities of open-source data for exploring surface changes in a time-series context. While we presented results using ASTER data, hence at medium-resolution, a similar study should consider the use of high-resolution DigitalGlobe Worldview data for tracking fine topographic changes at active volcanoes and other areas of study. For example, Shean et al. (2019) used DEM time-series created from ASTER and sub-meter commercial imagery (DigitalGlobe WorldView-1/2/3 and GeoEye-1) to quantify the changes in glacier mass balance in the High-Mountain Asia region.

While optical stereo capabilities are common and easy to use, future work should also integrate DEMs from Interferometric Synthetic Aperture Radar (InSAR) to track changes in topography during the course of an eruption, or for other types of natural hazards such as earthquakes. The advantage of InSAR is that depending on its wavelength, the radar signal may penetrate the canopy cover and provide terrain elevations as opposed to surface elevations. This is particularly important in tropical regions where dense forests may hide the underlying topography, as evidenced in DEMs generated from optical imagery. DEM time-series from InSAR was used by McAlpin (2019) to derive high-accuracy estimates of effusive volume during the 2012-2013 Tolbachik fissure eruption (Kamchatka). Combining time-series of DEMs generated from various sensor types (e.g., InSAR, optical, lidar) would provide comprehensive overviews of the topographic and surface changes anywhere data is available.
Finally, I hope that this dissertation can provide directives for further research aiming to improve hazard assessment and disaster response work beyond that of volcanic hazards. The technologies for Earth observations are developing at an increasing rate, with new sensors being sent to space every year and offering more opportunities for work in the field of disasters management. It is important to build robust, reproducible and scalable methods that will be applicable to new datasets as they keep being produced, and to any type of natural or man-made disaster. It is also critical to maintain open lines of communication between scientists and stakeholders in order to not lose sight of what products can be actually effective for disasters assistance.

References


APPENDIX I

CHAPTER 5 SUPPLEMENTAL MATERIALS

Landsat-8 OLI Data Processing with ARCSI Software

In order to compress the data, Landsat images are typically provided with pixel values in digital number (DN), which can be converted to at-sensor radiance \( L_\lambda \) with unit \( \text{W.m}^{-2}\text{.sr}^{-1}\text{.}\mu^{-1} \) using the gain \( M_L \) and offset \( A_L \) constants provided in the metadata file:

\[
L_\lambda = (M_L \times \text{DN}) + A_L
\]

For the OLI bands, the gain and offset constants vary for each band and between images, and the radiometric calibration was carried out using the Atmospheric and Radiometric Correction of Satellite Imagery (ARCSI) Software, following the steps described hereafter:

For each OLI bands and for each Landsat-8 scene at Colima and Calbuco, the DN values were converted to at-sensor radiance \( L_\lambda \), which was then converted into Top of Atmosphere (TOA) reflectance \( \rho_{\text{TOA}} \).

The effects of the atmosphere and illumination factors were removed to calculate the surface reflectance \( \rho_\lambda \). In ARCSI, the atmospheric correction for converting TOA reflectance to surface reflectance is done using the 6S radiative transfer model, initially developed as Py6S algorithm by Wilson (2012). The 6S parameters include: (1) atmospheric profile, (2) aerosol profile or Aerosol Optical Thickness (AOT), (3) ground reflectance, (4) sensor geometry, (5) altitude and (6) wavelength.

The 6S outputs are a set of coefficients applied to the input image for converting at-sensor radiation to surface reflectance. To process our datasets of Landsat 8 images at Colima and
Calbuco in ARCSI, we used the Tropical and Mid-latitude Winter respective standard atmospheric profiles implemented in the software and the default ‘Green Vegetation’ ground reflectance model for both datasets.

The sensor geometry is provided in the metadata associated with each file, the wavelength value is band-dependent, and the altitude information were provided using an external DEM (i.e., NASADEM 30-m). Finally, the AOT values were derived using the Dark Object Subtraction (DOS) method (e.g., Chavez, 1996). The DOS method allows to create and apply a modelled atmosphere to earth observation data, and to estimate surface reflectance in the blue channel.

Taking the estimated surface reflectance as input, the 6S model is numerically inverted to determine the most likely AOT. For most region, a single AOT constant is not viable as the atmosphere is too variable, and failure to properly parameterize an AOT may result in large errors of atmospheric corrections (Wilson et al., 2014).

**NDVI Normalization Using Linear Regression**

The procedure consists in using a linear function: \( y = ax + b \), where \( x \) and \( y \) represent the NDVI values for NDVI images \( X \) and \( Y \). For each volcano, I used the time-series function of Google Earth to find areas with LULC that remained constant through time and created polygons to extract pixel values of unchanged LULC for all NDVI images. Then, for a given pair of NDVI images \( X \) and \( Y \), the linear least-square regression method is used to determine the slope (\( a \)) and intercept (\( b \)) coefficients (Figure I.1), and NDVI image \( Y \) can be predicted from NDVI image \( X \) by reapplying the linear function to the whole image, such that: \( Y_{pred} = aX + b \). Finally, the NDVI image differencing and ratio can be done between actual NDVI image \( Y \) and the predicted NDVI
image $Y_{pred}$. The Jupyter notebook I wrote and used to perform NDVI normalization can be found in my Github repository (Appendix II).

**Table I.1.** List of NDVI image pairs at Colima and Calbuco, with their respective linear regression functions. The functions were obtained from applying the NDVI normalization method described in the text.

<table>
<thead>
<tr>
<th>Volcano</th>
<th>NDVI image pair</th>
<th>Linear regression function</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Colima</td>
<td>2014-02-05 ($X$) – 2015-03-12 ($Y$)</td>
<td>$y = 1.02x - 0.0006$</td>
<td>0.90</td>
</tr>
<tr>
<td></td>
<td>2014-02-05 ($X$) – 2016-01-10 ($Y$)</td>
<td>$y = 0.79x + 0.18$</td>
<td>0.82</td>
</tr>
<tr>
<td></td>
<td>2015-03-12 ($X$) – 2016-01-10 ($Y$)</td>
<td>$y = 0.68x + 0.24$</td>
<td>0.80</td>
</tr>
<tr>
<td></td>
<td>2016-01-10 ($X$) – 2016-11-25 ($Y$)</td>
<td>$y = 0.51x + 0.28$</td>
<td>0.68</td>
</tr>
<tr>
<td>Calbuco</td>
<td>2015-02-13 ($X$) – 2015-10-11 ($Y$)</td>
<td>$y = 0.61x + 0.32$</td>
<td>0.64</td>
</tr>
<tr>
<td></td>
<td>2015-02-13 ($X$) – 2016-01-31 ($Y$)</td>
<td>$y = 1.05x - 0.04$</td>
<td>0.92</td>
</tr>
<tr>
<td></td>
<td>2015-02-13 ($X$) – 2016-03-19 ($Y$)</td>
<td>$y = 1.13x + 0.12$</td>
<td>0.87</td>
</tr>
<tr>
<td></td>
<td>2016-01-31 ($X$) – 2016-03-19 ($Y$)</td>
<td>$y = 0.57x + 0.39$</td>
<td>0.60</td>
</tr>
</tbody>
</table>

**Figure I.1.** Linear regression plot of NDVI values for unchanged LULC polygons at Colima. The slope (m), the intercept (b) and the $R^2$ values were obtained by applying the least-square regression technique to the pair of NDVI images: 2015-03-12 ($X$) – 2016-01-10 ($Y$), for which the results are presented in Chapter 5.
Figure I.2. Histograms of NDVI pixel values to compare actual and predicted NDVI images at Colima. These are the results for the NDVI image pair: 2015-03-12 (X) – 2016-01-10 (Y), for which the results are presented in Chapter 5.

Figure I.3. NDVI image differences at Colima. On the left the absolute NDVI difference between original image $Y$ (2016-01-10) and image $X$ (2015-03-12). On the right, the absolute NDVI difference between original image $Y$ and predicted image $Y_{pred}$. 
Figure I.4. Linear regression plot of NDVI values for unchanged LULC polygons at Calbuco. The slope (m), the intercept (b) and the $R^2$ values were obtained by applying the least-square regression technique to the pair of NDVI images: 2015-02-13 (X) – 2015-10-11 (Y), for which the results are presented in Chapter 5.

Figure I.5. Histograms of NDVI pixel values to compare actual and predicted NDVI images at Calbuco. These are the results for the NDVI image pair: 2015-02-13 (X) – 2015-10-11 (Y), for which the results are presented in Chapter 5.
Figure I.6. NDVI image differences at Calbuco. On the left the absolute NDVI difference between original image $Y$ (2015-10-11) and image $X$ (2015-02-13). On the right, the absolute NDVI difference between original image $Y$ and predicted image $Y_{\text{pred}}$.

Very-High-Resolution Optical Imagery Dataset

Table I.2. Acquisition dates, spatial resolution and specifications of the VHR optical images at Colima and Calbuco.

<table>
<thead>
<tr>
<th>Volcano</th>
<th>Pre/Post-eruption</th>
<th>Acquisition Date</th>
<th>Satellite</th>
<th>Spatial Resolution (m)</th>
<th>Specifications</th>
</tr>
</thead>
<tbody>
<tr>
<td>Colima</td>
<td>pre-eruption</td>
<td>2013-04-08</td>
<td>Pleiades-1A</td>
<td>0.5</td>
<td>pansharpened</td>
</tr>
<tr>
<td></td>
<td>post-eruption</td>
<td>2015-07-25</td>
<td>SPOT-6</td>
<td>2</td>
<td>pansharpened</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2016-01-10</td>
<td>Pleiades-1A</td>
<td>0.5</td>
<td>pansharpened</td>
</tr>
<tr>
<td>Calbuco</td>
<td>pre-eruption</td>
<td>2014-05-22</td>
<td>Worldview-1</td>
<td>1</td>
<td>panchromatic</td>
</tr>
<tr>
<td></td>
<td>post-eruption</td>
<td>2015-04-30</td>
<td>Pleiades-1A</td>
<td>0.5</td>
<td>pansharpened</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2015-12-01</td>
<td>Pleiades-1A</td>
<td>0.5</td>
<td>pansharpened</td>
</tr>
</tbody>
</table>

ALOS-2 Imagery Dataset

The series of command lines used in the GAMMA software for interferometry processing are compiled in shell scripts, which can be found in my Github repository (see Appendix IV), and the datasets processed for Colima and Calbuco are summarized in Table I.3.
Table I.3. Orbit and frame information for ALOS-2 datasets used to generate interferograms at Colima and Calbuco. The acquisition dates are written in the format YYYYMMDD, and the interferogram names show primary and secondary dates in YYMMDD format.

<table>
<thead>
<tr>
<th>Interferograms</th>
<th>Dates of Image Acquisition</th>
<th>Orbits</th>
<th>Frame Number</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Primary image</td>
<td>Secondary image</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20170126</td>
<td>20171214</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20170410</td>
<td>20170605</td>
<td>Descending</td>
</tr>
<tr>
<td></td>
<td>20160411</td>
<td>20160606</td>
<td>Descending</td>
</tr>
<tr>
<td>Colima</td>
<td>20170126</td>
<td>20170410</td>
<td>Descending</td>
</tr>
<tr>
<td></td>
<td>20150608</td>
<td>20150129</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20150525</td>
<td>20150429</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20140911</td>
<td>20150129</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20151125</td>
<td>20161123</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20150708</td>
<td>20151125</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20150429</td>
<td>20150708</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20150304</td>
<td>20150429</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20150204</td>
<td>20150304</td>
<td>Ascending</td>
</tr>
<tr>
<td>Calbuco</td>
<td>20150608</td>
<td>20150129</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20150525</td>
<td>20150429</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20140911</td>
<td>20150129</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20151125</td>
<td>20161123</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
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<td>20151125</td>
<td>Ascending</td>
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<td>20150304</td>
<td>20150429</td>
<td>Ascending</td>
</tr>
<tr>
<td></td>
<td>20150204</td>
<td>20150304</td>
<td>Ascending</td>
</tr>
</tbody>
</table>

Landsat-8 TIRS Data Processing

The gain and offset values for TIRS band 10, to obtain at-sensor spectral radiance from Equation 1, are constant for all Landsat 8 images and are provided in Table I.4.

Table I.4. Constants for computing at-sensor radiance and brightness temperature from Landsat 8 Thermal Infrared Sensor (TIRS) data. $M_L$: gain; $A_L$: offset; $K_1$ and $K_2$: band 10 calibration constants.

<table>
<thead>
<tr>
<th>TIRS Band</th>
<th>$M_L$</th>
<th>$A_L$</th>
<th>$K_1$ (W.m$^{-2}$.sr$^{-1}.µm^{-1}$)</th>
<th>$K_2$ (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Band 10</td>
<td>0.0003342</td>
<td>0.1</td>
<td>774.89</td>
<td>1321.08</td>
</tr>
</tbody>
</table>
**Table I.5.** Atmospheric correction parameters for Landsat-8 scenes at Colima. Values are interpolated based on the Mid-latitude Summer atmospheric profile. (https://atmcorr.gsfc.nasa.gov/; Barsi et al., 2005).

<table>
<thead>
<tr>
<th>Scene Acquisition Date and Time (UTC)</th>
<th>Transmittance (τ) (unitless)</th>
<th>Upwelling radiance ( (L_\lambda^\uparrow) ) (W.m(^{-2}).sr(^{-1}).µm(^{-1}))</th>
<th>Downwelling radiance ( (L_\lambda^\downarrow) ) (W.m(^{-2}).sr(^{-1}).µm(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>2014-01-04 – 17:19</td>
<td>0.73</td>
<td>2.24</td>
<td>3.63</td>
</tr>
<tr>
<td>2014-02-05 – 17:19</td>
<td>0.74</td>
<td>2.22</td>
<td>3.57</td>
</tr>
<tr>
<td>2014-03-25 – 17:18</td>
<td>0.72</td>
<td>2.44</td>
<td>3.91</td>
</tr>
<tr>
<td>2014-04-10 – 17:18</td>
<td>0.76</td>
<td>2.07</td>
<td>3.44</td>
</tr>
<tr>
<td>2014-12-06 – 17:18</td>
<td>0.72</td>
<td>2.46</td>
<td>3.93</td>
</tr>
<tr>
<td>2015-03-12 – 17:18</td>
<td>0.84</td>
<td>1.36</td>
<td>2.29</td>
</tr>
<tr>
<td>2015-04-13 – 17:17</td>
<td>0.61</td>
<td>3.29</td>
<td>5.17</td>
</tr>
<tr>
<td>2015-07-18 – 17:18</td>
<td>0.49</td>
<td>4.28</td>
<td>6.40</td>
</tr>
<tr>
<td>2015-08-03 – 17:18</td>
<td>0.54</td>
<td>3.86</td>
<td>5.69</td>
</tr>
<tr>
<td>2016-01-10 – 17:18</td>
<td>0.88</td>
<td>1.00</td>
<td>1.67</td>
</tr>
<tr>
<td>2016-11-25 – 17:18</td>
<td>0.67</td>
<td>2.90</td>
<td>4.55</td>
</tr>
<tr>
<td>2017-04-02 – 17:18</td>
<td>0.77</td>
<td>1.96</td>
<td>3.16</td>
</tr>
</tbody>
</table>

**Table I.6.** Atmospheric correction parameters for Landsat-8 scenes at Calbuco. Values are interpolated based on the Mid-Latitude Winter atmospheric profile. (https://atmcorr.gsfc.nasa.gov/; Barsi et al., 2005).

<table>
<thead>
<tr>
<th>Scene Acquisition Date and Time (UTC)</th>
<th>Transmittance (τ) (unitless)</th>
<th>Upwelling radiance ( (L_\lambda^\uparrow) ) (W.m(^{-2}).sr(^{-1}).µm(^{-1}))</th>
<th>Downwelling radiance ( (L_\lambda^\downarrow) ) (W.m(^{-2}).sr(^{-1}).µm(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>2015-02-13 – 14:35</td>
<td>0.89</td>
<td>0.76</td>
<td>1.27</td>
</tr>
<tr>
<td>2015-03-26 – 14:29</td>
<td>0.90</td>
<td>0.72</td>
<td>1.22</td>
</tr>
<tr>
<td>2015-04-11 – 14:29</td>
<td>0.90</td>
<td>0.70</td>
<td>1.21</td>
</tr>
<tr>
<td>2015-04-27 – 14:29</td>
<td>0.85</td>
<td>1.07</td>
<td>1.78</td>
</tr>
<tr>
<td>2015-09-02 – 14:29</td>
<td>0.94</td>
<td>0.37</td>
<td>0.64</td>
</tr>
<tr>
<td>2015-10-11 – 14:36</td>
<td>0.94</td>
<td>0.39</td>
<td>0.67</td>
</tr>
</tbody>
</table>
APPENDIX II

CODE AND DATA

The codes used for this project are archived on my Github at https://github.com/emacorps

Landsat-8, ASTER, and NASADEM data are freely accessible from NASA EarthData. TanDEM-X 90-m data was obtained from the German Aerospace Center (DLR) and can be downloaded on their website: https://download.geoservice.dlr.de/TDM90/. DigitalGlobe Worldview imagery was obtained through CAD4NASA, and therefore the raw data cannot be shared. The raw datasets from Pleiades-1A were purchased and therefore cannot be freely released.
APPENDIX III

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Stratigraphy, sedimentology and inferred flow dynamics from the July 2015 block-and-ash flow deposits at Volcán de Colima, Mexico

Author: Eloide Maceiros, Sylvain J. Charbonnier, Nick R. Varley, Lucia Capra, Zachary Atlas, Josep Cabré
Publication: Journal of Volcanology and Geothermal Research
Publisher: Elsevier
Date: 1 January 2018
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Authors’ Contributions to Chapter 3 Publication

My contribution as primary author consisted of:

- 95% of manuscript writing
- Field observations and sampling during a week-long field work at Colima
- Laboratory analysis including sample sieving, and grain componentry
- Statistical analysis of grain size and grain componentry results
- GIS analysis including generation of very-high-resolution digital elevation model (DEM),
  deposit volume measurements and topography analysis
- Developing the main interpretations, discussion and conclusions, which were often
discussed with second author and prior thesis advisor Dr. Charbonnier

Contribution from other authors included edits to the manuscript, both grammatical and content-related, and assistance during field work at Colima:

Nick Varley as director of the Colima monitoring observatory and therefore expert,
provided insights regarding the July 2015 eruption and past eruptive activity. He also provided us
with access to the volcano and guided us to the deposits.

Lucia Capra is an academic expert on Colima and owns research seismic, video and
infrasound monitoring equipment situated on the volcano.

Zach Atlas provided help with physical labor during field work

Josep Cabre, who volunteered at the Colima observatory, also helped with sampling during
field work, and collected two of the samples during a separate field campaign a few weeks later.