MULTI-SCALE ANALYSIS OF MULTIPARAMETER GEOPHYSICAL AND GEOCHEMICAL DATA FROM ACTIVE VOLCANIC SYSTEMS

by

Guillaume MAURI
Master, Blaise Pascal University, 2005
Licence, Blaise Pascal University, 2003
DEUG, Louis Pasteur University, 2002

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APPROVAL

Name: MAURI Guillaume
Degree: PhD of Science
Title of Thesis: Multi-scale analysis of multiparameter geophysical and geochemical data from active volcanic systems

Examining Committee:

Chair: Dr John Clague
Professor
Department of Earth Sciences

_________________________
Dr. Glyn Williams-Jones
Senior Supervisor
Assistant Professor,
Department of Earth Sciences

_________________________
Dr. Kirstie Simpson
Supervisor
SFU Adjunct and VP Minerals
Research, Geoscience BC

_________________________
Dr. Philippe Labazuy
Supervisor
Associate Physicist,
Université Blaise Pascal,
France

_________________________
Dr. Nicolas Fournier
External Examiner
Volcano Geodesist, Wairakei
Research Centre, GNS Science,
New Zealand

_________________________
Dr. Dirk Kirste
Internal Examiner
Assistant Professor,
Department of Earth Sciences

Date Defended/Approved: August 28th, 2009
ABSTRACT

Persistently active volcanoes present short and long term variation in their magmatic, hydrothermal, and/or hydrogeological systems. As magma is rarely accessible on the surface, investigation of the dynamic behaviour of the hydrothermal system is an indirect approach to study the underlying magmatic activity. Variations in volume, mass and flow direction of the water are expressed through change and generation of local disturbances of potential-fields, such as gravity or Self-potential. The source generating the potential-field signal is a non-unique solution, making it very difficult to model. This study uses a modern signal analysis technique, Multi-scale wavelet tomography, to accurately determine the depths of these sources. The accuracy of Multi-scale wavelet tomography on Self-potential data was tested on three volcanoes (Masaya, Stromboli, Waita) in comparison with water depths calculated by more traditional geophysical methods. Traditional inverse gravity modelling is also used to better constrain the non-unique solution of potential-fields. This study also investigates two persistently active volcanoes, Masaya and Kawah Ijen, through time-series and spatial surveys to monitor change occurring within them. This study shows that a well established and mature hydrothermal system can show limited surface expression as at Kawah Ijen volcano, while an apparently low intensity hydrothermal system can have an extensive and complex system beyond its active crater, such as at Masaya volcano. The hydrothermal system of Masaya is spatially controlled by a ring fault structure and has been stable between 2006 and 2009. In contrast, on Kawah Ijen, the intense and well-established hydrothermal system is completely self-sealed within the upper volcanic edifice and can only release the pressurized fluids and gas through the active crater. Nevertheless, between 2006 and 2008, the hydrological system showed significant vertical change due to seasonal effects. By integrating a wide variety
of distinct, complementary techniques to a number of persistently active volcanoes, over an extended period of time, it is possible through accumulation of baseline information, to characterize the components of signals detected on volcanoes. A more accurate understanding of the volcanic system as a whole, through accurate constraint of the different volcanic signals, is fundamental in improving volcano monitoring and hazard mitigation.

**Keywords:** Active volcanoes; Self-potential; gravity; wavelet analysis; hydrothermal system; volcano monitoring.
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**Geothermal**


**Geothermal anomaly**

“An abnormal temperature condition in relation to the overall temperature filed.”

**Geothermal energy**

“Heat that occurs naturally in the Earth and that can be extracted from the Earth’s internal heat. Geothermal energy is usable for a wide range of temperature and volume, e.g. nonelectric use (direct use of geothermal water, geothermal heat pumps) and electric use of stream or hot water.”

**Geothermal gradient**

“the rate of change of temperature in the Earth with depth measured in °C m⁻¹ or °C km⁻¹. The gradient differs from place to place depending on the heat flow in the region and the thermal conductivity of the rocks. The average geothermal gradient in the Earth’s crust approximate 25°C km⁻¹ of depth. Symbol: Γ. See also: temperature gradient.”

**Geothermal reservoir**

“Underground storage of water trapped in porous rock capable of providing hydrothermal (hot water and steam) resources. Geothermal reservoirs may be classified as being either of low enthalpy (<150°C) or high enthalpy (150°C). The latter term generally is applied to reservoirs that contain water or steam suitable for the generation of electric power, whereas low-enthalpy reservoirs are usually those where the energy is principally of interest for heating purpose. Liquid-dominated, vapor-dominated, or dry-steam reservoirs are recognized.”

**Geothermal system**

“Any regionally localized geological setting where naturally occurring portions of the Earth’s internal heat flow are transported close enough to the Earth’s surface by circulating steam or hot water to be readily harnessed for use; examples are the Geysers Region of northern California and the hot brine fields in the Imperial Valley of southern California.”
**Hydrologic systems**  
“A complex of related parts—Physical, conceptual, or both—forming an orderly working body of hydrologic units and their man-related aspects such as the use, treatment and reuse, and disposal of water and the costs and benefits thereof, and the interaction of hydrologic factors with those of sociology, economics, and ecology.”

**Hydrothermal**  
**[petrology]** (hy-dro-ther’m-al): “Of or pertaining to hot water, to the action of hot water, or to the product of this action, such as a mineral deposit precipitated from a hot aqueous solution, with or without demonstrable association with igneous processes; also, said of the solution itself. Hydrothermal “is generally used for any hot water but has been restricted by some to water of magmatic origin.”

**Hydrothermal alteration**  
“Alteration of rocks or minerals by the reaction of hydrothermal water with pre-existing solid phases. Open-system chemical reaction resulting from partial to complete chemical equilibrium between the host rock and hydrothermal fluids of a mineral deposit. Syn: wall-rock alteration.”

**Hydrothermal deposit**  
“A mineral deposit formed by precipitation of ore and gangue minerals in fractures, faults, breccias openings, or other spaces, by replacement or open-space filling, from aqueous fluids ranging in temperature from 50° to 700°C but generally below 400°C, and ranging in pressure from 1 to 3 kilobars. The fluids are of diverse origin. Alteration of host rocks is common.”

**Hydrothermal system**  
“A ground-water system that has a source (or area) of recharge, a source (or area) of discharge, and a heat source.”

**Hydrothermal water**  
“Subsurface water whose temperature is high enough to make it geologically or hydrologically significant, whether or not it is hotter than the rock containing it. It may include magmatic water and metamorphic water, water heated by radioactive decay or by energy release associated with faulting; meteoric water that descends slowly enough to acquire the temperature of the rocks in accordance with the normal geothermal gradient but then rise more quickly so as to retain a distinctly above-normal temperature as it approaches the surface; meteoric water that descends to and is heated by cooling intrusive rocks; water of geopressed aquifers, and brine that accumulate in an areas of restricted circulation at the bottom of the sea.”
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1.1 Introduction

This thesis focuses on shallow hydrothermal, hydrogeological and magmatic changes that take place within two volcanic edifices over a period of 4 years. As magma is often not directly accessible, hydrothermal surveys can be used as an indirect view of volcanic change. In order to characterize the correlation between hydrothermal and volcanic variations, this study investigates two volcanoes having similar volcanic surface expressions (low eruption frequency and long term degassing). How extensive and active are the hydrothermal system on each volcano? Based on a hydrothermal survey, what kinds of significant information can be found regarding current volcanic activity? How do hydrothermal systems change in response to persistent volcanic activity? Do volcanoes with similar volcanic surface expressions, have similar hydrothermal behaviour?

Volcanic activity is a broad and complex topic so that to even partially understand a single volcano, the volcanological community needs to apply a lot of manpower, skills and a wide range of scientific tools over a period of at least several decades. According to the Smithsonian Institutions, Global Volcanism Program, there are ~1400 active volcanoes on Earth excluding those on the deep sea floor. Even a lifetime is generally insufficient to completely understand a single volcano because the majority of them can remain quiescent without any sign of volcanic activity for several hundreds or even thousands of years. Fortunately, some volcanoes behave in the opposite way, with “persistent volcanic activity” relative to human lifetimes. These volcanoes are characterized by continuous or very high eruption frequency or passive degassing, such as Stromboli, Italy, which has had a volcanic explosion approximately every 15 min for at least the last 2000 years ago. Another well-known example is Kilauea volcano, Hawaii, which has produced lava flows without any significant interruption since 1983 and is still erupting at this time. Persistently active
volcanoes have the advantage of behaving like any other volcano, except that they are acting continuously, allowing scientists to work on them at any time and to improve methods and knowledge much faster that if the community had to wait decades to hundreds of years for the next eruption of a quiescent volcano. Although this thesis does not claim to find the holy grail of understanding all volcanoes on Earth, by focusing on a few key persistently active volcanoes, this thesis brings new insight on how volcanic activity changes and how they are expressed through measurable phenomena.

Another key point for understanding volcanoes is the number of techniques and scientists that are studying it. Except for a few tens of volcanoes (e.g., Stromboli, Etna, Kilauea, Merapi, St Helens, Sakura-Jima or Piton de la Fournaise volcanoes), which are well covered with scientific instruments and surveyed by an army of scientists, the majority of the active volcanoes around the world, are poorly monitored with only at best one recording seismometer or observer to monitor their level of activity. The main reason for this is generally the lack of funding, which is quite often due to the lack of political will to support such work as well as the difficulty of safely accessing most of the volcanoes.

Even though a network of seismometers is still the best method, at this time, to survey and investigate volcanic activity with limited funding and trained scientists, other techniques may be used to help improve our understanding of persistently active volcanoes. Skilled volcanologists are few and many of these volcanoes are located in developing countries rather than within one of the G-20, which typically have the research groups in volcanology. However, these developing countries may have large manpower resources, thus rather than spending a lot of money on scientific instruments that require high levels of skill to operate, it may be often better to use simple inexpensive techniques. Although they may be labour intensive, these techniques may nevertheless yield important information on changes in the volcanic activity.

For almost any choice that we have to make during our lifetime, we must face the dilemma of the tradeoff between what we want to do and what is possible when considering safety, available funding, time and manpower. In
volcanology, we face the same choice when considering the right approach to investigate a volcano. As mentioned before, persistently active volcanoes can be active over tens to thousands of years, however, “persistently” does not mean that there is always lava or magma at the surface all the time. For example, activity can be characterized by continuous passive degassing with only ephemeral explosions. In fact, the magma is quiet often not directly accessible, or only during infrequent eruptions. However, near the surface, on persistently active volcanoes, there are a few things, which are always present: underground magmatic mass flux and, underground hot and cold water. Large changes of these magmatic or water bodies will be directly linked to changes in volcanic activity, and these changes will generate potential field variations on the surface, such as gravity, electric generation and magnetism. These geophysical phenomena are measurable and can be quantified.

As new hot magma rises towards the surface, it will exsolve gases (e.g., H$_2$O, CO$_2$, SO$_2$, H$_2$S), which is a very effective process to transfer heat from the magma to the surface. When hot gases reach an underground water table, the gas may dissolve in the water and the heat will be transferred to the water. Under a sustained gas/heat flux, cold water may become hot/boiling water and increase its fluid pressure and volume. The increasing pressure within the water will cause it to rise and this displacement (for hot, boiling, or cold water) will generate an electrical current at depth, which will induce an electrical current at the ground surface. This phenomenon and the method to measure it, is called Self-potential. When used over time in survey loop profiles, Self-potential can be an efficient tool to characterize changes occurring in the water. The Self-potential method is inexpensive, relatively easy and fast to perform and to process the results, but it can only give qualitative information about the surface distribution of the water table (such as, is the water hot, cold and what is the horizontal spatial distribution). New developments over the last two decades, in signal processing by wavelet analysis of Self-potential, now make it possible to acquire information on the water depth.
In this study, shallow hydrothermal systems are assumed to be within the first kilometre below the topographic surface. Any hydrothermal system deeper than 1 km, is considered a deep structure, while shallow magma is assumed to be at a depth less than a few kilometres below the topographic surface. Hot rising water requires hot gas, or hot rock as a heat supply and these gases can only come from shallow magma. It therefore becomes possible to investigate underground volcanic change by studying the vertical movement of the hot water. An important decrease in depth of the hot water is a possible indication that more hot gas is heating it and that more or new magma is underlying the hot water body. Thus, large vertical changes of hot water may be attributed to changes occurring within the volcanic edifice. Of course there are limitations as other phenomena can have similar effects and as for any scientific study, interpretation has to be supported by other scientific methods, such as soil CO₂ gas concentrations, ground temperature, deformation or gravimetry.

To investigate changes within the magma, this thesis also looks at micro-gravimetry variations to characterize changes in mass or density of the underground magma. Fresh, hot, gas-rich magma will be less dense than cold, degassed and partially crystallised magma. If new magma approaches the surface, it will generate a gravity anomaly on the surface above the magmatic intrusion. In addition, the gravimetry method (Bouguer mapping and dynamic surveys) will be complimented by other methods such as high-resolution GPS positioning to control external parameters affecting gravity measurements. Inverse and forward modelling are used to complete and model the underground magmatic structure.

Each method, assumption, hypothesis and model is fully described and discussed in this thesis. This work has one main purpose, to improve traditional methods and to contribute to the understanding of persistently active volcanoes. To achieve this objective, this study focuses on two persistently active volcanoes: Masaya volcano in Nicaragua and Kawah Ijen volcano in Indonesia. These volcanoes have a number of similarities; they both have continuous passive degassing and only a few eruptions in more than 150 years, but with their own
characteristics, which make them unique. Masaya and Kawah Ijen are fully described in the following sections of this chapter.

This study covers 4 years of work on these two active volcanoes, between 2006 and 2009, which allows us to acquire both spatial and time-series data in order to investigate the behaviour of these persistently active volcanic systems. The main body of this thesis is organized in three parts: The Introduction, Methodology and studies of the persistently active volcanoes. Each chapter is written in manuscript format to facilitate future publication and thus there may be some repetition. The structure of the thesis is as follows: Chapter 1, the introduction, includes the geological setting of both volcanoes. Chapter 2, 3 and 4 describe the Self-potential method, the gravimetry method and the multi-scale wavelet tomography (MWT), respectively. Chapter 5 is a case study of depth constraints of shallow hydrothermal systems by Self-potential and multi-scale wavelet tomography on three volcanoes: Stromboli, Waita and Masaya. Chapter 6 investigates the structure and dynamics of the hydrothermal system of Masaya volcano. Chapter 7 investigates the structure and dynamics of the hydrothermal system of Kawah Ijen volcano. Chapter 8 is the conclusion, with a summary of the hydrothermal, hydrogeological dynamisms and structures of these active volcanoes. Several appendices, which provide more detail about the work and topics related to it, complete the main body of this thesis and are organized in three parts: Appendix 1 is the investigation of Self-potential changes and depth estimates of the summit hydrothermal system of Piton de la Fournaise volcano, La Réunion. This study, which was the centre of my Master’s project, has been completely re-analysed with the multi-scale wavelet tomography methodology (as described in Chapter 4) to improve depth determination and interpretation of the hydrothermal dynamism of this volcano. Appendix 2 describes the Physical and chemical processes within Hydrothermal Systems. Appendix 3 presents an introduction of the mathematical tools required in order to understand the wavelet transform. Appendix 4 includes figures and tables for Masaya volcano. Appendix 5 is a preliminary investigation of short-term gravity variations on Masaya
volcano. Appendix 6 includes figures and tables for Kawah Ijen volcano. Finally, all the data and analyses are included on an accompanying DVD.

1.2 Geological Setting of Masaya Volcano

Figure 1-2: View from the South of Masaya volcano in 2006. Background: On the left, view of Ninidiri cone. On the right, view of Masaya cone.
1.2.1 Continental & Regional Scale

Masaya volcano (11.984°N, 86.161°W, 635 m) is located in southwestern Nicaragua, Central America, about 20 km south of the capital, Managua, and north-west of Masaya city (Fig. 1-2). Masaya volcano is a basaltic shield volcano with a caldera structure (~6 km by 11.5 km) [Williams, 1983b]. It is part of the Central American Volcanic Front (CAVF) [Stoiber and Carr, 1973], which for its Nicaraguan part includes 18 different volcanic edifices (Fig. 1-3). The CAVF is oriented NW-SE and is parallel and linked to the Mesoamerican subduction trench, where the Cocos plate subducts beneath the Caribbean plate [Carr et al., 1982]. Western Nicaragua is made up of three distinct parts [Walker et al., 1993]:

1. Cretaceous-Tertiary basins: the basement, which consists of a Pre-Cretaceous to Cretaceous ophiolitic suite, is overlain by the Cretaceous-Tertiary basin, which is made up of different marine formations. Regional deformation and isostatic readjustment created the Nicaraguan basin [van Wyk de Vries, 1993].

2. Tertiary volcanic terrains: The active Quaternary volcanic range is responsible for recent formations within the Nicaraguan basin, which among them includes the basaltic ignimbrite of the Las Sierra Caldera [van Wyk de Vries, 1993].

The tectonic setting of Central America shows that the northern boundary of the Caribbean Plate (oriented NNW-SSE) moves along the Montagua-Cayman strike-slip fault, which generates a counter-clockwise rotation of the crustal component [Burkart and Self, 1985; Manton, 1987; La Femina et al., 2002]. That movement has an average velocity, for the Central America rotation, of around 14 mm/year [DeMets, 2001]. The rotation has generated extension with the formation of the Managua Graben, opening N-S, by generation of a pull-apart basin, which overlie a magmatic complex [Girard et al., 2005]. The faults are normal sinistral faults and are located on the western border of the graben. Among these, the main faults are the Nejapa-Mirafl ores fault and Punta Huete
fault (Fig. 1-4). The Cofradrias and the Mateare faults are on the eastern border of the graben [Sebesta, 1997].

**Figure 1-3:** Structural map of Central America with the subduction trench and the main faults. The dots represent the Central American Volcanic Front (CAVF) [Girard et al., 2005]. Shaded box highlights Masaya volcano. *Reprinted from Journal of Volcanology and Geothermal Research, vol. 144(1-4), Girard, G., and B. Van Wryk de Vries, “The Managua Graben and Las Sierras-Masaya volcanic complex (Nicaragua); pull-apart localization by an intrusive complex: results from analogue modelling”, Pages 37-57, Copyright (2005) Elsevier.*
The volcanic trend is oriented NNW-SSE parallel to the Montagua-Cayman strike-slip fault. Inside it, Masaya caldera lies within the larger Las Sierras Caldera (~ 15 km by 15 km). Las Sierras is a basaltic ignimbrite shield volcano formed during the Quaternary [van Wyk de Vries, 1993; Sebasta, 1997], which appears to lie partially within another large caldera, Las Nubes caldera (Fig. 1-4). However, the distinctions are not clear and some authors believe that these are the same feature [Girard et al., 2005]. All the caldera formations and thus also the volcanic trend, are controlled by both the subduction trench and the counter-clockwise rotation. The nearest volcanoes to Masaya (except the two previous calderas) are Apoyo caldera to the southeast and the Apoyeque caldera to the northwest [Walker et al., 1993].

1.2.2 Masaya volcano evolution

The historical evolution of Masaya volcano can be described as follows:

i. Formation of Proto-Masaya, a small shield volcano with basaltic lavas (thin pahoehoe lava flows 1-5 m thick) and tephra [Walker et al., 1993; Maciejewski, 1998] coming from violent explosion where ash fall was deposited around the Proto-Masaya edifice. Later pre-caldera eruption formations were between 35 and 7 ka [Bice 1980; Walker et al., 1993].

ii. Caldera formation due to large basaltic plinian and large ignimbrite eruptions, which occurred between 2250 and 6500 B.P. [McBirney, 1956; Bice, 1980; Williams, 1983a,b; van Wyk de Vries, 1993; Perez et al., 2009]. The total of the estimated erupted volume is ~ 8 km³. The formation of the caldera is estimated between 6.5 and 2.25 ka [Williams, 1983a,b; Bice, 1985], based on tephra, ignimbrite and proximal pyroclastic surge deposits.
Figure 1-5: Masaya caldera. In red the crater names, in blue, the volcanic cone names and in burgandy the name of faults and hydrothermally active areas. The burgandy dashed lines are the fault and fissure structures. Black dashed line is the inferred limit of the caldera. Dashed circle is the estimated position of the ring faults along which the cinder cones are distributed. Star refers to the solfatara.
iii. Post-caldera activity was largely dominated by effusive eruptions and only small volumes of scoria and ash fall deposits, which are interlayered between lava flows. A complete stratigraphic description was made by Williams (1983b). The eruption centers are distributed along a ring-fault structure (Fig. 1-5), which is cut on its east part by the Cofradrias fault [Williams, 1983b, Walker et al., 1993]. A scar of the ring fault is expressed through the San Pedro Fault (Fig. 1-5, 1-6, 1-7), which cuts the west flank of Nindiri cone. The main cones are Masaya cone (Fig. 1-8), Nindiri cone (Fig. 1-8, 1-9), Cerro Montoso, Media Luna, Arenal, and Comalito cone (Fig. 1-11, 1-12). The caldera floor topography decreases to the east, where the Laguna de Masaya is located (Fig. 1-5, 1-13).

iv. Recent activity at Masaya volcano consists of cone and pit crater formation, lava flows and passive degassing. The first recorded observation of an eruption of Masaya volcano was in 1529. At this time, Nindiri cone and Masaya cone were already the largest cones in the caldera [Maciejewski, 1998] (Fig. 1-5, 1-14, 1-15). The largest eruption since 1529 was the 1772 fissure eruption that formed on the North flank of Masaya cone and flowed from its summit toward the North, passing on the West side of Comalito cone (Fig. 1-15) [Rymer et al., 1998; Roche et al., 2001]. For the last ~150 years, the main activity of Masaya volcano has been continuous and persistent passive degassing (Fig. 1-5, 1-8), sometimes interrupted by vent collapse and vent clearing, lava lakes (Fig. 1-14, 1-15) and small explosions related to hydrothermal-magma interaction. All the volcanic activity is currently confined to Santiago crater on the east part of Nindiri cone. The only surface expression of hydrothermal activity is the Solfatara of Comalito [Lewicki et al. 2003; Chiodini et al., 2005; Pearson et al., 2008]
**Figure 1-6:** Oblique aerial view of Nindiri cone, Masaya cone and Comalito cone. Masaya city can be seen overlooking Laguna de Masaya.
Figure 1-7: View in 2007 from the upper part of San Pedro Fault on the west flank of Nindiri cone. Background: Cerro Montozo. Foreground: San Pedro fault. Both are on the ring fault structure. San Pedro Fault curves toward Cerro Montozo.

Figure 1-8: View in March 2007 from the South of degassing Nindiri cone (left) and Masaya cone (middle and left). The actual crater of Masaya is in the left part of its cone.
Figure 1-9: View from the centre of Caldera of the North flank of Nindiri cone.

Figure 1-10: View in 2008 from the north-west flank of Masaya cone. Background: Cerro Montozo cone on the left and Arenal cone on the right. The caldera wall is in the background.
**Figure 1-11:** Arenal cone is located in the north part of the caldera.

**Figure 1-12:** Comalito seen from the North Masaya rim in 2008. The Comalito solfatara is located on the left side of the cone.
Figure 1-13: Laguna de Masaya located in the eastern part of the Masaya Caldera. The lake is oriented North-South and is the reference point for all Self-potential measurements.

1.2.3 Masaya volcano pit crater structures

The structure of Masaya volcano (Fig. 1-5, 1-14) is characterized by two main basaltic cones: Nindiri forms the North-West part and Masaya occupies the Eastern part. The summit of each cone hosts pit craters: San Fernando crater for Masaya Cone, while Nindiri cone is more complex, with 2 pit craters (Santiago and San Pedro, Fig. 1-13, 1-14, 1-15), which are on either side of the subsided Nindiri lava lake [Harris, 2008]. The pit crater formations result from vertical collapse of the crater floors associated with magma drainage from beneath the crater [Roche et al., 2001]. The pit crater structures are themselves very complex with different crater formations, which where successively infilled by eruptive deposits. Deposits consist of alternating layers of lava and scoria, which are cut by faults having circular and sub-circular shapes (Fig. 1-16).
Figure 1-14: Summit area of Nindiri cone (Masaya volcano) with at left (NorthWest): San Pedro crater, middle: Subsided Nindiri Lava lake and at right (East): Santiago crater. (March 2007).
Figure 1-15: Pit crater structures on Masaya (right) and Nindiri (left) cones [Roche et al., 2001]. Reprinted from Journal of Volcanology and Geothermal Research, vol. 105(1-2), Roche, O., B. Van Wryk de Vries, and T.H. Druitt, “Sub-surface structures and collapse mechanisms of summit pit craters”, Pages 1-18, Copyright (2001) Elsevier.
1.2.4 Volcanic activity of Masaya volcano 2005-2009

1.2.4.1 Explosive activity

Current volcanic activity is concentrated in the Santiago crater and characterized by continuous degassing, since its formation in 1853. Since the beginning of the 19th century, 27 eruptions were observed and several others are uncertain [Smithsonian Institution]. In 2001, a small explosion occurred after several months of the vents being plugged by collapse of the inner wall of the crater. Similar, but smaller vent-clearing explosions occurred in:

- March 30th, 2005: small explosion with ash generated at the Central vent, with Volcanic Explosivity Index (VEI) of 1 [GVN-2005]
- April 29th, 2008: Small explosion in Santiago crater with ash plume. [GVN-2008]
- June 18th, 2008: Small explosion with ash plume [GVN-2008]

1.2.4.2 Santiago Crater change

The magmatic activity has evolved since the observation of Santiago crater in 2006. In March 2006, the main vent showed a red glow only at night (Fig. 1-17). In October 2007, in the North part of the crater, the small vent, not visible during daytime in March 2006, became larger (~3 or 4 times wider), with red incandescence easily visible during the day (Fig. 1-19). At night, the large main vent, responsible for the majority of degassing, did not “glow”. From these observations, it appears that both vents (small and large) are connected to the magma. However, the disappearance of incandescence in the large vent and its appearance in the small vent in 2007 (Fig. 1-19), suggests a very complex structure immediately beneath the crater. The large vent may have a deeper or more complex connection with magma, but remains the main degassing path of the magmatic gas, while the small vent seems to be connected at very shallow depth to the magma.
The different connected paths may explain the strong incandescence and the low level of degassing in comparison to the large vent. During March 2008 and March 2009, no glow was visible in any of the vents (neither in daylight, nor night, Fig. 1-18, 1-20).
In March 2008, both vents inside Santiago crater had shown a change in shape and size in comparison to the previous years (Fig. 1-19). At the same time, the degassing appeared to be slightly more intense, obscuring the vents. The main vent seemed to be slightly wider, and the second vent had grown significantly, at least 5 or 6 times wider than in 2006. From the summit it was quite hard to estimate with accuracy the actual diameter of both vents (Fig. 1-19). This was due to the significant gas volume escaping from them and due to the distance between vent and crater rim (~ 300 to 400 m).

During one month of observation, no glowing was ever observed during daylight or night. It seems that the magma had moved down, perhaps not very far, but with a network of caverns and tubes, it is not possible to be constrain. From visual observation, the quantity of gas in the crater was higher than the two previous years. In March 2009 (Fig. 1-20), the vent seems to appear similar as in 2007; however, no glowing was visible in daylight or at night.

1.2.5 Masaya volcano in the local myths, legend and anecdotes

It seems that as far in the history as it is possible to go back, Masaya volcano has always been considered to be the centre of mystical energy and believe [Lola C. R., 2008]. Before the time of Spanish conquistadors, local communities were performing bloody sacrifices to keep away the anger of the volcano. With the settlement of the Spanish conquistadors and the spread of the Christian religion, local priests and Christian culture transformed Masaya volcano into the gate to hell and so called “El Boca del Inferno” (“Hells-mouth”) [Lola C. R., 2008]. Still today, religious processions are commonly made along the road up to the summit of the Nindiri north flank where the large wooden Christian cross overlooks the active Santiago crater on Nindiri cone, which is believed to protect the community from the “gate of hell”. In my own opinion, if Santiago crater is really the gate of Hell, then the entrance of hell is very quiet and pretty.
1.3 Geological setting of Kawah Ijen volcano

Figure 1-21: Kawah Ijen volcano from Paltuding in 2008.

1.3.1 Continental & Regional Scale

Kawah Ijen volcano (Fig. 1-21) is located on the eastern border of Java island, Indonesia, South-East Asia (8°3.5’ S, 114°14.5’ E). The country is highly populated (~237 million people and ~134 people/km²) [ICSB, 2006]. The North shore of Java Island is bordered by the Java Sea, while the south Java shore is bordered by the Indian Ocean (Fig. 1-22). All Indonesian islands are along the
convergence boundary due to subduction of the Indo-Australian plate beneath the Eurasian plate. Java Island, and thus Kawah Ijen, is on the Eurasian plate, ~250 km north of the Sunda subduction trench [Moore et al., 1980; Global Volcanism Program].

Figure 1-22: Geographical setting of Kawah Ijen volcano, which is located in South-East Asia, on the eastern side of Java Island, Indonesia. Red box corresponds to inset box.

The subduction started ~50M years ago [Simandjuntak et al., 1996] and moves ~7 cm/year perpendicular toward Java Island [Malod et al., 1996; McCaffrey, 1996]. The volcanic arc linked to this subduction is localised on the north side of the trench and makes up Sumatra, Java and the majority of the islands in the area.

Kawah Ijen is one of the 41 active volcanoes of Java and one of the 144 Holocene volcanoes present in Indonesia. Kawah Ijen is located on the south part of the Ijen caldera complex (Fig. 1-23), which is built upon a Mesozoic and Cenozoic basement, ~20 km thick [Sujanto et al., 1988; Hamilton, 1979; Simandjuntak et al., 1996]. The eastern part of Java consists mainly of the products from the Ijen caldera complex and these volcanic deposits are dominant among the quaternary bedrock [Sujanto et al., 1988].
The Ijen caldera complex includes a large caldera structure and 17 post-caldera volcanic cones (Fig. 1-23) [Delmelle et al., 1994; Berlo, 2001; Van Hinsberg, 2001]. The caldera is ~15 km in diameter, with its floor decreasing in elevation northward and converging toward the Blawan water fall (north rim of the caldera), which cuts the Caldera wall [Delmelle et al., 1994]. The highest elevation within the Ijen caldera complex is at ~2600 m a.s.l., located on the south-east rim on the summit of Gunung Merapie (volcanic edifice #3, Fig. 1-23). The main hydrologic structure is the Banyupahit River, which goes across the caldera from the South to Blawan water fall (Fig. 1-23). The first springs of the Banyupahit River are located on the upper west flank of Kawah Ijen volcano and
are due to seepage from the acid lake, which discharges toxic hyper-acidic water. Several secondary springs and small rivers discharge fresh water into the Banyupahit River, leading to a more neutral pH, but nevertheless with a high level of toxicity [Delmelle et al., 1994].

1.3.2 Ijen caldera evolution

The historical evolution of Ijen caldera complex is poorly known. However, it seems that the Ijen caldera complex formed during the Pleistocene together with two other main volcanic centers (Tengger and Ijang) on Java Island [Sitorus, 1990]. The evolution of the Ijen caldera complex can be defined in 4 main periods (Fig. 1-23).

1.3.2.1 Pre-caldera stage

The first stage, around 0.29 Ma ±0.003 consisted of a large strato-volcano building event with sequences of stratified pyroclastic flows, pyroclastic airfall deposits and basaltic to dacitic lava flows [Sitorus, 1990]. These can be found on the northern part of the caldera complex, within the caldera rim. Outcrops are also present outside the south caldera rim, near the village of Songgong [Sujanto et al., 1988; Sitorus, 1990; Berlo, 2001; van Hinsberg, 2001].

1.3.2.2 Ijen caldera stage

There is currently, no exact date for the formation of the Ijen caldera, however, the main event is thought to have occurred between 300,000 to 50,000 years ago [Sitoru, 1990]. Deposits from this event are still present inside and outside the caldera, as thick layers of pumice air fall and pumice pyroclastic flows
[Kemmerling, 1921, van Bemmelen, 1949; Sitorus, 1990; Berlo, 2001; van Hinsberg, 2001]. Near the Caldera rim, the total deposit thickness ranges from 100 to 150 m.

1.3.2.3 Post-caldera stage

The post-caldera stage can be spread over 5 different periods, which include a shallow lacustrine sedimentary period, a resurgent dome period, rim volcanoes and a cinder cone trend (Fig. 1-23).

- The first main post-caldera formation is a non-volcanic structure, with the development of a lake inside the caldera. Sedimentary layers are still present in the north-north-east part of the caldera as interbedded volcaniclastic shales and silts, lahars and volcanic air fall deposits [Sujanto et al., 1988; Sitorus, 1990]. The presence of volcanic ash fall suggest that the lake was present during the reactivation of volcanic activity within the caldera.

- The second period consists of volcanic activity within the caldera and growth of a resurgent volcanic dome: Gunung Blau volcano [Sitorus, 1990].

- The third period is characterized by the construction of four rim volcanoes: Ringgghi, Jampit, Ranteh and Gunung Merapie. The volcanic products have chemical compositions ranging from basalt to andesite and are found as pumice air fall, pyroclastic flow, surge deposits, lahars and lava flows [Kemmerling, 1921, van Bemmelen, 1949; Sitorus, 1990; Berlo, 2001; van Hinsberg, 2001]. The work by Carn et al. [1999] suggests that these volcanic cones are along a tectonic lineament. However, only three of the four rim volcanoes are located along the south rim of the caldera (Jampit, Ranteh and Gunung Merapie, Fig. 1-23), suggesting a significant ring fault
structure, which is probably linked with the caldera formation and shallow magma chamber. It is also possible that both the tectonic lineament and the ring fault structure are linked. The first could have developed into a semi-ring fault structure during caldera formation, by increasing the fracturing in this part of the strato-volcano. Petrological evidence [Berlo, 2001] suggests that Blau (the resurgent dome) and the four rim volcanoes are related to the remnant magma body of the pre-caldera formation.

Figure 1-24: View of the West side of Ijen Caldera in 2008. From foreground to background: Telaga Weru, Kukusan, Blau, Widodaren, Kawah Ijen, Gunung Merapi, Papak and Ringgih. Photo taken from Genteng in 2008 (Fig. 1-23).
• ~50,000 years ago, the lake emptied through the north rim of the caldera. Based on the stratigraphy [Sitorus, 1990], the drainage of the lake occurred in two different events.

• Finally, ~25,000 years ago [Sitorus, 1990], the last period consists of the formation of a volcanic cinder cone trend, oriented east-west, which consists of twelve cinder cones. The most easterly volcanic edifice is Cemara volcano (Fig. 1-23). This trend seems to be more or less parallel to the tectonic lineament [Carn et al., 1999] and suggests a migration of volcanic activity from the west to east across the caldera. The youngest of these cinder cones is Kawah Ijen volcano, which developed on the west flank of Gunung Merapie [Berlo, 2001; van Hinsberg, 2001].

1.3.3 Kawah Ijen volcano

Kawah Ijen volcano is the youngest volcanic cone of the Ijen Caldera Complex and the only one showing signs of activity (Fig. 1-24). Kawah Ijen is a volcanic composite cone, ~700 m high, built on the west flank of Gunung Merapie (Fig. 1-23, 1-25). It summit peak, located on its south crater rim, reaches ~2450 m a.s.l. The volcano has a single elliptical crater, ~1.2 km x 1 km (elongated east-west) and ~500 m deep (Fig. 1-26). About one third of the crater volume is occupied by a large hyper acid lake of ~30 million m$^3$ (pH ~0.00) [Delmelle et al., 1994, 2000; Takano et al., 2004; Löhr et al., 2005]. The edifice consists of alternating layers of ash, scoria and lava flow deposits, which include basalt to andesite magma compositions [Berlo, 2001]. Layers of volcanic deposits are from a few centimeters to several meters thick [van Hinsberg, 2001].
Figure 1-25: Digital elevation model of Kawah Ijen volcano. The Banyupahit acid river flows from the upper flank of Kawah Ijen, around several cinder cones toward the north.
Figure 1-26: Digital elevation model of Kawah Ijen crater. Top: Cross section of the crater. The red elevation line at 2100 m a.s.l. is the approximate level of the acid lake. Bottom: View of the crater. The blue triangles are the springs present in the summit area, above the lake level. The pink polygon represents the location of the sulphuric dome (Solfatara). Blue diamond is the location of the seismic station. The green line is the cross section view of the crater. Bathymetry data from Takano et al. [2004].
1.3.3.1 Constructive period

The first period of the Kawah Ijen history seems to have been dominated by cone building ending with a summit lava flow, that flowed south [van Hinsberg, 2001], however, no specific date is known for the end of this period. Remnants of this lava flow are present on the south flank of the crater and above the sulphur dome (solfatara) [van Hinsberg, 2001].

Figure 1-27: Youngest lava flow on Kawah Ijen. This lava infilled the previous valley of the Banyupahit River. Now the river flows along the lava flow on the west flank of the volcano.
Figure 1-28: Acid crater lake on Kawah Ijen, pH ~0.00, T~37°C in 2007. Background on the left: sulphuric dome; on the right: Lower topographic elevation of the crater rim, where a man-made dam was built to prevent flooding of the Banyupahit river. Beyond the dam is located the upper acid springs of the Banyupahit river.

1.3.3.2 Destructive period

After this building phase, Kawah Ijen entered a destructive phase, which is still ongoing. The actual crater shape is the result of this destruction period and corresponds to a western shift of the eruptive center [van Hinsberg, 2001]. Several lava flows are present on the north flank of the cone and their compositions range from basalt to dacite [van Hinsberg, 2001]. Another lava flow, more than ten meters thick, is present on the west flank of the cone and starts at
the summit (Fig. 1-27). This lava flow, ~9 km long, flowed within the paleovalley and now forms a rim of the current river valley. At one point during the destruction period, the Kawah Ijen crater became sufficiently impermeable to allow for formation of a crater lake. The origin of the lake is probably due to the combination of several processes, such as erosion and sedimentation processes and development of hydrothermal alteration within the crater. Strong hydrothermal alteration and hydrothermal mineral formations will decrease the permeability of the rock. Although the onset of lake development and hyper acidification is unknown, historic reports of its existence were made in 1789 [Hengeveld, 1920].

1.3.3.3 Volcanic activity

During the constructive period of the composite cone, the eruptive style was dominated by magmatic eruptions with lava flows and probably low explosive eruptions with deposition of lapilli (scoria) and ash. During the destructive period, the typical eruptive activity was dominated by phreatic eruptions within the acid lake. Secondary eruption styles are phreato-magmatic and magmatic. The oldest eruption reported was in 1796, while other eruptions are reported to have occurred in 1817, 1917, 1936, 1950, 1952, 1993, 1994, 1999 [Hengeveld, 1920; van Hinsberg, 2001; Takano et al., 2004, Global Volcanism Program]. The only reported magmatic eruption was in 1817 and since the early 20th century, only phreato-magmatic eruptions have occurred [Kusumanidata, 1979; Takano et al., 2004; Global Volcanism Program]. In 1917, following a regional earthquake, the acid lake partially drained, generating a water surge through the Banyupahit River [Hengeveld, 1920; van Hinsberg, 2001]. To prevent future surges, several dams were built over the years on the west rim of crater, which overlooks the valley of the Banyupahit River (Fig. 1-28).
Figure 1-29: Sulphur deposits on the shoreline of acid lake, in 2008. Deposits consist mainly of tiny spherules of sulphur, which rise from the bottom of the lake, drift on the surface of the lake and eventually reach the shoreline.

1.3.3.4 Present day

For at least the last 200 years, the summit of Kawah Ijen has hosted one of the largest and most hyper acid lakes in the world, along with an active sulphur
The main aspect of the volcanic activity is the continuous degassing (Fig. 1-21, 1-24). It is still unknown when this continuous degassing started or if it occured at the same period as lake formation.

1.3.3.4.1 Hyper acid lake

The Ijen acid lake is ~1 km E-W x 800 m N-S x ~180 m deep [Takano et al., 2004] (Fig. 1-26, 1-28). The depth of the lake has varied over time, since 1922 (first bathymetry) [Taverne, 1923], with a decrease in lake volume from 43.5 x 10^6 to ~30 x 10^6 m^3 [Takano et al., 2004]. Changes in lake morphology appear to be controlled mainly by landslides, which fill the crater, but also to a lesser degree by sedimentation and chemical precipitation. Lake shore sediments consist of native sulphur layers (Fig. 1-29) interbedded within fine and coarse-altered sedimentary deposits [van Hinsberg, 2001]. Ash and fine and coarse clasts are also present and come from the different phreato-magmatic and phreatic eruptions which took place within the lake. The hyper acid lake water (pH ~0.00) [Delmelle et al., 1994, 2000; Mauri et al., 2007] is oversaturated in chemical elements, mainly dominated by S\text{composite}, F, Cl\text{2}, Al, Fe, As, HCl and HF. The total dissolved solid concentration (TDS) is above 100 g/kg. An underwater native sulphur layer is inferred to exist at the bottom of the lake [Delmelle et al., 1994, 2000]. The complete chemical composition of the lake water can be found in the work of Delmelle et al. [1994, 2000] and Takano et al. [2004].

1.3.3.4.2 Sulphure dome

In the south-east part of the crater near the lake shore is an active sulphur dome, which covers an area of ~ 40 m by 40 m (Fig. 1-28, 1-30, 1-31). The dome is ~20 m high and made of thick layers (several centimeters to meters) of well crystallized native sulphur and glassy sulphur flows (Fig. 1-32). Older sulphur
deposits are present on the eastern side of the current dome. The degassing center of the sulphur dome appears to have migrated several meters toward the east based on the relative localization of the fumaroles [van Hinsberg et al., 2009], however, no information has been published on the migration of the sulphuric dome. The sulphuric dome is densely covered by fumaroles, with a wide range of temperature over years (250-600°C)[Vigouroux et al., 2007].

Figure 1-30: View of the sulphur dome in 2007. Fumarole temperatures ranged between 250 and 600°C.

Sulphur mining has been active for more than 150 years and was reported during the early period of Dutch colonization. Steel and ceramic pipes are used
by a local mining company to increase sulphur condensation (Fig. 1-30, 1-31). The miners extract ~14 tonnes per day of solid native sulphur. One of the consequences of the mining is increased deposition of sulphur due to sulphuric gas condensation inside the numerous pipes. Several monitoring surveys have been made since at least the 1970’s. The fumarole temperatures were measured at ~216 °C in 1971, ~244 °C in 1979 [Allard, 1986], between 169 °C and 244 °C in 1993 [Delmelle et al., 2000a] and ~219 °C in 1996 [Takano et al., 2004]. However, most of these surveys were from the pipe openings and not directly from the fumaroles. The impact on the gas is a decrease of the SO₂ gas released to the atmosphere, which will only marginally affect the gas flux measurements made by UV spectrometer (COSPEC and FLYSPEC) [Mauri et al., 2007; Vigouroux et al., 2007; van Hinsberg et al., 2009].

**Figure 1-31:** Close up view of the sulphur dome in 2007. T_{Fumaroles} ranging: 250-600°C.
Figure 1-32: Crystallized native sulphur at the outlet of one of the pipes on the sulphuric dome (2007).

1.3.3.4.3 Banyupahit River

The Banyupahit River is the main river in eastern Java Island. With its highest springs directly connected to the hyper acid lake of Kawah Ijen, the stream is highly concentrated in chemical elements, some of which are still at toxic concentrations when the river crosses the countryside to the Java Sea. At more than 5 km from the spring on Kawah Ijen, the river still has a pH < 1 with a strong yellow-greenish colour, due to the high levels of Cl, F and other elements (Fig. 1-33). The smell of H$_2$S from the river is strong. Between 2005 and 2008, there was no change in water temperature (18-20°C, springs and river). The upper spring area is covered by a gypsum sheet more than 200 m along the acid stream (Fig. 1-34).
In the literature, the Banyupahit river is well studied in comparison to Kawah Ijen volcano. The river has a strong impact along its shore, both on ecology and population. High concentrations of Al, F, Cl and Mg are the cause of several types of disease among the population and the plants and animals. A complete study of the impact of toxicology can found in the detailed works of Delmelle et al. [2000b], Heikens et al. [2005a, 2005b, 2005c], Löhr et al. [2005, 2006; 2007] and van Rotterdam et al. [2008a].
1.3.4 Volcanic activity of Kawah Ijen 2005-2009

Between 2005 and 2008, the lake temperature was at a constant temperature of ~37 °C, measured during each survey (measurements made at depth of more than 30 cm below the water surface). The lake over this same period was consistently a green-blue in colour, becoming progressively lighter along the eastern shore, likely due to fresh water influx (Fig. 1-35). Commonly present on the surface of the lake are extensive sulphur sheets drifting with the wind (Fig. 1-36). Such phenomena were previously observed [Delemelle et al., 1994; van Hinsberg, 2001], and are due to accumulation of tiny porous spheres

Figure 1-34: Gypsum waterfall in 2008. Yellow colour is from particles of sulphur within the gypsum crystals.
of native sulphur (Fig. 1-29), which seem to come from the underwater degassing [Mauri et al., 2007; Vigouroux et al., 2007].

![Discoloration of the lake is present every year. Photo from 2007. Main colour is a green-blue colour, which is lighter along the eastern shore, due to input of fresh water.](image)

**Figure 1-35:** Discoloration of the lake is present every year. Photo from 2007. Main colour is a green-blue colour, which is lighter along the eastern shore, due to input of fresh water.

In 2006, fumaroles were measured at the output of the pipes between 220 and 250 °C [Mauri et al., 2007], however, because hot gas will expand and cool inside the pipe, these temperatures underestimate the true fumarole temperatures. In 2007, the direct fumarole temperatures were measured at up to 600 °C, while the gas temperatures at the pipe output were ranged between 150 and 270°C [Vigouroux et al., 2007]. The fumarole temperatures were up to 580 °C in 2008 [Van Hinsberg, 2009]. Average gas flux, measured by UV-spectrometer (FLYSPEC), was ~300 t/day from August 2005 to July 2008 [Mauri
et al., 2007; Vigouroux et al., 2007; van Hinsberg et al., 2009].

Over 2006 to 2009, no eruption was reported, however, local miners state that the volcano seems to have one eruption per year in August. As no one is working during the wet season, it is hard to confirm. Eruptions are described by the miners as being the shape of water explosions, which reach several tens of meters in length, suggesting only phreatic eruptions of small intensity.

Figure 1-36: Sulphur sheet made of tiny balls of sulphur (yellowish colour) on the lake and along the shoreline. Foreground gypsum deposit (white-grey colour), partially covered by sulphur deposits. Photo from 2007.
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2 CHAPTER 02: SELF-POTENTIAL METHODS

2.1 Self-Potential

Self-potential (SP) is an electrical method to measure the distribution of the natural electrical potential at the ground surface. Developed by Japanese researchers in the early 20th century, the SP method has been most commonly used since 1931 for mineral exploration [Broughton, 1931; Ewing, 1939; Poldini, 1939]. It has also been effectively used for water resource management by mapping the extension of water tables and hydrothermal systems. In hydrogeological studies, SP is used to determine the distribution of saturated ground water and to determine the extent of contaminated ground water plumes [Naudet et al., 2004]. In petroleum and mineral exploration, SP is often used to characterize electrical bedrock properties in vertical bore hole cross-sections. In geothermal exploration, Self-potential can be used to rapidly locate hydrothermal areas, prior to applying more expensive and complex techniques such as ground resistivity and drilling. Since the 1970s, SP has been used to survey volcanic activity by qualitatively monitoring hydrothermal change. However, with the recent development of new computer models, SP data can now be analyzed to acquire quantitative information on the depth and flow direction of the water or ore body generating the measured anomaly.

Self-potential is based on the variation of the natural electrical potential through the ground. The natural electrical current is generated by two main phenomena: electrokinetic and thermoelectric coupling [Corwin & Hoover, 1979]. However, other phenomena can also generate this current, such as rapid fluid
disruption, ground resistivity contrasts, redox reactions, human activity, and the telluric field. In nature, the SP signal is generated by a complex interaction between each of these.

Figure 2-1: SP equipment, with high impedance multi-meter, insulated copper wire, hammer, shovel, GPS and copper anode with copper sulphate. Ijen crater, Indonesia in 2006.

2.1.1 SP equipment

The SP equipment consists of two electrodes connected together by a copper cable through a high impedance multi-meter to the ground. A number of different types of electrodes exist such as Pb/PbCl$_2$ [Petiau, 2000; Perrier and Pant, 2005] or Ag/AgCl [Timm and Möller, 2001; Byrdina et al., 2003] and have been used with clay (kaolinite and illite), which is relatively expensive and more commonly used for continuous measurements. The clay allows the electrolyte to stay longer in the electrode and thus to maintain a good contact over time between the electrode and the soil. In this study, all measurements are made with copper-copper sulphate (Cu/CuSO$_4$) non-polarizing anodes (Fig. 2-1a, 2-2), a 300 meter long insulated copper cable and a high impedance multi-meter (200 Mega-Ohm, Fig 2-1) [Finizola, 2002; Mauri et al., 2009]. The copper sulphate
powder is composed of 99% copper sulfate pentahydrate and the copper sulfate electrolyte is saturated. The copper-copper sulfate (Cu/CuSO₄) non-polarizing anodes are inexpensive to make, have good quality response over a day and are the most common electrode type used in SP surveys.

**Figure 2-2:** Self-potential electrode with copper rode. Top: Schematic of the copper anode made at Simon Fraser University based on a smaller model from the Laboratory Magma & Volcan in Clermont-Ferrand (France). Dimension in millimetre. Bottom: dimension in centimetres.
2.1.2 Field procedure

To obtain reproducible data and allow comparison of the data over time during the survey, each SP measurement is made with a sampling step of 20 meters and the position of each point is recorded with a GPS. Measurements are made in a small hole at a depth where there is moist soil to allow for a good contact (e.g., ~5 to 40 cm deep). The potential difference between the electrodes is measured twice per day. To avoid crystallization of copper sulfate within the porous wood plug, all electrodes are emptied, filled with fresh water and reassembled in the same pot of water. As such, the electrode will remain under the same influence of residual copper sulfate solution and will help to avoid polarization of the electrode. Each day before starting, the copper sulfate electrolyte is changed and the copper electrodes are cleaned with sand paper.

There are two common techniques to perform the SP mapping [Corwin and Hoover, 1979]:

- The total-field technique consists of keeping one electrode at the same reference position during the entire acquisition; this requires a wire cable long enough for the entire profile [Corwin and Hoover, 1979].
- The leapfrog technique, also called the gradient technique, uses the same reference point over several hundred meters (depending on the length of the wire cable) and restarting at the last measured point, which becomes the new local reference for the next few hundred meters and so on, until that profile is completed [Corwin and Hoover, 1979; Revil et al., 2002; Mauri et al., 2009].

In this study, we use the leapfrog technique as it is most effective given the size of the survey area. Sections of the profile are joined by adding the different sections together and drift corrected as described in section 2.3. Permanent reference points are preferentially chosen along the shoreline of fresh water (e.g., lake, river, spring).
2.2 Self-Potential theory

2.2.1 Sources of phenomena:

An electrical current is generated due to the differential displacement of ions (anions and cations) through a volume (rock, fluid, etc.), which is characterized by intensity (I in Ampere), a difference of potential (Tension, U in Volt) and by the resistance (R in Ohm) of the volume inside which the ions are moving:

\[ U = R \cdot I \]  
\[ (2-1) \]

When the current is described as the sum of an electrical contribution [Nourbehecht, 1963; Zlotnicki and Nishida, 2003; Yasukawa et al., 2005; Jardini et al., 2006], the equation of the electrical current can be written as :

\[ J_i = \sum L_{ij} \cdot X_j \]  
\[ (2-2) \]

where \( J_i \) is the electric current (A m\(^{-2}\)), \( X_j \) is the driving Force (electrical potential gradient, temperature gradient, pressure gradient, etc.) and \( L_{ij} \) is the cross coupling coefficient associated with the driving force \((i \neq j)\), also called the current coupling coefficient. \( X_j \cdot L_{ij} \), expressed in A m\(^{-2}\), characterizes each of the phenomena generating the electrical current (electrical potential gradient, temperature gradient, pressure gradient, resistivity contrast, etc.).

Each mineral in a rock has a specific polarity, depending on its crystallographic structure and chemical composition, which will generate an electrostatic field, attracting free ions of opposite charge to the mineral surface. The natural electric charge of the mineral will control the amount of free ions that
can be attracted to its surface, and will increase in force when the ions approach the mineral surface. Thus, more energy is required to separate ion layers from the molecular structure of the mineral. Conversely, with increasing distance from the surface, the electrostatic force decreases and less energy is required to liberate the ions. The ion structure is organized in layers, called the Helmholtz double layer [Revil et al., 2003] (Fig. 2.3), along which electrical generation occurs. The capacity of a rock (or mineral) to generate electricity is called the Zeta potential ($\zeta$). These layers of ions (cations and anions) are affected by several parameters including fluid circulation through the rock, chemical composition and pH of the fluid and temperature of the fluid and rock.

2.2.1.1 Electrokinetic coupling (EK)

Electrokinetic coupling is the most important source of natural electrical potential. When fluid (water, hydrothermal fluid, etc.) flows through a porous medium, it generates both a pressure gradient at the head of the flow [Corwin & Hoover, 1979] and an electric potential gradient due to the interaction between the pore surface and pore fluid [MacInnes, 1961]. The stronger the flow pressure, the more energy is transferred to the ions to remove them from the surface of the mineral.

The fluid movement in the pores interacts with the Helmholtz double layer. The outer layer, also called the diffuse layer (Fig. 2-3), is generally formed by positively charged ions (cations) [Avena and De Pauli, 1996; Revil and Leroy, 2001], although it can sometimes be negatively charged [Guichet and Zuddas, 2003]. The type of ions (cations or anions) depends on the polarity of the mineral, but generally speaking, the crystallographic structure of the mineral has sites of $\text{OH}^-$ and $\text{O}_2^-$, which allow them to retain cations.

Thus, when a fluid moves through porous rock, the flow pressure may be strong enough to release cations ($\text{Ca}^{2+}$, $\text{Na}^+$, etc.) from their electrostatic link with the $\text{OH}^-$ (and $\text{O}_2^-$) site. The positive charges will be carried by the flow generating
a difference in the ion balance. The electrical diffuse layer along the flow, which contains the free ions, creates a macroscopic current density and an electrical field, called the Streaming potential [Zablocki, 1976; Zohdy et al., 1973; Corwin & Hoover, 1979]:

Figure 2-3: Helmholtz double layer on the surface of a mineral, which is affected by fluid flow. The Zeta potential ($\xi$) is the quantification of the electrical generation along the Helmholtz double layer (shear plane between moving ions and static ions) [Revil et al., 2003]. Reprinted from Water Resource Research, vol. 39(5), Revil, A., V. Naudet, J. Nouzaret, and M. Pessel, “Principles of electrography applied to self-potential electrokinetic sources and hydrogeological applications”, Pages 1114, doi:10.1029/2001WR000916, Copyright (2003) AGU.

2.2.1.1.1 Streaming Potential

The Streaming potential is the mathematical expression that links the physical process of fluid flow through a porous rock with the electricity generation
measured on the surface. The Streaming potential, $E$, can be expressed as [Pride, 1994; Revil, 1999a,b; Jouniaux et al., 2000]:

$$E = \frac{\rho \varepsilon \xi}{4\pi \eta} \Delta P$$  \hspace{1cm} (2-3)

where $\rho$ is the electrical resistivity, $\varepsilon$ is the dielectric constant, $\xi$ is the zeta potential, $\eta$ is the viscosity of the pore fluid and $\Delta P$ is the pressure drop along the flow path. Among these parameters, the Zeta potential ($\xi$) is the most important and depends directly on the mineral electrical charge as well as the pH and chemical composition of the water (Fig. 2-4).

2.2.1.1.2 Zeta Potential ($\xi$)

The Zeta potential, $\xi$, is the capacity of a rock (or mineral) to generate electricity. The Zeta potential characterizes the voltage across the Helmholtz double layer, when a fluid flows through a porous/permeable medium. The Zeta potential depends mainly on the secondary rock mineralogy due to alteration (zeolite, talc, etc) [Perrier et al., 2003], and the pH of the fluids (Fig. 2.4). Other parameters include the temperature and the conductivity of the fluids [Ishido et Mituzani, 1981; Ishido et al., 1983; Lorne et al., 1999; Jouniaux et al., 2000; Hase et al, 2003; Zlotnicki and Nishida, 2003; Aizawa et al, 2008]; however, it is less sensitive to Major element rock composition [Aizawa et al, 2008]. The Zeta potential is measured in mV and can be negative or positive, although, it is most often negative. A number of studies have shown that the Zeta potential may change differently depending the pH and the secondary mineral composition [Guichet and Zuddas, 2003; Zlotnicki and Nishida, 2003; Aizawa et al., 2008]. The Zeta Potential, $\xi$, can be expressed as [Ishido et Mituzani, 1981;
Ishido et al., 1983; Lorne et al., 1999; Jouniaux et al., 2000; Guichet and Zuddas, 2003; Hase et al, 2003; Revil, 2003; Aizawa et al, 2008]:

\[
\frac{\Delta \phi}{\Delta P} = \frac{-\varepsilon \xi}{\sigma \eta}
\]  

(2-4)

with \(\sigma, \eta, \varepsilon\) being the electric conductivity, viscosity of the pore fluid and the dielectric constant, respectively. \(\Delta \phi\) is the measured electrical potential and \(\Delta P\) is the variation of water flow pressure. The mineralogy of the rock will change with alteration and the formation of secondary minerals. Thus, in hydrothermal systems, where there are hot boiling fluids (one or two phases: gas and/or liquid), the alteration processes are very strong and the Zeta potential can be strongly modified by the mineral alteration.

Rock will be dissolved and modified at depth due its interaction with the hydrothermal fluids, and the hydrothermal fluid composition will in turn be modified (chemical composites and pH, Table 2-1), generally becoming more acidic. The uprising hydrothermal fluids will lose their heat along the way, which will trigger the precipitation at shallow depths of the chemical elements present within it. The precipitated elements and fluids will generate further alteration of the surrounding rock and formation of secondary mineral deposits (e.g., zeolite, talc, etc.) [Ishido et Mituzani, 1981; Ishido et al., 1983; Lorne et al., 1999; Jouniaux et al., 2000; Guichet and Zuddas, 2003; Aizawa et al, 2008]. When the alteration becomes sufficiently advanced, it will directly affect the value of the Zeta potential, modifying the Streaming potential and with it, the SP generation.
2.2.1.1.3 Porosity of the rock

The Zeta potential, $\xi$, is dependent on the electric conductivity, which depends on the rock porosity. The higher the porosity of the rock, the higher the electrical conductivity, expressed in Siemens per meter (S m$^{-1}$). The total electrical density, $J$, (in A m$^{-2}$) can thus be defined, depending on the porosity, as [Revil et al., 2004]:

$$J = -\sigma (\Delta \varphi - E \cdot \Delta P)$$  \hspace{1cm} (2-5)
where $\sigma$ is the electrical conductivity of porous rock (S m$^{-1}$), $\varphi$ is the electrical potential (V), $E$ is the electrokinetic coupling coefficient (V Pa$^{-1}$) and $\Delta P$ is the fluid pressure (Pa). Similarly, the electrokinetic coupling ($E$) can be expressed as [Revil et al., 2004]:

$$E \equiv -\frac{L}{\sigma} \quad (2-6)$$

where $L$ is the electrokinetic coupling term (A Pa$^{-1}$ m$^{-1}$).

As rock porosity increases, the surface contact between fluid and mineral surface will increase, leading to an increase in the Zeta potential by increasing the charge across the Helmotz double layers and finally increasing the SP generation (Table 2-1).

<table>
<thead>
<tr>
<th>$\sigma_f$ (in S m$^{-1}$)</th>
<th>100–200 $\mu$m</th>
<th>200–400 $\mu$m</th>
<th>400–630 $\mu$m</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.001</td>
<td>−189</td>
<td>−203</td>
<td>−168</td>
</tr>
<tr>
<td>0.004</td>
<td>−106</td>
<td>−99</td>
<td>−85.3</td>
</tr>
<tr>
<td>0.010</td>
<td>−25.5</td>
<td>−22.6</td>
<td>−20.6</td>
</tr>
<tr>
<td>0.050</td>
<td>−2.05</td>
<td>−1.98</td>
<td>−1.95</td>
</tr>
</tbody>
</table>

Electrokinetic coupling is the most important source of electrical generation; however, its origin depends on both physical and chemical parameters, notably the Streaming potential, Zeta potential, porosity and mineral composition of the rock, fluid and pH. At the head of the flow, the generated current will be positive (Fig. 2-5) due to the higher concentration on cations, while at the back of the flow, the diffuse layer has fewer cations [Pride, 1994; Revil, 1999a,b; Jouniaux et al., 2000]. In hydrogeological environments where gravitational down-flow of water is dominant, the effect of electrokinetic coupling on the self-potential measurement results in negative SP anomalies and an inverse SP/altitude gradient [Jackson and Kauahikaua, 1987; Finizola, 2004]. In hydrothermal systems, by contrast, where the fluids are dominantly up-rising, the effect of electrokinetic coupling will result in positive SP anomalies [Corwin & Hoover, 1979; Zlotnicki et al., 1998; Finizola, 2004]. When the SP anomalies are only generated by the electrokinetic effect, they will delineate the limit between the saturated zone and the vadose zone.

The saturated zone is the area of ground where all the pores of the rock are full of fluid (meteoric, sea water or hydrothermal), while the vadose zone is the area where pores are not saturated in fluid. If the SP anomaly is only due to electrokinetic process, the SP anomaly range will be more or less proportional to the thickness of the vadose zone [Jackson and Kauahikana, 1987]. Typically this area is above the saturated zone where the vertical displacement of fluids occur.
2.2.1.2 Thermoelectric coupling

Thermoelectric coupling is the second major source of SP signals. When the rock has a temperature gradient across it, a voltage gradient will be present [Corwin & Hoover, 1979]. This phenomenon, called the Soret effect, is due to the differential thermal diffusion of ions between those within the pore fluid and those within the rock matrix [Heikes and Ure, 1961]. As with electrokinetic coupling, thermoelectric coupling is defined by one coefficient: the thermoelectric coupling coefficient, as the ratio of the voltage to the temperature difference, $\frac{\Delta V}{\Delta T}$, (Fig. 2-5). In areas where high temperature is concentrated at shallow depths (e.g., thermal fluids in a fault zone), the electrical generation can be strong enough to be detected by SP measurements. Generally, SP anomalies due to the thermoelectric effect are of shorter wavelength and smaller amplitude (typically a few ten's of mV) than anomalies generated by the electrokinetic effect.

2.2.1.3 Rapid fluid disruption (RFD)

Rapid fluid disruption (RFD) is an ephemeral phenomenon occurring over short periods of time, which characterizes a transition state of the fluid between two stable states. It occurs when liquid is vaporized by heat flux from a magma body or is removed by transport from hot gas [Johnston et al., 2001]. The RFD is one of the main phenomena in the hottest regions, such as volcanic vents and near magmatic intrusions (dyke or sill) and occurs at any depth. However, because it characterizes an ephemeral process (i.e., rapid vaporization of a water body), RFD can generally be detected only by continuous SP measurements. If vaporization due to sustained heat flux becomes continuous, the system will reach a new steady state and RFD will no longer have an effect. The electrokinetic effect is considered as the origin of this sustained process.
During a spatial SP survey, which generally occurs over days or weeks, it becomes very difficult to determine if a SP anomaly is the result of RFD, which generally occurs on a scale of hours or less. Thus, during long SP mapping campaigns, where mapping of structures is important, RFD is generally included within the electrokinetic effect rather than as an independent process. In areas without known heat sources, the RFD does not exist and electrokinetic coupling dominates and occurs at shallow depths. Commonly, the rapid vapour transition and up-rising of boiling gas-fluid generates a positive anomaly that can be up to several hundred mV, but only over a very short period of time.

2.2.1.4 Effect of ground resistivity contrast

As electricity flows through the rock, it is directly affected by the rock resistivity ($\rho$), and inversely by the rock conductivity ($\sigma$), which will control how easily the ions can move through the rock (equation 2-3; 2-4, respectively). In the case of homogeneous ground, the significance of the rock resistivity value is minimal. However, over several km of SP profiles, the rock properties are generally not homogeneous and thus the ground resistivity may significantly change and impact SP generation. The variation in ground resistivity contrast will affect the SP generation by the same order of magnitude (eq. 2-3). Recent studies [Minsley et al., 2007] have shown that ground resistivity contrasts may be significant in the SP generation and thus in inverse or forward modeling of the source generating the SP anomaly. To avoid these problems, a survey of the resistivity at each measurement site is suggested to prevent electrical artifacts due to the local high resistivity of subsurface [Finizola, 2004; Nishida and Tomiya, 1987]; however, it is not always possible to perform a resistivity survey due to logistical constraints.
Contrasts in resistivity will be due to physical differences in the rock properties, e.g., mineral composition, porosity, permeability and presence of water flow within the pores. Commonly, low ground resistivity is associated with underground water flow, ore bodies, or strong rock alteration by earlier hydrothermal alteration [e.g., Revil et al., 2004; Finizola et al., 2006; Revil et al., 2008]. In the case of active volcanoes, the probability of large ore deposits is small; however, the presence of water and strong hydrothermal alteration is very likely. As water flow is present in areas with higher porosity and thus low resistivity, one can reasonably assume that ground resistivity contrasts, on volcanoes, are associated with water flow and thus the electrokinetic effect. Similarly, rock alteration on volcanoes is often associated with ongoing or earlier hydrothermal activity. At the scale of the volcanic edifice, or at the scale of the SP sampling step (commonly 10 to 50 m), electrical generation by ground resistivity contrasts will be a secondary phenomena in comparison to the electrokinetic effect. As ground resistivity contrast is commonly associated with underground water flow, its effect on SP generation can be associated with electrokinetic SP generation, even if they are two different processes.
2.2.1.5 Acid fluids & the electrochemical effect

When the chemical composition and concentration of fluids (e.g., water table, lake, or hydrothermal fluids) are heterogeneous, chemical differences will affect the electrical current generation by changing the charge balance between the rock surface and the water flow. In the natural environment, when water is highly enriched in dissolved chemical elements (Fig. 2-6) (e.g., acid lake, or polluted water table with a TDS of 100s g kg\(^{-1}\)), the heterogeneity of element concentrations will generate a polarization of the fluid, which will influence the electrical measurement [Zlotnicki and Nishida, 2003].

![Figure 2-7: Sketch of SP signal generation/ fluid displacement in volcanic area in heterogeneous ground. Modified after Mauri [2004].](image)

2.2.1.6 SP signal generation in volcanic areas
In a volcanic edifice, such as a basaltic or composite cone, the volcanic structure is always heterogeneous. The edifice may be composed of alternating layers of more or less permeable scoria, highly fractured lava, volcanic breccias, ash and altered layers. The less permeable layers may consist of lava flows with a low fracture density as well as poorly to extremely altered layers [Stimac et al., 2003].

Hydrothermal fluids may spread through the edifice via the more permeable layers and preferentially across fractures (Fig. 2-7). The heat flux from the shallow magma chamber (or shallow intrusion), will heat the deep hydrothermal fluids, which may boil and partially transform into a gas phase. These boiling fluids (hot liquids and gas) move upwards to the surface along the fractures due to the dilatation of fluid by heat [Johnston et al., 2001]. During the upward migration, the boiling fluids lose energy by heat transfer to surrounding environment (cooler hydrothermal fluids and rock) [Merlani et al. 2001; Chiodini et al., 2005]. When these boiling fluids cool, they cease to rise, stabilizing at a given level, before eventually sinking back down due to gravity (Fig. 2-7). In the layer where the boiling fluids stabilize, hydrothermal fluid convection will develop between the rising boiling fluids and the cooler pre-existent fluids in the rock layers. As the fluids (cold and hot) are rich in ions (mainly anions), and there is a difference of kinetics and temperature, the differential displacement of ions generates an electromagnetic field induced by the current density.

2.2.2 Noise phenomena

In nature, an electrical current is typically generated by the three main phenomena discussed above; however, other phenomena may generate a small natural current or disturb the natural current [Corwin and Hoover, 1979].
Conductive mineral deposits generate negative SP anomalies above the deposit [Sato and Mooney, 1960], which can be up to 100 mV or more in amplitude and seldom exceed a few hundred meters in width. The conductive minerals are pyrite, pyrhotite, chalcopyrite, chalcocite, magnetite, covellite and graphite (Fig. 2-8) and are typically found in hydrothermal areas [Corwin & Hoover, 1979].
2.2.2.2 The Telluric current

The Telluric current is a long-period current, generated inside the Earth by temporal variations of the natural magnetic field. Keller and Frischknecht [1966] determined that the period of the telluric current is between 10 and 40 s and may be of several hundred mV km\(^{-1}\) over resistive terrain. In areas of uneven topography, such as a volcano, the lateral resistivity variation of the ground is important as is the variation of topography. However, Telluric current variations will only be significant for continuous SP measurements over several days or months. In SP mapping, the telluric variation may affect the reference station, however, any effect will be corrected by closing the profile in a loop and applying a drift correction over the entire loop.

2.2.2.3 Non-thermal subsurface water streaming

Non-thermal subsurface water streaming generates important variations in SP measurements. Rainfall percolating through the soil will create an important electrical current, which may mask or increase the noise of the current generated by deeper phenomena (e.g., hydrothermal circulation, water table). This type of noise is characterized by accentuation of the topographic effect [Poldini, 1939; William et al., 1976]. As the effect is difficult to constrain, it is best to avoid it where possible by only making SP measurements at least several hours or a day following rain.

2.2.2.4 Human activity

Human activity is an important problem for SP measurements, when the study area is close to population centers. Power lines, underground power lines, corrosion of pipelines or buried metallic debris can generate an electrical current, which can strengthen the anomaly by tens or hundreds of mV km\(^{-1}\), up to a few
kilometers from the source. The electrical signal may be steady, or take the shape of individual spikes or pulses or irregular variations, or a series of sinusoidal or square waves [Corwin & Hoover, 1979]. Furthermore, as seen during our fieldwork, any disturbed surface (e.g., road, etc.) creates a strong contrast of the ground resistivity, permeability and porosity and thus a variation of the Zeta potential, leading to increased noise in the signal.

2.2.2.5 Chemical heterogeneity

Heterogeneity can be expressed through different forms, such as ground resistivity contrast (discussed previously), but can also correspond to chemical heterogeneity within the electrolyte, or the ground. Electrochemical effects near the measurement site correspond to variations of the physical and chemical parameters (e.g., chemistry, temperature or moisture) within the ground and generates electrical noise (Fig. 2-9), which can have spatial wavelengths as short as a few centimeters and amplitudes generally of ±10 mV [Corwin & Hoover, 1979]. This noise is lower in areas where the soil is moist and uniform. The non-polarizing electrodes (copper-copper sulfate) that we use are more easily affected by this type of noise.

Contamination of electrolytes and moisture are important sources of noise during the field measurements. To achieve a good SP signal, the electrochemical contact between the soil ions and the electrolyte is necessary [Corwin & Hoover, 1979]. For that, the first step is to choose a site which is sufficiently moist; however, in areas, such as hydrothermal zones or highly contaminated ground water where the flux of fluids are rich in chemical species and after a long period of contact with the soil, the electrolyte within the electrode will become contaminated by these chemical species. This contamination will result in a drift of the SP measurement due to polarization of the electrode. To avoid this contamination and drift, it is necessary to constrain the potential between both
electrodes, change the electrolyte and clean the electrode if the SP value differs from the initial value. Another influence linked with the hydrothermal area, is the temperature increase of the electrolyte by the hot hydrothermal fluid. This temperature increase will create a potential across the electrode pair. For the saturated copper-copper sulfate electrodes, the temperature coefficient is \( \sim 0.5 \text{ mV} \degree \text{C}^{-1} \) [Ewing, 1939; Poldini, 1939].

![Figure 2-9: Solfatara of Kawah Ijen (August 2006), with ground covered by several tens of cm layer of native sulphur. Photo b courtesy A.E. Williams-Jones.](image)

### 2.2.2.6 Noise reduction

The noise can be estimated through direct analysis of the data. Noise due to poor contact between electrode and ground can generate up to several ten’s of mV and can be easily avoided by checking at each measurement point the contact resistance of the electrode with the ground using the high impedance multi-meter. A poor contact is characterized by a large increase of the resistance, often several orders higher than a good contact (< 100 kOhm on ash, < 1 MOhm on soil and scoria, < 10 MOhm on lava flows).
The main noise on recorded data is the drift of the measurement over the survey, which is controlled and corrected when the measurement profiles are closed in a loop. Generally, a large loop is made to support the secondary loops, which are corrected in reference to it.

The secondary sources of noise are estimated in amplitude and size by making comparison with the large structures of the main SP signal. Secondary noise is always due to local disruptions of the electrical signal and generally has small wavelengths of less ~80 m in volcanic areas [Finizola et al., 2002]. Traditionally, any small anomalies defined by less than 4 consecutive values (for a 20 m step) must be considered as noise, because there is insufficient information to distinguish between noise and a real structure. Thus, with a measurement step of 20 m, it is possible to find only the structures which have a SP signal longer than 60 m. Secondary noise can also be detected by a simple cross correlation with the field notes, which should include information about human activity or other information that may be useful.

2.3 SP Data analysis & interpretation:

2.3.1 Analysis of SP data

2.3.1.1 The corrections

Since each SP field measurement is the differential potential between two points, it is necessary to make corrections for all data. The first step in the correction is the reference correction, which is merely normalization of each data point to the same reference. The second step is to correct any drift occurring during the measurement period. As the measurements are made in the shape of a loop, theoretically, the value of the first point and the last point must be equal. However, because the measurements are not made at the same time, there is always a drift. When measurements are correctly made and the environmental
conditions are good, the drift should be very small (a few mV). The drift value is equal to the difference between the first value of the loop and the last value, which spatially are the same point:

\[ N_{SPi} = SP_i + i \times C_{Incrementdrift} \]  

(2-7)

with \( N_{SPi} \) the corrected SP value of the drift; \( SP_i \) is the SP value for point \( i \); \( C_{Incrementdrift} \) is the value of the gradient of Drift correction per point and \( I \) is the total of increment:

\[ C_{Incrementdrift} = \frac{-\left( SP_{lastSP} - SP_{firstSP} \right)}{I} \]  

(2-8)

with \( SP_{firstSP} \) and \( SP_{lastSP} \) the first SP value and the last SP value of the loop, respectively; \( SP_{lastSP} \) and \( SP_{firstSP} \) must have the same GPS coordinate. For example, with a measurement loop of 10 points, the point #1-Ref and #10 are the same spatial point. Along the loop, there are 2 reference points: point #1-Ref and point #6.

Finally, if there is a point along the profile, which is located at a constant fresh water area (e.g., stream, lake, etc.), which is not necessarily equal to 0 mV on the raw values, this point should be considered as the absolute reference point. As a consequence, all the other points should be shifted in value to accommodate the point at 0 mV. A reference point is chosen along the shore line as there is no SP generation in fresh water (no interaction between water and surface of rock). By definition, the difference between 2 points on surface water is always 0 mV, except if there is a very strong indication of chemical composition heterogeneity. The SP data are typically analyzed by different methods: SP profiles, SP maps, SP/elevation gradient, SPS surface [Corwin &
Hoover, 1979; Aubert et al., 1990; Finizola et al., 2004] and SP processed by mathematical techniques as described below.

<table>
<thead>
<tr>
<th>point A</th>
<th>raw SP</th>
<th>Ref correction</th>
<th>loop corrected SP</th>
<th>N pt X</th>
<th>Point 1 &amp; 10 are the same spatial point</th>
</tr>
</thead>
<tbody>
<tr>
<td>#1-Ref</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td></td>
<td>Difference 10-1 = 5</td>
</tr>
<tr>
<td>#2</td>
<td>2</td>
<td>2+0 = 2</td>
<td>2.9</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>#3</td>
<td>3</td>
<td>3+0 = 3</td>
<td>4.8</td>
<td>2</td>
<td>Loop Corrected pt A = Corrected Ref SP pt X + [(0 – (-8)) / 9 * N pt X]</td>
</tr>
<tr>
<td>#4</td>
<td>-1</td>
<td>-1+0 = -1</td>
<td>1.7</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>#5</td>
<td>-5</td>
<td>-5+0 = -5</td>
<td>-1.4</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>#6</td>
<td>-10</td>
<td>-10+0 = -10</td>
<td>-5.6</td>
<td>5</td>
<td>Reference point</td>
</tr>
<tr>
<td>#7</td>
<td>-8</td>
<td>-8 + -10 = -18</td>
<td>-12.7</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>#8</td>
<td>3</td>
<td>3 + -10 = 7</td>
<td>13.2</td>
<td>7</td>
<td></td>
</tr>
<tr>
<td>#9</td>
<td>6</td>
<td>6 + -10 = -4</td>
<td>3.1</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>#10</td>
<td>2</td>
<td>2+ -10 = -8</td>
<td>0.0</td>
<td>9</td>
<td>SAME GPS corrdinate that Point #1-</td>
</tr>
</tbody>
</table>

Table 2-2: Example of synthetic SP profile, structure of the different correction. Loop corrected SP are calculated following the equations (2-8) and (2-9).

2.3.1.2 SP profiles and SP map

Self-potential data are most often analyzed as SP profiles which transect the study area in segments that are considered representative of the area. To obtain a global view of the SP signal, the data are interpolated through gridding to generate a SP map [e.g., Corwin & Hoover, 1979; Finizola et al., 2002; Finizola et al., 2004; Revil et al., 2008]. As seen in the Figure 2-5, in the
hydrogeological zone, the SP signal is increasingly negative with an increase in altitude along the slope.

### 2.3.2 Interpretation based on SP amplitude anomaly

As seen above, negative and positive SP anomalies do not mean the same thing in hydrogeological and hydrothermal areas. On active volcanoes, both areas are present. Typically, the hydrogeological area is localized on the lower slope and the shallow hydrothermal area is on the upper slope and summit of the volcanic edifice. An excellent example is Misti volcano, Peru [Finizola et al., 2004]; however, it is not always the case [Lénat, 2007].

The amplitude of SP anomalies varies from one volcano to another, but generally is from 100 to several 100s of mV [Azaiwa et al., 2008]. However, this can sometimes be as low as only a few 10s of mV in areas where the Electrokinetic effect is negated by very high acidity or very dry ground. In contrast, in areas with a strong heat flux and strong rainfall infiltration, SP anomalies can reach a few Volts. Thus, the wide range of anomaly amplitude shows that SP anomalies for a similar process are strongly influenced by environmental conditions, as well as the underground structure and typical SP anomaly range from few 10’s of mV to few 1000’s of mV. Based on the literature and field experience, the SP anomaly signature of a water table is generally between 50 to 150 mV.

### 2.3.3 Interpretation based on SP / elevation gradient

The interpolation of the SP data may be improved by analysis of the relationship between the SP and elevation, called the SP/elevation gradient (Ce) [Finizola et al., 2004]. Generally, an inverse SP/elevation gradient (negative Ce
number) characterizes a water table with down flow of water [Corwin & Hoover, 1979; Aubert and Atangana, 1996; Finizola et al., 2002; Lénat, 2007]. Traditionally, the SP/elevation gradient is between -1 and -10 mV m\(^{-1}\) [Ishido, 1981], however in volcanic areas, the SP/elevation gradient is generally between -1.5 and -4 mV m\(^{-1}\) [Aubert and Atangana, 1996]. A gradient plot with no clear trend may characterize an area of strong ground heterogeneity along the profile. A positive gradient, asymmetric gradient, or symmetrical gradient characterizes a hydrothermal environment where fluids are moving principally along a vertical or sub-vertical structure [Zlotnicki and Nishida, 2003; Lénat, 2007]. When SP signals are on the border between two areas, the SP/elevation gradient or SP signal rise up (or decrease down) very steeply (e.g., Fig. 6-12, 7-12). The break in slope on the SP profile is characteristic of the transition between the two areas. In the hydrothermal zone, the SP signal is not or only weakly dependent on topography, because the fluid movements are typically vertical. The positive SP anomalies are characteristic of the upward movement of hydrothermal fluid and the negative anomalies represent downward movement of hydrothermal fluid. However, due to the topographic effect, the SP/elevation gradient on each side of the hydrothermal system can be asymmetric [Lénat, 2007]. The SP/elevation gradient may also be made in a 2D-approach by using a SP map (which requires a grid structure of the SP data) and a high resolution DEM (Digital elevation model) [Finizola et al., 2004; Lénat, 2007]. A 2-D SP/elevation gradient may allow for better constraint on the underground water structures within the volcanic edifice. The main limitations of the 2-D approach are: a) the quality of the DEM, which must be without artifact; b) the SP profiles making the grid must be close enough to avoid generation of artifacts during the gridding of the data. The main difficulty in the field, when working on an active volcano, is the large variation of topographic elevation (e.g., cliffs, unstable slopes, deep valleys, etc.) and the density of the vegetation (e.g., dense forest, jungles, etc.), which make it very difficult to do proper SP mapping.
2.4 2-D analysis of Self-potential data

2.4.1 Analysis Methods

This study uses Muti-scale wavelet tomography to model the depth of the sources generating the Self-potential signal; however, other methods exist:

- Depth of the electrical source based on the curve of the SP anomaly is one of the earliest SP analysis techniques:
  - Based on a few SP points characterizing the anomaly [Yungul, 1950; Paul, 1965, Bhattacharya and Roy, 1981].
  - Matching the curve between a SP anomaly and synthetic SP signal [Atchuta Rao and Ram Babu, 1983; Ram Babu and Rao, 1988; El-Araby, 2004].

- Depth based on least-squares technique include:
  - Iterative calculations which use non-linear least-square optimization of both field and synthetic data [Shalivahan et al., 1998; Abdelrahman and Sharafeldin, 1997; Abdelrahman et al., 1997].
  - Constrained nonlinear least-squares optimization techniques [Asfahani and Tlas, 2002, 2005].

- Depth based on derivative calculation of the SP anomaly following resolution of the mathematical expression of the anomaly and requires resolution of both linear and non linear equations. This method seems to be able give both the structural order and depth of the source generating the anomaly [Abdelrahman et al., 1998, 2003; Srigutomo et al., 2008].
- Depth of water based on Fournier Method using the SP/elevation gradient, SP data and piezometer data to calculate the surface of the water table. A complete description can be found in [Fournier, 1989; Birch, 1993].

- Depth of water based on the SPS surface is the limit between the Saturated and Vadose zones. Calculation of the SPS requires a SP profile along with at least 2 piezometers [Aubert, 1994; Aubert et al., 2000]. This technique is very sensitive to noise affecting the SP data.

- Spectral analysis in wave number-domain is based on the Fourier and Hilbert transform and is a spectral analysis that should be able to give depth of the source generating the anomaly [Atchuta Rao et al., 1982; Sundararajan et al., 1990; Sundararajan and Srinivas, 1996; Sundararajan et al., 1998].

- Depth based on Green’s function. Almost all the techniques described here use the Green’s function at one point or another in the calculation. Recent work has used the Green function to decompose the SP signal in order to localize in 3D the source generating the SP anomaly. This type of modeling is significantly affected by resistivity contrasts [Minsley et al., 2007; Mendonça, 2008; and references within].

- Depth based on distribution of electrokinetic generation uses synthetic SP signals by grid distribution of water flow, zeta potential, permeability, or porosity [Yasukawa et al., 2002, 2003; Bolève, 2007; Jardani et al., 2007, 2008].

- Neural Networks is a mathematical technique developed for seismic analysis and has been applied, with some success, to other geophysical techniques such as SP. It is based on a layered system, where the
complexity is determined by the user in terms of a number of input parameters. The number of layers is a function of the complexity of the problem. The technique uses results from a synthetic SP signal to resolve each of the different layers of the analysis to eventually determine the depth of the source generating the SP anomaly [Heshman and Mansour, 2009].

- **Dipolar occurrence probability (DOP)** is a mathematical technique [Patella, 1997; Saracco et al., 2004] based on the occurrence of a probability function and scanning function. The DOP works well if the source is a dipole or monopole and if its nature (dipole or monopole) is known.

- **Continuous Wavelet transform** uses wavelet analysis based on the Poisson family wavelet to calculate the depth of the source generating the SP anomaly. When the depth is calculated on the correlation of several wavelet analyses, the method is called multi-scale wavelet tomography [Sailhac and Marquis, 2001; Saracco et al., 2004; Mauri, submitted 2009]. This technique is fully described in Chapter 4 of this thesis.

### 2.4.2 Synthetic Self-potential signal

Synthetic signals are often as important as the real signal measured in the field and allow us to test and validate the signal analysis methods before applying them to real data. This study use synthetic Self-potential signals based on punctual sources in order to validate the Multi-scale wavelet methodology (Chapter 4).

The synthetic Self-potential signals are calculated with code written in MATLAB and based on two types of source equations. The first type of electrical source is a punctual dipole, where input variables are the depth, horizontal
position, length of the dipole and charge of the dipole. Equation 2-9 [Telford et al., 1995] describes the potential distribution of a dipole:

\[ V = q \left[ \frac{1}{\sqrt{x^2 + z^2_i}} - \frac{1}{\sqrt{(x-a)^2 + z^2_2}} \right] \]

with the charge, \( q \), of the dipole in V m\(^{-1}\). The horizontal distance, \( x \), is in m and corresponds to the distance from the left side of the dipole to the measured point on the surface. The length of the dipole, \( a \), is in m. The depth of the left corner of the dipole, \( z_1 \), is in m. The depth of the right corner of the dipole, \( z_2 \), is in m.

All the synthetic signals used in this thesis are considered to be within a homogenous medium, which allows us to consider that there is no effect from resistivity contrast and thus to use the equation 2-9. Several examples of synthetic SP signals can be found in Chapter 4 (Fig. 4-2, 4-10, 4-11) and Chapter 5 (Fig. 5-2). While these synthetic signals are very simple models, they are not used to model the water flow within a volcanic edifice, but rather to investigate the capacity and the accuracy of the Multi-scale wavelet tomography method to locate the source of an electrical signal.

2.5 Conclusion

The Self-potential method is a passive electrical method, which allows for the investigation of the horizontal extension of underground water flow. SP can be used to differentiate aquifers (having a down flow driven by the gravity effect) from the hydrothermal fluids (up rising fluids, which are supported by high fluid pressure and may be boiling). Due to ambiguity between one water type and another, SP is commonly combined with soil CO\(_2\) concentration and ground
temperature in order to better characterize the hydrothermal fluids. Field examples are described and discussed in Chapter 5, 6, 7 and Appendix A1. Although, several techniques of 2-D analyses have been briefly described here. This study uses the Continuous wavelet transform method on the SP profiles (Chapter 5, 6, 7 and Appendix 1). This signal processing technique has been chosen for its capacity to combine multi-scale analyses and depth calculation, allowing us to obtain reliable information on the depth of the underground water in order to investigate the dynamic behaviour of the water.

2.6 References


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3 CHAPTER 03: PUNCTUAL SURVEYS AND CONTINUOUS GRAVITY MEASUREMENT IN VOLCANIC ENVIRONMENTS.

3.1 Introduction

The variation of mass or density in the subsurface can be measured by gravity and micro-gravity monitoring techniques. In volcanic environments, the dynamic behaviour of volcanoes is often characterized by variations of mass and/or density within the volcanic edifice. On one hand, the variation of mass may be due to an uprising of fresh hot magma to shallow depths within the volcanic edifice (e.g., magma chamber, dyke intrusion, magmatic column, lava lake, etc.). On the other hand, a variation of density may be due to physical changes within the magma body (e.g., crystallization, cooling, degassing). A variation of mass or density of magma generates a gravity change of several tens to several hundred μGal and can be measured by two main methods: the spatial survey (Bouguer mapping) and the temporal survey (dynamic and continuous). This chapter gives a brief overview of the gravity phenomenon, the principle of gravimetry and the technical characteristics of the LaCoste & Romberg Gravimeter G127 [LaCoste & Romberg, 2004], used in this study. Finally, this methodology chapter will focus on the common types of gravity applications in volcanology and the methods of data correction used in this study.
3.2 Gravity theory

Gravity has been known since Galileo Galilei (1564-1642) first studied the relationship between the weight and speed of dropped objects. The universal law of gravitation was developed in the last half of the 1600s by Sir Isaac Newton (Mathematical Principles of Natural Philosophy, 1685-1687). In 1735-45, the main gravitational relationships were discovered by Pierre Bouguer (French mathematician and astronomer). However, it was only during the first half of the twentieth century that the first stable portable gravimeter was developed by LaCoste & Romberg (1939) [Torge, 1989; Telford et al., 1995]. Since the 1980s, gravity surveys have also been made by satellite observation [Kahn, 1983]. Since middle of 1990s, with the development of advanced computer modeling, gravimetry studies have become more accurate through forward and inverse modelling of the internal sources (within the Earth and extra-planetary objects), which generate the gravity anomalies [Becker et al., 1995].

The principle of gravity was developed by Newton, through Newton’s Law of gravitation (first law) and states that between two masses \((m_1, m_2)\), there is always an attractive force, \(F\), such that \(F\) is directly proportional to the product of the masses and inversely proportional to the square of the distance, \(r^2\), between the centers of mass:

\[
F = \gamma \left( \frac{m_1 \cdot m_2}{r^2} \right) \vec{r}
\]

\(F\) is the force on \(m_2\), \(\vec{r}\) is a unit vector directed from \(m_2\) toward \(m_1\), \(r\) is the distance between \(m_1\) and \(m_2\), and \(\gamma\) is the universal gravitational constant \((6.674 \times 10^{-8} \text{ N cm}^2 \text{ g}^{-2})\). The force, \(F\), is in \(\text{N (kg m s}^{-2}\). From equation 3-1, the force of gravitation, \(F\), can be expressed through the total gravity acceleration, \(g_{\text{total}}\), which is the acceleration of the mass, \(m_2\), due to the presence of \(m_1\) in space:
\[ g_{total} = \frac{F}{m_2} = \left( \frac{\gamma m_1}{r^2} \right) \hat{r} \]  

(3-2)

with \( g_{total} \) in N kg\(^{-1} \leftrightarrow 10^{-5} \text{ m s}^{-2} = 10^{-3} \text{ Gal} = 1 \text{ mGal} = 10^3 \mu\text{Gal}.

The total acceleration of gravity, \( g \), is a 3 component force, where the vertical component, at least near the surface of the Earth (and other planetary bodies), is the strongest component.

\[ g_{total}\hat{r} = g_x\hat{x} + g_y\hat{y} + g_z\hat{z} \]  

(3-3)

To accurately measure the acceleration of gravity generated by the Earth, the vertical component must be isolated, which is done by placing the gravimeter horizontally (perpendicular to the vertical component). A simplification of the vertical component of gravity acceleration, \( g_z \), is called simply \( g \).

### 3.2.1 Value of \( g \) and gravity corrections:

The value of the total acceleration of gravity, \( g \), corresponds to the average acceleration of gravity on Earth, chosen at the reference sea level (Geodetic reference system 1967), and is equal to 978,031.846 mGal [Telford et al., 1995]. The local variation of \( g \) can be measured with a gravimeter (different types exist), which allows us to make measurements of \( g \) with a precision on Earth of \( 10^{-2} \text{ mGal} = 10 \mu\text{Gal} = 10^{-7} \text{ m s}^{-2} \). Some gravimeters have a higher resolution, however, they are generally too heavy or costly for field measurements.

Globally, at the scale of Earth, \( g \) is maximal at the poles and minimal at the equator and varies locally due to the heterogeneity of the lithosphere. The intensity of \( g \) is a function of 5 parameters, which are variations of:
✓ latitude,
✓ elevation,
✓ density,
✓ topography of the surrounding terrain,
✓ earth tides.

The general model of the geodetic Earth is the reference spheroid, which is the shape of the Earth (determined by geodetic and satellite measurements) and approximately the average altitude of the Earth (about sea level, called the geoid). When \( g \) is measured on the surface of the Earth (equation 3-2), \( m_1 \) is the mass, \( M \), of the Earth, \( r \) is the radius of the Earth, \( R \), and \( \mathbf{r} \) is a vector unit, which extends downward toward the center of the Earth. Mass and distance are the main parameters which must be corrected for the area where the measurement is made. As Earth is an ellipsoid, the gravity acceleration, \( g \), will vary with the latitude, \( \phi \) in degree, and decrease from the equator toward the pole. The average value of \( g \), corrected for latitude is [Telford et al., 1995]:

\[
g = 978,031.846 \times (1 + 0.005,278,895 \times \sin^2 \phi + 0.000,023,462 \times \sin^4 \phi)
\]

(3-4)

with the gravity, \( g \), in mGal and the latitude, \( \phi \), in degree.

The latitude correction, \( C_{g_{lat}} \), can be simplified as:

\[
C_{g_{lat}} = \frac{\Delta g_{lat}}{\Delta \text{latitude}} = 0.811 \sin(2\phi)
\]

(3-2)

with \( C_{g_{lat}} \) in mGal km\(^{-1} \) and \( \phi \) in degree.

As the surface of the Earth is not a smooth ellipsoid, a local topographic correction, called the *Free-air correction*, \( C_{g_{f-a}} \), must be made to adjust the acceleration of the gravity [Telford et al., 1995]:
\[ C_{g_f-a} = -0.3086 \times \Delta \text{elevation} \] (3-6)

where the free-air correction, \( C_{g_f-a} \), is in mGal.

As the density of the lithosphere is not uniform, the local density variation, \( \Delta \rho \), can be corrected through the Bouguer correction, \( C_{g_{Boug}} \) (mGal), for a layer of thickness, \( h \) [Telford et al., 1995]:

\[ C_{g_{Boug}} = 2 \pi \gamma \Delta \rho \Delta h = 0.04192 \Delta \rho \Delta h \] (3-7)

where \( \Delta \rho \) is in g cm\(^{-3} \), \( \gamma \) is the universal gravitational constant (\( \gamma = 6.674 \times 10^{-11} \) N m\(^2\) kg\(^{-2} \) ↔ 6.672 \times 10^{-13} \) N cm\(^2\) g\(^{-2} \)), \( \Delta h \) is in meter.

Quite often, for Bouguer mapping and spatial forward or inverse modeling, the Free-air correction and Bouguer correction are combined in the Elevation correction, \( C_{g_{elevation}} \) (mGal), which is the combination of equation 3-6 and 3-7 [Telford et al., 1995]:

\[ C_{g_{elevation}} = (0.3086 - 0.04192\Delta \rho)\Delta h \] (3-8)

with the density variation, \( \Delta \rho \), in g cm\(^{-3} \) and \( \Delta h \) in m.

As seen in equation 3-2 and 3-3, the acceleration of gravity is a 3 dimensional phenomenon, where the vertical component is isolated during the measurement. However, when significant change in topography occurs near the measurement station, topographic changes (positive or negative) will have an impact on the vertical gravity acceleration. This local topography effect can be corrected through the terrain correction, \( C_{g_{terrain}} \), calculated using the Hammer disk [Hammer, 1939], where each topographic point is grouped with its neighbour to form a sector of disk, which is associated to an average value. The result is a
disk divided into sectors (equation 3-9) and the total terrain correction, $C_{g_{terrain}}$, is the sum of all the $\delta g_{terrain}$ (equation 3-10) [Telford et al., 1995]:

$$\delta g_{Terrain}(r, \theta) = \gamma p \theta \left( \left( r_0 - r_i \right) + \left( r_i^2 + \Delta h^2 \right)^{\frac{1}{2}} - \left( r_0^2 + \Delta h^2 \right)^{\frac{1}{2}} \right)$$

(3-9)

with $\delta g_{terrain}$ the terrain correction of one sector of the hammer disk with the polar coordinate $(r, \theta)$, the inner sector boundary, $r_i$, and the outer boundary, $r_0$, are in m. The polar angle, $\theta$, is in radians. The variation in elevation between the gravity station and the sector, $\Delta h$, is in m. The density variation, $\Delta \rho$, in is in g cm$^{-3}$. The smaller the sector surface is, the more accurate will be the associated topographic correction. The sector surface increases with increasing distance from the gravity station. The outer boundary, $r_0$, must be chosen by the user, based on how variable the topography is around the gravity station (plains, mountains, cliffs, valleys, etc). Generally, an outer bound, $r_0$, of a few km is sufficient.

$$C_{g_{terrain}} = \sum_r \sum_{\theta} \delta g_{terrain}(r, \theta)$$

(3-10)

The terrain correction, $C_{g_{terrain}}$, is in mGal, the distance between the station and the surface, $r$, is in m and the angle, $\theta$, is in radians. $C_{g_{Terrain}}$ is the sum along $r$ and $\theta$ of every point of polar coordinate $(r, \theta, \Delta h)$, with the difference in elevation, $\Delta h$, between the surface point and the gravity station [Telford et al., 1995]. Any topographic point above or below the elevation of the gravity station generates an error on the vertical component of $g$, which is corrected in equation 3-9.

The last significant parameter, but also the most important, is the tide, which is the vertical deformation of Earth’s surface through the gravitational attraction of the Moon and the Sun. The tide consists of two components: the
Earth tides and the oceanic tides [Jaggar, 1923; Hamilton, 1973; Dzurisin, 1980; El Wahabi et al., 2000]. When the gravity station is near the shore of a water body (e.g., lake, ocean), the effect of the vertical displacement of water must be corrected in the vertical gravity acceleration, $g$. On any land surface, near the shore or in land, the earth tide must corrected from the vertical gravity acceleration, $g$. Both tides (earth and ocean) can generate gravity errors on the measurements ranging from a few $\mu$Gal to several hundred’s of $\mu$Gal. The tide correction depends on the relative position of the Sun and the Moon with respect to the gravity station. As Earth orbits around the Sun and the Moon orbits around Earth, following specific wavelength frequency and at a known distance from Earth, it becomes possible to accurately calculate the synthetic tide value for any place and time on the surface of Earth. Several software programs (free or licensed) exist; this study uses QUICK TIDE PRO from Microg Lacoste [MicrogLacoste] for Bouguer and dynamic gravity surveys. All continuous gravity data are corrected for tide by using TSOFT [Vauterin, 1997], which allows for correction and handling of large quantities of data at once. Both software programs have an accuracy of 1$\mu$Gal.

Other corrections should be made when the atmospheric conditions, such as temperature and pressure, change significantly, [Merriam, 1992; El Wahabi et al., 1997; Warburton and Goodkind, 1997; Andò & Carbone, 2001, 2004]. Other sources of noise can be due to a change of the internal pressure or temperature of the gravimeter, and also due to wind and earthquakes (of any intensity). Seismic noise can generate gravity noise when the frequency of the seismic wave is within the frequency range of the gravimeter [Iyer and Hitchcoc, 1974]. Sometimes, an earthquake or significant shock to the gravimeter may induce a temporary or permanent tare, thus corrections must be done to adjust all the data to the same tare reference [Davies, 2000].
3.3 LaCoste & Romberg Gravimeter

A LaCoste & Romberg Model G-127 land gravity meter with liquid electronic level and Aliod 100 Upgrade with Digital Readout to 0.01 mGal [LaCoste & Romberg, Inc., 2003, 2004] was used in this study. The gravimeter remains at a constant temperature of 50.1 ±1 °C in order to avoid any calibration drift. With the Aliod, the gravimeter can be used with a precision of 0.01 mGal in the range of measurements between -50,000 and +50,000 μGal. The feedback Aliod 100 system records 2 measurements every second and uses a 40 second digital FIR (finite impulse response) low pass filter to remove high frequency background noise (common seismic background on Earth) [LaCoste & Romberg, 2003].

All LaCoste & Romberg meters employ a zero-length spring to make the gravity measurement. With the zero-length spring, the size of the spring can be adjusted to stay in a range of measurements of the gravimeter. The tension of the spring is proportional to its actual length and when all external forces of attraction are removed, the spring returns to its zero-length reference. Tests on various LaCoste & Romberg gravimeters [e.g., Rymer, 1989; Ander et al., 1999; Andò and Carbone, 2004] show that the precision of the instrument over several months of measurements is about ± 15 μGal.

3.4 Method of investigation by gravimetry

3.4.1 Spatial gravity surveys

A spatial gravity investigation is generally the first step for any gravity work in a new area and is made with Bouguer Mapping. Generally, only one measurement campaign is necessary to obtain a density image of the subsurface. Bouguer mapping allows us to determine the underground structures and localize sources of the gravity anomaly, such as:
Magmatic systems (i.e., chambers, dykes and conduits) [Metaxian, 1994; Fournier et al., 2004; Gailler et al., 2009; Jordan et al., 2009],

ore bodies and hydrothermal systems [Hunt et al., 2009],

karst [Jacob et al., 2009],

The maximal depth that can be investigated depends on the extent of coverage of the area and of the sampling step of the measurement [Telford et al., 1995]. A Bouguer mapping grid is interpolated by Kriging into a regular grid and interpreted through a contour map of the gravity variations [e.g., Metaxian, 1994; Telford et al., 1995; Fournier et al., 2004; Gailler et al., 2009; Hunt et al., 2009]. The smaller the sampling step of the grid, the smaller the gravity anomaly is that can be detected. Generally, the measured anomaly ranges from a few hundred μGal to several mGal.

3.4.2 Temporal gravity surveys

Temporal gravity surveys consist of making micro-gravity measurements by re-occupying the same network of stations over a period of time (generally over several years). Temporal gravity changes generally range between a few μGal to several hundred μGal and characterize the dynamic behaviour of the anomaly source, such as a water table, magmatic system, etc. The change can be due to a change in mass (volume) or in density of the source. Temporal gravity surveys can be made at different time scale; continuous gravity measurements have a sampling rate ranging from one minute to a few hours and allow us to characterize very short variations due to short term change. The dynamic gravity survey includes any kind of temporal survey with a sampling rate longer than a day and allows us to characterize events of medium to long period, which can be expressed by phenomena having a wavelength from a few days to several years.
A general overview of dynamic phenomena that can be observed by temporal gravity survey for sources at shallow depths (first few kilometers below the topographic surface):

- magmatic recharge (medium to long period) [Dzurisin et al., 1980; Jachens and Eaton, 1980; Carbone et al. 2003],
- magmatic crystallization or cooling (long period), degassing of magma / magmatic convection (short period) [Rymer, 1994; Metaxian et al. 1997; Williams-Jones et al., 2003; De Zeeuw-van Dalfsen, 2005],
- magmatic tide, lava lake variation (short period) [Jaggar, 1923; Nakagawa and Shimozuru, 1970],
- water table variation (medium to long period) [Kroner et al., 2004; Imanishi et al., 2006; Naujoks et al., 2008],
- hydrothermal fluid circulation (short to long period) [Todesco and Berrino, 2005; Todesco et al., 2008],
- karst formation (medium to long period) [Jacob et al., 2009],
- ore deposit formation (medium to long period) [Sander and Cawthorn, 1996; Sandrin et al., 2007].

In addition to underground sources, there are also:

- Seasonal effects (medium to long period) [Merriam, 1992; El Wahabi et al., 1997; Warburton and Goodkind, 1997; Andò and Carbone, 2004],
- Atmospheric variations (short to long period) [Beldon and Mitchell, 2009; Mitchell and Beldon, 2009],
- Earth and oceanic tides [El Wahabi et al., 2000].

### 3.4.3 Gravimetry application in Volcanology

Volcano gravity studies in conjunction with deformation studies are used to characterize underground volcanic structures and dynamics over time. Gravimetry surveys are a useful tool to monitor volcanic activity and help to
forecast future eruptions [e.g., Rymer and Brown, 1986; Rymer, 1994; Rymer, 1996 ; Gottsmann et al., 2003; Kauahikaua and Miklius, 2003; Locke et al., 2003; Carbone et al, 2006; Battaglia et al, 2008; Carbone et al, 2008 ; Crider et al., 2008; Williams-Jones et al., 2008; Carbone et al., 2009].

Underground volcanic structures are characterized by contrasts of density within the medium, such as:

- voids within a cavern or conduit [e.g., Metaxian, 1994; Rymer et al., 1998b; Rymer and Williams-Jones, 2000],
- cumulates within a (active or old) magma chamber [Metaxian, 1994; Battaglia et al., 2003a,b; Gottsmann et al., 2008; Gailler et al., 2009; Jordan et al., 2009],
- volcanic structure boundaries (buried caldera rims or buried craters) [Metaxian, 1994; Battaglia et al., 2003a,b; Gottsmann et al., 2008; Gailler et al., 2009; Jordan et al., 2009],
- buried volcanic edifices beneath sedimentary rock [Battaglia et al., 2003a, 2003b; Cassidy et al., 2009; Jordan et al., 2009].

Volcanic structures can be found by spatial Bouguer mapping, where the density and the extent of the gravity station network will determine the minimal dimension of the structure that can be characterized. Contrasts of density due to rock type variation will generate gravity anomalies ranging from a few hundred μGal to several mGal.

Volcano dynamics over time will generate changes in gravity, when the change in mass or density occurs within the volcanic edifice. Typically, changes in mass or density are of much smaller size than that due to the density contrast of underground volcanic structures and thus generate gravity variations ranging from a few μGal to several hundred μGal. Volcano dynamics over time include volcanic phenomena, such as:
- magmatic dyke intrusion (through surface deformation) [Dzurisin et al., 1980; Jachens and Eaton, 1980; Carbone et al., 2006, 2007, 2008, 2009],
- change in height of the magmatic column within the shallow magmatic conduit [e.g., Rymer, 1994; Rymer and Williams-Jones, 2000; Williams-Jones et al., 2003],
- change in density of the magma within the shallow magmatic conduit (due to magma renewal, cooling and degassing processes) [Dzurisin et al., 1980; Rymer, 1994; Rymer et al., 1998a; Rymer and Williams-Jones, 2000; Williams-Jones et al., 2003]
- large changes in mass, density within a shallow magmatic chamber [Dzurisin et al., 1980; Jachens and Eaton, 1980; Rymer, 1994; Rymer et al., 1998a],
- volcanic tide within a lava lake or magma within an open magmatic conduit [Jaggar, 1923; Nakagawa and Shimozuru, 1970],
- changes in the hydrothermal system [Berrino et al., 1984; Bonafede and Mazzanti, 1998; Battaglia et al., 2006; Tikku et al., 2006; Hunt et al., 2009; Hunt and Graham, 2009].

Quite often, changes in gravity occur with other quantitative physical or chemical changes, which allow us to constrain the volcanic changes. These physical or chemical changes are commonly:

- surface deformation [Berrino et al., 1984; Bonafede and Mazzanti, 1998; Bonafede and Mazzanti, 1998; Rymer et al., 1998a; Rymer and Williams-Jones, 2000; Battaglia et al., 2003a,b],
- electrical variation [Paoletti et al., 2008; Hunt et al., 2009],
- magnetic change [Guillen et al., 2008; Paoletti et al., 2008; Dutra et al., 2009]
- variation in degassing flux [Williams-Jones et al., 2003]
- seismic activity [e.g., Carbone et al., 2006, 2008].
3.5 Field methodology, analysis method and modeling method

3.5.1 Measurement methodology

This study uses a LaCoste & Romberg Model G gravimeter (G127) with a feedback system [LaCoste & Romberg Inc., 2003] as described previously. At each gravity station, measurements are made with an average sampling rate of 1 minute over a period of at least 5 minutes (often 8 min) after stabilization of the signal. Results are automatically recorded on Palm OS PDA and written in a field notebook as backup, with a brief bedrock description (for Bouguer mapping). At the same time or following the gravity measurement, the GPS coordinates are recorded. For temporal gravity surveys, differential GPS is used to assess to a few centimeters of vertical accuracy, this allows us to remove gravity variation due to elevation change. For Bouguer mapping, the GPS localization are obtained by handheld GPS, which allows us to get only ~7-10 m horizontal resolution and ~20 m vertical resolution. While the vertical accuracy is very poor with the hand GPS, the final vertical position will be determined by digital DEM having vertical accuracy of a few meters. The DEM elevation also allows us to avoid discrepancies between the DEM elevations used for data processing (e.g., modeling and terrain correction) and the GPS elevation value, and thus helps to reduce error when doing terrain correction and modeling process.

Daily measurements are made by starting and ending each day at the same reference station to control drift and tares which could occur during the day. When there are strong elevation variations (more than 200 m), an intermediate gravity reference station is made, to allow for recalibration of the mass counter ID within the detection windows (100 mGal) of the Alid 100 system. The intermediate reference station will be occupied before returning to the reference station. In order to control the linearity of the counter ID and avoid measurement artefacts, at each reference station (daily and intermediate), a counter ID calibration is done over a full range of the Alid 100 system detection windows.
3.5.2 Methodology for the correction process

The types of correction that must be applied to the gravity data depend on the type of survey which is made, such as Bouguer mapping, dynamic survey or continuous measurements.

3.5.2.1 Bouguer mapping

In the case of the Bouguer mapping, the correction processes are as follows:

- **Earth tide correction** (and oceanic tide if required) of each measurement for each station.
- **Counter ID drift correction** is probably not necessary with all gravimeters. However, with G-127, a full turn of the ID counter wheel is not exactly of 1 mGal. The graduation of the wheel does not match with the real shift affecting the ID counter value when the wheel is turn.
- The shift is very small of -0.0443 mGal by turn (44.3 μGal), unfortunately this shift become a big problem when doing a survey of an area with large topographic variation, which require to adjust the ID counter of several tens of mGal between different areas. The shift must be calculated for each new geographical in order to avoid to miss correct it. A test has been done at Simon Fraser University (SFU) within the Physical Volcanology lab (Fig. 3-1). Data sets were collected at the same station by shifting the ID counter of a few mGal between each set of measurements. In the test at SFU, 2 data sets, which included in 39 data points, were collected over 2 days by period of 4 hours each day.
Figure 3-1: Drift of the counter ID of G-127. Test made at Simon Fraser University in November 2007. a) Blue diamond symbols represent the residual gravity corrected of tide. The black line is the linear trend (and its equation above it) associated to it. The orange triangle symbols are the residual gravity after correction of the ID counter drift. b) Zoom on the residual gravity after correction of the ID counter drift. The residual gravity does not show any linear trend.
The shifting of the ID counter has been done by step of a few mGal each time over a range of 72 mGal. Residual gravity data (blue diamond on Fig. 3-1) were corrected for tide. No time dependent drift were corrected (see section 3.5.2.3 with the continuous measurement), because over the period of measurement, the drift will be of only a few μGal, which is of several order smaller than the drift from the counter ID itself. The drift from the counter ID is linear allowing us to correct it by calculating the best fit (Fig. 3-1). Similar drift was found on Kawa Ijen volcano, during the gravity survey (see section 7).

Daily drift correction is based on the variability of gravity values of the reference station (daily and intermediate). Correction of the drift is applied to all the stations of the same day only if, over this day, the variability of the reference station value is bigger than 80 μGal. Another daily drift is the field campaign drift correction is the same type of correction as that for the daily drift, except that this time the gravity value of the reference station is observed over the entire period of the field campaign (generally 2 to 3 weeks). Correction is made by applying a linear correction each day.

Single gravity value per station. As each station has at least 5 gravity values (and several ten’s to hundred for the reference station), an average value is calculated for each station based on the mean of the data set available for the station.

Latitude correction to correct each station for the Earth curvature, as described previously.

Terrain correction using the Hammer method, as described previously in value of g and gravity corrections section.

Even elevation plane brings all the stations to the same elevation level by applying the elevation correction (Free air and Bouguer correction) at each station.

Traditionally, to obtain an anomaly map that is easiest to read, all the stations are normalised by subtracting the Station reference value from them. Thus, the reference station will have a 0 mGal value and the others
will have a positive or negative gravity value (in mGal) in comparison to this station.

✓ Finally, Kriging is calculated based on all the gravity station GPS position and DEM elevation and Final gravity value. The more homogeneous the distribution of the station over the map and higher the density of station is over the map, the better will be the interpolation of the gravity anomaly grid. This study used the Surfer 8 software from Golden Inc. to generate Kriging and anomaly maps.

✓ Generally, several planar sections at different elevations (m a.s.l.) are calculated to observe how gravity changes.

3.5.2.2 Dynamic gravity survey

In the case of the dynamic gravity survey, only the first four steps are similar to the Bouguer correction.

✓ *Earth tide correction.*

✓ *Daily drift correction* is this time necessary if the variability of the gravity station is greater than 20 μGal. This is due to the fact that dynamic gravity survey is used to investigate very small gravity variations, generally less than 150 μGal.

✓ *Single gravity value per station.*

✓ *Reference station at 0 μGal value.* As the reference station is supposed to be invariant over time, and in order to compare the change occurring at each station over time, all the stations are normalized by subtracting the Station reference value from them. As such, the sign of the gravity value of each station corresponds to the relative change (- for a decrease and + for an increase).

✓ *Vertical elevation correction* is to counterbalance the vertical deformation that could happen during the occupation site between the year of reference survey and the year of the actual survey. Based on the
differential GPS survey, the variation of vertical elevation is calculated. If the variation is at least two times larger that the vertical accuracy of the differential GPS, the elevation change can be considered as real and must be corrected. The correction value is based on the equation of the free air correction, equation 3-7.

- Time series data are generally displayed on plot of time vs. gravity.

### 3.5.2.3 Continuous gravity survey

The correction process of a continuous gravity survey is quite different from the two previous gravity methods. In the case of the continuous gravity survey, the gravity station is permanently occupied by the gravimeter, which records data every minute or less. As the gravimeter is both the reference station and the recording station overtime, several parameters must be constrained, such as atmospheric pressure and temperature, ground deformation and if possible a seismometer to extract seismic noise (tremor, earthquake). A second continuous gravimeter outside the active volcanic area should be installed to characterise any gravity change of non volcanic origin. During the continuous measurement, the gravimeter must be kept in a horizontal position as much as possible, by checking the horizontal level at least twice a day and adjusted if necessary. The correction process regroups:

- **Earth tide correction**.
- **Jump correction** corresponds at any tare that can affect the gravimeter over the recording period. This will generally be due to seismic noise (e.g., rock fall, earthquake and tremor).
- **Field campaign drift**, which is very important over several days or more
- **Atmospheric pressure** correction may be necessary if the gravity record shows any kind of correlation with the atmospheric pressure [Andò and Carbone, 2001; Carbone et al., 2003; Andò and Carbone, 2006], such as:
\[ C_{g_{\text{Pressure}}} = -0.365 \times \Delta P \]  

(3-3)

with the correction of atmospheric pressure, \( C_{g_{\text{Pressure}}} \), in \( \mu \text{Gal} \) and the pressure variation, \( \Delta P \), in mbar.

✓ **Atmospheric temperature** correction may be necessary if the gravity record shows signs of the atmospheric temperature effect. Normally, this correction is not necessary, because the internal thermostat of the gravimeter is always constant and above the atmospheric temperature. However, if necessary, the correction must be in two steps. First, the time delay between temperature and gravity variation must be determined by time-cross correlation. The Neuro Fuzzy approach can then be applied to correct the gravity data from the temperature effect [Andò and Carbone, 2001; Carbone et al., 2004; Andò and Carbone, 2006].

✓ **Reference station at 0 \( \mu \text{Gal} \) value.** To allow a more accessible reading of the gravity variation, all the data are corrected by subtracting the value of the first gravity value, which will become after correction a 0 \( \mu \text{Gal} \) value.

✓ Data are generally displayed through a time vs. gravity plot.

An example of continuous gravity measurement is shown on Fig. 3-2 and 3-3. Data were recorded with the G-127 gravimeter (from LaCoste & Romberg) at Simon Fraser University (Burnaby, BC, Canada) in the Physical Volcanology lab (+49.915N, -122.915E, 349 m a.s.l) during 7 days (from December 14\textsuperscript{th} to 21\textsuperscript{th}, 2007, \sim 167.22 hours). Sampling rate was 1 min. Atmospheric pressure variations were recorded with a HoboMicro station from ONSET Corporation. The levels of the gravimeter were monitored several times a day and adjusted when necessary. No significant change in the level occurred (level data are not presented in this example.). Comparisons between the deformation levels (X and Y) of the meter and the residual gravity have shown no significant correlation.
Figure 3-2: Raw, earth tide and residual continuous gravity in December 2007. Sampling step is 1 min. From top to bottom, atmospheric pressure (black line), atmospheric temperature (red line), raw gravity (blue line), synthetic tide (earth and oceanic tides, in green line) and residual gravity (purple line). The residual gravity has been corrected of tide, spikes, jumps, drift and pressure effects.
Figure 3-3: Spectral frequency structure of raw gravity, earth tide and residual continuous gravity on December 2007. Sampling step is of 1 min. From top to bottom, atmospheric pressure (black line), atmospheric temperature (red line), raw gravity (blue line), synthetic earth tide (green line) and residual gravity (purple line). The residual gravity has been corrected of tide, spike, jump, drift and pressure effect.
Pressure accuracy was 3 mbar and temperature accuracy was 0.5 °C. After correction for earth tide, oceanic tide, drift and pressure, the residual gravity data (purple line on Fig. 3-2) shows variation smaller than 20 μGal. Raw gravity, tide (earth and oceanic), atmospheric temperature, atmospheric pressure and residual gravity data were analysed by Fast Fourier Transform in order to compare their frequency spectra (Fig. 3-3). Tide (earth and oceanic) and Raw data have each of them two main frequency components of period of ~ 24 and 13 hours. Temperature variations show a main time period over ~42 hours (Fig. 3-3). Pressure signals do not show any significant frequency component, but rather a random variation having period of variations smaller than 28 hours.

The residual gravity (purple line, Fig. 3-3) presents only one significant frequency component of a period of ~42 hours. Thus, all influence from tide (earth and oceanic) has been removed from the residual gravity signal, which do not show one main frequency component. The only frequency left, which may be significant, has a period of ~ 42 hours. This period of 42 hours is also the main frequency component of the temperature signal; however, the power spectrum of the temperature signal for the 42 hours period is of very small amplitude (Fig. 3-3) in comparison to the tide, or raw gravity signals. Their main amplitude frequency is ~ 28 and ~ 15 (unitless), for period of ~ 24 and 13 hours. Finally, it seems possible to efficiently remove any influence from tide (earth and oceanic) and pressure variation, which could affect the gravity signal. The temperature change may have an influence on the continuous gravity measurement, variations should be small, because the residual gravity (not corrected of temperature) shows changes of amplitude smaller than 20 μGal (Fig. 3-2), which is traditionally below the accuracy of the measurement for the G-127 gravimeter.
3.5.3 Forward and inverse modeling of gravity data’s

Gravity data are also used in modeling by looking at variations of mass and density required to generate the observed anomaly. Modeling can be done through a forward or inverse approach. This study uses the forward approach by assuming a reasonable value of the significant parameter controlling the variation of mass and density, such as porosity, density (gas, magma, rock) and volume. This study used Grav3D software from UBC, which allow us to do 3D processing [Li, and Oldenburg, 1998; Gailler et al., 2009]. Results of the modeling are described in Chapter 5 and 6.

3.6 Conclusion

Gravity surveys can be used to characterize change in both volume/mass and density occurring in the ground over time or space. Each source (e.g., punctual object, extended object) will be characterized at the surface by its own gravity signature, controlled by its own properties (e.g., contrast of density, volume) and its distance to the gravimeter. While gravity investigation records the relative gravity variation of the sum of all sources, it becomes very difficult to characterize the signal associated with each of these sources within a complex structure. However, when combined with other geophysical, or geochemical studies (e.g., deformation, gas flux, seismic, resistivity), it becomes possible to better constrain the gravity model and thus obtain good insight into the origin of this variation. This study applies all types of gravity surveys on two active volcanoes in order to investigate change in volume or density which occurs within them (Chapter 7, Appendix 5 and 6). In this study gravity models are compared with water depth from Multi-scale wavelet tomography applied on Self-potential data to investigate the underground water dynamism (Chapter 7).
3.7 Reference List


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## Mathematical Conventions

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \phi )</td>
<td>potential field</td>
</tr>
<tr>
<td>( \Delta \phi )</td>
<td>variation of the potential field</td>
</tr>
<tr>
<td>( \varphi )</td>
<td>electrical potential</td>
</tr>
<tr>
<td>( \varphi_0 )</td>
<td>electrical potential in ( z = 0 )</td>
</tr>
<tr>
<td>( \sigma )</td>
<td>electrical charge</td>
</tr>
<tr>
<td>( t )</td>
<td>time</td>
</tr>
<tr>
<td>( \vec{x} )</td>
<td>translation vector</td>
</tr>
<tr>
<td>( \vec{z} )</td>
<td>vertical vector</td>
</tr>
<tr>
<td>( \nabla^2 )</td>
<td>Laplacian ( \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2} )</td>
</tr>
<tr>
<td>( \mathbb{R} )</td>
<td>Real space</td>
</tr>
<tr>
<td>( \mathbb{R}^2 )</td>
<td>Real space in two dimensions</td>
</tr>
<tr>
<td>( P )</td>
<td>Poisson kernel symbol</td>
</tr>
<tr>
<td>( \hat{P} )</td>
<td>Poisson kernel symbol in Fourier Space</td>
</tr>
<tr>
<td>( L^2 )</td>
<td>Fourier Space in two dimensions</td>
</tr>
<tr>
<td>( FT(f) )</td>
<td>Fourier transform of the function ( f ).</td>
</tr>
<tr>
<td>( i )</td>
<td>imaginary number ( (i^2 = -1) )</td>
</tr>
<tr>
<td>( F' )</td>
<td>the first derivative of the function ( F )</td>
</tr>
<tr>
<td>( F'' )</td>
<td>the second derivative of the function ( F )</td>
</tr>
<tr>
<td>( e )</td>
<td>exponential</td>
</tr>
<tr>
<td>( D_z )</td>
<td>dilation operator</td>
</tr>
<tr>
<td>( \int )</td>
<td>integral in one dimension</td>
</tr>
<tr>
<td>( \iint )</td>
<td>integral in two dimensions</td>
</tr>
</tbody>
</table>
4.1 Introduction of Continuous Wavelet Transform

The principles of wavelet theory were developed during the 1980s by Grossmann & Morlet in order to resolve questions in seismic signal analysis [Grossman and Morlet, 1984; Grossmann, 1986; Grossmann et al., 1987; Tchamitchian, 1989; Grossmann et al., 1989]. Since then, wavelets and their applications have been used in a wide variety of domains, from fluid mechanics to image processing in astronomy and telecommunications [Daubechies et al., 1992; Farge, 1992; Meyer and Xu, 1997; Addison, 2002]. Wavelet analysis is a multi-scale analysis method that can be used to determine power and frequency spectrums and the spatial distribution of a processed signal. They can be applied for image processing, optimization and image restoration, as well for spatial or temporal signals. Each wavelet is a mathematical equation which defines its specific shape (normally a wave), amplitude, wavelength and frequency [Combes et al., 1989; Mallat, 1989; Chui, 1992; Daubechies et al., 1992; Farge, 1992; Mallat and Hwang, 1992; Guillemaın, 1994; Perrier et al., 1995; Holschneider, 1996; Addison, 2002; Holschneider, 2000, 2003, 2005]. Traditionally, when applied to a signal, multi-scale analysis is made by a series of dilations of the wavelet. In other words, the wavelet is stretched (dilation > 1) and squeezed (dilation < 1), which will change its amplitude and wavelength, but not its surface. Dilation of the wavelet will thus act as band-filter, where small dilation values act as a high-frequency pass-filter and large dilation values will have the effect of a low-frequency pass-filter. Typically, processing of a signal by a wavelet will result in a grid of correlation coefficients, along axes defined by the analyzed signal and the dilation coefficient [Daubechies et al., 1992; Farge, 1992; Meyer and Xu, 1997; Addison, 2002]. The result will give information about the frequency pattern and the structure of the analyzed signal. As with Fourier transform (FFT), wavelet analysis decomposes the signal through a specific succession of coefficients which characterize it. The wavelet transform can be inverted and thus the signal can be recreated using the same wavelet and the specific series of coefficients.
characterizing the signal. These properties make the wavelet a very powerful tool for compressing files and signals. There are many kinds of wavelet, each developed for a specific type of analysis [Mallat and Hwang, 1992; Daubechies et al., 1992; Holschneider, 1996; Addison, 1999; Keller et al., 1999; Addison, 2002; Holschneider, 2000, 2003, 2005]. Among the different families of wavelets, the main groups are the discrete wavelet and the continuous wavelet [Daubechies et al., 1992; Farge, 1992; Meyer and Xu, 1997; Addison, 2002], each of which has its own specifics.

This study use the continuous wavelet transform on potential field data (electric, magnetic or gravimetric data) in order to localize the depth and horizontal position of the source generating the anomalies measured in the field. When wavelets based on the Poisson kernel family are applied to a potential field, it becomes possible to determine the depth and horizontal position of the sources generating the different anomalies [Alexandrescu et al., 1995; Moreau, 1995; Moreau et al., 1997, 1999; Sailhac et al., 2000; Sailhac and Marquis, 2001; Gibert and Pessel, 2001; Fedi and Quarta, 1998; Fedi et al., 2004, 2005; Saracco et al., 2004, 2007, 2009; Boukerbout and Gilbert, 2006; Cooper, 2006]. In this study, the continuous wavelet transform was chosen over other techniques (e.g., FFT) because of its capacity to simultaneously perform both multi-scale analysis and depth determination, which allows us to use all the information making the analysed signal without requiring other information (e.g., well depth, anomaly frequency, etc.).

4.2 Mathematical expression of an Electrical field

4.2.1 Diffusion equation with potential fields

Any potential field, such as an electrical field, acts following the diffusion equation:
\[
\frac{\partial \phi}{\partial t} - \Delta \phi = \sigma
\]  

(4-1)

where \( \phi \) is the potential field, \( \varphi \) is the electrical potential and \( \sigma \) is the electrical charge. At a given time \( t \), the electrical potential, \( \varphi \), can be considered as constant: \( \Delta \varphi = 0 \). Thus the diffusion equation of the potential field, \( \phi \), can be written as the Poisson equation:

\[
\frac{\partial \phi}{\partial t} = \sigma
\]  

(4-2)

This assumption is appropriate, when the time required to collect the electrical data is small in comparison to the variation occurring within the source generating the recorded electrical signal [Saracco et al., 2004].

The surface of the Earth is considered as the elevation reference. At the local scale, such as the field scale, if the topographic variations are small (generally must be less than 150 m) [Saracco et al., 2004], the distortion of the topographic variation on the electrical signal can be considered as negligible. In reality, the topographic surface is never constant, but its effect will ultimately be corrected by adding the topographic elevation to the calculated depth of the source.

For a 2D problem, within a half space, the potential field, \( \phi \), depends on the translation vector, \( \vec{x} \), and the vertical vector, \( \vec{z} \). The translation vector, \( \vec{x} \), corresponds to the measured profile in the field. Depending on the value of \( z \), there are different ways to describe \( \phi \).

v. On the surface of the ground, \( \vec{z} = 0 \), therefore the diffusion equation can be expressed through the Poisson equation:
\[ \phi(x,0) = \phi(x,0) = \varphi_0 \]  

(4-3)

within the defined domain, the electrical potential, \( \varphi_0 \), is equal to the electrical potential, \( \phi(x,0) \), in \( z = 0 \).

- Above the ground, \( z > 0 \), there is no source of electrical variation and potential field, \( \phi \), can described as a constant:

\[ \Delta \phi(x,z) = 0 \]  

(4-4)

4.2.2 Laplace equation with potential fields

When above the surface of the ground, \( z > 0 \), variation of the potential field, \( \Delta \phi \), in 2D problems, can be described by the Laplace equation, which has several implications:

For \( z > 0 \):

\[ \frac{\partial^2 \phi}{\partial x^2} + \frac{\partial^2 \phi}{\partial z^2} = 0 = \nabla^2 \phi \]  

(4-5)

- **First implication**, the electrical potential, \( \Delta \phi \), is still described by the Poisson equation.

- **Second implication**, as the potential field is zero, \( \Delta \phi(x,z) = 0 \), all functions derived from the potential field are both harmonic [Stewart et al., 1999] and holomorphic functions, meaning that \( x \) and \( z \) can be Real or Complex.
4.2.3 Poisson Equation with potential fields

As the potential field is a solution of Laplace equation 4-5, it can also be described through the Poisson equation:

\[ \nabla^2 \phi = -4 \pi \rho \quad (4-6) \]

where \( \rho \) is the charge density and \( \phi \) is potential field.
In the case of \( z > 0 \), the potential field is above the ground, thus there is no charge generation (\( \rho = 0 \)) and can be expressed as in (4-5).

Consequently, from (4-6) and (4-5), and when above the ground (\( z > 0 \)), the \( \Delta \phi(x,z) \) is only defined by \( \phi_0 \) (diffusion equation as \( z = 0 \)), which can be expressed following Poisson Kernel equation, \( P \).

\[ P(\vec{X}) = C_{n+1} \left( 1 + \left| \vec{X} \right|^2 \right)^{(-n+1)/2} \quad (4-7) \]

where \( n \) is the order of the dimension and \( C_{n+1} \) is the constant associated to the dimension of order \( n=1 \). In 2D problems, the vector \( \vec{X} = jx + ky \), thus (4-7) is in the real space, \( \mathbb{R}^2 \) with \( x \in \mathbb{R} \) and \( y \in \mathbb{R} \) [Moreau et al., 1995, 1997, 1999]:

\[ P(\vec{x},\vec{y}) = \frac{1}{1 + \left| jx + ky \right|^2} \quad (4-8) \]

If \( P \) is transformed into Fourier Space, \( x \overset{FT}{\longrightarrow} u \) and \( y \overset{FT}{\longrightarrow} v \), (4-8) will become [Moreau et al., 1997]:

128
\[ \hat{P}(u, v) = e^{-2\pi i \sqrt{u^2 + v^2}} \] (4-9)

Equation (4-9) is thus the Poisson kernel equation, \( \hat{P} \), in Fourier space, \( L^2 \), with \( u \in L \) and \( v \in L \).

### 4.3 Potential field expressed by Poisson Kernel in Fourier Space

I will now discuss why the Poisson Kernel, \( \hat{P} \), in Fourier space, has only one solution within Fourier space [Roddier, 1971; Moreau, 1995, 1997, 1999].

#### 4.3.1 Fourier Transform

In order to simplify notation, let us consider the Poisson kernel, \( P \), as being a general function, \( f \), in normal space, which will have \( F \) as its equivalent in Fourier Space. By the Laplace equation (4-5), in 3 dimensions:

\[ \nabla^2 f(x, y, z) = 0 \] (4-10)

where \( f \) is the simplified notation of Poisson kernel in Real space. The Fourier transform of equation (4-10) will therefore be:

\[ FT(f)(x, y, z) = F(u, v, z) \] (4-11)

where:
\[ F(u,v,z) = (2i\pi u)^2 F(u,v,z) + (2i\pi v)^2 F(u,v,z) + \frac{\partial^2 F(u,v,z)}{\partial z^2} = 0 \] (4-12)

As the square of the imaginary number, \( i^2 = -1 \), \( FT(f) \) becomes:

\[ F(u,v,z) = -\left(2\pi \sqrt{u^2 + v^2}\right)^2 F(u,v,z) + \frac{\partial^2 F(u,v,z)}{\partial z^2} = 0 \] (4-13)

Equation (4-13) thus becomes:

\[ \left(2\pi \sqrt{u^2 + v^2}\right)^2 F(u,v,z) = \frac{\partial^2 F(u,v,z)}{\partial z^2} \] (4-14)

### 4.3.2 Simplification of the Potential field equation in Fourier Space

As in Fourier space, the potential field, \( F(u,v,z) \), is a constant for \( z=0 \) and can be written as:

- \( C_{ct} = F(u,v,z = 0) \) (4-15)

Equation (4-14) can then be simplified as follows:
where \( F' \) and \( F'' \) are the first and second derivatives of the function \( F \), respectively. \( C \) here is the function defined by equation (4-16).

Thus, based on equation (4-14), it becomes possible to write the following equivalence:

\[
F'' = FC^2 \iff F'F'' = FC^2 F'
\]  

(4-19)

Both sides can be raised to the integral order (4-19):

i. \[
\int F''F' = \frac{1}{2} (F')^2 + C_{cst1}
\]  

(4-20)

ii. \[
\int FF'C^2 = \frac{1}{2} (F)^2 C^2 + C_{cst2}
\]  

(4-21)

Where \( C_{cst1} \) and \( C_{cst2} \) are two constants.

Using equations (4-19), (4-20) and (4-21), equation (4-14) becomes:

\[
\frac{(F')^2}{2} + C_{cst1} = \frac{F^2 C^2}{2} + C_{cst2}
\]  

(4-22)

Equation 4-22 can then be simplified to:
\[
\frac{(F')^2}{F^2} = C^2 \tag{4-23}
\]

Replacing \(C\) by its value (4-16), equation (4-23) becomes:

\[
\sqrt{\frac{(F')^2}{F^2}} = \left(2\pi\sqrt{u^2 + v^2}\right) \tag{4-24}
\]

### 4.3.3 The Unique Solution of the Potential field equation in Fourier Space

Equation (4-24) has 2 solutions, one positive and one negative. To resolve these, (4-24) must be integrated at the next level:

i. \( \int \frac{F'}{F} = \log F \) \tag{4-25}

ii. \( \int Cdz = zC \) \tag{4-26}

The solutions are:

\[
\log F = zC \quad \text{or} \quad \log F = -zC \tag{4-27}
\]

and

\[
\log F = 2\pi z\sqrt{u^2 + v^2} \quad \text{or} \quad \log F = -2\pi z\sqrt{u^2 + v^2} \tag{4-28}
\]

Both equations of (4-28) are then passed to the exponential:
The general solution is thus:

\[ F = C_{\text{cst}} e^{2\pi \sqrt{u^2 + v^2}} + C_{\text{cst}} e^{-2\pi \sqrt{u^2 + v^2}} \]

(4-30)

while \( z \to +\infty \), \( e^{2\pi \sqrt{u^2 + v^2}} \to +\infty \). In the natural world, a potential field cannot have an infinite energy, thus \( C_{\text{cst}} e^{2\pi \sqrt{u^2 + v^2}} \) cannot exist. Finally when \( z > 0 \), the only solution of \( F \) is:

\[ F = C_{\text{cst}} e^{-2\pi \sqrt{u^2 + v^2}} \]

(4-31)

### 4.4 Toward the Derivative Poisson Kernel Equation

#### 4.4.1 Dilation Operator of the Poisson Kernel

As in equation (4-11), \( F \) is the Poisson kernel equation within the Fourier space, thus \( F \) in equation (4-31) can be replaced by its general equation (4-14), such that:

\[ F(u, v, z) = C_{\text{cst}} e^{-2\pi \sqrt{u^2 + v^2}} \]

(4-32)

where \( C_{\text{cst}} \) is a constant.
For $z > 0$, the Poisson kernel equation (4-9) is in Fourier space: [Moreau et al., 1997].

$$\hat{P}(u,v) = e^{-2\pi \sqrt{u^2 + v^2}} \quad (4-33)$$

It now becomes possible to introduce the dilation operator, $D_z$, which can be expressed for any elevation, $z$, such as [Goupillaud et al., 1984; Moreau et al., 1997, 1999; Boukerbout and Gilbert, 2006]:

- $D_z g(x) \equiv \frac{1}{a} g\left(\frac{x}{a}\right)$ in Real space
- $D_z = \frac{B}{z} \rightarrow B_z \quad (4-34)$
- $D_z e^{[\theta]} = e^{[\theta]} = e^{[z\theta]}$ in Fourier space

By inserting equations (4-32), (4-33) and (4-34) into (4-31), the Potential field equation can be expressed as dependent on the Poisson Kernel in Fourier dilated $z$, such as:

$$F(u,v,z) = F(u,v,z = 0) D_z \hat{P}(u,v) \quad (4-35)$$

In 2D horizontal $(u,v)$ and 1D vertical $(z)$, with $z > 0$ and on any point of coordinate $(u,v,z)$; the Potential field expressed through the Poisson Kernel equation in Fourier Space is equal to the Continuous Wavelet Transform equation. In equation (4-35), $F(u,v,z)$ is the wavelet transform in Fourier space by the scalar product of the Potential field, $F(u,v,0)$, measured by the Poisson wavelet, $\hat{P}(u,v)$, which is dilated by a factor, $D_z$. 
4.4.2 From 2D horizontal toward 1D horizontal

When a Potential field (electric, gravimetric, magnetic) is measured in the field, the measured profile is a 1D horizontal problem (following a vector $\vec{x}$), whereas the inversion will be a 2D problem (1D horizontal & 1D vertical), depending on the vector $(\vec{x}, \vec{z})$.

In a 2D horizontal problem, the potential field, $\phi$, is expressed by $x$ and $y$ in normal space, which becomes $u$ and $v$ in Fourier space. Equation (4-35) is expressed in 2D horizontal $(u,v)$ and 1D vertical $(z)$, which leads to 3D parameter equations.

In a 1D horizontal problem, the Potential field, $\phi$, is expressed by $x$ and $y = 0$, while in Fourier space, the Potential field, $\phi$, is only expressed by $u$ and $v = 0$, as:

\[
\sqrt{u^2 + v^2} = \sqrt{u^2} = |u|
\]  

(4-36)

Equation (4-35) thus becomes in **1D horizontal space**:

\[
F(u, z) = F(u, z = 0) D_z \hat{P}(u)
\]

(4-37)

with :

- \[
\hat{P}(u) = e^{-2 \pi |u|}
\]

(4-38)

and

- \[
F(u, z) = C_{cd} e^{-2 \pi z |u|}
\]

(4-39)
4.4.3 General equation of the horizontal derivative of Poisson Kernel family in 1D horizontal space

One of the first rules of the wavelet is that the wavelet must answer the Laplace equation (4-5). This means that the wavelet must be zero and that the sum of the wavelet coefficients of the entire space must be zero [Moreau et al., 1997; Sailhac et al., 2000].

For \((u,v) = (0,0)\), any wavelet of equation, \(M\), must be:

\[
\begin{align*}
\cdot & \quad M(u,v) = 0 \\
\cdot & \quad \int M(u,v) \, du \, dv = 0 
\end{align*}
\] (4-40, 4-41)

However, the Poisson kernel, \(\hat{P}\), in Fourier space does not answer the Laplace equation for \((u,v) = (0,0)\):

\[
\hat{P}(0,0) = e^{-2\pi \sqrt{u^2 + v^2}} = e^0 = 1 \neq 0
\] (4-42)

Consequently, the Poisson kernel, \(P\), itself is not an admissible wavelet. However, all the derivatives of the Poisson kernel, \(\hat{P}\), answer the conditions of the Laplace equation and thus for any order of derivative, \(\alpha\), such as, \(\alpha = i + j \geq 1\):

\[
\frac{\partial^\alpha \hat{P}(u,v)}{\partial u^n \partial v^m} = (-2\pi)^\alpha (u^n + v^m)e^{-2\pi \sqrt{u^2 + v^2}} = 0
\] (4-43)

with \(n\) the derivative order of \(u\) and \(m\) the derivative order of \(v\), the derivative of \(\hat{P}\) is \(d^n \hat{P}\).
In this study, the Poisson kernel family is applied only along the vector, \( \vec{u} \), within the vertical plan \((u,z)\). Thus, the Poisson kernel and its derivatives are functions of \( u \) only, not \( v \). Within the horizontal plane, defined by the vector, \( \vec{u} \), the general equation of the horizontal derivative of the Poisson kernel is:

\[
\frac{\partial^n \hat{P}(u)}{\partial u^n} = (2i\pi u)^n e^{-2\pi|\hat{u}|} \tag{4-44}
\]

where \( n \) is the order of derivative and \( u \) is the horizontal component (position along the profile).

### 4.4.4 General equation of the Vertical derivative of Poisson Kernel family

The Poisson kernel equation, within the half-plane \((z>0)\), is only dependent on \( u \) and \( v \) (4-33). However, previous studies [Moreau, 1995; Sailhac et al., 2000; Saracco et al., 2007] have shown that the vertical derivative of the Poisson Kernel, \( \frac{\partial^m \hat{P}(u,z)}{\partial z^m} \), at the order \( m \), is the Hilbert transform of the horizontal derivative of the Poisson Kernel, such that:

\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = -H \left[ \frac{\partial^m \hat{P}(u)}{\partial u^m} \right] = i[\text{sign}(u)] \frac{\partial^m \hat{P}(u)}{\partial u^m} \tag{4-45}
\]

There are two possibilities, when \( u > 0 \) and when \( u < 0 \):

- \( \forall m \in \mathbb{N}, u > 0 \)
\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = i[1](i2\pi u)^m e^{(-2\pi|u|)}
\tag{4-46}
\]

\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = (-1)(i2\pi u)^{m-1} \left[ 2\pi |u| \right] e^{(-2\pi|u|)}
\tag{4-47}
\]

In the case of \( u > 0, \ u = |u|, \) and thus the equation becomes:

\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = -(i2\pi u)^{m-1} \left[ 2\pi |u| \right] e^{(-2\pi|u|)}
\tag{4-48}
\]

- \( \forall m \in \mathbb{N}, u < 0 \)

\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = i[-1](i2\pi u)^m e^{(-2\pi|u|)}
\tag{4-49}
\]

\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = (-1)(i^2)(i2\pi u)^{m-1} \left[ 2\pi u \right] e^{(-2\pi|u|)}
\tag{4-50}
\]

\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = (-1)(-1)(i2\pi u)^{m-1} \left[ 2\pi u \right] e^{(-2\pi|u|)}
\tag{4-51}
\]
\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = (i2\pi u)^{m-1} [2\pi |u|] e^{-2\pi |z|} \quad (4-52)
\]

In the case of \( u < 0 \), \( u = -|u| \), and thus the equation becomes:

\[
\frac{\partial^m \hat{P}(z)}{\partial z^m} = -(i2\pi u)^{m-1} [2\pi |u|] e^{-2\pi |z|} \quad (4-53)
\]

For both \( u > 0 \) and \( u < 0 \), the vertical derivative of Poisson Kernel of order \( m \) is the same for equation (4-48) and (4-53). In order to simplify notation, \( \frac{\partial^m \hat{P}(z)}{\partial z^m} \) will be called \( g_z^m \) and \( \frac{\partial^n \hat{P}(x)}{\partial u^n} \) will be called \( g_u^n \). Finally, the general equation of the vertical derivative of the Poisson Kernel of order \( m \) is the equation (4-55):

Thus, the Derivative Poisson kernel equation at an \( (n+m)^{th} \) order, with \( n \) the order of the horizontal derivative (4-44) and \( m \) the order of the vertical derivative (4-55) can be expressed, in Fourier space [Moreau et al., 1997; Fedi et al.,1998; Sailhac et al., 2000; Sailhac and Marquis, 2001; Fedi et al., 2005; Cooper, 2006] as:

\[
g_u^n = (2i\pi u)^n e^{-2\pi |u|} \quad (4-54)
\]

\[
g_z^m = \frac{(2\pi |u|)^{m-1}}{(2\pi |u|)} e^{-2\pi |z|} \quad (4-55)
\]

When applying the dilation, \( a \), to the derivative (horizontal or vertical) of the Poisson kernel, the operator of dilation, \( D_a \), is inserted in the Derivative equation following (4-34), which allows us to modify equation (4-54) and (4-55), as (4-56) and (4-57), respectively:
\begin{align}
g_u^n &= (2i\pi u a)^n e^{-2\pi |u|} \\
g_z^m &= (2\pi i u a)^{(m-1)} (-2\pi |u a|) e^{-2\pi |u|}
\end{align}  

\subsection*{4.4.5 Law of the derivative order of the Poisson derivative}

As described in the literature [Moreau et al., 1997; Fedi and Quarta, 1998; Sailhac et al., 2000; Sailhac and Marquis, 2001; Fedi et al., 2005; Cooper, 2006], the type of singularity that can be accurately detected (dirac, monopole, dipole, quadripole) by a wavelet from the Poisson family depends on the structural order, \( \alpha \), of the singularity. The structural order of a dirac (a signal consisting of just one point) is \( \alpha = -1 \), of a monopole is \( \alpha = -2 \), of a dipole is \( \alpha = -3 \) and of a quadripole is \( \alpha = -4 \).

In order to characterize a singularity having a structural order, \( \alpha \), the analyzing wavelet must be of a derivative order, \( n \), which is equal or greater than \( \alpha \) [Moreau et al., 1997; Fedi and Quarta, 1998; Sailhac et al., 2000; Sailhac and Marquis, 2001; Fedi et al., 2005; Cooper, 2006]:

\[ n \geq -(1 + \alpha) \]  

with \( n \) and \( \alpha \) included in the Real space of dimension \( m \), \( \mathbb{R}^m \).
Figure 4-1: Poisson kernel wavelet family in Fourier space with their real and imaginary parts. V1 to V4 are the vertical derivative of order from 1 to 4 (see equations 4-63 to 4-66). H1 to H4 are the horizontal derivative of order from 1 to 4 (see equations 4-59 to 4-62). Each wavelet is calculated over 1024 points on a frequency from 0 to 2.5. The negative part of the frequency axis is the symmetric construction to give the wavelet its full shape.
4.5 Poisson wavelet family of derivative order from 1 to 4

This study uses wavelets having a derivative order of 2 or 3, however, in order to give a more general view of the wavelet family, the first 4 derivatives of each type (vertical and horizontal) are given in Table 4-1 based on equations (4-56) and (4-57).

<table>
<thead>
<tr>
<th>Derivative</th>
<th>Order</th>
<th>Name</th>
<th>Imaginary part</th>
<th>Real part</th>
<th>Equation</th>
<th>Figure</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal</td>
<td>1</td>
<td>H1</td>
<td>Yes</td>
<td>No</td>
<td>eq. 4-59</td>
<td>4-1</td>
</tr>
<tr>
<td>Horizontal</td>
<td>2</td>
<td>H2</td>
<td>No</td>
<td>Yes</td>
<td>eq. 4-60</td>
<td>4-1</td>
</tr>
<tr>
<td>Horizontal</td>
<td>3</td>
<td>H3</td>
<td>Yes</td>
<td>No</td>
<td>eq. 4-61</td>
<td>4-1</td>
</tr>
<tr>
<td>Horizontal</td>
<td>4</td>
<td>H4</td>
<td>No</td>
<td>Yes</td>
<td>eq. 4-62</td>
<td>4-1</td>
</tr>
<tr>
<td>Vertical</td>
<td>1</td>
<td>V1</td>
<td>No</td>
<td>Yes</td>
<td>eq. 4-63</td>
<td>4-1</td>
</tr>
<tr>
<td>Vertical</td>
<td>2</td>
<td>V2</td>
<td>Yes</td>
<td>No</td>
<td>eq. 4-64</td>
<td>4-1</td>
</tr>
<tr>
<td>Vertical</td>
<td>3</td>
<td>V3</td>
<td>No</td>
<td>Yes</td>
<td>eq. 4-65</td>
<td>4-1</td>
</tr>
<tr>
<td>Vertical</td>
<td>4</td>
<td>V4</td>
<td>Yes</td>
<td>No</td>
<td>eq. 4-66</td>
<td>4-1</td>
</tr>
</tbody>
</table>

Table 4-1: Summary of the different wavelets based on the Poisson family.

- **Horizontal Derivative of order 1**
  
  \[ g_u^1 = (2i\pi ua) e^{-2\pi |u|} \quad (4-59) \]

- **Horizontal Derivative of order 2**
  
  \[ g_u^2 = -4(\pi ua)^2 e^{-2\pi |u|} \quad (4-60) \]

- **Horizontal Derivative of order 3**
  
  \[ g_u^3 = -8(i)(\pi ua)^3 e^{-2\pi |u|} \quad (4-61) \]

- **Horizontal Derivative of order 4**

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\[ g_u^4 = 16(\pi uu)^4 e^{-2|uu|} \] (4-62)

- **Vertical Derivative of order 1**

\[ g_1^1 = (-2\pi |uu|) e^{-2|uu|} \] (4-63)

- **Vertical Derivative of order 2**

\[ g_2^2 = -4\pi^2 i (uu) |uu| e^{-2|uu|} \] (4-64)

- **Vertical Derivative of order 3**

\[ g_3^3 = 8\pi^3 (uu)^2 |uu| e^{-2|uu|} \] (4-65)

- **Vertical Derivative of order 4**

\[ g_4^4 = (2\pi i uu)^3 (-2\pi |uu|) = 16\pi^4 i (uu)^3 |uu| e^{-2|uu|} \] (4-66)

### 4.6 Example of an electrical signal

Examples of analyses on a synthetic electrical signal generated by three dipoles are shown in Figures 4-2 to 4-9 for each of the 8 wavelets (H1 to H4 and V1 to V4) described earlier. In each of the figures, the upper part (e.g., 4-2a) is the synthetic signal without noise, while Gaussian noise has been added to the lower part of the Figure (e.g., 4-2b). The signal to noise ratio is the same for each analysis “b)” and is fixed at a ratio of 0.9 (i.e., noise represents 10% of the signal).
Figure 4-2: Continuous wavelet analysis with the first horizontal derivative (H1, see equation 4-59) of three dipoles at different depths (z = -50 m, z = -200 m, z = -125 m). Analyses are made with 500 dilations on a range from 4 to 20. a) Electrical signal without noise and a sampling step of 4 m. b) Same signal as a) with added Gaussian noise and a Signal noise ratio of 0.9 (10% noise).
Figure 4-3: Continuous wavelet analysis with the second horizontal derivative (H2, see equation 4-60) of three dipoles at different depths (z = -50 m, z = -200 m, z = -125 m). Analyses are made with 500 dilations on a range from 4 to 20. a) Electrical signal without noise and a sampling step of 4 m. b) Same signal as a) with added Gaussian noise and a Signal noise ratio of 0.9 (10% noise).
Figure 4-4: Continuous wavelet analysis with the third horizontal derivative (H3, see equation 4-61) of three dipoles at different depths (z = -50 m, z = -200 m, z = -125 m). Analyses are made with 500 dilations on a range from 4 to 20. a) Electrical signal without noise and a sampling step of 4 m. b) Same signal as a) with added Gaussian noise and a Signal noise ratio of 0.9 (10% noise).
Figure 4-5: Continuous wavelet analysis with the forth horizontal derivative (H4, see equation 4-62) of three dipoles at different depths (z = -50 m, z = -200 m, z = -125 m). Analyses are made with 500 dilations on a range from 4 to 20. a) Electrical signal without noise and a sampling step of 4 m. b) Same signal as a) with added Gaussian noise and a Signal noise ratio of 0.9 (10% noise).
Figure 4-6: Continuous wavelet analysis with the first vertical derivative (V1, see equation 4-63) of three dipoles at different depths (z = -50 m, z = -200 m, z = -125 m). Analyses are made with 500 dilations on a range from 4 to 20. a) Electrical signal without noise and a sampling step of 4 m. b) Same signal as a) with added Gaussian noise and a Signal noise ratio of 0.9 (10% noise).
Figure 4-7: Continuous wavelet analysis with the second vertical derivative (V2, see equation 4-64) of three dipoles at different depths \((z = -50 \text{ m}, z = -200 \text{ m}, z = -125 \text{ m})\). Analyses are made with 500 dilations on a range from 4 to 20. a) Electrical signal without noise and a sampling step of 4 m. b) Same signal as a) with added Gaussian noise and a Signal noise ratio of 0.9 (10% noise).
Figure 4-8: Continuous wavelet analysis with the third vertical derivative (V3, see equation 4-65) of three dipoles at different depths (z = -50 m, z = -200 m, z= -125 m). Analyses are made with 500 dilations on a range from 4 to 20. a) Electrical signal without noise and a sampling step of 4 m. b) Same signal as a) with added Gaussian noise and a Signal noise ratio of 0.9 (10% noise).
Figure 4-9: Continuous wavelet analysis with the fourth vertical derivative (V4, see equation 4-66) of three dipoles at different depths (z = -50 m, z = -200 m, z = -125 m). Analyses are made with 500 dilations on a range from 4 to 20. a) Electrical signal without noise and a sampling step of 4 m. b) Same signal as a) with added Gaussian noise and a Signal noise ratio of 0.9 (10% of noise).
The synthetic electrical signal was calculated with a sampling step of 4 m for three dipoles as follows: the first dipole is a horizontal dipole 20 m in length at a depth of 50 m; the second dipole is a 20 m long vertical dipole at a depth of 200 m; the third dipole 20 m in length is oblique (oriented at 225° or 45° clockwise) at a depth of 125 m.

The full results are summarised in Tables 4-2, 4-3 and 4-4 in order to compare the accuracy of each wavelet (H1 to H4 and V1 to V4) for the electrical signal with (SNR of 0.9) and without (SNR of 1) noise. For an electrical signal without noise, a dipole can be localized (on X and Z) by applying a continuous wavelet to its electrical signature using any of the 8 wavelets. These wavelets have a vertical accuracy better than 5 m on each of the dipoles, except V1 and H1 which have an error between 9 to 20 m at depths greater than 50 m. Their horizontal accuracy is better than 10 m.

For an electrical signal with 10% noise (SNR of 0.9), the vertical accuracy of any of the 8 wavelets decreases with increasing depth of the dipole. At 50 m depth, the accuracy will be better than 20 m, except for V1 (Table 4-3). At 125 m depth, the accuracy will be better than 10 m, however, for H1 and H3, the vertical accuracy is better than 15 m (Table 4-5). For a dipole at a depth of 200 m, the accuracy of any of the wavelets becomes quite poor, between 13 and 56 m (Table 4-4), however, V4 and H4 seem to be less sensitive to the noise, with an accuracy between 10 to 13 m. Their horizontal accuracy follows a similar pattern; when the dipole is at less than 130 m depth, the error is less than 20 m. At 200 m depth, the horizontal error is between 20 and 60 m. Only H4 and V4 are accurate to better than 10 m. An explanation of this increase of resolution for an increasing depth source from 50 m to 150 m deep is the frequency of the electrical anomaly which is analysed. Wavelet analysis is a signal analysis technique where each electrical anomaly is defined by its own wavelength (number of data points characterizing the anomaly). In the same idea, noise has its own frequency (e.g., Gaussian noise in this example, which is by definition the inverse of the sampling step, Fig 4-9b). Thus, when an electrical anomaly is generated by a very shallow source (e.g., dipole D-1 at 50 m deep, Fig 4-9), its
wavelength will be defined by a shorter distance (which corresponds to a higher frequency) than the wavelength for a deeper source (e.g., D-3 at 125 m deep, Fig. 4-9). However, both anomalies while not too different in depth, will have similar anomaly amplitudes. The result is that the shallowest source (D-1) will be more affected by noise than the deeper source. However, the other extreme is also true. A very deep source, even if it’s associated anomaly is defined by a large wavelength (e.g., D-2 at 200 m deep, Fig. 4-9), will have a lower amplitude than a shallower source (e.g., D-2), which will make it more susceptible to background noise. The only way to increase the resolution of the wavelet analyses, and thus the depth calculation, will be to decrease the size of the sampling step. That will limit the noise effect on the signal (in the case where the noise is not Gaussian), but also allow for a better constraint on the measured anomaly by increasing the number of points which define it without changing its property (e.g., wavelength, amplitude and frequency).

Based on the example of these 3 dipoles, the average vertical error is 25 m ± 20 m (Table 4-5). The average horizontal error is 15 m ± 15 m. Horizontal and vertical wavelets of the same derivative order show similar accuracy on depth and horizontal determination of a dipole (Table 4-6).

In comparison to the other wavelets from the Poisson family, the first vertical (V1) and horizontal (H1) wavelets are less accurate than the other wavelets for determination of the depth of a dipole (Table 4-3, 4-4 and 4-4). Mathematically, this difference is easily described by the law of the derivative order (section 4.4.5), which requires that the level of the derivative order of the wavelet is equal or greater than the structural order of the singularity.
<table>
<thead>
<tr>
<th>Dipole Orientation</th>
<th>HORIZONTAL</th>
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<tr>
<td>Position in m</td>
<td>-1600</td>
</tr>
<tr>
<td>Depth in m</td>
<td>50</td>
</tr>
<tr>
<td>NOISE</td>
<td>None</td>
</tr>
<tr>
<td>Signal Noise Ratio</td>
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<table>
<thead>
<tr>
<th>Wavelet type</th>
<th>Calculated value in m</th>
<th>Error in m</th>
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<tbody>
<tr>
<td>H1</td>
<td>-1602 2 -53 4 2 3</td>
<td></td>
</tr>
<tr>
<td>H2</td>
<td>-1601 1 -46 2 1 4</td>
<td></td>
</tr>
<tr>
<td>H3</td>
<td>-1602 2 -52 2 2 2</td>
<td></td>
</tr>
<tr>
<td>H4</td>
<td>-1600 0 -52 1 0 2</td>
<td></td>
</tr>
<tr>
<td>V1</td>
<td>-1595 8 -46 8 5 4</td>
<td></td>
</tr>
<tr>
<td>V2</td>
<td>-1600 1 -52 4 0 2</td>
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<td>-1601 2 -51 1 1 1</td>
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<tr>
<td>V4</td>
<td>-1599 1 -53 1 1 3</td>
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<tr>
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<tbody>
<tr>
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</tr>
</tbody>
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<table>
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<tr>
<th>Wavelet type</th>
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<th>Error in m</th>
</tr>
</thead>
<tbody>
<tr>
<td>H1</td>
<td>-1601 4 -53 6 1 3</td>
<td></td>
</tr>
<tr>
<td>H2</td>
<td>-1598 14 -61 26 2 11</td>
<td></td>
</tr>
<tr>
<td>H3</td>
<td>-1602 3 -35 6 2 15</td>
<td></td>
</tr>
<tr>
<td>H4</td>
<td>-1599 9 -52 21 1 2</td>
<td></td>
</tr>
<tr>
<td>V1</td>
<td>-1585 3 -77 22 15 27</td>
<td></td>
</tr>
<tr>
<td>V2</td>
<td>-1602 1 -47 6 2 3</td>
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<td>V3</td>
<td>-1605 18 -56 33 5 6</td>
<td></td>
</tr>
<tr>
<td>V4</td>
<td>-1596 2 -39 9 13 11</td>
<td></td>
</tr>
</tbody>
</table>

**Table 4-2:** Localization results (Z and X) of a synthetic horizontal dipole at 50 m depth and at -1600 m along the profile. Analyses were made by continuous wavelet analysis as described in the text using 8 wavelets (H1 to H4 and V1 to V4). Upper section is the electrical signal without noise (SNR 1) and the lower sector is the electrical signal with 10% noise (SNR 0.9).
<table>
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<tr>
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<tbody>
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<td>None</td>
</tr>
<tr>
<td>Wavelet</td>
<td>Calculated value in m</td>
</tr>
<tr>
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<td>X Z</td>
</tr>
<tr>
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</tr>
<tr>
<td>H2</td>
<td>6 5 -202 9 6 2</td>
</tr>
<tr>
<td>H3</td>
<td>-1 3 -203 17 1 3</td>
</tr>
<tr>
<td>H4</td>
<td>9 1 -202 2 9 2</td>
</tr>
<tr>
<td>V1</td>
<td>4 4 -204 11 4 4</td>
</tr>
<tr>
<td>V2</td>
<td>1 4 -199 14 1 1</td>
</tr>
<tr>
<td>V3</td>
<td>9 1 -201 3 9 1</td>
</tr>
<tr>
<td>V4</td>
<td>9 1 -203 4 9 3</td>
</tr>
<tr>
<td>NOISE</td>
<td>YES</td>
</tr>
<tr>
<td>type X σx Z σz</td>
<td>X Z</td>
</tr>
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<tr>
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<tr>
<td>H3</td>
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<tr>
<td>H4</td>
<td>57 7 -210 20 57 10</td>
</tr>
<tr>
<td>V1</td>
<td>1 5 -237 7 1 3</td>
</tr>
<tr>
<td>V2</td>
<td>20 8 -224 18 20 24</td>
</tr>
<tr>
<td>V3</td>
<td>54 7 -213 20 54 13</td>
</tr>
<tr>
<td>V4</td>
<td>30 1 -195 13 30 5</td>
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</tbody>
</table>

**Table 4-3:** Localization results (Z and X) of a synthetic vertical dipole at 200 m depth and at 0 m along the profile. Analyses were made by continuous wavelet analysis as described in the text using 8 wavelets (H1 to H4 and V1 to V4). Upper section is the electrical signal without noise (SNR 1) and the lower section is the electrical signal with 10% noise (SNR 0.9).
<table>
<thead>
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</thead>
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</tr>
<tr>
<td>Depth in m</td>
<td>125</td>
</tr>
<tr>
<td>NOISE</td>
<td>None</td>
</tr>
<tr>
<td>Signal Noise Ratio</td>
<td>1</td>
</tr>
<tr>
<td>Wavelet</td>
<td></td>
</tr>
<tr>
<td>type X</td>
<td></td>
</tr>
<tr>
<td>σX</td>
<td></td>
</tr>
<tr>
<td>Z</td>
<td></td>
</tr>
<tr>
<td>σZ</td>
<td></td>
</tr>
<tr>
<td>Χ</td>
<td></td>
</tr>
<tr>
<td>Ζ</td>
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</table>

<table>
<thead>
<tr>
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<th>σX</th>
<th>Z</th>
<th>σZ</th>
<th>Χ</th>
<th>Ζ</th>
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</thead>
<tbody>
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<td>0</td>
<td>1</td>
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<tr>
<td>H2</td>
<td>1599</td>
<td>4</td>
<td>-123</td>
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<td>1</td>
<td>2</td>
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<td>H3</td>
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<td>2</td>
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<td>H4</td>
<td>1598</td>
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<td>V2</td>
<td>1600</td>
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<td>-124</td>
<td>7</td>
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<td>V3</td>
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<td>3</td>
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<td>V4</td>
<td>1601</td>
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<td>-129</td>
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<td>1</td>
<td>4</td>
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<table>
<thead>
<tr>
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</tr>
</thead>
<tbody>
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<td>X</td>
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<td>Z</td>
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<table>
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<th>Z</th>
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<th>Z</th>
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<td>17</td>
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<td>H2</td>
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<td>-123</td>
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<td>47</td>
</tr>
<tr>
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<td>8</td>
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<td>V1</td>
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<td>-123</td>
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<td>2</td>
</tr>
<tr>
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<td>2</td>
</tr>
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<td>-125</td>
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<td>18</td>
<td>0</td>
</tr>
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<td>5</td>
<td>-120</td>
<td>35</td>
<td>4</td>
<td>5</td>
</tr>
</tbody>
</table>

**Table 4-4:** Localization results (Z and X) of a synthetic oblique dipole (45°) at 125 m depth and at 1600 m along the profile. Analyses were made by continuous wavelet analysis as described in the text using 8 wavelets (H1 to H4 and V1 to V4). Upper section is the electrical signal without noise (SNR 1) and the lower section is the electrical signal with 10% noise (SNR 0.9).
In this example, the electrical signal is generated by three electrical sources, which have the structural order of a dipole \( n = -2 \). To respect the mathematical relationship between derivative order, \( n \), and structural order, \( \alpha \), \( n \) must answer the equation 4-59. The derivative order of V1 and H1 \( (n =1) \) is smaller than the requirement to characterize a dipole \( (\alpha = -3) \).

\[ -(-3+1) = 2 \leq n \]  
\[ (4-67) \]

Thus the derivative order, \( n \), of the wavelet must be equal to or greater than 2.

### 4.7 Limitations of real wavelet analysis

The analyzing wavelet used must answer several requirements which in fact control the limitations of the method. The wavelet must at all times respect:

- **The Shannon/Nyquist theorem**, which requires that the sampling frequency, \( f_s \), of the wavelet is at least twice the minimum sampling frequency, \( f_m \), which can completely describe the wavelet.

\[ F_s \geq 2 f_m \]  
\[ (4-68) \]

Each wavelet has its own minimum sampling frequency, \( f_m \). The minimum sampling frequency of one wavelet also depends on the dilation applied to it. In order to make this limitation easier to follow, a baseline has been chosen such that it allows us to respect the Shannon/Nyquist theorem at any dilation. The values in Table 4-8 give a safe minimum value for the wavelet at the most common dilations.
Figure 4-10: Continuous wavelet analysis with the forth horizontal derivative (H4, see equation 4-62) of one dipole at a depth, z = -50 m. Analyses are made with 500 dilations on a range from 5 to 20. Top: Analyses by continuous wavelet, extrema and depth calculation. Middle: Electrical signal without noise (SNR = 1) and a sampling step of 4 m. Bottom: Topography variation along the profile is a positive slope of 20°. Red diamond represents the dipole generating the electrical signal.
• **The range of dilation** must be chosen to characterize, as much as possible, the frequencies which form the analysed signal. When applied, a dilation acts as a type of pass-band filter, thus small dilation (i.e., dilation of 2) acts as a high frequency pass-filter. In contrast, a large dilation (i.e., higher than 10) acts as a low frequency pass-filter. The choice of the dilation range should be first based on Table 4-8, in order to insure that the analyzed signal contains enough points to be properly analyzed. The choice of the range of dilation should then be chosen to highlight the frequencies present within the analysed signal.

• **The number of dilation** controls the vertical resolution, thus it is strongly recommended to always use from a few to several hundred dilations. For example, the analyses of the synthetic signal shown here (Fig. 4-2 to 4-9) were made for a range of dilation from 4 to 20 and include 500 dilations. This allowed us to characterize the most significant information coming from the analysed signal.

• **The length of the signal**, as distance or time scale, is not significant during the analyses, however, the number of data points that defines the signal is of utmost importance. As described above, with the Shannon theorem, the signal must be long enough to be analysed by a wavelet at different dilations. Also, the sampling of the signal is very important for the result of the continuous wavelet transform. An anomaly, which is defined by too few data points, will not show up well in the analyses because of this poor sampling resolution. This study strongly suggests that any analyzed profile be resampled at a lower step to increase its resolution. However, the resampling should not be overdone and the Nyquist frequency should always be respected. Although decreasing the sampling step will increase the resolution of the anomalies, it will also increase the influence of the noise on the results of the continuous wavelet transform.
Figure 4-11: Continuous wavelet analysis with the forth horizontal derivative (H4, see equation 4-62) of one dipole at a depth, z = -50 m. Analyses are made with 500 dilations on a range from 5 to 20. Top: Analyses by continuous wavelet, extrema and depth calculation. Middle: Electrical signal without noise (SNR = 1) and a sampling step of 4 m. Bottom: Topography variation along the profile is a symmetrical curve slope in hill shape with a high/ base ratio of 1/16. Red diamond represents the dipole generating the electrical signal.
This study uses only resampling methods that are equal or less than 5 times the true sampling of the data set. If necessary, a smoothing of the data may be done to reduce the noise present in the analysed signal. The synthetic signals shown here have a sampling step of 4 m and contain 1251 data points (Fig. 4-2 to 4-9).

- **Real wavelet analyses** cannot give information regarding the phase, nor the structural order of the source generating the analyzed signal. Information on the phase helps to quantify the different frequencies making the analysed signal. In addition, the phase will also show the converging extremas, such as those obtained from the correlation coefficient of the real wavelet. Theses extrema converge toward the source of the signal and can give the depth of the source, as well as the structural order of the source. Phase information can only be acquired through use of the complex wavelet, which can be done by combining the horizontal and vertical wavelets, as described in the literature [Moreau et al., 1995; Fedi and Quarta, 1998; Sailhac et al., 2000; Sailhac and Marquis, 2001; Saracco et al, 2004, 2007, 2009; Fedi et al., 2005; Boukerbout and Gibert, 2006; Cooper, 2006].

- Although this study is not based on complex wavelets, I strongly encourage future studies to use them when analysing magnetic data, where knowledge of the structural order can be of utmost importance, as well as the depth. The depth investigation of electrical sources, as in this study, does not require knowledge of the structural order, because electricity generation is always the result of a combination of micro-sources of different order, which act together as a macro-source.
<table>
<thead>
<tr>
<th>Z in m</th>
<th>Synthetic Signal</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>SNR</td>
<td>1</td>
<td>0.9</td>
</tr>
<tr>
<td>ERROR</td>
<td>without noise</td>
<td>with noise</td>
</tr>
<tr>
<td>average</td>
<td>6</td>
<td>23</td>
</tr>
<tr>
<td>σ</td>
<td>5</td>
<td>21</td>
</tr>
<tr>
<td>min</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>max</td>
<td>25</td>
<td>97</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>X in m</th>
<th>Synthetic Signal</th>
<th></th>
</tr>
</thead>
<tbody>
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<td>with noise</td>
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<tr>
<td>σ</td>
<td>4</td>
<td>15</td>
</tr>
<tr>
<td>min</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>max</td>
<td>15</td>
<td>64</td>
</tr>
</tbody>
</table>

**Table 4-5:** Summary of the error on depth and horizontal localisation from results obtained with the 8 wavelets. Signal noise ratio, SNR, of 0.9 represent 10% noise, when SNR of 1 is without noise.

- **Topography effect** on depth calculation may be in some cases, a significant limitation to the continuous wavelet analyses. A topographic variation along the profile will directly affect the electrical measurement and thus will induce a distortion of the signal characterising the anomaly. Consequently, strong topography variations may significantly increase the error on the depth calculation. The synthetic signals presented in Figures 4-2 to 4-9 are made with a flat topography (0 m a.s.l) and the associated error without noise is commonly below 10 m. For topography changing along the profile and a signal without noise, the error on the depth can reach 15 to 20 m (Fig. 4-10 and 4-11), however, the continuous wavelet analysis is still able to localize the source generating the signal. It is not possible to accurately correct the signal of topographic effect without previously knowing its depth, thus the best way to reduce its effect is to try to make the field measurement as much as possible along the same topographic contour.
<table>
<thead>
<tr>
<th></th>
<th>Horizontal wavelet</th>
<th>Vertical wavelet</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dipole</td>
<td>Error on Z</td>
<td>σz</td>
</tr>
<tr>
<td>ALL depth</td>
<td>15</td>
<td>20</td>
</tr>
<tr>
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<td>12</td>
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<td>200</td>
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<td>27</td>
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<td>125</td>
<td>15</td>
<td>18</td>
</tr>
<tr>
<td>Without noise</td>
<td>Error on Z</td>
<td>σz</td>
</tr>
<tr>
<td>50</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>200</td>
<td>8</td>
<td>7</td>
</tr>
<tr>
<td>125</td>
<td>5</td>
<td>4</td>
</tr>
<tr>
<td>With Noise</td>
<td>Error on Z</td>
<td>σz</td>
</tr>
<tr>
<td>50</td>
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<td>31</td>
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<tr>
<td>125</td>
<td>23</td>
<td>20</td>
</tr>
</tbody>
</table>

**Table 4-6:** Comparison of error on calculated depth between horizontal wavelets (H1 to H4) and vertical wavelet (V1 to V4).

<table>
<thead>
<tr>
<th></th>
<th>Horizontal wavelet</th>
<th>Vertical wavelet</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dipole</td>
<td>Error on X</td>
<td>σx</td>
</tr>
<tr>
<td>ALL depth</td>
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<tr>
<td>50</td>
<td>5</td>
<td>6</td>
</tr>
<tr>
<td>200</td>
<td>18</td>
<td>20</td>
</tr>
<tr>
<td>125</td>
<td>7</td>
<td>7</td>
</tr>
<tr>
<td>Without noise</td>
<td>Error on X</td>
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<td>200</td>
<td>6</td>
<td>3</td>
</tr>
<tr>
<td>125</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td>With Noise</td>
<td>Error on X</td>
<td>σx</td>
</tr>
<tr>
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<td>6</td>
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<tr>
<td>200</td>
<td>27</td>
<td>24</td>
</tr>
<tr>
<td>125</td>
<td>10</td>
<td>7</td>
</tr>
</tbody>
</table>

**Table 4-7:** Comparison of error on calculated horizontal localization between horizontal wavelets (H1 to H4) and vertical wavelet (V1 to V4).
Figure 4-12: Statistical approach through histogram of the error a) on depth and b) horizontal localisation of the dipole discussed in text and in Table 4. The error (in blue) for an electrical signal without noise (SNR of 1) and a total of 117 analyses. In yellow is the error for an electrical signal with 10 % noise (SNR of 0.9) and a total of 144 analyses.
4.8 Multi-scale tomography

As seen previously, each wavelet based on the Poisson kernel family can be used to determine the depth and horizontal position of the source that generates the measured potential field. However, depending on the structure and the intensity of the source, noise intensity or topographic variation, each of these wavelets will be more or less effective in accurately determining the location of the source. Generally, when the analysis is performed, no apriori information is known about the source, (e.g., its depth, is it a dipole or monopole, how much real noise is present, is it a point source or laterally extensive, etc.). In a very simple case, with one source generating a very strong signal (as in the synthetic example, Fig. 4-2 to 4-9), it is not necessary to know what kind of wavelet that could be more effective to localize it. Unfortunately, in the real world, this simple case generally does not occur and analysis by one wavelet can give erroneous results or generate artefacts due to poor sampling, excessive noise or large topographic change. In order to be as accurate as possible and to avoid depth artefacts, this study used the multi-scale wavelet tomography (MWT) approach, which consists of making repeated analyses of the signal with a number of different wavelets and changing the analyzed wavelet from one analysis to another. By not choosing one wavelet over another, and by statistically determining the depth of the source, it becomes possible to reduce the risk of generating artefacts and thus increase the accuracy of the results. Artefact depth is due to the response of the wavelet to noise having the same structure as it or by topographic effects that have excessively distorted the signal. Each wavelet has its own shape (Fig. 4-1), thus each wavelet should be sensitive to a specific noise structure (or topographic distortion) and so generate an associated artefact. However, a different wavelet which encounters the same noise, will probably not respond in the same way and so will not generate the same artefact. In contrast, a real anomaly will be picked up by all wavelets (from the Poisson family) or at least by several of them. Thus, if similar results are obtained
from several different wavelets, it is reasonable to assume that the depth and position characterize a real source.

<table>
<thead>
<tr>
<th>Dilation</th>
<th>Number of Points</th>
</tr>
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<tbody>
<tr>
<td>1</td>
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</tr>
<tr>
<td>5</td>
<td>286</td>
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<tr>
<td>10</td>
<td>576</td>
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<tr>
<td>15</td>
<td>858</td>
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<tr>
<td>20</td>
<td>1092</td>
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<tr>
<td>25</td>
<td>1502</td>
</tr>
<tr>
<td>30</td>
<td>1716</td>
</tr>
</tbody>
</table>

Table 4-8: Minimum number of points that must contain the analyzed signal in order to respect the Shannon/Nyquist Theorem.

The more wavelets used, the more accurate should be the depth and horizontal position of the source. However, this will clearly require more processing time, thus, in order to balance accuracy and efficiency, this study always uses 4 wavelets. Any depth which is found by at least 3 of the 4 wavelets is considered significant. A complete control on the accuracy of the depth cannot be obtained, because the real depth of the source is unknown, however, a qualitative approach can be done by looking at the standard deviation of all depths obtained by all wavelets for a similar horizontal position. Based on the assumption that the final depth, which is the mean of all calculated depths, is statically close to the real depth, the standard deviation can be used as a quality control tool. The standard deviation of all depths in this case, expresses the scattering of the depth calculations. Thus, wide spread clusters have a large standard deviation, characterize poor accuracy and are associated with a poor confidence on the depth value of the source. In contrast, a well focused cluster of depth values, with a very small standard deviation can be considered to be of high confidence for the source depth. Another way to better constrain the uncertainty of the depth by the wavelet analyses is to combine it with other independent techniques (e.g., gravity, seismic), which could help to obtain a
better view of the underground structures and thus help to isolate the source of the measured anomaly and its depth.

Furthermore, the scattering may also be used as a qualitative indicator of how punctual is the source dimension. A source (of the analyzed signal) very well constrained in space and having a small volume dimension in comparison to its depth, should be easiest to accurately localize. In contrast, a source which is widely extended and poorly constrained in space, will be harder to localize in space (depth and horizontal plane).

For example, a synthetic electrical signal consisting of 3 dipoles at different depths and orientation was generated (Fig. 4-2 to 4-9; Table 4-1 to 4-3). First the signal without noise was analyzed by 8 wavelets (V1 to V4 and H1 to H4). Subsequently, a Gaussian noise was added to the synthetic electrical signal, such that the noise amplitude represents 10% of the signal amplitude. Finally, the final depths and associated standard deviation were calculated for each of the dipoles (Table 4-5). There is no significant difference in either the depth or horizontal localization (Table 4-6 and 4-7) when using wavelets from the horizontal wavelet group (H1 to H4) or the vertical wavelet group (V1 to V4). When using the results from the wavelet analyses, the vertical and horizontal accuracies are quiet similar, ~ 10 m or less, when there is no noise and ~ 20 m or less, with noise (Table 4-6 and Fig 4-12).

The Multi-scale Wavelet Tomography (MWT) approach is applied on different volcanoes in this study. Chapter 5 will present a comparison between the MWT depth results and other traditional geophysical methods on Stromboli, Waita and Masaya volcanoes. In Chapter 6, MWT is applied to Self-potential mapping, over time, in order to investigate the depth variation of the hydrothermal system fluids present within Masaya volcano. In Chapter 7, the MWT is used on Self-potential signals to characterize water table depth variations over time at Kawah Ijen volcano. In Appendix A-1, the example of the hydrothermal dynamism links with volcanic activity is presented for Piton de la
Fournaise volcano, with the MWT applied on Self-potential to track hydrothermal water displacement.

4.9 Multi-scale Wavelet Tomography MATLAB code (MWTmat)

The multi-scale wavelet tomography code used in this study was written in MATLAB and should work with any MATLAB version of 2007 or later. In order to work properly, the MATLAB version must have the databases and mapping toolboxes. The wavelet toolbox is not used in this code, as the wavelet equations (see above) are directly incorporated in the code. The wavelet transform is also performed within the code.

The MWTmat code is organized in three distinct sections, which allow us to run each of the main parts independently of the others. Each code generates specific file formats, which are recognized by the other components.

✔ The first part of MWTmat code is the Matrix generation of the wavelet transform of the analyzed signal (section 4.9.1).

✔ The second part of MWTmat code is the display and extraction of the lines of extrema (section 4.9.2).

✔ The last part of the MWTmat code is the selection, extension of the lines of extrema and depth calculation (section 4.9.3).

Each part of the MWTmat code is relatively straightforward to use with descriptive text displayed throughout the data processing, allowing the user to know what he/she has to do.
4.9.1 Generation of the continuous wavelet transform

- The input data must be in an Excel spreadsheet (compatible Microsoft office 2003) consisting of a single sheet. The data must be organized in three columns:
  - 1st column is the distance in m;
  - 2nd column is the topographic elevation in m;
  - 3rd column will be the data in units of the users choice. The choice of the unit should be selected in order to best fit the data amplitude.

<table>
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<tr>
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<th>4</th>
<th>33.341</th>
<th>0</th>
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</tr>
</thead>
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<tr>
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<td>0</td>
<td>66</td>
</tr>
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<td>0</td>
<td>0</td>
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</tr>
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<td>0</td>
<td>0</td>
</tr>
<tr>
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<td>0</td>
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</tr>
<tr>
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<td>37.325</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
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<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
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<td>0</td>
<td>0</td>
</tr>
<tr>
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<td>0</td>
<td>0</td>
</tr>
<tr>
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<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>4.3527</td>
<td>40.762</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>4.3848</td>
<td>41.461</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 4-9: Example of automatic trend file with the first 13 rows of the file B2. The first column is the distance coordinate, the second column is the dilation coordinate, and the third column is the correlation coefficient. On the sixth column, the first row is the mean value of the correlation coefficient, while the second row is the calculated strength of the extrema (as described in the text).
The number of points that define the analyzing wavelet is always defined to be equal to the length of the input signal (analyzed file). The dilation operator applied on the wavelet will act as a band filter by modifying the frequency of the wavelet. It is therefore possible that the wavelet may not be defined by sufficient points to characterize its frequency ("shape", see previous section on limitations). If the wavelet is not correctly sampled, the result of the analysis will be wrong. In order to avoid errors, the analyzing wavelet must always be well defined at each of the selected dilations (Shannon/Nyquist theorem). A safety limit function of the dilation has been calculated for all of the wavelets described in this study (Table 4-8).

<table>
<thead>
<tr>
<th>-60</th>
<th>433.341</th>
<th>-139.74</th>
<th>19.557</th>
<th>202.21</th>
</tr>
</thead>
<tbody>
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</tr>
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<td>-60</td>
<td>35.313</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
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<td>35.98</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
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<td>36.65</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>37.325</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>38.005</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>38.688</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>39.375</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>40.067</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>40.762</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>-60</td>
<td>41.461</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 4-10: Example of automatic trend file with the first 13 rows of the file B2_manual. The first column is the distance coordinate, the second column is the dilation coordinate, and the third column is the correlation coefficient. The first three rows of the fourth and fifth columns are the distance and dilation coordinates of the manual trend. On the sixth column, the first row is the mean value of the correlation coefficient, while the second row is the calculated strength of the extrema (as described in the text).
As the number of points of the wavelet is always the same as the input file, I strongly recommended that the number of points of the input signal always be bigger than the number in the Table 4-8 for the largest dilation applied to the wavelet. In order to facilitate the application of this rule, the code will tell the user what the maximum dilation which can be applied on the wavelet to analyse the profile (input file).

For example, an input signal consisting of 600 points may be analyzed by any of the 8 wavelets (described previously) on any range of dilation from 1 to 10 (Table 4-8). However, for dilations above 10, none of the wavelets will be properly defined and thus dilation above 10 cannot be used to analyze the input signal.

One way to apply greater dilation operators on the wavelet, is to resample the input signal with a smaller sampling step in order to increase the number of points that define it and thus the number of points that define the wavelet. Table 4-8 was calculated for the wavelets V1 and H1, and similar calculations have been done for the other wavelets and the results show similar value, which allow us to use Table 4-8 as a rule for any of the wavelet. The user can calculate the minimum number of points required for any dilation by using the data in the table or by using the following equation:

\[
N_{\text{brofpoints}} = 56,884 \times Dilation + 0.2299
\]  

(4-69)

- **Choice of the wavelet** can be made among the 8 wavelets described in this study (Table 4-1, section 4.5), which include the first four derivative order of both horizontal and vertical derivatives of the Poisson family (V1, V2, V3, V4, H1, H2, H3, H4). There is no rule to determine which wavelet will work best for a specific input signal, however, in theory, the derivative order of the wavelet must always be bigger than the structural order of the source generating the signal (equation 4-67, section 4.6).
Figure 4-13: Continuous wavelet analysis with the forth vertical derivative (H4, see equation 4-66) of three dipoles at different depths (z = -50 m, z = -200 m, z= -125 m) with added Gaussian noise and a Signal noise ratio of 0.9 (10% noise). Analyses are made with 500 dilations on a range from 4 to 20. a) Extrema lines extraction without extrema length filter. b) Extrema lines extraction with a 50% extrema length filter. Only lines longer than 50% of the length of the image are extracted.
As electrical or magnetic signals are generally dipole or monopole, the derivative order of the wavelet should be equal or greater than 1 for a monopole or a 2 for a dipole. However, the structural order of the source is generally unknown, thus it is best to assume that the structural order of the source is of the greater type. For example, with a dipole, the input signal should be analyzed with a wavelet having a derivative order equal or greater than 2, such as the wavelets V2, V3, V4, H2, H3 or H4.

**Dilation range** refers to the minimum and maximum dilation value. Any kind of dilation value can be used from 1 to any number above which respects the Shannon/Nyquist theorem (Table 4-8). Numbers below 1 are not suitable for this version of MWTmat code because the edge of the wavelet will not be properly defined by the support matrix within the code. Dilations act as a “zoom in” (small dilation) and “zoom out” (large dilation), which correspond to high band pass and low band pass filters, respectively. When performing the analyses for the first time on a new input signal, I strongly recommend taking a large range of dilations in order to obtain a full scan of the frequencies present in the input signal. The maximum recommended dilation will be displayed on the screen to help the user to make the best choice.

**Significance of the dilation range on the result and interpretation**

It should be recalled that MWT is a type of signal processing and that dilations are like filters, thus all the high frequencies will be present on the bottom of the figure (smallest dilation) and the lowest frequencies detected by the analysis will be on the top of the figure (large dilation). When an anomaly is detected by the wavelet over the range of dilations, it will be characterized by lines of extrema converging toward each other (at least two of them).

If the anomaly consists of a single frequency of low value, the anomaly will be detected across the range of dilations until its own frequency is reached (dilation) and will be represented by extrema forming a perfect cone shape structure (e.g., Fig. 4-2). The extrema will be calculated in the part 2 of the code.
However, it is important to understand them before selecting the appropriate range of dilation.

If two anomalies of high frequency are close enough to each other (e.g., their edges merge), the output figure of the MWT analyzes will perfectly show each of them at small dilations (until the one corresponding to the frequency). For large dilations, it is quite possible (depending on the wavelet and how the anomalies merge) that the results show merging or overlap of one of the extrema with an extrema from the next nearest anomaly (e.g., extrema within blue circle 1, Fig. 4-13b), or one of the extrema may completely overlap the extrema of the second anomaly. Commonly, this type of merging is easily visible on the output figure, because the slope of extrema line (for one or both anomalies) will be broken at the dilation (frequency) where they merge. Thus, to avoid merging of extrema which will distort the extrema and thus increase the error on the depth, it is best to chose a maximum dilation that is smaller than the dilation where the merging starts. Another way to avoid merging influence in the depth calculation will be described in the part 3 of the code.

- **Different approach to zoom in or out:**

  If the user wants to zoom in further on the smallest dilation, one can resample the input signal with a smaller sampling step. In contrast, if the user considers that the input signal has too many points characterizing the anomaly described by the input signal, rather than using large dilation values to characterize it with the wavelet, the user could resample the input signal with a larger sampling step, which will allow use of a smaller dilation on the wavelet to characterize the anomaly. However, I do not recommend doing so because valuable small scale information present in the signal could be lost.

- **The Number of dilations** which are applied on the wavelet to scan the input signal over the range of dilation, controls the resolution and helps to increase the accuracy of the depth calculation. The number of dilation should always be of at least a few hundred dilations in order for good resolution.
Typically, analyzes done in this study are made with 500 dilations. There is no maximum limit, except that higher the number of dilations, the longer the processing time required and the larger the output file.

- Output files will save the result of the MWT analyses and the important information about the parameters of the analysis. The first part of the MWTmat code generates 3 files:
  - The 1st file saves the matrix including all the correlation coefficients within a .dat file. Data are recorded in ASCII, with double precision (16 digits) and TAB delimited.
  - The 2nd file saves the parameters of the analyses within a text file.
  - The 3rd file saves the parameters in .dat file, which will be needed for the other part of the MWTmat code

Example of Info file:

```
=============================================  
Wavelet type, enter the corresponding number:  
VERTICAL Wavelet V=1 : 1                           
VERTICAL Wavelet V=2 : 2                           
VERTICAL Wavelet V=3 : 3                           
VERTICAL Wavelet V=4 : 4                           
HORIZONTAL Wavelet H=1 : 5                         
HORIZONTAL Wavelet H=2 : 6                         
HORIZONTAL Wavelet H=3 : 7                         
HORIZONTAL Wavelet H=4 : 8                         
Get OUT : 9                                       
--------------------------------------            
Wavelet Selected = 4                              
Dilation min = 1                                 
Dilation Max = 14                                
Number of Dilations = 500                        
```
The output Figure shows the matrix of correlation coefficients across the scale of dilations over the entire profile, which allows the user to see if the analyses detected something. The figure does not need to be saved because it will be automatically generated from the coefficient matrix (output file #1) within the other part of the code.

4.9.2 Extraction of the extrema

The extraction of the extrema from the matrix of correlation coefficients is done in the second part of MWTmat code.

The input data correspond to the two .dat files generated from the first part of the MWTmat code, which are the:
- correlation coefficients .dat file
- parameter .dat file
- Extraction of the extrema line and filter to reduce extrema artefacts.

All the extrema present within the correlation coefficient matrix are extracted, however, some are artefacts, while others characterize the analyzed signal. In order to remove as many of the extrema artefacts as possible, a manual filter is included in the code. The filter is based on the length of the extrema in comparison to the total height of the figure (Fig. 4-13). Artefacts are commonly made of a small number of correlation coefficients and are located within the lower dilations. This filter avoids overloading the figure with useless information, however, the user must be careful not to remove extrema characterizing the real signal. Generally it is quite obvious what is noise and what it is not. Unfortunately, sometimes when noise is strong and anomalies are small in amplitude, or asymmetric, it may be difficult to remove the artefact without removing extrema characterizing the real signal. Generally, selecting a filter of 10 to 13 % (of the vertical length = number of dilation) is sufficient, but user interpretation is required to make the right choice.

- The export file of the second part of the MWTmat code will generate 4 files. The two first files are .csv (comma delimited) files and will be named by the user. These files are the coordinates (dilation, distance, correlation coefficient) of all the extrema. One file includes all maxima extrema and the second file all the minima extrema. The last two files are .dat files and they automatically receive the same name as the associated .csv files. They contain the length of each of the extrema, one file for the maxima extrema and one for the minima extrema.

4.9.3 Depth calculation

This part of the MWTmat code allows the user to select extrema and to calculate as many depths as the user wants.
The import files will allow the user to load the previous files in the following order:

- The correlation coefficient .dat file
- The maxima extrema .csv file. The associated .dat file will be automatically loaded.
- The minima extrema .csv file. The associated .dat file will be automatically loaded.
- The parameter .dat file.

Much later, before the end of processing, the user will have the possibility to load the input data file .xls, which will be plotted below the final figure of the depth analysis (e.g., Fig. 4-3a).

- The input file of the signal (same as in part 1). The topographic signal will be used to calculate the depth in m a.s.l, while the signal will be plotted at the end of the processing.


The user has two possibilities to calculate the depth. The first is by using the automatic extrema linear trend and the second is by using a manual extrema linear trend.

To clarify, manual here means that the user will be able to graphically select 3 points on any extrema line to generate its linear trend. In contrast, automatic trend corresponds to the generation of the linear trend of each of the extrema, by applying a linear regression based on all the correlation coefficient defining the extrema. Both methods have their advantages and disadvantages. The automatic trend is faster to use to acquire depths and very efficient when the extrema have a very good linear shape. The manual trend allows the user to have more flexibility to avoid merging sections of 2 extremas. However, manual trends may be often somewhat less accurate and are very strongly influenced by the user interpretation.

One must remember that the dilation acts as a band filter and that the small dilation correspond the high frequency and the high dilation correspond to
the low frequency. Thus, if a very small wavelength anomaly (high frequency signature) is on the edge of a larger wavelength anomaly (lower frequency) than itself, it is strongly possible that the extrema characterizing each of them will be connected at each other. The transition from one to another is always expressed by a break in the slope of the extrema (extrema within circle 2, Fig. 4-13a). Thus, the linear trend should not be calculated over entire extrema, but only on the part above the break in the slope, which corresponds to the larger anomaly. From an objective approach, each extrema should be isolated on the range of dilation that characterize it, however, because on a even input signal, several tens of extrema can be found, it becomes unrealistic to perform a single analysis for each of them. Thus, the manual trend allows the user to decide what sections better characterize the anomaly on the extrema. Generally, this is quite obvious, although it may sometimes be difficult to be certain. If the extrema has a poor linear trend, this also indicates the fact that the wavelet did not properly localize the anomaly and only minimal confidence should be attributed to this extrema. In this case, the user should not use the extrema at all. Another type of wavelet may be able to better characterize this anomaly and its associated extrema. This is one of the reasons that the MWT approach uses several wavelet types to better characterize the extrema and thus the depth of the source associated to the anomaly.

Even in manual mode, the "automatic" trend will be made by the program, allowing the user to use both automatic and manual trends to calculate the depth. I strongly encourage the user to use both manual and automatic trend for accurate depth calculations.

- **Name and generation of the trend files.**
  
The trend file is a .dat file and is organized as shown in Table 4-9 (for automatic trend) and Table 4-10 (for manual trend).
  
  - **The automatic trend** will associate a predefined name for each of the extrema. All the maxima extrema will have a name starting by "B" following by a number (e.g., B_1 or B_15). All the minima extrema will have a name starting by
“E” following by a number. The extrema number increase from the left to right of the figure. The choice of B and E is completely arbitrary.

- **The Manual trends** will allow the user to generate as many trends as she/he want by selecting the name of the automatic extrema file. The new trend (manual) will be saved in a new file having the name of the original file with the added “_manual” (e.g., B_3 will become B_3_manual). To generate more trends of the same extrema without erasing the previous manual trend for this extrema, the user will just have to select the file having the “Name_manual”, which will become “Name_manual_manual” (e.g., B_3_manual will become B_3_manual_manual; E_10_manual_manual will become E_10_manual_manual_manual).

<table>
<thead>
<tr>
<th>Arretnes #1</th>
<th>Arretnes #2</th>
<th>Intersection_ X</th>
<th>REAL_DEPTH Z</th>
<th>RSQUARE_FIT_ONE</th>
<th>RSQUARE_FIT_TWO</th>
</tr>
</thead>
<tbody>
<tr>
<td>E.dat</td>
<td>B1.dat</td>
<td>-1600</td>
<td>-52.816</td>
<td>0.999485</td>
<td>-0.00664715</td>
</tr>
<tr>
<td>E_Manu.dat</td>
<td>B1.dat</td>
<td>-1600</td>
<td>-50.2331</td>
<td>0.999755</td>
<td>-0.00664715</td>
</tr>
<tr>
<td>E3.dat</td>
<td>B2.dat</td>
<td>10.0405</td>
<td>-201.81</td>
<td>0.997408</td>
<td>0.997414</td>
</tr>
<tr>
<td>E3_Manu.dat</td>
<td>B2_Manu.dat</td>
<td>12.4095</td>
<td>-187.899</td>
<td>0.999441</td>
<td>0.999529</td>
</tr>
<tr>
<td>B2_Manu.dat</td>
<td>E3.dat</td>
<td>12.9433</td>
<td>-189.924</td>
<td>0.999529</td>
<td>0.997408</td>
</tr>
</tbody>
</table>

**Table 4-11**: Output file results of the depth calculation by MWTmat code. The units are in the same unit as the inputs file. In this example everything is in m.

- **Calculating depth**
  The depth calculation is straight forward. At each new depth calculation, the user has 3 choices: Two automatic trends, two manual trends or hybrid (1 manual trend and 1 automatic trend).
  - **Two Automatic trends** allow the user to select 2 automatic trend files (file with name of type “E_*” or “B_*”, but not the file having “_Manual”). If a
manual trend file is selected, each file will nevertheless contain automatic information within the first 3 columns of the file (Table 4-12).

- **Two Manual trends** allows the user to select 2 manual trend files (with name finishing by "_manual"). The first three rows of the fourth and fifth column are used to calculate the trend, while the three first columns (coordinates of the entire extrema) are used to calculate the R-square of the trend. If an automatic trend file is selected, an error message will be displayed and the user will return to the menu for depth calculation.

- **Combining 1 manual with 1 automatic trend** allows the user to calculate the depth by combining manual trend with automatic trend, in order to give all the possibilities on the depth calculation.

The result will be directly plotted on the figure of the correlation coefficient map (e.g., Fig. 4-3a) and the information relative to the depth will be displayed on the command window of Matlab. The user will thus have the possibility to save the depth in the output result file. In order to keep the figures with only the significant information, the user may delete the trends and symbol from the figure by selected the line (either symbol) on the figure and by selecting delete. Be careful to not delete important information by doing that. The depth calculations can be stopped at any time.

- **Plotting the input signal** before exiting, the user can plot the input signal below the figure result (e.g., Fig 4-3a). The signal input file is the same excel file used in part 1 of MWTmat code.

- **Export files**
The output result file is a comma delimited .txt file and organized as shown in Table 4-11. Output files can be used for more depth calculation; the results will be appended to the bottom of the files.
Example of display information when doing depth calculation:

<table>
<thead>
<tr>
<th>Do you want to use Automatic Trend or Manual or both?</th>
</tr>
</thead>
<tbody>
<tr>
<td>2 AUTOMATIC TRENDS = 1</td>
</tr>
<tr>
<td>2 MANUAL BASED TRENDS = 2</td>
</tr>
<tr>
<td>1 Manual &amp; 1 Automatic TREND = 3</td>
</tr>
<tr>
<td>Stop analyse =&gt; 4</td>
</tr>
<tr>
<td>choice ???1</td>
</tr>
<tr>
<td>E.dat</td>
</tr>
<tr>
<td>B2.dat</td>
</tr>
</tbody>
</table>

Position on Distance in meter:
-1.6000e+003

Depth in meter:
-52.8160

E.dat
R-Square arrete #1:
0.9995

B1.dat
R-Square arrete #2:
-0.0066

Do you want to save those results? Yes:1 No: =>2

4.9.4 Working with the results

A lot of depth calculations can be obtained from the analyses of one input file by using one wavelet (in this study, typically from a few tens to hundreds of depth values). Depending on how many linear trends (automatic or manual) are used, for a single source, the result files can contain up to several tens of depths. Thus, in order to be objective, the user should apply a statistical approach to obtain, a single depth value, a single horizontal value and the associated
standard deviation for each of the sources. Here, standard deviation (\( \sigma \)) represent the scattering of the data.

The statistical interpretation should be first made with a spreadsheet (e.g., Excel) for each wavelet analysis. All the depths having similar horizontal positions can then be grouped together, however, to avoid artefacts, the user should always use the raw depth values to calculate the mean and standard deviations.

### 4.10 Conclusion

On one hand, wavelet analysis is a powerful tool for signal analysis. On the other hand, Potential fields, commonly used to investigate the structure of and change occurring within volcanic edifice, is a problem with a non unique solution. The non unique solution makes the localisation of the Potential field source often a struggle to obtain. Developed previously by the work of Moreau [1997, 1999], the Poisson family is a series of wavelets, which allows us to localize the depth of the source responsible for the Potential field. This study developed a code in MATLAB, called MWTmat, which allows us to analyse Potential fields using real wavelets from the Poisson Family. This study uses Self-potential data in order to localize the depth of the water flowing within volcanic edifice, however, the code can also be used on gravity, or magnetic data. In order to avoid depth artefacts and minimize the error on the calculated depth, this study developed a new methodology, called Multi-scale wavelet tomography, is based on cross-correlation of calculated depths from multiple wavelet analyses using several wavelets from the Poisson family. This study uses 4 wavelets to calculate the water depth, where only depths found with at least 3 of the 4 wavelets are considered significant. Uncertainty on calculated depth is based on a statistical approach using the scattering of the results from all the analysis. Small uncertainties, typically smaller than 10 m, allow us to have
a good confidence on the depth. In contrast, the larger is the uncertainty, the lower the confidence on the depth. This study shows that the MWTmat code is efficient to detect sources located at least 250 m deep and works well even with significant topography variation along the analysed profile.

4.11 References


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and characterization of archeological structures, *Geophys. J. Int.*, 171, 1, 87–103


5 CHAPTER 05: DEPTH CONSTRAINTS OF SHALLOW HYDROTHERMAL SYSTEMS BY SELF-POTENTIAL AND MULTI-SCALE WAVELET TOMOGRAPHY. **

5.1 Abstract

In studies of active hydrothermal systems, the depth of the hydrothermal system is generally required, but rarely known via traditional geophysical exploration techniques. While previous studies have shown that continuous wavelet transform algorithms applied to Self-potential data can theoretically determine the depth of the hydrothermal fluids, this study uses multi-scale wavelet tomography with multiple wavelets, field measurements and geophysical models to accurately determine this depth. On Stromboli, Waita and Masaya volcanoes, multi-scale wavelet tomography of field measurements gives reproducible depth results, supported by independent geophysical measurements and models, and accurately locates the main water flow path at shallow depths. Unlike other traditional geophysical methods, multi-scale wavelet tomography using Self-potential data is a low cost tool to rapidly determine depths of the shallowest hydrothermal structures. This approach has the potential to significantly enhance our ability to locate geothermal systems and monitor active volcanoes.

5.2 Introduction

In volcano monitoring and geothermal exploration, the depth of the hydrothermal fluid is of significant importance in understanding changes in the hydrothermal system and how these changes relate to the underlying magma body. Hydrothermal systems are highly variable in shape and state of activity making them difficult to model. Potential field techniques (e.g., gravity, magnetism, Self-potential) enable indirect geophysical observations of subsurface structures and are commonly used to infer the horizontal extension of hydrothermal systems [e.g., Revil et al., 1999; Battaglia et al., 2006]. However, none of these give accurate depths of the main fluid cells that form the hydrothermal system. To achieve this, we apply signal processing algorithms to potential fields in order to locate the main water flow within volcanoes.

Wavelet transform is a method that allows us to characterize and locate discontinuities or abrupt changes in a measured signal. The depth localization of hydrothermal fluids was first attempted on Piton de la Fournaise volcano (La Reunion Island) by applying multi-scale wavelet tomography (MWT) to Self-potential (SP) data using complex wavelets based on the derivative of the Poisson kernel of different orders [Saracco et al., 2004]. Our study uses four real wavelets, whose derivative order varies from 2 to 3, and which are used separately to determine depths, along a profile, of the main sources generating the potential field measured on the surface [Moreau et al., 1997, 1999; Fedi et al., 1998; Sailhac et al., 2000, 2001; Gibert and Pessel, 2001; Saracco et al., 2004, 2007]. Due to the difficulties in obtaining good empirical data from exploration drill holes coincident with SP surveys in active hydrothermal systems, this study applies MWT to SP data from three volcanoes, Stromboli volcano (Italy), Waita volcano (Japan) and Masaya volcano (Nicaragua), for which other independent geophysical data exist.
5.3 Self-potential

The Self-potential (SP) method is a passive electrical technique which allows for measurement of the difference in potential between two points. The electrical potential is due to a natural current flowing through the rock, caused by several physical and chemical phenomena [Corwin and Hoover, 1979; Ishido and Mizutani, 1981; Zlotnicki et al., 2003; Lénat, 2007; Aizawa et al., 2008].

On active volcanoes or geothermal areas, SP anomalies are typically generated by two main processes. The first is the electrokinetic effect, caused by the displacement of fluid through a porous medium which disrupts the electrical charge balance between polarized minerals and the free ions within the pore fluid. This ionic displacement generates a current that can be several hundred’s of mV in amplitude [Corwin and Hoover, 1979; Zlotnicki et al., 2003]. The second source is typically the thermoelectric effect, which occurs due to a thermal gradient through the rock [Zlotnicki et al., 2003]. This thermal gradient increases the energy of the ions leading to a differential displacement between the ions, generating an electrical current, typically on the order of several tens of mV. Other chemical parameters such as pH and redox reactions will influence current generation [Aizawa et al. 2008], as well as changes in porosity and permeability which control the water flow carrying the ions [Jouniaux et al., 2000]. Heterogeneous ground resistivity may also have an effect on the electrical potential measured on the surface [Minsley et al., 2007].

However, previous studies on Stromboli, Waita and Masaya volcanoes have shown that the heterogeneities of the volcanic edifice have only a minor effect on independent geophysical data, therefore the SP modelling can be simplified assuming a homogeneous medium [Yasukawa et al., 2003; Revil et al., 2004; Finizola et al., 2006; MacNeil et al., 2007].
Figure 5-1: a) Geographic setting of Stromboli volcano, Italy. Red line is the approximate location of the 2004 SP profile [Finizola et al., 2006]. b) Geographic setting of Waita volcano, Kyushu Island, Japan. Green dashed line is Takenoyu fault and the red line is the approximate location of the 1995-1996 SP profiles [Yasukawa et al., 2003]; c) Geographic setting of Masaya volcano, Nicaragua. Black dashed line denotes the caldera margin. Red line is the approximate location of the 2007 and 2008 SP profiles (this study). Contours in m.
On active volcanoes and geothermal systems, the electrokinetic and thermoelectric effects are considered to be the most significant source of SP signals [Corwin and Hoover, 1979; Sailhac and Marquis, 2001; Zlotnicki et al., 2003; Revil et al., 2004; Lénat, 2007; Aizawa et al., 2008].

5.4 Multi-scale wavelet tomography (MWT)

The wavelet transform is a method which allows characterization of the time-frequency (spatial position) behaviour of a signal, or characterization and localization of discontinuities in the space and time domain. A full description of continuous wavelet transform (CWT) can be found in the pioneering work of Grossmann and Morlet [1984]. Multi-scale wavelet tomography (MWT) is a method based on CWT and potential field theory, where the final result is a depth rather than frequency. In this study, MWT is used with a combination of four real wavelets of the derivative of the Poisson kernel (horizontal and vertical derivatives). This Poisson wavelet family was developed to investigate the origins of potential fields [Moreau et al., 1997, Sailhac et al., 2000, 2001; Gibert and Pessel, 2001; Fedi et al.,1998, 2005; Cooper, 2006; Saracco et al., 2004, 2007] and is based on the combination of wavelet theory and the dilation property of the Poisson kernel solution of the Laplace equation. Depths are determined from MWT of the potential-field signal, where maxima and minima of extrema lines (amplitude of wavelet coefficient) converge toward the source generating the measured potential-field. In the case of water flow, the strongest electrical signals are generated at the top of the flow and at the edges between saturated and unsaturated areas [Zlotnicki et al., 2003].

In continuous wavelet processing, a singularity (anomaly) from a signal, $s$, is typically described by a local exponent obtained from the behaviour of the wavelet transform across the range of dilations. In the case of potential fields, the singularity characterizes dipoles ($\alpha = -3$) or monopoles ($\alpha = -2$). For a singularity with a homogeneous distribution order of $\alpha \leq 0$, the derivative order $n \in \mathbb{N}$ of the
wavelet must be \( n \geq -(1 + \alpha) \). If the signal, \( s \), generated by the singularity, confirms the homogeneous property of the wavelet transform [Moreau et al., 1997; Fedi et al., 1998; Sailhac et al., 2000, 2001], then the wavelet transform, \( L_{(b,a)} \), of the signal, \( s \), by a wavelet, \( g \), is:

\[
W_{(b,a)}(L,s) = L_{(b,a)} s = a^{-n} \int g\left(\frac{r - b}{a}\right) s(r) dr^n , \quad \text{with} \quad n \geq -(1 + \alpha) \tag{1}
\]

where \( n \in \mathbb{N} \), \( b \) is the translation parameter and \( a \) is the dilation parameter. The general equation of the horizontal derivative of order \( n \) of Poisson kernel family, \( H_n(u) \) [Moreau et al., 1997, 1999; Sailhac et al., 2000; Fedi et al., 2005], is in the frequency domain:

\[
H_n(u) = (2\pi u)^n \times \exp(-2\pi |u|) \tag{2}
\]

with \( u \) being the Fourier transform of the distance, \( x \), in the frequency domain. In the frequency domain, the vertical component does not change. However, as the analysed signal, \( s \), is expressed by \( u \), it is necessary to express the vertical derivative of order \( n \) of the Poisson kernel family in the domain frequency by using the horizontal component \( u \). One must then apply a Hilbert transform as described in the work of Sailhac et al. [2000] and Saracco et al. [2007]. The general equation of the vertical derivative of order \( n \) of the Poisson kernel family, \( V_n(u) \), in the domain frequency is:

\[
V_n(u) = -2\pi |u|(2\pi iu)^{(n-1)} \times \exp(-2\pi |u|) \tag{3}
\]

with \( i \) being the imaginary number. The result of the continuous wavelet transform, \( L_{(b,a)} \), is a matrix of correlation coefficients \( W_{(b,a)}(L,s) \) in the half-plane \( (b, a) \), where singularities of the signal result from convergence of extrema lines (Fig. 1, lines of minima and maxima) [Grossmann and Morlet, 1984]. Each
singularity consists of at least one line of maxima and minima. The lines of extrema converge (with \( z < 0 \)) [Moreau et al., 1997; Fedi et al., 1998; Sailhac et al., 2000, 2001; Gibert and Pessel, 2001] toward the singularity, which is the mathematical expression of the source, in this case the hydrothermal fluid cell.

The matrix of correlation coefficients, \( W_{(b,a)}(L,s) \), corresponds to the space above the ground surface, \( (z_0 > 0) \), however, projection below surface \( (x, z < 0) \) is possible due to the Poisson kernel properties. The depth determination is thus based on the electrical potential equation which can be deduced from Maxwell equations under a quasi-static limit [Saracco et al., 2004]. Potential field theory allows us to obtain the Poisson kernel function, where only its derivatives are admissible for wavelet processing [Moreau et al., 1997, 1999]. If the potential field in any plane \( z_0 = z \) and \( z_0 > 0 \) is known, then the potential can be calculated in the entire half-space \( (z > 0) \), without any assumptions about the sources. This is accomplished through field measurements.

A more detailed description of the mathematical methodology can be found in the work of Moreau et al. [1997, 1999], Fedi et al. [1998], Sailhac et al. [2000, 2001], Gibert and Pessel [2001] and Saracco et al. [2007].

In order to demonstrate the capability of multi-scale wavelet tomography to localise sources, a synthetic Self-potential signal has been generated from three dipoles at specific depths and orientations and processed by MWT using the second vertical derivative of Poisson kernel \( (V2) \) (Fig. 2, Table 1). Analysis of the synthetic SP signal (sampled with a 4 m step) was made with 500 dilations ranging from 4 to 20.

The first dipole is oriented horizontally at 50 m depth, the second vertically oriented at 200 m depth and the third dipole is inclined at 45 degrees at a depth of 125 m. Each dipole is a rod with a length of 20 m. Figure 2a shows the synthetic SP signal without noise (Signal/noise ratio is 1). In order to simulate a natural signal, Gaussian noise was added to generate synthetic signals with signal/noise ratios (SNR) of 0.90 and 0.80 (10 and 20% noise, respectively).
Figure 5-2: Calculated depths of three theoretical dipoles at different depths and orientations (see text and Table 1). From left to right: D-1) horizontal dipole at Z = -50 m; D-2) vertical dipole at Z = -200 m; D-3) dipole with a 45° dip at Z = -125 m. Multi-scale wavelet tomography results with the second derivative (V2) of the Poisson kernel over 500 dilations on a range from 4 to 20. 

2a) Result for a signal without noise. 2b) Results with a signal/noise ratio of 0.8. Top: B to B4 and E to E5 represent maxima and minima lines, respectively. Middle: cone shaped structures with intersects at the depth of each source. Bottom: synthetic Self-potential signal generated by the three dipoles, with SNR of 1.0 (a) and 0.8 (b). Ss is the synthetic signal depth, Cd is the calculated depth and σ is one standard deviation.
Table 5-1: Synthetic source depths calculated by multi-scale wavelet tomography of synthetic Self-potential of 3 dipoles with different orientation and depth. The Synthetic signal is the reference depth used to generate the synthetic Self-potential signal. A Signal-Noise ratio (SNR) of 1, of 0.9 and of 0.8 was applied on the Synthetic Self-potential signal to investigate noise effect on the calculated depth. Synthetic signal (Ss) shows the depth and position of each dipole. Calculated depth (Cd) is the depth calculated by the multi-scale wavelet tomography with the wavelet V2. $\sigma$ is one standard deviation.

<table>
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<th>$\sigma X$ in m</th>
<th>$Z$ in m</th>
<th>$\sigma Z$ in m</th>
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<td>-50</td>
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Figure 2b shows the result of the MWT processing on the synthetic SP signal with a SNR of 0.80. The results show a good behaviour of the wavelet analysis with the increasing noise (Table 1). Variations and errors on depth calculations are due to two main sources. First, noise in the data, due to a heterogeneous medium [Sailhac et al., 2001], will distort the lines of extrema for small dilation values and thus will affect the point of convergence, as seen with the synthetic SP signal (Fig. 2, Table 1). The associated error due to the heterogeneous medium is calculated based on the quality of the best fit of the
cone shaped structure (Fig. 2). The sampling step also has a significant impact on the measured signal and on how well the signal may characterize the source. This study uses SP data collected with the common sampling steps of 20 or 50 m, which have proven sufficient to obtain reproducible depths. The error due to field measurement noise can be further limited by processing the signal with several wavelets (this study uses 4) in order to statistically constrain a depth. As discussed previously, heterogeneity in subsurface resistivity can sometimes be a significant source of electrical generation [Minsley et al. 2007]. However, previous studies on both Stromboli [Revil et al., 2004] and Waita volcanoes [Yasukawa et al., 2003] have shown that SP signals are principally due to the electrokinetic effect on these volcanoes. Furthermore, these studies have shown that the main subsurface resistivity changes were associated with the boundary of underground water flow and so their effects on SP generation can be regrouped with the effect of the water flow. On Masaya volcano, the resistivity contrast is associated with the upper edge of the water table [MacNeil et al., 2007]. We can therefore assume that the medium for each of these three volcanoes may be simplified as a homogeneous resistivity medium. Previously published models on each volcano will allow us to investigate and constrain the significance of depths calculated by MWT of Self-potential data.

5.5 Geologic Setting

5.5.1 Stromboli volcano

Stromboli volcano is an island stratovolcano, located in the Tyrrhenian Sea, Italy (38.178°N, 15.212°E) (Fig 1a). It forms the northernmost part of the Aeolian archipelago and is one of the world’s most active volcanoes, with persistent eruptive activity for at least the last 2000 years [Gillot and Keller, 1993]. The sub-aerial eruptive history of Stromboli volcano began ~100,000
200 years ago [Hornig-Kjarsgaard et al., 1993] and has resulted in a strong structural-controlled morphology, mainly controlled by a regional NE-SW trend of dyke injection [Pasquaré et al., 1993]. The volcanic stratigraphy of Stromboli is principally composed of pyroclastic ash and scoria deposits and the main hydrothermal system is located in the summit area ($T_{\text{ground}} \sim 80^\circ\text{C}$, $\text{CO}_2\text{gas flux} \sim 10,000 \text{ g m}^{-2} \text{ d}^{-1}$, Self-potential anomaly $\sim 250 \text{ mV}$ in amplitude).

**Figure 5-3:** Top: Comparison between MWT-calculated depths of hydrothermal fluids (squares and diamonds) and the electrical resistivity model [from Finizola et al., 2006] of a NE-SW profile across Stromboli volcano, Italy. V2, V3, H2 and H3 are second and third order of the vertical and horizontal derivatives. Error bars represent one standard deviation, see Table 2. Bottom: SP profile across the volcano. Modified from Finizola et al. [2006].
Stromboli shows a wide range of ground resistivity (< 100 $\Omega$ m$^{-1}$ to > 3,000 $\Omega$ m$^{-1}$) and previous SP studies correlate the structural boundaries with the extension of the complex hydrothermal system within the volcano [Finizola et al., 2002, 2003, 2006; Revil et al., 2004]. This study uses a SP profile acquired in May 2004 across the volcano (NE-SW), 3820 m in length with a sampling step of 20 m (Fig. 1a, 3) and a signal/noise ratio of 0.90. The profile was complimented by a resistivity profile, ground temperature and soil CO$_2$ flux measurements [Finizola et al., 2006].

5.5.2 Waita volcano

Waita volcano is located in the central part of Kyushu Island, Southern Japan (33.140°N, 131.165°E) (Fig. 1b), in the western part of Hohi Geothermal Region (HGR) [Yasukawa et al. 2003]. Waita volcano is situated within the Beppu-Shimabara Graben [Kamata, 1989]. The main fault structures are related to the Takenoyu fault (normal fault, oriented NW-SE), while secondary normal faults are oriented EW [Ikeda, 1979]. Significant hydrothermally altered deposits are distributed along the Takenoyu fault, suggesting fault-controlled flow path [Yasukawa et al. 2003]. Waita volcano hosts more than 20 hot springs (ranging from 36.5 to 97.8°C) and total hot water discharge is estimated at 524 l min$^{-1}$, with the total heat discharge reaching 1352 kj s$^{-1}$ [Kawamura, 1985]. Previous studies of the hydrothermal system have shown that hydrothermal fluids are rising from depth under Waita volcano and mixing with meteoric water flowing from the top of the volcanic edifice [Yasukawa et al., 2003].
Figure 5-4: Comparison between MWT-calculated depths of hydrothermal fluids (squares and diamonds) and the fluid velocity distribution model (Model 3, case 2, from Yasukawa et al. [2003]) with the Self-potential profile of Waita volcano, Japan. A'-B-C is a NW-SE profile across the summit. V2, V3, H2 and H3 are second and third order of the vertical and horizontal derivative. Error bars represent one standard deviation, see Table 2. Modified from Yasukawa et al. [2003].

The hydrothermal fluids then flow laterally westwards within the volcanic edifice along the Takenoyu fault. The SP profile used in this study is oriented NW-SE across the summit of Waita volcano and was made in 1995 and 1996 [Yasukawa et al, 2003]. The SP data consists of a 2000 m long profile with a sampling step of 50 m, and a 1600 m long segment based on forward modeling [Yasukawa et al, 2003] (Fig. 1b, 4) and has a signal to noise ratio of 0.99. Rock resistivity along the profile range from 1 to 300 Ω m⁻¹.

5.5.3 Masaya volcano

Masaya volcano is located in southwestern Nicaragua (11.984°N, 86.161°W), Central America, about 20 km south of the capital, Managua. Masaya
is a basaltic shield volcano (Fig.1c) with a caldera structure (6 km by 11.5 km) [Williams, 1983]. Post-caldera activity is largely dominated by effusive eruptions with only small volumes of scoria and ash interlayered between lava flows; a complete stratigraphic description can be found in Williams (1983). Recent eruptive centers are distributed along a ring fault, which is cut on its east part by the Cofradrias fault [Williams, 1983; Walker et al., 1993]. The main cones are Masaya, Nindiri, Cerro Montoso, and Comalito (Fig. 1c). For the last ~ 150 years, volcanic activity has been centered on Santiago crater (east part of Nindiri cone) and dominated by persistent open vent degassing, infrequent lava lake formation and small vent-clearing explosions [Rymer et al., 1998; Roche et al., 2001; Williams-Jones et al., 2003].

Previous geophysical studies with transient electromagnetic methods have shown that the north flank of Nindiri cone hosts a shallow water table while the south flank hosts an underground vapour dominated zone [MacNeil et al., 2007], which likely represents a small hydrothermal system. Two SP profiles were collected across the Nindiri cone in 2007 and 2008 (this study, Fig. 1c, 5), with a sampling step of 20 m. The signal/noise ratio of this Nindiri profile was of ~0.99 in both 2007 and 2008. All data are referenced to the Laguna de Masaya via a SP mapping survey more than 15 km in length. The SP profiles (2007 and 2008) show the presence of two distinct underground water structures (aquifer and hydrothermal system).

An extended positive SP anomaly (~200 mV), coincident with the vapour dominated water body [MacNeil et al., 2007], is interpreted as the expression of a small hydrothermal system located on the south flank of Nindiri cone and within Nindiri crater (Fig. 5). The positive SP anomaly is independent of the topographic variation (blue profiles on Fig. 5), which is typically a sign of uprising fluid and thus hydrothermal activity. Through the north flank of Nindiri cone, groundwater flows northward and is characterized by a low Self-potential anomaly (~100 mV), which is controlled by topographic change (SP/elevation gradient from -0.18 to -0.80 mV m⁻¹, Fig. 5).
<table>
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<tr>
<td>$\eta_2$</td>
<td>4</td>
<td>780</td>
<td>8</td>
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<td>368</td>
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<tr>
<td>$\eta_3$</td>
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<td>1150</td>
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<td>-111</td>
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<tr>
<td>$\eta_4$</td>
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<td>1560</td>
<td>11</td>
<td>-98</td>
<td>21</td>
<td>406</td>
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<td>$\eta_5$</td>
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<td>2330</td>
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<td>11</td>
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<td><strong>Masaya volcano in 2008, Figure 5b</strong></td>
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<td></td>
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<tr>
<td>$\eta_1$</td>
<td>3</td>
<td>283</td>
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<td>$\eta_{2a}$</td>
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<td>632</td>
<td>8</td>
<td>-65</td>
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<td>2649</td>
<td>43</td>
<td>-105</td>
<td>60</td>
<td>385</td>
</tr>
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</table>

**Table 5-2:** Source depths calculated by multi-scale wavelet tomography of Self-potential profiles on Stromboli, Waita and Masaya volcanoes. Number of wavelets used in MWT to locate the source along profile. $\sigma$ is one standard deviation. *Structure names from Finizola et al. [2006]
5.6 Results

In order to investigate the accuracy of multi-scale wavelet tomography calculated depths of hydrothermal fluids, this study uses a combination of 4 real wavelets based on the Poisson kernel function: the second and third vertical derivative (V2 and V3, respectively, equation 3) and second and third horizontal derivative (H2 and H3, respectively, equation 2).

Water flow is generally considered to have a dipolar behaviour ($\alpha = -2$), which in the case of an extended source can be considered as 2 monopoles ($\alpha = -1$) [Sailhac et al., 2001]. Using equation 1, the wavelets (second and third derivatives) have an order of homogeneous distribution of $n = 2$ and $n = 3$. Each analysis, for each data set, was made with each wavelet over 500 dilations with a range of dilation from 1 to 20. Only depths found with at least three of the four wavelet analyses are considered significant. For each wavelet and each profile, a depth estimate is obtained by calculating the best fit of the cone shaped structure (Fig. 2). For each depth estimate, the associated error is based on the quality of the best fit.

The final localization of each source (horizontal and vertical) is based on the mean depth of all the raw data from all wavelets analysed and their associated uncertainty ($\sigma$). Results for the three volcanoes are synthesised in Table 2.

As discussed previously, noise may be a source of error in the depth calculation. Calculated signal/noise ratio (SNR) on SP profile for both Waita and Masaya (2007 and 2008) volcanoes is of 0.99, while the SNR of the SP profile of Stromboli volcano is of 0.90. Even with a SNR as low as 0.8, the MWT-calculated depths from the synthetic SP profile (Fig. 2 and Table 1) show uncertainties less than 25 m. Thus the error due to noise on the calculated depths from the three volcanoes is considered to be negligible.
Figure 5-5: Comparison between the MWT-calculated depths of water table (squares and diamonds) and the TEM water table model (from MacNeil et al. [2007]) with the Self-potential profile of Masaya volcano, Nicaragua. V2, V3, H2 and H3 are second and third order of the vertical and horizontal derivative. Error bars represent one standard deviation, see Table 2.
<table>
<thead>
<tr>
<th>Method</th>
<th>elevation in m a.s.l.</th>
<th>South flank</th>
<th>North flank</th>
<th>Upper North flank</th>
<th>Lower North Flank</th>
</tr>
</thead>
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<tr>
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<td></td>
<td>σ</td>
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<tr>
<td>TEM</td>
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<td>271</td>
<td>342.9</td>
<td>200</td>
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<td></td>
<td>σ</td>
<td>67</td>
<td>85</td>
<td>28</td>
<td>16</td>
</tr>
<tr>
<td>Difference of elevation in m</td>
<td>144</td>
<td>62</td>
<td>-2</td>
<td>97</td>
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</tr>
</tbody>
</table>

Table 5-3: Comparison of water depth between multi-scale wavelet tomography of Self-potential profiles and TEM model on Masaya volcano. σ is one standard deviation of the scatter of the calculated depths. TEM water model is from MacNeil et al. [2007].

On Stromboli volcano, multi-scale wavelet tomography of the 2004 SP profile identified 18 different depths, in 5 groups (Fig. 3, Table 2) that characterise 5 sources: 5 depths from H2, 5 depths from H3, 4 depths from V2 and 4 depths from V3. Each of these 5 sources covers a specific area which is horizontally less than 80 m wide and vertically spread over 35 to 150 m (Table 2).

On Waita volcano, the multi-scale wavelet tomography of the 1995-1996 SP profile identified 19 different depths, from 5 sources (Fig. 4, Table 2): 5 depths from H2, 5 depths from H3, 5 depths from V2 and 4 depths from V3. Each of these 5 sources covers a specific area which is horizontally less than 50 m wide and vertically spread over 40 to 150 m (Table 2).

On Masaya volcano, across the active Nindiri cone, multi-scale wavelet tomography of the 2007 and 2008 SP profiles identified 23 and 26 different depths, respectively, from 6 sources (Fig. 5, Table 2). In 2007, 6 depths are from H2, 6 depths are from H3, 5 depths are from V2 and 6 depths are from V3. The 2007 depths are spread over 6 different locations (η1 to η6, Fig. 5a). In 2008, 6 depths are from H2, 6 depths are from H3, 7 depths are from V2 and 7 depths are from V3. The 2008 depths are organized in 7 groups (η1, η2a, η2b, η3, η4, η5, η6, Fig. 5b). The horizontal and vertical positions of the each of the 5 groups (η1, η3, η4, η5, η6, Fig. 5) are very similar. Between 2007 and 2008, the only
difference is for the source $\eta_2$ (Table 2, Fig. 5); in 2008, two sources, $\eta_2a$ and $\eta_2b$, were located in proximity to $\eta_2$ ($\Delta x_{\eta_2a - \eta_2} = 148 \text{ m}, \Delta z_{\eta_2b - \eta_2} = 50 \text{ m}; \Delta x_{\eta_2b - \eta_2} = 14 \text{ m}, \Delta z_{\eta_2b - \eta_2} = 38 \text{ m}$). As both $\eta_2a$ and $\eta_2b$ sources where calculated with 4 wavelets and are only 12 m vertically from each other (Table 2), we consider them as equivalent to the $\eta_2$ source.

### 5.7 Discussion

On Stromboli volcano, previous studies [Finizola et al., 2002; 2003, 2006] have shown the presence at shallow depth, from NE to SW, of three conductive structures (C1, C2 and C3) and three resistive structures (R1, R2, R3) (Fig. 2). Our MWT results show 5 sources at relatively shallow depths (from 60 to 300 m) below the topographic surface with almost all located within the low resistivity regions (Fig. 2, Table 2): one source is in the C2 conductive structure, three sources in the C3 structure and one source is inside the R1 resistive structure. Strong correlations between SP, ground temperature, soil CO$_2$ concentrations and electrical tomography support the existence of strong hydrothermal activity (C3) on the summit of the volcano [Finizola, 2006] (Fig. 2).

The depths calculated by MWT correlate very well with the electrical tomography anomalies (Fig. 2) and show the presence of 3 shallow sources inside the C3 structure (Fig. 2). Two sources are located on the summit and a third is on the upper SW flank. Thus, the two shallow sources (60 to 130 m below the topographic surface) likely characterize the main hydrothermal fluid cells (Fig. 2, Table 2) that shape the summit hydrothermal system (C3) of Stromboli. Similar correlations are seen with the calculated depths in the C2 conductive structures (Fig. 2, Table 2). This study also found one source (with each of the four wavelets) within the R1 resistive structure (Fig. 2). In R1, there is a small low resistivity anomaly at shallow depth, embedded in a high resistivity background [Finizola et. al., 2006]. This low resistivity area is only correlated with high CO$_2$
concentrations (> 4000 ppm) and its origin is associated with a developed vegetation area in this part of the island. However, given the four MWT-derived depths found for this source and the high CO₂ background, the R1 structure may represent an area of low-level hydrothermal activity surrounded by less permeable deposits.

On Waita volcano, the 5 groups of calculated depths also characterize five sources (#1 to #5, section A'-B on Fig. 3, Table 2) and correlate with the main flow structures defined by Yasukawa et al. [2003]. From left to right on this section, the first source (#1, Table 2) correlates with the upper part of the western water flow (Fig. 3). The second source (#2) is at the edge of the water flow with an impermeable layer. While the flow is not intense, the differential displacement is clearly sufficient to generate an electrical anomaly and be localized by the MWT. The third source (#3), beneath the NW side of the summit (Fig. 3), is near the main convergence of flow inside the summit area of the hydrothermal structure. The fourth source (#4) is spatially less well constrained. However, it is also at the intersection between two water flow paths: an eastward water flow from the west summit flank and a westward flow from the lower part of the east flank (Fig. 3). The fifth source (#5) is found on the SE boundary of the main water flow structure flowing westward.

As stated by Yasukawa et al. [2003], the water flow inside Waita volcano is westward and controlled by the regional Takenoyu fault. The depths calculated by MWT of the SP data show that 4 of the 5 sources (#1, #3, #4, #5) are found in the shallowest part of this westward flow. All are along the main flow direction and/or at the intersection of converging water flows, which likely supply the hydrothermal system.

On Masaya volcano, 6 groups of depths calculated by MWT characterize 6 sources in 2007 and 2008. The horizontal position and depths of these sources are very consistent between 2007 and 2008 (Fig. 5, Table 2), with a horizontal coefficient correlation \((r^2_x)\) of 0.998 and a vertical coefficient correlation \((r^2_z)\) of 0.81. Calculated water depths show a constant northward flow on the north flank of Nindiri cone and a southward flow on the south flank. The calculated depths
(Table 2) indicate that, during 2007 and 2008, the water table is shallow (mean of 92 ± 34 m below the topographic surface). In 2004, MacNeil et al. [2007] made transient electromagnetic (TEM) measurements through the caldera, including a profile across Nindiri cone (Fig. 1). Their study showed the presence of a low resistivity (<103 Ω.m⁻¹) layer which they inferred as the water table (Fig. 5). This TEM data shows a very similar trend with the water table becoming progressively shallower with increasing elevation. Although there is some discrepancy between the TEM modelled water table and the MWT calculated depths, when the mean depths (and uncertainties) from both methods are compared (Table 3), the depths of the water tables are indistinguishable.

As with all geophysical methods, especially potential field techniques, there are a number of limitations that must be considered in the application of Multiscale wavelet tomography [Sailhac and Marquis, 2001]. MWT on SP data cannot uniquely determine the nature of the object associated with the SP anomaly and calculated depth. As is commonly the case with many inverse models, MWT assumes a homogeneous medium, which is a significant simplification. Large heterogeneity in resistivity of the medium may noise the electrical signal generated by the water flow [Minsley et al., 2007]. Significant contrast resistivity, when not associated with underground water flow, will result in an error on the calculated water depth. However, this error should still be within the uncertainty of the MWT depth, which range from a few meters to several ten’s of meters (Table 5-1; Table 5-2) and thus will not really affect the accuracy of the depth. More often, underground water flows are within the most permeable layer, which also have the lowest ground resistivity. The resistivity contrast will be associated with the border of underground flow (e.g., Stromboli volcano, Fig. 5-3) [Finizola et al., 2006] and thus be the edge of the underground water flow. In this case, significant resistivity contrast should not have any effect on the water depth. It is therefore important that SP surveys be made in conjunction with electrical tomography or electro-magnetic surveys to characterise the principle resistivity structures and how they are associated with the MWT-calculated depths. Multiscale wavelet tomography of SP data has the advantage of being a rapid signal
analysis technique which allows one to obtain reproducible results. As shown by this study, MWT on SP data can be used even on a signal with significant noise (SNR of 0.8) and when a combination of at least four different wavelets of the Poisson kernel family is used, it becomes possible to reduce the number of artefacts and thus the uncertainty on the calculated depth. Finally, the accurate depth constraints from MWT of SP, when applied to time-series data, may be used to determine vertical change in the water flow.

5.8 Conclusion

This study demonstrates that multi-scale wavelet tomography, based on the Poisson wavelet family and applied to Self-potential data, can reproducibly locate the boundaries of hydrothermal and hydrological structures, even with signal/noise ratio reaching 0.8. The MWT-calculated depths for Stromboli, Waita and Masaya volcanoes correlate well with structures identified by independent geophysical methods. Traditionally, a SP survey a few km in length (e.g., profiles used in this study) can be made in a few days and complete data processing by MWT can be done in a single day. The vertical accuracy is sufficient to locate the top of the main hydrological structures and help to optimize drilling in the case of geothermal exploration. In comparison at other geophysical techniques, the MWT applied on SP data combines the multi-scale analyses, depth calculation and the fast processing of the data of the MWT with the low cost and easy doing of the SP technique. When applied to SP data from active volcanoes, MWT-derived depths can characterize the hydrothermal and/or hydrological systems and their relationship to the underlying magmatic plumbing system. Multi-scale wavelet tomography analysis of time-series Self-potential data has the possibility to significantly improve volcano monitoring by accurately determining changes in the hydrothermal system which may be precursors of imminent volcanic activity.
5.9 Acknowledgements

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5.10 References


6 CHAPTER 06: STRUCTURE AND DYNAMICS OF THE HYDROTHERMAL COMPLEX OF MASAYA VOLCANO, NICARAGUA

Figure 6-1: Nindiri crater on Masaya volcano in March 2006. Foreground: San Pedro crater and the Nindiri lava lake. Middle: Santiago crater and beyond it, on the left, Masaya cone. Background: Mombacho volcano in the distance.
6.1 Abstract

Masaya volcano, Nicaragua, is a persistently active volcano characterized by continuous degassing for more than 150 years. This study highlights the existence of an extensive hydrothermal system throughout the volcano, which is spatially controlled along a ring fault. The most intense hydrothermal activity is found on the North flank of Masaya cone, while the active Comalito solfatara is of lower intensity in comparison to it. In the close vicinity of the active vent of Santiago crater, the Nindiri hydrothermal system is established throughout the entire center and south part of Nindiri cone. In contrast, the center of the north part of the caldera hosts several aquifers. Water depths have been found to be always less than 150 m deep and often less than 100 m below the topographic surface. Between 2006 and 2009, this study shows that the hydrothermal systems are stable without significant change, due to the constant volcanic activity expressed on the surface by the continuous passive degassing. Type and localization of underground water flows were determined by combining Self-potential, soil CO$_2$ concentration and ground temperature measurements. Water depth was determined by Multi-scale wavelet tomography on Self-potential data using wavelets from the Poisson family. The method of Multi-scale wavelet tomography on Self-potential has proven to be an efficient way to locate and investigate the vertical distribution of underground water on the active Masaya volcano. The extended hydrothermal system through Masaya and the significant ground CO$_2$ anomalies indicates that all the gas does not escape through the open vent of Santiago crater and that the ring fault structure has a significant control on the extent of the hydrothermal system.

6.2 Introduction

Over the last 150 years, Masaya volcano (11.984°N, 86.161°W, 635 m), a basaltic shield volcano in Nicaragua, has continuously been characterized by a significant passive magmatic degassing (SO$_2$ flux > 500 t d$^{-1}$) [Williams-Jones et
al., 2003; Chapter 1-1]. Unlike other basaltic shield volcanoes, which are characterized by frequent magmatic eruptions (e.g., Kilauea, Piton de la Fournaise), Masaya volcano does not have frequent emission of juvenile magma (by explosive, either effusive eruptions). Masaya consists of a caldera (~ 11.5 x 6 km) hosting several cinder cones organized along a ring fault structure [Rymer et al., 1998a, 1998b; Roche et al., 2001; Acocella, 2007]. The area delimited by the ring fault (dashed black line, Fig. 1-5 in Chapter 1), is called in this study the inner ring fault to facilitate discussion. The caldera is partially filled by numerous lava flows [Williams, 1983b; Walker et al., 1993; Rymer, 1998; Roche et al., 2001]. The main cinder cones are Masaya and Nindiri cones (Fig. 6-1). All volcanic activity (small explosions and continuous degassing) currently originate from Santiago crater inside the Nindiri cone. The presence and frequent renewal of significant amounts of fresh magma near the surface at Masaya, is physically expressed through the continuous degassing. Previous geophysical investigations have supported that fact through long term gravity, SO$_2$ gas flux variations and volcanic tremor measurements [Métaxian et al., 1997; Rymer et al., 1998; Williams-Jones et al., 2003]. The continuous presence of fresh magma within a shallow magmatic reservoir is an extraordinary persistent source of heat and gas. Volcanic gas, whatever the species, is a very good heat carrier, which will control the heat distribution within the volcanic edifice. In the vicinity of the magma reservoir, the heat will be mainly transferred to the country rock by conduction. However, rocks are poor thermal conductors, thus it cannot efficiently or reasonably transfer heat through the entire edifice, while gas can. Sustained gas and heat flux within the ground will generate and support hydrothermal fluid circulation and thus hydrothermal systems. In the case of an open system such as at Masaya volcano, where the magma is directly in contact with the atmosphere, gases are easily evacuated from the magma through the open conduit and thus only a small percentage of the total gas flux may escape through the rock forming the edifice. On Masaya volcano, the volcanic activity is traditionally monitored through gas flux (SO$_2$, CO$_2$, etc.), gravity and seismic surveys [Métaxian et al., 1997; Rymer et al., 1998; Williams-Jones et al., 2003].
Only a few previous studies on Masaya volcano, have investigated the change in magmatic activity through the hydrothermal activity present within it [Lewicki et al., 2003; Pearson et al., 2008]. However, these studies were of limited spatial extent, but over long time periods. They focused on the solfatara of Comalito cone, a few kilometers from the active crater Santiago (Fig. 1-5, Chapter 1) and not near the main center of activity. Previous work on the hydrogeology of Masaya volcano has shown that the caldera floor hosts aquifers flowing toward the Laguna de Masaya (Eastern part of the Caldera) and following the topographic drop in elevation [MacNeil et al., 2007]. A low resistivity layer, assumed to be associated with presence of underground water, was found at shallow depth within the upper slope of Nindiri cone [MacNeil et al., 2007]. However, to date, it is not clear how the underground water is distributed between pure aquifers and hydrothermal systems. Previous geophysical work on Masaya volcano has focused on magmatic change over time through dynamic gravity and seismic studies [Metaxian et al., 1997; Rymer et al., 1998a, 1998b; Williams-Jones et al., 2003]. On a very active volcano, due to a persistent and significant gas/heat flux, it is common to find well developed hydrothermal systems. However, on Masaya volcano, where the main gas/heat flux escapes not through the rock, but directly to the atmosphere, the hydrothermal activity may not have enough gas/heat input to occur, or at least to be detectable. The knowledge of the location and structure of the hydrothermal system, if any, is of significant importance to volcano monitoring. In some circumstances, the hydrothermal system can have a direct effect on the trigger of eruptions [Browne et al., 2001, Sasai et al., 2002; Zlotnicki and Nishida, 2003]. This study investigates the presence and location of hydrothermal systems on Masaya volcano, how they relate to each other and how they change over time and interact with the other systems (magmatic activity and aquifer) within the volcano. Between February 2006 and March 2009, 4 geophysical surveys were performed to collect ~ 70 km of Self-potential profiles, ~ 7.5 km of soil CO₂ concentration and ~ 6.5 km of ground temperature. Survey profiles have been used to investigate both spatial and temporal variations. A regular sampling step of 20 m
was used during each survey. In order to investigate change in the hydrothermal system, water depth calculations were made by applying Multi-scale wavelet tomography to the Self-potential data [Mauri et al., Submitted 2009, Chapter 5 and Appendix 1]. A total of 56 analyses by Multi-scale wavelet tomography on the Self-potential data allows us to obtain 25 time-series data on water depth including 82 depth values.

6.3 Methodology

6.3.1 Self-potential

The Self-potential (SP) method is a passive electrical technique which allows us to detect the presence of underground water. Aquifers and hydrothermal systems are characterized by their own specific electrical signatures. Self-potential data may not always be sufficient to clearly characterize one from the other and thus soil CO2 gas concentration and ground temperature are required [Finizola et al., 2004; Finizola et al., 2006; Lénat, 2007]. Electrical generation occurs when water flows through a rock (porous medium) and when a temperature gradient is applied to a rock. Other sources exist, but are much smaller than these two. A complete description of the sources of Self-potential signals and field methodology can be found in Chapter 2.

The measurements were made with a copper/copper sulphate electrode, a 300 m wire copper cable and a high impedance multi-meter (~ 200 MOhm, Fig. 6-2a). GPS position and notable geologic structures were recorded. All profiles, except the Laguna road profile, are closed in a loop and are corrected for drift, which was never more than 40 mV. Electrode polarization was controlled at least twice a day and never exceeded more than 2 mV of polarization, and generally was close to 0 mV, which allows us to have good confidence in the data.
6.3.2 Soil CO$_2$ gas concentration

The soil CO$_2$ gas concentration method is a passive method, which allows us to measure the concentration of CO$_2$ gas within the ground. Atmospheric CO$_2$ concentrations range from ~350 to 500 ppm. CO$_2$ is the first magmatic gas to escape from the magma, rising through magma and country rock along the main structural faults [Finizola et al., 2004; Chiodini et al., 2005; Bruno et al., 2007; Finizola et al., 2006; Revil et al., 2008]. Magmatic gas can generate CO$_2$ gas concentration up to more than 20,000 ppm (20%) within the soil pores space. However, some vegetation can also release CO$_2$ through their root systems and may generate up to a few 1000’s ppm of CO$_2$. Measurement of soil CO$_2$ gas concentrations were made by inserting a probe 60 cm deep in the ground and by pumping the gas from the ground to a sensor. The device used was a CO$_2$-meter of type GM70, with GMP221 probe by Vaisala Inc. (Fig. 6-2b), which can accurately record concentration up to 20,000 ppm. The sampling step measurement was 20 m. Atmospheric CO$_2$ concentration was controlled several times each hour, in order to detect and control any drift on the device calibration.
6.3.3 Ground temperature

The ground temperature method is a passive method, which allows us to obtain a thermal image of the ground. Hot/boiling gas and hydrothermal fluids lose their temperature by conduction to the surrounding ground and by condensation of the gas. Ground temperature, at more than 30 cm depth, is protected from daily temperature variation due to the poor thermal conduction of soil. When there is no magmatic heat source, the temperature of the ground is commonly below the atmospheric temperature (> 20 °C). When the ground is heated by hot rising gas or hot/boiling hydrothermal fluids, the ground temperature can reach more than 300°C [Finizola et al., 2004; Chiodini et al., 2005; Bruno et al., 2006; Finizola et al., 2006; Revil et al., 2008]. Generally speaking, ground temperature is linked with CO₂ gas anomalies and allows detection of the main fracture paths followed by the gas. Ground temperature measurements were made with a temperature probe (K-type chromel-alumel probe and generic probe, Fig. 6-2b) having an accuracy of ~0.2°C; measurements were made at a depth greater than 30 cm. Ground temperature measurement were made with a sampling step of 20 m in the same spot as the SP measurement and following it. Atmospheric temperature was measured at least each 100 m but often every 20 m.

6.3.4 Multi-scale wavelet tomography

Multi-scale wavelet tomography is a mathematical method based on continuous wavelet analyses using the Poisson kernel family. The complete description of the Multi-scale wavelet tomography (MWT) the computing code and methodology can be found in Chapter 4. Each Self-potential profile is analyzed with the 4 same real wavelets, which are the second and third vertical derivative (V2, V3) and the second and third horizontal derivative (H2, H3).
These 4 wavelets allow us to detect monopole and dipole sources [Moreau et al., 1997, 1999, Mauri et al., Submitted 2009a, b]. Analyses were made by re-sampling each profile to a sampling step of 4 m and by applying 500 successive dilations of the analyzing wavelet on a range from 1 to 20. All the calculated depths of the water are presented in Appendix A5, however, in this study only depths found with at least 3 of the 4 wavelets are considered to be significant and thus presented here. All the results can be found in Appendix A5.

6.4 SP/elevation gradient

The self-potential method generally allows us to differentiate a aquifer from a hydrothermal system by evaluating at the SP/elevation gradient. However, it is not always unambiguous and more information from soil CO₂ concentration and ground temperature are required to differentiate one from the other. In order to be accurate and avoid mistakes, the SP/elevation gradient must be calculated on a SP profile made along the main slope direction, which very often matches with the water flow direction.

An example is given on Fig. 6-3a for a aquifer on Masaya volcano. In the example a), a aquifer, associated with the SP anomaly #1 (presented below) flows northward in the same direction as the slope (over about 600 m along the SP profile). Along the North-South direction, the SP/elevation is -1 mV m⁻¹, which is within the normal range of SP/elevation gradients (-1 to -10 mV m⁻¹) [Lénat, 2007]. However, toward the west and east, the SP profile crosses the outside border of the aquifer, which is expressed through a very strong negative gradient (-58.8 mV m⁻¹ toward the west and -12.6 mV m⁻¹ toward the east). Theses gradients are higher than the normal range of SP/elevation gradients for aquifers and are due to the fact that the gradient does not result directly from the depth variation of water flow, but rather to the border of the horizontal extension of the aquifer.
Figure 6-3: Self-potential/elevation signature of aquifers and hydrothermal systems on Masaya volcano. The plot on the left side is the SP/elevation gradient, while the right side plot is the SP along distance on the profile. Blue lines refer to the Cross section Nindiri profile (Fig. 6-7). The green lines refer to Masaya crater profile (Fig. 6-8). The orange lines refer to North Road profile (Fig. 6-10). The purple lines refer to Comalito profile (Fig. 6-9). The gradient is calculated based on the 2008 survey. The other anomalies can be found in Appendix A5, Fig. 13-1. Note the break in the horizontal axes on plot a) and d) on the right column.
Sometimes the gradient can even be positive, which corresponds to a horizontal limit of the aquifer associated with an increase of the topography (Fig. 6-3b). When a SP profile perpendicularly crosses a slope in comparison to its main slope direction, a aquifer signal may look like a hydrothermal system, with an asymmetric gradient (aquifer on Fig. 6-3b). In this case, if no other SP profiles or other data (e.g., CO\textsubscript{2} concentration and ground temperature) are available, the SP anomaly should be considered as being a aquifer until more information is available.

In this study, the hydrothermal system is considered to be the source of the SP anomaly when the SP/elevation gradient is symmetrically positive along an approximately horizontal surface to its associated SP anomaly, or when the SP/elevation gradient is asymmetric [Lénat, 2007] along a slope. The presence of areas with no clear gradient (such as toward \(+\infty\) or a scatter of points) is considered to be due to hydrothermal activity. Hydrothermal systems should have a soil CO\textsubscript{2} concentration clearly above the atmospheric concentration (~400 ppm). CO\textsubscript{2} ground concentration between 400 and 1000 ppm are considered of magmatic origin only if the surrounding vegetation does not contain significant concentration of trees in order to avoid organic CO\textsubscript{2} generation. Extended network of tree roots can release significant amounts of CO\textsubscript{2}, which could locally increase the CO\textsubscript{2} ground concentration. Hydrothermal systems may induce, when very close to the surface, a ground temperature anomaly from a few degrees to several hundred of degrees above the atmospheric temperature. When all three anomalies (SP, CO\textsubscript{2} concentration and ground temperature) are present, it can be interpreted with confidence that a hydrothermal system is present.

Two examples of SP/elevation gradients of hydrothermal systems, with anomaly #10 and #13, 14 and 15, are presented on Figure 6-3c-d. The hydrothermal signature is expressed as an asymmetric SP/elevation gradient (positive and negative gradient value) or a very strong gradient value (Fig. 6-4c-d). Each of these anomalies is correlated with ground temperature and CO\textsubscript{2} anomalies.
6.5 Results

6.5.1 Hydrothermal system and aquifer at the Caldera scale

Over 4 years of Self-potential surveys during the dry season (in March in 2006, 2007, 2008 and 2009), the Self-potential study has allowed us to locate 22 permanent anomalies (Fig. 6-4, 6-5). Although the entire caldera could not be mapped each year, the active cones (Ninidiri on Fig. 6-6 and Fig. 6-7, Masaya on Fig. 6-8) were repeatedly surveyed. In 2007 and 2008, the survey also covered Comalito cone (Fig 6-9). The main road (North and Laguna Road, Fig. 6-10) was fully mapped in 2007, 2008 and partially in 2006 (Fig. 6-11).

In 2007 and 2008, the mapped area, including the area surrounded by loop profiles) represents about a third of the caldera and shows SP anomalies throughout. In 2007 and 2008, a ground temperature survey was made in conjunction with the Self-potential survey around Comalito and along the north flank of Masaya cone. In 2008, the temperature profile was extended on Masaya crater and Nindiri crater. Thermal anomalies were only found on Masaya cone and Comalito cone and thus for clarity, only profiles having temperature anomalies are shown in this study (Fig 6-8 and 6-9).

Fully corrected data are presented in Appendix A03. In 2008, a soil CO₂ gas concentration survey was conducted around the active cone (Fig. 6-5, 6-6, 6-8 and 6-9). The different hydrothermal systems and aquifer structures are described in order of their geographical localisation and named by its location or by the SP anomaly number associated with it.
Figure 6-4: Self-potential map of Masaya caldera in 2008. All the data are referenced to the Laguna de Masaya. Numbers indicate the different anomalies characterizing aquifers or hydrothermal systems. The dashed circle represents the inferred position of the ring fault. Sampling step is 20 m.

Figure 6-5: CO$_2$ map of Masaya caldera in 2008. Atmospheric CO$_2$ was ~350 ppm. Numbers indicate the different anomalies found by Self-potential on Figure 6-4, which characterize aquifers or hydrothermal systems. The dashed circle represents the assumed position of the ring fault. Sampling step is 20 m.
6.5.2 Nindiri cone

The different SP profiles on Nindiri cone show 4 distinct underground water structures (hydrothermal system and aquifer), which can be characterized by the SP and CO$_2$ surveys. No thermal anomaly was found along Nindiri cone. Each of these structures are expressed through a spatially well defined area, which are the anomalies #1, #2 on Fig 6-6 and Fig. 6-17 and #3 on Fig. 6-6 and Fig. 6-8.

- The upper north Nindiri slope aquifer (anomaly #1, Fig 6-6, 6-7) is localized over an area, ~800 m northward in length (Fig. 6-7) and ~ 300 m wide (east-west, Fig. 6-6) on the north slope of Nindiri cone. This aquifer is characterized by an inverse SP/elevation gradient of -0.3 mV m$^{-1}$ northward and -1.6 mV m$^{-1}$ eastward (Fig. 13-1a). A SP amplitude of 40 mV (E-W) and ~140 mV (N-S) (Fig 6-6, 6-7) was measured. No ground temperature or CO$_2$ anomalies were found.

- The Nindiri hydrothermal system, SP anomaly #2, spreads throughout Nindiri crater and extends southward through the cone and the caldera floor (Fig. 6-4, 6-5, 6-6 and 6-7). The hydrothermal system of Nindiri is characterized by a strong and persistent SP amplitude (~ 200 mV, Fig. 6-5 and 6-6) and on each side of the San Pedro fault (Ring fault) by anomalous CO$_2$ concentrations (between 1000 to 2000 ppm, Fig. 6-5).

- A second peak of CO$_2$ is found near the old parking lot on the south side of the active Santiago crater. No CO$_2$ anomaly was found along the South slope of the cone and no CO$_2$ gas concentration data are available for the inside part of Nindiri crater. Within the crater, on the south slope of the cone and the caldera floor at its bottom, the electrical signal shows a high variation of the
SP/elevation gradient, from no correlation to -3.7 mV m\(^{-1}\) (Fig. 13-1b), along a north-south and east-west section.

The gradient is the strongest on the north side of Nindiri crater, which marks the northern border of the hydrothermal system and coincides with the subsidence ring fault of the lava lake. On a wide area from the lava lake toward the caldera floor in the south, the hydrothermal system is stable in intensity, almost independent of the topographic variation and is supported by a constant SP/elevation gradient of only ~+0.2 mV m\(^{-1}\) (Fig. 13-1b). While the hydrothermal border is clear on the north side of Nindiri crater, it was not detected on the south by this study.

The west and east border of the hydrothermal system of Nindiri seems to match with structural limit of the ring fault cutting the cinder cone. On the west side (Fig. 6-11), the SP anomaly #2 (associated to the hydrothermal system) breaks at the San Pedro fault. Along the west section of the SP anomaly (#2, Fig. 13-1b) there is no clear SP/elevation gradient; however, on the east side, the gradient is ~+1.4 mV m\(^{-1}\). SP amplitude is ~320 mV North-south and ~240 mV East-west (Fig. 6-6, 6-8). On the east side of the cone, the SP anomaly #2 stops when the south slope of Nindiri cone meets the south slope of Masaya cone (Fig. 6-5), which marks a structural limit between the two cones. In 2009, on this side of Nindiri cone, the SP anomaly #2 merged with anomaly #3, suggesting that the source responsible for the electrical generation may have grown in intensity or volume.

- The junction hydrothermal system (anomaly #3, Fig. 6-6 and 6-8) is at the junction between the east side of Nindiri and the west side of Masaya cone and extends northward on the upper slope of Masaya cone (#3, Fig. 6-6). The hydrothermal system shows high concentrations in ground CO\(_2\) (up to ~11,000 ppm, #3 on Fig. 6-8), however, further north on the old crater rim of Masaya cone, the CO\(_2\) concentration drops to ~450 ppm (#3 on Fig. 6-8).
Figure 6-6: Self-potential, CO₂ and topography of the survey profile around Nindiri crater between 2006 and 2009. All the Self-potential profiles are referenced to the same base station on the north rim of Nindiri, which is assumed to be stable over time and to which a 0 mV value has been attributed to facilitate the comparison of the different surveys. The purple numbers are the names of the anomalies. Sampling step is 20 m.
Figure 6-7: Self-potential and topography of the survey profile across Nindiri cone in 2007 and 2008. All the Self-potential profiles are referenced to the Laguna de Masaya, which is assumed to be stable over time and to which a 0 mV value has been attributed to facilitate the comparison of the different surveys. The purple numbers are the names of the anomalies. Sampling step is 20 m.
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**Table 6-1**: SP/elevation gradient of the different SP anomalies found in 2008. North, south, east, west and center are the local direction in comparison at the center of each of the anomalies. In blue are the negative values, in orange the positive values and in purple the strong negative gradient. Nan means that no SP data exists in this direction for the associated SP anomaly. Undefined means that the SP vs. elevation data shows no significant trend, which could be translated in SP/elevation gradient.
The hydrothermal system is characterized by a stronger SP/elevation gradient from north-west to south-east (-4.3 mV m\(^{-1}\) on North-east of Nindiri crater on Fig. 6-7 and +4.5 mV m\(^{-1}\) on west of Masaya crater on Fig. 13-1c) than in its perpendicular direction (South-west to north-east) with -1.3 mV m\(^{-1}\) on south-east of Nindiri crater (Fig. 13-1c) and -1.8 mV m\(^{-1}\) on the north-west side of Masaya crater. In addition, the south section on Masaya crater is very similar to the north-west section (north-east part of Nindiri part of Masaya crater) of the hydrothermal area (-1.3 mV m\(^{-1}\) on Fig. 13-1c). The hydrothermal system is characterized by a SP amplitude of ~100 to ~150 mV (Fig. 6-6 and 6-8). No thermal anomaly was found along the different SP profiles for this hydrothermal system. The junction hydrothermal system is within a well restricted area at the junction between the two cinder cone and the presence of significant CO\(_2\) concentration associated with strong SP/elevation gradient suggests a zone of higher fracturing, perhaps due to the ring fault.

6.5.3 Masaya cone

- The **north-east Masaya hydrothermal system of Masaya crater**, anomaly #4, is spread along ~320 m of the profile survey (Fig. 6-8). It is characterized by an electrical signature of ~80 mV amplitude with a strong SP/elevation gradient of -2.3 mV m\(^{-1}\) on the north side and +11.4 mV m\(^{-1}\) on the north-east side of the crater (Fig. 13-1d). The SP amplitude can reach up to ~100 mV (Fig. 6-8). The hydrothermal system is characterized by a low soil CO\(_2\) concentration ranging from 300 to 1200 ppm (Fig. 6-8).

- The **south-east Masaya Hydrothermal system** (SP anomalies #5 and #6) is characterized by a large electrical anomaly, which in 2006 and 2007 consisted of two small anomalies. Since 2008, theses anomalies have merged together within a large electrical structure (Fig. 6-8). The hydrothermal system is ~460 m across and does not show any temperature anomalies, however, a low
CO₂ anomaly was found in 2008 (between 360 and 960 ppm, Fig. 6-8). The electrical structure consists of an asymmetric SP/elevation gradient, of +6 mV m⁻¹ on the east rim of the crater and -7.3 mV m⁻¹ (Fig. 13-1e) on the south-rim of Masaya crater. The SP amplitude is from ~50 to 150 mV (Fig. 6-8).

- The San Juan Hydrothermal system (SP anomaly #7) is ~260 m across the south rim of Masaya cone, within the San Juan vent (Fig. 6-8) and is characterized by a strong SP amplitude (~180 mV), which is very constant every year. The SP/elevation gradient is asymmetric (Fig. 13-1f) with a value of +1.4 mV m⁻¹ on the east rim, while it reaches -9.9 mV m⁻¹ on the south-east part of the crater rim. A ground temperature survey did not show any anomaly, however, CO₂ surveys showed sign of abnormal concentrations ranging from 350 to 900 ppm. The west border of the hydrothermal system (within the south-east part of the crater rim ) is very well delimited by a strong negative gradient (-9.9 mV) and a drop of ~180 mV, which suggest that the hydrothermal system may be confined within the San Juan vent (#7 on Fig. 6-8).

- The south-west Masaya aquifer (SP anomaly #8) is within a small topographic depression (anomaly #8 on Fig. 6-8) and is characterized by a positive SP curve (~80 mV of amplitude). The SP/elevation gradient is associated with a symmetrical negative gradient between -1.7 and -1.9 mV m⁻¹ (Fig. 13-1g). No CO₂ concentration or ground temperature anomaly were found in this area.

- The Masaya North slope hydrothermal system of Masaya cone (SP anomaly #10, 11 and 12) is well developed (Fig. 6-9). Extending over more than 1100 m northward from the rim of the old infilled Masaya crater, the hydrothermal system has a strong SP anomaly amplitude (max at ~300 mV), with ground temperature above 50 °C and average soil CO₂ concentration of ~10,200 ppm (with a peak at ~20,000 ppm). The hydrothermal system shows a strong
gradient evolving from negative near the summit (-9.3 m mV⁻¹, Fig 6-3c, Fig. 13-1h) toward a strong positive gradient (+11.5 mV m⁻¹, Fig. 13-1i) over the area of strongest SP, CO₂ and temperature anomalies.

Further down on the slope, the positive gradient decreases northward until the lower part of the slope (+4.9 and 2.2 mV m⁻¹). At the same time, the temperature decreases below atmospheric and CO₂ concentration drops to only a few 1,000 ppm (peak at ~6,200 ppm, Fig. 6-9). On the lower part of the cone, the ground temperature (30 to 40°C) is slightly above atmospheric temperature (~32 °C) and CO₂ concentration does not exceed 2,000 ppm. The SP/elevation gradient is asymmetric, with a positive gradient (+4.9 mV m⁻¹) on its higher part and negative (-2.7 mV m⁻¹) in its lower part (Fig. 13-1j). The center of this large hydrothermal system is located below the upper part of the slope and its intensity decreases towards the north.

6.5.4 Comalito cone

Comalito cone hosts an extended hydrothermal system, the Comalito hydrothermal system, characterized by the anomalies # 13, 14 and 15 (Fig. 6-9 and 6-10). The hydrothermal system has a strong SP anomaly amplitude (~250 mV) and strong ground temperature ~45 °C (a max at 75 °C in 2007 and 61 °C in 2008, Fig 6-9). However, the CO₂ degassing seems to be restricted mainly to the north-east area of the cone (average ~4,200 ppm and max at ~21,200 ppm). The hydrothermal system shows a broad variety of strong SP/elevation gradients (North : -7.2 mV m⁻¹, South: -56.9 mV m⁻¹, East: +38 mV m⁻¹, West: +20.7 mV m⁻¹, center of the area: -10.5 mV m⁻¹, Fig. 13-1k).
Figure 6-8: Self-potential, CO₂, temperature and topography of the survey profile around Masaya crater between 2006 and 2009. All the Self-potential profiles are referenced to the same base station on the north rim of the crater of Masaya, which is assumed to be stable over time and to which a 0 mV value has been attributed to facilitate the comparison of the different surveys. The purple numbers are the names of the anomalies. Sampling step is 20 m. The star indicates the beginning and end of the profile.
Figure 6-9: Self-potential, CO₂, temperature and topography of the survey profile across the north flank of Masaya cone until the solfatara of Comalito in 2007 and 2008. All the Self-potential profiles are referenced to the same base station at the Laguna de Masaya, which is assumed to be stable over time and to which a 0 mV value has been attributed to facilitate the comparison of the different surveys. The purple numbers are the names of the anomalies. Sampling step is 20 m.
6.5.5 Caldera floor

The SP mapping on the caldera floor currently covers the north road, the Laguna road (Fig. 6-4) and part of the forest near the east slope of Masaya cone.

- The **east-side North-road aquifer** (SP anomaly # 16) is crossed by the North road (Fig.6-10) over ~300 m and it is characterised by a SP anomaly of ~120 mV in amplitude. This aquifer is located near the gravity station A1 [Rymer, 1998; Williams-Jones et al., 2003] at the intersection between North Road and the trail leading to Comalito cone. Due to low elevation variations, the SP/elevation gradient is very strong and negative toward the 3 directions where the profile goes north, south and east (-31.7 to -35.6 mV m\(^{-1}\), Fig. 13-1l). These strong gradients are very probably due to the sharp limit of the water table (within water saturated volcanic deposit), which is in a medium probably more porous and permeable than unsaturated volcanic deposit surrounding it. Thus, with only this profile, it is not possible to clearly describe the direction of underground water flow.

- The **1772 lava aquifer** (SP anomaly #17) is ~660 m long and is localized across the 1772 lava (Aa) flow. Its SP amplitude is ~100 mV high and the topographic elevation changes only slightly (less than 20 m), which is responsible for a very strong SP/elevation gradient. The gradient seems to decrease from the north (+23.9 mV m\(^{-1}\)) toward the south (-13.1 mV m\(^{-1}\)) and between them, an intermediate gradient at 21.8 mV m\(^{-1}\) (Fig. 13-1m). For now, due to a lack of CO\(_2\) data, the aquifer #17 cannot be considered as being a hydrothermal system, however, with so strong positive gradient, it may be one.
• The **North-Nindiri lower-slope aquifer** (SP anomaly #9) is characterized by a SP amplitude of ~80 m along the direction of the steepest slope (toward north) and ~150 mV perpendicular to it (Fig.6-7, 6-10). Similarly, the SP/elevation gradient is strongest perpendicular to the slope direction, with a very strong negative gradient (-58.8 mV m\(^{-1}\) toward West and -12.6 mV m\(^{-1}\) toward East, Fig. 6-3, 13-10). Along the slope direction, the SP/elevation gradient is of -1 mV m\(^{-1}\) (Fig. 6-3, 13-10).

• Along Laguna road, 5 large SP anomalies were found each year at the same place and with the same amplitude. However, the SP profiles are not corrected for drift from the Laguna de Masaya until Comalito (in 2007 and 2008, Fig.6-10) or to the east summit of Masaya cone through the forest (in 2006, Fig.6-11), because they are not closed in a loop or to a second natural surface water. Thus, no accurate gradient can be calculated from these profiles. The anomalies along the Laguna road, labeled # 18, 19, 20, 21, 22 (Fig.6-10) and #23 (Fig.6-11), may be a succession of aquifers flowing toward the lake, well delimited hydrothermal systems, or a combination of both. The aquifer seems to be the most logical choice, due to the presence of the large lake and the significant variation of elevation between Comalito (~300 m a.s.l) and the lake at ~120 m a.s.l.
Figure 6-10: Self-potential survey along the North Road and the Laguna Road in 2007 and 2008. All the Self-potential profiles are referenced at the same base station to the Laguna de Masaya, which is assumed to be stable over time and to which a 0 mV value has been attributed to facilitate the comparison of the different surveys. SP profiles are not corrected of drift. Sampling step is 20 m. The purple numbers are the names of the anomalies.
Figure 6-11: Self-potential survey along the Laguna Road, through the forest until the east rim of Masaya crater in 2006. All the Self-potential profiles are referenced to the same base station at the Laguna de Masaya, which is assumed to be stable over time and to which a 0 mV value has been attributed to facilitate the comparison of the different surveys. SP profiles are not corrected of drift. Sampling step is 20 m. The purple numbers are the names of the anomalies.
Figure 6-12: Water depth between 2006 and 2009 on Nindiri cone. Horizontal GPS coordinate and depth of the water are obtained by Multi-scale wavelet tomography applied on the Self-potential survey loop. Triangles represent depths calculated from the MWT analysis of the Nindiri survey loop (Fig. 6-6). Squares and diamonds are depths calculated from the MWT analysis of the Nindiri cross-section SP profile (Fig. 6-7). All depths are in Table 6-2.
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**Table 6-2:** Water depth calculated from the Multi-scale wavelet tomography on Self-potential data for Masaya volcano. SP anomaly number is expressed by the # following the number of the anomaly, N_C is for Nindiri crater profile, N_Cross is for Nindiri cross-section profile, M_C is for Masaya crater profile, Com_P is for Comalito profile and North_R is for North road profile. Nbr of wavelet corresponds to the number of wavelets which found the depth at this position (X, Z). σ is the standard deviation calculated from the raw data (not shown here) and represents the scatter of values. Type column: W means there is a aquifer associated to the depth value; H means, that the depth is associated to a hydrothermal system.

### 6.6 Dynamic of the hydrothermal system

Signal analysis of the Self-potential data by Multi-scale wavelet tomography has allowed us to obtain a large number of depth values. From the raw results, a total of 531 depths have been calculated. A majority of these depths are quiet similar for a similar horizontal GPS coordinate. In order to avoid artifacts among the depths, only the calculated depth found with at least 3 of the 4 analyzing
wavelets are considered significant and thus are presented here to describe the dynamism of the water and hydrothermal system within Masaya volcano. Among the 531 depths obtained using the 4 analyzing wavelets over the 4 years of survey (from all the profiles), it has been possible to regroup them in 173 individual depths found with at least 3 of the 4 analyzing wavelets (all the individual depths are presented in Appendix A5 in tables 13-10). By comparing the horizontal GPS coordinate associated to each depth, it has been possible to group them in 25 time-series data sets, which are well localized in space (GPS, depth) and time (from 2006 to 2009). Thus, it becomes possible to characterize the water depth evolution over the period from 2006 to 2009 (Fig. 6-7, 6-8, 6-9, 6-10). Each depth has an error bar (standard deviation) which represents the scattering of all the raw calculated depths associated to this depth.

The scattering value associated with each depth allows us to estimate how much confidence it is possible to have on the depth. The mean value of the error bar for the 173 depths is 41 m (min value of 29 m and max of 56 m). This average is twice the average (the scattering) of depth from the individual wavelet analysis, which is 18 m and so will increase the difficulty to characterize depth variation over time.

6.6.1 Nindiri cone

- The upper north Nindiri slope aquifer (SP anomaly #1) is localized at a depth of ~ -134 m on the north part of Nindiri crater. It's elevation along the summit north rim is ~406 m a.s.l. (Fig. 6-12a,b) and decreases toward the North, following the topographic slope, to reach ~360 m a.s.l. (depth of ~ -100 m, Fig. 5 in Chapter 5). Between 2006 and 2009, the aquifer depth shows no variation within the scattering of the data (Fig.6-12a,b and Fig. 5 on Chapter 5).
The Nindiri hydrothermal system (SP anomaly #2), which is spread through the South of Nindiri cone, has been localized in four different areas, which can be regrouped into two zones (Fig.6-12c,d). The first zone corresponds to the Nindiri cinder cone (San Pedro Fault, Fig.6-12c; inside the south crater rim, diamonds on Fig.6-12d; the outer cone slope, triangle on Fig.6-12c), where the hydrothermal system is at an approximately constant elevation of 440 m a.s.l. Due to large surface topography variation, the equivalent hydrothermal depth ranges from 50 m to 200 m below the topographic surface. As the hydrothermal system extends toward the south on the caldera floor (SP anomaly #2, Fig. 6-6), its elevation decreases to ~ 350 m a.s.l. During 2007 and 2008, within the bottom of the cone, the calculated depths indicate a rise of the hydrothermal system from 300 to 280 m a.s.l. (square on Fig.6-12c). However, this change is not confirmed by the other part of Nindiri hydrothermal system. Thus, this study cannot conclude on a change affecting the Nindiri hydrothermal system.

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<td>7 27</td>
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<tr>
<td>Between 50 and 100 m</td>
<td>10 38</td>
</tr>
<tr>
<td>Between 100 and 150 m</td>
<td>5 19</td>
</tr>
<tr>
<td>Between 150 and 200 m</td>
<td>3 12</td>
</tr>
<tr>
<td>deeper than 100 m</td>
<td>1  4</td>
</tr>
</tbody>
</table>

**Table 6-3:** Number of depths for both water and hydrothermal systems based on the result of Table 6-2. Statistics of the calculated depth are from the Multi-scale wavelet tomography on Self-potential data for Masaya volcano.
• The **junction hydrothermal system** (SP anomaly #3) has depths on three different areas at an average elevation of ~470 m a.s.l. (Fig. 6-12e, 6-13a, b). The south part of the hydrothermal system is located on the east rim of Nindiri crater, while the two others are on Masaya cone (Fig. 6-6a, b). They seem to show a trend oriented South-west to North-east, but that is very likely only due to the way that the SP profiles are organized on the summit area. Within the scattering of the data, there is no clear variation of the depth of the hydrothermal system.

### 6.6.2 Masaya cone

• The **North-east Hydrothermal system of Masaya crater** (SP anomaly #4) has depths on the east side of Masaya crater at ~470 m a.s.l. (Fig. 6-13f). Between 2006 and 2009, the elevation of the hydrothermal system is constant, which indicates that the hydrothermal system is stable over time.

• The **South-east Hydrothermal system** (SP anomaly #6) of Masaya cone is at ~480 m a.s.l. (Fig. 6-13e) and has no significant variation within the scatter of the data (depth values range between 90 m to 170 m below the topographic surface).

• The **San Juan Hydrothermal system** (SP anomaly #7) is found within the crater of San Juan vent at a depth of 450 m a.s.l. (Fig. 6-13d). The hydrothermal system seems to show some variation in elevation over the years. The variation indicates a change of ~70 m from year to year between 2006 and 2008, with a rise in 2007 (~500 m a.s.l.) before returning to its previous level in 2008 (~430 m a.s.l.) and since is stable.
Figure 6-13: Water depths between 2006 and 2009 around Masaya crater. Horizontal GPS coordinates and depths of the water are obtained by Multi-scale wavelet tomography applied to the Self-potential survey loop. Triangles represent depths calculated from the MWT analysis of the Masaya survey loop (Fig. 6-8). All depths are shown in Table 6-2.
However, because no sign of depth variation can be found on the nearby south-west hydrothermal system, it is difficult to be certain that the San Juan hydrothermal system really moved toward the surface in 2007.

- The **South-west aquifer** (SP anomaly #8) was not located by the MWT analysis of the SP survey profiles used in this study because not enough MWT were able to characterize it (only depth found with at least 3 of the 4 wavelets are considered to be significant).

- The **Masaya North slope hydrothermal system** of Masaya cone (SP anomaly #10, 11 and 12) is present at very shallow depths below the topography surface and was localized at 3 different places along the slope (Fig. 6-14e, f, g). The depth remains constant along the slope at ~60 m below the surface, while the elevation decreases following the topographic slope from 530 to 330 m a.s.l. The scattering of the data indicate that the hydrothermal is probably even closer to the surface, but the MWT analysis is not sufficiently accurate to localize very shallow depths (within the first 10 m). The stability of the shallow depth along the slope and over time (2007 and 2008) is coherent with the strong CO$_2$ concentration (up to 20,000 ppm) and high ground temperature (Max 62°C) found during the 2007 and 2008 surveys (Fig. 6-9, Fig. 6-14e, f, g).

### 6.6.3 Comalito cone

The Comalito hydrothermal system is depth located on 5 different positions along the West side of Comalito cone with an average depth of ~260 m a.s.l ($\sigma = 25$ m, Fig. 6-14a, d, c, Fig. 6-15f, g) and ~250 m ($\sigma = 19$ m, Fig. 6-15f, g). Over 2007 and 2008, there is no significant change, even with the small scattering of the data ($\sigma$ is ~24 m). The homogeneous depth seems to indicate that the hydrothermal system activity is quite similar, while the SP, CO$_2$ and
ground temperature surveys show significant variations along the west side of the cone.

Thus, one possibility is that the MWT analysis of SP profiles cannot be accurate enough to detect the shallowest fluid body, or perhaps the main hydrothermal body is at ~260 m a.s.l. (~70 m deep below the topographic surface) and the gas can generate fumaroles of low temperature (Max ~75°C, Fig. 6-9) on sporadic spot, where the ground permeability is higher on small area.

6.6.4 Caldera floor

- The **east-side North-road aquifer** (SP anomaly #16) is located in two different spots at an elevation of 260 m a.s.l. (~45 m deep, Fig. 6-15b, c). Within the scattering of the data, the depth of the water table does not change.

- The **1772 lava aquifer** (SP anomaly #17) is depth located on both sides (east and west) of the 1772 lava flow, however, only the west side was found with 4 wavelet analysis (Fig. 6-15a and east side as only 2 wavelet analysis, Table 13-10). The water is at ~240 m a.s.l. ($\sigma$ is ~55 m at a depth of ~65 m, Fig. 6-20a).

- The **North-Nindiri lower-slope aquifer** (SP anomaly #9) is found in 3 different spots (Fig. 6-15d, e) at ~280 m a.s.l. ($\sigma$ is ~50 m at a depth of ~62 m, Fig. 6-15d, e). The aquifer seems to be near a horizontal plane, while the topography decreases northward, thus the aquifer seems to not depend of the shallowest bed rock structure, which is a lava flow (Aa). In addition, there is no change of the depth of the water between 2007 and 2008.
Figure 6-14: Water depths between 2007 and 2008 across the north flank of Masaya cone until the solfatara of Comalito cone. Horizontal GPS coordinates and depths of the water are obtained by Multi-scale wavelet tomography applied on the Self-potential survey loop. Triangles represent depths calculated from the MWT analysis of the Comalito profile (Fig. 6-9). All depths are in Table 6-2.
Figure 6-15: Water depth between 2007 and 2008 along North road. Horizontal GPS coordinate and depths of the water are obtained by Multi-scale wavelet tomography applied on the Self-potential survey loop. Triangles represent depths calculated from the MWT analysis of the North road (Fig. 6-10). Squares and diamonds are depths calculated from the MWT analysis of the Nindiri cross-section SP profile (Fig. 6-7). All depths are in Table 6-2.
6.7 Discussion and Conclusion

Masaya volcano hosts a hydrothermal system complex, at least along the main volcanic cinder cones (Nindiri, Masaya and Comalito cones), which is controlled by the structural limits, namely the ring fault and the crater boundaries (Fig. 6-16). Each hydrothermal system (based on the Self-potential mapping) covers an area from ~100 m to 1 km along the SP profile. Unfortunately, due to limited SP coverage, the hydrothermal systems are often only spatially well constrained in one direction (generally parallel to the ring fault). Only a few SP profiles run perpendicular to the ring fault structure (SP anomaly #2 on the San Pedro Fault and #3 along the east side of Nindiri crater on Fig. 6-6; SP anomaly #2 on Nindiri cross section on Fig. 6-7; SP # 3 on the west side of Masaya cone on Fig. 6-8). Thus, the south boundary of the South Nindiri hydrothermal system (SP #2, Fig. 6-6 and 6-7) is not reached by this study. The same issue occurs with the east boundary of both North-east Masaya and South-east Masaya hydrothermal system (SP #4, 5, 6, Fig. 6-8), which may be extend through the entire old infilled Masaya crater. The only indication of the east rim of Masaya cone is from the Forest profile in 2006 (SP #5 and #23, Fig. 6-11), with SP anomalies #5 and #23; however, this SP profile (Fig. 6-11) is not drift corrected (it was not closed with a second control point) and no definitive interpretation can be properly done from it. The main hydrothermal systems principally cover the inside parts of each cone and more often their external side along the ring fault (Fig. 6-16) rather than their side within the inner ring fault (#4, #10, #11 and #12 Fig. 6-8 and 6-9). The strongest hydrothermal systems are located over the ring fault on the north flank of Masaya cone and within Nindiri cone (#2, #3, south and center area, Fig. 6-6, 6-7).

The inner ring fault, at least in its south section, hosts waters tables (SP#1, #9 North Nindiri aquifers on Fig. 6-7 and # 17 the 1772-lava-flows aquifer on Fig. 6-10), which are not spatially well delimited by this study for the same reason as mentioned above. However, the water table, at least for the North Nindiri aquifer is flowing northward, following the topography decrease.
Over the 4 years of surveys (March 2006 to March 2009), the 82 water depths, characterizing either hydrothermal systems or aquifers, were found to be at relative constant depths (Table 6-2) within the scattering of the data (the mean $\sigma$ is $\sim 41$ m, calculated for all the 173 depth, Table 13-6 to Table 13-10). On one hand, a stable depth may indicate a steady state within both the hydrothermal and the aquifer over the 4 years, which seems to correlate well with the steady state of Masaya volcanic activity. No major eruption or large change on its surface occurred over the same period of time.

On another hand, the relative constant water depth is mainly due to the scattering of the data. The average of the several water depth values present significant variation over time (plot #a, b and c on Fig. 6-12; plot #a, b, d and e on Fig. 6-13; plot #a on Fig. 6-14 and plot #a, b, c and g on Fig. 6-15). Unfortunately, scattering ($\sigma$) of the data range from minimum value of 29 m to a maximum of 56 m, thus it is only possible to characterize depth changes greater than 60 (twice the minimum value of $\sigma$), and only depth variation greater than 120 m (twice the maximum $\sigma$) can be considered. Some large scattering on the calculated depth may be related to large underground resistivity contrasts, which are not associated with the underground water and thus noise the component of signal generated by the water flow only.

Thus, based on both the variation of the water depth and the large scattering of some of the data, it seems that both the hydrothermal and the aquifer are stable over the period from 2006 to 2009. Although the water depth shows no significant variation, this information can be used as approximation to look at the spatial distribution of depth over different ranges, such as water depth less than 50 m deep, between 50 and 100 m deep and so on below the topographic surface.
Figure 6-16: Spatial localization of the hydrothermal system complex and aquifers found within Masaya caldera and for which depths have been obtained. Blue squares are the depth positions of aquifers. Orange diamonds are the depth positions of hydrothermal systems. Depth values can be found in Table 6-2 and Fig. 6-12 to 15. Green dashed lines represent the last main lava flows. Red dashed curves represent visible fissure vents. Black dashed circles are the limit of actual and previous craters on Masaya and Nindiri cone. Brown lines indicate areas having high levels of ground CO₂ from a previous study [St-Amand, 1999].
Over 4 years, this study had regrouped depth values within 25 time series data sets, which include 82 depth values. From the 82 depth values (Table 6-3), cross-correlated with the SP, CO₂ and temperature survey, 26 depth values belong to aquifers and 56 to hydrothermal systems (Fig. 6-16). From a statistical point of view, the majority of the depth values for both water types are between 50 to 100 m deep (38% of the water-table depth values and 46% of the hydrothermal-system depth values). Approximately 60% of the water depth values (for both) are less than 100 m deep and ~82 of the depth values (for both) are less than 150 m deep (Table 6-3). Thus, both aquifer and hydrothermal systems, within the study area on Masaya volcano could be considered to be at relative shallow depths, when topographic effect is not corrected.

Similarly, when corrected for topographic variation, the elevation in m a.s.l. of the underground water depth shows good correlation within the same hydrothermal system or aquifer. A good example is the South-Nindiri hydrothermal system (Fig. 6-12c and d), which is found on the south and east side of the cone at a depth of ~430 +/- 40 m a.s.l.. In addition, depth values obtained from different profiles crossing a aquifer or hydrothermal system give similar depths (aquifer on North Nindiri cone, SP #9 and #1 Fig. 6-12b and 6-15d and e). Thus, based on depth comparison between different profiles and different SP anomalies characterizing a same water structure, the MWT shows a relatively good capacity to give reproducible and coherent depth values on both aquifer and hydrothermal systems. The approach of MWT using a combination of wavelet seems to works quite well to locate the main water structure and to avoid artifacts, as shown before in comparison studies between depth determination by MWT on SP and traditional geophysical models (e.g., SP Nindiri-cross-section profile in Chapter 05, Fig. 5) [MacNeil et al., 2007]. The limitation of the method is the depth accuracy, which, for this study, has an average scattering of 41 m and strongly restricts the capacity to detect water depth change over time. A better constraint could be obtained by increasing the gridding of the Self-potential mapping, which allows to obtain parallel SP profiles and thus to increase the capacity to locate each of the water structures.
Between, 2006 and 2009, within Masaya volcano, neither the hydrothermal systems nor the aquifers showed any sign of significant depth variations (Fig. 6-12 to 6-15). The hydrothermal systems spread all along the south border of the ring fault (area covered by this study, Fig. 6-14), while the inner ring fault hosts aquifers. Several strong Self-potential anomalies were found along the Laguna road, however, this study cannot currently determine their origin. This study has shown that significant amounts of CO$_2$ (up to 20%) are present along the ring fault (Fig 6-16 and Fig. 6-5) and correlate well with previous unpublished studies [St-Amand, 1999], which found similar CO$_2$ anomaly on Arenal cone, on the north side of the ring fault. Thus, a more extensive Self-potential, soil CO$_2$ and ground temperature mapping over the entire ring fault should be made in the future in order to better understand the distribution of and extent of the hydrothermal system complex. While the majority of degassing occurs through the open conduit of Santiago crater, the presence of large amount of CO$_2$ at distances from the active crater, large SP anomalies and ground temperature anomalies, are proof that underground magmatic activity is spread at shallow depth far outside Nindiri cone through the caldera floors.
6.8 References


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7 CHAPTER 07: WATER FLOW STRUCTURES AND DYNAMICS OF THE PERSISTENTLY ACTIVE KAWAH IJEN VOLCANO (JAVA, INDONESIA)

Figure 7-1: Kawah Ijen crater and its acid lake in 2007.
7.1 Abstract

Kawah Ijen volcano is a persistently active volcano characterized by strong continuously passive degassing since more than 150 years, by a large hyper acid lake, and by an intense sulphur dome. While Kawah Ijen shows all the signs of a well established hydrothermal system, this study shows that the hydrothermal system is self-sealed within the upper volcanic edifice and releases its pressurized gas only through the active crater. The upper part of the volcanic edifice hosts an extensive hydrological system consisting of three main aquifers: the East aquifer; the West aquifer and the South aquifer. The South aquifer appears to be isolated from the acid lake and the hydrothermal system. In contrast, the East aquifer is very likely the most important source of fresh water recharge to the acid lake. The acid lake itself discharges via seepage to the West aquifer, which passes through the upper west flank of the volcanic edifice. The West aquifer also sustains the acid water spring feeding the Banyupahit river. This insight on the hydrogeological structures has been gained by combining Self-potential, ground temperature and dynamic gravity surveys. Field surveys were made every year between 2006 and 2008 during the dry season. Processing of the Self-potential profiles by Multi-scale wavelet tomography algorithms and forward gravity models were used to investigate the vertical displacement of the aquifer. From 2006 to 2008 the aquifers were affected by large vertical variations ranging from 10 to several 10’s of meters, with the deepest level in 2007. During the 2006/2007 monsoon, the South-East Asia has been affected by an El Niño event, which was responsible for a large drought in Java island. In 2007, the origin of the large decrease of the aquifer in Kawah Ijen volcano is very likely due to seasonal effects caused by the 2006/2007 El Niño. While the East aquifer feeding the acid lake is affected by the large decrease of water from year to year, the acid lake acts as a buffer to support the West aquifer and its related acid spring. Dynamic gravity surveys combined with Self-potential surveys processed by Multi-scale wavelet tomography allow for new insight on the dynamic of an active volcano, such as Kawah Ijen. Understanding of non-
volcanic phenomena is of significant importance in the investigation of volcanic change.

7.2 Introduction

Some volcanoes are characterized by persistent volcanic activity over several hundreds of years. As on many volcanoes, magma does not always reach the surface, thus it is necessary to investigate their subsurface structures indirectly. Kawah Ijen volcano, in Eastern Java, Indonesia, is one of these persistently active volcanoes, characterized by strong degassing and the world’s largest volcanic acid lake. Previous studies of Kawah Ijen have focused on the chemical composition of the acid lake, sulphur gases, rock composition, the acid spring and acid river, with the objective of characterizing its activity and its impact on local populations [section 2.3]. However, to date, no studies exist of the subsurface potential field changes, such as the natural electric or gravity change over time. This study focuses on geophysical variations, which are due to subsurface change. On one hand, the change in underground water depth (hydrothermal and aquifer) will generate variations of the local electrical current [Chapter 2], which may be linked to shallow magmatic change. On the other hand, gravity variations may be related to mass, volume or density change [Chapter 3] in the first few kilometres of the upper volcanic edifice. Thus, gravity changes can be due to shallow variations of water bodies, or magmatic phenomena, such as magmatic intrusion or drainage, or change in density during magmatic differentiation. Underground water depth can be determined by applying signal processing such as Multi-scale wavelet tomography (MWT) to Self-potential data [Chapter 4] or by 3D gravity modeling. In this study, shallow structures are considered to be within the first kilometre below the volcano summit. This study investigates, during the period of 2006-2009, the change occurring in the aquifer, hydrothermal system and shallow magmatic reservoir
through a combination of geophysical surveys, such as dynamic gravity, and self-potential. These geophysical methods are complemented by multi-scale wavelet tomography and 3D forward modeling in order to characterize the dynamic changes occurring on Kawah Ijen volcano. Across Java, the 2006/2007 monsoon was affected by El Niño, which caused a significant drought over the region [McPhaden, 2008; Luo et al., 2008]. How are the hydrogeological and hydrothermal structures related to each other and the acid lake? How do changes in these shallow water structures relate to both magmatic and climatic variations?

7.3 Hydrothermal systems

Hydrothermal systems are complex structures that can vary over a wide range of physical and chemical properties [Osinski et al., 2001; Stimac et al., 2003; Finizola et al., 2002]. A hydrothermal fluid is a fluid consisting of a single liquid phase or a combination of liquid and gas phases. A hydrothermal system is a natural system made of at least one convective cell of hydrothermal fluids and, in the context of this study, an expression of underground magmatic activity.

Hydrothermal fluids are composed of a wide range of chemical elements, which can vary significantly in concentration [Osinski et al., 2001; Stimac et al., 2003; Chiodini et al., 2005]. The most important chemical components are, in decreasing order, H₂O, CO₂, S_{composite}, CH₄, H₂, Ar, O₂, N₂, H₂, HCl and HF. The water is by far the largest component and can be magmatic or meteoric in origin.
Hydrothermal systems are fluids moving through a rock structure, thus the extension and hydrothermal flow paths are strongly controlled by the rock porosity and permeability of the area [Osinski et al., 2001; Finizola et al., 2002; Yasukawa et al., 2003, 2005]. On volcanoes, permeability and porosity can be highly variable in space and time. Permeability will be controlled by bedrock structures, volcanic structures and by faults and fractures. Rock alteration and deposition of hydrothermal minerals will change the porosity and permeability of the volcanic edifice [Delmelle et al., 1994; Osinski et al., 2001]. Hydrothermal systems are very different from one another; they can be relatively cold or
extremely hot, with wide ranges of heat flux (0.5 and 2.3 W m$^{-1}$ °C$^{-1}$) and surface temperature (~20 to 30 °C, but can reach > 600 °C). Hydrothermal systems can be enriched or depleted in chemical elements and concentration. They can be of low (~0.00) to high (~14) pH and can vary in size (from ~ km$^2$ to several km$^2$) and intensity of the activity. Hydrothermal systems are supplied in water by both the water vapour escaping from the magma and meteoric water. Large changes in the amount of water from each will have a direct impact on the hydrothermal activity.

The life time of a hydrothermal system depends on the underlying magmatic activity, allowing for prolonged activity over several tens of thousands of years [Osinski et al., 2001]. Finally, on active volcanoes, hydrothermal systems can consist of several levels of convection cells in open or closed systems [Osinski et al., 2001; Finizola et al., 2002]. More insight on hydrothermal systems can be found in the Chapter 1, section 1.3 and in Appendix 2.

7.4 Kawah Ijen volcano

Kawah Ijen volcano (Fig. 7-1) is located on the eastern border of the island of Java, Indonesia (8°3.5’ S, 114°14.5’ E) and is in the context of subduction of the Indo-Australian plate beneath the Eurasian plate [section 1.2.]. Kawah Ijen volcano is part of the Ijen volcanic complex (Fig. 7-2), which is a volcanic caldera ~15 km in diameter [Dellmelle et al., 1994]. The caldera floor decreases in elevation northward until the Blawan waterfall, which cuts the caldera wall. The Ijen caldera is constituted of 17 post-caldera volcanic edifices (Fig. 7-2). The highest point in the caldera is the summit of Gunung Merapie, with an elevation of ~2600 m a.s.l. Inside Ijen caldera, three different groups of volcanic edifice are present. In the center of the caldera is located the resurgent dome of Blaw. Along the caldera rim, four volcanic cones have formed (Ringgghi, Jampit, Ranteh and Gunung Merapie), which seem to be located a tectonic lineament [Carn et al, 1999]. The last and youngest group consists of a trend of
twelve cinder cones indicating migration of volcanic activity from the west side of the caldera toward the east. Kawah Ijen volcano is the youngest of these cinder cones and is built on the west flank of Gunung Merapi [Berlo, 2001; Van Hinsberg, 2001].

Kawah Ijen is subject to a tropical climate, with a dry season and monsoon season. Traditionally in Indonesia, the monsoon occurs from September to April, with a peak of rainfall in December [Vimont et al., 2009], while the dry season is from May to August. Within the Ijen caldera, during the monsoon, the rain intensity can reach 100 mm/day [Löhr et al., 2005]. The total rainfall during the monsoon on Ijen caldera has been measured at ~2400 mm/year [Löhr et al., 2005]. Over the period of 2006/2007, South-East Asia and Australia were affected by an El Niño event, which started around November 2006 and ended around April 2007 [McPhaden, 2008; Luo et al., 2008]. In Indonesia and especially in eastern Java, the El Niño consequences are expressed through a significant drought over this period [McPhaden, 2008; Luo et al., 2008; Robertson et al., 2009]. An average decrease of precipitation from September 2006 to November 2006 (first part of the monsoon period) was calculated to be ~1.5 mm/day over Java [Luo et al., 2008]. Over this period, no information regarding precipitation change related to the drought is available for Ijen caldera.

The Kawah Ijen crater (at 2400 m a.s.l.) is ~1.2 km x 1 km in diameter, oriented west-east and ~180 m deep (Fig.7-3) [Delmelle et al., 1994, 2000; Takano et al., 2004]. The main part of the crater is occupied by the world’s largest and most hyperacid lake (pH ~0.00, ~1.1 km x 0.9 km x 180 m and ~30 x 10^6 m^3, Fig. 7-3) [Delmelle et al., 1994, 2000; Löhr et al., 2005].
Figure 7-3: Bathymetry of the acid lake modified from Takano et al. [2004]. a) Kawah Ijen crater. b) West-east cross-section of the crater. c) South-north cross-section of the crater. Blue triangles are springs. Blue diamond is the seismic station. Purple polygon is the sulphuric dome. a-a’ line is the West-east cross-section. b-b’ line is the South-north cross-section. The purple contour line is the lake level.
Near the south-east shore of the lake, the crater has a sub aerial sulphuric dome (40 m x 40 m x 20 m), which is actively mined by a local company for its sulphur deposits (> 14 t/day) [Vigouroux et al., 2007; Section 1.2]. The sulphuric dome has hot fumaroles which reach between 580 and 650 °C (2007-2008). The lake temperature (37 °C) was constant between 2006 and 2009. The chemical composition of the lake was studied by several groups [Demelle et al, 1994, 2000; Takano et al., 2004], which found that the acid lake and the Banyupahit River are saturated in chemical elements with a total dissolved solid concentration (TDS) of more than 100 g/kg. The lake composition is mainly made of H₂O, S_{composite}, F, Cl₂, Al, Fe, As, HCL and HF. The complete composition can be found in Delmelle et al. [2000].

The hydrogeographical structure of Ijen caldera is characterized by the hyperacidic Banyupahit River, which flows from the western upper part of Kawah Ijen volcano and northward along the slope of the caldera floor, mixing with other secondary streams near Blawan. The Banyupahit River is fed by several springs on the west flank of Kawah Ijen [Van Hinsberg et al., 2008]. The upper part of the springs result from seepage of the acid lake, while the lower springs are likely due to seepage of the deep hydrothermal system and some natural springs from the Papak cinder cone. From the spring to the estuary on the north shore of eastern Java, the discharge of the acid lake through the Banyupahit River is an important threat to the ecosystem, agriculture and humans living along it. High levels of toxic elements (Al, F, Cl, Mg) generate illness and disease among plants, animals and humans. [Berlo, 2001; Van Hinsberg, 2001; Delmelle et al, 2000; Heikens et al., 2005a, 2005b, 2005c; Löhr et al., 2005, 2006; 2007; VanRotterdam et al., 2008a].

Kawah Ijen volcano is a persistently active volcano characterized by magmatic (ash, lava flow, scoria), phreato-magmatic (fine and coarse lapili with fresh, lake sediments and old material) and phreatic eruptions [Delmelle et al, 1994; Carn et al., 1999]. Present volcanic activity consists of phreatic eruptions with partial explosion of the lake. The objective of this study is to investigate how hydrothermal surveys of Kawah Ijen can help volcano monitoring? What can we
learn about the volcanic activity of Kawah Ijen based on the changes in underground water or hydrothermal system?

### 7.5 Dynamic gravity survey

Gravity variations are due to change in mass, volume, or density inside the shallow surface of the Earth. On an active volcano, gravity variations are generally the consequence of magmatic change, such as magma displacement or vesiculation [Rymer, 1989, 1994, 1996, 1986; Rymer and Williams-Jones, 2000; Williams-Jones and Rymer, 2002; Carbone et al., 2003]. However, gravity variations can be also generated by the rise and fall of the water table, by lake level variations or change in volume of the hydrothermal system of Kawah Ijen. A complete description of the methodology can be found in Chapter 3.

Strong degassing activity can support hydrothermal activity and express a strong underground magmatic activity. Kawah Ijen also has very large lake (~ 30 x 10⁶ m³), which should be supplied by at least one aquifer. To investigate the location and magnitude of gravity variations on Kawah Ijen, a network of 12 stations were installed in August 2006 (Fig. 7-4). Of these stations, 8 are spread around the crater rim, one is near the summit at Pondok (2200 m a.s.l.), one at Paltuding, ~ 2 km from the summit (~ 1900 m a.s.l.), one station ~ 4 km along the Banyupahit River and finally a reference station at the Kawah Ijen Observatory, ~12 km from the summit and outside the caldera. Since 2006, once a year each summer (dry season), a dynamic gravity survey was made. Measurements over all the stations were made over a period of 1 to 2 days.
Figure 7-4: Self-potential profile and gravity stations on Kawah Ijen volcano between 2006 and 2008. Block dots are the Self-potential profiles. Green crosses are the Self-potential reference points (0 mV), water springs and rivers. Red diamonds are the gravity stations. The global reference station (IJ-05) is off the map and outside the caldera toward the east. Blue stars indicate areas of interest.
Each survey day, the Pondok and Paltuding stations were used as local reference stations and occupied on the way to and from the summit in order to control any drift or shift of the gravimeter.

Surveys were all made with the same gravimeter, a LaCoste & Romberg land gravity meter (G-127), equipped with a liquid electronic level and Aliod 100 feedback system, with a Digital Readout to 0.01 mGal [LaCoste & Romberg, Inc., 2003, 2004]. The gravimeter is kept at a constant temperature of 50.1 ±1 ºC to ensure a linear response of the spring and mass. With the Aliod system, the gravimeter can be used with a precision of 0.01 mGal in the measurement range between -50 and +50 mGal [LaCoste & Romberg, Inc., 2003, 2004]. The Aliod records gravity at 2 Hz.

Each station measurement consists of at least 5 measurements taken every minute. Due to important elevation variations (more than 400 m), it is necessary to change the device counter in order to stay within the measurement window of the Aliod (100 mGal). To control the accuracy of the Aliod, during each survey a counter calibration line is made at each of the references stations (Fig. 14-1). Finally, each gravity station is precisely located with a differential GPS, allowing for centimetres vertical accuracy. A complete description of gravimetry theory and application to volcanology can be found in Chapter 3. GPS position and results of the different gravity station are presented in Table 7-1 and Appendix A6 in Tables 14-1.

7.6 Self-potential survey

7.6.1 Origin of self-potential

Hydrothermal systems and aquifers can be localized and differentiated from each other using the Self-potential method (SP), which is a passive measurement of the natural electrical current in the ground. Self-potential allows us to measure the difference of electrical potential between two points. There are several natural processes which generate a variation of the electrical potential
[Corwin and Hoover, 1979; Zlotnicki et al., 2003; Lénat, 2007; Aizawa et al., 2008]. On an active volcano such as Kawah Ijen, all the following sources of electrical generations are possible. By order of importance: the physical displacement of ions, which include electrokinetic, thermoelectric and rapid fluid disruption. Chemical reactions may have a significant effect and include pH of the fluid, chemical heterogeneity of the fluid, ore deposits and redox effects. Minor effects on electrical generation are the magnetic field, animal habitats and manmade structures. A complete description of the Self-potential method can be found in Chapter 2.

The electrokinetic effect is the generation of electrical current by a differential displacement between the ions within the water flow and the electrical charge present on the surface of minerals [Corwin and Hoover, 1979; Zlotnicki et al., 2003]. Electrical generation can reach several hundred mV. Since the electrokinetic effect is directly linked to water flow, there is a SP/elevation inverse gradient [Zlotnicki et al., 2003; Lénat, 2007]. Traditionally, in a hydrogeologic environment with down flow of water (aquifer), the SP/elevation gradient ranges from -10 to \( \leq 0 \) mV/m [Zlotnicki et al., 2003; Lénat, 2007]. In the hydrothermal environment, where hot water is rising up through the ground, the SP/elevation gradient is more complex. Depending on the shape and the localization of the uprising hot fluid, the SP/elevation gradient can be symmetrical, asymmetric or with no clear pattern [Lénat et al., 2007]. Traditionally, the SP anomaly linked to a hot rising fluid will have a bell shaped structure will a positive SP/elevation gradient when it is centered symmetrically within the volcanic edifice.

Electric generation by the electrokinetic effect is directly affected by the Zeta potential, which is the capacity of a mineral to generate electricity based on its polarity. Traditionally, the Zeta potential is negative in value (in mV), however, previous studies have shown that it can also be positive [Ishido et al., 1981; Revil et al., 1999a,b; Guichet et al., 2003; Aizawa et al., 2008]. The control of the Zeta potential does not depend on the major minerals, but rather the secondary minerals making up the rock [Guichet et al., 2003; Aizawa et al., 2008]. The secondary mineral formation will be controlled by the environmental conditions.
surrounding the rock such as: hydrothermal activity, redox conditions, pH, temperature and pressure. On volcanoes, such as Kawah Ijen, hydrothermal activity is intense for a long time which will lead to rock alteration and formation of mineral/ore deposits, which could change the Zeta potential over long periods of time.

7.6.2 Field methodology

Since 2006, two main SP profiles were surveyed every year (Fig. 7-4). The main survey profile is along the crater rim and the second is the South profile. Since 2007, a South-North radial profile has been made, starting from the South Crater peak, near the seismic station through Pondok until Paltuding and then to the spring or Banyupahit river. The sampling step was 20 m. Measurements are made with non polarized copper/copper sulphate electrodes. Electrode polarity is controlled at least twice a day. Recordings are made with a high impedance multi-meter (~ 200 MOhm). Each station is GPS localized and measurements are always taken in a shallow hole (~20 cm deep) to ensure sufficient moisture in the ground. Each record of the electrical potential (in mV) is also complimented by a record of the contact resistance (kOhm) between the electrode and the ground. Contact resistance is a means to determine if the SP measured is due to natural generation and not due to a contact problem between electrode and moist ground. All Self-potential profiles are closed in a loop or directly connected to natural water source. Three different springs are used as 0 mV reference points: the Banyupahit spring, Inner crater spring and the Paltuding spring. In 2007, the Banyupahit river was also used as reference where the river crosses the road. The Kawah Ijen acid lake was not used as reference due to significant signs of chemical heterogeneity at its surface and the risk of electrical generation by redox and electrolysis phenomena which could happen inside it. The springs and river are considered as good references, due to the constant water flow and their lower chemical concentration in comparison to the lake.
Self-potential measurements are complimented by a ground temperature surveys to detect any thermal anomaly. Ground temperatures are taken at more than 20 cm depth, to try to avoid any atmospheric influence. Unfortunately, in some areas it was not possible to dig so deep. Thus, all temperature measurements are also complimented by atmospheric temperature to avoid erroneous interpretation. All data from the Self-potential and ground temperature surveys are presented in the CD.

7.6.3 Effect of very low pH on Self-potential generation

On Kawah Ijen volcano, the pH of the water flow (aquifer and hydrothermal system) is very low (lake and river at pH = ~0.00) and could be of significant importance for electrical generation. Previous studies have shown that pH is the expression of the balance in ions within the water fluid, which will affect electrical generation [Ishido et al., 1981; Jouniaux et al., 2000; Guichet et al., 2003; Zlotnicki et al., 2003; Aizawa et al., 2008; Chapter 2]. In some cases, such as when the pH is ~3, it appears that the Zeta potential becomes null and electrical generation stops. However, it is still not clear what happens when the pH is near zero, as at Kawah Ijen. Our study shows, that SP generation is well established over time on Kawah Ijen, but the amplitude of the anomalies are much lower than on other volcanoes. Thus, in this study the low pH does not appear to stop the electrical generation, but may in some ways (not determined here) affect the electrical generation.

7.7 Multi-scale wavelet tomography

Multi-scale wavelet tomography is based on the continuous wavelet transform, which is a signal processing approach to investigate the origin and structure of a measured signal [Moreau et al., 1997, 1999; Sailhac et al., 2001;
Saracco et al., 2004; Mauri et al., 2009a,b]. A complete description of the theory and application on electrical signals can be found in Chapter 4, with more examples in Chapter 5, 6 and Appendix A1.

Figure 7-5: Image of Multi-scale wavelet tomography of the 2006 crater profile on Kawah Ijen (see Fig. 7-4). Analysis was made with the second horizontal derivative of the Poisson kernel family (H2) over a range of dilation from 1 to 15. Dilation and Correlation coefficient are unitless. E1 and B1 are minima and maxima extrema, respectively. Red diamonds are calculated depths.

In this study, Multi-scale wavelet tomography (MWT) uses a wavelet based on the derivative of the Poisson kernel family and is applied to the Self-potential profile to investigate the depth evolution of the water generating the electrical anomaly. These wavelets are the second and third vertical derivative (V2 and V3, respectively) and second and third horizontal derivative (H2 and H3, respectively, Fig. 7-5). Each analysis was made with each wavelet over 500 dilations on a range of dilation from 1 to 20. We only consider depths found with at least three
of the four wavelet analysis as significant. All the results are presented in Table 7-2, Fig. 7-14 and Appendix A6 (Fig. 14-2 and Table 14-2 to 14-5).

7.8 Forward Gravity model

In this study, forward gravity modeling is performed with Grav3D software (from UBC) [Li, and Oldenburg, 1998; Gailler et al., 2009] and by applying a statistical approach via cross correlation with the measured gravity field data. Due to the number of variables, this study uses very simple models which approximate the topographic area surrounding the gravity station.

- Based on the DEM (Fig. 7-3), and because topography is assumed invariable, the topography around each gravity station is assumed to be flat over a surface corresponding to a rectangle of ~500 m (radial direction from the rim toward the slope of the cone) by ~600 m (tangential direction to rim, Fig. 14-7). The dimension of the model is made large enough to avoid edge effects on the gravity station (Fig. 14-7).

- The rectangle is assumed to be an isotropic porous medium with water in the pore space. The front side of the block is set as the crater rim (Fig. 14-7), which is a cliff ~200 m high on the East side of the Ijen crater (Fig. 7-3b, 7-4). On the west side of the crater rim, the cliff is ~100 m lower than near the gravity station (Fig. 7-3b). The water table limit is considered to be at the edge of the cliff, thus at the edge of the model column (Fig. 14-7). The gravity map generated is always constant in dimension, in order to keep the gravity station in the middle of its front side, which matches with the edge of the water block.

- The density of the water within the model is the multiplication of the water density (1 g cm\(^{-3}\)) by the porosity of the medium (with a ratio from 0 to 1), which gives a full range from 0 to 1 g cm\(^{-3}\). Density of the acid lake is assumed to be of ~1.10 g cm\(^{-3}\) [D elmelle et al., 1994, 2000; Takano et al.,
and the density contrast between magma and gas is assumed to be 2.3 g cm\(^{-3}\), based on the measured gas content [Vigouroux et al., 2007; Van Hinsberg et al., 2009].

**Figure 7-6:** Calculation of water height from gravity value for different models at different porosity values. Model A assumes a top of the water column at 100 m below the surface. Models B, C and D assume a top of the water column at 50, 20 and 10 m below the surface, respectively. Polynomial trends (of second order) have been calculated for each of the models.
In order to investigate magmatic origin (Table 7-3) or lake level variation (Fig. 14-9) as possible sources of the measured gravity model, several models were made. As neither porosity, change of water height, nor depth of the top of the water column is known, the forward model was made in two distinct steps.

First, a series of models of different constant depths of the top of the water column were generated over the full range of both porosity and water column heights (Fig. 14-3). Over each year of gravity survey, the only thing that can be assumed to be constant over time is the porosity of the rock. This assumption can be considered right here, because no hydrothermal activity was detected by the Self-potential survey (See after section 7.9.2), suggesting that no hydrothermal alteration may have changed the property of the rock. Thus, to estimate the porosity through a statistical approach, the measured gravity was compared to each of the models (Fig. 14-3). By cross-correlation between all models and gravity data sets, an average porosity and its associated standard deviation were been calculated. In this way, no assumptions were made regarding the water column height. Thus, for each model having a specific depth of the top of the water column, a porosity range was statistically calculated (Table 14-2).

The second step of the forward modeling is the estimation of water column height change from year to year. From the calculated porosity associated to each model, new model data sets were calculated by changing the height of the water column. Each measured gravity data set is then compared to each gravity model (Fig. 7-6, Table 7-4). Examples of some of the models are presented in Appendix A6 (Fig. 14-3 to 14-10).
Figure 7-7: Dynamic gravity surveys between 2006 and 2008 on the summit area of Kawah Ijen volcano. See Figure 7-4 and Table 14-1 for station locations. Data are presented in Table 7-2. Error bar is one standard deviation.
The main limitation of this methodology to estimate water column depth and porosity of the medium, is that the results are strongly dependent on the number of models and the number of measured gravity data sets. Thus, results will only give a general estimate of the different depths of the water table. This study will only use these results as a first approximation to investigate the dynamics of the water table. All results are presented below.

7.9 Results

7.9.1 Dynamic gravity survey

Between 2006 and 2008, dynamic gravity surveys were made on 12 stations (Fig. 7-4, Table 14-1). All gravity measurements were corrected for earth tide, drift and tares to obtain the residual gravity signal. All stations are referenced to the Observatory station (IJ-05) and show the same pattern of gravity variations. Between 2006 and 2007, a general decrease of gravity ranging from -20 to -210 $\mu$Gal occurred (Fig. 7-7, Table 7-1). In 2008, in comparison to 2006, the gravity signal increased from 10 to 134 $\mu$Gal. All the gravity stations show more or less the same pattern, except IJ-04 at Paltuding, which has an opposite behavior (Fig. 7-7, Table 7-1). Over the period 2006 to 2008, differential GPS measurements showed no significant deformation, however, due to technical problems, there is no deformation data available for 2007.

A total range of gravity variations between 20 to 210 $\mu$Gal, (Fig. 7-7, Table 7-1), found on Kawah Ijen from one year to another, is related to change in mass, volume, or density within the ground, such as change in density of magma, or change in mass/volume of a magma body or of a aquifer. Unfortunately, with no deformation values for 2007, a quantitative approach cannot be used, such as described in the work of Rymer and Williams-Jones [2000]. However, the fact that no significant deformation was seen between 2006 and 2008, allows us to consider the possibility that the gravity variation may not have detected any magmatic change. If the origin of the gravity variation is due to change within the
underground shallow water table along the crater rim, those change should be detected by Self-potential measurements.

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Table 7-1: Gravity variations between 2006 and 2008. All gravity stations are referenced to the station IJ-05 located outside the Ijen caldera at the Volcanol observatory.

7.9.2 Self-potential data

Between 2006 and 2008, all measurements were made at the end of July to beginning of August, during the dry season. Self-potential surveys covered a total ~ 34 km of profile, with a sampling step of 20 m (Fig. 7-4). Similar references at the spring and river (Fig. 7-4) allow us to accurately compare the data through SP maps (Fig. 7-8), SP profiles (Fig. 7-9, 7-10, 7-12 and 7-13) and SP/elevation gradient (Fig. 7-9, 7-11, 7-12 and 7-13). Each year, a Self-potential map was generated from all collected data profiles (Fig. 7-7). All the Self-potential and ground temperature data can be found in Appendix A6, from Table 14-2 to 14-10.
Figure 7-8: Self-potential maps from 2006 to 2009. Distance and elevation contours are in m. All the data are referenced to the water spring and river (pink triangle). Dashed line is the Banyupahit river. Black dots are the SP profiles. 1 and 2 represent hydrothermal areas.
Figure 7-9: SP profiles of Crater rim (top) and its SP/elevation gradient (bottom). The numbers (1, 2, 3, 4, 5) represent the section of different SP/elevation gradient.
7.9.2.1 The Crater rim profile

The first profile is the Crater rim profile, which is 3550 m long (Fig. 7-8, 7-9) and shows significant SP variations over time in the east and north parts. Due to unstable and very steep slopes, the crater profile could not completely cover the crater and thus, the north-west part of the crater is not covered. In order to keep each profile closed inside a loop, the north-east side of the crater rim profile was complemented by a second profile going along the rim in the opposite eastward direction (Fig. 7-10). The closing point of the north loop is located at a large boulder (~1.5 x 1.5 m) near the crater entrance in the south-east part of the crater rim. The parallel profile along the north-east rim allows us to control and correct any drift occurring during the measurements (Fig. 7-10). While the profile will normally be close to the north rim water spring (inside the talweg on the north crater rim), this spring dries out during the dry season. As shown in Figure 7-10, the two parallel profiles have similar anomalies in both localization and amplitude.

7.9.2.2 The External south profile

The second SP survey profile is the External south profile, which is on the upper part of the south flank of Kawah Ijen (Fig 7-4). The profile follows the path from the Dam to the upper entrance of Pondok and then toward the crater entrance via the main trail. The main part of the profile is at low elevation change with an increase of 120 m in elevation over a 2000 m from south-west to south-east (Fig. 7-12). This profile is characterized by significant SP variations, which occur on the upper part of the east flank (toward the east crater rim).
7.9.2.3 The Seismic station-Pondok-Paltuding profile

This SP survey profile was started in 2007 along a north-south orientation. It starts near the seismic station at the south crater peak and continues down along the main slope near Pondok to Paltuding and terminates, outside of Kawah Ijen volcano (Fig. 7-4, 7-13). The Seismic-Pondok-Paltuding profile (SPP) is characterized by a single large anomaly (water table #1), which covers the south slope of Kawah Ijen volcano. Further south from Paltuding, the SPP profile is characterized by three other anomalies on the caldera floor, between Paltuding and the Banyupahit River (Fig. 7-13). Two of these anomalies seem to indicate aquifer flow (water table #2 and #3), while a large extended anomaly, called Undefined, cannot be clearly characterized due to lack of clear SP/elevation.
gradient. More surveys need to be done along this Undefined SP anomaly in order to determine if it represents an aquifer or a hydrothermal system.

7.9.2.4 Kawah Ijen crater

In 2006 and 2007, a Self-potential profile was made inside the crater, from the crater entrance, down to the sulphuric dome and to the inner crater spring (~60 m away from the dome, Fig. 7-4, 7-8). Inside the crater, no significant anomaly was found. The maxima of SP were found near the lakeshore and near the sulfuric dome. However, their close proximity does not allow us to characterize which component of the SP signal comes from each.

Between 2006 and 2008, ground temperature surveys showed no thermal anomaly outside the crater with a background temperature ~ 20°C. Inside the crater, thermal anomalies are only found on the sulphuric dome with a very strong gradient. Ground temperature can increase over a meter by more than 150°C, reaching 300°C on the lower part of the dome. No detailed ground temperature, or SP measurements were done on the sulfuric dome for safety reasons (i.e., fumarole temperatures reaching 600 °C and a very dense gas plume).
Figure 7-11: Comparison of SP/elevation gradient of the Crater rim profile between 2006 and 2008 (see Fig. 7-9). The numbers (1, 2, 3, 4, 5) represent the sections of the different SP/elevation gradient.
Figure 7-11 continued:

SP/Elevation Gradient in 2008

- 0-660m: -0.25 mV/m; R=0.48
- 680-1640m: -0.25 mV/m; R=0.56
- 1660-2100m: -0.40 mV/m; R=0.50
- 2120-2780m: No Gradient
- 2800-3560m: -0.99 mV/m; R=0.61

SP/Elevation Gradient in 2009

- 0-700m: -0.26 mV/m; R=0.71
- 720-1640m: -0.36 mV/m; R=0.91
- 1660-1640m: -0.60 mV/m; R=0.56
- 2060-2800m: No Gradient
- 2820-3120m: -0.78 mV/m; R=0.97
- 3140-3500m: -0.91 mV/m; R=0.08

Figure 7-11 continued:
Figure 7-12: Top: Self-potential, temperature and topography profiles of the External South profile between 2006 and 2009. Bottom: SP/elevation gradient associate to the External South profile. The numbers (1, 2, 3, 4, 5) represent the sections of the different SP/elevation gradients.
Figure 7-13: Top: Self-potential, temperature and topography profiles of Seismic station-Pondok-Paltuding profile (SPP). Bottom: SP/elevation gradient associated to the SP profile.
7.9.3 Depth estimates by Multi-scale wavelet tomography of Self-potential data.

Results of the MWT of SP data include a total of 189 depth values from 2006 to 2009 (47 in 2006, 43 in 2007, 47 in 2008 and 52 in 2009, Fig. 14-2a, 14-2b, 14-2c), however, only depths found with at least three of the four wavelets are considered significant, which allows us to obtain 26 water depths (8 in 2006, 9 in both 2007, 2008 and 2009, Fig. 14-2d, Table 14-11, 14-11 and 14-12). By using only depths found with at least 3 wavelets that are spatially well located over the 3 years (along the profile), this study can characterize 6 data sets (Fig. 7-14, Table 7-2), which are called, by their relative position from the crater: North, North-east, East, South-east, South and South-west.

As Self-potential anomalies are generated by subsurface water flow, the MWT-calculated depths represent the depth of the water. Between 2006 and 2009, four of the six anomalies show very similar patterns (North-east, East, South-east, South, Fig 7-14, Table 7-2), with a decrease of elevation of the top of the water in 2007 (in comparison to 2006) and an increase of the elevation in 2008 (in comparison to 2007). In 2008 (Fig 7-14, Table 7-2), the water depth is below (North-east, South-east) or similar (North, East and South) to the water level of 2006. Water variations generally range from 10 to 100 m (Fig 7-14, Table 7-2) year to year. Only the South-east water depth shows variations great than 150 m, however, its standard deviation (data scattering) is ~120 m. Thus, little confidence can be given to these depth values.

The North and South-West water depth series present a different pattern. In the case of the North water depth series, no trend can be characterized (Fig 7-14, Table 7-2), because the error bar is always equal, or greater than the depth variation. Similarly, the South-West water depth series show a constant increase of depth, the topography itself increases by ~60 m, which is more than the calculated depth variation.
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Table 7-2: Water depth calculated by Multi-scale wavelet tomography (MWT) on the Crater rim Self-potential profile between 2006 and 2008. σ characterizes the scattering of the data.
Figure 7-14: Water depth variation between 2006 and 2009 around the crater rim of Kawah Ijen volcano. Depths calculated by Multi-scale wavelet tomography (MWT) on the Crater rim Self-potential profile, as described in section 7-7. Only depths obtained with 3 of the 4 wavelets are presented here (Table 7-2). Error bars characterize the scattering of the data.
This strong topography variation characterizes a poor spatial localization with strong topography variation, thus the South-west water depth series cannot determine the water depth variation over the period 2006 to 2009.

Fortunately, the standard deviation of ~ 54 % of the depths are less than 30 m, ~ 79 % of the standard deviations are less than 50 m (Table 7-2). Even if the standard deviations are quite significant, there are at least in ~ 79 % of the cases, smaller than the associated variation of the water depth, which allows us to have some confidence in the result of the multi-scale tomography.

The depth variations of 4 of the 6 water depth series is supported by the statistical depth variation based on the 189 calculated depths, which show similar patterns (Fig. 7-15). The majority of the calculated depths drop by ~100 m between 2006 and 2007 and rise by an average increase of ~ 50 m. Thus, the 2008-level is at an intermediate depth level between the 2006 and 2007 depths (Fig. 7-15), while the 2009-level return to similar level than the 2006 water depths.

12% of the depth values have a standard deviation less than 14 m. 37% of the depth values have a standard deviation between 14 and 26 m. Only 29% of the depth values have a standard deviation between 26 and 50 m. Depths values from 2006 to 2009 characterize vertical variations of several ten's of meters from one year to another (Fig. 7-14).

7.10 Aquifer and hydrothermal structures on Kawah Ijen volcano

The summit of Kawah Ijen has a complex water structure. As shown in Figures 7-9, 7-12 and 7-13, the crater rim and the south slope mainly host aquifer structures characterized by negative or weakly positive SP values. Theses structures are broken down as follows: a) Small hydrothermal anomaly (South-East with the section 1-2 and South-West with section 4-5, Fig. 7-12); b) East aquifer (Fig. 7-9, anomaly #3 and #4); c) Ephemeral North aquifer (Fig. 7-9,
anomaly #5); d) Lake seepage on the west (Fig. 7-9, anomaly #1); e) South aquifer (water table #1 on Fig. 7-13 and section 3 on Fig. 12).

7.10.1 Small hydrothermal anomalies (south-east and south-west)

Outside the crater of Kawah Ijen volcano, the only hydrothermal expressions are found along the upper part of the south flank of the volcano. The hydrothermal areas are located on the south-east and the south-west side (Fig. 7-12, section 4-5) along the External south profile. Both South-east and South-west hydrothermal anomalies are of small extent (~ 300 to 500 m long) and with low asymmetric SP anomalies (~20 mV, Fig. 7-12). The South-east hydrothermal area is characterized by an asymmetric SP/elevation gradient (section 1-2 on Fig. 7-12). On the surface, on the outcrop, the South-east hydrothermal area is characterized by important hydrothermal deposits and rock alteration. The fact that there is no ground temperature or strong SP anomaly (Fig. 7-12) suggests that the hydrothermal fluid system was reduced due likely to a decrease of activity in this part of the volcano. Furthermore, as Kawah Ijen still shows strong hydrothermal and degassing activity inside the crater, another explanation is that the strong formation of hydrothermal deposits has self-sealed the hydrothermal fluid pathway somewhere between the surface on the South-east flank of the upper cone and the crater hydrothermal system, which is about ~ 200 m lower in elevation and at ~500 m toward the north of the South-east hydrothermal area (Fig. 7-12, Fig. 7-8).

The South-west hydrothermal area is characterized by an asymmetric SP/elevation gradient (section #3 and 4 on Fig. 7-12) and does not show any sign of surface hydrothermal alteration along the profile, nor any relevant ground temperature anomalies. The same kind of explanation as for the South-east anomaly is possible, where deeper hydrothermal fluids could be blocked by self-sealing.

Finally, neither of the South-east and South-west hydrothermal areas show evidence of active shallow hydrothermal system, but rather the electrical
surface expression of deeper (somewhere in the first 1 km) hydrothermal fluid circulation. A future increase of the electrical signature of these hydrothermal areas could indicate infiltration of hydrothermal fluid toward shallower depth from deeper parts of underlying hydrothermal system.

7.10.2 East aquifer

The East aquifer is located on the east side of the Kawah Ijen crater (Fig. 7-9, section 3-4) and flows from the west summit of Gunung Merapie down its slope and is discharged into the acid lake. It is possible that part of the east aquifer also feeds the crater hydrothermal system, although no information currently exists to support this hypothesis. The East aquifer seems to be the main source of water for the acid lake, due the very strong topographic decrease in elevation from Merapie to the acid lake (~ 500 m, Fig. 7-3b) and due to the persistent evidence of underwater springs along the east shore of the lake (Fig. 1-35, section 1.3.4, Fig. 7-4).

The East aquifer is characterized by a constant negative SP/elevation gradient (section 3-4 on Fig. 7-9 and 7-11) and a persistent negative anomaly. Over the period 2006-2008, the East aquifer has shown some variations of its electrical signature, mainly concentrated on the north-east part of the aquifer (Fig. 7-9). This electrical variation also coincides with variation in the water depth by MWT of the East source, with changes of ~50 m from one year to another (Fig. 7-14).
7.10.3 Ephemeral north aquifer

The Ephemeral north aquifer is located inside the north rim of Kawah Ijen crater (Fig. 7-8 and section 5 on 7-9). It is not clear how the Ephemeral north aquifer interacts or connects with the acid lake, or crater hydrothermal system. The thalweg on the north crater rim (distance 3125 m on topographic profile, Fig. 7-9) has deposits of evaporite minerals (white and yellow), which appear to indicate a seasonal spring. A spring was previously reported on colonial Dutch topographic maps of the summit crater [Berlo, 2001; Van Hinsberg, 2001]. However, over the summers of 2006, 2007 and 2008, no water spring was found on the thalweg. In addition, the electrical signature of the Ephemeral North aquifer has changed significantly from year to year, with an average of ~ ±75 mV/year (section 5 on Fig. 7-9 and Fig. 7-11). Similarly, the East aquifer (section 3 and 4 on Fig. 7-9 and Fig. 7-11) is characterized by a variable SP/elevation gradient from 2006 to 2008, which shows the instability of this water table. Two pieces of evidence suggest that the water flow is from east to west: first, the topographic surface indicates that surface water flow from east to west, following the elevation change. Second, the MWT-calculated depth indicates that the water depth within the east rim (Fig. 7-14a) is at a similar or shallower depth than the North-rim water table (Fig. 7-14b). The aquifer within the North rim is therefore probably supplied by the East aquifer.

7.10.4 Lake seepage on the west

Lake seepage occurs through the west side of Kawah Ijen crater (section 1 on Fig. 7-9). Several acid water springs are found on the external upper part of the west volcano flank. The upper spring position is controlled by the lake level, which controls the level of the top of seepage. The seepage is clearly visible on each Self-potential profile (first 200 m, section 1 on Fig. 7-9) and is characterized by a constant positive anomaly over 2006 to 2008. Unfortunately, no clear water
depth was found on this section by the MWT processing, this does not mean that no water is present, but only that water depth was not characterized by the analyses.

7.10.5 South aquifer

The South aquifer is located on the external south flank of Kawah Ijen volcano (map Fig. 7-8, Fig. 7-13 with water table #1) and flow from the south peak of the crater rim toward the south until at least Paltuding (caldera floor). The electrical signature of South aquifer is very constant over 2007 and 2008 (water table #1, Fig. 7-13) and has a bell shaped inverse SP/elevation gradient, which is made by a low SP/elevation gradient (~0.08 mV/m) and is delimited by a strong gradient (-0.66 and +0.52 mV/m ). The amplitude of the SP anomaly is ~50 to 60 mV (Fig. 7-13). The South aquifer has a water spring located on the lower south flank of Kawah Ijen cone (green cross above Paltuding on Fig. 7-4). The South aquifer can be summarized as having a consistent electrical signature (2007 and 2008), that water flows away from the volcano toward south and toward the upper part of the water table ~300 m above the lake surface. Thus, it seems that the South aquifer is not likely connected with the acid lake or the crater hydrothermal system, because of the electrical structure and the lack of anomalous ground temperature. In addition, local people use the spring located at the lower part of the South aquifer near Paltuding, as drinking water with no reports of problems.
7.11 Dynamic activity of Kawah Ijen volcano

7.11.1 Signs of hydrothermal activity

Four years of Self-potential and ground temperature investigations on Kawah Ijen volcano have shown no significant sign of shallow hydrothermal outside the crater (Fig. 7-9, 7-12 and 7-13). The main hydrothermal activity appears completely restricted within the active crater, as evidenced by the sulphuric dome and the acid lake.

On the upper part of the east flank of Kawah Ijen, each outcrop shows significant evidence of important hydrothermal deposits and rock alteration, implying prior hydrothermal activity. The east part of the External south profile shows significant hydrothermal areas, associated with a weakly hydrothermal SP anomaly, but with no ground temperature anomaly (Fig. 7-8). Thus, within the east flank, a prior hydrothermal system was likely present for a period of time sufficient to strongly alter the bed rock and potentially form hydrothermal ore deposits. Current hydrothermal activity is still present within the east flank, but no evidence is present on the surface.

7.11.2 No apparent subsurface magmatic change

Over the summers of 2006, 2007 and 2008, significant gravity variations were observed ($\Delta G_{2006-2007} = -10$ to $-210 \, \mu\text{Gal}$ and $\Delta G_{2007-2008} = +0$ to $130 \, \mu\text{Gal}$) across the entire gravity network (Fig. 7-7). At the same time, no significant deformation was observed between 2006 and 2008 by the differential GPS survey.

Furthermore, based on field observations, between 2006 and 2007, the fumarole temperatures increased by at least one hundred degrees, reaching an average of $\sim600 \, ^\circ\text{C}$ since 2007 [Mauri et al., 2007; Vigouroux et al., 2007; Van Hinsberg et al., 2009] and were still above $300 \, ^\circ\text{C}$ in 2008. Over the same period, no significant change in the degassing flux was observed (average being $\sim300 \, t$
day$^{-1}$ of SO$_2$) [Mauri et al., 2006, Vigouroux et al., 2007; Van Hinsberg et al., 2009]. Lake levels do appear to change from year to year, from a few centimeters to less than 30 cm; unfortunately, no accurate measurements were made in 2006 and 2007. In the summer 2008, the lake level was measured with the differential GPS, but, repeat measurements are necessary before a definite statement can be made. As Kawah Ijen is an active volcano, it is normal to observe some variations, however, the lake level variation, degassing flux and fumaroles temperatures do not suggest any significant magmatic change.

To investigate if a magmatic origin could be responsible for the measured gravity variation, a simple 3D model was made using Grav3D [Li, and Oldenburg, 1998; Gailler et al., 2009], in order to determine if a drop or rise of magma within a dyke structure (Fig. 7-15) could explain the observed gravity. When time and density contrast are fixed, gravity models become a 7 dimensional problem (X, Y, Z, height, length, thickness, strike). However, without any information on the depth of a magma storage area within the volcanic edifice, any kind hypothesis could be presented. Therefore this study estimates (to a first order) the depth of any magmatic body based on the surface structure of the volcanic edifice and on the measured gravity amplitude. Kawah Ijen host both a large acid lake (with constant T at ~37 °C) and very active solfatara (~300 °C < T < ~600 °C). As neither the lake, nor the solfatara show signs of significant variation associated with underground magmatic change (constant gas flux, temperature, no deformation), it is reasonable to assume that even the shallowest magmatic change must be deep enough to not directly affect the surface. This study assumes that the magmatic change, if it occurred, should take place at least several hundred meters beneath the bottom of the crater (Fig 7-16), but should be less than 1 kilometer to avoid models involving excessively large magmatic volumes.
Figure 7-16: Schematic of the setting of a magmatic dyke for forward models. The 1600 m a.s.l. line represents the top of the gravity model. The lines at 2161 m a.s.l (lake level) and 2400 m a.s.l. (summit station CI-10, CI-20, CI-26), represent the two different elevations used to calculate the synthetic anomaly maps. Blue triangle represent gravity stations.
<table>
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<th>Dyke Eastward</th>
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**Table 7-3**: Dimension of magma body and associated synthetic and measured gravity anomalies on station CI-10, CI20, CI26 and CI-35.

The magmatic volume has been assumed to be small, in order to fit with the lack of deformation measured on the surface, thus 3 types of simple shapes have been chosen (Table 7-3):

- a cubic shape 100 m long (volume of $1.0 \times 10^6$ m$^3$),
- a northward elongated dyke of 200 m by 60 m eastward and 100 m thick (volume of $0.6 \times 10^6$ m$^3$),
- a eastward elongated dyke by 300 m, northward of 60 m and 100 m thick (volume of $1.8 \times 10^6$ m$^3$).
- A rectangular shape by 300 m eastward, 200 m northward and high of 100 m (volume of $6.0 \times 10^6$ m$^3$).
Given the measured content of CO$_2$ (~720 tons/day), H$_2$O (~3,900 tons/day) gas and SO$_2$ (gas: ~200 tons/day; deposit: ~14 tons/day) and their relative abundance within an Andesite magma [Vigouroux et al., 2007; Van Hinsberg et al., 2009], the magma/gas contrast density has been estimated at 2.3 g cm$^{-3}$.

Results of the forward gravity models generated through Grav3D software are summarized in Table 7-3. Synthetic data are calculated for a magma body replacing a void. Thus, the important point is not so much the sign of the value, but rather the absolute amplitude of the gravity variation. Even though these different models are very basic shapes, they are based on reasonable assumptions (magma volume, depth) and thus give an idea of the order of the gravity amplitude that should be detected on the surface.

From each model, the synthetic gravity anomaly has been calculated at the coordinate and elevation of each station located on each “corner” of the crater (Table 7-3, Fig. 7-4). An example of the anomaly map can be found in the Appendix A6 (Fig. A5-9, A5-10 and A5-11).

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**Table 7-4:** Parameters of the models used to estimate the height of the water column.
The results show that synthetic gravity values (for station CI-10, CI-20, CI-26 and CI-35) are of the same order of magnitude as the measured gravity (Table 7-3, Fig. 7-17, 14-11, 14-12, 14-13); however, due to limited known parameters and data, the models do not replicate the observed data on the different stations (CI-10, CI-20, CI-26 and CI-35). Thus, these models cannot prove that the gravity variations measured are due to magmatic change within the volcanic edifice, but cannot refute it neither. In addition, as mentioned above, no other information supports magmatic input, thus this study assume that the gravity variation are not due to magmatic change in volume or that the measurement did not record it. Numerous hypotheses can be made to explain this; one is the influence from the change of a deeper and wider magma body. However, shallower changes should be more appropriate to explain the measured gravity because they will be of smaller volume than a deeper source. The other possibility to explain the gravity variation is a large variation of the volume of the aquifer within shallowest layers (first 300 m) below the topographic surface.
Figure 7-17: Elevation of the top of the water table calculated by inverse modeling as described in the text at each of the gravity stations between 2006 and 2008. Error bars represent the scattering of the data. Model parameters are presented on Tables 7-3 and 7-4.
7.11.3 Large vertical variations of water tables

Water is probably the main component of gravity change on Kawah Ijen volcano where it is present as:

- the acid lake,
- the acid water spring,
- the aquifer.

As a source of gravity anomalies, the acid lake could be the most important. The second will be the water tables. The acid lake has a volume of $\sim 30 \times 10^6$ m$^3$ [Delmelle et al., 1994, 2000; Takano et al., 2004] and should have seasonal variation and yearly cycles [Takano et al., 2004]. A simple forward gravity model was made for a vertical variation of the lake by up to 10 m (Fig. 14-8) and gravity anomaly associated to this change at station CI-35 (assumed at $\sim 2161$ m a.s.l., Fig. 7-4) only 10 m above the lake level. Another anomaly map was generated at an elevation of 2400 m a.s.l., near the elevation of stations CI-10, CI-20 and CI-26 (Fig. 7-4). The lake density is assumed to be $\sim 1.1$ g cm$^{-3}$ [Delmelle et al., 1994, 2000]. A variation of just 1 m will generate a gravity variation of only $\sim 28 \mu$Gal on station CI-35 (Fig. 14-8a), the closest to the lake. Much farther from the lake, the stations CI-10, CI-20 and CI-26 (Fig. 14-8a) will be affected by a gravity variation smaller than 4 $\mu$Gal. Even without accurate measurements of the lake level in 2006 and 2007, visual observations suggest that the lake surface did not change over a meter, perhaps a few ten’s of centimetres. Thus, a volume change of the acid lake is not likely the source of the gravity variation measured between 2006 and 2008.

The other source of large amounts of water is the ground aquifers, however, almost no quantitative information is available for it. Any model can thus only be of a first order and simple in shape.
Figure 7-18: Comparison of the water depth variation on Kawah Ijen volcano from three different models. Model A with a porosity of 0.16 (diamond) and Model D with a porosity of 0.40 (square) are generated from the inverse modeling of gravity data, while the third model (triangle) is from Multi-scale wavelet tomography on SP profile. Model A and D have a porosity of 0.16 and 0.40 respectively. Error bars represent the scattering of the data. No MWT depths are associated with gravity stations CI-34 and CI-35.
In the same approach, as models of magmatic variations as possible sources of the gravity change, several basic models of water table variations have been made using Grav3D as described previously in section 7.8. It is possible as a first order approximation to roughly estimate the porosity of the ground. Of course the result is only an estimation and does not give the real value of the porosity. Through cross-correlation between the observed gravity variations (by station over 3 years) and the synthetic gravity (Table 14-14), and based on the different models (section 7.8, Table 14-14 and Fig. 7-5), the estimated calculated porosities range between 0.1 and 0.45 (Table 7-4 and Fig. 7-6). From the end member of the estimated porosity value from each model (A, B, C, D, Table 7-4), the water height variations have been calculated using the height of the water as a function of the gravity (polynomial fit of second order from each model, Fig. 7-6).

From the measured gravity and associated error bar (Table 7-1), the results of the water height variation indicate a vertical variation of several ten's of meters (Fig. 14-3, 14-4, A5-5 and 14-6). Models A and B, which assume a top of the water column at -100 and -50 m below the surface, respectively, are used for the gravity stations CI-10, CI-20, CI-26 and CI-36, which are on top of a 200 m high cliff overlooking the lake (Fig. 7-17). Models C and D have a top of the water column at -10 and -20 m below the topography, respectively. They have been used to model the gravity measured at station CI-34 (top of the dam stairs) and CI-35 (on the dam), which are only at several metres from the lake level (Fig. 7-17). The main parameter, for constant porosity seems to be the depth of the top of the water column. The deeper the water column, the longer it must be to generate the measured gravity anomaly. The height of the water column that changes over time ranges from 10 to ~150 meters, however, the majority of the models show variations smaller than 100 m from year to year (Fig. 7-17).

Similar results have been obtained from the Multi-scale wavelet tomography (MWT) on the SP profile around the crater rim (Fig. 7-14, table 7-2). The water depth variation from the MWT have been graphically compared which each of the models. Due to the number of models, only the most significant
(Model A and D, Fig. 7-18 and 7-19) are presented in this Chapter, the others are in Appendix A6 (Fig. 14-11, 14-12 and 14-13). All models show roughly similar amplitudes of the water depth variation every year, as with the calculated depth by MWT. However, the elevation of the water table is always shallower when calculated from the measured gravity than with the MWT. Model A with a rock porosity of 0.16, is the only model showing very similar elevation than the water depth calculated by MWT on SP data. The best fit is obtained with stations CI-20 and CI-36 (Fig. 7-18 and 7-19), which match quite well with the pattern of the water depth called East and South, respectively. Based on the 4 gravity stations (CI-10, CI-20, CI-26 and CI-36) and the MWT depth of 3 locations (North-east, East and South), between 2006 and 2008 (Fig. 7-18), a water depth comparison has been calculated. The comparison of average water depth shows a difference averaging at 50 m, with a standard deviation of 30 m. Minimum difference is of 10 m and the maximum difference is of 100 m. Very similar averages difference are found when each gravity station is compared to its closest water depth MWT source. For the gravity station CI-10, CI-26 and CI-36, the model A indicates a water table a few ten’s of meters shallower than the water depth by MWT.

Among the possible explanation for the discrepancy of water depth between the different models (from MWT and inverse gravity), we have:

- The shape and geometry of the water table. The simple gravity model assumes that the water table stops abruptly near the edge of the cliff, however, this is very likely not the case. Much more likely is that the top of the water table is inclined with a significant slope. Unfortunately, there is no possibility to measure it. An inclined water sheet will not have exactly the same Self-potential signature as a flat water sheet. Thus the fact that depth from MWT is always deeper may be explained by the existence of an inclined water sheet. An inclined water sheet should have of course show difference in the gravity variation.
Figure 7-19: Comparison of the water depth variation on Kawah Ijen volcano from three different models. Model A with a porosity of 0.32 (diamond) and Model D with a porosity of 0.40 (square) are generated from the inverse modeling of gravity data, while the third model (triangle) is from Multi-scale wavelet tomography on SP profile. Model A and D have a porosity of 0.32 and 0.40 respectively. Error bars represent the scattering of the data. No MWT depths are associated with gravity stations CI-34 and CI-35.
Another source of error on Self-potential measurements is an effect of the uneven topography surface, which will affect the potential measured and thus increase the error on the depth value [section 4].

The gravity model is based on a estimate of the porosity, which could be wrong and so could significantly affect the depth of the water found by the gravity model.

On the west side of the crater, beneath the Dam, the seepage is a constant process and the elevation of the acid spring on the External west upper flank shows several metres of variation over time. These variations could be due to main factors:

- Large change of lake level should significantly affect the water pressure within the west crater flank.
- Chemical reaction taking place within the rock making the west crater flank could generate change in porosity or permeability of the rock, which will have a direct affect on the volume of water flowing through the west flank. These chemical reactions could be due to both the high level of acidity of the water lake and the high level of chemical elements present within it (section 1.3.3.4.1 in Chapter 1).

The gravity model, using both Grav3D and measured gravity, indicates a maximum range of water variation of a few ten’s of metre, when the porosity is between \( \sim 0.21 \) and \( \sim 0.45 \) (model D and C, Fig. 7-15). No accurate data of the lake level was recorded during 2006 and 2007, however, from visual comparison, the lake did not seem to have changed significantly over this period (probably less than 1 m). Thus, large change of the lake level may not be the origin of the change of the west water table. The continuous presence of a large extended gypsum waterfall all along the acid spring area (Fig. 1-34) indicates a significant interaction between the acid lake and the west crater flank. Based on the chemical composition of the lake water and the spring water found within the basement of the Dam [Delmelle et al., 2000a,b], the concentration in ppm of the
main elements does not show any significant variations. The same observation
can be made about the concentration of Ca and SO$_4$ within the acid water, which
are the main constituents of gypsum (CaSO$_4$.2H$_2$O). Thus, it is not possible to
conclude what is the origin of the water depth variation detected by the gravity
variation between 2006 and 2008.

On the east side of Kawah Ijen crater is a ~200 m high cliff hosting the
East water table, which flows from the summit of Gunnung Merapie toward
Kawah Ijen crater. Underwater springs are constantly present along the east
shore of the lake and the SP/elevation gradient supports the existence of this
water table. It is still not clear how the East aquifer and the Ephemeral north
aquifer are linked, however, it seems likely that a part of the water making the
East aquifer discharges within the Ephemeral north aquifer (SP signal, Fig. 7-9
and water depth by MWT on Fig. 7-14a, b and c). All models from both MWT
(Fig. 7-14b, c and f) and gravity (CI-26, CI-20 and CI-36, Fig. 7-17, 7-18 and 7-
19) indicate that the East aquifer has elevation variations of large amplitude from
one year to another. Gravity models allow us to suggest that the porosity of the
rock making the East flank ranges from ~0.15 to ~0.30.

While no precipitation or well data are available for Kawah Ijen over the
period from 2006 to 2009, regional studies on climate change [McPhaden, 2008;
Luo et al., 2008; Robertson et al., 2009; Vimont et al, 2009] have shown that Fall
2006 and Winter 2007 were affected by an El Niño event, which caused a
significant drought across eastern Java and thus over Kawah Ijen volcano. As fall
and winter are the monsoon period, the effect of such a drought should be
significant on the recharge of the aquifer. The MWT on the SP data have shown
that July 2007 is characterized by a deeper level of the underground water table
located around the crater rim. The drop of the water level is of several ten's m
(average of change ~50 m, $\sigma$ of 30 m, Table 7-2; Fig. 7-15).
Figure 7-20: Summary of the water flow structures on Kawah Ijen volcano based on SP and gravity survey, and on the water depth model by MWT and forward gravity models. The entire summit of Kawah Ijen is affected in one way or another by underground water flow (blue arrow). The main flows are the East aquifer flowing toward the acid lake and the South aquifer flowing toward the caldera floor. The question mark symbol indicates the existence of the Ephemeral north aquifer, for which the flow direction is still unknown. Numbers 1 and 2 mark the localization of weak or deep hydrothermal activity on the south-east and south-west flank of the volcanic edifice. Dashed green line is the Banyupahit river, which start on the upper west flank and is from the seepage of the lake through the crater rim.
After summer 2007 (Dry season in Java), the water table level during summer 2008 and summer 2009 have shown a water level rising up to return to a similar level of 2006. Thus, large climatic change, such as El Niño, which brought significant drought across Java, seems to have large impact on the water table depth, which can be detected by MWT on SP data.

The origin of the large variation of water volume making the East water table is very likely due to the seasonal influence on the rainfall. Kawah Ijen is in a tropical environment, with a dry season and a monsoon. The geophysical surveys are done every year during second half of July or during August, which are within the dry season. Disruption, such as El Niño, or shifts of the end of the monsoon could explain such variation of water flow. The lake itself, ~30 x 10^6 m^3 [Delmelle et al., 1994, 2000; Takano et al., 2004], likely act as a buffer, keeping the West water table (feeding the seepage) from having larger variation like the East water table does.

7.12 Conclusion

On Kawah Ijen volcano, which is well known for its strong sulphuric dome activity, large hyperacid lake and continuous passive degassing, no significant sign of hydrothermal activity was found outside the crater. Long term effects of intense hydrothermal alteration and possible preferential fracturing seems to have allowed the intense shallow hydrothermal system to completely seal itself within the upper volcanic edifice, leaving only the crater as a path to release its over pressurized fluids and gas.

Within the upper part of Kawah Ijen volcano, the vertical variation of the aquifer is a very significant component of the dynamics of the volcanic edifice. Aquifers have been affected by important vertical changes during 2007, while the later years are very similar to the 2006 water table level. Both Multi-scale wavelet tomography on Self-potential profiles and forward gravity models by Grav3D indicate vertical water table variations ranging from 10 to several 10’s m from 2006 to 2008. Even though this study cannot currently give the exact depth of the
water within the crater rim, it does bring quantitative information on water change and thus new insight into the dynamics of water present within Kawah Ijen volcano. These variations do not currently appear to be connected to the volcano activity, but rather to seasonal weather. Monsoon 2006/2007 was affected by a significant drought across Java due to El Niño effect, which correlate well the strong increase of the water table depth. Following the summer 2007, the aquifer have been recovering and the calculated water table depth rose back to the 2006-level.

Understanding the potential field generated from non-volcanic sources allows us to better isolate and detect potential field signal generated from volcanic source. However, non volcanic sources, such as aquifer, are controlled by seasonal weather and climate change, which change over long term periods and thus require long time series data. Self-potential and dynamic gravity surveys, combined with Multi-scale wavelet tomography allow us to access at the dynamic of this non volcanic signals and allows us to better understand the phenomena present on any active volcano.

7.13 References:


Heikens, A., S. Sumarti, M. van Bergen, B. Widianarko, L. Fokkert, K. van Leeuwen, and W. Seinen (2005a), The impact of the hyperacid Ijen Crater


LaCoste&Romberg, Inc. (2003),

LaCoste&Romberg, Inc. (2004),


8 CONCLUSIONS:

This study of persistently active volcanoes by a multi-disciplinary approach, shows that even if potential fields are not unique solutions to a given problem, their combination can be used to assess and better constrain the solution in order to determine the source of the potential field measured. This study use two types of potential-field (Self-potential and dynamic gravity) and combines them with traditional geochemical (ground CO₂ concentration and ground temperature) methods, but also with a modern signal analysis technique, multi-scale wavelet tomography. The results of this multi-disciplinary approach combined with Multi-scale wavelet tomography shows that:

✓ The extension and expression of a hydrothermal surface on a persistently active volcano does not necessarily represent its maturity, or intensity of the hydrothermal system. Masaya volcano, which has only limited hydrothermal surface expression (e.g., fumaroles, hot ground), nevertheless hosts an extensive hydrothermal system. In contrast, Kawah Ijen volcano, the apparently large and well established hydrothermal system exemplified by a very intense sulphur dome and the world’s largest acid lake suggesting, is almost completely constrained within the active crater. The hydrothermal system seems to have been self-sealed within the volcanic edifice.

✓ Both Masaya and Kawah Ijen volcanoes show numerous signs of stable volcanic activity between 2006 and 2009. On Masaya volcano, its volcanic stability is translated through the stability of the depth of its hydrothermal system, which spreads throughout the main cinder cones (Nindiri, Masaya and Comalito cone). On Kawah Ijen volcano, no significant signs of magmatic change where found during the 3 years of survey.
On Kawah Ijen volcano, significant depth variations of the water tables were characterized by both potential fields (Self-potential and gravity). Models of the vertical water displacement, from 2006 to 2008, by Multi-scale wavelet tomography of Self-potential data and gravity modeling allowed us to estimate the vertical water displacement at between 10 and several 10s of m change from year to year. The causes of these water changes are very likely due to seasonal weather changes.

On Masaya volcano, the self-potential mapping, combined with ground CO₂ concentration and ground temperature has allowed us to show that the hydrothermal system is spread widely through the main volcanic cone, and beyond the active crater. The result of this study indicates that the spreading of the hydrothermal systems is strongly controlled by the ring fault structure cutting across the caldera floor.

With respect to methodology, this study shows that by combining different potential fields (i.e., Self-potential and gravity) and processing them by Multi-scale wavelet tomography and traditional gravity modeling, it is possible to constrain the origin of the source generating the measured potential field.

Multi-scale scale wavelet tomography is an efficient tool when several wavelets from the Poisson family are used to calculate the depth of the water table (aquifer and hydrothermal system). The study of Waita volcano, Stromboli volcano and Masaya volcano shows that when Multi-scale wavelet tomography is applied to Self-potential data, it can successfully characterize the depth as accurately as other traditional geophysical methods (e.g., resistivity, water flow models). The main limitation of the Multi-scale wavelet tomography is the sampling step of the Self-potential profile, as well as the Signal/Noise ratio. Fortunately, both of these can be controlled during field measurement by choosing a sampling step of 20 m or smaller. The smaller the anomaly to be investigated; the smaller must be the sampling step. Noise can be limited by applying rigorous and objective field methodology as used in this study.
Multi-scale wavelet tomography (MWT) requires a thorough understanding of and skills in signal processing, geophysics signal and wavelet processing. In order to avoid false interpretations, results from wavelet analyses must always be interpreted through both a signal processing approach and traditional geophysical analysis methods. MWT applied on self-potential is an efficient combination of low cost and fast field technique (SP) with a efficient and robust signal analysis method (wavelet transform), which allows us to obtain significant information about the depth of the water (in this study). Similar result could be obtain by other geophysical technique (i.e., ground resistivity, TEM), however, those traditional geophysics techniques are more expensive to buy and require more manpower (and equipment) and are more time consuming to obtain similar result on the underground structure. Even if the MWT applied on SP data can only give depth accuracy at best from several meters to several ten’s of meters, its accuracy is still at least as good as the accuracy of more traditional geophysical method (i.e., ground resistivity, TEM). Thus, MWT is an efficient signal processing technique to investigate underground water variations (depth of aquifers and hydrothermal systems) which can significantly help on volcano monitoring when available funding, either manpower are an issue.

Further work or topics suggested by this study include continued surveying and expansion of the Self-potential mapping and analysis with Multi-scale wavelet tomography (MWT) for a deeper imaging of the hydrothermal systems on both volcanoes. On Masaya volcano, Self-potential surveying with the MWT is an efficient tool to gain insight on the dynamics of the hydrothermal system complex. Large changes or signs of disruption within its hydrothermal system may indicate signs of magmatic change within the caldera. In addition, a full assessment of the hydrothermal coverage throughout the caldera, but mainly on the northern part, should be able to accurately image the extension of the ring fault. The ring fault has been examined by many previous authors, as well as by this study, however, non clearly define its exact extension through the caldera. Self-potential, with ground CO₂ concentration and ground temperature, can allow
us to accurately characterize this ring fault and help to investigate its role, if any, in the volcanic activity of Masaya volcano.

On Kawah Ijen volcano, dynamic gravity and Self-potential surveys processed by Multi-scale wavelet tomography along the crater profile is an efficient means to investigate vertical change of the water table. Further surveys should focus on the effect of cycles of vertical water variation on the acid lake and on the volcanic activity.

In any volcano monitoring, time-series baseline data is required to investigate volcanic change, and isolate its associated potential field signals and surface expression. On both Kawah Ijen and Masaya volcanoes, this study led to a useful database, characterizing the structure and dynamism of hydrogeological system present on each of volcano. Building on that, future work may be able to detect volcanic signals among the non-volcanic background, in order to better investigate the volcanic change.
9 APPENDICES A1: SELF-POTENTIAL CHANGES AND DEPTH ESTIMATES OF THE SUMMIT HYDROTHERMAL SYSTEM OF PITON DE LA FOURNAISE VOLCANO, LA RÉUNION**.

9.1 Abstract:

Piton de la Fournaise volcano, La Réunion, is a persistently active basaltic shield volcano, which shows no evidence of persistent hydrothermal activity on its surface. However, the volcano hosts an extensive hydrothermal system within its summit cone, which has been regularly monitored by the Self-potential electrical method since the early 1980’s. By applying multi-scale wavelet tomography to Self-potential data, it is possible to obtain reproducible depth measurements of the main hydrothermal fluid cells. Over a period of 13 years, from 1993 to 2005, 8 complete-loop Self-potential surveys were analyzed by multi-scale tomography to obtain a total of 192 depths. This study suggests that the shape and location of the hydrothermal system changes substantially over this period. Vertical displacement of the main Self-potential sources, associated with the main hydrothermal fluid cells, is observed on the order of a few hundred metres at the transition between the period of quiescence (1992-1998) and the resumption of eruptive activity in 1998. From 1998 to the end of this study in 2005, the hydrothermal system was consistently located at relatively shallow

depths. The significant vertical displacements and shallow depths of the hydrothermal fluids suggest that the hydrothermal system is supported by significant and persistent heat flux that maintains it near the top of the summit crater rim. The hydrothermal system is thus directly affected by the high eruptive rate of Piton de la Fournaise and is an indirect indicator of the shallow magmatic activity occurring within the summit cone. Multi-scale tomography of Self-potential data is an efficient and accurate method to investigate the relationship between the dynamic behaviour of hydrothermal systems and persistent magmatic activity. When used in conjunction with long term volcano monitoring, this approach can play an important role in detecting the precursory signals leading to changes in volcanic activity.

9.2 Introduction

Piton de la Fournaise volcano, La Réunion, hosts a well developed hydrothermal system inside its summit cone [Lénat and Bachelery, 1990; Michel and Zlotnicki, 1998; Fontaine et al., 2002; Saracco et al., 2004]. Since the 1980's, the volcano has been well studied by Self-potential (SP), electrical resistivity and other geophysical methods [Malengreau et al., 1994, Lénat et al. 2000; Peltier et al., 2005, 2006, 2007; Brenguier, 2007; Peltier et al., 2009].

The high level of eruptive activity since 1998 [Peltier et al., 2005, 2006] and the frequent Self-potential surveys make Piton de la Fournaise an excellent location to investigate hydrothermal dynamism [Malengreau et al., 1994; Lénat et al. 2000]. Previous studies have shown that the summit cone is characterized by several strong Self-potential anomalies spread around Bory and Dolomieu craters [Malengreau et al., 1994; Lénat et al., 2000].

Multi-scale wavelet tomography of Self-potential data is based on the Poisson kernel family, which allows for the combination of wavelet theory with potential field properties in order to locate subsurface water flow and hydrothermal fluids [Moreau et al., 1997, 1999; Sailhac and Marquis, 2001;
Saracco et al., 2004; Mauri et al., submitted]. Saracco et al. [2004] used multi-scale wavelet tomography and dipolar probability of Self-potential data (1993, 1998 and 1999) on Piton de la Fournaise volcano to show the presence of a geophysical discontinuity linked with underground water flow, although no clear trend was visible. Through multi-scale wavelet tomography using 4 different wavelets and applied to 8 complete Self-potential surveys over 13 years (1993 to 2005), this study investigates the change in depth of the hydrothermal system and how its dynamic behaviour relates to changes in volcanic activity.

9.3 Geological setting

Piton de la Fournaise is a basaltic shield volcano, located in the Indian Ocean, which over the last 500,000 years, has been shaped by several caldera and flank collapses (Fig. 9-1) [Bachèlery, 1981; Lénat and Bachèlery, 1990; Merle and Lénat, 2003; Carter et al., 2007; Oehler et al., 2008; Michon et al., 2009]. Its highest peak is a large summit cone (2632 m a.s.l.; 5,000 years old), located in the Enclos Fouqué caldera (Fig. 9-1), which hosts two recent craters: Bory and Dolomieu, the latter formed in 1930. The summit cone consists of alternating lava flows and scoria layers, which are cut by a well developed dyke intrusion structure [Fontaine et al., 2002; Battaglia et al., 2005; Peltier et al., 2005, 2006, 2007, 2009; Brenguier, 2007].

The structure of Piton de la Fournaise volcano is strongly affected by different faults and fractures [Carter et al., 2007] which converge toward and/or are located within the summit cone. A local and circular fault network (vertically dipping faults) is present inside the summit cone and affects the shape of Bory and Dolomieu craters. A large fault system, at the edifice scale, is connected with the rift and shear zones [Lénat and Bachelery, 1990; Carter et al., 2007]. The southern and northern parts of this large fault structure converge toward the center of the summit cone. Finally, the east side of summit cone is affected by
conjugate faults in response to the eastern spreading of the eastern volcano flank [Carter et al., 2007; Michon et al., 2009].

The magma storage system forms a complex structure of pipes and caverns located within and immediately beneath the summit cone, which has been imaged by 3D wave seismic tomography [Brenguier et al., 2007] and dyke intrusion by inverse deformation modelling [Peltier et al., 2005, 2006, 2007, 2009]. Recent studies by Peltier et al. [2005, 2009] have shown that the internal structure is dominated by 3 distinct dyke structures. Within the summit cone, magma injection follows two main paths along the summit dyke pathway located below Dolomieu crater and a second dyke located on the eastern part of the summit cone. Both of these structures are thought to originate near the top of the shallow magma reservoir (located at ~0 m a.s.l.) [Peltier et al., 2009].

Piton de la Fournaise is very active with 18 eruptions between 1997 and March 2005 [Peltier et al., 2009]. During this period, the strongest volcanic activity was in 2003, when 4 eruptions and 1 shallow intrusion occurred [Peltier et al., 2005, 2009]. However, prior to this period, Piton de la Fournaise had no eruptive activity for 5 years (1993 to 1997) and only 3 eruptions between 1998 and 1999. From 2004 to March 2005, the volcano erupted 4 more times. The presence of large and dynamic dyke intrusion structures within the summit cone and the high eruption rate suggests that the heat flux from these shallow magmatic intrusions is very probably quite strong and stable over time since at least 1999.
Figure 9-1: Map of Piton de la Fournaise volcano, Réunion, France and location of the main hydrothermal zones (diamond). Self-potential (SP) profile around the summit crater is in red, with the red crosses showing the distance along the profile. BRS is the SP Bory reference station.
9.4 Hydrothermal system of Piton de la Fournaise volcano

Volcanic hydrothermal systems are complex systems with a wide range in shape, intensity and chemical composition [e.g., Osinski et al., 2001; Merlani et al., 2001; Chiodini et al., 2005]. The hydrothermal fluids consist of one or two phases (liquid and/or gas), are dominated by meteoritic and magmatic water, and generally develop into convective fluid cells, which shape the hydrothermal system [e.g., Osinski et al., 2001; Finizola et al., 2002]. Piton de la Fournaise volcano hosts a shallow hydrothermal system which is characterized by strong Self-potential anomalies (several hundred’s of mV) [Malengreau et al., 1994; Zlotnicki et al., 1994; Lénat et al., 2000] suggesting that it is located within the summit cone. The hydrothermal system appears to have a zone of water recharge in the eastern side of the summit cone (the lowest part of the summit rim), which is characterized by a strong and persistent negative Self-potential anomaly over the 13 years (EAST, Fig. 9-1). The limit between the recharge zone and the rising hydrothermal fluids is believed to pass through the eastern part of Dolomieu crater [Saracco et al., 2004].

Based on field observations, modelling of the magmatic plumbing system [Peltier et al., 2005, 2006, 2009; Brenguier et al., 2007], Self-potential mapping [Malengreau et al., 1994; Lénat et al., 2000; Saracco et al., 2004], and our understanding of hydrothermal systems [Osinski et al., 2001; Merlani et al., 2001; Chiodini et al., 2005], it is reasonable to make the following assumptions about the hydrothermal system of Piton de la Fournaise volcano. The hydrothermal fluids are heated and recharged by hot gas and magmatic water from the shallow magmatic intrusions within the summit cone. The gas phase of these partially boiling hydrothermal fluids expands and increases the fluid pore pressure. This causes the hydrothermal fluids to rise and spread through the summit cone along the main fault structures. Upon reaching the surface, the surrounding pressure and temperature drop and the hot hydrothermal fluids cool by heat transfer to the surrounding environment, causing the gas phases to condense and the fluid
pressure to decrease. When the fluid pressure drops below the sum of the gravitational and resistance force of the surrounding rock, the cooler hydrothermal fluids will cease to rise and start to sink along the bedrock and fault structures. In addition, any surface or meteoric water will recharge the hydrothermal system with fresh water, adding to the cooler sinking fluids. When these hydrothermal fluids return to the bottom of the hydrothermal system, their proximity to hot magmatic dykes or reservoirs will increase their temperature, which will initiate a new convective cycle. The hydrothermal system of Piton de la Fournaise volcano hosts a well established hydrothermal system [Malengreau et al., 1994; Lénat et al., 2000; Saracco et al., 2004] that is probably shaped by several hydrothermal convective cells spread through the summit cone.

9.5 Self-potential method

Self-potential is a passive electrical method that measures the natural electrical current present in the ground [Ewing, 1939; Poldini, 1939; Corwin and Hoover, 1979; Ishido and Mizutani, 1981; Zlotnicki et al., 1994; Zlotnicki and Nishida, 2003; Lénat, 2007]. SP surveys are commonly made using two copper electrodes, consisting of a copper rod in a saturated copper-sulphate solution, connected to a 500 m insulated wire cable and a high impedance (100 MOhm) multi-meter [e.g., Finizola et al., 2002].

A natural electrical current is generated by a number of different processes, with the two most important being the electrokinetic and thermoelectric processes. The electrokinetic process occurs when water flows through a porous medium. Several parameters will affect the electrokinetic effect such as the chemical composition of the water flow and the zeta potential, which is the electrical expression of the interaction between ions along the Helmholtz double layer [Hase et al., 2003; Aizawa, 2008]. The pressure of the flow will remove ions from the surface of minerals constituting the rock and thus generate
a differential displacement between the ions present in solution and the polarized mineral, which is located along the Helmholtz double layer [Avena and de Pauli, 1996; Guichet and Zuddas, 2003; Hase et al., 2003]. In the case of a common negative value of zeta potential, a positive electrical current is generated at the head of the water flow. The stronger the water flow or fluid pressure, the stronger the differential ion displacement will be, leading to an increase in the electrical current. In hydrogeological environments, where the water flow direction is typically controlled by gravitational forces, the water will generally flow down. Previous studies [e.g., Corwin and Hoover, 1979; Zlotnicki and Nishida, 2003; Lénat, 2007] have shown that there is an inverse relationship between the topographic surface and the electrical anomaly, called the topographic effect. The further the top of the water table is from the topographic surface, the more negative will be the electrical potential. In hydrothermal environments, where the fluids are generally rising due to fluid pressure, electrokinetic processes will generate a positive SP anomaly, which will be measured above the rising point. The SP/elevation gradient can be symmetrical or asymmetric to the center of the anomaly, depending on the shape of the topographic surface [Lénat, 2007]. Self-potential anomalies generated by the electrokinetic effect are typically several ten’s to several hundreds of mV in amplitude. A related effect, rapid fluid disruption (RFD) [Johnston et al., 2001], is an ephemeral phenomena which characterizes changes in state of the water to vapour phase and will be expressed by an increase of the rising water flux.

The thermoelectric process occurs when heat flux (e.g., due to magmatic intrusions) is applied to a rock generating a thermal gradient. An increase in temperature will increase the energy of the free ions inside the pores of the rock. The differential displacement of these ions will thus generate an electrical current. Self-potential anomalies generated by thermoelectric processes typically range from a few to several ten’s of mV in amplitude.

Another phenomena is the effect of heterogeneous ground resistivity on electrical potential, which is commonly considered as a secondary effect [Sailhac
and Marquis, 2001; Saracco et al., 2004]. However, a previous study [Minsley et al., 2007] has shown that resistivity contrasts, when not spatially associated with water flow (electrokinetic effect), may be of significant importance in some instances. In this study, however, we investigate a situation where the main resistivity contrasts have been associated with underground water flow or hydrothermal systems.

A more detailed description of self-potential electrical generation can be found in the works of Ewing [1939], Poldini [1939], Corwin and Hoover [1979], Zlotnicki and Nishida [2003], Avena and de Pauli [1996], Johnston et al. [2001], Guichet and Zuddas [2003], Hase et al. [2003], Lénat [2007] and Aizawa [2008]. As classical Self-potential studies can only give the surface projection of the underground water structure, one approach to obtain information about the depth of the water is to process SP data by multi-scale wavelet tomography. The depth estimate of the different hydrothermal cells can be determined by using multi-scale wavelet tomography of the temporal Self-potential data through several analysing wavelets [Mauri et al., submitted]. The vertical movement of the SP source associated with the hydrothermal cells can then be evaluated in order to understand the temporal behaviour of the hydrothermal system.

9.6 Multi-scale wavelet tomography

Multi-scale wavelet tomography (MWT) is a signal processing method based on continuous wavelet transform [e.g., Grossmann and Morlet, 1984; Saracco, 1994]. When multi-scale wavelet tomography is used with Poisson kernel family and applied to potential field data, it allows for depth determination of the object generating the measured potential-field anomaly [Moreau et al., 1997, 1999; Sailhac et al., 2000; Sailhac and Marquis, 2001; Gibert and Pessel, 2001; Saracco et al., 2004; Fedi et al., 2005; Cooper, 2006; Fedi et al., 2007; Mauri et al., submitted]. The Poisson kernel family includes all wavelets that are
a derivative of the Poisson kernel as described in the work of Moreau et al. [1997]. Each of these wavelets of derivative order $n \in \mathbb{N}$ can detect any singularity which has a homogeneous distribution order of $\alpha \leq 0$. For example, a monopole has a homogeneous distribution order of $\alpha = -2$, while a dipole has $\alpha = -3$. For a singularity with a homogeneous distribution order of $\alpha \leq 0$, the derivative order $n \in \mathbb{N}$ of the wavelet must be $n \geq -(1 + \alpha)$ [Moreau et al., 1997; Fedi et al., 1998; Sailhac et al., 2000; Sailhac and Marquis, 2001; Fedi et al., 2005; Cooper, 2006]. The general equation of the horizontal Poisson kernel family is in the frequency domain [Moreau et al., 1997, 1999; Sailhac et al., 2000; Fedi et al., 2005]:

$$H_n(u) = (2\pi u)^n \times \exp(-2\pi |u|) \quad (1)$$

with $u$ the Fourier transform of the distance, $x$, in the frequency domain. Similarly, by using the Hilbert transform [Sailhac et al., 2000; Saracco et al., 2007], the general equation of the vertical Poisson kernel family can be expressed as a function of the horizontal component $u$ in the frequency domain.

$$V_n(u) = -2\pi |u|(2\pi u)^{(n-1)} \times \exp(-2\pi |u|) \quad (2)$$

where $n \in \mathbb{N}$ is the derivative order of the Poisson kernel family and $i$ is the imaginary number.

In the case of Self-potential data, multi-scale wavelet tomography allows us to locate water tables and hydrothermal systems [Sailhac and Marquis, 2001; Saracco et al., 2004; Mauri et al., submitted] where the medium is considered to be homogeneous. This is an assumption where water flow is responsible for both the electrokinetic effect and the resistivity contrast [Revil et al., 2004]. As the electrokinetic effect is the principle component of electrical potential generation [Corwin and Hoover, 1979; Zlotnicki and Nishida, 2003; Revil et al., 2004;
Finizola et al., 2006], we can assume that the electrical potential from resistivity contrast is negligible [Mauri et al., submitted].

To obtain source depths, the Self-potential profile is analyzed over a range of dilations (several hundred) of the analyzing wavelet, which allows us to detect and characterize all the singularities shaping the analysed signal [Moreau et al., 1997, 1999; Saracco et al., 2004, 2007]. The result of the multi-scale wavelet tomography over a full range of dilations on the analysed signal is an image of correlation coefficients (amplitude of wavelet coefficient) (Fig. 9-2), where minima and maxima of the correlation coefficient are organized on lines of extrema. These lines of extrema are organized in pairs or triplets, which converge in a cone shape structure (Fig. 9-2). A complete description of the theory can be found in the work of Moreau et al. [1997, 1999] and Sailhac et al. [2000].

This study uses four real wavelets based on the Poisson kernel family [Moreau et al., 1997, 1999; Sailhac and Marquis, 2001], which are the second and third vertical derivative (V2 and V3, respectively) and second (n = 2) and third (n = 3) horizontal derivative (H2 and H3, respectively).

Theses wavelets allow us to locate dipole ($\alpha = -2$) and monopole ($\alpha = -1$) sources, which correspond to the electrical anomalies generated by water flow through bedrock (Fig. 9-2) [Sailhac et al., 2001; Zlotnicki and Nishida, 2003; Lénat, 2007]. Each analysis for each profile was made with 500 dilations on a range of dilation from 1 to 15. Only depths found with at least three of the four wavelet analyses are considered significant in this study.
9.7 Results

The Self-potential data set for Piton de la Fournaise consists of 8 complete profiles made around the summit craters of Dolomieu and Bory over 13 years (1993 to 2005) (Fig. 9-3). The profiles were made in June 1993, December 1995, March 1998, December 1999, April 2002, April 2003, November 2003 [Malengreau et al., 1994; Lénat et al., 2000; Labazuy, unpublished data] and March 2005 (this study) (Fig. 9-3). The sampling step of these surveys was 12.5 m or 25 m along the summit loop (~3.7 km in length). All SP profiles were referenced to the Bory Reference Station (BRS) on the south rim of the Bory crater (Fig. 9-1). SP profiles prior to 1993 and that of 2001, are not considered...
here as they had excessively large sampling steps or did not completely encircle the summit craters. Over the 13 years of Self-potential surveys, 5 of the 8 surveys (March 1998, December 1999, April 2002, April 2003 and November 2003) were made less than a few weeks before, during or after an eruption on the summit cone. During the March 2005 survey, an eruption was ongoing in the north-east part of the Enclos Fouqué caldera (Fig. 9-1).

As these data were collected by several authors over the years, it is difficult to constrain all the sources of uncertainty. However, along the SP survey loop, location error will produce uncertainties less than 10mV. Closure errors over the 3.7 km are less than 100 mV; in 2005, the closure error was less than 50 mV. All SP surveys were conducted during the dry season and never during or following rain to avoid water infiltration effects. Drift of the electrodes were controlled twice a day. The traditional leapfrog technique was used in 2005 survey. Error and noise on data acquisition are considered to be negligible in comparison to the amplitude, size and persistence of the SP anomalies measured over 13 years.

Several persistent anomalies are present on the SP profiles (Fig. 9-3). There are 6 main positive SP anomalies [Malengreau et al., 1994]: the Bory north anomaly, BN; Bory south anomaly, BS; Dolomieu south anomaly, DS; Dolomieu south-east anomaly, DSE; Dolomieu east anomaly, DE; and the Dolomieu north anomaly, DN. In addition, one large negative anomaly (EAST) is located on the eastern part of the summit crater (Fig. 9-3). Everything further east of the Dolomieu south-east (DSE) and Dolomieu north (DN) Self-potential anomalies is linked to the EAST negative SP anomaly (Fig. 9-3), which has an inverse SP/elevation gradient and thus characterizes sinking water. To the south-east of the EAST anomaly, the well established positive Dolomieu east anomaly (DE) appears to be linked with the south-east fault (EF) transecting the Dolomieu summit crater (Fig. 9-3).
Figure 9-3: The 8 Self-potential profiles used in this study from 1993 to 2005. a) Topographic elevation. SWF is the south-west fault and EF is the east fault. b) Self-potential profile for 1993 showing the location of the hydrothermal zones (BS, DS, DSE, DE, DN, BN and BS). EAST is the eastern water recharge zone. c) Self-potential profiles between 1995 and 2005.
<table>
<thead>
<tr>
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<th>$\sigma x$ in m</th>
<th>Depth Z in m</th>
<th>$\sigma Z$ in m</th>
<th>Elevation in m a.s.l.</th>
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Table 9-1: Source depths of the hydrothermal zones calculated by multi-scale wavelet tomography of Self-potential profiles on Piton de la Fournaise volcano between 1993 and 2005. Number of wavelets used in MWT to locate the source along profile. $\sigma$ is one standard deviation.

<table>
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<tr>
<th>Month</th>
<th>Wavelets</th>
<th>SP Value</th>
<th>Amplitude</th>
<th>Depth 1</th>
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Although the BRS reference station is located on the summit cone and as such, its absolute SP value may vary over time, any change will not affect the relative amplitude of the SP anomaly or the signal structure of the profiles. Furthermore, multi-scale wavelet tomography of Self-potential profiles is effective with relative SP data [Mauri et al., submitted], which consider peak to peak amplitude rather than absolute amplitude. Consequently, the MWT-derived depths for the 8 SP profiles are accurate independently of the choice of the reference point.

In order to characterise the effect of noise in the SP profile on the MWT depth calculation, the signal/noise ratio (SNR) of each of the 8 profiles was calculated. A SNR of 1 means that there is no noise. In a related study, Mauri et al. [submitted] show that a synthetic SP signal with SNR between 0.8 and 0.9 will
have errors on the MWT-calculated depths less than 25 m for objects located at 200 m depth. All of the 8 SP profiles in this study have a SNR between 0.92 and 0.99, which allows us to be confident that noise should not have a significant impact on calculated depth.

In total, 6 main fluid zones were located with each of the four wavelets on each of the 8 SP profiles for a total of 192 MWT-calculated depths (Fig. 9-4, Table 9-1). The uncertainty of the depths (diamonds, Fig. 9-2) is due to multiple best fit possibilities and is expressed as the error bar of the depth value (Fig. 9-4). In order to better evaluate temporal variations, the depth values for each of the 6 hydrothermal cells (DN, DS, DSE, DE, BN, BS) are also presented as an average depth value (with its associated standard deviation) (Table 9-1, Fig. 9-5). Furthermore, an arbitrary average depth boundary equal to the global average of the 192 depths (2370 m a.s.l. or 189 m below the topographic surface) is shown (Fig. 9-4, 9-5). This average depth also corresponds to the median elevation of the summit cone, which rises from ~2200 m a.s.l. to around 2600 m a.s.l. [Michon et al., 2009]. For this study, we consider all depths above this boundary (~2400 m a.s.l.) as shallow and those below it as deep.

<table>
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**Table 9-2**: Minima, maxima, mean and $\sigma$ of the depth estimate per year of the 6 hydrothermal zones on Piton de la Fournaise volcano between 1993 and 2005. $\sigma$ is one standard deviation.
In 1993, depth values are spread equally on either side of the average depth boundary, from 2100 m to the surface (Fig 9-4). In 1995, depth values are more scattered, but consistently below the average depth boundary, ranging from ~1500 m to ~2400 m a.s.l. (Table 9-2). In 1998, the majority of depth values are above the average depth boundary, from a few ten's of meters below the surface to 2150 m a.s.l. In 1999, as in 1993, depth values scatter on either side of the average depth boundary, from ~2600 m down to ~2200 m (Fig. 9-4). From April 2002 to 2005, depth values are spread around the average depth boundary from ~2150 m to near the surface (Fig. 9-4).

9.8 Discussion

Based on the 8 Self-potential profiles (1993-2005), the MWT-calculated depths form distinct groupings that are horizontally well correlated with the 6 SP anomalies (BS, DS, DSE, DE, DN, BN, Fig. 9-2, 9-3, 9-4). Previous geophysical work by Lénat et al. [2000] has shown that the summit cone of Piton de la Fournaise hosts a large hydrothermal system and the 6 SP anomalies are very similar to those found by Malengreau et al. [1994] between 1983 and 1988. In order to evaluate the temporal change of each of these zones, the average depth for each year is presented in Figure 5 (Table 9-1). In these depth calculations, the uncertainty generally is small (Fig. 9-4), however, the scatter can be large (from an 18 m to 270 m, Table 9-1).
Figure 9-4: Multi-scale wavelet tomography-calculated depths for the 1993 to 2005 SP profiles. V2, V3 are the second and third vertical derivative wavelet, respectively. H2 and H3 are the second and third horizontal wavelet, respectively. Each year, there are 24 depth calculations found along the profile. Error bars represent the uncertainty on the best fit intersection (see Fig. 9-2). Dashed grey line is the global average depth based on 192 values found over the 13 years. In addition, the grey line is the median elevation of the summit cone.
Figure 9-4 continued:
In fact, the scatter of the depth values increases with the depth value itself, however, it is possible to interpret change in depth of the hydrothermal fluid as a whole and how they relate to volcanic activity on Piton de la Fournaise volcano. In order to identify, over time, similar behaviour of the hydrothermal fluid cells, a correlation coefficient \( r^2 \) was calculated by taking each time-series depth of the 5 hydrothermal fluid cells (DS, DSE, DE, DN, BN) and comparing it with the time-series depths of the Bory South (BS) hydrothermal cell. Although the choice of Bory South as reference is arbitrary, the correlation \( (r^2) \) between BS and each of the other hydrothermal fluid cells is high (DS: 0.83, DSE: 0.77, DE and DN: 0.98 and BN: 0.94) suggesting that all hydrothermal zones have a common dynamic behaviour (Table 9-1 and Fig. 9-5).

From its base (~2000 m a.s.l.) to its highest point (2600 m a.s.l.), the summit cone of Piton de la Fournaise volcano is cut by two shallow magmatic dyke structures (Dolomieu and Proximal dyke pathways) [Peltier et al., 2009], which feed each of the summit eruptions. In June 1993, about 10 months after the last eruption prior to the period of quiescence, all the hydrothermal fluids were calculated to be near the median elevation of the summit cone (~2400 m a.s.l., Table 9-2, Fig 9-5, 9-6). By 1995, 26 months after the August 1992 eruption, the hydrothermal fluid depths had dropped to ~2100 m a.s.l. (Table 9-2, Fig. 9-4b, 9-5, 9-6) suggesting a cooling and depressurization of the hydrothermal system. With the restart of eruptive activity in March 1998 [Peltier et al., 2009], frequent eruptions took place on the summit cone (above 2200 m a.s.l., red bars on Fig. 9-4c, 9-5, 9-6). Magma migration along dyke structures through the summit cone [Peltier et al., 2009] increased the heat flux through the upper part of the summit cone. This is supported by the upward movement, by almost 400 m, of the hydrothermal fluids to ~2400 m a.s.l. (maxima at 2600 m a.s.l., Table 9-2, Fig 9-4d).
Figure 9-5: Map of the position of the main hydrothermal structures found by SP and MWT on SP profiles along the survey loop around the summit Bory-Dolomieu craters and their time-series depth evolution over the period from 1993 to 2005. Dashed grey line represents both the average global depth over the 13 years and the median elevation of the summit cone. Red bars denote eruptive events that took place above 2200 m a.s.l. on the summit cone.
Figure 9-6: 3D view of the depths of the hydrothermal fluid cells within the summit cone of Piton de la Fournaise volcano in 1993, 1995 and 1998. Shaded contour is the topographic surface of the summit cone (Gauss Laborde Réunion coordinates, meters). Contours are at 285 m.
From 1999 to 2005 (end of this study), the eruptive activity was high (17 eruptions happens above 2200 m a.s.l.) [Peltier et al., 2009] implying a potentially high and constant heat flux inside the summit cone. During the same time, the hydrothermal fluid depths are always located near the 2400 m a.s.l. limit (Table 9-2). The depth of the hydrothermal system is thus directly affected and controlled by the volcanic activity taking place within the summit cone of Piton de la Fournaise volcano.

9.9 Conclusion

The hydrothermal system of Piton de la Fournaise volcano consists of 6 distinct and persistent hydrothermal zones spread within the summit cone and located around the Bory-Dolomieu craters. With the renewed eruptive activity in 1998, the hydrothermal elevation varied by ~400 m and since 1998 until 2005 (end of this study), the hydrothermal zones remained at ~2400 m a.s.l., near the top of the summit cone. As a complement to traditional geophysical survey methods, multi-scale wavelet tomography of Self-potential data can be a useful tool to investigate dynamic change in the hydrothermal system and such gather information about the internal magmatic activity. On Piton de la Fournaise volcano, the summit Self-potential survey loop requires ~2 to 3 days of measurement and full processing by multi-scale wavelet tomography can be done in one day. Thus, a survey every few months or less could be performed and thus could significantly improve our understanding of the short-term changes within the hydrothermal system and the different process affecting the summit cone of Piton de la Fournaise volcano.
9.10 Acknowledgements

This work would not have been possible without the continued support of the Piton de la Fournaise Observatory and the University of St Denis, La Réunion, in particular T. Staudacher and P. Bachelery. We thank T. Spurgeon for his help and discussion and A. Peltier, L. Letourneur and T. Dorsch for their help during field work. We thank the CNRS-France for financial support through the national scientific program ACI CatNat and ANR volcarisk.

9.11 References


10.1 Abstract

An understanding of the physical and chemical processes inside hydrothermal systems is presented through an evaluation of the scientific literature. Hydrothermal systems are localized mainly in volcanic regions and more generally where the geothermal gradient is abnormally hot. Hydrothermal systems are also an important localization of ore deposit and metals.

This study is based on geophysical (self-potential) and/or geochemical (gas measurement, mineral composition) analyses of many hydrothermal systems in volcanic edifices around the world and in one impact crater in the Canadian Artic.

Hydrothermal systems are complex systems, which are controlled in shape and chemical composition by the structure and the composition of rock which host it. Their heat flux can be highly variable between, 0.5 and 2.3 W m$^{-1}$ °C$^{-1}$ and is transferred from the shallow magmatic intrusion mainly by conduction. The surface temperature in the soil is generally between 20 and 30°C, but the temperature of gas fumaroles can be more than 250°C.

The hydrothermal fluids are a mix of hot gas and “cooler” water, with chemical composition dominated, in decreasing order, by H$_2$O, CO$_2$, S$_{total}$ and in trace elements by Ar, O$_2$, N$_2$, CH$_4$, H$_2$, He, HCl and HF. The pH of fluids is typically more and less acid (0-7 pH), but below that of meteoric water. Hydrothermal systems can occur as a series of fluid convection cells with both
vertical and horizontal fluid transfer. Fluid circulation is controlled both by faults and by the porosity and impermeability of the rock.

The evolution of the hydrothermal system is controlled mainly by self-sealing phenomenon, which generates a major modification in the shape and structure of system as well as fluid transfer and convection type.

10.2 Introduction

Hydrothermal systems are linked with areas that have a geothermal gradient hotter than the average geothermal gradient of the crust. The hydrothermal system is composed mainly of rock, minerals and fluids (hot and cool). The main area of high geothermal gradients are more often located in volcanic zones, such as oceanic ridges, volcanic arcs, within any volcanoes and also where the upper crust thins (e.g., grabbens or impact crater structures).

What are the types of fluid present inside the hydrothermal system? How does the fluid move inside the hydrothermal system? What is the type of heat transfer? What is the chemical composition of fluids? The hydrothermal system being an active process, what is the evolution of hydrothermal system? How long is the life of the hydrothermal system?

The physical and chemical processes, which take place inside the hydrothermal system, are principally studied by measurements of gas composition as it passes through the soil or fumaroles, temperature measurements [Ghidini et al., 2005], self-potential measurements [Finizola et al., 2002; Yasukawa et al., 2003] and altered rock composition [Osinski et al., 2001; Stimac et al., 2004]. The physical and chemical processes in the hydrothermal system will here be approached by analysis of 5 case studies in order to obtain a better knowledge of the hydrothermal system.
10.3 Structure of the hydrothermal system at several convection levels: example of Stromboli volcano, Italy

Temperature, fluid and gas are the main elements that compose the hydrothermal system. Traditionally, in volcanic areas, the hydrothermal structure is determined by Resistivity, Self-potential and CO₂ degassing measurements (Fig. 10-1), which allows one to constrain the spread of the hydrothermal system.

The Self-potential (SP), which is an electrical method, allows characterization of the circulation of subsurface fluids [Finizola et al., 2002]. The positive SP anomalies are due to upward movement of hot hydrothermal fluids, while the negative SP anomalies characterize infiltration of meteoric water.

The CO₂ degassing measurement is a geochemical method, which measures the concentration (or flux) of CO₂ in the ground. CO₂ is generally the first gas to exsolve from the magma, as it rises to the surface.

Finizola et al. [2002] describe the hydrothermal system of Stromboli volcano, as a composite hydrothermal system that is composed of 4 systems: 3 major and 1 minor (Fig. 10-1 and 10-2). The limits of these systems are controlled by the structural faults of the volcano, which consist of craters, flank landslides and ancient collapsed calderas. The Self-potential and CO₂ degassing measurements allow determination of different rates and type of flow within the hydrothermal system.

The first type of limit is the fault zones which have a high permeability. These are characterized by the downward transfer of fluids controlled by gravity and by the rise up of magmatic gas (H₂O, CO₂). The SP signal is marked by negative anomalies and the CO₂ concentration is high [Finizola et al., 2002].

The second type of process is the self-sealing of the hydrothermal system. The transfer of hot mineral-rich fluids through the rock generates a strong alteration, with the formation of hydrothermal minerals (e.g., calcite, gypsum, etc.). The first effect of alteration is the increase in the permeability, but when the alteration is very important, the altered rock will become impermeable. The self-
sealing is characterized by a positive SP anomaly and by absence of magmatic CO₂ gas.

The existence of high heat flow in the subsurface (70ºC at 30 cm depth) [Finizola et al., 2002] and the important presence of hydrothermal minerals on the ground, shows the existence of hydrothermal fluid convection cells. Theses fluids are supplied by meteoric water, which increase in temperature as the sink.

The transfer of heat, by conduction through the impermeable layer, is due to the gas and fluid from the deep hydrothermal system below the impermeable layer.

The study of Stromboli volcano by Finizola et al. [2002] shows that the distribution of the hydrothermal system is not focused on the active crater, but is shifted to areas where the permeability is important. The internal heat source is therefore not the magmatic conduit of the active crater, but rather due to a shallow magmatic intrusion from the conduit. The hydrothermal system in Stromboli volcano is complex and structured with 4 distinct hydrothermal systems (Fig. 10-3), which are either supplied only by meteoritic water or from a combination of meteoritic water and magmatic gases and fluids. Figure 10-3 shows the structure of hydrothermal system determined by Finizola et al. [2002].

10.4 Life of a hydrothermal system: example of the Impact crater of Haughton, Northern Canada.

The impact of meteors generates high energy with the transformation of kinetic energy into heat. The three potential sources of heat for hydrothermal system formation are the shock-generated melt, the hot breccias and the elevated geotherms due to target rock uplift (~2 km for Haughton crater) [Osinski et al., 2001].

The impact crater of Haughton, ~ 23.4 ± 1.0 Ma [Omar et al., 1987], was originally ~20 km in diameter [Robertson and Sweeney, 1983] and more than 1750 m deep [Osinski et al., 2001]. The crater structure is composed of a faulted annulus (radius ~5.5 to 8.5 km), formed of interconnected concentric and radial faults. Inside the crater, the lithology is composed at depth of gneiss, limestone, sandstone, covered by lithic clasts (formed from clasts of underlying rocks) and on the top by a series of post-impact lacustrine sediments.
More than three types of intra-breccia alteration are present: sulphide and carbonate mineralization, sulphate mineralization and carbonate mineralization. For more details see Osinski et al. [2001]. These alteration minerals are strong evidence for the presence of a hydrothermal system in the Haughton crater. Osinski et al. [2001], propose a post-impact hydrothermal model, defined by three distinct stages: Early, Main and Late.

The Early stage begins immediately after the impact. The temperature in the rock was at least 650°C and the trapped pore water from uplifted sedimentary rock reacted with hot polymict impact breccias to induce the first mineral precipitation. During this stage, the hydrothermal system was dominated by vapour at a temperature of 236°C (maximum enthalpy of dry stream) [Nicholson, 1993]. However, beneath the hot breccia (at deeper levels), the hydrothermal system is composed of vapour and liquid. With time, there was a rapid decrease in confining pressure and temperature (to ~340°C) in the system, due to water infiltration through the ring fault system [Osinski et al., 2001]. The principal mineral formed during this first stage is quartz.

The Main Stage is characterized by a temperature between 100 to 200°C, due to the progressive cooling of the polymict impact breccias. The hydrothermal system was composed of both vapour (mainly CO₂ and H₂S) and liquid phase. The hydrothermal system was characterized by a simple convective system. During this stage, the principal minerals to form were calcite, celestite, barite and marcacite [Osinski et al., 2001].

The Late Stage is characterized by temperatures below 100°C and the final hydrothermal alteration of breccias is marked by selenite and calcite. The hydrothermal fluids are cooler, more oxygenated, due to the increased contribution of groundwater [Osinski et al., 2001].

The hydrothermal system of an impact crater is a good example of the life and death of a hydrothermal system, unfortunately, the duration of the Haughton hydrothermal system is still unknown. Many other examples of deep
hydrothermal systems exist among the studies on geothermal reservoirs [e.g., Stimac et al., 2008, Acuna et al., 2008].

**10.5 Hydrothermal pathways: example of fluid flow patterns**

Traditionally, the fluid pathways in volcanic hydrothermal systems are interpreted as vertical movement downward and upward. The example of Waita volcano is not conventional. Waita volcano, however, is located in south-west Japan, central Kyushu island, in the western part of the Hohi geothermal region.

![Figure 10-4: Geographic setting of Waita volcano, Kyushu Island, Japan. Green dashed line is Takenoyu fault and the red line is the approximate location of the 1995-1996 SP profiles [Yasukawa et al., 2003] See in Chapter 5 on Figure 5-4 and Fig. 10-7.](image-url)
Yasukawa et al. [2003] developed a model of hydrothermal flow within the Waita volcano, which shows discharge flow through the western flank of the volcano and recharge by meteoric water on the eastern flank. The study of Yasukawa is based on a numerical block model of the fluid flow from Self-potential [Yasukawa et al., 2003] (Fig. 10-4) and resistivity surveys [Widarto et al., 1992]. The resistivity surveys show that the ground resistivity is very weak (>= 300 Ohm m) and thus the permeability is high (0.1 to 30 mD = 0.1 to \(30 \times 10^{-15}\) m²). Figure 10-5 shows the final model of Yasukawa et al. [2003].

![Figure 10-5: Top: fluid velocity distribution model (Model 3, case 2, from Yasukawa et al. [2003]). Bottom: Self-potential profile of Waita volcano, Japan. A'-B-C is a NW-SE profile across the summit. Modified from Yasukawa et al. [2003].](image-url)
The hydrothermal system is recharged by meteoric water from the eastern flank (Fig. 10-5). Fluids then move through the volcano and their temperature increases by conduction with the hot rock and perhaps also by mixing with deep hot hydrothermal fluids. Finally, the hot hydrothermal fluids are discharged through the western flank to the surface, where many heat sources are found. However, although the choice of models parameters can be debated, as well as the self-potential measurement and their source, the model is close enough to reality and the location of the heat source (more than 20 source) on Waita volcano flanks [Yasukawa et al., 2003].

In volcanic areas, the displacement of hydrothermal fluids can be other than just vertical. A hydrothermal system is a complex system and fluid movement is mainly controlled by the permeability of the rock and faults [Yasukawa et al., 2003, Teng and Koike, 2007]. In the example of Waita volcano, the faults are principally regional and are parallel to the SP profiles, which allows development of models which are independent of structural influences.

### 10.6 Hydrothermal minerals and importance of porosity and permeability on hydrothermal reservoir formation: Example of Tiwi geothermal field.

Hydrothermal fluids move through the rock via spaces, which are formed by the open pores and fractures present within the rock. Porosity and permeability of any rock is controlled by the chemical and mineral composition of the rock and by the temperature and the pressure in and around the rock.

The Tiwi geothermal field is located on the north-eastern flank of Mt. Malinao, within the Bicol volcanic arc in south-eastern Luzon, Philippines, [Gambill and Beraquit, 1993]. The lowermost reservoir in the Tiwi geothermal field, is composed (~ 600 m thick) of intercalated mudstone, limestone and
andesitic wacke, which are underlain by hot (>315°C) schist (quartz-muscovite) basement rock. The geothermal field is cut by four faults oriented NE-SW (Fig. 10-8), which are the Kagumihan (semi-permeable barrier), Tiwi (delimits the area into two pressure domains: low and high) and Takla and Naglagbong faults (main conduit for meteoric recharge) [Protacio et al., 2001]. On the top and side of the field, where the temperature is low, the main alteration mineral is smectite. In the hydrothermal reservoir, the main minerals are chlorite, quartz, illite and epidote. In the southwest of field, advanced argillic alteration is characterized by alunite, diaspore, pyrophyllite, anhydrite and pyrite [Stimac et al., 2004].

The life of the Tiwi geothermal field can be organized into three main stages. The first stage began more than 279,000 years ago, with an initial episode of intrusive activity with high temperatures (>300°C), boiling fluids, hydraulic fracturing and formation of chalcedinite, sericite, quartz, adularia, epidote and pyrite.

The second stage began 200,000 years ago and was characterized by a long period of quiescence, with a temperature about 235°C [Moore et al., 2000]. The current stage began between 50,000 and 10,000 years ago with a new episode of intrusion, and current activity is marked by a drop in fluid pH, high CO₂ content in the hydrothermal system, deposition of sericite, illite and deposition of wairakite, epidote, calcite and actinolite.

Stimac et al. [2004] showed by petrographic image analysis that the porosity declines with depth in the Tiwi geothermal field. The decrease of porosity is due to different physical and chemical factors. The chemical reactions between the hot hydrothermal fluids and the rock are the hydrothermal mineral deposition and the dissolution of rock. The pore characteristics are controlled by the dissolution of the rock and the type of hydrothermal mineral deposition.

The physical decrease of porosity is due to compaction by overburden pressure of the bedrock due to burial and reaction with hot fluids. The porosity decreases from 10% to 2.5% and the breccias and sandstone are more affected by this phenomenon than the lava [Stimac et al., 2004].
The permeability is controlled by the shape and the size of pores, by faulting and the porosity. As a function of depth, the relationship between open fracture frequency, pore type and porosity is not the same. However, at shallow depths, the porosity is inversely correlated with the open fracture frequency. At 2000 m depth (middle depth), the correlation between them is normal. However, in the deepest section both porosity and open fracture frequency are uncorrelated. Stimac et al. [2004] suggest that the inverse correlation at shallow sections of the hydrothermal reservoir (1000-2500 m) is due to illite sealing of permeability zones. In the deepest section, this inverse correlation is due to sericitization which increases the impermeability.

In fact permeability and porosity, which allows the movement of hydrothermal fluids, are controlled by the type of depositing mineral, which themselves are controlled by the reaction between the fluid and rock compositions. Illite and sericite are mainly deposited since the renewal of activity that began between 50,000 and 10,000 years ago. The Tiwi geothermal field is currently in a phase of self-sealing on the top and on the bottom of reservoir and the hydrothermal fluid reservoir is locked in the middle section.

10.7 Chemical composition, gas flux and thermal flux of a hydrothermal system.

The chemical composition of a hydrothermal system is controlled by both rock and fluid compositions. The rock composition may be highly variable, as the hydrothermal system can be present in any type of bedrock, but igneous rocks are always presents, because magmatic intrusion are generally the source of heat. The composition of hydrothermal fluids is also variable, as seen with Tiwi geothermal field and Impact crater of Haughton. The hydrothermal fluid comes from mixing of liquid and/or vapour [Osinski et al., 2001, Finizola et al., 2002, Chiodini et al., 2005, Boyce et al., 2007].
Chiodini et al. [2005] investigate 8 different volcanic systems to determine a budget of flux (gas and thermal), and chemical composition of hydrothermal system to estimate heat released from volcanic and hydrothermal systems. The temperatures of fumaroles for these 8 volcanic systems are between 68 and 197°C, whereas the temperature on the soil surface is between 20 and 30°C, but at 50 cm it can reach as high as 95 °C (Fig. 10-6). Consequently, the majority of gas is condensed in the ground and the heat is transferred by conduction. The thermal gradient in the ground of a volcanic system is between 34 and 405°C m⁻¹. The composition of the hydrothermal gas was measured at the hottest fumaroles. The main source of the hydrothermal gas comes from the degassing of magma at depth. The main gases are, in decreasing order, H₂O (33 to 100% by vol), CO₂ (1 to 16%), S_{total} (0.0003 to 0.3524%). The other minor gases are (in trace amounts), in decreasing order, Ar, O₂, N₂, CH₄, H₂, He, HCl and HF, for the values seen in Chiodini et al. [2005].

The gas flux measured on the different volcanic systems is between 2.69 x 10⁻⁵ and 2.08 x 10⁻² mol m⁻² s⁻¹ [Chiodini et al., 2005], while the liquid (H₂O) flux is between 1.55 x 10⁻³ and 1.21 x 10⁻² mol m⁻² s⁻¹ and the CO₂ flux is between 2.16 x 10⁻⁵ and 1.08 x 10⁻³ mol m⁻² s⁻¹. The molar ratio X_{H₂O}/X_{CO₂} varies over a large range between 0.5 and 410.

After the models of Chiodini et al. [2005], the transfer of heat in the hydrothermal system is dominantly by conduction, with an output about of 98.4%; the rest is by advection. In the soil, the average thermal conductivity is 1.14±0.4 W m⁻¹ °C⁻¹ (Fig. 10-7), but varies between 0.5 and 2.3 W m⁻¹ °C⁻¹.

In fact, the movement of hydrothermal fluids is controlled by the phase of the fluid. The vapour phase rises, while the movement of steam (H₂O_{liq}) is downward. The steam condenses near the surface of the soil and the heat, which is generated by condensation, is transferred by conduction to the surface with a linear temperature gradient [Chiodini et al., 2005].
Figure 10-6: Simulated and measured temperature of soil at Solfatara in Campi Flegrei caldera (Italy), Stefano hydrothermal crater on Nisyros volcanic island (Greece) and solfatara of Comalito cinder cone at Masaya volcano (Nicaragua, see section 1.2 of chapter 1). After Chiodini et al. [2005]. Reprinted from Journal Geophysical Research, vol. 110, Chiodini, G., D. Granieri, R. Avino, S. Caliro, A. Costa, and C. Werner, “Carbon dioxide diffuse degassing and estimation of heat release from volcanic and hydrothermal systems”, Pages B08204, doi:10.1029/2004JB003542, Copyright (2005) AGU.
Figure 10-7: Estimation of thermal conductivity (K in W m\(^{-1}\) °C\(^{-1}\)) for 30 surveys at 15 different volcanic systems. K is linear among the measured ratio \(\nabla T/\phi_{\text{Gas}}\) and the log ratio \(X_{\text{H}_2\text{O}}/X_{\text{GAS}}\) of the fumaroles fluids. Error bars refer to an uncertainty of ±30% in the estimation of \(\nabla T/\phi_{\text{Gas}}\) and to the effect of 50% condensation of the original steam. After Chiodini et al. [2005]. \(\nabla T\) is the thermal gradient in °C m\(^{-1}\). \(\phi_{\text{Gas}}\) represent gas flux in mol m\(^{-2}\) s\(^{-1}\). Reprinted from Journal Geophysical Research, vol. 110, Chiodini, G., D. Granieri, R. Avino, S. Caliro, A. Costa, and C. Werner, “Carbon dioxide diffuse degassing and estimation of heat release from volcanic and hydrothermal systems”, Pages B08204, doi:10.1029/2004JB003542, Copyright (2005) AGU.

10.8 Discussion and Conclusion

Shallow hydrothermal systems are most often located in volcanic areas. Their shape, structure and level of activity are directly influenced by the shape, structure and activity of volcano or volcanic field.
Hydrothermal systems are composed of rock, altered rock, hot and cool aqueous liquids and gas. The system is recharged mainly by meteoric water and magmatic gas. The temperature inside the hydrothermal system is highly variable, from one system to other, and between the different elements (rock, gas and liquid). The hottest is the gas, which can be $\geq 250^\circ$C for the fumaroles. While on the surface, the temperature of the soil is close to 20 to 30$^\circ$C, in the ground, the thermal gradient can be very high, between 34 and 405 $^\circ$C m$^{-1}$. This variation of temperature is mainly caused by the depth of condensation of gas to liquid. The rising fluids are hot and gas rich, which condense near the surface (less than 1 m from the surface) while the denser liquid sinks by gravity. In fact, the heat flow is transferred mainly from depth to the surface by conduction, with the exothermal reaction of condensation of gas flux and the value of heat flux can be very variable 0.5 and 2.3 W m$^{-1}$ $^\circ$C$^{-1}$.

The structure and shape of the hydrothermal system can be very complex, and controlled mainly by lithology and faults (regional and local). With time, modification of the chemical composition of the rock by the hydrothermal fluid, which causes a strong interaction with the rock, strongly modifies the shape and the structure itself. The hot (and more or less acid) fluids dissolve and change the mineral compositions and the propriety of rocks at depth on or near the surface. The deposition of hydrothermal minerals will initially increase the permeability and porosity of the rock, allowing a better circulation of fluid. However, with significant increase in concentration of impermeable hydrothermal minerals in the rock, the permeability will decrease, until it forms a self-sealing layer within hydrothermal system.

The chemical composition of fluids are by decrease order: H$_2$O (33 to 100% in vol), CO$_2$ (1 to 16%), S$_{\text{total}}$ (0.0003 to 0.3524%) and in trace elements: Ar, O$_2$, N$_2$, CH$_4$, H$_2$, He, HCl and HF. The pH of the fluid (gas and water) is controlled by its chemical composition and an increase CO$_2$ concentration in the fluid will generate a drop of it pH.

The hydrothermal minerals are mainly: Illite and sericite, followed by wairakite, epidote, calcite and actinolite, halcedinites, sericite, quartz, adularia,
epidote and pyrite. The type of mineral deposit is controlled primarily by the chemical composition of the original rock fluid chemistry and temperature.

The hydrothermal system can have a single level of fluid convection: the fluid is vaporised at depth and rises until the sub-surface where it condenses. The steam water moves downward across the fault and the rock and is recharged with gas at depth. The hydrothermal system can also be constituted by several convection levels, which are separated by self-selling impermeable layers. Furthermore, the flux orientation of the hydrothermal fluid can be horizontal, vertical or a combination of both.

The “life” of a hydrothermal system is controlled by the influx of heat from depth from magmatic intrusions and may be present for many tens of thousands of years after cooling of the heat source.

10.9 References


Finizola, A., F. Sortino, J.F. Lénat, and M. Valenza (2002), Fluid circulation at Stromboli volcano (Aeolian Islands, Italy) from self-potential and CO₂


Teng, Y., and K. Koike (2007), Three-dimensional imaging of a geothermal system using temperature and geological models derived from a well-log


### 11 APPENDICES A3: GENERAL OVERVIEW OF MATHEMATICAL BACKGROUND

#### 11.1 Mathematical Conventions

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\phi$</td>
<td>potential field</td>
</tr>
<tr>
<td>$\Delta \phi$</td>
<td>variation of the potential field</td>
</tr>
<tr>
<td>$\varphi$</td>
<td>electrical potential</td>
</tr>
<tr>
<td>$\varphi_0$</td>
<td>electrical potential in $z = 0$</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>electrical charge</td>
</tr>
<tr>
<td>$t$</td>
<td>time</td>
</tr>
<tr>
<td>$\vec{x}$</td>
<td>translation vector</td>
</tr>
<tr>
<td>$\vec{z}$</td>
<td>vertical vector</td>
</tr>
<tr>
<td>$\nabla^2$</td>
<td>Laplacian $\frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2}$</td>
</tr>
<tr>
<td>$D$</td>
<td>Space of distribution</td>
</tr>
<tr>
<td>$\mathbb{R}$</td>
<td>Real space</td>
</tr>
<tr>
<td>$\mathbb{R}^*$</td>
<td>Real space without zero</td>
</tr>
<tr>
<td>$\mathbb{R}^+$</td>
<td>Real space with only positive number and zero</td>
</tr>
<tr>
<td>$\mathbb{R}^2$</td>
<td>Real space in two dimensions</td>
</tr>
</tbody>
</table>
Complex space

$P$  Poisson kernel symbol

$\hat{P}$  Poisson kernel symbol into Fourier Space

$L^2$  Fourier Space in two dimensions

$FT(f)$  Fourier transform of the function $f$

$i$  imaginary number ($i^2 = -1$)

$LT$  Laplace transform

$LT^{-1}$  Laplace transform inverse

$F'$  is the first derivative of the function $F$

$F''$  is the second derivative of the function $F$

$e$  is the exponential

$D_z$  dilation operator

$\int$  integrals in one dimension

$\iint$  integrals in two dimensions

$\langle \cdot, \cdot \rangle$  is the Cauchy number

$\| \|$  is the norm

$| |$  is absolute value
11.2 Fourier Transform setting

11.2.1 General equation

Fourier Transform is a mathematical operation, which allows us to express spatial or time-series data in the frequency domain [Roddier, 1971; Max, 1985].

There are two main types of Fourier transform:
- Discrete Fourier Transformation
- Continuous Fourier Transform

For the purpose of this thesis only the Continuous Fourier Transform is presented.

Continuous Fourier Transform can be expressed as
\[ \forall u \in \mathbb{R} \text{ and } f : \mathbb{R} \rightarrow \mathbb{C} \]
where:

\[
\hat{f}(u) = FT\left[ f(x) \right] = \int_{-\infty}^{+\infty} f(x)e^{(-2\pi i xu)} \, dx
\]  

(A3-1)

Inverse Continuous Fourier Transform can be expressed as \( \forall u \in \mathbb{R} \) and \( \hat{f} : \mathbb{C} \rightarrow \mathbb{R} \), where:

\[
 f(x) = FT^{-1}\left[ \hat{f}(u) \right] = \int_{-\infty}^{+\infty} \hat{f}(u)e^{(2\pi i xu)} \, du
\]  

(A3-2)
11.2.2 Main property of Fourier Transform

For $f$, $g$, $h$, three integrable functions, and $\hat{f}, \hat{g}, \hat{h}$, their Fourier transform [Max, 1985; Roddier, 1971]. $\forall x \in \mathbb{R}, \forall u \in \mathbb{C}$ There is $x \rightarrow u$.

11.2.2.1 Linearity

$\forall a, b \in \mathbb{C}$ and if $af(x) + bg(x) = h(x)$, thus:

$$a \hat{f}(u) + b \hat{g}(u) = \hat{h}(u) \quad (A3-3)$$

11.2.2.2 Translation

$\forall x_0 \in \mathbb{R}$ and if $f(x - x_0) = h(x)$, thus:

$$\hat{h}(u) = \hat{f}(u)e^{-2\pi iux_0} \quad (A3-4)$$

11.2.2.3 Modulation

$\forall u_0 \in \mathbb{R}$ and if $h(x) = e^{2\pi iux_0}f(x)$, thus:

$$\hat{h}(u) = \hat{f}(u - u_0) \quad (A3-5)$$

11.2.2.4 Scaling

$\forall a \in \mathbb{R}^*$ and if $h(x) = f(ax)$, thus:
\[ \hat{h}(u) = \frac{1}{|a|} \hat{f}\left(\frac{u}{a}\right) \]  

(A3- 6)

if \( a = -1 \), thus :

- \( h(x) = f(-x) \)  

(A3- 7)

and

- \( \hat{h}(u) = \hat{f}(-u) \)  

(A3- 8)

11.2.2.5 Conjugation

If \( h(x) = \overline{f(x)} \), then \( \hat{h}(u) = \overline{\hat{f}(-u)} \)  

(A3- 9)

11.2.2.6 Convolution

If \( h(x) = (f \ast g)(x) \), then \( \hat{h}(u) = \hat{f}(u) \ast \hat{g}(u) \)  

(A3- 10)

11.3 Laplace Equation

11.3.1 Definition

The Laplace transform is noted here as: \( LT \)

\[ \forall t \in \mathbb{R}^+ \, , \, \forall p \in \mathbb{C}^+ \, , \, \text{such that} \, \, p = a + i\omega \, , \, \text{with} \, \, \forall a, \omega \in \mathbb{R} \]

For a function, \( f \), such as \( \forall t < 0 \, , \, f(t) = 0 \), there is a complex function, \( g \), which
is its Laplace transform and noted as [Max, 1985; Roddier, 1971]:

\[
g(p) = LT \left[ f(t) \right] = \int_{0}^{+\infty} f(t) e^{-pt} dt
\]

(A3- 11)

The Laplace transform is an extension of Fourier transform within the Conjugate domain.

This can be proven by doing a change of variable.

- If \( P \) is a complex number, such as: \( p = a + i\omega \), when its real part is null \( a = 0 \), \( p = i\omega \) and if \( \hat{F} \) is the Fourier Transform of the function \( f \).

The Laplace transform \( g(i\omega) \) of \( f(t) \) is equal at the Fourier transform \( \hat{F}\left(\frac{\omega}{2\pi}\right) \) of \( f(t) \).

\[
g(p) = g(i\omega) = \int_{0}^{+\infty} f(t) e^{-i\omega t} dt = \int_{0}^{+\infty} f(t) e^{-\frac{i\omega}{2\pi} \frac{2\pi}{2\pi} dt} = \hat{F}\left(\frac{\omega}{2\pi}\right)
\]

(A3- 12)

The equality is true if \( p = a + i\omega \), with \( a = 0 \)

\[
f(t) \xrightarrow{LT} g(i\omega) \iff \hat{F}\left(\frac{\omega}{2\pi}\right) \xrightarrow{FT^{-1}} f(t)
\]

(A3- 13)

If \( p = a + i\omega \), with \( a \neq 0 \), the equation becomes:
\[ g(a + i\omega) = \int_{0}^{+\infty} f(t) e^{-(a+i\omega)t} dt \]  

(A3- 14)

\[ = \int_{0}^{+\infty} f(t) e^{-(a+i\omega)t} dt = \int_{0}^{+\infty} e^{-at} f(t) e^{-i\omega t} dt \]  

(A3- 15)

### 11.3.2 Laplace Transform inverse

Laplace transform inverse is noted here as: \( LT^{-1} \)

\( \forall t \in \mathbb{R}^+, \forall p \in \mathbb{C}^+ \), such as \( p = a + i\omega \), with \( \forall a, \omega \in \mathbb{R} \)

For a function, \( f \), such as \( \forall t < 0, f(t) = 0 \), there is a complex function, \( g \), which is its Laplace transform and noted as [Max, 1985; Roddier, 1971]:

\[ f(t) = LT^{-1} \left[ g(p) \right] = \frac{1}{2i\pi} \lim_{i \to \infty} \int_{a_0-i\infty}^{a_0+i\infty} g(p) e^{pt} dp \]  

(A3- 16)

Where \( a_0 \in \mathbb{R} \) and \( a_0 > a \), as defined above. So that the contour path is within the region of convergence of \( g(p) \). In the case of \( a_0 = 0 \) and if \( a \in \mathbb{R}^+ \), thus the equation is similar at the inverse Fourier transform (equation A3-13).

### 11.3.3 Properties of Laplace transform

The Laplace transform has the same properties as the Fourier transform (see above) for [Max, 1985; Roddier, 1971]:

i. linearity
ii. change of unity
iii. translation
iv. convolution
v. derivative
vi. integrable

11.4 Homogeneous functions

If $V, W$ are two vectorial spaces and $f$ is a function between these two vectorial spaces and if $F$ is a field as:

$$f : V \rightarrow W$$

then $f$ is a homogeneous function of degree $n$, if the following equation is true:

$$f(\alpha x) = \alpha^n f(x)$$

where:

- $\alpha \in F$
- and
- $x \in V$

For example, if $g$ is a homogeneous function of degree 5 and if $\forall x \in V, \forall y \in V$ with $g : V \rightarrow W$, then:

$$g(\alpha x, \alpha y) = \alpha^5 g(x, y) \quad \text{(A3-17).}$$

1.1.1 The distribution function

A distribution or generalized function is a mathematical object that generalizes functions and probability distributions. The properties presented below are from Roddier [1971], but can be found in any advanced mathematics text.

It is important to recall these properties in order to understand the Hilbert
transform.

For notation:

\[ \langle T, \varphi \rangle \] is the scalar that results from the distribution \( T \) of a function \( \varphi \).

The main properties are:

1.1.1.1 Addition

For a locally summation function \( f(x), g(x) \), the distribution respects the addition operator.

\[
\langle f + g, \varphi \rangle = \int (f + g) \varphi \, dx = \langle f, \varphi \rangle + \langle g, \varphi \rangle
\]  \hspace{1cm} (A3-18)

1.1.1.2 Multiplication by a scalar

For any scalar \( \lambda \in \mathbb{C} \), thus:

\[
\langle \lambda T, \varphi \rangle = \lambda \langle T, \varphi \rangle
\]  \hspace{1cm} (A3-19)

1.1.1.3 Translation of the distribution

For the local summation function \( T(x) \) and a constant \( a \), the translated of \( T(x) \) is \( T(x-a) \), as:

\[
\langle T(x-a), \varphi(x) \rangle = \langle T(x), \varphi(x+a) \rangle
\]  \hspace{1cm} (A3-20)

1.1.1.4 Transposition of the distribution

For the local summation function \( T(x) \), the distribution of \( T(-x) \) is:
\[ \langle T(-x), \varphi(x) \rangle = \langle T(x), \varphi(-x) \rangle \]  \hspace{1cm} (A3-21)

1.1.1.5 Unit Replacement

A variable \( x \), which is expressed in one unit, can be changed to another unit by multiplying \( x \) by a constant \( a \neq 0 \). For the local summation function, \( T(x) \):

\[ \langle T(ax), \varphi(x) \rangle = \frac{1}{a} \langle T(x), \varphi\left(\frac{x}{a}\right) \rangle \]  \hspace{1cm} (A3-22)

**Note:** This property will be useful when applying the dilation on a wavelet.

1.1.1.6 Multiplication of the distribution

For a local summation function \( f(x), g(x) \), the product will not always be a local summation function. \( D \) is the space of distribution where within it, any function is indefinitely derivable.

For the local summation function \( T(x) \) and a function \( \psi \), which is infinitely derivable, \( \forall \varphi \in D \), the multiplication is:

\[ \langle \psi T, \varphi \rangle = \langle T, \psi \varphi \rangle \]  \hspace{1cm} (A3-23)

thus \( \forall \psi \varphi \in D \).
11.4.1 Derivative of the distribution

*Main property:* Any distribution is infinitely derivable

For the local summation function $T(x)$ and for $\forall m \in \mathbb{N}$, thus:

$$\langle T^{(m)}, \varphi \rangle = (-1)^m \langle T, \varphi^{(m)} \rangle$$  \hspace{1cm} (A3-24)

11.4.2 Cauchy value

*The Cauchy Value is noted as v.p. and is used to define the principal value of a distribution.*

Cauchy values are defined as:

$$\left\langle \text{v.p.} \frac{1}{x}, \varphi(x) \right\rangle = \text{v.p.} \int_{-\infty}^{\infty} \varphi(x) \frac{dx}{x}$$  \hspace{1cm} (A3-25)

11.4.3 Sign of the distribution

The sign of a distribution $x$ is noted $\text{sgn} \, x$:

$$\text{sgn} \, x = \frac{|x|}{x}$$  \hspace{1cm} (A3-26)
11.5 Hilbert Transform definition

11.5.1 Overview

The Hilbert Transform is a mathematical tool based on the properties of the distribution (see previous section) [Roddier, 1971].

The Hilbert Transform, $HT$, is a linear transformation, which allows us to create the imaginary part of any pure real number function. The Hilbert Transform and the inverse Hilbert Transform ($HT^{-1}$) are an anti-involution:

$$HT^{-1}(HT(u)) = -u \text{ [Bracewell, 1966].}$$

Notation:

For any function, $f$, there is a function, $F$, which is the Hilbert Transform of $f$ [Roddier, 1971].

For $g$ a function, $\forall t \in \mathbb{R}^+$

If $\tilde{F}(t) = HT[f(t)] = g(t)$, thus $HT^{-1}[g(t)] = f(t) \text{sgn } t$ \hspace{1cm} (A3-27)

In the Spatial domain:

The Hilbert Transform $\tilde{f}$ of a function, $f$, is $\forall t \in \mathbb{R}^+$, equals [Roddier, 1971]:

$$\tilde{f}(t) = HT[f(t)] = \frac{1}{\pi t} * f(t) \hspace{1cm} (A3-28)$$

In the Frequency domain, the Hilbert Transform $\tilde{F}$ of the Fourier Transform $\hat{F}$ of the function, $f$, is [Roddier, 1971]:

$$f \xrightarrow{FT} \hat{F} \xrightarrow{HT} \tilde{F}, \forall u \in \mathbb{C}^+, $$
\[ HT[\hat{F}(u)] = -i \cdot \text{sgn}(u)\hat{F}(u) \]  

(Causal signal definition):
A causal signal is a signal where:
\[ \forall t < 0, \ f(t) = 0 \]
\[ \forall t \geq 0, \ f(t) \neq 0 \]

(Theorem about linear filter):
A necessary and sufficient condition to ensure that a linear filter becomes causal (which can be physically generated) is that the imaginary part of its transfer function is the Hilbert Transform of its real part.

In Equation:
\[ \forall t \geq 0, \ FT[f(t)] = F(v) \]  

(A3-30)

- \[ \text{Im} F(v) = -TH\left[ \text{Re} F(v) \right] \]  
  and  
- \[ \text{Re} F(v) = TH\left[ \text{Im} F(v) \right] \]  

(A3-31)

(A3-32)

11.5.2 Properties of HT

11.5.2.1 Linearity
For two functions [Roddier, 1971], \( f \) and \( g \), \( \forall t \geq 0 \):

\[ HT\left(af(t) + bg(t)\right) = aHT(f(t)) + bHT(g(t)) \]  

(A3-33)
The sum of the Hilbert Transform of two functions is equal to the Hilbert Transform of the sum of the two functions.

11.5.2.2 Convolution product

For two functions, \( f \) and \( g \), \( \forall t \geq 0 \):

\[
HT(f(t) \ast g(t)) = HT(f(t)) \ast HT(g(t))
\]  (A3-34)

The convolution product of the Hilbert Transform of two functions is equal to the Hilbert Transform of the convolution product of the two functions.

11.6 Hilbert Space

Hilbert space is any real or complex inner product space: \( \langle y | x \rangle \)

where:

- \( \langle \ | \ \rangle \) is the Dirac notation, as in Quantum mechanics.
- \( x, y \) are the element of the vectorial space: \( L_2 \), and is a finite-dimensional inner product space and square-summation.

11.7 Hermitian matrix

Notation of the relationship between two elements of the vectorial space and their conjugate inside vectorial space.

\[
\langle y | x \rangle = \overline{\langle x | y \rangle}
\]

where \( \overline{\langle x | y \rangle} \) is the complex conjugate of \( \langle y | x \rangle \)

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11.7.1 Banach space

Banach space is a complete vector space, which include Real $\mathbb{R}$ and Complex $\mathbb{C}$ number with a norm.

The main property is on the operation on the vector:

$$\|x + y\|^2 + \|x - y\|^2 = 2\left(\|x\|^2 + \|y\|^2\right)$$  \hspace{1cm} (A3- 35)

- For a Real space

$$\langle y|x \rangle = \frac{1}{4}\left(\|x + y\|^2 - \|x - y\|^2\right)$$  \hspace{1cm} (A3- 36)

ii. For a Conjugate space

$$\langle y|x \rangle = \frac{1}{4}\left(\|x + y\|^2 - \|x - y\|^2 + i\|x + iy\|^2 - i\|x - iy\|^2\right)$$  \hspace{1cm} (A3- 37)

11.7.2 Minkowsky inequality

The modulus of the sum is smaller than the sum of the modulus.

$$\|x + y\| < \|x\| + \|y\|$$  \hspace{1cm} (A3- 38)

11.7.3 Reflexivity:

$$\varphi(x) = \langle u|x \rangle$$  \hspace{1cm} (A3- 39)
where if \( x \in L_2 \) and if \( u \) is the result of \( \varphi(x) \), thus \( u \in L_2 \).

### 11.7.4 Fourier Transform of Hilbert space

i. **Fourier transform of a function in Hilbert space**

\( \forall f(x) \in L_2 \), the Fourier transform of \( f(x) \) is \( \hat{f}(u) \) [Max, 1985; Roddier, 1971], with \( \hat{f}(u) \in L_2 \) and:

\[
\int |f(x)|^2 \, dx = \int |\hat{f}(u)| \, du
\]

(A3-40)

ii. **Fourier transform of a scalar of two functions in Hilbert space**

\( \forall f(x) \in L_2 \) and \( \forall g(x) \in L_2 \), their Fourier transform \( \hat{f}(x) \) and \( \hat{g}(x) \) are also inside the Hilbert space:

- \( \forall \hat{f}(x) \in L_2 \)
- \( \forall \hat{g}(x) \in L_2 \)

### 1.2 References:


12.1 Introduction

Masaya volcano (11.984°N, 86.161°W, 635 m) is a basaltic shield volcano in Nicaragua, which is characterized by a continuously passive magmatic degassing (SO$_2$ flux > 500 t d$^{-1}$) [Williams-Jones et al., 2003; Chapter1-1]. Unlike other basaltic shield volcanoes, Masaya volcano does not have frequent emission of juvenile magma (by explosive, or effusive eruptions). Masaya caldera (~ 11.5 x 6 km) hosts several cinder cones distributed along a ring fault [Rymer et al., 1998a, 1998b; Roche et al., 2001; Acocella, 2007].

Previous geophysical investigations have shown that the main gravity anomaly is located within Nindiri cone and more precisely beneath the active Nindiri crater [Métaxian et al., 1997; Rymer et al., 1998; Williams-Jones et al., 2003]. The continuous presence of fresh magma within a shallow magmatic reservoir and within the open conduit (network of pipes and caverns) within the active Santiago crater makes Masaya volcano an excellent location to investigate short term variations affecting the magma at shallow depths. Short term change in magmatic activity has been investigated over a short period of time by continuous gravity measurement within Nindiri crater, near the active Santiago crater.
12.2 Continuous gravity methodology

Gravity variations are due to changes in mass and density within the ground. On active volcanoes, significant variations of gravity are generated by change in mass, density, or volume of the magma, or hydrothermal system. Changes in volume are generally the consequence of magma/hydrothermal displacement [See Chapter 3 for more details]. Variations of density occur only within the magma, within a shallow magma storage where a magmatic differentiation or degassing process is taking place [See Chapter 3 for more details]. The continuous measurements were made using a LaCoste & Romberg Model G-127 (Fig. 12-1a), with Aliod feed-back system, allowing us to obtain an accuracy of ~0.01 mGal. A complete description of the continuous gravity method can be found in Chapter 3. The continuous gravity measurements were complimented by atmospheric temperature and pressure measurements with a Hobo Micro station (HB21-002) from Onset Company (Fig. 12-1a). The HWS Barometric pressure sensor has a accuracy of 3 mbar and the S-TMB-M002 temperature sensor has an accuracy of 0.2 °C between 0 to 50°C and can record temperatures from -40 to 100°C. Ground variations were controlled by the electronic levels of the gravimeter to insure that only the vertical gravity acceleration is measured. GPS position was measured by differential GPS with a Leica GPS system 500 allowing us to obtain centimetre accuracy. All the instruments were set together at the same place, in 2006 and in 2007, within a concrete box on the north side of Nindiri crater and at ~ 60 meters from the edge of Santiago crater (Fig. 12-1a, b).
Figure 12-1: Concrete box for the continuous gravity measurements in March 2006 and 2007. a) Location map of the concrete box on Nindiri cone (red triangle). b) Photo of the north plateau of Nindiri crater. The concrete box is at the bottom of the back side of the tumulus of lava, c) Gravimeter, and weather station within plastic bag to protect them from volcanic gas and dust, d) gravimeter G-127, with battery, palm and weather station just on its left side. UTM GPS coordinate are zone P16, N590442, E1325247.

The area was found to show the strongest gravity anomaly zone in a previous study by Williams-Jones et al. [2003]. The sampling step is at 2 Hz averaged every 5 minutes. Tilt levels of the gravimeter were controlled at least twice a day in order to keep recording only the vertical component of the
gravitational acceleration. The synthetic earth tide has been calculated with Quick Tide Pro. from MicroG LaCoste Inc. Pressure effects have been corrected by a factor of -0.365 per mbar as described in the work of Ando and Carbone [Andò and Carbone, 2001; Carbone et al., 2003; Andò and Carbone, 2004, 2006]. In order to avoid the influence of atmospheric temperature change (ranging between 24 and 32 °C) on the gravimeter, the internal temperature of the instrument was continuously kept at 50.2 °C. However, previous studies [Andò and Carbone, 2001; Carbone et al., 2003; Andò and Carbone, 2004, 2006] have shown that large atmospheric temperature variation (a few ten's of °C) can affect the gravimeter, even when its internal temperature is constant. Over a few days, temperature variations were less than 8 °C and even if they are small, they may nevertheless have an indirect effect on the gravity measurement (e.g., surface contraction/ dilation of the concrete floor or steel leg of the gravimeter). However, due to lack of information on the real effect of small temperature variations, no temperature corrections have been applied to the gravity data.

12.3 Short term dynamic gravity variation

In March 2006 and March 2007, continuous gravity measurements were made on Masaya volcano near the active Santiago crater (N+11.987, E-86.169; elevation 526.65 m, Fig. 12-1). The continuous measurement site is a concrete box, built within Nindiri crater on the north upper plateau, near a tumulus of lava ~60 m (Fig. 6-3) of the north-east border of Santiago crater. The box itself is built within the ground allowing the gravimeter to be better protected from wind and weather. On March 1st and 2nd, 2006 a period of ~22 hours was recorded (Fig. 12-2a), which was completed in March 6th and 7th, 2006 by a period measurement long of ~27 hours (Fig. 12-2b). In March 2007, continuous measurements were again made within the concrete box at the exact same place.
(north-east corner within the box) as the previous years, In 2007, a period of ~9 days was recorded (~222 hours, Fig. 12-3). The continuous gravity survey was also attempted in March 2008. However, due to technical problems with the data logger, no data had been collected. After correction for earth tide, spikes, instrumental tares, drift and atmospheric pressure influence, the two sets of data (~24 hours each, Fig. 12-2) of March 2006 and the 9 days periods of March, 2007 have shown presence of short term gravity variation of ~ 60 to 80 μGal and having a wavelength of ~ 25 to 28 hours (Fig. 12-3). The residual data were not corrected for oceanic loading, because the large lakes (Nicaragua Lake and Managua Lake) are more than 40 km from the gravity station and thus their effect is considered as negligible. Laguna de Masaya is too small and too far to be of any affect on the data’s.

Unfortunately, residual gravity data sets from March 2006 do not cover the full period of these short term gravity (Fig. 12-2). However, the residual gravity of March 2007 shows several complete recorded periods of this short term variation (Fig. 12-3). Comparison of the residual gravity signal from 2006 and 2007 shows significant similarity in both wavelength and amplitude (Fig. 12-4). It seems that the short term gravity anomaly, ~25 hours in period, consists of two smaller period anomalies, which overlap and give the asymmetrical shape of a peak and a plateau (Fig. 12-3).
Figure 12-2: Raw gravity, earth tide and residual continuous gravity on March 2006. Sampling step is of 5 min. From top to bottom, atmospheric pressure (black line), atmospheric temperature (red line), raw gravity (blue line), synthetic earth tide (green line) and residual gravity (purple line). The residual gravity has been corrected for tide, tares, instrumental drift and pressure effect.
Figure 12-3: Raw gravity, earth tide and residual continuous gravity, March 2007. Sampling interval is 5 min. From top to bottom, atmospheric pressure (black line), atmospheric temperature (red line), raw gravity (blue line), synthetic earth tide (green line) and residual gravity (purple line). The residual gravity has been corrected for earth tide, tares, instrumental drift and pressure effects.
As such, the recorded gravity signal must consist of a number of different effects, such as external parameters (pressure, temperature, deformation), tides (earth, oceanic) or gravitational variations from magmatic activity. When analyzed within the frequency domain, we can see the significance and the presence of each of these signals shaping the recorded gravity data. In order to investigate the origin of this short term gravity variation and to determine if the residual signal is due to an uncorrected parameter (such as atmospheric temperature or ocean loading), each of the signals have been analyzed through a Fast Fourier Transform (FFT) [Fournier, 2003] to compare their frequency signatures.

Figure 12-4: Comparison of residual continuous gravity between 2006 and 2007. Purple line is the residual gravity in 2007. The blue and red lines are, respectively, the residual gravity of March 1\textsuperscript{st} and 2\textsuperscript{nd}, 2006, and 6\textsuperscript{th} and 7\textsuperscript{th}, 2006.
Figure 12-5: Frequency structures of raw gravity, earth tide, atmospheric pressure and temperature, and residual gravity during March 2007. Note the different amplitude axes for temperature and pressure, while raw gravity, earth tide and residual gravity are presented on a similar axe range. Frequency scale is in log Hertz.
On Masaya, the ocean loading is considered to be negligible because the major lakes (Lake Nicaragua and Lake Managua) and Pacific ocean are more than 40 km from the gravimeter and thus only gravity (raw and residual), synthetic earth tide, atmospheric temperature, and atmospheric pressure were analyzed by FFT (Fig. 12-5).

The atmospheric temperature signal is characterized by several frequency spikes, which with decreasing amplitude are: ~25, ~22, ~12, ~56 and ~6 hour periods (red line, Fig. 12-5). The main period of ~25 h and ~12 h represent the daily temperature variation. The 56 h periods appears to change over 2 to 3 days.

The atmospheric pressure signal consists of 4 frequencies (black line, Fig. 12-5). Those frequencies have a time-periodicity of ~32, ~25, ~22 and ~12 hours, respectively, the period of 12 and 25 hours are the most important.

The raw gravity shows 3 main frequencies (period in hours, purple line on Fig. 12-5). The first and strongest, is ~12 hours. The second and third peaks are much smaller in amplitude with ~22 and ~28 hours, respectively.

The earth tide has 2 main frequency peaks, the largest and strongest at ~12 hours. The second peak has a period of ~25 hours (green line, Fig. 12-5).

The residual gravity (corrected for tide, drift, instrumental tares and atmospheric pressure, but not atmospheric temperature) has a main peak of ~25 hours. A secondary frequency of ~12 hours is also present.

Assuming that both earth tide and pressure effect are fully removed from the raw gravity, the frequency spikes obtained on the residual gravity (Fig. 12-5) must characterize real gravity variations or external effects other than earth tide. One main limitation of the FFT is the redundancy of information. Any frequency present inside the analyzed signal will be found through the FFT at its own frequency, but also at any of its component frequencies (component of order 2 for the 12 hours period and component of order 4 for the 6 hours period). For example, the FFT of the 2007 residual gravity signal shows frequency spikes at

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6, 12, 22, 25 and 56 hours (Fig. 12-5). Due to the limited quantity of data, only 9 days, it is very possible that the period of 22 and 25 hours characterize the same thing, which could be averaging at 24 h period. Thus, 24 h, 12 h are both a multiple of 6 h. Similarly, frequencies are found on the temperature frequency signal; it seems very likely that the frequency spike characterize effects of temperature variation on the residual gravity. The main temperature variations over a day are generally of 12 h, with the maximum during the afternoon and the minimum temperature during the last part of the night. Thus, as suggested by previous studies on other volcanoes [Andò and Carbone, 2001; Carbone et al., 2004; Andò and Carbone, 2006], the continuous gravity measurement of March 2007 shows a daily effect of the temperature variation on the residual gravity data. Unfortunately, the actual data set of continuous measurement is not large enough to efficiently apply the Neuro Fuzzy technique to correct for atmospheric temperature effect on the gravimeter [Andò and Carbone, 2001; Carbone et al., 2004; Andò and Carbone, 2006]. However, this study do not dismiss the possibility of magmatic tide within the magmatic conduit or deformation of the crater floors due to underground volcanic/magmatic change, or short term hydrothermal variations [Tikku et al., 2006]. More continuous measurement on Masaya volcano are required in order to better understand the measured gravity signal.

12.4 Discussion and Conclusion

Continuous gravity measurements, done over 9 days in 2007 in Nindiri crater, show the presence of two cycles of gravity variation of ~60 to 90 μGal over periods of 12 hours. The present continuous gravity data set, unfortunately, does not allow us to conclude on the origin of the short gravity change occurring over 12 and 24 hours. However, this study does not exclude the possibility of large effects of daily atmospheric temperature variation. A magmatic tide acting as one of the components of the measured anomaly is still a possibility, but for now it can not be supported by this study.
13 APPENDICES A5: THE STUDY OF MASAYA VOLCANO
Figure 13-1: Self-potential/elevation signature of water table and hydrothermal system on Masaya volcano. The plot on the left side is the SP/elevation gradient; while on the right side the plot is the SP along distance on the profile. Blue symbols refer to the Crosssection Nindiri profile (Fig. 6-3). The green symbols refer to Masaya crater profile (Fig. 6-6). The orange symbols refer to North Road profile (Fig. 6-7). The purple symbols refer to Comalito profile (Fig. 6-7). The gradient is calculated for the 2008 survey. The other anomalies can be found in Appendix A3-01. Note the break on the horizontal axis of plot a) on the right column.
Figure 13-1 continued:
Figure 13-1 continued: Note the break on the horizontal axis of plot k) on the right column.
Figure 13-1 continued: Note the break on the horizontal axis of plot o) on the right column.
Figure 13-2: Calculated water depth around Nindiri crater from Multi-scale Wavelet tomography on the SP Nindiri survey loop between 2006 and 2009. Gold symbols are the average value of the water depth based on all depths having similar distance $x$. Calculations were made using the vertical and horizontal derivative of $2^{nd}$ or $3^{rd}$ order of the Poisson kernel equation (as described in Chapter 5) Green symbols are wavelets of $2^{nd}$ order of derivative. Red symbols are wavelet of $3^{rd}$ order of derivative. Squares represent wavelets based on the vertical derivative. Diamonds represent wavelets based on the horizontal derivative. Error bar characterize the scattering of the data.
Figure 13-3: Calculated water depth around Masaya crater from the Multi-scale Wavelet tomography on the SP Masaya survey loop between 2006 and 2009. Gold symbols are the average value of the water depth based on all depths having similar distance x. Calculations were made using the vertical or horizontal derivative of 2\textsuperscript{nd} or 3\textsuperscript{rd} order of the Poisson kernel equation (as described in Chapter 5) Green symbols are wavelets of 2\textsuperscript{nd} order of derivative. Red symbols are wavelet of 3\textsuperscript{rd} order of derivative. Squares represent wavelets based on the vertical derivative. Diamonds represent wavelets based on the horizontal derivative. Error bar characterize the scattering of the data.
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<th>Name</th>
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<th>Years</th>
<th>Nbr of Wavelet</th>
<th>Distance x in m</th>
<th>σ x in m</th>
<th>Depth σ in m</th>
<th>σ Z in m</th>
<th>Elevation a.s.l.</th>
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**Table 13-1:** Water depths obtained from Multiscale Wavelet Tomography analysis of the Nindiri SP survey loop in 2006, 2007, 2008 and 2009. N_C is Nindiri crater SP profile. N_cros is Nindiri cross-section SP profile. H is hydrothermal system and W is aquifer. UTM zone is P16. # indicate the Self-potential anomaly number associated with the depth. Wavelets used for analysis were the vertical or horizontal derivative of 2nd or 3rd order of the Poisson kernel. More details can be found in Chapter 4 for methodology and Chapter 6 for the study of Masaya volcano.
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### Table 13-2: Water depths obtained from MWT analysis of the Masaya crater SP survey loop in 2006, 2007, 2008 and 2009. M_C is Masaya crater SP profile. Com_P is Comalito SP profile. H is hydrothermal system and W is aquifer. UTM zone is P16. # indicate the Self-potential anomaly number associated with the depth. Wavelets used for analysis were the vertical or horizontal derivative of 2<sup>nd</sup> or 3<sup>rd</sup> order of the Poisson kernel. More details can be found in Chapter 4 for methodology and Chapter 6 for the study of Masaya volcano.

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**Table 13-3:** Water depths obtained from MWT analysis of the Comalito SP profile in 2007 and 2008. Com_P is Comalito SP profile. North-R is North road SP profile. H is hydrothermal system and W is aquifer. UTM zone is P16. # indicate the Self-potential anomaly number associated with the depth. Wavelets used for analysis were the vertical or horizontal derivative of 2nd or 3rd order of the Poisson kernel. More details can be found in Chapter 4 for methodology and Chapter 6 for the study of Masaya volcano.
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<th>Depth $\sigma_Z$ (m)</th>
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**Table 13-4:** Water depths obtained from MWT analysis of the North road SP profile in 2007 and 2008. North-R is North road SP profile. H is hydrothermal system and W is aquifer. UTM zone is P16. # indicates the Self-potential anomaly number associated with the depth. Wavelets used for analysis were the vertical or horizontal derivative of 2nd or 3rd order of the Poisson kernel. More details can be found in Chapter 4 for methodology and Chapter 6 for the study of Masaya volcano.
Figure 13-4: Oblique aerial view of Nindiri cone, Masaya cone and Comalito cone. Masaya city can be seen overlooking Laguna de Masaya (See Chapter 1).
14 APPENDICES A6: THE STUDY OF KAWAH IJEN VOLCANO
Figure 14-1: ID Counter drift measured on Kawah Ijen Volcano in 2006, 2007 and 2008, expressed through the $G_{\text{meas}}/\Delta I_D$ vs. $\Delta G_{\text{theoric}}$ (real ID counter-reference at 1336). Top: gravity station HW01 at Pondok near the summit of Kawah Ijen. Bottom: gravity station IJ04 at Paltuding (see Chapter 7, fig. 7-4). The drift of the ID counter is of $\sim$0.039 mGal/turn of the ID counter.
Figure 14-2: Structure and dynamic behaviour of the water tables by depth estimates by Multiscale wavelet tomography of Self-potential data between 2006 and 2008. H2, H3, V2 and V3 are the wavelets from the horizontal, either vertical) derivative of the Poisson kernel Family. Error bars represent the scattering of the data as described in Chapter 4 and 7. a) All water depths calculated in 2006. b) All water depths calculated in 2007. c) All water depths calculated in 2008. d) Water depths found with at least 3 of the 4 wavelets.
Figure 14-3: Calculation of water height over time based on the model A at two different porosities and each of the gravity station on Kawah Ijen Volcano (see Chapter 7, Table 7-5). Error bars represent the scattering of the data.
Figure 14-4: Calculation of water height over time based on the model B at two different porosities and each of the gravity station on Kawah Ijen Volcano (see Chapter 7, Table 7-5). Error bars represent the scattering of the data.
Figure 14-5: Calculation of water height over time based on the model C at two different porosities and at the site of two gravity stations (CI-34 and CI-35 near the Dam) on Kawah Ijen Volcano (see Chapter 7, Table 7-5). Error bars represent the scattering of the data.

Figure 14-6: Calculation of water height over time based on the model D at two different porosities and at the site of two gravity stations (CI-34 and CI-35 near the Dam) on Kawah Ijen Volcano (see Chapter 7, Table 7-5). Error bars represent the scattering of the data.
Figure 14-7: Screen capture of model A of a water column at porosity 0.32 during forward modelling. Anomaly map dimension is set to have the gravity site on position B (bottom map) at the edge of the water column. Water column dimension is of 400 m by 300 m and high of 100. Top of the water column is at 100 m below the surface symbolized by the anomaly map (top map, see Chapter 7, Table 7-5). Error bars represent the scattering of the data. Gravity values are in mGal.
Figure 14-8: Results of forward modelling of a lake having a thickness of 10 m and a density of 1.1 during forward modelling (such as Ijen acid lake, see Chapter 2.2 and Chapter 7-6.). Anomaly map dimension is set to have the gravity site on position A at the edge of the west side of the lake. Lake dimension is 1000 m by 640 m as Ijen acid lake. Top: anomaly map is at 2 m above the lake level (such as gravity stations CI-34 and CI-35 on the top of the Dam stair and on the Dam, respectively). Bottom: anomaly map is at 241 m above the lake surface (such as gravity stations CI-10, CI-20 and CI-36 along Kawah Ijen Rim, see Chapter 7). Gravity values are in mGal. Coordinate are in UTM zone 50L and match with the Ijen lake coordinate.
Figure 14-9: Results of forward modelling of a magmatic dyke with a density contrast between magma and gas of 2.3 g cm\(^{-3}\). Anomaly map dimension is set to have the gravity site along its edge. Magmatic dyke dimension is of 60 m by 300 m and 100 m high. Top of magmatic column is at 1600 m a.s.l, such as described in Chapter 7 (Table 7-4). a) The anomaly map calculated at an elevation of 2161 m a.s.l (~ gravity station CI-34 and CI-35). b) The anomaly map calculated at an elevation of 2400 m a.s.l (~ gravity station CI-10, CI-20 and CI-26, see Chapter 7 and Fig A5-12). Gravity values are in mGal.
Figure 14-10: Results of forward modelling of a magmatic dyke (cubic shape) with a density contrast between magma and gas of 2.3 g cm$^{-3}$. Anomaly map dimension is set to have the gravity site along its edge. Magmatic dyke dimension is of 100 m by 100 m and 100 m high. Top of magmatic column is at 1600 m a.s.l, such as described in Chapter 7 (Table 7-4). a) The anomaly map calculated at an elevation of 2161 m a.s.l (~ gravity station CI-34 and CI-35). b) The anomaly map calculated at an elevation of 2400 m a.s.l (~ gravity station CI-10, CI-20 and CI-26, see Chapter 7 and Fig A5-12). Gravity values are in mGal
Figure 14-11: Comparison of water depth calculations between from Model B and Model C to MWT on Kawah Ijen Volcano (see Chapter 7) over 2006 to 2008. Triangles represent the water depth calculated by MWT on Self-Potential data. Diamonds and squares are water depths by forward gravity Model B and C, as described in Chapter 7. Model B and C have a porosity of 0.15 and 0.45, respectively.
Figure 14-12: Comparison of the water depth variation on Kawah Ijen volcano from three different models. Model B (diamond) and Model D (square) are generated from the inverse modelling of gravity data, while the third model (triangle) is from Multi-scale wavelet tomography on SP profile. Model B and D have a porosity of 0.15 and 0.40, respectively.
Figure 14-13: Comparison of the water depth variation on Kawah Ijen volcano from three different models. Model B (diamond) and Model D (square) are generated from the inverse modelling of gravity data, while the third model (triangle) is from Multi-scale wavelet tomography on SP profile. Model B and D have a porosity of 0.27 and 0.40, respectively.
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Table 14-1: Gravity stations UTM coordinate on Kawah Ijen volcano.

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Table 14-2: Calculated water depths by Multi-scale wavelet tomography (MWT) on the crater rim Self-potential profile in 2006. $\sigma$ characterizes the scattering of the data
Table 14-3: Calculated water depth by Multi-scale wavelet tomography (MWT) on the crater rim Self-potential profile in 2007. $\sigma$ characterizes the scattering of the data

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Table 14-4: Calculated water depth by Multi-scale wavelet tomography (MWT) on the crater rim Self-potential profile in 2008. $\sigma$ characterizes the scattering of the data

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**Table 14-5:** Calculated water depth by Multi-scale wavelet tomography (MWT) on the crater rim Self-potential profile in 2009. σ characterizes the scattering of the data.
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**Table 14-6:** Parameters of the model used to estimate the ground porosity.