Cross-section electrical resistance tomography of La Soufriere of Guadeloupe lava dome
Nolwenn Lesparre, Bartlomiej Grychtol, Dominique Gibert, J.C Komorowski, Andy Adler

To cite this version:

HAL Id: insu-01086403
https://hal-insu.archives-ouvertes.fr/insu-01086403
Submitted on 24 Nov 2014

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L’archive ouverte pluridisciplinaire HAL, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d’enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.
Cross-section electrical resistance tomography of La Soufrière of Guadeloupe lava dome

Nolwenn Lesparre,1,* Bartłomiej Grychtol,2 Dominique Gibert,3,4 Jean-Christophe Komorowski3 and Andy Adler1

1Systems and Computer Engineering, Carleton University, Ottawa, Canada. E-mail: nolwenn.lesparre@irsn.fr
2Medical Physics in Radiology, German Cancer Research Center, Heidelberg, Germany
3Institut de Physique du Globe de Paris, Sorbonne Paris Cité, Univ Paris Diderot, UMR F-7154 CNRS, Paris, France
4Geosciences Rennes, Univ Rennes 1, UMR F-6118 CNRS, Rennes, France

Accepted 2014 March 18. Received 2014 March 18; in original form 2013 July 11

SUMMARY

The electrical resistivity distribution at the base of La Soufrière of Guadeloupe lava dome is reconstructed by using transmission electrical resistivity data obtained by injecting an electrical current between two electrodes located on opposite sides of the volcano. Several pairs of injection electrodes are used in order to constitute a data set spanning the whole range of azimuths, and the electrical potential is measured along a cable covering an angular sector of ≈120° along the basis of the dome. The data are inverted to perform a slice electrical resistivity tomography (SERT) with specific functions implemented in the EIDORS open source package dedicated to electrical impedance tomography applied to medicine and geophysics. The resulting image shows the presence of highly conductive regions separated by resistive ridges. The conductive regions correspond to unconsolidated material saturated by hydrothermal fluids. Two of them are associated with partial flank collapses and may represent large reservoirs that could have played an important role during past eruptive events. The resistive ridges may represent massive andesite and are expected to constitute hydraulic barriers.

Key words: Image processing; Tomography; Electrical properties; Volcanic hazards and risks.

1 INTRODUCTION

La Soufrière of Guadeloupe volcano belongs to the active part of the volcanic arc forming the Lesser Antilles and caused by the subduction of the North American Plate beneath the Caribbean Plate. The La Soufrière lava dome (Fig. 1) is dated 1530 A.D. (Boudon et al. 2005; Boichu et al. 2005, 2008; Legendre 2012).

La Soufrière lava dome is located in the horseshoe-shaped ‘Amic’ crater formed 3100 B.P. by a St Helens-type edifice collapse and directed blast event (Boudon et al. 1987). Since its formation, La Soufrière had six phreatic eruptions (1690, 1797–1798, 1809–1812, 1836–1837, 1956, 1976–1977) located in different sectors of the northern and eastern sides of the lava dome (Fig. 2). The last 1976–1977 event is considered a failed magmatic eruption (Feuillard et al. 1983; Komorowski et al. 2005; Villemant et al. 2005; Boichu et al. 2008, 2011) caused by the intrusion of a small volume of andesitic magma whose ascension stopped at about 3 km beneath the dome summit (Villemant et al. 2005; Boichu et al. 2011). Since then, this magma body sporadically releases acid gases in the hydrothermal reservoirs and produces episodic chlorine spikes in the ‘Carbet’ hot spring located on the northeast side of the volcano (Boichu et al. 2011). The thermal energy released in the shallower parts of the volcano drove thermal convection of hydrothermal fluids. This crisis was particularly intense and forced the evacuation of 73 000 inhabitants over 6 months.

Following the 1976–1977 crisis, both the volcanic and seismic activities reduced gradually until 1992 when a notable increase in shallow low-energy seismicity and in the flux of summit fumarolic activity was observed (Komorowski et al. 2005). One possible interpretation of this event is attributed to a reorganization of the fluid circulation pattern inside the lava dome as a response to progressive sealing of the hitherto active flow paths. Both the intense hydrothermal activity and the heavy rains (≈5 m yr⁻¹) supplying the hydrothermal shallow reservoirs favour fluid mineralization by magmatic gas and the formation of clayey material that progressively fills and blocks open fractures in the edifice decreasing its...
macroporosity (Zlotnicki et al. 1994; Villemant et al. 2005; Salaün et al. 2011). The resulting sealing causes fluid confinement and overpressurization that eventually leads to the opening of new flow paths inside the edifice (Salaün et al. 2011). Another possibility is that this seismic and fumarolic reactivation characterized with a new pulse of marked chlorine degassing reflects injections of magmatic fluids and heat from the magmatic reservoir to some shallower level in the hydrothermal system below the summit (Fournier 2006).

The past eruptive history of La Soufrière indicates that somewhat different scenarios have to be considered for the future, depending on the nature of the event: collapse, phreatic eruption or magma ascent (Komorowski et al. 2008). The evaluation of hazards for each scenario depends on both the impact and the likelihood of each type of event, and some of them suggest important societal impacts in case of renewed activity. For this reason, multiparameter monitoring is conducted by the local volcano observatory (IPGP/OVSG). Permanent networks monitor seismicity and ground deformation. Thermal springs and fumaroles are also sampled and analysed on a fortnightly base (Villemant et al. 2005). Beside these routine measurements, it is important to perform complementary geophysical studies to obtain an ever more precise view of the inner structure of the volcano and derive models necessary to better understand the monitoring data. Knowledge of the inner structure is necessary to set reliable initial conditions to flank destabilization models (Le Friant et al. 2006) and to estimate the amount of fluid contained in shallow hydrothermal reservoirs that may supply thermal and explosive energy in case of phreatic explosion and provoke lahars as observed during the 1976–1977 crisis.

Over the last decade, La Soufrière of Guadeloupe has been subject to several geophysical imaging experiments including self-potential mapping (Zlotnicki et al. 1994), electrical resistivity (Nicollin et al. 2006) and very low-frequency survey (Zlotnicki et al. 2006). Such methods were particularly sensitive to the presence of fluids circulating in the volcano hydrothermal system. The self-potential study evidenced the structural heterogeneity of La Soufrière lava dome and the circulation of fluids (Zlotnicki et al. 1994). Negative anomalies present on the northern part of the dome emphasized the presence of vertical conduits where meteoric fluids circulated mostly downward (Zlotnicki et al. 1994). A smooth positive anomaly enclosed the dome (except on the northwestern part) and underlined the crater Amic wall and was correlated to ancient or active fumarolic areas, signalling the existence of upward flowing fluids (Zlotnicki et al. 1994). The very low frequency electromagnetic survey (Zlotnicki et al. 2006) allowed to characterize the state of the main fault systems located on the volcano: hydrothermally active faults appear electrically conductive, and clayed, sealed or opened faults have higher resistivity values (Zlotnicki et al. 2006). Profiles of apparent electrical resistivity display high resistivity contrasts confirming the
heterogeneous structure of the dome (Nicollin et al. 2006). Some profiles also indicate the presence of vertical conductive conduits interpreted as paths of an upward circulation for hydrothermal fluids. Local 1-D geoelectrical soundings (Nicollin et al. 2006) show that highly conductive material surrounds the dome and could constitute a continuous layer below the basis of the dome.

The dome density distribution was also studied using gravity measurements (Gunawan 2005) jointly inverted with seismic data (Coutant et al. 2012). These studies confirmed the heterogeneous structure of the dome. Recently, a cosmic muon radiography method has been developed (Gibert et al. 2010; Lesparre et al. 2010; Marteau et al. 2012) and applied to La Soufrière lava dome (Lesparre et al. 2012) to perform direct imaging of its density.
slice electrical resistance tomography

2.1 Data

The data analysed in this study constitute a subset of so far unprocessed measurements acquired during a larger electrical resistivity survey performed in 2003 December (Nicollin et al. 2006). The data were acquired with a multi-electrode resistivity meter connected to a 945-m-long main cable equipped with 64 plugs connected to stainless steel electrodes. Either plug 1 or 64 located at the extremities of the main cable is connected to an auxiliary long wire in order to place the corresponding electrode on the opposite side of the lava dome. Both the remote electrode and one electrode plugged onto the main cable are used to inject an electrical current forced to cross the innermost parts of the volcano (Fig. 4). The main cable was moved to successively occupy three circular segments, each of them covering about one third of La Soufrière circumference to form an almost closed loop (Fig. 4). The entire data set counts a total of 298 measurements obtained with 13 pairs of current electrodes combined with a number of pairs of potential electrodes that varies between 12 and 30 (Table 1). The maximal distance between current electrodes is of 940 m in the north–south direction and of 820 m in the east–west direction. The electrode positions fall nearby
Figure 4. Map showing the 13 pairs of current electrodes (circles) used in this study. The curved lines crossing the volcano correspond to the main current lines (calculated using a 2-D homogeneous model) joining the corresponding current electrodes. The electrodes used for acquisitions are represented with different symbols and colours that correspond to a specific pair of stimulating electrodes represented by circles.

Table 1. Description of the measuring electrodes configuration for each pair of stimulating electrodes.

<table>
<thead>
<tr>
<th>Current electrode pair</th>
<th>Cable location</th>
<th>Profile length (m)</th>
<th>Mean distance between V electrodes (m)</th>
<th>Number of dipoles</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>SE</td>
<td>851</td>
<td>29</td>
<td>29</td>
</tr>
<tr>
<td>2</td>
<td>SE</td>
<td>851</td>
<td>29</td>
<td>29</td>
</tr>
<tr>
<td>3</td>
<td>SE</td>
<td>836</td>
<td>56</td>
<td>15</td>
</tr>
<tr>
<td>4</td>
<td>SE</td>
<td>841</td>
<td>30</td>
<td>28</td>
</tr>
<tr>
<td>5</td>
<td>SE</td>
<td>784</td>
<td>56</td>
<td>14</td>
</tr>
<tr>
<td>6</td>
<td>SE</td>
<td>754</td>
<td>54</td>
<td>14</td>
</tr>
<tr>
<td>7</td>
<td>N</td>
<td>788</td>
<td>39</td>
<td>20</td>
</tr>
<tr>
<td>8</td>
<td>N</td>
<td>850</td>
<td>28</td>
<td>30</td>
</tr>
<tr>
<td>9</td>
<td>N</td>
<td>842</td>
<td>28</td>
<td>30</td>
</tr>
<tr>
<td>10</td>
<td>W</td>
<td>849</td>
<td>39</td>
<td>21</td>
</tr>
<tr>
<td>11</td>
<td>W</td>
<td>733</td>
<td>61</td>
<td>12</td>
</tr>
<tr>
<td>12</td>
<td>W</td>
<td>854</td>
<td>33</td>
<td>26</td>
</tr>
<tr>
<td>13</td>
<td>W</td>
<td>831</td>
<td>28</td>
<td>30</td>
</tr>
</tbody>
</table>

Figure 5. Finite element model used to compute the forward problem; x and y coordinates are oriented positively eastwards and northwards, respectively. The orange vertical column (A) is obtained through a vertical extrusion of the elements (A) in Fig. 7. The blue elements (B) located in the exterior domain of the cross-section are obtained through a combination of vertical extrusion and a nearest neighbour extrapolation of elements B in Fig. 7.

Figure 6. Apparent resistivity acquired from the different electrode profiles surrounding the dome. Colours correspond to the configuration number given in Fig. 4 and Table 1.

a slightly inclined plane with an elevation decrease of 230 m from north to south (Fig. 5). The elevations of the electrode loop vary between 1146 and 1337 m with an average of 1270 m, that is, about 200 m bellow the summit.

The primary data set is formed by \( K = 298 \) \( n \)-tuples \( \{I_k, V_k, C^-_k, C^+_k, P^-_k, P^+_k\} \), where \( I_k \) is the electrical current for current electrode positions \( C^-_k, C^+_k \) and \( V_k \) is the voltage measured between electrodes \( P^-_k \) and \( P^+_k \). The current was automatically adjusted between 20 and 100 mA to ensure good signal-to-noise ratio. More details concerning the measurement procedure are given by Nicollin et al. (2006), and a discussion about the assessment of the data quality may be found in Nicollin et al. (2007). The noise present in the data is mainly multiplicative with an estimated signal-to-noise ratio of about 90 per cent.

Since the distance between the potential electrodes may greatly vary from one \( n \)-tuple to another, \( V_k \) spans several orders of magnitude. Consequently, a normalization is performed by converting the \( \{I_k, V_k\} \) pairs into apparent resistivity \( \rho_{app,k} = \beta_k V_k/I_k \) (Fig. 6) where the geometrical factor \( \beta_k \) is computed with a 3-D model of the volcano (Fig. 5; Lesparre et al. 2013). The resulting apparent resistivities vary between 12 and 1360 \( \Omega\cdot m \). These values agree with the pseudo-sections obtained by Nicollin et al. (2006) where the apparent resistivites are mainly comprised in the 10 and 2000 \( \Omega\cdot m \) range. Apparent resistivites up to 10,000 \( \Omega\cdot m \) are reported by these authors for several shallow areas where massive andesitic rock and cavities are present. Because of the particular electrode set-up used in this study, these shallow high-resistivity regions are not expected to significantly influence our data which are aimed to mainly sample the innermost parts of the lava dome. Discrepancies between the apparent resistivites derived in this study and those obtained...
by Nicollin et al. (2006) may also be partly explained by the fact that these authors used a flat geometry to compute their geometrical factors instead of a full 3-D model like in this study (Fig. 5).

2.2 SERT inversion

The limited amount of data available are not suitable to perform a full 3-D reconstruction of the conductivity structure inside the volcano and, in this study, we perform a slice tomography to reconstruct the conductivity distribution in a cross-section limited by the ring of electrodes (Fig. 7). This approach is similar to the slice impedance tomography of the human thorax (Adler et al. 2012) where the lung cavities provoke high contrasts of electrical conductivity (Vogt et al. 2012). In this study, SERT is implemented by defining the unknown conductivity distribution \( \sigma_{z-D} \) on a coarsely meshed 2-D cross-section in order to reduce the number \( W \) of conductivity values to invert (Fig. 7).

The cross-section conductivity \( \sigma_{z-D} \) is subsequently used to construct the full 3-D conductivity distribution \( \sigma_{3-D} \) necessary to solve a forward 3-D finite element model (Fig. 5). This is achieved by using a coarse-to-fine matrix \( M \) that maps the conductivity \( \sigma_{z-D} \) of each element of the cross-section (Fig. 7) onto each of the \( N \) elements of the 3-D model (Fig. 5). For elements vertically located either above or below the cross-section, the mapping is performed through a vertical extrusion that makes the conductivity distribution vertically invariant (i.e. 2-D). This case is illustrated with the elements labelled A in Fig. 7 that gives the extruded vertical column \( A \) in the 3-D model of Fig. 5. Elements located outside the extruded cross-section have their conductivity values assigned with a nearest neighbour criteria. This case is illustrated with the 3-D elements labelled B in Fig. 5 whose conductivity is inherited from element B of the cross-section (Fig. 7). In this study, the number of conductivity values \( \sigma_{2-D} \) to invert is \( W = 2690 \) and \( \sigma_{z-D} \) is defined on 33 174 nodes forming the \( N = 170 491 \) elements of the 3-D model (Fig. 5).

Both the forward modelling and the inversion are implemented with the open-source EIDORS software initially dedicated to medical applications (Polydorides & Lionheart 2002; Adler & Lionheart 2006) and recently augmented with geophysical functionalities (Lesparre et al. 2013). The meshing is performed with NETGEN (Schöberl 1997) and uses a digital elevation model with a mesh of 5 m (Fig. 5). Point electrodes are used to represent the steel rods used on the field. A refined meshing is implemented near the current electrodes to account for the sharp gradient of the electrical potential (Rucker et al. 2006).

The forward model solution gives the electrical potential \( u(x, y, z) \), which is further transformed into apparent resistivity \( \tilde{\rho}_{app} \) for a given distribution of the electrical conductivity \( \sigma(x, y, z) \). Insulating conditions are imposed on the boundaries \( \Gamma \) of the model volume \( \Omega \) excepted at the current electrodes where a Neumann condition is imposed to represent the injected electrical current. The equations to be solved read,

\[
\nabla \cdot (\sigma \nabla u) = - \nabla \cdot j \quad \text{in} \quad \Omega \in \mathbb{R}, \quad (1)
\]

\[
\sigma \left( \frac{\partial u}{\partial n} \right) = j \cdot n \quad \text{on} \quad \Gamma, \quad (2)
\]

where \( j \) is the source current density and \( n \) denotes the outward normal on \( \Gamma \) (Rucker et al. 2006).

Because of the large range (10–1000 \( \Omega \) m) spanned by the apparent resistivity data, the fit is made on \( \log(\tilde{\rho}_{app}) \) instead of \( \tilde{\rho}_{app} \). Inverted parameters correspond to log-conductivity which is the natural quantity that appears in the integral equation relating \( \sigma \) to \( \tilde{\rho}_{app} \). Working in the log domain also makes the usage of either conductivity or resistivity equivalent (Tarantola 2006), and it also appears in asymptotic formulations of high-contrast conductivity imaging as shown by Borcea et al. (1999), Günther et al. (2006), Marescot et al. (2008) and Lesparre et al. (2013).

The inverse problem aims to recover the distribution of the logarithm of the conductivity \( \varsigma = (\varsigma_1, \varsigma_2, ..., \varsigma_B) \) in the cross-section able to reproduce the logarithm of the apparent resistivity data \( \log \theta = (\log \theta_1, \log \theta_2, ..., \log \theta_B) \). The inversion is iteratively performed with a conjugate gradient method. Although SERT strongly reduces the ill-posedness of the inversion, the inverse problem remains strongly underdetermined and a regularization of the Jacobian is performed through a filtering via a singular value decomposition (SVD; Marescot et al. 2008; Lesparre et al. 2013). In practice, the SVD cut-off used to construct the pseudo-inverse \( J^\dagger \) of the Jacobian is chosen according to an L-curve criterion (Hansen 2001).

The main stages of the inversion procedure are:

1. Estimation of the 3-D forward problem from a given distribution of \( \varsigma' \): \( \log \tilde{\theta}' = f(\varsigma') \), with \( i \) the iteration number;
2. Computation of the Jacobian to estimate the sensitivity changes to the sought values defined by the coarse 2-D mesh

\[
J_{i,k} = \frac{\partial \log \tilde{\theta}_i}{\partial \varsigma_k}; \quad (3)
\]

3. The direction of the perturbation \( \delta' \) affected to the sought values is estimated from the SVD-regularized pseudo-inverse of the Jacobian

\[
\delta' = (J^\dagger)(\log \theta - \log \tilde{\theta}); \quad (4)
\]

4. Perturbations are then added to the previous values \( \varsigma' \), with a step length \( \alpha' \)

\[
\varsigma'^{i+1} = \varsigma'^i + \alpha' \delta'; \quad (5)
\]

5. Inverted values defined on the 2-D coarse mesh \( \varsigma' \) are interpolated using the matrix \( M \) to reconstruct the 3-D forward model;

Figure 7. Meshing of the conductivity cross-section used in the SERT inversion. The green stars represent the electrodes. The orange and blue elements labelled A and B are used as examples to illustrate the construction of the 3-D forward model shown in Fig. 5.
(6) Computation of the 3-D forward model with some trial values for $\alpha'$ in order to estimate the corresponding values of $\log \tilde{g}^{i+1}$:

$$\log \tilde{g}^{i+1} = f(\zeta' + \alpha' \delta');$$

(7) Residuals $\log g - \log \tilde{g}'$ corresponding to the different values of $\alpha'$ are compared to estimate the appropriate value for $\alpha'$, which retained value is the one corresponding to the minimum value of $\sigma$.

(8) If convergence is not achieved, return to step 1.

The starting model of the inversion is initialized with a resistivity of 106 $\Omega$ m which corresponds to the average of the measured apparent resistivities. The first iteration is done with 13 trial values for $\alpha$, varying between 0.2 and 0.5. For next iterations, the trial values correspond to $\frac{1}{2} \alpha^{i-1}$; $\alpha^{i-1}$; $2 \alpha^{i-1}$, where $\alpha^{i-1}$ represent the step length used at the previous iteration. This procedure efficiently reduces the number of the forward model computations which is the most time-consuming part of the inversion (Marescot et al. 2008). In practice, 10 iterations are performed and the convergence is mainly obtained during the first two iterations.

2.3 Results

The resistivity cross-section $\sigma_{2-D}$ obtained from the SERT inversion of the data set shown in Fig. 6 is displayed in Fig. 8. Resistivity values span a range of two orders of magnitude, from 10 to 1000 $\Omega$ m. The L-curve cut-off used to obtain the results of Fig. 8 corresponds to a residual-to-roughness ratio $\lambda = 0.0077$, and the global rms error is about 20 per cent.

The cross-section model of conductivity used in the inversion constitute an important simplification and only the main structures labelled R1 and C1–C6 in the reconstructed cross-section shown in Fig. 8 are discussed hereafter. These structures remain stable during a sequence of independent inversions performed with different values of the control parameters (i.e. number of iterations, cut-off $\lambda$ in the L-curve, meshing). These tests show that several structures located near the boundary of the cross-section may significantly vary both in size and conductivity contrast from one inversion to another. Such variations are typical of poorly resolved domains where non-uniqueness may produce the appearance of small-scale structures with opposite conductivity contrasts (i.e. one conductive and one resistive) nearby stable large-scale structures. This is for instance the case of the two resistive anomalies located on the western and southern sides of the R1 resistive structure (Fig. 8; Yasin et al. 2011).

3 INTERPRETATION OF MAIN STRUCTURES

3.1 Resistive ridge R1

Structure R1 is the only major resistive structure of the cross-section with an average resistivity of $\approx 400 \Omega$ m. The most resistive part of R1 is localized beneath the southwestern side of the lava dome and its northern end coincides with a series of promontories visible on the summit plateau, the most prominent being the Dolomieu peak located near the northern end of R1 (PD on Fig. 9). The R1 structure is located beneath the 1-D resistivity soundings L, M, O and P of Nicollin et al. (2006) who give a resistivity range of 230–500 $\Omega$ m for the basement of their 1-D models. VLF soundings performed above the northern part of R1 by Zlotnicki et al. (2006) give a resistivity range of 90–250 $\Omega$ m. The agreement between the resistivity found for the R1 structure and those derived from geoelectrical soundings performed on the top part of the dome indicate that R1 corresponds to the root of a resistive body that vertically extends up to the summit of the lava dome. This a posteriori validates the vertically extruded resistivity model used in this study.

The R1 structure also coincides with the dense region RS3 visible on the western side of the east–west muon radiography shown in the top part of Fig. 3. Both the high resistivity and density are typical of a massive lava body, and this agrees with the fact that the peaks visible at the summit are likely to be the top parts of extruded vertical lava spines. The southern half of R1 is curved eastwards and may be associated with the deep part of a bulge located on the southern flank of the lava dome.

The R1 ridge seems to constitute an efficient barrier that prevents hydrothermal fluids to flow on the west side of the lava dome. This may explain the absence of any activity during the successive phreatic crises that punctuated the eruptive history of the volcano. However, the chemical tracing performed by Bigot et al. (1994) sustains the existence of an hydraulic pathway below the R1 structure and linking the ‘Tarissan’ pit (GT on Fig. 9) to the ‘Bains Jaunes’ hot springs located about 1km south–west of the dome. This indicates that the basis of the R1 body is probably highly fractured or altered below the lava dome.

3.2 Conductive structures C1 and C3

We join the interpretation of the C1 and C3 conductive structures (Fig. 8) because both are associated with major surface fractures
that have been active during the 1976–1977 crisis such as ‘Fracture Faujas’ (FF) under C1 and ‘Fracture 1956’ (F56), ‘Fracture 8 juillet 1976’ (FJ56), and ‘Fracture 30 août 1976’ (F76) under C3 (Figs 2 and 9; Komorowski et al., 2005). These structures have an average resistivity of $\approx 50 \Omega \cdot m$ in agreement with the C I-D sounding of Nicollin et al. (2006) who report a basement resistivity of $35 \Omega \cdot m$ at the eastern edge of C3. The VLF sounding performed by Zlotnicki et al. (2006) in the same area gives significantly higher resistivities in the 200–800 $\Omega \cdot m$ range. This discrepancy may be explained by the fact that VLF soundings are limited to shallow structures which, in the considered area, are constituted by rockfall deposits. C1 is in the axis of the low-density RF4 region of the ‘Roche Fendue’ muon radiography (bottom of Fig. 3), and C3 corresponds to the low-density domain RS4 in the ‘Ravine Sud’ radiography (top of Fig. 3). Both the low density and the low resistivity of C1 and C3 indicate that these regions are likely to be filled with altered unconsolidated material saturated with hydrothermal fluids.

Both C1 and C3 are associated with vent collapses that occurred during the explosive opening of fractures during historical eruptions. These events ejected a mass of non-juvenile debris from the dome that flowed for a short-run out in nearby valleys (Sheridan 1980; Feuillard et al. 1983; Komorowski et al. 2005). The collapse associated with C1 lead to the ‘Faujas’ rockslide that occurred on 1798 April 26 during the 1797–1798 phreatic eruption (FF on Fig. 9). Later, on 1837 February 12, a new fracture opened nearby the ‘Faujas’ rockslide and released a lahar (blue arrow 1 in Fig. 9) that invaded the ‘Ravine Amic’, a gully located on the northwest side of the dome (Fig. 2), and the ‘Noire’ river (Hapel-Lachênaie et al. 1978; Komorowski et al. 2005). These events support the hypothesis that C1 is a reservoir likely to release significant amount of fluid and energy. The connection of C1 with surface fractures [‘Fracture du Nord Ouest’ (FNO) and ‘Fente du Nord’ (FN); Figs 2 and 9] is sustained by the presence of a negative anomaly of spontaneous potential (Zlotnicki et al. 1994) typical of downward-going fluid flow. Although located in a presently inactive region of the summit plateau of La Soufrière, this reservoir seems still active as observed during the 1976–1977 crisis when ephemeras vents appeared in several fractures.

The explosive collapse associated with C3 occurred on 1976 August 30 during a particularly intense phreatic event where the ‘Tarissan’ crater (GT on Fig. 9) that was reported active at the end of the 17th century during the 1680 eruption (Boudon et al. 1988; Komorowski 2008). No activity was reported on the summit plateau in this area during the 1956 and 1976–1977 crisis, and only a moderate vent was observed at the eastern edge of C2 on the flank of the lava dome during the 1976–1977 eruption. The fact that C2 remained inactive during the intense 1976–1977 crisis can be a clue that the C2 reservoir is isolated from C1 and C3 by a hydrological barrier corresponding to the ridge of moderate resistivity (i.e. $\approx 100 \Omega \cdot m$) visible on the southern and western sides of C2 (Fig. 9).

3.3 Conductive structure C2

The C2 conductive structure is located in a presently inactive part of the lava dome. Its northern edge is limited by the ‘Fracture du Nord-Est’ (FNE on Fig. 9) that was reported active at the end of the 17th century during the 1680 eruption (Boudon et al. 1988; Komorowski 2008). No activity was reported on the summit plateau in this area during the 1956 and 1976–1977 crisis, and only a moderate vent was observed at the eastern edge of C2 on the flank of the lava dome during the 1976–1977 eruption. The fact that C2 remained inactive during the intense 1976–1977 crisis can be a clue that the C2 reservoir is isolated from C1 and C3 by a hydrological barrier corresponding to the ridge of moderate resistivity (i.e. $\approx 100 \Omega \cdot m$) visible on the southern and western sides of C2 (Fig. 9).

In 2009 November, a moderate landslide (Fig. 1) occurred after several days with heavy rains on the eastern extremity of C2. A clear negative anomaly of spontaneous potential (Zlotnicki et al. 1994) is associated with C2 and supports the existence of a downward fluid flow from the summit plateau down to the C2 reservoir which is likely to be connected to the hot Carbet spring (blue point labelled CE in Fig. 2) and the Carbet–Echelle fumarolic field (label 6 on Fig. 2; Zlotnicki et al. 2006; Komorowski 2008). Field observations of the landslide deposit and the landslide scar show the abundant and pervasive of plastic bluish-grey hydrothermally altered clay-rich formations (Fig. 10) that form part of the internal units of the dome. These clay units are present in situ in the dome and constitute low-strength low-friction layers that promoted land-sliding following an exceptionally intense rainfall event that occurred on 2009 November 19 and 20 (Météo France 2009).
3.4 Conductive structures C4–C6

The conductive structures C4–C6 appear as peripheral structures not connected with the central regions of the dome. Their resistivity of $\approx 40 \, \Omega \cdot m$ agrees with the A, B and H 1-D soundings of Nicollin et al. (2006) who find a resistivity of $\approx 30 \, \Omega \cdot m$ for the basement of their 1-D models. The materials forming these structures are likely to be unconsolidated debris fallen from the steep slopes of the lava dome. This constitutes potentially unstable volumes that may produce significant landslides during heavy rains or earthquakes.

4 CONCLUDING REMARKS

The data analysed in this study are well adapted to get information concerning the innermost resistivity structure of La Soufrière dome (Fig. 4). The reconstructed resistivity cross-section shows that the interior of the lava dome contains three main conductive domains (C1, C2, C3 in Fig. 8) and one resistive structure (R1 in Fig. 8).

Considering the resistivity values of these structures together with the densities obtained by cosmic muon radiography (Fig. 3) we may conclude that C1, C2 and C3 are reservoirs filled with unconsolidated material and conductive hydrothermal fluids. This description is coherent with the activity observed during the successive phreatic eruptions that occurred since the creation of the lava dome 500 yr ago.

Similarly, R1 is interpreted as a massive lava body that vertically extends through the whole height of the lava dome and which seems to constitute a barrier that, up to now, blocked eruptive activity on the southwest flank of the volcano. However, the chemical tracing performed by Bigot et al. (1994) suggests that this barrier is fractured at least at its basis level.

The presently active reservoir C3 is located inside the southeastern quarter of the dome and, accounting for the fact that both structures were active in 1976–1977, a connection seems to exist with the C1 reservoir located in the northwestern quarter. These two reservoirs may contain a significant amount of fluids and thermal energy that could be released in case of rapid deflation caused by landslide or overheating at the base of the dome as was observed at several instances for Soufrière Hills at Montserrat since the beginning of the magmatic eruption in 1995 (Komorowski et al. 2005). Oriented blasts and mud flows released by the reservoirs may invade nearby rivers—‘Carbet’, ‘Matylis-Galion’ and ‘Rivière Noire’—over several kilometres.

The C3 and C1 structures fall nearby the most important and historically active fractures that transect the dome in half. Moreover these structures are a continuous of the en-echelon normal La Ty fault (Fig. 2) that propagates from the southeast to the northwest through the ‘Fracture 30 août 1976’ (F76 on C3 structure, Fig. 9), through the summit craters and fractures, and the ‘Fente du Nord’ (FN on the eastern edge of the C1 structure, Fig. 9). La Soufrière lava dome is thus characterized by this heterogeneous geometry of low density and high-conductivity fluid-saturated and hydrothermally altered areas. The clear recurrent link of these structures to explosive historical activity and major surface deformation associated with recent phreatic and still-born magmatic eruptions suggest that areas within La Soufrière lava dome are prone to slope instability and partial edifice collapse, overpressurization leading to explosive vent fracturing, as well as the genesis of significant volume of water from acid perched hot aquifers that will generate mobile and potentially damaging mudflows (lahars). Hence these results have important implications for continuing multiparameter monitoring of La Soufrière volcano, hazard scenario definition, risk assessment and crisis management.

ACKNOWLEDGEMENTS

We benefited from the help of Florence Nicollin (Rennes 1 University) and colleagues from the Guadeloupe Volcano and Seismological Observatory (IPGP/OVSG) in acquiring the electrical resistivity measurements. Comments made by two anonymous referees helped us to improve the manuscript. The recent eruptive history of the volcano has been improved in the framework of the CASAVA ANR project. The EIDORS open-source software is available at http://eidors3d.sourceforge.net. The work of BG was supported by a Research Fellowship from the Alexander von Humboldt Foundation. Financial support was provided by NSERC Canada. This is IPGP contribution 3513.

REFERENCES


Slice electrical resistance tomography


**APPENDIX: COSMIC MUON RADIOPHGRAPHIES**

Density radiography using cosmic muons is a novel method that uses the attenuation of the flux of muons crossing a geological body to determine its density structure (Gibert *et al.* 2010). Measurements are made with telescopes equipped with detection matrices that allow to count the number of muons coming from about one thousand lines of sight as shown in Fig. A1 (Marteau *et al.* 2012). For each line of sight, the number of detected muons that crossed the volcano is compared with the incident flux in order to deduce the amount of matter—also called the opacity (in g cm\(^{-2}\))—encountered by the particles along their trajectory (Lesparre *et al.* 2010). The opacity values are converted into average density along the lines of sight to produce the radiographies of Fig. 3.

One advantage of muon density radiography is the straight ray geometry of the acquisition (Fig. A1) which allows to locate the density heterogeneities visible on the radiographies. Because of the cone-like geometry of the rays, the density structures visible on Fig. 3 are averaged along more or less oblique lines of sight. The dotted lines visible on the radiographies bound the regions where the corresponding rays pass within ±15 m above or below the electrode ring that defines the SERT cross-section of Fig. 7.

*Figure A1.* View of the lines of sight scanned by the cosmic muon telescope when located at ‘Ravine Sud’ (top, altitude 1168 m) and ‘Roche Fendue’ (bottom, altitude 1263 m). The data acquired at the ‘Ravine Sud’ station produced the east–west radiography shown in the top part of Fig. 3, and the north–south radiography for the data of ‘Roche Fendue’ is shown at the bottom of Fig. 3. The black stars represent the electrodes and the green surface represent the regions sounded by the SERT. The lines of sight that appear in red are entirely comprised in a volume of ±15 m above and below the SERT cross-section. See Lesparre *et al.* (2012) for a detailed description of the muon tomography experiments on La Soufrière.